EM38 FOR MEASURING AND MAPPING SOIL MOISTURE
IN A CRACKING CLAY SOIL

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A thesis submitted in the fulfillment of the requirements for the degree of

Doctor of Philosophy

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New South Wales, Australia

June 2008
This thesis is dedicated

To

My Parents
Declaration

Except as stated herein this thesis contains no material which has been accepted for the award of any higher degree or diploma by the University or any other institution.

To the best of my knowledge and belief this thesis contains no material previously published or written by another person except where due acknowledgment is made in the text.

I certify that any help received in preparing this thesis and all sources used, have been acknowledged in this thesis.

Md Bilal Hossain
Undertaking the PhD programme at University of New England, Armidale in the past three years was a truly exceptional experience in my life. However, I could not achieve this work without the blessings of ‘Almighty’ and support of different organizations and my near and dear ones.

First of all I would like to express my sincere indebtedness to the ‘Almighty’, for his kindness to allow me to successfully complete my PhD course.

The course was done while I have been enjoying study leave from my home university, Bangladesh Open University. Thanks are due to that organization.

Throughout the course of this project I was privileged to have support from a number of people, to whom I am deeply grateful. First and foremost, I wish to acknowledge the expert guidance I received from my principal supervisor, Dr. Paul Frazier. I am very grateful to Paul for the time and effort he has invested in helping me to achieve this goal and for being an inspiration at every step of the way. Our time spent together was not only a chance to share technical information but to learn his philosophy and passion for life. Additionally, I should also appreciate his kind cares about my family and personal life.

I equally thank my co-supervisor, Dr. David Lamb, for his support and encouragement of excellence in all phases of my research particularly his expertise on the EM38 instrument and modelling. I also thank my co-supervisor Dr. Peter Lockwood. His profound understanding and insight about soils and soil moisture encouraged me to progress this research. Thanks are also due to Dr. John Louis for his assistance in developing MAPLE code.

I also thank George Henderson, Cate MacGregor, Leanne Lisle, Allan Mitchell, Derek Schneider and David Edmonds for their continued technical support both in the field and in the laboratory during the PhD project. Thanks to Tania Marshall and Geraldine Cronin for their timely help whenever required.
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ABSTRACT

Soil moisture content in the crop root zone varies both spatially and temporally. Moisture in this zone is critical to plant development and health, so understanding the variability and dynamics of moisture distribution in this zone is crucial for optimal crop management. The EM38 electromagnetic induction sensor, a tool for measuring apparent electrical conductivity in soils, can be used to infer a range of soil properties, including soil moisture. The ability to configure an EM38 for on-the-go sensing and mapping of soil apparent electrical conductivity ($\text{EC}_a$), if related to key soil properties such as moisture content, means high resolution soil maps can be produced that may significantly aid in the management of agricultural fields.

This thesis describes an investigation of the EM38 to quantify soil moisture variability in a paddock of homogenous Black Vertosol soil (cracking clay). The research project comprised three main components: (i) an assessment of two ‘conventional’ soil moisture probes; (ii) an investigation of the relationship between soil moisture, and its vertical distribution, and EM38-derived $\text{EC}_a$ readings; and (iii) a brief investigation of the potential of a mobile on-the-go EM38 survey to assess temporal and spatial patterns of paddock-scale soil moisture.

Data from a soil moisture neutron probe (SMNP) and a Diviner 2000 capacitance probe were compared against volumetric soil moisture content ($\theta_v$) determined via soil core samples and both wet and dry pit-based samples. Pit-based samples proved to be more reliable for soil moisture determination when compared to core-sampling. The Diviner 2000 was shown to provide useful soil moisture information to a depth of only above 0.6 m ($R^2 = 0.81 - 0.92$), but the SMNP was successful at all depths down to 1.2 m ($R^2 = 0.94 - 0.97$).

A model linking $\theta_v$ to EM38 derived $\text{EC}_a$ was developed and subsequently verified. Comparison of the model predictions with multi-height EM38 measurements showed that the EM38 depth response function was not perturbed by the depth profile of soil moisture content and the subsequent model showed that $\text{EC}_a$ for both dipole orientations explained 99% of the variation observed in $\theta_v$ averaged to a depth of 1.2 m. The calibration results show that the on-ground EM38 measurement could measure $\theta_v$ with an accuracy of $\pm 0.007 \text{ m}^3/\text{m}^3$ in either mode. However, the horizontal mode of EM38 was found to be
better for predicting $\theta_v$. The forward-propagation models of Rhoades & Corwin (1981) and Slavich (1990) were refined and tested for their ability to directly predict the vertical profile of soil moisture content. The Slavich model, incorporating both vertical and horizontal dipole configurations, was found to produce the best predictions with an error of approximately 10%. A complex inversion process involving Thikonov regularization was also tested and was found to consistently under-predict $\theta_v$ by approximately 50 – 75%.

The capture and comparison of multiple on-the-go EM38 surveys showed that soil moisture was the primary driver of temporal variation in the EM38 derived $EC_a$ at this site. Maps of derived $\theta_v$ values were correlated to site topography and the inclusion of multi-temporal EM38 survey data gave the most accurate representation of topographic effects (drainage).
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CHAPTER – ONE

INTRODUCTION
1.1 INTRODUCTION

Understanding soil variability is critical for humans. We rely on soil to produce our food, support our shelters and purify our drinking water. However, understanding soil variability is difficult as most of the soil remains hidden from our view. Until recently our knowledge of soil variability came from extrapolating site-based (profile or pit) soil descriptions across a landscape using ancillary data such as topographic maps, vegetation changes, geological maps and surface soil colour. These traditional techniques use few samples sites (Few), detailed site-based descriptions (Detailed) and extensive extrapolation across a landscape (Extrapolation). Recent developments in sensor and navigation technology have permitted a shift from Few-Detailed-Extrapolation soil mapping methods to Many-Single-Interpolation methods. These sensors capture thousands or hundreds of thousands of individually located points (Many); record a single or few soil characteristics (Single); and are interpolated between points rather than extrapolated across a landscape (Interpolation).

Soil mapping has a history as old as humanity, in the precivilised era, areas of particular importance were identified/mapped by special marks such as votive stones (kudurus) or relayed in verbal stories or ancient art works but were not mapped explicitly. Detailed soil mapping in the ancient era was documented by Dilke (1961; 1971). The early Egyptians were perhaps the first civilization to produce land resource maps. They developed units of measurement (cubit or forearm length) and surveying devices (such as mekhet, a plum-line sighter) to measure land areas and produce land maps. The Romans produced soil maps of greater detail and accuracy than previous civilisations with devices that allowed them to measure distance and land slopes with a high level of accuracy (described in Dilke 1985; Dorling & Fairbairn 1997).

Modern soil mapping techniques have developed rapidly in the past 100 years, Birch (1952) considers that modern soil mapping began in the 1950’s with the use of precise survey equipment (chain survey, prismatic compass, plane table survey with spirit-level, site rule, measuring tape) and detailed methods of soil description and analysis. Air photo interpretation made it possible to extend soil type boundaries beyond the sample area, based on topographic information and other photographic features (e.g., texture, colour, pattern, association etc) (Jensen 2007). The digital revolution of soil mapping started in 1960’s where a number of satellites
(e.g., LANDSAT, SPOT, ERS, JERS, RADARSAT) and computer-based mapping software, primarily GIS, were developed and used for soil mapping. These technologies allowed the rapid development of soil mapping capabilities using the Few-Detailed-Extrapolation mapping technique.

Proximal sensors (e.g., ultrasonic displacement sensor, NIR reflectance sensor, ground penetrating radar, Soil Director, VERIS, EM31, EM38) coupled with differential global positioning system (DGPS) permitted the shift to much higher resolution soil maps as they provide thousands of samples at a sub-paddock scale. In addition to image-based information, the proximal sensors provide complimentary data sources for detailed soil property mapping with sensors physically closer (1 m or less) to the object (Barnes et al. 2003). In particular electromagnetic induction (EMI) sensors such as the commercially-available EM31 and EM38, have been found to be useful inferring (indirectly), then mapping a wide range of soil features including soil moisture, clay content and salinity (Halvorson & Rhoades 1974; Cameron et al. 1981; Williams & Baker 1982; McNeill 1986; Wollenhaupt et al. 1986; Williams & Hoey 1987; Kachanoski et al. 1988; McBride et al. 1990; Doolittle et al. 1994; Jaynes et al. 1994; Jaynes 1996; Davis et al. 1997; Brevik & Fenton 2002; Hezarjaribi & Sourell 2007).

The role of moisture in the top one or two metres of the soil is widely recognized as a key parameter in agriculture. The moisture content of this soil layer is a primary control on the success of agriculture and regulates partitioning of precipitation into runoff and ground water storage. Thus, soil moisture is critical in determining crop productivity through its impact on germination and growth. However, soil moisture varies both spatially (horizontally and vertically) and temporally due to the heterogeneity of soil characteristics, topography and microclimate (Engman 1991; Wood et al. 1993). Development of the means to monitor soil moisture in agricultural fields in a way that accurately encapsulates the two-dimensional (horizontal) spatial variability as well as vertical (depth-dependent) variation that is both time and cost-effective is critical for effective soil moisture management.

There are a number of techniques available to determine soil moisture. The in situ status of soil moisture can be determined on a gravimetric or volumetric basis using traditional soil sampling and analysis (Gardner 1986). Alternatively, soil moisture can
be determined on a volume basis with the quantitative measurement of some physical or chemical properties of a soil such as hydrogen content, dielectric constant, electrical conductivity or magnetic susceptibility that ultimately relates to soil moisture content (e.g., Holmes 1956; Bell et al. 1987; Sheets & Hendrickx 1995). Due to the destructive nature of gravimetric determination, other techniques based on volumetric principles are preferred for repeated in situ measurement of soil moisture. For example, soil moisture neutron probes (SMNP) and capacitance probes are probably the most widely used instruments for irrigation scheduling around the world. While these in situ techniques can provide accurate information on soil moisture, the spatial range of the sensors is limited to tens of centimetres and extension of the information to a paddock-scale can be problematic if the sensors are poorly located.

To help overcome the limitations of in situ techniques, EMI sensors have been coupled with DGPS and a mobile platform to provide whole-of-paddock coverage. Although these sensors were originally developed to detect soil salinity, they have been found to provide useful information on other soil properties including soil moisture. If the information on soil conductivity produced by these sensors can be securely tied to variations in soil moisture they may provide a useful means to detect spatial and temporal variability in paddock-scale soil moisture.

Cracking clays also known as Vertosols (Isbell 1996) or Vertisols (Soil Survey Staff 1975) are important soils worldwide because of their relatively high fertility and potential for agricultural production. These soils are often irrigated to maximise production potential. However, they pose particular problems for water management as their shrink-swell behaviour restricts the value of in situ probe with little previous work being conducted on cracking clay soils.

While EMI sensors have been used to identify variations in soil moisture, the theory underlying soil moisture determination with EMI sensor has yet to be developed fully.

1.2 THESIS AIM AND OBJECTIVES

The principal aim of this thesis was to explore the potential of EM38 for soil moisture determination at a paddock-scale in cracking clay soils. This required comparing EM38 to other more established techniques for soil moisture measurement. To date there have been no published systematic comparisons of EM38, SMNP, capacitance
and gravimetric measurements of soil moisture in cracking clays. To achieve the principal aim the key thesis objectives were to:

1. Compare a single core and a pit soil sampling technique for gravimetric soil moisture determination for the purpose of sensor calibration in a cracking clay soil.

2. Develop and compare calibration models (sensor count to volumetric soil moisture) for the SMNP and Diviner 2000 capacitance *in situ* probes in a cracking clay soil.

3. Determine whether EM38-derived multi-height apparent electrical conductivity ($EC_a$) values are in fact an integration of the accepted depth-response function of the EM38 and volumetric moisture content for deep Vertosols.

4. Calibrate the EM38 to estimate volumetric soil moisture.

5. Develop models to determine soil moisture content at depth using both forward propagation and inverse matrix procedures.

6. Characterise the spatial and temporal variability of volumetric moisture content predicted from multiple EM38 surveys.

1.3 THESIS ORGANISATION

This thesis is divided into seven chapters, with the relevant literature being reviewed at the start of each experimental chapter:

**Chapter One:** provides a brief overview on soil mapping and a description of the primary aim and objectives of the thesis.

**Chapter Two:** describes the study area including location, climate, soil and geology, vegetation and land use information.

**Chapter Three:** describes and identifies the best method of soil sampling for sensor calibration. The chapter also gives a comparative study of the performance the SMNP and Diviner 2000 *in situ* probes for determining soil moisture in this particular
cracking clay soil. The view to compare the performance of the SMNP and the Diviner 2000 is to identify the most accurate sensor for determining soil moisture from point measurements for this soil. Subsequently this point measurements by the comparatively accurate sensor would be used for the evaluation of the performance of models developed from EM38 measurements. Following three chapters are related to soil moisture measurements with EM38 sensor.

**Chapter Four:** assesses the theoretical basis for the measurement of soil moisture with an EM38 sensor.

**Chapter Five:** investigates the ability of the EM38 sensor to estimate vertical variations in the soil moisture profile.

**Chapter Six:** establishes the relationship between $EC_a$ and soil moisture content in the cracking clay study soil and illustrates the potential of the EM38 sensor to characterize the spatial and temporal variability of soil moisture using multiple surveys.

**Chapter Seven:** provides the key conclusions from the study and discusses recommendations for future study.
CHAPTER – TWO

STUDY AREA
2.1 LOCATION

The experiment was conducted at Clark’s Farm, one of the experimental properties of the University of New England, Armidale, Australia. Armidale is located in northern New South Wales, Australia (30° 30' S 151° 40' E) (Figure 2.1) with altitude ranging from 970 metres at the floor of the valley to approximately 1100 metres (ASL). Clark’s Farm is located 6 km north-west of the Armidale CBD and immediately adjacent to UNE (30° 31.7' S 151° 37' E).

Figure 2.1 The location of the study area (Clark’s Farm) (Google Imagery 2008).
2.2 SOILS AND GEOLOGY

According to the Australian Soil Classification (Isbell 1996) the soils of Clark’s Farm fall under the orders Vertosol, Dermosol, Tenosol and Chromosol. An extensive soil survey was conducted on Clark’s Farm by the Department of Agronomy and Soil Science of UNE (Lockwood 2008, unpublished report) and found that the farm contains a range of soil types (Figure 2.2) with two principal parent materials namely the Armidale Beds (conglomerate, greybilly, ferruginous, siliceous sandstone and claystone) and Tertiary Basalt. Basalt is found on the hills in the north and east of the farm, and basaltic clay colluvium has covered some of the Armidale Beds in the centre and south. A 2 ha block of the farm was selected for the study area. The selected study area was mapped as containing Black Vertosols (Figure 2.2) (Isbell 1996). The important characteristics of Black Vertosols are the dominant colour of the soil is black (Figure 2.3), high clay content (>30%) throughout the solum (Table 2.2), shrink-swell character that allows the soil to crack when it dries. Cracks greater than 5 mm are typical (Figure 2.3). Black Vertosols are the typical soil type found on basalt in lower slope positions in this environment.

Two soil pits were dug, one each at the top and bottom slopes of the site, to assess if soil morphology varied with slope position. The soil profiles were described using standard Australian descriptors (McDonald et al. 1990) and were found to be very similar (Table 2.1 and Figure 2.4). Both soil profiles were classified as Self-mulching Black Vertosols (Isbell 1996). Two soil cores to a depth of approximately 1.5 m were extracted from the site at top and bottom slope locations. These cores were sectioned into 0.1 m lengths and prepared for laboratory analysis of soil physical and chemical properties namely texture, volumetric moisture content ($\theta_v$), pH and electrical conductivity (EC). Both profiles showed similar characteristics with high clay content (60 – 70%), neutral to alkaline pH values and low EC values (Table 2.2). Leaching down the soluble salts through the hill slopes and tends to accumulate in lower slope positions caused the higher EC contents in the bottom site than the top site.
Figure 2.2 A sketch of Clark’s Farm showing the boundary of soil types and the approximate location of the study area (Lockwood 2008, unpublished report)
Figure 2.3 Photographs revealing the characteristics of the Black Vertosol soils of the study area (A) dominant black colour of the soil profile (B) surface cracking during dry periods (C) vertical cracks through the profile (D) width of vertical cracks

Table 2.1 Morphological description of soil profile following the Standard Australian Descriptors (McDonald *et al.* 1990).

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<td>Heavy clay</td>
<td>Heavy clay</td>
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<td>Roots in cracks, slickensides &amp; CaCO&lt;sub&gt;3&lt;/sub&gt; present</td>
<td>Roots in cracks, slickensides &amp; CaCO&lt;sub&gt;3&lt;/sub&gt; present</td>
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Table 2.2 Physical and chemical properties of two representative soil core profiles taken from the top and bottom slope locations of the study area

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<th>Fine sand (%)</th>
<th>Silt (%)</th>
<th>Clay (%)</th>
<th>pH (1:5)</th>
<th>EC_{1.5} (mS/m)</th>
<th>$\theta_v$ (m$^3$/m$^3$)</th>
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Figure 2.4 Soil profile photographs and descriptions from (A) the top slope pit and (B) the bottom slope pit of the study area.
2.3 CLIMATE

The long-term climate statistics of the Clark’s Farm site were considered to be similar to those of the city of Armidale. Hence, figures from the Bureau of Meteorology for Armidale were used to describe the Clark’s Farm climate.

2.3.1 Rainfall

The rainfall pattern of Armidale is summer dominant. In summer, thunderstorms often produce heavy rainfall events in the afternoon and early evenings. The average annual rainfall of the region is 791.5 mm over 140 years (1857 – 1997). January has the highest average monthly rainfall fall with 104.5 mm (Figure 2.5) and an average of 10 rainy days (BoM 2008a). The lowest average monthly rainfall is in May (44.4 mm). The annual total rainfall for the study period varied notably from the long-term average rainfall. In the 2006 it was below the average (620.8 mm) whereas it was 907.8 mm in 2007. The highest and lowest monthly rainfall during the study period was 173.8 mm in February 2008 and 2.4 mm in May 2006 respectively (BoM 2008b).

![Figure 2.5 Monthly rainfall (mm) for the long-term average of 140 years (1857 – 1997) and the study period years of 2006, 2007 and 2008 (BoM 2008a, 2008b)]
2.3.2 Temperature

Armidale is in the temperate climatic zone of Australia. The average minimum and maximum temperatures range from 0.3 °C to 16.4 °C respectively in cooler months (May – July) and 12.2 °C to 27.1 °C respectively in warmer months (December – February) (BoM 2008a). In winter, temperature often falls below -5 °C over-night, however, occasionally it drops as low as -10 °C. The coldest month is July with average maximum and minimum daily temperature of 12.2 °C and 0.3 °C respectively. In summer it is warm to hot with an average daily maximum temperature of 26.6 °C. The temperature rarely exceeds 35 °C (less than 1 day a year), however, it exceeds 30 °C on an average of 20.9 days per year (BoM 2008a). The hottest month is January with average maximum and minimum daily temperatures of 27.1 °C and 13.4 °C respectively. The monthly average maximum and minimum temperature for the long term and the study period is illustrated in Figure 2.6. Both maximum and minimum average temperatures for the study period 2007 and part of 2008 were a little below the average. In 2006, the temperature pattern was similar to the long-term average.

![Figure 2.6](image-url)

**Figure 2.6** Monthly temperatures (°C) for the long-term average (1857 – 1997) and the study period years of 2006, 2007 and 2008 (BoM 2008a, 2008b)
2.4 VEGETATION

The current vegetation of Clark’s Farm is naturalized pasture. The pasture is dominated by long established summer and winter pastures namely *Microlaena spp.*, *Lolium rigidum*, *Fescue spp.*, *Ostrodanthonia spp.*, *Chloris truncata*, *Phalaris minor*, *Bromus spp.*, *Bothriochloa macra*, *Paspalum dilatatum* and *Poa pratensis*. The study area contains some weeds (*Saffron thistle*, *Trifolium repens* and *Vulpia spp.*). The pasture is usually at its optimum in autumn (March – May) due to favourable growing condition (Kemp & Dowling 1991).

2.5 LAND USE

Clark’s Farm was used as commercial farm until it was purchased by UNE in 1948. Among other enterprises cropping and grazing were practiced in the more fertile parts of the farm since 1880s. The cropping pattern of the area was generally an oats-wheat-clover-pasture rotation. Orchards were developed and commercialised between 1935 and 1950. However, successive hailstorms and subsequent poor returns led to most of the trees being pushed out in 1980 (Lockwood 2008, unpublished report).
CHAPTER – THREE

COMPARISON OF TWO FIELD SENSORS FOR SOIL MOISTURE MEASUREMENT IN A BLACK VERTOSOL SOIL
3.1 INTRODUCTION

The soil moisture neutron probe (SMNP) as an in situ measurement instrument for soil moisture determination was first introduced in 1950s (Gardner & Kirkham 1952; Van Bavel et al. 1956). The SMNP has been shown to be a robust instrument suited to local soil moisture determination (Moutonnet et al. 1988; Heng et al. 2001). However, limitations that include safety regulations costly licensing and training of users make it difficult to use in particular situations, particularly in remote and unattended areas (Evett 2000b).

Capacitance sensors were introduced as an alternative to the SMNP for rapid assessment of soil moisture content. They became popular in the 1990s, following advances in microelectronics and have been used extensively in soil water monitoring for irrigation management (Buss 1993) and soil water resources research (Fares & Alva 2000; Paltineanu & Starr 2000a, 2000b).

In this chapter a comparison of the SMNP and Diviner 2000 capacitance probe for soil moisture determination was undertaken in a Black Vertosol soil. In addition two field soil sampling techniques for soil moisture measurement and calibration of the probes was carried out.

3.2 METHODS OF SOIL MOISTURE MEASUREMENT

Methods of soil moisture determination are often classified into direct and indirect methods (Schmugge et al. 1980; Topp & Ferré 2002; Muñoz-Carpena 2004). Direct methods involve taking the weight of a soil sample before and after oven drying. The removal of moisture from a soil sample in this process is commonly referred to as the thermo-gravimetric or gravimetric method, expressed on a weight basis as gravimetric moisture content (kg/kg). Measurement of soil bulk density then enables the gravimetric moisture content to be converted to volumetric moisture content which is expressed as a volume of water in a volume of undisturbed soil (m$^3$/m$^3$). Indirect measurement methods are based on the estimation of some physical or chemical properties of a soil such as dielectric constant (relative permittivity), electrical conductivity, heat capacity, hydrogen content or magnetic susceptibility that ultimately relates to soil moisture content (Topp & Ferré 2002). Indirect methods
either measure soil moisture on a volumetric basis or measure soil water potential. In indirect methods, absolute moisture content is estimated by a calibrated relationship with some other measurable variables (Muñoz-Carpena 2004).

Gravimetric or volumetric soil moisture measurements can be presented as either a dimensionless ratio or percentage. It is, therefore, necessary to show symbolically whether it was measured as the ratio of two masses or of two volumes. Gravimetric moisture content and volumetric moisture content are related through the soil bulk density (Equation 3.1).

\[
\theta_v \times \rho_w = \theta_g \times \rho_b
\]  

(3.1)

where, \(\theta_v\) is volumetric moisture content, \(\theta_g\) gravimetric moisture content, \(\rho_w\) density of water (the numerical value of \(\rho_w\) is 1.0 Mg/m\(^3\)) and \(\rho_b\) bulk density.

### 3.2.1 Gravimetric and Volumetric Moisture Determination

The gravimetric technique involves the collection of soil samples from the field, weighing the sample, removal of soil moisture and determination of the mass of moisture content in relation to the mass of dry soil. The sample is weighed, dried in an oven with a temperature between 100 °C and 110 °C until the weight remains constant (Reynolds 1970a, 1970b; Gardner 1986; Topp & Ferré 2002). Moisture content can be determined by using the following equation:

\[
\theta_g = \frac{\text{mass of water}}{\text{mass of oven dried soil}} = \frac{M_w}{M_s}
\]  

(3.2)

Although this technique provides the advantage of accurate measurement (±0.01 kg/kg), low cost and easy operation (Muñoz-Carpena 2004), it is slow (minimum 2 days per measurement) and impossible to repeat in the same location since samples are extracted and subsequently destroyed (Zazueta & Xin 1994).

Once the gravimetric moisture content (\(\theta_g\)) is determined the volumetric moisture content (\(\theta_v\)) is then calculated from Equation 3.1. The bulk density of the soil for Equation 3.1 is calculated from the ratio of the mass and volume of the particular soil sample and so requires undisturbed soil cores to be taken.
3.2.2 Neutron Thermalization Soil Moisture Determination

The neutron thermalization technique, an in situ indirect method of moisture determination has been extensively used both in scientific research (e.g. Burrows & Kirkham 1958; Greacen et al. 1981; McKenzie et al. 1990; Kamgar et al. 1993; Corbeels et al. 1999; Evett et al. 2002b; Heng et al. 2002; Yao et al. 2004) and industrial applications (Cull 1979; Johnson & Borough 1992; Hanson & Dickey 1993; Kjelgren et al. 2000).

The SMNP, often called a ‘Neutron Probe’, consists of a probe, a pulse counter or ratescaler, a cable that connects the probe and ratescaler and a transport shield with display and keyboard (Figure 3.1) (Bell 1987). The probe is a long and narrow cylinder, containing a fast neutron source and a slow neutron detector. The fast neutron source is a mixture of radioactive americium (²⁴¹Am) and non-radioactive beryllium (Be). The radioactive americium emits an alpha particle and the Be nucleus absorbs the alpha particle and finally emits a fast neutron \([^{9}\text{Be}(\alpha,n)^{12}\text{C}]\). The fast neutron source is welded into double encapsulations of stainless steel and the detector is a tube filled with helium \((^3\text{He})\) or boron trifluoride \((\text{BF}_3)\) gas. The source is positioned either directly beneath the detector or is centred on the detector (Figure 3.1). The cable includes several cable stoppers or metal tags that mark soil measurement depths. The shield consists of a block of high density polyethylene.

The principles of the SMNP are that the high-energy neutrons (mean energy of 5 MeV) are expelled at a rate of \(~ 10^{27} \text{ s}^{-1}\) from the radioactive source of the probe into the surrounding soil, where collisions with nuclei of atoms in the soil, predominantly those of hydrogen in soil water take place. The collisions cause the neutrons to scatter, slow and lose energy. The neutrons then slow to ‘thermal’ energy levels and produce a ‘cloud’ of thermal neutrons that are then absorbed by other nuclear reactions. The density of this cloud is directly proportional to the hydrogen content and hence soil moisture content within the soil matrix around the source. From the cloud a small fraction of thermalized neutrons \((~0.025\text{eV})\) are scattered back to the detector of which an even smaller fraction (rate is \(~ 10^3 \text{ s}^{-1}\)) of thermal neutrons are measured by the detector. The back scattered thermal neutrons collide with the boron nuclei of the detector tube \((\text{BF}_3)\) and produce an alpha particle that in turn creates an electrical
pulse. The electrical pulse from the detector is measured and their mean count rate is displayed.

Figure 3.1 Schematic diagram of a neutron probe (Bell 1987; and Evett 2000a)

Although hydrogen in the soil matrix, acts as the principal thermalizer of neutrons, some other elements in the soil can scatter fast neutrons, or cause collisions and capture or absorb neutrons which in turn creates slow ‘thermal’ neutrons. For example, aluminium, silicon and oxygen are able to scatter neutrons with a little loss of energy. Moreover, C, N and O can also act as neutron thermalizers but they are weaker than hydrogen. To thermalize a neutron, approximately 19 collisions with H atoms are required where about 120, 140 and 150 collisions are required with C, N and O atoms respectively (Hignett & Evett 2002). Organic matter and some clay minerals (e.g. illite and vermiculite) also affect slow neutrons measurements as they contain significant amounts of hydrogen.

Moisture determination is regulated by the concentration of the hydrogen nuclei around the access tube and water is the main source of hydrogen in the soil matrix. Each soil element has a different ‘scattering cross section’ and ‘capture cross section’ and these are constants for a unit volume of soil and proportional to bulk density
 Hydrogen has a large scattering cross section that causes a large energy loss of neutrons during their collision (Gardner & Kirkham 1952). On the other hand, neutrons can be captured by some other soil elements namely potassium, chlorine, fluorine, iron (Lal 1974; Carneiro & De Jong 1985) boron (Wilson 1988), cadmium and manganese as they have a large capture cross section. In the interest of soil water management, changes in the concentration of soil carbon, nitrogen and elements other than hydrogen have a minor effect on the thermal neutron concentration. Changes in hydrogen and oxygen content in the soil matrix are mainly due to changes in soil moisture content. Therefore, in most cases, the measurement of soil moisture content by neutron probe represents the soil volumetric moisture content.

![Zone of Influence](image)

**Figure 3.2** CPN 503 Moisture meter showing the zone of influence (Charlesworth 2005)

Van Bavel *et al.* (1956) and Glasstone & Edlund (1957) defined the zone of influence of the neutron probe measurement as the volume over which the average moisture content is calculated and depends on the probe design and the amount of the moisture in the soil. The zone of influence is usually a spherical zone originated from the source of the probe. The measurement volume depends on soil moisture with a radius...
of about 0.15 m in a wet clay soil and up to 0.5 m in dry soil (Figure 3.2) (Van Bavel et al. 1956). In multi-depth moisture measurement, the volume of soil measurement is critical. The vertical resolution is more critical particularly for changes in the depth distribution of soil moisture over time. About 95% of the measured slow neutrons of a soil specified volumetric moisture content ($\theta_v$, m$^3$/m$^3$) are from the radius, $r$ (cm) of the zone of influence (Olgaard 1965). The $r$ can be expressed by following equation:

$$r = \frac{100}{1.4 + 10\theta_v}$$  \hspace{1cm} (3.3)

### 3.2.3 Neutron Probe Calibration

A SMNP cannot measure absolute moisture content automatically. For accurate measurement of the soil moisture content, calibration is necessary for different soils and soil depths (Van Bavel et al. 1961). Though, a neutron probe comes with the factory calibration for use in common soil types and for routine soil moisture determination (Dickey 1990a; Dickey 1990b) it is recommended to produce different calibration equations for different types of soil for better water management (Tyler 1988; Klenke & Flint 1991). Calibration of the SMNP requires the correlation of measured count ratio ($n$) (see Equation 3.5) against values with independent volumetric moisture content, $\theta_v$ (m$^3$/m$^3$). In most studies, $\theta_v$ has been found to be a linear function of count ratio (Greacen et al. 1981). The count ratio is typically regressed on $\theta_v$, rather than the other way around, on the assumption that $\theta_v$ measurements involve less error than measurements of $n$ as suggested by Greacen et al. (1981). Greacen et al. (1981) also claimed that regression of $n$ on $\theta_v$ produces a more accurate calibration than $\theta_v$ on $n$ as the estimation of sampling variance of count ratio is impossible. Hence, the calibration equation should be as follows:

$$n = a + b\theta_v$$  \hspace{1cm} (3.4)

where, $n$ is the ratio of neutron count rate in the soil to the count rate in a standard medium (200 L water in a drum), $\theta_v$ is the volumetric moisture content and calibration coefficients $a$ & $b$ are the intercept and slope respectively. $n$ is also defined as

$$n = x / x_s$$  \hspace{1cm} (3.5)
where, \( x \) is the neutron count measured in the soil and \( x_s \) is the standard count taken in a 200 L water filled drum.

Soil bulk density has a significant influence on neutron probe calibration (Holmes 1966; Lal 1974). Bulk density affects the slope coefficient of the calibration equation as the amount of absorbed neutrons increases with increased bulk density (Holme 1966). According to Holmes (1966) the macroscopic absorption cross section of thermal neutrons changes with changes of soil bulk density. On the contrary Olgaard & Haahr (1968) reported that the bulk density actually affects the transport cross section of both fast and slow neutrons. The bulk density effect is pronounced while measuring moisture by a neutron probe in Vertosols because they shrink and swell (Berndt & Coughlan 1976; McIntyre 1984; Kirby et al. 2003).

