COMPARISON OF FIELD AND LABORATORY MEASURED HYDRAULIC PROPERTIES OF SELECTED DIAGNOSTIC SOIL HORIZONS

by

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A dissertation submitted in accordance with the requirements for the Magister Scientiae Agriculturae degree in Soil Science in the

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November 2009

Table of contents

Declaration

I declare that this dissertation hereby submitted by me for the qualification in Magister Scientiae Agriculturae degree in Soil Science at the University of the Free State is my own independent work and has not previously in its entirety or part been submitted to any other University. I also agree that the University of the Free State has the sole right to the publication of this dissertation.

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Signature Date

Acknowledgements

Taking a look back at the end of this long journey, I can't help but wonder if I would be standing at the finishing line, which once seemed so far away, without the people who have been there with me along the way to give me direction, guidance, and encouragement. They have carried me through the times where it just seemed too hard to go on.

Southern African Development Community/Implementation and Coordination of Agricultural Research and Training in the SADC Region (SADC/ICART) programme hosted at the University of the Free State, South Africa, for the financial support that enabled me pursue and accomplish my Bachelors of Science (Honours) and Master of Science degree studies.

The University of Malawi through Bunda College of Agriculture for the study leave, financial, administrative, human resource and material support during the entire study period.

Prof. L.D. Van Rensburg my supervisor for his enduring patience, full commitment to my success, and unconditional efforts of providing the best possible academic training have been the source of my conviction, confidence, and motivation to reach higher standards. As a good teacher and mentor, he saw my strengths and weaknesses, and pushed me to reach my full potential.

Prof. M. Hensley my co-supervisor stood firmly by me during the times of confusion and frustration. He devoted a tremendous amount of time and effort to help me redefine a set of achievable goals and then made sure every step I took was a step closer to those goals. His painstaking attention to detail and accuracy in research, specifically during the development of this dissertation are two qualities I will spend the rest of my career trying to achieve.

The staff members in the Department of Soil, Crop and Climate Sciences, especially Mr R. Snetler, Mr W. Hoffmann, Mr N. Nhlabatsi, Mr G. Madito, and Mr E. Yokwane for their assistance during field and laboratory experiments.

Lastly, my thanks and appreciation should go to my fiancé, Orpah Tsokonombwe, for her much needed love and support through good and difficult times and relatives and friends for whatever they have done and patience to help me accomplish this work.

Dedication

I dedicate this dissertation to my parents Gregory Divala Chimungu and Mary Theodora Chimungu and my fiancé Orpah Tsokonombwe.

List of figures

List of tables

Abstract

An adequate characterization of soil hydraulic properties is a necessary solution for agriculturally and environmentally oriented problems such as irrigation, drainage, runoff and pollutants movement. The three approaches to determine hydraulic properties of soils are field measurements, laboratory measurements and mathematical models. *In situ* measurements, though representative, have the inherent limitation of being costly and time consuming. Laboratory and mathematical techniques are more convenient but require extensive comparison to field results as bench mark for evaluation. The objective of this study was to characterize the hydraulic properties of Bainsvlei and Tukulu form soils utilizing the above mentioned three approaches and to compare the results.

The laboratory methods selected were hanging water column and pressure plate apparatus. Undisturbed soil samples were used to determine θ-h relationships at 0-100 kPa suctions and disturbed soil samples up to 1500 kPa. The water retention characteristics for both soils were generally well defined with little variability between replicates. The main variations were due to texture differences between the horizons. The θ-h relationships were used to estimate textural and structural domains using empirical pore class limits and derivative curves. The suction value separating the structural domain from the textural domain varies from horizon to horizon. The boundary between soil pore categories cannot be taken as a fixed value for all soils and all types of soil use. The measured water retention data corresponded well with the fitted curve via the van Genuchten (1980) model, indicating that the model can be successfully used to describe θ-h relationships for Bainsvlei and Tukulu soils.

Soil water sensors were calibrated using undisturbed soil samples in climate controlled room for five horizons of a Bainsvlei form soil and three horizons of a Tukulu form soil. Soil water sensors and circuitry show extremely low sensitivity to temperature fluctuations. Horizon specific calibration is essential to get accurate water content estimates from the sensors if used in different soil horizons. Our study demonstrate that horizon specific calibrations of the water sensors improves the accuracy of soil water content monitoring compared with the manufacturer’s generic calibration equation for the soils tested in this study.

Hydraulic conductivity was obtained by measuring the hydraulic head and water content of the Bainsvlei soil form *in situ* with tensiometers and horizon specific calibrated ECH_2O EC-20 probes, respectively. The profile was characterized with several relations of hydraulic conductivity and varied with depth. The reason for this was attributed to heterogeneous nature of the profiles due to variation in particle size distribution. The van Genuchten (1980) model laboratory method was used to predict K-θ relationships utilizing laboratory determined θ-h relationships. The K-θ relationships predicted from the θ-h relationships of the soil cores corresponded well with those determined by the instantaneous profile field method for water contents which they have in common. Thus it appears that this laboratory method is applicable to the soils studied, but the accuracy of the predicted values is quite sensitive to the matching factors. Thus, accurate measurement of these parameters is necessary for its successful use.

The instantaneous profile field method is regarded as a reference method to measure *in situ* unsaturated hydraulic conductivity for both homogenous and layered soils (Hillel *et al.*, 1972). There are, however, several site or profile characteristics that may limit this method (Bouma, 1983). Our studies show that it is not applicable on duplex soils with slow permeable C-horizons i.e. the Tukulu form profile at Paradys, because of negative hydraulic gradients within the profile due to impaired internal drainage. There is a need to adapt this method to duplex soils.

Overall results indicate that from a practical perspective, the prediction of K-θ relationship from laboratory determined water retention data can be a viable alternative for determining the hydraulic properties of diagnostic horizons. The prediction of DUL using θ-h relationship has been found to be satisfactory.

CHAPTER 1

Motivation and objectives

1.1 Motivation

Quantitative knowledge about soil hydraulic properties such as water retention and hydraulic conductivity has traditionally been an important factor for assessing the suitability of land for irrigation and rain fed agriculture and trafficability (Schaap, 2005). In modern agricultural, environmental and engineering practices, varying degrees of quantitative aspects about soil hydraulic properties are needed for determining the soil water holding capacity, infiltration, percolation, and runoff rates, or for quantifying the transport of pollutants in soil (Dane & Topp, 2002). Furthermore, the soil water retention relationship can be used in mathematical models for estimation of unsaturated hydraulic conductivity (Mualem, 1976; van Grnuchten, 1980). Soil hydraulic properties are physical soil properties that depend mainly on soil structure, soil texture, organic matter content, cation exchange properties and bulk density (Hillel, 1998). Therefore they vary both vertically (horizons/layers in the profile) and horizontally in each plot. Thus, knowledge of soil hydraulic properties with respect to horizons is a prerequisite to understand the overall hydrological behaviour of a soil profile (ISO, 2009).

Owing to their relative importance in many disciplines, including environmental engineering, soil physics (Hopmans *et al.*, 2002), agricultural and environmental issues (Vachaud & Dane, 2002), a wide variety of methods are being developed and improved to effectively determine soil hydraulic properties. These properties are difficult to measure and therefore require the use of both direct and indirect methods to adequately describe them accurately. Several field methods, laboratory methods and theoretical models for such determinations exist, each having their own limitations (Stephens, 1994). *In situ* determinations are generally preferred owing to the large volume of soil tested and the preservation of soil structure during the experiments (Green *et al.*, 1986). *In situ* measurements, though more representative of actual conditions, have the disadvantage of being costly and time consuming, whereas laboratory and mathematical processes are perceived to be more convenient and offer many advantages compared to *in situ* techniques. However it still requires extensive comparisons between field and laboratory results to determine the validity of the latter for a range of different soils. This study attempts to make a contribution specifically in this connection, illustrated by the results for two very different soils.

The capability of accurately predicting water movement within the soil profile has been the object of extensive research. Although these techniques have been tested on some specific soils, the inherent variability between different soils, which affect soil hydraulic properties, accentuate the need for additional experimentation. Furthermore very few if any systematic relevant studies have been conducted on Bainsvlei and Tukulu soils (Soil Classification Working Group, 1991). Characterization of soil hydraulic properties for these soils would provide valuable information on soil hydrologic processes as affected by management and the pedological characteristics of these soils. This study is intended to fill that gap by providing detailed soil descriptions and step by step procedures for determining soil hydraulic properties.

Modern soil water sensor and electronic data storage technology provides a new basis to study hydraulic processes at a time scale as never before. There is great variety of soil water sensors on the market, especially capacitance based sensors. Due to variation in soil texture and salinity the manufacturers' generic calibration for ECH$_2$O EC-20 probes results in approximately ± 3-4% accuracy for most medium to fine textured mineral soils (Foley & Harris, 2007). However it has been suggested by the manufacturers that a more rigorous calibration is necessary when the sensors are used where accuracy is of paramount importance.

1.2 Objectives of the study

In light of the above discussion the main objective of the study was to characterize and compare field and laboratory determined soil hydraulic properties of two contrasting soil forms. This overarching objective was evaluated in four independent studies outlined below. Each study was carried out with its own set of specific objectives.

Study 1: This study was entitled: **"Laboratory characterization of the water retention characteristics of a Bainsvlei form soil and a Tukulu form soil" (Chapter 3).** The specific objectives of this study were: (i) to describe and classify the morphological, physical and chemical characteristics of a Bainsvlei form soil and a Tukulu form soil in detail by means of profile descriptions and relevant analytical data;

(ii) to determine the soil water retention characteristics (θ-h relationships) for the different horizons of two soils; (iii) to described the θ-h relationships using the van Genuchten (1980) model; (iv) to relate the θ-h relationships of the different horizons to their morphological and other characteristics such as structure, texture and bulk density; (v) to use the θ-h relationships to separate the porosity into matrix and structural domains.

Study 2: This study was entitled: "**Calibrating ECH$_2$O EC-20 sensors for measuring soil water content" (Chapter 4).** The specific objectives of this study were: (i) to evaluate the effects of soil temperature on ECH$_2$O EC-20 probes in different soils at variable water content; (ii) to derive specific calibration equations for eight horizons of two soils and compare horizon specific calibrations with generic calibrations.

Study 3: This study was entitled: **"Comparing laboratory and field determined hydraulic conductivity for a Bainsvlei form soil" (Chapter 5).** The general objective of this study was to investigate and compare the hydraulic properties of the Bainsvlei form soil using field and laboratory methods. Specific objectives were: (i) to characterize the *in situ* K-θ relationships for the profile; (ii) to predict the K-θ relationships with van Genuchten (1980) model; (iii) to estimate drained upper limit for the profile.

Study 4: This study was entitled: "**Comparing laboratory and field determined hydraulic conductivity values for a Tukulu form soil" (Chapter 6).** The general objective of this study was to investigate and compare the hydraulic properties of the Tukulu form soil using field and laboratory methods. Specific objectives were: (i) to characterize the *in situ* K-θ relationships for the profile; (ii) to predict the K-θ relationships with van Genuchten (1980) model; (iii) to estimate drained upper limit for the profile.

1.3 Layout of thesis

This thesis consists of seven chapters. Chapter one deals with the motivation and objectives of the study. Chapter two reviews the literature relevant to soil hydraulic properties, theory, measurement and estimation techniques, major application of the soil hydraulic properties, and calibration of ECH$_2$O-EC 20 probes. Detailed materials and methods and results pertinent to the experiments conducted to achieve the research objectives are presented in chapters three, four, five, and six. Summary and major conclusions are presented in chapter seven. The format of the chapters is in the form of intact papers for submission to journals. As a result this format leads to some duplication in the chapters.

1.4 References

DANE, J. & TOPP, G. 2002. Method of soil analysis, Part 4, Physical methods. SSSA., Book Series No 5, Madison, Wisconsin.

FOLEY, J. & HARRIS, E., 2007. Field calibration of thetaprobe (ML2X) and ECH_2O probe (EC-20) soil water content sensors in black vertisol. *Aust. J. Soil Res.* 45, 233-236.

GREEN, R., AHUJA, L. & CHONG, S., 1986. Hydraulic conductivity, diffusivity, and sorptivity of unsaturated soils: field methods. *In:* A. Klute, (ed.). Method of soil analysis, Part I, Physical and mineralogical methods Monograph No. 9. ASA., Madison, Wisconsin.

HILLEL, D., 1998. Environmental soil physics. Academic Press Inc, New York.

HOPMANS, I., SIMUNEK, J., ROMANO, N. & DURNER, W., 2002. Simultaneous determination of water transmission and retention properties inverse methods. *In:* J. Dane, G. Topp, (eds.). Methods of soil analysis, Part 4, Physical methods. SSSA., Book Series No 5, Madison, Wisconsin, 963-1008.

INTERNATIONAL ORGANIZATION FOR STANDARDIZATION (ISO), 2009. Determination of the soil water retention characteristics. Version 1.3, FUTMON-soil moisture workshop, International Organization for Standardization, Geneva, 1-12. *(Available at www.iso.ch)* 1-12.

MUALEM, Y., 1976. A new model for predicting the hydraulic conductivity of unsaturated porous media. *Water Resour. Res.* 12, 513-522.

SCHAAP, M.G., 2005. Models for indirect estimation of soil hydraulic properties. *In:* M. Anderson (ed.). Encyclopaedia of hydrological sciences, John Wiley & Sons, New York, 1-8.

SOIL CLASSIFICATION WORKING GROUP, 1991. Soil classification a taxonomic system for South Africa. Department of Agricultural Development, Pretoria, South Africa.

STEPHENS, D.B., 1994. Hydraulic conductivity assessment of unsaturated. *In:* D.E. David & T.J. Stephen (eds.). Hydraulic Conductivity and Waste Contaminant Transport in Soil, ASTM STP 1142, American Society for Testing and Materials, Philadelphia.

VACHAUD, G. & DANE, J., 2002. Instantaneous profile method. *In:* J. Dane, & G. Topp, (eds.). Methods of soil analysis, Part 4, Physical methods. SSSA., Book Series, No 5, Madison, Wisconsin, 937-962.

VAN GENUCHTEN, M. T., 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Sci. Soc. Am. J.* 44, 892-898.

CHAPTER 2

Literature review

2.1 Introduction

The main focus of this chapter is essentially to give a short review of the theoretical background of relevant soil hydraulic properties and processes. Hence, field, laboratory and theoretical methods for estimating the hydraulic properties have been reviewed and special attention was also given to answer the question on how soil porous systems can be separated into textural and structural domain using soil water retention characteristics. Soil water monitoring by continuous logging soil water sensors was central to the study; it was therefore decided to compliment the study with a section on the calibration of ECH$_2$O EC-20 probes in the laboratory.

2.2 Soil hydraulic properties

Water movement within the soil profile is an important component of agricultural and environmental and the understanding of it will help to solve problems related to irrigation, subsurface drainage contributions to groundwater, growth of saline seeps, and water disposal. Adequate and effective management of soil and water therefore often necessitates characterization of water retention and hydraulic conductivity functions of the area concerned. These functions collectively are referred to as soil hydraulic properties (Klute & Dirksen, 1986).

2.2.1 Soil water retention characteristic

Soils differ in their capacity to retain water against gravity. The water binding properties of soils are represented by the relationship called soil water retention characteristics, which is coded as θ-h relationship in this thesis. This is the relationship between amount of water in the soil and the potential energy with which it is bound by the soil (Jury *et al.*, 1991). The θ-h relationship is a unique function for each soil, because of variation in soil particle size distribution and structure. Both of these factors influence the θ-h relationship by affecting the pore size distribution and the number of a given size pore in each size class (Dexter, 2004).

The θ-h relationship is an important soil property that is needed for the study of plant available water, infiltration, drainage, hydraulic conductivity, irrigation scheduling, water stress on plants, and solute movement (Kern, 1995). In non swelling soils it reflects the geometry of the pores and this geometry, in turn, determines to a large extent the hydraulic conductivity. Since the pressure difference across an air-water interface is inversely proportional to the equivalent radius of the interface, the θ-h relationships can be converted into an equivalent pore size distribution (derivative curve), and the water content at any given suction is equal to the porosity contributed by the pores that are smaller than the equivalent diameter corresponding to that suction, (Jury *et al.*, 1991). The spatial patterns of water retention characteristics are important factors for studying the response of vegetation and hydrological systems in climate change (Dolph *et al.*, 1992). Soil particle size distribution strongly affects the θ-h relationship at suction heads > 100 kPa and to a lesser extent, at lower suctions where soil structure is also important (Hillel, 1998).

This section will introduce laboratory methods for determing the θ-h relationship and will also described the parameters and functions that describe the θ-h relationship and then briefly review how the soil porous system can be separated into textural and structural domain using θ-h relationship.

2.2.1.1 Methods of characterizing soil water retention characteristic

In literature, there are several methods available used to obtain measurements of θ-h relationship. It is impossible to cover all the soil water retention measurements methods that are presently in use. A complete discussion of measuring methods of θ-h relationships is given by Dirksen (1999). This section presents direct methods for measuring θ-h relationships in the laboratory (i.e. hanging water column and pressure plate apparatus). In the laboratory, the θ-h relationship may be measured on replicated samples over range of water contents. Virtually the entire range from water saturated soil to very dry soil may be covered by using a hanging water column and pressure plate apparatus (Klute, 1986; Dirksen, 1999; Bohne, 2005). According to Jury *et al.* (1991) equilibrium water contents are usually obtained by exposing saturated, undisturbed soil sample to a hanging water column at suctions < 10 kPa. The use to of the hanging water column is confined to this range because of the limitation in the length of hanging water column and due to possible cavitation (Jury *et al.,* 1991). However, the pressure plate apparatus is normally used for the suction range of 30-1500 kPa (Jury *et al.,* 1991). According to Reeve and Carter (1991) the precision of pressure plate apparatus is very good, with a coefficient of variation of 1-2% attainable. However, clogging of the ceramic plates by soil particles or alga growth can occur after repeated use, reducing the efficiency of the plate (Townend *et al.*, 2001) and other problem is time taken to reach equilibrium if at all is reached because conductivities are so low at the dry end.

The θ-h relationship in low suction range of 0-100 kPa is strongly influenced by soil structure and its natural pore size distribution (Hillel, 1998). Hence, undisturbed soil samples are recommended to be used (Dirksen, 1999). It is acceptable to use disturbed samples at suctions greater than 100 kPa, provided that the disturbance consists only in breaking off small pieces of soil and not in compressing or remoulding the soil.

2.2.1.2 Mathematical description

Measured θ-h pairs are often fragmentary, and usually constitute relatively few measurements over the θ range of interest. According to Bohne (2005) mathematical functions provide a valuable tool to smooth measured data and simplify interpolation between measured data points and fitted parameters used to predict unsaturated hydraulic conductivity. Or & Wraith (2002) state that mathematical expression describing the θ-h relationship should: (i) contain as few parameters as necessary to simplify its estimation, and (ii) describe the θ-h relationship at the limits (i.e. both wet and dry ends) while closely fitting the nonlinear shape of θ-h data. But the choice of function depends on soil type, application, and personal preference (Kosugi *et al.,* 2002).

The use of parametric models in soil water retention studies has several advantages. For example, they allow for a more efficient representation and comparison of the θ-h relationships of different soils and horizons (Marion *et al.,* 1994). They are also more easily used in scaling procedures for characterizing the spatial variability of soil hydraulic properties across the landscape (Kosugi *et al.,* 2002). And, if shown to be physically realistic over a wide range of water contents, analytical expressions provide a method for interpolating or extrapolating to parts of the θ-h relationship for which there is missing data (Bohne, 2005) and appropriateness to application in unsaturated flow models (van Genuchten *et al.,* 1991).

Several mathematical functions have been proposed to empirically describe the complete θ-h relationship (Leiji *et al.*, 1999). Notable among those are the equations proposed by Brooks & Corey (1964) and by van Genuchten (1980). Several researchers for example van Genuchten & Nielsen (1985) and Fare *et al.* (2000), have shown that the van Genuchten (1980) equation give good description of the observed retention data of large number of soils. The van Genuchten (1980) equations (VG) are given below:

$$S_e = \frac{(\theta-\theta_r)}{(\theta_s-\theta_r)} = \left[\frac{1}{1+(\alpha h)^n}\right]^m \qquad 2.2$$

where θ_r and θ_s are the residual and saturated water contents respectively, S_e is the dimensionless value water content, and α, n, and m are parameters directly dependent on the shape of the θ-h curve. A considerable simplification is gained by assuming that $m = 1 - 1/n$ (van Genuchten, 1980). Thus, the parameters required for estimation are θ_r, θ_s, α and n. θ_s is usually known and is easy to obtain experimentally with good accuracy, in many cases. Note that θ_r may be taken as water content at a suction of 1500 kPa or air dry or a similar value (van Genuchten, 1980). No physical meaning is attached to the parameters α, n and m.

Estimation of parameters from the observed retention data requires: (i) sufficient data points, at least five to eight θ-h pairs (ISO, 2009), and (ii) a program for performing nonlinear regression (Or & Wraith, 2002). Recent versions of computer spreadsheets provide a relatively simple and effective mechanism to perfom nonlinear regression. A more complete discussion of the computational steps required for fitting a model to the observed data using a commercially available spreadsheet is given in van Genuchten *et al.* (1991), Kosugi *et al*. (2002), and Seki (2007). In this study, RETC computer program was adopted. Marion *et al*. (1994) and Fare *et al.* (2000) found that the RETC computer program coupled with the VG model produced promising fits for the measured retention data in their respective studies.

The VG model has been used by several researchers to describe θ-h relationship with different soil types (Lorentz *et al.*, 2001; Fare *et al.*, 2000; Zhang *et al.*, 2007). However, although the van Genuchten (1980) model is almost 30 years old, after an extensive literature search, there are very few, if any studies on the use of the VG model on Bainsvlei and Tukulu soils.

2.2.1.3 Separation of porosity into textural and structural domains

Soil structure has been traditionally considered as one of the main attributes of soil quality (Dexter, 2004) and the qualitative role of soil structure in soil hydrology is well documented in literature on the pedon scale (Kutilek, 2004). However, quantitative evaluation has only recently begun to generate a large amount of interest and practical application. A quantitative evaluation of the bi-modal porosity by Durner (1994) and of the realated soil hydraulic functions has improved the understanding about the influence of the soil structure on soil hydraulics and hydrology.

The description of a soil porous system is a basic requirement for the quantification of the role of soil structure on soil hydraulic properties. With respect to hydraulic functions Kutilek (2004) distinguishes three categories of pores: submicroscopic pores which are so small they perclude clusters of water molecules forming fluid particles or flow paths; micropores is where the shape of the interface between air and water is determined by the configuration of the pores and the forces on the interface (this category is subdivided into two sub-categories; textural and structural domain); macropores of a size that capillary minisci are not formed across the pore.

According to Nimmo (1997) and Kutilek (2004) to quantitatively define the influence of soil structure on the hydraulic properties of soil, it is useful to only deal with the category of micropores and its subdivisions into textural and structural domains (i.e. assuming θ-h relationship is the sum of two components, one textural (θs_1) and the other structural (θs_2)). It is obligatory to define carefully the terminology because other authors use different experimental approaches and use some of the terms for different components of the soil pore system. Textural domain is the pore space within soil aggregates or within blocks of soil if aggregates are not present (Elhers *et al.*, 1995; Kutilek, 2004). Textural domain is little affected by soil management. Structural domain is the pore space between the micro-aggregates and also between incipient aggregates (Kutilek, 2004). Their morphology depends upon soil genesis and soil management factors such as tillage, compaction and cropping (Dexter, 2004).

The boundary between the two domains given in literature is usually between 15 to 30 µm (Marshall, 1959; Luxmoore, 1981; Skidmore, 1985; Kutilek, 2004), and it is dertermined either by tools of the soil micromorphology (Pagliai *et al.*, 2004) or from the retention curve as their derivative (Kutilek, 2004; Kutilek *et al.,* 2006; Kutilek & Jendele, 2008; Dexter *et al.*, 2008) and using empirical pore class limits (Marshall, 1959; Luxmoore, 1981; Skidmore, 1985).

Based on the knowledge of the θ-h relationships (∂θ/∂log(h) and log(h)) (i.e. derivative curve) Kutilek & Jendele, (2008) reported that the boundary between the structural and textural domain can be represented by the minimum between the two principal peaks. The separation of the soil pore system into these categories produces what are called bi-modal pore size distributions (Dexter *et al*., 2008).

The pore radius is calculated from the θ-h relationships using the capillary theory that relates pore radius (r) to the suction (h) at which the pore drains;

$$r = \frac{2\gamma cos\beta}{gh\rho_w} = \frac{0.1490cm^2}{h} \qquad 2.3$$

where r is the pore radius (µm), γ is the surface tension between the water and air (ergs cm^{-2}) (72.7), β is the contact angle (assumed to be zero), ρ_w is the density of water (g cm^{-3}), g is the acceleration due to gravity (cm s^{-2}) and h is the soil water suction in cm.

According to Marshall (1959) and Skidmore (1985) a pore radius of 15 µm corresponding to 10 kPa suction and it is frequently arbitrarily chosen as a boundary separating textural domains from structural domain. Luxmoore (1981) has however stated that the operational boundary definitions between pore categories based on suction and equivalent pore radius may not necessarily be the best choice for all soils. However, because of its simplicity and applicability it was adopted in this study.

Several studies (e.g. Tuller & Or, 2002; Kutilek *et al*., 2006) found that the suction value (h_a) separating textural and structural domains varies for various soil taxons and they concluded that a fixed boundary between the textural and structural domains does not exist in natural soils because it mainly depends on soil genesis and soil management.

Several studies have revealed that quantification of pore size distribution over a wide range of soil horizons enable better description of soil functions including retention, hydraulic conductivity (Nimmo, 1997; Droogers *et al.*, 1998, Kosugi *et al.,* 2002), root growth (Glinski & Lipiec, 1990), agriculture management effects (Lipiec & Hatano, 2003) and evolution of soil (Richard *et al.*, 2001). Few, if any, comprehensive literature exist on the separation of soil porosity into textural and structural domains for South

African soils. This information is vital for improving our understanding on the quantitative relationships between morphological characteristics of soil structure and soil hydraulic functions.

2.2.2 Unsaturated hydraulic conductivity

Several agriculturally important water-flow processes involve unsaturated flow, viz. infiltration, evaporation, and the flow of water to plant roots as well as the transport of water and solutes beyond the root zone. To understand and describe these and other processes, the hydraulic properties that govern water movement in the soil must be quantified (i.e. hydraulic conductivity). This soil property varies over many orders of magnitude not only between different soils but also for the same soil as a function of water content or suction. This makes the soil hydraulic conductivty function one of the most important physical soil properties, but also very difficult to measure accurately (Hillel, 1998; Dirksen, 1999).

Hydraulic conductivity describes the ease of water flow in the soil. It is a constant of proportionality between the flux of water and its driving force the hydraulic gradient. In saturated conditions the saturated hydraulic conductivity reflects the number of pores and their arrangement. Hydraulic conductivity also depends on the water content and soil water suction. If the soil is unsaturated, part of the pores will be empty, hence hydraulic conductivity is reduced as water content decreases.

Much has been published on the determination and/or estimation of the unsaturated hydraulic conductivity, including reviews (Klute & Dirksen, 1986; Green *et al.,* 1986; van Genuchten *et al.,* 1992; Dane & Topp, 2002). Because the literature is so extensive that it is neither necessary nor possible to give a complete review and evaluation of all available methods. Instead, this setion present a review of methods that will be utilized in this study (Section 2.2.2.1 and 2.2.2.2). This includes some very recent work by several researchers. Detailed theoretical concepts and equations associated with specific methods are given in the discussion of individual methods in separate chapters. The physical principles involved in unsaturated movement of water and its measurement are discussed in more detailed and elementary level in soil physics textbooks (e.g. Jury *et al.,* 1991; Hillel, 1998).

2.2.2.1 Instantaneous profile method (field method)

Soil water dynamics have been described using the Darcy–Buckingham equation, which relates the soil water flux density to the total soil water potential gradient through the soil hydraulic conductivity. The soil hydraulic conductivity (K), for rigid porous systems in general is taken as a function of only soil water content (K-θ), is a fundamental parameter for their hydraulic characterization, and for the use of the Darcy–Buckingham approach. Its determination, mainly under field conditions, is therefore an important issue, described in detail by Klute (1986) and Dirksen (1999).

Of the several available methods for estimating K-θ relationships in the field, the instantaneous profile method or internal drainage method is regarded as reference method to measure unsaturated hydraulic properties for both homogeneous and layered soils (Hillel *et al.*, 1972). The instantaneous profile method is based on the Darcy-Buckingham analysis of transient soil water content and hydraulic head profiles during vertical internal drainage. The history of soil internal drainage description started with Richards (1956), who first studied the drainage flux method in the field. Watson (1966) improved upon the method by replacing the computation of differences in time and depth by the presumably accurate "instantaneous profile method" in laboratory studies. Thereafter several contributions were made based on field measurements by van Bavel *et al.* (1968) and Davidson *et al.* (1969). Hillel *et al.* (1972) published a procedure to calculate soil hydraulic conductivity *in situ*.

Although the instantaneous profile method is regarded as a standard method to measure *in situ* unsaturated hydraulic properties, Tseng & Jury, (1993) have shown that the instantaneous profile method may give significant errors in heterogenous soils even if the instrument errors are neglected. In general there are two aspects that affect the applicability of instantaneous profile method; (i) problems encoutered during measurement of hydraulic conditions, such as soil water suction, and (ii) limitations imposed by the profile characteristics (Baker *et al.,* 1974). Occurrence of relatively slowly permeable soil horizons, such as commonly occuring plough pan layers or certain genetic horizons, may result in failure to calculate hydraulic conductivity due to negative hydraulic gradients (Paige & Hillel, 1993). Because θ and h determination is needed over a long period of time, the instantaneous profile method is time- and equipment- intensive, and thus costly, especially if there are several sites to be monitored. Simplified methods have been developed which require fewer *in situ* measurements but depend on flow approximations (Libardi *et al.,* 1980; Sisson *et al.,* 1980). Their use is mainly limited because the founding assumptions were not validated on critical field measurements (Reichardt, 1993).

