VARIABILITY OF INFILTRATION

IN THE NORTHERN GUINEA SAVANNA ALFTSOL OF NIGERIA

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DECLARATION

I, hereby declare, that the content of this thesis is a record of my research work. The subject matter herein has not been presented in any previous publications for a higher degree.

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CERTIFICATION

This thesis entitled VARIABILITY OF INFILTRATION IN THE NORTHERN GUINEA SAVANNA ALFISOL OF NIGERIA, by WUDDIVIRA HASSAN NAKKA, meets the regulations governing the award of the degree of Master of Science of the Ahmadu Bello University, Zaria and is approved for its contribution to scientific knowledge and literary presentation.

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DEDICATION

To my father and mother, Mr. and Mrs. Wuddivira Zamanduniya whose love, patience, prayers and exemplary life is a challenge to me.

ABSTRACT

The ring infiltrometer method was used to measure infiltration rate of a savanna soil on a 60m transect, at 2m intervals. This was done with the aim of evaluating the magnitude of spatial variation of infiltration over a field considered uniform interms of soil type. Philip and Kostiakov's equations were employed for predicting infiltration rate of the soils in the field. Curve fitting by regression techniques showed that infiltration of the soils are adequately described by both equations ($R^2 > 0.98$). There was relatively high variation in infiltration rate with a mean of 27.4mm/hr and a standard deviation of 15.25mm/hr. These values were also log-normally distributed.

Infiltration rate was poorly correlated with bulk density (r = 0.32) and porosity (r = 0.32) but it was significantly correlated with macroporosity (r = 0.38*) at 5% level of significance. Further analysis using autocorrelograms, showed that the steady-state infiltration rates for the 30 points/locations were spatially independent at the sampling intervals of 2m.

The mean and standard deviation values of saturated hydraulic conductivity were 44.78mm/hr and 37.27mm/hr respectively. These values gave rise to wide variation in conductivity measured between points. The variation was spatially independent with a log-normal frequency distribution. Thus, it can be unreliable to use single determined infiltration or hydraulic conductivity value as representative of the field.

Due to the spatial independence of the data for these

parameters, 10 and 58 samples are required to estimate the mean steady-state infiltration rate and hydraulic conductivity for the study area, respectively. These sample numbers are obtained within ± 10% of the population mean at the 95% confidence level assuming normal distribution.

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CHAPTER ONE

1.0 INTRODUCTION

Lack of adequate water supply is a principal constraint to crop production in the Nigerian Savanna (Jones and Wild, 1975; Virmani et al., 1980; Lal, 1991; Maduakor, 1991; Oluwasemire, 1997). Therefore, there is the need to carefully manage scarce water resources for sustainable agricultural productivity in this region (Lal, 1994).

The rainfall in this region is aggressive and erratic, thereby making drought a natural phenomenon (Lal, 1991; Hoogmoed et al., 1991; Owonubi et al., 1991). The impact of the raindrops from the intensive rainfall on the poorly structured soils (Wilkinson, 1975; Adeoye, 1986), can be minimized through appropriate management of soil, crop and water (Lal, 1991). If the scarce water is to be optimally utilized, a sound knowledge of the physical and hydraulic properties of the soil becomes essential (Sivakumar and Wallace, 1991). One principal determinant property is its infiltration capacity.

Infiltration is one of the most important processes controlling the fate of water over large areas of land in the field. It is regarded as the key input in the development of a sound soil water management and conservation practices. Apart from playing a vital role in the distribution of water, it determines the amount of runoff that will occur over the soil, and hence soil erosion hazards (Babalola, 1986). Proper knowledge of the rate at which different soils will take in water under different conditions, is therefore important for planning water management activities.

Infiltration and hydraulic properties have been shown to vary

widely over space and time (Nielsen et al., 1973; Babalola, 1978; Folorunso and Olorunju, 1986; Umorou and Babalola, 1986). This has been attributed to surface and profile heterogeneities in physical properties of seemingly similar soils. The spatial differences in infiltration, call for site specific measurements even in a quite uniform field. Therefore, for a realistic planning and use of water resources, the spatial variability of infiltration must be taken into account.

There have been some earlier works done on infiltration and related properties of Samaru soils (Wilkinson, 1975; Aremu, 1979; Kowal, 1970a; Ahmed, 1982; Babaji, 1982; Adeoye, 1983; Kureve, 1991). Kureve (1991) was the only one that skeletally considered spatial variability of infiltration, in his study on the water intake characteristics of a Savanna soil.

Soil variability has been recognized to occur over distances of only meters in West Africa (Folorunso and Olorunju, 1986; Umouru and Babalola, 1986; Chase et al, 1987; Kureve, 1991). In order to obtain accurate picture of local soil hydrology, many measurements are required. This justifies the need for this study.

The study which has the goal of providing information to enable a farmer decide, on whether or not, his field can be given uniform agronomic management such as irrigation, has the following objective:-

To measure infiltration and evaluate the magnitude of its spatial variation over a field considered to be uniform for agricultural purposes.

CHAPTER TWO

2.0 LITERATURE REVIEW

2.1 WATER MOVEMENT IN SOILS AND RELATED EQUATIONS

Movement of water in soils can be divided into two categories, namely (a) non saturated system, where the three phases of matter (soil - air - water) exist, and (b) saturated system where only the solid and the liquid (soil - water) states of matter exist.

Most of the water movement in soil occurs when both water and air are present in the pores i.e. unsaturated flow. This is true even where water moves into the soil from ponded water at the surface. This flow is so - called (unsaturated flow) because water entering at the soil surface does not saturate the soil below and so moves through unsaturated soil. Saturated flow could also be important. Water movement during drainage and in the ground water system are obvious cases of saturated flow.

The saturated flow is easier to deal with because the hydraulic conductivity (K) is a constant in this system. But the value of K is not a constant in unsaturated flow because it decreases as the water content decreases or as the soil water potential (negative) decreases (Yong and Warkentin, 1966). The continuity equation and Darcy's Law describe water movement in the soil for both saturated and unsaturated conditions.

Infiltration is an example of unsaturated flow in the field and is described by three variables, namely (a) infiltration rate, i, (same as Darcian flux through the soil surface), time, t and cumulative infiltration, I. The three are related via the equation

$$i=\frac{dI}{dt} \quad \ldots\ldots\ldots\ldots\ldots\ldots (1)$$

2.1.1 Darcy's law

The Darcy's law expresses movement of water on a macroscopic dimensional perspective. This is so because of the complication of microscopic description due to variation, irregularity, tortuosity, discontinuity and complexity of pores. For one dimensional flow, Darcy's equation is

$$J=-K\frac{dH}{dx} \quad \ldots\ldots\ldots\ldots\ldots\ldots (2)$$

where J is the flux, $cm^3.cm^2.s^{-1}$; dH is the hydraulic head difference causing flow; x is the directional coordinate along which flow takes place; K is a proportionality constant known as hydraulic conductivity, cm s^{-1}.

From (2) above, we can substitute J = Q/At, so that Darcy's equation in terms of discharge rate Q becomes,

$$Q=-KA\frac{dH}{dx} \quad \ldots\ldots\ldots\ldots\ldots\ldots (3)$$

The unit area A = πR^2 (cm^2)

2.1.2 Richard's equation

The magnitude or direction of unsteady flow processes vary with time and can therefore not be sufficiently described by Darcy's law, which takes care of steady flow processes. The law of conservation of matter expressed in the continuity equation when employed, solve such flow problem. The continuity equation in one dimensional form is:

$$\frac{\delta\theta}{\delta t} = -\frac{\delta J}{\delta x} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (4)$$

where Θ is volumetric wetness. Inserting Darcy's equation (2) into the continuity equation (4), gives the general flow equation.

$$\frac{\delta\theta}{\delta t} = (K\frac{\delta H}{\delta x}) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (5)$$

Equation (5) is known as Richard's equation.

But,

$$H = H_p + H_g \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (6)$$

where H_p is negative in an unsaturated soil and can be expressed as a suction head ψ, and H_g = - x (datum plane is at the soil surface, Z = 0).

Hence, $H = \psi$ - x and putting into (5), gives

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta x}(K\frac{\delta(-\psi - x)}{\delta x}) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (7)$$

so that,

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta x}(-K\frac{\delta\psi}{\delta x} - K) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (8)$$

if we assume a unique relation between Θ and ψ, we can expand $\delta\psi/\delta x$ by the chain rule as follows

$$\frac{\delta\psi}{\delta x} = \frac{\delta\psi}{\delta\theta} \cdot \frac{\delta\theta}{\delta x} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (9)$$

where $d\theta/dx$ is wetness gradient.

But differential water capacity C (Θ), is given by

$$C(\theta) = \frac{\delta\theta}{\delta H_p} = -(\frac{\delta\theta}{\delta\psi}) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (10)$$

therefore equation (9) becomes,

$$\frac{\delta\psi}{\delta x}=\frac{1}{C}(\theta)\left(\frac{\delta\theta}{\delta x}\right) \quad \ldots\ldots(11)$$

putting equation (11) into (8), we obtain

$$\delta\theta=\frac{\delta}{\delta x}\left(\frac{K}{C(\theta)}\cdot\frac{\delta\theta^{-K}}{\delta x}\right) \quad \ldots\ldots(12)$$

Setting D (θ) = K (θ)/C(θ) and substituting into (12), we get,

$$\frac{\delta\theta}{\delta t}=\frac{\delta}{\delta x}\left(D(\theta)\frac{\delta\theta^{-K}}{\delta x}\right) \quad \ldots\ldots(13)$$

where D is soil water diffusivity.

2.2 INFILTRATION

Infiltration is the downward entry of water into the soil (Sundara and Nagarajarao, 1976). It has also been defined as the process of water entry into the soil through the surface followed by vertical downward movement (Philip, 1957; Hillel, 1971; Babalola, 1986). It is essentially a surface phenomenon (Hillel, 1971).

Infiltration rate is defined as the flux of water passing through the surface into the soil profile. Infiltration capacity on the other hand, is defined as the maximum rate at which a given soil, when in a given condition, can absorb rain as it falls (Horton, 1940; Hillel, 1980).

Falling drops of water from rain or irrigation may be intercepted by vegetation or may fall directly on the soil surface. The water that reaches the ground, may evaporate back to the atmosphere, filters downward into the soil or runoff. Runoff starts only when surface detention has been satisfied but infiltration commences immediately the first drop of water touches

the ground surface and continues until all depressions containing water after the cessation of rainfall are emptied.

The water that infiltrates into the soil may percolate into the ground water system, or it is redistributed within the soil and is held for plant use. Water entry into the soil is therefore much affected by surface conditions, inherent soil properties and mode of water supply (Hillel, 1980).

Infiltration of water into a soil is an unsaturated flow phenomenon. The two driving forces that cause water to move into the soil are gravity and matric suction or matric potential. The pull of gravity is constant but the matric potential weakens as the moisture content of the soil increases.

Equation (13) shows the contribution of the wetness gradient and gravity to infiltration rates. The term that predominates (whether suction gradient or gravity) depends on the initial and boundary conditions and on the stage of the process considered. For instance, the suction gradients can be much greater than the gravitational gradient when infiltration takes place into an initially dry soil. Infiltration rate decreases rapidly as infiltration proceeds and then more slowly to a somewhat constant rate known as the final or equilibrium infiltration capacity or steady state infiltrability (Hillel, 1980; Babalola, 1986; Ghildyal and Tripathi, 1987).

2.3 SOIL PROPERTIES AFFECTING INFILTRATION

Infiltration of rain or irrigation water is largely contingent on soil texture, soil structure and its stability, profile

stratification and the antecedent moisture content (Babalola, 1986). The first three factors, primarily influence porosity and pore-size distribution or pore geometry and have overriding influence on infiltration (Jones and Wild, 1975; Wilkinson, 1975; Ojanuga, 1987; Owonubi et al., 1991).

