EFFECT OF LAND USE ON PHYSICAL AND HYDROLOGICAL CHARACTERISTICS OF KABETE SOILS

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DEDICATED TO MY LATE FATHER, JOHN; MY MOTHER MARY, MY WIFE CHRISTINA AND MY SON STANLEY.

DECLARATION

I hereby declare that this thesis is my original work and has not been presented for a degree in any other University. All quotations have been distinguished by quotation marks and all sources of information specifically acknowledged.

Signature ____________

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This thesis has been submitted for examination with our approval as University Supervisors.

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ABSTRACT

Soil physical and hydrological properties of kabete soils were evaluated at three different land use sites namely forest, grassland, and cropland sites. The purpose of this work was to determine the effect of selected land uses on physical and hydrological characteristics of the kabete soils in Kenya.

Soil samples for this study were collected from the College of Agriculture and Veterinary Sciences (CAVS), field station. Soil samples for the determination of bulk density, organic matter content, particle size distribution, soil aggregate stability, soil erodibility indices, pore size distribution, soil moisture release characteristics, profile water holding capacity, available water capacity, and saturated hydraulic conductivity, were collected from 0, 10, 30, 60 and 100 soil depth at four different representative areas on each site. Soil samples were then taken for laboratory analysis of physical and hydrologic soil properties using standardized laboratory procedures.

The bulk density for the cropland had an average value of 1.16 g cm^{-3} as opposed to forest and grassland sites with average values of 1.02 and 1.1 g cm^{-3} respectively. Organic matter content showed a high value of 4.04% at the forest site, grassland and cropland sites had organic matter content values of 3.51 and 2.77% respectively. Particle size distribution, soil aggregate stability and soil erodibility

indices showed significant differences between the three sites. The forest site showed higher readily available water content than the other two sites. The cropland site had the highest amount of non readily available water than grassland and forest sites. The three sites showed different patterns of soil moisture release characteristic curves depending on the soil texture and organic matter content of each site. Field capacity, permanent wilting point, available water capacity, and saturated hydraulic conductivity for the three sites varied significantly at the soil surface as well as at the 100 cm soil depth.

1.0 INTRODUCTION

1.1 Soil Physical Characteristics

Soil physical characteristics are responsible for the transport of air, heat, water, and solutes through the soil. Soils particularly those in the tropics, show great diversity in texture, structure, type of clay, organic matter, and rooting depth. These variations result in significant differences in infiltration rate, erodibility, moisture holding capacity, drainage characteristics, aeration, susceptibility to and recovery from compaction, and general response to soil management and manipulation.

Infiltration is influenced by particle size distribution, surface area, bulk density, pore size distribution and hydraulic conductivity. Properties like soil type, soil profile, biological factors, macrostructure within the soil and vegetal cover affects infiltration indirectly. Soil surface characteristics affect soil susceptibility to wind and water erosion, water infiltration rate, and crusting. The susceptibility to wind and water erosion is influenced by the size, number, and stability of the surface soil clods or aggregates. Water stable aggregates, help to minimize soil dispersion, thereby maintaining higher infiltration rates and decreasing run-off aand erosion by water.

. Stable soil aggregates also decrease the potential of soil crusting that may hinder seedling emergence.

Several soil physical characteristics can change with management. In the tropics for example severe desiccation due to high temperatures at the soil surfaces may be followed by abrupt changes due to high-intensity rainstorms. Bulk density and organic matter content, soil aggregate stability and erodibility indices are soil physical characteristics that deteriorates with cultivation rendering the soil less permeable and more susceptible to run-off and soil erosion.

1.2 Soil Hydrological Characteristics

Knowledge of the pattern of water movement in agricultural and forest soils is essential to solving the problem of irrigation scheduling, land drainage,soil and water conservation, nutrient transport, run-off pollution, and ground water contamination. The water in the soil is dynamic. It is constantly moving from one point to another in response to differences in hydraulic potential created by percolation , evaporation, irrigation, rainfall, plant use, and temperature.

Knowledge of infiltration rate of water into the soil under field condition is very useful in designing farm irrigation system, estimating the run-off, and determining the time required to irrigate a given field up to a desired soil depth. The measures of reclamation of saline or sodic soils are greatly influenced by infiltration rates of a particular soil. Infiltration rate is a very dynamic property and changes within season, soil, and crop management. Soil texture, ground slope, depth of water table, and biological activities also influence the rate of infiltration.

1.3 The Problem

Water is the limiting natural factor to crop production in East African countries. Improving the management and conservation of soil and water for increased crop production becomes the primary aim of agricultural research. Rainfall, the only source of moisture available in most of these regions is unpredictable and may not occur when needed by crops. When it does occur, however, it is usually of short duration and high intensity and much of it is lost as run-off. Rains often stop before crops have had sufficient moisture to take them to maturity. In these areas of erratic rainfall which often occur in brief intense storms, the physical and hydrological characteristics of the soil can be major factors in deciding whether a crop is a success or failure for they determine how much of water from an intense storm is absorbed by the soil, and how much water from the storm is then available to the plants to enable them to survive until the next rainfall.

The proportion of rain water that can be stored in the root zone for crop use is significantly influenced by the soil particle size distribution, organic matter content, and mineralogical composition. Closely related to soil and water retention capacity are the soil water transmission characteristics. These characteristics determines the soil water behavior in the rooting zone of plants and therefore the supply of water to the plant roots. Drastic yield reductions of shallow rooted crops can be caused by periodic occurrence of dry spells at critical stages of growth due to low available water holding capacity of many soils.

Information of soil physical and hydrological characteristics are essential in the development of improved soil, water and crop management systems. When integrated with improved crop varieties, fertilization, and plant protection, these soil characteristics can aid in the development of economically viable farming systems which can increase and stabilize agricultural production. The purpose of determining the soil physical and hydrological characteristics for the major soils is to facilitate methodological decision - making in planning development strategies and selecting appropriate land use and management practices that are concerned with care and maintenance of soil resources that underpin agricultural productivity. The objective of this study was therefore to determine the effect of selected land uses on physical and hydrological characteristics of the Kabete soils.

1.4 Objectives

The specific objectives of the study were.

a) To determine soil physical properties of Kabete soils.

b) To determine soil hydrological properties of Kabete soils.

c) To determine the effect of cropland, forest and grassland land uses on the physical and hydrological charateristics of the Kabete soils

2.0 REVIEW OF LITERATURE

2.1 Introduction

Soil is a three phase system that is made up of solid, liquid and gaseous material (see Fig 1). The solid phase may be mineral or organic. The mineral portion consists of various sizes, shapes, and chemical compositions. The organic fraction on the other hand, includes residues in different stages of decomposition as well as live active organisms (Hillel, 1971).

Plants that are the basis of all agriculture, grow within the complex soil system. If the solid phase of the soil contains sufficient nutrients that can be released to the plants, the soil is said to be fertile. If the pore spaces between the solid particles are so distributed as to provide ample water storage for plant growth and at the same time permit adequate aeration to plant roots, the soil is considered to have favorable water air relationships (Kesse, 1968).

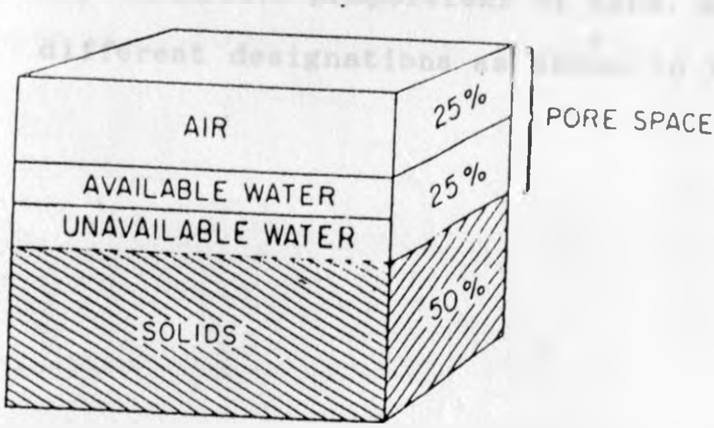


Figure 1. A cultivated loam soil

From the soil, plant roots receive mechanical support, essential elements, water, and oxygen. Soil physical properties largely determine the soil water supplying capacity to plants while soil chemical properties determine the soil nutrient supply. Both physical and chemical properties determine root extension and the volume of soil that serves as reservoir for both water and essential nutrients for plants (Hillel, 1971)

2.2 Soil Texture

Texture refers to the relative proportion of primary particles of sand, silt, clay and other skeletal material in the soil body.

Quantitatively, soil texture refers to the relative proportions of various sizes of particles in a given soil. The traditional method of characterizing particle sizes in soils is to divide these particles into sand (2.0 - 0.05 mm), silt (0.05 - 0.002 mm) and clay (below 0.002 mm) size ranges. Soils with different proportions of sand, silt, and clay are given different designations as shown in figure 2 (Hillel, 1971)

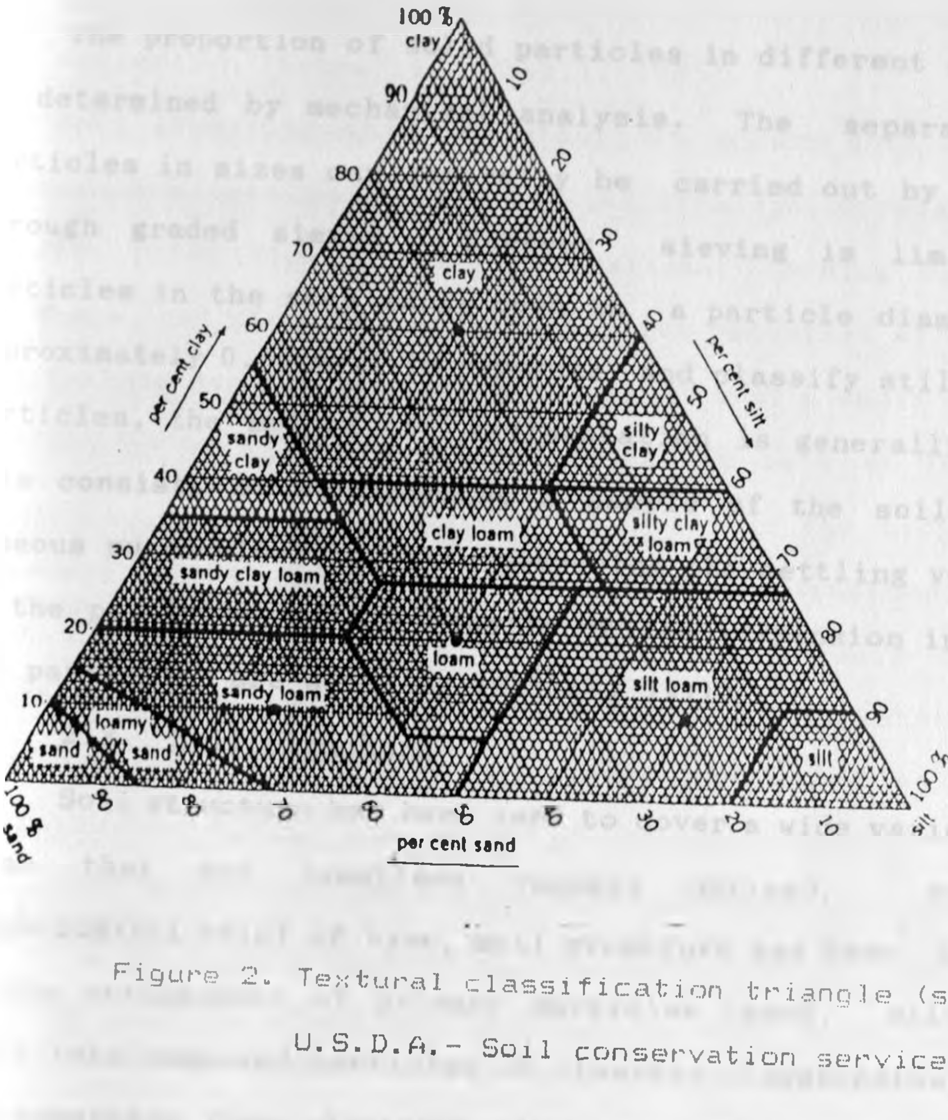


Figure 2. Textural classification triangle (source U.S.D.A.- Soil conservation service)

The proportion of solid particles in different sizes is determined by mechanical analysis. The separation of particles in sizes can generally be carried out by sieving through graded sieves. However sieving is limited to particles in the sand range down to a particle diameter of approximately 0.05 mm. To separate and classify still finer particles, the method of sedimentation is generally used. This consists of dispersing a sample of the soil in an aqueous suspension, and then measuring the settling velocity of the particles or the density of the suspension in which the particles are settling (Hillel, 1971).

2.3. Soil Structure

Soil structure has been used to cover a wide variety of ideas that are sometimes vaguely defined. From a morphological point of view, soil structure has been defined as the arrangement of primary particles (sand, silt, and clay) into compound particles or clusters (aggregates) that are separated from adjoining clusters and have properties unlike an equal mass of unaggregated primary soil particles (USDA, 1955). From an adaphological point of view, however, several factors associated with structure are more significant than aggregate size and shape. Some of these factors include the pore size distribution that results from aggregates, the stability of the aggregates when wet and their ability to reform upon drying, and the hardness of the aggregates. These factors are all more closely connected with soil - plant relations than are the size and shape of the aggregates (Kemper and Chepil, 1965).

Soil structure and its stability govern soil - water relationship, aeration, crusting, infiltration, permeability, run-off, inter flow, root penetration, leaching of plant nutrients, and therefore, the productive potential of a soil. Soil texture can not be changed, at least over a short period of time, by an economical means, since it is a an inherent property. Successful management of soils in the tropics as elsewhere depends on the management of soil structure. Soil structure is therefore an important but vaguely defined property. From the point of view of soil management, soil structure may be regarded as "the property of a soil that regulates a continuous array of various sizes of inter connected pores, and their stability and durability, governs retention and movement of water, regulates gaseous diffusion from and into the atmosphere and control root proliferation and development. It is a complex phenomenon indeed. Soil structure, therefore both directly and indirectly promotes or controls phenomena such as wind and water erosion, waterlogging and aeration, leaching of plant nutrients, soil temperature, trafficability of farm machinery, root penetration and development, and crop yield (Lal, 1979).

The following types of soil structure are commonly recognized as described by Davies, 1972.

(i) Granular and crumb - Usually found in top soils when well weathered and cultivated.

(ii) Angular blocky - There are frost tilths of heavy soils, particularly silty clays.

(iii) Angular blocky (Polished faces) - These are sub soils of some heavy clays.

(iv) Sub-angular blocky - Top soils or sub soils of well structured soils.

(v) Coarse prismatic - Sub soils of heavy clays e.g. carboniforous clay.

(vi) Fine prismatic - Sub soils of brickearths, and well structured clays.

(vii) Platy structure - Plough pans, under slipping tractor Wheels, sub soil of soils in recent sediments.

