ESTIMATING UNSATURATED HYDRAULIC CONDUCTIVITY OF A SANDY CLAY LOAM SOIL FROM BULK DENSITY DATA

BY

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A DISSERTATION SUBMITTED IN PARTIAL FULFILMENT

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ABSTRACT

The Gardner model, $K = K_{sat}/[1 + (bh_m)^n]$, with three parameters (K_{sat}, b, n), was calibrated and validated with field data for estimating the unsaturated hydraulic conductivity for a sandy clay loam soil in semi-arid Dodoma, Tanzania.

Calibration of the model was achieved by measuring the *in situ* saturated and unsaturated hydraulic conductivity at different levels of bulk density (1.43, 1.64, 1.85 Mg/m^3) and matric suction (0.0, 0.2, 0.5, 0.8 and 1.0 kPa) using the Disc Permeameter method which is a surface technique. K_{sat} was obtained from the graph of $\log(K_{sat})$ versus bulk density. The parameters b and n were obtained by solving the simultaneous equations obtained by measuring K at different soil matric suctions. For the sandy clay loam soil at Hombolo, K_{sat} was obtained from the linear equation, $\log(K_{sat}) = 5.644 - 2.164\varrho_b$ and values of b and n are 0.66 and 1.32, respectively.

The model was validated using the Disc Permeameter method by measuring the *in situ* unsaturated hydraulic conductivity at different bulk densities (ranging between 1.3 and 2.1 Mg/m^3) and different matric suctions (0.2, 0.5, 0.8 and 1.0 kPa within the same range of bulk density) from the plots

which initially were not used in the model calibration measurements. The unsaturated hydraulic conductivity was estimated from the same values of bulk densities used in measuring the *in situ* unsaturated hydraulic conductivity by using the calibrated Gardner model. A coefficient of determination (R^2) = 0.997 was obtained when the estimated unsaturated hydraulic conductivity values were plotted against the ones measured in the field, indicating that easily measured soil characteristics such as bulk density for a particular soil, can be used to estimate the hydraulic conductivity of that particular soil.

DECLARATION

I, SHAKWAANANDE ROLAND NATAI, do hereby declare to the Senate of the Sokoine University of Agriculture that this dissertation is my own original work and that it has not been submitted, in whole or in part, for a degree in any other University.

Signature: ...Natai...

Date: ...14/02/1997...

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DEDICATION

This dissertation is affectionately dedicated to my husband Roland and my children, Daniel and Victory for their love, moral support, encouragement and patience throughout this study.

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LIST OF SYMBOLS & ABBREVIATIONS

SYMBOL	DESCRIPTION	UNITS
A	specific surface area	m^2/m^3
a	empirical constant	-
b	empirical constant	-
Cl	clay content	%
CSIRO	Commonwealth Scientific and Industrial Research Organization	-
CTFTW	Canadian Training Fund for Tanzanian Women	-
D	hydraulic diffusivity	m^2/s
da	height of the drain above the impervious layer	m
D_b	soil bulk density	Mg/m^3
E_t	rate of evapotranspiration in a vegetated field	m/s
g	acceleration due to gravity	m/s^2
H	hydraulic head (being the sum of gravitational and matric suction heads)	m
$\Delta h/\Delta l$	summation of the driving forces in the direction of flow or hydraulic gradient	m/m
∇H	hydraulic gradient	-
h_m	matric potential	Pa
h_p	fluid pressure head of phase p in the soil	m

J_p	flux of phase p (water, non-aqueous phase liquid or air)	m/s
K	hydraulic conductivity	m/s
k'	pore shape factor ranging from 2 to 2.5	-
K est	estimated hydraulic conductivity	mm/day
K meas	measured hydraulic conductivity	mm/day
$K(\theta)$	unsaturated hydraulic conductivity	m/s
K_p	relative permeability of the medium to phase p	m/s
K_r	residual hydraulic conductivity	m/s
K_{sp}	saturated hydraulic conductivity of phase p in the medium	m/s
K_{sat}	saturated hydraulic conductivity	m/s
K_{sat},e	estimated saturated hydraulic conductivity	m/s
L	length of the soil column	m
L_e	actual path length or the distance traversed by an average parcel of water flowing through the soil pores	m
n	empirical constant	-
NAIL	non-aqueous phase liquid	-
O	organic matter content	%
q	volume flux or bulk velocity of flow rate	m/s
$q/\pi r^2$	the steady-state flow rate	m/s
r	radius of the ring	m

S	distance between two adjacent drains (tubes or ditches)	m
S	sand content	%
S	effective saturation	m
So	sorptivity	$m/s^{1/2}$
T	time of measurement	days
t	time	s
V_v	the change in fractional volume of voids	m^3
dW/dt	rate of total diminution of soil water content in the root zone	m/s
x	empirical constant	-
z	height above the water table	m
Zr	vertical depth coordinate (in this text taken up to the root zone)	m

Greek symbols

α	scaling factor	-
θ	volumetric water content	m^3/m^3
θ_n	initial volumetric moisture content (expressed as decimal fractions)	-
θ_o	volumetric moisture content at the supply potential (expressed as decimal fraction)	-
θ'	effective water content	m^3/m^3
ϱ_b	bulk density	Mg/m^3

ϱ_p	density of phase p	Mg/m^3
ϱ_w	density of water	Mg/m^3
ϱ^*	density of the phase used as the reference	Mg/m^3
Φ_e	effective porosity	m^2/m^2
η	viscosity of water	kg/ms
Ψ	suction head or potential of soil water	Pa
τ	soil matric suction	Pa

1. INTRODUCTION

Hydraulic conductivity (K) is a measure of a soil's ability to transmit water (Amoozegar and Warrick, 1986). It is defined by Darcy's law (Trout *et al.*, 1982) which states that, in an isotropic porous medium the specific flow rate is proportional to the negative head gradient and is expressed by the following equation:

$$q = - K \, \Delta h/\Delta l \qquad [1]$$

where:

q is volume flux or bulk velocity of flow (m/s)

K is proportionality constant termed the hydraulic conductivity (m/s)

$\Delta h/\Delta l$ is the summation of the driving forces in the direction of flow or hydraulic gradient (m/m)

Hydraulic conductivity depends on the size and distribution of pores in the soil, and therefore on bulk density and texture. It is generally assumed to remain constant for a given material and location, though clogging of pores by clay migration can alter it markedly in saturated flow experiments (Campbell, 1985; Paige and Hillel, 1993).

Hydraulic conductivity, being a soil property which determines the behaviour of soil water flow systems (Clothier and Smettem, 1990), has vital field applications for efficient management of resources and maintenance of the environmental quality. It is widely used in different situations ranging from solute transport, irrigation, drainage, water conservation, infiltration and runoff control, groundwater recharge and pollution, drain spacing design, as well as in crop modelling (Hillel, 1980b; Kinzelbach, 1986). Some examples of its use include:

(a) In considering the transport and reactions of chemicals in the soil, Jury and Fluhler (1992) showed that the equation of flow (flux) of a nonaqueous phase liquid (NAIL) in an unsaturated soil is given as:

$$J_p = -k_p K_{sp} (\delta h_p/\delta z + \varrho_p/\varrho^*) \quad \text{..........................[2]}$$

where: J_p is the flux of phase p (water, NAIL or air) (m/s)

k_p is the relative permeability of the medium to phase p (m/s)

K_{sp} is the saturated hydraulic conductivity of phase p in the medium (m/s)

ϱ_p is the density of phase p (Mg/m^3)

ϱ^* is the density of the phase used as the reference (Mg/m^3)

h_p is the fluid pressure head of phase p in the soil (m)

z is the height above the water table (m)

(b) In considering the available water for crop production, the hydraulic conductivity in relation to wetness can be applied to the analysis of drainage and evapotranspiration in actual field management. Hillel *et al.* (1972) showed that the actual rate of evapotranspiration (E_t) in a vegetated field can be obtained from the relation:

$$dE_t/dt = (dW/dt)_{zr} - (K\ \delta H/\delta z)_{zr} \qquad [3]$$

where: The first term on the right hand side of the equation is the rate of total diminution of soil water content in the root zone (to depth Z_r per unit time) (m/s)

The second term is the flow rate (flux) across the bottom of the root zone (being the product of the hydraulic conductivity by the hydraulic gradient operating across the Z_r plane) (m/s)

Z_r is the vertical depth coordinate here taken up to the root zone (m)

(c) In considering the evaporation from bare-surface soils, Hillel (1980b) described the steady-state upward flow of water from a water table through the soil profile to an evaporation zone at the soil surface by the following equations:

$$q = K(\Psi)(d\Psi/dz-1) \qquad [4]$$

$$\text{or } q = D(\theta)d\theta/dz - K(\Psi) \qquad [5]$$

where: Ψ is suction head (Pa)

D is hydraulic diffusivity (m^2/s)

θ is volumetric wetness (m^3/m^3)

From these examples, it is evident that hydraulic conductivity is of considerable importance in many aspects of agricultural and urban life.

A considerable number of methods have been developed to determine K both in the laboratory and in the field. Such methods include: the constant head method, falling head method, auger hole method, piezometer method, unsteady drainage-flux method, crust-imposed steady flux method (Amoozegar and Warrick, 1986; Green *et al.*, 1986; Klute and Darken, 1986) and the disc permeameter method (White *et al.*, 1992). Although these field methods determine hydraulic conductivity in the real condition, they have one big disadvantage in that they disturb the soil and cause hydrological discontinuities. This is very important if the site is to be used over a long period of time. When a soil sample is taken from a field it is no longer a representative of the site since the site is now disturbed. When the site is disturbed, pore connectivity changes, resulting into a reduction/removal of overburden pressures. Additionally, field methods have been found to be laborious, tedious, time-consuming, expensive, and restricted to a limited measurement range (Libardi *et al.*, 1980; Ragab *et al.*, 1981; Mualem, 1986; Vereecken *et al.*, 1990).

In order to avoid some or all of the limitations mentioned above, researchers have attempted to develop mathematical models (called *Pedotransfer* functions) which assist in estimating hydraulic conductivity functions from easily measurable soil physical properties such as texture, bulk density and organic carbon or organic matter content (McCuen *et al.*, 1981; Alexander and Skaggs, 1987; Vereecken *et al.*, 1990).