2.3.1 Porosity

Pore space typically accounts for 40 to 60% of the total volume of soil (Olson, 1985). The number, size, and continuity of pores, control both the amount of the storage capacity of available water and the rate of flow of water through the soil to the plant-root-soil interface (Bouma, 1983; Olson, 1985; Grismer, 1986).

For water to enter the soil profile, there must be pathways by which it can be conveyed. These pathways called pores could be channels created by structural arrangement of soil particles or burrowing activities of soil fauna (e.g. earthworms and ants) or cracks, fissures and root channels. The pathways contribute to total porosity and affect infiltration and hydraulic conductivity of a soil (Kureve, 1991).

Soil pores vary in size, but for convenience, they can be grouped into 3 categories: (1) larger pores that are air-filled at field capacity; (2) medium pores that are small enough to hold water against gravity, yet permit water to be withdrawn by roots; and (3) small pores from which plants are unable to remove water (Arkin and Taylor, 1981). Pores that are smaller than 60 to 30 um have been found to be water-filled at matric potential of - 5 to - 10 KPa i.e at field capacity (Russell, 1973); The large pores will be air-filled. Pores small enough to hold water against gravity,

yet able to make it available to the roots, had a diameter of 0.5 to 0.2 um and a potential of - 1500 KPa (Arkin and Taylor, 1981).

The large pores (Macropores) have been reported to increase infiltration capacity, flux and conductivity by several orders of magnitude (Dixon and Peterson, 1971; Ehlers, 1975; Wilkinson, 1975; Wilkinson and Aina, 1976; German and Beven, 1981; Ahmed, 1982; Watson and Luxmoore, 1986). Macropores also act as escape route of entrapped air during infiltration under ponded condition (Dixon and Peterson,1971; German and Beven, 1981; Watson and Luxmoore, 1986).

Pore-size distribution has been recognized to have profound influence on infiltration (Kowal and knabe, 1972; Wilkinson, 1975; Bauder et al., 1981; Morin et al., 1981; Olson, 1985; Parlange et al., 1985; Adeoye, 1986; Babalola, 1986; Grismer, 1986). A change in pore-size distribution that skewed toward larger pores results in an increase in infiltration as much as five times for typically fine-textured soils. Similarly, a corresponding decrease is noticed in infiltration rate when pore-size distribution skewed toward small pores (Grismer, 1986). The factors that have been held responsible for changing the pore-size toward successively smaller pores are soil compaction and surface crusting (McIntyre, 1958; Bateman, 1963; Ohu and Folorunso, 1989). Their resultant effect is a net decrease in infiltration rate and capacity (Morin et al., 1981).

2.3.2. Soil texture

Permeability was found to increase exponentially with an

increase in particle size. The progressive decrease in infiltration capacity as the soil gets finer confirms this assertion (Kohnke, 1968).

Kohnke (1968) found infiltration rate on loamy sand to be 10 times higher than on clay loam, while it was 3 times higher in silt loam than in clay loam. This trend has been attributed to (i) the abundance of large and continuous aeration pores in coarse-textured soils, and (ii) the tendency of colloidal particles in fine-textured soils to swell upon hydration there by reducing the cross-sectional area available for flow (Babaji, 1982; Babalola, 1986).

Except in well developed structures, the particles in fine-textured soils tend to adhere and pack closely there by reducing pore-sizes and restricting water movement and infiltration. But infiltration is enhanced in coarse-textured soils because the soil tend to be porous with more conducting pores (Babaji, 1982).

2.3.3 Soil structure

All things being equal, high infiltration is associated with well aggregated soil's while structurally weak soils are characterized by low infiltration.

Low infiltration leading to runoff, erosion and insufficient water use, is a critical problem in arid and semi-arid regions (Adeoye, 1986; Babalola, 1986; Maduakor, 1991). A reduction in the soil infiltration rate is caused mainly by slaking and subsequent crusting of the soil surface exposed to the beating action of rain and sprinkler drops (Kowal and Kassam, 1978; Adeoye,

1986; Babalola, 1986;). In the Nigerian Savanna, this process is strongly influenced by the fact that rainfall is intense (Kowal and Knabe, 1972) and soils are generally poorly structured and unstable (Jones and Wild, 1975; Wilkinson, 1975; Ojanuga, 1987). Hillel (1968) reported that a soil that is highly porous and of "Open" structure, has initial infiltrability that exceeds that of soil whose surface is crusted.

Removal of layers suspected to be crusted, resulted in significant increase in infiltration at Samaru (Babaji, 1982; Adeoye, 1986). Higher aggregation which was confined to surfaces protected from raindrop impact with mulch, leads to higher infiltration as opposed to the bare surfaces (Adeoye, 1986). Wilkinson (1975) submitted that since Samaru soils have little structure in the surface layer, variation in infiltration of water into these soils is principally a function of worm and termite activity.

2.3.4 Soil profile characteristics

Impeded water movement during infiltration may be caused by zones of low permeability like clay pan and hard pan in the profile. The presence of layers which differ in texture or structure from overlying soil (profile stratification) may also present a bottleneck to water movement during infiltration.

A clay layer within the profile may impede flow because of its low hydraulic conductivity. Sand layer retards the wetting front where unsaturated conductivity prevail, owing to the lower

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unsaturated conductivity of the sand. Water will flow into a dry sand layer only after the pressure head has build-up sufficiently for water to move into sand and fill the large pores of sand.

2.3.5 Antecedent moisture content and infiltration

Infiltration rate tends to be higher in a soil which was initially dry. Water moves slowly into the soil when moisture conditions are high (Ghosh and Maity, 1976; Babaji, 1982; Babalola, 1986). Ward (1967) attributed this to reduction of the soil capacity to receive any additional moisture, reduction of the hydraulic gradient across the wetting front and swelling of soil particles.

Slaking and swelling was found to be influenced by initial water potential (Lal and Greenland, 1978). The influence of initial moisture content became less and ultimately negligible as infiltration progresses. In other words, the initial moisture content does not influence the equilibrium rate except that soils which are already wet when rainfalls upon them, attain final infiltration rate more quickly (Philip, 1957).

2.4 INFILTRATION EQUATIONS

Infiltration of water into soils depends on so many factors. These include physical properties of the soil, type of vegetal cover, condition of surface crust, intensity of rainfall, size of raindrops, and water quality (Esen, 1987; Owonubi et al., 1991). Several equations (models) have been proposed and used to describe

infiltration with time (Hjelmfelt and Cassidy, 1975; Viessman et al., 1977; Shaw, 1983; Garg, 1984).

Basically, an infiltration equation expresses the total quantity of water infiltrated into the soil (I) or the infiltration rate (i) as a function of time (t) (Ghosh, 1985). The use of these equations for different soils and conditions has been limited by lack of adequate knowledge of the equation parameters (Sundara and Nagarajarao, 1976). Infiltration equations have been extensively used to characterize measured infiltration data. These characterizations are used to form a picture of existing conditions and to determine the criteria necessary for good water management. Meaningful planning and execution of drainage and irrigation projects can therefore be achieved by the use of appropriate equations.

Also, in modelling the water balance of a field, the infiltration component is a very crucial factor determining both runoff and infiltration capacity. Infiltration equations therefore serve as a viable means of quantifying infiltration to fit into the model. All the infiltration equations reflects the tendency of infiltration rate to decrease rapidly with time as infiltration progresses.

2.4.1 Green-ampt equation

The Green-Ampt equation proposed by Green and Ampt (1911) was the first attempt to model infiltration. The equation can be written as follows:

$$i = K_s \left(1 + \frac{H_o - H_f}{L_f}\right) \text{ ---------- (14)}$$

in which i = dI/dt, the infiltration rate in $mm.hr^{-1}$; t = time in hr; K_s = saturated hydraulic conductivity in $mm.hr^{-1}$; H_o = pressure head at the entry surface (depth of water in the infiltrometer in mm); H_f = pressure head at the wetting front; L_f = length of the wetted Zone; I = cumulative infiltration in mm.

From equation (14) above, it is clear that as the L_f increases i decreases and at infinite time, L_f becomes very large rendering the term $H_o - H_f/L_f$ insignificant. Therefore, i which is the same as flux approximates K_s ($i = K_s$)

At large time, $I = Kt + \delta$ ---------- (15)

δ which is small at $t = \infty$, is regarded as a constant.

The Green-Ampt equation has a theoretical basis (Esen, 1987). The parameters defining the equation can be determined independently by field and laboratory tests (Bouwer, 1966). Several semi-empirical equations together with available soils and agronomic information can also be used (Rawls and Brakensiek, 1983; Rawls et al., 1983). However, independent determination of the parameters may inaccurately predict the cumulative infiltration as a function of time (Brakensiek and Onstad, 1977; Esen, 1989).

2.4.2 Kostiakov's equation

The Kostiakov's infiltration equation suggested an exponential type of equation relating cumulative infiltration (I) to time (t). He used two matching coefficients M and n for each individual

infiltration event. The form of the equation is:

$$I = Mt^{n} \text{ ---------- (16)}$$

Where M and n are constants. The value of M depends on the range of time considered. M is indicative of the average infiltration rate at initial short time period of infiltration, and n is related to structural stability of the soil (Swartzendruber and Huberty, 1958). The two parameters are used as measures of soil properties influencing infiltration rate.

Kostikov's equation has been shown by Dixon et al. (1978), to satisfy the evaluation criteria used in predicting and controlling rain water infiltration in crop and rangelands, better than other equations. This has been attributed to its simplicity (Sundra and Nagarajarao, 1976) and its application under wide range of conditions in the real world soils (Dixon, 1976; Ghosh, 1985). The use of Kostiakov's equation is practical and gives better result than the theoretically founded equation of Philip (1957) and Green and Ampt (1911) (Canarache, 1974). To corroborate this fact, this equation has been found by many reseachers to reasonably describe field infiltration data (Sundara and Nagajarao, 1976; Aremu, 1979; Ahmed, 1982; Babaji, 1982; Ghosh, 1983; Ghosh, 1985; Adeoye, 1986; Babalola, 1986; Kureve, 1991).

2.4.3 Horton equation

The Horton equation (Horton, 1940), is one of the most popular empirical models simulating infiltration of water into soil. The Horton equation is commonly expressed as:

$$i = i_c + (i_o - i_c)\, e^{-kt} \quad \text{---------- (17)}$$

where i is the infiltration rate at time t in $mm.hr^{-1}$; i_o is the initial infiltration rate in $mm.hr^{-1}$; i_c, the final infiltration rate in $mm.hr^{-1}$; and K, a rate constant in $mmhr^{-1}$.

The performance of the Horton model was observed to be excellent in estimating infiltration volume. And it shows definite superiority over Kostiakov's model. But it was still limited by the unavailability of a direct method for determining i_o, i_c and K, since the conventional way of measuring infiltration rate is only approximate (Esen, 1987).

2.4.4 Philip's equation

Philip's (1957) solved equation (15) for a vertical flow system by analytical methods subject to initial boundary conditions. Mathematically, these conditions are stated as

$$t = O,\ Z > O,\ \theta = \theta_o$$

$$t = O,\ Z = O,\ \theta = \theta_i \quad (18)$$

where the soil is assumed to be infinitely deep with constant initial wetness θ_o throughout the system at time Zero, which subsequently has plane Z = O maintained at moisture content θ_i. His solution took the form of a power series of the square root of time. The equation is thus:

$$I = S_1 t^{1/2} + S_2 t + S_3 t^{3/2} \quad \text{---------- (19)}$$

In which S is a soil parameter called "sorptivity" (ie a measure of capillary uptake of water); and A is a parameter depending upon the ability of the soil to transmit water.