SMNP calibration is also affected by the access tube materials and access tube sizes (Lal 1974; Tyler 1988; Klenke & Flint 1991; Grimaldi et al. 1994). Aluminium access tubes are widely used for neutron probe counts. However, they are not recommended for saline or alkaline soils due to the corrosive potential of those soils. Stainless steel tubes were found to be very effective, however, they absorbs neutron and decrease the sensitivity of the instrument by about 2% (Hignet & Evett 2002). Polythene tubes are cheap and easy to access, however, H and C in the polythene can increase the probe count by \( \sim 20\% \) (Hignet & Evett 2002). Allen et al. (1993) used both aluminium and polyvinyl chloride (PVC) tubes in their study and reported that the root mean square error (RMSE) increased by 0.003 m\(^3\)/m\(^3\) for PVC tubes over aluminium tubes. The chlorine atoms that prevail in the PVC tube can capture neutrons and lower neutron counts (\( \sim 50\% \)) when compared to aluminium tubes (Allen & Segura 1990; Allen et al. 1993). Usually the diameter and wall thickness of the access tube vary regardless of the components of the tube. The recommended outside diameter and wall thickness of the access tube are 44 to 56 mm and 1.6 mm respectively (Hignet & Evett 2002). Larger diameters may increase the radius of the zone of influence which in turn decreases the vertical resolution of the instrument (Stone 1990).

SMNP calibration is generally performed by taking neutron counts in at least two contrasting moisture conditions, wet (field capacity) and dry. There are three basic techniques for determining the soil calibration equation: theoretical calibration; laboratory calibration; and field calibration (Bell 1987). The latter two procedures are
well established (Holmes 1956; Van Bavel et al. 1961; Greacen & Hignett 1979; Greacen et al. 1981; Bell 1987; Hignett & Evett 2002; IAEA 2003) and widely used (Hignett et al. 1980; Jayawardane et al. 1984; Tyler 1988; Allen & Segura 1990; McKenzie et al. 1990; Evett & Steiner 1995; Corbeels et al. 1999; Evett et al. 2002a; Evett et al. 2002b; Gaze et al. 2002; Heng et al. 2002; Kamilov et al. 2003; Fares et al. 2004; Yao et al. 2004). Field calibration is the most common method of calibration and provides the least methodological error (Hignett & Evett 2002). A typical example of field calibration equation for dry and wet sites is given by (Hignett & Evett 2002) (Figure 3.3).

Figure 3.3 Moisture content in wet and dry soils and the calibration line from a wet-site/dry-site SMNP calibration (Hignett & Evett 2002)

SMNP calibrations have not often been reported for Vertosol soils. Accurate calibration in swelling clay soils is critical especially for severely cracking clay soils (Richards 1965; Stirk 1972). Vertosols by definition contain high amounts of clay i.e. a heavy clay soil (Isbell 1996). The heavy clay soil exhibits shrink-swell characteristic upon drying and wetting and produces cracks during drying of the profile. The main problem of soil moisture estimation in a Vertosol is associated with its shrink-swell
characteristics as this complicates bulk density determination (Grecean & Hignett 1979; Kirby et al. 2003). In shrinking and swelling clay soils, the mass of soil per set volume changes with changes in moisture content (Murray & Quirk 1980; Yule & Ritchie 1980). Bulk density decreases with drying of the profile as large cracks are formed (Bronswijk 1990; Bronswijk & Evers-Vermeer 1990). The changes of soil volume severely affect soil mechanical properties (Yong & Warkentin 1975) particularly soil pores (Berndt & Coughlan 1976) which in turn affect the bulk density-water content relationship (Allbrook 1992). Hence, it is necessary to measure bulk density values for every measurement of gravimetric moisture content in order to convert the values to volumetric moisture content and establish the calibration equation. Changes in bulk density have been shown to cause a relative error of ≤ 30% in measured volumetric moisture content (Gardner et al. 1990).

Field measurement of soil volume change is complex. After Fox (1964), Mitchell (1992) proposed the Soil Shrinkage Characteristics Curve to express volume changes using the relationship between gravimetric moisture content and the specific volume of oven-dry soil. Peng et al. (2005) proposed a sigmoidal four-phase shrinkage curve for the whole range of moisture content. Peng’s curve was based on the relationship between the void ratio (volume of soil pores per unit volume of soil solid, m$^3$/m$^3$) and moisture ratio (volume of soil water per unit volume of soil solid, m$^3$/m$^3$). The shrinkage curve can also be shown by plotting the specific soil volume (reciprocal of bulk density) against gravimetric moisture content (Figure 3.4) (Bronswijk 1991a, 1991b; Kim et al. 1992; Tariq & Durnford 1993; Groenvelt & Grant 2001). The four-phase are:

i) Phase of structural shrinkage, this phase belongs to the wettest part of the moisture range, from the saturation to swelling limit. In this phase as water drains, water-filled macro pores are aerated and little soil volume change occurs with moisture change.

ii) Normal shrinkage phase, lies between swelling limit and air entry point (Sposito & Giraldez 1976). In this phase soil volume change is proportional to the water loss (McGarry & Malafant 1987; Tariq & Durnford 1993) and hence is called proportional shrinkage (Groenvelt & Grant 2001).
iii) Residual shrinkage phase, where the decrease of soil volume is lower than water loss and the phase ranging from air entry to shrinkage limit

iv) Phase of zero shrinkage, ranging from the shrinkage limit to the dry end. In the zero shrinkage phase, soil volume remains almost constant. No volume change occurs in this phase even if soil moisture is dried totally.

Figure 3.4 Typical plot of specific volume on gravimetric moisture content in a Vertosol to show the shrinkage characteristics curve with four shrinkage phases (after Corbeels et al. 1999).

Although effective measurement of bulk density is difficult in Vertosols particularly in dry conditions, it may help minimize sampling error during SMNP calibration. A relative error of up to 30% may be derived while estimating volumetric moisture content without considering bulk density changes in a shrink-swell soil (Gardner et al. 1990; Fares et al. 2004). To avoid this relative error, soil bulk density could be corrected for shrink-swell effects using the empirical relationship developed by Fox (1964) or bulk density values should be measured for every single measurement of gravimetric moisture content in each depth interval (e.g., Corbeels et al. 1999;
Fares et al. (2004) to calculate volumetric moisture content. Fox (1964) assumed that there is moisture content in a cracking clay soil at which the cracks are just closed. If the soil dries below this point, then the soil peds shrink in three dimensions (3D) (i.e. in addition to shrinking vertically, they also shrink in the horizontal dimensions and create the cracks). However, if the soil wets up past the crack-closure moisture content, then the soil can only expand in one dimension (1D), namely upwards. So there are two modes of soil shrinkage in the field, with 1D shrinkage occurring at the wet end of the curve, and 3D shrinkage at the dry end. Both of these modes are assumed to be within the normal or 1:1 shrinkage zone, since the equations of Fox (1964) (Equations 3.6 and 3.7) were developed based on the assumption that the change in soil volume is the same as the change in soil moisture volume. Greacen & Hignett (1979) proposed that “one-dimensional shrinkage does not occur on a field basis and that any change in field bulk density occurs in a three-dimensional mode”. In their study they corrected field bulk density using the empirical 3D shrinkage model of Fox (1964) to account for any failure to adequately sample cracks. Subsequently, Greacen et al. (1981), Hulme (1987), Hodgson & Chan (1987), Grismer et al. (1995), Corbeels et al. (1999) followed the same model to correct bulk density for shrink-swell effects.

$$\rho_{b3D} = \rho_{bRef} / \left[ \left( \rho_{bRef} / \rho_bS \right) + \left( \rho_{bRef} \times \theta_g \right) + \varepsilon \right]^{0.33}$$  \hspace{1cm} (3.6)$$

$$\rho_{b1D} = \frac{1}{\left( \frac{1}{\rho_{bS}} + \theta_g + \left( \frac{\varepsilon}{\rho_{bRef}} \right) \right)}$$  \hspace{1cm} (3.7)$$

where, \( \rho_{b3D} \) and \( \rho_{b1D} \) are bulk density corrected for three-dimensional and one-dimensional shrinkage, \( \rho_{bS} \) is the density of soil solids, \( \theta_g \) gravimetric moisture content, \( \rho_{bRef} \) is the reference bulk density, the bulk density of the point where the cracks are only just closed (the break point at the change from the 1D to 3D mode) and \( \varepsilon \) is the air filled porosity.

Greacen & Hignett (1979) also pointed out that calibration in cracking clay soils can be improved significantly by avoiding sampling error, adjusting neutron counts and considering cracks during bulk density measurement. In early work, Greacen & Schrale (1976) proposed the following model to correct neutron counts for bulk density effects in cracking clay soils.
\[ n' = \left( \frac{\overline{\rho}_{3D}}{\rho_{3D}} \right)^{0.5} \times n \]  

(3.8)

where \( n' \) is the neutron count ratio corrected for bulk density effect, \( n \) is the neutron count ratio (\( n \)), \( \overline{\rho}_{3D} \) is the mean bulk density corrected for 3D shrinkage for the soil profile of the calibration site and \( \rho_{3D} \) an individual determination of \( \rho_{3D} \) bulk density at a particular depth.

Conflicting experimental evidence has been found from studies conducted by different authors who corrected for bulk density and neutron count ratio prior to calibrating neutron probe counts in cracking clay soils using the empirical method proposed by Fox (1964) and Greacen & Schrale (1976). Greacen et al. (1981), Jayawardane et al. (1984) and Hodgson & Chan (1987) conducted studies for neutron probe calibration in Australian cracking clay soils. They found slightly improved \( R^2 \) values of neutron probe calibration after correcting the bulk density and neutron counts. In agreement, Corbeels et al. (1999) found a similar calibration improvement in Vertisols (Vertosols in Australian Soil Classification) in Morocco. On the other hand, Hulme (1987) did not find any improvement after correction for bulk density and the neutron count ratio for neutron probe calibration in an Australian Vertosol. Mitchell (1990) reviewed his own work and concluded that no bulk density correction is necessary for moderately swelling silt clay soils. To evaluate Mitchell’s findings Grismer et al. (1995) conducted an experiment in a silty clay soil in California and did not find noticeable improvement of the standard error of neutron probe measurements after correction for bulk density which supports the findings of Hulme (1987). Fares et al. (2004) compared the results between two types of volumetric moisture contents calculated from a function of gravimetric moisture content with depth specific bulk density and with a published average bulk density in a shrinking-swelling clay soil. They found 20% relative error in estimating volumetric moisture content with the method where the reference bulk density was used. Their results are in agreement with Gardner et al. (1990) who reported that the failure to account for bulk density correction may cause \( \leq 30\% \) residual error.

### 3.2.4 Capacitance Techniques for Moisture Determination

Soil moisture measurement with the capacitance technique was first introduced in the early 1930s (Smith-Rose 1933). However, the capacitance probe was first developed
commercially in the 1980s and tested under both field conditions (Bell et al. 1987) and laboratory conditions (Dean et al. 1987). Since then, the capacitance technique has been considered a viable alternative to the SMNP for soil moisture measurements. With the advent of microelectronics in the 1990s capacitance probes developed rapidly (Paltineanu & Starr 1997) and became popular for measuring in situ soil moisture content (Fares & Alva 2000). Capacitance sensors have been used extensively in soil water monitoring in a wide range of soil types for irrigation management (Buss 1993; Alva & Fares 1999) and soil water research (Starr & Paltineanu 1998b, 1998a; Fares & Alva 2000; Paltineanu & Starr 2000a, 2000b; Girona et al. 2002; Fares et al. 2004).

Different types of equipment are currently available for capacitance-based soil moisture measurement. Most of the systems consist of probes, oscillator equipment, a data logger and associated software. A probe is the main part of the instrument that consists of one or more electrodes. The design and configuration of the electrode varies depending on the use and application of the instrument. The electrodes of the probe can be two or more parallel rods (Chernyak 1964; Campbell 1990; Gardner et al. 1998), one or more pairs of cylindrical metal rings (Kuráz et al. 1970; Kuráz 1982; Dean et al. 1987; Mead et al. 1994; Ayars et al. 1995; Evett & Steiner 1995; Paltineanu & Starr 1997), a combination of one circular ring and one rod (Ould Mohamed et al. 1997; Chanzy et al. 1998) or a single sensor (Evett et al. 2002b; Geesing et al. 2004; Sentek 2004) (Figure 3.5). The probe of the Diviner 2000, a capacitance probe consists of one cylindrical ring capacitance sensor at the end of a rod (Sentek 2004). The equipment also includes a display unit (Figure 3.6) that automatically records the volumetric soil moisture content at each 0.1 m depth increment.

The capacitance technique for soil moisture determination involves the measurement of the soil dielectric constant by measuring the capacitance between two electrodes of a probe placed into the soil profile. The dielectric constant is a measure of the capacity of a non-conducting material to transmit electromagnetic waves or pulses. In essence the capacitance probe determines the velocity of an electromagnetic wave or pulse through the soil (Muñoz-Carpena 2004). The theory of the capacitance technique is described by Dean et al. (1987) and Whalley et al. (1992). When the
probe is activated a high frequency electrical field (>100 MHz) is created around the sensor in the soil-water-air matrix around the PVC access tube and forms the dielectric of a capacitor, a pair of cylindrical electrodes of the probe and completes an oscillation circuit (Dean et al. 1987). A resonant LC (\(L=inductance, C=capacitance\)) circuit in the probe that includes the ensemble of the access tube, soil in the zone of influence around the access tube and the air space between the probe and access tube, as one of the capacitive elements. Hence, changes in the resonant frequency depend on the changes in the capacitance of the soil-access tube system. The capacitance change of the soil-water-air matrix is mainly governed by the soil moisture content as the dielectric constant of water (\(K_{aw}\)) is 81, much higher than that of other soil components i.e. soil minerals and soil air valued of 2-5 and 1 respectively (Evett & Steiner 1995; Muñoz-Carpena 2004). Therefore, the measurement of the dielectric constant (capacitance of the soil system) of a soil is primarily determined by the volumetric soil moisture content.

![Two Electrodes](image1)

![Paired Cylindrical Electrodes in PVC Access Pipe](image2)

Figure 3.5 Schematic diagram of different capacitance probe design (Starr & Paltineanu 2002)

The capacitance of the soil-access tube system is measured in terms of resonant frequency around the probe. The resonant frequency decreases with the increases in the moisture content of the soil volume to be measured. The resonant frequency, \(F\), is given by Dean et al. (1987):

\[ F = \frac{1}{2\pi \sqrt{LC}} \]
where, $L$ is the inductance of the coil in the LC circuit, respectively $C_b$ & $C_c$ are the base and collector capacitance of the oscillator circuit and $C$ is the capacitance measured from the soil-access tube fringe.

The capacitance of the system is directly proportional to the moisture content of the measured soil. However, the soil dielectric constant inside the access tube should not be changed. The relationship between measured capacitance and dielectric constant is given by Dean et al. (1987):

\[ C = gK_a \]  \hspace{1cm} (3.10)

where $K_a$ is the apparent dielectric constant of the soil-access tube system $g$ is a geometrical constant that depends on the probe geometry, hence, it is difficult to calculate the geometrical constant for other than simple electrode geometry. Therefore, an overall relationship between dielectric constant and volumetric moisture content is:

\[ F = f(\theta_v) \]  \hspace{1cm} (3.11)

The relationship must be determined empirically by a calibration equation against a standard technique.

Due to wide variation of dielectric properties of soil constituents (i.e. $K_{\text{water}} = 81$ & $K_{\text{air}} = 1$) measurement of the dielectric constant by capacitance probes is extremely sensitive to air space around the access tube. In addition, capacitance probes have a very narrow radial range of sensitivity that originates from the centre of the probe. Paltineanu and Starr (1997) obtained 99% sensitivity within a 0.1 m radius and 92% sensitivity within a 0.03 m radius of outside of the access tube. Hence, even small variations in the soil properties close to the access tube such as air gaps between soil and access tube have to be taken into consideration.

In recent years portable capacitance sensors have been found to be a suitable alternative to the SMNP (Williams 2002; Charlesworth 2005). The first prototype of portable capacitance-type sensor was proposed by de Plater (1955). Recently, several
unique portable sensors have been developed commercially for field application. The use of a portable capacitance probes is very straightforward. The sensor of the probe is lowered into the access tube and soil moisture is measured from a number of depths of one profile within few seconds (e.g., Bell et al. 1987; Tomer & Anderson 1995; Evett et al. 2002b; Heng et al. 2002; Geesing et al. 2004; Groves & Rose 2004). Other probes involve placing an array of sensors at predefined depths into an access tube and logging the out frequency at a regular intervals (e.g., Buss 1993; Paltineanu & Starr 1997; Baumhardt et al. 2000; Fares et al. 2004; Kelleners et al. 2004; Hamdhani et al. 2005; Polyakov et al. 2005) and in some sensors by installing the instrument directly into the soil (e.g., Robinson & Dean 1993; Hilhorst & Dirksen 1994; Ould Mohamed et al. 1997; Seyfried & Murdock 2004).

Calibration is generally done by estimating the Scaled Frequency ($SF$) of the probe. The $SF$ is generated by comparing the probe response in the access tube in the soil to the probe responses in air and water. A calibration equation is then determined by regressing $SF$ against volumetric moisture content. The standard regression equation for capacitance probe calibration is derived by a non-linear power model (Sentek 2004):

$$\theta_v = a \cdot (SF)^b$$  \hspace{1cm} (3.12)

where $a$ and $b$ are the fitting coefficients, $\theta_v$ is the volumetric moisture content and $SF$ is the scaled frequency. Scaled frequency is a normalized value such that $SF$ in air = 0 and $SF$ in water = 1 and is calculated using the following equation as outlined by Sentek (2004).

$$SF = \frac{(F_A - F_S)}{(F_A - F_W)}$$  \hspace{1cm} (3.13)

where $F_A$ is the frequency reading in the PVC access tube while suspended in air, $F_S$ is the reading in the PVC access tube installed in the soil at a particular depth, and $F_W$ is the reading in the PVC access tube suspended in the water bath.

It was reported by Hulme (1997) that the uneven distribution of soil water will affect the scaled frequency of the capacitance probe which necessitated the regression of $SF$ on $\theta_v$. Consequently, the calibration equation should be created by regressing $SF$ on $\theta_v$.
as reported by Wu (1998), Hulme (1997), Heng et al. (2002), Kelleners et al. (2004) rather than $\theta_v$ on $SF$.

The manufacturer company (Sentek, Australia) of the capacitance probes (Diviner 2000, EnviroSCAN) performed sensor calibration using the non-linear power model. In previous studies the calibration of capacitance probe was done using a two parameter power model (e.g., Paltineanu & Starr 1997; Morgan et al. 1999; Baumhardt et al. 2000; Heng et al. 2002; Geesing et al. 2004; Groves & Rose 2004) and a three-parameter power model (e.g., Baumhardt et al. 2000; Fares et al. 2004; Polyakov et al. 2005) and found to give acceptable precision of measurement. In addition, a linear model was also attempted in several studies to calibrate the capacitance probe (e.g., Bell et al. 1987; Evett & Steiner 1995; Tomer & Anderson 1995; Ould Mohamed et al. 1997; Evett et al. 2002b). The form of the calibration equations for linear and power model is as follows:

$$SF = a + b\theta_v$$  \hspace{1cm} (3.14)

and

$$SF = a\theta_v^b$$  \hspace{1cm} (3.15)

Capacitance probes come with a factory calibration based on a wide range of soils. Although factory calibration is generally not sufficient for scientific studies, it can be used for estimating relative water change (Tomer & Anderson 1995; Paltineanu & Starr 1997; Alva & Fares 1998; Starr & Paltineanu 1998a; Alva & Fares 1999; Sentek 2001). Soil specific calibration of the probe is needed for accurate work because the dielectric constant is influenced not only by soil moisture content but also by soil physical and chemical characteristics such as texture, shrinking & swelling, organic matter content and salinity (Paltineanu & Starr 1997; Muñoz-Carpena 2004). The need for soil specific calibration of capacitance sensors is also supported by other studies (Baumhardt et al. 2000; Lane & Mackenzie 2001; Morgan et al. 2001; Fares et al. 2004; Polyakov et al. 2005).

The basic principles for the calibration of capacitance sensors are similar to those of the SMNP. Although few scientific studies on the calibration of the Diviner 2000 capacitance probe (eg. Evett et al. 2002b; Geesing et al. 2004; Groves & Rose 2004)
have been done, the calibration procedure is the same as for the EnviroSCAN capacitance probe that has been documented in many studies (e.g., Bell et al. 1987; Evett & Steiner 1995; Mead et al. 1995; Paltineanu & Starr 1997; Morgan et al. 1999; Baumhardt et al. 2000; Kelleners et al. 2004).

![Diagram of soil sampling positioning](image1)

![Diagram of improper access tube installation](image2)

Figure 3.7 Positioning of soil sampling (Sentek 2001)
Figure 3.8 Effect of improper access tube installation (Sentek 2001)

Although Australia has a wide range of Vertosols, the calibration of capacitance probes in this soil type is not well documented. Due to the sensitivity of capacitance probes to air gaps, common in cracking clay soils, the relationship between scaled frequency and volumetric moisture content may be poor. Nonetheless, capacitance probes are widely used in agriculture for irrigation scheduling on Vertosols, for example by the cotton industry. In Australia, Sentek (2001) developed the best-fit calibration equation of EnviroSCAN capacitance probes using data taken from four cracking clay soils at different times. Hulme (1997) conducted a study for the calibration of EnviroSCAN capacitance probes on cracking clay soils and found an $R^2$ value of 0.58 with standard error (where SF is regressed on $\theta_v$) of 0.051 m$^3$/m$^3$. Fares et al. (2004) conducted calibration in different depths of a red-brown cracking clay soil in Australia. They found good relationships ($R^2 = 0.83$) between scaled frequency and volumetric water content at different depths.
3.2.5 **Comparison of SMNP and Capacitance Techniques**

There is limited research that compares the comparative performance of neutron probes and capacitance probes for soil moisture assessment and no clear trend of superior performance has emerged. Evett & Steiner (1995), Heng et al. (2002) and Evett et al. (2002b) found the neutron probe better than the capacitance probe for routine soil moisture measurement. Evett & Steiner (1995) compared six neutron scattering gauges and four capacitance probes and found that the capacitance probes provide poor precision while the neutron scattering gauges provided acceptable precision. They also concluded that the poor precision of capacitance probes may arise from the difficulties during the installation of access tubes (Figure 3.8), error in soil sampling (Figure 3.7) and large measurement intervals (0.1524 m) rather than sampling volume, as suggested by Bell et al. (1987). Heng et al. (2002) conducted a study in a coarse clay loam soil with a considerable amount of gravel particularly below 0.5 m to compare the Troxler SMNP, Diviner 2000 capacitance probe and Time Domain Reflectometer (TDR). Their study was carried out based on individual plots (3.4 m x 5 m) and different irrigation treatments (furrow & drip) with three replications for each instrument. They found that the Diviner 2000 provided the largest standard error and lowest precision of the three probes. However, they suggested that using more access tubes and measuring sites for the Diviner probe may increase measurement precision. Evett et al. (2002b) compared SMNP and capacitance probes in four continents and concluded that capacitance sensor overestimate moisture content near saturation and underestimate near wilting point compared to neutron probes in Australian soil. In contrast, Tomer & Anderson (1995), Ould Mohamed et al. (1997) and Evett et al. (2002c) obtained better results with a capacitance probe in comparison with a neutron probe. Ould Mohamed et al. (1997), calibrated the capacitance probe in a silty clay loam soil by making direct contact between the electrodes and soil while the neutron probe was calibrated using a theoretical calibration process. Their study suggests that the lack of air gaps between electrodes and the soil improved the results of the capacitance probe. This result is the consistent of the findings of Tomer & Anderson (1995). Evett et al. (2002c) compared neutron and capacitance probes in silty clay, clay and clay loam soils and found the capacitance probe was more accurate than the neutron probe.
An important component of comparing both probes was the performance of their measurements at the 0.1 m depth. Both the neutron and capacitance probes have a measurement error at the 0.1 m depth due to the erratic loss of neutrons and the effect of air above the soil surface respectively. Theoretically, when the neutron probe measurements are taken close to the surface some of the thermal neutrons are lost causing an error in moisture measurements (IAEA 1970; Bell 1987; Falleiros et al. 1993). Greacen et al. (1981) proposed different correction factors for different depths close to the soil surface to adjust for the loss of neutrons during surface measurements. Another approach to minimize neutron loss from the surface measurements was to calibrate topsoil measurements separately to the rest of the profile. Reedy & Scanlon (2003) used a correction factor rather than doing separate calibration for surface measurements. On the other hand, Cull (1979), Jayawardane et al. (1984), Hodgson & Chan (1987), Evett (2000a), Hignett & Evett (2002) and Fares et al. (2004) did a separate calibration for the top soil.

Wu (1998) reported that capacitance probe readings taken from 0.1 m were affected by the air above the surface and did not reflect the real value of moisture content. Consequently, he found a poor relationship for upper 0.1 m measurements. Evett & Steiner (1995) found a similar result, very low coefficient of determination ($R^2 = 0.01 - 0.2$) at the 0.1 m depth and a good coefficient of determination ($R^2 = 0.68 - 0.71$) at the 0.41 – 1.02 m depth group in four experimental sites. In contrast, Robinson & Dean (1993) and Fares et al. (2004) did separate calibrations for 0.1 m depths and found high $R^2$ values of 0.94.

A summary of the scientific studies that compare the performance of capacitance probes and neutron probes is listed in Table 3.1. Moreover, a comparative evaluation criterion of capacitance probe and neutron probe is displayed in Table 3.2.
Table 3.1 Summary of scientific studies comparing neutron probe and capacitance probe performance

<table>
<thead>
<tr>
<th>Author(s)</th>
<th>Soil type</th>
<th>Results</th>
</tr>
</thead>
<tbody>
<tr>
<td>Heng <em>et al.</em> (2002)</td>
<td>Clay loam with high content of gravel</td>
<td>SMNP gave better precision than CP</td>
</tr>
<tr>
<td>Evett <em>et al.</em> (2002b)</td>
<td>Medium clay, Silt loam to Silty clay loam</td>
<td>CP overestimated and underestimated moisture content compared to SMNP in near saturation and wilting points respectively.</td>
</tr>
<tr>
<td>Evett <em>et al.</em> (2002c)</td>
<td>Silty clay loam, clay, calcic clay loam</td>
<td>CP more accurate than SMNP</td>
</tr>
<tr>
<td>Ould Mohamed <em>et al.</em> (1997)</td>
<td>Silty clay loam</td>
<td>CP more accurate than SMNP</td>
</tr>
<tr>
<td>Tomer &amp; Anderson (1995)</td>
<td>Fine sand</td>
<td>Both SMNP and CP gave good correlation with moisture content $R^2 = 0.97$ and 0.92 respectively.</td>
</tr>
<tr>
<td>Evett and Steiner (1995)</td>
<td>Fine sandy loam</td>
<td>Poor correlation for CP gauges ($R^2 = 0.68$ to 0.71) and SMNP gauges ($R^2 = 0.97$ to 0.99).</td>
</tr>
</tbody>
</table>
Table 3.2 Evaluation criteria for neutron probe and capacitance probe (Charlesworth 2005)

<table>
<thead>
<tr>
<th>Features</th>
<th>Neutron probe</th>
<th>Capacitance probe</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reading range</td>
<td>Wide range of water content</td>
<td>Wide range of water content</td>
</tr>
<tr>
<td>Stated accuracy</td>
<td>± 0.5% when calibrated</td>
<td>± 0.1% to ± 0.5% when calibrated</td>
</tr>
<tr>
<td>Measurement volume</td>
<td>~ 0.15 m and ~ 0.5 m sphere radius for wet and dry soil respectively</td>
<td>0.1 m sphere radius</td>
</tr>
<tr>
<td>Installation method</td>
<td>Aluminium or PVC access tube, sensor embedded in access tube</td>
<td>PVC access tube, sensor embedded in access tube, sensors inserted into the soil</td>
</tr>
<tr>
<td>Logging capability</td>
<td>No</td>
<td>Portable probes have no logging capability while others have.</td>
</tr>
<tr>
<td>Soil types not recommended</td>
<td>None</td>
<td>None</td>
</tr>
<tr>
<td>Field Maintenance</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>Safety hazards</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>Application</td>
<td>Irrigation, research, consultants</td>
<td>Irrigation, research</td>
</tr>
</tbody>
</table>

3.3 SCOPE OF THIS CHAPTER

It is clear that SMNP and capacitance probes have great potential to monitor soil moisture in fields. However, several questions remain to be answered concerning the value of each sensor across a range of soil types; considering variability in key soil characteristics and soil specific calibration sampling methodologies. While there have been several studies to compare the performance of SMNP and capacitance probes in a range of soils none of these studies has been conducted on Vertosols (Table 3.1).
Therefore, the aim of this chapter was to compare the performance of SMNP and capacitance probes on a Vertosol soil with specific objectives to:

1. Compare the performance (accuracy, precision and efficiency) of single core and pit soil sampling techniques to measure profile depth specific variation in soil bulk density and soil moisture content for the purpose of sensor calibration

2. Develop and compare calibration models (sensor count to volumetric soil moisture) using the SMNP and Diviner 2000 capacitance probe.

3.4 MATERIALS AND METHODS

A SMNP (Borat Longyear CPN, CPN 503 DR Hydroprobe) and a CP (Sentek, Diviner 2000) were calibrated against gravimetrically determined water contents in a Black Vertosol soil at Clark’s Farm, UNE (Chapter 2)

Two methods of calibration were compared:

I. A single core calibration method, in which samples for gravimetric moisture and bulk density were taken from the core extracted when an access tube was installed, and

II. A pit calibration method, in which samples for gravimetric moisture and bulk density were taken as intact cores by excavating around an access tube.

3.4.1 Single Core Sampling and Analysis

For the single core calibration method, 18 access tubes were installed for both the neutron and capacitance probes measurements throughout the study area. Of the 18 core sites, 10 sites were selected using a systematic sampling procedure (20 m × 20 m grid), two were selected based on the topography of the study site and the remaining six were selected using Spatial Response Surface (SRS) sampling design applied to an EM38 survey (Lesch 2005) (Figure 3.9). Before generating the SRS sampling design with ESAP software (Lesch et al. 2000), 10 m buffer zones around the other 12 sites were created. Buffering was done to maintain a minimum distance between existing and new sampling sites.
Cores were obtained using a stainless steel sampling tube with 50 mm inside cutting diameter and 56 mm outside diameter driven by a hydraulic ram (Figure. 3.10) to a depth of approximately 1.7 m. Sentek PVC access tubes (56 mm outside diameter and 51.5 mm inside diameter) were fitted immediately into the core holes (Figure 3.10). The lower part of the access tube was sealed with a PVC conical cap. Access tubes were pushed down the hole to approximately 0.25 m deeper than the greatest depth (1.5 m) to be measured to allow sufficient fringe zones for the measuring probes. Care was taken during access tube installation to ensure that the tubes were fitted well to the soil to prevent vertical leaking of surface water around the tube and to minimize air gaps.

The length of the extracted core was compared to the core hole depth to determine if core compaction had occurred. No measurable compaction was recorded. A plastic top-cap was screwed on the top of the tubes to prevent water access (Figure 3.11).