Vereecken (1988) described in detail 16 models. However, not all the 16 models could be applied in this research. Three of the models, namely the Mualem model (Mualem, 1976), the Burdine model (Burdine, 1953) and the Gardner model (Gardner, 1958) whose merits and limitations have been explained in detail in section 2.3, were selected and compared mainly because of their simplicity. Finally the Gardner model (Gardner, 1958) because of (a) its simplicity and flexibility in describing the measured data and (b) hydraulic conductivity does not become infinite when matric potential becomes zero, was adopted in this work.

These models despite being used in many parts of the world (Schuh and Sweeney, 1986; Michiels *et al.*, 1989; Vereecken *et al.*, 1989; Paige and Hillel, 1993), have so far not been used in Tanzania because they have not been calibrated for Tanzanian soils. However, limited research on hydrological properties (infiltration rate, bulk density and moisture retention characteristics) has been done in studies associated with soil erosion and the effect of different cultivation techniques on soil moisture conservation by Ngatunga (1981) and MaCartney *et al.* (1971), respectively.

Therefore, the objectives of this study were to calibrate and to validate with field data, the Gardner Model (Gardner, 1958) for estimating K, for a sandy clay loam soil in semi-arid localities. This will facilitate the use of the model to estimate K values at different periods and assist soil physics and agronomy research conducted in these localities.

2. LITERATURE REVIEW

2.1 Factors affecting hydraulic conductivity

The factors which affect the hydraulic conductivity of a soil are those related to the nature of the soil, such as soil structure, soil texture (Table 1), and the soil water content (Hillel, 1980a; Campbell, 1985; Ghildyal and Tripathi, 1987). Coarse textured soils having a high porosity have a corresponding high hydraulic conductivity while the heavy clays which have finer pores have low hydraulic conductivity. Generally, the hydraulic conductivity of light to heavy textured soils ranges from 10^{-1} to 10^{-6} mm/s (10^{-2} to 10^{-7} cm/s) (Ghildyal and Tripathi, 1987).

Table 1: Hydraulic conductivity of different Indian soil types

Texture	Hydraulic conductivity (mm/day)
Patharchatta sandy loam	234.6
Haldi loam	129.0
Beni silty clay loam	88.1
Phoolbagh clay loam	53.0

Source: Ghildyal and Tripathi (1987).

Where: Patharchatta, Haldi, Beni and Phoolbagh are Indian local names of the different places where the hydraulic conductivity was measured.

Hydraulic conductivity primarily depends on the size, distribution and tortuosity (i.e. the geometry) of the conducting pores. Anything that alters soil structure, such as compaction, alters hydraulic conductivity. Compaction increases the soil bulk density and changes the volume, size and shape of the conducting pores, exerting an effect of smaller magnitude which in turn results into a drastic reduction of the soil hydraulic conductivity (Kayombo and Lal, 1993). For example, Hillel (1980a) defined the equation of continuity according to Darcian's outflow of water in a compacted soil as follows:

$$\delta V_v/\delta t = - K\ \delta^2 h/\delta z^2 \quad \text{[6]}$$

where: V_v is the change in fractional volume of voids (m^3)

t is time (s)

h is pressure head of soil water (m)

From this equation, the hydraulic conductivity (K) is seen to be greater when the soil is highly porous, fractured, or aggregated than when it is tightly packed and dense.

Generally, hydraulic conductivity varies logarithmically with soil water content. Hence, it follows a log normal distribution. Because of the above reason, hydraulic conductivity ranges over many orders of magnitude between soil saturation and soil dryness. The high reduction in unsaturated conductivity with water content is caused by two main factors as discussed below:

(a) During drying, the soil pores are filled with air which makes them no longer effective channels for the flow of water. As a result the effective porosity is reduced (Ghildyal and Tripathi, 1987). In addition, as the soil desaturates, the path of flow becomes much more tortuous (Hillel, 1980a). Both tortuosity and the reduction in effective porosity are accompanied by a lowering of the hydraulic conductivity of a soil. The relationship of both tortuosity and reduction in effective porosity with hydraulic conductivity, K, can be explained by the following equation:

$$K = \varrho_w g/k'\eta A^2 \ (L/L_e)^2 \ (\Phi_e/(1-\Phi_e)^2) \quad \text{[7]}$$

where: ϱ_w is density of water (Mg/m^3)

g is acceleration due to gravity (m/s^2)

k' is pore shape factor ranging from 2 to 2.5

η is viscosity of water kg/ms

A is specific surface area (m^2/m^3)

L is length of the soil column (m)

L_e is actual path length or the distance traversed by an average parcel of water flowing through the soil pores (m)

Φ_e is effective porosity or area of conducting channel per unit area of the cross-section (m^2/m^2)

(b) The soil matric suction increases with drying which empties the water content of the macro-pores. This greatly reduces the effective cross-sectional area of the flow. In this case, the water has to move through micropores which conduct water much less readily than the macropores and, consequently, the conductivity is reduced (Ghildyal and Tripathi, 1987). This can be explained by the following equation:

$$q = - K(\tau) \nabla H \qquad [8]$$

where: τ is soil matric suction (Pa)

∇H is hydraulic gradient

Hence, as the soil matric suction increases, the soil hydraulic conductivity decreases, and vice versa.

Another important factor which affects hydraulic conductivity is spatial variability of the soil hydraulic conductivity (Warrick and Nielsen, 1980; Ahuja *et al.*, 1984; 1989). The hydraulic conductivity values obtained from two locations only a few metres apart can be quite different due to soil heterogeneity (Table 2).

Table 2: Spatial distribution of K_{sat} from effective porosity (Φ_e) measurements in different soils

Soil	Soil type	K_{sat} (in terms of scaling factors (α))
Cecil soil	clayey soil	1.925
Lake-land soil	sandy soil	1.076
Norfolk soil	fine-loamy	1.560
Pima soil		1.545
Renfrow soil		1.833
Hawaii soil		1.586

After: Ahuja *et al.*, 1984, 1989.

Hydraulic conductivity also varies with time. When the soil is ponded with water for some time, there is an increase in hydraulic conductivity. Vereecken (1988) reported increased saturated hydraulic conductivity at every measured time interval in an area that had been ponded for some time. (Table 3)

Table 3: Temporal distribution of K_{sat} in one of the Belgian soils

Soil layer	Soil type	K_{sat} (mm/day)		
		T_1	T_2	T_3
Plough layer	clayey silt loam	715	3041	3321

After: Vereecken, 1988.

Where: T_1 is the initial time of measurement

T_2 is T_1 + 3days

T_3 is T_1 + 6 days

From this information, an enormous amount of data is required to accurately represent the hydraulic conductivity of a soil. Moreover, a large number of measurements are inevitable in characterizing the variability and distributions of hydraulic conductivity (Warrick and Nielsen, 1980). These two requirements make hydraulic conductivity measurements both laborious and time consuming.

2.2 Direct field methods for determining K

The most obvious way of determining K is by taking *in situ* measurements. The most frequently used methods include: the Unsteady Drainage-Flux method (also referred to as the Instantenous profile method), the Guelph Permeameter method, the Auger hole method, and the Disc Permeameter method (Watson, 1966; Amoozegar and Warrick, 1986; Green *et al.*, 1986; Klute and Darken, 1986; White *et al.*, 1992).

Among the four methods, the first three methods can be used to measure K at any desired depth down the soil profile with a minimum disturbance to the natural profile. The disc permeameter method is best only for surface measurements. If measurements at depth are to be made using this method, a soil pit must be dug. This will cause a major site disturbances.

2.2.1 The disc permeameter method

The disc permeameter (Perroux and White, 1988) is a relatively new instrument for measuring *in situ* hydraulic properties of soils. It enables rapid measurements of hydraulic conductivity, sorptivity, macroscopic capillary length and characteristic pore size with minimal soil disturbance. It can also be used to quantitatively assess the contribution of preferential flow paths, such as biopores, to field infiltration. Its principal application is in measurements on surface soils, particularly in relation to soil management and land degradation studies (White and Sully, 1987; White and Perroux, 1987, 1989; Commonwealth Scientific and Industrial Research Organization [CSIRO], 1988; Smettem and Clothier, 1989; Ankeny *et al.*, 1991).

In the study conducted by Vauclin and Chopart (1992) to characterise the hydrodynamics of the surface layers of a gravel soil subjected to three different types of cropping profile management (direct sowing of maize and cotton, ploughing before each crop and recently ploughed fallow) in Ivory Coast using the Multi-disc infiltrometry with controlled suction, this method was reported to be simple and could be easily employed in hydrological studies in drainage basins or agrophysical studies in rural areas. Also its portability and ease and speed of installation in the field enabled multiple repetitions to be carried out at very reasonable cost in order to determine the variability of the environment within the area studied. From the study conducted by Cook and Broeren (1994) to measure the flow rate at different potentials on a Kokotau silt loam soil at Waingawa in New Zealand, disc permeameter method was observed to offer a rapid means of measuring the surface hydraulic properties of a soil with minimum site disturbances.

The method for determining soil hydraulic properties in the field using the disc permeameter is given by White *et al.* (1992). This method is mainly used for taking measurements on the soil surface. The disc permeameter method is based on the Wooding (1968) method of analysis of three-dimensional flow from a shallow circular pond or surface disc (Figure 1). The principle behind this method is based on monitoring the infiltration/flow rate of water from the disc into the soil.