Several researchers have used the instantaneous profile method on different types of soils (Field *et al.,* 1983; Paige & Hillel, 1993; Marion *et al*., 1994; Fare *et al.,* 2000; Zhang *et al.,* 2007). These workers have proven the feasibility of the instantaneous profile method for determining the unsaturated hydraulic conductivity functions of soils in the field. Given the spatial variability of soil hydraulic conductivity, it is necessary to determine soil hydraulic properties in the area of interest (i.e. Bainsvlei and Tukulu soils), to ensure that the flow dynamics in the these soils are well understood and this is particularly important for evaluation of soil and water conservation and sustainability of its use.

2.2.2.2 Prediction of unsaturated hydraulic conductivity based on soil water retention (laboratory method)

The direct measurement of unsaturated hydraulic conductivity still remains notoriously sophisticated and difficult (Hillel, 1998; Ippish *et al.,* 2006). One alternative to direct measurement of the unsaturated hydraulic conductivity is to use theoretical methods which predict the conductivity from more easily measured θ-h relationships (Dirksen, 1999). Execution of these predictive conductivity models requires independently measured θ-h relationships and some matching factors. Measured input retention data may be given either in tabular form, or by means of closed-form analytical expressions which contain parameters that are fitted to the observed data (Leiji *et al.,* 1999). Because of the shortcomings of direct measurement procedures and their simplicity and ease of use, predictive models for hydraulic conductivity are gaining popularity in numerical studies of unsaturated flow (Kosugi *et al.,* 2002). Results thus far suggest that predictive models work reasonably well for many coarse-textured soils, but that predictions for many fine-textured and structured field soils remain inaccurate (van Genuchten & Nielsen, 1985; Leiji *et al.,* 1999).

While a large number of analytical soil water retention functions have been proposed in literature, only a few functions can be easily incorporated into the predictive pore-size distribution models to yield relatively simple analytical expressions for the unsaturated hydraulic conductivity function (van Genuchten & Nielsen, 1985). Notable ones are the models by Brooks-Corey/Burdine (Brooks & Corey, 1964) and van Genuchten/Mualem (van Genuchten, 1980). The model of van Genuchten/Mualem is the most frequently used and it is given by:

$$K(\theta) = K_s S_e^l \left[1 - \left(1 - S_e^{1/m}\right)^m\right]^2 \qquad 2.6$$

where K_s is the saturated hydraulic conductivity, l is an empirical constant assumed equal to 0.5 (Mualem, 1976), S_e was previously defined in equation 2.2, and $m = 1 - 1/n$ (van Genuchten, 1980). It is common practice, therefore, to use measured K_s to match the measured and calculated values of hydraulic conductivity. Any spreadsheet tool or parameter optimization program may be used to calculate Equation 2.6. Furthermore, the RETC computer program (van Genuchten *et al.,* 1991), is a useful and versatile tool to calculate K-θ and to estimate unknown parameters in Equation 2.2 and to calculate Equation 2.6.

Several investigators have used the RETC computer program to predict K-θ relationships for different soils (Marion *et al.,* 1994; Fare *et al*., 2000). Marion *et al.* (1994) found that the RETC computer program (van Genuchten *et al.,* 1991), produced promising predictions of the K-θ relationship based on water retention data. They used the instantaneous profile method data as a basis for collecting their data.

Several investigators have tested the van Genuchten-Mualem model by comparing the predicted to measured values of hydraulic conductivity (Dane, 1980; Field *et al*., 1983; van Genuchten & Nielsen, 1985; Paige & Hillel, 1993; Marion *et al.,* 1994; Zhang *et al.,* 2007). Dane (1980) reported a reasonable correspondence between instantaneous profile method-measured K-θ and predicted K-θ on sandy to loam textured soils. Paige & Hillel, (1993) also found that hydraulic conductivities predicted by the van Genuchten-Mualem based on soil core data agreed closely with those obtained by the instantaneous profile method for corresponding ranges of suction and water content in a fine sandy loam. These results demonstrate that all estimation procedures, however, needs the results of direct measurements as benchmarks for validation. Futhermore, the results of these studies are not meant to represent characteristics for all soils including South African soils. It is necessary to evaluate the applicability of this model on South African soils in particular the Tukulu and Bainsvlei soils.

2.3 Calibration of ECH_2O EC-20 capacitance water sensors

The standard method of soil water content measurement involves taking a sample of the soil, weighing it before any water is lost, and drying it in an oven before weighing it again. Because water content is determined by direct weighing, this method is called gravimetric. However, the use of gravimetric sampling is destructive, labour intensive, is often slow, not timely and may be costly. Many alternative methods have been developed that take advantage of the relatively high permittivity of water and after calibration, they are a good measure of soil water content. During the past decades, dielectric methods have become popular, such as Time Domain Reflectometry (TDR), impedance and capacitance methods (Saito *et al.*, 2008).

Capacitance methods have become very popular because of their low cost, among other reasons such as lack of radiation hazard, allowing continuous monitoring, repeatability, and applicability to a wide range of soil types (Seyfried & Murdock, 2001). Many of these capacitance sensors have the additional benefit of being loggable-readings at short intervals, for example, 15 minute intervals so that volumetric water content (θv) change during short duration events, such as during internal drainage experiments, can be monitored with ease. Sensors that operate at low frequencies (<100 MHz), are often criticized for being more susceptible to soil environmental variables (Chandler *et al.*, 2004). Studies continue to determine the effects of soil environment variables like temperature, electrical conductivity, and soil type that affect the accuracy and reliability of the measurement (Seyfried & Murdock, 2001; Chandler *et al.*, 2004).

Several manufacturers produce capacitance type sensors for commercial use. Some are intended for long term data acquisition with sensors fixed in place, while others are intended to be portable with measurements triggered manually by the user (Paltineanu & Starr, 1997). The capacitance sensors that utilize access tubes can cause substantial measurement errors, because of uncertainty in measurement volume, and annular air gaps around sensors (Or & Wraith, 2002). Hence, buried probe designs (e.g. ECH_2O EC-20 probes) seem to perform more reliably at present than those inserted into soil access tubes (Or & Wraith, 2002).

The ECH_2O EC-20 probes measure volumetric water content (θv) by measuring the dielectric constant (ε) of the bulk soil. The soil bulk dielectric constant is dominated by the dielectric constant of soil water ($\varepsilon_w \approx 80$) because of its large magnitude relative to that of soil solids ($\varepsilon_s \approx 5$) and soil air ($\varepsilon_a \approx 1$). Thus, when the amount of water changes in the soil, the ECH_2O EC-20 probe will measure a change in capacitance (from the change in dielectric constant) that can be directly correlated with a change in water content. Circuitry inside the ECH_2O EC-20 probe changes the capacitance measurement into proportional millivolt (mV) output (Kelleners *et al.,* 2005). ECH_2O EC-20 probe mV output is directly converted to θv using a generic calibration given by the manufacturer as follows:

$$\theta_v = 0.000695mV - 0.29 \tag{2.7}$$

The manufacturers expected their generic calibration equation to be independent of soil type, soil density, temperature, and salinity, but it proved not to be quite as 'universal'. For low bulk densities, specific mineralogical properties, clays, organic soils, etc., and for accurate results, soil specific calibration is necessary. Several studies from independent researchers have proven that the generalized calibration of these ECH_2O EC-20 probe sensors is not accurate for all soils (Evett & Steiner, 1995; Paltineanu & Starr, 1997; Baumhardt *et al.,* 2000; Cepuder *et al*., 2002; Foley & Harris, 2007). Thus, it is important to calibrate each sensor for the specific soil horizon in which the sensor will be used. There is also ample evidence from other studies to support the determination of soil specific calibrations to improve sensor accuracy (Baumhardt *et al.,* 2000; Lane & Mackenzie, 2001; Geesing *et al.,* 2004; Czarnomski *et al.,* 2005). Czarnomski *et al*. (2005) found that soil specific calibration of the ECH_2O EC-20 probe improve accuracy of the sensors to ±1-2%.

ECH_2O EC-20 calibration generally follows the standard procedure for calibrating capacitance probes outlined by Paltineanu & Starr (1997) which is commonly performed in a temperature controlled room, with distilled water and disturbed soil which is uniformly packed around the sensor. Unfortunately, conditions such as these do not exist in the field, and thus the results obtained are, at best, a rough estimate of the field condition. However, according to Weitz *et al.* (1997) and IAEA (2008) only reliable calibration of the sensors can be done by using the undistubed soil cores that can replicate *in situ* physical conditions.

2.4 Concluding remarks

The preceding review briefly describes the importance of characterization of soil hydraulic properties and some of the existing measurements and analytical techniques, and their limitations. There is a relative abundance of literature dealing with the theory and application of soil hydraulic properties, but there is very few, if any that give detailed descriptions of soil hydraulic properties determination on Bainsvlei and Tukulu soils. This information is needed for improving understanding of the effects of soil management or land use on soil profile hydrology. This study is intended to help fill this gap.

ECH$_2$O EC-20 probes have been used extensively in recent years to monitor soil water content at different depths and locations (Bandaranayake *et al.*, 2007). However, after an extensive literature search, we did not find any work on the use of the continuous logging probes time independent process such as the instantaneous profile method.

2.5 References

BAKER, F., VENEMAN, P. & BOUMA, J., 1974. Limitations of instantaneous profile method for field measurements of unsaturated hydraulic conductivity. *Soil Sci. Soc. Am. Proc*. 38, 885-888.

BANDARANAYAKE, W., PARSONS, L., BORHAN, M. & HOLETON, J., 2007. Performance of a capacitance-type soil water probe in a well drained sandy soil. *Soil Sci. Soc. Am. J.* 71, 993-1002.

BAUMHARDT, R.L., LASCANO, R.J. & EVETT, S.R., 2000. Soil material, temperature, and salinity effects on calibration of multisensor capacitance probes. *Soil Sci. Soc. Am. J.* 64, 1940–1946.

BOHNE, K., 2005. An introduction to applied soil hydrology. Lecture notes in GeoEcology. 35447, Catena VERLAG GMBH Reiskirchen, Germany.

BROOKS, R. & COREY, A., 1964. Hydraulic properties of porous media. Hydrology Paper, No 3, Colorado State Univ. 1-15.

CEPUDER, P., KOCH, M., HAUER, G. & ZARTL, A., 2002. Experiences with different soil water measuring systems on diverse locations in Lower Austria. Biennial Report. Project number 302-D1-AUS-11184. Presented to the International Atomic Energy Agency, Vienna, Austria.

CHANDLER, D., SEYFRIED, M., MURDOCK, M. & MCNAMARA, J., 2004. Field calibration of water content reflectometers. *Soil Sci. Soc. Am. J.* 68, 1501-1507.

CZARNOMSKI, N., MOORE, N., PYPKER, T., LICATA, J. & BOND, B., 2005. Precision and accuracy of three alternative instruments for measuring soil water content in two forest soils of the pacific northwest. *Can. J. For. Res.* 35, 1867-1876.

DANE, J. & TOPP, G., 2002. Method of soil analysis, Part 4, Physical methods. SSSA., Book Series No 5, Madison, Wisconsin.

DANE, J., 1980. Comparison of field and laboratory determined hydraulic conductivity values. *Soil Sci. Soc. Am. J.* 44, 228-231.

DAVIDSON, J.M., STONE, L.R., NIELSEN, D.R. & LA RUE, M.E., 1968. Field measurement and use of soil water properties. *Water Resour. Res.* 5, 1312–1321.

DEXTER, A., 2004. Soil physical quality Part I. Theory, effects of soil texture, density, and organic matter, and effects on root growth. *Geoderma* 120, 201-214.

DEXTER, A., CZYZ, E., RICHARD, G. & RESZKOWSKA, A., 2008. A user friendly water retention function that takes account of the textural and structural pore spaces in soil. *Geoderma* 143, 243-253.

DIRKSEN, C., 1999. Soil physics measurements. GeoEcology paperback, Catena Verlag GMBH, 35447 Reiskirchen, Germany.

DOLPH, J., MARKS, D. & KING, G., 1992. Sensitivity of the regional water balance in the Columbia River basin to climate variability: Application of a spatially distributed water balance. *In:* R.J. Naiman, (ed.). Watershed management: Balancing sustainability and environmental change. Springer-Verlag, New York.

DROOGERS, P., STEIN, A., BOUMA, J. & DE BOER, G., 1998. Parameters for describing soil macroporosity derived from staining patterns. *Geoderma* 83, 293-308.

DURNER, W., 1994. Hydraulic conductivity estimation for soils with heterogeneous pore structure. *Water Resour. Res.* 30, 211-223.

ELHERS, W., WENDROTH, O. & DE MOL, F., 1995. Characterizing pore organisation by soil physical parameters. *Adv. Soil Sci*. 257-275.

EVETT, S.R. & STEINER, J.L., 1995. Precision of neutron scattering and capacitance type soil water content gauges from field calibration. *Soil Sci. Soc. Am. J.* 59, 961–968.

FARE, A., ALVA, A., NKEDI-KIZZA, P. & ELRASHIDI, M., 2000. Estimation of soil hydraulic properties of a sandy soil using capacitance probes and guelph permeameter. *Soil Sci.* 165, 768-777.

FIELD, J., PARKER, J. & POWELL, N., 1983. Comparison of field and laboratory measured and predicted hydraulic properties of a soil with macropores. *Soil Sci.* 138, 385-396.

FOLEY, J. & HARRIS, E., 2007. Field calibration of thetaprobe (ML2X) and ECH_2O probe (EC-20) soil water content sensors in black vertisol. *Aust. J. Soil Res.* 45, 233-236.

GEESING, D., BACHMAIER, M. & SCHMIDHALTER, U., 2004. Field calibration of a capacitance soil water probe in heterogeneous fields. *Aust. J. Soil Res.* 42, 289-299.

GLINSKI, J. & LIPIEC, J., 1990. Soil physical condition and plant roots. CRC Press, Boca Raton, Florida, USA.

GREEN, R., AHUJA, L. & CHONG, S., 1986. Hydraulic conductivity, diffusivity, and sorptivity of unsaturated soils: field methods. *In:* A. Klute, (ed.). Method of soil analysis, Part I, Physical and mineralogical methods Monograph No 9. ASA., Madison, Wisconsin, 771-798.

HILLEL, D., 1998. Environmental soil physics. Academic Press Inc, New York:

HILLEL, D., KRENTOS, V. & STYLIANOU, Y., 1972. Procedure and test of an internal drainage method for measuring soil hydraulic characteristics *in situ*. *Soil Sci.* 114, 395-400.

INTERNATIONAL ATOMIC ENERGY AGENCY (IAEA), 2008. Field estimation of soil water content-A practical guide to methods, instrumentation and sensor technology. Training course series No. 30. International Atomic Energy Agency, Vienna, Austria.

IPPISCH, O., VOGEL, H.J. & BASTIAN, P., 2006. Validity limits for the van Genuchten-Mualem model and implications for parameter estimation and numerical simulation. *Adv. Water Resour.* 29, 1780-1789.

INTERNATIONAL ORGANIZATION FOR STANDARDIZATION (ISO), 2009. Determination of the soil water retention characteristics. Version 1.3, FUTMON-soil moisture workshop, International Organization for Standardization, Geneva, 1-12. http:// www.iso.ch (accessed 23/07/2009).

JURY, W., GARDNER, W. & GARDNER, W., 1991. Soil physics. John Wiley & Sons, New York.

KELLENERS, T., ROBINS, D., SHOUSE, P., AYARS, J. & SKAGGS, T., 2005. Frequency dependence of the complex permittivity and its impact on dielectric constant sensor calibration in soil. *Soil Sci. Soc. Am. J.* 69, 63-66.

KERN, J.S., 1995. Evaluation of soil water retention models based on basic soil physical properties. *Soil Sci. Soc. Am. J.* 59, 1134-1141.

KLUTE, A. & DIRKSEN, C., 1986. Hydraulic conductivity and diffusivity: Laboratory methods. *In:* A. Klute, (ed.). Method of soil analysis, Part 1, Physical and Mineralogical methods. Agronomy Monograph No 9. ASA., Madison, Wisconsin, 687-700.

KLUTE, A. (ed.). 1986. Water retention: Laboratory methods. *In:* Methods of soil analysis, Part 1, Physical and mineralogical methods, Agronomy Monograph No 9. ASA., Madison, Wisconsin, 636-662.

KOSUGI, K., HOPMANS, J. & DANE, J., 2002. Parametric models. *In* J. Dane, & G. Topp, (eds.). Methods of soil analysis, Part 4, Physical methods. SSSA., Book series No 5, Madison, Wisconsin, 739-757.

KUTILEK, M. & JENDELE, L., 2008. The Structural porosity in soil hydraulic Functions. *Soil & Water Res.* 3, S7-S20.

KUTILEK, M., 2004. Soil hydraulic properties as related to soil structure. *Soil & Till. Res.* 79, 175-184.

KUTILEK, M., JENDELE, L. & KYRIAKOS, P., 2006. The influence of uniaxial compression upon pore size distribution in bi-modal soils. *Soil Till. Res.* 86, 27-37.

LEIJI, F., NEMES, A. & SCHAAP, M., 1999. The UNSODA unsaturated soil hydraulic database, Version 2.0. http:// www.ussl.ars.usda.gov/models/unsoda.htm (Accessed 13/05/2009).

LIBARDI, P., REICHARDT, K. NIESLEN, D., & BIGGAR, J., 1980. Simple field method for estimating soil hydraulic conductivity. *Soil Sci. Soc. Am. J.* 44, 3-7.

LIPIEC, J. & HATANO, R., 2003. Quantification of compaction effects on soil physical properties and crop growth. *Geoderma* 116, 107-136.

LORENTZ, S., GOBA, P., & PRETORIUS, J., 2001. Hydrological process research: Experiments and measurements of soil hydraulic characteristics. *Water Research Commission (WRC) Report* No 744/1/01, Pretoria.

LUXMOORE, R., 1981. Micro-, meso- and macroporosity of soil. *Soil Sci. Soc. Am. J.* 45, 241-285.

MARION, J., OR, D., ROLSTON, D., KAVVAS, M. & BIGGAR, J., 1994. Evaluation of methods for determining soil-water retentivity and unsaturated hydraulic conductivity. *Soil Sci.* 158, 1-13.

MARSHALL, T., 1959. Relations between water and soil. Farnham Royal: Commonwealth Bureau of Soils Harpeden. Technical Communication, 50.

MUALEM, Y., 1976. A new model for predicting the hydraulic conductivity of unsaturated porous media. *Water Resour. Res.* 12, 513-522.

NIMMO, J., 1997. Modelling structural influences on soil water retention. *Soil Sci. Soc. Am. J.* 61, 712-719.

OR, D. & WRAITH, J., 2002. Soil water content and water potential relationships. *In:* A. Warrick, (ed.). Soil physics companion. CRC press, Florida, 49-84

PAGLIAI, M., VIGNIZZI, N. & PELLEGRINI, S., 2004. Soil structure and the effect of management practices. *Soil tillage Res.* 79, 131-134.

PAIGE, G. & HILLEL, D., 1993. Comparison of three methods for assessing soil hydraulic properties. *Soil Sci.* 155, 175-189.

PALTINEANU, I. & STARR, J., 1997. Real-time soil water dynamics using multisensory capacitance probes: Laboratory calibration. *Soil Sci. Soc. Am. J.* 65, 1576-1585.

REEVE, M.J. & CARTER, A., 1991. Water release characteristic. *In:* K.A. Smith & C.E. Mullins (eds.). Soil analysis: Physical methods. Marcel Dekker, New York, 111-160.

REICHARDT, K., 1993. Unit gradient in internal drainage experiments for the determination of soil hydraulic conductivity. *Sci. Agric. Piracicaba,* 50(1), 151-153.

RICHARD, G., COUSIN, I., SILLON, J., BRUAND, A. & GUERIF, J., 2001. Effect of soil compaction on the porosity of a silty soil: influence on unsaturated hydraulic properties. *Euro. J. Soil Sci.* 52, 49-58.

RICHARDS, L.A., GARDNER, W.R. & OGATA, G., 1956. Physical processes determining water loss from soil. *Soil Sci. Am. Proc.* 20, 310-314.

SAITO, T., FUJIMAKI, H. & INOUE, M., 2008. Calibration and simultaneous monitoring of soil water content and salinity and capacitance and four-electrode probes. *Am. J. Environ. Sci.* 4, 683-692.

SEKI, K., 2007. SWRC fit- a nonlinear fitting program with a water retention curve for soils having unimodal and bimodal pore structure. *Hydrol Earth Syst. Sci.* 4, 407-437.

SEYFRIED, M., & MURDOCK, M., 2001. Response of a new soil water content sensor to variable soil, water content, and temperature. *Soil Sci. Soc. Am. J.* 65, 28-34.

SISSON, J., FERGUSON, A. & VAN GENUCHTEN, M., 1980. Simple method for predicting drainage from field plots. *Soil Sci. Soc. Am. J.* 44, 1147-1152.

SKIDMORE, E., 1985. Soil porosity. Lecture notes for College of Soil Physics. ICTP, Trieste.

TOWNEND, J. REEVE, M.J. & CARTER, A., 2001. Water release characteristic. *In:* K. A. Smith & C.E. Mullins (eds.). Soil analysis and environmental analysis: Physical methods. (2nd ed.). Marcel Dekker, New York, 95-140.

TSENG, P. & JURY, W., 1993. Simulation of field measurements of hydraulic conductivity in unsaturated heterogeneous soil. *Water Resour. Res.* 29, 2087-2099.

TULLER, M. & OR, D., 2002. Unsaturated hydraulic conductivity of structured porous media. A review. *Vadose Zone J.* 1, 14-37.

VAN BAVEL, C.H.M., STIRK, G.B. & BRUST, K.J., 1968. Hydraulic properties of a clay loam soil and the field measurement of water uptake by roots: interpretation of water content and pressure profiles. *Soil Sci. Soc. Am. Proc.* 32, 310-317.

VAN GENUCHTEN, M. & NIELSEN, D., 1985. On describing and predicting the hydraulic properties of unsaturated soils. *Ann. Geo-phys.* 3, 615-628.

VAN GENUCHTEN, M. T., 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Sci. Soc. Am. J.* 44, 892-898.

VAN GENUCHTEN, M., LEIJI, F. & LUND, L., 1992. Indirect methods for estimating the hydraulic properties of unsaturated soils. Proc. Int. Workshop. Riverside, California: University of California.

VAN GENUCHTEN, M., LEIJI, F. & YATES, S., 1991. The RETC code for quantifying functions of unsaturated soils. EPA/600/2-91/065.

WATSON, K.K., 1966. An instantaneous profile method for determining the hydraulic Conductivity of unsaturated porous materials. *Water Resour. Res.* 2, 709-715.

WEITZ, A., GRAUUEL, W., KELLER, M. & VELDKAMP, E., 1997. Calibration of time domain reflectomeetry technique using undisturbed soil samples from humid tropical soils of volcanic origin. *Water Resour. Res.* 33, 1241-1250.

ZHANG, S., LOVDAHL, L., GRIP, H. & TONG, Y., 2007. Soil hydraulic properties of two loess soils in China measured by various field-scale and laboratory methods. *Catena* 69, 264-273.

CHAPTER 3

Laboratory characterization of the water retention characteristics of a Bainsvlei form soil and a Tukulu form soil

3.1 Introduction

3.1.1 Motivation

The soil water retention characteristic describes the relationship between the series of the water contents of a soil (θ) from very wet to very dry, and the matric suction (h) at which the water is held at each θ value. It is therefore sometimes described as the θ-h relationship. It is a physical soil property which describes the soil porous system. It depends basically on soil structure, texture, organic matter content, and bulk density. It will therefore vary both vertically (diagnostic horizons/layers in the profile) and horizontally in any field. Stratified sampling according to diagnostic horizons or specific layers is a prerequisite to determine the overall hydrological behaviour of a soil profile (ISO, 2009). Because of this and the importance of the hydrological behaviour of soils to agriculture, forestry, hydrology, engineering and pollution, research concerning the soil water retention characteristics of soil horizons is important. Information on soil water retention is needed for example: (i) to determine plant available water in the soil (the portion of water that can be readily absorbed by plants roots) (van Rensburg, 1988); (ii) to evaluate soils for irrigation purposes; (iii) to estimate the soils' pore size distribution (Kutilek, 2004); (iv) to check changes in the structure of a soil, e.g. caused by tillage, mixing of soil layers (Kutilek, 2004); (v) to predict other soil physical properties (e.g. hydraulic conductivity) (Mualem, 1976; van Genuchten, 1980); (vi) to provide inputs in most water balance and hydrological models (Bennie *et al.,* 1994). Consequently, information about water retention characteristics can be useful for farmers and governments as a planning tool for development and investment strategies.

Several *in situ* and laboratory methods for determining the θ-h relationship are now available. *In situ* methods are considered to be more representative than laboratory methods for determining water retention characteristics (Marion *et al.*, 1994). This is largely because of the larger volume of soil involved. The former procedure also avoids changes in pore size distribution and continuity which may be caused during the collection and transportation of undisturbed samples. Such changes may substantially influence the measured θ-h relationship, particularly near the wet end (h < 10 kPa) where most water flow occurs (Kutilek, 2004). Reliable results can, however, be obtained if samples are carefully collected and handled. The routine use of field methods is often inhibited by cost and time considerations, including the cost of instrumentation, especially when large areas have to be characterized for their hydraulic properties. The θ-h relationships are therefore frequently determined in the laboratory. As large numbers of representative, carefully collected samples from different locations can be dealt with simultaneously in the laboratory, this method is considered a valuable alternative to *in situ* measurements (Marion *et al.,* 1994).

The θ-h relationship is strongly dependent on soil pore geometry. To characterize the soil pore system, it is useful to consider the θ-h relationship as consisting of two components, the structural and textural domains (Nimmo, 1997). The structural domain consists of the pore space between aggregates, while the textural domain consists of the pores within aggregates or within blocks of soil if aggregates are not present (Dexter *et al.,* 2008). The separation of pore space into these two categories produces what are called bi-modal pore size distributions (Dexter *et al.,* 2008).

The separation of porosity into textural and structural domains is determined either by direct or indirect methods. Image analysis of soil thin sections is a direct method whereby the pores are visually described and in some instances quantified (Pagliai *et al.*, 2004). The most common indirect methods are: firstly, by subdividing the θ-h curve into two portions using empirical pore class limits (fixed boundary method); secondly, via the derivative curve of the θ-h relationship. The derivative curve is defined here as the relationship between the function dθ/dlog(h) and log(h) (Kutilek, 2004). The curve provides information about the frequency distribution of pores of different sizes, a peak indicating a high value and a trough a low value. It is convenient to express θ relative to saturation (S) where $S = \theta/\theta_s$ and θ_s is the water content at saturation. The function used by Kutilek therefore becomes dS/dlog h.

The existence of bi-modal pore size distributions has consequences for water retention modelling (Kutilek, 2004). Using the bi-modal approach offers great flexibility and allows detailed description of the θ-h relationship near saturation (Coppola *et al.*, 2009). There is a wide body of literature relevant to the separation of pores into structure and textural domains based on the derivative curve of the θ-h relationship (Kutilek, 2004; Kutilek *et al.,* 2006; Kutilek & Jendele, 2008; Dexter *et al.,* 2008). Although the derivative curve procedure is frequently used it is more complicated than using empirical limits of pore classes (e.g. Luxmoore, 1981; Ahuja *et al.,* 1984). Both procedures will be used in this study.

Many South African soils are structured to some extent, with the degree of structural development varying through individual diagnostic horizons down to the parent material. The proposal by Kutilek (2004) for making a distinction between the matrix and structural domains of soil porosity provides a possible avenue towards the quantification of soil structure. Soil structure has traditionally been described using subjective, qualitative terms by pedologists when making soil profile descriptions. Although this procedure produces valuable results it is fraught with all the weaknesses associated with all subjective qualitative observations in science. A reliable soil profile description together with the associated analytical results and location coordinates is considered to be the best pedotransfer function that a soil scientist can provide for agriculture, hydropedology, environmental studies, engineering, and many other studies in which soil is involved. It would therefore be very valuable if a procedure could be developed to quantify soil structure in a reliable way. Despite the extensive international studies like those of Kutilek (2004) and Dexter *et al.* (2008) no detailed systematic studies in this connection have yet been conducted in this connection for South African soils. Local related studies regarding the applicability of such procedures for South African, and all African soils, especially in view of the need to promote food production in Africa, would therefore be valuable.

3.1.2 Objectives of the study

The objectives of this study were: (i) to describe and classify the morphological, physical and chemical characteristics of a Bainsvlei form soil and a Tukulu form soil in detail by means of profile descriptions and relevant analytical data; (ii) to determine the soil water retention characteristics (θ-h relationships) for the different horizons of the two soils; to described the θ-h relationship using the van Genuchten (1980) model; (iii) to relate the θ-h relationships of the different horizons to their morphological and other characteristics such as structure, texture and bulk density.

3.2 Procedure

3.2.1 Site location and pedological characteristics

The soil samples used in this study were collected from soil profiles at two experimental field stations of the University of the Free State; at Kenilworth (29^{o}01’ 00’’ S, 26^{o}08’ 00’’ E, altitude 1354 m) and at Paradys (29^{0}13’ 25’’ S, 26^{o}12’ 08’’ E, altitude 1417 m). The Kenilworth soil is classified as a Bainsvlei form, *Amalia family* (Soil Classification Working Group, 1991). At Paradys, the soil is a Tukulu form, *Dikeni family* (Soil Classification Working Group, 1991). The Kenilworth Bainsvlei soil is a very suitable agricultural soil in this semi-arid climate as the solum is deep (2000 mm) and drains freely while the plinthic horizon dams water within the lower part of the profile which is within range to be tapped by plant roots during frequent dry spells (Bennie *et al.,* 1994). The Paradys Tukulu soil has some limitations for crop production using conventional techniques. The solum (A and B horizons) is shallow (500 mm) and is underlain by a prismatic, smectite rich, hydromorphic layer that shows signs of wetness. The latter can be an advantage using an appropriate crop production technique such as in-field rainwater harvesting (Hensley *et al.*, 2000; Fraenkel, 2009) and appropriate drought resistant deep-rooted crops. The objective of choosing the two sites was to obtain a range of normal agricultural soils that occur in this semi-arid region.