Each core was sectioned into 16 sub-samples, 0 – 0.05 m and then every 0.1 m. Each sub-sample was placed in a pre-weighed, sealed tin and transported for laboratory analysis. To avoid moisture loss during sampling a plastic sheet was used to cover the cores and each core was sectioned and sealed immediately following extraction. All samples were then taken to the UNE Agronomy laboratory for moisture content and bulk density determination and subsequent soil analysis. Moisture content samples were oven-dried at 105 °C for approximately one week until they reached a constant weight. Bulk density was measured from samples taken from different depths of each core. Sub-sample volumes were determined by multiplying the core cross sectional area by sub-sample length. Volumetric moisture content was estimated using Equation 3.1. For subsequent statistical analysis, data from 17 cores were taken as the first core was treated as test core and the sampling depth of the core was not consistent with the others. In addition bulk density was corrected for 3D shrinkage ($\rho_{b3D}$) using Equation 3.6 to see if there is any effect of soil shrinkage on bulk density measurements and neutron probe calibration. The corrected volumetric moisture content for 3D shrinkage ($\theta_{3D}$) was also calculated from the $\theta_s$ and corresponding $\rho_{b3D}$ values using same principles as in Equation 3.1.

On the same day as core extraction 16-second neutron probe readings were obtained thrice from each tube at each soil sampling depth. Regressing the neutron count ratio
against the volumetric moisture content at each specific depth using simple linear regression was performed to calibrate the neutron probe. The neutron count ratio corrected for bulk density effects \((n')\) was calculated using Equation 3.8 and \((n')\) was also regressed on \(\theta_{3D}\). Standard readings were taken before and after each measurement to check the instruments drift.

Similarly, the Diviner 2000 probe was placed into each access tube and readings were taken three times. Before taking the Diviner 2000 readings, the probe was normalized using Equation 3.13 as each sensor responds differently to air and water. By default the Diviner 2000 takes readings from 0.1 m depth increments. The Diviner 2000 output is volumetric moisture content generated from SF using the manufacturers default calibration (Equation 3.16). Therefore SF was back-calculated from the Diviner readings by applying the default calibration in reverse:

\[
SF = 0.2746 \times (\theta_D)^{0.3314} \tag{3.16}
\]

where, \(\theta_D\) is the volumetric moisture content measured by the Diviner 2000 probe. The three calculated SF values of each depth were then averaged and paired with corresponding \(\theta_v\) to develop the calibration equation.

---

Figure 3.9 Spatial distribution of access tubes used in the single core calibration method.
Figure 3.10 (A) Extracting the soil core with a hydraulic ram to install the PVC access tube (B) Installing PVC access tube into the hole.

Figure 3.11 PVC access tubes covered with a cap

3.4.2 Pit Method Sampling

Two artificially wet plots (5 m × 3 m) and two artificially dry plots (5 m × 3 m) were prepared to create two extreme moisture conditions i.e. wet and dry for instrument calibration. A wet and dry plot was established at the approximate top and bottom elevations of the study site. To establish the dry plots, rain exclusion shelters (a ‘greenhouse’ set up with a frame and clear plastic sheeting) were established for approximately six months prior to measurement. To exclude horizontal soil water movement, black plastic sheets were inserted vertically in a 1.2 m trench around the upslope side of the shelter. Existing deep-rooted pasture (Lolium rigidum Chloris...
truncate, Phalaris minor, Bromus spp., Bothriochloa macra) was allowed to use available soil moisture from the profile. The dry plots were initially watered and fertilized to stimulate pasture growth.

The wet plots were located 10 m apart from the dry plots at the same slope and contour. Water was initially applied to the plot using a fire tender. Wetting was augmented by heavy rainfall (90.4 mm in 8 days) 7 days prior to sampling.

In each plot two single core access tubes were established to allow SMNP and Diviner 2000 measurements as described in Section 3.4.1. Subsequent to core extraction two 1.8-m-deep sampling pits were excavated in each plot using a back-hoe (4 pits in total with 8 sampling locations) approximately 0.4 m away from the access tubes.

At each pit, 3 undisturbed ring samples (73 mm diameter and 36 mm depth) were collected from the topsoil (0.1 m). Then at depths of 0.2, 0.4, 0.6, 0.8, 1.0 and 1.2 m, 3 larger ring samples (99 mm diameter and 78 mm depth) were collected by first manually exposing each level without disturbing the underlying soil and then extracting the samples from each site. Core rings were protected with a steel cap and driven vertically and evenly into the soil using a slide hammer. Samples were taken in a ring pattern within a distance of 0.13 m from the access tubes. Each sample core was trimmed to the ring volume and then weighed and placed in the oven for drying. Laboratory analysis and data processing was identical to the procedure outlined in Section 3.4.1.

3.4.3 Statistical Analysis

Linear regression of \( n \) on \( \theta_v \), \( n' \) on \( \theta_{3D} \) and \( SF \) on \( \theta_v \) were done using JMP statistical software (SAS 2005). In order to find out if there were any significant differences between depths for the regression lines of \( n \) and \( \theta_v \) and \( SF \) on \( \theta_v \) for both sampling methods, the slopes and the intercepts of the regression equation of all depths of two methods were compared following ‘Comparison of Regression Lines’ using Statgraphics Centurion software (Statpoint 2005). All other regressions were done by using same software. A Tukey’s significant mean difference test was performed to quantify the difference among the bulk densities at different depths of the profile using JMP statistical software (SAS 2005). The accuracy of the predicted moisture
content regression model was tested by the value of the root mean square error (RMSE). The RMSE was explained as follows:

\[
RMSE = \sqrt{\frac{1}{n} \sum_{i=1}^{n} (x_i - \hat{x}_i)^2}
\]  

(3.17)

where, \(x_i\) are gravimetrically measured moisture content, \(\hat{x}_i\) are predicted moisture content by the regression model and \(n\) is the observation number of the data. Therefore, the RMSE represents the average variance between measured and predicted moisture content. Volumetric units of moisture content (m\(^3\)/m\(^3\)) were used to compare the prediction accuracy of different models.

3.5 RESULTS

3.5.1 Bulk Density

As expected soil bulk density measured by the pit method increased with increasing soil depth at all sites, however, the mean bulk densities for depths down to 1.0 m were not significantly different while the bulk density of 1.2 m was only significantly different with those of 0.1, 0.2 and 0.4 m depths (Table 3.3). There was no significant difference between depths for bulk densities measured using the single core method. For the pit samples the specific volume (reciprocal of bulk density) and gravimetric moisture content of this experiment showed a positive relationship (Table 3.4) at each depth confirming the shrink-swell characteristics of this soil, however, it did not follow the “normal shrinkage” curve completely, as indicated by slopes of < 1 Mg/m\(^3\).

On the other hand, the relationship between specific volume and moisture content was not statistically significant for the single core method. Regression lines analysis between the specific volume and \(\theta_g\) revealed that the there was a significant difference in bulk densities measured by the two methods at 0.1, 0.6, 0.8 and 1.0 m depths (Figure 3.12 and Table 3.5). Figure 3.13 also demonstrated that the bulk densities derived using the pit method resemble more closely to the theoretical 3D line at upper depth (0.1 m) and are closer to the 1D shrinkage curve of Fox (1964) at deeper depths (0.8 and 1.2 m).
Table 3.3 Depth wise minimum, maximum and mean bulk densities and maximum and minimum gravimetric moisture content for the single core and the pit calibration sites. Mean bulk density at depths not connected by same letter in the homogenous group column are significantly different (P<0.05).

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>$\theta_g$ kg/kg</th>
<th>Bulk Density (Mg/m$^3$)</th>
<th>Homogenous Group</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Max</td>
<td>Mean</td>
<td>Min</td>
</tr>
<tr>
<td>Pit Method</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.565</td>
<td>0.383</td>
<td>0.221</td>
</tr>
<tr>
<td>0.2</td>
<td>0.604</td>
<td>0.428</td>
<td>0.280</td>
</tr>
<tr>
<td>0.4</td>
<td>0.597</td>
<td>0.435</td>
<td>0.301</td>
</tr>
<tr>
<td>0.6</td>
<td>0.588</td>
<td>0.442</td>
<td>0.319</td>
</tr>
<tr>
<td>0.8</td>
<td>0.532</td>
<td>0.434</td>
<td>0.333</td>
</tr>
<tr>
<td>1.0</td>
<td>0.506</td>
<td>0.402</td>
<td>0.319</td>
</tr>
<tr>
<td>1.2</td>
<td>0.450</td>
<td>0.368</td>
<td>0.262</td>
</tr>
<tr>
<td>Single Core Method</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.493</td>
<td>0.350</td>
<td>0.254</td>
</tr>
<tr>
<td>0.2</td>
<td>0.498</td>
<td>0.374</td>
<td>0.245</td>
</tr>
<tr>
<td>0.4</td>
<td>0.433</td>
<td>0.349</td>
<td>0.268</td>
</tr>
<tr>
<td>0.6</td>
<td>0.457</td>
<td>0.346</td>
<td>0.286</td>
</tr>
<tr>
<td>0.8</td>
<td>0.464</td>
<td>0.350</td>
<td>0.307</td>
</tr>
<tr>
<td>1.0</td>
<td>0.430</td>
<td>0.339</td>
<td>0.230</td>
</tr>
<tr>
<td>1.2</td>
<td>0.473</td>
<td>0.344</td>
<td>0.240</td>
</tr>
</tbody>
</table>
Table 3.4 Relationship between specific volume and gravimetric moisture content at individual sampling depths quantified by linear regressions for both calibration methods. The slopes and intercepts are for the regression equation: \( \text{SP} = a + b\theta_g \), where \( \text{SP} \) and \( \theta_g \) are specific volume and gravimetric moisture content; \( b \) & \( a \) are slope and intercept of regression equation and \( N \) is the number of observation, respectively.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Slope</th>
<th>Intercept</th>
<th>( R^2 )</th>
<th>RMSE</th>
<th>P-value</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pit Method</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.329</td>
<td>0.770</td>
<td>0.800</td>
<td>0.027</td>
<td>0.003</td>
<td>8</td>
</tr>
<tr>
<td>0.2</td>
<td>0.500</td>
<td>0.680</td>
<td>0.780</td>
<td>0.040</td>
<td>0.004</td>
<td>8</td>
</tr>
<tr>
<td>0.4</td>
<td>0.554</td>
<td>0.642</td>
<td>0.700</td>
<td>0.046</td>
<td>0.009</td>
<td>8</td>
</tr>
<tr>
<td>0.6</td>
<td>0.610</td>
<td>0.592</td>
<td>0.760</td>
<td>0.037</td>
<td>0.004</td>
<td>8</td>
</tr>
<tr>
<td>0.8</td>
<td>0.701</td>
<td>0.535</td>
<td>0.890</td>
<td>0.020</td>
<td>0.000</td>
<td>8</td>
</tr>
<tr>
<td>1.0</td>
<td>0.799</td>
<td>0.491</td>
<td>0.950</td>
<td>0.014</td>
<td>0.000</td>
<td>8</td>
</tr>
<tr>
<td>1.2</td>
<td>0.664</td>
<td>0.538</td>
<td>0.950</td>
<td>0.011</td>
<td>0.000</td>
<td>8</td>
</tr>
<tr>
<td>Single Core Method</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.347</td>
<td>0.832</td>
<td>0.110</td>
<td>0.072</td>
<td>0.195</td>
<td>17</td>
</tr>
<tr>
<td>0.2</td>
<td>0.473</td>
<td>0.742</td>
<td>0.220</td>
<td>0.071</td>
<td>0.056</td>
<td>17</td>
</tr>
<tr>
<td>0.4</td>
<td>0.159</td>
<td>0.882</td>
<td>0.002</td>
<td>0.135</td>
<td>0.852</td>
<td>17</td>
</tr>
<tr>
<td>0.6</td>
<td>0.659</td>
<td>0.653</td>
<td>0.120</td>
<td>0.077</td>
<td>0.168</td>
<td>17</td>
</tr>
<tr>
<td>0.8</td>
<td>0.598</td>
<td>0.638</td>
<td>0.130</td>
<td>0.065</td>
<td>0.154</td>
<td>17</td>
</tr>
<tr>
<td>1.0</td>
<td>0.693</td>
<td>0.581</td>
<td>0.330</td>
<td>0.052</td>
<td>0.017</td>
<td>17</td>
</tr>
<tr>
<td>1.2</td>
<td>0.822</td>
<td>0.555</td>
<td>0.220</td>
<td>0.103</td>
<td>0.058</td>
<td>17</td>
</tr>
</tbody>
</table>

Table 3.5 Significance test of regression equations of \( \text{SP} = a + b\theta_g \) by both calibration methods. RL = Number of regression lines, N = Number of total observations, * indicates significantly different at 95% confidence level

<table>
<thead>
<tr>
<th>Comparative Regression Equations</th>
<th>P-value</th>
<th>RL</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Slope</td>
<td>Intercept</td>
<td></td>
</tr>
<tr>
<td>Pit_{0.1} vs Single core_{0.1}</td>
<td>0.948</td>
<td>0.022*</td>
<td>2</td>
</tr>
<tr>
<td>Pit_{0.2} vs Single core_{0.2}</td>
<td>0.919</td>
<td>0.084</td>
<td>2</td>
</tr>
<tr>
<td>Pit_{0.4} vs Single core_{0.4}</td>
<td>0.633</td>
<td>0.114</td>
<td>2</td>
</tr>
<tr>
<td>Pit_{0.6} vs Single core_{0.6}</td>
<td>0.919</td>
<td>0.038*</td>
<td>2</td>
</tr>
<tr>
<td>Pit_{0.8} vs Single core_{0.8}</td>
<td>0.879</td>
<td>0.050*</td>
<td>2</td>
</tr>
<tr>
<td>Pit_{1.0} vs Single core_{1.0}</td>
<td>0.746</td>
<td>0.032*</td>
<td>2</td>
</tr>
<tr>
<td>Pit_{1.2} vs Single core_{1.2}</td>
<td>0.785</td>
<td>0.067</td>
<td>2</td>
</tr>
</tbody>
</table>
Chapter: Three

Comparison of Field Moisture Probes

(a) Specific Volume vs. Gravimetric Moisture Content

(b) Specific Volume vs. Gravimetric Moisture Content

(c) Specific Volume vs. Gravimetric Moisture Content

0.1 m

0.2 m

0.4 m
Comparison of Field Moisture Sensors

Gravimetric Moisture Content (kg/kg) vs. Specific Volume (m$^3$/Mg)

- **0.6 m**
- **0.8 m**
- **1.0 m**
Figure 3.12 Comparison of regression lines (a-g) between gravimetric moisture content and specific volume of different depths using the two calibration methods.
Figure 3.13 Relationship between bulk density and gravimetric moisture content for a particular depth layer of (a) 0.1 m, (b) 0.8 m and (c) 1.2 m from pit samples. The corresponding 3D and 1D line are the theoretical relationships of Fox (1964) derived using the values of $\rho_{bRef} = 1.3 \text{ Mg/m}^3$, $\rho_{bS} = 2.65 \text{ Mg/m}^3$, $\theta_g = 0.48 \text{ kg/kg}$ and $\varepsilon = 0.03 \text{ m}^3/\text{m}^3$. 
3.5.2 Neutron Probe Calibration

Neutron probe data were calibrated using both \( n' - \theta_{3D} \) and \( n - \theta_r \) data. Bulk density corrected data (\( n' \) and \( \theta_{3D} \)) did not improve the results in most of the cases. Consequently, uncorrected data sets were used to compare between depths and methods. Comparative results of neutron probe calibration with both data sets are illustrated in Table 3.8 and Figure 3.14.

Effects of Depths on Calibration

Calibration of the neutron probe was found to be statistically significant (\( P < 0.01 \)) at each depth for both the pit and the single core calibration methods (Table 3.6). For each method the slopes and intercepts of groups of depths were compared to find out whether there was a significant difference between the regression lines at different depths (Table 3.7). The regression lines for the top 0.1 m in both methods were significantly different in both slope and intercept in comparison to the other depths. Hence, a separate calibration was performed for the top 0.1 m depth. There was a significant difference in slope and intercept for depths from 0.2–0.8 m, but the regression lines for depths between 0.2–0.6 m were not significantly different for either method. Therefore, data from the depths 0.2–0.6 m were pooled. The deep samples from depth groups 0.8–1.2 m, for which the regression slopes and intercepts were not significantly different, were also pooled.

A single calibration was also done for the entire profile to compare with the separate depth group calibrations (Table 3.8). As expected, lower precision (low \( R^2 \) value) and accuracy (high RMSE value) were observed using the combined single calibration equation in both the single core (\( R^2 = 0.54, \ RMSE = 0.025 \ m^3/m^3 \)) and pit methods (\( R^2 = 0.85, \ RMSE = 0.027 \ m^3/m^3 \)). Thus, separate calibration equations (0.1 m, 0.2–0.6 m and 0.8–1.2 m) were used to quantify the volumetric moisture content for the profile.
Table 3.6 Significance test of regression of $n$ on $\theta_v$ in each individual depth for both the single core and the pit calibration methods. The slopes and intercepts are for the calibration equation: $n = a + b\theta_v$, where $n$ and $\theta_v$ are neutron count ratio and volumetric moisture content; $b$ & $a$ are slope and intercept of calibration equation and $N$ is the number of observation, respectively.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Max $\theta_v$ $m^3/m^3$</th>
<th>Min $\theta_v$ $m^3/m^3$</th>
<th>Slope</th>
<th>Intercept</th>
<th>$R^2$</th>
<th>RMSE $m^3/m^3$</th>
<th>P-value</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Pit Method</td>
<td>Single Core Method</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.578</td>
<td>0.512</td>
<td>0.257</td>
<td>0.254</td>
<td>0.830</td>
<td>0.028</td>
<td>0.76</td>
<td>8</td>
</tr>
<tr>
<td>0.2</td>
<td>0.599</td>
<td>0.522</td>
<td>0.324</td>
<td>0.267</td>
<td>0.654</td>
<td>0.140</td>
<td>0.70</td>
<td>8</td>
</tr>
<tr>
<td>0.4</td>
<td>0.601</td>
<td>0.448</td>
<td>0.374</td>
<td>0.241</td>
<td>0.591</td>
<td>0.180</td>
<td>0.60</td>
<td>8</td>
</tr>
<tr>
<td>0.6</td>
<td>0.609</td>
<td>0.503</td>
<td>0.410</td>
<td>0.321</td>
<td>0.495</td>
<td>0.233</td>
<td>0.60</td>
<td>8</td>
</tr>
<tr>
<td>0.8</td>
<td>0.583</td>
<td>0.490</td>
<td>0.446</td>
<td>0.318</td>
<td>0.382</td>
<td>0.297</td>
<td>0.51</td>
<td>8</td>
</tr>
<tr>
<td>1.0</td>
<td>0.568</td>
<td>0.517</td>
<td>0.429</td>
<td>0.249</td>
<td>0.416</td>
<td>0.279</td>
<td>0.60</td>
<td>8</td>
</tr>
<tr>
<td>1.2</td>
<td>0.534</td>
<td>0.524</td>
<td>0.405</td>
<td>0.204</td>
<td>0.004</td>
<td>0.287</td>
<td>0.47</td>
<td>8</td>
</tr>
</tbody>
</table>


Table 3.7 Significance of differences in slope and intercept of regression lines for count ratio ($n$) regressed against volumetric moisture content ($\theta_v$) within selected depth groupings for both the pit and the single core calibration methods. RL = Number of regression lines, N = Number of total observations, ** and * indicate significantly different at 99% and 95% confidence level respectively.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Pit Method</th>
<th>Single Core Method</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Slope</td>
<td>Intercept</td>
</tr>
<tr>
<td>0.1 - 1.2</td>
<td>0.000**</td>
<td>0.000**</td>
</tr>
<tr>
<td>0.1 - 0.2</td>
<td>0.037*</td>
<td>0.004**</td>
</tr>
<tr>
<td>0.2 - 0.8</td>
<td>0.012*</td>
<td>0.010*</td>
</tr>
<tr>
<td>0.2 - 0.6</td>
<td>0.127</td>
<td>0.102</td>
</tr>
<tr>
<td>0.8 - 1.2</td>
<td>0.837</td>
<td>0.981</td>
</tr>
</tbody>
</table>

Effects of Methods on Calibration

The calibration equations for two different methods are presented in Table 3.8 and Figure 3.14. Linear regression was performed between volumetric moisture content and neutron count ratio for all calibration equations in the different depth groups. The calibration equations for each depth groups were highly significant ($p < 0.001$) for both the single core and the pit methods. The precision of calibration for the pit method in each depth group was greater than that of the single core method. High precision (high $R^2$) of neutron probe measurement was observed in top layer for both methods, however, the accuracy for top layer calibration is lower (high RMSE) than the calibration of other layers. The range of coefficients of determination ($R^2$) was 0.54 to 0.76 for the single core method and 0.94 to 0.97 for the pit method. The value of $R^2$ for top layer (0.1 m) of the pit method was 0.97. For the single core method, the top layer again accounted for the strongest relationship ($R^2 = 0.76$) between moisture content and count ratio than that of other depth groups.
Comparison of regression lines was performed to investigate the difference between group specific regressions of the two methods using an ANOVA (Statpoint 2005). The comparison revealed that both the slope and intercept of the regression equations of the single core method were significantly different ($P < 0.05$) from those of the pit method above 0.6 m and the intercepts were different below 0.8 m ($P < 0.05$) (Table 3.9).

Table 3.8 Calibration equations for the neutron probe for the two calibration methods. The slopes and intercepts are for the calibration equation: $n = a + b\theta_v$, where $n$ and $\theta_v$ are neutron count ratio and volumetric moisture content; $b$ & $a$ are slope and intercept of the calibration equation and $N$ is the number of observation, respectively.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Slope</th>
<th>Intercept</th>
<th>$R^2$</th>
<th>RMSE</th>
<th>P-value</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Pit Method: Uncorrected</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.830</td>
<td>0.028</td>
<td>0.97</td>
<td>0.024</td>
<td>0.000</td>
<td>8</td>
</tr>
<tr>
<td>0.2-0.6</td>
<td>0.609</td>
<td>0.169</td>
<td>0.94</td>
<td>0.015</td>
<td>0.000</td>
<td>24</td>
</tr>
<tr>
<td>0.8-1.2</td>
<td>0.400</td>
<td>0.288</td>
<td>0.95</td>
<td>0.005</td>
<td>0.000</td>
<td>24</td>
</tr>
<tr>
<td>0.1-1.2</td>
<td>0.699</td>
<td>0.125</td>
<td>0.85</td>
<td>0.027</td>
<td>0.000</td>
<td>56</td>
</tr>
<tr>
<td></td>
<td>Pit Method: Corrected</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.961</td>
<td>-0.010</td>
<td>0.95</td>
<td>0.032</td>
<td>0.000</td>
<td>8</td>
</tr>
<tr>
<td>0.2-0.6</td>
<td>0.680</td>
<td>0.162</td>
<td>0.89</td>
<td>0.024</td>
<td>0.000</td>
<td>24</td>
</tr>
<tr>
<td>0.8-1.2</td>
<td>0.466</td>
<td>0.285</td>
<td>0.96</td>
<td>0.006</td>
<td>0.000</td>
<td>24</td>
</tr>
<tr>
<td>0.1-1.2</td>
<td>0.728</td>
<td>0.147</td>
<td>0.75</td>
<td>0.039</td>
<td>0.000</td>
<td>56</td>
</tr>
<tr>
<td></td>
<td>Single Core Method: Uncorrected</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.581</td>
<td>0.189</td>
<td>0.76</td>
<td>0.023</td>
<td>0.000</td>
<td>17</td>
</tr>
<tr>
<td>0.2-0.6</td>
<td>0.331</td>
<td>0.324</td>
<td>0.61</td>
<td>0.016</td>
<td>0.000</td>
<td>51</td>
</tr>
<tr>
<td>0.8-1.2</td>
<td>0.295</td>
<td>0.354</td>
<td>0.54</td>
<td>0.014</td>
<td>0.000</td>
<td>51</td>
</tr>
<tr>
<td>0.1-1.2</td>
<td>0.446</td>
<td>0.278</td>
<td>0.54</td>
<td>0.025</td>
<td>0.000</td>
<td>119</td>
</tr>
<tr>
<td></td>
<td>Single Core Method: Corrected</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.885</td>
<td>0.127</td>
<td>0.83</td>
<td>0.022</td>
<td>0.000</td>
<td>17</td>
</tr>
<tr>
<td>0.2-0.6</td>
<td>0.541</td>
<td>0.284</td>
<td>0.61</td>
<td>0.019</td>
<td>0.000</td>
<td>51</td>
</tr>
<tr>
<td>0.8-1.2</td>
<td>0.608</td>
<td>0.288</td>
<td>0.86</td>
<td>0.010</td>
<td>0.000</td>
<td>51</td>
</tr>
<tr>
<td>0.1-1.2</td>
<td>0.614</td>
<td>0.264</td>
<td>0.43</td>
<td>0.030</td>
<td>0.000</td>
<td>119</td>
</tr>
</tbody>
</table>
Table 3.9 Significance test of regression equations \((n = a + b\theta_v)\) comparison between the pit and single core calibration method. RL = Number of regression lines, N = Number of total observations, ** and * indicate significantly different at 99% and 95% confidence level respectively.

<table>
<thead>
<tr>
<th>Comparative Regression Equations</th>
<th>P-value</th>
<th>RL</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Single core(<em>{0.1}) vs Pit(</em>{0.1})</td>
<td>0.026*</td>
<td>0.000**</td>
<td>2</td>
</tr>
<tr>
<td>Single Core(<em>{0.2-0.6}) vs Pit(</em>{0.2-0.6})</td>
<td>0.000**</td>
<td>0.000**</td>
<td>2</td>
</tr>
<tr>
<td>Single Croc(<em>{0.8-1.2}) vs Pit(</em>{0.8-1.2})</td>
<td>0.061</td>
<td>0.000**</td>
<td>2</td>
</tr>
</tbody>
</table>

![Graph showing count ratio vs. volumetric moisture content](image)

- **R\(^2\) = 0.97**
- **R\(^2\) = 0.95**
- **R\(^2\) = 0.83**
- **R\(^2\) = 0.76**

The graph shows the relationship between count ratio and volumetric moisture content for different depth ranges, with regression lines indicating the goodness of fit for each range.
Figure 3.14 Calibration equations for the neutron probe at different depths i.e. (a) 0.1 m (b) 0.2 – 0.6 m and (c) 0.8 – 1.2 m for both the pit and single core calibration methods with corrected and uncorrected data.
3.5.3 Diviner 2000 Calibration

The calibration of Diviner 2000 was performed by doing regression analyses to establish the relationship between volumetric moisture content and the scaled frequency (SF) of the Diviner. Calibration equations were explored using linear models and non linear power models. Since the linear model produced better results than the power model in every case the results with linear model were used to compare results between depths and methods. Notwithstanding, a comparative performance of Diviner 2000 calibration using both linear and power models is presented in Table 3.14.

Effect of Depths on Calibration

The calibration equations were statistically significant at all depths for the single core method (Table 3.10). For the pit method, the calibration equations were highly significant for the upper three depths (0.1, 0.2 and 0.4 m), and the equations for 0.6 m and 1.0 m depths were statistically significant while the calibration for depths 0.8 m and 1.2 m were not significant. The response of the scaled frequency to the water content decreased with depths (Table 3.10). As the regression equations for two depths below 0.6 m were not significant, the calibration of Diviner 2000 was considered only for depths up to 0.6 m. In the single core calibration method the responses of scaled frequency to moisture content were statistically significant at every single depth. Consequently all depths were considered for the calibration process.

The regression lines at each depth for both methods were compared with each other, and selected comparisons are presented in Table 3.11. The regression line of the top depth (0.1 m) was found to be significantly different from those of other depths (0.2, 0.4 and 0.6 m). There was no significant difference of regression lines of depth groups 0.2–0.6 m. Therefore, a separate calibration was done for top 0.1 m depth and a pooled regression was performed for rest of the depths (0.2–0.6 m) for the pit method. For the single core calibration method, there was no significant difference between the regression lines for different depths. Consequently, one combined calibration equation was established to cover entire profile for the single core method.
Table 3.10 Significance test of regression of \(SF\) on \(\theta_v\) in each individual depth for both the pit and the single core calibration methods. The slopes and intercepts are for the calibration equation: \(SF = a + b\theta_v\), where \(SF\) and \(\theta_v\) are scaled frequency of Diviner 2000 readings and volumetric moisture content; \(b\) & \(a\) are slope and intercept of calibration equation and \(N\) is the number of observation, respectively.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Max (\theta_v) (m^3/m^3)</th>
<th>Min (\theta_v) (m^3/m^3)</th>
<th>Slope</th>
<th>Intercept</th>
<th>(R^2)</th>
<th>RMSE (m^3/m^3)</th>
<th>P-value</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pit Method</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.578</td>
<td>0.257</td>
<td>0.293</td>
<td>0.046</td>
<td>0.93</td>
<td>0.012</td>
<td>0.000</td>
<td>8</td>
</tr>
<tr>
<td>0.2</td>
<td>0.599</td>
<td>0.324</td>
<td>0.527</td>
<td>-0.091</td>
<td>0.96</td>
<td>0.013</td>
<td>0.000</td>
<td>8</td>
</tr>
<tr>
<td>0.4</td>
<td>0.601</td>
<td>0.374</td>
<td>0.559</td>
<td>-0.105</td>
<td>0.87</td>
<td>0.023</td>
<td>0.000</td>
<td>8</td>
</tr>
<tr>
<td>0.6</td>
<td>0.609</td>
<td>0.410</td>
<td>0.449</td>
<td>-0.042</td>
<td>0.56</td>
<td>0.034</td>
<td>0.032</td>
<td>8</td>
</tr>
<tr>
<td>0.8</td>
<td>0.583</td>
<td>0.446</td>
<td>0.224</td>
<td>0.092</td>
<td>0.33</td>
<td>0.019</td>
<td>0.138</td>
<td>8</td>
</tr>
<tr>
<td>1.0</td>
<td>0.568</td>
<td>0.429</td>
<td>0.063</td>
<td>0.183</td>
<td>0.64</td>
<td>0.002</td>
<td>0.018</td>
<td>8</td>
</tr>
<tr>
<td>1.2</td>
<td>0.534</td>
<td>0.405</td>
<td>-0.051</td>
<td>-0.231</td>
<td>0.08</td>
<td>0.012</td>
<td>0.458</td>
<td>8</td>
</tr>
<tr>
<td>Single Core Method</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.512</td>
<td>0.254</td>
<td>0.176</td>
<td>0.096</td>
<td>0.42</td>
<td>0.015</td>
<td>0.004</td>
<td>17</td>
</tr>
<tr>
<td>0.2</td>
<td>0.522</td>
<td>0.267</td>
<td>0.147</td>
<td>0.102</td>
<td>0.50</td>
<td>0.011</td>
<td>0.001</td>
<td>17</td>
</tr>
<tr>
<td>0.4</td>
<td>0.448</td>
<td>0.241</td>
<td>0.148</td>
<td>0.099</td>
<td>0.41</td>
<td>0.006</td>
<td>0.005</td>
<td>17</td>
</tr>
<tr>
<td>0.6</td>
<td>0.503</td>
<td>0.321</td>
<td>0.142</td>
<td>0.102</td>
<td>0.36</td>
<td>0.009</td>
<td>0.010</td>
<td>17</td>
</tr>
<tr>
<td>0.8</td>
<td>0.490</td>
<td>0.318</td>
<td>0.147</td>
<td>0.102</td>
<td>0.40</td>
<td>0.008</td>
<td>0.006</td>
<td>17</td>
</tr>
<tr>
<td>1.0</td>
<td>0.517</td>
<td>0.314</td>
<td>0.139</td>
<td>0.106</td>
<td>0.38</td>
<td>0.010</td>
<td>0.009</td>
<td>17</td>
</tr>
<tr>
<td>1.2</td>
<td>0.524</td>
<td>0.280</td>
<td>0.138</td>
<td>0.108</td>
<td>0.39</td>
<td>0.011</td>
<td>0.008</td>
<td>17</td>
</tr>
</tbody>
</table>

Table 3.11 Significance tests of differences between regression lines in different depths for both the single core and the pit calibration methods (\(SF\) were regressed on \(\theta_v\)). RL = Number of regression lines, \(N\) = Number of total observations, ** and * indicate significantly different at 99% and 95% confidence level respectively.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>P-value</th>
<th>RL</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pit Method</td>
<td>Slope</td>
<td>Intercept</td>
<td></td>
</tr>
<tr>
<td>0.1 - 0.6</td>
<td>0.044*</td>
<td>0.039*</td>
<td>4</td>
</tr>
<tr>
<td>0.1-0.2</td>
<td>0.000**</td>
<td>0.000**</td>
<td>2</td>
</tr>
<tr>
<td>0.2-0.6</td>
<td>0.769</td>
<td>0.744</td>
<td>3</td>
</tr>
</tbody>
</table>

| Single Core Method | P-value | RL | N  |
| 0.1 - 1.2 | 0.996 | 0.500 | 7 | 119 |
Effect of Methods on Calibration

Calibration of Diviner 2000 was done using a linear regression model (Table 3.12 and Figure 3.15). The coefficient of determination for all equations of the two calibration methods ranged from 0.08 at 1.2 m depth to 0.93 in the top layer of the pit method. The R$^2$ values were 0.93 and 0.83 for depth groups 0.1 m and 0.2-0.6 m respectively for the pit method. On the other hand, the combined calibration equation for the single core method to a depth of 1.2 m produced an R$^2$ value of 0.42.