Davies, 1972 went further to explain that structures in sub soils are more permanent and usually larger than in the plough layer. The formation of soil structure is brought about by wetting and drying, by freezing and thawing, and by cultivations. In addition the growth of roots, the activities of earthworms, bacteria and other soil organisms also encourage structure. By these factors soil particles are joined together, or broken down into distinct units, their size and shape depending mostly on the type of soil, but also on the system of management.

2.4. Flow of water in Saturated and Unsaturated Soil

Water in the soil is dynamic. It is constantly moving from one point to another in response to difference in hydraulic potential created by percolation, evaporation, irrigation, rainfall, plant water use, and temperature. The saturated and unsaturated flow processes are therefore generally complicated and difficult to describe quantitatively, because they often entail changes in the state and content of soil water during the flow. Such changes involve complex relations among the variable water content, suction and conductivity which may be affected by hysterisis (Ponlovassilis, 1962).

In saturated soil, the driving force or hydraulic potential gradient is due to difference in height and water pressure between two points and the saturated hydraulic conductivity, K, is independent of the hydraulic potential gradient. Under unsaturated conditions, however, the water in the soil is subjected to a sub-atmospheric pressure, or suction whose gradient constitutes, a driving force. The unsaturated hydraulic conductivity, K(0), unlike saturated hydraulic conductivity, therefore, varies with the soil water content which itself varies with matric potential.

When a suction is developed, the first pores to empty are the largest ones, which are the most conductive, leaving water to flow only in the smaller pores.
The empty pores must be circumvented, so that, with

desaturation, the tortuosity increases. In coarse - textured soils, water may sometimes remain almost entirely in capillary wedges at the contact points of the particles, thus forming separate and discontinuous pockets of water. In aggregated soils, too, the large interaggregate spaces resulting into high conductivity at saturation become barriers to liquid flow from one aggregate to its neighbours. Because of these reasons the transition from saturation to unsaturation generally entails a steep drop in hydraulic conductivity.

2.4.1. Equations governing saturated and unsaturated flow

The flow of water in saturated soil is described by Darcy's law which states that the water flux density q, is proportional to, and in the direction of the driving force which is the hydraulic potential gradient.

In one dimension, this equation is given by :

$$q = - K \frac{dh}{dx} \qquad (1)$$

where q is the water flux density in cm S^{-1}, dh/dx is the hydraulic potential gradient in cm cm^{-1}, and k is the saturated hydraulic conductivity in cm S_{-1}

Darcy's law, though originally conceived for saturated flow only, has been extended to unsaturated flow, with the provision that the conductivity is now a function of water content (Richards, 1931).

Thus, Darcy's law for unsaturated soil is given by

$$q = - k(\theta) \frac{dh}{dx} \qquad (2)$$

However, to obtain the general flow equation and account for transient as well as steady state flow processes, the equation of continuity must be introduced. The equation of continuity states that the time rate of change of water content in a volume element of soil is equal to the divergence of the water flux density, q. It is given, in one dimension, as follows:

$$\frac{d\theta}{dt} = - \frac{dq}{dx} \qquad (3)$$

Assuming that there is a defined relation between the water content, θ, and the hydraulic potential H, and that this relation is not dependent upon position, then substituting equation (2) into equation (3) yields the form of equation for the Darcy - based theory of water flow in unsaturated soils either in the hydraulic potential form:

$$c\ (\theta) \frac{dh}{dt} = \frac{d}{dx} [k\ (\theta) \frac{dH}{dx}] \qquad (4)$$

or in the water content form:

$$\frac{d\theta}{dt} = \frac{d}{dx} \underline{D\theta \frac{d\theta}{dx}} \qquad (5)$$

where the water capacity C(θ) is defined as:

$$C(\theta) = \frac{d\theta}{dH} \qquad (6)$$

and the soil water diffusity D (θ) as:

$$D\ (\theta) = K\ (\theta) \frac{dH}{dO} = \frac{K\ (\theta)}{C\ (\theta)} \qquad (7)$$

2.4.2. Measurement of Saturated Hydraulic Conductivity

Saturated hydraulic conductivity indicates the water entry and water movement in soils. The available field methods are laborious and time consuming and require large quantities of water or presence of water table and are of very limited applicability on cracking clay soils (Klute, 1965). This is because in field, much of the water entry and distribution takes place in unsaturated soil and/or negative suctions. The laboratory measurements under such conditions imposed are slow and complex and is impossible to simulate field conditions directly especially for swelling soils (Bridge *et al*, 1970).

It has been shown that the rate of flow of water down a tube of radius r is proportional to r^4. A similar result follows for flow down cracks and other more complicated shapes of channels. As a consequence, in a soil with any structural development, the saturated hydraulic conductivity is often dominated by the size and the way the small number of large cracks and/or pore are connected (despite the much greater number of smaller pores per unit area) (Hillel, 1971).

Three methods are available for measuring saturated hydraulic conductivity. These are:

(i) Laboratory measurements on cores taken in the field

(ii) Field measurements above the water table.

(iii) Field measurements below the water table.

A meaningful measurement of K can only be obtained in the laboratory on a soil sample which is large enough to include a representative number of pores/cracks.

2.5. Soil Moisture Release Characteristics

The relationship between matric potential and moisture content of a draining soil is called moisture release characteristics. Several empirical equations have been proposed to describe the soil moisture release characteristics for some soils within limited ranges of matric potential (Visser, 1966; Gardener, et al, 1970).

Hillel (1971), stated that in a saturated soil at equilibrium with free water at the same elevation, actual pressure is atmospheric and so hydrostatic pressure and suction (tension) are zero. If a slight sub-atmospheric pressure is applied to water in a saturated soil no outflow may occur until the suction is increased beyond a certain critical at which the largest pore of entry begins to empty. As suction is further increased, more water is drawn out of the soil and more of the relatively large pores which cannot retain water against the suction applied, will empty out. A gradual increase in suction will result in emptying of progressively smaller pores, until at high suction values only the narrow pores retain water. The amount of water remaining in the soil at equilibrium is a function of the sizes and volume of the water filled pores and hence it is a function of the matric suction called soil moisture retention curve or soil moisture characteristics curve (Childs, 1940). Moisture characteristics curve covers a range of soil moisture from saturation at atmospheric pressure (zero suction) to the oven dry condition. The most important part of the curve is the lower suction (higher potentials) range because it shows what energy water is held in the available range. Gardner (1971), suggested that the most important part of the curve is in the range 0 - 1 bar for most soils, and 0-3 bars for heavy clay soils.

The soil physical properties (i.e. structure, texture and macrostructure) affect water retention only up to suctions of 2 or 3 bars. At higher suctions, water retained by an undisturbed sample does not differ significantly from that retained by material ground to pass a 2 mm - sieve (Hillel, 1971). Water characteristics curves determined with sieved samples frequently are different from those determined with relatively undisturbed cores of the same soil. The difference is pronounced in moist soils where the shape of characteristic curve is determined largely by pore water. Degree of aggregation has a distinct effect on the pore size distribution in soil. At higher water content, aggregation has pronounced influence on the soil moisture characteristic curve. Effect of aggregation decreases as water is removed from soils because many pores are emptied and film water becomes more important in influencing the shape of the soil moisture characteristic curve. The amount of water held in films, is a function of specific surface and consequently is largely determined by soil texture and not by soil aggregation (Gavande, 1968).

Soil moisture characteristic curve is significantly influenced by soil texture. The greater the clay content, the greater the water content at any particular suction, and the more gradual the slope of the curve (Hillel, 1971).

In a sandy soil, most of the pores are relatively large and once these large pores are emptied at a given suction, only a small amount of water remains. In a clayey soil, however,the pore size distribution is more uniform and more of the water is adsorbed, so that increasing matric suction causes a more gradual decrease in water content (Hillel, 1971). Soil structure affects the shape of the soil moisture characteristic curve (especially in the low suction range). Effect of compaction upon a soil is to decrease the total porosity, and especially to decrease the volume of the large interaggregate pores. Hence saturation water content and initial decrease of water content with the application of low suction are reduced. Again the volume of intermediate size pores is likely to be somewhat greater in compact soil (as some of the originally large pores have been squeezed into intermediate size by compaction); while intra aggregate micropores remain unaffected and thus the curves for the compacted and uncompacted soil may be nearly identical at high suction range. At very high suction range, water is held mainly by adsorption and retention is a textural rather than a structural attribute of soil (Taylor and Ashcroft 1972).

2.5.1 Hysteresis in Soil Moisture Release Characteristics

The relation between matric potential and soil wetness is not generally a unique and single valued one. This relation can be obtained in two ways of either desorption or sorption. Each of these two methods will yield a continuous curve, but the two curves will in general not be identical.

The equilibrium soil wetness at a given suction is greater in desorption (drying) than in sorption (wetting). This dependence of the equilibrium content and state of soil water upon the direction of the process leading up to it is called hysteresis (Hillel, 1971). This type of behavior is attributed to:

(i) The geometric non - uniformity of the individual pores which result in the bottle neck effect;

(ii) The contact angle effect caused by surface roughness, the presence and distribution of adsorbed impurities on the solid surfaces and the mechanism by which liquid molecules adsorb or desorb when the interface is displaced (Hillel, 1971),

(iii) Entrapped air, which decreases the maximum possible water content of the soil; and

(iv) Swelling, shrinking, or aging phenomena which result in differential changes of soil structure depending on the melting and drying history of the soil (Hillel and Mottles, 1966).

Few soils, if any, are completely stable structurally. Most particularly, clayey soils of low humus content (as may occur in arid regions) exhibit structural changes related to wetting - drying cycles (including slaking, swelling, and particle-reorientation). This is especially marked in disturbed (compacted or crushed) soil samples (Hillel and Mottles, 1966).

2.6 Field Capacity

Field capacity, (FC) is not a constant value but represents the range of moisture contents retained in the soil when macropores have been drained by gravity (Russell, 1973). Loveday (1974) referred to it as the amount of water remaining in a well - drained soil when velocity of down ward flow into unsaturated soil has become small. Rich (1971), referred to it as the percentage of water remaining in a soil two or three days after having been saturated and after free drainage has practically ceased. Recent definition by Pidgeon, 1972 states that field capacity is the amount of water that a well - drained soil would hold against gravitational forces or the amount of water remaining when downward drainage has markedly decreased. Bear et al, (1968), reported that field capacity depends on soil texture and structure. Soil structure has the major role of determining field capacity under wet conditions. Field capacity varies with moisture content of soil when wetting begins and on depth of wetting (Loveday, 1974).

Taylor (1972), pointed out that, the foregoing concept of field capacity does not express any exact water content in the soil and assumes that water in excess of the supposed field capacity value quickly drains away. This approach overlooks the observation that soil water is not held so tightly by the soil matrix as such but that some of the soil moisture can be used by plants while it remains in contact with the plant roots.

2.6.1. Permanent Wilting Point

The permanent wilting point (PWP) concept is yet another soil constant that has for long been a source of argument for many researchers concerned with the soil - water - plant relationship (Hillel, 1971; Marshall, 1959; Kohnke, 1968; and Taylor, 1972). Richards and Weaver (1943) indicated that a suitable mean figure at which PWP occurs is 15 atmospheres. Permanent wilting point is dependent on soil profile features and is determined by amount of water in soil at various depths and by the rate at which water moves to plant root. It involves not only surface soil but any soil depth in which plant roots are growing (Taylor, 1972).

Permanent wilting point was defined by Marshall (1959), as the moisture level at which plants wilt and fail to regain their original cell turgidity even when placed in a humid environment for a specified period of time. Although the wilting phenomenon was key to the permanent witting point, allowance was given for the duration of the recovery period.

The permanent wilting point is the soil moisture condition at which the ease of release of water to plants is just simply too small to counter balance the transpirational loses (Kohnke, 1968). Taylor (1972), defined the PWP as a dynamic range of soil water percentages over which the rate of water supply to plants is not enough to prevent wilting. The 15 bar percentage is commonly used in place of the permanent wilting point or the permanent point percentage, (PPP).

2.7. Water Infiltration Rates

Infiltration is the volume of water passing into the soil per unit area per unit time and has the dimensions of velocity. Russell (1973), terms the rate of entry of water into soil as its infiltration rate, and this rate he adds, is initially high for all soils if they are dry. But once they are wet, the rate is dependent on the distribution, continuity, and stability of the coarse pores.

The rate of water entry into the soil fluctuates widely between soil types, and also wide differences can be found within a single soil type, depending upon the soil moisture levels and management practices (Parr and Bertrand, 1960). Very permeable soils will have infiltration rates as high as 35 cm/hr, while soils of low permeability will have rates of 0.03 cm/hr or less (Russell, 1973). Lal (1976), generalized that the water - intake rate of tropical soils under their natural vegetation cover is high. However, the removal of vegetation and introduction of mechanized tillage operations results in disturbance and exposure of soil and cause a rapid decline in infiltration rate.

Studies conducted on Alfisol in Western Nigeria indicated that the infiltration rates of eroded soil were considerably improved by fallowing with grass and leguminous cover crops (unpublished data by Lal). However, the improvement was rather slow and the soil did not attain the infiltration rate that existed prior to clearing even after three years on continuous fallowing.

When water is flooded on the surface of a dry soil core, the initial percolation rate falls sharply and may take many hours to become steady (Pereira, 1955). Reeve (1953), gave times from 50 to 100 hours to reach such an equilibrium. The reduction of percolation rate is an important measure of soil stability. In his assessment of structure in tropical soils, Pereira (1955), found soil in the poorest structural condition, in each soil type, transmitted water at rates exceeding 22.8 cm per hour. Water run-off before the soil reaches its maximum capacity is due to the failure of the heavy rain to infiltrate into surfaces which have slumped and sealed under rainfall impact. Pereira (1955),recorded high infiltration rates of 14.7 cm/hr from soil newly broken from napier grass, and low rates of 7.61 cm/hr, from over-tilled continuous arable land.

2.7.1. Factors affecting infiltration rates

Horton (1940), suggested the following factors affecting infiltration rate: Soil type, soil profile, biological factors and macro-structure within the soil, vegetal cover, Antecedent moisture content, and surface sealing.

Horton (1940), was of the opinion that infiltration rate is governed mainly by conditions at or near the surface.

Lewis and Powers (1938), listed a large number of factors affecting infiltration rates and they divided them into two major groups:-

(i) Those factors influencing the infiltration rate at a given time and point, such as texture, structure and organic matter.

(ii) Those factors influencing the average infiltration rate over considerable area and period of time such as slope, vegetation and surface roughness.

Duley and Russell (1939), studied the influence of vegetative factors by leaving crop residues at the soil surface which revealed the following:-

(i) Infiltration rate was greatly increased.

(ii) Evaporation from the surface soil was reduced.

(iii) Water erosion was reduced; and

(iv) Wind erosion was reduced.

Musgrave (1955) summarized the major factors that affect intake of water by soil as follows:

(i) Surface conditions and the amount of protection against the impact of rain.