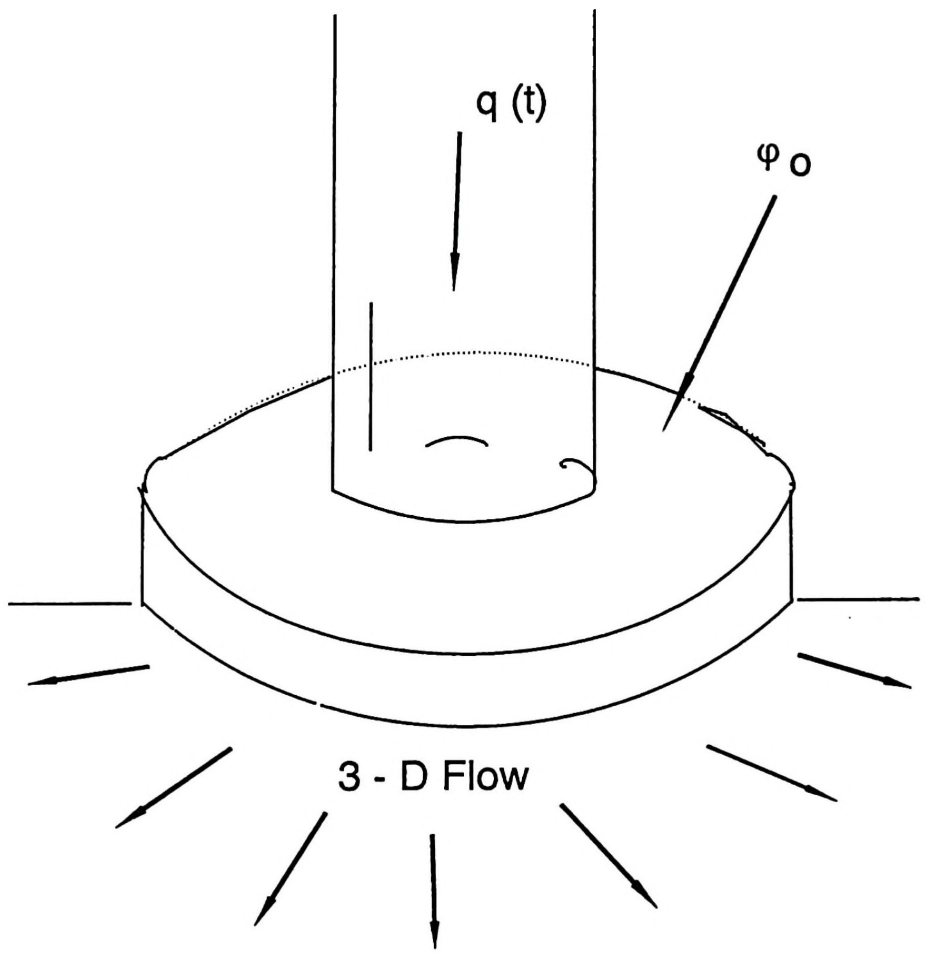


Figure 1 Three dimensional flow from a shallow pond or surface disc. After CSIRO (1988)

To determine the hydraulic conductivity of the soil by this method, the following measurements are required: the sorptivity of the soil, the steady-state flow rate, the initial volumetric moisture content and the volumetric moisture content at the supply potential.

The procedure of taking measurements in the field using this method is described in detail in Chapter 3, Section 3.3.

Even though these methods are most frequently used to measure *in situ* K or hydraulic characteristic, they are time-consuming, expensive, difficult and laborious (Libardi *et al.*, 1980; Ragab *et al.*, 1981; Mualem, 1986; Vereecken *et al.*, 1990). Mathematical models (*pedotransfer* functions) which assist in estimating hydraulic conductivity functions from easily measured soil physical properties like soil texture, soil bulk density, soil organic matter or soil organic carbon, can be employed to avoid some or all of these limitations.

2.3 Different mathematical models for estimating K

Several investigators have proposed empirical models for estimating hydraulic conductivity functions from easily measured soil physical properties. Vereecken (1988) did a detailed analysis of about 16 different models. Examples of three of them are explained below:

1. The **Mualem Model** (Mualem, 1976):

$$K_r = S_e^x \left| \int_0^{\theta'} d\theta'/h_m^{1+b} / \int_0^{\theta's} d\theta'/h_m^{1+b} \right|^2 \quad \text{[9]}$$

where: K_r is residual hydraulic conductivity (m/s)

S_e is effective saturation (m)

θ' is effective water content (m^3/m^3)

h_m is matric potential (Pa)

$x = 0,\ b = 0.$

2. The **Burdine Model** (Burdine, 1953):

$$K_r = S_e^x \int_0^{\theta'} d\theta'/h_m^{2+b} / \int_0^{\theta's} d\theta'/h_m^{2+b} \quad \text{[10]}$$

These models take account of the varying pore sizes occurring in a porous medium, and the hydraulic conductivity is determined by integration of elementary pore domains, represented by a pore radius. Integrating over these pore domains gives a mean pore size, resulting in a uniform pore size model.

Advantage:

The knowledge of the pore size distribution, which is essential in these models, can be obtained from the soil water characteristic curve, by making use of the capillary flow over a given matric potential.

Disadvantages:

(a) These models are too strict (on pore size distribution) in describing the measured data,

(b) They have complicated mathematical formulations which cannot be evaluated easily for different values of the matric potential,

(c) When matric potential (h_m) becomes zero, hydraulic conductivity becomes infinite.

3. The **Gardner Model** (Gardner, 1958):

$$K = K_{sat}/(1 + (bh_m)^n) \quad \text{..............................} [11]$$

where: K is unsaturated hydraulic conductivity (m/s)

K_{sat} is saturated hydraulic conductivity (m/s)

b and n are soil dependent parameters.

This is an empirical power function model with three parameters (K_{sat}, b, n). It gives steady state analytical solutions for the unsaturated flow equations (Gardner, 1958).

Vereecken (1988) listed the following as the most important advantages of this model:

(a) It is the most flexible model in describing measured data.

(b) Hydraulic conductivity (K) does not become infinite when matric potential (h_m) becomes zero.

(c) Gardner (1958) determined the values of coefficients (parameters) of certain formulas for several soil classes. Therefore, those values may serve as a guideline for first approximation of hydraulic conductivity when no data are available except the soil classification.

(d) Compared to theoretical models, the Gardner model has a simple mathematical formulation in a closed form which can be evaluated very easily for different values of matric potential.

(e) The moisture retention characteristic is not needed as in the case of the theoretical models. With regard to mathematical structure, it has the mathematical form of the best model for the moisture retention characteristic.

(f) It has been extensively used by other researchers (e.g. Vereecken, 1988; Vereecken *et al.*, 1989, 1990).

The main disadvantage is that, most of these models are site specific since the parameters vary with soil type. For example, Table 4 shows different values of b and n in Eq. [12] for different soil types as obtained by Gardner (1958) and Vereecken (1988). It is, therefore, important to calibrate these models for each site of interest.

Table 4: Values of b and n (Eq.[11]) for different soil types

Soil type	b	n
Clayey silt loam*	0.17	1.9
Sandy silt loam*	0.36	1.5
Sandy soil*	0.1	2.04
Fine sandy loam**	1.0	3.0
Loamy soil*	7.25	1.5
Loamy sand*	0.01	2.48

After: Gardner (1958)** and Vereecken (1988)*

2.4 Explanation of the Gardner (1958) model parameters

According to Vereecken (1988), the parameters of the Gardner (1958) Model can be obtained from easily measured soil physical properties such as soil texture, soil organic matter content and bulk density.

From the Gardner equation (Eq.[11]), the regression equations for the parameters are as follows:

$$\log (K_{sat}, e) = A - B \log (Cl) - C \log (S)$$

$$- D \log (O) - E (\varrho_b) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots [12]$$

where: K_{sat}, e is estimated saturated hydraulic conductivity

Cl is clay content in % (< 2 μm),

S is sand content in % (> 50 μm),

O is organic matter content (%),

ϱ_b is bulk density (Mg/m^3),

A, B, C, D and E are constants.

$$\log (b) = F - G (S) + H (Cl) \quad \ldots\ldots\ldots\ldots\ldots\ldots [13]$$

where: b is soil dependent parameter,

F, G and H are constants.

$$\log (n) = I - J \log (Cl) - K \log (Si) \quad \ldots\ldots\ldots\ldots [14]$$

where: n is a soil dependent parameter,

Si is silt content in % (2 - 50 μm),

I, J and K are constants.

Therefore:

(a) K_{sat} is dependent on soil texture, soil organic matter content and soil bulk density.

(b) b and n are dependent on soil texture.

(c) K is related to K_{sat} and, therefore, is dependent on soil texture, soil organic matter content, and soil bulk density.

The Gardner (1958) model, due to (a) its simplicity and flexibility in describing the measured data, (b) that hydraulic conductivity does not become infinite when matric potential becomes zero, and (c) its advantage of obtaining its parameters from easily measured soil physical properties like soil texture, was adopted in estimating K during the current study.

3. MATERIALS AND METHODS

3.1 Description of the experimental area

3.1.1 Location

The research was conducted on the research farm of Hombolo Agricultural Research Station in Dodoma. The farm is located at about 58km North-East of Dodoma municipality at $5^0 53'$S Latitude; $35^0 55'$E Longitude and 1037m above mean sea level. It lies on the middle of a long uniform slope of about 2% (Dregne, 1990; Mahoo and Kaaya, 1993). Figures 2 and 3, respectively, indicate a schematic map of Dodoma Rural and Urban districts with some rainfall isohyets, and the Hombolo experimental site.

3.1.2 Climate

The area experiences a semi-arid type of climate. Temperatures show seasonal variations although the transition is not sharply marked. The months of June to August are the coolest. The coldest month is July with an average temperature of 19.6^0C and the hottest month is November with 24.8^0C. The average temperature for the year is 22.7^0C. For approximately 40 days in each year the mid-day temperatures exceed 32^0C.

Annual rainfall records from 1974 to 1991 show that rainfall at Hombolo varies from 350 mm to 814 mm with a mean of 588 mm (Ngana, 1993). The rainy season in this area is usually from November to April and the heaviest rains occur between December and March. From May to October the area is extremely dry. Total rainfall from December, 1993 to March, 1994, during the period when the experiment was in progress, was 582.6 mm (Table 5).