3.2.1.1 Profile description

The soil profiles were described following the procedure specified by the South African Agricultural Research Council-Institute for Soil Climate and Water. The profiles were classified according to the Soil Classification a Taxonomic System for South Africa (Soil Classification Working Group, 1991). The profiles at both sites were described *in situ* from a fresh face in soil pits.

3.2.2 Soil sampling and storage

Three undisturbed core samples with a volume of 641 cm^3 (10.3 cm inner diameter and 7.7 cm height) were collected from each selected horizon from both sites. A core sampler was mounted vertically on the soil surface and forced in using a 12 ton hydraulic jack to ensure sampling with minimum disturbance (Grossman & Reinsch, 2002). Immediately after taking the core samples both ends were trimmed carefully and sealed with masonite boards to prevent any soil disturbance during transportation. The undisturbed samples were used for the determination of bulk density and the θ-h relationship at low suctions (< 100 kPa). Disturbed soil samples (28 cm^3) were used for the determination of the θ-h relationship at higher suctions (> 100 kPa) using a pressure plate apparatus, and for soil analyses. For the Bainsvlei profile samples were collected at the midpoint of each of the six selected horizons (0-250, 250-420, 420-700, 700-1200, 1200-1450, and 1450-1850 mm) identified when making the profile description (Table 3.1); and similarly for the Tukulu soil the horizons were 0-270, 270-500, 500-800 mm (Table 3.2). The samples were transported carefully to avoid disturbance. The samples were stored in the laboratory at room temperature until time for measurement.

3.2.3 Soil analysis

Each of the selected diagnostic horizons were chemically and physically analyzed for exchangeable cations (Ca, Mg, Na, K), $pH_{(water)}$, and particle size distribution, using procedures described by The Non-Affiliated Soil Analysis Work Committee (1990).

3.2.4 Measurement of water retention characteristics

3.2.4.1 Sample saturation

The bulk density (D_b) was determined after drying the samples at 105^oC for 24 hrs. To determine the θ-h relationship using the undisturbed soil samples the first step was to saturate the soil cores using vacuum saturation chambers. Saturation chambers are vessels filled with water in which a soil sample in a retaining ring can be inundated for saturation. Deaired water was used for saturation. Water was deaired by continual stirring in a container attached to a vacuum source ± 60 kPa. The deaired water was then let in gradually into a companion chamber where deaired soil samples were placed, until the water level was just below the top of the soil samples. It was found that 24 hours was enough to reach saturation. The gravimetric saturated water content of the samples was then determined by weighing the sample immediately after taking it out of the saturation chamber. The volumetric saturated water content (θ_s) value for each sample was later calculated by multiplying the gravimetric water content by the D_b value.

3.2.4.2 Desorption measurements

The θ-h relationships were measured using a combination of desorption techniques in order to provide a detailed description of the whole water retention characteristic. The hanging water column tension cup method (Dirksen, 1999) was used at lower suctions 0 - 10 kPa, and starting as close to saturation as possible to try to accurately identify the air entry suction. The material used in the tension cups was diatomaceous earth. The pressure plate apparatus (Jury *et al.*, 1991) was used to define the water retention characteristic at higher suctions; using undisturbed core samples at suctions between 10 and 100 kPa and disturbed samples for suctions between 100 and 1500 kPa.

The samples on the hanging water column apparatus were equilibrated until no more outflow occurred. The ceramic plates with their air entry pressures and corresponding equilibration times were as follows: for h = 10 - 100 kPa, 100 kPa plates and 5 days were used; for h = 100 - 300 kPa, 300 kPa plates and 7 days were used; for 500, 1000, and 1500 kPa, 1500 kPa plates and 10 days were used. After equilibration, the samples were weighed to determine the water content corresponding to the suction or pressure applied. The gravimetric water content was converted to volumetric water content by multiplying it by the relevant D_b value.

A pore size distribution characteristic was determined from the θ-h relationship using 10 kPa as an empirical boundary between structural and textural domains. The total pore space was taken as the volumetric water content at saturation (0 kPa). The pore sizes were categorized as, structural domain (θs_2) and textural domain (θs_1), with equivalent pore radii of >15 μm and < 15 μm respectively, based on relevant suggestions by Marshall (1959), Skidmore (1985) and Kutilek (2004), and within the ranges proposed by Luxmoore (1981) presented in Table 3.6. The subdivision used defines pores draining between 0-10 kPa, and 10-1500 kPa, as belonging to the structural and matrix domains, respectively. The water held by the larger pores, i.e. the structural domain pores, was computed by subtracting the volumetric soil water content at 10 kPa from total porosity, with the remaining water logically being allotted to the textural domain pores.

The derivative curve procedure used to separate the porous system into structural and textural domains is described in Section 3.4.3.2.

The analyses described below were performed using average water retention characteristics of the replicated samples collected from each selected horizon. Thus, the water retention characteristics of nine horizons were determined, each result being the mean of three replicates.

3.2.5 Mathematical description of the soil water retention characteristic

The van Genuchten model (Equation 3.1) was fitted to each of the nine observed water retention characteristics using the parameter optimization computer program RETC (van Genuchten *et al.*, 1991). The parametric θ-h model described by van Genuchten (1980) is written as:

$$S_e = \frac{(\theta - \theta_r)}{(\theta_s - \theta_r)} = \left[\frac{1}{1+(\alpha h)^n}\right]^m \qquad 3.1$$

Where the subscripts r and s denote residual and saturated volumetric water content, respectively, S_e is the dimensionless water content, α, n, and m are empirical parameters that determine the shape of the curve. The value θ_s was fixed as the measured water content at saturation and θ_r was fixed at the water content at 1500 kPa suction, consistent with the suggestion of van Genuchten (1980). The RETC computer program was then used to determine the van Genuchten parameters (α & n) using the measured values of θ and h; as suggested by van Genuchten m = 1- 1/n.

3.2.6 Statistical analysis

Willmott (1982) statistics were used to assess the deviation between measured θ-h values and those fitted by the van Genuchten model (VG) (van Genuchten, 1980). The statistical parameters used for the assessment were: coefficient of determination (R^2); index of agreement (D); root mean square error (RMSE) and the relationship to its unsystematic component $RMSE_u$, i.e. $RMSE_u$/RMSE.

3.3 Results and discussion

3.3.1 Profile attributes of the soils sampled

Detailed soil profile descriptions are presented in Tables 3.1 and 3.2 for the Bainsvlei and Tukulu soils respectively. A summary of the chemical and physical properties of two soil profiles are presented in Tables 3.3 and 3.4, respectively.

All layers of the Bainsvlei soil had very low silt contents, ranging from 4 to 5.3%, similar sand contents (>67% in each case), and clay contents of around 8 to 22%. Generally bulk density is fairly uniform down the soil profile, ranging from 1.65 to 1.68 Mg m^{-3}. The massive structure is relatively uniform throughout the profile. The Bainsvlei form soil profile can be described as being relatively homogenous with depth. The Tukulu soil is characterized by a high clay content (55%) in the prismatic structured C horizon (500-800 mm), compared to the sandy loam A and sandy clay loam B horizons with a clearly defined B/C transition. The D_b increases with depth, ranging from 1.67 to 1.71 Mg m^{-3}. These variations in bulk density and clay content influence the water retention properties, especially in the wet region (Figure 3.3).

Table 3.1 Profile description of the Bainsvlei form soil

Map/photo :	2926 Bloemfontein	Soil form and family:	Bainsvlei *Amalia*
Latitude + Longitude:	29° 1' 00''/26° 08' 00''	Surface rockiness:	None
Altitude:	1354 m	Occurrence of flooding:	None
Terrain unit:	Lower foot slope	Wind erosion:	Slight wind
Slope:	1%	Water erosion:	None
Slope shape:	Straight	Vegetation/Land use:	Agronomic field crops
Aspect:	North-west	Water table:	None
Microrelief:	None	Described by:	M. Hensley & J. Chimungu
Parent Material Solum:	Origin single, aeolian	Date described:	14/06/09
Underlying Material:	Sandstone (Feldspathic)	Weathering of underlying material:	Weak physical to moderate chemical
		Alteration of underlying material:	Ferruginised

Horizon	Depth (mm)	Description	Diagnostic horizon
A	0-250	Moist state; dry colour: yellowish red (5YR5/6); moist colour: reddish brown (5YR4/4); texture: fine loamy sand; structure: apedal massive; consistence: friable; few fine normal pores; water absorption: 1 second; few roots; gradual smooth transition.	Orthic
B1	250-420	Moist state; dry colour: red (2.5YR4/8); moist colour: red (2.5YR4/6); texture: fine sandy loam; structure: apedal massive; consistence: friable; few fine normal pores; water absorption: 1 second; few roots; gradual smooth transition.	Red apedal
B2	420-700	Moist state; dry colour: yellowish red (5YR5/8); moist colour: red (2.5YR4/6); texture: fine sandy loam; few fine faint black illuvial humus mottles; structure: apedal massive; consistence: friable; common fine normal pores; water absorption: 1 second; few roots; gradual wavy transition.	Red apedal
B3	700-1200	Moist state; dry colour: yellowish red (5YR4/6); moist colour: reddish brown (5YR4/4); texture: fine sandy clay loam; common fine faint black illuvial humus mottles; structure: apedal massive; consistence: slightly firm; common fine normal pores; water absorption: 1 second; clear wavy transition.	Red apedal
B4	1200-1450	Moist state; dry colour: strong brown (7.5YR5/8); moist colour: strong brown (7.5YR5/6); texture; fine sand; few fine faint black illuvial humus mottles; structure: apedal massive; consistence: friable; water absorption: 1 second; few roots; clear wavy transition.	Non diagnostic; yellow brown Aeolian sand
B5	1450-1850	Moist state; moist colour: strong brown (7.5YR4/6); texture: fine sandy loam; many medium distinct grey and yellow reduced iron oxide mottles; many medium distinct red and black oxidised iron oxide mottles; structure: apedal massive; consistence: friable; water absorption: 1 second(s); few roots; gradual smooth transition.	Soft plinthic
C	1850-2220	Similar to B5 with patches of weathered feldspathic sandstone; colour of mottles similar to B5 but more prominent.	

Table 3.2 Profile description of the Tukulu form soil

Map/photo :	2926 Bloemfontein	Soil form and family:	Tukulu *Dikeni*
Latitude + Longitude:	29^0 13' 25''/26^0 12' 08''	Surface rockiness:	None
Altitude:	1417 m	Occurrence of flooding:	None
Terrain unit:	Midslope	Wind erosion:	Slight wind
Slope:	1%	Water erosion:	None
Slope shape:	Straight	Vegetation/Land use:	Agronomic field crops
Aspect:	South	Water table:	None
Microrelief:	None	Described by:	M. Hensley & J. Chimungu
Parent Material Solum:	Origin binary, aeolian, solid rock	Date described:	12/06/09
Underlying Material:	Sandstone (Feldspathic)	Weathering of underlying material:	Moderate physical, moderate chemical
		Alteration of underlying material:	Ferruginised

Horizon	Depth (mm)	Description	Diagnostic horizon
A	0-270	Moist state; dry colour: reddish brown (2.5YR5/4); moist colour: reddish brown 2.5YR4/4; texture: fine sandy loam; structure: apedal massive: consistence: friable; few fine pores; common roots; gradual transition.	Orthic
B1	270-500	Moist state; dry colour: reddish brown (2.5YR5/4); moist colour: reddish brown (2.5YR4/4); texture: fine sandy clay loam; structure: apedal massive becoming weak subangular blocky towards transition: consistence: friable; few fine pores; few clay cutans; very few fine pore; common roots; clear, tonguing transition.	Neocutanic
C1	500-800	Moist state; dry colour. Dark greyish brown 2.5YR4/2; moist colour: grey 2.5Y5/0; texture: clay; common distinct grey and yellow reduced iron oxide mottles: few prominent black oxidised iron oxide mottles; structure: prismatic; consistence: firm; few slickensides; common clay cutans; very few roots; gradual transition.	Unconsolidated material, with signs of wetness
C2	800-±1350	Allows water to enter, hence the presence of roots. It chops out of the profile relatively easily up to ±1350 mm, getting harder towards the bottom and probably very impermeable slightly deeper.	

Table 3.3 Summary of chemical and physical characteristics of a Bainsvlei form soil.

	A	B1	B2	B3	B4	B5
Physical properties						
Coarse sand (2 - 0.5 mm) (%)	0.4	0.3	0.3	0.3	0.3	0.6
Medium sand (0.5 - 0.25) (%)	7.1	5.2	5.4	4.1	3.3	6.0
Fine sand (0.25 - 0.106 mm) (%	61.4	55.1	53.8	44.9	64.3	48.3
Very fine sand (0.106 - 0.53mm) (%)	16.8	15.1	15.5	18.0	17.3	17.0
Silt (%)	4.0	4.0	6.0	8.0	4.0	6.0
Clay (%)	8.0	18.1	18.0	22.1	8.1	20.1
Bulk density (Mg m^{-3})	1.66	1.68	1.66	1.67	1.68	1.67
Chemical properties						
pH(water)	5.2	5.1	6.3	6.1	6.7	6.5
Ca ($cmol_c.kg^{-1}$)	7.2	15.3	11.0	16.2	11.3	11.1
Mg ($cmol_c.kg^{-1}$)	4.3	10.2	10.4	9.4	11.7	10.3
K ($cmol_c.kg^{-1}$)	2.5	3.6	2.2	2.6	2.1	2.0
Na ($cmol_c.kg^{-1}$)	4.1	3.8	2.7	2.5	2.4	2.7
Clay –S value	-	Eutrophic*	-	-	-	-

**clay-S value is calculated as follows:* $\frac{\sum Exchangeable\ cations x100}{\%\ clay}$

Table 3.4 Summary of chemical and physical characteristics of a Tukulu form soil.

	A	B1	B2
Physical properties			
Coarse sand (2-0.5 mm) (%)	3.1	1.6	1.3
Medium sand (0.5-0.25) (%)	2.8	2.7	1.6
Fine sand (0.25-0.106 mm) (%)	42.4	42.2	23.2
Very fine sand (0.106-0.53mm) (%)	25.6	19.6	9.9
Silt (%)	6.0	3.0	8.0
Clay (%)	18.6	28.9	54.8
Bulk density (Mg m^{-3})	1.67	1.68	1.71
Chemical properties			
pH(water)	5.8	6.1	6.5
Ca ($cmol_c.kg^{-1}$)	35.3	40.8	62.7
Mg ($cmol_c.kg^{-1}$)	18.8	20.8	54.7
K ($cmol_c.kg^{-1}$)	12.8	6.8	12.9
Na ($cmol_c.kg^{-1}$)	2.4	2.5	3.4

3.3.2 Water retention characteristics

The water contents at different suctions for the different horizons in the two soil profiles are presented in Appendix (A & B). The data are averages of three replicates. Standard deviations are given in parentheses. Plots of the average θ-h relationships are shown in Figures 3.1 and 3.2 for the Bainsvlei and Tukulu form soils, respectively.

Figure 3.1 (for the Bainsvlei form soil) shows that the shape of the curves for the different horizons are fairly similar. This was expected in view of the textural, structural and mineralogical homogeneity of the profile. There were slight differences between the curves, particularly at high suctions. The reason for this is that at these suctions particle size distribution is the dominant factor controlling water retention, and there is slight variation in clay content down the profile (Table 3.3).

For the Tukulu form soil the saturated water content (θ_s), decreased very slightly from the A to the B to the C horizon, reflecting the associated small increases in bulk density. The θ-h relationships of A and B horizon show a gentle slope from 0 up to 1 kPa and a relatively steep slope from 1 to 1000 kPa. The θ-h relationship of C-horizon is very different and more or less fairly flat up to 10 kPa, followed by a gentle slope thereafter. This shape can be attributed to the high clay content (Table 3.4) in this soil horizon (Hillel, 1982). In clay soils, since there is large proportion of micropores, these pores hold most of the water resulting in a more gradual decrease in soil water content with increase in suction than in coarse textured horizons (Hillel, 1998), like the A and B horizons here. This also reflects the more strongly developed structure (prismatic, Table 3.2) caused by the swelling clay, and the slightly higher bulk density (Table 3.4).

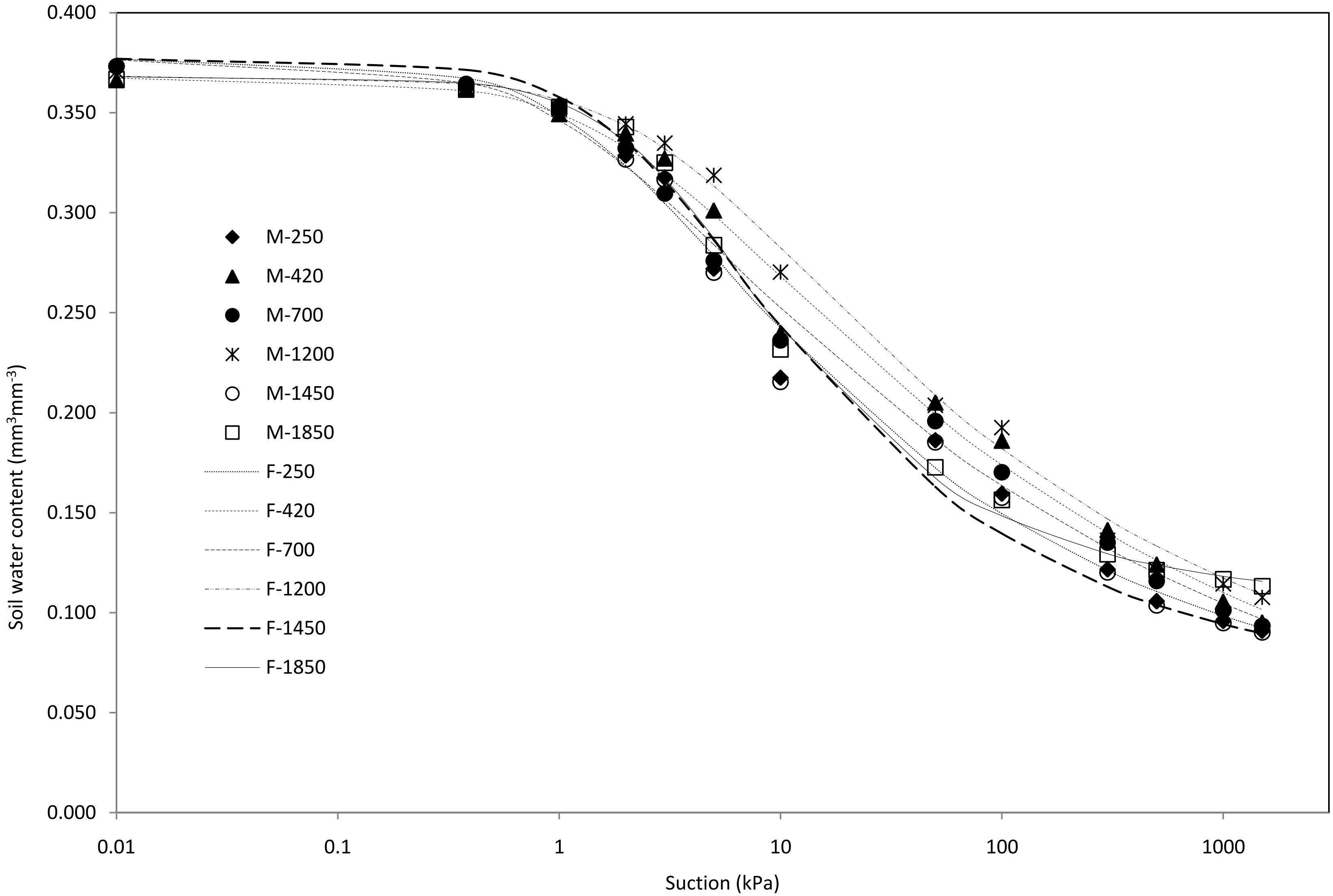


Figure 3.1 Comparison of measured (M) and fitted (F) retention curves for the Bainsvlei form soil.

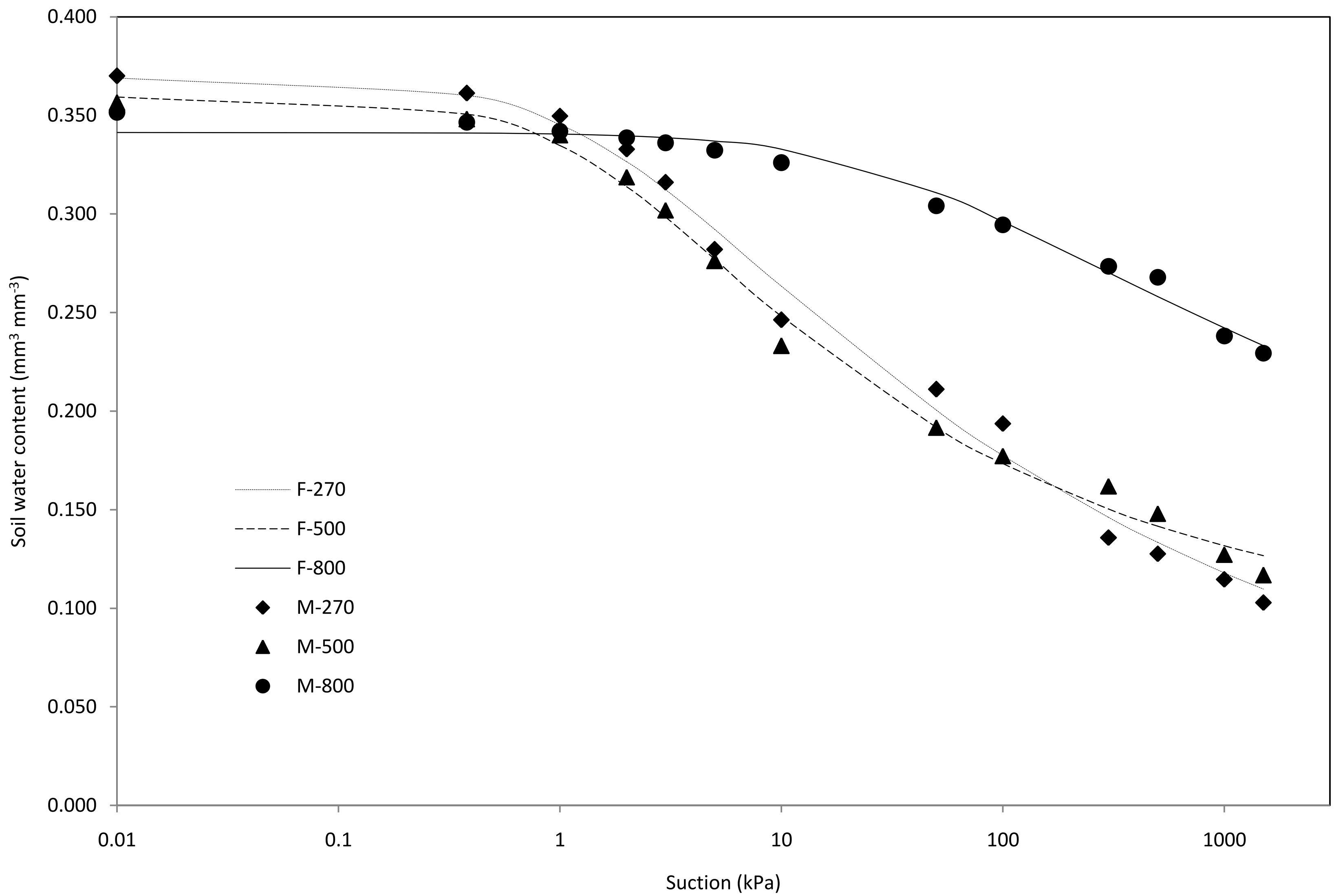


Figure 3.2 Comparison of measured (M) and fitted (F) retention curves for the Tukulu form soil.

The measured θ-h relationships are compared with fitted curves by the VG model in Figures 3.1 and 3.2. It is clear from the Figures that the VG model performed quite well for each retention data set. This fact is also clear from the statistical results presented in Table 3.5. Relevant are: the high coefficients of determination (R^2); high index of agreement (D); small root mean square error (RMSE) values; absence of bias indicated by the high $RMSE_u$/RMSE values throughout. There are only small differences in the VG parameters (α & n) values for all horizons within the Bainsvlei soil profile (Table 3.5). This is to be expected in view of the textural, structural and mineralogical homogeneity of the profile. However, there is a large difference between the α and n values for the A and B horizons of the Tukulu soil compared to the C-horizon. The cause is presumably the large difference between the particle size distribution, structure and clay mineralogy of the C compared to the A and B horizons.

In spite of the overall satisfactory fit by the VG model there is one h value for which the fit is almost invariably unsatisfactory for all nine horizons. It is the all important 10 kPa suction value (see Figures 3.1 and 3.2) which probably approximates in many soils the boundary between the 'structural porosity' and the 'textural porosity' (Kutilek, 2004; and Figures 3.3 and 3.4). It is very likely that a statistical analysis comparing the measured and fitted θ values for this h value would give a significant difference. Particularly for hydrological modelling the amount water held at suctions above h = 10 kPa is important since this is the main part of the soil water that can drain out of a profile and therefore contribute to groundwater recharge, stream flow and leaching of chemicals.

To summarize, satisfactory overall fitting of the data was obtained by the VG model. These findings are consistent with the results of van Genuchten & Nielsen (1985) which showed that the VG model gives a good description of measured retention data for a large number of soils.

Table 3.5 Parameters of the van Genuchten model and the results of the Willmott statistical test to describe the extent to which the model fitted the measured data.

Site	Soil depth mm	θ_r mm^3mm^{-3}	θ_s mm^3mm^3	α	n	R^2	D-index	RMSE	$\frac{RMSEu}{RMSE}$
Kenilworth	0-250	0.091	0.374	0.58	1.281	0.991	0.995	0.009	1.00
	250-420	0.095	0.366	0.43	1.199	0.990	0.998	0.009	1.00
	420-700	0.093	0.373	0.73	1.194	0.995	0.997	0.006	1.00
	700-1200	0.108	0.371	0.28	1.235	0.996	0.989	0.006	1.00
	1200-1450	0.098	0.377	0.58	1.280	0.991	0.998	0.009	1.00
	1450-1850	0.113	0.366	0.30	1.528	0.997	0.994	0.005	1.00
Paradys	0-270	0.102	0.369	0.60	1.177	0.991	0.996	0.008	1.00
	270-500	0.116	0.356	0.62	1.272	0.993	0.997	0.006	1.00
	500-900	0.229	0.351	0.04	1.095	0.980	0.998	0.005	1.00

3.3.3 Separation of porosity into textural and structural domains

3.3.3.1 Based on empirical pore class limits (fixed boundary)

For the θ-h relationships of the Bainsvlei and Tukulu form soils, the hypothetical structural domain of porosity was separated from the textural domains (Figures 3.3) at a suction of 10 kPa (Marshall, 1959; Kutilek, 2004). The limits adopted in this study are within the range suggested by Luxmoore (1981) and are presented in Table 3.6. It is significant to note that Luxmoore uses the terms 'drainage' and 'gravitational driving force for water dynamics' for the suction range in which 10 kPa occurs. This indicates that he thinks that the h value separating the water which will drain freely from a soil profile and that which will be retained against the force of gravity (i.e. approximating the field drained upper limit or DUL) should be in the range 0.3 to 30 kPa. Although the use of fixed boundaries has been criticised by Kutilek *et al.* (2006) and Kutilek & Jendele (2008) on a theoretical basis, the simplicity and current applicability of this procedure it has been adopted in this study.

Table 3.6 Suggested characteristics of three soil porosity classes (Luxmoore, 1981)

Soil porosity class	Suction ranges (kPa)	Pore diameter range (µm)	Dominant phenomena
Micro	>30	<10	Evapotranspiration; matric pressure gradient for water redistribution
Meso	0.3 to 30	10 to 1000	Drainage; hysteresis; gravitational driving force for water dynamics
Macro	< 0.3	>1000	Channel flow through profile from surface ponding and/or perched water table

What is immediately apparent in Figure 3.3 is that, although the dominant structure in all horizons is massive, except for the Tukulu C-horizon, there are considerable differences in the shapes of structural domains. These differences can be attributed here mainly to variation in particle size distribution. It is logical and also generally stated in the literature that coarse textured soils have larger pores due to the large individual particles, whereas the finer textured soils have small pore sizes. Thus, because the large pores drain quickly (< 10 kPa) these are accounted for here as contributing to the structural domain of porosity. The term ‘structural’ used by Kutilek (2004) and several authors seems rather unsatisfactory. A better description may be ‘easily drainable porosity’.

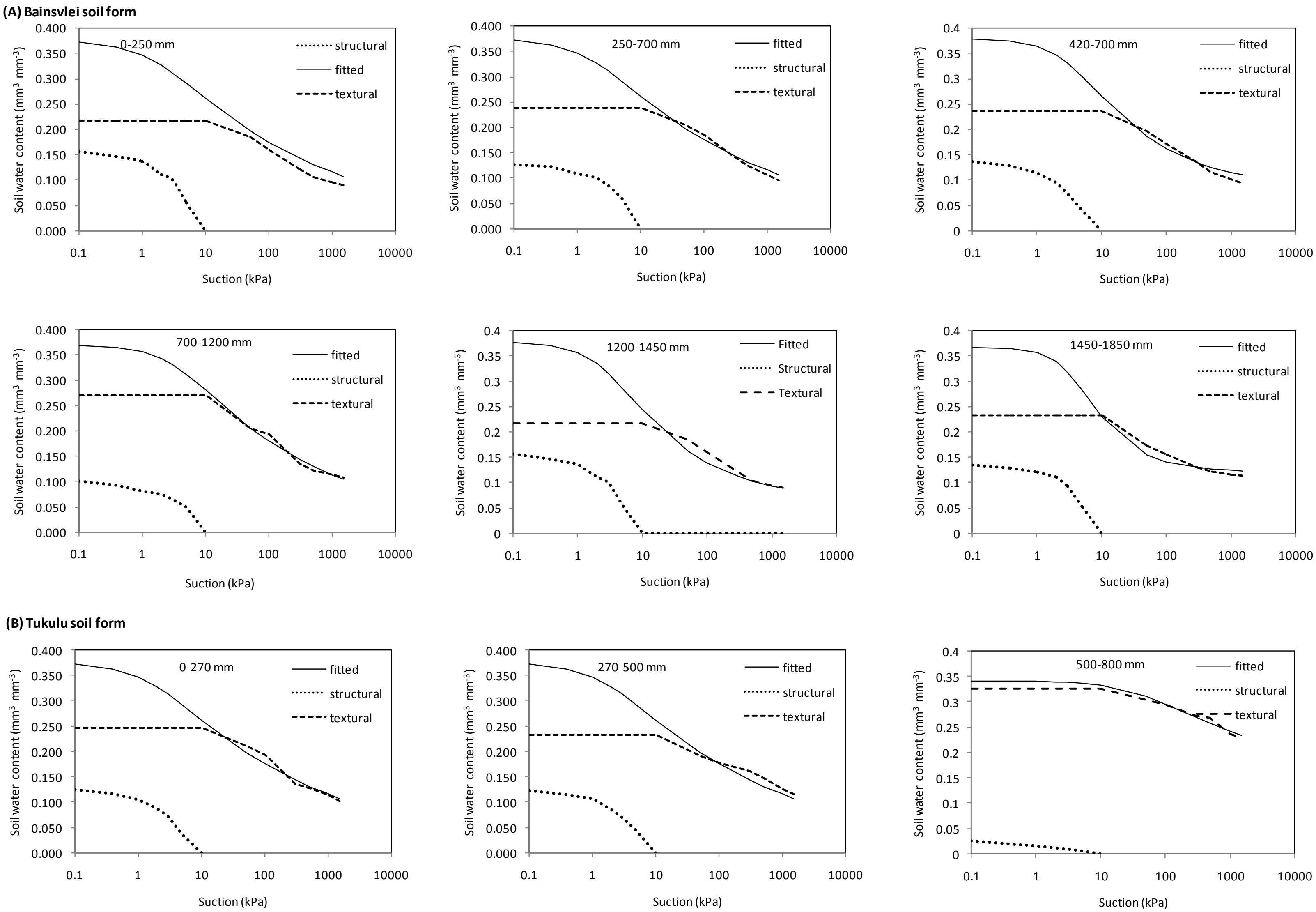


Figure 3.3 Separation of the porosity into textural and structural domains computed assuming h = 10 kPa (A) Bainsvlei and (B) Tukulu soil form.

Quantitative results from (Figure 3.3) are presented in Table 3.7 giving the two domains of porosity using h = 10 kPa as a fixed boundary. Based on the well known capillary function relating mean pore radius to suction (Section 3.4.3.2) one can conclude from the results for the nine soil horizons, that there were in all the horizons, excepting the C horizon in the Tukulu soil, a large fraction of pores that are sufficiently large (> 15 µm) to drain easily. These pores have been defined as belonging to the structural domain of porosity by Kutilek (2004). A general downwards trend of structural domain of porosity with increasing clay content can been seen in Table 3.7. Horizons with a low clay content have high percentage of total porosity occupied by these coarser pores, for example for the Bainsvlei form soil the 0-250 mm and 1200-1450 mm horizons have 8% clay and the structural domain ranges between 41 to 42% whereas in the 700-1200 horizon with 22% clay the structural domain decreases to 27%. The effect of clay is better understood by examining the more clayey 500-800 mm C-horizon of the Tukulu form soil where the proportion of structural domain pores is very low (7%) compared to the A and B horizons with 34%. It is interesting to note that Gardner (1988) showed that for clay soil structural porosity ranges from 0.03 to 0.07 $mm^3\ mm^{-3}$, which is consistent with our findings in Table 3.7 for the Tukulu C-horizon. In general textural porosity dominates the porosity in the soils studied here, i.e. generally > 65% of total porosity. This is consistent with findings of Elhers *et al.* (1995) that in soil with weak or little aggregation, the pore system is dominated by textural pores. From the above discussion, it is clear that in most soils structural pores are present to some extent regardless of the type of structure.

These results support the use of a bimodal description of the θ-h relationship because of its flexibility with regard to describing the significant influence of structure as well as texture of a soil on its pore size distribution and therefore on its hydraulic properties. Thus, for many soils these relationships cannot be accurately predicted on the basis of texture alone, because it is the dominant factor at high suctions only. It is generally stated in literature that to effectively and reliably predict hydraulic properties, the predominant effects of soil hydraulic behaviour near saturation should be accurately accounted for, hence the importance of the ‘structural’ or ‘easily drainable’ porosity (Coppola *et al.,* 2009).

Table 3.7 Quantitative values of structural and textural porosity from Figure 3.3.

Soil form	Depth (mm)	V_p ($mm^3 mm^{-3}$)	θ_{s2}(mm^3 mm^{-3})	θ_{s1}(mm^3 mm^{-3})	% θ_{s2} of V_p	Clay content (%)
Bainsvlei	0-250	0.374	0.157	0.217	42	8
	250-420	0.366	0.127	0.239	35	18
	420-700	0.375	0.137	0.238	37	18
	700-1200	0.372	0.100	0.271	27	22
	1200-1450	0.373	0.152	0.221	41	8
	1450-1850	0.368	0.135	0.232	37	20
Tukulu	0-270	0.368	0.124	0.246	34	19
	270-500	0.366	0.123	0.233	34	29
	500-800	0.355	0.026	0.326	7	55

V_p = total porosity: θ_{s2}= structural porosity: θ_{s1} = textural porosity

3.3.3.2 Using the derivative curve technique

An alternative way of separating structural pores from textural pores is to study the shapes of the derivative curves presented in Figure 3.4. The derivative curves (i.e. dS/d log(h) vs log h) have been calculated and results are presented graphically in Figure 3.4. The graphs provide information about the frequency distributions of different pore sizes for the samples reported in Figures 3.1 and 3.2. The value of h is inversely proportional to the radius of the largest pores holding water at a particular suction. This is according to the well known capillary function r = C/h, where r is the pore radius of the pores emptied at the particular suction value h, and C is a constant determined by the relationship $2\gamma\cos\beta/\sigma_w g$ where: γ is the surface tension between water and air; β is the contact angle between water and the solid surface of the soil matrix; σ_w is the density of water; g is the acceleration due to gravity. When r is expressed in µm and h in cm the value of C is 0.1490 cm^2 (Seki, 2007). Kutilek *et al.* (2006) suggest that the boundary between the structural and textural domain from the derivative curves like those presented in Figure 3.4 can be represented by the minimum suction value, and the corresponding equivalent pore radius, between two peaks. They designate this suction value as h_a. These minima occur in four of the horizons of the Bainsvlei soil (0-250 mm; 250-420 mm; 420-700 mm; 1200-1450 mm) at h_a values of around 10 kPa (log(h) = 1). The equivalent pore radius is therefore approximately 15 µm. It seems significant that in one horizon (700-1200 mm), with a slightly higher clay content (Table 3.3), the minimum is around h = 30 kPa (log(h) = 1.5). No explanation is available for the shape of the derivative curve for the 1450-1850 mm horizon. These results for the Bainsvlei soil support in a general way the choice of h = 10 kPa as a fixed boundary between structural and textural pores in Figure 3.3. For the Tukulu soil the minima in the derivative curves occur at h values of 30 kPa (log(h) = 1.5) and 100 kPa (log(h) = 2) for the A and B horizons, respectively. It seems to be significant that both the field description of the structure (both massive), and textures (19-22% clay) of the 700-1200 mm Bainsvlei horizon and 0-270 mm Tukulu horizon are very similar and that their derivative curves are also very similar (Figure 3.4), both with troughs at h_a values around 30 kPa (log(h) = 1.5). In contrast in the Tukulu horizon with considerable more clay (29%) and structure described as becoming weak sub-angular blocky towards the transition (transition to the C-horizon), the trough in the derivative curve (Figure 3.4) occurs at higher suction around 100 kPa (log(h) = 2). The only visible significant feature of the derivative curve of the Tukulu C-horizon is a peak at h ≈ 100 kPa (log(h) = 2), and a decline thereafter. This expression of a completely atypical frequency of different pore sizes, compared to all the other horizons, is to be expected in view of the very different structure and texture of this horizon. As can been seen from Figure 3.4, the multimodality of pore size distribution is evident in most horizons excepting for the C-horizon of the Tukulu soil. This indicates the existence of a heterogeneous pore system in the horizons. This multimodality of the pore size distribution may be the result of specific particle size distributions or to the formation of a secondary pore system (structural domain) by various soil genetic processes such as soil aggregation or biological soil forming factors (Durner, 1994). This is consistent with the report by Kosugi *et al.* (2002) that states that undisturbed soils occasionally exhibit derivative curves with more than one inflection point. These findings also demonstrate that derivative curves provide a logical procedure for partitioning soil porosity, preferable to using a fixed boundary. Kutilek & Jendele (2008) consider that a fixed boundary between structural and matrix pore cannot be expected to occur in natural soils.

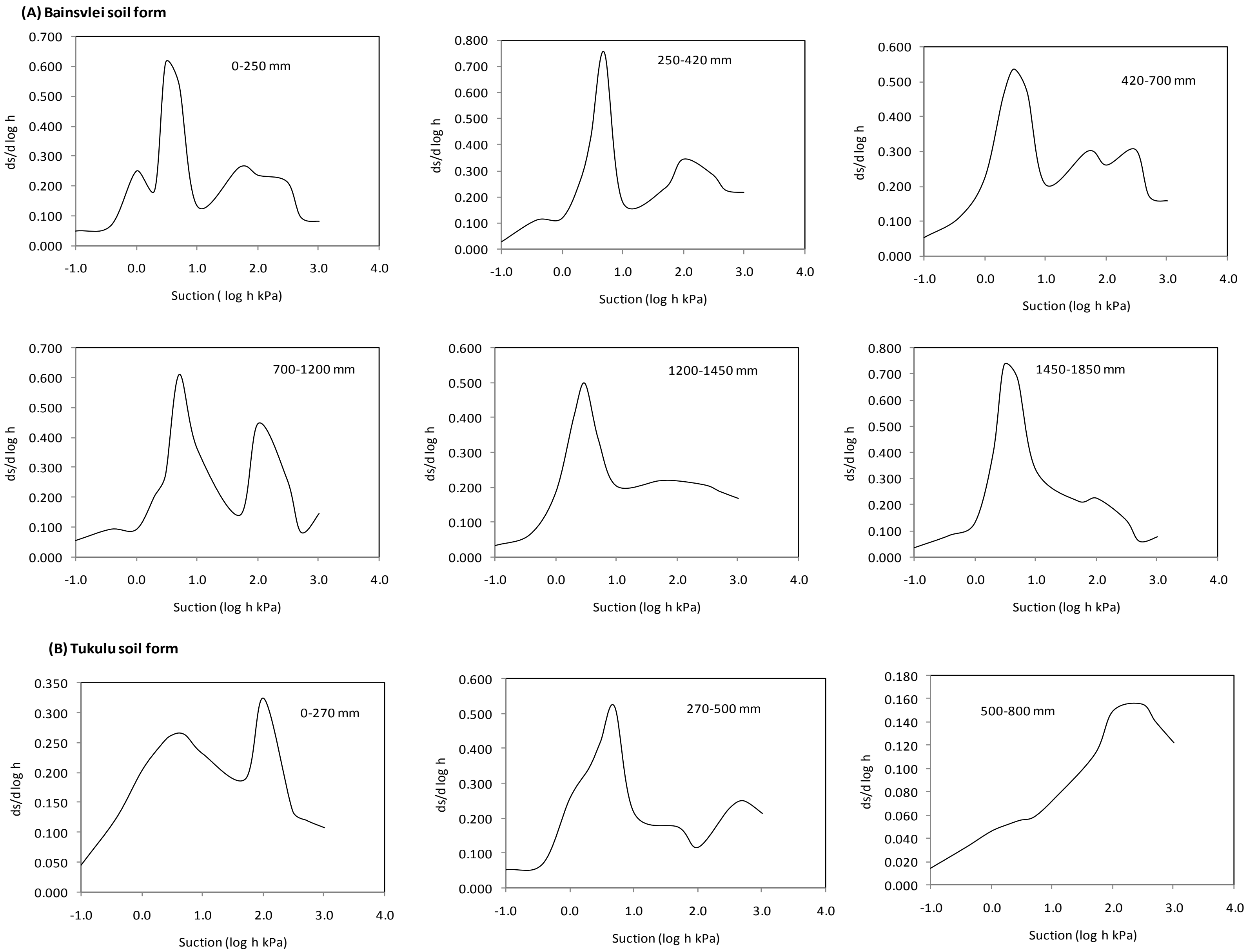


Figure 3.4 Separation of pore volume using the derivative technique (A) Bainsvlei and (B) Tukulu soil form.

3.4 Conclusion

The two soil profiles were successfully characterized and classified, thereby providing the basic information needed to fulfil requirements of a valuable pedotransfer function, the appropriate vehicle to carry the soil information needed to optimize the biological productivity of these ecotopes. The soil at Kenilworth was classified as Bainsvlei form *Amalia family* and at Paradys was classified as Tukulu form *Dikeni family*. The detailed profile descriptions proved useful in providing a picture of the various horizons of the two soils and differences in horizons, mainly regarding their colour and structure. The soil analysis was useful to distinguish soil layers in terms of texture within the profile.

The θ-h relationships, successfully determined using the standard procedures, demonstrate significant variation as shown in Figures 3.1 and 3.2 for all the horizons. The water retention characteristics for both soils were generally well defined with little variability between replicates. The main variations were due to particle size distribution differences between the horizons. The van Genuchten model (van Genuchten, 1980) applied via the RETC computer program (van Genuchten *et al.,* 1991), satisfactorily fitted the θ/h relationship measured in the laboratory. However, the model fails to accurately fit the curve at h = 10 kPa, which probably approximates in many soils the boundary between the 'structural porosity' and the 'textural porosity'. The RETC computer program (van Genuchten *et al.,* 1991), coupled with the van Genuchten model, created favourable retention curves in regard to the laboratory data. However, this method fails to save time and expense because of the raw data inputs. It is however, useful in interpolation of soil hydraulic property curves based on limited data.

The structural and textural domains were defined based on the principal of a fixed boundary (h = 10 kPa), from θ-h relationships. We also tested the theory using the derivative curve of θ-h relationships. The suction value (h_a) at the minimum between two peaks of the derivative curve represents the boundary between the structural and textural domains of the soil pore system. The value of h_a ranges from 10 kPa to about 30 kPa, corresponding to equivalent radius r = 15 µm and 5 µm. The boundary between soil pore categories cannot be taken as a fixed value for all soils. In this study, we have shown that the fixed boundary method although has been heavily criticised by several authors, offers promise for separating the structural and the textural domains of soil porosity. The structural domain varies with clay content. It has been shown that increase clay content in a given soil is associated with a decrease in the structural domain.

3.5 References

AHUJA, L., NANCY, J., GREEN, R. & NIESLEN, D., 1984. Macroporosity to characterize variability of hydraulic and effects of land management. *Soil Sci. Soc. Am. J.* 48, 670-699.

BENNIE, A., HOFFMAN, J. C. & VREY, H., 1994. Storage and use of rain water in soil for stabilising plant production in semi areas (Afr). *Water research commission (WRC), Report* No 227/1/94. Pretoria.

BROOKS, R. & COREY, A., 1964. Hydraulic properties of porous media. Hydrology Paper, Vol. 3, Colorado State Univ. 1-15.

COPPOLA, A., BASILE, A., COMEGNA, A. & LAMADDALENA, N., 2009. Monte carlo analysis of field water flow comparing uni- and bimodal effective hydraulic parameters. *J. Cont. Hydrol. 104,* 153-165.

DEXTER, A., CZYZ, E., RICHARD, G. & RESZKOWSKA, A., 2008. A user friendly water retention function that takes account of the textural and structural pore spaces in soil. *Geoderma* 143, 243-253.

DIRKSEN, C., 1999. Soil physics measurements. GeoEcology paperback, Catena Verlag GMBH, 35447 Reiskirchen, Germany.

DURNER, W., 1994. Hydraulic conductivity estimation for soils with heterogenous pore structure. *Water Resour. Res.* 30, 211-223.

ELHERS, W., WENDROTH, O. & DE MOL, F., 1995. Characterizing pore organisation by soil physical parameters. *Adv. Soil Sci.* 257-275.

FRAENKEL, C.H., 2009. Spatial variability of selected soil properties in and between map units. M.Sc. thesis, University of the Free State, Bloemfontein, South Africa.

GARDNER, E., 1988. Soil water. *In:* I. Fergus, (ed.). Understanding soils and soil data. Brisbane: Austrslian Society of Soil Science, 153-185.

GROSSMAN, R. & REINSCH, T. 2002. Core method. *In:* J. Dane, & G. Topp, (eds.). Methods of soil analysis, Part 4, Physical methods. SSSA., Book series No 5, Madison, Wisconsin, 207-209.

HENSLEY, M., BOTHA, J.J., ANDERSON, J.J., VAN STADEN, P.P & DU TOIT, A., 2000. Optimising rainfall use efficiency for developing farmers with limited access to irrigation water. *Water Research Commission (WRC) Report* No 878/1/00, Pretoria.

HILLEL, D., 1982. Fundamentals of soil physics. Academic Press Inc, New York.

HILLEL, D., 1998. Environmental of soil physics. Academic Press Inc, New York.

INTERNATIONAL ORGANIZATION FOR STANDARDIZATION (ISO), 2009. Determination of the soil water retention characteristics. Version 1.3, FUTMON-soil moisture workshop, International Organization for Standardization, Geneva, 1-12. http:// www.iso.ch (accessed 23/07/2009).

JURY, W., GARDNER, W. & GARDNER, W., 1991. Soil physics. John Wiley & Sons, New York.

KOSUGI, K., HOPMANS, J. & DANE, J., 2002. Parametric models. *In:* J. Dane, & G. Topp, (eds.). Methods of soil analysis, Part 4, Physical methods, SSSA., Book series No 5 Madison, Wisconsin, 739-757.

KUTILEK, M., & JENDELE, L., 2008. The Structural Porosity in Soil Hydraulic Functions. *Soil & Water Res.* 3, S7-S20.

KUTILEK, M., 2004. Soil hydraulic properties as related to soil structure. *Soil & Till. Res.*, 79, 175-184.

KUTILEK, M., JENDELE, L. & KYRIAKOS, P., 2006. The influence of uniaxial compression upon pore size distribution in bi-modal soils. *Soil & Till. Res.* 86, 27-37.

LUXMOORE, R., 1981. Micro-, Meso- and macroporosity of soil. *Soil Sci. Soc. Am. J.* 45, 241-285.

MARION, J., OR, D., ROLSTON, D., KAVVAS, M. & BIGGAR, J., 1994. Evaluation of methods for determining soil-water retentivity and unsaturated hydraulic conductivity. *Soil Sci.* 158, 1-13.

MARSHALL, T., 1959. Relations between water and soil. Farnham Royal: Commonwealth Bureau of Soils Harpeden. Technical Communication, 50.

MUALEM, Y., 1976. A new model for predicting the hydraulic conductivity of unsaturated porous media. *Water Resour. Res.* 12, 513-522.

NIMMO, J., 1997. Modelling structural influences on soil water retention. *Soil Sci. Soc. Am. J.* 61, 712-719.

PAGLIAI, M., VIGNIZZI, N. & PELLEGRINI, S., 2004. Soil structure and the effect of management practices. *Soil tillage Res*. 79, 131-134.

SEKI, K., 2007. SWRC fit- a nonlinear fitting program with a water retention curve for soils having unimodal and bimodal pore structure. *Hydrol Earth Syst. Sci*. 4, 407-437.

SKIDMORE, E., 1985. Soil porosity. Lecture notes for College of Soil Physics. ICTP, Trieste.

SOIL CLASSIFICATION WORKING GROUP, 1991. Soil classification a taxonomic system for South Africa. Department of Agricultural Development, Pretoria, South Africa.

THE NON-AFFILIATED SOIL ANALYSIS WORK COMMITTE, 1990. Handbook of standard soil testing methods for advisory purposes. SSSSA., Pretoria, South Africa.

VAN GENUCHTEN, M. & NIELSEN, D., 1985. On describing and predicting the hydraulic properties of unsaturated soils. *Ann. Geo-phys.* 3, 615-628.

VAN GENUCHTEN, M. T., 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Sci. Soc. Am. J*. 44, 892-898.

VAN GENUCHTEN, M., LEIJI, F. & YATES, S., 1991. The RETC code for quantifying functions of unsaturated soils. EPA/600/2-91/065.

VAN RENSBURG, L.D., 1988. The prediction of soil induced crop water stress for selected soil-plant atmospheric systems *(Afrikaans).* M.Sc. Thesis, University of the Free State, Bloemfontein, South Africa.

WILLMOTT, C.J., 1982. Some comments on the evaluation of model performance. *Am. Met. Soc.* 63, 1309-1313.

CHAPTER 4

Calibrating ECH_2O EC-20 sensors for measuring soil water content

4.1 Introduction

4.1.1 Motivation

Accurate soil water content measurements are required for measurements of crop water use, hydraulic characteristics of soils and in hydropedological studies. The gravimetrical method is generally used as standard for other indirect methods, but it is not attractive for routine measurements because it destructive and time consuming. Several non destructive methods have been devised to measure and monitor volumetric soil water content (θv). Widely accepted *in situ* methods include radioactive methods (IAEA, 2008), however these probes cannot be left unattended and as it is difficult to automate their measurements they are not suitable for making continuous measurements. Alternative techniques have been developed to estimate θv that take advantage of the relatively high permittivity of water, such as Time domain reflectometry (TDR) (Dirksen, 1999; Or & Wraith, 2002). However, the high cost of these sensors has lead to the development of alternative means of measuring θ.

Cheaper alternatives to TDR have recently widely become used, e.g. the ECH_2O EC-20 probes (Decagon Devices Inc., Pullman, Washington, USA). The ECH_2O EC-20 probe is a relatively low cost instrument compared to TDR systems, and it has proved to be useful in measuring soil water content *in situ* (Bandaranayake *et al.,* 2007). The ECH_2O EC-20 sensor uses capacitance to measure the dielectric permittivity of the surrounding medium (Kelleners *et al.,* 2005) and the final output from the sensor is a millivolt (mV) value that is converted to volumetric water content (θv) using a generic calibration equation supplied by the manufacturer. This calibration equation has since been shown not to be truly generic. Several studies have shown that the factory calibrations of these capacitance systems are not accurate for all soils (Baumhardt *et al.,* 2000; Evett & Steiner, 1995; Paltineanu & Starr, 1997; Foley & Harris, 2007). It is therefore important to calibrate each system for the specific soil in which the sensors will be used. The frequency of oscillation in these capacitance systems is affected not only by soil water content, but also by clay content and type, bulk electrical conductivity, salinity, and temperature (Baumhardt *et al.,* 2000).

A low measurement frequency is one of the primary factors that make these sensors liable to be influenced by the environmental factors mentioned above, and they often therefore require temperature correction and soil specific calibration (Seyfried & Murdock, 2001; Chandler *et al.*, 2004). According to the manufacturer, high electrical conductivity of the soil, high clay content, high quartz content, and high soil organic matter affect the response of ECH_2O EC-20 probes. Accordingly, the manufacturer recommends that probe users conduct a soil specific, and preferably horizon specific, calibration for best accuracy θv measurements. Adopting this procedure will not only produce more accurate, site specific calibrations, but will also help to identify problems with installation of the probes (IAEA, 2008). Recent studies by Czarnomski *et al.* (2005) indicate that soil specific calibration increases accuracy to ±1-2% for all soils. Foley & Harris (2007) found that the generic calibration supplied by the manufacturer underestimated water content by 10-20 % in clay soil. The generic calibration for the ECH_2O EC-20 probes is nominally valid at 20°C, but the manufacturers suggest that they have only a small and in most cases negligible temperature dependence. However, the temperature effect has been shown to depend on soil type by Seyfried & Murdock, (2001); and water content by Campbell *et al.* (2006) with fine textured soils and that high water contents generally resulted in the ECH_2O EC-20 sensors being more sensitive to temperature.

4.1.2 Objectives of the study

The objectives of this study were: (i) to evaluate the effects of soil temperature on ECH_2O EC-20 probes in different soils at variable water content; (ii) to derive specific calibration equations for eight horizons of two agricultural soils and compare these with generic calibrations. The data obtained in this study will be used for the determination of soil hydraulic characteristics of these two soils in subsequent studies.

4.2 Materials and method

4.2.1 Sensor description

The ECH_2O EC-20 probe is a ruler type capacitance soil water sensor 3.2 cm wide and 20 cm long. It requires an excitation current of 2 to 5 V at 3 mA for approximately 10 milliseconds. The output of the sensor is approximately proportional to the input. The sensor output ranges from 250 mV to 1000 mV with a 2.5 V excitation. ECH_2O EC-20 sensors change by approximately 10 mV per % change in θv. They use capacitance to measure the dielectric permittivity of the surrounding soil medium (Kallener *et al.,* 2005). The circuitry inside the sensor transforms the capacitance measurement into an output in millivolt (mV). The measurement principle of the ECH_2O EC-20 probes is reported in detail by Decagon Devices, Inc.

4.2.2 Site and soil description

The two soils used in this study were from two experimental field stations of the University of the Free State, at Kenilworth (29^{o}01' 00'' S, 26^{o}08' 00''E, altitude 1354 m) and at Paradys (29^{0}13' 25'' S, 26^{o}12' 08'' E, altitude 1417 m). The soil at Kenilworth is classified as a Bainsvlei form *Amalia family* (Soil Classification Working Group, 1991). This soil consists of an orthic A horizon which grades into a red apedal B with underlying soft plinthic material. At Paradys, the soil is Tukulu form *Dikeni family* (Soil Classification Working Group, 1991). These two soils represent two extremes of South African agricultural soils. The Bainsvlei soil represents a large area of land being utilized successfully for rain fed cropping while the Tukulu soil represents soils that have a marginal potential for rain fed cropping with conventional production techniques, but that can be utilized with the in-field rainwater harvesting production technique. Generalized soil profile descriptions of the two soils and depths at which ECH_2O EC-20 sensors were installed are presented in Table 4.1.

ECH2O EC-20 probes were installed by making a pilot hole using a metal blade of same dimensions as the probe. Later on the probe was inserted into the pilot hole just made with the long axis of the probe horizontal and flat plane horizontal. Using two fingers the soil was compacted around the edges of the probe to ensure that there is perfect contact between the soil and probe (no gaps around the probe).

Table 4.1 Soil horizons where ECH_2O EC-20 sensors were installed in the field.

Soil form	Probe No	Horizon	Depth(mm)	Bulk density (Mgm^{-3})	Soil description
Tukulu (*Dikeni*)	Probe 1	A	100	1.67	Fine sand loamy, massive structure
	Probe 2		265		
	Probe 3	B	275	1.68	Fine sandy clay loam, massive structure becoming blocky towards transition
	Probe 4		495		
	Probe 5	C	505	1.71	Clay, prismatic structure with few slickensides
	Probe 6		795		
Bainsvlei (*Amalia*)	Probe 7	A	85	1.66	Fine loamy sand
	Probe 8		245		
	Probe 9	B1	263	1.68	Fine sandy loam, massive structure
	Probe 10		415		
	Probe 11	B2	695	1.66	Fine sandy loam, massive structure
	Probe 12	B3	713	1.67	Fine sandy clay loam, massive structure
	Probe 13		1195		
	Probe 14	B4	1463	1.67	Fine sandy loam, massive structure.
	Probe 15		1845		

4.2.3 Calibration experiment

In most previous laboratory calibration studies repacked soil samples were used. In contrast we used undisturbed soil samples in order to maintain *in situ* physical conditions, as recommended by Weitz *et al.* (1997) and IAEA (2008). The undisturbed soil samples were taken at the field sites of the internal drainage experiments and at the depths as shown in Table 4.1.

The undisturbed soil samples were collected in polyvinyl chloride (PVC) cylinders pressed horizontally from the sidewall of an exposed pit, using a hydraulic device. The columns were 103 mm in diameter and 200 mm long. The PVC pipes were perforated manually beforehand with random holes (2 mm in diameter) at a high density (two holes per cm^2) to ensure maximum evaporation potential from the core during calibration. In the laboratory the columns were vacuum saturated, the saturation procedure took about three days. ECH_2O EC-20 probes were installed at each end of two 200 mm samples joined together in order to make allowance for the critical sphere of influence of the probes. The core assembly was then taken to a climate controlled room and suspended on load cells, and allowed to dry gradually by evaporation until no detectable changes in mass were observed; the drying cycle took about 27 days. The mass of the core assembly was recorded automatically using a data logger (CR10x, Campbell Sci.) connected to the load cells. The water content within the core assembly was monitored by the ECH_2O EC-20 probe readings in millivolts using Decagon's Em5 data logger with a 2.5-V excitation. Four temperature sensors were installed randomly into the cores. These sensors were connected to a HOBO8 datalogger. All the dataloggers were programmed to measure the output of the load cells, the EC-20 probes and the temperature sensors at 30 minute intervals. The air temperature in the chamber was oscillated between 5^{o}C and 30^{o}C. The actual soil temperature ranged from 7.8^{o}C to 22.9^{o}C. This temperature range approximates the daily or seasonal soil water temperature fluctuations close to the soil surface. The soil water content was determined gravimetrically via the load cell readings, and converted to volumetric soil water content using an independently determine bulk density value for the respective soil horizons as indicated in Table 4.1. For each selected temperature level, the soil was given ample time (approximately 48-50 hours) to equilibrate, before changing to another level.

4.2.4 Selection of sub-data sets

The data set was grouped into three data sub-sets, viz: (i) lower temperature (7.8^{o}C) marked by rectangles in Figure 4.1; (ii) higher temperature (22.86^{o}C) marked by ovals; (iii) the transition period in-between, either increasing or decreasing. An example showing how the data was utilized to obtain three independent data sub-sets, the mV output of probe No 1, and temperature over the entire drying cycle are depicted in Figure 4.1. The temperature data in Figure 4.1 is presented as follows: lower values are marked with rectangles; higher values as ovals; transitional values as the lines joining the higher and lower values.

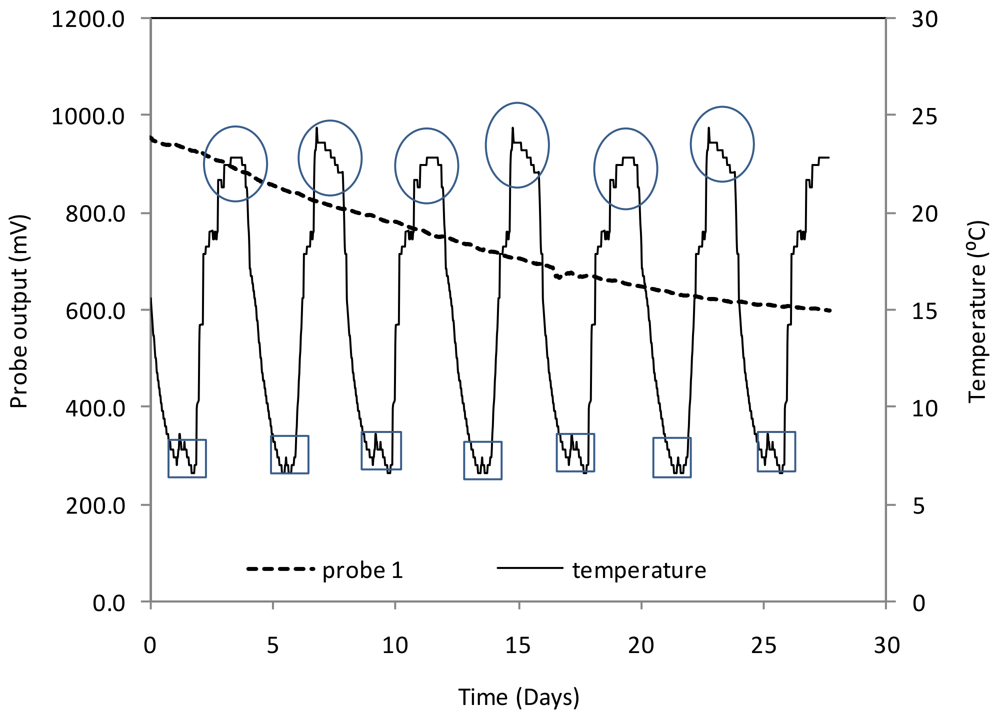


Figure 4.1 An illustration of how the three independent sub data sets were obtained using probe 1 (mV) and the corresponding soil temperature.

4.2.5 Development of the calibration equations

The θv for each soil column was linearly regressed against probe output (mV) at each temperature level. Comparisons of the regression parameters (slopes and intercepts) (low vs. high temperature) were undertaken with GenStat (2008). In order to demonstrate their prediction patterns they were plotted on the same graph (Figure 4.2). Finally a horizon specific calibration equation was developed for each probe based on the results from the entire calibration experiment. Regression equations were sought that fit the data points smoothly and accurately. All curve fittings were accomplished using CurveExpert 1.4. The precision of the sensors was evaluated by the coefficient of determination (R^2) that describes the proportion of the variance in the observed data that can be explained by the fitted equation.

4.2.6 Comparison and validation of water content predictions

The data sets provide the opportunity to test the validity of the generic equation supplied by manufacturers for some of South African soils. Willmott (1982) statistics were used to compare the θv values obtained using the horizon specific calibrations determined for each probe, against the results with the generic calibration equation supplied by the manufacturer.

4.3 Results and discussion

4.3.1 Effect of soil temperature

Calibration curves of fifteen ECH_2O EC-20 probes at two levels of temperature (22.9^oC and 7.8^oC) are shown in Figure 4.2. No significant variation was determined for all sensors at the temperature levels tested. Statistical comparisons between the slopes of the calibration curves for individual ECH_2O EC-20 probe/temperature combinations showed no significant differences between 22.9^oC and 7.8^oC temperature calibration curves. The lack of significant differences between calibration curves for different temperature levels was not surprising, considering the results in Figure 4.2. However, we were generally surprised by the similarity of calibrations between the two different soils. It is important to note that the temperature range was limited to values between 7.8^oC and 22.9 ^oC. It is therefore not clear that these findings will hold true for extreme temperatures. These findings are in agreement with those of Paltineanu and Starr (1997) who reported that temperature effects on capacitance sensors might be negligible within the temperature range of 10 to 30°C. Seyfried and Murdock (2001) also reported that there were no identified temperature effects on the measured values of θv for coarse textured soils. The practical implications of these results are twofold. First, the effect of temperature on ECH_2O EC-20 probes is negligible for many applications in which temperature changes are small. Therefore, we can express the volumetric water content as a function of probe output (mV). Second, although individual probes may vary in terms of their response to temperature, influence on θv determinations for individual probes is expected to be negligible. Differences in the soils studied did not influence the relationship between θv and soil temperature. It has been reported for some time that different soil types can affect the dielectric constant of water and therefore estimates of θv based of dielectric constant (Dobson *et al.*, 1985). The primary cause of the differences between soils is usually attributed to the effects of solid-liquid interactions at the soil particle surfaces that restrict the rotational freedom of adsorbed water molecules (Seyfried & Murdock, 2001). This water is considered to have a dielectric constant much lower than that of free water (Dobson *et al.*, 1985). Hence, a temperature effect is expected to be more pronounced in soils with more bound water (e.g. > clay) due to the liberation of bound water as the temperature increases. The results for this study show no evidence for this.

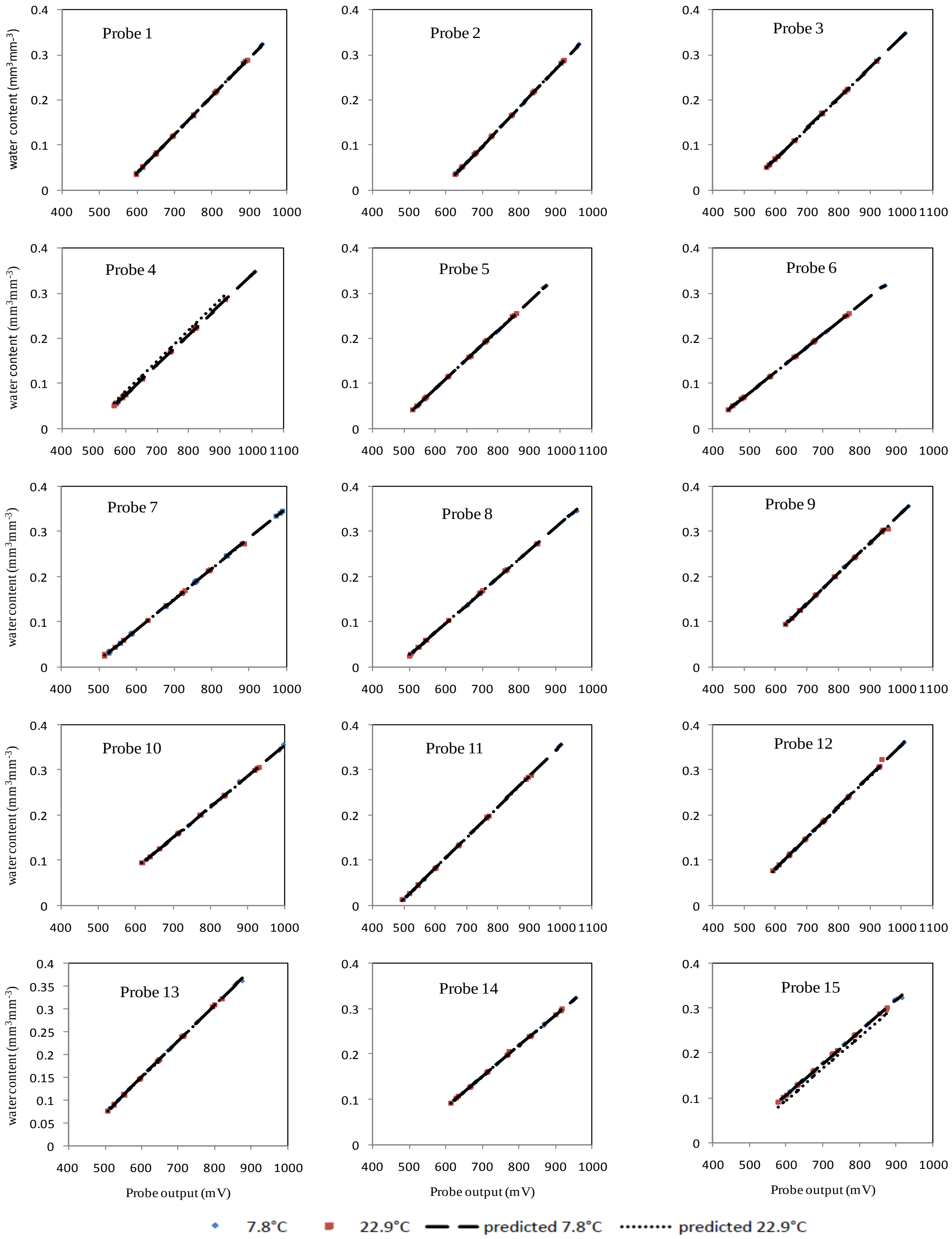


Figure 4.2 The response of fifteen ECH$_2$O EC-20 probes to water content and temperature variations in the different horizons of two soil profiles.

4.3.2 ECH_2O EC 20 probe calibrations

In this study the calibration of probes produced a linear relationship between probe output and θv (Figure 4.2). The coefficients of determination (R^2) for all the probes were close to unity (Table 4.2). All the equations were free of any bias, as reflected by high $RMSE_u$/RMSE.

Table 4.2 Calibration equations for the fifteen ECH_2O EC-20 sensors in different soil layers. All coefficients are significant at the P = 0.01 level (n = 1329).

Probe no	Soil texture	Intercept	Slope	R^2	RMSE	$\frac{RMSE_u}{RMSE}$
Probe 1	Sandy loam	-0.473	0.00085	0.999	0.000523	0.986
Probe 2	Sandy loam	-0.496	0.00085	0.999	0.000106	0.987
Probe 9	Sandy loam	-0.326	0.00067	0.999	0.000645	0.996
Probe 10	Sandy loam	-0.327	0.00068	0.999	0.000257	0.987
Probe 11	Sandy loam	-0.327	0.00068	0.999	0.000295	0.998
Probe 14	Sandy loam	-0.327	0.00068	0.999	0.000568	0.987
Probe 15	Sandy loam	-0.318	0.00071	0.999	0.000805	0.985
Probe 3	Sandy clay loam	-0.332	0.00067	0.999	0.000593	0.995
Probe 4	Sandy clay loam	-0.327	0.00067	0.999	0.000350	0.992
Probe 12	Sandy clay loam	-0.326	0.00068	0.999	0.000304	0.987
Probe 13	Sandy clay loam	-0.328	0.00080	0.999	0.000601	0.987
Probe 7	Loamy sand	-0.327	0.00068	0.999	0.000251	0.986
Probe 8	Loamy sand	-0.327	0.00071	0.999	0.000952	0.989
Probe 5	Clay	-0.300	0.00065	0.999	0.000246	0.987
Probe 6	Clay	-0.241	0.00064	0.999	0.000119	0.989

The data for each texture class was pooled and fitted to determine one calibration for each textural class (Table 4.3). Textural specific calibration equations had R^2 values lower than those of the horizon specific equations (Table 4.3). On average, the R^2 value for horizon specific equations was 0.999, while the R^2 value for the textural specific equations was 0.924. Correspondingly the horizon specific calibration resulted in lower average RMSE values than the textural specific calibration (0.0004 vs. 0.021). These findings suggest that the textural specific calibration is less suited to describe θ_v in the present soils. Using it would result in the underestimation or overestimation of θv. It seems that is preferable to calibrate these sensors for a particular soil horizon. But it is interesting to note that the clay soil has the largest and sandy soil has smaller intercept. Similar trends were also noticed for slope, R^2 and RMSE.

Table 4.3 Calibration equations of the ECH_2O EC-20 sensors for different soil texture classes. All coefficients were significant at P = 0.05 level.

Soil texture	Intercept	Slope	R^2	RMSE
Clay	-0.231	0.00058	0.905	0.027
Sandy clay loam	-0.273	0.00062	0.859	0.026
Sandy loam	-0.347	0.00070	0.941	0.022
Loamy sand	-0.324	0.00068	0.993	0.009

4.3.3 Comparing generic and horizon specific calibration

The RMSE, D-index, and $RMSE_u$/RMSE for the comparison of θv measured gravimetrically with those estimated using either the calibration equations supplied by the manufacturer or horizon specific calibrations are summarized in Table 4.4. Manufacturers' calibration equations had a much lower and therefore unfavourable RMSEu/RMSE values and a larger average RMSE values as compared to the horizon specific calibration equations, i.e. 0.724 vs. 0.989, 0.0468 vs. 0.0005 respectively. Low RMSEu/RMSE values (< about 0.80) indicate the presence of considerable bias in the prediction procedure. This is clearly visible in Figure 4.3. Foley & Harris (2007) also reported that manufacturers calibration equations rarely produce RMSE values that will be comparable to those obtained with horizon specific calibration equations.

Table 4.4 Comparison between generic and horizon-specific calibrations.

Probe No	Horizon-specific calibration			Generic calibration		
	RMSE	$RMSE_u$/RMSE	D-index	RMSE	$RMSE_u$/RMSE	D-index
Probe 1	0.000523	0.986	0.999	0.0686	0.689	0.858
Probe 2	0.000106	0.987	0.999	0.0894	0.675	0.790
Probe 3	0.000593	0.995	0.999	0.0598	0.665	0.912
Probe 4	0.000350	0.992	0.999	0.0551	0.654	0.924
Probe 5	0.000246	0.987	0.999	0.0434	0.689	0.944
Probe 6	0.000119	0.989	0.999	0.0185	0.978	0.989
Probe 7	0.001155	0.986	0.999	0.0462	0.678	0.950
Probe 8	0.000952	0.989	0.999	0.0293	0.876	0.978
Probe 9	0.000645	0.996	0.999	0.0573	0.588	0.892
Probe 10	0.000257	0.987	0.999	0.0469	0.677	0.923
Probe 11	0.000295	0.998	0.999	0.0458	0.678	0.957
Probe 12	0.000304	0.987	0.999	0.0467	0.675	0.936
Probe 13	0.000601	0.987	0.999	0.0301	0.765	0.968
Probe 14	0.000568	0.987	0.999	0.0469	0.709	0.912
Probe 15	0.000805	0.985	0.999	0.0190	0.877	0.983

To further compare the results the θ_v values predicted by the two procedures are plotted in Figure 4.3. The manufacturers' calibration equations consistently overestimated θv, except for probes 6 and 13 which underestimated θv. This result confirms and further clarifies the statistics in Table 4.4 providing further justification for the use of horizon specific calibration equations. It is interesting to note that these findings are consistent with previous studies by Foley & Harris, (2007) and Czarnomski *et al.* (2005) who also found that the manufactures' calibrations consistently overpredict θ_v. The lower level of accuracy of manufacturers' generic calibration equations over the entire range of the drying cycle makes them unsuitable for use with irrigation scheduling and in drainage studies, since these require a high level of accuracy. The manufacturers' generic calibration equation did not achieve their advertised accuracy of ±4%. However, horizon specific calibrations produced an excellent level of accuracy across the entire drying cycle, to within ± 2% of the measured θv values (Figure 4.3).

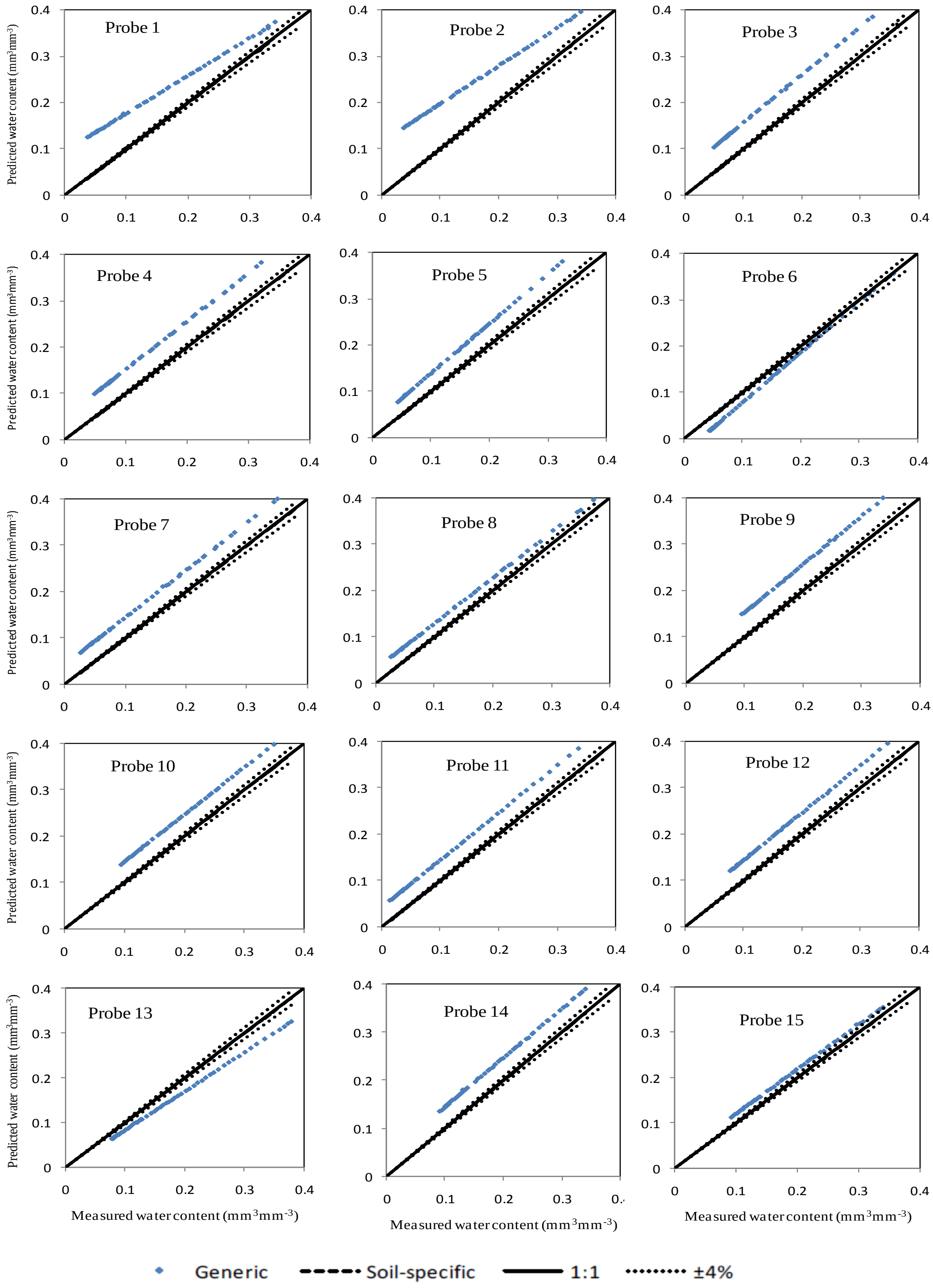


Figure 4.3 Comparison of results for fifteen ECH$_2$O EC-20 sensor responses using horizon specific calibration with those obtained using generic calibration.

4.4 Conclusion

Calibration of ECH_2O EC-20 probes was conducted for five horizons of a Bainsvlei form soil and three horizons of a Tukulu form soil in the laboratory. ECH_2O EC-20 probe sensors and circuitry show extremely low sensitivity to temperature fluctuations. Based on finding of this study temperature influence can be ignored in most field applications, because temperature dependence plays a minor role in probe output. Horizon specific calibration is essential to get accurate water content estimates from ECH_2O EC-20 probes if used in different soil horizons. Our study demonstrates that horizon specific calibration of ECH_2O EC-20 sensors improves the accuracy of soil water content monitoring of the sensors as compared to the use of manufacturer generic calibration equation for the soils tested in this study. This is especially important to monitor θv for irrigation scheduling, drainage studies and for the determination of the hydraulic characteristics of soils.

4.5 Reference

BANDARANAYAKE, W., PARSONS, L., BORHAN, M. & HOLETON, J., 2007. Performance of a capacitance-type soil water probe in a well drained sandy soil. *Soil Sci. Soc. Am. J.* 71(3), 993-1002.

BAUMHARDT, R., LASCANO, R. & EVETT, S., 2000. Soil material, temperature, and salinity effects on calibration of multisensor capacitance probes. *Soil Sci. Soc. Am. J.* 64(6), 1940-1946.

CAMPBELL, C., CAMPBELL, G., COBOS, D. & BISSEY, L., 2006. Calibration and evaluation of an improved low cost soil moisture sensor. Htpp://*www.decagon.com*.pdf (accessed 24/08/09)

CHANDLER, D., SEYFRIED, M., MURDOCK, M. & MCNAMARA, J., 2004. Field calibration of water content reflectometers. *Soil Sci. Soc. Am. J.* 68, 1501-1507.

CZARNOMSKI, N., MOORE, N., PYPKER, T., LICATA, J. & BOND, B., 2005. Precision and accuracy of three alternative instruments for measuring soil water content in two forest soils of the pacific northwest. *Can. J. For. Res.* 35(8), 1867-1876.

DIRKSEN, C., 1999. Soil physics measurements. GeoEcology paperback, Catena Verlag GMBH, 35447 Reiskirchen, Germany.

DOBSON, M., ULABY, F., HALLIKAINEN, M. & ELRAYERS, M., 1985. Microwave dielectric behaviour of wet soil, II. Dielectric mixing models. *IEEE Trans. Geosci. Remote Sens.* 23, 35-46.

EVETT, S.R. & STEINER, J.L., 1995. Precision of neutron scattering and capacitance type soil water content gauges from field calibration. *Soil Sci. Soc. Am. J.* 59, 961–968.

FOLEY, J. & HARRIS, E., 2007. Field calibration of thetaprobe (ML2X) and ECH$_2$O probe (EC-20) soil water content sensors in black vertosol. *Aust. J. Soil Res.* 45, 233-236.

GENESTAT for teaching edition, 2008. Release 7.22 TE. VSN International. Oxford, UK.

INTERNATIONAL ATOMIC ENERGY AGENCY (IAEA)., 2008. Field estimation of soil water content-A practical guide to methods, instrumentation and sensor technology. Training course series No. 30. International Atomic Energy Agency, Vienna, Austria.

KELLENERS, T., ROBINS, D., SHOUSE, P., AYARS, J. & SKAGGS, T., 2005. Frequency dependence of the complex permittivity and its impact on dielectric constant sensor calibration in soil. *Soil Sci. Soc. Am. J.* 69, 63-66.

OR, D. & WRAITH, J., 2002. Soil water content and water potential relationships. *In:* A. Warrick, (ed.). Soil physics companion. CRC press, Florida, 49-84

PALTINEANU, I. & STARR, J., 1997. Real-time soil water dynamics using multisensory capacitance probes: Laboratory calibration. *Soil Sci. Soc. Am. J.* 65, 1576-1585.

SEYFRIED, M. & MURDOCK, M., 2001. Response of a new soil water content sensor to variable soil, water content, and temperature. *Soil Sci. Soc. Am. J.* 65, 28-34.

SOIL CLASSIFICATION WORKING GROUP, 1991. Soil classification a taxonomic system for South Africa. Department of Agricultural Development, Pretoria, South Africa.

WEITZ, A., GRAUUEL, W., KELLER, M. & VELDKAMP, E., 1997. Calibration of time domain reflectomeetry technique using undisturbed soil samples from humid tropical soils of volcanic origin. *Water Resour. Res.* 33, 1241-1250.

WILLMOTT, C.J., 1982. Some comments on the evaluation of model performance. *America meteorological society*, 63, 1309-1313.

CHAPTER 5

Comparing laboratory and field determined hydraulic conductivity for a Bainsvlei form soil

5.1 Introduction

5.1.1 Motivation

Knowledge of the hydraulic conductivity characteristics of unsaturated soil (the K-θ relationship) is fundamental to studies involving water balances, irrigation and rain fed agriculture, movement of pollutants, and all water transport processes occurring in the vadose zone (Ross & Parlange, 1994). Historically this has been an important factor for assessing the suitability of land for irrigation agriculture and for trafficability. Owing to its importance, a sustained research effort towards the estimation of hydraulic conductivity has resulted in the development of several laboratory, field and theoretical methods (van Genuchten *et al.,* 1992).

Laboratory techniques for determining the K-θ relationship involve setting up either steady or transient state flow systems in which field extracted undisturbed samples are tested. However, it is now known that laboratory tests cannot fully duplicate field conditions. Among field methods, the instantaneous profile method (Hillel *et al.,* 1972) has come to be accepted as a standard *in situ* procedure (Marion *et al.,* 1994) in spite of certain inherent limitations (Paige & Hillel, 1993). Advantages of this method over the other methods include the following; (i) it provides *in situ* measurements of the K-θ relationship, and because it is non-destructive (Vachaud & Dane, 2002) it eliminates the disturbance of soil structure, and the resultant changes in the relationship due to sampling; (ii) a relatively large volume of soil is sampled when measuring the K-θ relationship (Green *et al.,* 1986), thereby reducing the variability of the measurements, and providing a better estimate of the larger scale hydraulic properties (e.g. in this study it was conducted on a monolith of 32 m^3); (iii) hydraulic properties are measured simultaneously at multiple depths within the soil profile (Hillel *et al.,* 1972).

The instantaneous profile method has been successfully used by several researchers on different soil types Davidson *et al.* (1969), Paige & Hillel (1993) and Zhang *et al.* (2007). It requires frequent and simultaneous measurements of soil water suction and water content at several depths within the soil profile under natural internal drainage conditions. Tensiometers equipped with electric pressure transducers facilitate the automatic acquisition of soil matric potential, while water content is commonly monitored by frequency domain probes, time domain reflectometry (TDR) probes or neutron water meters. However, TDR technology is relatively expensive and neutron water meters have a potential danger of radiation (Or & Wraith, 2002). The latter procedure also cannot log water content measurements continuously (Tyndale-Biscoe & Malano, 1995). In this respect, the use of capacitance-based sensors such as the ECH_2O EC-20 sensors (Decagon Devices, Inc., Pullman, WA) has the considerable advantage of being cheaper compared to TDR (Bandaranayake *et al.*, 2007). Their use is becoming popular overseas. However, after an extensive literature search, I did not find any work on the use of ECH_2O EC-20 sensors in an instantaneous profile method experiment on South African soils. There is therefore a knowledge gap in this respect with regard to soil hydraulic properties of local soils.

Hydraulic properties need to be well understood so that the data obtained provide reliable inputs to numerical simulation models for evaluating alternative soil and water management practices (Mallants *et al.*, 1997). However, the routine use of the instantaneous profile method is often inhibited by cost considerations, especially when large areas have to be characterised for their hydraulic properties. Often K-θ relationship is estimated using predictive models based on the soil water retention characteristics. The accuracy of such models in predicting the actual soil K-θ relationship is seldom verified by direct measurements of hydraulic conductivities, mainly due to the considerable large timeframe necessary to evaluate precisely several magnitudes of hydraulic conductivities at different water contents. In order to contribute to the understanding of the accuracy of K-θ estimates using predictive models based on soil water retention data, this study presents a comparison between the predicted K-θ using a water retention based model and direct measurements of unsaturated hydraulic conductivity of a Bainsvlei soil.

Considerable research has been addressed to the developing and testing of economical methods for determining the K-θ relationship (Ahuja *et al.*, 1984). Among the most attractive methods for the estimation of the K-θ relationship has been to seek statistical relationships with soil texture, bulk density, organic matter content, clay mineralogy, and soil structure (van Genuchten *et al.*, 1992). However, the use of pedotransfer functions (PTF's) of this sort is only valid for interpolation, and not for extrapolation (Wosten *et al.*, 2001). Other investigators have sought to derive soil hydraulic properties from closed form equations based on laboratory measured water retention data of undisturbed soil cores (Brooks & Corey, 1964; Campbell, 1974; van Genuchten, 1980; Kosugi, 1996). The equation of van Genuchten (1980) is the most widely used, because it describes the whole retention range with a continuous curve. A few studies have compared laboratory predicted K-θ relationships using the van Genuchten-Mualem model with *in situ* results (e.g. Field *et al.,* 1983; Paige & Hillel, 1993; Zhang *et al.*, 2007). A finding common to all these studies is that there is reasonable correspondance between instantaneous profile measured and the laboratory-predicted K-θ relationship for freely drained sandy to loam soils. With well-structured, finer textured soils correspondance has generally been poor. However, the prediction peformance of the van Genuchten- Mualem model compared to field measured K-θ has not been widely tested on South African soils. Furthermore, detailed comparison between the laboratory-prediction and field measured K-θ relationship will help to identify strengths and weakness of each method, and rationalize their applicability.

In this study we will also consider how the drained upper limit (DUL) can be predicted from laboratory determined θ-h relationships. *In situ* measurements of DUL are expensive, time consuming and technically difficult (Saxton *et al.,* 1986). This information is required to estimate the available water reserve of a soil profile, a critical input required by soil-water balance models and agronomic studies (Hensley *et al.*, 2000; Gebregiorgis & Savage, 2006). In attempt to improve DUL predictions it is necessary to understand that the DUL of a soil profile depends on the hydraulic conductivity of the most limiting layer in the profile. For this reason in this study DUL will be predicted separately for each horizon. For convenience this method will be termed the laboratory method. Results obtained using the laboratory method are compared in this study with those obtained using the *in situ* method proposed by Ritchie, (1981).

5.1.2 Theory

Water movement in the unsaturated zone of the soil occurs due to gradients in the hydraulic potential H, which is composed of the sum matric and gravitation potentials.

$$H = h + z \tag{5.1}$$

Where h is the soil water suction (mm), representing the matric potential; and z the depth, positive in downward direction (representing the gravitational potential).

According to Darcy's Law, water movement through a one dimensional unsaturated vertical column is mathematically expressed as (Jury & Horton, 2004):

$$q = -K(\theta)\frac{\partial H}{\partial z} \qquad 5.2$$

where q is the soil water flux (mm $hour^{-1}$), K(θ) is the unsaturated hydraulic conductivity (mm $hour^{-1}$) at a specified value of θ, and $\frac{\partial H}{\partial z}$ is the hydraulic head gradient. Accordingly, if measurements such as water flux q; gradient of hydraulic head at each depth $\frac{\partial H}{\partial z}$ and the water content (θ) at each depth at a particular time can be made in a soil profile draining from saturation then the hydraulic conductivity can be calculated as the unknown in Equation 5.2. The partial derivative in equation 5.2 is required for the mathematical description of transient (time-dependent) flow. However, because the measurement of h and θ are made simultaneously it is acceptable to consider the system at steady state at that time, hence, presumably the name 'instantaneous profile method' used by Hillel *et al.* (1972). For a system at steady state a normal derivative is acceptable. The instantaneous profile method works well for soils where the water table is absent or too deep to affect drainage. Although the method is fairly time consuming and expensive, it is often used as a standard for comparisons of other field and laboratory methods (Marion *et al.*, 1994).

5.1.3 Objectives of the study

The general objective of the study was to investigate and compare the hydraulic properties of the Bainsvlei form soil using field and laboratory methods. Specific objectives were: (i) to characterize the *in situ* K-θ relationships for the profile; (ii) to predict the K-θ relationships with the van Genuchten (1980) model; (iii) to estimate the drained upper limit for the profile.

5.2 Materials and methods

5.2.1 Site location and pedological characteristics

The study was conducted at Kenilworth (29^{o}01’ 00’’S, 26^{o}08’00’’E, altitude 1354 m) the experimental field station of the University of the Free State. The soil of the profile studied is classified as Bainsvlei form, *Amalia family* (Soil Classification Working Group, 1991). A detailed profile description is given in Table 5.1. Each of the selected diagnostic horizons were chemically and physically analyzed for exchangeable cations (Ca, Mg, Na, K), pH$_{(water)}$, and particle size distribution, using procedures described by The Non-Affiliated Soil Analysis Work Committee (1990). Field saturated hydraulic conductivity (K_{fs}) was determined using double ring procedure on each horizon. Particle size distribution, bulk densities and field saturated hydraulic conductivity (K_{fs}) values for the horizons and some relevant chemical properties are given in Table 5.2.

Table 5.1 Profile description of the Bainsvlei form soil.

Map/photo :	2926 Bloemfontein	Soil form and family:	Bainsvlei *Amalia*
Latitude + Longitude:	29º 1' 00''/26º 08' 00''	Surface rockiness:	None
Altitude:	1354 m	Occurrence of flooding:	None
Terrain unit:	Lower foot slope	Wind erosion:	Slight wind
Slope:	1%	Water erosion:	None
Slope shape:	Straight	Vegetation/Land use:	Agronomic field crops
Aspect:	North-west	Water table:	None
Microrelief:	None	Described by:	M. Hensley & J. Chimungu
Parent Material Solum:	Origin single, aeolian	Date described:	14/06/09
Underlying Material:	Sandstone (Feldspathic)	Weathering of underlying material:	Weak physical to moderate chemical
		Alteration of underlying material:	Ferruginised

Horizon	Depth (mm)	Description	**Diagnostic horizon**
A	0-250	Moist state; dry colour: yellowish red (5YR5/6); moist colour: reddish brown (5YR4/4); texture: fine loamy sand; structure: apedal massive; consistence: friable; few fine normal pores; water absorption: 1 second; few roots; gradual smooth transition.	Orthic
B1	250-420	Moist state; dry colour: red (2.5YR4/8); moist colour: red (2.5YR4/6); texture: fine sandy loam; structure: apedal massive; Consistence: friable; few fine normal pores; water absorption: 1 second; few roots; gradual smooth transition.	Red apedal
B2	420-700	Moist state; dry colour: yellowish red (5YR5/8); moist colour: red (2.5YR4/6); texture: fine sandy loam; few fine faint black illuvial humus mottles; structure: apedal massive; consistence: friable; common fine normal pores; water absorption: 1 second; few roots; gradual wavy transition.	Red apedal
B3	700-1200	Moist state; dry colour: yellowish red (5YR4/6); moist colour: reddish brown (5YR4/4); texture: fine sandy clay loam; common fine faint black illuvial humus mottles; structure: apedal massive; consistence: slightly firm; common fine normal pores; water absorption: 1 second; clear wavy transition.	Red apedal
B4	1200-1450	Moist state; dry colour: strong brown (7.5YR5/8); moist colour: strong brown (7.5YR5/6); texture; fine sand; few fine faint black illuvial humus mottles; structure: apedal massive; consistence: friable; water absorption: 1 second; few roots; clear wavy transition.	Non diagnostic; yellow brown Aeolian sand
B5	1450-1850	Moist state; moist colour: strong brown (7.5YR4/6); texture: fine sandy loam; many medium distinct grey and yellow reduced iron oxide mottles; many medium distinct red and black oxidised iron oxide mottles; structure: apedal massive; consistence: friable; water absorption: 1 second(s); few roots; gradual smooth transition.	Soft plinthic
C	1850-2220	Similar to B5 with patches of weathered feldspathic sandstone; colour of mottles similar to B5 but more prominent.	

Table 5.2 Summary of chemical and physical characteristics of the Bainsvlei form soil.

Physical properties						
	A	B1	B2	B3	B4	B5
Coarse sand (2 - 0.5 mm) (%)	0.4	0.3	0.3	0.3	0.3	0.6
Medium sand (0.5 - 0.25) (%)	7.1	5.2	5.4	4.1	3.3	6.0
Fine sand (0.25 - 0.106 mm) (%)	61.4	55.1	53.8	44.9	64.3	48.3
Very fine sand (0.106 - 0.53mm) (%)	16.8	15.1	15.5	18.0	17.3	17.0
Silt (%)	4.0	4.0	6.0	8.0	4.0	6.0
Clay (%)	8.0	18.1	18.0	22.1	8.1	20.1
Bulk density (Mg m^{-3})	1.66	1.68	1.66	1.67	1.68	1.67
*K_{fs} (mm $hour^{-1}$)	60.0	26.0	43.0	40.0	-	39.0
Chemical properties						
pH(water)	5.2	5.1	6.3	6.1	6.7	6.5
Ca ($cmol_c.kg^{-1}$)	7.2	15.3	11.0	16.2	11.3	11.1
K ($cmol_c.kg^{-1}$)	2.5	3.6	2.2	2.6	2.1	2.0
Mg ($cmol_c.kg^{-1}$)	4.3	10.2	10.4	9.4	11.7	10.3
Na ($cmol_c.kg^{-1}$)	4.1	3.8	2.7	2.5	2.4	2.7
Clay –S value	-	Eutrophic*	-	-	-	-

* *Field saturated hydraulic conductivity-determined by the double ring procedure.*

**clay-S value is calculated as follows:* $\frac{\sum exchangeable\ cations\ x\ 100}{\%\ clay}$

5.2.2 Field measurements

The internal drainage experiment was carried out on a (4 m x 4 m) x 2 m depth monolith following the procedure described by Hillel *et al.* (1972) and Hillel (1998). The monolith was vertically isolated with a plastic sheet to inhibit lateral water movement. Earth bunds were constructed around the plot to act as dikes for flooding the plots and to provide support to the covering plastic sheet used to prevent evaporation. The monolith was instrumented with a set of ECH_2O EC-20 probes and tensiometers. The sensors were installed at the following depths: 85, 245, 263, 415, 433, 695, 713, 1195, 1463, and 1854 mm. The depth intervals were based on a soil profile description (Table 5.1). Tensiometers with pressure transducers were used to measure the soil water suction (h) at each depth whereas ECH_2O EC-20 probes were utilized for obtaining the volumetric water contents (θ).

Prior to the drainage test, the monolith was continuously irrigated for three consecutive days until it was presumed to be at field saturation. The drainage test was initiated when free water disappeared from 50% of the soil surface. The soil surface was covered with a layer of thick pink (5 cm cotton wool) to minimise extreme temperature fluctuation at the surface during subsequent drainage, and to stop evaporation. Both θ and h were then monitored for a period of 480 hours (20 days). Measurements were recorded on data loggers at 15 minute intervals for the entire drainage period. The ECH_2O EC-20 probes were calibrated for each horizon as described in Chapter 4.

5.2.3 Data processing

The step-by-step handling of the field data of changes in θ and h, and their subsequent processing was carried out as described by Hillel *et al.* (1972), Hillel (1998) and Vachaud & Dane (2002). The procedure and data is presented in the Appendix C to F. Field measurements showing the changes of θ and H with time are presented in Figures 5.1 and 5.2. From the soil water content-time curves the soil water flux (q) through each selected horizon was calculated (Equation 5.3). The hydraulic conductivity, at each depth and for different soil water contents was calculated by dividing fluxes by the corresponding hydraulic gradients $\left(\frac{dH}{dz}\right)$. A semi-log graph (Fig. 5.3) shows the relationship of K to the θ at various depths. Because of failure of tensiometers during the experiment h was estimated using θ-h relationships from the water retention data using *in situ* determined values of θ. Calculations were made for a 480 hour drainage period. Data from hours 0.5, 1, 8, 24, 48, 72, 120, 252, and 480 were analyzed to obtain specific points on the K-θ curve.

5.2.4 Prediction of hydraulic conductivity based on the soil water retention characteristic (laboratory method)

Prediction of unsaturated hydraulic conductivity requires a model containing the same parameters of both the water retention function and the hydraulic conductivity function. Fitting the retention model (Equation 5.4) to measured data (Chapter 3) yields values of parameters (Table 5.4) that were used to predict hydraulic conductivity. The estimated retention parameters for the van Genuchten (1980) retention model are:

$$S_e = \frac{(\theta - \theta_r)}{(\theta_s - \theta_r)} \quad 5.3$$

$$S_e = \left[\frac{1}{1+(\alpha h)^n}\right]^m \quad 5.4$$

where S_e is the dimensionless soil water content, θ is soil water content ($mm^3\ mm^{-3}$), θ_r and θ_s are residual and saturated water contents ($mm^3\ mm^{-3}$), respectively, h is the soil water suction (cm), and α, n and m are constants that define the shape of the curve. The retention data were obtained from desorption measurements on undisturbed cores (Chapter 3) collected at depths of 125, 335, 564, 950 and 1650 mm. Mualem (1976) developed an equation relating the relative hydraulic conductivity $\left(\frac{K(\theta)}{K_s}\right)$ to knowledge of the soil water retention curve. Van Genuchten combined these equations to produce Equation 5.5;

$$K(\theta) = K_{fs} S_e^l \left[1 - \left(1 - S_e^{1/m}\right)^m\right]^2 \quad 5.5$$

where K_{fs} is the saturated hydraulic conductivity, l is an empirical constant assumed equal to 0.5 (Mualem, 1976), and $m = 1 - 1/n$ (van Genuchten, 1980). From Equation 5.5 it is obvious that these predictions require known values of field saturated hydraulic conductivity (K_{fs}). In this study it was measured on each horizon using the double ring apparatus (Table 5.2). Since θ values for the K_{fs} determination are not available the decision was made, following the suggestion proposed by (Basile *et al.*, 2006), to use the field measured θ values at zero time, (coinciding with the placing of the plastic sheet on the surface) during the internal drainage experiments, for each horizon as the θ matching point of K_{fs}. Estimation of van Genuchten retention parameters (Table 5.4) for the θ-h relationship and prediction of K-θ (Equation 5.5) was carried out using the nonlinear parameter estimation RETC programme (van Genuchten *et al.*, 1991).

5.3 Results and discussion

5.3.1 Instantaneous profile method (field method)

Internal drainage patterns: Figure 5.1 presents the water contents in the monolith during the internal drainage test during the 480 hours of the experiment. Each curve on the graph represents the water content at a given depth. The graphs indicate that the water content throughout the profile decreases rapidly at the start of the drainage test, but the rate tapers off after approximately 100 hours. The rapid initial drainage reflects how freely drained the soil profile is. The water content varies with depth and reflects variation in particle size distribution (Table 5.2). The wettest zone occur in the B3 horizon with 22% clay (fine sandy clay loam) followed by B5 with 20% clay (fine sandy loam), then B1 and B2 with 18% clay (fine sandy loam) and A with 8% clay (loam fine sand). The discrepancy in Figure 5.1 is that for the B1, although it has the same clay content as the B2 (Tables 5.1 & 5.2), it retains more water than the B2 horizon. This is probably due to subsoil compaction in the B1 caused by tillage as indicated by slightly higher bulk density (Table 5.2) that has caused a reduction in the total pore space and a change of the pore connectivity. A relevant observation is the presence of fine faint black alluvial humus mottles in the B2 horizon which are conspicuously absent in the B1 horizon (Table 5.1).

The drainage curves presented in Figure 5.1 for the different horizons, various mathematical models (Table 5.3) were used to fit the measured data to characterize the drainage profile using Curve Expert 1.4.

Table 5.3 Mathematical description for soil water content–time data for different Bainsvlei form horizons.

Horizon	Model type	Equation	Coefficients	R^2
A	Exponential Association (3)	y=a(b-exp(-cx))	a = -0.066 b = -3.344 c = 0.016	0.993
B1	MMF Model	y=(a*b+c*x^d)/(b+x^d)	a = 0.314 b = 18.730 c = 0.053 d = 0.306	0.987
B2	MMF Model	y=(a*b+c*x^d)/(b+x^d)	a = 0.305 b = 41.385 c = -0.075 d = 0.369	0.970
B3	MMF Model	y=(a*b+c*x^d)/(b+x^d)	a = 0.331 b = 30.275 c = 0.189 d = 0.509	0.976
B4	Harris Model	y=1/(a+bx^c)	a = 3.086 b = 0.078 c = 0.417	0.978

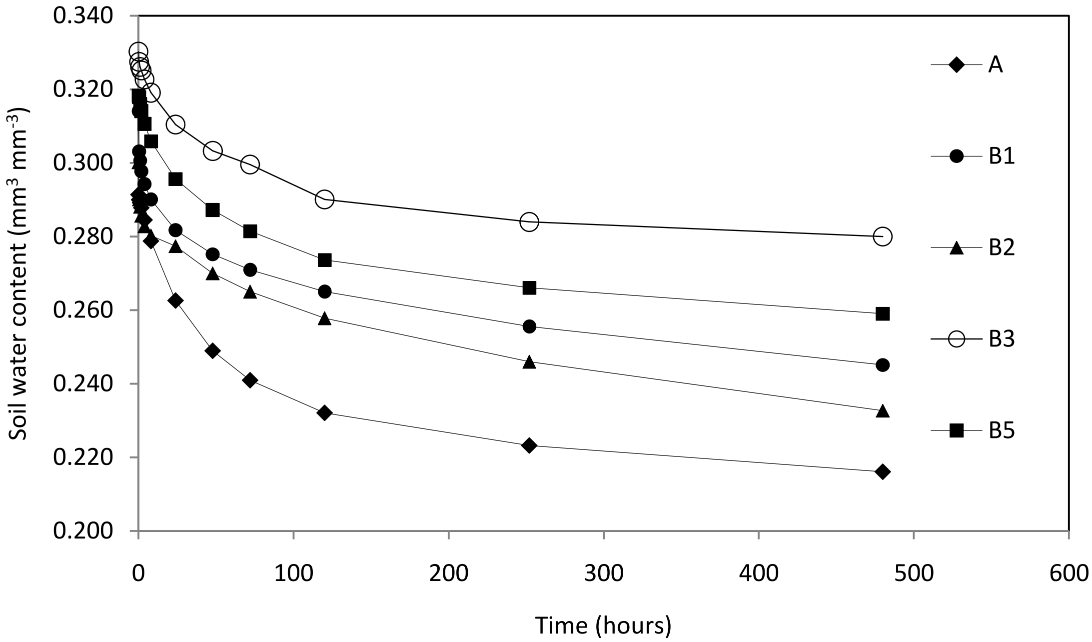


Figure 5.1 Changes in soil water content (θ) with time in the draining Bainsvlei form soil profile.

Hydraulic head (H) profiles (Figure 5.2) were determined by adding h values to elevation heads, i.e. gravitational potential, z in Equation 5.1. Figure 5.2 shows a more or less linear relationship between hydraulic head and depth and therefore a positive gradient and hence a downward flow of water from the upper part of the soil profile to the bottom for the 480 hours drainage period. The hydraulic gradients were derived from Figure 5.2 by integration (Appendix F). The hydraulic gradients obtained by integration varied with depth, however none of them really satisfied the unit gradient assumption. This was consistent with earlier findings that unit hydraulic gradient cannot occur, even for a homogenous soil profile. It is interesting to note that a similar behaviour has been observed by several authors Ahuja *et al.* (1984), Jones & Wagenet (1984), and Reichardt (1993). Therefore, based on the results of the current study, simplified methods based on the unit gradient assumption are not applicable for this soil. This further justifies the use of the instantaneous profile method for the reliable determination of the K-θ relationship of a soil profile.

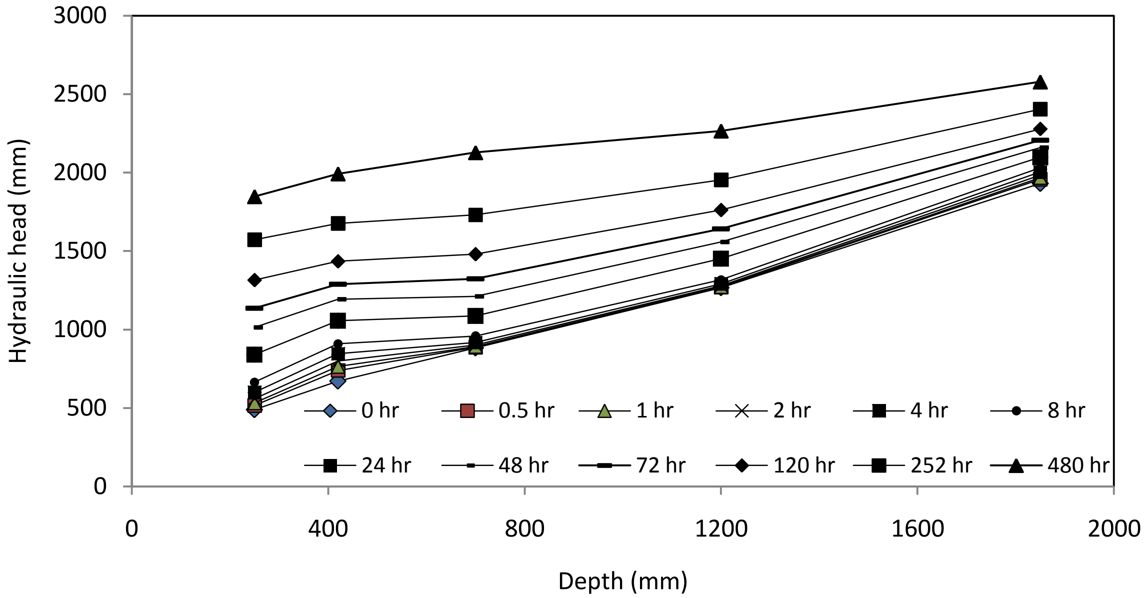


Figure 5.2 Changes in hydraulic head (H) with time in the draining Bainsvlei form soil profile.

Hydraulic conductivity was calculated at known times and water contents $(\bar{\theta})$ by dividing the flux, q, by the change in hydraulic head, $\left(\frac{dH}{dz}\right)$ from Equation 5.2. Detailed calculations are presented in Appendix F. Calculated values of K are plotted against the soil water content in Figure 5.3; each depth is represented by a different symbol. The K-θ relationship varies with depth because of the differences in soil physical properties with depth particularly soil texture and porosity.

Empirical relationships between field measured hydraulic conductivities and soil water content of the form $K = ae^{b\theta}$, where K is hydraulic conductivity, θ is soil water content, and a and b are empirical coefficients characterizing the soil (Hillel *et al.*, 1972) were developed for the four depths. Results are presented in Table 5.4 and Figure 5.3 as solid lines. The values of R^2 were significant (P = 0.01) for all depths. The coefficients for all depths differ markedly. This is consistent with the separate measured K-θ characteristics observed in Figure 5.3, hence not a single curve but a family of curves will suffice to characterize the profile as a whole.

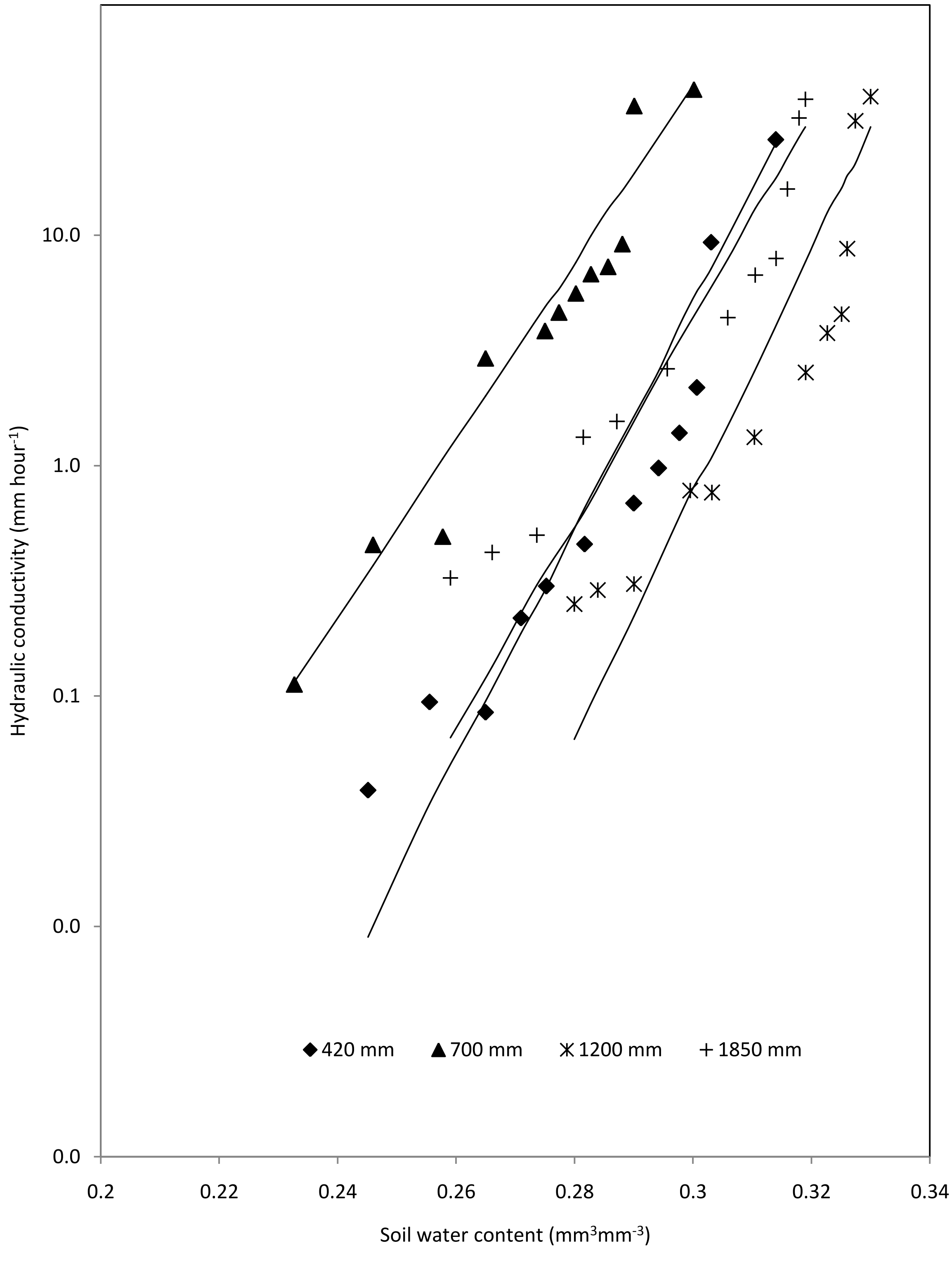


Figure 5.3 Hydraulic conductivities at different depths of the Bainsvlei form soil profile.

Table 5.4 Empirical relationship between field-measured hydraulic conductivity (mm $hour^{-1}$), and soil water content (mm^3 mm^{-3}) for the Bainsvlei form soil (n = 12).

Depth (mm)	K-θ relationship	R^2
420	$K = 7.44x10^{-15}e^{113.8\theta}$	0.976
700	$K = 1.13x10^{-10}e^{98.0\theta}$	0.905
1200	$K = 4.34x10^{-16}e^{118.4\theta}$	0.850
1850	$K = 6.69x10^{-13}e^{99.2\theta}$	0.984

5.3.2 Prediction of the K-θ relationship based on the water retention characteristic (laboratory method)

Water retention characteristics, from which the hydraulic conductivities were predicted are given in Chapter 3 and the retention parameters and statistical indicators, i.e. results of the Willmott test of the reliability, are presented in Table 5.5.

Table 5.5 Parameters of the van Genuchten model and statistical indicators.

Soil depth mm	θ_r mm^3mm^{-3}	θ_s mm^3mm^3	α	n	R^2	D-index	RMSE	$\frac{RMSEu}{RMSE}$
0-250	0.091	0.374	0.58	1.281	0.991	0.995	0.009	1.00
250-420	0.095	0.366	0.43	1.199	0.990	0.998	0.009	1.00
420-700	0.093	0.373	0.73	1.194	0.995	0.997	0.006	1.00
700-1200	0.108	0.371	0.28	1.235	0.996	0.989	0.006	1.00
1200-1450	0.098	0.377	0.58	1.280	0.991	0.998	0.009	1.00
1450-1850	0.113	0.366	0.30	1.528	0.997	0.994	0.005	1.00

In order to validate the results obtained by the van Genuchten (1980) model laboratory method they are compared with K-θ relationships obtained by the instantaneous profile field method in Figure 5.4. The graphic results were quantified by calculating the root mean square error (RMSE), index of agreement (D-index) and the relationship to its unsystematic component $RMSE_u$ ($RMSE_u$/RMSE) (Willmott, 1982). The results of these statistical tests are presented in Table 5.6. Furthermore, with a view to interpreting these statistics more easily, the results obtained by the two methods were compared in Figure 5.4 over the range of water content values observed during the drainage experiment. The values of D-index and $RMSE_u$/RMSE were generally very high, thus indicating good agreement between the two methods and that the predictions were not biased. This was to be expected; given the reasonably accurate description of retention data obtained using the van Genuchten (1980) model (Table 5.5). Considerable deviations do however; occur at K values around 5mmh^{-1} for all the depths, approximately in the region where there is a clearly defined change of slope. An explanation for this is not available at present. These results are consistent with earlier findings by Paige and Hillel (1993) who found that hydraulic conductivities predicted using the van Genuchten model agreed closely for corresponding ranges of water content in a fine sandy loam.

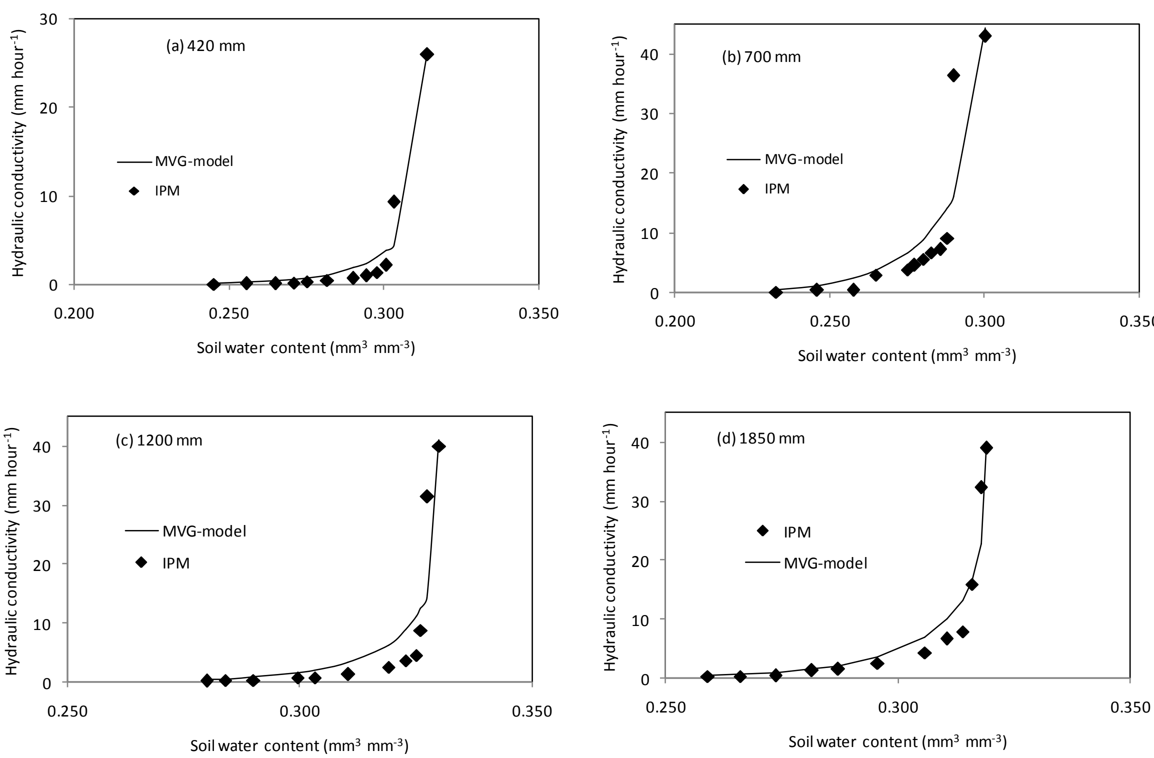


Figure 5.4 Comparison of hydraulic conductivity-water content relationships obtained from the instantaneous profile field method (IPM) and the van Genuchten (1980) model (MVG) laboratory method.

Table 5.6 Comparison of hydraulic conductivity-water content relationships obtained from instantaneous profile field method and the van Genuchten (1980) model laboratory method based on the water retention curve

Soil depth (mm)	RMSE	D-index	$RMSE_u$/RMSE
420	1.050	0.994	0.981
700	1.449	0.996	0.886
1200	1.237	0.967	0.875
1850	1.062	0.997	0.964

5.3.3 Estimation of the drained upper limit (DUL)

5.3.3.1 Using the water retention characteristic

Knowledge of the DUL value is useful for many purposes and is frequently used in agronomic, engineering and environmental applications. Results of DUL estimations based on the θ-h relationships, with θ at h = 10 kPa for the DUL value of each horizon are presented in Table 5.7. The results are compared with the field determined DUL values made during this study. The results reveal a slight discrepancy between the *in situ* and laboratory estimates for 1200 mm depth, viz. 306 mm and 296 mm respectively. There are furthermore differences in the individual horizons. It may be significant that the estimates from the θ-h relationship in horizons are less than the field determined values for all the horizons. Nevertheless, given the potential of soil natural variability and these differences are within the plausible range of spatial variability.

It has long been realised that to relate the DUL of soils with a specific point on the θ/h curve has shortcomings (Ritchie, 1981; Romano & Santini, 2002). It has been reported that the value of h at DUL varies widely in different soil profiles. Some investigators have used tensiometers to measure soil water suction values during field tests on freely drained, fairly homogenous soil profiles with different textures. Values for different soils were found to vary from 0.1 kPa to 0.9 kPa in determinations on 14 fairly homogenous South African soil profiles varying in clay content from 1 to 33% (Beukes, 1984); 0.5 to 35 kPa (Romano & Santini 2002). Gebregiorgis & Savage, (2006) found a DUL estimate based a water retention curve procedure, using 33 kPa as the equilibrium value, to give a higher water content than the field measured value. These results support our findings that any attempt to estimate DUL values using a specific h value for the whole profile cannot be recommended. These results also suggest that if accuracy is of paramount importance such as in soil-water balance calculations, laboratory-estimated DUL values should be used with caution, field measurements being preferable (Ritchie, 1981; Ratliff *et al.*, 1983; Romano & Santini, 2002).

Table 5.7 Comparison of *in situ* determined vs. laboratory estimated DUL values

Horizon	Laboratory (mm)	In situ (mm)
A	54.	56
B1	41	42
B2	66	67
B3	135	141
1200 mm (depth)	296	306

5.3.3.2 Using the K-θ relationship

Romano and Santini (2002) suggested that DUL 'should be more related to the hydraulic conductivity than to the soil water matric potential'. Support for this suggestion is provided by the results in Figure 5.4 which show that there is almost perfect agreement between measured and predicted K-θ values at the lower end, i.e. DUL end, of the K-θ relationship when K has decreased to a very low value. Since field measured DUL determination are laborious and expensive, using the laboratory method as describe here offers an attractive alternative, at least for freely drained soils.

5.4 Conclusion

Our study suggests that the most accurate method for determining the K-θ relationships and DUL probably depends upon the intended application. The best method will depend upon such additional factors as available human and financial resources, and existing data. For our purpose the instantaneous profile method was successfully used to determine unsaturated hydraulic conductivities at four depths in the top 1850 mm depth profile of a Bainsvlei form soil. The K-θ showed variability from depth to depth to the extent that not a single curve but a family of curves would suffice to characterize the profile as whole, reflecting differences in soil physical properties especially particle size distribution. The results also demonstrate that the ECH_2O EC-20 probe is a practical tool if properly calibrated that can be used effectively to determine *in situ* hydraulic properties at different depths in a soil profile.

The van Genuchten (1980) model used together water retention data (laboratory method) provided reliable estimates of unsaturated hydraulic conductivity that were very close to field measurements obtained from the internal drainage experiment. Hence, from a practical perspective, the prediction of K-θ relationship from water retention data can be a viable alternative for determining the hydraulic properties of soil profiles similar to the one examined in this study. It appears that the accuracy of the predicted K is quite sensitive to the shape of the θ-h relationship, the field saturated water content value and the K_{fs} value. Thus, accurate measurement of these parameters is necessary for successful use. Results also showed that for this profile the laboratory method was able to provide a reliable prediction of DUL. The prediction of DUL using θ-h relationship has been found to be satisfactory.

5.5 References

AHUJA, L., BARNES, B., CASSEL, D., BRUCE, R. & NOFZIGER, D., 1984. Effect of assumed unit gradient during drainage on the determination of unsaturated hydraulic conductivity and infiltration parameters. *Soil Sci.* 145, 235-243.

BANDARANAYAKE, W., PARSONS, L., BORHAN, M. & HOLETON, J., 2007. Performance of a capacitance-type soil water probe in a well drained sandy soil. *Soil Sci. Soc. Am. J.* 71, 993-1002.

BASILE, A., COPPOLA, A., MACSELLIS, R. & RANDAZO, L., 2006. Scaling approach to deduce field unsaturated hydraulic properties and behavior from laboratory measurements on small cores. *Soil Sci. Soc. Am. J.* 5, 1005-1016.

BEUKES, D. 1984. The effect on certain soil properties on matric at, and time duration to, field capacity. *S. Afr. J.Plant Soil.* 1, 125-131.

BROOKS, R. & COREY, A., 1964. Hydraulic properties of porous media. Hydrology Paper, Vol. 3, Colorado State Univ. 1-15.

CAMPBELL, G., 1974. A simple method for determining unsaturated hydraulic conductivity from moisture retention data. *Soil Sci.* 117, 311-317.

DAVIDSON, J., NIELSEN, D. & LA RUE, M., 1969. Field measurement and use of soil water properties. *Water Resour. Res*. 5, 1312-1321.

FIELD, J., PARKER, J. & POWELL, N., 1983. Comparison of field and laboratory measured and predicted hydraulic properties of a soil with macropores. *Soil Sci.* 138, 385-396.

GEBREGIORGIS, M. & SAVAGE, M. 2006. Field, laboratory and estimated soil-water content limits. *Water SA.* 32(2), 155-162.

GEBREGIORGIS, M. & SAVAGE, M., 2006. Field, laboratory and estimated soil-water content limits. *Water SA.* 32, 155-162.

GREEN, R., AHUJA, L. & CHONG, S., 1986. Hydraulic conductivity, diffusivity, and sorptivity of unsaturated soils: field methods. *In:* A. Klute, (ed.). Method of soil analysis, Part I, Physical and mineralogical methods Monograph No. 9. ASA., Madison, Wisconsin, 771-798.

HENSLEY, M., BOTHA, J., ANDERSON, J., VAN STADEN, P. & DU TOIT, A., 2000. Optimising rainfall use efficiency for developing farmers with limited access to irrigation water. *Water Research Commission (WRC) report* No 878/1/00, Pretoria.

HILLEL, D., 1998. Environmental of soil physics. Academic Press Inc, New York.

HILLEL, D., KRENTOS, V. & STYLIANOU, Y., 1972. Procedure and test of an internal drainage method for measuring soil hydraulic characteristics *in situ*. *Soil Sci.* 114, 395-400.

JONES, A. & WAGENET, R., 1984. *In situ* estimation of hydraulic conductivity using simplified methods. *Water Resour. Res.* 20, 1620-1626.

JURY, W. & HORTON, R. 2004. Soil physics 6th edition. John Wiley & Sons, New York.

KOSUGI, K., 1996. Lognormal distribution model for unsaturated soil hydraulic properties. *Water Resour. Res.* 32, 891-901.

MALLANTS, D., JACQUES, D., TSENG, P., van GENUCHTEN, M.T. & FEYEN, J., 1997. Comparison of three hydraulic property measurement methods. *J. Hydrol*. 199, 295-318.

MARION, J., OR, D., ROLSTON, D., KAVVAS, M. & BIGGAR, J., 1994. Evaluation of methods for determining soil-water retentivity and unsaturated hydraulic conductivity. *Soil Sci.* 158, 1-13.

MUALEM, Y., 1976. A new model for predicting the hydraulic conductivity of unsaturated porous media. *Water Resour. Res. 12*, 513-522.

OR, D. & WRAITH, J., 2002. Soil water content and water potential relationships. *In:* A. Warrick, (ed.). Soil physics companion. CRC press, Florida. 49-84.

PAIGE, G. & HILLEL, D., 1993. Comparison of three methods for assessing soil hydraulic properties. *Soil Sci.* 155, 175-189.

RATLIFF, L., RITCHIE, J. & CASSEL, D. 1983. A survey of field-measured limits of soil water availability as related to laboratory-measured properties. *Soil Sci. Soc. Am. J. 47*, 750-775.

REICHARDT, K., 1993. Unit gradient in internal drainage experiments for the determination of soil hydraulic conductivity. *Sci. Agric. Piracicaba*, 50(1), 151-153.

RITCHIE, J., 1981. Soil water availability. *Plant Soil. 58*, 327-338.

ROMANO, N. & SANTINI, A., 2002. Field. *In:* J. Dane, & G. Topp, (eds.). Methods of soil analysis, Part 4, Physical methods, Book Series No 5, SSSA., Madison, Wisconsin, 721-738.

ROSS, P. & PARLANGE, J., 1994. Investigation of a method for deriving unsaturated hydraulic properties from water content profiles. *Soil Sci.* 157, 335-340.

SAXTON, K., RAWLS, W., ROMBERGER, J. & PAPENDICK, R., 1986. Estimating generalized soil-water characteristics from texture. *Soil Sci. Soc. Am. J.* 50, 1031-1036.

SOIL CLASSIFICATION WORKING GROUP., 1991. Soil classification a taxonomic system for South Africa. Department of agricultural development, Pretoria, South Africa.

THE NON-AFFILIATED SOIL ANALYSIS WORK COMMITTE, 1990. Handbook of standard soil testing methods for advisory purposes. SSSSA., Pretoria, South Africa.

TYNDALE-BISCOE, P. & MALANO, H., 1995. A laboratory comparison of some currently available soil moisture monitoring devices. *Agric. Eng. Aust.*, 1, 18-24.

VACHAUD, G. & DANE, J., 2002. Instantaneous profile method. *In:* J. Dane, & G. Topp, (eds.). Methods of soil analysis, Part 4, Physical methods, SSSA., Book Series, No 5, Madison, Wisconsin, 937-962.

VAN GENUCHTEN, M. T., 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Sci. Soc. Am. J.* 44, 892-898.

VAN GENUCHTEN, M. T., LIEJI, F. & LUND, L., 1992. Indirect methods for estimating the hydraulic properties of unsaturated soils. Proceedings of international workshop on unsaturated hydraulic properties. University of California, Riverside: CA.

VAN GENUCHTEN, M., LEIJI, F. & YATES, S., 1991. The RETC code for quantifying functions of unsaturated soils. EPA/600/2-91/065.

WILLMOTT, C.J., 1982. Some comments on the evaluation of model performance. *Am. Met. Soc.* 63, 1309-1313.

WOSTEN, J., PACHEPSKY, Y. & RAWLS, W., 2001. Pedotransfer function: bridging the gap between available basic soil data and missing soil hydraulic characteristics. *J. Hydrology*, 251, 123-150.

ZHANG, S., LOVDAHL, L., GRIP, H. & TONG, Y., 2007. Soil hydraulic properties of two loess soils in China measured by various field-scale and laboratory methods. *Catena* 69, 264-273.

CHAPTER 6

Comparing laboratory and field determined hydraulic conductivity for a Tukulu form soil

6.1 Introduction

6.1.1 Motivation

The Tukulu form soil being studied here is a duplex soil. The C-horizon in which signs of wetness occur has a prismatic structure and the B/C transition is clearly defined. Large areas of duplex soils occur in South Africa. For example it is estimated that in 35% of the area of Free State Province, duplex soils are dominant (Hensely *et al.,* 2006). A large fraction of this area is suitable for in-field rain water harvesting (IRWH). Because IRWH is a suitable technique for semi-arid areas, the hydraulic properties of these soils are important. This information is particularly essential for the understanding of the soil water behaviour, including for example, the movement of water and solutes within the profile and in studies of water uptake by plant roots. There are, however, few published observations on the hydraulic properties of duplex soils (Olsson & Rose, 1978).

A literature study and necessary theoretical aspects of this subject are dealt with in Chapter 5.

6.1.2 Objectives of the study

The general objective of the study was to investigate and compare the hydraulic properties of the Tukulu form soil using field and laboratory methods. Specific objectives were: (i) to characterize the *in situ* K-θ relationships for the profile; (ii) to predict the K-θ relationships with the van Genuchten (1980) model; (iii) to estimate the drained upper limit.

6.2 Materials and methods

6.2.1 Site location and pedological characteristics

The study was conducted at Paradys ($29^{0}13'$ $25''$ S, $26^{o}12'$ $08''$ E, altitude 1417 m) an experimental field station of the University of the Free State. The profile studied is classified as Tukulu form, *Dikeni family* (Soil Classification Working Group, 1991). Table 6.1 provides a detailed description of the soil profile. Each of the selected diagnostic horizons were chemically and physically analyzed for exchangeable cations (Ca, Mg, Na, K), $pH_{(water)}$, and particle size distribution, using procedures described by The Non-Affiliated Soil Analysis Work Committee (1990). Field saturated hydraulic conductivity (K_{fs}) was determined using double ring procedure on each horizon. Table 6.2 presents some of the relevant soil properties of the diagnostic horizons of the soil.

Table 6.1 Profile description for the Tukulu form soil

Map/photo :	2926 Bloemfontein	Soil form and family:	Tukulu *Dikeni*
Latitude + Longitude:	29° 13’ 25’’/26° 12’ 08’’	Surface rockiness:	None
Altitude:	1417 m	Occurrence of flooding:	None
Terrain unit:	Midslope	Wind erosion:	Slight wind
Slope:	1%	Water erosion:	None
Slope shape:	Straight	Vegetation/Land use:	Agronomic field crops
Aspect:	South	Water table:	None
Microrelief:	None	Described by:	M. Hensley & J. Chimungu
Parent Material Solum:	Origin binary, aeolian, solid rock	Date described:	12/06/09
Underlying Material:	Sandstone (Feldspathic)	Weathering of underlying material:	Moderate physical, moderate chemical
		Alteration of underlying material:	Ferruginised

Horizon	Depth (mm)	Description	**Diagnostic horizon**
A	0-270	Moist state; dry colour: reddish brown (2.5YR5/4); moist colour: reddish brown 2.5YR4/4; texture: fine sandy loam; structure: apedal massive: consistence: friable; few fine pores; common roots; gradual transition.	Orthic
B1	270-500	Moist state; dry colour: reddish brown (2.5YR5/4); moist colour: reddish brown (2.5YR4/4); texture: fine sandy clay loam; structure: apedal massive becoming weak subangular blocky towards transition: consistence: friable; few fine pores; few clay cutans; very few fine pore; common roots; clear, tonguing transition.	Neocutanic
C1	500-800	Moist state; dry colour. Dark greyish brown 2.5YR4/2; moist colour. grey 2.5Y5/0; texture: clay; common distinct grey and yellow reduced iron oxide mottles: few prominent black oxidised iron oxide mottles; structure: prismatic; consistence: firm; few slickensides; common clay cutans; very few roots; gradual transition.	Unconsolidated material, with signs of wetness
C2	800-±1350	Allows water to enter, hence the presence of roots. It chops out of the profile relatively easily up to ±1350 mm, getting harder towards the bottom and probably very impermeable slightly deeper.	

Table 6.2 Summary of chemical and physical characteristics of Tukulu form soil profile.

	A	B1	C1
Physical properties			
Coarse sand (2 - 0.5 mm) %	3.1	1.6	1.3
Medium sand (0.5 - 0.25) %	2.8	2.7	1.6
Fine sand (0.25 - 0.106 mm) %	42.4	42.2	23.2
Very fine sand (0.106 - 0.53mm) %	25.6	19.6	9.9
Silt %	6.0	3.0	8.0
Clay %	18.6	28.9	54.8
Bulk density (Mg m^{-3})	1.67	1.68	1.71
*K_{fs} (mm $hour^{-1}$)	10.0	6.7	3.6
Chemical properties			
pH(water)	5.8	6.1	6.5
Ca ($cmol_c.kg^{-1}$)	35.3	40.8	62.7
Mg ($cmol_c.kg^{-1}$)	18.8	20.8	54.7
K ($cmol_c.kg^{-1}$)	12.8	6.8	12.9
Na ($cmol_c.kg^{-1}$)	2.4	2.5	3.4

**K_{fs} is field saturated hydraulic conductivity-determined by double ring procedure.*

6.2.2 Field measurements

The internal drainage experiment was carried out on a (4 m x 4 m) x 1 m depth monolith following the procedure described by Hillel *et al.* (1972) and Hillel (1998). The monolith was isolated vertically by a plastic sheet to inhibit lateral water movement. Earth bunds were constructed around the plot to act as dikes for flooding the plots and to provide support to the covering plastic sheet used to prevent evaporation. The monolith was instrumented with a set of ECH_2O EC-20 probes and tensiometers. The sensors were installed at the following depths: 100, 265, 275, 495, 505, and 795 mm, corresponding to horizon boundaries. The depth intervals were based on a soil profile description (Table 6.1). Tensiometers with pressure transducers were used to measure the soil water suction (h) at each depth whereas ECH_2O EC-20 probes were utilized for obtaining the volumetric water contents (θ).

Prior to the drainage test, the monolith was continuously irrigated for three consecutive days until it was presumed to be at field saturation. The drainage test was initiated when free water disappeared from 50% of the soil surface. The soil surface was covered with a layer of thick pink (5 cm cotton wool) to minimise extreme temperature fluctuation at the surface during subsequent drainage, and to stop evaporation. Both θ and h were then monitored for a period of 624 hours (26 days). Measurements were recorded on data loggers at 15 minute intervals for the entire drainage period. The ECH_2O EC-20 probes were calibrated for each horizon as described in Chapter 4.

6.2.3 Data processing

The step-by-step handling of the field data of changes in θ and h, and their subsequent processing was carried out as described by Hillel *et al.* (1972), Hillel (1998) and Vachaud & Dane (2002). Field measurements showing the changes of θ and H with time are presented in Figures 6.1 and 6.2. From the soil water content-time curves the soil water flux (q) through each selected horizon was calculated. Calculations were made for a 624 hour drainage period. Data from hours 0.5, 1, 8, 12, 24, 48, 72, 120, 264, 480 and 624 were analyzed to obtain specific points on the K-θ curve.

6.2.4 Prediction of hydraulic conductivity based on soil water retention characteristic (laboratory method)

The relevant theory is presented in Chapter 5. Retention data is presented in Chapter 3.

6.3 Results and discussion

6.3.1 Instantaneous profile method (field method)

Vertical textural and structure properties (Tables 6.1 & 6.2) strongly influenced water movement (internal drainage) through the soil profile and drainage out of the C-horizon. The mechanism of water redistribution under natural conditions is poorly understood for the duplex soil being studied. There is a sharp change in texture and structure at the B/C transition at a depth of 500 mm (Table 6.1). The B horizon has a fine sandy clay loam texture and apedal massive structure which becomes weak subangular blocky just above the transition. The C-horizon has a clay texture and a prismatic structure when dry. This sharp transition inhibits water flow down the profile. It also results in intermittent water-logging in the C-horizon as evidenced by the grey mottles on that horizon. Similar conditions are reported in other duplex soils, together with some degree of lateral flow (Rutherglen, 2006). The drainage rate from the profile was relatively low for the entire drainage period (624 hours). This is attributed to the inability of the C-horizon to conduct water flowing out of the upper solum (A and B horizons). The measured water contents for the A and B horizons show a steady but relatively slow loss of water after field saturation, as indicated by the exponential decrease in the drainage curves of the A and B horizons in Figure 6.1. The C-horizon, however, shows little change of water content. This is attributed to its high content of expanding clay (55%). The drainage out of the A and B horizons in Figure 6.1 is attributed to lateral flow followed by leakage out of the monolith along its edges between the plastic sheet and earth wall. This is the only reasonable way to explain how approximately 320 litres of water flowed out of the A and B horizons during the 624 hours drainage period without passing through the C-horizon, which only lost about 38 litres of water during this period. This process was not expected when designing the experiment. Although it shows that the results of the instantaneous profile method of determining the K-θ relationship in this case are invalid, the measurements made will nevertheless be presented as they are of value for other purposes. If the whole landscape was saturated, drainage from the profile would presumably occur at the rate determined by the C-horizon as shown in Figure 6.1, approximately at an average of 0.001 mm $hour^{-1}$ over the 624 hour period. This prolonged period of wetness, which would probably occur very infrequently in this semi-arid area, is presumably the cause of the hydromorphic mottling in the C-horizon.

The drainage curves presented in Figure 6.1 for the different horizons, various mathematical models (Table 6.3) were used to fit the measured data to characterize the drainage profile using Curve Expert 1.40.

Table 6.3 Mathematical description for soil water content–time data for different Tukulu form horizons.

Horizon	Model type	Equation	Coefficients	R^2
A	MMF Model	y=(a*b+c*x^d)/(b+x^d)	a = 0.300	0.963
			b = 9.506	
			c = 0.231	
			d = 0.396	
B	MMF Model	y=(a*b+c*x^d)/(b+x^d)	a = 0.313	0.956
			b = 16.701	
			c = 0.227	
			d = 0.425	
C	Harris Model	y=1/(a+bx^c)	A =2.918	0.987
			B = 0.001	
			C = 0.405	

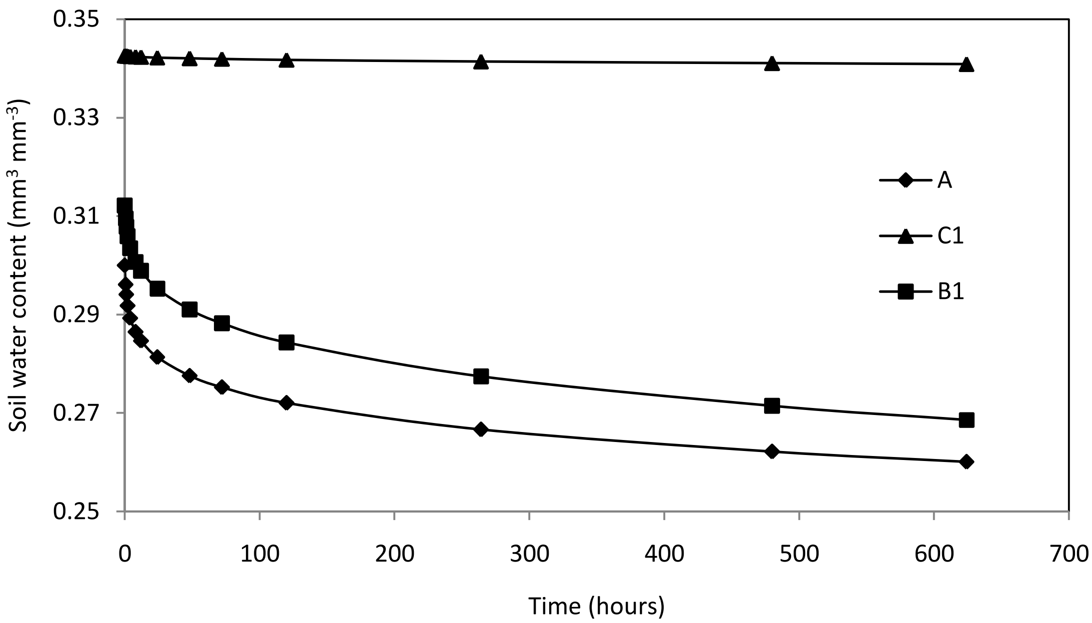
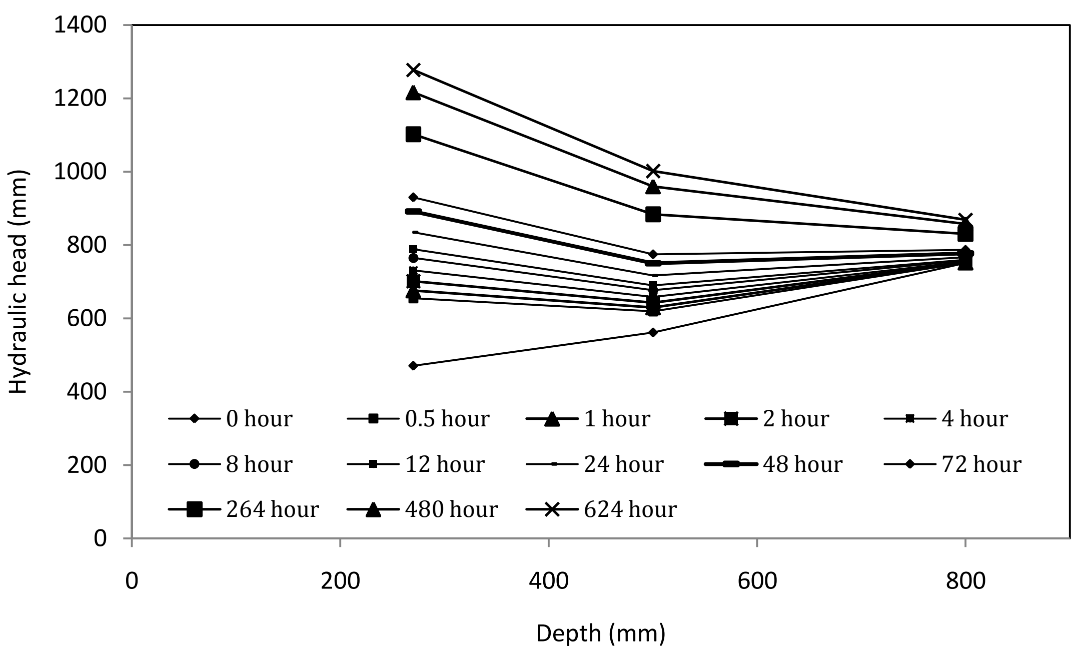


Figure 6.1 Selected measured changes in soil water content (θ) with time in the different horizons of the Tukulu form soil profile as it appears to drain naturally from field saturation.



Figures 6.2 Measured hydraulic head (H) changes with time in the Tukulu form soil profile as it appears to drain naturally.

Figure 6.2 illustrates hydraulic head profiles that are in fact invalid, but nevertheless show to some extent the differences in soil texture and layering as well as the impaired internal drainage as depicted in Figure 6.1. Graphical differentiation of the curves in Figure 6.2 yielded negative dH/dz values rendering it impossible to determine hydraulic characteristics of the soil profile using the instantaneous profile method. It is interesting to note that a similar behavior was observed by Paige & Hillel (1993) in their studies on a Typic Dystrochrept soil. These authors called this behaviour an anomaly and attributed it to abrupt changes in texture and bulk density down the profile.

To solve the problem highlighted above the unit gradient assumption is invoked (Hillel, 1998). It allows K-θ to be set equal to the water flux (q) (Equation 5.2). As discussed above drainage during the course of the experiment allowed sufficient resolution to calculate 'apparent hydraulic conductivities' at only two depths (i.e. 270 mm & 500 mm). Calculated values of K are plotted against the θ in Figure 6.3 for each of the two depths. The K-θ relationship varies with depth in the profile because of differences in particle size distribution and structure. It is important to note that the flux calculated for the 270 mm and 500 mm depths comprises both lateral (towards the edges of the monolith) and vertical water movement, the latter being minimal considering the presence of the restrictive C-horizon.

Empirical relationships between 'apparent field measured hydraulic conductivities' and soil water content of the form $K = ae^{b\theta}$, where K is hydraulic conductivity, θ is soil water content, a and b are empirical coefficients characterizing the soil (Hillel *et al.,* 1972) were developed for the two depths. Results are presented as solid lines in Figure 6.3 and the equations are given in Table 6.4. The values of R^2 were significant (P = 0.05) for the two depths and the coefficients differed markedly. This is consistent with the measured 'apparent K-θ characteristics' in Figure 6.3.

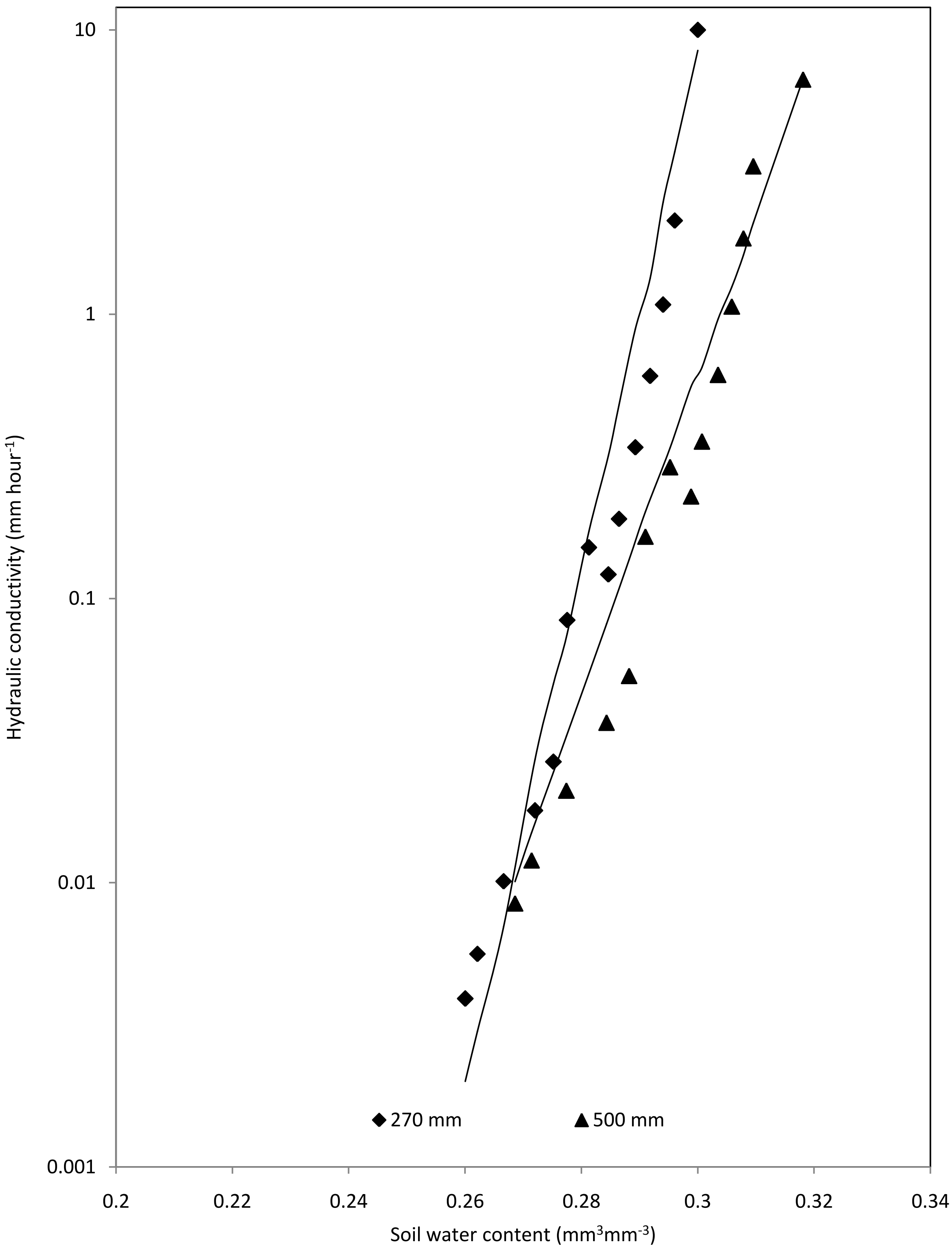


Figure 6.3 The 'apparent K-θ relationships' for the different depths of the Tukulu form soil profile.