(ii) Internal characteristics of the soil mass, including pores, depth and thickness of the permeable portion, degree of swelling of clay and colloids, content of organic matter, and degree of aggregation.

(iii) Soil moisture content and degree of saturation.

(iv) Duration of rainfall or application of water.

(v) Season of the year and temperature of the soil and water.

2.8. Factors causing the deterioration of soil physical and hydrologic properties

Soil structure deteriorates from poor management practices. These includes, continuous cultivation of crops without additions of organic matter, or crop rotations with fallow periods and cultivating soils at high moisture contents. Russell (1973), reported that the use of heavy machinery for field operations has also been observed to adversely affect soil structure, especially if soil moisture content is too high at the time of operation. Soil erosion can also enhance the deterioration of soil structure (Lal, 1975), and this is particularly true where some weak soil structure has inherent low aggregate stability.

Cultivation without crop rotations, fallow periods or without adequate additions of organic materials has generally been observed by many researchers (Russel, 1973; Sanchez, 1976) to drastically reduce the original organic matter content of the soils often leading to a deterio- ration of soil aggregate stability. This may render such soils much more susceptible to erosion and to compaction from heavy machinery resulting in the development of higher bulk densities and increasing the impedance to root penetration.

2.8.1. Soil physical properties under forestry grassland and cultivated land

In most tropical regions a rapid decline in crop production follows the clearing of forests and savannas and the development of the land for food production (Ruthen-berg,1976). Much of the decline can be attributed to loss of plant nutrients and increased aggressivity of weeds (Kang et al.,1977).The intervention of erosion, especially when large scale schemes are developed is also frequently very important (Greenland and Lal,.1977). However, a more general problem occurs in almost all regions:the deterioration of soil physical conditions during crop production (Ahn,1968).

The rate of organic matter decline after clearing a forest for arable agriculture has attracted a great deal of attention (Kononova,1968). Time studies have shown that the rate of decline in soil organic matter is highest just after clearing and reduces or stabilizes with time. Kononova (1966) reported that humus content in the temperate soil decreased by 43% during the first 13 years after clearing and by 9% during the subsequent 32 years of cultivation.

2.8.2. Impacts of deteriorating soil physical properties on crop yield.

The organic matter content in a soil is dependant on the rate of addition and loss of organic materials in the soil.Under virgin conditions a dynamic equilibrium is established between the gains and losses so that the organic matter content is maintained at a constant level. Sanchez(1976). Continuous cultivation frequently leads to decline in the soil organic matter. This decline in organic matter can generally be attributed to two causes. First, clearing and cultivation of land results to reduced rate of addition of fresh organic material (Greenland and Nye, 1979).

Aina (1979) studied the changes in physical and chemical properties of an Alfisol under different management practices. He found that after 10 years, soils under bush fallows had 4.3% organic carbon while the corresponding soil under grass fallow and under continuous cultivation had 3.8% and 1.1% organic carbon respectively.

The higher the bulk density, the more compacted the soil and the lower the pore space (Hillel, 1982). Compaction in soil leads to poor soil physical properties such as low infiltration, poor aeration and high impedance to root penetration. Bulk density for surface layer of fine textured soils generally ranges from 1.0 to 1.6 g\cm^3. In sand and sandy loams, the bulk density may run as high as 1.8 g\cm^3 because sand particles tend to pack closely and remain in close contact.

Cultivation has been shown to cause soil densities to rise both in tropical and temperate soils (Aina,1979). The increase in bulk density was mainly shown to affect the top soil significantly and not the sub soil.

The stability of aggregates, the size distribution and the quantity are important soil parameters that affect tilth and therefore indirectly affect crop growth. Soils with poor aggregate stability results to poor structural stability. Poor structural conditions may cause yields declines through a direct influence on root growth, deteriorated drainage conditions and hence poor aeration, and poor nutrition (Greenland, 1981).

The degradation of aggregates is highly correlated with the soil organic matter content (Harris et al.,1966). Russel (1978), noted that the disturbance of soil due to ploughing has a direct effect on the stability of aggregates.

Sessanga (1982), reported that the saturated hydraulic conductivity was best correlated with the coarse fraction. Continuous cultivation causes deterioration in the soil structure leading to low porosity and poor pore continuity. This leads to a reduction in the percent composition of transmission pores in the soil (Greenland,1981). It is the transmission pores (>50 um) that influence most of the conductivity soil. Continuous cultivation therefore leads to decline in the soils saturated hydraulic conductivity.

The soil moisture retention, release and storage capacity are influenced by various soil properties among which the most important include soil particle size composition, clay mineralogy and the organic carbon content (Stevenson, 1974). De Jong *et al.*, 1983 showed that both soil texture and soil organic matter highly influenced the soil moisture retention characteristics. Their results indicated that organic matter increased water retention at low suctions but had little effect on the rates of release at high suction. Addition of organic matter may not always increase retention of soils but can leave it unchanged or lessen its capacity to retain moisture.

2.9 Nitosols

Nitosols are clayey red soils of the tropics that have an argillic B horizon with shiny ped surfaces but without abrupt textural changes (Fitzpatrick, 1986). These soils have clay distribution where the per centage of clay does not decrease from its maximum amount as much as 20% within 150 cm of the surface. FAO-UNESCO 1988, recognized three major types of these soils as Eutric Nitosols which have a base saturation of

50% or more; Dystric Nitosols with a base saturation of less than 50% and Humic Nitosols that have a base saturation of less than 50% and an Umbric A horizon. These soils are usually deep with a well formed sub-angular blocky or granular structure which inparts good root room and water storage. They are among the most fertile soils in the tropics and are extensively used for a wide range of crops. They do need fertilizer particularly phosphorous which is rapidly fixed (Fitzpatric, 1986).

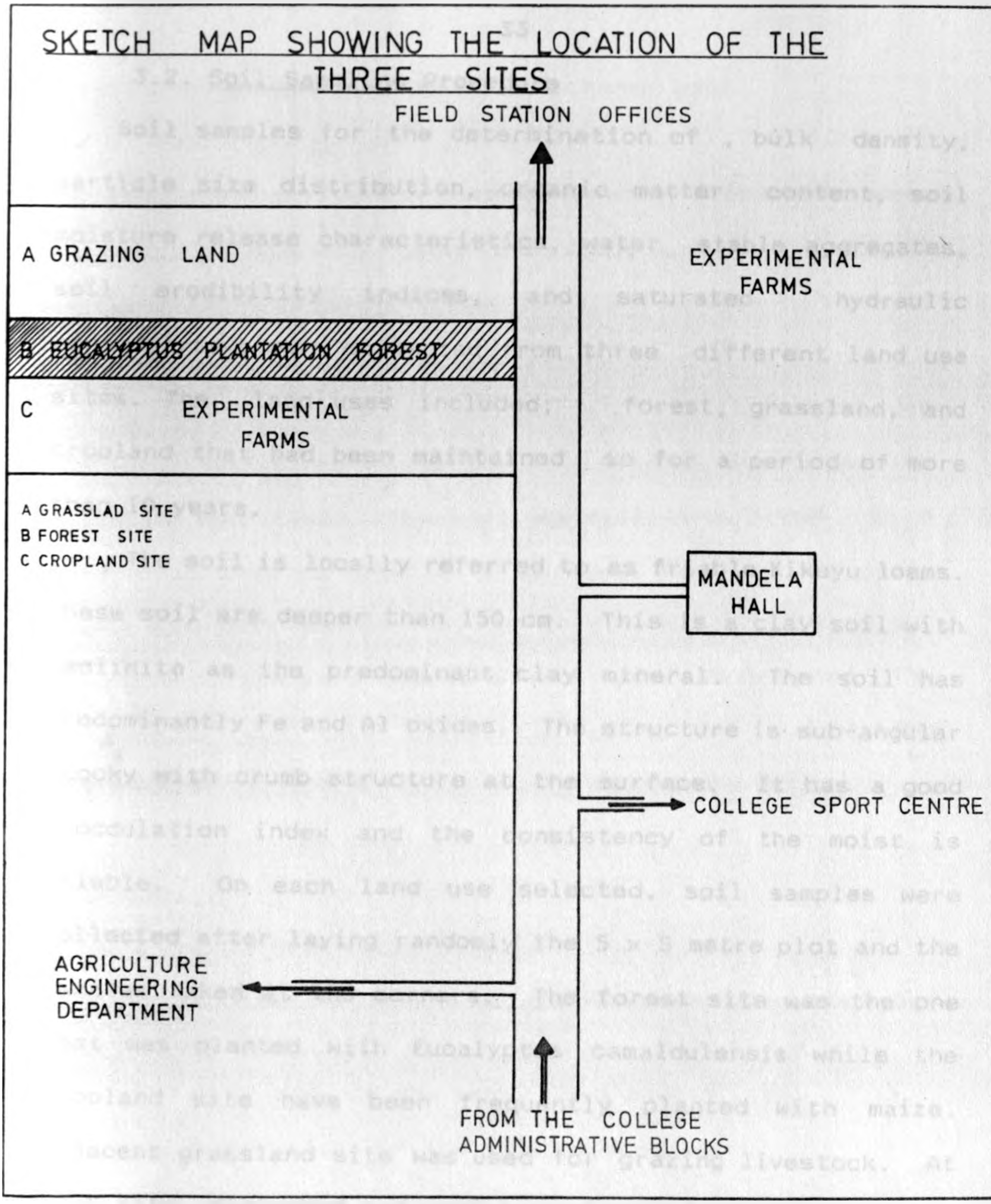
FAO UNESCO (1988) defines argillic B horizon as one that contains illuvial layer-lattice clays. This horizon forms below an elluvial horizon but it may be at the surface if the soil has been partially truncated.

3.0. MATERIALS AND METHODS

3.1. Introduction

The soil samples for this study were collected from the University of Nairobi’s field station at Kabete.The attached map clearly shows the location of the three land use referred as cropland, forest and grassland sites. Kabete is situated North of Nairobi city. Kabete lies at latitude of 1x 15’S and longitude of 36x 44’E, at an altitude of 1930 m above sea level; in an agro-climatic zone III (referred to as semi-humid) (Sombroek *et al.*, 1982). It has a bimodal distribution of rainfall, with long rains from early March to late May and the short rains from October to December with three months of dry period between the two rainy seasons. Mean annual rainfall of Kabete is 925 mm based on 27 years period (Taylor and Lawes, 1971).

The soils of Kabete are eutric nitosols (Sombroek *et al*, 1982) or Rhodic Paleadult according to United States Department of Agriculture Soil Taxonomy (Soil survey staff, 1985). The soils are dark red clay top soil, developed on tertiary trachytic lava. The soils are very deep and well drained.



SKETCH MAP SHOWING THE LOCATION OF THE THREE SITES

3.2. Soil Sampling Procedure

Soil samples for the determination of , bulk density, particle size distribution, organic matter content, soil moisture release characteristics, water stable aggregates, soil erodibility indices, and saturated hydraulic conductivity; were collected from three different land use sites. The land uses included; forest, grassland, and cropland that had been maintained so for a period of more than 10 years.

The soil is locally referred to as friable Kikuyu loams. These soil are deeper than 150 cm. This is a clay soil with kaolinite as the predominant clay mineral. The soil has predominantly Fe and Al oxides. The structure is sub-angular blocky with crumb structure at the surface. It has a good flocculation index and the consistency of the moist is friable. On each land use selected, soil samples were collected after laying randomly the 5 x 5 metre plot and the samples taken at the corners. The forest site was the one that was planted with Eucalyptus camaldulensis while the cropland site have been frequently planted with maize. Adjacent grassland site was used for grazing livestock. At each site disturbed and undisturbed soil samples were taken at 0, 10, 30, 60, and 100 cm depths from four representative sampling areas that formed the corners of a 5 by 5 m plot. The undisturbed soil samples were taken in core samplers of 5.0 cm outside diameter and 4.5 cm long. The disturbed soil samples were, however, taken in polythene bags.

3.3. Laboratory Analysis

Laboratory analysis of the required properties were carried out using standardized laboratory procedures of soil analysis

3.3.1. Bulk Density

The bulk density at various depths of the three land use sites were determined using the undisturbed core samples replicated four times. The soil samples were dried in a conventional oven at 105xC to a constant mass and weighed. The soil was then removed from the core samplers. The weight and volume of the core samplers were then determined. The bulk density was then calculated using the following relationship:

$$d_s = \frac{M_T - M_C}{V_s} \quad (1)$$

Where,

d_s = Bulk density, g cm^3

M_T = Mass of the core sampler and the oven dry soil, gm.

M_C = Mass of the empty core sampler, gm.

V_s = Volume of the soil sample, cm^3

3.3.2. Particle Size Distribution.

The particle size distribution of the soil samples collected from the different land use sites was determined by the hydrometer method as described in the Manual series No.1

of the International Institute of Tropical Agriculture [IITA], December 1979. In this method, 51.0 g of air dry soil which has been passed through a 2 mm sieve was weighed. The soil was then transferred to a "milk shake" mix cup after being treated with hydrogen peroxide to remove the organic matter, and 5.0 % Sodium hexametaphosphate along with 100 cc. of distilled water. The soil was then mixed with a stirring rod and let to settle for 30 minutes. The soil suspension was then stirred for 10 minutes with a multimix machine. The suspension was transferred from the cup to the glass cylinder. The top of the cylinder was then covered and inverted several times until all soil was in suspension. The cylinder was then placed on a flat surface and the time noted. Immediately the soil hydrometer was placed into the suspension. The hydrometer was slid down slowly into the suspension until it floated. The first reading on the hydrometer was recorded at 40 sec. The hydrometer was removed and the temperature recorded.

After the first hydrometer reading, the suspension was let to stand for 3 hours and the second reading taken: the temperature was also taken. The second reading indicates the percentage of 2 micron (total) clay in the suspension.

The percentages of sand, clay, and silt were then calculated using the formulae below.

$$SAND = 100.0 - [H_1 + 0.2 (T_1 - 68) -2] 2 \quad (2)$$

$$CLAY = [H_2 + 0.2 (T_2 - 68) - 2.0] 2 \quad (3)$$

$$SILT = 100.0 - (\% SAND + \% CLAY) \quad (4)$$

Where;

H_1 = Hydrometer reading after 40 seconds.

T_1 = Temperature reading after 40 seconds.

H_2 = Hydrometer reading after 3 hours.

T_2 = Temperature reading after 3 hours.

0.2 (T2 - 68) = Temperature correction to be added to hydrometer reading (T in degrees fahrenheit).

-2.0 = Salt correction to be added to hydrometer reading.

3.3.3. Organic Matter Content

The organic matter content of the soil samples from the different land use sites was determined by the loss-on-ignition as described by Ball (1964).

Four replicates of 10 gm soil samples were dried overnight in a conventional oven at 105xC, re-weighed and then ignited in a muffle furnace at 375xC plus/minus 5xC for 16 hours. The organic carbon and hence organic matter content was then calculated from the loss- on - ignition as follows:

$$C = 0.458L - 0.4 \quad (5)$$

$$O.M = 1.72C \quad (6)$$

Where;

C = The organic carbon.