For very large times, the mathematical manipulation of equation (19) becomes too cumbersome (Ghosh, 1985), but at small times, the series can be reduced to the following 2 - term equation

$$I = St^{1/2} + At \quad \text{---------- (20)}$$

From equation (20), the infiltration rate becomes

$$i = \tfrac{1}{2} St^{-1/2} + A \quad \text{---------- (21)}$$

At large times, infiltration rate tends to A. Sorptivity S which is defined in relation to a fixed initial state θ_o and an imposed boundary condition (Hillel, 1980), is proportional to the increase in soil water content during infiltration. Hence, sorptivity gets less important with increase in time of infiltration because matric potential gets less important as the soil approaches saturation. Philip's model has physical basis but its theoretical nature has made it to give inferior results when compared to practical Kostiakov's equation (Canarache, 1974; Esen, 1987). The two are the most widely used.

2.5 SPATIAL VARIABILITY OF INFILTRATION

Infiltration is an important process controlling the fate of water over large areas of land. Infiltration is influenced by so many factors, which are themselves not constant due to inherent

heterogeneities of field soils. Field soils are therefore likely to vary greatly in properties both in time and space (Jury et al., 1991). Meaningful prediction of field phenomena is achieved by so many in-situ measurements, even in a quite uniform field (Warrick and Nielsen, 1980; Russo and Bressler, 1981; Vieira et al., 1981).

Soil variability particularly in West Africa have been observed to occur over distances of only meters (Folorunso and Olorunju, 1986; Umouru and Babalola, 1986; Chase et al., 1987; Kureve, 1991). This calls for the need to understand the spatial variability of infiltration in order to make planning and use of water resources meaningful.

The spatial variability of field - measured soil properties, has been reported by many workers (Nielsen et al., 1973; Biggar and Nielsen, 1976; Babalola, 1978; Sharma et al., 1980; Vieira et al., 1981; Folorunso and Olorunju, 1986; Umouru and Babalola, 1986; Kureve, 1991). Infiltration and other hydraulic properties have been found to vary greatly spatially (Sharma et al., 1980; Warrick and Nielsen, 1980; Russo and Bressler, 1981).

In a tropical soil of Western Nigeria, very high coefficient of variability was obtained in hydraulic values estimated in the field (Babalola, 1978). Samaru soils gave similar results with the variation in conductivity of the order of 10 - 100 fold (Kureve, 1991). Coefficient of variability (CV) for hydraulic properties was reported to be high (> 100%) (Warrick and Nielsen, 1980). While variability for other physical properties such as particle size fractions and bulk density was intermediate (CV of 10 - 100%)

and low (CV of < 10%), respectively. Kureve (1991) obtained less variation in bulk density and porosity in Samaru soils.

Infiltration values for an alfisol in Ibadan and Samaru were found to be normally distributed (Umouru and Babalola, 1985; Kureve, 1991). These values have been shown world wide to have a log - normal frequency distribution (Sharma et al., 1980; Warrick and Nielsen, 1980; Sisson and Wierenga, 1981). Therefore, to formulate reasonable and realistic management practices in agroecosystems, we have to face squarely the inescapable problem of spatial variability. Also, it is only when spatial variability is taken into account that mathematical models to simulate water movement in field soils can be made realistic (Hornsby et al., 1983).

Classical statistical methods, such as analysis of variance (ANOVA), have been used to measure the variability of soil across the field. They were also used to calculate the number of samples necessary to estimate the mean within some specified confidence interval, from the frequency distribution data (Biggar and Nielsen, 1976; Mulla, 1989). By this means, a complete description of the variability of soil properties is not achieved. This is because the distance between observations is not accounted for by the calculated variance which solely depends on the theoretical frequency distribution (Saddiq et al., 1985).

Some degree of dependency between sample values especially of soil physical properties have been shown to exist. This dependency is a function of distance between sample values. It is therefore

anticipated that closer measurements in the field will give similar magnitudes, while values differing by a greater order of magnitude will be obtained from measurements farther apart (Warrick and Nielsen, 1980; Hajrasuliha et al., 1980; Vieira et al., 1981). For soil variability to be statistically meaningful, therefore, it has to take into account spatial dependency.

Geostatistical techniques have been developed to measure the spatial dependency of many soil properties. Webster and Cuanalo (1975) used a time series method to measure the spatial variability of soil properties. They recorded these properties at equal intervals along a transect and then computed correlograms to express spatial dependence. The sampling interval at which the spatial series is repetitive and the maximum sampling distance for which the observation remain correlated, can be determined using autocorrelograms. The autocorrelogram is a process of "self-comparison" that expresses the linear correlation between a spatial series and the same series at a further interval of space. The autocorrelation function, r(h) of a spatial series at Lag h can be calculated by the following (Davis, 1973):

$$r(h) = \frac{C(h)}{S^2} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (22)$$

$$C(h) = \left(\frac{1}{n-h-1}\right) \sum_{i=1}^{n-h} (x_i - S)(x_{i-h} - S) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (23)$$

where C(h) is autocovariance and S^2 is variance. n is the number of

observations; x_i is the value of the observation at i^{th} position; and $x_{(i+h)}$ is the value of the observation at the $(i+h)^{th}$ position. The lag h is defined as the amount of offset or relative distance between the elements being compared. Autocorrelogram are produced by plotting autocorrelation r(h) against lag h (intervals).

Autocorrelograms were used to study spatial dependency of twelve soil properties measured along different transect. These properties were observed to be correlated with each other over a distance that varied from a few centimeters to several tens of meters (Gajem, et al., 1981). The steady state infiltration rate of a soil was found to be spatially dependent within a 50-m range (Vieira et al., 1981). The latter authors suggested a reduction of number of measurements from 1280 to 128 samples as no information about infiltration rate was lost by the reduction across the field.

It is very glaring that autocorrelograms show the degree of dependency of infiltration rate on the distance between pairs of measurements. Autocorrelograms can therefore help to deduce the possibility of applying uniform agronomic management to a field.

2.6 INFILTRATION MEASUREMENT

Measurement of the rate of water entry into the soil is better done by the means of infiltrometers, which are basically of two types: (i) Ring infiltrometers and (ii) Sprinkler infiltrometers.

2.6.1 Ponded water infiltration: ring infiltrometers

The conventionally used method for measuring surface infiltration rates in the field is the ring infiltrometer. The ring infiltrometer is a metal cylinder inserted a small distance into the soil leaving some segment of the cylinder protruding above the soil. Water is ponded to a fixed level in the portion of the ring above the soil. As the water level falls in the ring, the level is measured at specified time intervals. The reading of the level of water in the inner ring, is facilitated by means of a plastic ruler taped to the inside wall. The depth of water in the ring can be maintained by adding measured volumes of water. The average infiltration rate for each time can be calculated, and an infiltration capacity curve relating infiltration rate to elapse time can be drawn (Schultz, 1976). Infiltration rate at anytime can be read from this curve.

Of the various methods which use ponded water, the double ring infiltration method, using concentrically placed rings to facilitate "One dimensional" vertical infiltration into the soil is preferred. Measurements are made on the rate of fall of water in

the inner ring, while water infiltrating into the soil between the two rings ensures one-dimensional flow (FAO, 1979).

The double ring infiltrometer method of measuring infiltration rate is inexpensive, simple, very easy to learn, and rings can easily be made locally, or even "On-the-spot" by ponding water in between small earthen dikes lined with plastic sheets. Despite these merits, ring infiltrometers have the following draw backs that can make the measured infiltration rate to be considerably higher (Tricker, 1978) or lower (Clemens, 1981) than the actual rates.

i) The act of driving the ring into the ground tends to compact the soil and cuts off natural paths for air and water flow (Macropores), thus reducing infiltration (Dixon, 1975).

ii) The fine particles dislodge and suspended when water is poured into the ring, settle out producing a layer of fine material that tends to reduce infiltration and "seal" the ring (Clemmens, 1981).

iii) The method is not suitable for soils with macropores, where flow is not one-dimensional, or for soils with fragile surface crusts which will break when the ring's are pushed into the ground (Hoogmoed et al., 1991).

iv) It is difficult to maintain accurate water head when manually pouring in water. This will lead to over estimated results

particularly when measuring scale is a float. A constant head device such as marriote bottle will remove this problem.

v) In dry climates, infiltration rates especially in low permeable soils, may be over estimated due to contribution from evaporation losses.

vi) Air may become entrapped during ponding, creating internal air cushions which impedes infiltration. As a result, infiltration rate may drop, appear stationary or seen to increase after the escape of the encapsulated air.

vii) The tendency to over estimate infiltration rate does exist when the wetting front goes beyond the end of the ring (Tricker,1978).

viii) Properly seated rings used with a buffer generally measure infiltration rates less than the true infiltration rates (Clemmens, 1981).

2.6.2 Infiltration of simulated rain: Sprinkler infiltrometer

This method is meant to simulate rainfall and sprinkler irrigation conditions. However, each rainfall event has its characteristics. Intensity can change swiftly while it is fairly constant under simulated rainfall. The use of rainfall simulators may not therefore duplicate a true natural situation. The aggressive nature of the rainfall in the semi-arid region with

intensity of over 300 mmh^{-1} during 6 minutes as found by Hoogmoed (1986), can be very difficult to simulate. Another limiting factor to the use of rain simulator is that a lot of water and a large power generator are required for a rainfall simulation covering just an area of $10m^2$ or more (Hoogmoed et al., 1991).

Various types of rainfall simulators or sprinkling infiltrometers have been reported (Brockman, 1975; Zegalin and White, 1982). There are basically two systems of sprinkler infiltrometer used to produce raindrops: the nozzle and the drop-former type. The nozzle is the most widely used.

In the savanna region, simulators have to be made to produce realistically sized drops to hit the soil surface at a realistic terminal velocity, if they are to duplicate the aggressive rain in this region. Surface seals are formed when soils of the semi-arid region are exposed to the impact of drops. On drying, these soils form crusts. The sealing and subsequent crusting of these soils are the main cause of decrease in infiltration rate in this region (McIntyre, 1958; Morin and Benyamini, 1977; Agassi et al., 1985; Ben-Hur et al., 1985).

2.6.3 Disc permeameter

The disc permeameter designed by Perroux and Whites (1988) can supply water to insitu soils at a readily selectable positive and negative pressures. It is therefore developed to measure

infiltration rate under positive or negative pressure heads.

The simplicity, versatility, rapidity and ease of determination and sampling of infiltration rates of subsoils of this equipment, provides a good prospect for its future acceptance. Other methods of measuring infiltration rate are Basin and Furrow (Bondurant, 1957).

CHAPTER THREE

3.0 MATERIALS AND METHODS

3.1 SITE DESCRIPTION

The trials were conducted on the Institute for Agricultural Research (IAR) farm at Samaru (Latitude 11°11'N and Longitude 7°38'E). The soils are fairly representative of the loess plain unit of most agriculturally important lands of Northern Nigeria (Kowal, 1970b). The soil has been classified in the U.S soil Taxonomy as Typic-Haplustalf. The FAO-Unesco soil map of the world, classified it as Nitosol/Luvisol (Awujoola, 1979). The plains are undulating with gentle and broad slopes lying at an altitude of between 610m - 931m above sea level. The rainy season which normally begins from April and terminates in October has an annual mean rainfall of 1,060 mm.

The surface soil is predominantly loamy sand in texture with low structural stability (Wilkinson, 1975), and a high tendency to form crusts with consequent reduction in infiltration rates (Lawes, 1966). The presence of a B_t horizon in the profiles is thought to have a significant influence on the movement of water in the soil (Wild, 1972; Jones, 1975; 1976).

3.1.1 Physical and Chemical Properties

Disturbed soil samples from the experimental site were randomly collected from five different points with the help of an auger. The samples were collected from the top 0-5 cm, 5-10 cm and 10-15 cm. The samples from these depths were air-dried and sieved through a 2mm sieve for the determination of some physical and chemical properties of the upper soil layers.

Particle size distribution of the soil was determined by the *hydrometer method (Day, 1965). The textural class determination* were carried out using the USDA textural triangle.