Separate calibration of individual depths showed only a slight improvement at one depth in model accuracy and precision for both the single core and pit methods (Tables 3.10 and 3.12).

Comparison of regression equations showed that slopes of the regression lines were significantly different between two methods (Table 3.13) and that the pit method gave improved R$^2$ in comparison to the single core method.

Calibration was also explored using a two-parameter power model. Compared with the linear model, the power model did not improve either the precision (R$^2$) or accuracy (RMSE) of measurement in any of the calibration equations (Table 3.14).

Table 3.12 Calibration equations for Diviner 2000 for the two methods. The slopes and intercepts are for the calibration equation: SF = a + b$\theta_v$, where SF and $\theta_v$ are scaled frequency and volumetric moisture content; b & a are slope and intercept of the calibration equation and N is the number of observation, respectively.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Slope</th>
<th>Intercept</th>
<th>R$^2$</th>
<th>RMSE</th>
<th>P-value</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1</td>
<td>0.293</td>
<td>0.046</td>
<td>0.93</td>
<td>0.013</td>
<td>0.000</td>
<td>8</td>
</tr>
<tr>
<td>0.2-0.6</td>
<td>0.527</td>
<td>-0.087</td>
<td>0.83</td>
<td>0.023</td>
<td>0.000</td>
<td>24</td>
</tr>
<tr>
<td>0.8</td>
<td>0.224</td>
<td>0.092</td>
<td>0.33</td>
<td>0.019</td>
<td>0.138</td>
<td>8</td>
</tr>
<tr>
<td>1.0</td>
<td>0.063</td>
<td>0.183</td>
<td>0.64</td>
<td>0.002</td>
<td>0.018</td>
<td>8</td>
</tr>
<tr>
<td>1.2</td>
<td>-0.051</td>
<td>-0.231</td>
<td>0.08</td>
<td>0.485</td>
<td>0.458</td>
<td>8</td>
</tr>
<tr>
<td>0.1-1.2</td>
<td>0.149</td>
<td>0.101</td>
<td>0.422</td>
<td>0.011</td>
<td>0.000</td>
<td>119</td>
</tr>
</tbody>
</table>
Table 3.13 Significance test of regression equations ($SF = a + b\theta_v$) comparison between both pit and single core calibration method. RL = Number of regression lines, N = Number of total observations, ** and * indicate significantly different at 99% and 95% confidence level respectively.

<table>
<thead>
<tr>
<th>Comparative Regression Equations</th>
<th>P-value</th>
<th>RL</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Single Core $0.1-1.2$ vs Pit$_{0.1}$</td>
<td>0.000**</td>
<td>0.329</td>
<td>2</td>
</tr>
<tr>
<td>Single Core $0.1-1.2$ vs Pit$_{0.2-0.6}$</td>
<td>0.000**</td>
<td>0.000**</td>
<td>2</td>
</tr>
</tbody>
</table>

Table 3.14 Comparison of calibration equations developed using simple linear regression and power regression model for Diviner 2000 for two methods. $SF = a + b\theta_v$ and $SF = a\theta_v^b$ are linear and power model of calibration equations where $SF$ and $\theta_v$ are scaled frequency and volumetric moisture content; $b$ & $a$ are slope intercept of the calibration equations and $N$ is the number of observation, respectively.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>$R^2$</th>
<th>RMSE</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pit Method</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.93</td>
<td>0.013</td>
<td>0.034</td>
</tr>
<tr>
<td>0.2-0.6</td>
<td>0.83</td>
<td>0.023</td>
<td>0.076</td>
</tr>
<tr>
<td>Single Core Method</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1-1.2</td>
<td>0.422</td>
<td>0.011</td>
<td>0.028</td>
</tr>
</tbody>
</table>

![Graph](a)
Chapter: Three

Comparison of Field Moisture Probes

3.5.4 Comparison of Performance of Neutron Probe and Diviner 2000

Since the pit method of calibration was found to be better and the response of the Diviner 2000 below the depth 0.6 m was not significant (p > 0.05), the performance of two probes were compared based on $\theta_v$ measurement using the calibration equations for 0.1 m and 0.2-0.6 m depths for the pit method.

To assess the repeatable performance of the neutron probe and the Diviner 2000, soil moisture measurements with both probes were analysed from the 15 access tubes across the study site, 5 times during 2006 and 2007. The volumetric moisture content was calculated from the corresponding count ratio and scaled frequency data using the calibration equations in Tables 3.8 and 3.12. The moisture content data were then
paired according to depths. The minimum, maximum, mean $\theta_v$, standard deviation (SD) and difference of mean and SD of $\theta_v$ measured by two probes are presented in Table 3.15.

Table 3.15 shows that the Diviner measures higher water content with higher SD at all depths (0.1, 0.2-0.6, 0.1-0.6 m). The differences of mean $\theta_v$ and SD between the two probes at different depths range from 0.012 to 0.016 m$^3$/m$^3$ and 0.011 to 0.030 m$^3$/m$^3$ respectively. The highest difference of the variability (SD) was found in 0.2-0.6 m depth group with the value of 0.030 m$^3$/m$^3$.

Comparing the average moisture content up to 0.6 m depth, the $\theta_v$ measured by the neutron probe was normally distributed (skewness and kurtosis were −0.346 and -0.589 respectively) with a standard error of 0.005 m$^3$/m$^3$ while the $\theta_v$ measured by Diviner 2000 was not normally distributed (skewness and kurtosis were -0.051 and -1.375 respectively) with a 0.007 m$^3$/m$^3$ standard error. A variance check test was performed to identify whether there was a significant difference between the variability of $\theta_v$ measurements given by both probes. The P-value of the variance check test was 0.005 and indicated the variability of $\theta_v$ measured by the neutron probe was statistically different (P<0.05) from that of the Diviner 2000 measurement. Figure 3.16 shows that a lower variation of $\theta_v$ was associated with the neutron probe measurements than the Diviner 2000 measurements.

Table 3.15 Minimum, maximum, mean and standard deviation (SD) for volumetric moisture content measured by neutron probe and Diviner 2000 at different depths. Difference of mean- $\theta_v$ and SD between two probes is also shown.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>$\theta_v$ by NP (m$^3$/m$^3$)</th>
<th>SD</th>
<th>$\theta_v$ by Diviner (m$^3$/m$^3$)</th>
<th>SD</th>
<th>Diviner-NP</th>
<th>$\theta_v$ mean</th>
<th>SD</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1</td>
<td>0.403</td>
<td>0.528</td>
<td>0.498</td>
<td>0.047</td>
<td>0.381</td>
<td>0.591</td>
<td>0.513</td>
</tr>
<tr>
<td>0.2-0.6</td>
<td>0.354</td>
<td>0.584</td>
<td>0.486</td>
<td>0.054</td>
<td>0.258</td>
<td>0.599</td>
<td>0.499</td>
</tr>
<tr>
<td>0.1-0.6</td>
<td>0.407</td>
<td>0.581</td>
<td>0.501</td>
<td>0.047</td>
<td>0.414</td>
<td>0.589</td>
<td>0.516</td>
</tr>
</tbody>
</table>
Figure 3.16 Plotting of Standard deviations to quantify the variability of $\theta_v$ measured by Neutron Probe and Diviner 2000.

3.6 DISCUSSION AND CONCLUSION

3.6.1 Bulk Density and Gravimetric Moisture Content

The slopes of the regression equations of specific volume and $\theta_v$ were less than 1 for both calibration methods indicating that the shrinkage behaviour of the data did not fall into the normal shrinkage zone i.e. soil volume change was not exactly equal to soil moisture change (Figure 3.4). This is due to the cracking through the soil profile. Table 3.5 and Figure 3.12 show that the bulk densities measured using the two methods (pit and single core) were significantly different for most depths. As expected the bulk densities of the pit method were found to be better than those of single core method. This was due to the more rigorous and painstaking sampling method that better considers the crack volume present in the profile.

The gravimetric moisture content (pit method) at upper depth (0.1 m) fitted the 3D theoretical shrinkage curve of Fox (1964) and was approaching the 1D shrinkage curve with increasing depths (Figure 3.13). The results indicated that the 3D shrinkage occurred at low moisture content at the upper depth when vertical cracks were open and 1D shrinkage occurred when vertical cracks were closed. This is in
agreement of the results obtained by Berndt & Coughlan (1977) and Hodgson & Chan (1987). The results also agree with Bronswijk (1990), who found that soil in the field generally shows a 3D shrinkage pattern.

3.6.2 Pit versus Single Core Methods

The pit calibration method was better for calibrating both the neutron probe and the Diviner 2000. The poor performance of the single core calibration method compared to the pit method for both probes was probably due to the poorer bulk density estimates. In fact, the methods of the calibration in this study were differentiated based on the soil sampling procedure for determining volumetric moisture content. The samples for the single core method neither completely covered the zones of influence of the probes nor the intensity of sampling volume resulting in errors in bulk density and volumetric moisture content determination. On the other hand, in the pit method, replicated samples were taken from less disturbed cores which reduced error in the sampling process and hence is considered a better method for establishing calibration equations.

The response of the scaled frequency of the capacitance probe to the moisture content decreased with increasing depth and did not show a significant relationship for depths below 0.6 m in the pit method. This is in agreement with the findings of Hulme (1997) who calibrated a capacitance probe in cracking clay soil and found similar responses with depth at two of his experimental sites. The reasons for this variation with depths are still unknown, however, it could be assumed that there were some variations of soil structure in deeper depths of the profile that influences the dielectric response resulting in an insignificant relationship between moisture content and scale frequency. A number of studies have shown that the soil dielectric constant is sensitive to soil texture and structure (Wobschall 1978; Kuráž 1981; Hallikainen et al. 1985; Bell et al. 1987). In addition, the SMNP samples quite large soil volumes, comparable to the size of peds even at depth. However, the Diviner samples a very small volume, smaller than the size of the peds in the subsoil. Both the SMNP and the Diviner sample the volume of soils in relation to the ped size and as peds get larger with depth this may lead to a non significant calibration for Diviner at great depths (Hulme 2008, personal communication).
Separate calibrations for different depth groups were found to give better soil moisture estimates for both probes. The regression of the neutron probe at the 0.1 m depth was significantly different from all other depths. The higher RMSE value (0.02 m³/m³) was also associated with top 0.1 m depth. This is likely to be due to neutron escape into the atmosphere during the measurement from the top 0.1 m (Evett 2000a; Hignett & Evett 2002). This result is similar to those found by previous workers (e.g., Cull 1979; Jayawardane et al. 1984; Hodgson & Chan 1987; Fares et al. 2004). In agreement with other studies, this study found calibration improvements if separate calibration equations were developed for different depths or groups of depths. The differences of calibration equations in different depths is likely to be due to the variations in bulk density-moisture content relationship, clay content, and organic matter content and CaCO₃ (McKenzie et al. 1990). Correction of neutron counts for bulk density effects did not improve the precision and accuracy of the pit calibration hence the uncorrected calibration equation was recommended for predicting moisture content by neutron probe.

For the Diviner 2000, top layer (0.1 m) calibration gave the best results for any depth and either sampling method. Results from the study conducted by Robinson & Dean (1993) and Fares et al. (2004) are in agreement with those of this study.

### 3.6.3 SMNP versus Diviner 2000

The neutron probe gave the best soil moisture estimates for each depth group. This result was consistent with findings of Evett & Steiner (1995), Evett et al. (2002b), and Heng et al. (2002). The reasons for the relatively poor performance of Diviner 2000 in this study are likely to be related to the high clay content and cracking nature of the study site. The soil properties that most influence dielectric response are soil structure and texture (Tomer & Anderson 1995) and the 2:1 type clay minerals of Vertosols that could affect the performance of capacitance sensors. The second likely cause is the small sphere of influence of the Diviner 2000 that is likely to be highly sensitive to soil voids and cracks (Evett & Steiner 1995; Wu 1998). Another important cause is that the spatial variability of moisture content caused by local soil structure variability that may affect the Diviner 2000 more than the SMNP. The soils of this study area are strongly uniform in texture both down the profile and across the study site. However, clay mineralogy may be different due to the variation of parent materials in different
topography. Generally the 2:1 type clay minerals are prominent in bottom slopes (Smith 1959, Kalmer & Yaalon 1984), leads to the severe cracking, than in upper slopes. Consequently, scattered results may be expected from the capacitance measurements due to the sensitivity to soil voids and cracks as supported by other authors (Evett & Steiner 1995; Wu 1998; Baumhardt et al. 2000). Similar results to this study have also been recorded with the repeated measurements of the both probes (Section 3.5.4) where the Diviner gave more scattered results (SE = 0.007 m$^3$/m$^3$) than the neutron probe (SE = 0.005 m$^3$/m$^3$).

Table 3.15 shows that small variations in soil moisture content produce large variations between the two measurements. The lowest difference of variability is associated with the maximum moisture content at depths 0.1-0.6 m depth. This is probably due to the sensitivity of capacitance probe to the cracks that develop when the profile becomes dry. The difference between $\theta_v$ measurement with the two probes increases with decreasing moisture content (Table 3.15). This is in agreement with the findings of Tomer & Anderson (1995) and Ould Mohamed et al. (1997) who found that the differences between the two probes decrease with increased moisture content.

In conclusion the SMNP provided better estimates of soil moisture content both in precision and accuracy for the complete profile down to 1.2 m in comparison to the Diviner 2000. Additionally, pit calibration is necessary for SMNP calibration in such a shrink-swell Vertosol.
CHAPTER – FOUR

FIELD DETERMINATION OF SOIL MOISTURE IN THE ROOT ZONE BY MULTI-HEIGHT EM38 MEASUREMENTS
4.1 INTRODUCTION

There are a number of classes of non-invasive soil moisture probing instruments that rely on electrical or electromagnetic phenomena, including soil resistivity, electrical capacitance, intrinsic magnetism and electromagnetic induction (McNeill 1980, 1986; Dean et al. 1987; Whalley & Stafford 1992; Basson et al. 1993; Evett & Steiner 1995; White & Zegelin 1995; Paltineanu & Starr 1997; Lund et al. 1999; Golovko & Pozdnyakov 2007). This and subsequent chapters will be confined to the phenomenon of electromagnetic induction. Electromagnetic induction sensors typically integrate the below-ground response to interrogate electromagnetic fields, and if the depth-response function of the probe is known, the potential exists to invert the integrated response to give a depth profile of the driving soil attribute (for example electrical conductivity). This chapter investigates the use of the EMI instrument (e.g., EM38) to detect and describe the soil moisture depth profile.

4.2 THE EM38

The EM38 is a widely-used example of an electromagnetic instrument for soil sensing developed by Geonics Ltd. (Ontario, Canada). It comprises two electrical coils, one a transmitter (Tx) and the other a receiver (Rx), placed 1 metre apart in a wooden (Canadian Cherry) frame (Figure 4.1).

The transmitter coil is excited with a sinusoidal current at a frequency of 14.6 kHz. This creates a time-varying magnetic field in the vicinity of the coil (Figure 4.2).

Another widely-used instrument, not used in the present work but referred to in other work, is the EM31. It works on exactly the same principle except the inter-coil spacing is 3.67 m (McNeill 1996). When activated on the ground the time-varying magnetic field induces eddy currents (approximated as circular electrical current loops) in the soil. The magnitude of the eddy currents is proportional to the electrical conductivity of the soil in that layer of the soil (Figure 4.2). Each current loop generates a secondary magnetic field proportional to the value of the current flowing within the loop. A fraction of the secondary magnetic field from each loop is intercepted by the receiver coil of the instrument, the sum of these signals is amplified and formed into an output voltage. The process of electromagnetic induction results in a secondary magnetic field which is 90° out of phase with the primary field. The
sensor coil is designed to measure this out-of-phase component, hence the notion of quadrature.

Figure 4.1 The photograph of the EM38 showing the locations of transmitter coil (Tx), receiver coil (Rx), and inter-coil spacing (s).

Figure 4.2 Schematic diagram of the Geonics EM38 unit showing locations of transmitter and receiver coils, the spatial structure of the primary magnetic field during the peak current phase within the transmitter coil and that of the secondary magnetic fields generated in response eddy currents generated in the conductive medium. In this vertical dipole mode of operation, the primary field lines shown ($H_P$) have cylindrical symmetry around the vertical axis (Lamb et al. 2005)
At low induction numbers (that is, a value for the ratio of the distance between transmitter and receiver coils, to conductor skin-depth), the apparent conductivity ($\sigma_a$) in the vicinity of the transmitter coil is determined by the ratio of the magnitudes of the out-of-phase secondary to primary magnetic fields using the following equation (McNeill 1980):

$$\sigma_a = \frac{4}{i\omega \mu_0 \sigma^2} \left( \frac{H_s}{H_p} \right)_Q$$  \hspace{1cm} (4. 1)

where $\left( \frac{H_s}{H_p} \right)_Q$ is the ratio of the out-of-phase secondary to primary magnetic fields (subscript ‘$Q$’ denotes quadrature, that is, $90^\circ$ out of phase)

$\sigma_a$ = apparent electrical conductivity  
$\omega = 2\pi f$ ($f$ = frequency in Hz)  
$\mu_0$ = permeability of free space = $4\pi \times 10^{-7}$ kg m/s$^2$.A$^2$  
$s$ = distance between transmitter and sensing coils or inter-coil spacing = 1 m  
$i = \sqrt{-1}$ (which merely denotes quadrature phase)  
$A$ = area of the (transmitter) coil

In the context of soil probing the apparent electrical conductivity in Equation 4.1 is given the symbol EC$\sigma_a$

In cylindrical coordinates, the axial and radial components, respectively, of the primary magnetic field in the vicinity of a magnetic dipole, such as that formed by the small transmitter coil in the EMI unit, are given by (Wait 1982):

$$H_{Pz} (r,z) = \frac{IA}{4\pi} \left[ \frac{3z^2}{(r^2 + z^2)^{3/2}} - \frac{1}{(r^2 + z^2)^{3/2}} \right]$$  \hspace{1cm} (4. 2)

$$H_{Pr} (r,z) = \frac{IA}{4\pi} \left[ \frac{3rz}{(r^2 + z^2)^{3/2}} \right]$$  \hspace{1cm} (4. 3)

where, $H_{Pz}$ is the axial component of the magnetic field  
$H_{Pr}$ is the radial component of the magnetic field  
$z$ is the axial distance, along the axis of the field symmetry, from the coil  
r is the radial distance, perpendicular to the axis of field symmetry, from the coil  
$A$ is the area of the (transmitter) coil
The configuration of the sensor, and resulting primary magnetic field lines, shown in Figure 4.2, is referred to as the vertical dipole mode of operation as this is defined by the axis of cylindrical-symmetry of the primary magnetic field lines. Placing the unit on its side, where the axis of symmetry of primary field lines is now horizontal, produces a configuration corresponding to what is known as horizontal dipole mode. The approximation in (Equation 4.1) holds for both horizontal as well as vertical dipole modes (McNeill 1980).

A key assumption in understanding the nature of the integrated response of the surface measurement of EMI instruments like the EM38 is that individual, below ground ‘current loops’ are not influenced by others near by (McNeill 1980). Consequently, the net secondary magnetic field at the receiver is the sum of the independent secondary magnetic fields from each of the individual current loops. This gives rise to the notional depth-response of the EMI sensor according to the relative contributions of secondary magnetic fields arising from different depths directly below the sensor. For vertical and horizontal dipole configurations (Kaufman 1983 after McNeill 1980), these contributions are given respectively as

\[
\phi^V(z) = \frac{4z}{(4z^2 + 1)^{3/2}}
\]

(4.4)

\[
\phi^H(z) = 2 - \frac{4z}{(4z^2 + 1)^{1/2}}
\]

(4.5)

Here \( z \) is the ratio of axial distance below the sensor, \( z \), and inter-coil spacing, \( s \) (1 m). Both of these expressions, shown graphically in Figure 4.3, are developed from the notion that the sensor is placed, regardless of vertical or horizontal dipole mode of operation, on the surface of a conductive half-layer, whereby there is no conductive medium above the surface \( z > 0 \) and a conductive medium below the surface \( z < 0 \). This is a reasonable assumption given the EMI instrument is placed on top of the ground surface, in air. Note that increasing the inter-coil spacing, for example to that of an EM31 unit \( s \approx 3.67 \) m ‘stretches’ the x-axis by a similar factor.

The exact amplitude and phase of the secondary field will differ from those of the primary field as a result of soil properties (e.g., clay content, moisture content and salinity), spacing of the coils and their orientation, frequency, and distance from the
soil surface (Hendrickx et al. 2002). The response of the instrument depends on the sensing depth of the soil and top soil profile texture (Davis et al. 1997). Soil profiles with higher average electrical conductivity yield a greater instrument response as shown in Figure 4.4 (Sudduth et al. 2001). Clayey soils have a higher electrical conductivity than coarser textured soils. In Figure 4.4, the thicker circles illustrate soils that are better conductors of electrical current. The higher EMI readings persist on the shallow clayey topsoil (b in Figure 4.4) whereas less EMI response is found on the deeper clayey topsoil (a in Figure 4.4).

Figure 4.3 Relative responses of EM38 to the secondary magnetic field in a homogenous profile at different depths for the vertical (—) and horizontal (---) dipole configurations. Curves calculated from McNeill (1980).

Although Equations 4.4 and 4.5 imply no limit to the penetration depth of EM38, the ‘effective’ measurement depth of operation is accepted as 1.5 m and 0.75 m as 70% of the integrated response generated from those depths for vertical and horizontal dipole operations respectively (McNeill 1980). Rhoades & Corwin (1981), on the other hand report that the EM38 is primarily responsive to a depth of 1.2 m compared to other
depths. Notwithstanding this discrepancy, the effective measurement depths of EM38 is accepted as being appropriate to agriculture as the depths represent the root zone area (Corwin & Lesch 2005a).

Figure 4.4 Principle of operation of EM38 in soils. The thicker circles indicate soils that are better conductors of electrical current (a) deeper clayey topsoil giving less response (b) shallow clayey topsoil giving higher response (Davis et al. 1997)

Figure 4.5 Schematic diagram of the primary and secondary magnetic fields of the EM38 when it is in operation. Tx, Rx, Hp, Hs and dashed circle <⋯> are the transmitter coil, the receiver coil, the primary magnetic field, the secondary magnetic field and current loop respectively (after Norman 1990)
4.2.1 Theory of EC\textsubscript{a} Measurement of Soils

The theory of apparent soil electrical conductivity (EC\textsubscript{a}) measurement is based on the Archie’s empirical law (Archie 1942) for saturated rocks and sand soil.

\[
EC_a = a \times \sigma_w \times \varphi^m
\]  

(4.6)

where,

- \(a\) = empirical constant
- \(\sigma_w\) = electrical conductivity of the porous media solution (dS/m),
- \(\varphi\) = the porosity (m\(^3\)/m\(^3\)) and
- \(m\) = the cementation exponent (often called porosity exponent, \(m\) was observed to be higher in cemented rock. The average value of \(m\) for typical reservoir rocks is often taken as 2)

As described earlier, the electromagnetic induction process involves the flow of eddy currents in the soil. Corwin & Lesch (2005a) described three pathways of current flow that contribute to the EC\textsubscript{a} of a soil namely i) a solid-liquid phase pathway primarily via exchangeable cations associated with clay minerals, ii) a liquid phase pathway via salts contained in the soil moisture occupying the large pores and iii) a solid phase pathway via soil particles that are in direct and continuous contact with one another (Figure 4.6).

The study of Rhoades et al. (1989) revealed that the movement of electrons through a soil solution is complex and they identified that soil structure does not provide enough direct particle-to-particle contact to form a continuous pathway for current flow. Hence, they identified the major factors that affect soil EC\textsubscript{a} measurement as being: the electrical conductivity of the soil particles (EC\textsubscript{s}); the electrical conductivity of the soil solution associated with discontinuous pores (EC\textsubscript{ws}); the electrical conductivity of the mobile soil solution associated with large, continuous pores (EC\textsubscript{wc}); the volumetric soil moisture content of the small, discontinuous pores (\(\theta\textsubscript{ws}\)); the total volumetric content of moisture in the soil (\(\theta\textsubscript{w}\)); and the volumetric content of soil particles (\(\theta\textsubscript{s}\)). Among those the vital factors influencing EC\textsubscript{a} are \(\theta\textsubscript{w}\) (Rhoades et al. 1976; Nadler 1982); EC\textsubscript{s}, influenced by cation exchange capacity (Shainberg et al. 1980); \(\theta\textsubscript{s}\) per unit of soil, initially influenced by soil texture and bulk density.
(Rhoades & Corwin 1990); and \( EC_{ws} \) and \( EC_{wc} \), influenced by the amount of dissolved salts in the soil solution (Malicki & Walczak 1999).

![Diagram](image)

Figure 4.6 The cross section of a soil showing the three-phase electrical conductance pathways for the \( EC_a \) measurement (Corwin & Lesch 2005a)

The theoretical model of three-phase electrical conductance was initially developed by Rhoades et al. (1989) and the Dual-Pathway Parallel Conductance (DPPC) model is related according to:

\[
EC_a = \left[ \frac{\left( \theta_{ss} + \theta_{ws} \right)^2 \cdot EC_{ws} \cdot EC_{ss}}{\theta_{ss} \cdot EC_{ss} + \theta_{ws} \cdot EC_{ws}} \right] + (\theta_w - \theta_{ws}) \cdot EC_{wc}
\]  
(4.7)

where, \( \theta_{ss} \) = volumetric moisture content of the surface-conductance (m\(^3\)/m\(^3\))

\( \theta_{ws} \) = volumetric soil moisture content in the soil-moisture pathway (m\(^3\)/m\(^3\))

\( \theta_w \) = total volumetric moisture content (m\(^3\)/m\(^3\))

\( EC_{ws} \) = specific electrical conductivity of the soil-moisture pathway (dS/m)

\( EC_{wc} \) = specific electrical conductivity of the continuous-liquid pathway (dS/m)

\( EC_{ss} \) = electrical conductivity of the surface-conductance (dS/m)

Rhoades et al. (1989) subsequently simplified Equation 4.7 and developed five more equations to measure \( EC_a \), quantitatively linking \( EC_a \) to soil physical and chemical properties such as moisture content (PW: percent moisture on a gravimetric basis),
bulk density ($\rho_b$ in Mg/m$^3$), saturation percentage (SP), average EC of the soil moisture assuming equilibrium (EC$_w$ in dS/m), and EC of the saturation extract (EC$_e$ in dS/m). Assuming the density of water is 1 Mg/m$^3$, the relationships are, specifically:

$$\theta_w = (PW \cdot \rho_b)/100$$

(4.8)

$$\theta_{ws} = 0.639 \theta_w + 0.011$$

(4.9)

$$\theta_{ss} = \rho_b/2.65$$

(4.10)

$$EC_{ss} = 0.019(SP) - 0.434$$

(4.11)

$$EC_w = \left[ \frac{(EC_e \times \rho_b \times SP)}{100 \times \theta_w} \right]$$

(4.12)

Equations 4.7 to 4.12 quantitatively link EC$_a$ to soil physical and chemical properties such as moisture content, soil salinity, saturation percentage, and bulk density (Corwin & Lesch 2003; Corwin & Lesch 2005a).

4.2.2 Calibrating an EMI Sensor for Soil Moisture Determination

The EMI technique (EM31 and EM38) was initially introduced for measuring and mapping soil salinity (Halvorson & Rhoades 1974; Cameron et al. 1981; Williams & Baker 1982; McNeill 1986; Wollenhaupt et al. 1986). However, a combination of ease of use and robustness has seen it extended to quantifying and mapping soil moisture content. If the soil in a field has a profile thickness that is larger than the effective depth of measurement of the EM38, then the EC$_a$ will be primarily due to variations in clay content or volumetric moisture content (McBratney et al. 2005). Brevik & Fenton (2002) found soil moisture to be the single most important edaphic factor among others (e.g., soluble salts, clay content and soil temperature) that influence EC$_a$ determination. Brevik et al. (2006) illustrated that the soil EC$_a$ can vary at any single location with changes in soil moisture content. The relationship between soil moisture content and electrical conductivity has been established by many investigators (Kachanoski et al. 1988; Sheets & Hendrickx 1995; Hanson & Kaita 1997; Khakural et al. 1998; Brevik & Fenton 2002; Reedy & Scanlon 2003; Brevik et al. 2006; Hazarjaribi & Sourell 2007).
Kachanoski et al. (1988) developed the relationship between ECₐ and moisture content, in a non-saline soil with low concentrations of dissolved electrolytes. They found that the ECₐ explained 96% of the spatial variation in the soil moisture (surface to 0.5 m depth) in a 1.8-ha study area. However, they did not find any significant relationship between ECₐ and moisture content at depths below 0.5 m.

Kachanoski et al. (1990) compared the values of moisture content determined by neutron probe with integrated ECₐ values captured by EM31 and EM38 from a 660-m transect of fine-textured moderately calcareous soil. They found that the bulk soil electrical conductivity explained more than 80% of the soil moisture content at scales ≥ 40 m. Both of the studies done by Kachanoski et al. (1988, 1990) involved calibration of the EMI sensors using ‘indirect measurements’, namely time-domain reflectometry (TDR - Kachanoski et al. 1988) and neutron probe (Kachanoski et al. 1990). It is asserted that the double-calibration process (i.e., calibrating the TDR/neutron probe to volumetric soil moisture first) introduces an unnecessary error in the evaluation of the accuracy of the EMI sensor to predict moisture content. Moreover, Kachanoski et al. (1990) obtained EMI readings from a distance of 2 m from the neutron probe access tubes and at different times to the reading of the neutron probes.

Sheets & Hendrickx (1995) performed a rigorous experiment to measure soil total moisture content of the profile with EMI technique. They installed 65 neutron probe access tubes at 30 m intervals along a 1950 m transect to measure moisture content and compared the reading with the response of an EM31 unit. They found a comparatively weaker relationship between ECₐ and moisture content than that found by Kachanoski et al. (1988) and Kachanoski et al. (1990) (r² = 0.64 compared to 0.96 and > 0.80). This was probably due to the double calibration process of EMI meter (EM31) and the mismatch of soil sampling (samples were not taken within the vicinity of EMI sensor) and effective penetration depths of the EM31 than EM38 (EM31 has a greater effective penetration depth than EM38). Furthermore, Sheets & Hendrickx (1995) took EMI measurements 10 m away from the point of neutron probe measurements.