L = The loss-on-ignition.

O.M = The organic matter content.

3.3.4. Soil moisture release characteristics.

The bottom of the undisturbed soil core samples collected from the selected land use sites for the determination of soil moisture release characteristics was covered with a cloth membrane held in position by a strong elastic band. The samples were then saturated by capillary action for three days. The outside of the samples were then dried, weighed, and subjected to 0.1, 0.33, and 0.7 bar pressure in a pressure plate apparatus. The samples were weighed after a minimum equilibration period of 4, 8 and 16 hours at each suction respectively. After the final weighing the soil was transferred quantitatively to pre- weighed containers, oven dried at 105^{0}C to a constant mass and weighed. The volumetric water content at any given suction was then calculated as follows:

$$Pvt = \left(\frac{MT - Me - Ms}{Ms}\right) 100\ ds \qquad (7)$$

Where, Pvt = Percentage volumetric water content at T bar pressure

MT = Mass of the core sampler and wet soil at Tbar pressure, g.

Me = Mass of the empty core sampler with the saturated cloth membrane and the elastic band, gm

Ms = Mass total oven dry soil, gm

d_s = Dry bulk density of the soil gcm^{-3}

Loose but otherwise undisturbed soil samples collected at the same depths were used for the pressure plate measurements at a higher suction range. These samples were subjected to 1, 5, and 15 bar pressure. For each depth, four replicated samples were placed in circular plastic retaining rings of 2.7 cm diameter and 1.1 cm depth, resting on a saturated pressure plate. The samples were saturated on the plate by maintaining excess water on the plate surface prior to applying pressure. Equilibrium was considered to have been attained when no further outflow greater than 0.5 ml was measured over a period of twelve hours. The samples were quickly transferred into pre-weighed containers, weighed, oven dried at 105^0C to a constant mass, and weighed. The volumetric water content was then calculated as follows:

$$Pvt = \frac{M_t - M_c}{M_s} (100\ d_s) \qquad (8)$$

Where:

Pvt= Percentage volumetric water content at T bar pressure

M_t = Mass of container + wet soil,g.

M_c = Mass of the container plus oven dry soil, g.

M_s = Mass of total oven dry soil, g.

3.3.5. Water Stable Aggregates

The portion of disturbed soil samples for the determination of water stable aggregates was air dried, passed, without forcing , through 4.75 and 2 mm sieves. The portion remaining on the 4.75 mm and that passing through 2 mm sieves were discarded. A 25 gm sample was taken from the fraction retained on the 2 mm sieve and added to the top of a set of 2, 1, 0.5, 0.25, and 0.06 mm sieves fixed on Endecotts test sieve shaker model EFL2 with a wet sieving attachment. The sample was then wetted with water and allowed to stand for 10 minutes. The water was then turned on to give a fine spray on the sample and then the shaker was switched on for 10 minutes. The fraction of the sample retained on each sieve at the end of the shaking period was quantitatively transferred to pre-weighed evaporating dishes , oven dried at 105^0C to a constant mass, weighed to 0.01 g, and expressed as a percentage of the total sample on oven dry basis.

3.3.6. Degree of Aggregation

The degree of aggregation was determined by dry sieving with a set of sieves with openings similar to those used in water stable aggregate analysis and using 25 g of air dry soil sample taken from the soil fraction passing through the 4.75 mm sieve.

3.3.7. Soil Erodibility Indices

The particle size distribution was determined using chemically and non - chemically dispersed samples as outlined by the manual series No 1 of IITA as described in section 3.3.2. The non-chemically dispersed samples were dispersed without adding any reagents. The two particle size distributions were used for the determination of the clay ration (Cr), dispersion ratio (Dr), erosion ratio (Er), colloid - moisture equivalent ratio (Cmr) and erosion index (Ei) as follows:

$$Cr = \frac{\%\ \text{sand}}{\%\ (\text{silt} + \text{clay})} \qquad (9)$$

$$Dr = \frac{\%(\text{Silt} + \text{clay})\ \text{undispersed sample}}{\%\ (\text{silt} + \text{clay})\ \text{dispersed sample}} \qquad (10)$$

$$Cmr = \frac{\text{Colloidalcontent}}{\text{moisture equivalent}} \qquad (11)$$

$$Er = \frac{\text{Dispersion ratio}}{\text{colloidal content/moisture equivalent ratio}} \qquad (12)$$

$$Ei = \frac{\text{Dispersion ratio}}{\text{clay/half water holding capacity}} \qquad (13)$$

Where colloidal content is the percentage sum of clay and organic matter content, while the moisture equivalent is the moisture content held at field capacity.

3.4. Saturated Hydraulic Conductivity

The constant head method as outlined by Black (1965) was used in the determination of a saturated hydraulic conductivity. The bottom of each soil sample was capped with cheese cloth filter using rubber bands while the top was trimmed to the ring volume. A second empty ring was connected to the top of the sample and the two rings firmly sealed at the junction using water proof adhesive tape. Samples were then placed in a tray and tap water was introduced to cover the sample rings up to about 1 cm from the top of the upper ring and left to saturate for at least 24 hours at room temperature. Soil cores, were then mounted vertically and supported on porous outflow surfaces connected to funnels leading to water receivers below each sample. A shallow column of water was maintained over the soil surface by a siphon tube from constant level reservoir. The system was given approximately 10 minutes to attain both a steady water column level over the soil surface and a steady water flow through the soil core. It was also established that no air bubbles were in the system to ensure constant and consistent water flow at appropriate moment. A water receiver was then placed under the funnel and simultaneously a stop clock was started. Collection of the out flowing water continued for the predetermined time (t) of 60 minutes. The quantity of water (Q) collected was measured in cm^3 and the shallow water column height (h) above the soil surface was measured in cm. Saturated hydraulic conductivity K was then calculated using Darcy's equation as follow;

$$K = \frac{Q}{At} \times \frac{L}{H} \qquad (14)$$

Where:

K =Saturated hydraulic conductivity (cm/min)

Q = Quantity of water collected after time t (cm^3)

A = Area of the ring (cm^2)

t = Time used in the experiment (1hr)

L = Length of the soil in the core through which water travelled (cm)

H = l + h (constant head) in cm

Each depth for the whole process was replicated four times.

3.5 Statistical analysis

The soil samples which were replicated four times had their results analysed using a 3 x 3 x 5 factorial design for bulk density, organic matter content, and saturated hydraulic conductivity. The aggregate stability for both wet and dry sieving were analysed using completely randomized design without blocks. Significant differences of these data have been reported in the results and discussion chapter accordigly. The data were analysed using statistical package (SPSS+ and SYSTAT packages)

4.0. RESULTS AND DISCUSSION

4.1. Soil physical characteristic

4.1.1. Bulk Density

The variation of the average bulk density with depth for the forest, grassland, and cropland sites at Kabete are shown in table 1. The bulk density for the forest site varied with depth from 1.04 g cm^{-3} at the surface to 1.02 g cm^{-3} at 100 cm soil depth. Except for the bulk density value of 0.98 at 60 cm soil depth, the bulk density values were fairly uniform with depth and were not significantly different (at $P < 0.05$).

The bulk density for the grassland varied from 1.05 g cm^{-3} at the soil surface to 1.08 g cm^{-3} at 100 cm soil depth. The bulk density values increased with soil depth from 1.05 g cm^{-3} at the soil surface to 1.21 g cm^{-3} at 60 cm soil depth before decreasing to 1.08 g cm^{-3} at 100 cm soil depth. However, this increase in bulk density values did not give a significant difference (at $P < 0.05$).

The bulk density values for the cropland site increased uniformly from 1.09 g cm^{-3} at the soil surface to 1.28 g cm^{-3} at 100 cm soil depth. This increase in bulk density did not give a significant difference ($P < 0.05$).

A comparison of the bulk densities for the three sites showed that the forest site had the lowest bulk density at all soil depths followed by grassland and cropland sites respectively except at 30 and 60 cm soil depths where the bulk densities for the cropland site, were lower than those of the grassland site. Forest soils generally have surface horizons lower in density than agricultural soils. The lower bulk densities observed at the forest site could be attributed to the accumulation of humus in the forest soil. The low value observed at 30 and 60 cm soil depths at the cropland site could be attributed to the fact that during ploughing the organic matter is incorporated at those two depths. The same result was observed by Coile (1940), where he reported that because of the accumulation of litter and formation of humus, forest soils have lower densities of the A layer (particularly the A1) than agricultural soils; however, soil density depends more upon the physical structure of the soil than upon land use. The higher bulk value at 60 cm soil depth at the grassland site compared to 100 cm soil depth had no concrete explanation as this may have been due to the differences existing within the soil. Cultivation that has been going at the cropland site may have contributed to the higher value recorded at 100 cm soil depth.

Table 1. The variation of average bulk density with depth for forest, grassland and cropland sites.

Soil profile depth cm	Bulk density, $g\ cm^{-3}$ Forest site	Grassland site	Cropland site
0	1.04	1.05	1.09
10	1.03	1.05	1.10
30	1.04	1.18	1.14
60	0.98	1.21	1.18
100	1.02	1.08	1.28

4.1.2. <u>Organic Matter Content</u>

The variation of the average organic matter content with depth for the forest, grassland, and cropland sites are shown in table 2. The organic matter content for the forest, grassland, and cropland sites decreased from 9.6%, 5.9%, and 4.6% at the soil surface to 1.7%, 2.5%, and 1.7%, at 100 cm soil depth respectively. The statistical analysis done on organic matter showed that the variation of organic matter with depth at each site from 0 cm to 100 cm soil depth was significantly different at both 1 and 5% level. The variation of organic matter with depth for the three sites also showed significant differences at both levels.

The values presented in table 2 shows the effect of different land uses on the soil organic matter content. The high organic matter content recorded at the soil surface on the forest site could be attributed to the addition of dead plant and animal residue to the soil surface which occurs each year. The annual accumulation of organic matter deposited on the forest floor varies with age, species, stand density, site quality, and climate. One of the most obvious effects that vegetation has on the soil supporting it is the deposition of dead plant matter, leaves, twigs, fruits, branches and even dead stemwood. The result of this is an increase in organic matter in the top soil. The increase in organic matter in the top soil is well seen in table 2 where at the 10 cm soil depth, the forest site lead by having 4.2% when compared to the grassland and the cropland sites which had both 3.5%. Another reason for the high organic matter content at the forest site could be due to the fact that when forest trees occupy a site for many years, they send their roots well into the subsoil. During that period considerable amounts of organic materials are returned to the soil in the form of leaf and litter fall, and decaying roots.

Further more observations on the data presented in table 2 indicate that the cropland site had less organic matter content at the top 30 cm soil depth when compared to the other sites. The same observation has been reported by Jones (1972), and Mwonga & Mochoge (1986), where they found that soil organic matter is most affected by cultivation.

The decline of organic matter is coupled with nitrogen decline and deterioration of soil structure and aggregate stability which both results in poor rainfall acceptance and higher soil erosion intensities (Okigbo and Lal, 1979). Contradictory data have been found concerning the effects of cultivation on the distribution of soil organic matter. Keeney and Bremmer (1964), concluded that cultivation did not result in marked changes in relative proportion of hydrolysable forms of nitrogen. Results in this study supports that of Porter et al, (1964), who showed that cultivation led to general decrease of organic matter content. Though the original condition of the soil at the three sites were unknown in terms of their physical and chemical properties, the difference in the land use may have contributed to the difference on the recorded organic matter content values at the three sites where the forest lead by recording 9.6% at the soil surface while cropland had a value of 4.6%

The quantity of organic matter contained in soils is important from many points of view. With respect to erodibility, its greatest effect is on structure. The organic fraction of the soil has a greater capacity for absorbing and storing water than the mineral fraction, but its most important effect is in forming water - stable aggregates that increase the porosity and permeability of the soil. Soils with low organic matter content, like those found on the cropland site are, therefore, subject to comparatively rapid erosion than the soils on the forest and grassland sites. The forest site soils had the highest organic matter content and were therefore expected to contain more water stable aggregates and hence be less erodible.Increasing amounts of organic matter in soils is usually accompanied by increased aggregate porosity, lowered aggregate density, and a narrower range in aggregate size distribution, which result in lowered soil bulk density. Changes in stability of soil structure are more pronounced than changes in organic matter.

Table 2. The variation of organic matter content with depth for the forest, grassland, and cropland sites

Soil profile depth cm	organic matter content, %		
	Forest site	Grassland site	Cropland site
0	9.6	5.9	4.6
10	4.2	3.5	3.5
30	2.6	3.1	1.9
60	2.1	2.6	2.1
100	1.7	2.5	1.7

4.1.3. Particle Size Distribution

Table 3 shows the particle size distribution for the forest, grassland, and cropland sites. For the forest site sand decreased gradually with depth and ranged from 47.8% at the soil surface to 33.1% at 100 cm soil depth. Silt on the other hand, decreased randomly with depth from 36.9% at the soil surface to 14.9% at 100 cm soil depth. The textural classification for the forest site changed from loam at the soils surface to clay loam at both 10 and 30 cm soil depth and finally to clay at both 60 and 100 cm soil depth.

The grassland site showed a different pattern of particle size distribution. Sand increased from 33.1% at the soil surface to 41.0% at 10 cm soil depth before decreasing with soil depth to 39.8% at 100 cm soil depth. Silt, on the other hand, varied randomly from 30.3% at the soil surface to 39.2% at 100 cm soil depth. The clay content decreased gradually with depth from 36.6% at the soil surface to 21.0% at 100 cm soil depth. The textural classification for the grassland site changed from clay loam at 0, 10, and 30 cm soil depths to loam at the 100 cm soil depth.

The sand content on the cropland site decreased randomly from 16.8% at the soil surface to 15.3% at 100 cm soil depth. Silt showed a gradual decrease with depth from 35.0% at the soil surface to 19.5% at 100 cm soil depth.The clay content showed a similar trend to that of the forest site by increasing with depth from 48.2% at the soil surface to 65.2% at 100 cm soil depth. .The textural classification for this site indicated clay throughout the profile.

Sand, silt, and clay have been different for the three sites mainly because of the inherent characteristics that existed before. The texture of a soil controls its drainage, water storage, workability properties, and suitability for different crops. Texture also plays a major part in determining the soil "structure" which is the arrangement of individual particles into larger units or aggregates which are also referred to as structures or peds. Because soil structure controls the agriculturally vital processes of water movement and root growth it has received a great deal of attention in recent years with changes in farming practices. The increasing weight and power of tractors on many arable farms has made soil structure more vulnerable to damage. Looking at the cropland site where the profile is dominated by clay, it can be said that generally soils containing a lot of clay are stable and do not collapse or "slake" when wet although they will readily compact under the weight of implements. This is shown in table 1 where the cropland site showed higher bulk density values when compared to the other sites.

Table 3. Soil particle size and distribution for the forest, grassland and cropland site.