Table 5: Monthly rainfall distribution at Hombolo**

Month	Rainfall (mm)
December	2.9
January	293.7
February	137.4
March	150.1
April	20.8
May	3.2

** From December, 1993 to May, 1994.

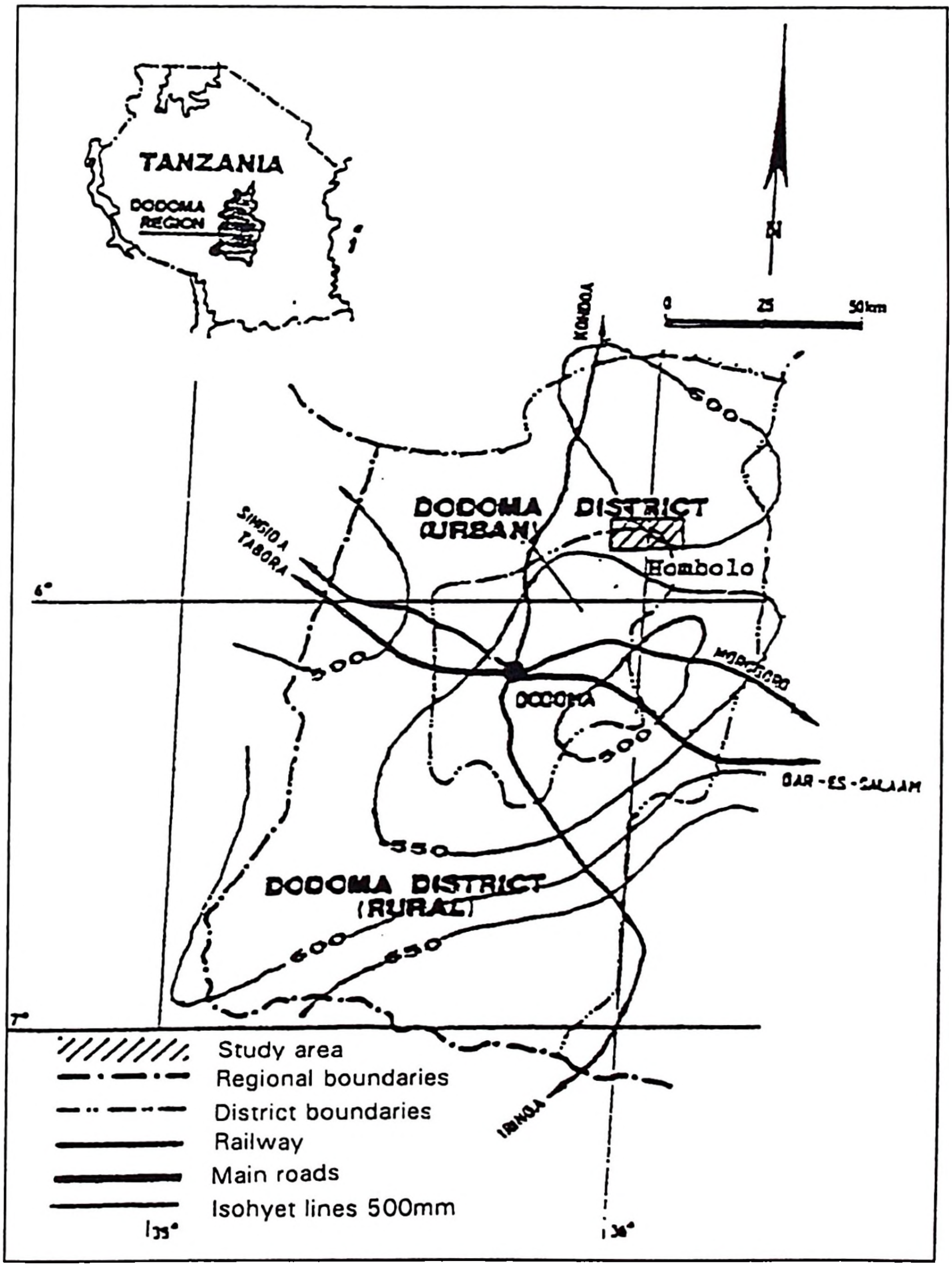


Figure 2. Dodoma Rural and Urban district.
Source: AGRAR - UND HYDROTECHNIK GMBH (1984)

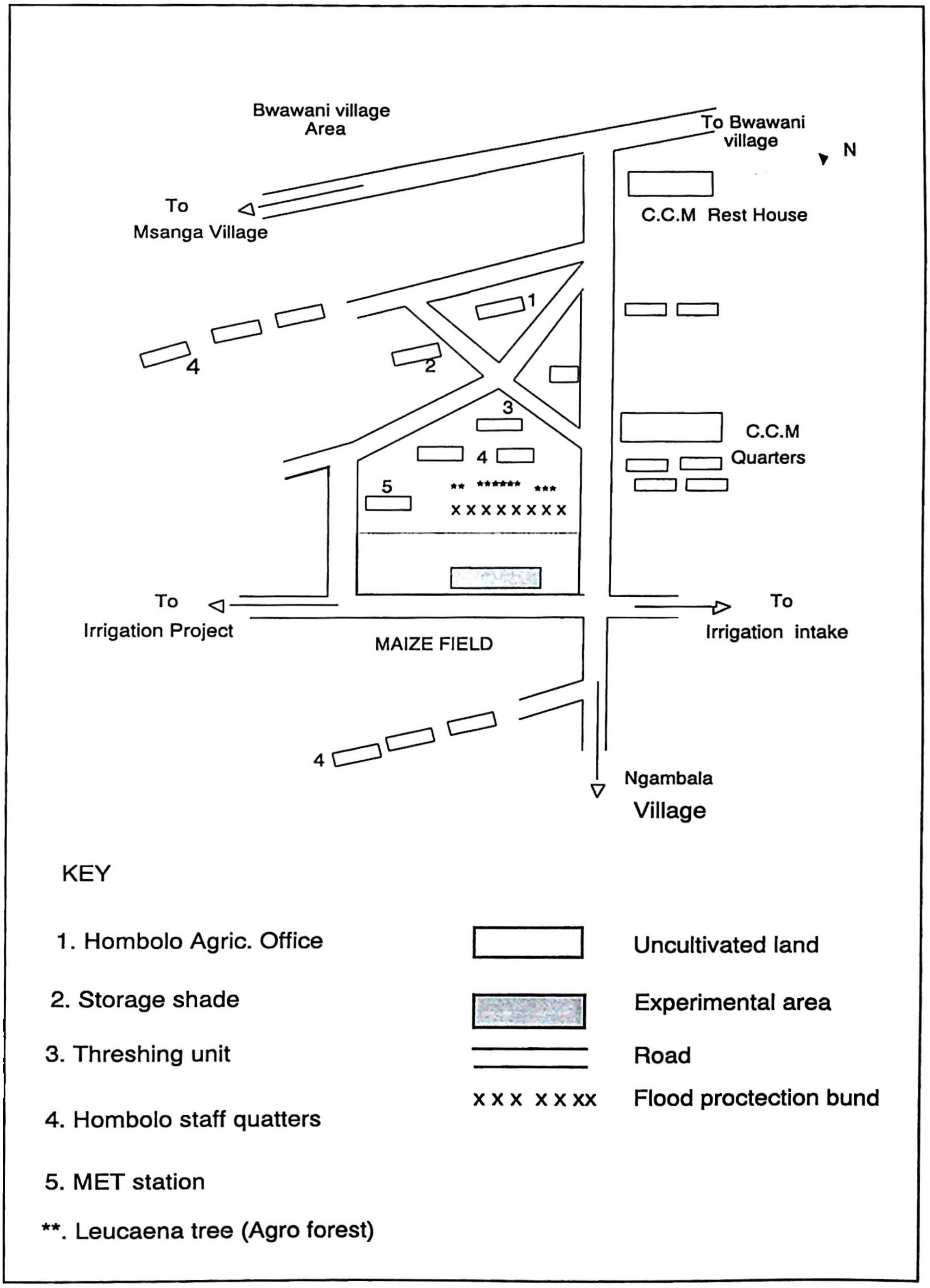


Figure 3. Location of the Experimental site. Adapted from Kayombo (1993)

The relative humidity is low. Early morning values may be in the range of 84 to 93% but in the afternoons strong winds in combination with frequent bright sunshine bring relative humidity to very low values. The average midday minimum of 34% is reached in October. During the rainy season dew occurs. The mean annual dew point lies between 16°C and 17°C (Griffiths, 1972).

Potential evaporation is high. When the driest period sets in and the force of wind increases considerably, evaporation also increases and reaches its highest monthly average in October.

3.1.3 Soils

From the detailed survey and analysis of the soils of the experimental area (Mahoo and Kaaya, 1993), the soils are classified as Typic Ustorthent in the US Soil Taxonomy and as Dystric Regosol in the FAO-UNESCO system. The profile is fairly deep (>100 cm) with soil texture ranging from sandy loam on the surface to a sandy clay loam subsoil. The structure of the surface horizon is weakly developed showing a marked evidence of sheet erosion on the surface. The profile is characterized by an ochric epipedon. The sand fraction of nearly the whole profile is dominated by quartz minerals. The parent material of the soil is Si-rich gneiss with granite. The moisture and temperature regimes of the soil are ustic and thermic, respectively. The profile is also characterized by a light brown surface horizon which changes to strong brown, reddish yellow and light brown with depth. The fertility of the soils of the experimental area as shown by the levels of organic matter content, total N, available P and Cation Exchange Capacity (CEC) of these soils is generally very low. Organic matter content ranged from 0.76 to 1.06%, total N ranged from 0.04 to 0.06%, available P ranged from 5.5 to 5.9 mg/kg and CEC ranged from 12.0 to 16.4 cmol(+)/kg. The levels of Mg^{2+}, Ca^{2+} and K^{+} (about 0.9, 5.4 and 0.8 cmol(+)/kg) are rated as low, medium and high, respectively. Soil pH on average is 5.9. This is within the suitable range for availability of most plant nutrients (Brady, 1984).

3.1.4 Vegetation

The experimental plots were located on a piece of land which had been under grass fallow for the past 6 years. The native vegetation in the surrounding area consisted of scattered *Adansonia digitata* (baobab tree) and *Hyperrhenia-Acacia* bush with grass land. When the plot was left for fallowing, the indigenous grass types such as *Trichodesma zeylanicum* (Late weed), *Borhavia diffusa* (Tar vine), *Chrolis gayana* (Rhodes grass) and *Cydon dactylon* (Star grass) were allowed to recolonise the land. However, all the time prior to fallowing, the land was under intensive cultivation, used for growing cowpeas, sorghum, sunflower, groundnuts and pearl millet. Sometimes these crops were intercropped or grown in rotation.

3.2 Soil sampling and analysis

Soil sampling was done at random from each experimental plot. The samples were analysed to determine some important physico-chemical properties (see Table 6 section 4.1).