Hanson & Kaita (1997) conducted an experiment to observe the response of moisture content in soils of three salinity levels. The R² values for the horizontal dipole
orientation were 0.81, 0.89, 0.92 for low, medium and high salinity level, respectively, and for the vertical dipole orientation were 0.76, 0.94 and 0.95. The EM38 was calibrated against neutron probe-based estimates of moisture content. Moreover, they conducted EM38 measurements 0.3 m away of the aluminium neutron probe access tubes which would be expected to distort the EMI measurements (Voltman 2000; Lamb et al. 2005).

Khakural et al. (1998) conducted a study along four transects (420 m × 10 m each) to relate ECₐ and moisture content (determined by SMNP). They found a relatively low relationship between ECₐ and moisture content with R² values of 0.71 and 0.52 for vertical and horizontal measurements respectively. Again these results are problematic owing to the proximity to the aluminum neutron probe access tubes and the use of neutron-probe calibrations in inferring the moisture content.

Reedy & Scanlon (2003) conducted a study in an artificially-engineered barrier soil profile to predict moisture content and to observe spatial and temporal variance of moisture content by EM38. Here the soil moisture values were determined using a neutron probe. They found that the EM38 could explain 80% and 99% of the moisture content variance when moisture contents were averaged vertically and spatially (i.e. across the surface of the soil volume) in all depths of the soil profile respectively. Although they found a very good relationship (R² = 0.99) between ECₐ and spatially averaged soil moisture content, again they followed the double calibration process in an artificially built soil profile that did not account for natural field variation.

Sherlock & McDonnell (2003) found a good relationship (R² = 0.70) between ECₐ and gravimetrically determined moisture content from soil samples taken at a depth of 0.2 m. The same relationship was observed by Brevik et al. (2006). Interestingly, although Brevik et al. (2006) calibrated EM38 with gravimetrically determined soil moisture content, the R² values (0.50 to 0.90 in different sites) were not as good as Kachanoski et al. (1988), Kachanoski et al. (1990), Hanson & Kaita (1997) and Reedy & Scanlon (2003) where calibration of EMI meter in all cases was with moisture contents derived from other moisture probes. The inferiority of their results may stem from the fact that Brevik et al. (2006) collected soil samples along a circle with about a 3 m radius centered at the point where EM38 readings were taken. The other possible explanation is that they paired vertical and horizontal dipole
measurements with moisture content measured at depths of 0.9 m and 0.75 m, respectively, assuming this specific depth layer accounts for the maximum instrument response. However, this relies on the assumption that there is no significantly higher moisture levels above this layer that would contribute to a loss of EM38 response (see discussion associated with Figure 4.4) for both operational modes below the depths mentioned above.

Hezarjaribi & Sourell (2007) calibrated the EM38 to predict moisture content in a non-saline loamy sand soil. They found that the total available moisture content explained 56% and 35% of the variance with temperature-corrected vertical and horizontal EC$_a$ values respectively. The poor performance of the EM38 in their study was attributed to the elevated survey (0.3 m above the ground) and depth of moisture content determination (0.6 m). It is quite possible that 14% and 43% cumulative response was lost for vertical and horizontal dipole orientations, respectively when the EM38 was elevated at 0.3 m height (Rhoades & Corwin 1981) accounting for reduced accuracy of prediction.

### 4.3 SCOPE OF THIS CHAPTER

It is clear that good relationships between volumetric soil moisture and EC$_a$ can exist. However there are a number of limitations in this previous work that include:

(i) not directly calibrating the EMI (here EM38) sensor to soil moisture measurements, but relying on the use of other instruments to infer the moisture content (with their own attendant calibration inaccuracies),

(ii) direct (or indirect) calibration using measurements not co-located with EMI sensor, and

(iii) it was never verified whether the soil/site investigated had a depth-related moisture profile that may have perturbed the natural depth-response function of the EMI sensor and hence contributed to inaccuracies observed in correlating sensor response to moisture values at depth.

Furthermore, assuming (iii) above is relevant, then none of the aforementioned work has exploited the depth-response function of the EM38 in inferring the underlying
moisture content over and above simply relying on the fact that 0-0.80 m or 0 – 1.2 m is ‘the region of maximum response’.

Consequently, this chapter has two main objectives, the first relates to the incorporation of the depth response function of the instrument to assess the instrument’s response to underlying moisture content, and the second addressing the actual calibration of the EM38 instrument to measurements of soil moisture content. To this end, the experiments in the following sections were performed in a ‘deep Vertosol’ (described in Chapter 2) to reduce the impact of both vertical and spatial (i.e. the surface of the soil volume) variability on the response of the instrument under investigation. Specifically, the objectives of this chapter were to:

1. Confirm that the EM38-derived $EC_a$ values are in fact an integration of the accepted depth-response function of the EM38 and $\theta_v$ for deep Vertosols, and

2. Compare the $EC_a - \theta_v$ relationship using direct $\theta_v$ measurements derived from soil core and pit-based methods.

This chapter is therefore divided into two specific sections. The first part deals with collecting multi-height EM38 measurements with a view to addressing objective 1, and the second part addresses objective 2.

4.4 MULTI-HEIGHT EM38 MEASUREMENTS

Combining the EM38 depth response function with depth-specific measurements of the volumetric moisture content should give a more accurate correlation with the instrument response (i.e. measured $EC_a$) if the dominant driver of local electrical conductivity (EC) is moisture (as discussed in Section 4.2). Furthermore, raising the EM38 above the ground, when collecting measurements, is equivalent to ‘shunting’ the volumetric moisture content profile lower down the EM38 depth response function. Table 4.1 illustrates the link between depth-specific volumetric moisture content ($\theta_v$), the sensor height above ground, and the measured integrated sensor response.

The key underlying assumption is that “local EC at depth, $EC(z)$ is directly proportional to the volumetric moisture content at depth, $\theta_v(z)$” and this assumption is
reasonable in a situation where the concentration of soluble salts is constant with depth. *In situ* soil electrical conductivity is dependent on the concentration of charge carriers (i.e. dissolved ions from salts) as well as the size and connectivity of water filled soil pores which provide pathways for electrical current. As a dry soil wets up, progressively larger pores become water filled and connections between water filled pores increase, so that soil moisture content is the primary control on electrical conductivity. Once the soil is close to saturation and all but the largest pores have filled, further water addition has only a small effect on electrical conductivity and in this situation the EM38 response is primarily dependent on soil salinity.

If the depth-response function of the EM38 is not perturbed by the vertical $\theta_v$ profile, and the assumption of McNeill (1980) that horizontal eddy currents induced by the primary field do not influence each other (discussed earlier in Section 4.2), the instruments response should be proportional to the addition of the combined depth response function and $\theta_v$ according to

$$EC_{a(measured)} \approx \sum z \times k \times \theta_v(z) \times \phi^{V,H}(z)$$  (4.13)

where, $EC_{a(measured)}$ is the instrument (integrated) response in mS/m, $k$ is the constant of proportionality between local EC at depth and $\theta_v$ at the same depth, assuming the relative contributions of all other EC-driving parameters remain fixed at depth, and $\phi^{V,H}(z)$ is the depth response function for either vertical or horizontal dipole orientation (Equations 4.4 and 4.5).

Assuming the relationship between local EC and $\theta_v$ does not change with depth, then

$$EC_{a(measured)} \approx k \sum z \times \theta_v(z) \times \phi^{V,H}(z)$$  (4.14)

which for brevity will be reduced to

$$EC_{a(measured)} \approx k \times EC_{a\theta}$$  (4.15)

where,

$$EC_{a\theta} = \sum z \times \theta_v(z) \times \phi^{V,H}(z)$$  (4.16)
Table 4.1 The data combination process for linking depth-specific $\theta_v$, the depth response function ($\phi^V(z)$ for vertical dipole orientation and $\phi^H(z)$ for horizontal dipole orientation) and the integrated sensor response $EC_{a0}$ with position of sensor at varying heights above ground. Note $\theta_v$ is limited to 1.2 m soil depth.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>$\phi^V(z)$ or $\phi^H(z)$ X</th>
<th>Sensor Height (m)</th>
<th>0</th>
<th>0.2</th>
<th>0.4</th>
<th>0.6</th>
<th>0.8</th>
<th>1.0</th>
<th>1.2</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>0</td>
<td>2</td>
<td>$\theta_{0.025}$</td>
<td></td>
<td></td>
<td></td>
<td></td>
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4.4.1 Materials and Methods (Multi-height EM38)

The multi-height EM38 measurements were performed on the field site (Chapter 2) using both the single core and the pit methods to determine volumetric moisture content (describe earlier in Section 3.4). Measurements were performed over 17 core sites and 8 pit sites. To facilitate multi-height EM38 measurement, a
specially-designed polymer-plastic ‘ladder’ was used (Figure 4.7). The ladder was constructed from lengths of 42 mm diameter polyvinyl chloride (PVC) pipe segmented at heights corresponding to the ground surface (0 m), 0.2, 0.4, 0.6, 0.8, 1.0 and 1.2 m. The EM38 measurements were collected in both vertical and horizontal dipole configurations with the sensor axis (that is, a line between the sensor and receiver coils) orientated east-west. All subsequent EM38 measurements were taken from 7 heights (up to 1.2 m). Prior to each measurement the EM38 was nulled according to the protocol of Geonics Limited (2003).

Figure 4.7. Polymer ladder device for multi-height EM38 measurements (note- EM38 in the horizontal dipole configuration).

For the soil core sites, EM38 measurements were performed after the extraction of the soil cores. As the core sites were subsequently used for neutron-probe evaluations of \( \theta_v \), the Sentek PVC access tubes were inserted in them. A check of the integrated EM38 response (i.e. on-ground level) before removal of the cores and following removal and insertion of the access tubes showed no discernible changes in the instrument readings.

When the pit calibration process was used, the multi-height measurements were conducted immediately before excavation of the pits.
The soil cores were also tested (at each depth strata) for electrical conductivity of soil suspension (EC$_{1:5}$) as it indicates the concentration of dissolved salts, using the methods outlined by Rayment & Higginson (1999).

### 4.4.2 Results and Discussion (Multi-height EM38)

The EC$_{1:5}$ in soil with depth was found to be, within the uncertainty associated with measurement, constant to a depth of 1.2 m throughout the study area which indicated that the level of dissolved salts at depths is constant to a depth of 1.2 m for this soil. An example of the distribution of EC$_{1:5}$ at depths from two core sites is presented in Figure 4.8. EC typically increases down a soil toposequence as soluble salts are leached to the bottom of a hillslope.

![Figure 4.8 Distribution of EC$_{1:5}$](image)

Figure 4.8 Distribution of EC$_{1:5}$ (indication of level of dissolved salts) in soil at different depths for two sites of the study area. ‘Top Slope’ and ‘Bottom Slope’ refer to the relative locations in the field site. Vertical bars are representing 20% error of the data.

A plot of EC$_{\text{a(measured)}}$ versus predicted EC$_{\text{a}}$ (denoted as EC$_{\text{a(0)}}$) at different sensor heights above the ground for a single core site is given in Figure 4.9. The data include both horizontal and vertical dipole orientations. Not surprisingly, the integrated EM38 response (measured in mS/m) decreases monotonically as the sensor is progressively raised above the ground. In Figure 4.9 progressively raising the sensor...
above the ground corresponds to the data points (one for each dipole orientation) starting at the top right and finishing at the lower left in sequence. The measured $EC_a$ values explained 99% & 97% of the variance observed in the $EC_{a0}$ values with root-mean-square error (RMSE) values of 1.38 mS/m and 2.57 mS/m for horizontal and vertical dipoles, respectively. The $R^2$ and RMSE values were typical of all 17 test-core locations investigated, as was the sequential trend in data points with increasing sensor height. The entire dataset of $EC_{a(\text{measured})}$ values, for every sensor height at the 17 core locations, was plotted against respective $EC_{a0}$ values in Figure 4.10. The coefficients of determinations ($R^2$) for entire data sets (17 cores and 7 heights each) were 0.87 and 0.83 with RMSE of 6.01 mS/m and 8.06 mS/m for horizontal and vertical dipoles, respectively. Similarly, the combined $EC_{a(\text{measured})}$ versus $EC_{a0}$ for the pit sites is given in Figure 4.11.

![ECa(measured) versus ECaθ values for vertical and horizontal dipole configurations for a single core site. Sensor heights sequentially progress from 0 (ground level- top right data point) through to maximum height (1.2 m - lowest left data point) in each plot](image)

Figure 4.9 $EC_{a(\text{measured})}$ versus $EC_{a0}$ values for vertical and horizontal dipole configurations for a single core site. Sensor heights sequentially progress from 0 (ground level- top right data point) through to maximum height (1.2 m - lowest left data point) in each plot.
**Figure 4.10** EC\textsubscript{a(measured)} versus EC\textsubscript{aθ} values for vertical and horizontal dipole configurations for all core sites.

**Figure 4.11** EC\textsubscript{a(measured)} versus EC\textsubscript{aθ} values for vertical and horizontal dipole configurations for all pit sites.
The pit calibration site data yielded an increase in $R^2$ which is not surprising given the improved precision in measuring the $\theta_v$ (discussed earlier in Section 3.5). In all analyses, the response of $EC_{a\theta}$ to $EC_{a(measured)}$ was statistically highly significant ($P<0.0001$).

A key outcome of Figures 4.9 – 4.11 is that the plots are all effectively linear. This supports the underlying assumption that the relative contributions of all $EC_a$-driving parameters at depth (for example, salinity, moisture content) remain the same. Any small, systematic deviation from linearity is likely to be attributed to small variations in the depth-related distribution of soluble salts in the profile (Figure 4.8). Another likely determinant of the relationship between $EC_a$ and $EC_{a\theta}$ is the soil pore size distribution. Electrical conductivity through the liquid phase pathway is dependent on continuous chains of water filled pores. The maximum radius of water filled pores (assumed to be cylindrical capillaries) is inversely proportional to the metric potential, which in turn is a function of moisture content. For a given moisture content a soil with more pores small enough to be filled will have a higher EC than one with more unfilled pores. Moreover, the layered earth model whereby the contribution of each layer at depth adds linearly is also verified, as is the fact that the depth-response function of the EM38 is not perturbed by the depth profile of the moisture (or ion) content. The significance of the constant $k$ is also evident. It points to the direct link between local EC, $EC(z)$ and $\theta_v$ at depth, $\theta_v(z)$. For the pit sites, for example, the values of $k$ (unitless) are $31.74 \pm 1.60$ and $49.30 \pm 2.69$ for horizontal and vertical dipole orientations, respectively. The physical significance of $k$ lies in the connection between the local $\theta_v$ and electrical conductivity at depth. In the absence of any dissolved electrolytes, water has an intrinsic electrical conductivity of $5.5 \mu S/m$ (e.g., Marshall 1987) and this generally increases linearly with increasing concentrations of ions (e.g., Lide 2007). The slope of the conductivity-concentration curves for water varies with the specific acid, base or salt. Thus, $k$ must represents the contribution of dissolved ions in water as well as the soil-water and soil-soil interfacial characteristics to the electrical conductivity. The linearity in Figure 4.11 suggest these contributing factors to be relatively consistent throughout the deep Vertosol soil profiles investigated here within the ‘penetration range’ of the EM38.
The actual magnitude of $k$ is found to vary depending on the dipole orientation of the EM38 however, the ratio of the $k$-values for horizontal and vertical dipole configurations is approximately 0.6 (core and pit samples). This value is similar to the ratio of the horizontal to vertical integrated response of the EM38, calculated by integrating Equations 4.4 and 4.5 with respect to depth ($z$), for depths exceeding 0.5 m (refer to Figure 5.1, Chapter 5). This assertion is also supported by Corwin and Rhoades (1982). Thus, $k$ is attributed to a combination of the soil conductivity characteristics described above and the relative response functions of the EM38 in the specific dipole configuration to which it is employed.

The multi-height EM38 measurements confirm the veracity of the underlying assumptions related to the integrated response of the EM38 sensor over the deep Vertosol soils at this field site. It is therefore, possible to consider a field calibration process that involves using the on-ground, integrated response of the EM38 sensor to underlying moisture content.

### 4.5 FIELD CALIBRATION OF ON-GROUND EM38 MEASUREMENTS

As discussed in Section 4.2, the known depth-response function of the EM38 has been used by previous researchers as the basis to calibrate the sensor to underlying ‘average’ EC (i.e. below the sensor) over specific depth ranges, even though the actual depth-response function of the sensor was never checked for perturbations. Previous research has employed depth ranges in the vicinity of 0-0.8 m (e.g., Rhoades & Corwin 1981; Amezketa 2006) and 0 – 1.2 m (e.g., Rhoades & Corwin 1981; Corwin & Rhoades 1982, 1984). Given that the earlier work in Section 4.4 supports the notion that the depth response function of the EM38 is not perturbed by the soil moisture profile for our deep Vertosol, integrating the depth response functions of the EM38 (Equations 4.4 and 4.5) over these depth ranges indicates that for the vertical dipole orientation, 47% and 62% of the integrated sensor response originates from the soil between the surface and 0.8 and 1.2 m, respectively. Similarly in the horizontal dipole orientation, 71% and 79% of the integrated sensor response originates from the soil down to these depths. A useful comparison with previous work is therefore to repeat the field calibration process to these depths.
4.5.1 Materials and Methods (Field Calibration)

For the field calibration work the EM38 sensor was placed on the ground surface, again with the sensor itself orientated east-west, and each set of measurements were preceded by the standard nulling procedure (Geonics Limited 2003). Again, data derived from core and pit calibration approach were used to calibrate the EM38 response to underlying $\theta_v$.

The core calibration sites comprised of 17 locations distributed around the field site. For the first calibration dataset, EM38 measurements were conducted prior to extraction of the cores and subsequent determination of $\theta_v$ following the procedure described in Section 3.4.1. The average $\theta_v$ values over the depth range 0 – 0.8 m ($\bar{\theta}_{0.8}$) and 0 – 1.2 m ($\bar{\theta}_{1.2}$) were determined using the derived depth-profiles. The pit calibration sites comprised two ‘wet’ and two ‘dry’ sites, as described in Section 3.4.2. Similarly with the pit method, a calibration dataset was first created by conducting an EM38 measurement prior to the excavation of the pits. The field calibration equations relating EM38 response to $\bar{\theta}_{0.8}$ and $\bar{\theta}_{1.2}$ (both from the single core and the pit methods) were validated on 6 subsequent revisits to the randomly selected 8 core sites from an option of 14 (throughout 2006 and 2007). As the cores from these sites had already been extracted during the core-calibration step, the $\theta_v$ depth profiles were determined using the neutron probe calibrated earlier in Section 3.5.2.

Calibration equations (core and pit) were derived using simple linear regression within JMP statistical software (SAS 2005).

The pit calibration equations were evaluated against the validation datasets using residual plots (plotting the residuals generated from the regression analysis between predicted and measured $\theta_v$, and measured $\theta_v$), calculating the ‘root mean squared error’ (RMSE) and conducting a ‘lack-of-fit’ test. The RMSE measures the accuracy of the prediction described in Section 3.4.3. The ‘lack-of-fit’ test is used to check the adequacy of the model applied to repeated measures of independent variables (measured $\theta_v$). Owing to the inherent inaccuracy of determining soil moisture content, $\theta_v$ data were rounded down to 2-decimal points (Gunst & Mason 1980) prior to conducting the ‘lack-of-fit’ test using JMP statistical software (SAS 2005).
4.5.2 Results and Discussion (Field Calibration)

Single Core Calibration

An example of the \( \theta_v \) depth profiles used in the calibration of the EM38 responses for the single core sites is given in Figure 4.12. The distribution of \( \theta_v \) at depths of the two core sites show a trend with lower \( \theta_v \) in the top 0.4 m depth and increasing \( \theta_v \) down the profile. A similar pattern was observed through the study area. The variability of \( \theta_v \) over the study area at two depth groups (\( \bar{\theta}_{0.8} \) and \( \bar{\theta}_{1.2} \)) is not significantly different, however, the variability of \( \bar{\theta}_{0.8} \) is slightly higher (CV = 12.99%) than that of \( \bar{\theta}_{1.2} \) (CV = 11.99%).

![Figure 4.12 Depth profiles of \( \theta_v \) for one core site each of top and bottom slope and the mean \( \theta_v \) at depths for 17 core sites of the study area.](image)

Scatter-plots of \( \bar{\theta}_{0.8} \) and \( \bar{\theta}_{1.2} \) versus EC\(_a\) are given for both horizontal and vertical dipole orientations in Figure 4.13. The calibration equations derived from the single core method, and associated statistics, for both horizontal and vertical dipole orientations are summarized in Table 4.2.
Figure 4.13 Scatter-plot of average volumetric moisture content at (a) 0-0.8 m depth ($\bar{\theta}_{0.8}$) and (b) 0-1.2 m depth ($\bar{\theta}_{1.2}$) derived from the single core calibration method versus EC$_a$ measured by EM38 in both vertical (EC$_a$-V) and horizontal (EC$_a$-H) dipole orientations.

The calibration equations summarized in Table 4.2 are all statistically significant (P < 0.01), however, the R$^2$ values are lower for the vertical dipole orientation, and the lowest R$^2$ value is derived using $\theta_i$ data integrated over the shallower depth ($\bar{\theta}_{0.8}$). This result is not surprising since only 47% of the integrated sensor response is estimated to originate from the soil between the surface and 0.8 m depth and 62%
from between the surface and 1.2 m depth (Section 4.2). Similarly in horizontal dipole orientation, a larger component, namely 71% and 79% of the integrated sensor response originates from the soil down to these respective depths. However, the inaccuracy is further compounded by the findings of Chapter 3 where the single core method of deriving $\theta_v$ is itself relatively inaccurate.

Table 4.2 Derived linear calibration equations for prediction of $\theta_v$ by EM38 using the single core calibration data where $\bar{\theta}_{0.8}$, $\bar{\theta}_{1.2}$ are average $\theta_v$ at 0.8 and 1.2 m depths respectively, $EC_{a\cdot H}$ and $EC_{a\cdot V}$ are on-ground EM38 measurements in horizontal and vertical dipole modes respectively and N is the number of observation

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Figure 4.14 Depth profiles of replicated average $\theta_v$ at 1.2 m depth for the 4 pit sites. The pit ‘Wet plot 1’ is considered an outlier here as it shows poor ‘wetting down’ following application of surface water (discussed earlier in Section 3.4.2).
Pit Calibration

An example of the $\theta_v$ depth profiles used in the calibration of the EM38 responses for the pit sites is given in Figure 4.14. Scatter-plots of $\bar{\theta}_{0.8}$ and $\bar{\theta}_{1.2}$ versus EC$_a$ for both horizontal and vertical dipole orientations for all four pits are given in Figure 4.15.

Figure 4.15 Scatter-plot of average $\theta_v$ at (a) 0-0.8 m depth ($\bar{\theta}_{0.8}$) and (b) 0-1.2 m depth ($\bar{\theta}_{1.2}$) derived from the pit calibration method versus EC$_a$ measured by EM38 in both vertical (EC$_a$-V) and horizontal (EC$_a$-H) dipole orientations. Data were taken from 4 pits.
In Figure 4.15 the atypical behaviour of ‘wet pit 1’ is evident in the EM38 responses (circled in red). As discussed earlier, the $\theta_v$ profiles for this particular pit (Figure 4.14) indicate that it is significantly drier at lower depths. In horizontal dipole orientation the EM38 response is dominated by the $\theta_v$ closer to the surface with approximately 29% of the integrated response originating below 0.8 m. The drier profile at depth does not overly affect the horizontal dipole EM38 response when the $\theta_v$ is averaged down to 1.2 m. There is a small ‘under-response’ (i.e. compared to its expected response if the particular pit had a wetting profile similar to the other 3 pits) of the EM38 response in horizontal mode when the $\theta_v$ is averaged down to 0.8 m. This under-response translates to shifting of the EC$_a$ points to the left in Figure 4.15. However, for the vertical dipole orientation more than 50% (i.e. approximately 53%) of the integrated response originates below 0.8 m. and so the under-response is more evident in the vertical dipole data and the displacement of the EM38 data is largest in the data where the $\theta_v$ is averaged down to only 0.8 m.

The combination of wet soil overlying relatively dry soil below 0.8 m in wet pit 1 (Figure 4.14) was an artifice resulting from incomplete redistribution of the large amount of water applied to the surface over a short period of time during the wetting up procedure. No similar extreme contrasts at depth were found over the course of the project at the access tubes monitoring rainfed soil moisture profiles.

Removing the data from this pit on the basis of its atypical $\theta_v$ profile significantly improves the linearity of the EM38 response (Figure 4.16).

There is a considerable improvement in the $R^2$ and RMSE values associated with using the pit calibration method compared to the single core method (Table 4.2). Model regression equations and associated regression statistics are summarized in Table 4.3.

The calibration equations summarized in Table 4.3 are, again, all statistically significant ($P < 0.0001$). The $R^2$ value is slightly lower for the vertical dipole orientation when using $\theta_v$ data integrated over the shallower depth ($\bar{\theta}_{0.8}$) although this is not likely to be a significant difference. Again, this is the result of the smaller component of the sensor’s integrated response coming from this shallow soil section.
Figure 4.16 Scatter-plot of average $\bar{\theta}$, at (a) 0-0.8 m depth ($\bar{\theta}_{0.8}$) and (b) 0-1.2 m depth ($\bar{\theta}_{1.2}$) derived from the pit calibration method versus EC$_a$ measured by EM38 in both vertical (EC$_{a-V}$) and horizontal (EC$_{a-H}$) dipole orientations. Data were taken from 3 pits.
Table 4.3 Derived linear calibration equations for the prediction of \( \theta_v \) using the pit calibration data where \( \bar{\theta}_{0.8}, \bar{\theta}_{1.2} \) are average \( \theta_v \) at 0.8 and 1.2 m depths respectively, \( EC_{a-H} \) and \( EC_{a-V} \) are on-ground EM38 measurements in horizontal and vertical dipole modes respectively and \( N \) is the number of observation.

<table>
<thead>
<tr>
<th>EM38 Dipole Orientation</th>
<th>Regression Equation</th>
<th>( R^2 )</th>
<th>RMSE (( m^3/m^3 ))</th>
<th>P-value</th>
<th>( N )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal</td>
<td>( \bar{\theta}_{0.8} = 0.003EC_a + 0.273 )</td>
<td>0.99</td>
<td>0.009</td>
<td>&lt;0.0001</td>
<td>18</td>
</tr>
<tr>
<td></td>
<td>( \bar{\theta}_{1.2} = 0.002EC_a + 0.318 )</td>
<td>0.99</td>
<td>0.010</td>
<td>&lt;0.0001</td>
<td>18</td>
</tr>
<tr>
<td>Vertical</td>
<td>( \bar{\theta}_{0.8} = 0.004EC_a + 0.071 )</td>
<td>0.98</td>
<td>0.014</td>
<td>&lt;0.0001</td>
<td>18</td>
</tr>
<tr>
<td></td>
<td>( \bar{\theta}_{1.2} = 0.003EC_a + 0.150 )</td>
<td>0.99</td>
<td>0.007</td>
<td>&lt;0.0001</td>
<td>18</td>
</tr>
</tbody>
</table>

Model Validation Using Pit Data

Owing to the superior calibration produced using the pit data, the linear regression models of Table 4.3 were validated against the measurements from eight randomly selected core sites revisited 6 times during 2006 and 2007. The results are summarized in Figure 4.17. Here the \( \theta_v \) values predicted using the regression models are directly compared to the actual \( \theta_v \) values determined using multi-depth neutron probe measurements and the neutron probe calibration equations derived earlier in Section 3.5.2.
Figure 4.17 Scatter-plot of predicted average volumetric moisture content using EM38 measurements and the regression equations of Table 4.3 with the actual average volumetric moisture content, $\theta_v$ derived using neutron probe measurements. Data for (a) 0-0.8 m depth ($\theta_{0.8}$) and (b) 0-1.2 m depth ($\theta_{1.2}$). The 1:1 line (grey dotted line) on each graph indicates the region where predictions would agree with measurements.

It is evident in Figure 4.17 that the predicted average $\theta_v$ is more closely in agreement with the actual average $\theta_v$ when the values are averaged over the larger depth (i.e. down to 1.2 m) and that the closest agreement occurs when using the horizontal dipole orientation, again related to the fact that the integrated response is more accurately reflected within this larger depth of soil. The RMSE of prediction, RMSEP (RMSE for validation data sets) of 0.044 m$^3$/m$^3$ achieved with the horizontal dipole orientation and averaging the volumetric moisture content down to 1.2 m (Figure 4.17(b)) represents an error of only 9% considering the range of moisture contents encountered in the field site. By comparison, for the vertical orientation validation dataset (also Figure 4.17(b)) it can be seen that 12.5% prediction error was encountered with these measurements.

Whilst not surprising that the graph of Figure 4.17(a) reflects a poorer representation of the EM38 response by the shallower depth of $\theta_v$ measurement it is interesting that
the predicted values underestimate the measured values for the lower $\theta_v$ and overestimates for the higher $\theta_v$.

Although there are mixtures of underestimates and overestimates of $\theta_v$ with these models, all models to predict $\theta_v$ at 0-0.8 m and 0-1.2 m depths by EC$_a$-H and EC$_a$-V are evaluated with the residual plot analysis. The residual analysis depicted that there is no relation between residuals and measured $\theta_v$ ($R^2 = 0$ and $P > 1.0$) for all models and the residuals were normally distributed (Skewness ranges from 0.07 to -0.72 and Kurtosis ranges from -0.51 to -0.73). Finally a lack-of-fit test was performed to test the adequacy of the models. The lack-of-fit test was found statistically non significant in any models ($P > 0.729, 0.730, 0.725, 0.713$ for $\theta_{1.2-H}, \theta_{1.2-V}, \theta_{0.8-H}, \theta_{0.8-V}$ respectively) which confirmed that there was no lack-of-fit of the fitted models indicated that the fitted models were adequate to predict $\theta_v$.

4.6 CONCLUSION

When calibrating the EM38 to undertake direct estimates of underlying $\theta_v$, a pit method is recommended owing to the improved accuracy in determining the actual $\theta_v$ values. In the deep Vertosol soil of this fieldsite, the relationship between measured EC$_a$ and that predicted using a simple function combining the depth response function of the EM38 and the volumetric moisture content was observed to be linear, confirming that the EM38 depth response function was not perturbed by the local $\theta_v$ profile and that the basic drivers of EM38 response other than moisture content (i.e. ion content, soil-soil and soil-water interfacial processes) remained consistent with depth and across the sampled locations within the fieldsite. The constant of proportionality between the measured and predicted values of EC$_a$, $k$, was also found to be linked to the integrated depth response function of the EM38, hence is determined by the dipole orientation employed when using the EM38 to conducting $\theta_v$ measurements.

When used as a method of estimating $\theta_v$ from surface-only measurements of EC$_a$, the EM38 was found to provide more accurate measurements (~9% error) when compared to average $\theta_v$ measured down to 1.2 m depth and when used in horizontal dipole configuration. However, the results of this chapter also indicate the inherent uncertainties in making estimates of ‘average’ $\theta_v$ and the significance of the
interaction between the well-known (and now considered reliable) EM38 depth-response function and an unknown depth profile of $\theta_v$. The obvious step is to now examine methods of extracting out the depth profile of $\theta_v$ from the EC$_a$ values rather than simply estimating the $\bar{\theta}_{1,2}$ and this is the subject of the next chapter.
CHAPTER – FIVE

PREDICTING MULTI-DEPTH SOIL MOISTURE BY SURFACE AND MULTI-HEIGHT EM38 MEASUREMENTS
5.1 INTRODUCTION AND SCOPE OF THIS CHAPTER

The results of Chapter 4 established the stability of the depth response function of the EM38 when used to determine the volumetric moisture content ($\theta_v$) in the deep Vertosol soils of our field site. On the basis that $\theta_v$ remains the dominant factor in the response of the EM38 (assuming ions etc are all mobilized by $\theta_v$), and that the depth response function is stable and known, then the depth response function of the EM38 could, in principle be used to extract the depth profile of $\theta_v$. Consequently, this chapter will discuss the underlying theory behind processing EM38 measurements for creating depth-profiles of $\theta_v$ and evaluate a number of processes, previously published in the literature ostensibly to determine depth EC$_a$ profiles, for converting above-ground EM38 measurements to depth profiles of $\theta_v$.