Table 6.4 Empirical relationship between field-measured hydraulic conductivity (mm hour^{-1}), and soil water content (mm^3 mm^{-3}) for the Tukulu soil (n=14).

Depth (mm)	K-θ relationship	R^2
270	$K = 1.63x10^{-26}e^{205.0\theta}$	0.952
500	$K = 7.68x10^{-18}e^{129.9\theta}$	0.987

6.3.2 Prediction of the K-θ relationship based on the water retention characteristic (laboratory method)

Water retention characteristics, from which the hydraulic conductivities were predicted, are given in Chapter 3. The van Genuchten parameters and statistical indicators, i.e. results of the Willmott test of the reliability of the model, are presented in Table 6.5.

Table 6.5 Parameters of the van Genuchten model and statistical indicators.

Soil depth mm	θ_r	θ_s	α	n	R^2	D-index	RMSE	$\frac{RMSEu}{RMSE}$
0-270	0.102	0.369	0.60	1.177	0.991	0.996	0.008	1.00
270-500	0.116	0.356	0.62	1.272	0.993	0.997	0.006	1.00
500-900	0.229	0.351	0.04	1.095	0.980	0.998	0.005	1.00

It has been shown in Section 6.3.1 that it has not been possible to determine reliable values of the K-θ relationship for the soil profile as a whole, or for the different depths separately, by the instantaneous profile method. The graphs presented in Figure 6.3 are termed 'apparent' because it was necessary to invoke the unit gradient assumption which does not infact reflect what actually happens under natural conditions (Reichardt, 1993). In contrast, the undisturbed core samples used to determine the water retention characteristics will reflect the true K-θ relationship (assumimg reliable prediction by the van Genuchten model) specific for each horizon, separated from the profile as a whole. Since there is no comparable data from the instantaneous profile method, meaningful comparisons were not possible. The parameters in Table 6.5 were then used to predict the hydraulic conductivity using the van Genuchten from field saturation to drained upper limit. Graphs showing predictions of K-θ are presented in Figure 6.4.

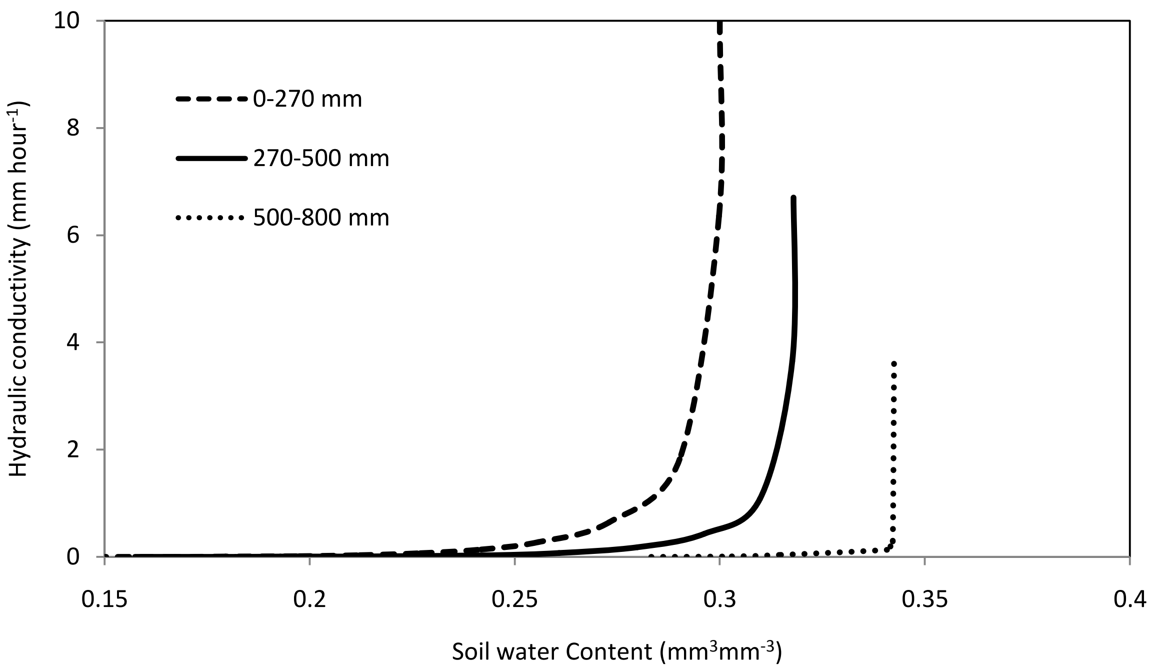


Figure 6.4 van Genuchten model predicted hydraulic conductivity-water content relationships based on laboratory water retention data.

The results in Figure 6.4 show that separate K-θ relationships may be described for the various horizons in the soil profile. Hydraulic conductivity is observed to decrease more rapidly with decreasing θ in the A and B horizons than in the C-horizon. The low conductivity of the C-horizon is known to be important in restricting profile drainage as shown in Figure 6.1.

6.3.3 Estimation of drained upper limit (DUL) from K-θ relationship predicted from water retention data

The K-θ relationships for the three horizons are presented in Figure 6.4. The hydraulic conductivity of the C-horizon controls drainage out of this profile under natural conditions. This concept will be used to estimate what the DUL value would be after 624 hours of draining under natural conditions, i.e. if the whole landscape had been saturated and no evapotranspiration occurred during this period. Each horizon will be considered separately. The data in the appendix show that θ in the C-horizon at 624 hours is 0.340 mm^3mm^{-3} (equivalent to 102 mm) field measurement of the drainage rate from the C-horizon at 624 hours was 0.001 mm $hour^{-1}$ by average. This means that the drainage rate from the A and B horizons must also have been 0.001 mm $hour^{-1}$ at 624 hours.

Interpolating this drainage rate into the K-θ curves of the A and B horizons provide an estimate of their θ values at 624 hours (i.e. DUL) under natural conditions already specified. It is important to note that field measured drainage rate was used to make this prediction. Alternative procedure is to use the hydraulic predicted by the van Genuchten model. This is a more convenient procedure as if reliably it eliminates the need for the laborious field measurements. The van Genuchten model predicted hydraulic conductivity change dramatically at $\theta = 0.34\ mm^3mm^{-3}$ for the C-horizon (Figure 6.4) at this θ hydraulic conductivity is 0.1 mm $hour^{-1}$. The van Genuchten predicted θ for the A and B horizons at this hydraulic conductivity value as 65 mm and 60 mm respectively. DUL for the profile for this procedure is therefore 227 mm. Results by two procedures are presented in Table 6.6. The results reveal a slight discrepancy between the estimation using the measured drainage rate from the C-horizon at 624 hours and using the drainage rate predicted by the van Genuchten model, i.e. at the point in Figure 6.4 where there is as a abrupt change in the C-horizon viz. 189 mm and 227 mm respectively. There are furthermore differences in the individual horizons.

Table 6.6 Estimation of drained upper limit (DUL) from predicted K-θ relationships

Horizon	**Depth (mm)**	**Estimated DUL (mm)**	
		A*	B*
A	0 – 270	45	65
B	270- 500	41	60
C	500-800	102	102
	Total for the profile	**189**	**227**

A using the measured drainage rate from the C-horizon at 624 hours*
B using the drainage rate predicted by the van Genuchten model, i.e. at the point in Figure where there is a sharp change in the C-horizon.*

6.4 Conclusion

The instantaneous profile field method is regarded as a reference method to measure *in situ* unsaturated hydraulic conductivity for both homogenous and layered soils (Hillel *et al.*, 1972). There are, however, several site or profile characteristics that may limit this method (Bouma, 1983). Our studies show that it is not applicable on duplex soils with slowly permeable C-horizons like the Tukulu form profile at Paradys, because of negative hydraulic gradients within the profile due to impaired internal drainage. Water retention characteristics were used to predict K-θ relationships for the three horizons. It was not possible to validate this prediction, because it was not possible to determine reliable values of the K-θ relationship *in situ*.

It was possible to make a reasonably reliable prediction of DUL for the profile using water retention data and the van Genuchten model. It has not been possible to validate this prediction in the present study.

6.5 References

BOUMA, J., 1983. Use of soil survey data to select measurement techniques for hydraulic conductivity. *Agriculture Water Management,* 6, 177-190.

HENSLEY, M., LE ROUX, P.A.L., DU PREEZ, C.C., VAN HUYSSTEEN, C., KOTZE, E. & VAN RENSBURG, L.D., 2006. Soils: The Free State agricultural base. *South African Geographical Journal*, 88, 11-21.

HILLEL, D., 1998. Environmental soil physics, Academic Press Inc, New York.

HILLEL, D., KRENTOS, V. & STYLIANOU, Y., 1972. Procedure and test of an internal drainage method for measuring soil hydraulic characteristics *in situ*. *Soil Sci.* 114, 395-400.

OLSSON, K. & ROSE, C., 1978. Hydraulic properties of a red brown earth determined from in situ measurements. *Aust. J. Soil. Res.* 16, 169-180.

PAIGE, G. & HILLEL, D., 1993. Comparison of three methods for assessing soil hydraulic properties. *Soil Sci.,* 155, 175-189.

REICHARDT, K., 1993. Unit gradient in internal drainage experiments for the determination of soil hydraulic conductivity. *Sci. Agric. Piracicaba*, 50(1), 151-153.

RUTHERGLEN, C.H., 2006. Small farm: Infiltration and drainage of soils. Agriculture notes, Department of primary industries, state of Victoria, Australia.

SOIL CLASSIFICATION WORKING GROUP, 1991. Soil classification a taxonomic system for South Africa. Department of Agricultural Development, Pretoria, South Africa.

VACHAUD, G. & DANE, J., 2002. Instantaneous profile method. *In:* J. Dane, & G. Topp, (eds.). Methods of soil analysis, Part 4, Physical methods, SSSA., Book Series, No 5, Madison, Wisconsin, 937-962.

VAN GENUCHTEN, M. T., 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Sci. Soc. Am. J.* 44, 892-898.

CHAPTER 7

Summary and recommendations

7.1 Summary

An understanding of hydrological processes is essential for assessing water resources as well as the changes to the resource caused by changes in the land use or climate. There is a relative abundance of literature dealing with the theory and application of soil hydraulic properties, but there are very few, if any systematic studies that give detailed description of soil hydraulic properties on Bainsvlei and Tukulu form soils. This information is needed for improving understanding of the effects of soil management or land use on soil profile hydrology. This study was intended to help fill this gap by measuring soil hydraulic properties of these soils and utilize the results from different approaches and compare the results to validate their applicability.

The two soil profiles were successfully characterized and classified, hereby providing the basic information needed to fulfil requirements of a valuable pedotransfer function and the appropriate vehicle to carry the soil information needed to optimize the biological productivity of these ecotopes. The soil at Kenilworth was classified as Bainsvlei form and at Paradys was classified as Tukulu form. The detailed profile descriptions proved useful in providing a picture of the various horizons of the two soils and differences in horizons, mainly regarding their colour and structure. The soil analysis was useful to distinguish soil layers within the profile.

The soil water retention characteristics were successfully determined using the hanging water column and pressure plate apparatus. The water retention characteristics for both soils were generally well defined with little variability between replicates. The main variations were due to texture differences between the horizons. The clay rich horizons were characterized by high water retention resulting in flat shapes in figures of water content vs. soil water suction. The measured water retention data corresponded well with fitted curve by the van Genuchten model. Thus, it appears that this mathematical model is applicable to the soils studied.

The structural and textural domains were defined basing on the principal of fixed boundary (h = 10 kPa) from θ-h relationships. The suction value separating the structural domain from the matrix domain varies from horizon to another horizon. Hence, the boundary between soil pore categories cannot be taken as a fixed value for all soils. Although the fixed boundary method has been heavily criticised by several authors, offers promise for separating the structural and the textural domains of soil porosity. The structural domain varies with clay content. It has been shown that increase clay content in a given soil is associated with decrease in structural domain.

In contrast to most previous studies that derived laboratory calibrations from repacked soil samples, in this ECH_2O EC-20 probe were calibrated using undisturbed soil core to replicate *in situ* physical conditions. Horizon specific calibration was found to be essential to get sensible water content estimates from ECH_2O EC-20 probes if used in different soil horizons. Our study demonstrate that horizon specific calibration of ECH_2O EC-20 sensors improve the accuracy of soil water content monitoring of the sensors as compared to with the use of manufacturer generic calibration equation for the soils tested in this study.

The instantaneous profile field method was used to determine the unsaturated hydraulic conductivities of the two soil profiles. Water content and soil water suction were monitored at frequent intervals using ECH_2O EC-20 probes and tensiometers. The resulting data from the soil profiles provided the necessary information for calculation of hydraulic conductivity. The Bainsvlei form soil was characterized by several relations of hydraulic conductivity, varying with depth. The reason for this was attributed to heterogeneous of the profiles due to variation in clay content. The instantaneous profile method is attractive when applied to relatively homogeneous soil in which free vertical drainage can be expected to occur after “saturation”. However, it is worth recalling that when it is applied for duplex soils, it must be suitably verified and, in general, employed with caution and in a critical spirit. The findings from instantaneous profile field method demonstrated that the ECH_2O EC-20 capacitance probes, if properly calibrated, is a practical tool that can be used effectively to determine *in situ* hydraulic properties at different soil depths.

The van Genuchten (1980) model used together water retention data (laboratory method) provided reliable estimates of unsaturated hydraulic conductivity that were very close to field measurements obtained from the internal drainage experiment. Hence, from a practical perspective, the prediction of K-θ relationship from water retention data can be a viable alternative for determining the hydraulic properties of Bainsvlei form soil profile examined in this study.

7.2 Recommendations

For future studies on comparison of field and laboratory measured hydraulic properties of selected diagnostic soil horizons, the following aspects can be recommended;

It is suggested that for additional studies using the methods described (i) the instantaneous profile will require a measuring devices to accurately measure soil water suction within the soil profile (ii) extensive measurements of water retentivity relations in the laboratory and field are necessary for further comparison and evaluation

The instantaneous profile method is attractive when applied to relatively homogeneous soil profile where one-dimensional vertical flow can be expected to after saturation. However it is not applicable to stratified soil profiles with slowly permeable layer and there is a need to devise a procedure for soil profile with slowly permeable layers.

Field and laboratory studies should continually be carried out, particularly on South African soils in order to quantitatively to assess the field description of soil structure and improve on understanding bi-modal pore size distributions.

To explore the possibility of predicting drained upper limit from hydraulic conductivity, as it has pointed out by Romano & Santini, (2002), DUL should be more related to the hydraulic conductivity of a soil than to the soil water suction.

Appendices

Appendix A Soil water retention data for Bainsvlei form soil at Kenilworth. Values represent average volumetric water contents (mm^3 mm^{-3}) of three replicate measurements and standard deviations are given in parentheses.

Suction (kPa)	0-250 mm	250-420 mm	420-700 mm	700-1200 mm	1200-1450 mm	1450-1850 mm
0	0.374*(0.007)*	0.366*(0.001)*	0.373*(0.005)*	0.371*(0.002)*	0.373*(0.004)*	0.367*(0.003)*
0.38	0.364(*0.009)*	0.362*(0.002)*	0.364*(0.002)*	0.362*(0.002)*	0.363*(0.003)*	0.362*(0.002)*
1	0.354*(0.003)*	0.349*(0.002)*	0.351*(0.009)*	0.352*(0.003)*	0.353*(0.003)*	0.353*(0.004)*
2	0.329*(0.006)*	0.340*(0.003)*	0.332*(0.013)*	0.344*(0.001)*	0.327*(0.009)*	0.343*(0.002)*
3	0.317*(0.007)*	0.327*(0.006)*	0.310*(0.012)*	0.335*(0.005)*	0.316*(0.011)*	0.325*(0.013)*
5	0.272*(0.008)*	0.301*(0.018)*	0.276*(0.006)*	0.319*(0.018)*	0.270*(0.009)*	0.284*(0.010)*
10	0.217*(0.012)*	0.239*(0.014)*	0.236*(0.005)*	0.270*(0.014)*	0.215*(0.008)*	0.232*(0.011)*
50	0.186*(0.009)*	0.205*(0.004)*	0.196*(0.002)*	0.204*(0.003)*	0.185*(0.010)*	0.173*(0.008)*
100	0.159*(0.003)*	0.186*(0.001)*	0.170*(0.002)*	0.193*(0.008)*	0.157*(0.009)*	0.156*(0.007)*
300	0.122*(0.003)*	0.141*(0.006)*	0.135*(0.008)*	0.136*(0.009)*	0.120*(0.007)*	0.129*(0.010)*
500	0.106*(0.002)*	0.124*(0.008)*	0.116*(0.017)*	0.121*(0.010)*	0.104*(0.008)*	0.121*(0.008)*
1000	0.096*(0.003)*	0.106*(0.002)*	0.101*(0.009)*	0.114*(0.008)*	0.095*(0.007)*	0.117*(0.005)*
1500	0.091*(0.007)*	0.095*(0.008)*	0.093*(0.006)*	0.108*(0.007)*	0.091*(0.006)*	0.113*(0.008)*

Appendix B Soil water retention data for Tukulu form soil at Paradys. Values represent average volumetric water contents (mm^3 mm^{-3}) of three replicate measurements and standard deviations are given in parentheses.

Suction (kPa)	0-270 mm	270-500 mm	500-800 mm
0	0.370*(0.007)*	0.356*(0.009)*	0.352*(0.003)*
0.38	0.361*(0.006)*	0.348*(0.008)*	0.346*(0.003)*
1	0.350*(0.008)*	0.340*(0.009)*	0.342*(0.006)*
2	0.333*(0.010)*	0.319*(0.009)*	0.339*(0.007)*
3	0.316*(0.008)*	0.302*(0.009)*	0.336*(0.009)*
5	0.282*(0.011)*	0.276*(0.010)*	0.332*(0.007)*
10	0.246*(0.002)*	0.233*(0.009)*	0.326*(0.008)*
50	0.211*(0.002)*	0.192*(0.009)*	0.304*(0.009)*
100	0.194*(0.002)*	0.177*(0.009)*	0.295*(0.007)*
300	0.136*(0.007)*	0.162*(0.009)*	0.273*(0.008)*
500	0.128*(0.009)*	0.148*(0.008)*	0.268*(0.009)*
1000	0.115*(0.009)*	0.127*(0.013)*	0.238*(0.010)*
1500	0.103*(0.008)*	0.117*(0.012)*	0.229*(0.009)*

Appendix C Water content (mm^3 mm^{-3}) at selected times for Bainsvlei form soil.

Hours	0-250 mm	250-420 mm	420-700 mm	700-1200 mm	1450-1850 mm
0	0.291	0.314	0.300	0.330	0.318
0.5	0.290	0.303	0.290	0.327	0.318
1	0.289	0.301	0.288	0.326	0.316
2	0.288	0.298	0.286	0.325	0.314
4	0.285	0.294	0.283	0.323	0.311
8	0.279	0.290	0.280	0.319	0.306
24	0.263	0.282	0.277	0.310	0.296
48	0.249	0.275	0.275	0.303	0.287
72	0.241	0.271	0.265	0.300	0.281
120	0.232	0.265	0.258	0.290	0.274
252	0.223	0.256	0.246	0.284	0.266
480	0.216	0.245	0.233	0.280	0.259

Appendix D Hydraulic head (mm) at selected times for Bainsvlei form soil.

Hours	0-250 mm	250-420 mm	420-700 mm	700-1200 mm	1450-1850 mm
0	490	672	883	1267	1931
0.5	515	739	887	1269	1958
1	532	764	892	1272	1968
2	558	799	900	1278	1983
4	599	845	917	1291	2003
8	664	909	959	1317	2031
24	840	1057	1086	1452	2097
48	1012	1191	1210	1558	2160
72	1136	1289	1323	1641	2206
120	1314	1434	1479	1761	2277
252	1571	1675	1729	1952	2404
480	1847	1992	2128	2265	2578

Appendix E Calculation of water flux for Bainsvlei form soil.

Time (hours)	z (mm)	$\Delta\theta/\Delta t$ (mmh^{-1})	$dz(\Delta\theta/\Delta t)$ (mmh^{-1})	$q=\sum dz(\Delta\theta/\Delta t)$ (mmh^{-1})
0.5	250	0.00278	0.69600	0.69600
	420	0.02195	9.21942	9.91542
	700	0.02018	14.12740	24.04282
	1200	0.00561	6.72720	30.77002
	1850	0.00055	1.01935	31.78937
1	250	0.00157	0.39300	0.39300
	420	0.00481	2.01852	2.41152
	700	0.00399	2.79160	5.20312
	1200	0.00279	3.34320	8.54632
	1850	0.00391	7.23350	15.77982
2	250	0.00138	0.34463	0.34463
	420	0.00297	1.24677	1.59140
	700	0.00242	1.69540	3.28680
	1200	0.00094	1.12380	4.41060
	1850	0.00194	3.58437	7.99497
4	250	0.00163	0.40706	0.40706
	420	0.00175	0.73489	1.14196
	700	0.00144	1.00625	2.14821
	1200	0.00122	1.46490	3.61311
	1850	0.00174	3.22733	6.84043
8	250	0.00144	0.35959	0.35959
	420	0.00105	0.43979	0.79939
	700	0.00064	0.44573	1.24511
	1200	0.00090	1.08465	2.32976
	1850	0.00116	2.14415	4.47391
24	250	0.00101	0.25302	0.25302
	420	0.00052	0.21830	0.47131
	700	0.00018	0.12443	0.59574
	1200	0.00054	0.65070	1.24644
	1850	0.00064	1.18458	2.43101
48	250	0.00057	0.14233	0.14233
	420	0.00027	0.11338	0.25571
	700	0.00010	0.06857	0.32428
	1200	0.00030	0.35787	0.68216
	1850	0.00035	0.65505	1.33721

72	250	0.00033	0.08310	0.08310
	420	0.00018	0.07501	0.15810
	700	0.00042	0.29298	0.45108
	1200	0.00015	0.18423	0.63531
	1850	0.00024	0.43837	1.07368
120	250	0.00009	0.02274	0.02274
	420	0.00006	0.02554	0.04827
	700	0.00007	0.05155	0.09982
	1200	0.00010	0.12121	0.22103
	1850	0.00008	0.14768	0.36871
252	250	0.00007	0.01679	0.01679
	420	0.00007	0.03003	0.04682
	700	0.00009	0.06256	0.10938
	1200	0.00005	0.05569	0.16506
	1850	0.00006	0.10608	0.27115
480	250	0.00003	0.00775	0.00775
	420	0.00005	0.01919	0.02694
	700	0.00006	0.04075	0.06770
	1200	0.00002	0.02072	0.08842
	1850	0.00003	0.05703	0.14545

Appendix F Calculation of hydraulic conductivity for Bainsvlei soil.

z (mm)	t (h)	ΔH/Δz (mm mm^{-1})	q=∑dz(Δθ/Δt)	K(θ) (mmh^{-1})	θ(mm^3mm^{-3})
420	0.5	1.0636	9.9154	9.322342	0.303
	1	1.1059	2.4115	2.180689	0.301
	2	1.1469	1.5914	1.387602	0.298
	4	1.1714	1.1420	0.974838	0.294
	8	1.1641	0.7994	0.686705	0.290
	24	1.0323	0.4713	0.45657	0.282
	48	0.8534	0.2557	0.299655	0.275
	72	0.7247	0.1581	0.218155	0.271
	120	0.5682	0.0483	0.084957	0.265
	252	0.4968	0.0468	0.094252	0.256
	480	0.6902	0.0269	0.039036	0.245
700	0.5	0.6601	24.0428	36.42435	0.290
	1	0.5676	5.2031	9.167165	0.288
	2	0.4500	3.2868	7.304422	0.286
	4	0.3161	2.1482	6.795258	0.283
	8	0.2222	1.2451	5.603393	0.280
	24	0.1288	0.5957	4.624718	0.277
	48	0.0843	0.3243	3.845741	0.275
	72	0.1540	0.4511	2.929153	0.265
	120	0.2024	0.0998	0.493081	0.258
	252	0.2410	0.1094	0.453767	0.246
	480	0.6021	0.0677	0.112434	0.233
1200	0.5	0.9797	30.7700	31.40817	0.327
	1	0.9760	8.5463	8.75636	0.326
	2	0.9698	4.4106	4.547919	0.325
	4	0.9597	3.6131	3.764891	0.323
	8	0.9181	2.3298	2.537501	0.319
	24	0.9374	1.2464	1.329642	0.310
	48	0.8918	0.6822	0.764904	0.303
	72	0.8141	0.6353	0.78035	0.300
	120	0.7214	0.2210	0.306409	0.290
	252	0.5718	0.1651	0.288658	0.284
	480	0.3526	0.0884	0.250735	0.280

1850	0.5	0.9832	31.7894	32.33143	0.318
	1	0.9940	15.7798	15.8756	0.316
	2	1.0068	7.9950	7.940909	0.314
	4	1.0174	6.8404	6.723223	0.311
	8	1.0198	4.4739	4.387038	0.306
	24	0.9224	2.4310	2.635515	0.296
	48	0.8593	1.3372	1.55623	0.287
	72	0.8075	1.0737	1.329655	0.281
	120	0.7378	0.3687	0.499745	0.274
	252	0.6456	0.2711	0.419991	0.266
	480	0.4467	0.1455	0.325652	0.259

Appendix G Water content ($mm^3\ mm^{-3}$) at selected times for Tukulu form soil.

Hours	0-270 mm	270-500 mm	500-800 mm
0	0.300017	0.312125	0.342591
0.5	0.296063	0.309559	0.342589
1	0.294064	0.30788	0.342555
2	0.291817	0.305886	0.342511
4	0.289296	0.303522	0.342453
8	0.286476	0.300728	0.342376
12	0.284675	0.298867	0.342319
24	0.281319	0.295249	0.342199
48	0.277589	0.291013	0.342039
72	0.275222	0.288216	0.341923
264	0.26664	0.277448	0.341393
480	0.262149	0.271466	0.341038
624	0.260064	0.268619	0.340854

Appendix H Hydraulic head (mm) at selected times for Tukulu form soil.

Hours	0-270 mm	270-500 mm	500-800 mm
0	471.136	561.647	751.01
0.5	654.579	619.463	751.682
1	676.209	630.003	752.205
2	701.286	642.716	753.129
4	730.606	658.156	754.756
8	765.228	677.064	757.593
12	788.492	690.124	760.127
24	834.681	716.772	766.737
48	891.174	750.459	777.537
72	930.38	774.434	786.472
264	1101.66	883.622	830.592
480	1216.38	959.95	857.145
624	1277.69	1001.62	869.136

Appendix I Calculation of soil water flux for Tukulu form soil.

Time (hours)	z (mm)	$\Delta\theta/\Delta t$ (mmh^{-1})	$dz(\Delta\theta/\Delta t)$ (mmh^{-1})	$q=\sum dz(\Delta\theta/\Delta t)$ (mmh^{-1})
0.5	0-270	0.00791	2.13489	2.13489
	270-500	0.00513	1.18036	3.31525
1	0-270	0.00400	1.07946	1.07946
	270-500	0.00336	0.77234	1.85180
2	0-270	0.00225	0.60669	0.60669
	270-500	0.00199	0.45862	1.06531
4	0-270	0.00126	0.34033	0.34033
	270-500	0.00118	0.27186	0.61219
8	0-270	0.00070	0.19035	0.19035
	270-500	0.00070	0.16066	0.35101
12	0-270	0.00045	0.12157	0.12157
	270-500	0.00047	0.10701	0.22858
24	0-270	0.00056	0.15102	0.15102
	270-500	0.00060	0.13869	0.28971
48	0-270	0.00031	0.08393	0.08393
	270-500	0.00035	0.08119	0.16512
72	0-270	0.00010	0.02663	0.02663
	270-500	0.00012	0.02680	0.05343
120	0-270	0.00007	0.01793	0.01793
	270-500	0.00008	0.01864	0.03657
264	0-270	0.00004	0.01011	0.01011
	270-500	0.00005	0.01099	0.02110
480	0-270	0.00002	0.00561	0.00561
	270-500	0.00003	0.00637	0.01198
624	0-270	0.00001	0.00391	0.00391
	270-500	0.00002	0.00455	0.00846

Appendix M Calculation of hydraulic conductivity based on unit gradient assumption.

Depth	Time (hours)	θ (mm^3 mm^{-3})	q-flux (mmh^{-1})
0-270	0	0.300017	10
	0.5	0.296063	2.13489
	1	0.294064	1.07946
	2	0.291817	0.60669
	4	0.289296	0.34033
	8	0.286476	0.19035
	12	0.284675	0.12157
	24	0.281319	0.15102
	48	0.277589	0.08393
	72	0.275222	0.02663
	120	0.272034	0.01793
	264	0.26664	0.01011
	480	0.262149	0.00561
	624	0.260064	0.00391
270-500	0	0.318125	6.7
	0.5	0.309559	3.31525
	1	0.30788	1.85180
	2	0.305886	1.06531
	4	0.303522	0.61219
	8	0.300728	0.35678
	12	0.298867	0.22858
	24	0.295249	0.28971
	48	0.291013	0.16512
	72	0.288216	0.05343
	120	0.284326	0.03657
	264	0.277448	0.02110
	480	0.271466	0.01198
	624	0.268619	0.00846