Site	Depth cm	Particle size distribution (%) sand	silt	clay	Textural classification
F	0	47.8	36.9	15.4	LOAM
	10	44.5	28.0	27.5	CLAY LOAM
	30	41.0	29.4	29.6	CLAY LOAM
	60	39.1	20.0	41.0	CLAY
	100	33.1	14.9	52.0	CLAY
G	0	33.1	30.3	36.6	CLAY LOAM
	10	41.0	27.4	31.6	CLAY LOAM
	30	34.0	38.0	28.1	CLAY LOAM
	60	39.0	35.4	25.6	LOAM
	100	39.8	39.2	21.0	LOAM
C	0	16.8	35.0	48.2	CLAY
	10	16.3	30.5	53.2	CLAY
	30	18.3	21.0	60.7	CLAY
	60	16.8	20.0	63.2	CLAY
	100	15.3	19.5	65.2	CLAY

F = Forest

G = Grassland

C = Cropland

4.2 Soil Structural Characteristics

4.2.1. Soil Aggregate Stability

The particle size distribution of aggregates after wet sieving for the forest, grassland, and cropland sites are shown in tables 4, 5, and 6. The proportion of soil water stable aggregates less than 0.5 mm in diameter has been used as an index of soil erodibility (Bryan, 1974; Rai et al,1954). The percentage of the aggregates less than 0.5 mm for the forest site increased from 15.3% at the soil surface to 21.0% at 100 cm soil depth, while those of the grassland increased from 13.3% at the soil surface to 15.7% at 30 cm soil depth before decreasing to 11.1% at 100 cm soil depth. The particle size distribution may have been a probable cause to this. The percentage of the aggregates less than 0.5 mm for the cropland decreased from 24.4% at the soil surface to 18.4% at 100 cm soil depth. This decrease could be attributed to the high clay content that increased with depth from 48.2% at the soil surface to 65.2% at 100 cm soil depth on the cropland. The soil erodibility of the forest and grassland sites increased with depth, while that of the cropland decreased with depth. This is qualified by the trend on how the relative proportions of both forest and grassland site behaved. However the relative erodibility of the cropland site was higher than that of the forest and grassland sites at all depths.

Soils that breakdown into many small aggregates or primary particles are considered more erodible than soils that break down into intermediate size aggregates or remain stable. Soil structure is a very important factor in soil erosion because it largely determines the rate at which water can enter into the soil as well as the resistance of soil particles to detachment by rainfall impact and subsequent removal by surface runoff. The stability of any given soil structural organization is important in relation to soil erodibility. This arises firstly in relation to the ease of detachment of particles from aggregates and secondly in relation to the detached smaller particles of clay and silt which are likely to be washed into the coarser pores of the existing structure and cause a decrease in its hydraulic conductivity.

The statistical analysis for the 0.5 mm aggregates showed significant differences between the forest, grassland, and cropland sites at the 1 % level. The difference between 0.5 mm aggregates was due to inherent differences in the soils at the three sites. Looking at their three means as shown in appendix 7, the forest, grassland, and cropland sites had 17.24%, 13.57% ,and 21.19% respectively. As indicated earlier this shows that the erodibility of cropland site was higher than the rest, followed by forest and lastly the grassland site.

Tables 7, 8 and 9 show the particle size distribution of aggregates after dry sieving for the forest, grassland, and cropland sites. These values show the degree of soil aggregation which is an indication of the soils susceptibility to wind erosion. The distribution of aggregates of size less than 0.5 mm for the forest site decreased from 16.1% at the soil's surface to 11.3% at 10 cm soil depth before starting to increase again to 24.1% at 100 cm soil depth. For the grassland site the aggregates increased with depth from 5.1 % at the soil surface to 29.7% at 30 cm soil depth before starting to decrease up to 23.7% at 100 cm soil depth. The cropland site showed a general trend in which the aggregates decreased with depth from 13.2% at the soil's surface to 3.2% at the 100 cm soil depth.

The statistical analysis carried out for the three sites showed that there was a significant difference between the three sites at 5 % level. The means for the three sites as shown in appendix 8 was 19.5%, 19.03%, and 6.03 for forest, grassland, and cropland sites respectively. The data indicates that the differences between the forest and grassland sites in terms of susceptibility to wind erosion was more or less the same.

Table 4. Particle size distribution of aggregates after wet sieving for the forest site

Soil profile depth cm	Particle size % retained at each sieve set.					
	5-2mm	2-1mm	1-0.5mm	0.5-0.25mm	0.25-0.06mm	<0.06mm
0	9.4	12.8	15.2	15.3	14.1	33.3
10	3.0	5.4	17.0	15.7	17.1	41.8
30	0.4	3.8	10.7	18.4	23.7	43.1
60	0.6	2.0	9.0	15.9	28.0	44.5
100	0.5	2.9	11.9	21.0	26.6	37.2

Table 5. Particle size distribution of aggregates after wet sieving for the grassland site

Soil profile depth cm	Particle size % retained at each sieve set					
	5-2mm	2-1mm	1-0.5mm	0.5-0.25mm	0.25-0.06mm	<0.06mm
0	6.0	8.1	8.7	13.3	16.3	47.6
10	6.2	13.3	16.0	12.9	17.0	34.6
30	2.5	7.1	16.1	15.7	19.0	39.7
60	5.5	11.2	9.1	15.1	18.4	40.8
100	10.5	12.4	8.2	11.1	19.1	38.7

Table 6. Particle size distribution of aggregates after wet sieving for the cropland site

Soil profile depth cm	Particle size % retained at each sieve set					
	5-2mm	2-1mm	1-0.5mm	0.5-0.25mm	0.25-0.06mm	<0.06mm
0	0.9	3.3	7.5	24.4	26.1	37.7
10	0.0	2.3	13.7	21.9	20.4	41.7
30	0.1	1.3	6.9	21.3	24.5	45.7
60	0.1	0.7	7.5	20.1	24.4	47.2
100	0.0	1.3	6.8	18.4	19.2	52.3

Table 7. Particle size distribution of aggregates after dry sieving for the forest site

Soil profile depth cm	Particle size % retained at each sieve set					
	5.2mm	2-1mm	1-0.5mm	0.5-0.25mm	0.25-0.06mm	<0.06mm
0	25.7	23.5	21.6	16.1	8.1	5.2
10	29.2	30.2	20.2	11.3	4.7	4.6
30	17.2	22.1	21.1	21.7	13.7	4.2
60	12.2	18.1	20.9	24.5	18.6	5.8
100	6.3	19.7	20.4	24.1	22.0	7.5

Table 8. Particle size distribution of aggregates after dry sieving for the grassland site

Soil profile depth cm	Particle size % retained at each sieve set				
	5-2mm	2-1mm	1-0.5mm	0.5-0.25mm	0.25-0.06mm	<0.06mm
0	52.8	24.8	11.9	5.1	1.6	3.9
10	27.7	25.0	23.1	15.5	5.3	5.5
30	9.5	12.3	20.8	29.7	21.4	6.3
60	18.1	17.1	17.2	21.4	18.3	7.8
100	11.2	13.6	17.0	23.7	22.8	11.6

Table 9. Particle size distribution of aggregates after dry sieving for the cropland site

Soil profile depth cm	Particle size % retained at each sieve set					
	5-2mm	2-1mm	1-0.5mm	0.5-0.25mm	0.25-0.06mm	<0.06mm
0	23.6	28.5	28.1	13.2	5.5	0.73
10	50.1	29.5	13.9	4.3	1.7	0.2
30	46.9	28.7	18.4	4.7	0.9	0.4
60	47.9	28.0	15.7	4.8	2.4	1.2
100	58.5	24.9	10.4	3.2	2.0	0.9

4.2.2. Soil Erodibility Indices

The most useful information on structural stability of soil is provided by the ease and degree of dispersion of soil particles, together with the degree of aggregation. The measurement of dispersion and aggregation are complementary in that particles of soil not dispersed by water remain aggregated or clustered in granules which favor high infiltration of rain water. Tables 10, 11 and 12 show the soil erodibility indices for the forest, grassland, and cropland sites. The dispersion ratio for the forest, grassland, and cropland sites varied between 24.7, 55.7, and 30.0 at the soil surface to 49.4, 57.0, and 38.4 at the 100 cm soil depth respectively. The dispersion ratio for forest and cropland increased with depth. The dispersion ratio for grassland, however, decreased with depth from 55.7 at the soil surface to 37.2 at 30 cm soil depth before increasing to 57.0 at 100 cm soil depth. The dispersion ratio is an index of the ease with which soil particles can be brought into suspension by the action of rain drops or runoff water, and therefore the greater the ratio, the more easily the soil can be dispersed and the higher the erosion rate. Comparison of the dispersion ratio values for the three sites at the soil surface shows that the grassland site was more susceptible to erosion followed by cropland and the forest sites.

The colloid - moisture equivalent ratio shown in tables 10, 11 and 12 expresses the relative permeability of soil to water. The values for this ratio increased with depth for the forest and cropland sites and it ranged from 0.58 and 1.26 at the soil surface to 1.26 and 1.38 at 100 cm soil depth respectively. The percolation rate of these two sites could therefore be assumed to increase with depth. For the grassland site, however, the values decreased with depth and varied between 1.08 at the soil surface to 0.57 at 100 cm soil depth. The percolation rate for this site could therefore be assumed to decrease with depth.

Since the erosion rate increases directly with dispersion ratio and inversely with colloid moisture equivalent ratio, the erosion ratio was also evaluated as shown in tables 10, 11, and 12. The erosion ratio for the forest site increased with depth from 42.6 at the soil surface to 56.7 at 30 cm soil depth before decreasing to 39.2 at the 100 cm soil depth. The erosion ratio for the grassland and cropland sites increased with soil depth except for an odd value of 36.1 for the grassland site. The clay content for forest and cropland sites increased with depth as discussed in section 4.1.3. and therefore the erosion ratio decreased with the increase in clay content. The clay content for the grassland site however varied randomly with depth.

Table 10. Soil erodibility indices for the forest site.

Soil profile depth cm	Silt + Clay (undispersed) %	Moisture Equivalent(ME) %	Colloidal content /ME %	Clay ratio	Dispersion ratio	Erosion ratio	Erosion Index
0	12.98	42.8	0.58	0.91	24.73	42.6	50.3
10	18.56	44.4	0.71	0.80	33.47	47.1	37.3
30	26.42	40.5	0.79	0.69	44.78	56.7	48.7
60	27.92	38.2	1.13	0.64	48.81	43.2	40.8
100	33.06	42.7	1.26	0.49	49.43	39.2	31.6

Table 11. Soil erodibility indices for the grassland site.

Soil profile depth cm	Silt+Clay (Undispersed) %	Moisture Equivalent(ME) %	Colloidal Content /ME %	Clay ratio	Dispersion ratio	Erosion ratio	Erosion Index
0	37.30	39.2	1.08	0.49	55.72	51.6	46.6
10	18.34	40.8	0.86	0.69	31.04	36.1	31.5
30	24.61	47.2	0.66	0.51	37.24	56.4	41.0
60	24.57	47.5	0.59	0.64	40.22	68.2	46.5
100	34.30	40.9	0.57	0.66	56.96	99.9	88.8

Table 12. Soil erodibility indices for the cropland site.

Soil profile depth cm	Silt+clay (Undispersed) %	Moisture Equivalent (ME) %	Colloidal Content /ME %	Clay ratio	Dispersion ratio	Erosion ratio	Erosion Index
0	25.0	41.9	1.26	0.20	30.04	23.8	17.6
10	28.0	43.3	1.29	0.19	33.44	25.9	18.4
30	29.5	44.1	1.42	0.22	36.10	25.4	17.3
60	32.0	45.3	1.44	0.20	38.45	26.7	18.0
100	32.5	48.2	1.38	0.18	38.36	27.8	16.9

4.3. Soil Hydrological Characteristics

4.3.1. Pore Size Distribution

The type and arrangement of soil particles determines the amount and nature of the pores. Tables 13, 14 and 15 and figures 3, 4 and 5 show the variation in pore size distribution with depth for the forest, grassland, and cropland sites. The total pore space for the forest, grassland, and cropland increased randomly with depth from 62.5%, 61.3%, and 56.6% at the soil surface to 66.5%, 65.5%, and 57.5% at 100 cm soil depth. The cropland site had lower total pore space than the forest and grassland sites at all depths. The macropores made up of very freely drained and freely drained pores for the forest and grassland sites increased randomly with depth from 19.7% and 22.1% at the soil surface to 22.1% and 24.6% at the 100 cm soil depth. The percentage macropores for the cropland site, however, increased from 14.7% at the soil surface to 15.4% at 10 cm soil depth before decreasing drastically to 9.3% at the 100 cm soil depth. The capillary pores for the three sites behaved in a similar manner. However the forest site had high capillary pores retaining more available water content than grassland and cropland site at all depths. Conversely, the cropland site showed a higher capillary pores retaining unavailable water than the forest and grassland sites at all depths. The low macropores and capillary pores retaining available water for the cropland site could be attributed to structural degradation due to cultivation.

Table 13. Pore size distribution for the forest site.

Soil profile depth cm	Total pore space	Pore Size Distribution (%) Macro Pores: Very freely drained+freely drained	Capillary Pores: Retaining available water	Capillary Pores: Retaining un available water
0	62.5	19.7	22.8	20.0
10	61.3	16.9	19.0	25.4
30	64.3	23.8	17.6	22.9
60	68.4	30.2	14.7	23.5
100	66.5	22.1	20.1	24.3

Table 14. Pore size distribution for the grass land site.

Soil profile depth cm	Total pore space	Pore Size Distribution (%)		
		Macropores	Capillary Pores	
		Very freely drained+freely drained	Retaining available water	Retaining unavailable water
0	61.3	22.1	11.5	27.7
10	64.2	23.4	9.1	31.7
30	61.9	14.7	17.8	29.4
60	59.3	11.8	16.1	31.4
100	65.5	24.6	12.6	28.3

Table 15. Pore Size Distribution for the crop land site.