To determine aggregate sizes, field samples from 0-5, 5-10 and 10-15 cm layers were gently crushed to pass through an 8 mm - opening sieve and then air-dried. Sub samples of the air-dry material were sieved using nest of sieves with mesh of 5, 2.36, 2,0.5 and 0.25 mm. The nest of sieves were shaken on a horizontally, reciprocating mechanical shaker for 15 min. The result of each sieve analysis was expressed as the mean weight diameter.

Soil P^H was determined in water and in 0.01 M $CaCl_2$ (1: 2.5 soil/solution ratio), using a pH meter with a glass electrode, after equilibrating for 30 minutes.

The organic carbon content of the soils was determined by the Walkley-Black procedure (Allison, 1965). The soil carbon was

oxidized with potassium dichromate in the presence of H_2SO_4 which supplied the heat for the reaction.

Cation exchange capacity was determined using ammonium acetate (P^H 7.0) saturation followed by displacement of the absorbed NH_4^+ by the acid-Nacl method described by Chapman (1965).

3.1.2 Field infiltration measurements

A total of 30 infiltration runs were carried out along a 60 m transect at 2 m intervals, between February and April, 1997. The measurements were made by the means of a double ring infiltrometer.

The double ring infiltrometer, is a set consisting of a large outer ring and a small inner ring. Three sets of different diameters were used. The first set has an outer ring of 60 cm and an inner ring of 35 cm diameter. The second set has an outer ring and an inner ring of 58 cm and 33 cm diameter respectively, while the third set has an outer ring and an inner ring of 55 cm and 30 cm diameter, respectively (Eijeikamp Agrisearch Equipment No. 09.04). Generally, the set includes an adjustable measuring bridge and float with measuring rod. Other accessories include a withdrawel clamp, impact absorbing hammer, driving plate and stop watch. The predetermined head of water in the inner ring was easy to maintain because of the conspicuous mark left on the ring by small pieces of crop residues, as the water in the ring recedes.

At each site, the cutting edges of both the inner and the outer rings were well seated on the soil surface and then the driving plate was attached by means of an adjustable screw. The process was done with care to avoid soil disturbance in the inner ring.

Infiltration was commenced by first applying water into the outer ring at a shallow depth. The water therein was allowed to infiltrate into the soil for sometimes. This was done with the view of providing a buffer to discourage lateral flow and encourage one - dimensional vertical flow (Clemmens, 1981). Immediately, water was added to the depth of 10 cm in the inner ring to simulate field heads usually encountered during irrigation. This was done after mounting the float and measuring rod over the inner ring using the adjustable measuring bridge. The initial reading on the measuring rod was recorded against the reference time (t = 0). Water was constantly allowed to recede to 7 cm in the inner ring and swiftly added to maintain the predetermined head.

The infiltration rate which was high initially, was read at one minute interval for the first 5 minutes, and at 5 minutes interval for the next 55 minutes, and 10 minutes thereafter for 60 minutes.

3.1.3 Soil sampling

At each site, samples were taken from adjacent soil for the determination of initial moisture content. The withdrawal clamp was used to carefully remove the double rings on termination of each infiltration run. Three undisturbed samples were taken from each site using core rings of dimension 50 mm diameter and 20 mm height (Eijelkamp Agrisearch Equipment No. 07.03.02.53). In taking samples, the core ring was fixed into the holder and gently but firmly pushed into the soil and dug out. A sharp spatular was used to trim the protruding soil on both ends of the ring, making sure they are not smeared. The rings are immediately capped to prevent desiccation.

Immediately, following core sampling, three soil samples were taken in moisture cans for the determination of final moisture content. Both the core samples and the samples in the moisture can were immediately weighed in the laboratory. While the samples in the moisture can were dried in the oven after weighing, the core samples were used to determine pore size distribution, saturated hydraulic conductivity and bulk density.

3.1.4 Pore - size distribution

Pore-size distribution is an important factor influencing infiltration. The number, size and continuity of pores have been

found to control the rate of flow of water through the soil to the plant-root-soil interface (Olson, 1985).

Pore-size distribution was measured on undisturbed samples collected from sites where infiltration run was carried out. The sample rings were covered with a piece of a nylon cloth held in place with rubber bands to prevent soil loss. The samples were saturated over night, weighed and placed in the sand box apparatus. They were pressed lightly and subjected to 0.1 bar suction, by lowering the surface of the reservoir to 100 cm below the level of the center of the samples. The suction was imposed and left for 24 hours during which time all water held at this tension was sucked. The samples were then weighed. After that, the samples were saturated over night and weighed. They were then subjected to suctions equivalent to 0.1, 0.3, and 1.0 bars in a low pressure plate extractor. First, a suction of 0.1 bar was imposed and maintained until the cessation of flow from the out flow tube for an equilibrium to be established. The samples were then taken out, weighed to determine loss in weight and replaced in the chamber. The procedure was repeated with suctions 0.3 and 1.0 bars. Finally the samples were oven dried at 105°C for 24 hours and weighed. Bulk density (Db) was calcutated from the formula:

$$Db = \frac{ovendryweight}{internalvolumeofthering} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (24)$$

Total porosity (F) was calculated from the formula;

$$F = \frac{W_w - W_d}{Bulk volume of soil} \quad \text{.....(25)}$$

where W_w is the weight of saturated soil and W_d is the weight of oven dry soil.

The volume of water extracted at a specified suction from a given volume of soil represents the volume of pores of the size indicated by that tension (Vomocil, 1965). Capillary rise equation in it's simplified form was used for the calculation of pore size. The equation is written as;

$$h = 3/d \text{ (Baver et al., 1972)} \quad \text{----------} \quad (26)$$

where h = suction level with which the water in the pore is in equilibrium (cm)

d = pore diameter (mm)

The volume of pores effective in water retention is given by the volume of water held between successively applied suctions, so that the effective pore diameter at 0.1 and 0.3 bar are 0.03 mm and 0.01 mm respectively (obtained by substitution in the equation given above). The difference between the weight of a sample at 0.1 bar and at 0.3 bar gave the weight of water and hence its volume (density of water is approximately = 1.0 g/cm^3) held between these suctions. This therefore, gives the volume of pores with diameter

of 0.01 to 0.03 mm.

This method is based on the assumption that capillary theory is valid for soil pores. As such, it is useful in comparing information about different soils rather than exact pore-size distribution (Vomocil, 1965).

3.1.5 Saturated hydraulic conductivity

The hydraulic conductivity was measured by the constant head method of Klute (1986), using the ICW laboratory permeameter (Eijikelkamp Agrisearch No. 09.02). The equipment operates on the principle that water is caused to flow through a saturated soil column by the pressure difference created on both sides of a well saturated soil sample.

The caps from the ring were removed and the samples were saturated over night in a basin of water. This was done by first of all covering the blunt end of the ring with a piece of nylon cloth (approximately 8 cm in diameter) which was held in place by means of a rubber band, to disallow soil loss. A specially meshed container was used to hold the ring which was in turn placed inside a plastic container. The container containing the sample was then inserted into the permeameter after establishing a constant head. A tube previously filled with water was used as a junction connecting the inside of the ring holder and the water in the

permeameter. This ensured flow of water into a burette. Depending on the ease with which water flows through the sample, the time at which a conveniently chosen volume is attained in the burette is taken using a stop watch. The height of water inside the ring holder and outside was measured and the saturated hydraulic conductivity (K_s) was calculated from the formula;

$$Ks=\frac{V.L}{At(\Delta H)} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots (27)$$

where V is the volume of water collected (cm^3); L is the length of soil column (cm); A is the cross-sectional area of the sample (equivalent to area of core ring) (cm^2); t is the time in minutes; ΔH is the hydraulic head difference (cm).

3.1.6 Curve fitting techniques

This technique employs the principle of finding an equation which best fits the data with minimum deviation. This is achievable through regression analysis accomplished by the least squares method. The least squares involves finding minimum sum of squares of deviation. The deviation implies the difference between the experimental infiltration rate (f_i) at time t_i and estimated infiltration rate f'_i at time t_i computed from a functional relationship. Sum of least squares is given by:

$$S = \sum_{i=1}^{N} (f_i - f^{\wedge} i)^2 \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (28)$$

where S = sum of square deviations; N = total number of observations. Regression analysis was used to estimate infiltration rates from Philip's and Kostiakov's equations. The equations are first of all linearized before fitting them into equation (28) above.

3.1.7 Steady state infiltration rate

Steady state infiltration rate is reached when cumulative infiltration (I) versus time (t) curve gives a linear relationship or dI/dt = constant.

3.1.8 Spatial analysis

The sampling interval at which the spatial series is repetitive and the maximum sampling distance for which the observations remain correlated, were determined by computing the autocorrelation function (r(h)) from equation (22) using the statgraphics software (STSC, 1986).

3.1.9 Sample numbers

The need for estimating sample numbers necessary to estimate the mean within some specified confidence interval, has been justified

by the heterogenous nature of soil across the field. If it is assumed that the samples are independent over distance, the number of samples necessary for the estimation (N) is given by

$$N = x_\alpha^2 \, \delta^2/d^2 \text{ ---------- (29)}$$

where δ = standard deviation; x_α = Normalized deviate at chosen probability level (tabulated alone or as the two tailed, student's t value with infinite degrees of freedom); d= percentage of the mean estimate.

Note that $x_\alpha = (x - \mu)/\delta$

CHAPTER FOUR

4.0 RESULTS AND DISCUSSIONS

4.1 SOIL PROPERTIES

The physical and chemical properties of the upper soil layers are presented in Table 1. Values of percent sand, silt, and clay for each depth are given in Table 1. The texture of the soil at the three depths is loam.

The sand content decreases with depth from 36% at the surface (0 - 5 cm) to 30% at the 10 - 15 cm. Clay content on the other hand increases with depth from 16% at the 0 -5 cm to 22% at the 10 - 15 cm depth. The silt content which was high in all the depths ranging from 46 to 48%, showed no regular pattern within the depths.

The higher sand fraction at the surface could be due to low clay content at that layer as a result of clay eluviation. The high silt content of the soil could indicate the influence of aeolian deposition on the area (Kowal, 1968).

Aggregation which was expressed as mean weight diameter, increse with increase in depth. This could be attributed to increase in clay content with depth.

The pH of the soil is presented in Table 1. The pH of both 0 -5 cm and 10 -15 cm is 6.1, indicating that these layers are slightly acidic in reaction. The 5-10 cm layer is moderately acidic with a pH of 5.8.

All the layers examined contained low organic carbon (Table 1). The organic carbon content ranges from 4.79 gkg^{1} at the 10 -15

cm layer to 5.98 gkg^{-1} at the surface 0 - 5 cm layer indicating decrease in organic carbon content with depth. This trend was also observed by Babaji (1982), Esu (1982) and Kparmwang (1987), although they reported higher values than the ones obtained here.

Cation exchange capacity (CEC) of savanna soils are influenced by organic matter contents rather than clay contents(Jones and Wild, 1975). This is because clay content is low and the clay is mainly kaolinitic and its contribution to CEC is frequently very small (Jones and Wild, 1975). Cation exchange capacity values for the depths are presented in Table 1. The CEC was generally low and it decreases with depth. This is expected because organic matter decreases with depth. The uppermost 0 - 5 cm layer has the highest CEC (4.39 cmol kg^{-1}), while the 10 - 15 cm layer has the lowest (3.28 cmol kg^{-1}).

Table 1: Some physical and chemical properties of the upper soil layers at the experimental site

Soil property	Soil depth (cm)		
	0 - 5	5 - 10	10 - 15
Particle size Analysis (%)			
Sand	36	**34**	30
Silt	48	**46**	48
Clay	16	**20**	22
Textural Class	Loam	Loam	Loam
Mean weight diameter (mm) of aggregates	2.38	2.76	2.58
pH (water)	6.10	5.80	6.10
pH ($CaCl_2$)	4.30	4.70	5.10
Organic Carbon (gkg^{-1})	5.98	5.39	4.79
CEC ($cmol\ kg^{-1}$)	4.39	3.69	3.28

4.2 PORE SIZE DISTRIBUTION

The pore size distribution and the mean pore volumes drained at the various suctions at all sites are presented in Table 2. The coefficient of variability which increases with decreasing soil water potential, could be attributed to a high variability in capillary pores with desaturation (Babalola, 1978; Hamblin, 1985).