5.2 EM38 AND DEPTH PROFILING

5.2.1 EM38 Depth Response Function Revisited

The principles of operations and the depth response function of EM38 have been previously discussed (Section 4.2). The relative contribution of the response of the EMI sensor at each depth is given in Equations 4.4 and 4.5.

Since the EC$_a$ measured by the EM38 on the ground in either dipole configuration is a reflection of the integrated depth-response of the EM38, the cumulative response curve of both dipole orientations of the instrument can be determined by integrating Equations 4.4 and 4.5 with respect to depth ($z$). The cumulative response function for vertical and horizontal dipole configurations are therefore given by

$$R^V(z) = \frac{1}{(4z^2 + 1)^{1/2}}$$  \hspace{1cm} (5.1)

and

$$R^H(z) = (4z^2 + 1)^{1/2} - 2z$$  \hspace{1cm} (5.2)

where, $R^V(z)$ and $R^H(z)$ are cumulative response of the EMI instrument in depths for vertical and horizontal orientations respectively. These functions are plotted in Figure 5.1.
Recalling the key assumption of McNeill (1980), that the medium below the sensor is homogenous and of low induction number ($N_B << 1$), the secondary magnetic field is a very simple (and linear) function of soil electrical conductivity (Equation 4.1). Also, the linear model developed by McNeill (1980) was based on the assumption that the current flow within the horizontally stratified medium is entirely horizontal. Under these assumptions McNeill (1980) suggested that the sub-surface soil information at discrete depths could be determined by conducting measurements with the instrument at different heights. Borchers et al. (1997) discussed and improved the initial model, suggesting that if the instrument is held at a given height $h$ above the surface, the apparent conductivity reading in both vertical and horizontal dipole configurations takes the form

Figure 5.1 Cumulative responses of EM38 of all soil electrical conductivity at different depths for the vertical (—) and horizontal (---) dipole configurations. Curves calculated from McNeill (1980).
\[ EC_{a-V,H} = \int_0^\infty \phi^{V,H}(z + h) \, EC(z) \, dz \] (5.3)

where, \( EC_{a-V,H} \) is the apparent electrical conductivity measured by the EMI instrument, \( h \) represents the height of the instrument placed above the ground, \( EC(z) \) is the conductivity at depth \( z \) and \( \phi^{V,H} \) are, respectively the relative contributions of the sensitivity function of the vertical and horizontal the instrument in vertical and horizontal dipole configurations (Equations 4.4 and 4.5).

### 5.2.2 Depth Profiles Using ‘Forward Propagation’ Models

Numerous workers have utilized the depth response function of the EM38 in conjunction with multiple-height \( EC_a \) measurements and/or a combination of horizontal and vertical dipole configurations in order to stratify the soil electrical conductivity at depth. Rhoades & Corwin (1981), Corwin & Rhoades (1982, 1984), Wollenhaupt et al. (1986), Slavich (1990) and Cook & Walker (1992) all used empirical linear models to convert their measurements into \( EC_a \) depth profiles in what can collectively be termed ‘forward propagation’ (FP) models. A ‘multiple regression coefficient model’ was developed by Rhoades & Corwin (1981) to predict the \( EC_a \) in 0.3 m steps up to a depth of 1.2 m. In their study they regressed multi-height EMI measurements with \( EC_a \) measured by a four-electrode probe at different depths. Their model was based on the following equations.

\[
EC_{0-0.3} = \alpha_1 EC_a1 + \alpha_2 EC_a2 + \alpha_3 EC_a3 + \alpha_4 EC_a4
\] (5.4)

\[
EC_{0.3-0.6} = \alpha_1 EC_a1 + \alpha_2 EC_a2 + \alpha_3 EC_a3 + \alpha_4 EC_a4
\] (5.5)

\[
EC_{0.6-0.9} = \alpha_1 EC_a1 + \alpha_2 EC_a2 + \alpha_3 EC_a3 + \alpha_4 EC_a4
\] (5.6)

\[
EC_{0.9-1.2} = \alpha_1 EC_a1 + \alpha_2 EC_a2 + \alpha_3 EC_a3 + \alpha_4 EC_a4
\] (5.7)

where subscripts 0-0.3, 0.3-0.6, 0.6-0.9, 0.9-1.2 represent the electrical conductivity of respective depths in metres, \( EC_{a1}, EC_{a2}, EC_{a3} \) and \( EC_{a4} \) represent the apparent electrical conductivity measured by the EM38 at 0, 0.3, 0.6, 0.9 and 1.2 m height above the surface and \( \alpha_1, \alpha_2, \alpha_3, \alpha_4 \) are the regression coefficients. The regression equations allowed them to reconstruct the \( EC_a \) depth profile from multi-height EMI
measurement in both horizontal and vertical dipole configurations, yielding a very high precision of measurement ($R^2 = 0.99$) for all depth groups.

Corwin & Rhoades (1982, 1984) were able to combine both horizontal and vertical dipole configuration measurements into a single model by adjusting the $EC_a$ values according to the relative sensing volumes of horizontal versus vertical dipole configurations. In this ‘established coefficient model’ they were able to predict the $EC_a$ values of both composite and successive depth intervals from 0.3 m down to 1.2 m from surface-only measurements. Their study also included a comparison of the ‘established coefficient model’ with the ‘multiple regression coefficient model’ of Rhoades & Corwin (1981). It was found that the ‘multiple regression coefficient model’ gave higher $R^2$ values (0.98 compared to 0.57 to 0.97). However, a modified version of the ‘established coefficient model’ described by Corwin & Rhoades (1984), subsequently improved the $R^2$ values ($R^2 = 0.99$ in most cases) over that of their original model. Rhoades et al. (1989) further modified the ‘established coefficient model’ using a statistical procedure to predict $EC_a$ for similar composite and successive depth intervals where their $R^2$ values of their study ranged from 0.90 to 0.99 in different soils. Slavich (1990) developed a ‘modeled coefficient approach’ for predicting $EC_a$ of composite depth profile of 0.05 m intervals. In this model Slavich (1990) established the relationship between a simulated $EC_a$ profile and calculated $EC_{a-V}$ and $EC_{a-H}$ readings. The simulated $EC_a$ profile was generated from different mean $EC_a$ values to create possible field $EC_a$ profiles of 0.05 m interval using a simulation process involving a cubic spline interpolation method. Calculated $EC_{a-V}$ and $EC_{a-H}$ were derived using following equations.

$$EC_{a-V} = \sum_{i=1}^{N_v} ECA_i (R^V_i - R^V_{(i-1)})$$ \hspace{1cm} (5. 8)

$$EC_{a-H} = \sum_{i=1}^{N_h} ECA_i (R^H_i - R^H_{(i-1)})$$ \hspace{1cm} (5. 9)

where $N_v$ and $N_h$ are number of layers to measurement depths in vertical and horizontal dipole configurations respectively, $ECA_i$ is the mean $EC_a$ value of the synthetic profile of the particular soil segment in $i$th depth layer and $R^V_i$ & $R^H_i$ are
the vertical and horizontal cumulative depth-response function for the \(i\)th depth layer (Figure 5.1). The model for reconstructing the \(E_C\) profile is then

\[
E_{C_i(0-z)} = \alpha_1 E_{C_i-V} + \alpha_2 E_{C_i-H} + c \tag{5.10}
\]

where \(E_{C_i(0-z)}\) is the electrical conductivity of the particular depth, \(E_{C_i-V}\) and \(E_{C_i-H}\) are the vertical and horizontal electrical conductivity calculated following Equations (5.8) and (5.9) and \(\alpha_1, \alpha_2\) and \(c\) are the regression coefficients.

The above-mentioned studies were limited to specific sample sites. Cook & Walker (1992) developed a ‘mathematical coefficient model’ using a least square minimization technique to determine the \(E_C\) for specific depths and claimed the model as non site-specific. They assumed that the previous models for \(E_C\) depth profiles were based on the linear regressions where the minimization of measurement-system response outside the depth interval was ignored. They added a damping parameter \((\lambda)\) in their model to minimize the sum of the response outside the desired interval and sum of the coefficients. However, comparison of this model with those cited beforehand showed mixed results in terms of prediction accuracies.

### 5.2.3 Depth Profiles Using ‘Inversion Processes’

The multi-height response function (Equation 5.3) also allows prediction of the vertical distribution of electrical conductivity of the soil profile \(E_C(z)\) by following a Simple Least Squares’ inversion procedure. For a stratified medium model (Figure 5.2) it is assumed that soil has been divided into \(M\) layers \((z_1, z_2, \ldots, z_M)\) with specified thickness \(\Delta z\), volumetric moisture content, \(\theta_{vj}\), electrical conductivity \(E_{C_j}\) and magnetic permeability \(\mu_j\). It is also assumed that the volumetric moisture content, conductivity and magnetic permeability are uniform in each specific layer. For this linear model it is again assumed that the magnetic permeability of the soil is equal to that of free space: \(\mu_j = \mu_0\) where \(j = 1, 2, \ldots, M\).

The diagram of Figure 5.2 is consistent with that outlined earlier in Chapter 4 (Table 4.1, Equation 4.16) when describing the integrated response of the EM38 in response to a layered \(\theta_v\).
Figure 5.2 Schematic diagram of the M-layered stratified subsurface model. Tx and Rx represent the transmitter and receiver coil of the EM38 respectively, $\Delta z$ is the specified thickness of the layer; $\theta$, $\mu$ and EC are the volumetric moisture content, magnetic permeability and apparent electrical conductivity respectively; $M$ is the number of layers and $\text{EC}_{a-V}$ or $\text{H}(h)$ is the measured EM38 response at height $h$.

In the situation described here in Figure 5.2 the column vector $\Theta$ (vectors and matrices are denoted with bold case letters in the text) contains the local volumetric moisture content of each layer and the column vector $\text{EC}_a$ comprises the apparent conductivities measured by the EM38 at different heights above the ground for both vertical and horizontal dipole configurations, as follows

$$\Theta = \begin{bmatrix} \theta_{v0} \\ \theta_{v1} \\ \ldots \\ \theta_{vM} \end{bmatrix}$$

(5.11)
where, for convenience, Equation 5.12 represents both the horizontal and vertical dipole configurations. The system of linear equations connecting the volumetric moisture content profile of the soil and the apparent conductivity measurements of the EM38 is established following Section 4.4 by incorporating the instrument response function using Equation (5.3) as a \((n \times m)\) matrix \(\vartheta\) according to

\[
\begin{align*}
EC_a &= \left[ \begin{array}{c}
EC_{\alpha-V}(h_1) \\
EC_{\alpha-V}(h_2) \\
\vdots \\
EC_{\alpha-V}(h_n) \\
EC_{\alpha-H}(h_1) \\
EC_{\alpha-H}(h_2) \\
\vdots \\
EC_{\alpha-H}(h_n)
\end{array} \right] \\
\text{(5.12)}
\end{align*}
\]

where \(EC_a\) is the vector defining the volumetric moisture content-predicted \(EC_a\). In keeping with Equation 5.12, the vector \(\vartheta\) can be written as

\[
\vartheta = \begin{bmatrix}
A \\
B
\end{bmatrix}
A \text{ and } B \text{ are the matrices of response function of EMI instrument for vertical and horizontal dipole configurations, respectively, and the elements of the } A \text{ and } B \text{ entities are themselves}
\]

\[
A_{ij} = \varphi^V (z_j + h_i) \Delta z
\quad \text{(5.14)}
\]

and

\[
B_{ij} = \varphi^H (z_j + h_i) \Delta z
\quad \text{(5.15)}
\]

where,

\[
z_j = (j - 1) \Delta z \quad \text{for } j = 1, 2, \ldots, M
\]
\( h_i = (i-1) \Delta z \) for \( i = 1, 2, \ldots, N \).

\( \varphi^v, \varphi^h \) are the instrument response function for vertical and horizontal dipole configurations discussed earlier in Section 4.2.

Following the nomenclature of Equations 4.14 and 4.15, the \( \mathbf{EC}_a \) column vector can be related to the volumetric moisture content column vector by

\[
\mathbf{EC}_a = \mathbf{K.EC}_{a0} + \mathbf{C} \tag{5.16}
\]

where \( \mathbf{K} \left( \begin{array}{c} k_v \\ k_h \end{array} \right) \) and \( \mathbf{C} \left( \begin{array}{c} c_v \\ c_h \end{array} \right) \) are respectively the constant of proportionality between local EC at depth (depending on whether using vertical or horizontal dipole configurations) and \( \theta_v \) at the same depth (Section 4.4), and \( \mathbf{C} \) is the numerical intercept on the respective calibration curve associating \( \mathbf{EC}_a \) with \( \mathbf{EC}_{a0} \).

**Creating Depth Profiles of Volumetric Moisture Content by Simple Matrix Inversion**

In theory the process of extracting the depth-related \( \theta_v \) profile would involve multi-height \( \mathbf{EC}_a \) measurements, followed by conversion to \( \mathbf{EC}_{a0} \) using relevant \( k \) and \( c \) values, then using Equation 5.13 by calculating the inverse, \( \mathbf{\Theta} \), of the column vector \( \mathbf{\Theta} \) and using

\[
\mathbf{\Theta}^{-1}.\mathbf{EC}_{a0} = \mathbf{\Theta}.\mathbf{\Theta}
\]

therefore,

\[
\mathbf{\Theta}^{-1}.\mathbf{EC}_{a0} = \mathbf{\Theta} \tag{5.17}
\]

where \( \mathbf{\Theta}^{-1}.\mathbf{\Theta} = \mathbf{I} \) (identity matrix).

The vector \( \mathbf{\Theta}^{-1} \) is given by

\[
\mathbf{\Theta}^{-1} = \begin{bmatrix} A^{-1} \\ B^{-1} \end{bmatrix}
\]

where \( A^{-1} \) and \( B^{-1} \) are the matrix inversions of \( A \) and \( B \), calculated using, for example Gaussian reduction.
However, the inversion process described above yields highly unstable or ill-conditioned matrices. In practice, this means that when the inverse matrix is applied to the measured multi-height ECa data, sensible predictions are only possible for a very small range of ECa around ‘perfect’ values (Thikonov & Arsenin 1977). In this work, the process is further compounded by the fact that the coefficients $k_V$ and $k_H$ are themselves determined from fitting regression equations to the multi-depth calibration equations derived from a number of pit calibration sites and the uncertainly in each value is close to ±10% (Section 4.4). The effect of the ill-conditioned process is illustrated as follows using a representative core sample site (core site #2) from the field data described earlier in Chapter 4 (Section 4.4.2). Table 5.1 is the matrix $A$ associated with Equation 5.14 for vertical dipole configuration and EM38 heights (depths) of 0, 40, 60, 80, 100 and 120 cm above (below) the ground.

Application of the matrix $A^{-1}$ (Table 5.2) to converted multi-height EM38 measurements, ECaθ-V, was found only to succeed in reproducing the multi-depth $\theta_v$ data ($\Theta \Theta \Theta \Theta$) when the $k_v$ value derived specifically from the location that the multi-height EM38 and multi-depth $\theta_v$ data used in the first place. This graphically illustrates the issue of instability. For a given set of multi-height EM38/multi-depth $\theta_v$ measurements, the prediction of $\Theta \Theta \Theta \Theta$ was found to become nonsensical ($\theta_v (z) \pm 10^5 \text{ m}^3/\text{m}^3$) when any of the ECa values at height above ground were deviated by ±10%, considered a suitable instrument uncertainly in ECa measurements (Heath et al. 1999) or when the $k_v$ value used deviated from the actual site-specific value by the expected ±10% uncertainty observed earlier in Section 4.4.2.

**Thikonov Regularization**

The process of ‘Thikonov regularization’ has been proposed as a method of stabilizing the inversion process (Thikonov & Arsenin 1977; Hansen 1992; Groetsch 1993), and minimize the sensitivity to input data errors. The Thikonov regularization procedure minimizes the residual norm $\| \Theta - EC_{a0} \|$ and some desirable property of the discrete differential operator $\| L_n \Theta \|$, where, $n$ denotes the order of differentiation.

The form of normalizing the standard least squares is given by

$$\min_{\Theta} \{(\Theta)\} = \min_{\Theta} \| \theta \Theta - EC_{a0} \|^2$$

(5.18)
In the regularization process a parameter $\alpha$ is also introduced to enforce the stability of the process. In associating all functions, the solution for minimizing the sensitivity to errors in the sense of Thikonov regularization is given by

$$
\min_{\Theta} \left\{ \Theta, \alpha \right\} = \| \theta \Theta - EC_{\alpha \theta} \|^2 + \alpha \| \nu_\alpha \Theta \|^2
$$

subject to $\alpha \geq 0$.

Table 5.1 EM38 sensitivity matrix $A$ (vertical dipole configuration) for heights/depths of 0 – 1.2 m above/below ground. This table is a reproduction (in different format) of the data presented in Table 4.1.

<table>
<thead>
<tr>
<th>Depth</th>
<th>0.2 m</th>
<th>0.4 m</th>
<th>0.6 m</th>
<th>0.8 m</th>
<th>1.0 m</th>
<th>1.2 m</th>
</tr>
</thead>
<tbody>
<tr>
<td>Height</td>
<td>0.2 m</td>
<td>0.640329</td>
<td>0.761823</td>
<td>0.629690</td>
<td>0.476404</td>
<td>0.357771</td>
</tr>
<tr>
<td></td>
<td>0.4 m</td>
<td>0.761823</td>
<td>0.629690</td>
<td>0.476404</td>
<td>0.357771</td>
<td>0.273100</td>
</tr>
<tr>
<td></td>
<td>0.6 m</td>
<td>0.629690</td>
<td>0.476404</td>
<td>0.357771</td>
<td>0.273100</td>
<td>0.213064</td>
</tr>
<tr>
<td></td>
<td>0.8 m</td>
<td>0.476404</td>
<td>0.357771</td>
<td>0.273100</td>
<td>0.213064</td>
<td>0.169836</td>
</tr>
<tr>
<td></td>
<td>1.0 m</td>
<td>0.357771</td>
<td>0.273100</td>
<td>0.213064</td>
<td>0.169836</td>
<td>0.138040</td>
</tr>
<tr>
<td></td>
<td>1.2 m</td>
<td>0.273100</td>
<td>0.213064</td>
<td>0.169836</td>
<td>0.138040</td>
<td>0.114134</td>
</tr>
</tbody>
</table>

The inversion matrix $A^{-1}$ is reproduced in Table 5.2 below.

Table 5.2 The inverted EM38 sensitivity matrix $A^{-1}$ (vertical dipole configuration) for heights/depths of 0 – 1.2 m above/below ground. Note, values are rounded off to nearest 4th decimal place and height column and depth row layout are the same as Table 5.1.

<table>
<thead>
<tr>
<th></th>
<th>-22.9227</th>
<th>641.7041</th>
<th>-4728.6432</th>
<th>14330.4232</th>
<th>-20804.6298</th>
<th>14396.8594</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>641.7041</td>
<td>-14085.3498</td>
<td>94528.6613</td>
<td>-268403.3494</td>
<td>365023.2375</td>
<td>-233673.7399</td>
</tr>
<tr>
<td></td>
<td>-4728.6432</td>
<td>94528.6613</td>
<td>-583332.8194</td>
<td>1446900.4832</td>
<td>-1557058.0964</td>
<td>614117.4481</td>
</tr>
<tr>
<td></td>
<td>14330.4232</td>
<td>-268403.3493</td>
<td>1446900.4831</td>
<td>-2496358.9986</td>
<td>155723.7945</td>
<td>2943623.1714</td>
</tr>
<tr>
<td></td>
<td>-20804.6298</td>
<td>365023.2375</td>
<td>-1557058.0962</td>
<td>155723.7945</td>
<td>8465457.1371</td>
<td>-13616013.1940</td>
</tr>
<tr>
<td></td>
<td>14396.8594</td>
<td>-233673.7399</td>
<td>614117.4481</td>
<td>2943623.1714</td>
<td>-13616013.1940</td>
<td>17837532.7007</td>
</tr>
</tbody>
</table>
The choice of $\alpha$ and $L$ is important for a balance solution of the problem. The most common choice of $L$ is $L_0$ and $L_2$, where $L_0$ represents the identity matrix $I$ (first derivative operator) that controls the fluctuation of the moisture content of soil profile. The other option is $L_2$, the second derivative operator that enforce the smoothness of the moisture profile. The first and second derivative forms of $L$ are shown below

\[
L_0 = \begin{bmatrix}
1 & 0 & 0 \\
1 & 0 & 0 \\
1 & 0 & 0 \\
\end{bmatrix}
\quad \text{and} \quad
L_2 = \begin{bmatrix}
1 & -2 & 1 \\
1 & -2 & 1 \\
1 & -2 & 1 \\
\end{bmatrix}
\quad (5.20)
\]

The regularization parameter $\alpha$ can be chosen using a variety of methods such as discrepancy principles (Morozov 1984), generalized cross validation (Wahba 1990) and the L-curve criterion (Hansen 1992). Use of L-curve criterion is a common approach to choose the appropriate value of the parameter $\alpha$. The L-curve can be generated by plotting the norm of the regularized solution $\|L_n \Theta\|$ versus the norm of the corresponding residual $\|\Theta - EC_{\alpha_0}\|$ (Lawson & Hanson 1974; Hansen 1992; Borchers et al. 1997). The value of $\alpha$ is also important to minimize both norms. The large value of $\alpha$ effectively minimizes the norm of the residual and the small value of $\alpha$ minimizes the norm of the solution (Hansen 1992; Borchers et al. 1997). The value of the parameter $\alpha$ should be the lowest distance between the L-shaped corner and the origin of the axis that produce a balance between norms $\|\Theta - EC_{\alpha_0}\|$ and $\|L_n \Theta\|$. Following the L-curve criterion the value of $\alpha$ was chosen from the distance to the corner of the L-curve where maximum curvature of the L-curve exists. The equation of finding the $\alpha$ value is given as:

\[
D^2 = \left\|\Theta - EC_{\alpha_0}\right\|^2 + \left[\alpha \left\|L_n \Theta\right\|^2\right]^2
\]

\[
\min_{\alpha} D^2 = \min_{\alpha} \left\{\left\|\Theta - EC_{\alpha_0}\right\|^2 + \left[\alpha \left\|L_n \Theta\right\|^2\right]^2\right\}
\quad (5.21)
\]

A number of workers have utilized the Thikonov regularization procedure to reconstruct the depth profiles of soil property (e.g., $EC_a$) from multi-height EMI measurements (e.g., Borchers et al. 1997; Hendrickx et al. 2002; Deidda et al. 2003).
Borchers et al. (1997) used a linear inverse problem (where the assumption \( N_B << 1 \)

, described in Chapter 4, is valid) with a second order Thikonov regularization technique

(Thikonov & Arsenin 1977) to determine the soil \( \text{EC}_a \) profile from multi-

height field EMI measurement. They compared their results of \( \text{EC}_a \) values derived

with that obtained using models proposed by Cook & Walker (1992) and Rhoades

(1993). The solutions were substantially far from the measured \( \text{EC}_a \) profiles, however,

produced better results than those obtained using the forward propagation models of

Cook & Walker (1992) and Rhoades (1993). Hendrickx et al. (2002) conducted a

study to compare the linear and non-linear inverse problem (where the assumption \( N_B

<<1 \) breaks down) for predicting an \( \text{EC}_a \) profile. The non-linear inverse model was

found to outperform the linear inverse model in a situation where the EM38 values

were below 100 mS/m. However, the error difference between results produced using

both linear and nonlinear models was only 1% when data were evaluated for

horizontal measurements. Deidda et al. (2003), Borchers et al. (1997) and Hendrickx

et al. (2002) all performed similar work and in all studies the linear inverse model

produced \( \text{EC}_a \) profiles remarkably different from the expected conductivity profile.

5.3 OBJECTIVES OF THIS CHAPTER

Both the forward propagation and inverse matrix approaches described above offer

the potential to produce depth profiles of soil parameters. In the previous work the

parameter of interest has either been ‘local’ electrical conductivity, or salinity. The

former parameter is an obvious direct driver of integrated, above-ground EMI

response, and the latter, in consideration of the fieldsite over which the work was

conducted was considered the primary contributor to the local, at-depth electrical

conductivity. This present work concerns itself with underlying volumetric soil

moisture and in the previous chapter (Chapter 4) it was demonstrated that \( \theta_v \) is a key

driver of \( \text{EC}_a \) in the deep Vertosol soils of this field’s sites. The objective of this

chapter is therefore to demonstrate the use of both forward propagation and inverse

matrix procedures for determining \( \theta_v(z) \). To this end, the forward propagation

methods of Rhoades & Corwin (1981) and Slavich (1990) will be tested along with

the inverse matrix, Thikonov Regularization procedure described by Borchers et al.

(1997).
5.4 MATERIALS AND METHODS

5.4.1 Soil Sampling and $\theta_v$ Determination

The two forward propagation models were conducted using soil samples as well as surface and multi-height EM38 measurements acquired at the four pit sites described earlier in Sections 3.4.2. and 4.4.1. Average values of $\theta_v$ were calculated for ‘specific depth’ intervals of 0-0.4, 0.4-0.8, 0.8-1.2 m and for ‘composite depths’ of 0-0.4, 0-0.8 and 0-1.2 m. For the Tikhonov Regularization, $\theta_v$ were used for depths of 0.1, 0.2, 0.4, 0.6, 0.8, 1.0, 1.2 m. The forward propagation models were subsequently validated using 8 randomly selected core locations (from the 14 core site locations on offer) with 6 revisits (similar to that used earlier in Section 4.5.1). As in the previous chapter, $\theta_v$ from each of the randomly selected sites were determined using the neutron probe and calibration equations developed in Section 3.5.2. The Thikonov regularization-generated $\theta_v$ profile was compared directly with the actual measured pit $\theta_v$ profile for the particular pit over which the multi-height EM38 data was collected.

5.4.2 EMI Measurements

The EM38 measurements of $EC_a$ were acquired at heights of 0, 0.2, 0.4, 0.6, 0.8, 1.0 and 1.2 m above the pit sites. Each measurement was recorded as the average of 3 separate measurements (in the same visit) for each height for both horizontal and vertical dipole orientations. Multi-height EM38 measurements were also collected above 8 core locations (same height intervals) for validation of each forward propagation model. As a means of testing the validity of the equations linking $EC_a$ and local moisture-driven $EC_{a\theta}$, models were derived and their performance under validation compared using both $EC_a$ measurements as well as converted $EC_{a\theta}$ values. Based on the earlier work in Section 4.4.2 (specifically Figure 4.11), equations used to convert measured $EC_a$ to $EC_{a\theta}$ were

\[
EC_{a\theta-V} = 0.018 EC_{a-V} - 0.163
\]  
\[ (5.22) \]

\[
EC_{a\theta-H} = 0.028 EC_{a-H} - 0.176
\]  
\[ (5.23) \]
5.4.3 Forward Models for Depth-Specific $\theta_v$ Prediction

*Rhoades and Corwin Model*

The forward propagation model of Rhoades & Corwin (1981) was applied to the pit calibration data. Both the multi-height EC$_a$ and subsequently converted EC$_{a0}$ values were regressed with the volumetric moisture content at composite depth groups of 0-0.4, 0-0.8 and 0-1.2 m and successive depth groups of 0-0.4, 0.4-0.8 and 0.8-1.2 m. The model equations generated were

$$\bar{\theta}_{\psi(a-z)} = \alpha_1 EC_{a0-V,H1} + \alpha_2 EC_{a0-V,H2} + \alpha_3 EC_{a0-V,H3} + \alpha_4 EC_{a0-V,H4}$$  \hspace{1cm} (5.24)

and

$$\bar{\theta}_{\psi(a-z)} = \alpha_1 EC_{a-V,H1} + \alpha_2 EC_{a-V,H2} + \alpha_3 EC_{a-V,H3} + \alpha_4 EC_{a-V,H4}$$  \hspace{1cm} (5.25)

where $\bar{\theta}_{\psi(a-z)}$ is the average $\theta_v$ (0-0.4, 0-0.8, 0-1.2, 0.4-0.8, 0.8-1.2 m), EC$_{a-V,H1}$, EC$_{a-V,H2}$, EC$_{a-V,H3}$, EC$_{a-V,H4}$ are EM38 measurements and EC$_{a0-V,H1}$, EC$_{a0-V,H2}$, EC$_{a0-V,H3}$, EC$_{a0-V,H4}$ are the converted EM38 measurements using Equations (5.22) and (5.23) at 0, 0.4, 0.8, 1.2 m height respectively and $\alpha_1$, $\alpha_2$, $\alpha_3$, $\alpha_4$ are the regression coefficients.

The multiple regression analyses for this model were conducted using the JMP statistical software (SAS 2005). Comparitative statistics were generated using Statgraphics software (Statpoint 2005).

*Slavich Model*

The Slavich (1990) model equation was employed in two forms; one each to accommodate EC$_a$ and EC$_{a0}$:

$$\bar{\theta}_{\psi(a-z)} = \alpha + \beta_1 EC_a + \beta_2 EC_a$$  \hspace{1cm} (5.26)

$$\bar{\theta}_{\psi(a-z)} = \alpha + \beta_1 EC_{a0} + \beta_2 EC_{a0}$$  \hspace{1cm} (5.27)

where $\bar{\theta}_{\psi(a-z)}$ is the average $\theta_v$ down to a given depth group (0-0.4, 0-0.8, 0-1.2, 0.4-0.8, 0.8-1.2 m), $\alpha$, $\beta_1$ and $\beta_2$ are the modelled coefficients, EC$_a$, EC$_{a-H}$ are the actual
surface EM38 measurements for vertical and horizontal dipole configurations and \( \text{EC}_{a0-V}, \text{EC}_{a0-H} \) are EC\(_a\) values converted using Equations (5.22) and (5.23) for the specific dipole configuration. Comparison of fitted regression lines analysis was performed using Statgraphics software (Statpoint 2005).

### 5.4.4 Inverse Model for Predicting Multi-Depth \( \theta_v \)

A matrix inversion process employing the Thikonov regularization procedure described in Section 5.2.3 (Equations 5.16 - 5.21) and Equations (5.22) and (5.23) was applied to the pit calibration data. The procedures of ‘no regularization’, first-order and second order Thikonov regularization inversion were conducted using code written in the mathematical software- MAPLE (Maplesoft 2003). A transcript of the code developed specifically for this project is provided in Appendix A.

### 5.5 RESULTS AND DISCUSSION

#### 5.5.1 Multi-depth \( \theta_v \) Prediction Using Forward Models

**Rhoades & Corwin Model Calibration**

Following the Rhoades & Corwin model the multiple linear regression equations for \( \theta_v \) at depth groups 0-0.4, 0-0.8, 0-1.2, 0.4-0.8 and 0.8-1.2 m of the profile and multi-height EC\(_a\) values for both horizontal and vertical measurements are summarized in Table 5.3. The equations for the same soil depths and multi-height EM38 measurements converted to EC\(_a\) are given in Table 5.4. Statistically significant relationships (p < 0.05) were observed in all cases.

The EC\(_a\) for both vertical and horizontal dipole orientations predicts moisture content at different depths with accuracy ranges from 0.014 m\(^3\)/m\(^3\) to 0.028 m\(^3\)/m\(^3\) (Table 5.3). The lowest precision (\( R^2 = 0.89 \)) of calibration was associated with deepest depth group (0.8 – 1.2 m) when it was measured in vertical mode. In the same dipole orientation the highest precision (\( R^2 = 0.98 \)) was acquired in top depth group (0 - 0.4 m). This is a direct result of the depth response function of the EM38, where in vertical mode, the maximum response occurs at ~ 0.4 m depth which decreases to zero at surface and again subsequently decreasing with increasing depth below ~ 0.4 m depth (Figure 4.3 in Chapter 4). In the case of horizontal dipole configuration
measurements, higher precision was observed ($R^2 = 0.97$ to $0.98$) in all depth groups. This is attributed to the fact that significantly more of the integrated response of the instrument occurs from the surface to a depth of 1.2 m in the horizontal dipole configuration (~80%) compared to vertical dipole configuration (~62%).