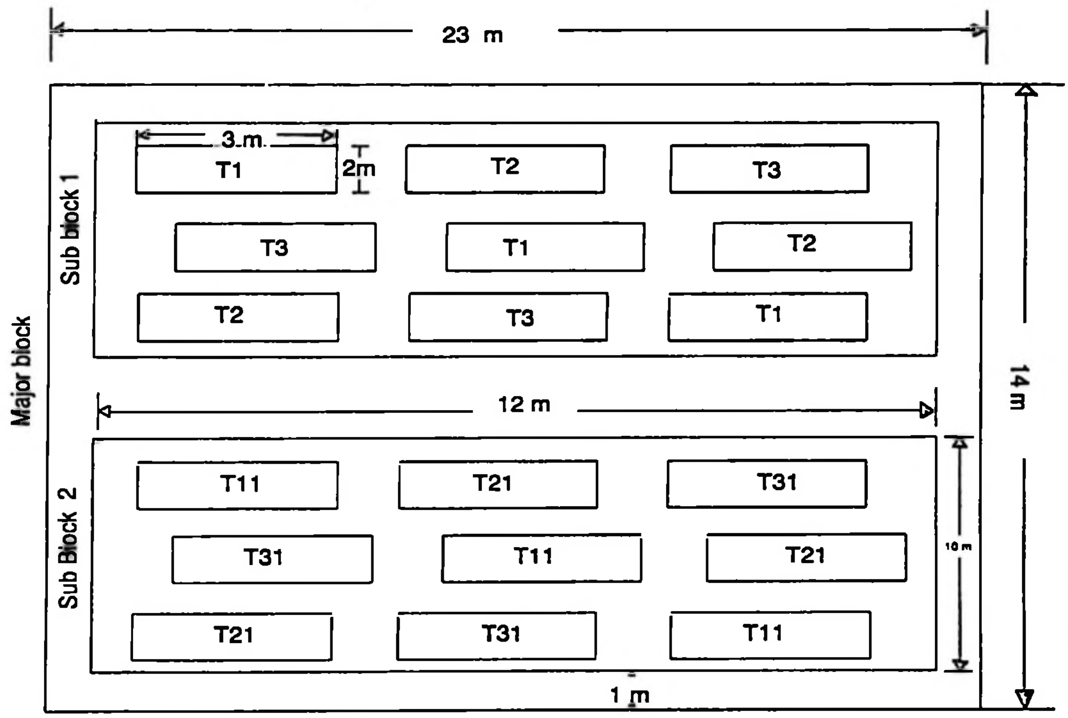
3.3 Measurement of hydraulic conductivity

3.3.1 Experimental Layout

The layout comprised one major block of 23 x 14 m with two sub-blocks of 12 x 10 m each were demarcated (Figure 4).

Sub-block 1 with treatments T1, T2 and T3 was used for the determination of K at different suctions (0.0, 0.2, 0.5, 0.8 and 1.0 kPa) for the calibration of the Gardner model for estimating K for sandy clay loam soil in semi-arid Dodoma, Tanzania.

Sub-block 2 with treatments T11, T21 and T31 was used for the determination of K at different bulk densities (ranging between 1.3 - 2.1 Mg/m^3) for the validation of the model.



KEY

T1 = Ploughed plot   Bulk Density = 1.43 Mg/m^3

T2 = Undisturbed plot   Bulk density = 1.64 Mg/m^3

T3 = Compacted plot   Bull density = 1.85 Mg/m^3

T11, T21, T31 = Ploughed, Undisturbed and compacted plots, resectively; D_b ranging between 1.3 to 2.1 Mg/m^3

Figure 4. Experimental layout

3.3.2 Treatments

The treatments were randomly assigned to different plots (Figure 4) based on varying bulk densities. Three plots of 3x2 m each were taken to represent three field conditions with different soil bulk densities. Treatment 1 represented the newly ploughed field, Treatment 2 the moderately compacted field and Treatment 3 the highly compacted field. Every treatment was replicated three times.

Treatment 1 (T1): cultivated down to 10cm depth to a bulk density of 1.43 Mg/m^3. Loosening of the soil was achieved by ploughing, scooping off of the 10cm surface soil and then putting it back as loose soil (Plate 1).

Treatment 2 (T2): This treatment was imposed on an undisturbed soil with a bulk density of 1.64 Mg/m^3.

Treatment 3 (T3): Soil compacted to a bulk density of 1.85 Mg/m^3. Compaction was achieved by hammering the loosened 10cm surface soil (Plate 2) using a 15.5 kg hammer (Plate 3) dropped from 55 cm height, as described by Hillel (1980a). About seven hammerings were performed per plot to attain the desired bulk density.

Treatments T11, T21 and T31 were, respectively, ploughed, undisturbed and compacted plots with bulk densities ranging between 1.3 - 2.1 Mg/m^3.

Plate 1 Scooped soil to be put back as loose soil

Plate 2 Compaction process

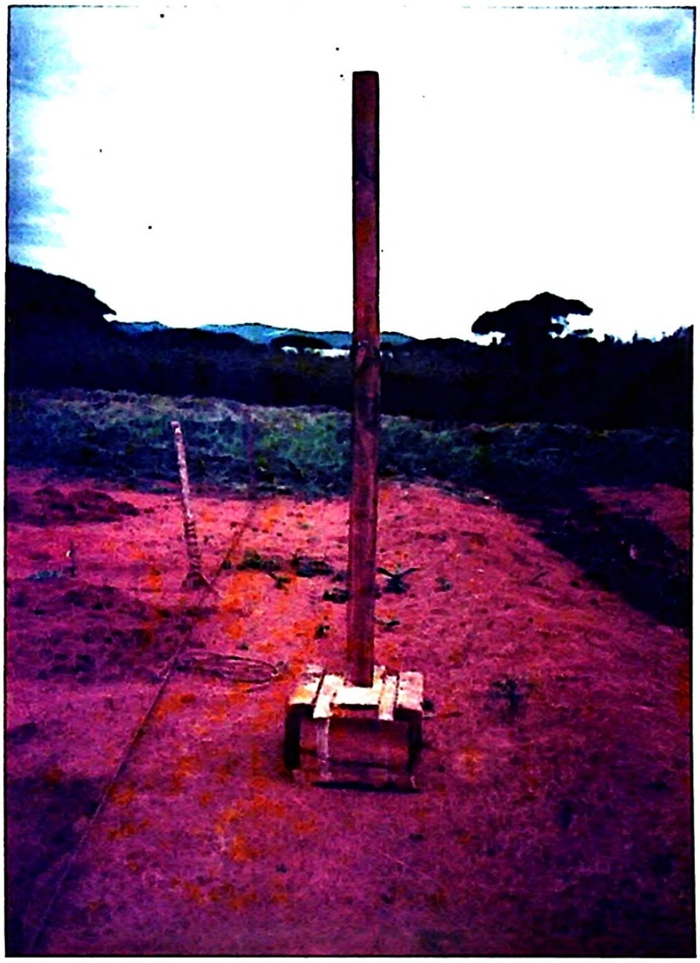


Plate 3 The hammering device

3.3.3 Regression equation for K_{sat} for sandy clay loam soils of semi-arid Dodoma

The regression equation, (Eq.[12]), for the parameter K_{sat} of the Gardner Model (Gardner, 1958) for sandy clay loam soils of semi-arid Dodoma was re-arranged as follows:

$$\log (K_{sat}, e) = L - E \varrho_b \qquad [15]$$

where: K_{sat}, e is estimated saturated hydraulic conductivity

$L = A - B \log (Cl) - C \log (S) - D \log (O)$

A, B, C, D and E are constants

Cl is clay content in % (< 2 μm)

S is sand content in % (> 50 μm)

O is organic matter content (%)

ϱ_b is bulk density (Mg/m^3)

Hence, from Eqs. [11 and 15]:

1. b and n were obtained directly from Eq. [11] by solving the simultaneous equations obtained by measuring K at different matric potentials.
2. K_{sat} was obtained from a graph of log (K_{sat}) against Bulk density [Eq.15].

3.3.4 Measurements

Hydraulic conductivity (K) was measured at different suctions for the calibration of the Gardner model for estimating K (on sub-block 1). K values were also measured at different bulk densities for the validation of the model (on sub-block 2). Hence, measurements taken were:

(a) at five different levels of matric suction (0.0, 0.2, 0.5, 0.8 and 1.0 kPa) for every treatment for the measurement of K at different suctions. At every potential, full measurements were made for every treatment and its replicates. So there were five sets of measurements in this section.

(b) at different levels of bulk density ranging between 1.3 and 2.1 Mg/m^3 for the measurement of K at different bulk densities. Measurements were made on plots which had similar soil characteristics but initially were not being used in the measurements (a) (sub-block 2). The measurements were made 18 times at different bulk densities and at different suctions (0.2, 0.5, 0.8 and 1.0 kPa) within the same range of bulk density (Appendix 3).

3.3.5 Principle of measurement

The disc permeameter, which is a surface technique for measuring hydraulic conductivity developed by CSIRO (1988) was employed. The principle behind this technique is that the infiltration/flow rate of water from the disc into the soil is measured. The design of the permeameter used to make measurements in this study is shown in Figure 5. The potential is set by adding or withdrawing water from the bubbling tower using a plastic syringe with a length of thin tubing connected to it. Principally, the initial potential set in the bubbling tower is not equal to the soil matric potential of the soil in which the measurement is taken. However, when the infiltration attains steady state or constant rate, the potential set equals the soil matric potential of that area.

The difference which exists between the two different initial suctions, is taken care of by subtraction of the volumetric moisture content at the initial potential from the volumetric moisture content at the supply potential.

The water reservoir is filled by placing a vacuum in the one way valve or stopcock at the top of the reservoir.

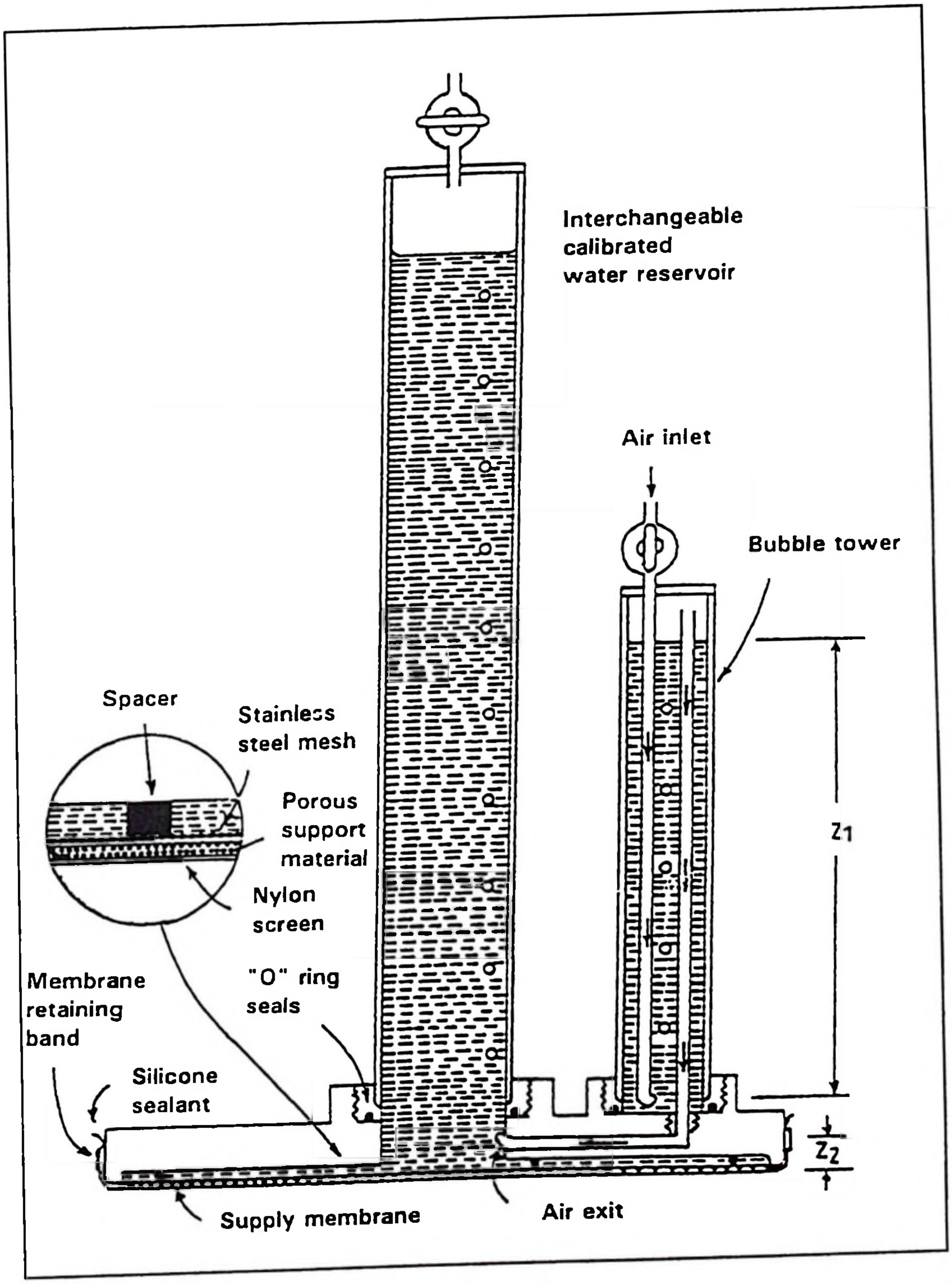


Figure 5 The disc permeameter for unsaturated measurement. After CSIRO (1988)

3.3.6 Field measurements

On the prepared surfaces of the plots, of every treatment and every replicate for every measurement the disc permeameters were set to make measurements. Before setting the permeameters, the water was filled in the reservoirs and the potential was set in the bubbling towers. Two core samples were taken approximately 25 cm from the centre of the disc measurement surface. These samples gave the initial soil moisture content (at the supply potential) which was determined by the weight loss from oven drying at 105°C for 24 h.

Rings of the same radius as the discs and 5 mm in height were placed on the soil surfaces. Wet fine sand was placed in these rings and levelled. The sand was necessary to provide good contact between the disc permeameter and the soil. Then the discs were placed on these beds of sand and cumulative infiltration was monitored until the flow rates were constant (Plate 4).

Plate 4 Monitoring the infiltration rate

Immediately after measurement of the steady-state flow rate (which was attained after 1 h, 3 h and after 6 h depending on the potential set in the bubbling tower), the discs were removed from the soil surfaces, the sand caps were quickly scraped aside, and soil moisture content samples (0.2 - 0.3 cm thick) were taken.

The readings or measurements obtained were used in the equation which was described by Wooding (1968) to compute the hydraulic conductivity (K) values. The Wooding (1968) equation is as follows:

$$K = q/\pi r^2 - 4bS^2/\pi r(\theta_o - \theta_n) \quad [16]$$

where: $q/\pi r^2$ is the steady-state flow rate (m/s)

r is the radius of the ring (m)

b is a constant

S_o is sorptivity (m/s^{½})

θ_o is the volumetric moisture content at the measurement/supply potential

θ_n is the initial volumetric moisture content

(θ_o and θ_n are expressed as decimal fractions).

3.4 Data analysis

Statistical analysis was performed following simple regression-correlation analysis procedures as described by Snedecor and Cochran (1989).

4. RESULTS AND DISCUSSION

4.1 Soil analytical data

Some important soil physico-chemical properties are presented in Table 6. The soil is a coarse-textured sandy clay loam with bulk densities ranging between 1.20 and 1.80 Mg/m^3. Brady (1984) reported a similar range of bulk densities for the soils of North America which had similar physical properties as those studied at Hombolo. Because of the large size of the spaces between the particles, the passage of water and air is rapid. Therefore, they possess good drainage and aeration. Brady (1984) noted that in North America soils with similar physical and hydrological properties were prone to drought. Coarse-textured soils generally have low organic matter content. This property has the tendency of reducing the capacity of the soil to retain nutrients against leaching. This in turn leads to poor soil fertility. The soil had a pH value of 5.9 (Table 6), indicating that the soil is acidic. This acidity in combination with the observed low levels of organic carbon, total N, total exchangeable bases and available P, indicates that the soil of the experimental plots is generally poor in fertility. These results conform to the results obtained by Mahoo and Kaaya (1993) at the same location.

Table 6: Some physico-chemical properties of the top soil (0-25cm) at the experimental plots

Property	Magnitude
Particle size	
Sand (%)	70
Silt "	4
Clay "	26
Texture	SCL*
pH in water	5.9
in KCl	4.4
Organic carbon (%)	0.33
Total N (%)	0.04
Available P (mg/kg)	3.00
Exchangeable bases	
Na (me/100g)	2.82
K "	1.92
Ca "	2.19
Mg "	2.17
Effective CEC (me/100g)	9.10

* SCL = Sandy clay loam

4.2 Hydraulic conductivity results for model calibration

4.2.1 Plot of K versus matric suction

The plot of K versus suction for each bulk density is presented in Figure 6. A bulk density of 1.43 Mg/m^3 was taken to represent a newly ploughed field, while bulk density of 1.64 and 1.85 Mg/m^3 represent moderately compacted and highly compacted conditions, respectively.

General observation from the graph shows that, as both bulk density and suction increase, hydraulic conductivity decreases. Hydraulic conductivity is high in soil with low bulk density (1.43 Mg/m^3) as compared to soils with high bulk densities (1.64 Mg/m^3 and 1.85 Mg/m^3) and especially at low suctions (at saturation). This means that ploughing has positive effects on soil bulk density and hence on soil hydraulic conductivity. Ploughing changes the hydrophysical behaviour of the soil as it loosens it making it more porous leading to the development of a network of macropores which in turn leads to increased water flow (Warkentin, 1971; Vauclin and Chopart, 1992).

Use of heavy machinery tends to cause soil compaction. Tightly compacted and dense soils have higher soil bulk densities than highly porous, fractured or aggregated ones (Warkentin, 1971; Kayombo and Lal, 1993). Increases in soil bulk density due to soil compaction result from decreases

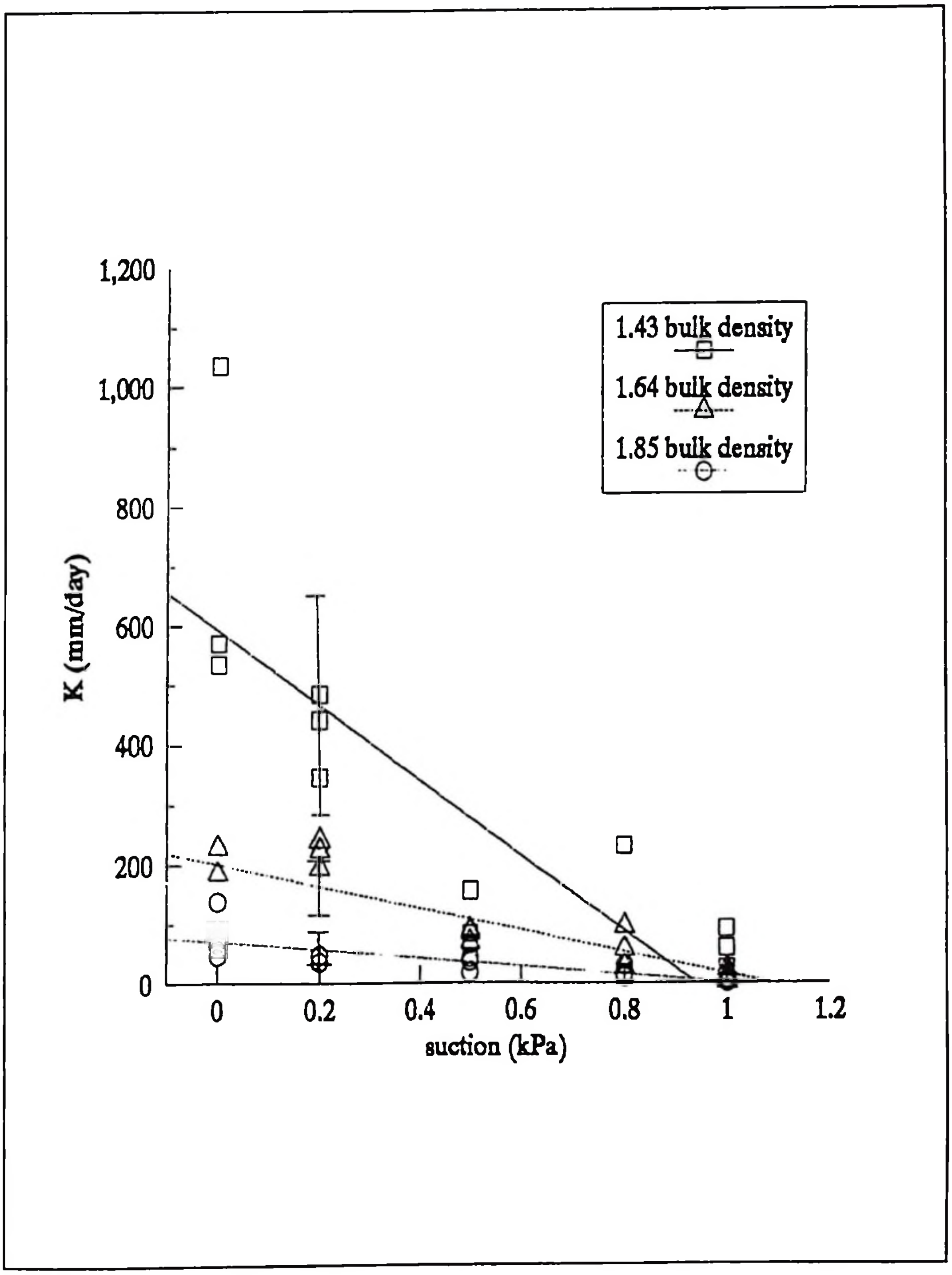


Figure 6 K (mm/day) versus Suction (kPa)

in the volume, size and shape of the soil conducting pores which in turn results in the reduction of the soil macroporosity. Decreases in soil macroporosity are accompanied by high tortuosity of water flow. Both a decrease in soil structural macroporosity as well as high tortuosity of water flow result in the reduction of soil water infiltration rate (Hillel, 1980a; Kayombo and Lal, 1993) which in turn results into the reduction of soil hydraulic conductivity.

Figure 6 indicates that the compacted soil (bulk density 1.85 Mg/m^3) had the lowest hydraulic conductivity. A poorly granulated or compacted soil has few total pore space as compared to well-granulated or ploughed soil. This results into poor or low rate of water movement. At field capacity which is approximately at 0.5 kPa suction, the rate of water movement was very low for compacted sandy clay loam soil of Hombolo. This condition ultimately, due to reduced infiltration, increased surface run-off and poor drainage, tends to result in soil erosion, reduced soil fertility and reduced crop yields (Vauclin and Chopart, 1992; Kayombo and Lal, 1993).

At field capacity the loose soil had the lowest bulk density (1.43 Mg/m^3) and, at the same time, the highest hydraulic conductivity compared to the more compact ones (bulk densities of 1.64 Mg/m^3 and 1.85 Mg/m^3). Crumbly and/or granular structured soils (ie. ploughed soils) have low soil bulk densities. Because of being crumbly and well pulverized, these soils allow ease movement of soil water. These soils are the ones which are desired for good crop growth. Good farm practices such as organic matter incorporation, crop rotation, agroforestry, use of animal drawn implements, inter-cropping and timely cultural practices can improve soil structure. This in turn improves soil hydraulic conductivity (Doneen and Westcot, 1984; Vauclin and Chopart, 1992).

Hydraulic conductivity also is observed to decrease in all soils as the soil matric suction increases. An increase in soil matric suction results in the withdrawal of water from the soil macropores. Since soil macropores are better conductors of water than soil micropores, an increase in matric suction greatly reduces the effective cross-sectional area of the flow. This in turn results in the reduction of soil hydraulic conductivity (Campbell, 1985; Ghildyal and Tripathi, 1987). In the loose soil with bulk density of 1.43 Mg/m^3, hydraulic conductivity decreased rapidly with an increase in matric suction. This is because the plot which was represented by this bulk density was the ploughed plot. So it was easy for water to flow or drain rapid through such a porous soil. In the soil with bulk density of 1.64 Mg/m^3 the hydraulic conductivity is observed to decrease gently and in the soil with bulk density of 1.85 Mg/m^3 hydraulic conductivity is observed to decrease very slowly. This is because the soil in this plot was compacted. So it was not easy for water to pass through the tightly held together soil particles in the compacted plot as in the well pulverized particles in the ploughed plot (Warkentin, 1971; Vauclin and Chopart, 1992; Kayombo and Lal, 1993).

At high soil matric suctions, hydraulic conductivity was low in all soils. However, the standard errors of estimates (Appendices 4, 5, 6) indicate that, the three different patterns of water flow are significantly different from one another at low suctions (at saturation), but at high suctions they are not, instead they are almost similar. This is because at high soil matric suctions water content in the soil is very low. During this time, the little water which is in the soil is tightly held to the soil particles. The force or energy required to remove it from the surfaces of the soil particles is great. In this case it is very difficult for the growing plants to remove water from the soil particles (Brady, 1984). The movement of water in the soil in this case is no longer through macropores but in micropores, where it moves by capillarity. Incorporation of organic matter which can improve soil structure can be employed. This can improve the soil's water-holding capacity which in turn improves soil water flow and finally soil hydraulic conductivity. (Hillel, 1980b; Doneen and Westicot, 1984; Brady, 1984; Ghildyal and Tipathi, 1987).

4.2.2 Values of b and n for the experimental soil

The constants b and n were computed by solving simultaneous equations obtained by measuring K at different soil matric suctions (Appendix 1) using Equation 11 (Gardner, 1958). The average values for b and n were 0.66 and 1.32, respectively (Table 7). Gardner (1958) worked with a sandy soil, a sandy loam, a fine sandy loam and a silt clay loam, while Vereecken (1988) worked with a clayey silt loam, a sandy silt loam, a sandy soil, a loamy soil and a loamy sand and both authors observed that large n values occur mainly in sandy materials and smaller values in finer textured soils. However, the values of b and n they obtained (Table 4), were slightly different from those obtained in this work. This is because the soil types they worked with were different from the soil type (sandy clay loam) which was I worked with in the study reported herein.

Table 7: Values of b and n for the experimental soil

Parameter	Pair of matric potentials (kPa)			Average
	0.2 - 0.5	0.2 - 0.8	0.2 - 1.0	
b	0.62	0.59	0.78	0.66($\pm$ 0.059)
n	1.56	1.05	1.35	1.32($\pm$ 0.148)

4.2.3 Plot of log (K_{sat}) versus bulk density

The plot of log (K_{sat}) versus bulk density (ϱ_b) is presented in Figure 7.

Log (K_{sat}) was significantly $R^2 = 0.649$, ($P < 0.01$) and negatively correlated with the bulk density (Figure 7, Appendix 7). This means that about 65% of the variations in Log (K_{sat}) can be explained by variations in soil bulk density. The values of L and E in Eq. [15] for the sandy clay loam soil used in this study were 5.644 and - 2.164, respectively. Vereecken (1988) observed the same trend and almost the same values when he plotted the graph of log (K) versus log (h_m) for a wide variety of Belgian soils: loamy soil, sandy loam, sandy soil, clayey silt loam, loamy sand and sandy silt loam which he worked with.

Using the obtained values of L and E, K_{sat} can be estimated at any value of bulk density for sandy clay loam soil in semi-arid Dodoma.

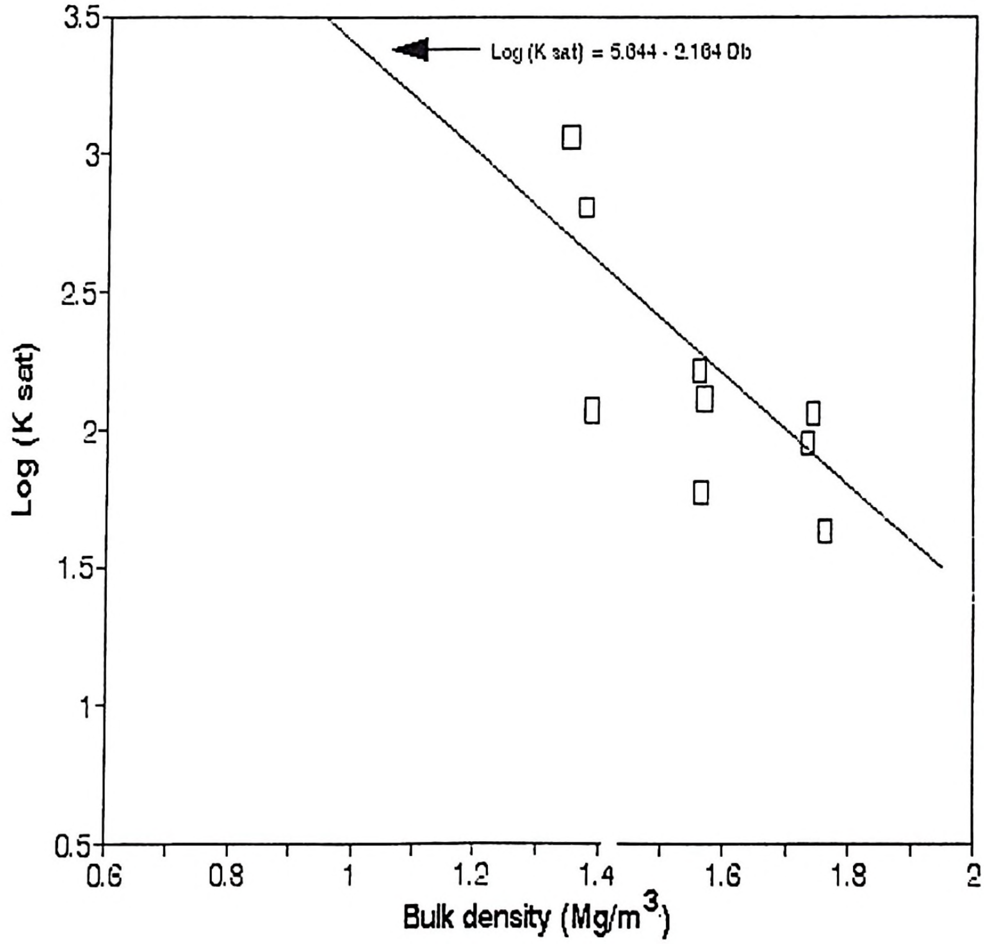


Figure 7 Log (K_{sat}) versus bulk density (Mg/m^3)

4.3 Hydraulic conductivity results for model validation

4.3.1 Plot of estimated K versus measured K

Using the calibrated Gardner model and the values of different bulk densities at which K was measured *in situ*, K was estimated and the graph of estimated K versus measured K was plotted (Figure 8).

There was a significant positive relationship ($P < 0.01$) between the estimated and the measured K (Appendix 8). There was also an indication that up to 99% of the variations in estimated K could be explained by variations in measured K ($R^2 = 0.997$). Therefore, the adopted model can predict the actual surface K of the sandy clay loam soil of semi-arid Dodoma to a 99.7% level of accuracy. This is in agreement with the findings obtained by Vereecken (1988) and Vereecken *et al.* (1989, 1990) when they compared measured with estimated hydraulic conductivity points for a loamy soil, a sandy loam, a sandy soil, a clayey silt loam, a loamy sand and a sandy silt loam soil in Belgium.