At the 10 KPa suction equivalent to pore diameter of 30 μm, the volume of pores effective in water retention gave low variation (CV = 22.19%) between measured points (Table 2). The higher suctions (30 KPa and 100 KPa) equivalent to pore diameters of 10 μm and 3 μm, indicated wider variations (CV = 46.57% and 49.67%, respectively) in volume of pores drained between measured points. This may be attributed to reduction in pore volume as a result of soil shrinkage as water is removed (Kowal, 1970b). These values conform with the findings reported by Babalola (1978), Babaji 1982), and Kureve (1991)

Table 2: Summary table for selected soil physical properties for 30 sites

	$\bar{x}$	δ	CV
Bulk density (Mg/m^3)	1.39	0.06	4.03
Total porosity (%)	48.40	2.19	4.53
Macroporosity (%)	20.20	4.48	22.19
Volume (%) of pores at 3 different suctions with equivalent pore diameter of 30 μm, 10 and 3μm			
10 KPa (0.1bar) 30μm	20.20	4.48	22.19
30 KPa (0.3bar) 10μm	6.54	3.05	46.57
100 KPa (1bar) 3μm	5.52	2.74	49.67

$\bar{x}$ = Mean
δ = Standard deviation
CV = Coefficient of variability

4.3 BULK DENSITY, POROSITY AND MACROPOROSITY

The mean bulk density of all the 30 sites is presented in Table 2. The mean bulk density is 1.39 Mg m^{-3} and the standard deviation is 0.06 g cm^{-3}. The coefficient of variability between measured points was 4.03%, which is on the lowerside. The frequency distribution of the soil bulk density for the entire sampled area is given in figure 1. The graph shows a histogram plot of the data with the best - fitting normal distribution superimposed on it. The values are normally distributed. The data for each site are shown in Appendix 2.

The normal frequency distribution function which describes the bulk density values is

$$f = \frac{1}{\delta\sqrt{2\pi}} \exp\left[-(x - \bar{x})^2/2\delta^2\right] \quad \text{---------- (30)}$$

where f is the frequency, $\bar{x}$ the mean, δ the standard deviation of the mean, and x the random variable.

When the bulk density data was subjected to spatial analysis and autocorrelograms plotted (Appendix 1), it was clear that there was no separation distance at which the autocorrelation coefficient exceeded the 95% confidence limit. This shows that the bulk density result for the 30 sites were spatially independent. It is therefore not possible to estimate the value of an observation from its nearest neighbour.

Total porosity is high with a mean of 48.40%, (Table 2). It

exibits a poor correlation with bulk density with a coefficient of correlation, r = -1.00 (Table 4). Total porosity like bulk density, has low coefficient of variability (4.53%) (Table 2) between measured points.

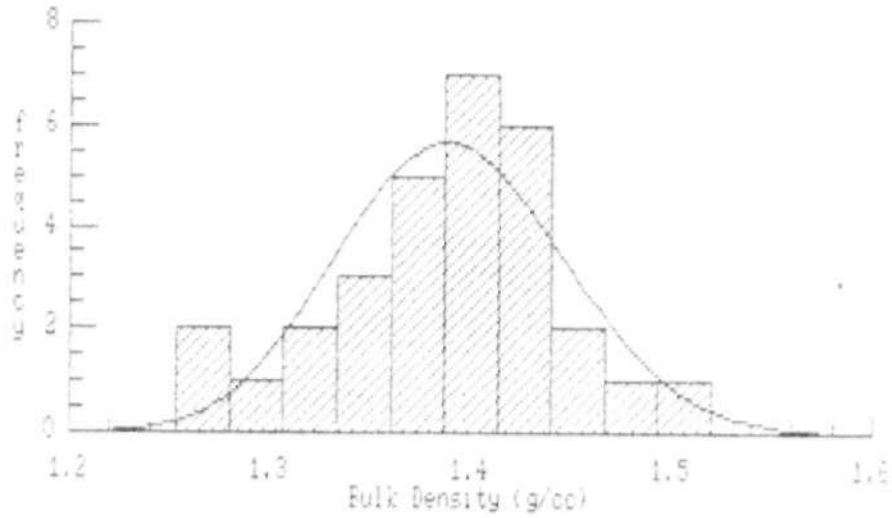


Fig.1: Frequency Distribution for Bulk Density Values

Figure 2 depicts the total porosity histogram and the normal frequency curve for 30 sites. Frequency distribution for total porosity which is adequately described by equation (5.0), shows a normal distribution. Total porosity values for each site are shown in Appendix 2. The autocorrelogram plot (Appendix 1) showed spatial independence of observations.

The result of macroporosity determination is presented in Table 2 and it shows a medium coefficient of variability (22.19%) between measured points. These are pores with diameter > 30 μm which are drained at suction lower than 10KPa. Therefore, the trend is still maintained because it is showing lower CV than those drained at higher suctions (Table 2).

Subjecting this result to spatial analysis revealed no degree of dependency as shown in Appendix 1. In other words, it is not possible to estimate from one neighbour the expected value of another neighbour in a pair of observations.

4.4 SATURATED HYDRAULIC CONDUCTIVITY (K_S)

Saturated hydraulic conductivity showed very high variability (CV = 83.23%) between measured points (Table 3) but properties such as bulk density and porosity generally had CVs of less than 10% (Table 2). Less precise methods for measuring K_s and more importantly the inherent heterogeneity in the pore geometry both vertically in the profile and spatially in the land scape, are the ones accounting for most of these variations (Hamblin, 1985).

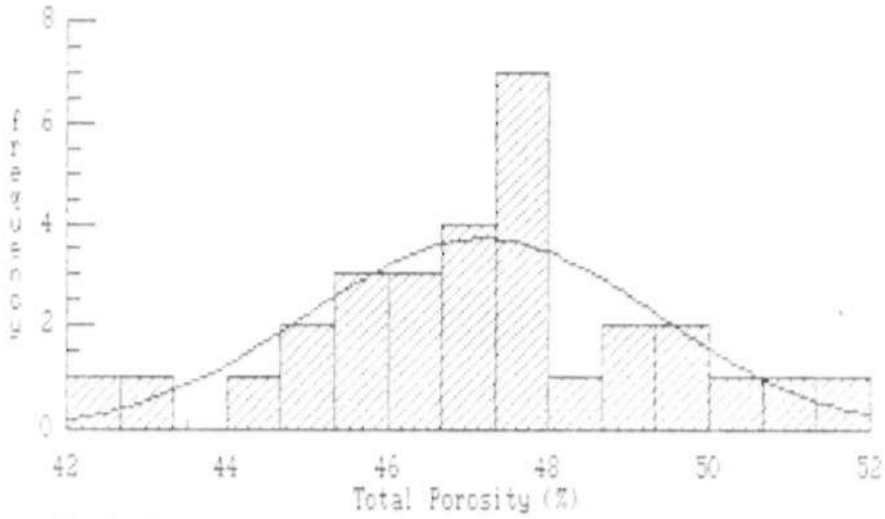


Fig.2: Frequency Distribution for Total Porosity Values

Similar findings were documented in literature (Nielsen et al., 1985;Kureve, 1991). Mean saturated conductivity for all the 30 sites was 44.78 mm/hr with a standard deviation of 37.27 mm/hr (Table 3). These figures are higher than Kowal (1970b), Ahmed (1982), and Kureve (1991).

The frequency function of a log - normal distribution is

$$f = \frac{1}{\delta(x-ß\sqrt{2\Pi})} \exp\{-[\ln(x-ß)-x]^2/2\delta^2\} \quad \text{---------- (31)}$$

for $x > ß$ and $f = 0$ for $x < ß$, where ß is any constant to allow ln (x-ß) to be normally distributed. Figure 3 depicts the distribution of hydraulic conductivity values measured for 30 sites at 2 meters interval which is described by equation (5.10). Apparently, values of the hydraulic conductivity are log-normally distributed.

A plot of autocorrelation function against lag indicated that the observations are spatially independent because the autocorrelograms (Appendix 1) were all close to zero. To estimate the population mean and be within ± 10% of its estimated value at the 95% confidence level, 58 average samples must be obtained.

Bulk density which is thought to be an important index in predicting saturated conductivity (Hamblin, 1985; Puckett et al., 1985) was found to be poorly and negatively correlated ($r = -0.11$) with saturated conductivity as shown in Table 4. The umpredictable nature of pore-size distribution which differ even at few centimeters of separation (Dixon and Peterson, 1971; German and Beven, 1981; Hamblin, 1985; Grismer, 1986; Watson and Luxmore,

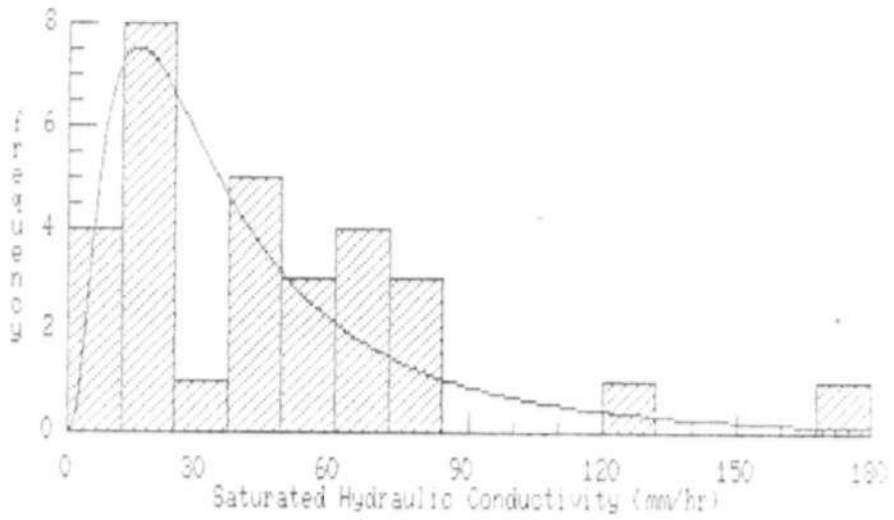


Fig.3: Frequency Distribution for Saturated Conductivity Values

Table 3: Steady State Infiltration Rate, Saturated Hydraulic Conductivity and Estimated Soil Parameters for Infiltration Equations from Curve Fitting.

		$\bar{x}$	δ	CV
Steady infiltration (mm/hr		27.40	15.25	55.66
Saturated conductivity (mm/hr)		44.78	37.27	83.23
Estimated soil parameters for infiltration equations from curve fitting				
Kostiakov	M	62.69	24.27	38.71
	n	0.48	0.07	13.51
	R^2	0.997		
Philip	A	0.82	14.18	1736.12
	s	60.89	22.48	36.91
	R^2	0.999		

1986) may be responsible for this poor relationship. There was also a poor correlation (r = 0.32) between hydraulic conductivity and macroporosity and between hydraulic conductivity and moisture content (r = 0.14) as given in Table 4. This is not surprising inview of the high variability of the soil hydraulic properties.