Soil profile depth cm	Total pore space	Pore Size Distribution (%)		
		Macropores	Capillary Pores	
		Very freely drained + freely drained	Retaining available water	Retaining unavailable water
0	56.6	14.7	11.0	30.9
10	58.7	15.4	12.3	31.0
30	58.3	14.2	15.4	28.7
60	59.2	13.9	11.6	33.7
100	57.5	9.3	8.2	40.0

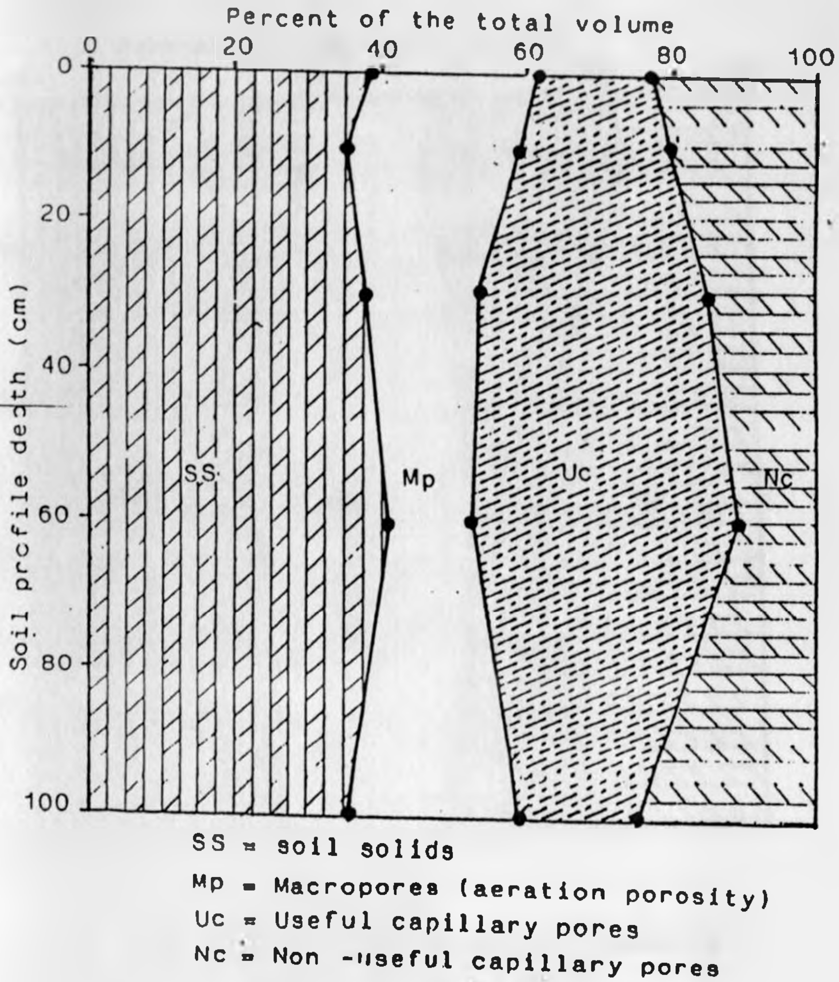
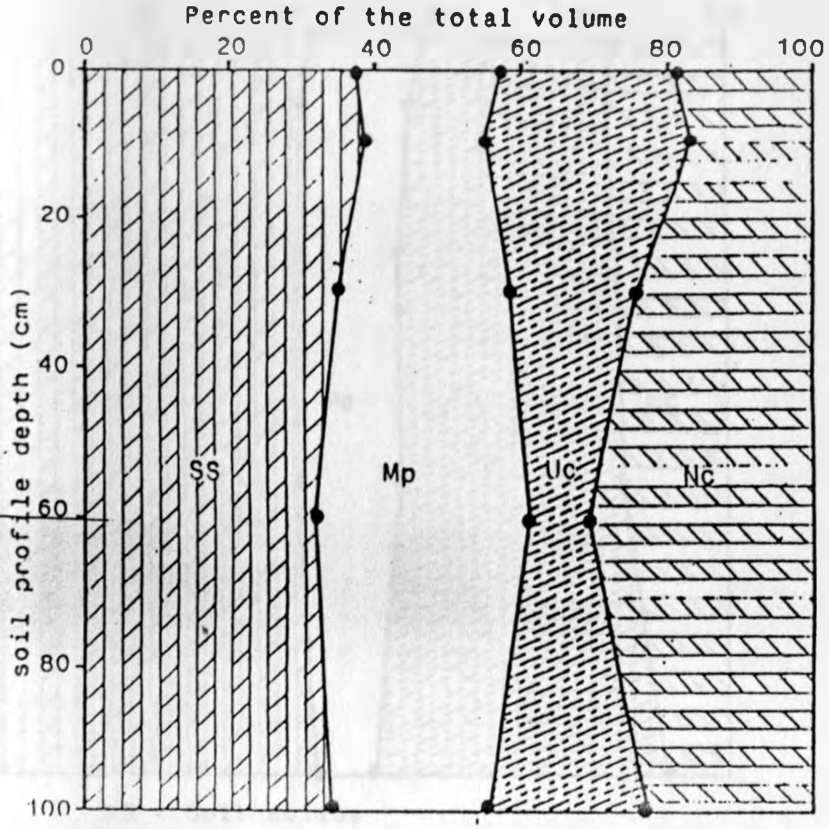
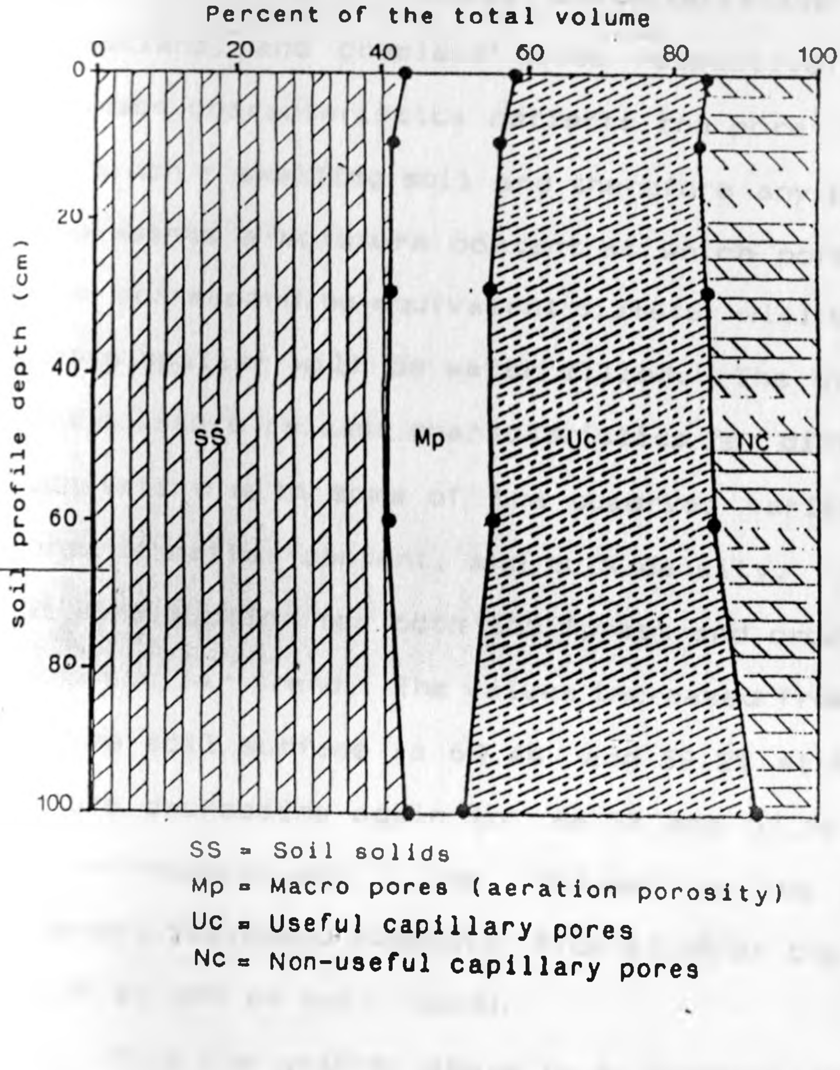


Figure 5. Variation of pore size distribution with depth for the forest site



SS = Soil solids
Mp = Macropores (aeration porosity)
Uc = Useful capillary pores
Nc = Non-useful capillary pores

Figure 4 Variation of pore size distribution with depth for the grassland site



gure 5. Variation of pore size distribution with depth for the cropland site

4.3.2. Soil Moisture Release Characteristics

Tables 16, 17, and 18 and figures 6, 7 and 8 show the average moisture release characteristics for the forest, grassland, and cropland sites respectively. The moisture release characteristics reflects the pore size distribution in a non - swelling soil and therefore any point on the curve represents a moisture content at which pores of larger than the corresponding equivalent diameter will be air filled, and those smaller will be water filled. The difference between the moisture release characteristics for different depth were consistent with some of the observed variation in texture, organic matter content, and bulk density. Moisture content at zero suction for both the forest and cropland site showed some similar trend. The values increased from 62.5% and 56.6% at the soil surface to 68.4% and 59.2% at 60 cm soil depth before decreasing again to 66.5% and 57.5% at 100 cm soil depth respectively. The values for the grassland site, however, increased randomly from 61.3% at the soil surface to 65.5% at 100 cm soil depth.

From the graphs, there is a tendency for the P^F curves to be approximately linear especially at the forest and cropland sites over the range of 3.0 to 4.18. For the cropland site (which is predominantly occupied by clay as shown in section 4.1.3), the graphs show that the suction increased more gradually with decreasing water content and at any given suction there is more water retained in the soil. Usually, as the soil dries out, the suction of the plant root must increase more rapidly than does the soil suction itself. The energy level of the water in the plant, depends not only on the energy level of the soil water, but also on the rate of water uptake, especially at soil suctions above 1 bar. Those with large channels of entry, in which only gentle interface curvatures can be maintained will empty at low suctions, while those with narrow channels of entry, supporting interfaces of sharp curvature, will not empty until larger suctions are imposed. Hence, as the soil water suction is increased, the soil moisture content is reduced, the larger pores emptying at the lower suctions and the smaller pores at the higher suctions. It should also be emphasized that the availability of water to plants is indicated by the shape of the percentage moisture/log tension curve rather than the position of the curve relative to the abscissa. The amount of water remaining in the soil at equilibrium is a function of the sizes and volumes of the water-filled pores and hence it is a function of the matric suction. The amount of water retained at relatively low values of matric suction (say, between 0 and 1 bar of suction) depends primarily upon the capillary effect and the pore- size distribution, and hence is strongly influenced by the structure of the soil. On the other hand, water retention in the higher suction range is due increasingly to adsorption and is thus influenced less by the structure and more by the texture and specific surface of the soil material. Soil structure, also affects the shape of the soil moisture characteristic curve, particularly in the low suction range. The effect of compaction upon a soil is to decrease the total porosity and, especially, to decrease the volume of the large interaggregate pores. The slope of the soil moisture characteristic curve, which is the change of water content per unit change of matric potential, is generally termed as the differential water capacity. This is an important property in relation to soil moisture storage and availability to plants. The actual value of differential water capacity depends upon the wetness range, the texture, and the hysteresis effect. The observed variation in the measured water content in the lower suction range is partly due to structural differences between replicate samples and partly to small differences in particle size distribution and organic matter content. Water content retention in the higher suction range is mainly related to texture; this is why smaller variability was observed between replicate samples in the higher suction range as the texture at each sampling depth on each site was fairly uniform.

Table 16. Percentage volumetric moisture content for different suctions of the forest site

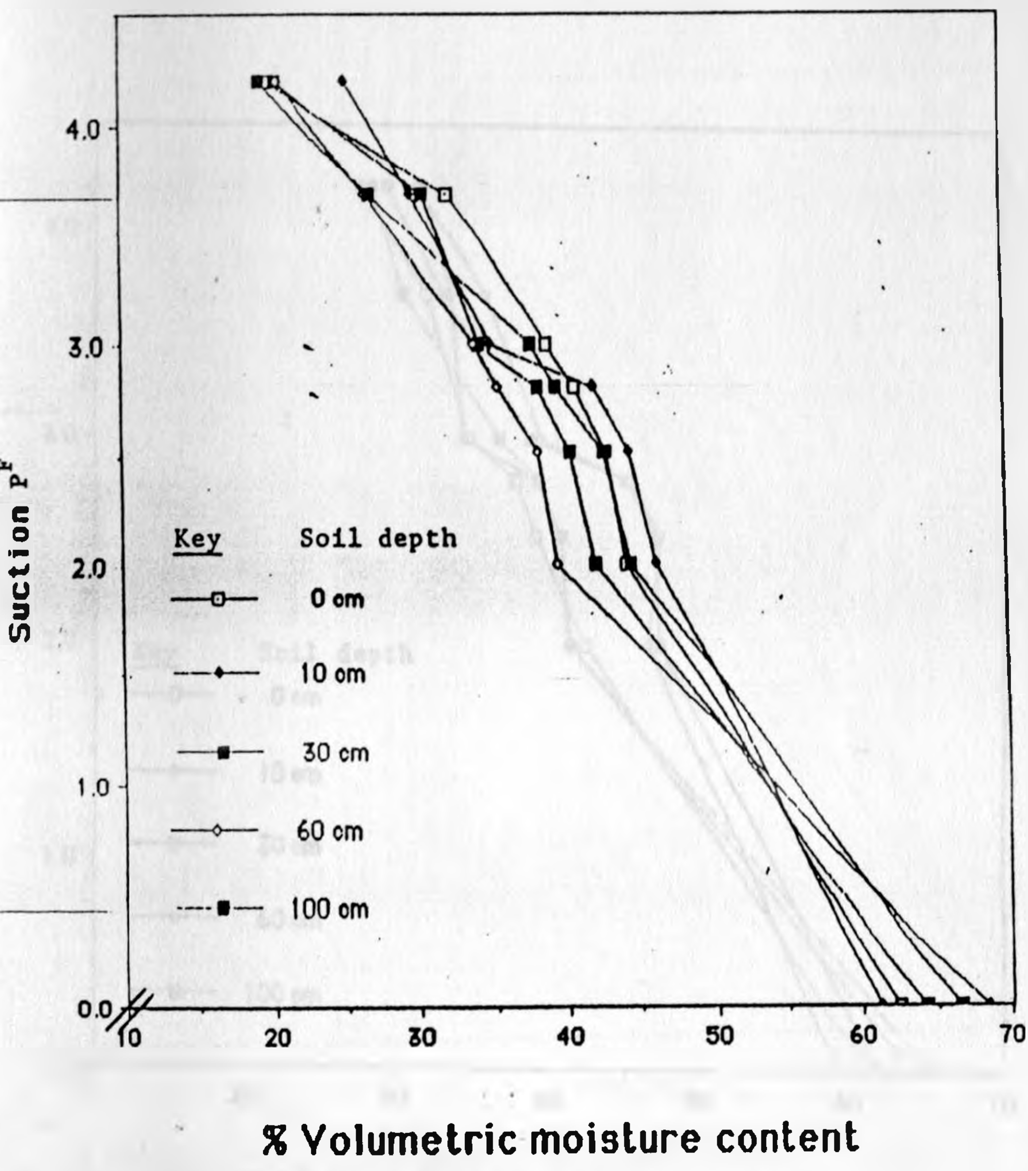
Soil profile depth cm	Suction in Bars						
	0	0.1	0.33	0.7	1	5	15
0	62.5	44.2	42.8	40.8	38.9	32.4	20.0
10	61.3	46.2	44.4	42.1	35.0	30.0	25.4
30	64.3	42.0	40.5	38.5	34.2	30.8	20.7
60	68.4	39.5	38.2	35.6	33.9	26.8	20.9
100	66.5	44.4	42.7	39.6	37.7	27.0	19.5

Table 17. Percentage volumetric moisture content for different suctions of the grassland site.

Soil profile depth cm	Suction in Bars						
	0	0.1	0.33	0.7	1	5	15
0	61.3	42.4	39.2	37.7	34.6	32.0	27.7
10	64.2	41.3	40.8	39.2	34.4	33.6	31.7
30	61.9	47.4	47.2	45.1	38.7	33.1	29.4
60	59.3	46.6	47.5	45.4	39.4	35.8	31.4
100	65.5	41.4	40.9	39.3	36.6	30.4	28.3

Table 18. Percentage volumetric moisture content for different suctions of the cropland site

Soil profile cm	Suction in Bars depth						
	0	0.1	0.33	0.7	1	5	15
0	56.6	42.4	41.9	40.2	36.3	32.8	30.9
10	58.7	43.7	43.3	41.8	38.1	33.3	31.0
30	58.3	44.5	44.1	42.5	35.0	32.1	28.7
60	59.2	47.2	45.3	44.8	41.0	36.5	33.7
100	57.5	49.9	48.2	47.5	44.7	42.6	40.0


Suction pF
4.0
3.0
2.0
1.0
0.0
10
20
30
40
50
60
70
% Volumetric moisture content
Key
Soil depth
0 cm
10 cm
30 cm
60 cm
100 cm

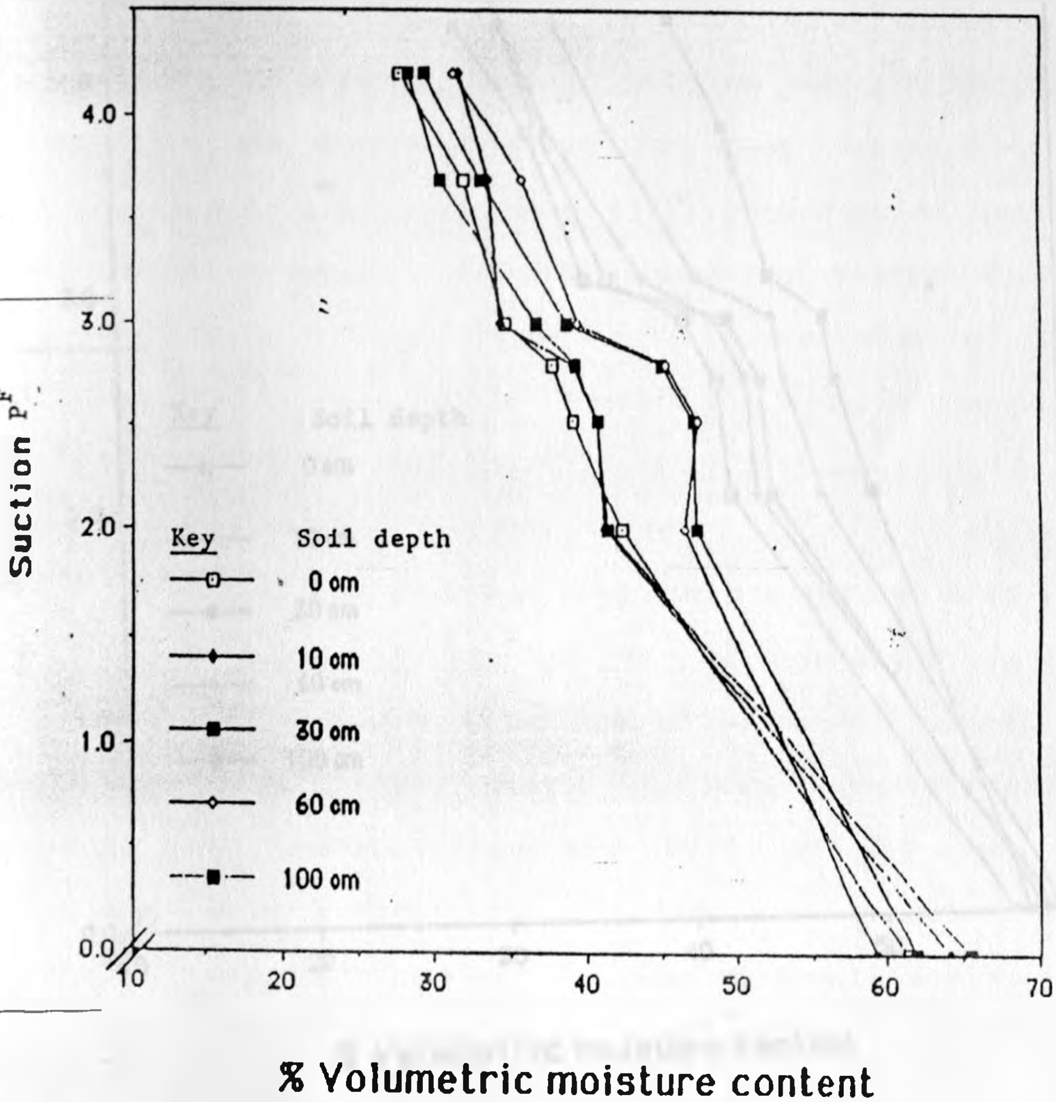


Figure 7 Average soil moisture release characteristics for the grassland site.

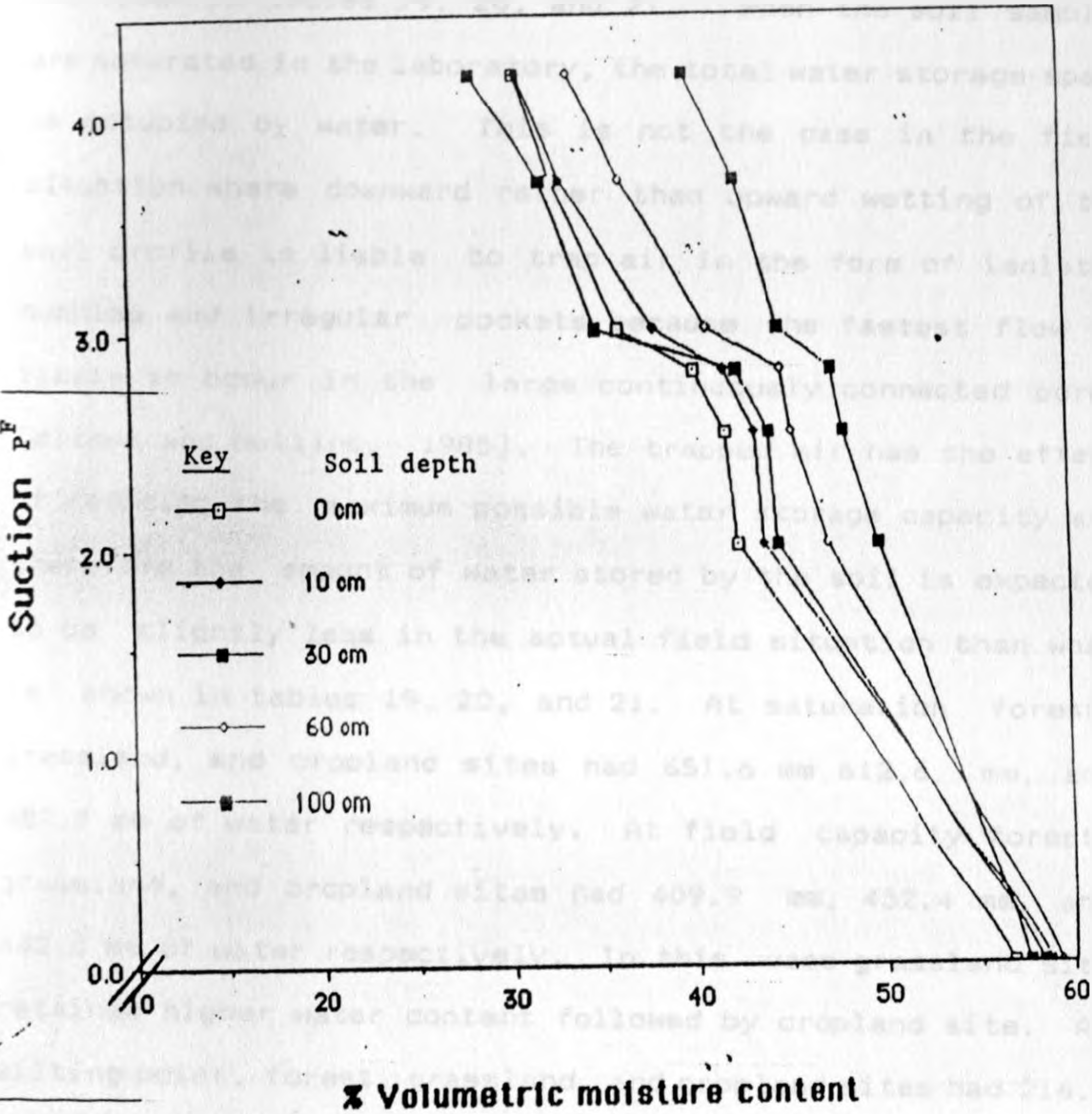


Figure 8. Average soil moisture release characteristics for the cropland sites.

4.3.3 Profile Water Holding Capacity

The running totals for the maximum profile water storage capacity at saturation, field capacity, and permanent wilting point for the forest, grassland, and cropland sites are shown in tables 19, 20, and 21. When the soil samples are saturated in the laboratory, the total water storage space is occupied by water. This is not the case in the field situation where downward rather than upward wetting of the soil profile is liable to trap air in the form of isolated bubbles and irregular pockets because the fastest flow is likely to occur in the large continuously connected pores (Kilewe and Mullins, 1985). The trapped air has the effect of reducing the maximum possible water storage capacity and therefore the amount of water stored by the soil is expected to be slightly less in the actual field situation than what is shown in tables 19, 20, and 21. At saturation forest, grassland, and cropland sites had 651.6 mm 612.6 mm, and 587.7 mm of water respectively. At field capacity forest, grassland, and cropland sites had 409.9 mm, 452.4 mm, and 442.0 mm of water respectively. In this case grassland site retained higher water content followed by cropland site. At wilting point, forest, grassland, and cropland sites had 216.5 mm, 304.9 mm, and 313.8 mm of water respectively.

Field capacity, is at times referred to as field carrying capacity, normal moisture capacity and the capillary capacity. While deep soils reach field capacity rather quickly, the presence of a water table near the soil surface will prolong the time required for drainage. Lack of homogeneity in the soil will also affect the water content at field capacity. For example a fine textured soil overlying a coarse textured soil will have a higher water content than a uniformly fine textured soil. Thus the field capacity of a soil is related to the conditions under which it is measured as well as to the characteristics of the soil itself. As field capacity has no fixed relationship to soil water potential, it can not be regarded as a soil moisture constant. The amount of water retained at field capacity decreases as the soil temperature increases (Richards and Weaver, 1944).

Table 19. Profile water holding capacity in mm of water for the forest site.

Soil profile depth cm	Water holding capacity in mm of water		
	At Saturation	At field capacity	At wilting point
0 - 10	62.5	42.8	20.0
10 - 30	185.1	131.6	70.8
30 - 60	378.0	253.1	132.9
60 - 100	651.6	405.9	216.5

Table 20. Profile water holding capacity in mm of water for the grassland site.

Soil profile depth cm	Water holding capacity in mm of water		
	At Saturation	At field capacity	At wilting Point
0 - 10	61.3	39.2	27.7
10 - 30	189.7	120.8	91.1
30 - 60	375.4	262.4	179.3
60 - 100	612.6	452.4	304.9

Table 21. Profile water holding capacity in mm of water for the cropland site.

Soil profile depth cm	Water holding capacity in mm of water		
	At saturation	At field capacity	At wilting point
0 - 10	56.6	41.9	30.9
10 - 30	174.0	128.5	92.9
30 - 60	348.9	260.8	179.0
60 -100	587.7	442.0	313.8

4.3.4. Available Water Capacity

The assessment of the available water content of a soil depends on the accurate measurement of the upper limit, field capacity, and the lower limit, the permanent wilting point of that particular soil. Not all the water remaining in the root zone can be taken up by the plant as rapidly as needed because it is held tightly by the soil particles. For any given soil there is an upper and lower limit to the amount of water that is available for plant use. These limits are those described earlier as "field capacity" and "permanent wilting point". Water that is present above field capacity drains off to lower horizons while that found below the permanent wilting point is not available for plant use. It is therefore more appropriate to speak of the available water capacity of the soil. The available water capacity is the amount of water held between the upper and the lower limits of soil water content expressed as percent by volume or more usefully in millimeters of water for a given soil depth. This is regarded as an index of the ability of a soil to store water and thus allow plants to maintain growth during dry periods.

Tables 22, 23 and 24 shows the running totals for the available water capacity for forest, grassland, and cropland sites respectively. The total available water capacity for the 100 cm soil profile depth at each site were, 435.1 mm and 189.4 mm, 307.7 mm and 147.5 mm, 273.9 mm and 128.2 mm of water at saturation and at field capacity for the forest, grassland, and cropland sites respectively. From tables 22, 23 and 24, the difference in soil moisture reserves available to crops can be appreciated. A crop rooting depth of 30 cm in forest, grassland, and cropland sites will have a reserve of 60.8 mm, 29.7 mm and 35.6 mm of available moisture respectively. However a crop rooting depth of 100 cm will have a reserve of 189.4 mm 147.5 mm and 128.2 mm respectively. It can therefore be seen that the amount of ater actually available to plants depends on the soils available water capacity and the plant rooting depth. For the forest site, it is observed that as trees extend their roots deep into the soil, they have more reserve at 100 cm soil depth than in the cropland site. The forest site had more available water at field capacity followed by grassland site and cropland site at the 100 cm soil depth. This is attributed to the fact that the forest and grassland sites had a dense soil cover. This increases infiltration and thereby reduces run-off; hence much of the water that moves into the soil later becomes available to plants. Proper land use therefore leads to the increase of available soil moisture required for the plant use.

Table 22. Available water storage capacity in mm of water for the forest site.

Soil profile depth cm	Available water capacity in mm of water At saturation	At field capacity
0 - 10	42.5	22.8
10 - 30	114.3	60.8
30 - 60	245.1	120.2
60 -100	435.1	189.4

Table 23. Available water storage capacity in mm of water for the grassland site.

Soil profile depth cm	Available water capacity in mm of water At saturation	At field capacity
0 - 10	33.6	11.5
10 - 30	98.6	29.7
30 - 60	196.1	83.1
60 -100	307.7	147.5

Table 24. Available water storage capacity in mm of water for the cropland site.

Soil profile depth cm	Available water capacity in mm of water At saturation	At field capacity
0 - 10	25.7	11.0
10 - 30	81.1	35.6
30 - 60	169.9	81.8
60 -100	273.9	128.2

4.4. Saturated Hydraulic Conductivity

The variation of the average saturated hydraulic conductivity with depth for the forest, grassland,and cropland sites are shown in table 25. Hydraulic conductivity at saturation is a characteristic property of soil water flow and is related to porosity and pore size distribution. The saturated hydraulic conductivities for the forest, grassland, and cropland sites decreased from 0.134, 0.229, and 0.014 cm/day at the soil surface to 0.0002 0.00054, and 0.0032 cm/day at the 100 cm soil depth respectively. The variation of soil hydraulic conductivity with depth at each site was significantly different at both the 1 and 5% levels.

The saturated hydraulic conductivity as shown in table 25 is very variable within a given soil and between the three sites. The variability in conductivity may be due to many factors. It may be due to soil being heterogeneous in the field or the fact that the samples used to conduct the experiment were in containers of 81.43 cm^3 volume and therefore the samples were not representative of the field conditions. Cracks, worm holes, and decayed root channels are present in the field and may affect flow in different ways, depending on the directions and conditions of the flow process. The presence of a decayed root channel, worm hole or tiny crack in the core obtained from the field, could have accounted for the high values obtained as they accelerated the flow rate of water. Other reasons may be that, the area for conductivity in the field is not fixed by any boundaries, but is dependent on the width, continuity, shape,and tortuosity of the conducting channels. The pore geometry thus affects the conductivity under field conditions. Lack of continuity of conducting channels caused by the boundary effect of the core containers could have accounted for the very low values. Saturating a soil without trapping some air is difficult. The entrapped air bubbles can block pore passages and this could also have contributed to the low values. The observed differences between the three sites may be attributed to the differences resulting from soil structure and texture. The hydraulic conductivity is usually greater if the soil is highly porous, fractured, or aggregated. Hence the high values observed in grassland and forest sites at the soil surface may be attributed to the high values of organic matter content reported in section 4.1.2. The low value of hydraulic conductivity at the soil surface may also be due to low organic matter as reported and the high value of bulk density at the soil surfaces as indicated in section 4.1.1., as values are usually low if the soil is tightly compacted and dense. The conductivity also depends not only on the total porosity, but also, and primarily, on the sizes of the conducting pores. For example, a gravelly or sandy soil with large pores can have a conductivity much greater than a clay soil with narrow pores though the total porosity of clay is generally greater than that of sandy soil. This is indicated in the cropland site which had its profile mainly occupied by clay.

Table 25. The variation of saturated hydraulic conductivity with depth for the forest, grassland, and cropland sites.

Soil profile depth (cm)	Saturated hydraulic conductivity cm/day		
	Forest site	Grassland site	Cropland site
0	0.134	0.229	0.014
10	0.0066	0.243	0.0051
30	0.0004	0.00063	0.00313
100	0.0002	0.00054	0.0032

5.0 CHAPTER V

5.1. CONCLUSIONS AND RECOMMENDATIONS

5.1.1. Soil physical characteristics

If productivity is to be maintained then an agricultural system that is able to preserve and maintain satisfactory physical conditions in the soil must be developed. Fertilizer alone or even with improved crop varieties and measures to control pests and diseases, will not maintain productivity if significant deterioration of physical conditions occurs. The bulk density data determined in this study showed that the cropland site was more compacted followed by grassland and forest sites. These data agrees with that of Page and Willard (1946), who found out that cultivation resulted in a loss of pore space and a corresponding increase in weight per unit volume of soil.

Due to intensive management that is taking place in forest sites, it is becoming apparent that the distinction between forest soils and agricultural soils is becoming progressively less evident. Although some properties acquired by soil during its development such as particle size distribution and mineralogical composition persist long after the forest cover has been removed and the soil subjected to cultivation, other characteristics such as organic matter content, tilth and aggregation properties,and the distribution of soluble and exchangeable ions, can undergo drastic change either naturally or because of management. The forest site showed more of organic matter at the soil surface followed by grassland and cropland sites. Organic matter contributes much to the productive capacity of soil. Nutrients are mineralized during decomposition of organic matter. Depending on the quality of organic matter they usually have high cation exchange capacity, high water holding capacity, the capacity to chelate cations, and the ability to improve the physical characteristics of soils.

The textural classification for the three sites indicated that generally there was a greater quantity of clay size particles associated with, cropland site compared to the grassland and forest sites. These differences are due to the existing inherent characteristics of the soils rather than the land uses. Changes of land use for a period of only 10 years cannot lead to such differences in clay content within the profiles up to a depth of 100 cm. Cultivation that has been going on for that period in the cropland sight might have increased the weight of soils as observed in the high bulk density reported. The increase could have also been caused by higher mean soil temperature during the growing season, thus increasing the rate of carbon weathering due to abrasion or re-orientation combined with displacement of particles resulting from tillage operations.

5.1.2. Soil Structural Characteristics

Water stable aggregates in the cropland site have been found to be higher than in the forest and grassland sites. The erodibility of the cropland site was therefore higher than the other sites. The lower values on organic matter reported on this site showed some influence on aggregation.

The degree of soil aggregation which is an indication of a soil's susceptibility to wind erosion as determined by dry sieving led to the conclusion that the forest and grassland sites were more susceptible to wind erosion than the cropland site. The soil at the forest site showed a weak structure and therefore the number of soil particles small enough to be moved by wind is very high. The state and stability of the structural units are principally determined by water, soil texture, organic cements, and desegregating processes.

The soil erodibility indices calculated on forest, grassland, and cropland sites brought the following conclusions:

(i) That the erodibility of the cropland site was higher than that of forest and grassland sites.

(ii) The dispersion ratio, which is an index of the ease with which soil particles can be brought into suspension by the action of raindrops or runoff water, showed high values on the grassland site than cropland discussed that the soil moisture characteristic curve was strongly affected by soil texture. In general the greater the clay content, the greater the water content at any particular suction, and the more gradual the slope of the curve. In sandy soil, most of the pores are relatively large, and once these large pores are emptied at a given suction only a small amount of water remains. In a clay soil, the pore size distribution is more uniform, and more of the water is absorbed, so that increasing the matric suction causes a more gradual decrease in water content.

The running totals for the maximum profile water storage capacity, led to the following conclusions:

(i) That the forest site had more water at saturation

(ii) That at field capacity the grassland site had more water and,

(iii) That the cropland site had more water at wilting point.

The available water capacity of the soil, regarded as an index of the ability of a soil to store water, showed that at field capacity, the forest site was more capable of storing water than the grassland and cropland sites.

Saturated hydraulic conductivity was highly variable within depth and between the three sites. Cracks, worm holes, decayed roots, entrapped air, and lack of true simulation of the actual field conditions are among the reasons causing this high variability within depths and between the sites. Structure and texture also contributed to the high variability obtained between the three sites.

5.2 RECOMMENDATIONS

Soil deterioration, attributed to cultivation has always been a problem in agriculture because maintenance of soil structure and fertility is essential in any cropping system. Where it may be possible to replace plant nutrients removed from the soil by cropping, the reclamation of eroded and physically degraded soils a rather difficult task and often uneconomical. An important factor in sustainable productivity of tropical soils, therefore is the maintenance of soil physical characteristics at optimum level.

Most of the projects undertaking soil conservation activities are situated in semi-arid areas. These are the areas with lower rainfall and most of it is unreliable. The few forest and grassland sites found in these areas are now turned into agricultural use due to population increase and land pressure. Much emphasis on soil management should therefore focus on maintaining good soil physical properties for their sustained use. The habit of changing the forest and grassland sites towards cultivation areas should be looked upon more carefully.

Cultivation systems or soil tillage which can keep soil fertility at a sufficient level will be very important for sustainable agriculture production. Any type of land use that adds organic matter, living or dead, results in a soil that is more open, porous and easily penetrated by soil. Planting trees, leaving crop residues, adding manures, rotation grazing are some of the cultivation systems that will increase infiltrations and reduce runoff. Organic matter loss is a primary cause of decreasing crop yield in the tropic. A decrease in soil organic matter can result in soil structure deterioration, lower plant nutrient reserves.

Much yet is to be learned about how to manage the water in the soil at the same time that one manages the plant root distribution so that the water is where it is needed as early as possible. Similar studies should be extended on the major soil types to determine the effect of selected land uses on soil physical, structural and hydrological characteristics.

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APPENDIX 1: SOIL PHYSICAL PROPERTIES FOR THE FOREST SITE

Soil Profile Depth cm	Bulk Density gcm^{-3}	Particle size distribution (%)			Organic matter content (%)	Textural classification
		Sand 2-0.1mm	Silt 0.05-0.002mm	Clay (0.002mm)		
0	1.04	47.8	36.9	15.4	3.6	Loam
10	1.03	44.5	28.0	27.5	4.2	Loam
30	1.04	41.0	29.4	29.6	2.6	Clayloam
60	0.98	39.1	20.0	41.0	2.1	Clay
100	1.02	33.1	14.9	52.0	1.7	Clay

APPENDIX 2: SOIL PHYSICAL PROPERTIES FOR THE GRASSLAND SITE

Soil Profile Depth cm	Bulk Density gcm^{-3}	Particle size distribution (%)			Organic matter content (%)	Textural classification
		Sand 2-0.1mm	Silt 0.05-0.002mm	Clay (0.002mm)		
0	1.05	33.1	30.0	36.6	5.9	Clayloam
10	1.05	41.0	27.4	31.6	3.5	clayloam
30	1.18	34.0	38.1	28.1	3.1	Clayloam
60	1.21	39.0	35.4	25.6	2.6	Loam
100	1.08	39.8	39.2	21.0	2.5	Loam

APPENDIX 3: SOIL PHYSICAL PROPERTIES FOR THE CROPLAND SITE

Soil Profile Depth cm	Bulk Density gcm^{-3}	Particle size distribution (%)			Organic matter content (%)	Textural classification
		Sand 2-0.1mm	Silt 0.05-0.002mm	Clay (0.002mm)		
0	1.09	16.3	35.0	48.2	4.6	Clay
10	1.10	16.3	30.5	53.2	3.5	Clay
30	1.14	18.3	21.0	60.7	1.9	Clay
60	1.18	16.8	20.0	63.2	2.1	Clay
100	1.28	15.3	19.5	65.2	1.7	Clay

APPENDIX 4: ANALYSIS OF VARIANCE FOR BULK DENSITY AT THE STUDY SITES

ANOVA

VO_1 Response

VO_2 Site

VO_3 Depth

VO_4 Replicates

SV	SS	DF	MS	F	Sign of F
Main effects	.956	8	.119	1.562	.213
VO_2	.448	2	.224	3.133	.068
VO_3	.209	4	.052	.685	.613
VO_4	.259	2	.129	1.687	.216
2-way interactions	1.871	20	.094	1.224	.345
VO_2 VO_3	.651	8	.081	1.065	.433
VO_2 VO_4	.445	4	.111	1.454	.262
VO_3 VO_4	.775	8	.097	1.267	.326

APPENDIX 5: ANALYSIS OF VARIANCE FOR ORGANIC MATTER CONTENTS AT THE STUDY SITES

ANOVA

VO_1 Response

VO_2 Site

VO_3 Depth

VO_4 Replicates

SV	SS	DF	MS	F	Sign of F
Main effects	146.829	8	18.354	59.703	0.000
VO_2	12.402	2	6.201	20.171	0.000
VO_3	132.985	4	33.246	108.147	0.000
VO_4	1.442	2	.721	2.345	.128
2-way					
interactions	36.162	20	1.808	5.882	.000
VO_2 VO_3	32.162	8	4.118	13.395	.000
VO_2 VO_4	.450	4	.113	.311	.829
VO_3 VO_4	2.769	8	.346	1.26	.389

APPENDIX 6: PORE SIZE DISTRIBUTION FOR THE FOREST GRASSLAND, AND CROPLAND SITES

SITE	PROFILE DEPTH cm	%TOTAL PORE SPACE	%MAC-RESPORE SPACE	%CAPI LLIARY SPACE	%USEFUL PORE SPACE	% NON USEFUL CAPIL-PORE SPACE	%VOL-UME OF THE SOIL
FOREST	0	52.5	19.7	42.8	22.8	20.0	37.5
	10	61.3	16.9	44.4	19.6	25.0	38.7
	30	64.[illegible]	28.8	45.2	17.6	28.6	35.7
	60	68.4	30.2	38.2	14.7	23.5	31.6
	100	66.5	22.1	44.4	20.1	24.3	33.5
GRASS-LAND	0	61.3	22.1	39.2	11.5	27.7	38.7
	10	64.2	23.4	40.8	9.1	31.7	35.8
	30	61.9	14.7	47.2	17.8	29.4	38.1
	60	59.3	11.8	47.5	16.1	31.4	40.7
	100	65.5	24.6	40.9	12.6	28.3	34.5
CROP-LAND	0	56.6	14.7	41.3	11.0	30.9	43.4
	10	58.7	15.4	43.3	12.3	31.0	41.3
	30	58.3	14.2	44.1	15.4	28.7	41.7
	60	59.2	13.9	45.3	11.6	33.7	40.8
	100	57.5	9.3	48.2	8.2	40.0	42.5

APPENDIX 7: ANALYSIS OF VARIANCE FOR AGGREGATE STABILITY (WET SIEVING)

CELL MEANS

VO_1 Response

by VO_2 Site

TOTAL POPULATION

17.34
(15)

VO_2	1	2	3
	17.24	13.57	21.19
	(5)	(5)	(5)

A N O V A

VO_1 RESPONSE

by VO_2 SITE

Source of Variation	Sum of squares	DF	Mean square	F	Sign of F
Main effects	145.300	2	72.65	15.037	0.001
VO_2	145.300	2	72.65	15.037	0.001
Explained	145.300	2	72.76	15.037	0.001
Residual	57.979	12	4.832		
Total	203.279	14	14.520		

APPENDIX 8: ANALYSIS OF VARIANCE FOR AGGREGATE STABILITY (DRY SIEVING)

CELL MEANS

VO_1 Response
by VO_2 Site

TOTAL POPULATION

14.85

(15)

VO_2	1	2	3
	19.50	19.03	6.03
	(5)	(5)	(5)

A N O V A

VO_1 RESPONSE

VO_2 SITE

Source of Variation	Sum of squares	DF	Mean square	F	Sign of F
Main effects	584.613	2	292.306	6.629	0.012
VO_2	584.613	2	292.306	6.629	0.012
Explained	584.613	2	292.306	6.629	0.012
Residual	529.170	12	44.097		
Total	1113.782	14	79.556		

APPENDIX 9: ANALYSIS OF VARIANCE FOR SATURATED HYDRAULIC CONDUCTIVITY

ANOVA

VO_1 Response

VO_2 Site

VO_3 Depth

VO_4 Replicate

SV	SS	DF	MS	F	Sign of F
Main effects	.344	8	.043	31.122	.000
VO_2	.193	2	.097	69.861	.000
VO_3	.148	4	.037	26.812	.000
VO_4	.003	2	.001	1.003	.389
2-way interactions	.117	20	.009	6.412	.000
VO_2 VO_3	.155	8	.019	14.024	.000
VO_2 VO_4	.008	4	.002	1.472	.257
VO_3 VO_4	.014	8	.002	1.270	.324