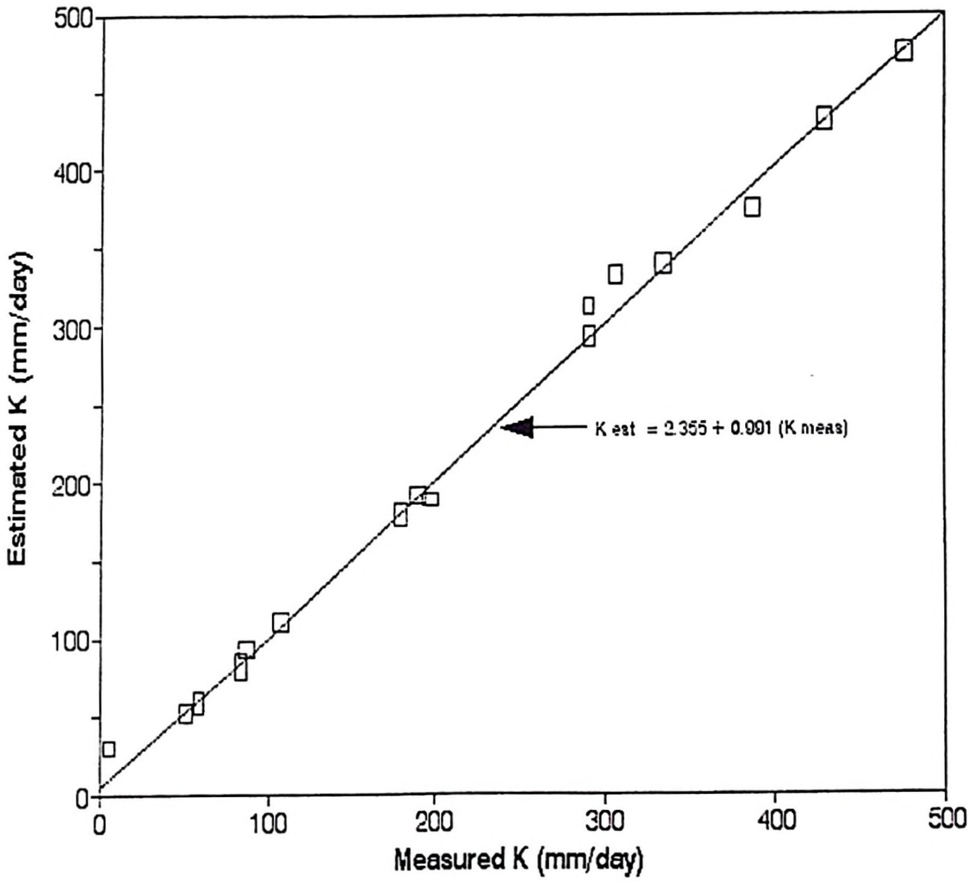


Figure 8 Estimated K versus measured K

From this it can be concluded that, simple soil physical properties like bulk density which is frequently characterized in soil survey studies, can be used to estimate to a reasonable level of accuracy the hydraulic conductivity of a given soil provided that the Gardner (1958) model is calibrated for that particular soil. Schuh and Sweeney (1986), Vereecken (1988), Ahuja *et al.* (1989), Michiels *et al.* (1989), Vereecken *et al.* (1989, 1990), Ravender *et al.* (1992), and Paige and Hillel (1993) observed the same.

The hydraulic conductivity values obtained in this research refer only to the surface soil, because the disc permeameter method is used for surface measurements only. In case disc permeameter measurements at depth are desired, a pit must be dug or other methods like the Auger hole method, the Guelph Permeameter method or the Instantenous profile method must be employed.

5. CONCLUSIONS AND RECOMMENDATIONS

5.1 Conclusions

Although the current study was conducted for only a short period, the following tentative conclusions can be drawn from it:

1. Saturated hydraulic conductivity can be readily estimated at any value of soil bulk density for sandy clay loam soils in semi-arid localities using the Gardner model.
2. Simple measurable soil characteristics, such as bulk density and matric suction, can be used to reliably obtain the parameters b and n of the Gardner model, and to accurately calibrate hydraulic conductivity of the surface soil.
3. The Gardner model can be used to accurately predict the surface hydraulic conductivity of sandy clay loams in semi-arid localities for which it was calibrated.

5.2 Recommendations

Since the study reported herein is based on only one year's research, it is difficult and unrealistic to attempt to make concrete recommendations. However, more studies need to be carried out in order to further calibrate and validate the Gardner model. At the same time, if the Gardner model is to be calibrated and validated at depth, methods other than the Disc permeameter must be employed. The Disc permeameter method is used for surface measurements only unless a soil pit is dug. However, digging could cause major soil disturbances.

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7. APPENDICES

Appendix 1: Values of K at different matric suctions and different bulk densities

Bulk density (Mg/m^3)	K (mm/day) Matric suction (kPa)				
	0	0.2	0.5	0.8	1.0
1.43	714.24 ±161.59	423.36 ±40.83	101.95 ±27.25	86.83 ±71.94	59.9 ±32.25
1.64	167.04 ±42.21	220.32 ±13.89	81.65 ±6.56	59.90 ±19.51	12.34 ±9.59
1.85	82.8 ±28.14	43.63 ±3.69	31.82 ±7.51	17.89 ±6.49	3.67 ±2.63

Appendix 2: Calculated values of log (Ksat) and their respective bulk densities (ϱ_b)

Ksat (mm/day)	log (Ksat)	Bulk density (Mg/m^3)
570.2	2.76	1.44
1036.8	3.02	1.42
121.0	2.08	1.43
103.7	2.02	1.65
57.0	1.76	1.63
121.0	2.08	1.64
69.9	1.84	1.85
60.5	1.78	1.84
32.8	1.52	1.86
241.43 (average)	2.1 (average)	1.64 (average)
113.55 (± s.e)	0.16 (± s.e)	0.06 (± s.e)

Appendix 3: Values of measured and estimated K

K measured (mm/day)	K estimated (mm/day)
7.60	23.63
54.43	53.38
54.43	54.97
66.53	65.67
80.35	78.44
83.81	82.01
95.04	96.95
95.04	97.96
146.88	143.50
164.16	150.02
155.52	152.78
302.40	305.45
302.40	319.33
319.68	333.84
345.60	349.01
406.08	394.36
440.64	435.86
483.84	476.37
200.25 (average)	200.75 (average)
35.73 (± s.e)	35.45 (± s.e)

Appendix 4

Regression Analysis - K (mm/day) Vs. Suction (kPa) - D_b = 1.43 Mg/m^3

Dependent variable: K (mm/day)

Independent variable: suction (kPa)

Parameter	Estimate	Standard Error	T Statistic	P-Value
Intercept	629.8414	73.8745	8.0159	0.0000
Slope	- 630.098	118.905	- 5.29916	0.0001

Analysis of Variance

Source	Sum of Squares	Df	Mean Square	F-Ratio	P-Value
Model	809927.0	1	809927.0	28.08	0.00-1
Residual	374952.0	13	28842.5		
Total (Corr.)	1.18488E6	14			

Correlation Coefficient = - 0.826772

R - Squared = 68.3553 Percent

Standard Error of Est. = 169.831

Appendix 5

Regression Analysis - K (mm/day) Vs. Suction (kPa) = 1.64 Mg/m^3

Dependent Variable: K (mm/Day)

Independent Variable: Suction (kPa)

Parameter	Estimate	Standard Error	T. Statistic	P-Value
Intercept	200.512	20.8577	9.6133	0.0000
Slope	- 184.522	33.5717	- 5.49635	0.0001

Analysis of variance

Source	Sum of Squares	Df	Mean Square	F-Ratio	P-Value
Model	69458.7	1	6945.7	30.21	0.00001
Residual	29889.6	13	2299.2		
Total Corr.)	99348.4	14			

Correlation Coefficient = - 0.836148

R-Squared = 69.9143

Standard Error of Est. = 47.95

Appendix 6

Regression Analysis - K (mm/day) Vs. suction (kPa) - D_b = 1.85 Mg/m^3

Dependent variable: K (mm/day)

Independent variable: Suction (kPa)

Parameter	Estimate	Standard Error	T. Statistic	P-Value
Intercept	70.7347	9.71952	7.2776	0.0000
Slope	- 69.5441	15.6441	- 4.44538	0.0007

Analysis of variance

Source	Sum of squares	Df	Mean Square	F-Ratio	P-Value
model	9866.22	1	9866.22	19.76	0.0007
Residual	6490.46	13	499.266		
Total (Corr.)	16356.7	14			

Correlation Coefficient = - 0.776654

R-squared = 60.3192 percent

Standard Error of est. = 22.3443

Appendix 7

Regression Analysis - Log (K_{sat}) Vs. Bulk density (Mg/m^3)

Dependent variable: Log (K_{sat})

Independent variable: Buld density (Mg/m^3)

Parameter	Estimate	Standard Error	T. Statistic	P-Value
Intercept	5.64394	0.990275	5.69937	0.0007
Slope	- 2.16365	0.600546	- 3.60281	0.0087

Analysis of Variance

Source	Sum of Squares	Df	Mean Square	F-Ratio	P-Value
Model	1.2415	1	1.2415	12.98	0.0087
Residual	0.66952	7	0.0956457		
Total (Corr.)	1.91102	8			

Correlation Coefficient = - 0.806011

R-Square = 64.9654 percent

Standard Error of Est. = 0.309266

Appendix 8

Regression Analysis - estimated K Vs. measured K

Dependent variable: K estimated

Independent variable: K measured

Parameter	Estimate	Standard	T Statistic	P. Value
Intercept	2.35489	3.42055	0.688456	0.5010
Slope	0.990765	0.0137597	72.0047	0.0000

Analysis of Variance

Source	Sum of Squares	Df	Mean Square	F-Ratio	P-Value
Model	383401.0	1	383401.0	5184.67	0.0000
Residual	1183.18	16	73.949		
Total (Corr.)	384584.0	17			

Correlation Coefficient = 0.998461

R-Squared = 99.6923 percent

Standard Error of Est. = 8.59936