Table 4: Correlation Table

Correlation coefficient (r)	Correlated parameters
0.11	Steady-state infiltration and saturated conductivity
0.32	Steady-state infiltration and Total porosity
0.33*	" " " and Macroporosity
-0.32	" " " and Bulk density
0.94**	" " " and Sorptivity M
0.81**	" " " and Parameter n
0.73**	" " " and Parameter A
0.58**	" " " and Sorptivity S
0.02	" " " and Initial moisture content
-0.04	" " " and Final moisture content
0.11	Saturated conductivity and Total porosity
0.32	" " and Macroporosity
-0.11	" " and Bulk density
0.04	" " and Sorptivity M
0.07	" " and Parameter n
0.03	" " and Sorptivity S
0.07	" " and Parameter A
0.31	Total porosity and M
-1.00	" " and bulk density
0.03	" " and n
0.11	" " and A
0.26	" " and S

4.5 INFILTRATION RATE

The infiltration rate at all 30 sites assumed a general trend of very high initial values which reduce very sharply within the first 10 minutes. Intake rates decreased only slightly after 1 hour thus the rate between 1¼ and 2 hours was taken as the equilibrium infiltration rate. Ahmed (1982), Adeoye (1986), and Kureve (1991) reported 3½ hours, 4 hours and 40 minutes respectively for the attainment of steady-state infiltration rate in Samaru soils.

The steady-state in all the runs was not attained through a steady reduction in infiltration rate but there was an evident rising and falling in a run. Entrapped air, slaking (Kowal, 1970a; Dixon and Linden, 1972; Beven and German, 1982; Babalola, 1986; Kureve, 1991), inherited heterogeneities of field soils (Hillel, 1980), water level in the infiltrometer and errors in reading and timing are the likely causes of the observations.

To compare the performance of kostiakov's and Philip's models in describing infiltration data of Samaru soils, curve fitting by regression (Table 3) was carried out and in both cases, R-squared value was > 0.98. This is an indication that both Kostiakov's and Philip's equations adequately describe the data. Therefore, either of the equation can suitably predict infiltration rates into Samaru soils.

The estimated parameters (M and n for Kostiakov's equation and A and S for Philip's equation) are given in Table 3. An attempt was made to determine which of the models gives the best fit

Standard deviation was calculated for the parameters in the two equations. The Kostiakov's parameters gave a standard deviation value of 17.16 while Philip's parameters gave a standard deviation value of 18.70. This clearly indicate the superior performance of the Kostiakov's model. Also when the two models were used to compute cumulative infiltration after 120 minutes of infiltration and compared with the measured cumulative infiltration, Philip's model gave the highest coefficient of variability (46.25%) between measured points (Table 5).

The steady-state infiltration rate of the 30 sites measured gave a mean of 27.4 mm/hr and a standard deviation of 15.25 mm/hr resulting in a relatively high coefficient of variability (55.66%) between measured points (Table 3). These values were too irregular to fit into a normal distribution and to be described by equation (5.0). Figure 4 depicts the distribution of steady-state infiltration rate which appear log-normally distributed and described by equation (5.10). The high variability of steady-state infiltration rate is at variance with the low CV reported by Umouru and Babalola (1986) and Kureve (1991) but similar to Folorunso and Olorunju's (1986) findings.

The high variability led to the subjection of the data to autocorrelation analysis as a means of expressing changes over distance and whether or not there is dependency among neighbouring observations. The autocorrelogram(Appendix 1) indicated that there was no separation distance at which the autocorrelation coefficient exceeded the 95% confidence limit. An obvious conclusion is that

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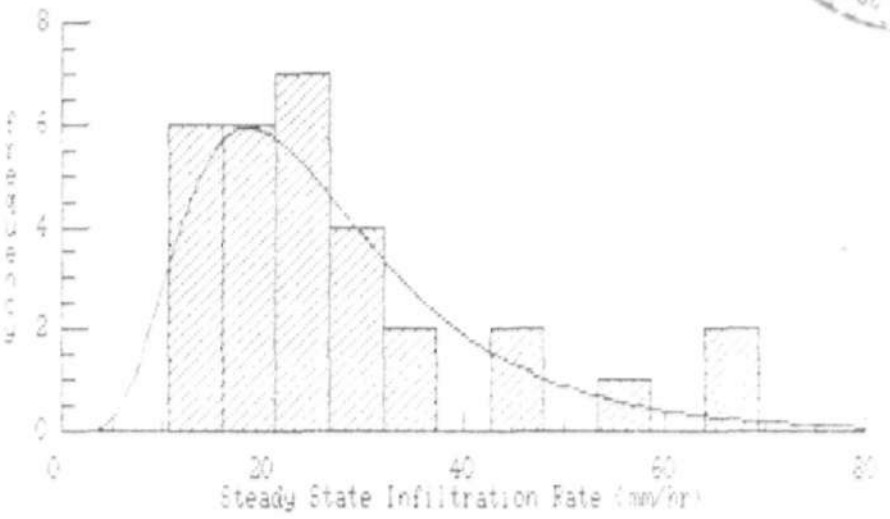


Fig.4: Frequency Distribution of Infiltration Rate

observation of steady-state infiltration rates for the 30 sites were spatially independent at the sampling intervals of 2 meters. This suggest that one cannot estimate the value of an observation from its nearest neighbour. This shows that, interpolation of infiltration to unsampled sites in the field will not be reliable. This agrees with the findings of Babalola (1986) and Kureve (1991). To obtain a reliable estimate of the mean infiltration rate for the field studied and be within $\pm$ 10% of the population mean at 95% probability, 10 samples are required.

From curve fitting results (Table 3), variation was observed in the parameters of the two models between measured points. This is expected because of soil heterogeneity and variations in surface conditions. The Kostiakov's time exponent n, gave the least variation in the field (CV = 13.51%) for all the parameters. The closeness of the mean value of the Kostiakov's time exponent (0.48) and the time exponent of Philip's equation (0.5) may be responsible for their similarity in describing the infiltration data.

A very striking observation is the one on the constant A, in Philip's equation. Negative values were obtained in 19 sites out of the 30 sites sampled (Appendix 2). This could be explained from the fact that the rates started from high values (about 360 to 240 mm/hr and reduced rapidly to about 66 to 12 mm/hr within 2 hours infiltration run. This rapid decrease resulted in a steep slope of the regression line to the extent that it's extension cuts the y-axis below zero giving a negative intercept. This indicates that

Philip's equation must have restriction on structurally poor soils, since negative values are obtained at longer periods when the first term is insignificant. Prediction of infiltration rate using Philip's equation therefore, is possible within a limited time range.

Deductions made on infiltration parameters and their physical meaning on Samaru soils, indicated that the time coefficient in both equation's reflect the same soil parameter; soil sorptivity.

The Kostiakov time exponent n, was poorly correlated with porosity (r = 0.27) but highly correlated with M (r = 0.75**) and S (r = 0.45**) (Table 4). The time coefficient of Kostiakov equation M, was found to be strongly correlated with sorptivity (r = 0.80**) indicating that it is also related to water intake. Steady-state infiltration rates and conductivity gave a poor correlation (r = 0.11) (Table 4). This was unusual because from the Richard's equation, at steady-state situations, matric potential becomes negligible and gravitational potential is the only one that is operational. If this potential is one for vertical flow, then infiltration at steady-state will solely be determined by the conductivity of the soil. Which is not so in this case. Experimental technique used here may be responsible for this poor relationship. The saturated hydraulic conductivity whose result is presented in Table 3, was determined on core samples of limited area (20 cm^2) compared to the ring infiltrometer area (706.5 cm^2) i.e about 2.8% of surface area through which infiltration rates were determined. The resultant effect is that

a single pore in the core samples could account significantly for the values of conductivity obtained (Dixon and Peterson, 1971; Watson and Luxmoore, 1986). Whereas infiltration measurements were an average of such limited areas enclosed by the ring infiltrometer. To get a more realistic relationship between these properties, sample size will have to be readdressed.

Table 5: Measured and Computed Values of Cumulative Infiltration Rate (I) after 120 minutes

	$\bar{x}$	δ	CV
I (mm)			
Measured	88.93	39.48	44.40
Computed by Kostiakov' equation	88.49	38.08	43.03
Computed by Philip's equation	87.75	40.58	46.25

Both initial and final moisture content presented in Table 6 were poorly correlated with infiltration rate (r = 0.23 and 0.04 respectively) (Table 4). This is true because initial moisture content has been found not to influence the equilibrium rate but only makes soil to attain final rate slowly or quickly depending on the moisture content (Lal and Greenland, 1978).

Macroporosity showed significant correlation (r = 0.38*) with infiltration rate (Table 4). This is expected because a change in pore-size distribution toward a narrow range of larger pores has been reported to increase infiltration rate of as much as five times for typical fine-textured soils (Grismer, 1986).

The autocorrelograms of Kostiakov's parameters M and n, and Philip's parameters A and S, are presented in Appendix 1. These autocorrelograms indicated spatial independence for all the parameters.

Table 6: Volumetic Moisture Content and Moisture Percentage (at 10 KPa, 30 KPa and 100 KPa)

	$\bar{x}$	δ	CV
Volumetric moisture (θV) content (cm^3/cm^3)			
Initial (θV_o)	0.018	0.014	77.74
Final (θV_1)	0.352	0.039	10.97
Moisture percentage			
10 KPa	28.24	3.39	12.02
30 KPa	21.87	4.65	21.26
100 KPa	16.39	4.15	25.31

CHAPTER FIVE

5.0 CONCLUSION

The study was aimed at evaluating the magnitude of spatial variation of infiltration rate over a field considered to be uniform for agricultural purposes. It was carried out with the ultimate aim of providing information to enable the farmer decide on whether or not, his field can receive uniform agronomic management, such as irrigation, without the risk of runoff or *ponding*.

The study involved the use of double ring infiltrometer to measure infiltration rates along 60m transect at 2m intervals. The experiment was sited at Samaru in the Northern guinea Savanna.

Surface crusting and soil compaction are factors that generally influence infiltration rate. The non correlation between porosity and infiltration in this study, suggests that the crust rather than the transmission zone controlled infiltration. There was relatively high variation in infiltration with the mean of 27.4mm/hr and a standard deviation of 15.25mm/hr. These values were also log-normally distributed. To obtain a reliable estimate of mean infiltration rate and be within $\pm$ 10% of the population mean at 95% probability, 10 samples are required.

The infiltration data was satisfactorily described by Kostiakov and Philip's equations ($R^2 > 0.98$), although Kostiakov's equation showed some superior performance over Philip's equation. This was evident from the standard deviation value of Kostiakov's parameters (17.16) which was lower than that of Philip's parameters

(18.70).

The time coefficient of Kostiakov's equation M, was strongly correlated ($r = 0.80^{**}$) with sorptivity parameter s, of Philip's equation. The s parameter was related to initial intake rates while A associated with infiltration rates at higher times. The Kostiakov's time exponent n, was poorly correlated with porosity ($r = 0.03$) and with hydraulic conductivity ($r = 0.07$). It is a capacity index with mean 0.48 for the field. The decay of both Kostiakov and Philip's equation at a comparable rate, may not be unconnected with the fact that n value is close to Philip's time exponent (0.5).

Prediction of saturated hydraulic conductivity was difficult because of its large coefficient of variability. The mean and standard deviation values of hydraulic conductivity were 44.78mm/hr and 37.27mm/hr giving a CV of 82.23% and a frequency distribution that is log-normal. The variation was spatially independent. Fifty-eight (58) average samples are therefore required to estimate the mean saturated conductivity within $\pm$ 10% of the population mean at 95% confidence interval. Porosity and bulk density thought to be good indications of conductivity, gave poor correlation with conductivity. Heterogeneity of the soil is the factor responsible for this poor relationship.

As a consequence of this study it can be concluded:

1. Variations in bulk density and porosity are normally distributed over space, while values of hydraulic conductivity and infiltration rates were log-normally distributed.

2. The high variation and spatial independence in infiltration rate indicate that one single farmer's field can exhibit wide variations in its ability to take water either from natural rainfall or from irrigation.
3. Even seemingly uniform fields manifest large variations in hydraulic conductivity values. Variations in bulk density, porosity and water content are much less. Thus, it can be very unreliable to use measurement of hydraulic conductivity from a single plot as a representative of a large area.
4. In the Northern Guinea Savanna where substantial irrigation is practiced, effective management packages should be designed to improve infiltration and reduced runoff. Such management packages should aim at improving the soil structure and preventing surface sealing through the use organic matter and mulches.