The results achieved using converted EC$_{a\theta}$ values (Table 5.4, for both vertical and horizontal dipole orientations) were consistent with those derived using EC$_a$ (Table 5.3) with the notable exception of the 0.8 – 1.2 m depth layer.

Table 5.3 Multiple linear regression equations between $\theta_v$ layers and multi-height EC$_a$ measurements. All equations are statistically-significant ($p < 0.05$). Coefficients apply to $\overline{\theta}_{(\theta_v - \cdot)} = a + b_1EC_{a0} + b_2EC_{a0.4} + b_3EC_{a0.8} + b_4EC_{a1.2}$, where $\overline{\theta}_{(\theta_v - \cdot)}$ is the average volumetric moisture content at a given depth group.

<table>
<thead>
<tr>
<th>Dipole Depth Config. (m)</th>
<th>Regression equations coefficients</th>
<th>$R^2$</th>
<th>RMSE $(\text{m}^3/\text{m}^3)$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$a$</td>
<td>$b_1$</td>
<td>$b_2$</td>
</tr>
<tr>
<td><strong>Composite depths</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.215</td>
<td>-0.001</td>
<td>0.023</td>
</tr>
<tr>
<td>0-0.8</td>
<td>0.344</td>
<td>0.004</td>
<td>-0.003</td>
</tr>
<tr>
<td>0-1.2</td>
<td>0.389</td>
<td>0.007</td>
<td>-0.019</td>
</tr>
<tr>
<td><strong>Specific depth</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.215</td>
<td>-0.001</td>
<td>0.023</td>
</tr>
<tr>
<td>0.4-0.8</td>
<td>0.473</td>
<td>0.010</td>
<td>-0.029</td>
</tr>
<tr>
<td>0.8-1.2</td>
<td>0.348</td>
<td>0.005</td>
<td>-0.020</td>
</tr>
<tr>
<td><strong>Composite depths</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.206</td>
<td>-0.015</td>
<td>0.019</td>
</tr>
<tr>
<td>0-0.8</td>
<td>0.253</td>
<td>-0.011</td>
<td>0.011</td>
</tr>
<tr>
<td>0-1.2</td>
<td>0.243</td>
<td>-0.005</td>
<td>0.003</td>
</tr>
<tr>
<td><strong>Specific depth</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.206</td>
<td>-0.015</td>
<td>0.019</td>
</tr>
<tr>
<td>0.4-0.8</td>
<td>0.300</td>
<td>-0.007</td>
<td>0.004</td>
</tr>
<tr>
<td>0.8-1.2</td>
<td>0.222</td>
<td>0.005</td>
<td>-0.013</td>
</tr>
</tbody>
</table>
Table 5.4 Multiple linear regression equations between $\theta_v$ layers and multi-height EC$_a$ measurements subsequently converted to EC$_{a\theta}$. All equations are statistically-significant (p < 0.05). Coefficients apply to $\bar{\theta}_{v(a-c)} = a + b_1$EC$_{a0} + b_2$EC$_{a0.4} + b_3$EC$_{a0.8} + b_4$EC$_{a1.2}$, where $\bar{\theta}_{v(a-c)}$ is the average volumetric moisture content at a given depth group.

<table>
<thead>
<tr>
<th>Dipole Depth Config. (m)</th>
<th>Regression equations coefficients</th>
<th>$R^2$</th>
<th>RMSE (m$^3$/m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>a</td>
<td>b$_1$</td>
<td>b$_2$</td>
</tr>
<tr>
<td>Composite depths</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.268</td>
<td>-0.123</td>
<td>1.195</td>
</tr>
<tr>
<td>0-0.8</td>
<td>0.33</td>
<td>-0.036</td>
<td>0.703</td>
</tr>
<tr>
<td>0-1.2</td>
<td>0.355</td>
<td>0.023</td>
<td>0.251</td>
</tr>
<tr>
<td>Specific depth</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.268</td>
<td>-0.123</td>
<td>1.195</td>
</tr>
<tr>
<td>0.4-0.8</td>
<td>0.392</td>
<td>0.051</td>
<td>0.209</td>
</tr>
<tr>
<td>0.8-1.2</td>
<td>0.404</td>
<td>0.14</td>
<td>-0.65</td>
</tr>
</tbody>
</table>

| Composite depths        |     |       |       |       |       |
| 0-0.4                   | 0.271 | 0.301 | 0.27 | 0.043 | -1.416 | 0.93 | 0.050 |
| 0-0.8                   | 0.319 | 0.22 | 0.306 | -0.144 | -0.973 | 0.93 | 0.040 |
| 0-1.2                   | 0.306 | 0.117 | 0.308 | -0.239 | -0.422 | 0.95 | 0.026 |
| Specific depth          |     |       |       |       |       |
| 0-0.4                   | 0.271 | 0.301 | 0.27 | 0.043 | -1.416 | 0.93 | 0.050 |
| 0.4-0.8                 | 0.367 | 0.139 | 0.342 | -0.332 | -0.529 | 0.94 | 0.030 |
| 0.8-1.2                 | 0.279 | -0.087 | 0.311 | -0.429 | 0.679 | 0.99 | 0.003 |

*Slavich Model Calbration*

The calibration outcomes for the Slavich Model using both EC$_a$ and converted EC$_{a\theta}$ measurements are summarized in Tables 5.5 and 5.6, respectively.

Using EC$_a$ data, the precision of the model was found to be similar for each depth group except the lowest depth group (0.8 – 1.2 m). The highest accuracy of calibration (0.011 m$^3$/m$^3$) was found at the 0 – 0.4 m depth groups with highest precision.
(R^2 = 0.99). On the other hand the lowest precision and accuracy was associated with 0.8 – 1.2 m depth group with R^2 = 0.86 and RMSE = 0.027 m^3/m^3 respectively.

Consistently strong relationship between \( \theta_v \) and EC_{a0} was also observed for all cases of composite depths (R^2 = 0.97 to 0.99) and for specific depth groups (R^2 = 0.78 to 0.99). The comparatively low R^2 value (0.78) was associated with 0.8-1.2 m depth group and with RMSE value of 0.027 m^3/m^3.

The Slavich model predicted \( \theta_v \) very well for both composite and specific depth groups. The performance of this model was found similar to the original model developed by Slavich (1990) when applied to EC_a. This is possibly due to the ability of the model to incorporate both the vertical and horizontal measurements of EM38 which in fact encompasses the maximum soil volume to predict \( \theta_v \).

In both models a comparatively lower R^2 value was observed between \( \theta_v \) at 0.8 – 1.2 m depth and apparent electrical conductivity (EC_a and EC_{a0}) of both dipole modes. This is not surprising since only 15% and 9% of the instrument’s response in vertical and horizontal modes respectively comes from this depth segment.

Table 5.5 The regression coefficients for the average volumetric moisture content at particular depths, \( \bar{\theta}_{v(z-v)} \) and EC_a measurements. Equation: \( \bar{\theta}_{v(z-v)} = a + b_1 \text{EC}_{a-V} + b_2 \text{EC}_{a-H} \). All results were statistically highly significant (p < 0.0001).

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>a</th>
<th>( b_1 )</th>
<th>( b_2 )</th>
<th>R^2</th>
<th>RMSE (m^3/m^3)</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Composite Depths</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.1</td>
<td>0.331</td>
<td>-0.004</td>
<td>0.006</td>
<td>0.99</td>
<td>0.015</td>
<td>8</td>
</tr>
<tr>
<td>0-0.2</td>
<td>0.370</td>
<td>-0.004</td>
<td>0.006</td>
<td>0.99</td>
<td>0.013</td>
<td>8</td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.376</td>
<td>-0.003</td>
<td>0.005</td>
<td>0.99</td>
<td>0.011</td>
<td>8</td>
</tr>
<tr>
<td>0-0.6</td>
<td>0.384</td>
<td>-0.002</td>
<td>0.004</td>
<td>0.99</td>
<td>0.011</td>
<td>8</td>
</tr>
<tr>
<td>0-0.8</td>
<td>0.387</td>
<td>-0.002</td>
<td>0.004</td>
<td>0.99</td>
<td>0.014</td>
<td>8</td>
</tr>
<tr>
<td>0-1.0</td>
<td>0.352</td>
<td>-0.001</td>
<td>0.003</td>
<td>0.98</td>
<td>0.014</td>
<td>8</td>
</tr>
<tr>
<td>0-1.2</td>
<td>0.310</td>
<td>0.0003</td>
<td>0.002</td>
<td>0.97</td>
<td>0.014</td>
<td>8</td>
</tr>
<tr>
<td>Specific Depth</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.376</td>
<td>-0.003</td>
<td>0.005</td>
<td>0.99</td>
<td>0.011</td>
<td>8</td>
</tr>
<tr>
<td>0.4-0.8</td>
<td>0.396</td>
<td>-0.001</td>
<td>0.003</td>
<td>0.96</td>
<td>0.018</td>
<td>8</td>
</tr>
<tr>
<td>0.8-1.2</td>
<td>0.224</td>
<td>0.004</td>
<td>-0.001</td>
<td>0.86</td>
<td>0.022</td>
<td>8</td>
</tr>
</tbody>
</table>
Table 5.6 The regression coefficients for the average volumetric moisture content at particular depths, $\overline{\theta}_{(a-c)}$ and EC$_a$ measurements. Equation: $\overline{\theta}_{(a-c)} = a + b_1$ EC$_{aV} + b_2$ EC$_{aH}$. All results were statistically highly significant (p < 0.0001).

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>a</th>
<th>b$_1$</th>
<th>b$_2$</th>
<th>R$^2$</th>
<th>RMSE (m$^3$/m$^3$)</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Composite Depths</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.1</td>
<td>0.335</td>
<td>-0.212</td>
<td>0.224</td>
<td>0.99</td>
<td>0.016</td>
<td>8</td>
</tr>
<tr>
<td>0-0.2</td>
<td>0.369</td>
<td>-0.191</td>
<td>0.202</td>
<td>0.99</td>
<td>0.015</td>
<td>8</td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.380</td>
<td>-0.156</td>
<td>0.176</td>
<td>0.99</td>
<td>0.011</td>
<td>8</td>
</tr>
<tr>
<td>0-0.6</td>
<td>0.390</td>
<td>-0.133</td>
<td>0.157</td>
<td>0.99</td>
<td>0.012</td>
<td>8</td>
</tr>
<tr>
<td>0-0.8</td>
<td>0.396</td>
<td>-0.106</td>
<td>0.135</td>
<td>0.99</td>
<td>0.012</td>
<td>8</td>
</tr>
<tr>
<td>0-1.0</td>
<td>0.368</td>
<td>-0.050</td>
<td>0.101</td>
<td>0.98</td>
<td>0.012</td>
<td>8</td>
</tr>
<tr>
<td>0-1.2</td>
<td>0.331</td>
<td>0.009</td>
<td>0.067</td>
<td>0.97</td>
<td>0.014</td>
<td>8</td>
</tr>
<tr>
<td>Specific Depth</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.380</td>
<td>-0.157</td>
<td>0.176</td>
<td>0.99</td>
<td>0.011</td>
<td>8</td>
</tr>
<tr>
<td>0.4-0.8</td>
<td>0.410</td>
<td>-0.064</td>
<td>0.101</td>
<td>0.97</td>
<td>0.016</td>
<td>8</td>
</tr>
<tr>
<td>0.8-1.2</td>
<td>0.263</td>
<td>0.176</td>
<td>-0.037</td>
<td>0.78</td>
<td>0.027</td>
<td>8</td>
</tr>
</tbody>
</table>

Performance of Rhoades & Corwin and Slavich Models with Validation Data

The performance of the Rhoades & Corwin and Slavich calibration equations for the validation data (data taken from core sites) using both EC$_a$ and EC$_{a0}$ models is summarized in Tables 5.7 and 5.8 and graphs of predicted versus actual $\theta_v$ using each of the models are given in Figure 5.3.

Regression lines of predicted $\theta_v$ (using EC$_a$ or EC$_{a0}$) vs actual $\theta_v$ were analysed to verify whether there is any significant difference for predicting $\theta_v$ from EC$_a$ and EC$_{a0}$. The P-value of the difference in slopes of ‘Predicted’ versus ‘Actual’ curves for both sets of model equations in Tables 5.7 and 5.8 were greater than 0.05 indicated that there were no significant difference at 95% or higher confidence level in predicting moisture content from EC$_a$ and EC$_{a0}$. This is not unexpected as the relationship between EC$_a$ and EC$_{a0}$ was found to be highly significant (P<0.0001) and linear (Section 4.4 in Chapter 4).
Table 5.7 Comparison of predicted versus actual volumetric moisture contents for the Rhoades & Corwin calibration equations using ECₐ and ECₐθ.

<table>
<thead>
<tr>
<th>Dipole Config.</th>
<th>Depth (m)</th>
<th>ECₐ</th>
<th>RMSEP (m³/m³)</th>
<th>R²</th>
<th>RMSEP (m³/m³)</th>
<th>R²</th>
<th>P-value for difference between ECₐ and ECₐθ predictors</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>ECₐ</td>
<td>RMSEP (m³/m³)</td>
<td>R²</td>
<td>RMSEP (m³/m³)</td>
<td>R²</td>
<td></td>
</tr>
<tr>
<td>Composite depths</td>
<td>0-0.4</td>
<td>0.57</td>
<td>0.086</td>
<td>0.73</td>
<td>0.063</td>
<td>0.244</td>
<td></td>
</tr>
<tr>
<td></td>
<td>0-0.8</td>
<td>0.61</td>
<td>0.063</td>
<td>0.73</td>
<td>0.047</td>
<td>0.000</td>
<td></td>
</tr>
<tr>
<td></td>
<td>0-1.2</td>
<td>0.53</td>
<td>0.037</td>
<td>0.63</td>
<td>0.053</td>
<td>0.126</td>
<td></td>
</tr>
<tr>
<td>Specific depth</td>
<td>0-0.4</td>
<td>0.57</td>
<td>0.086</td>
<td>0.73</td>
<td>0.063</td>
<td>0.244</td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.4-0.8</td>
<td>0.65</td>
<td>0.030</td>
<td>0.67</td>
<td>0.040</td>
<td>0.211</td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.8-1.2</td>
<td>0.58</td>
<td>0.060</td>
<td>0.56</td>
<td>0.081</td>
<td>0.111</td>
<td></td>
</tr>
</tbody>
</table>

Table 5.8 Comparison of predicted versus actual volumetric moisture contents for the Slavich calibration equations using ECₐ and ECₐθ.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>ECₐ</th>
<th>RMSEP (m³/m³)</th>
<th>R²</th>
<th>RMSEP (m³/m³)</th>
<th>R²</th>
<th>P-value for difference between ECₐ and ECₐθ predictors</th>
</tr>
</thead>
<tbody>
<tr>
<td>Composite depths</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.1</td>
<td>0.60</td>
<td>0.075</td>
<td>0.61</td>
<td>0.062</td>
<td>0.300</td>
<td></td>
</tr>
<tr>
<td>0-0.2</td>
<td>0.57</td>
<td>0.056</td>
<td>0.58</td>
<td>0.055</td>
<td>0.665</td>
<td></td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.68</td>
<td>0.045</td>
<td>0.69</td>
<td>0.049</td>
<td>0.756</td>
<td></td>
</tr>
<tr>
<td>0-0.6</td>
<td>0.73</td>
<td>0.045</td>
<td>0.71</td>
<td>0.045</td>
<td>0.999</td>
<td></td>
</tr>
<tr>
<td>0-0.8</td>
<td>0.72</td>
<td>0.049</td>
<td>0.69</td>
<td>0.043</td>
<td>0.575</td>
<td></td>
</tr>
<tr>
<td>0-1.0</td>
<td>0.63</td>
<td>0.046</td>
<td>0.62</td>
<td>0.044</td>
<td>0.777</td>
<td></td>
</tr>
<tr>
<td>0-1.2</td>
<td>0.50</td>
<td>0.066</td>
<td>0.49</td>
<td>0.049</td>
<td>0.420</td>
<td></td>
</tr>
<tr>
<td>Specific depth</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0-0.4</td>
<td>0.68</td>
<td>0.045</td>
<td>0.69</td>
<td>0.062</td>
<td>0.300</td>
<td></td>
</tr>
<tr>
<td>0.4-0.8</td>
<td>0.71</td>
<td>0.063</td>
<td>0.69</td>
<td>0.042</td>
<td>0.191</td>
<td></td>
</tr>
<tr>
<td>0.8-1.2</td>
<td>0.69</td>
<td>0.078</td>
<td>0.68</td>
<td>0.047</td>
<td>0.016</td>
<td></td>
</tr>
</tbody>
</table>
$R^2 = 0.57$ (R & C_H)
RMSEP = 0.086 m$^3$/m$^3$

$R^2 = 0.11$ (R & C_V)
RMSEP = 0.179 m$^3$/m$^3$

$R^2 = 0.68$ (S)
RMSEP = 0.045 m$^3$/m$^3$

\[0.10 \quad 0.20 \quad 0.30 \quad 0.40 \quad 0.50 \quad 0.60 \quad 0.70\]
\[0.10 \quad 0.20 \quad 0.30 \quad 0.40 \quad 0.50 \quad 0.60 \quad 0.70\]

$R^2 = 0.57$ (R & C_H)
RMSEP = 0.086 m$^3$/m$^3$

$R^2 = 0.11$ (R & C_V)
RMSEP = 0.179 m$^3$/m$^3$

$R^2 = 0.61$ (R & C_H)
RMSEP = 0.063 m$^3$/m$^3$

$R^2 = 0.72$ (S)
RMSEP = 0.049 m$^3$/m$^3$

$R^2 = 0.13$ (R & C_V)
RMSEP = 0.134 m$^3$/m$^3$
$R^2 = 0.53$ (R & C_H)
RMSEP = 0.057 m$^3$/m$^3$

$R^2 = 0.26$ (R & C_V)
RMSEP = 0.086 m$^3$/m$^3$

$R^2 = 0.50$ (S)
RMSEP = 0.066 m$^3$/m$^3$

$R^2 = 0.19$ (R & C_V)
RMSEP = 0.091

$R^2 = 0.71$ (S)
RMSEP = 0.063 m$^3$/m$^3$

$R^2 = 0.65$ (R & C_H)
RMSEP = 0.030 m$^3$/m$^3$

$R^2 = 0.19$ (R & C_V)
RMSEP = 0.091
Figure 5.3 Comparison of average volumetric moisture content, $\bar{\theta}_v$ predicted by the different FP models and that measured at depths (a) 0 – 0.4 m (b) 0 – 0.08 m (c) 0 – 1.2 m (d) 0.4 – 0.8 m and (d) 0.8 – 1.2 m. The symbols R & C_H, R & C_V and S represent the Rhoades & Corwin model (Horizontal), Rhoades & Corwin model (Vertical) and Slavich model. The 1:1 line (grey dotted line) on each graph indicates the region where predictions would agree with measurements.

The relative performances of each model for each depth layer, based on the graphs of Figure 5.3, are summarized in Table 5.9.

Figure 5.3 and the summary in Table 5.9 show that the Slavich model performed better as a $\theta_v$ predictor than other models in most of the depth groups. The Slavich model overestimated in higher moisture content and underestimated in lower moisture content; a trend also observed earlier in Chapter 4 with the on-ground, direct calibration of the EM38. In general the vertical mode model of Rhoades & Corwin performed poorest. The relative performance of each of the models, and including the dipole configurations employed for the Rhoades & Corwin model, is essentially determined by the matching of the respective model with the depth segments. Model performance, in general is maximized when the sensing volume of the soil in relation to the particular EM38 configuration used, is matched to the depth segment of interest. Whilst the ‘construction’ of the Slavich model offers most resilience to the
matching of depth segments (Table 5.9), the very sensitivity of the Rhoades & Corwin model to matching of depth segments (for example horizontal dipole configuration) ensures it performs best for the over mid-range depths.

Table 5.9 Summary description of FP model performance for predicting $\bar{\theta}_r$.

<table>
<thead>
<tr>
<th>Depth strata (m)</th>
<th>Rhoades &amp; Corwin Model (vertical)</th>
<th>Rhoades &amp; Corwin Model (horizontal)</th>
<th>Slavich Model</th>
<th>Model with lowest RMSEP</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-0.4</td>
<td>Consistently underestimated (all $\theta_v$)</td>
<td>Consistently underestimated (all $\theta_v$)</td>
<td>Good estimation with mix of underestimation ($\theta_v &lt; 0.5$ m$^3$/m$^3$) and overestimation ($\theta_v &gt; 0.5$ m$^3$/m$^3$)</td>
<td>Slavich</td>
</tr>
<tr>
<td>0 – 0.8</td>
<td>Consistently underestimated (all $\theta_v$)</td>
<td>Consistently underestimated (all $\theta_v$)</td>
<td>Good estimation with mix of underestimation ($\theta_v &lt; 0.45$ m$^3$/m$^3$) and overestimation ($\theta_v &gt; 0.45$ m$^3$/m$^3$)</td>
<td>Slavich</td>
</tr>
<tr>
<td>0 – 1.2</td>
<td>Consistently underestimated (all $\theta_v$)</td>
<td>Underestimated $\theta_v &gt; 0.42$ m$^3$/m$^3$</td>
<td>Overestimated $\theta_v &gt; 0.42$ m$^3$/m$^3$</td>
<td>Rhoades &amp; Corwin (horizontal)</td>
</tr>
<tr>
<td>0.4 – 0.8</td>
<td>Consistently underestimated (all $\theta_v$)</td>
<td>Consistently underestimated (all $\theta_v$)</td>
<td>Slightly underestimated (all $\theta_v$)</td>
<td>Rhoades &amp; Corwin (horizontal)</td>
</tr>
<tr>
<td>0.8 – 1.2</td>
<td>Good estimation</td>
<td>Consistently underestimated (all $\theta_v$)</td>
<td>Mix of underestimation ($\theta_v &lt; 0.52$ m$^3$/m$^3$) and overestimation ($\theta_v &gt; 0.52$ m$^3$/m$^3$)</td>
<td>Rhoades &amp; Corwin (horizontal)</td>
</tr>
</tbody>
</table>

5.5.2 Performance of TR Inversion Procedure

Of the inversion processed tested, the first-order Thikonov regularization of the multi-height EC$_a$ and EC$_{ab}$ data to predict $\theta_v$ at depths 0.1, 0.2, 0.4, 0.6, 0.8, 1.0 and 1.2 m yielded the best results for the two pit sites. These are given in Figure 5.4.
The TR model consistently under-predicted $\theta_v$ (~50 – 75%) at every depth irrespective of whether EC$a$ or EC$a_0$ were used. In fact the P-value for the regression lines of predicted $\theta_v$ (using EC$a$ and EC$a_0$) versus actual $\theta_v$ was 0.904 which indicated that there was no significant difference at 95% or higher significance level for predicting $\theta_v$ from EC$a$ and EC$a_0$ using TR. The ‘no regularization’ process was found to be highly unstable and yield significantly distorted values and second-order regularization processes also yield significantly distorted profiles. A mixture of significant under and over-prediction was also reported by Deidda et al. (2003) who attempted to reconstruct an EC$a$ depth profile using a TR inversion method. Deidda et al. (2003) also inverted the EC$a$ profiles using first order, second order and ‘no regularization’ processes and actually achieved relatively better results with no regularization. They concluded that Thikonov regularization process often distorted the conductivity profile. Similarly, Borchers et al. (1997) and Hendrickx et al. (2002) predicted the EC$a$ depth profile from multi-height EM38 measurements using least square inverse problem with second order Thikonov regularization and their modelled EC$a$ depth profile were generally outside the confidence region (95% level of significance) of measured EC$a$ profile with ±40% error. The results of this present work therefore, reinforce the observations of earlier worker; that the Thikonov regularization procedure, whilst achieving the aim of stabilizing the inversion process to produce a sensible (as opposed to ‘nonsensical’) output, nonetheless produces results of variable (and generally poor) accuracy. It is difficult to ascribe the poor performance of the Thikonov regularization process to any single aspect of the field data used in this work. Further work is required to examine the performance of the TR process in light of varying soil profile-shapes (irrespective of whether measuring EC$a$ or $\theta_v$).
Figure 5.4 Prediction of volumetric moisture content at different depths using a first-order Thikonov regularization inverse method for (a) Pit #1 and (b) Pit #2 sites. Inversion results for both ECₐ and ECₐθ are depicted.

5.6 CONCLUSION

The results of this chapter confirm that the depth profile of volumetric moisture content can be constructed from on-ground or multi-height EM38 measurements.
using forward propagation models. There is no significant difference in the ability to predict $\theta_v$ using $EC_a$ or $EC_{a0}$ values; thus direct measurements of $EC_a$ can be used without converting it to $EC_{a0}$. Results of this study claim that two forward propagation models; Rhoades & Corwin, but horizontal dipole mode only and Slavich model can be applied to reconstruct the depth profile of $\theta_v$ with accuracies (as determined by RMSEP) of approximately 10%. Of the two forward propagation models investigated, the Slavich models performed best overall, however, the results demonstrated that a key criterion for applying either model was in matching the model to the depth segment of interest. The procedure of inverting multi-height EM38 measurements using Thikonov regularization was found to yield a stable, however, quantitatively inaccurate estimation of the $\theta_v$ depth profile.
CHAPTER – SIX

MULTI-TEMPORAL EMI SURVEY OF Paddock-Scale Soil Moisture Variability
6.1 INTRODUCTION

EMI surveys offer the potential to map soil properties at high resolution (Buchter et al. 1991; Reese & Moorhead 1996; Waine et al. 2000; Huang et al. 2001; Wu 2002; Gallardo 2003; Hedley et al. 2004; Shukla et al. 2004; Corwin & Lesch 2005c; Terra et al. 2006; Balcovič et al. 2007). This study has so far concentrated on analyzing the vertical distribution of moisture in the soil profile (Chapters 3, 4 and 5). However, modern EMI surveying, when the sensor is integrated with a global positioning system (GPS) and datalogger has the potential to undertake rapid surveys across entire paddocks or regions. This chapter utilizes the understanding of how EM38 measures volumetric soil moisture content in deep Vertosol soils to explore the potential for rapid EM38 survey to delineate soil moisture zones across the study area.

6.2 MAPPING SOIL SPATIAL VARIABILITY WITH EM38

6.2.1 Measuring Soil Variability Using EMI Sensor

Spatial variability of soil characteristics within a field significantly affects crop production (Birrel et al. 1995; Verhagen et al. 1995) and soil management (Corwin et al. 2003). The variability of soil properties has been identified and mapped by many authors using point-based sampling and various geostatistical procedures (Buchter et al. 1991; Reese & Moorhead 1996; Huang et al. 2001; Wu 2002; Gallardo 2003; Shukla et al. 2004; Terra et al. 2006; Balcovič et al. 2007). However, mobile soil sensors offer the potential to sample thousands of points across a field and greatly increase the spatial resolution of soil variability maps. Electromagnetic conductivity measurements have been found to provide rapid, high resolution information on soil variability (Hendrickx et al. 1992; Johnson et al. 2001; Domsch & Giebel 2004) that relate to variability in salinity, texture and claypan depth to name a few (e.g., Cameron et al. 1981; Doolittle et al. 1994; Franzen & Kitchen 1999; Sudduth et al. 2001).

Halvorson & Rhoades (1974) identified saline seeps using EC$_a$ maps. Corwin & Lesch (2005b) proposed a protocol of EMI survey to characterize soil spatial variability and Corwin & Lesch (2005c) investigated the variability of more than 15 soil properties including soil moisture, bulk density, clay content, saturation extract,
saturation percentage and exchangeable cations. In other applications, the EMI technique has been used to quantify spatial variability and to produce EC$_a$ maps that relate to the chemical and physical properties of non-saline soils, including moisture content (Kachanoski et al. 1988; Waine et al. 2000; Johnson et al. 2001; Corwin & Lesch 2005c), depth to water table (Schumann & Zaman 2003; Sherlock & McDonnell 2003), clay content and textural classes (Williams & Hoey 1987; Lund et al. 1999; Waine et al. 2000; Neudecker et al. 2001; Anderson-Cook et al. 2002; Domsch & Giebel 2004), depth to clay pan (Doolittle et al. 1994; Sudduth et al. 1995), top soil depth (Davis et al. 1997; Sudduth & Kitchen 2006), cation exchange capacity and exchangeable Ca and Mg (McBride et al. 1990; Corwin & Lesch 2005c), soil organic carbon (Jaynes 1996), herbicide behavior in soil (Jaynes et al. 1994) and nitrogen application recommendations (Davis et al. 1997).

Several authors have found good relationships between EMI survey and soil nutrient zones. Johnson et al. (2001) found EC$_a$ useful to measure some soil physical (bulk density) and chemical properties (pH and organic matter content) and categorized soils based on the variation of soil variables. Heiniger et al. (2003) used EC$_a$ to map soil pH, texture, CEC and moisture holding capacity and to identify soil nutrient variability. A good relationship between nutrient level and soil EC$_a$ was also developed by Chang et al. (2001). Soil variability measurement by EC$_a$ value was also found suitable for variable rate fertilizer application (Robert 1993; Anderson-Cook et al. 2002).

### 6.2.2 Measuring Soil Moisture Variability Using EMI Sensor

Soil moisture is an important factor in determining crop growth and yield and it can vary dramatically within a field (Greminger et al. 1985; Famigletti et al. 1998; Hupet & Vanclooster 2002). As previously noted EC$_a$ is dependent on the clay, moisture and salinity in a soil (McNeill 1980). Though EMI sensors have been found to be useful to quantify the paddock-scale spatial variability of soil moisture content, it has mostly been used to monitor soil moisture content at a single point in time rather than examine temporal variability (Kachanoski et al. 1988; Sheets & Hendrickx 1995; Hanson & Kaita 1997; Khakural et al. 1998; Waine et al. 2000; Brevik & Fenton 2002; Reedy & Scanlon 2003; Brevik et al. 2006; Hezarjaribi & Sourell 2007). Moreover, previous work in this area used either a single on-the-go EMI survey or
point-based samples. Kachanoski *et al.* (1988) conducted a study in a section of 1.8 ha field. They sampled ECₜ values with EM38 and EM31 sensors and volumetric moisture content with Time Domain Reflectrometry (TDR) from 52 locations. In their study ECₜ explained 96% of the spatial variation in soil moisture content at the 0.5 m depth. Hedley *et al.* (2004) conducted a study to show the spatial relationship of a single EM38 survey with soil properties including soil moisture. They found a comparatively poor spatial relationship (R² = 0.42) between ECₜ and soil moisture compared to other soil properties such as percent clay (R² = 0.72). Similarly, Corwin & Lesch (2005c) did a single EM38 survey to delineate spatial variability of soil properties including soil moisture, however, they have found poor relation (R² = 0.13) between ECₜ and spatially distributed soil moisture.