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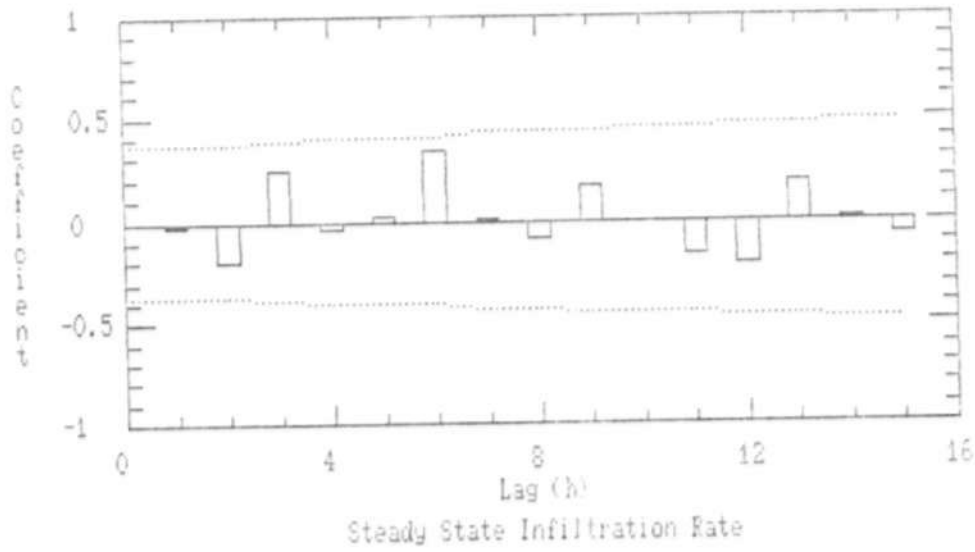
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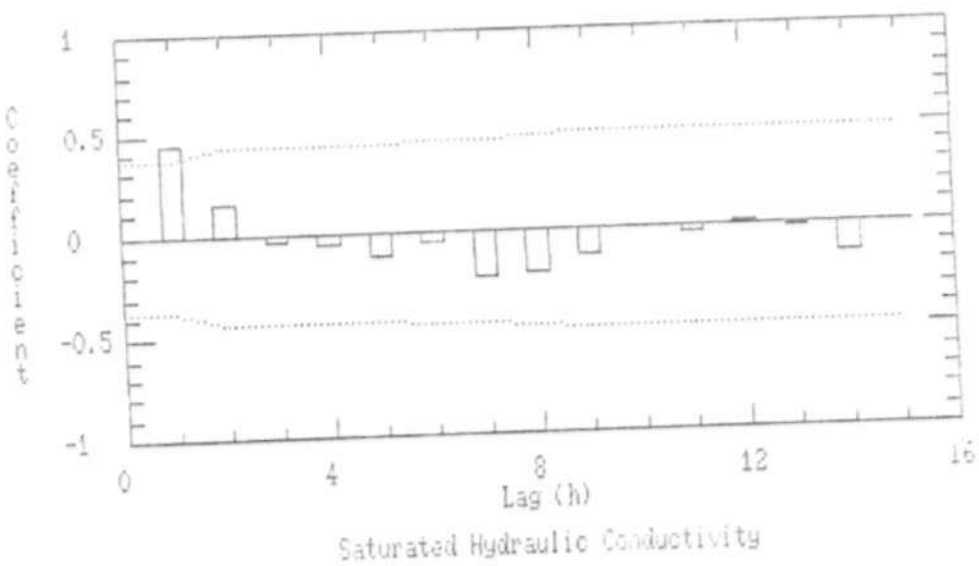
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APPENDIX 1

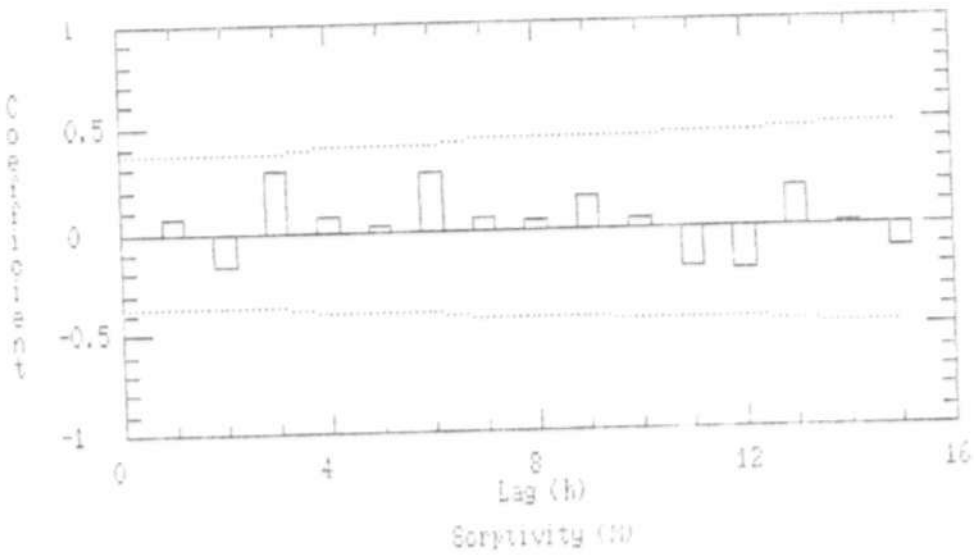
AUTOCORRELOGRAMS

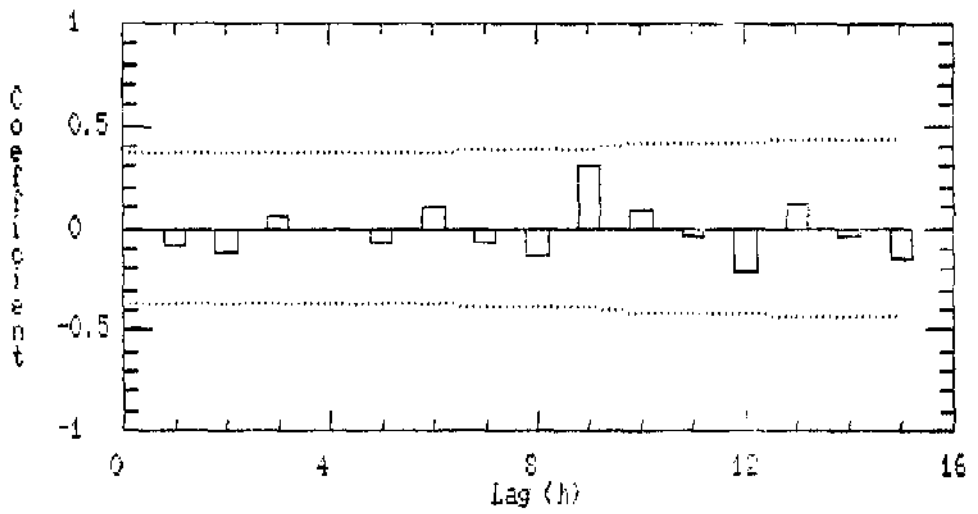


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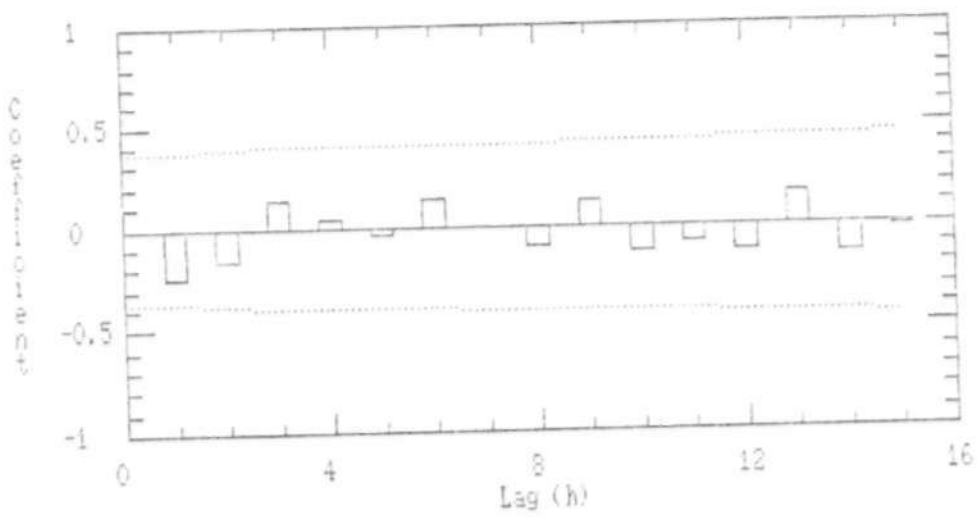


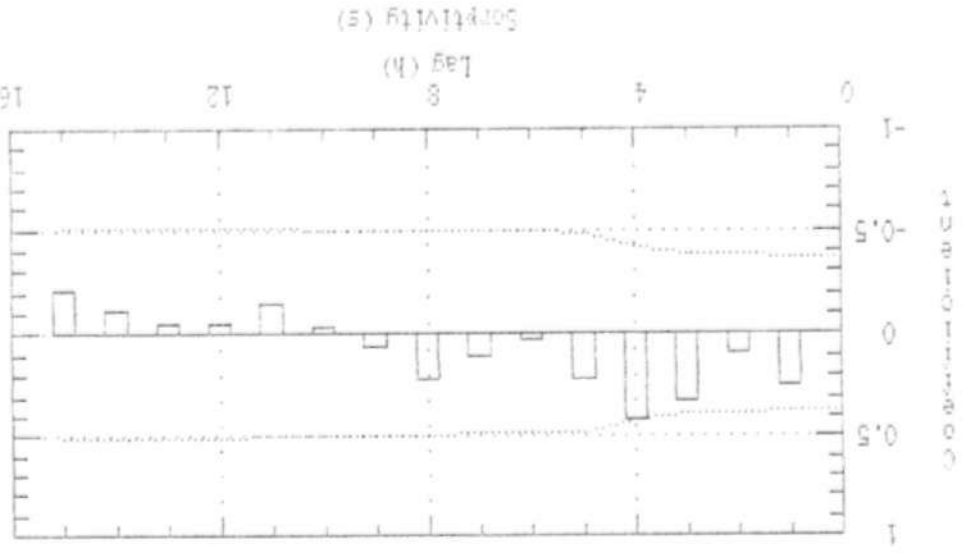
Saturated Hydraulic Conductivity

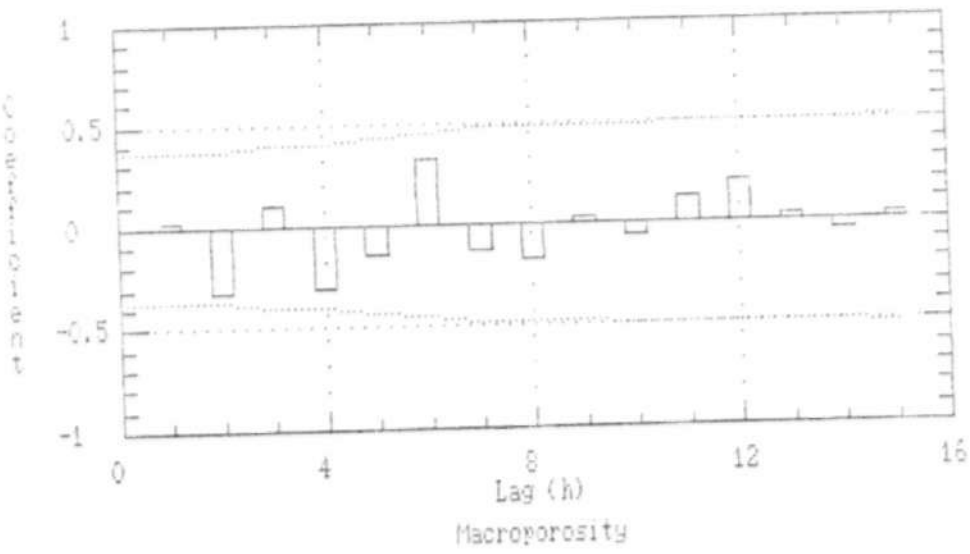

1
0.5
0
-0.5
-1
0
4
8
12
16
Lag (h)
Sorptivity (S)

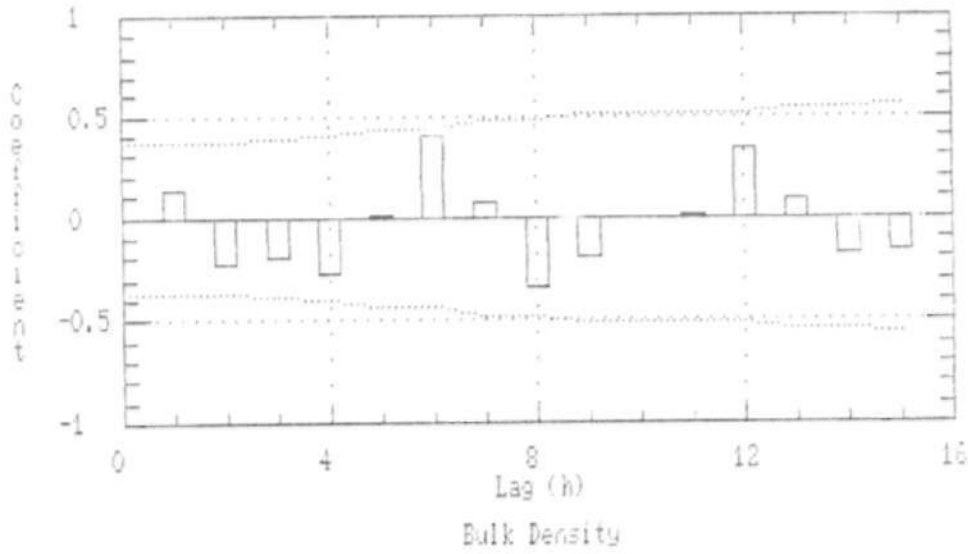


Structural Stability Index (n)


1
0.5
0
-0.5
-1
0
4
8
12
16
Lag (h)
Philip 'A' Parameter


Correlation
1
0.5
0
-0.5
-1
0
4
8
12
16
Lag (h)
Correlation (s)


1
0.5
0
-0.5
-1
0
4
8
12
16
Lag (h)
Macroporosity


1
0.5
0
-0.5
-1
0
4
8
12
16
Lag (h)
Bulk Density

APPENDIX 2

SOIL HYDRAULIC AND RELATED PHYSICAL PROPERTIES FOR 30 SAMPLED SITES

Estimated Soil Parameters for Infiltration Equations from Curve Fitting.

Site	Kostiakov			Philip		
	M	n	R^2	A	s	R^2
1	91.568	0.585	1.00	16.243	76.439	1.000
2	61.654	0.441	0.992	-2.369	62.641	0.999
3	63.959	0.506	0.998	-8.701	75.353	1.000
4	113.955	0.618	0.999	30.970	86.340	0.993
5	38.814	0.443	0.997	-7.818	46.610	0.998
6	63.489	0.522	0.995	-12.981	81.156	1.000
7	101.260	0.513	1.000	9.520	89.718	1.000
8	70.305	0.470	0.999	-2.789	71.770	0.998
9	79.82	0.405	0.998	-12.069	86.708	1.000
10	129.851	0.574	1.000	16.694	116.064	1.000
11	38.056	0.387	0.996	-10.180	46.312	0.999
12	38.820	0.468	0.995	-9.208	49.186	1.000
13	58.198	0.497	1.000	-9.910	70.476	1.000
14	80.856	0.560	0.996	-4.057	88.996	1.000
15	58.198	0.446	0.999	-0.296	55.641	0.999
16	45.024	0.418	0.999	-1.304	42.817	0.998
17	86.196	0.479	0.998	-0.515	83.981	0.999
18	62.797	0.455	0.991	-13.309	76.259	0.999
19	49.684	0.460	0.995	9.145	36.391	0.998
20	38.677	0.425	0.995	-7.366	44.727	0.999
21	35.350	0.445	0.997	-4.566	39.058	0.999
22	60.184	0.488	0.989	-12.381	86.271	0.998
23	88.515	0.627	0.994	48.831	35.558	1.000
24	48.040	0.476	0.997	7.660	37.260	0.995
25	35.124	0.371	0.994	-4.089	34.944	0.998
26	42.312	0.436	0.998	0.989	38.259	0.999
27	52.040	0.396	0.999	-4.050	51.406	0.999
28	55.266	0.500	0.995	13.688	37.668	0.999
29	46.890	0.453	0.996	3.970	39.567	0.999
30	45.164	0.462	0.996	3.747	39.213	0.999
$\bar{x}$	62.69	0.48	0.997	0.82	60.89	0.999
δ	24.27	0.07	0.003	14.18	22.48	0.002
CV	38.71	13.51	0.277	1736.12	36.91	0.16

Measured and Computed Values of Cumulative Infiltration (I)

		I, mm		
Site	Elapse Time, h	Measured	Computed by Kostiakov's Equation	Computed by Philip's Equation
1	2	140	137.36	140.59
2	2	84	83.70	83.85
3	2	87	90.83	89.16
4	2	180	174.89	184.04
5	2	51	52.76	50.28
6	2	85	91.17	88.81
7	2	149	144.50	145.92
8	2	98	97.38	95.92
9	2	103	105.69	98.49
10	2	195	193.30	197.53
11	2	48	49.77	45.14
12	2	51	53.70	51.14
13	2	78	82.13	79.85
14	2	113	119.20	117.75
15	2	81	79.74	78.10
16	2	62	60.16	57.94
17	2	121	120.14	117.74
18	2	81	86.33	81.23
19	2	74	68.34	69.75
20	2	50	51.93	48.52
21	2	47	48.12	46.10
22	2	76	84.41	79.24
23	2	151	136.70	147.95
24	2	71	66.82	68.01
25	2	46	45.42	41.24
26	2	60	57.24	56.08
27	2	70	68.48	64.60
28	2	85	78.16	80.65
29	2	67	64.19	63.90
30	2	64	62.21	62.95
$\bar{x}$		88.93	88.49	87.75
δ		39.48	38.08	40.58
CV		44.40	43.03	46.25

Steady State Infiltration Rate and hydraulic Conductivity.

Site	Steady-state infiltration rate (mm/hr)	Average conductivity (mm/hr)
1	48	44.22
2	24	21.20
3	30	11.67
4	66	66.34
5	12	63.95
6	18	53.57
7	48	36.55
8	30	73.47
9	36	42.80
10	66	39.67
11	12	24.90
12	18	55.13
13	24	17.42
14	36	83.31
15	24	12.37
16	18	16.97
17	30	48.99
18	18	68.92
19	24	18.59
20	12	13.13
21	12	13.43
22	18	6.81
23	54	13.97
24	24	39.50
25	12	10.23
26	12	7.93
27	18	65.15
28	30	74.38
29	24	169.71
30	24	129.00
$\bar{X}$	27.4	44.78
δ	15.25	37.27
CV	55.66	8.3.23

Volumetric Moisture Content (θV) (cm^3/cm^3)

Site	Initial (θV_0)	Final (θV_1)	Bulk Density (g/cm^3)	Total Porosity (%)	Macroporosity (%)
1	0.014	0.372	1.43	46.64	13.75
2	0.029	0.360	1.44	46.51	17.10
3	0.014	0.361	1.39	48.56	17.83
4	0.016	0.306	1.33	50.74	18.04
5	0.000	0.359	1.38	48.81	14.19
6	0.035	0.403	1.39	48.42	14.05
7	0.000	0.335	1.34	50.25	18.93
8	0.015	0.421	1.45	43.99	14.42
9	0.051	0.368	1.27	53.17	21.38
10	0.039	0.403	1.30	51.89	21.30
11	0.014	0.409	1.41	47.79	18.50
12	0.000	0.317	1.51	43.69	14.28
13	0.014	0.294	1.40	47.92	22.17
14	0.028	0.353	1.41	47.67	18.04
15	0.000	0.331	1.38	48.95	26.42
16	0.013	0.333	1.28	52.78	27.69
17	0.014	0.423	1.41	47.53	20.41
18	0.015	0.397	1.47	45.49	20.97
19	0.028	0.290	1.38	48.69	22.74
20	0.029	0.374	1.44	46.38	22.41
21	0.046	0.378	1.35	49.97	27.72
22	0.033	0.356	1.37	49.07	32.52
23	0.000	0.300	1.43	47.02	20.41
24	0.015	0.350	1.46	45.87	16.63
25	0.007	0.302	1.44	46.38	19.58
26	0.014	0.298	1.42	47.16	20.36
27	0.014	0.345	1.38	49.50	18.60
28	0.004	0.326	1.36	49.72	23.66
29	0.013	0.330	1.32	51.00	24.03
30	0.014	0.361	1.39	48.42	17.87
$\bar{X}$	0.018	0.352	1.39	48.40	20.20
δ	0.014	0.039	0.06	2.19	4.48
CV	77.74	10.97	4.03	4.53	22.19

Volume (%) of Pores at 0.1, 0.3, and 1.0 bar with Equivalent Pore Diameter of 30 μm, 10 μm, and 3 μm.

Site	10 KPa 30 μm	30 KPa 10 μm	100 KPa 3 μm
1	13.75	6.05	3.10
2	17.10	8.95	4.46
3	17.83	9.58	4.03
4	18.04	5.13	10.51
5	14.19	6.07	7.00
6	14.05	6.91	5.72
7	18.93	6.46	6.72
8	14.42	5.73	4.09
9	21.38	8.28	4.12
10	21.30	8.21	3.43
11	18.50	8.16	4.36
12	14.38	7.52	4.63
13	22.17	9.08	4.31
14	18.04	9.21	5.49
15	26.42	6.00	4.39
16	27.69	9.26	3.76
17	20.41	11.48	3.52
18	20.97	8.23	7.83
19	22.74	10.50	3.23
20	22.41	8.95	3.35
21	27.72	8.43	3.82
22	32.52	7.19	4.23
23	20.41	1.00	7.80
24	16.63	3.00	14.26
25	19.58	1.67	7.80
26	20.36	1.67	5.77
27	18.60	2.67	6.11
28	23.66	3.67	5.43
29	24.03	1.66	1.02
30	17.87	2.00	11.30
$\bar{x}$	20.2	6.42	5.52
δ	4.48	2.99	2.74
CV	22.19	46.58	49.69

Moisture Percentage at 10KPa, 30KPa, and 100KPa.

Site	10KPa	30KPa	100KPa
1	32.18	26.13	23.12
2	28.84	19.92	15.65
3	30.12	20.30	17.24
4	33.82	28.60	18.35
5	34.09	28.06	21.11
6	33.78	26.83	21.31
7	30.91	24.43	17.73
8	30.99	25.27	21.32
9	31.41	23.03	18.96
10	31.16	22.27	18.85
11	29.19	21.01	16.74
12	28.99	21.55	16.96
13	25.76	16.66	12.42
14	29.51	20.26	14.90
15	22.77	16.74	12.32
16	25.47	16.09	12.38
17	27.11	15.61	12.08
18	24.51	16.27	8.57
19	25.85	15.36	12.24
20	25.05	15.12	11.81
21	21.64	13.95	10.22
22	22.43	16.71	10.37
23	26.64	25.41	18.16
24	29.10	26.03	12.02
25	26.60	24.91	17.32
26	26.74	25.09	19.27
27	30.13	27.46	21.35
28	26.07	23.39	17.27
29	26.97	25.30	24.29
30	30.34	28.31	17.28
$\bar{x}$	28.24	21.87	16.39
δ	3.39	4.65	4.15
CV	12.02	21.26	25.31