### 6.3 Scope of this Chapter

Chapters 4 and 5 have demonstrated that ECₜ is strongly related to the θᵥ in the deep Vertosol soils of this study site. The aim of this chapter is to combine the relationships identified in Chapter 4 and relate multi-temporal, spatial, EM38 surveys to soil moisture status. To achieve this aim the chapter objectives were to:

- Re-confirm the influence of soil properties on ECₜ measurements and create 2-D maps of θᵥ using an on-the-go EM38 system
- Evaluate temporal variability in the spatial pattern of volumetric soil moisture content

### 6.4 Materials and Methods

#### 6.4.1 Soil Sampling and Analysis

Soils cores were taken from 17 locations within the study site (Figure 6.1). Each core location was measured using a differential global positioning system (DGPS) receiver (Trimble model TSCe, Trimble, Sunnyvale California, USA). Spatial coordinates for all data were projected to Universal Transverse Mercator (UTM), datum WGS84 and zone 56S. Each core was extracted to a depth of 1.2 m and sectioned into 0.1 m sub-samples (Section 3.4.1). Each sub-sample was preprocessed and analysed for soil θᵥ, pH and electrical conductivity (EC₁:5) and then averaged to a depth of 1.2 m for each
core. Clay content was determined from a composite sample of each sub-sample for each core. The method of determining $\theta_v$ was as described in Section 3.4.1. Soil pH and EC$_{1:5}$ were determined from a soil solution of 10 g air dried soil and 50 mL of deionised water giving a proportion of 1:5 of soil and water respectively (Rayment & Higginson, 1999). Soil clay content was determined using the pipette method of particle size analysis (Day, 1965).

### 6.4.2 EM38 Survey and Data Processing

On-ground EC$_a$ measurements were collected at each of the core sites (Figure 6.1) using the Geonics EM38 sensor in both vertical and horizontal dipole orientations to confirm the relationship between EC$_a$ and soil properties including $\theta_v$.

The on-the-go EM38 survey was conducted on the entire study area (2 ha) along pre-defined, 10 m transects (Figure 6.1). The 2-D survey was carried out in horizontal dipole orientation only (as the horizontal mode has been proved the best estimator of soil moisture, Chapter 4) at four different times (November 2007, December 2007, April 2008 and May 2008). The on-the-go EM38 system comprised the EM38 sensor mounted on a rubber sled, towed behind an all-terrain vehicle (ATV) (Figure 6.2) at a speed of approximately 8 km/hr. The EM38 unit itself was housed in a thermoplastic case (with Styrofoam insulation on either side and a cover over the top to reduce case temperature fluctuations). The EM38 housing and rubber sled combined maintains the EM38 sensor approximately 15 mm above the soil surface and this small spacing is found to have no noticeable impact of sensor readings (Brevik & Fenton 2003). The continuous data stream from the EM38 was fed into a Trimble TSCe data logger at 1-s intervals and simultaneously georeferenced using the Trimble DGPS receiver mounted at the back of the ATV. Prior to each survey, the EM38 was ‘nulled’ according to the protocol of Geonics Limited (2003). The elevation data were also recorded by the DGPS receiver during the continuous survey.

On-the-go EC$_a$ survey data were pre-processed to remove the interference from fence lines (located at the eastern side of the field), on-ground metals and sensor anomalies. Data were processed by removing extreme values (top and bottom 5% of the data) following the Cooperative Research Centre for Viticulture (CRCV) yield monitor processing protocol (Bramley & Williams 2001).
Figure 6.1 On-the-go EM38 survey track across the paddock and soil moisture neutron probe sampling locations (Background image derived from EC$_a$ survey).
6.4.3 Soil Moisture Content Estimation

The continuous EM38 survey EC$_a$ data (horizontal) were converted to $\bar{\theta}_{1,2}$ using the calibration equation for horizontal mode of Table 4.3 in Chapter 4 and the $\bar{\theta}_{1,2}$ values corresponding to each of the 14 core locations extracted and subsequently validated against $\bar{\theta}_{1,2}$ values determined using multi-depth (up to 1.2 m with 0.1 m strata) neutron probe measurements.

6.4.4 Statistical and Geostatistical Analysis

Descriptive statistics such as the mean, standard deviation (SD) and coefficient of variation (CV) were determined from the survey data. The frequency distribution of the estimated moisture content data was assessed for normality and the skewness and kurtosis coefficients were calculated. Linear regression and multiple stepwise regression models were developed between point EC$_a$ values and measured soil properties of 17 sampling points. A regression analysis was also performed between $\theta_v$ measured by neutron probe and $\theta_v$ predicted by on-the-go EM38 survey. All statistical analyses were performed using JMP statistical software (SAS 2005).

Semivariograms were calculated from the average volumetric moisture content estimated by the EC$_a$-H to a depth of 1.2 m (denoted as $\bar{\theta}_{1,2}$) to determine the degree of spatial correlation. The semivariograms were calculated using the VESPER software (Minasny et al. 2006). The settings for the semivariogram options for all datasets were: number of lags = 12; lag tolerance = 50%; maximum lag distance = 200 m; weighting = number of pairs/SD (Minasny et al. 2006). Each experimental semivariogram was fitted with linear, spherical, exponential and Gaussian models and the model of best fit was selected using the object function value. The model that produced the lowest objective function value was used as a fitted theoretical model. Once the experimental semivariograms had been calculated and the best fitting model was checked and then used to fit the data in each case. The results of the fittings were plotted and the nugget, sill and range values were recorded.
6.4.5 Map Production

The 2-D data were interpolated to a raster surface with 1 m by 1 m cells (pixels) using block kriging (10 m by 10 m local variograms) and the Vesper software package. The kriging analysis was configured according to the Southern Precision Agriculture Association (SPAA), Australia protocol (Bramley & Williams 2004). The interpolation was done both on $\theta_{1:2}$, $EC_a$ and elevation attributes. The interpolated values of the variables were then imported into ArcGIS software (Environmental Services Research Institute, California, USA) to create maps. $EC_a$ and moisture content maps were produced to show the temporal and spatial variability across the paddock and the elevation map was produced to compare the moisture variability with topography. All maps were produced with ArcGIS software using the configuration recommended by SPAA protocol (Bramley & Williams 2004).

![Continuous EM38 survey system in action](image)

Figure 6.2 A photograph of continuous EM38 survey system in action. The Trimble DGPS and datalogger are mounted on the front of the all-terrain vehicle (ATV). The DGPS antenna is mounted on the back of the ATV and the EM38 unit is mounted in a thermoplastic case on the rubber sled behind the ATV.

6.5 RESULTS AND DISCUSSION

6.5.1 Influence of Soil Properties on $EC_a$

A summary of the simple and step-wise regression statistics between on-ground $EC_a$ and average values of moisture content, clay content, $EC_{1:5}$ and pH is given in Table 6.1. The simple regression statistics indicate that of the four measured properties only
\( \theta_v \) was found to have a significant relationship with \( EC_a \), confirming \( \theta_v \) as the dominant factor influencing \( EC_a \) variability in this soil. This result was not unexpected as the study area is relatively small and has strongly uniform characteristics throughout paddock and profile (Figure 4.8 and Chapter 2).

The combined effect of the other soil properties using stepwise regression (Table 6.1) indicates that when clay and pH were added individually to \( \theta_v \) the \( R^2 \) values slightly improved for the horizontal dipole configuration measurements. The \( R^2 \) values improved for both operational modes when \( EC_{1.5} \) were added to the regression and it further increased with the addition of \( EC_{1.5} \) and pH in the multiple regression analysis \( (R^2 = 0.63 \text{ for both cases}) \). When all four properties \( (\theta_v, \text{ clay, } EC_{1.5} \text{ and pH} ) \) were regressed with \( EC_a \) the \( R^2 \) values increased from 0.46 to 0.63 for vertical measurements and from 0.57 to 0.65 for horizontal measurements and the regression equations were found to be significant. However, the multiple regressions without \( \theta_v \) were not significant.

Table 6.1 Simple and step wise regression analysis between \( EC_a \) and soil properties to show the influence of soil properties on \( EC_a \) measurements. The superscript * or ns indicate significance levels of 95% or non significant respectively.

<table>
<thead>
<tr>
<th>Regression Combinations</th>
<th>Vertical R²</th>
<th>Vertical P-value</th>
<th>Horizontal R²</th>
<th>Horizontal P-value</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Simple Regression</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( EC_a ) vs ( \theta_{1.2} )</td>
<td>0.46</td>
<td>0.003*</td>
<td>0.57</td>
<td>0.000*</td>
</tr>
<tr>
<td>( EC_a ) vs Clay</td>
<td>0.14</td>
<td>0.137ns</td>
<td>0.10</td>
<td>0.208ns</td>
</tr>
<tr>
<td>( EC_a ) vs pH</td>
<td>0.08</td>
<td>0.269ns</td>
<td>0.17</td>
<td>0.105ns</td>
</tr>
<tr>
<td>( EC_a ) vs ( EC_{1.5} )</td>
<td>0.18</td>
<td>0.085ns</td>
<td>0.14</td>
<td>0.147ns</td>
</tr>
<tr>
<td><strong>Stepwise Multiple Regression</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( EC_a ) vs ( \theta_{1.2} ) + Clay</td>
<td>0.46</td>
<td>0.013*</td>
<td>0.59</td>
<td>0.002*</td>
</tr>
<tr>
<td>( EC_a ) vs ( \theta_{1.2} ) +pH</td>
<td>0.46</td>
<td>0.013*</td>
<td>0.59</td>
<td>0.002*</td>
</tr>
<tr>
<td>( EC_a ) vs ( \theta_{1.2} ) + ( EC_{1.5} )</td>
<td>0.55</td>
<td>0.004*</td>
<td>0.62</td>
<td>0.001**</td>
</tr>
<tr>
<td>( EC_a ) vs ( \theta_{1.2} ) + ( EC_{1.5} ) + pH</td>
<td>0.63</td>
<td>0.004*</td>
<td>0.63</td>
<td>0.004*</td>
</tr>
<tr>
<td>( EC_a ) vs ( \theta_{1.2} ) + ( EC_{1.5} ) + Clay</td>
<td>0.55</td>
<td>0.013*</td>
<td>0.64</td>
<td>0.003*</td>
</tr>
<tr>
<td>( EC_a ) vs Clay + pH + ( EC_{1.5} )</td>
<td>0.63</td>
<td>0.013*</td>
<td>0.65</td>
<td>0.010*</td>
</tr>
<tr>
<td>( EC_a ) vs Clay + pH + ( EC_{1.5} )</td>
<td>0.29</td>
<td>0.197ns</td>
<td>0.21</td>
<td>0.361ns</td>
</tr>
</tbody>
</table>
6.5.2 Soil Moisture Survey Validation

The four EC\textsubscript{a} surveys conducted over the 6 months period (November 2007, December 2007, April 2008, May 2008) coincided with significantly different soil moisture conditions and the observed range of EC\textsubscript{a} values (i.e. mS/m) consequently varied. During the November - December period, the field site received over 100 mm of rainfall each month whereas the April - May period was significantly drier (Refer to Figure 2.5), the field site receiving only 8.0 mm of recorded rainfall falling in the 30 days prior to the April survey and 50.6 mm falling in the 30 days preceding the May survey. The highest EC\textsubscript{a} values were therefore, associated with the survey conducted in December 2007 with field moisture ranging from 0.4858 – 0.5565 m\(^3\)/m\(^3\), and the lowest EC\textsubscript{a} values obtained in April 2008, when the field moisture content ranged from 0.3898 to 0.4789 m\(^3\)/m\(^3\) (as measured by the neutron probe). Predicted and measured $\theta_{i,2}$ were significantly correlated for all surveys (P<0.001). A plot of predicted versus measured $\theta_{i,2}$ and summary statistics are illustrated in Figure 6.3.

Figure 6.3 shows that the predicted $\theta_{i,2}$ more closely resembled the actual $\theta_{i,2}$ when surveys were conducted with a relatively higher soil moisture profile. The November 2007 survey data predicted $\theta_i$ with an RMSEP of 0.007 m\(^3\)/m\(^3\) representing an error of approximately 1.5%. While lowest prediction accuracy was associated with the April 2008 survey with an error of approximately 10%. In dry conditions the EMI survey substantially underestimated the moisture content which is likely to be related to the distribution of moisture within the soil profile. Whilst the average moisture content determined from neutron probe or gravimetric sampling at 0.1 m intervals is not weighted according to depth however, the EM38-derived EC\textsubscript{a} values, especially those collected in horizontal dipole configuration, are significantly weighted to the moisture content closer to the surface. In dry conditions, owing to evaporation and pasture uptake the uppermost soil layers would be significantly drier.
Figure 6.3 Scatter-plot of predicted $\bar{\theta}_{1.2}$ using interpolated map values with the actual average volumetric moisture content, $\bar{\theta}_{1.2}$, derived using neutron probe measurements. Data for four different survey times. The 1:1 line (grey dotted line) on each graph indicates the region where predictions would agree with measurements.

### 6.5.3 Map Analysis

As mentioned in Section 6.4.2, the four EC$_a$ surveys were conducted with different soil moisture conditions. The maps in Figure 6.4 show that the field-scale spatial pattern of EC$_a$ remained relatively stable over the survey period, although, the magnitude of EC$_a$ values (i.e. mS/m) varied with soil moisture conditions. Again, the highest EC$_a$ values were associated with the survey conducted in December 2007 (field $\bar{\theta}_{1.2} \sim 0.4858 - 0.5565$ m$^3$/m$^3$ as measured by the neutron probe), and the lowest EC$_a$ values obtained in April 2008, (field $\bar{\theta}_{1.2} \sim 0.3898$ to 0.4789 m$^3$/m$^3$ as measured by the neutron probe).
Figure 6.4 Maps of soil EC$_a$ survey (horizontal mode) for four survey periods.
Descriptive statistics related to the survey dates are presented in Table 6.2. The standard deviations (SD) of the wet and dry surveys were significantly different (P < 0.0001) from each other and the two wet surveys were also significantly different from each other. However, the SD’s were the same for two dry surveys. Volumetric moisture content was less variable in the dry period compared to the wet period, the coefficient of variations (CV) ranged from 2.7 to 2.8% and 3.3 to 3.9% in dry and wet periods, respectively. The statistical distribution of $\bar{\theta}_{1,2}$ for each survey was normal as shown by the values of skewness and kurtosis. The normality of the data was tested by fitting the normal distribution and testing the goodness of fit. The P-value of Shapiro-Wilk test was < 0.0001 for all surveys indicates the normality of the data sets (SAS 2005).

The spatial distribution of soil EC$_a$, and the subsequently derived $\theta_v$ values is illustrated in map-form in Figures 6.4 and 6.5, respectively. These maps indicate that the spatial pattern of relatively high and low soil moisture areas is fairly stable throughout the survey period. A map of average moisture content, calculated by averaging the four individual survey datasets, and integrated with the digital elevation model of the field site is given in Figure 6.6. Since the main soil characters (texture, horizons etc) are relatively homogenous across the study area (Chapter 2) the spatial variability in moisture content is dominated by topographic variation. Sloping topography has caused soil moisture to flow down-slope, resulting in higher soil moisture in the lower-lying areas (Figure 6.6).

Quantitative measures of the relationship between the DEM and individual moisture content maps, as obtained by correlation and multiple regression analysis using the values for each 1 m$^2$ cell, showed that the December 2007 and April 2008 surveys had the highest ($R^2 = 0.45$) and lowest ($R^2 = 0.21$) respectively. This result is also consistent with the notion that wetter soils, owing to a more completely-filled profile, reflect more accurately in the EM38-derived EC$_a$ values, and thus moisture accumulation effects as dictated by topography are also accurately reflected. As the profile dries out (from the top down), the decoupling of the EM38 data produces a decoupling of the EC$_a$-derived $\theta_v$ from the topographical effects as well. However, the power of utilizing multi-temporal data, even when the less-accurate drier condition datasets are included become evident. Whilst the results indicate that the
relationship between elevation and $\theta_v$ increases with increasing average moisture content in the profile, the best relationship between elevation and $\overline{\theta}_{1.2}$ was observed ($R^2 = 0.50$) when $\overline{\theta}_{1.2}$ for all four dataset were incorporated into the multiple regression analysis.

Figure 6.5 Maps of volumetric soil moisture content across the field produced from horizontal EM38 survey conducted in November 2007, December 2007, April 2008 and May 2008.
Figure 6.6 Three dimensional (3D) view of moisture distribution across the field in relation to relative elevation of the field. The map was produced by draping the moisture content map with DEM to illustrate the distribution of volumetric moisture content. This 3D moisture content map was produced from a combined data sets averaged from 4 surveys conducted in November 2007, December 2007, April 2008 and May 2008.

Table 6.2 Statistical parameters for $\bar{\theta}_{12}$ predicted from EC$_a$ survey conducted in different times during 2007 and 2008 study periods. SD and CV are representing the standard deviation and coefficient of variation respectively.

<table>
<thead>
<tr>
<th>Survey</th>
<th>EC$_a$ (mS/m)</th>
<th>$\bar{\theta}_{12}$ (m$^3$/m$^3$)</th>
<th>SD (m$^3$/m$^3$)</th>
<th>CV (%)</th>
<th>Skewness</th>
<th>Kurtosis</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nov-07</td>
<td>76.1</td>
<td>0.4704</td>
<td>0.5364</td>
<td>0.4988</td>
<td>0.016</td>
<td>3.3</td>
</tr>
<tr>
<td>Dec-07</td>
<td>86.5</td>
<td>0.4948</td>
<td>0.5746</td>
<td>0.5286</td>
<td>0.021</td>
<td>3.9</td>
</tr>
<tr>
<td>Apr-08</td>
<td>27.2</td>
<td>0.3702</td>
<td>0.4134</td>
<td>0.3901</td>
<td>0.011</td>
<td>2.7</td>
</tr>
<tr>
<td>May-08</td>
<td>35.2</td>
<td>0.3882</td>
<td>0.4324</td>
<td>0.4074</td>
<td>0.011</td>
<td>2.8</td>
</tr>
</tbody>
</table>
6.6 CONCLUSION

The results of this study showed that moisture content is the major factor contributing to variation in the apparent electrical conductivity (EC$_a$) survey in the deep Vertosols of this study area.

The spatial distribution of soil EC$_a$ and the derived moisture content were observed to be temporally stable, although the absolute values of EC$_a$ did vary reflecting the absolute differences in moisture content. The EM38 derived EC$_a$ values produced the most accurate estimations of average moisture content (0 – 1.2 m) when the profile was wetter, however, the inclusion of multi-temporal datasets produced the most accurate representation of the topographical effects on moisture content.
CHAPTER – SEVEN

CONCLUSIONS AND RECOMMENDATIONS
7.1 CONCLUSIONS

The key finding of this research was that the apparent electrical conductivity (EC<sub>a</sub>) data produced by the commercially-available EM38 unit can be used to measure soil moisture in the root zone (0-1.2 m) of agricultural crops.

This study also found that the soil moisture neutron probes (SMNP) performed significantly better than the Diviner 2000 capacitance probe for measuring moisture content, θ<sub>v</sub>, in the Black Vertosol (cracking clay) soil. The SMNP was able to provide accurate soil moisture information at all soil profile depths. However, the Diviner 2000 capacitance probe was uncorrelated to gravimetrically determined soil moisture below 0.6 m. This result is consistent with previous studies that show mixed results for the two probes in cracking clay soil. The SMNP method measures a larger volume of soil than capacitance probe method, and is a more robust technique for soil moisture determination than the Diviner 2000 in the root zone area (1.2 m) of a cracking clay soil. As the SMNP performed well during the calibration work (Chapter 3) it indicates that the SMNP is well suited for profile based measurements. However, it is important to consider soil cracking while calibrating either the neutron probe or Diviner 2000 in cracking clay soils. In cracking clay soils, cracks are more prominent in surface soils and decrease with depth which affects the bulk density at depth. Therefore, it was found useful to do separate calibration at depths for both the SMNP and the Diviner 2000 in pit method. A pit-based calibration rather than calibration based on access tube cores analysis was found to provide more accurate volumetric soil moisture measurements and is recommended for these soils, even though it is more laborious, costly and time consuming. Therefore, the pit-based volumetric soil moisture content was subsequently used for EM38 sensor calibration.

In the deep Vertosol soil of this study site, the EM38 depth response function was not perturbed by the soil moisture profile and the results indicated that the depth-related soil moisture profile was the dominant contributor to the EM38 response. The constant of proportionality between the measured and predicted values of EC<sub>a</sub>, k, was defined and quantified and also found to be linked to the integrated depth response function of the EM38. This result confirms that the EM38 device can be used to measure soil moisture content. The regression of calibration between EC<sub>a</sub> and moisture content was found to be linear (Figure 4.16) and EC<sub>a</sub>-H explained 99%
variation with moisture content. The calibration points were derived based on the standard calibration procedure of two extreme moisture conditions (dry and wet profile) (Hignett & Evett 2002). Future work should verify the linearity of EM38 response to soil moisture content by including more intermediate moisture contents in the calibration, something that was beyond the scope of the present project.

In an attempt to discretize soil moisture into depth profiles, two forward propagation models (Rhoades & Corwin 1981 and Slavich 1990) and an inverse model using Thikonov regularization were trialled. Results of this study showed that measurement of $\theta_v$ at depth was possible with both on-ground and multi-height EM38 readings. The models were tested with both $\text{EC}_a$ and $\text{EC}_{a0}$ (derived from a function of EM38 depth response and local $\theta_v$) and both parameters provided almost identical results in all most all cases. The measurements provided by the EM38 in horizontal dipole configuration was found to explain more of the variation observed in the depth-related $\theta_v$ than those acquired from the EM38 in vertical dipole configuration. Both the forward propagation models were shown to provide useful measurements of depth specific soil moisture, while the inverse procedure was found to be inaccurate. In fact, the inverse procedure without imposing any regularization parameter is quite unstable. Imposing Thikonov regularisation parameters into the system improved the stability in the inverse system, however, failed to reconstruct the moisture profile within the range of accuracy although the results of this study were consistent with other workers. In order to improve the accuracy of soil moisture measurements through the inverse procedure, further research is required. Of the forward propagation models, the Slavich model, was found to be the best estimator for estimating depth-specific $\theta_v$ using EM38-derived $\text{EC}_a$ data.

On-the-go EM38 survey with subsequent calibration to volumetric moisture content proved to be a useful technique for understanding paddock scale soil moisture variability in the root zone. While soil moisture content varied considerably over the study period, zones of relatively high and low soil moisture content were temporally stable and corresponded to the topographic variation in the study site. The EM38-derived maps were more accurate when the soil was wetter; and the integration of multiple survey datasets yielded more accurate information about soil variability than a single survey.
The results of this study particularly for Chapters 4, 5 and 6 show that the EM38 is useful for soil moisture measurement and hence potentially useful in agricultural industry. Due to its ease of use, robustness, rapidity, large data capture and large volume of soil measured compared to other soil moisture sensors, EM38 has potential for irrigation planning and measuring water use efficiency.

7.2 RECOMMENDATIONS

This study was deliberately confined to a small study area with uniform soil, both spatially and at depth, to minimize the influence of factors other than soil moisture on EM38. While the results of the study are very positive, extension of the study to different areas with a range of agricultural soil types is required to generalise the results.

Further recommendations include:

1. Additional work to identify if increasing the number of the single core samples can achieve similar accuracy to the much more time consuming pit method.

2. Further calibration work following the pit-based sampling may be conducted in a range of soils covering a wide range of soil moisture contents to confirm the linear relationship between ECₐ and soil moisture.

3. One of the vital soil parameters that govern the response of EM38 measurement is salinity. Assessing the efficiency of using EM38 for estimating soil moisture in a high salinity condition is required. This may provide additional information as to the ‘robustness’ of the EM38 as a tool to measure soil moisture.

4. This study was conducted in a native pasture land, therefore, the results of this study may not be similar in an irrigated land of similar soil properties. In order to get maximum benefit from the results of this study, similar work may be conducted in an irrigated land of similar soil properties to recommend the EM38 to be a robust tool.

5. The inverse procedure substantially underestimated the moisture content while the first order regularization parameter was used. Other regularization
parameter e.g.; discrepancy principle may be applied to increase the accuracy of measurements. In addition to that model could be constructed from the data taken from multiple sites. Sites should have a wide range of soil properties and wide range of moisture content which might give a clear view of the performance of inverse procedure to measure soil moisture.

6. The efficacy of the on-the-go survey was not tested in crop land. Further study may be conducted in an irrigated land to show the spatial and temporal variability and to generate maps of soil moisture content at different depths following the models developed in Chapter 5.

7. As this study was restricted to the homogenous soil of the field site, it was assumed not to be useful to delineate the site-specific management zones. Conducting similar work in a relatively heterogeneous soil in combination with yield data would help interpret the relationship between EM38-derived $EC_a$ and yield variations that may lead to the adoption of site specific management practices.
REFERENCES


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Hulme, P. J. (2008, personal communication).


APPENDIX
Appendix-A

MAPLE Code for Inverting Moisture Profile From Converted EC<sub>a</sub> (EC<sub>a0</sub>) Using Thikonov Regularization

1. EMI Depth Profile Analysis

> restart:
> with(LinearAlgebra):
> with(plots):
> with(Optimization):
> with(stats):

2. Define EM38 Sensitivity with Depth Function

SV:=unapply(voff+4*d/(4*d^2+1)^(3/2),d):
SH:=unapply(hoff+2-4*d/(4*d^2+1)^(1/2),d):

3. Define Cumulative EM-38 Sensitivity Function [z to infinity]

> CSV:=z->int(SV(d),d=z..infinity);
> CSH:=z->int(SH(d),d=z..infinity):

\[
CSV := z \rightarrow \int_{z}^{\infty} SV(d) \, dd
\]

\[
CSH := z \rightarrow \int_{z}^{\infty} SH(d) \, dd
\]

4. Define EM Sensitivity versus Depth Vector

> sv:=matrix(7,7):
> sh:=matrix(7,7):
> dd:=matrix(7,7):
for i from 1 to 7 do
for j from 1 to 7 do
d:=(i-1)*0.2+(j-1)*0.2:
dd[i,j]:=d:
sv[i,j]:=SV(d):
sh[i,j]:=SH(d)
od:
K:=stackmatrix(sv,sh):
> print(`depth Matrix [dd]`);
evalm(dd);
print(`Stacked Apparent EMI Values asv & ash`);
evalm(ds);
print(`Sensitivity Matrix [K]`);
evalm(K);

5. Read EMI & $\theta$, Data

> n:=0:
h:=matrix(7,1):
asv:=matrix(7,1):
ash:=matrix(7,1):
vmc:=matrix(7,1):
s:=matrix(7,1):
data_file:='C:/Data.txt':
fopen(data_file,READ,TEXT):
for i from 1 to 1 do readline(data_file) od:
for i from 1 while feof(data_file) = false do
h[i,1]:=fscanf(data_file, "%e").[1];
asv[i,1]:=fscanf(data_file, "%e").[1];
ash[i,1]:=fscanf(data_file, "%e").[1];
vmc[i,1]:=fscanf(data_file, "%e").[1];
s[i,1]:=fscanf(data_file, "%e").[1];
n:=n+1:
od:
fclose(data_file):
s:= $\theta$;
ds:=stackmatrix(asv,ash):
6. Compute Predicted Forward EC\textsubscript{a} (EC\textsubscript{a\theta}) Values $pds=\{pasv | ash\}$

\begin{verbatim}
> pds:=evalm(K&*s):
pasv:=matrix(7,1):
pash:=matrix(7,1):
for i from 1 to n do
  pasv[i,1]:=pds[i,1]:
pash[i,1]:=pds[i+n,1]:
od:
evalm(pasv):
evalm(pash):
wfv:=fit[leastsquare[[x,y],y=a*x+b,\{a,b\}]]([[seq(asv[i,1],i=1..7)],[seq(pasv[i,1],i=1..7)]]):
wfv:=unapply(rhs(wfv),x):
wfh:=fit[leastsquare[[x,y],y=a*x+b,\{a,b\}]]([[seq(ash[i,1],i=1..7)],[seq(pash[i,1],i=1..7)]]):
wfh:=unapply(rhs(wfh),x):
> for i from 1 to n do
ds[i,1]:=wfv(asv[i,1]):
ds[i+n,1]:=wfh(ash[i,1]):
od:
\end{verbatim}

7. Compute VMC Depth Profile from EC\textsubscript{a\theta} - Direct Inversion with Generalised Inverse

\begin{verbatim}
> ps:=evalm(inverse(transpose(K)&*K)&*transpose(K)&*ds):
\end{verbatim}

8. Sensitivity / Condition Number

\begin{verbatim}
> eig:=eigenvals(transpose(K)&*K);
CN:=max(eig)/min(eig):
\end{verbatim}

9. Least Squares Formulation

\begin{verbatim}
> ss:=matrix(7,1,[s0,s20,s40,s60,s80,s100,s120]);
ll:=matrix(5,7,[[1,-2,1,0,0,0,0],[0,1,-2,1,0,0,0],[0,0,1,-2,1,0,0],[0,0,0,1,-2,1,0],[0,0,0,0,1,-2,1]]);
\end{verbatim}
ll:=matrix(7,7,[[1,0,0,0,0,0,0],[0,1,0,0,0,0,0],[0,0,1,0,0,0,0],[0,0,0,1,0,0,0],[0,0,0,0,1,0,0],[0,0,0,0,0,1,0],[0,0,0,0,0,0,1]]);

> evalm(ll&*ss);
> evalm(transpose(ll&*ss)&*(ll&*ss))[1,1];
> TR:=proc(alpha)
  global s0,s20,s40,s60,s80,s100,s120,err,xx,yy:
  s0:='s0':=s0:s20:='s20':=s20:s40:='s40':=s40:s60:='s60':=s60:s80:='s80':=s80:s100:='s100':=s100:s120:='s120':
  err:=evalm(transpose(K&*ss-ds)&*(K&*ss-ds))[1,1]+alpha*evalm(transpose(ll&*ss)&*(ll&*ss))[1,1]:
  sol_s:=solve({diff(err,s0)=0,diff(err,s20)=0,diff(err,s40)=0,diff(err,s60)=0,diff(err,s80)=0,diff(err,s100)=0,diff(err,s120)=0},{s0,s20,s40,s60,s80,s100,s120}): assign(sol_s);
  sol_s:=matrix(7,1,[s0,s20,s40,s60,s80,s100,s120]);
  yy:=evalm(transpose(K&*sol_s-ds)&*(K&*sol_s-ds))[1,1]:
  xx:=evalm(transpose(ll&*sol_s)&*(ll&*sol_s))[1,1]:
  dist:=xx^2+yy^2:
> alpha:=rhs(min_alpha[2][1]);
  alpha:=0.1;
TR(alpha);

s0:='s0':=s0:s20:='s20':=s20:s40:='s40':=s40:s60:='s60':=s60:s80:='s80':=s80:s100:='s100':=s100:s120:='s120':
  err:=evalm(transpose(K&*ss-ds)&*(K&*ss-ds))[1,1]+alpha*evalm(transpose(ll&*ss)&*(ll&*ss))[1,1]:
  sol_s:=solve({diff(err,s0)=0,diff(err,s20)=0,diff(err,s40)=0,diff(err,s60)=0,diff(err,s80)=0,diff(err,s100)=0,diff(err,s120)=0},{s0,s20,s40,s60,s80,s100,s120}): assign(sol_s);
  sol_s:=matrix(7,1,[s0,s20,s40,s60,s80,s100,s120]);
  transpose(s);transpose(sol_s);
  alpha:='alpha':
  s0:='s0':=s0:s20:='s20':=s20:s40:='s40':=s40:s60:='s60':=s60:s80:='s80':=s80:s100:='s100':=s100:s120:='s120':
> evalm(transpose(K&*sol_s-ds)&*(K&*sol_s-ds))[1,1];
evalm(transpose(ll&*sol_s)&*(ll&*sol_s))[1,1];
>P1:=pointplot([seq([h[i,1],sol_s[i,1]],i=1..n)],view=[0..140,0..100],color=red,title=`Moisture Content Depth Profile`,labels=`Depth (cm)` ,`Moisture Content`
Appendix

\[ P2 := \text{pointplot([seq([h[i,1],s[i,1]],i=1..n]),view=[0..140,0..100],color=blue,title=\text{`Moisture Content Depth Profile`},labels=[\text{`Depth (cm)`},\text{`Moisture Content (%)`}],labeldirections=[\text{HORIZONTAL},\text{VERTICAL}],axes=boxed,symbol=box,symbolsize=12,legend=\text{`Measured Moisture Content`}); } \]

display({P1,P2});
end: