A Method For The Simple And Rapid Determination Of Deep Drainage And Its Requirement

Erin Trainer

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Section 1: Review

ESTIMATING DEEP DRAINAGE AND ITS REQUIREMENT

Erin Trainer

University of Sydney, Faculty of Agriculture Food and Natural Resources, Sydney, NSW, Australia

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As world population continues to grow, and the demand for agricultural products increases, the efficient use of water resources is becoming an increasingly pressing issue. Inefficient use of water is recognized as a significant factor contributing to dryland and irrigation salinity. The mechanism through which this occurs is deep drainage. Without deep drainage excess salts accumulate in the root zone, resulting in soil salinity. However, if deep drainage occurs above this requirement there is the potential for the groundwater table to rise, mobilising salts and bringing them to the surface, also resulting in salinity. There must be a balance, where the amount of deep drainage occurring is in line with this requirement.

The allowable root zone salt concentration will determine the necessary deep drainage; expressed as a leaching requirement. Many methods have been developed to determine the leaching requirement; these vary in complexity and accuracy.

Deep drainage can be estimated through the use of physical and chemical techniques including lysimetry, water flux meters, Darcian flux calculations, soil water balance, chloride mass balance and chemical tracers. The more accurate of these methods are generally time-consuming, expensive, require a high degree of technical knowledge and are unable to describe soil spatial variability. This highlights the need for a method, which is fast, low cost, easy to use, able to account for the spatial variability of soil properties and achieve a reasonable degree of accuracy.

I. INTRODUCTION

In Australia, immediate action is needed on water quality and salinity issues. At least 2.5 million hectares, 5% of currently cultivated land, is affected by dryland salinity. Soil salinity can cause reduced plant growth and yield, and, in some cases, total crop destruction (Qadir *et al.*, 2000). Salinity is not just an agricultural issue; salinity is also detrimental in urban areas, causing damage to roads and buildings (Anonymous, 2004b).

Salinity refers to the presence of soluble salts in the soil or soil solution. Throughout the world areas of naturally formed saline soil can be found, especially where precipitation is low and evaporation high. This is termed primary salinity. Soil which is saline as a result of human activity is termed secondary salinity (Rose, 2004).

There are two main types of secondary soil salinity: irrigation and dryland. Dryland salinity is caused by the substitution of deep-rooted vegetation for agricultural crops. The latter use less water, causing a rise in the water table. Irrigation salinity results when applied irrigation water is the cause of an elevated groundwater table (Anonymous, 2004b).

There is a large degree of concern in Australia about irrigation-dependent agriculture (Willis *et al*., 1997). This concern is well founded considering that many ancient civilisations, including those of South America and Mesopotamia, employed irrigation which resulted in soil salinisation and ultimately the demise of their agricultural practices and an end to their civilisations (Szabolcs, 1989).

Groundwater (defined as water that is stored below the earth's surface) accounts for 22% of the freshwater on earth. It comprises a large portion of the freshwater supply of many regions in the world (Hillel, 1998). To ensure replenishment of the groundwater supply, deep drainage is required. Deep drainage is defined as the water flux below the depth to which plant roots extract water (Bond, 1998). A certain quantity of drainage, known as the leaching requirement, is needed for the removal of salts accumulated in the root zone. However, if excessive deep drainage occurs irrigation salinity may result, hence a balance must be found.

From the 1970s to the 1990s there was a 42% increase in world population, and a 300% increase in water usage (Wood, 2003). This population increase is expected to continue with Lutz *et al* (2001) predicting a 39% rise before the end of the 21st century. This will place an even greater strain on worldwide water resources. Coping with this will require more efficient use of water resources.

Attempting to meet the growing demand for agricultural products, arid and semi-arid areas, previously not cultivated, are being irrigated and brought into use. However, current irrigation practices are frequently inefficient, deliberately applying excess water to ensure leaching and prevent of salt accumulation. This results in water loss, fertiliser loss, land degradation, and an increasing need for drainage systems (Alsaeedi and Elprince, 1999).

Understanding the process of soil water flow is central to using water more efficiently and improving salinity management. Soil water plays a vital role in agricultural systems and is perhaps the most important factor in determining plant growth and crop yields, also playing a large role in practices which degrade soil quality (DeJong and Bootsma, 1996).

Agriculture is not the only reason an understanding of soil water flow is necessary; it is also required for effective environmental preservation. Disposal of waste, such as effluent, from dairy farms commonly occurs through application to the land. Without an understanding of soil water flow, lowering of soil quality and pollution of ground and surface water bodies can occur (Houlbrooke *et al*., 2004).

The following review considers deep drainage and why it is important; how much is required and what techniques can be used for its estimation. The estimation of this quantity is generally time-consuming, expensive, or of low accuracy. Many methods are available, including physical, chemical and modelling techniques. Physical methods include lysimetry, water flux meters and Darcian flux calculations. Chemical methods include the use of tracers and the chloride mass balance. Models incorporating the use of physical and chemical principles are widely used and have varying degrees of complexity and accuracy.

II. THE HYDROLOGICAL CYCLE

Figure 1 depicts the hydrological cycle, which is natural flow of water in the vicinity of the earth's surface. This is the continuous c irculation of water between the ocean, atmosphere and land (Freeze and Cherry, 1979). Evaporation of water, predominantly from oceans, can be taken as the starting point for this cycle. This water vapour is transported within the atmosphere until the conditions are right for precipitation. Precipitation may go back to the ocean directly or be transported onto land. Once on land, this water may be evaporated directly back to the atmosphere, or become surface runoff and find a path to a body or flow of water. It can penetrate the soil and be evaporated or taken for use by plants. Alternatively, it may enter the soil and be retained, or continue to flow downwards as drainage.

The driving force for this water movement is energy provided by the sun. This movement is not consistent and has a large amount of associated variability in both space and time (Zhang, 2002).

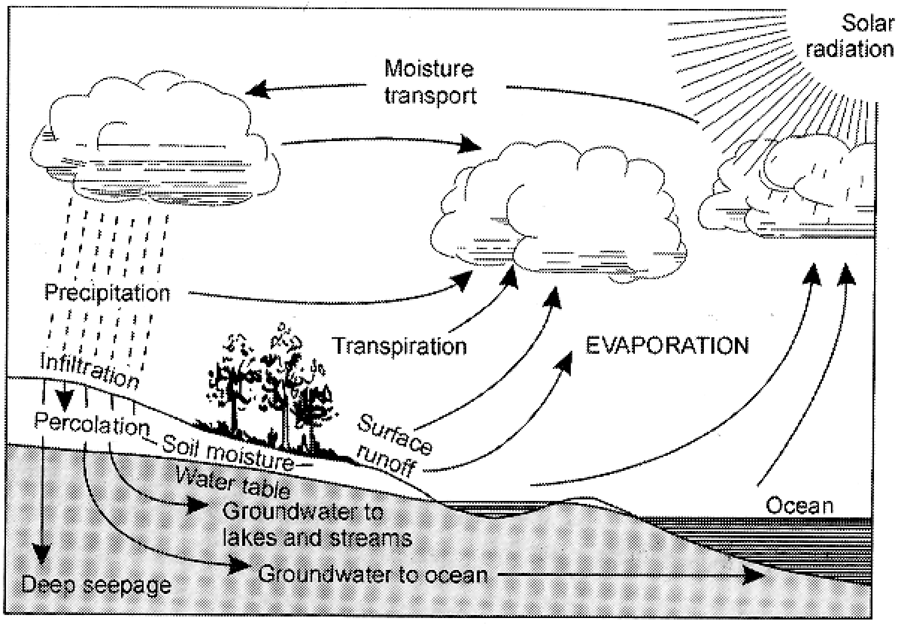

Figure 1 The hydrological cycle (Zhang, 2002)

III. WATER BALANCE

The various fates of precipitation can be used in the description of soil water flow. For a full description of this flow all of the hydrological inputs and outputs, and the variations in water storage in the soil, must be accounted for. To achieve this, a soil water balance, or budget, is used which can be derived from the law of conservation of mass, [Eq. (1)] (Hillel, 2004). A water balance can be derived for a surface water body, watershed, aquifer system or vadose zone.

The law of conservation of mass [Eq. (1)] states that matter can be neither created nor destroyed but can only change from one state or location to another:

$$\frac{accumulation}{time} = \frac{input}{time} - \frac{output}{time} \qquad (1)$$

When considering soil water flow, the vadose zone water balance is of concern. The vadose zone is defined as the geologic media between land surface and the regional water table (Fig. 2) (Stephens, 1996).

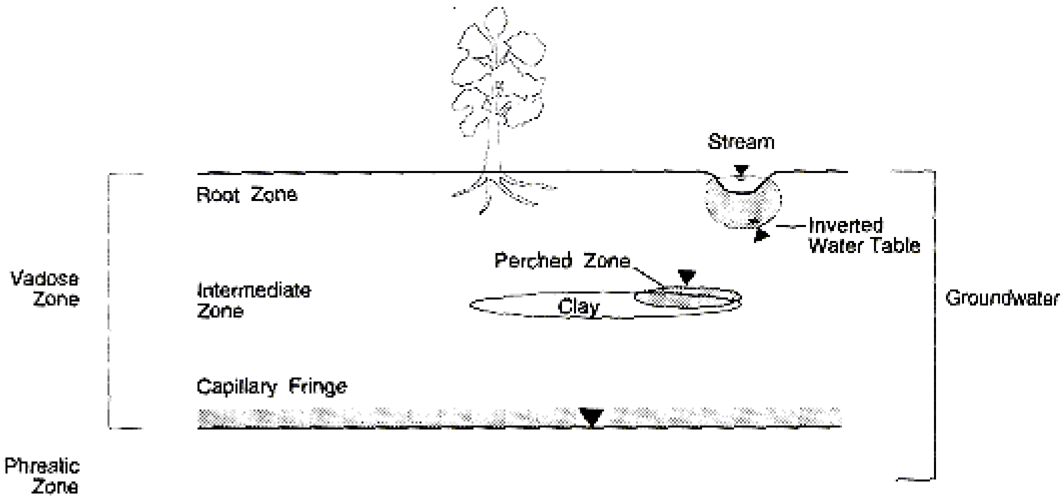

Figure 2 The vadose zone (Stephens, 1996)

The water balance for this area is illustrated in Fig. 3 and given in Eq. (2):

$$\Delta S = I + P - R - ET - DD \tag{2}$$

where ΔS is the change in soil water storage, I is the applied irrigation, P is precipitation, R is runoff, ET is evapotranspiration and DD is deep drainage.

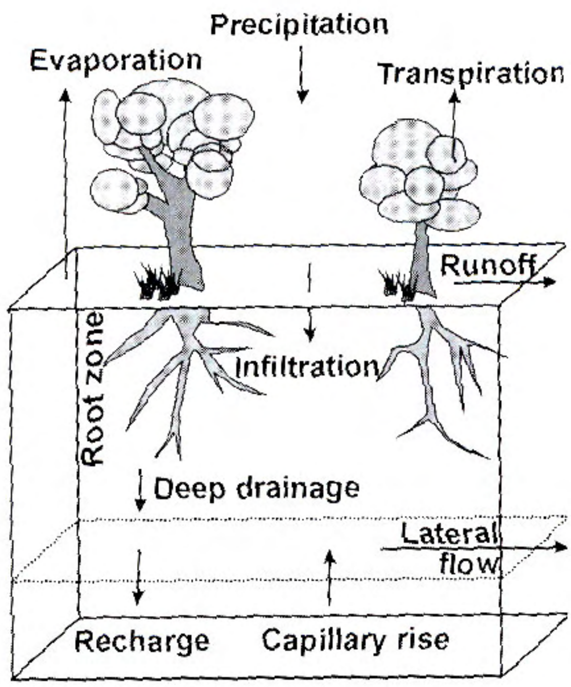

Figure 3 The soil water balance over a root zone (Zhang, 2002)

An ideal agricultural model would combine high productivity and efficiency, whilst being environmentally and economically sustainable. The ability to control the water made available to plants plays an important role in being able to achieve this. For this to be possible an understanding of the soil water balance is required.

Most modern agriculture uses some form of irrigation. As such, water is important from both an ecological and economic point of view. This is especially the case in Australia, where water is a scarce resource and its misuse has the potential to make land unsuitable for agricultural use. Other effects of misuse include the pollution of water reserves and soil salinisation.

To understand the workings of the soil water balance, and so enable implementation of efficient water management, the mechanisms of soil water flow must first be understood.

IV. SOIL WATER FLOW

A. DARCY'S LAW

Soil physical principles can be used to describe and understand the flow of water through soil. The total soil water potential is the sum of the gravitational, pressure (or matric) and osmotic potentials (Marshall *et al.*, 1996). When this potential differs between two points there is a tendency for water to move from high potential to low potential.

The gravitational potential (z) is associated with the energy change due to lifting the water in a gravitational field (Kutilek and Nielsen, 1994). When water enters the soil it is subject to attractive capillary and absorptive forces. The energy change due to these forces, which is by definition negative, is known as the matric potential (h). This is sometimes referred to as the pressure potential (Φ), although the pressure potential additionally accounts for the state where a positive potential arises due to submergence in the water table.

The addition of solutes to water also provides an energy difference, known as the osmotic potential (Marshall *et al.*, 1996). Various methods can be used in the field and laboratory to measure soil-water potential including tensiometers, piezometers and psychrometers (Kutilek and Nielsen, 1994).

The first quantitative description of water flow through a porous medium was developed by Henri Darcy in 1856, where a linear relationship between the flux density (the volume of water flow per unit cross-sectional area) and the potential gradient (change in potential per unit distance of flow) was discovered for saturated flow (Kutilek and Nielsen, 1994; Marshall *et al.*, 1996). The saturated hydraulic conductivity (K_s), constant at saturation, is the proportionality constant between these, and describes the soil's ability to conduct water.

The modern expression of Darcy's Law is:

$$q = \frac{Q}{A} = K_s \frac{\Delta H}{\Delta z} \qquad (3)$$

where q is the flux density of water (ms^{-1}), K_s is the saturated hydraulic conductivity (ms^{-1}), $\frac{\Delta H}{\Delta z}$ is potential gradient, or change in potential H, per change in height z and Q is the volume of water discharged (m^3/s) through A, the cross-sectional area of flow (m^2).

Darcy's Law can only be used to describe steady-state water flow and is valid only where flow is laminar, the soil saturated and rigid, and the flux and temperature are constant. Steady state flow described the state where the water volume in equals the water volume out.

Fluid flow may be laminar or turbulent. Laminar, or streamline, flow describes the occurrence of streams flowing in parallel, which do not interact. This is the dominant flow regime for low velocities and narrow tubes. Turbulent flow occurs as flow velocity increases, characterised by eddies and vortices, causing dispersion across the tube area (Coulson and Richardson, 1996). Due to the small size of soil pores, flow through soil is generally laminar (Hillel, 2004).

The majority of the processes in the field that involve soil and water, including the flow of water and solutes below the root zone, occur when the soil is unsaturated (Hillel, 1998). Darcy's law is not directly applicable to these circumstances.

In 1907 Buckingham identified that the hydraulic conductivity (K) is water content or potential dependent, and illustrated that Darcy's law could be taken further and also applied to unsaturated, but still steady-state, flow (Smiles, 2000; Stephens, 1996). This can be expressed as:

$$q = -K(h)\frac{dH}{dz} = -K(h)\left(\frac{dh}{dz} + 1\right) \qquad (4)$$

where H = h + z, and h is the matric potential.

Eq. (4), known as the Darcy-Buckingham equation, calculates the flux between two points if the potentials at these points and the hydraulic conductivity of the soil at this potential is known.

B. RICHARDS' EQUATION

Alone, Darcy's Law is able to calculate the flux under saturated, constant conditions. The Darcy-Buckingham equation allows flux to be calculated for unsaturated, but still constant, flow. More common are conditions of transient flow, where the extent and direction of flow is changing over time (Hillel, 2004).

The law of conservation of mass [Eq. (1)] applies not only to the whole soil but also to any definable volume. When no sinks for water are present, the change in soil water content, for a given volume, is equal to the difference between the water fluxes in and out of this volume. In 1931 Richards combined this idea of continuity with Darcy's law to describe the unsaturated flow of water, in three dimensions, through a soil volume given by:

$$\frac{\partial \theta}{\partial t} = \nabla.\left[K(h)\nabla h\right] + \frac{\partial K}{\partial z} \qquad (5)$$

where ∇ represents the requirement of differentiation in three physical dimensions and θ is the volumetric soil moisture content.

In the horizontal (x) dimension, this simplifies to:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x}\left(K(h)\frac{\partial h}{\partial x}\right) \qquad (6)$$

The general form of Richards' equation is given in Eq. (6), but many expressions for this equation can be used for calculating water flux during periods of unsteady flow (Hillel, 1998). This is a partial differential equation where the relationship between hydraulic conductivity and moisture content is highly non-linear; only for specific initial and boundary conditions can it be solved analytically. This results in numerical techniques being required to obtain solutions which are characteristic of field conditions (DeJong and Bootsma, 1996).

The water flux through soil is dependent on hydraulic conductivity, which is not constant. It is related to the structure and texture of the soil, being highly dependent not only on porosity but also the sizes of the conducting pores, and the tortuosity (pore geometry) (Marshall *et al.*, 1996). In the surface zone, hydraulic conductivity can vary over more than six orders of magnitude as the soil moisture content changes. This variation is less in the vadose zone but the margin for error here is still significant (Gee and Hillel, 1988).

Soil texture and structure are features of the soil that exhibit a high degree of spatial variability. This results in the hydraulic conductivity being variable not only temporally but also spatially. Hence the vadose zone, a heterogeneous region where flow is transitory, is often not stable and typically is spatially inconsistent (Gee and Hillel, 1988).

C. BYPASS FLOW

The theoretical aspects of water flow have been described. Implicit in this theory is that small pores will fill before larger pores, as a result of capillarity. In practice this is often not the case, a result of the presence of pores with physical properties that are significantly different to the surrounding soil, known as macropores. Their incidence and magnitude often lead to their exclusion when measurements are made (Bond, 1998).

Piston flow, where a front of water moves through the soil absorbing nutrients as it goes, is commonly used to describe water flow (Ryan and Noonan, 1995). Soils that have a well defined structure generally allow faster transport of solutes relative to the piston flow description than soils with poorer structures (Rice *et al*., 1986). It has been found that these structured soils contain macropores, which offer a preferred path for the soil solution to follow (Kanchanasut *et al*., 1978). This is often referred to as bypass flow.

Macropores are pores which are appreciably larger than the standard packing pores in a soil, and may be cylindrical channels or planar voids depending on the process of formation (Bouma, 1983). Dyes have been used to identify flow paths of water through soil, this can be seen in Fig. 4, where a soil cross section reveals that flow has not occurred uniformly, but has followed preferential paths.

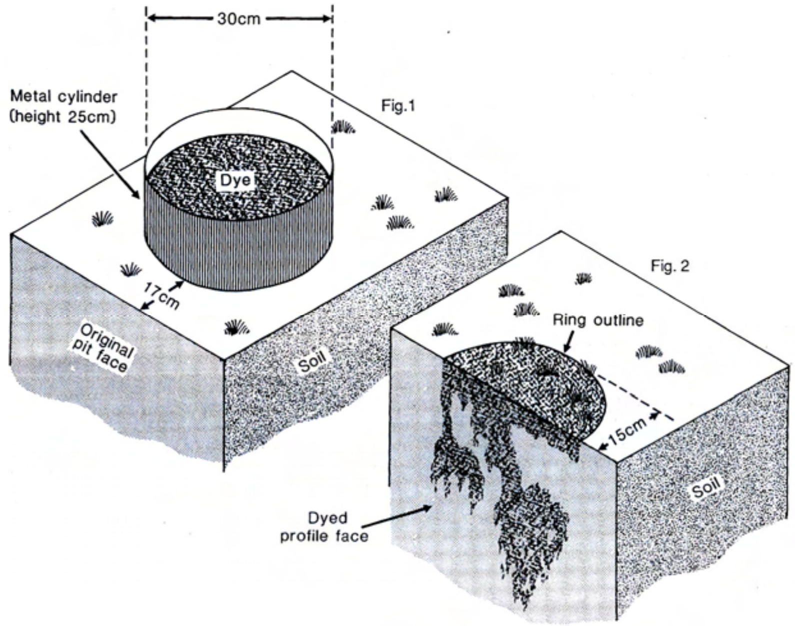

Figure 4 Path of dyed water flow through the soil (Ryan and Noonan, 1995).

When considering water movement through the soil, continuity of these pores is more important than size. These pores not only act as highly conducting for flow, but also as barriers to flow in other directions, meaning that water flow patterns in soils containing macropores are complex and difficult to quantify (Bouma, 1983).

D. Swelling Soils

So far considered is that hydraulic conductivity is a constant parameter (K_{sat}) once the soil is saturated. However, certain soils have the ability to shrink and swell over time along with changing moisture content. In swelling soils, as water is added pore space is not a temporally constant property, resulting in temporal variability of K_{sat}. Bouma and Wosten, (1979) observed a decrease in K_{sat} over a period of several months as the soil being studied continued to swell with the addition of water.

Factors that govern the swelling process include clay content, the type of clay minerals present and the concentration of solutes in the soil solution (Bouma, 1983). Swelling soils have a high content of swelling clay minerals, which increase up to 50% in size on the addition of water. Upon drying, tapered cracks form, and can have a width of 10cm at the surface, and sometimes reach over 1.5m in depth (Hubble, 1972).

Flow of water in soils that change volume with moisture content was first described in 1923 by Terzaghi, and this required using a mass balance for both the soil material and soil solution (Smiles, 2000). Darcy's law can again be applied, but flow must be considered relative to the soil particles and not to a fixed point.

Due to the addition or removal of a volume of water, wet soil is vertically displaced with reference to a certain point, giving rise to an overburden (Hubble, 1972). In swelling soils this overburden pressure must also be considered.

V. DEEP DRAINAGE

A. DEFINING DEEP DRAINAGE

Deep drainage is defined as the volume of water flux passing below the depth at which plant roots extract water (Bond, 1998). This is known as the root zone, a particular depth which varies according to plant species, variety and level of development, as well as soil and water table conditions (Humphreys, 2003). Deep drainage contributes directly to groundwater recharge and will equal the recharge if, below the root zone, no sinks for water exist; this is only true under steady-state conditions, with deep drainage and recharge constant. Otherwise, drainage will be delayed depending on the volume of flux, the depth to the groundwater and the ability of the soil to store water, and can only be regarded as potential recharge (Bond, 1998).

Over a field, even within a few metres, the deep drainage rate can vary significantly (Johnston, 1987) as a result of varied water application, inconsistent soil properties, (Rice *et al*., 1986) and differences in the type of vegetation canopy (Finch, 1998).

Recharge or drainage can occur via two modes:

1. Continual diffuse recharge distributed over an area as a result of infiltration throughout the entire vadose, and
2. Transient or irregular recharge which bypasses the major portion of the vadose zone as a result of short-term water infiltration through distinct flow channels (Gee and Hillel, 1988)

The drainage rate is closely related to the hydraulic conductivity of the soil. Soil hydraulic conductivity, and hence deep drainage, is influenced by soil properties, including clay content, clay mineralogy, organic matter content, exchangeable sodium percentage (ESP) and bulk density. The influence of these factors on deep drainage is complex and interrelated (Shaw and Thorburn, 1985).

B. THE IMPORTANCE OF DEEP DRAINAGE

By gaining an understanding of the processes through which soil-water flow occurs it is possible to better understand how much of the water applied in irrigation is being delivered to plants.

Deep drainage is a key indicator in water use efficiency. If in excess, a scarce resource is being lost, whilst if not occurring in sufficient quantities salt accumulation may occur, also poor management practice.

Deep drainage plays a significant role in the hydrological cycle and its scale can have positive and negative repercussions on and off-site. Without deep drainage, water would accumulate in the soil, aeration would be limited and salts would not be leached from the root zone. Where irrigation occurs, particularly in arid and semi-arid regions, evapotranspiration (ET) causes salts to build up in the root zone (Qadir *et al*., 2000), leading to a reduction in crop yields (Shaw and Thorburn, 1985). Deep drainage is required for the removal of these salts.

Groundwater, an important resource, provides water for irrigation, industry and domestic use in many parts of the world. For its future use, continually replenishment is required (Freeze and Cherry, 1979). Deep drainage is the main contributor to groundwater recharge.

Water that has infiltrated the soil may contain solutes. These solutes can be removed by plants, reacted or retained in the soil or transferred to the gaseous phase and removed to the atmosphere. If not they may be transported with deep drainage and so contribute to groundwater pollution (Pepper *et al*., 1996).

Although some deep drainage is required for efficient irrigation management, excessive drainage can lead to rising water tables, which may contain saline water, and cause secondary soil salinisation. Excess drainage is also a water loss, especially during periods of high rainfall, low ET and small soil water storage capacity. Therefore, knowledge of deep drainage is important for the evaluation of a sensible irrigation schedule. In many areas, the lack of fresh water means irrigating with saline water is the most feasible option. Again, for this to be considered practical, adequate leaching is required.

Ideally, a system will maximise biological productivity within the limit of water availability, consequently diminishing the water losses from runoff and deep drainage (Pratley, 1987).

C. Measurement Difficulties

Logistically, the direct measurement of actual deep drainage under field conditions is exceedingly difficult (Shaw and Thorburn, 1985). This is primarily due to deep drainage occurring below ground in areas not accessible to measurement. To gain access requires removing soil, consequently disturbing its natural state, altering natural water flow patterns and the deep drainage that occurs. This results in a reliance on estimating deep drainage, for which the methods available are numerous. The accuracy, ease of use and applicability of these methods varies greatly.

Generally, for the estimation of deep drainage soil hydraulic properties must be measured. The measurement of these properties also has considerable associated difficulties (Shaw and Thorburn, 1985)

In arid and semi-arid areas the difficulties become even greater when using methods based on physical parameters. Fluxes are small and not easy to detect. The high temporal variability requires measurements spanning several years for mean estimates to be obtained. This and the spatial variation attributable to the local topography and variable soil texture necessitates numerous sampling sites (Allison *et al.*, 1994), making the process time-consuming and expensive.

VI. MEASURING DEEP DRAINAGE

Techniques for estimating deep drainage may be divided into physical and chemical methods (Allison *et al.*, 1994). Physical methods may be further divided into direct and indirect methods. Models are also used, these incorporate physical or chemical techniques in the prediction of deep drainage.

A. PHYSICAL METHODS

Physical methods use the principles of soil physics to estimate deep drainage. This includes analysing water flows and measuring soil physical properties, which affect water flow.

1. Direct

Direct methods are those that attempt to actually measure the volume of water passing below the root zone. These methods include lysimetry and flux meters.

a. Lysimetry

A lysimeter is a device in which a container encloses a volume of soil to isolate it hydrologically from the surrounding soil on all sides and at its base Fig. 5 (Bond, 1998).

(Rosenberg, 1983) states that 'lysimeters provide the only direct measure of water flux from a vegetative surface. As such they provide the standard against which other methods can be tested and calibrated.' The containers are filled with soil and placed in the field to represent prevailing conditions. Water that passes through the lysimeter is collected and its volume measured.

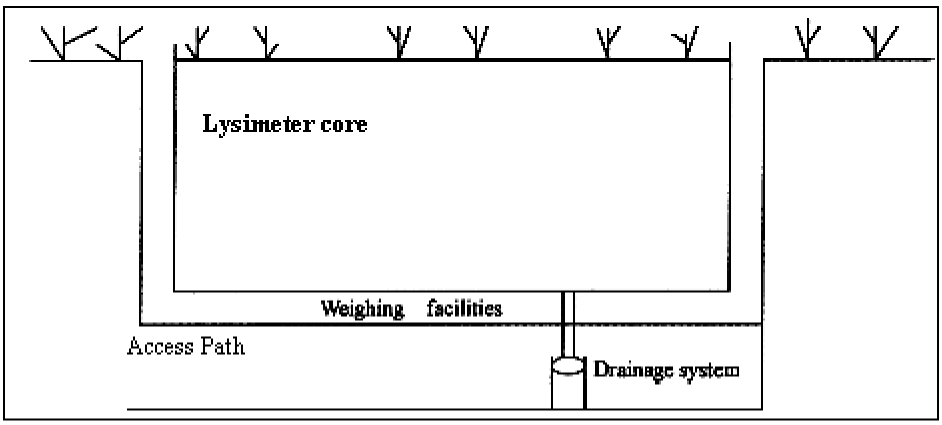

Figure 5 Schematic diagram of a lysimeter with weighing and drainage systems (adapted from (Liu *et al.*, 2002))

Lysimeters have the ability to measure all components in the water balance. They are most frequently used to measure evapotranspiration and drainage. Some are equipped with a weighing device and drainage system permitting the simultaneous measurement of these terms (Bond, 1998).

For the measurement of drainage, weighing and non-weighing lysimeters are used. Weighing lysimeters have scales incorporated into their design and determine drainage based on weight differential. Non-weighing, or interception lysimeters rely on the collection of water through a drainage system to make measurements.

Although lysimeters as deep as 18m have been constructed, due to cost and logistics their size is often limited to a few square meters with depth rarely greater than 3m (Allison *et al.*, 1994). Although considered quite accurate, there are many disadvantages of their use. The cost of construction and maintenance is substantial and their installation disturbs soil and vegetation in close proximity. The lysimeter bottom boundary conditions are different to that of the prevailing soil conditions, affecting water flow (Gee and Hillel, 1988). Their permanent and immobile nature, hinders their ability to describe spatial variability (Allison *et al.*, 1994).

b. Water Flux Meters

Soil water flux meters involve the interception of water flux within the soil and the calculation of its extent through the measurement of hydraulic head loss occurring across a known hydraulic resistance (Van Grinsven *et al.*, 1988).

Van Grinsven *et al.* (1988) described a water flux meter (Fig. 6) consisting of a porous plate whose suction is adjusted to maintain an identical matric potential to that of the surrounding soil, through which the flow of water is measured. All adjustments are made automatically from a remote source. This instrument is not a passive element, incorporating the ability to pull water into the device despite the existence of a larger hydraulic resistance.

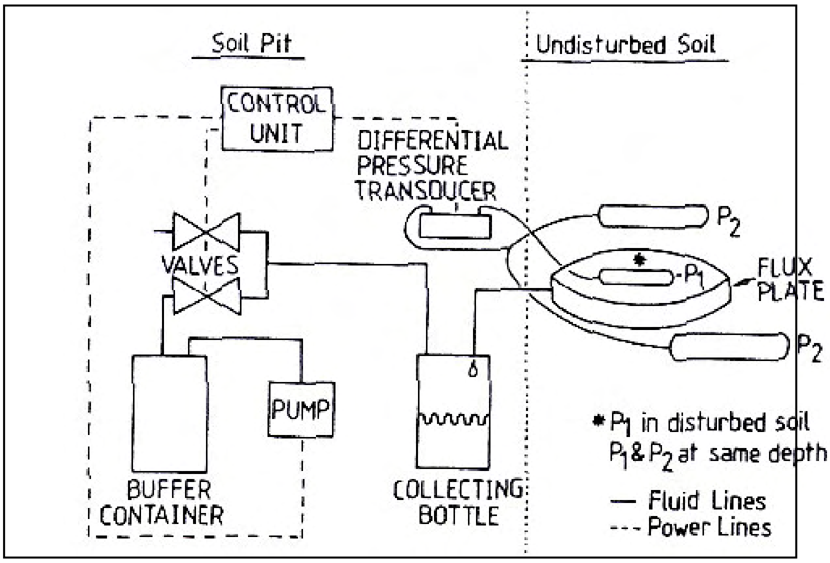

Figure 6 Schematic diagram of the water flux meter described by Van Grinsven *et al.* (1988)

The passive flux meter (PFM) is another example of a water flux meter. It has the ability, when placed in an area of flow, to simultaneously measure the cumulative dissolved solute fluxes and water fluxes occurring. This is possible through the inclusion of a self-contained device, which passively measures the volume of, without hindering, fluid flow. The device includes permeable hydrophobic and hydrophilic sorbents, which retain solutes in the solution that is passing through the device. The device also contains known amounts of soluble tracers, which are used to estimate the total flow through the device. Applications other than the measurement of recharge include estimating the arsenic contamination of groundwater (Clark *et al.*, 2005).

Masarik *et al.* (2004) developed a low-power automated equilibrium tension lysimeter (AETL), which can be used as a water flux meter. The success of this device depends on the lysimeter and soil potential being maintained in equilibrium, achieved through automated adjustment of lysimeter tension. The device also has the ability to measure the soil water matric potential, improving accuracy of the drainage measurement.

Although offering direct and relatively accurate measurements, water flux meters are not as commonly used as other methods for recharge estimation. This can be attributed to cost and the time consuming nature of installing and monitoring these instruments. They also require a relatively high degree of operator training and, on installation, disrupt the soil and the natural water flow paths within the soil (Bond, 1998).

2. Indirect

Physical indirect methods rely on the measurement or estimation of soil physical parameters, which along with soil physical principles, can be used to estimate the potential or actual deep drainage. Allison *et al*. (1994) divided these methods into three categories:

a. Soil water balance
b. Zero-flux plane method
c. Darcian flux calculations

a. Soil Water Balance

This is a frequently used method based on the soil water balance Eq. (2). Deep drainage is calculated by the rearrangement of this equation and the measurement or estimation of all other terms. Features of the soil water balance are non-uniform initial water contents and time dependent boundary conditions, so that numerical methods are often used to obtain solutions (DeJong and Bootsma, 1996).

Problems arise as measurements of soil water balance components are frequently inaccurate (Gee and Hillel, 1988). The error in the drainage term is determined from the errors of the input terms (Willis *et al*., 1997). As these errors accumulate the error in the drainage term becomes relatively large (Rice *et al*., 1986). The timescale of measurement can also impact on the results, and often leads to an underestimation of recharge (Sophocleous and Perry, 1985).

The least challenging components to obtain with relative accuracy are the irrigation and precipitation terms. The quantity of irrigation applied can be easily controlled, and the value of precipitation can be readily measured in the field, or is easily obtainable (Hartmann *et al*., 1980).

The precipitation term is frequently the largest term of the water balance, and therefore exerts a large influence over the reliability of the balance. When calculating average precipitation in an area, the spatial and temporal distribution of precipitation, the variability of rainfall events, and the data that is available must all be accounted for (Xu and Singh, 1998).

Evapotranspiration (ET) is usually the second largest term, and is more challenging to either measure or estimate. ET can be measured using a lysimeter (Bond, 1998) or estimated using complex models, such as the Penman-Monteith model, which estimates the potential evapotranspiration (PET). Commonly the actual ET (AET) is then calculated as a function of the PET and the dryness of the soil (Xu and Singh, 1998).

Runoff is water not absorbed by the soil or accumulated at its surface which runs downslope. Runoff can be divided into various components, which are analysed depending upon the nature of the area and the time frame of interest. For water balance modelling, (Dyck, 1985) has proposed four components:

- surface flow
- fast interflow

$\Rightarrow$ Fast components

- slow interflow
- baseflow

$\Rightarrow$ Slow components

Soil moisture content can be measured with a variety of instruments. These include neutron probes, capacitance probes (C-probes) and time domain reflectometry (TDRs). Measurement of the soil dielectric constant provides the basis for the functioning of the capacitance and TDR devices (Chanzy *et al.*, 1998). Neutron probes and TDRs permit repeatable in-situ determinations on undisturbed soil.

Zhang (2002) has assumed $\Delta S = 0$ on an annual basis. However for shorter time periods the change in water storage must be measured as it may be important relatively.

Hartmann *et al.* (1980) measured all of the terms in a soil water balance for a bare soil and a cropped soil to compare the effects. They determined evaporation and drainage terms through the use of the zero flux plane method. No irrigation was applied and no runoff occurred. The change in soil water storage was calculated with the use of neutron probes and precipitation was obtained from meteorological data.

b. Zero Flux Plane Method

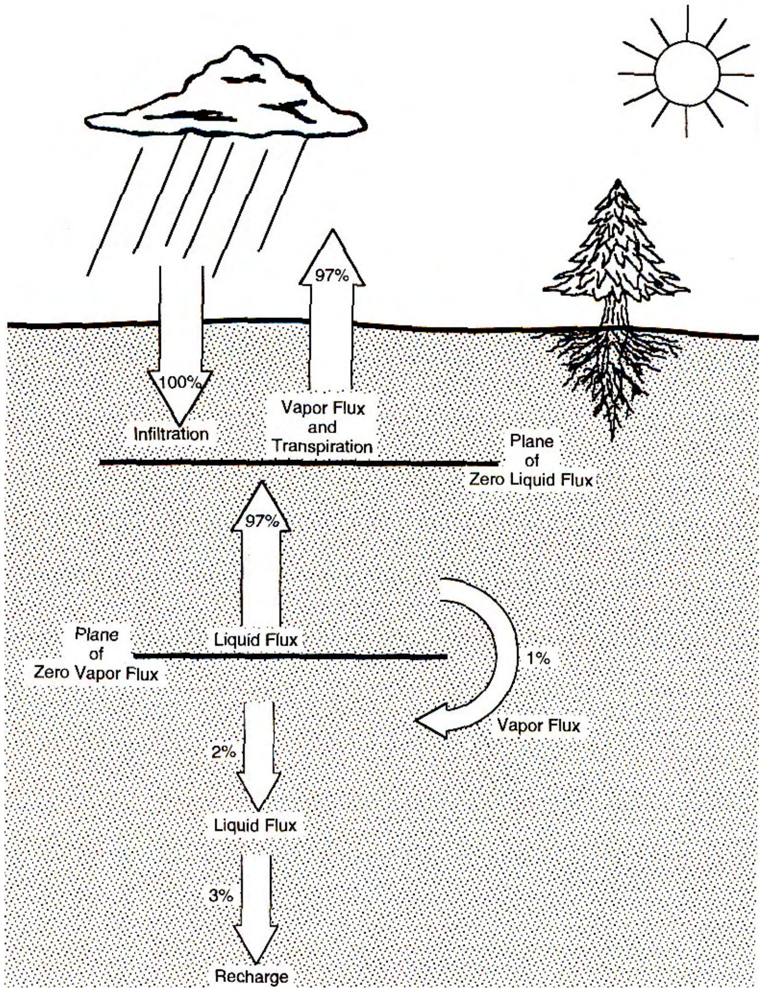

Figure 7 The plane of zero flux (Stephens, 1996)

Giesel (1970) first suggested the plane of zero flux method. After an infiltration event some water will flow upwards due to evapotranspiration and some will flow down, due to potential differences. There will be a particular depth, known as the plane of zero flux (Fig. 7), dividing these flows, where no flow occurs (Bond, 1998). The amount of drainage (downward flux) and evapotranspiration (upward flux) can be determined by the summation of changes in soil moisture content below or above this zero plane of flux respectively (Hartmann *et al*., 1980; Sophocleous and Perry, 1985).

The main problems are the occurrence of times when this method will fail. Under the influence of gravity it is possible for conditions of steady flow to exist. At this time the soil moisture content will remain constant, but drainage will be taking place. When this situation arises, a combination of neutron probe and tensiometer data can help to determine the drainage occurring (Gee and Hillel, 1988).

The method also proves unreliable at times of high infiltration. The hydraulic gradient becomes positive throughout the profile and no plane of zero flux exists. During these times a large amount of drainage, relatively, is occurring (Allison *et al*., 1994).

c. Darcian Flux Calculations

Flux calculations use Darcy's Law to determine the flux of water below the root zone. This requires knowledge of the hydraulic conductivity and potential gradient at a depth below the root zone.

Tensiometers can be used to determine the potential gradient throughout a profile (Gee and Hillel, 1988). Hydraulic conductivity, however, is one of the hardest parameters to establish in the field, and one of the most spatially variable (Nielsen *et al*., 1973) and its value is closely related to soil moisture content. For reliable estimates of deep drainage to be produced, accurate determination of the function relating these variables is required (Sophocleous and Perry, 1985).

Due to the difficulties in accurately determining the necessary parameters, this technique has several disadvantages. As water content decreases, hydraulic conductivity and matric potential vary almost exponentially, causing errors to increase in the flux calculations. This is particularly a problem when dry land soils are being investigated (Phillips, 1994).

Soils, which exhibit a uniform nature, may still demonstrate a large amount of lateral flow, indicating an anisotropic hydraulic conductivity. This is indicative of the bypass flow mechanism present in the recharge process, which contributes significantly to errors present in these calculations (Gee and Hillel, 1988).

Willis *et al*. (1997) used Darcian flux calculations, and found it yielded the lowest estimates of deep drainage compared to a chloride mass balance or a soil water balance. These calculations also contained the most significant amount of variation in the results.

B. CHEMICAL METHODS

Chemical methods utilise the presence of water-soluble substances, moving through the soil, as deep drainage occurs. By quantifying the flow of these substances, the volume of water occurring as deep drainage can be estimated.

In arid and semi-arid areas, water fluxes in the vadose zone are relatively small and hard to detect, the use of physical methods is fairly limited. Chemical methods are potentially more useful in these climates (Allison and Hughes, 1983).

Chemical methods also have the advantages of lower costs, and the ability to integrate the soil water flux in terms of its spatial and temporal variation (Cook and Herczeg, 1998).

1. Tracers

Tracers are waterborne substances that can be used to measure flux of water through the soil by monitoring their movement either visually or analytically (Gee and Hillel, 1988). Three types of tracers have been identified by Walker (1998):

a. Artificial Tracers: Applied by the researcher at or below the soil surface. The basis of their use being that the movement of the tracer is directly proportional to the movement of the water.

b. Historical Tracers: Have a high concentration at the soil surface due to some historical event, such as pollution, or a change in farming procedures. The principle is the same as that of artificial tracers. These include bomb tracers, which were present in rainfall at high concentrations after nuclear testing in the 1950s and 1960s.

c. Environmental Tracers: Exist naturally in the environment and their spatial pattern or mass balance in the landscape can be used for recharge estimation.

For a tracer to be considered suitable, it must exhibit certain traits. It should be mobile and soluble, such that it can move with water. It should be easily extractable and measurable (Walker, 1998). Conservative behaviour is required such that tracer flow rate and concentration do not vary as a result of interaction with the soil.

Knowledge of the rate of application is required for calculating the rate of movement through the soil. Also, artificial tracers should not be produced within the soil.

For a more detailed understanding of soil water flow, it can be useful to have a range of tracers which measure different time spans (Phillips, 1994). Allison *et al.* (1994) has suggested three methods for using tracer profiles in the vadose zone to estimate the recharge rate:

1- Using the position of the peak tracer concentration. If preferential flow is prevalent, an underestimation in recharge will occur.

2- Analysis of the shape of the tracer profile. This allows for flow mechanisms to be considered.

3- The total amount of tracer stored in the profile.

Rice *et al*. (1986) acknowledged that for an accurate determination of the field average of deep drainage, the frequency distribution of the individual measurements of tracer velocity must be determined.

a. Artificial Tracers

Using an artificial tracer means their one-off application and measurement over time of the tracer pulse. Application below the root zone and an adequate time gap between the application and measurement will yield the best results. Artificial tracers are most suitable for shallow rooting plants and areas where high recharge rates exist (Allison *et al*., 1994). They have the advantage that there is no uncertainty associated with the source of the tracer (Walker, 1998).

One of the simplest tracer methods is the use of dyes. Ryan and Noonan (1995) used red dye in water to infiltrate the soil. The dyed profile was then sliced vertically to expose the flow paths followed by the water (Fig. 4).

Chemical compounds are also used. Tilahun *et al*. (2005) applied potassium bromide such that bromide movement could be used to study the solute leaching through the soil.

b. Historical Tracers

Nuclear weapons testing in the 1950s and 1960s gave rise to an increase in the concentration of the radioactive isotopes ^{36}Cl and ^{3}H (tritium) in the atmosphere, resulting in their increase in rainfall and hence in soil water. They maintain the advantages of artificial tracers whilst the potential recharge required for their use is significantly less and longer time scales can be investigated.

Tritiated water acts almost identically to water, making it an ideal tracer (Phillips, 1994). However, its relatively short half life of 12.4 years, has meant it can no longer be used in the southern hemisphere (Walker, 1998). Tyler *et al*. (1992) successfully used tritium to estimate potential groundwater recharge.

^{36}Cl was produced incidentally by sea level atmospheric nuclear weapons testing of the 1950s. Its half-life is $3x10^6$ years and so will be useful for many years as a tracer. ^{36}Cl is not a preferred tracer when recharge is small in comparison to precipitation, as analysis of background concentrations is difficult (Allison *et al.*, 1994). Analysing samples for ^{36}Cl is expensive, and frequently where ^{36}Cl can be used the chloride ion (Cl^-), which is significantly cheaper, can also be used (Walker, 1998).

The use of nitrate (NO_3^-) in agricultural production has been increasing since the 1950s, resulting in an increase of its leaching below the root zone. The concentration of NO_3^- at depth, can then be used to indicate the position of the profile of recharge since the time of this increased use is a known factor (Allison *et al.*, 1994).

c. Environmental Tracers

Environmental or natural tracers are advantageous in their representation of a spatially uniform application and ability to assess the soil water flow over large time scales (Phillips, 1994). The results will be influenced by the mechanism used to describe flow and replication is usually lacking due to cost and logistical difficulties (Allison *et al.*, 1994).

^{14}C, ^{15}N, ^{18}O, ^{2}H, ^{13}C, and Cl^- are the most commonly used natural tracers. Meteoric chloride has been used in many studies (Allison and Hughes, 1983) and has the advantage of being omnipresent in the ground and soil water. It is highly soluble, only precipitates at high concentrations, and infrequently substitutes in soil minerals. The main difference between the geochemical cycle of chloride and water is that water is removed from the soil through evapotranspiration, whereas chloride, having a low volatility, remains and is concentrated (Heczeg and Edmunds, 2000).

For studies using chloride to be successful, the contributions from lithographic sources must be negligible or able to be estimated (Allison and Hughes, 1983). The main source of chloride is from sea salts, which is constantly deposited on land, either dissolved in precipitation or as dry fallout (Phillips, 1994).

Since ^{3}H, ^{2}H and ^{18}O are a part of the water molecule they will provide the most accurate results. However, having a lower volatility than water, evapotranspiration (ET) brings about their enrichment in the soil solution, making them less suitable in areas of high ET. They are also not appropriate when recharge is low; their use requires the calculation of a total volume. Here chloride is preferred, where only a concentration must be determined, enabling greater precision (Allison and Hughes, 1983).

3. Chloride Mass Balance

The chloride mass balance (CMB) uses the aqueous chloride in, and outflows to, a volume of soil to calculate water flux through this soil volume. The inflow generally used is precipitation, and groundwater the outflow. The balance can be performed in zero, one, two and three dimensions depending on the nature of the input information and the degree of complexity considered necessary (Cook and Herczeg, 1998). The conservative nature of the chloride ion - it neither leaches nor is absorbed by aquifer sediments - makes it well suited and frequently used for recharge studies (Subyani, 2004).

This balance has been expressed by Cook and Herczeg (1998):

$$C_P P = C_G R \tag{7}$$

where P and R are the precipitation and recharge rates respectively, C_P and C_G are the chloride concentrations in precipitation and recharge (or groundwater) respectively.

Using this balance to calculate the water flux across the plane of the watertable, several assumptions are required (Wood, 1999):

1) Chloride ion in the groundwater has been derived solely from precipitation
2) The chloride in the system has a conservative nature
3) There has been a constant chloride mass flux over time
4) Recycling, or concentrating of the chloride ion has not occurred in the aquifer

Gee *et al*. (2005) has used this balance with C_S, the chloride concentration in soil pore water, in the place of C_G. For this to be valid, the following assumptions are required:

1) A steady rate of input water and chloride
2) Any vertical flow of chloride below the root zone is constant
3) That the soil contains no sources or sinks of chloride
4) The flow of chloride can be described by piston flow, and hence a point concentration can be assumed as the soil average.

At high chloride concentrations and low recharge rates, using pore water is most reliable. There is a disproportionate increase in the error as the chloride concentration detection limit is approached. Anion exclusion also has the ability to adversely affect results; this is most noticeable in soils with high clay contents.

Subyani (2004) modified this simple balance to account for the spatial and temporal variation of these variables:

$$\overline{q} = \left(\overline{R}\overline{Cl}_r + \hat{\rho}_{RCl_r}\hat{\sigma}_R\hat{\sigma}_{Cl_r}\right) / \overline{Cl}_{gw} \qquad (8)$$

where $\overline{q}$ is the recharge rate (mm/year), $\overline{R}$ is rainfall (mm/year), $\overline{Cl}_r$ is the chloride concentration of rainfall, $\hat{\rho}_{RCl_r}$ the correlation coefficient between the rainfall and its chloride concentration, $\hat{\sigma}_R$ and $\hat{\sigma}_{Cl_r}$ the standard deviations of rainfall and its chloride concentration measurements respectively and $\overline{Cl}_{gw}$ the average groundwater chloride concentration.

This method can also be useful in determining recharge rates thousands of years in the past (Gee *et al*., 2005).

Willis *et al*. (1997) compared CMB modelling to Darcian flux calculations and a soil water balance when calculating the deep percolation below cotton. CMB modelling was found to provide the most reliable results, but was only able to provide estimates for the growing season.

C. MODELS

Modelling can be used to estimate deep drainage without disturbing the soil profile. For a model to be useful it should be selected in accordance with the intended application, the required output information and accuracy, and the input data and resources available (DeJong and Bootsma, 1996). For the successful use of any model, validation of the predicted values with measured data must also be performed (Wegehenkel, 2005).

A range of water-balance models have been developed for analysis of different time scales such as hourly, daily, monthly and yearly, which also differ in complexity (Xu and Singh, 1998). Thornthwaite (1948) developed one of the earliest models, balancing evapotranspiration with monthly averages of soil water storage and precipitation. Over time, models such as the aforementioned have been accepted, adapted and utilised for a variety of hydrological applications. Generally, models having smaller time scales, such as daily and hourly, will have more associated parameters and are more data intensive (Xu and Singh 98).

Often, simple soil water balance models are used to estimate recharge. However, obtaining values for the required land surface parameters often poses a significant difficulty. This then manifests as uncertainty in the recharge estimates (Finch, 1998). This has led to water balance models being combined with other models, Hough and Jones (1998) combined models of evapotranspiration and soil moisture content with a soil water balance model. This was advantageous as recharge could be predicted from readily available meteorological data (Rushton and Ward, 1979).

Simulation models for the soil water balance have been divided into three categories: budget, semi-dynamic, and dynamic (DeJong and Bootsma, 1996). Budget models use the concepts of field capacity and wilting point to determine water flow within a profile. Single budget models regard the profile as a storage vessel which, once full, overflows; this excess water is treated as runoff, or drainage below the root zone. Budget models become more complex when the profile is divided into segments where water will flow from upper to lower zones as their field capacities are reached. This layered profile modelling has been shown by Calder *et al*., (1983) to be an improvement over the single vessel model. The advantage of budget models is their minimal input requirement. Shortcomings do exist - the concepts of field capacity and wilting point have been recognised as arbitrary and not as intrinsic soil properties. In addition, groundwater capillary rise is ignored, as is water distribution within the soil, making these models viable only for deep free-draining soils (De Jong and Bootsma 96). Houlbrooke *et al*. (2004) combined a layered budget model with various irrigator patterns to predict the drainage pattern occurring under these irrigators.

Semi-dynamic models use the multi-layer concept, but also allow for a varying water flow rate. Baier (1979) describes a semi-dynamic model, Versatile Soil Moisture Budget (VSMB), which has six layers. A percolation coefficient is included to describe the amount of water flowing out of a layer prior to field capacity being reached.

Dynamic models are based on the physical factors that govern water flow in the soil. With the advent of increased computer power and greater knowledge surrounding numerical solutions to Richards' equation, the use of such models has become easier and more widespread (DeJong and Bootsma, 1996).

SWATRE is an example of such a model. It is a transient finite-difference model for the one-dimensional flow in the unsaturated zone, which accounts for the uptake of water by roots through the use of simple sink term. Boundary conditions used for the top of the system are rainfall, potential evaporation and potential transpiration over a twenty-four hour period. For the bottom of the system, pressure head, zero flux or free drainage is used (Belmans *et al*, 1983).

1. SaLF Model

The Salt and Leaching Fraction (SaLF) model is an empirical model, for the prediction of leaching in dryland areas (Shaw and Thorburn, 1985). It relies on drainage being associated with hydraulic conductivity, which can be correlated to various soil properties. If these soil properties are known, along with the quantity and salt concentration of applied water, the steady state leaching fraction at the bottom of the root zone, can be calculated (Triantafilis *et al.*, 2003). For this model to be created, data from 766 non-irrigated soils in Queensland, Australia, was collected and empirical relationships established. The model predicts the electrical conductivity of water at the bottom of the root zone at maximum field water content, which is used to calculate the leaching fraction.

For the calculation of the leaching fraction the steady state mass balance for the leaching of salts [Eq. (9)] was used.

$$LF = \frac{EC_i}{EC_0} = \frac{D_0}{D_i} \qquad (9)$$

where LF is the leaching fraction, EC_i and EC_0 the electrical conductivities of the input and drainage waters respectively, and D_i and D_0 the depths of drainage input waters respectively.

D. OTHER METHODS

1. Thermal Gradient Method

The thermal gradient method is a less used technique of estimating the drainage rate. Developed from heat diffusion theory it is based on the idea that the vertical movement of groundwater has an effect on heat flux to the surface. Soil thermal conductivity, however, has a non-linear relationship with moisture content, making it difficult to obtain a realistic thermal profile, leading to large errors in estimates (Gee and Hillel, 1988).

Sammis *et al.*, (1982) used this method and found it to be less reliable than both Darcian flux calculations and using tritium as a tracer, but considered that with further work this could be a viable, low-cost method.

2. Electromagnetic Techniques

Electromagnetic (EM) induction instrumentation is a non-contacting technique for the fast estimation of soil attributes. EM instruments measure the bulk soil electromagnetic conductivity (EC_a), a function of clay content and composition, porosity, salinity, volumetric water content and temperature (McNeill, 1980). Many of these attributes have been mapped using EM techniques. Triantafilis *et al*. (2001) used EM measurements to determine soil EC_a. This was used as input to the SaLF model and the deep drainage occurring under various irrigation regimes was predicted.

3. Measuring K_{sat}

Potential deep drainage can also be estimated through the measurement of K_{sat}. Knowing K_{sat}, and the number of days per year or season the soil is saturated; potential deep drainage can be estimated. Many methods exist for obtaining a value of the K_{sat}.

In the laboratory, measurements can be made on samples of dried or fragmented soil, which have been packed in a standard manner, or on undisturbed soil cores (Reynolds *et al*., 2002).

Field measurements can be made above or below the water table. Below the water table the augerhole method, or piezometer method can be used (Reynolds *et al*., 2002).

The auger hole method is possibly the simplest of methods available for this purpose. A hole is made, to a depth greater than the water table, and the hole emptied of water. Based on the dimensions of the hole, the period of time taken for the hole to refill can be used to calculate K_{sat} (Beers, 1983).

The piezometer method also requires boring a hole below the water table. A piezometer tube is then inserted to below the water table; water depth is allowed to equilibrate and then measured. Water is rapidly removed from the hole and the change in the water level, with respect to the water table, is measured over time. K_{sat} is calculated from this data (Reynolds *et al*., 2002).

Above the water table various infiltrometers and permeameters are utilised to make infiltration measurements.

Ring infiltrometers are open-ended cylinders, with a bevelled edge on the bottom end. Their use is not limited to the measurement of K_{sat}, but also includes matric flux potential, sorptivity and the effective Green-Ampt wetting front pressure head. Their size varies according to their intended use, but it is generally between 10–50cm in diameter and 5–20cm long. Various ring configurations, with corresponding analyses, can be used depending on the application. This includes single ring, double-concentric rings and dual or multiple rings, which may have a constant or falling head (Reynolds *et al.*, 2002).

Also used is the well (or borehole) permeameter. This method is based on the analysis of three-dimensional flow infiltrating the soil from a hole, augered to the depth under consideration. If smearing or compaction has occurred at the bottom of the hole this should be amended, using a spiked roller or quick-setting resin (Koppi and Geering, 1986). The permeameter, which has the ability to maintain a constant head of water, is placed in the hole, and then backfilled to prevent well collapse (Reynolds *et al.*, 2002).

A simpler version of this technique can be used: the augered hole is lined, filled with water and the depth of water measured over time. The analysis for this technique has been described by Philip (1993). Once steady state has been reached, the amount of water infiltrating over time is recorded and can be used to calculate K_{sat}. For simpler analysis, a Mariotte bottle can be used to maintain a constant head.

VII. ESTIMATING THE LEACHING REQUIREMENT

As previously discussed excess salts must be leached through the root zone. To calculate the salt accumulation in the root zone a salt balance may be written:

$$V_iC_i + V_gC_g + S_m + S_f - V_dC_d - S_p - S_c = \Delta S_{sw} \qquad (10)$$

where ΔS_{sw} = change in the mass of salt stored in the root zone V_i, V_g, and V_d are the volumes of irrigation, ground and drainage waters respectively. C_i, C_g and C_d are the corresponding salt concentrations. S_m, S_f, S_p and S_c are the mass of salt resulting from weathering of minerals or dissolution of salt deposits, addition from fertilisers, precipitation in the soil or removal by crops respectively.

When steady state conditions have been reached $\Delta S_{sw} = 0$. For short periods of time V_g, C_g, S_m, S_f, S_p and S_c may be considered negligible Eq. (10) simplifies to:

$$V_iC_i = V_dC_d \qquad (11)$$

Or equating annual depths (D) with volumes applied:

$$D_iC_i = D_dC_d \qquad (12)$$

where D_i and D_d are the depths reached by irrigation and drainage water respectively (Rhoades, 1974).

In 1954 the US Salinity Laboratory Staff (USSLS) (Richards, 1954) found linear relationships to exist between concentrations (C) and electrical conductivities (σ) spanning almost three orders of magnitude ($0.1 mScm^{-1}$ - $100\ mScm^{-1}$) (Smith and Hancock, 1986). As a result of these findings, concentration and electrical conductivity are frequently interchanged and the salt balance expressed in terms of electrical conductivities:

$$D_i\sigma_i = D_d\sigma_d \qquad (13)$$

Care should be taken, because concentration and electrical conductivity are not always directly linear and calculations made using these terms are not always equal (Smith and Hancock, 1986).

The leaching fraction (LF) is the fraction of applied water that passes below the root zone and can be calculated by:

$$LF = D_d/D_i = \sigma_i/\sigma_d \qquad (14)$$

As a result of its conservative nature, chloride can be used in place of electrical conductivity in this equation (Qadir *et al.*, 2000).

Using Eqs. (7) and (8), it can be seen that by varying the fraction of applied water that passes below the root zone the electrical conductivity, or salinity, in the root zone can be controlled.

Richards (1954) defined the leaching requirement (LR) as the minimum fraction of applied irrigation water which must pass below the bottom of the root zone to prevent the reduction of crop growth, or maintain a specified level of salinity (Alsaeedi and Elprince, 1999; El-Haddad and Noaman, 2001). The objective for efficient water use is to achieve a leaching fraction equal to the leaching requirement. Realistically this is not possible; commonly an acceptable LF is larger than the desired LR.

The desired LF can be attained via two methods. Sufficient water can be applied at all irrigations, or a leaching irrigation can be applied infrequently to remove salt build up from previous irrigations (Qadir *et al.*, 2000).

White (2003) has expressed that to maintain an electrical conductivity at the bottom of the root zone, when salt equilibrium has been reached:

$$LR = \frac{\sigma_i}{\sigma_d} = \frac{V_d}{V_i} \qquad (15)$$

The critical value of σ_d is set by the salinity tolerance of the crop, where Rhoades and Miyamoto (1990) give the expression:

$$\sigma_d(critical) = 5\sigma_{ct} - \sigma_i \qquad (16)$$

where σ_{ct} is the salinity tolerance of the crop.

Richards (1954) proposed a basic technique for calculation of the LR. This method views the root zone as one layer, within which salt composition is uniform, and the amount of salt irrigation water provided is removed via leaching from the bottom of this zone (Vanhoorn, 1981). Based on this, and expressing salinity in terms of electrical conductivity, to maintain salt equilibrium into and out of the root zone the leaching requirement becomes:

$$LR = \frac{\sigma_i}{\sigma_{fc}} = \frac{\sigma_i}{2\sigma_e} \qquad (17)$$

where σ_{fc} and σ_e are the electrical conductivities of the field capacity soil water the saturated paste in the root zone (which is ≈ 0.5 σ_{fc}, for a wide range of soils excluding sandy soils).

There are problems with this model. Solute concentration does not remain constant over depth and the LR is overestimated (Vanhoorn, 1981).

Generally, plant water consumption and soil moisture content vary with depth, causing flux to decrease down the root zone, increasing salt concentration (Smith and Hancock, 1986). The plant will experience an overall salt concentration dependent on water uptake at each depth and the corresponding salt concentrations.

To account for this a variety of more detailed models have been described. Rhoades (1974) developed a semi-empirical model, recognising that the plant will act in response to a salt concentration in the soil water (C_s), which lies between the salt concentrations of applied and drainage waters:

$$C_s = K\left(\frac{C_i + C_d}{2}\right) \qquad (18)$$

For low values of the leaching fraction, it is suggested that K≈0.8. Rearranging and substituting into Eq. (14), the leaching requirement becomes:

$$LR = \frac{C_i}{2.5C_s - C_d} \qquad (19)$$

Another method divides the soil into quarter fractions, with a 40:30:20:10 water uptake pattern occurring.

These models use the salinity of applied water, and crop salinity tolerance, as inputs. Recharge is estimated through the simultaneous solution to a one-dimensional salt mass balance and a one-dimensional steady-state water flow equation, which includes a sink term (Alsaeedi and Elprince, 1999). Neither of these models account for evaporation from the soil surface. It was stated by Raats (1974) that the irrigation rate, and salt concentration of irrigation water, are not equivalent to the flux and salt concentration of water at the soil surface. The rate of evaporation affects the relationship between these terms.

Held within soil pores is an amount of stagnant water, which is effectively dead pore volume, effective pore volume must be used to correct for this. Alsaeedi and Elprince (1999) developed a model accounting for these factors. This model was successfully able to predict the LR using the exponential or 40-30-20-10 pattern of root water uptake.

Smith and Hancock (1986) developed a model where no specific water uptake pattern was assumed. They found that

$$\frac{Ln(LR)}{LR-1} = \frac{C_s^*}{C_i} \tag{20}$$

where C_s* is the desired salt concentration, from crop tolerance guidelines and C_i is the salt concentration of the irrigation water.

The limiting condition for this equation is where LR = 1 and all water applied drains through the soil. This is an unrealistic scenario, and even high leaching requirements, which approach 1 are uncommon (Smith and Hancock, 1986).

Rhoades (1982) developed a method for calculating the leaching requirement depending on the irrigation regime in use.

$$F'_c = \frac{\sigma'_e}{\sigma_i} \tag{21}$$

where F'_c is the permissible average concentration factor for the root zone and σ'_e the crop salt tolerance.

$$LR = \frac{0.3086}{\left(F'_c\right)^{1.702}} \tag{22}$$

$$LR = \frac{0.1794}{\left(F'_c\right)^{3.0417}} \tag{23}$$

When the soil is allowed to dry out between successive irrigations Eq. (22) is used, and when soil is not given sufficient time to dry out between irrigations Eq. (23) is used.

To compare LR values obtained using different techniques Table I provides the LR calculated using these methods for an irrigation scheme of cotton grown in Narrabri. This scheme assumes that precipitation is 600 mm, and that the critical value for the electrical conductivity of the drainage water can be calculated using Eq. (16). Cotton has a salt tolerance equivalent to an electrical conductivity of 7.7 dS/m for a 10% crop reduction. The average water requirement for cotton is 7ML (700mm) per season (Dugdale *et al*., 2004). This is the irrigation input for this scheme. Water may be sourced from the Namoi river ($\sigma = 4$ dS/m) or borewater ($\sigma = 1.1$ dS/m) (Anonymous, 2004a).

Table I

Comparison of Leaching Requirements Calculated for Cotton Under a Mock Irrigation Scheme

Equation used	Water Quality (dS/m)	LR	Drainage (mm/yr)
17	4	0.116	81.16
20	4	0.225	134.85
22	4	0.140	84.27
23	4	0.024	14.68
17	1.1	0.029	17.65
20	1.1	0.001	0.55
22	1.1	0.030	17.85
23	1.1	0.0005	0.29

As can be seen in Table I for the same water input conditions, the method used significant effect on the value of the LR calculated. Using water from the Namoi River the drainage required varied from 14.68 mm/yr to 134.85 mm/yr, different by almost a factor of 10.

VIII. ESTIMATES OF DEEP DRAINAGE IN THE LITERATURE

Table II

Examples of Deep Drainage Values Obtained in the Literature, with Various Climates, Land Uses and Techniques Used for Estimation

Method	Land use and Soil Type	Climate Applied water (mm/yr)	Drainage (mm/yr)	Reference
Cl^- Tracer	Native vegetation Sandy	Semi -arid 335	0.1	(Allison and Hughes, 1983)
Cl^- Tracer	Cleared Sandy	Semi -arid 335	3	(Allison and Hughes, 1983)
Tritium tracer	Eucalypt plantation Podosol	Semi -arid	65	(Allison and Hughes, 1972)
SWATRE model	Grass Loamy sand	Dry	60	(Belmans *et al.*, 1983)
EM techniques	Irrigated agriculture Sand	Semi -arid	15	(Cook *et al.*, 1989)
CMB	Cleared Sand	Semi -arid 180	46	(Gee *et al.*, 2005)
Lysimeter	Cleared Sand	Semi -arid 180	62	(Gee *et al.*, 2005)
Environmental tracers	Irrigated agriculture	Arid	2.2	(Harrington *et al.*, 1999)
Zero flux plane method	Irrigated agriculture Vertisol	Sub-tropical Monsoon 1208	110	(Hodnett and Bell, 1986)
Darcian flux calculations	Grassland Sand	~600	63	(Holmes and Coville, 1970a)
Darcian flux calculations	Forest Podosol	~600	none	(Holmes and Coville, 1970a)
Lysimetry	Improved pasture, Podosol	~600	61	(Holmes and Coville, 1970b)
CMB	Sandy	Semi –arid 250-580	3	(Leaney and Hercezeg, 1999)
CMB	Medium clay	Semi –arid 250-580	50	(Leaney and Hercezeg, 1999)

Table II
Continued

Method	Land use and Soil Type	Climate Applied water (mm/yr)	Drainage (mm/yr)	Reference
AETL (flux meter)	Lab conditions Sand		650 (potential)	(Masarik *et al.*, 2004)
Lysimetry	Non-irrigated pasture Dermosol	Temperate 750	205	(Pakrou and Dillon, 2000)
Meteoric Cl^- tracer	Uncultivated Fine sand	Semi -arid	75	(Phillips, 1994)
Meteoric Cl^- tracer	Uncultivated Sandy clay loam	Semi -arid	115	(Phillips, 1994)
^{36}Cl tracer	Uncultivated Fine sand	Semi -arid	5	(Phillips, 1994)
^{36}Cl tracer	Uncultivated Sandy clay loam	Semi -arid	15	(Phillips, 1994)
CMB	Irrigated pasture Dermosol	Temperate 900 - irrigation		(Prendergast, 1995)
CMB	Irrigated pasture Dermosol	Temperate 900 - irrigation		(Prendergast, 1995)
CMB	Irrigated pasture Dermosol	Temperate 900 - irrigation		(Prendergast, 1995)
Br^- tracer	Cleared and irrigated Sandy loam	Semi-arid 1170	16.1mm/day (potential)	(Rice *et al.*, 1986)
Darcian flux calculations	Irrigated agriculture Fine sandy loam	Temperate 738	2.5-154 (potential)	(Sophocleous and Perry, 1985)
CMB	Irrigated agriculture Sandy	Arid	6.1 (actual)	(Subyani, 2004)
SODICS model	Cleared	Temperate 600	29-70	(Thorburn *et al.*, 1991)
SODICS model	Native Vegetation	Temperate 600	7	(Thorburn *et al.*, 1991)
Piezometer rise	Native woodland Sand	Temperate 830	260	(Thorpe, 1987)
Tritium as a tracer	Native woodland Sand	Temperate 830	195	(Thorpe, 1987)

Table II
Continued

Br^- Tracer	Experimental plot, Sandy loam	5.41mm/hr – simulated rainfall	20.8mm/hr	(Tilahun *et al*., 2005)
SaLF Model	Irrigated agriculture Heavy clay	Semi -arid, 600mm applied	20	(Triantafilis *et al*., 2003)
SaLF Model	Irrigated agriculture Heavy clay	Semi -arid, 1500mm applied	85	(Triantafilis *et al*., 2003)
Soil-water balance	Irrigated agriculture Vertisol	Semi -arid	236	(Willis *et al*., 1997)

The methods for estimating deep drainage, previously described, estimate a range of drainage values, even under the same climatic regime and irrigation schedule. Gee *et al*. (2005) found that on the same soil the CMB predicted drainage of 46 mm/yr, a lysimeter at the same site measured 62 mm/yr, a 35% increase. Phillips (1994) stated that, in the same region using meteoric chloride as a tracer gave results 7.5 and 15 times larger than those found when using ^{36}Cl as a tracer on a sandy clay loam and fine sand respectively. In semi-arid regions, and on differing soils, values obtained for different techniques varied between 0.1 mm/yr, using the CMB, and 236 mm/yr, using a soil water balance.

These findings highlight the difficulty of accurately estimating the value of drainage. Each method includes certain assumptions and may have a tendency to over- or under-estimate drainage that is occurring.

IX. DISCUSSION

As described previously there are a variety of methods available for quantifying the potential and actual deep drainage, each with their own advantages and disadvantages, these have been summarised in
Table III.

Table III

The Pro's and Con's of Techniques Available for Estimating Deep Drainage

Method	Pro's	Con's
Lysimetry	High relative accuracy.	Expensive. Requires technical knowledge. Immobile and time consuming in the order of years.
Flux Meter	Good relative accuracy.	Expensive. Immobile. Requires technical knowledge.
Soil-Water Balance	Required parameters can be easily obtained or estimated.	Requires high accuracy of the input data. Potential for large errors. Inapplicable in areas of low drainage.
Zero Flux Plane Method	Reasonable accuracy.	Cannot be used during times of high infiltration, or steady flow conditions. Large time and cost input.
Darcian Flux Calculations	Relatively fast and inexpensive.	Parameters used are hard to accurately measure and have a high spatial variability. Relatively large errors, especially when applied to dry areas
Chloride Mass Balance	Conservative nature of chloride ion. Relatively fast and inexpensive. Able to describe temporal variability of water fluxes.	Relatively low accuracy.
Tracers	Can make measurements over varying time spans. Relatively accurate in dry areas. Relatively inexpensive. Ability to describe spatial and temporal flux variation.	Low relative accuracy. Results can depend on interpretation method used. Hindered by the effects of bypass flow.

Table III
Continued

Method	Pro's	Con's
Models	Require minimal fieldwork. Account for spatial and temporal variability of soil properties. Varying degrees of complexity and accuracy. Minimal input requirements.	Can be time consuming, may require a large amount of technical knowledge if highly complex. Lower relative accuracy.

It can be seen that the methods able to provide the most accuracy are the least flexible, they generally have a large associated cost, time and knowledge requirement and are the least able to describe the spatial variability. Of the other possible methods inaccuracies are high.

Lysimeters, considered the most accurate method for determining deep drainage, are immobile, prohibitively expensive, require technical knowledge and are time consuming. Flux meters and the zero flux plane method, also considered relatively accurate, are again expensive, and require a high degree of technical knowledge.

Methods, which are cheap and fast such as Darcian flux calculations and the chloride mass balance, are significantly less accurate. However, Minasny and McBratney (2002) found that when all factors are considered, not just accuracy, methods giving lower accuracy in individual measurements can be considered as more efficient.

X. CONCLUSIONS

Deep drainage, or recharge is required for the leaching of excess salts accumulated in the root zone. However, excessive deep drainage can lead to off-site water contamination, rising water tables, waterlogging and secondary soil salinisation. Additionally, water, a valuable resource, is lost. This is highly inefficient and unsustainable. A comprehensive understanding of the mechanisms of soil water flow is required before deep drainage can be accurately quantified.

Over time, knowledge of soil-water flow processes has advanced, and an abundance of techniques for estimating deep drainage have been developed. These are applicable to a wide range of soil types and climates, providing a range of accuracies. As has been discussed, these methods are prohibitively expensive, time consuming, require technical knowledge, can be poor in accuracy, have a large input data requirement or are unable to describe soil spatial variability.

Evidently, there is need for a new method having the ability to describe the spatial variability of soil physical properties, is cheap, time efficient, and is simple and easy to use. A technique, which has the potential to fulfil this description, involves the measurement of K_{sat} using a falling head borehole. Conductivity is then input to a 'black-box' model, which provides required and potential deep drainage as outputs. Spatial variability is described by taking measurements on a statistically relevant number of holes. This will be the subject of my research paper.

Section 2: Research

The Development Of A User-Friendly Method Of Estimating Deep Drainage And Its Requirement

Abstract

Soil salinity is an increasingly pressing issue, so the need for tools able to quantify contributing factors such as deep drainage is of significant importance. A method for estimating potential deep drainage, which when compared to the leaching requirement offers insight to the efficiency of an irrigation regime was developed. To facilitate this, the falling-head lined-borehole technique for estimating sub-soil saturated hydraulic conductivity (K_{sat}) was adapted and tested for use on vertosols in the cotton growing regions of NW NSW. A method for the removal of a smeared soil surface using araldite was developed, along with a sampling regime possessing the ability to describe soil spatial variability.

This adapted method was tested in Field 11, Auscott Moree, which is representative of the soils of the area, having heavy clays derived from alluvial material and lighter textured soils derived from a prior stream channel. Results showed a significant difference in the potential deep drainage occurring on different soil types, indicating the methods ability to differentiate between soil types and reveal or confirm potentially leaky areas in a field.

This method was found to be a cost-effective, easy to use, simple to understand, have a reasonable degree of accuracy and describe soil spatial variability.

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1. Introduction

Worldwide, the human population utilises 54% of all accessible freshwater, and if per capita usage continues to increase at its current rate, within 25 years this will increase to 90% (UNESCO, 2005). The expected population rise from approximately six to eight billion people over this time is a significant contributing factor. Of water currently used by the human population, 69% is employed in agricultural production, such as irrigation (FAO, 2003).

Present irrigation practises are often inefficient, resulting in water loss via deep drainage (DD) (Alsaeedi and Elprince, 1999; Triantafilis et al., 2003). Deep drainage is defined as the volume of water flux below the depth at which plant roots extract water (Bond, 1998). It is considered to be one of the main drivers of dryland and irrigated salinity (Dunlop et al., 2004; Vervoort et al., 2002). Salinity can result from DD either being in excess, through the formation of a perched water table of saline water (Dunlop et al., 2004; Willis et al., 1997), or not being sufficient such that salts applied to the soil accumulate and are not leached from the root zone (Qadir et al., 2000). It is estimated that of the 270 million hectares used for agricultural production, 20-30 million have been affected by salt accumulation (FAO, 2003).

In NSW 93% of the land affected by shallow groundwater is agricultural land, with an 8.4-fold increase in the affected area predicted over the next 50 years (Boschma and Lodge, 2003).

Salinity is a major problem from both an agricultural and environmental point of view, as land is rendered unproductive (Anonymous, 2004d) and water sources may be polluted (Pepper et al., 1996), and therefore knowledge of occurring DD is integral to land preservation and efficient farm management.

Many methods are available to estimate DD (Gee and Hillel, 1988). These include lysimetry (Bond, 1998), the plane of zero flux method (Hartmann et al., 1980; Sophocleous and Perry, 1985), chloride mass balance (Cook and Herczeg, 1998; Subyani, 2004) and tracers (Allison et al., 1994; Walker, 1998). However these methods generally have a variety of associated problems such as large cost, immobility, inaccuracy and/or the requirement of technical knowledge. This means that estimating DD is generally inaccessible to farmers, and hence frequently goes unmeasured and unconsidered in irrigation management.

One method, which has not received much attention, is to estimate potential DD by measuring the saturated hydraulic conductivity (K_{sat}) (Reynolds et al., 2002b). A variety of techniques are available for estimating K_{sat}, including the falling-head lined-borehole technique (FHLBT), for which an approximate analysis was determined by Philip, (1993). Obtaining data using this method is relatively simple and easy, making it ideal for users without a technical background in this field.

Vertosols cover 88 million hectares of land in Australia largely in arid and semi-arid areas to the north and east of the continent (McKenzie et al., 2004). These soils are important for Australian agriculture, with a variety of applications in both dryland and irrigated farming, including cotton, rice, sugar-cane, wheat and the grazing of sheep and cattle (McKenzie et al., 2004). These are important soils in NW NSW where they are predominantly developed for irrigated cotton production (Chan and Hodgson, 1984).

The measurement of K_{sat} on vertosols using many of the conventional methods has been difficult and unreliable. This may be attributed to the presence of macropores, where the use of small samples or cross-sectional areas is unable to adequately represent K_{sat} (Zobeck et al., 1985). These cracks, endowing the soil with shrink/swell properties have a large impact on the soils' physical properties (Bridge and Ross, 1984). Gaining a full understanding of infiltration into cracking clay soils has been hindered by the interplay between soil properties, rainfall and topography. This difficulty is increased as infiltration may also vary between soils and soil states (Ross and Bridge, 1984).

This paper looks at the development of a method that may potentially be fast, cost effective, able to describe soil spatial variability, has a reasonable degree of accuracy and is easy to use without any prior technical knowledge. The method is intended for use by individuals, being accessible in terms of cost and ease of use, and putting research results directly into context. Johnson et al. (2003) and Bonny et al. (2005) found that user participation in the research process contributes to actual or potential impact of the research, and may enhance the ability of participants to continue the process of innovation and development. They found greater economic impacts when users were involved in the research process, particularly in the early stages. There exist a variety of motives for farmer participation. Soderqvist (2003) found that personal advantage, rather than social consequences, was the primary motivating factor. For these reasons it is considered that having a method which gets individuals involved, and demonstrates direct advantage to their agricultural practice, may have an increased chance of influencing inefficient water use.

With this motivation the method to be developed includes a field technique for determining the infiltration rate using a lined borehole and a model, which uses this data to determine K_{sat}, modelling this to estimate the potential DD, calculating the leaching requirement and evaluating irrigation efficiency by comparing the two. The objectives of this research are to:

1) Develop a field technique for measuring sub-soil K_{sat} using a lined borehole that is able to adequately describe soil spatial variability.

2) Model K_{sat} to estimate the potential DD occurring

3) Determine the drainage requirement of the soil and compare to the potentially occurring drainage.

2. Materials and Methods

2.1. Study sites

In the Namoi and Gwydir Valleys, vertosols (Isbell, 2002) are the predominant soil type, having a high, relatively uniform clay content of between 30% and 80% throughout their profile, with their structure dominated by lenticular peds (Ringrose -Voase et al., 2002). So and Onus, (1984) observed that the brown and grey vertosols of this area are predominantly fertile and self-mulching, with a small distribution of some low-yielding soils throughout the area. There has been a large amount of agricultural development on these self-mulching clays, generally for irrigated cotton production (Chan and Hodgson, 1984).

2.1.1. ACRI

For the method development component of the research the Australian Cotton Research Institute (ACRI) (Fig. 8) was used. The ACRI is located approximately 25km north-west of Narrabri (149°36’E and 30°12’S) (Weaver et al., 2005) in the Namoi River Catchment of North West NSW, a part of the Murray-Darling Basin (Namoi Catchment Management Authority, 2004).

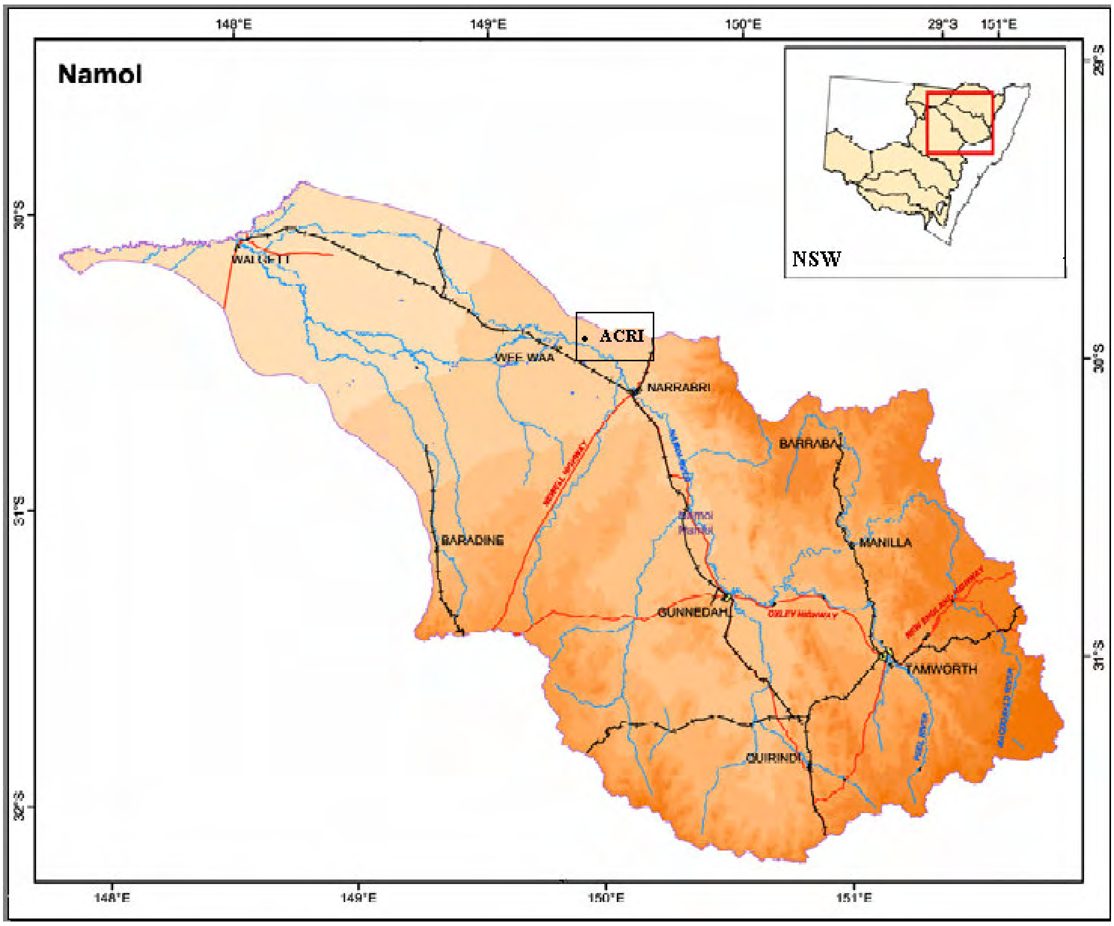


Fig. 8. ACRI location within the Namoi Catchment, adapted from Anonymous, (2004a).

In this area the annual mean daily maximum and minimum temperatures are 26.7 °C and 11.6 °C respectively and the mean annual rainfall is 643.3 mm (Bureau of Meteorology, 2005b) with rainfall dominant in the summer months (Namoi Catchment Management Authority, 2004). Grazing and dryland cropping are the most common land uses in the area, and irrigated cotton is a highly significant industry along waterways of the area. Plains of low gradient are characteristic of the landscape on and around the research station.

The soils on both study sites used at the ACRI are grey vertosols (Isbell, 2002), and were described by Northcote, (1966) as deep, grey, self-mulching cracking clays.

2.1.2. Field 11, Auscott Midkin

The second study site is field 11 of Auscott Midkin, Fig. 10, located at 29°18’ N and 149°45’E corresponding to the Australian map grid reference 6755660 N and 767020 E. Auscott Midkin is a large cotton farm at the lower end of the Gwydir River valley, approximately 25 km northwest of Moree and 7 km northwest of Ashley, a small town. Field 11 is approximately 243 ha in size and has been used for furrow-irrigated cotton for the last 25 years. An aerial photo of the field is shown in Fig. 9.

In the area the annual mean daily maximum and minimum temperatures are 27.6 ^{o}C and 11.7 ^{o}C respectively and the mean annual rainfall is 578.6 mm (Bureau of Meteorology, 2005a).

The property is situated within an alluvial plain landscape. Stannard and Kelly, (1968) identified three surface depositional systems in this landscape:

- Soils of the clay plains
- Soils of the prior stream formations
- Soils of the levee deposits of present streams

The soils on Auscott Midkin are predominantly vertosols formed from floodplain clay deposition, there is also a prior stream formation passing through the property. Field 11 is dominated by the heavy clay soils of the clay plains but also contains sandy soil types belonging to a prior stream channel, seen cutting across the field and in the southwest corner.

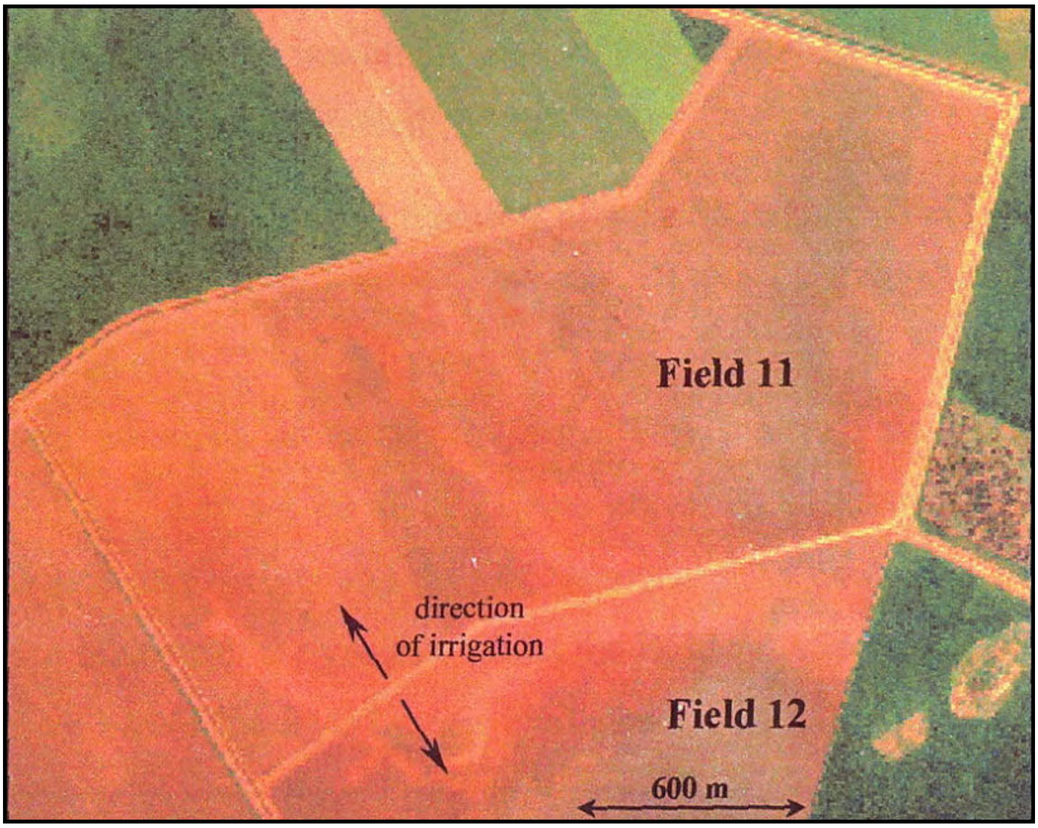


Fig. 9. Aerial photo of field 11, Auscott Midkin, with path of prior stream channel visually observable Huckel, (2001)

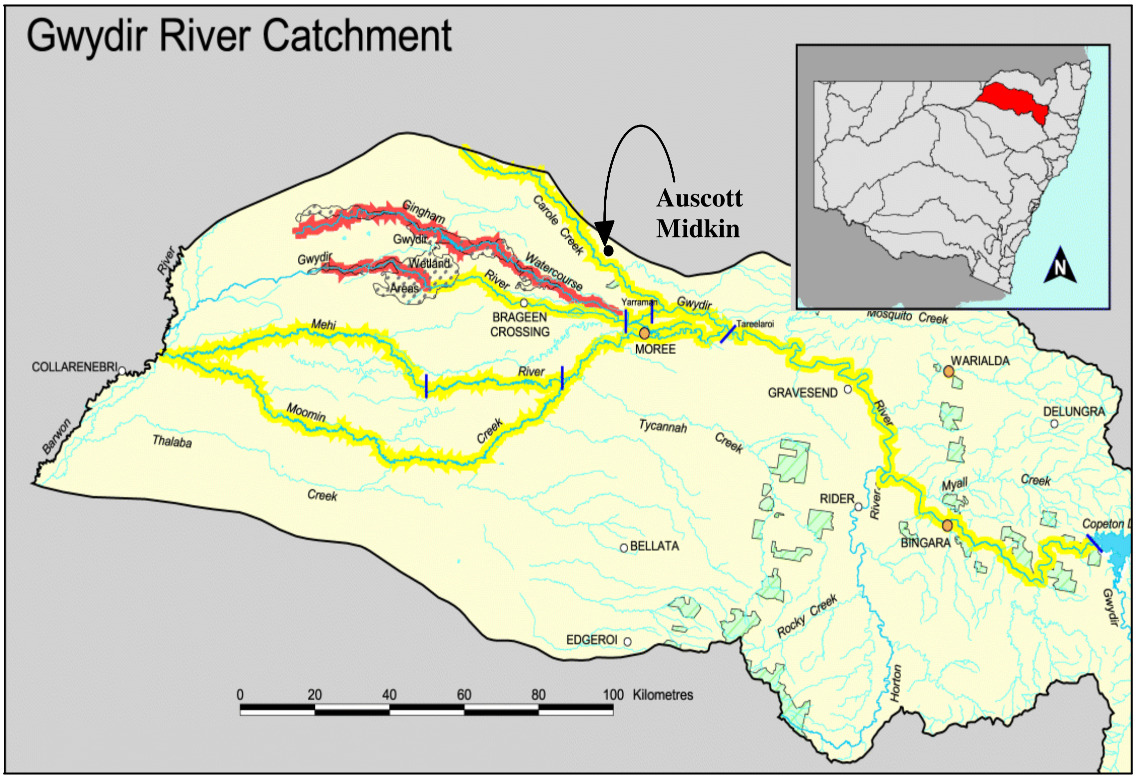


Fig. 10. Location of Auscott Midkin, adapted from Environmental Protection Agency, (2005).

Montgomery, (2003) has given a detailed pedological description of field 12, which is adjacent to field 11, and whose soils are derived from the same parent materials. Montgomery, (2003) found the soils of clay alluvial deposits to be black vertosols and those of the prior stream formations to be brown vertosols, the brown vertosols having a lighter texture throughout. Huckel, (2001) also found the soil of the alluvial clay deposits to be vertosols, but on the palaeo-channel within field 11 identified the presence of chromosols.

2.2. Methods

The aim of this research is to develop a method of estimating potential DD and through comparison with the leaching required assess irrigation efficiency. The steps necessary to achieve this are outlined in Fig. 4.

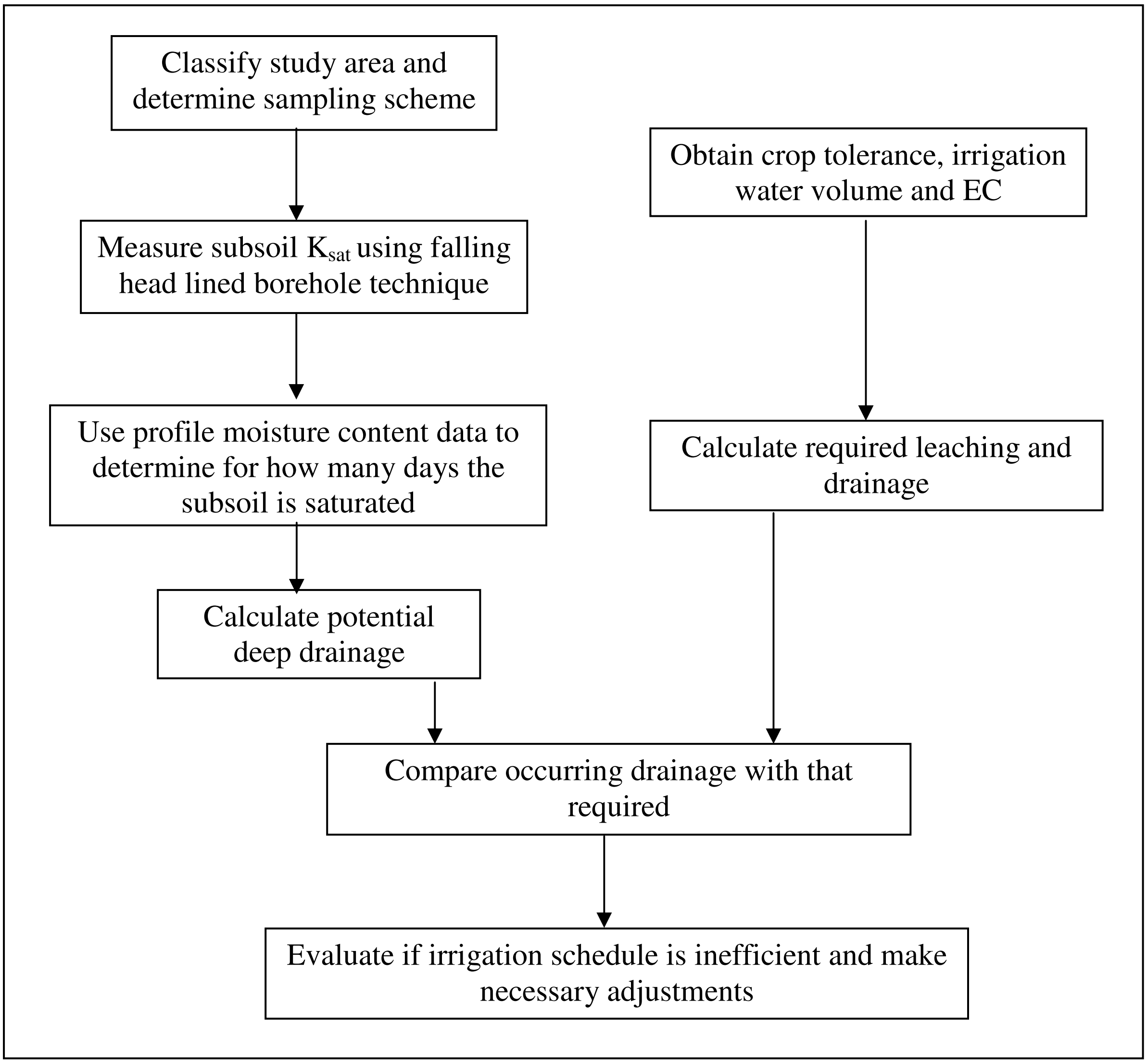


Fig. 11. Steps required for determining and comparing potential deep drainage its requirement

2.2.1. Measuring the sub-soil K_{sat}: The -head lined-borehole technique (FHLBT)

This method, the theory of which is described by Philip, (1993), measures sub-soil K_{sat}. A cylindrical hole is made to the desired measurement depth, into which a length of plastic pipe of equivalent diameter is inserted. Water is poured into this lined borehole, initial water depth recorded, and the depth of water over time monitored, as shown in Fig. 5, until steady-state infiltration is observed.

Fig. 12. Measuring the depth of water in a falling-head lined-borehole

The time interval required will depend on soil type and initial moisture content. To obtain a value of K_{sat} from this data, the infiltration process must be modelled. Philip, (1993) made three main assumptions:

1) **Equivalent spherical supply surface with hydraulic correction.** The actual three-dimensional flow from the circular surface at the base of the hole is replaced by spherically symmetrical flow from a sphere of equivalent surface area. This yields flow paths that are hydraulically more efficient, and a factor of $8/\Pi^2$ is introduced to account for this and provide balance. The difference between the actual flow shape and that assumed for this model can be seen in Fig.6.

2) **Three-Dimensional Green-Ampt Model.** The Green-Ampt analysis of water flow uses the idea of a step-function wetting front, which suggests there is a sharp boundary between the wet and non-wetted region. This is highly idealized, as soil moisture profiles do not have a distinct boundary, but are gradual, especially for soils with high clay content such as vertosols, shown in Fig.7. However it is considered that this does not affect the ability for useful information such as cumulative infiltration to be obtained.

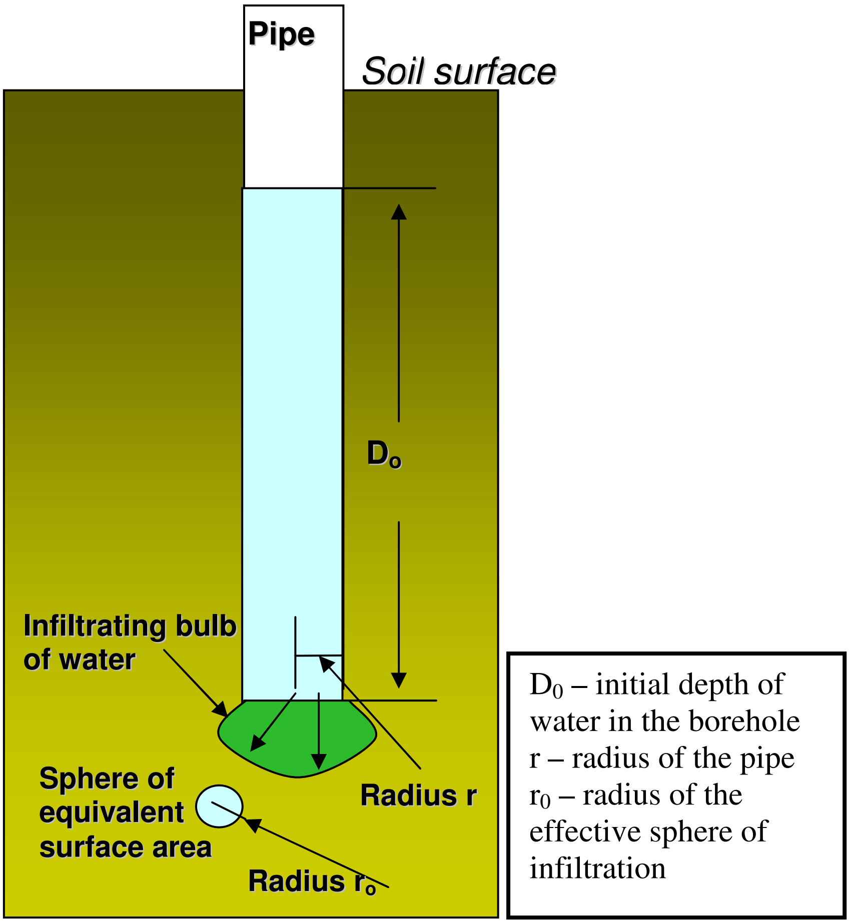


Fig. 13. Water infiltrates from the base of the borehole, radius r, but the modelled infiltration occurs from a sphere of equivalent surface area, with radius r_o

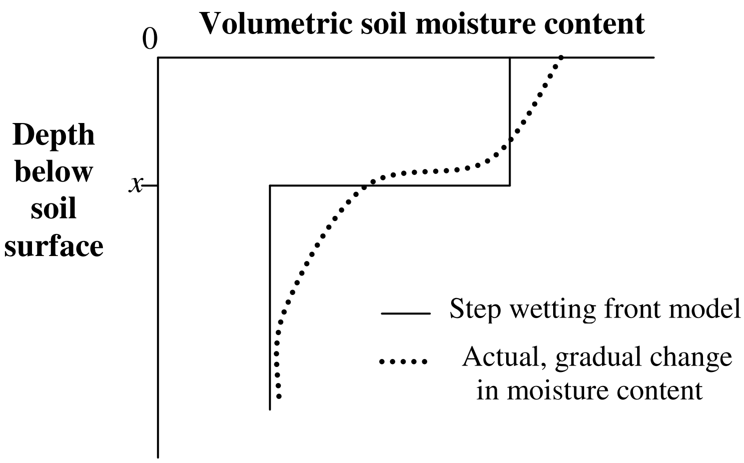


Fig. 14. Moisture content with depth for a gradual change in moisture content and for a step-wetting front model

3) **Pressure-Capillarity-Driven Flow Perturbed Symmetrically by Gravity.** Pressure and capillarity forces, causing spherical symmetrical flow, are considered the main driving forces to flow with gravity perturbing this flow downwards.

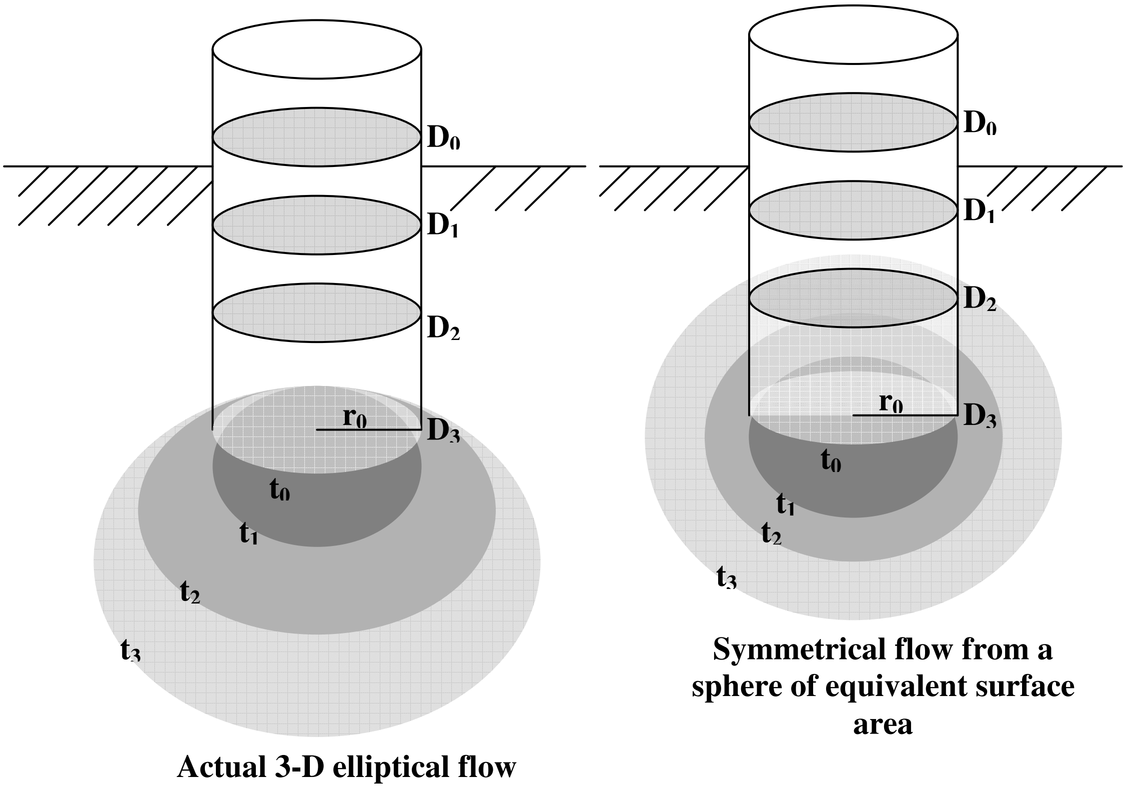


Fig. 15: Actual and assumed flow from base of borehole.
As time passes from t_0 to t_3 the sphere of infiltration increases and the depth in the holes decreases from D_0 to D_3, showing the effect of gravity on the actual flow occurring.

From these assumptions Philip, (1993) developed Equation 1 to calculate K_{sat}.

$$K_{sat} = \frac{-\frac{dD}{dt}\left(R_{\max} - r_0\right)}{R_{\max}\left(\frac{8C}{\Pi^2 r_0} + 1\right)} \quad (1)$$

Where K_{sat} is the saturated hydraulic conductivity (cm/min), $\frac{dD}{dt}$ is the steady-state infiltration rate (cm/min), r_0 is the initial radius of the sphere from which infiltration is being modelled (mm), C is the wetting front potential, which models capillarity (mm) and $R_{\max}$ is the final radius of the wetted bulb (mm), given by

$$R_{\max} = \left[r_0^{\,3} + \frac{3D_0 r_0^{\,2}}{\Delta\theta} \right]^{\frac{1}{3}} \tag{2}$$

where $\Delta\theta$ is the change in volumetric moisture content over the period of infiltration and D_0 is the initial water depth in the permeameter (mm). Values of C depend upon the soil type and structure and are given in Table 1.

Regalado and Munoz-Carpena, (2004) developed an empirical model to predict the wetting front suction C (mm), based on t_{med} and t_{max}, the times when the permeameter is half-full, and empty respectively, given in Equation 3.

$$C = \frac{1}{1000} \times \exp(a' + b' \sqrt{\frac{t_{med}}{t_{\max}}}) \tag{3}$$

where a’ = -13.503 and b’ = 19.678.

2.2.2. Estimating the duration of sub-soil saturation

The length of time over which the sub-soil is saturated can be determined by monitoring soil moisture content over time. Instruments such as a Capacitance probe (C-probe), shown in Fig. 10, or neutron probe can be used to achieve this. This only requires monitoring for a few irrigations, but more intense observation will provide more accurate results. This should indicate the number of days, for which the depth under consideration remains saturated, following irrigation. Fig. 9 shows an example of C-probe output, which can be used to determine time period of soil saturation.

Table 1

Values of C for various soil structures and textures (Reynolds et al., 2002b).

Soil structure and texture category	C (mm)
Compacted, structureless, clayey or silty materials eg. landfill caps and liners, laclustrine or marine sediments	1000.0
Soils that are both fine textured (clayey or silty) and unstructured; May also include some fine sands.	250.0
Most structured soils from clays to loams; includes unstructured medium and fine sands. The category most frequently applicable for agricultural soils.	83.3
Coarse and gravelly sands; may also include highly structured or aggregated soils, as well as soils with large and/or numerous cracks, macropores.	27.8

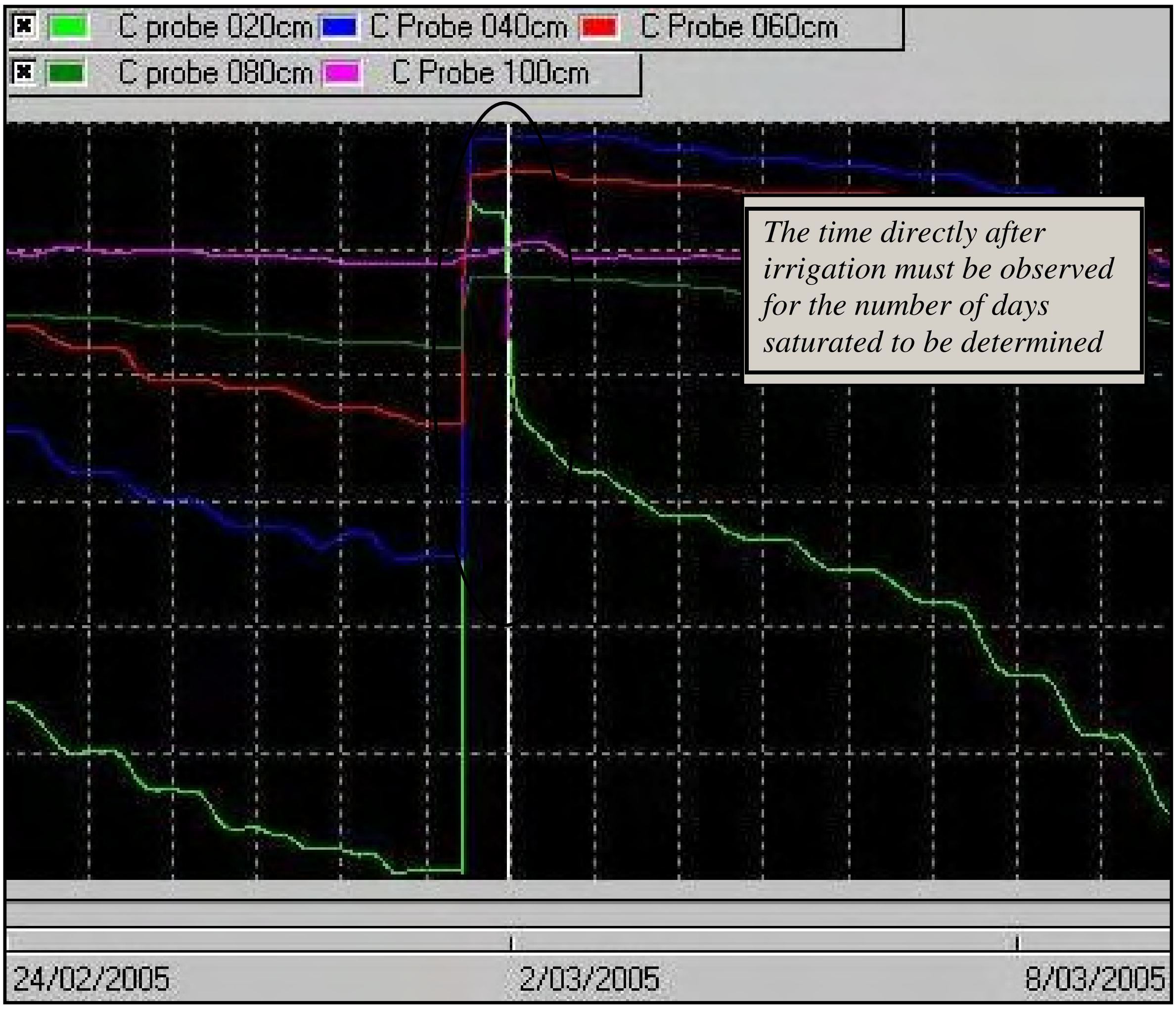


Fig. 16. Soil moisture content over time, as recorded by a C-probe

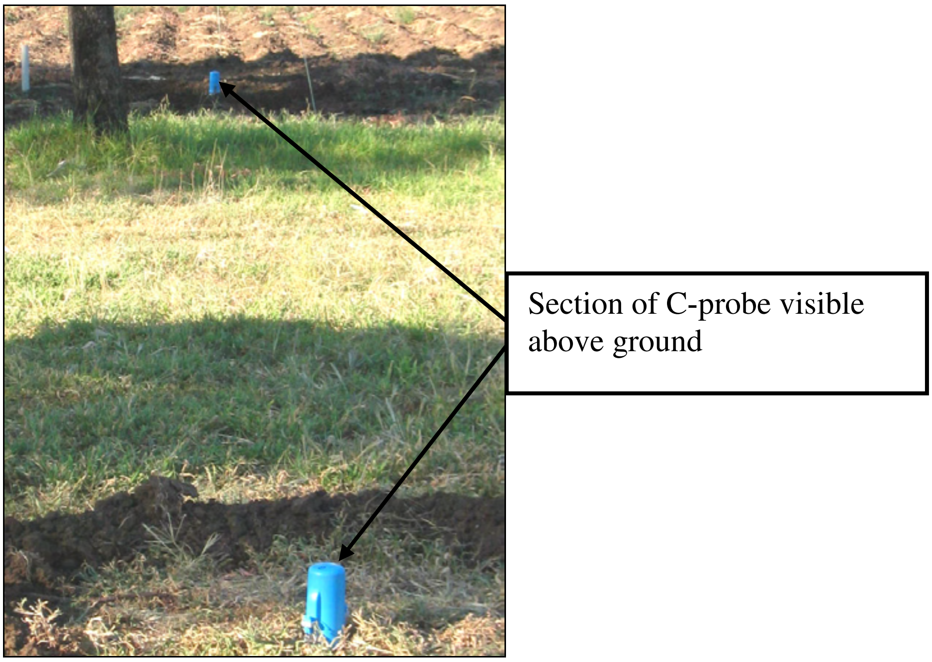


Fig. 17. C-probes in the ground at the ACRI, Narrabri

2.2.3. Calculating potential deep drainage

The potential DD, which may occur in a field, can be estimated if the saturated hydraulic conductivity of the sub-soil and the number of days the sub-soil is saturated for are known, using Equation 4

$$DD = K_{sat}.N_s \quad (4)$$

where N_s is the number of days for which the sub-soil is saturated.

2.2.4. Calculating the leaching and drainage requirement

The leaching fraction (LF) is the fraction of applied water that passes below the root zone and can be calculated by:

$$\mathrm{LF} = \frac{\mathrm{D_d}}{\mathrm{D_i}} = \frac{\sigma_\mathrm{i}}{\sigma_\mathrm{d}} \quad (5)$$

where D_i and D_d are the depths reached by irrigation and drainage water respectively and σ_i and σ_d are the electrical conductivities of the irrigation and drainage water respectively.

Richards, (1954) defined the leaching requirement (LR) as the minimum fraction of applied irrigation water which must pass below the bottom of the root zone to prevent the reduction of crop growth, or maintain a specified level of salinity (Alsaeedi and Elprince, 1999; El-Haddad and Noaman, 2001). It will vary depending on the crop grown and the volume and electrical conductivity of the irrigation water applied.

Ideally, when salt equilibrium has been achieved, the leaching requirement is equal to the leaching fraction with the electrical conductivity of the drainage water (σ_d) set by the crop tolerance, given by

$$\sigma_d(critical) = 5\sigma_{ct} - \sigma_i \quad (6)$$

where σ_{ct} is the salinity tolerance of the crop (Rhoades and Miyamoto, 1990).

Substituting this into Equation 5 yields;

$$LR = \frac{\sigma_i}{5\sigma_{ct} - \sigma_i} \quad (7)$$

The leaching requirement is the fraction of the applied water required to leach through the profile, so the actual volume of drainage required is given by;

$$DR = LR \times V_i \tag{8}$$

where DR is the drainage requirement and V_i is the volume of irrigation water applied

2.2.5. Classification of a field

Classification of a field is the segregation of areas sharing similar characteristics. This is performed such that areas with different characteristics may be identified and sampling developed to account for this.

Field classification can be performed by fuzzy k-means analysis (Bezdek, 1981) using FuzME, software developed by Minasny and McBratney, (2002b). The optimal number of classes is determined by monitoring three performance indicators; the Fuzziness Performance Indicator (FPI), the Modified Partition Entropy (MPE) (Roubens, 1982) and the compactness and separation validity index (S) (Xie and Beni, 1991), which are provided in the FuzME output. Plotting the indicators against the number of classes and observing the point where all indicators exhibit a local minimum determines the optimum number of classes.

3. Results

Smearing of a soil surface is a disruption with the potential to hinder water infiltration. The FHLBT has not previously been applied to soils where smearing was an issue, nor to soils with such a high degree of spatial variability. As such it needs to be determined whether this technique is applicable to these soils, and if so what amendments, if any, must be made.

3.1. Method Development: the practicalities of applying the FHLBT to vertosols of the cotton growing regions in NW NSW

Before this method was used, the practicalities of its use were explored. Fig.11 shows the process through which this occurred. This occurred at the ACRI in Narrabri.

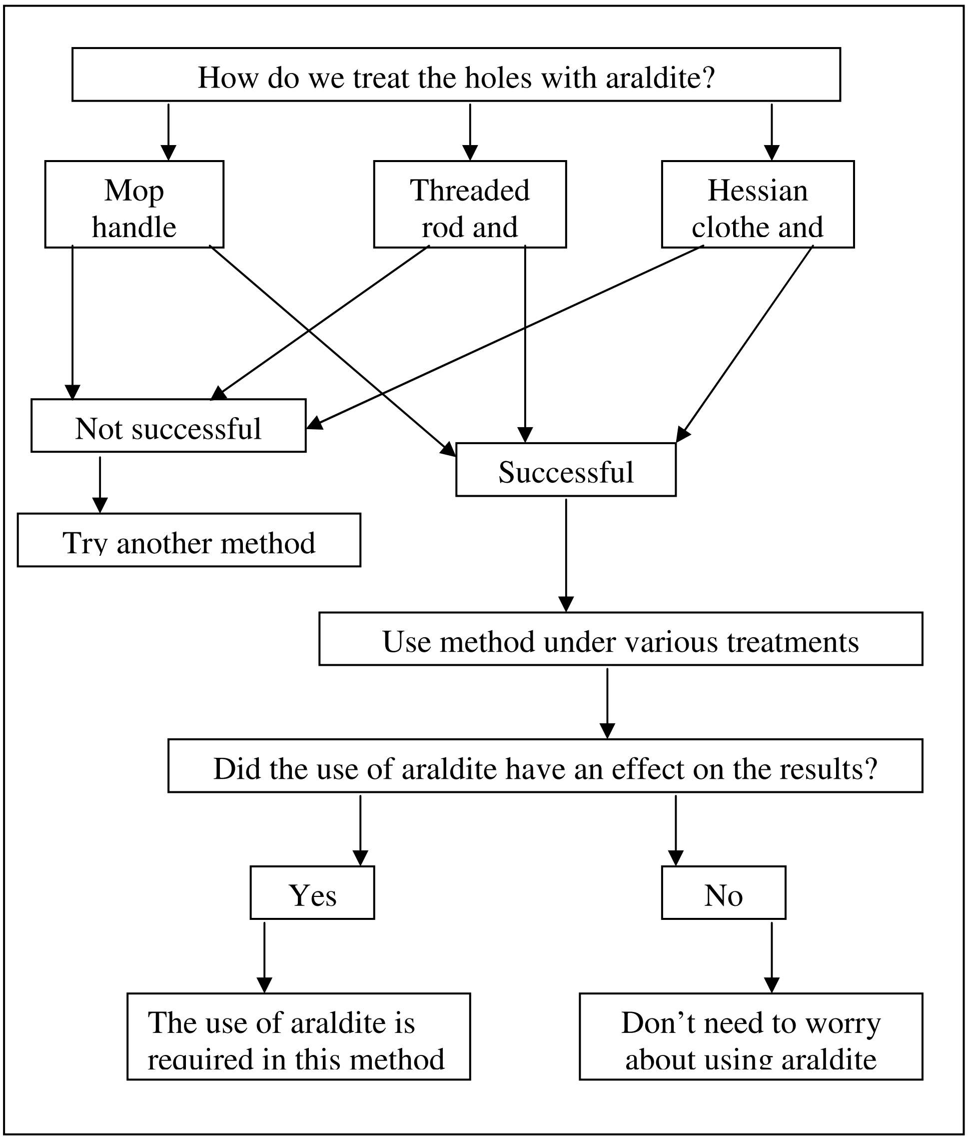


Fig. 18. Flowchart outlining plan for method development at the ACRI

3.1.1. Trials with araldite

It was initially thought a hand auger could be used for making the boreholes for infiltration, but this was likely to smear the base of the hole and hinder infiltration. The problem was removing the smeared layer and exposing a fresh soil surface for infiltration. Koppi and Geering, (1986) had success at exposing a fresh soil surface, revealing biological pores and structural surfaces, by using a quick-setting epoxy resin. The problem is how can the epoxy resin be physically removed from the bottom of a hole 1m deep and 50 mm in diameter. Ideas were tried until one was found to be successful:

- **Mop handle and cloth:** Initially, at a depth of 30cm, success was had using a mop handle and dish cloth to remove the resin. However, at greater depths this became unfeasible as the cloth would not remove all of the araldite, or would tear easily as it was hard to attach to the mop handle.

- **Threaded rod and washer:** The cloth was then attached with wire to a rod and washer, which was more successful. But there was still frequent cloth tearing and easily accessible rods were 1 and 3 m, both impractical lengths.
- **Hessian cloth and twine:** Folded squares of hessian cloth were tied onto a length of twine and the hessian pressed into the resin. This method proved to be the most successful.

Due to loose soil falling into the hole multiple araldite treatments were often required before a fresh surface was revealed. Generally two were sufficient, this agreed with the findings of Koppi and Geering, (1986). Fig. 12 shows the steps for treating a hole with araldite when removing a layer of soil from the bottom of the hole.

Fig. 19. Treating a hole with araldite using the hessian and twine method

3.1.2. The successful araldite treatment

Prepare 300 cm^2 of resin and pour carefully into borehole, trying to avoid getting resin on the walls of the hole. A 17cm^2 square of Hessian is folded into quarters and secured with twine. The hessian and twine are inserted into the hole, using a rigid implement to ensure good cohesion between the hessian and resin. The resin is left to set for 17 minutes from when it was first poured, and then removed firmly and carefully from the hole; visual inspection should be adequate to determine if further treatment is required.

This technique meets several of the proposed criteria; simple and easy to use, economical and time-efficient. However, the use of a hand auger was not found to be practical, being time consuming and labour intensive; an alternative is the use of a hydraulic ram. Holes treated and not treated with araldite were used such that it could be determined if this treatment was required.

3.1.3. Developing a sampling technique

At the ARCI, samples were taken on various land uses: pasture, cultivated but not irrigated and a cultivated, irrigated field. The cultivated, irrigated field was chosen as a lysimeter was recently installed in this field, for which the basic design is given by Brye et al., (1999), with the hope that the results of this study could be compared to those of the lysimeter. In the pasture site some holes were deliberately smeared with an auger for comparison.

The K_{sat} data for these sites shows skewed distributions; this can be seen in Fig. 13 (a) and (b), which displays distributions for treatments on the cultivated sites. This skew was evident in all of the treatments, so to normalize these distributions, a log transformation was applied, shown in Fig. 13 (c) and (d).

K_{sat} mean and standard deviations for each treatment were calculated using log-transformed data; a student's t-test was performed to determine if statistically these treatments could have come from the same population.

The mean K_{sat} values for each treatment, the potential DD and results of the statistical comparison are shown in

Table **2**. The distribution of the data in each treatment is given in Fig. 14.

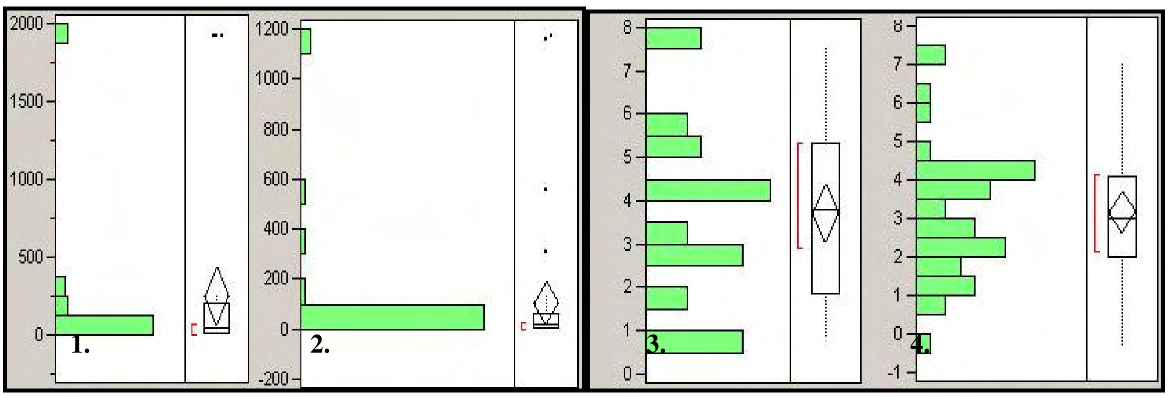


Fig. 20. The distribution of the unlogged K_{sat} data for the cultivated site for a) non-araldite treated sites and b) araldite treated sites and distributions of the logged data for the cultivated sites for c) non-araldite treated sites and d) araldite treated sites

Table 2

Comparison of results obtained under various treatments at the ACRI, treatments connected by the same letter are not statistically different.

Treatment		Number of holes	K_{sat} (mm/day)		Potential DD (mm/yr)	Statistical Comparison
			Mean	SD		
Pasture	Araldite	12	27.1	10.1	1056.9	A
	No araldite	41	23.6	5.2	920.4	A
Cultivated	Araldite	5	12.3	14.8	479.7	A B
	No araldite	8	4.2	5.5	163.8	B
	Smeared holes	7	2.0	1.3	78.0	B
Cultivated and irrigated	Lysimeter site	9	4.0	4.2	156.0	B

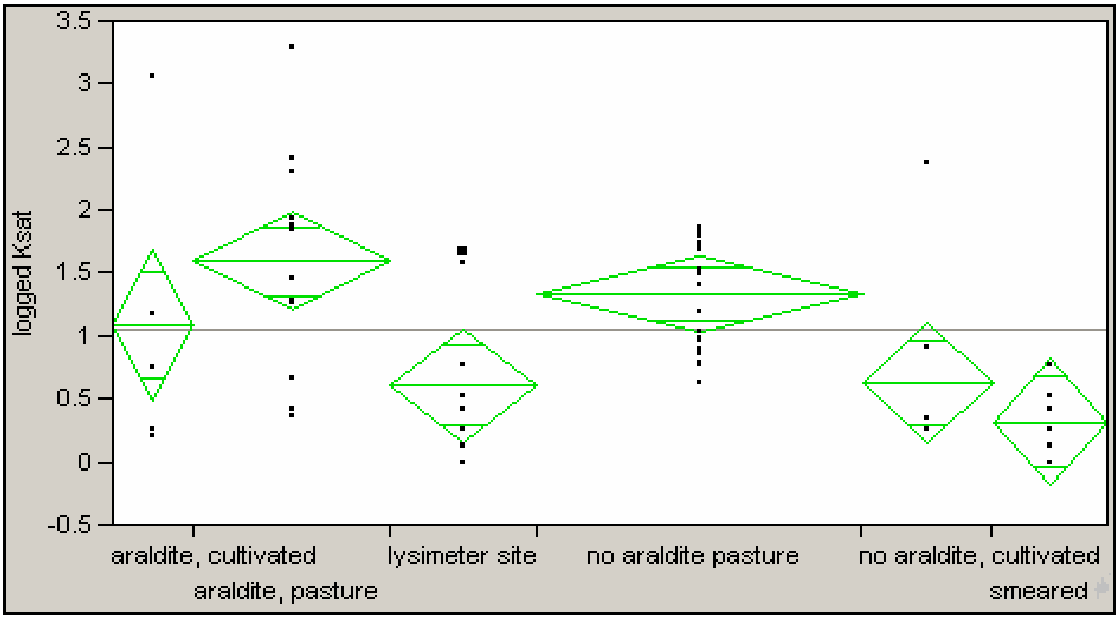


Fig. 21. Means and data distribution for each treatment

The students' t-test analysis showed no statistical difference between various treatments under cultivated conditions, including holes that were deliberately smeared. No statistical difference was observed in either the pasture or cultivated plots between holes treated with araldite and those that were not. Fig. 15 shows the soil surface at the base of the core removed from the borehole, no smearing is observed on this surface, so it is likely that the soil surface at the base of the hole is in a similar condition. This indicates that the soil surface at the base of the hole has not been smeared and treatment with araldite is unnecessary.

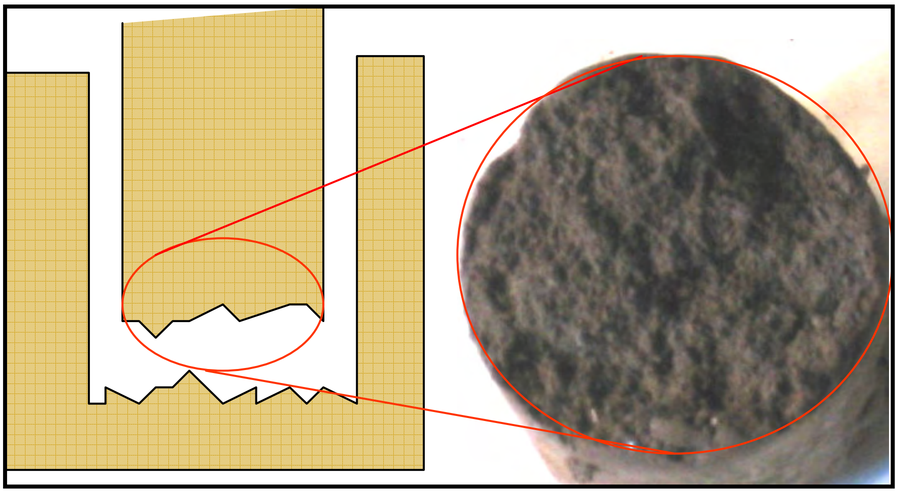


Fig. 22. Soil surface of core removed form borehole

Statistical difference was observed between pasture and cultivated sites. The pasture sites have a higher sample mean, potentially attributable to a greater presence of macropores at depth, not as frequent in the cultivated soil. Cracks were more visually apparent on the surface of the pasture compared to the cultivated site. If cracks originate at the surface, the higher infiltration observed in the pasture sites would be explained, as there would be no cracks at depth in the cultivated field and therefore no option for rapid bypass flow of water.

Due to the large amount of variation in the results it is necessary to have sample size whose mean provides an accurate representation of the infiltration occurring, so how many holes are required?

A population with n samples has a sample mean $\bar{x}$, an approximation of μ, the population mean, for the objective sample size. Once data has been collected, it is possible to determine how many samples are required to estimate, with a desired percentage of confidence, a mean within the range of μ-d and μ+d; d is representative of a limit within which the sample mean is to be found. Warrick, (1998) has described this method using Equation 9, to determine the required n.

$$n = Z_{0.5\alpha}^2 \frac{\sigma^2}{d^2} \qquad (9)$$

where n is number of samples required, $Z_{0.5\alpha}^2$ is the normalized difference from the mean, which can be obtained from the literature, σ^2 is the population variance, which can be equated to the sample variance (s^2) and d the tolerance.

Table 3

Number of holes required to be within 10% of the mean with 80% and 90% confidence for the various treatments tested

Treatment	Number of samples used	80% confidence	90% confidence
No araldite pasture	41	3	5
Araldite, pasture	12	7	12
Araldite, cultivated	5	30	50
No araldite, cultivated	8	37	61
Lysimeter site, cultivated	9	28	46

Table 4

Number of holes required to be within 10% of the mean with 80% and 90% confidence for an amalgamation of the cultivated and pasture areas

Treatment	Number of samples used	80% confidence	90% confidence
Pasture	53	6	9
Cultivated	22	32	53

Based on a log-normal distribution, using Equation 9 the number of holes required such that the average K_{sat} value will be within 10% of the population mean with an 80%, or 90% confidence was calculated, for all of the treatments, shown in Table 3, and for the pasture and cultivated sites, after amalgamating treatments that were not statistically different, shown in Table 4.

A sample mean within 10% of the population mean is considered adequate for the purposes of this research. There is a trade-off when selecting the required confidence for the estimate. The greater the desired confidence the more samples required; an 80% confidence level was considered reasonable, with the number of holes required being a realistic and practical amount.

3.2. Applying the method

Following the adaptation of this method for use on the vertosols of NW NSW it required testing in a field where previously collected data is available. Field 11 at Auscott Moree was chosen; previously EM surveys had been performed and potential DD estimated using the SaLF model (Huckel, 2001).

3.2.1. Soil classification of field 11

The field was classified with fuzzy k-means, creating two to seven classes inclusively, using the Diagonal metric, a fuzziness exponent of 1.3 and no extragrades. The basis for this classification was four variables, the EM31 and EM38 data collected in both the horizontal and vertical modes (Huckel, 2001).

Fig. 16 shows all three performance indicators exhibit a local minimum at three and five classes. Three classes was selected as most appropriate since the area is known to have three possible sources of parent material (Stannard and Kelly, 1968).

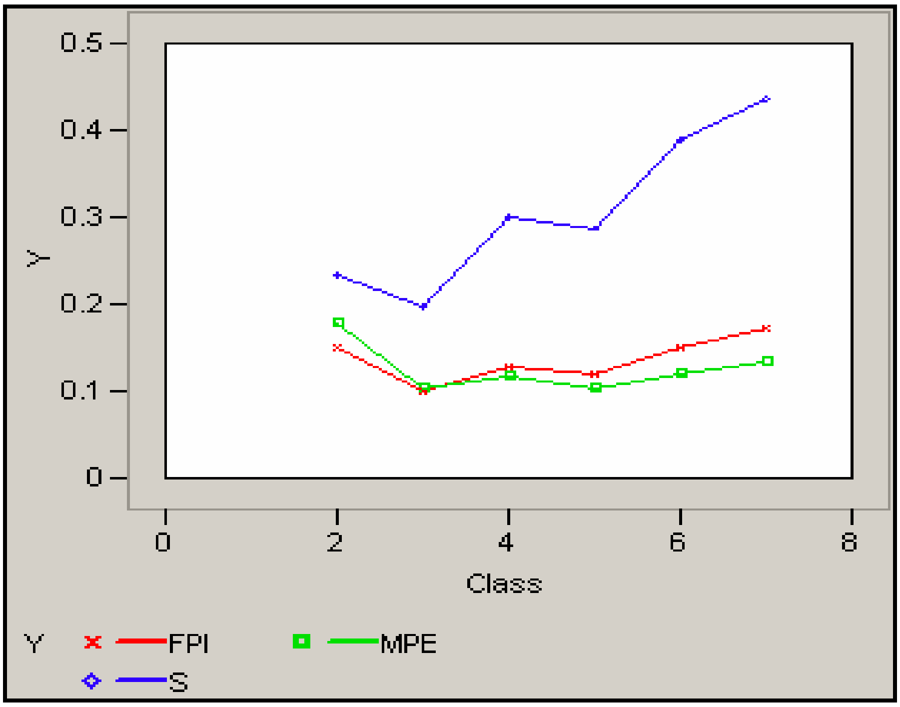


Fig. 23. Performance indicators vs number of classes from fuzzy k-means analysis output

The statistical package JMP5 (SAS Institute Inc., 2003) was used to create 3 classes, allocating each geographical location to a class based on bulk soil Electrical Conductivity (EC_a) readings, taken with Electromagnetic Induction Instrumentation (EMI). Using this information a class map of the field was developed, Fig. 14. Comparing this map with the aerial photo of the field, Fig. 2, it can be seen that class 3 coincides with where the prior stream channel can be observed and is derived from the soils of the prior stream channel. Table 5 provides the means and standard deviations of EC data, obtained with the EMI instruments in various modes, and used as the basis for developing classes.

Table 5

Class means and standard deviations of EM31 and EM38 EC_a readings in vertical and horizontal modes

	EM38v		EM31v		EM38h		EM31h	
Class	**Mean**	**Standard Deviation**	**Mean**	**Standard Deviation**	**Mean**	**Standard Deviation**	**Mean**	**Standard Deviation**
1	147.2	7.7	183.4	7.7	92.2	6.4	125.3	6.5
2	131.6	8.4	166.6	5.7	82.4	7.1	112.3	4.7
3	106.4	14.2	142.5	12.4	67.5	10.8	95.6	9.2

It can be seen in Table 5 that the mean EC_a values for class 3 are much lower for all EM variables than classes 1 and 2. As class 3 represents the soils of a palaeo-channel, this makes sense; these soils would have lower clay contents, and most likely a lower moisture content and therefore a lower EC_a.

A distribution of sites over the whole field was desired, for representation of all soil types present. 32 holes are required to describe soil variation (*2.2.4.*). It was decided that the sampling would be divided into six sites with five or six holes at each site.

The statistical package JMP5 (SAS Institute Inc., 2003) was used to create 6 classes, based on geographical location, within each soil class. A central point in each of these classes was then selected as the sample site. Huckel (2001) performed soil analysis, including particle size analysis, at 81 validation and calibration sites, to account for EC_a variation across field 11. At 0.9 m - 1.2 m depth the average clay content of these sites was 51 % (Huckel, 2001). To ensure adequate description of class 3 an extra five sites, selected from Huckel (2001)'s validation and calibration sites, located on the palaeo-channel, and having a lower than average clay content, were selected. The locations of these sites are shown in Fig. 17.

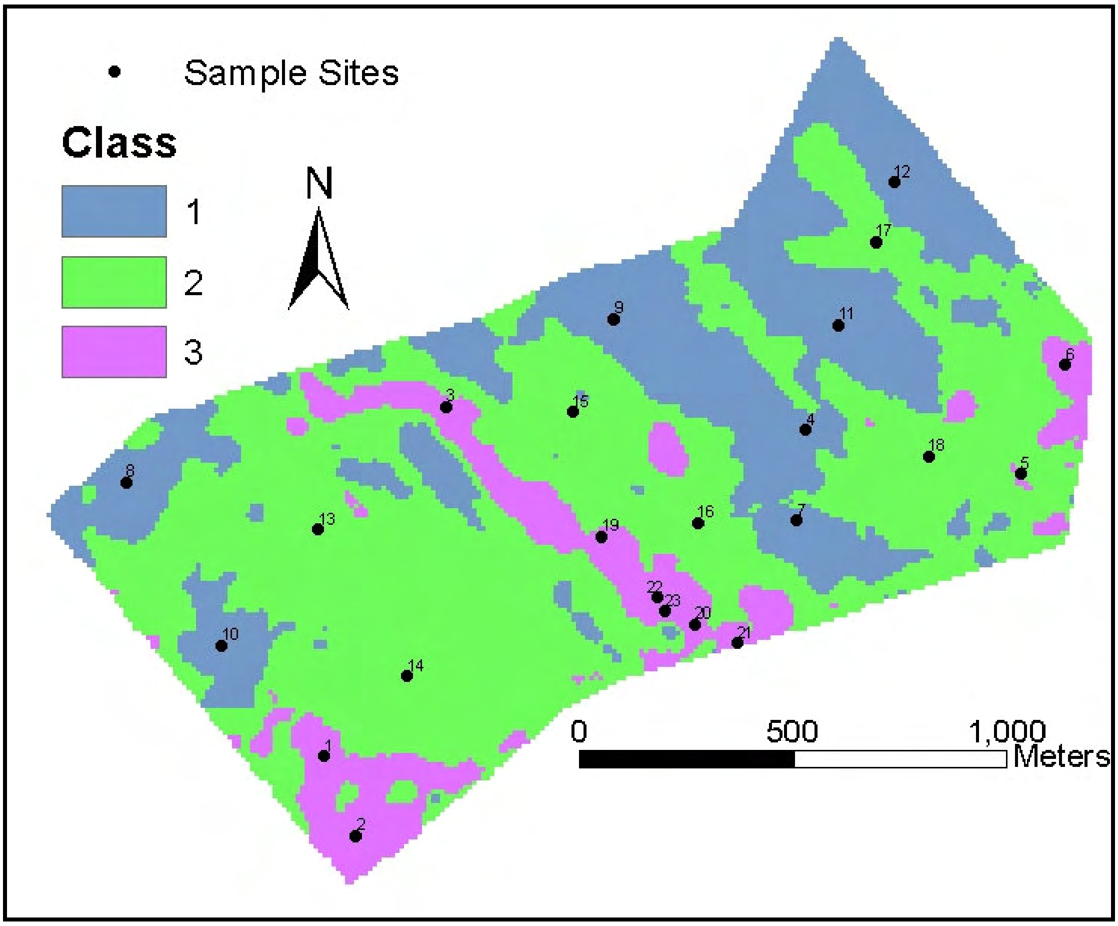


Fig. 24. Soil class map of field 11 with 3 soil classes with sample sites, based on EM readings

At each site soil was hand textured using the bolus method; Table 6 shows these textures and the texture range for each soil class.

Table 6
Textures of each sample site in field 11, Auscott Moree
Class 3 incudes 3 and 3*, * indicating site chosen directly on the palaeo-channel

Site	Class	Texture	Texture Range
1	3	Heavy Clay	Fine Sandy Clay Loam – Heavy
2	3	Light Medium Clay	Clay
3	3	Medium Clay	
5	3	Heavy Clay	
6	3	Light Medium Clay	
19	3*	Light Clay	Fine Sandy Clay Loam – Light
20	3*	Clay Loam	Medium Clay
21	3*	Fine Sandy Clay Loam	
22	3*	Sandy Clay	

Table 6 (continued)

23	3*	Light Medium Clay	
4	2	Light Medium Clay	Light Medium Clay– Heavy Clay
7	2	Medium Heavy Clay	
8	2	Medium Clay	
9	2	Light Medium Clay	
10	2	Medium Clay	
11	2	Light Medium Clay	
12	2	Heavy Clay	
13	1	Medium Clay	Light Medium Clay– Medium Clay
14	1	Medium Clay	
15	1	Medium Clay	
16	1	Medium Heavy Clay	
17	1	Light Medium Clay	
18	1	Medium Heavy Clay	

3.2.2. K_{sat}

At each site 5 replicates were made, the mean and standard deviation K_{sat} values for each site and class were determined and can be seen in Table 7.

dD/dt and D_o were measured in the field, r_o is 12.5 mm and $\Delta\theta$ was taken as 0.1 (discussed in *3.3.*) for all sites. For the ARCI sites and classes 1 and 2 in Field 11, C was 83.33, for class 3 in field 11, a C of 27.78 was used.

Table 7
Site and class K_{sat} mean and standard deviations
Class 3 incudes 3*, * indicates a site chosen directly on the palaeo-channel

Site	K_{sat} site mean (mm/day)	Standard deviation	Class	K_{sat} class mean (mm/day)	Standard deviation	Statistical Comparison
19	12.2	1.9	3 *			
20	8.2	1.6	3 *			
21	24.8	1.7	3 *	11.5	8.0	A
22	8.3	2.2	3 *			
23	4.1	1.3	3 *			
1	7.3	3.8	3	7.2		
2	1.4	1.6	3	*- includes all of*	7.2	A
3	1.4	1.3	3	*class 3*		

Table 7 (continued)

5	3.5	1.3	3			
6	0.9	3.2	3			
4	0.5	2.8	2			
7	0.6	2.0	2			
8	0.6	2.7	2			
9	0.3	3.2	2	0.5	0.2	B
10	0.3	2.2	2			
11	0.8	1.7	2			
12	0.4	1.8	2			
13	1.5	1.3	1			
14	0.7	2.	1			
15	0.4	3.1	1			
16	1.0	1.5	1	1.0	0.4	B
17	1.2	2.5	1			
18	1.2	1.7	1			

To determine if classes were from statistically different populations, a students t-test analysis was performed. Table 7 shows the results for field 11, where the same letter connects classes not exhibiting statistical difference.

For the sake of comparison,

Table **2** shows the K_{sat} values measured under various treatments at the ACRI, Narrabri.

3.2.3. Change in soil moisture content

To determine soil moisture content at the infiltrating surface presents a practical issue; it is at 1 m depth and the hole is only 5 cm in diameter. Using a moisture sensor is therefore not possible. Moisture content can be determined in the lab from a sample. However, collection of the wet sample is difficult; the pipe must first be removed, this causes excess water to empty into the hole, and not every site was sampled. Table 8 shows the change in volumetric soil moisture content that occurred during the infiltration process for those sites where a wet sample was able to be collected; the mean value was 0.1 cm^3/cm^3.

Table 8 Change in volumetric moisture content during infiltration for some of the sites in field 11

Site	Change in volumetric moisture content (cm^3/cm^3)
2	0.08
6	0.07
4	0.12
9	0.08
11	0.13
14	0.19
15	0.06
17	0.05
18	0.15
Mean	**0.10**

3.2.4. Number of days saturated

C-probe data collected from 28-10-2004 to 13-3-2005 on field 28 (T. Richards *pers. comm.*, 2005), Auscott Moree was used to estimate the number of days of sub-soil saturation after an irrigation event. These measurements were made to a depth of 0.7 m, the change in relative soil moisture content over time, at this depth, is shown in Fig. 18 for a period during which three irrigations were applied. For each irrigation event an estimate of the sub-soil saturation time was noted, an average these time periods was taken. The soil remains approximately saturated for roughly four days, after an irrigation event.

Using the industry standard of six irrigations per season there will be approximately 24 days over the irrigation season for which the sub-soil will be saturated.C-probe data was also collected on the lysimeter site at the ACRI from 1-12-2003 – 25-3-2004 (D. Richards *pers. comm.*, 2005) in 20cm increments to a depth of 120cm. The change in moisture content over time for a depth of 1m is shown in Fig. 19, during which time three irrigation events occurred. From this data the subsoil saturation time per irrigation is estimated to be an average of 6.5 days. For six irrigations per season, the soil will be saturated for 39 days.

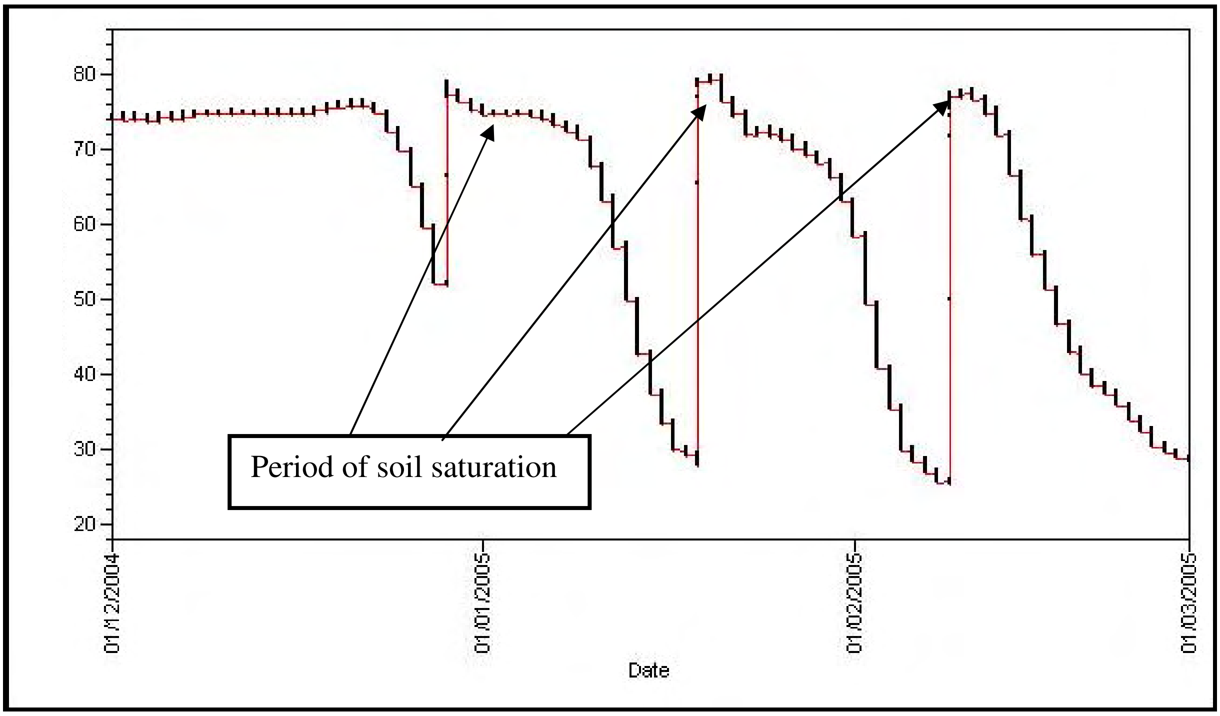


Fig. 25. Relative soil moisture content over time, measured with a C-probe at Auscott Moree, at 0.7 m depth

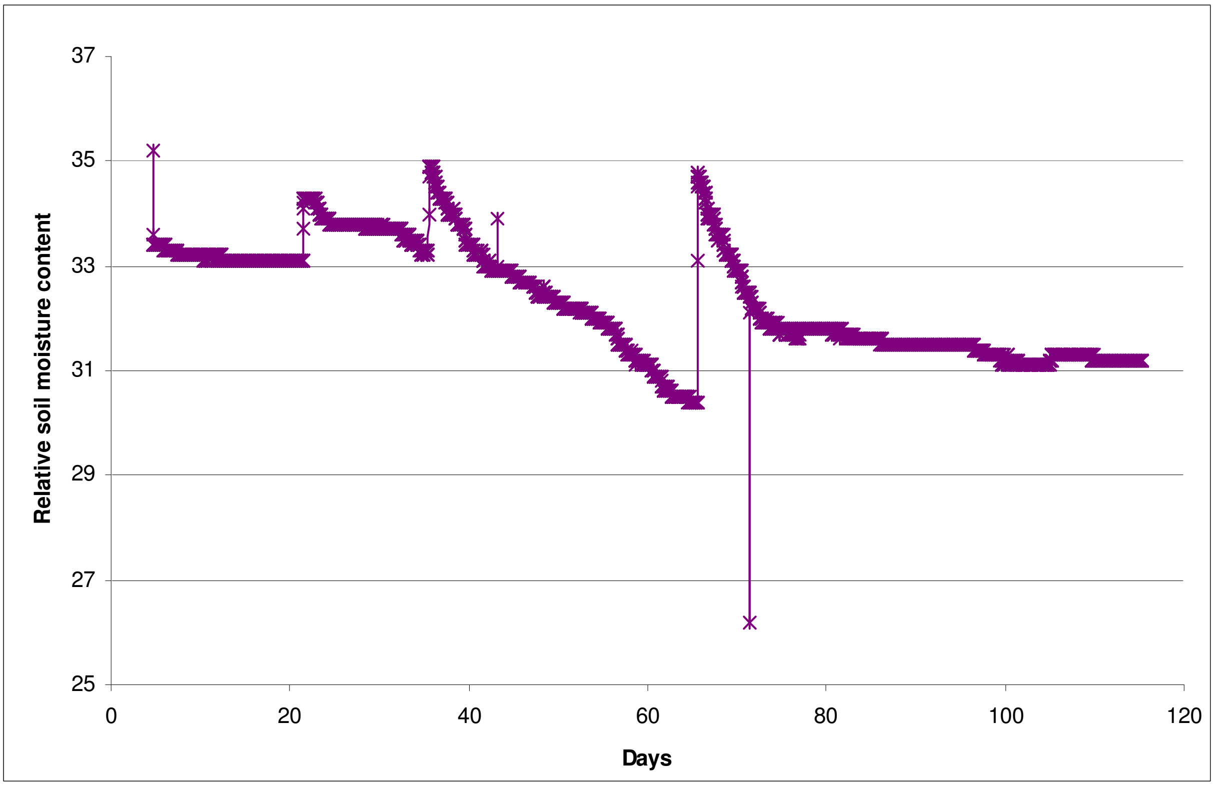


Fig. 26. Relative soil moisture content over time, measured with a C-probe at the ACRI, the lysimeter site over a 4 month period, at 1 m depth

3.2.5. Estimates of potential deep drainage

As the ACRI moisture content measurements were made at 1 m, the appropriate depth, sub-soil saturation of 39 days was used to estimate potential DD in field 11 (Table 9), and for the sites at the ACRI where araldite was not used (Table 10).

Table 9 Estimates of potential deep drainage occurring in field 11, and at the ACRI, over a season

Site	Potential DD (mm/season)	Class	Class mean (mm/season)
19	477.0	3 *	
20	321.0	3 *	
21	968.0	3 *	450.0
22	325.3	3 *	
23	158.7	3 *	
1	284.3	3	
2	55.4	3	
3	53.8	3	281.3
5	134.9	3	
6	34.3	3	
4	19.1	2	
7	21.8	2	
8	21.5	2	
9	13.3	2	19.3
10	13.3	2	
11	31.2	2	
12	15.2	2	
13	57.3	1	
14	27.3	1	
15	15.2	1	
16	39.0	1	39.0
17	48.4	1	
18	46.8	1	

Table 10 Estimates of potential deep drainage occurring in field 11, and at the ACRI, over a season

Site	**Mean potential DD (mm/season)**
Pasture	921.2
Cultivated	163.4
Lysimeter site	157.6

3.2.6. The leaching requirement

The leaching requirement can be calculated if crop salinity tolerance and the volume and electrical conductivity (EC) of applied irrigation water is known. Cotton has a salinity tolerance of 7.7 dS/m for a 10 % yield loss; in Narrabri there are two sources from which water can be obtained for irrigation: bore water which has an EC of 4 dS/m, and the Namoi river which in March 2004 had an EC of 1.1 dS/m (Anonymous, 2004b). In Moree, irrigation water may be sourced from the Gwydir river, which had an EC of 0.24 dS/m in January 2004 (Anonymous, 2004b). Table 11 shows the results when irrigating cotton with 700 mm.

The volume of drainage required will vary according to Equation 8 with changing input variables. Observing this variation allows detection of financial, environmental and practical limitations to irrigation. For a particular crop tolerance, the required drainage can be plotted against irrigation water EC and volume allowing observation of its response. Fig.s 20-22 show these changes for a range of critical EC conditions.

Table 11 Leaching and drainage requirement for cotton irrigated with 700 mm

Water Source	Leaching requirement	Drainage Requirement (mm)
Narrabri Bore water	**0.12**	**81.2**
Namoi	**0.03**	**20.6**
Gwydir	**0.006**	**4.4**

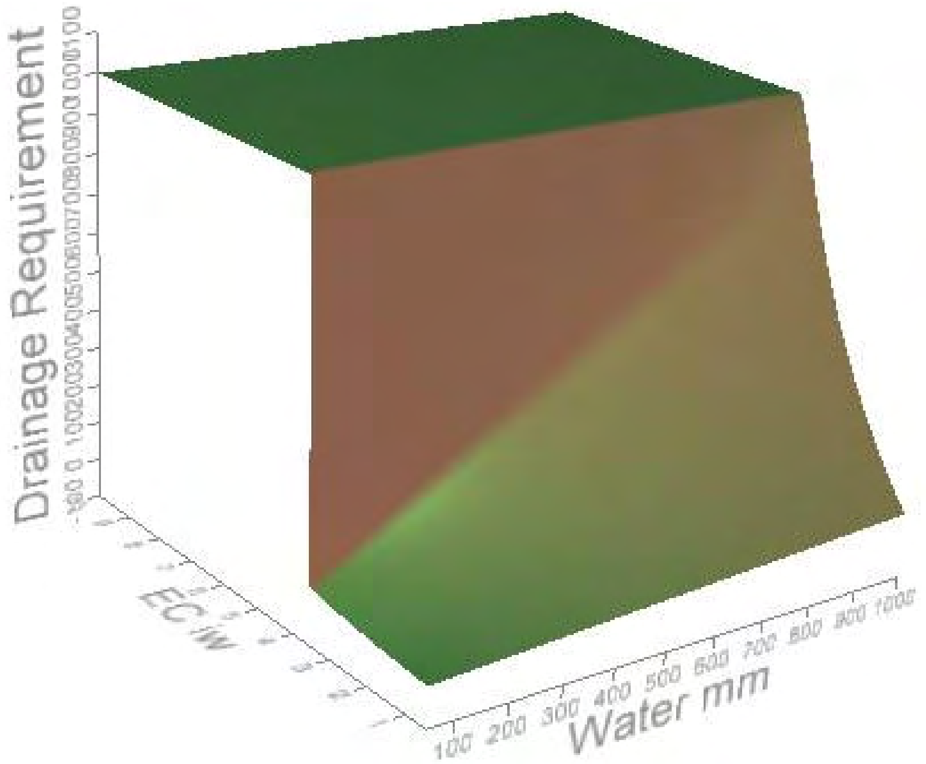


Fig. 27 Response of drainage requirement to changes in irrigation water volume and irrigation water electrical conductivity, for a critical EC of 1.5 dS/m

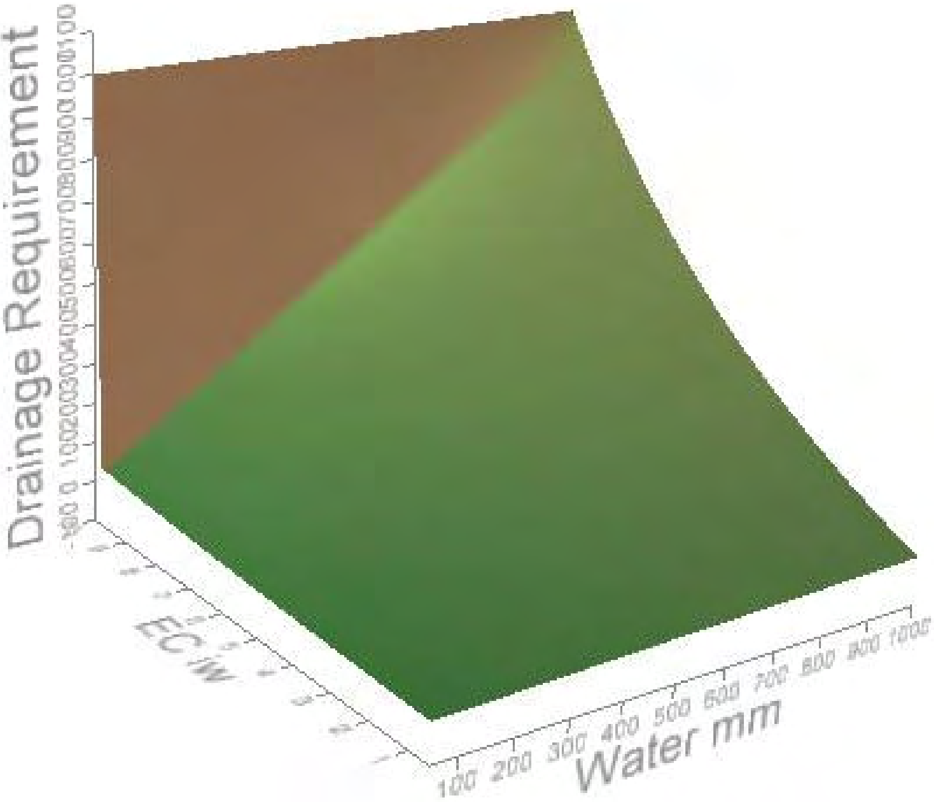


Fig. 28. Drainage response to changes in irrigation water volume and irrigation water electrical conductivity, for a critical EC of 4 dS/m

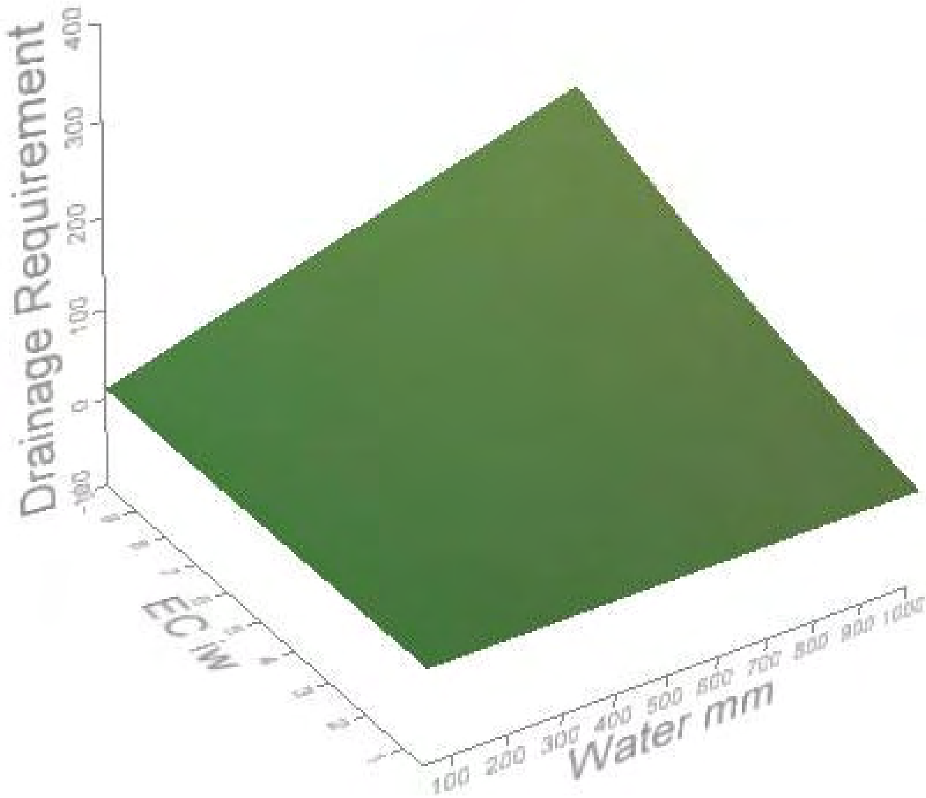


Fig. 29. Drainage response to changes in irrigation water volume and irrigation water electrical conductivity, for a critical EC of 10.5 dS/m

4. Discussion

4.1. Sensitivity of model to measured parameters

A sensitivity analysis was performed on the model developed by Philips', Equation 1, to determine the influence of dD/dt, C and $\Delta\theta$, on K_{sat}.

The results obtained in Moree and Narrabri showed dD/dt varying between 0.00005 and 1.57 cm/min, so dD/dt was varied over the range of 0.00001 and 10 cm/min. $\Delta\theta$, is by definition confined to vary between 0 and 1, however this range is unrealistic so for the purpose of this analysis it has been varied between 0.08 and 0.39. Reynolds et al., (2002b) reported C values to vary between 2.8 and 100, Table 1, for a variety of soil types; C was varied between 10 and 100 to account for agricultural soil types.

Fig. 23 shows that for C and dD/dt having constant values, as $\Delta\theta$ varies the variation in K_{sat} is relatively small, as it does not exert a large influence on K_{sat}. Within the tested range of $\Delta\theta$, K_{sat} varies by a factor of 1.15; however the variation in K_{sat} will be less, as $\Delta\theta$ variation will not be this much. K_{sat} variation of this magnitude is not considered significant, therefore the impact of Δθ on the estimated potential DD is considered negligible. This is fortunate as accurately determining Δθ is difficult if the method is to satisfy the criteria of quick, simple and easy. For calculating DD Δθ = 0.1 will be used, as this was the measured average, Table 8; this is sensible in terms of the likely moisture content change.

Fig. 24 shows the influence of dD/dt and C on K_{sat} for constant Δθ. For each factor increase in dD/dt there is a linear increase in K_{sat}, which makes sense in terms of Equation 1. dD/dt is a multiplying factor at the front of the equation.

Variation in C exerts an almost linear trend; for each factor increase in C, K_{sat} decreases to approximately 0.85 of its value.

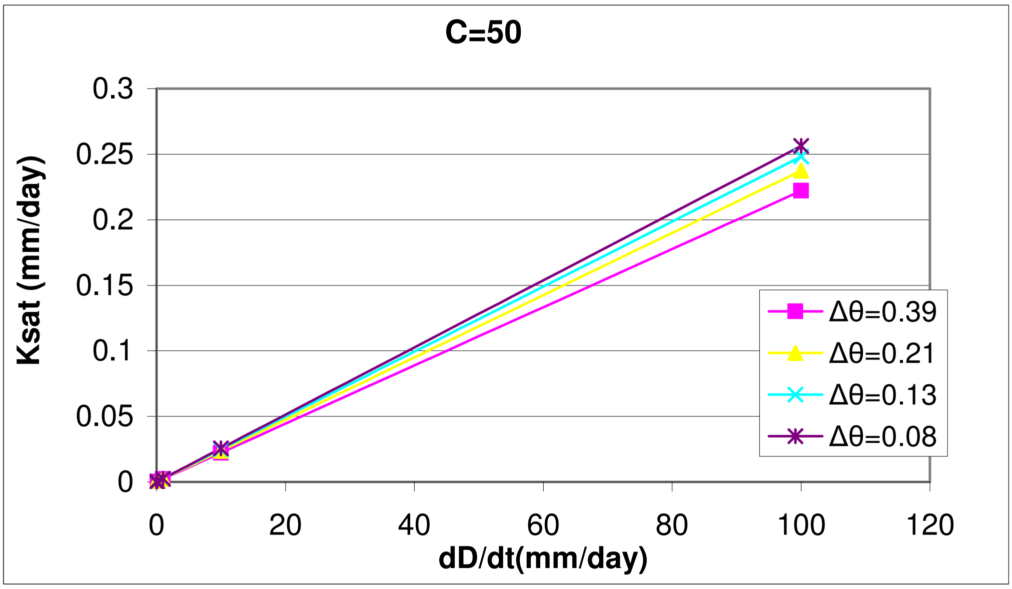


Fig. 30. K_{sat} vs dD/dt at various values of Δθ for constant C

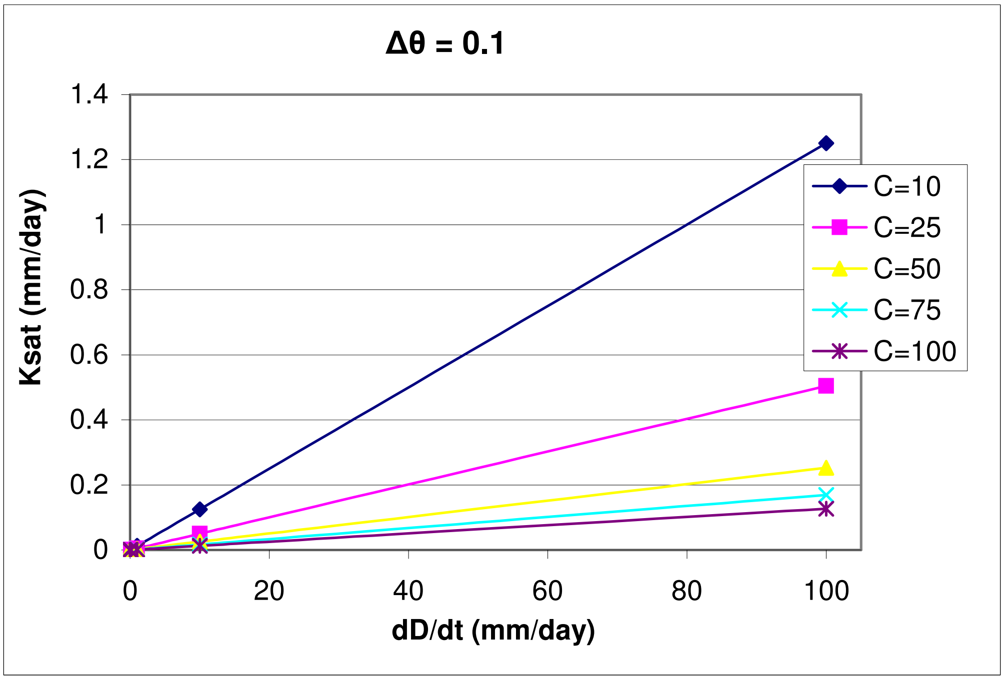


Fig. 31. K_{sat} vs dD/dt at various values of C for constant $\Delta\theta$

4.2. Comparing measured Ksat to previous studies

4.2.1. ACRI

At the lysimeter site estimates of K_{sat} were made using a ponded disc permeameter and 20cm diameter cores with a constant head (A. Ringrose-Voase *pers. comm.*, 2005), which can be compared to the FHLBT method, Table 12. The ponded disc permeameter is a standard technique used for the determination of K_{sat} in the field. The mean K_{sat} determined using the FHLBT was different by a factor 2 from that of the ponded disc permeameter, indicating this method is able to give reasonable in-field results. The K_{sat} determined using a core under a constant head of water is considerably larger in value, which may be attributed to varied conditions; this measurement being made in the lab, whereas the other measurements were taken in the field.

Table 12 Comparison of K_{sat} values obtained at the lysimeter site

Method of analysis	**K_{sat} - mean**	**Standard deviation**	**Number of replicates**
FHLBT (in-field)	11.2	17.5	9
Ponded disc permeameter (in-field)	22.5	11.3	4
Constant head core (laboratory)	444.2	764.7	3

4.2.2. Field 11, Auscott Moree

Montgomery (2003) measured the K_{sat} of a black vertosol and a brown vertosol in field 12, Auscott Moree, which is adjacent to and has the same parent material as field 11. These black and brown vertosols have the same parent material as classes 1 and 2, and class 3 in field 11, respectively. These K_{sat} values were obtained with a constant head well permeameter and by determining flux based on soil profile moisture content change over time.

Table 13 summarises these K_{sat} values and in comparison to the FHLBT, are much higher. This is not a direct indication of the accuracy of either method, but illustrates the large variation, which can occur through the use of various methods.

4.3. Deciding upon the duration of sub-soil saturation

In this study sub-soil moisture content data was not collected on field 11, and hence the estimated number of days saturated will not be at its most accurate. However the data used was collected on the same soil type, and likely to have drainage patterns, which are almost the same. The estimated period of saturation is, however, only approximate; an average of the saturation period for each irrigation.

Fig. 25 shows variation of hydraulic conductivity (K) with soil moisture content for a variety of soil types. As moisture content decreases below saturation, there is a rapid decline in K, as such the water flux and hence the DD occurring when the soil is unsaturated is negligible, relatively, and not accounted for.

Table 13 Estimates of K_{sat} made using a constant-head well permeameter and flux measurements on field 12 Auscott Moree (Montgomery, 2003)

Method	Soil (depth)	Mean K_{sat} (mm/day)
Constant head well permeameter	Black vertosol (85-115 cm)	14.4
	Black vertosol (135-165 cm)	16.8
	Brown vertosol (85-115 cm)	16.8
	Brown vertosol (135-165 cm)	93.6
Moisture content calculations	Black vertosol (100cm)	16.8
FHLBT	Class 1	1.0
FHLBT	Class 2	0.5
FHLBT	Class 3	7.2

4.4. DD variation across field 11 and how it compares to the literature

The sites located in class 3, exhibit the highest potential DD, and are statistically different from the values of classes 1 and 2. This is expected as the soils in this class are of a lower and therefore more conductive texture.

The sites located specifically on the palaeo-channel, having known lower clay contents, had the highest DD values, were statistically the same as the other class 3 samples, and different to those of other classes. The samples in class 3, but not specifically on the palaeo-channel, may not be directly on the prior stream channel, but in the transition between soil classes. The drainage at these sites was statistically different to the other classes, but lesser in value than the sites where clay content was known to be low.

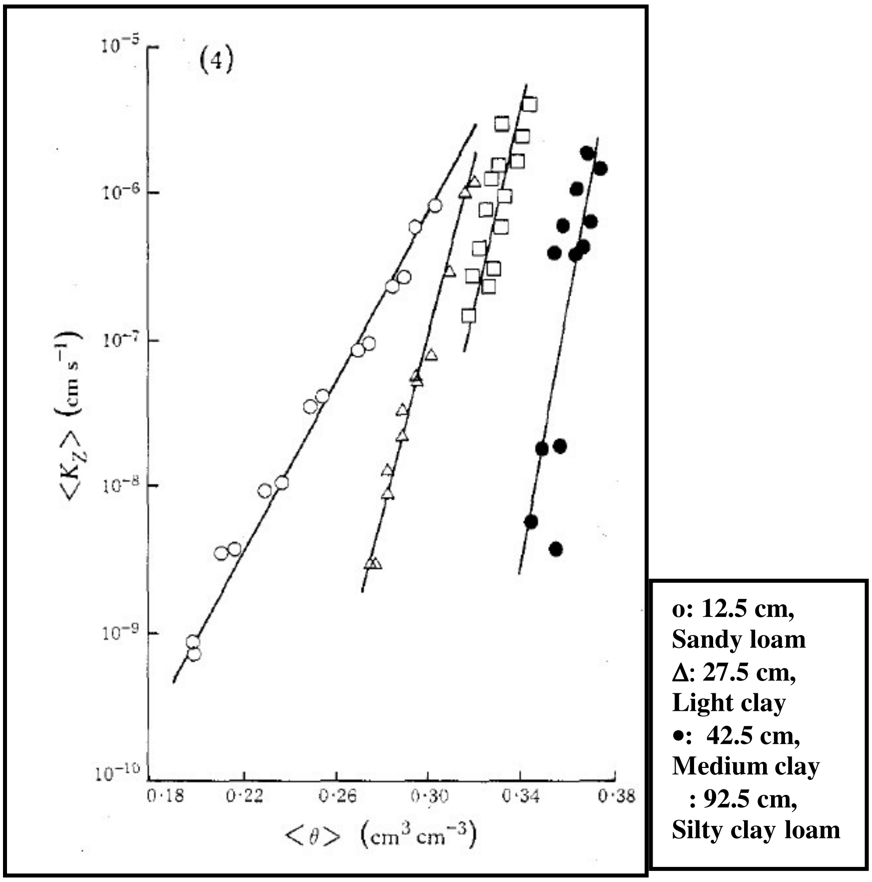


Fig. 32. Relationship between hydraulic conductivity and volumetric water content for a profile at different depths (Olsen and Rose, 1978).

In calculating potential DD, knowledge of the soil type is required such that the appropriate C (wetting front potential) value can be applied when evaluating K_{sat}. Prior knowledge is also useful such that potentially leaky areas, which may be missed in a random sampling regime, can be included in the sample set.

Montgomery (2003) made estimates of the DD occurring on field 12, adjacent to field 11, at Auscott Moree using a number of techniques. Components of the soil water balance were measured and the equation arranged and solved for DD. Darcian flux calculations were performed using K_{sat} measured with a constant-head well permeameter, using the chloride mass balance model SODICS developed by Rose et al. (1979), and based on water flux measurements from in situ water content measurements. These results are shown in Table 14.

The estimates made by Montgomery (2003) (Table 14) are of the occurring DD, estimates made with the FHLBT (Table 9), are for the DD that would potentially occur with the application of a specific volume of irrigation, and does not account for precipitation. The estimates made using darcian flux calculations (Table 14) are much larger, up to two orders of magnitude, than those estimated by this research for the same soil type. The K_{sat} values estimated by Montgomery (2003) were considerably larger than those found with the FHLBT, which is the variable causing this difference. The DD estimated with the SODICS model (Table 14) is very similar to that of the FHLBT, and the within the same order of magnitude as the estimates using the soil water balance equation.

Table 14

Estimates of deep drainage made on field 12, Auscott Moree by Montgomery (2003), for the black vertosol

Method		DD (mm)
DD by difference	Irrigation 1	24.5
	Irrigation 2	21.6
	Irrigation 3	26.5
	Irrigation 4	15.3
	Average	**22.0**
	Total for 6 irrigations	**131.8**
Darcian flux calculations *	K_{sat} (well permeameter)	2353.0
	K_{sat} (in situ moisture content)	860.0
SODICS (yearly average of a 17 year period)		34.5

* This drainage is for the period 12/11/96 – 17/3/97

Of the methods used by Montgomery (2003) the soil water balance calculation is considered the most accurate; that the FHLBT results are not too far from this is a positive indication that the method is giving results in the ballpark of the drainage occurring.

Huckel (2001) performed a survey of field 11 Auscott, Moree to determine clay content from electrical conductivity readings and DD using the SaLF model. The results for sites representative of the soil types present and the field average are shown in Table 15. The palaeo-channel only sites chosen in this research are sites where estimates of DD were made by Huckel (2001), these results are also displayed in Table 15.

All of the estimates made using the FHLBT are larger than those of the SaLF model, some sites show a large difference, while the class averages correlate well with representative soil sites. Since directly measuring DD is so difficult every method is an estimation that does not represent a right or wrong answer, but merely an estimation as good as the assumptions required for that method. Comparing different techniques, while useful, only provides a guide and cannot prove or disprove the use of another method.

Table 15

Summary of DD estimates in field made by Huckel (2001) using the SaLF model and comparison to FHLBT estimates where appropriate

Site	DD estimated with SaLF model (mm/year)	Site	Corresponding DD estimates with FHLBT (mm/year)
Whole field mean	25.5	Whole field mean	98.8
Representative Chromosol	213.1	Class 3	281.3
Representative Vertosol	18.3	Classes 1 and 2	29.2
Sample site 19	54.4	Sample site 19	477.0
Sample site 20	68.5	Sample site 20	321.0
Sample site 21	239.2	Sample site 21	968.0
Sample site 22	121.3	Sample site 22	325.3
Sample site 23	78.3	Sample site 23	158.7

4.5. DD at the lysimeter site

Unfortunately results are not yet available from the lysimeter experiment, and comparisons can only be made in any follow up research conducted.

At the ACRI the estimates of potential DD on the cultivated sites, both irrigated and not, are relatively close at 163 and 158 mm/season; however that of the pasture site is more than 5 times larger, 911 mm/year. This makes sense because in their natural state, vertosols form large macropores, which are able to rapidly conduct large volumes of water through the profile. Cultivation is detrimental to the soils natural structure, removing, these large pores and therefore the soils ability to conduct water as rapidly.

Weaver et al. (2005) used a chloride mass balance model to quantify the DD occurring on field C1, the lysimeter site at the ACRI, Narrabri. Table 16 compares these estimates with those made using the FHLBT.

Table 16

Comparison of deep drainage estimates with the FHLBT and chloride mass balance

Method	Estimated potential DD (mm/year)
FHLBT	157.6
Steady-state chloride mass balance (Weaver et al., 2005)	62 ± 21

The drainage estimates in Fig. 18 are approximately a factor of 2.5 apart; again this is reasonable correlation to a previously tested method, a positive indication of the accuracy and viability of the FHLBT.

4.6. DD with reference to the leaching requirement

Whilst ideally the drainage occurring would equal that required, a safer option is for the drainage occurring to be in excess of the requirement. The risk of salinity affecting the crop is decreased, and the possibility of growing other crops with a lower salinity tolerance at a future time is maintained.

Fig.s 20-22 show that for a given irrigation water volume as the EC of this water increases, the required drainage increases; as the critical EC, which is representative of crop tolerance, increases, the required drainage decreases.

The leaching requirement for continuous cotton in Narrabri was found to be 81 mm and 21 mm (Table 11), depending upon the water source. The estimated potential DD at the lysimeter site, which is used for cotton cropping, was 157 mm (Table 11), which is well above the possible required amount. This indicates that salt accumulation should not be a problem with the current irrigation scheme, and that irrigation is slightly in excess.

Using water from the Gwydir to irrigate continuous cotton cropping 46.5 mm (Table 11) of drainage is required. Field 11 displayed a drainage range of between 13 mm/season and 968 mm/season. However, the majority of the soils in the field are from classes 2 and 3, which have an average potential DD of between 20 mm/season and 40 mm/season. This indicates the drainage occurring is insufficient and salt accumulation inhibitive to plant growth may occur. Increasing the volume of water applied, or applying water of better, (i.e. less saline), quality are options that can be explored to rectify the situation.

4.7. Leaching requirement with reference to crop rotations

Frequently cotton is grown in rotation, and therefore not the only factor needing consideration when determining the leaching requirement. The crop having the lowest salinity tolerance will provide a lower limit for the drainage requirement. Crops commonly grown in rotation with cotton include wheat (*Triticum aestivum L.*), dolichos (*Lablab purpureus L.*), field pea (*Pisum sativum, L.*) and faba bean (*Vicia Faba L.*) (Hulugalle et al., 2001). Table 17 compares the drainage required for cotton with these crops, for 700 mm of irrigation, with an irrigation water quality of 3 dS/m.

As can be seen in Table 17 a crop with a low salinity tolerance, such as Faba bean can require significantly more drainage than cotton. To sustain the successful rotation of these crops the leaching requirement of the lower tolerance crop needs to be fulfilled.

Table 17

Salinity tolerance, leaching and drainage requirements of some crops commonly rotated with cotton, based on salinity tolerance data from

Crop	Salinity tolerance (dS/m)	Leaching Requirement	Drainage Requirement (mm)
Cotton	7.7	0.08	59.2
Wheat	6.0	0.11	77.8
Dolichos	6.8	0.10	67.7
Field pea	2.5	0.32	221.1
Faba bean	1.6	0.60	420.0

This may involve using a larger amount of irrigation water, changing the quality of irrigation water used or re-evaluating the crop schedule in place.

The rooting depth of other crops also needs to be considered, measurement of K_{sat} and soil moisture content may need to be taken at a different depth. Plants that root to a greater depth will use a larger amount of water and may act as a buffer against DD (Boschma and Lodge, 2003).

5. Conclusions

5.1. Conclusions

A method for the in-field measurement of sub-soil K_{sat} using the FHLBT was developed to be suitable for vertosols of the cotton-growing region of NW NSW. From this K_{sat}, potential DD was successfully modelled using the duration of sub-soil saturation, based on soil profile moisture content.

Field 11 at Auscott Moree was classified based on electrical conductivity readings and then surveyed with the FHLBT. Class 3 of field 11 represents the soil of a palaeo-channel and is considered a potentially "leaky area"; the results showed a significant difference between class 3 and the classes derived from alluvial deposition. These results support the theory that class 3 is a "leaky area" and demonstrates the capability of this method to observe differences in soil texture.

Comparison was made between the potential DD estimates and the drainage requirement, and these figures were considered realistic in value and useful for making a judgement concerning the irrigation schedule water-efficiency. If excess drainage is occurring irrigation volume can be reduced, or for insufficient drainage, increased. Altering the quality of applied water can also be an effective tool for bringing the occurrent drainage and its requirement closer together.

Before making management decisions the salinity tolerance of any rotation crops must also be taken into account, and the future use of the field considered. Cotton has a relatively high salinity tolerance. If the soil is raised to this salinity, switching to another crop will be considerably more difficult.

Results obtained with this method display a reasonable correlation to DD values reported in the literature for the same and similar sites, an indication that the method is able to provide values useful for describing the occurrent drainage. The method is low-cost, straightforward to use, easy to understand, able to provide a reasonable degree of accuracy and able to describe soil spatial variability.

Section 3:
Users Manual for Applying the Falling-Head Lined-Borehole Technique

Having an estimate of the potential deep drainage occurring in a field will be meaningless unless the drainage requirement is also known. For this to be calculated you must know how much irrigation water is applied per season and what the salinity (electrical conductivity dS/m) of this water is.

If the soil salinity is to be for a crop other than cotton then the default soil salinity must be changed.

Before conducting this fieldwork it is recommended that there is some form of prior knowledge about the field. This may be in the form of aerial photographs, EM surveys, yield mapping, soil survey and analysis etc., having the ability to differentiate soil types present within the field. It is also necessary that potential leaky areas be accurately located.

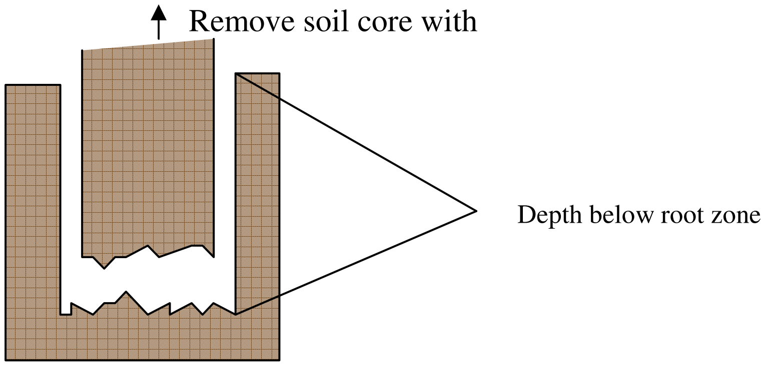
A sampling regime must then be developed. The field should be divided into areas representative of the soil types, and approximately 30 sample sites selected within each area. These sites should be spatially distributed within the class between 6 locations, with 5 samples at each location.

Potentially leaky areas will require extra surveying on top of this. The greater the number of samples, the greater the accuracy achieved, at least 20 sites over four extra locations is recommended, however this will depend on the extent of the area requiring further sampling.

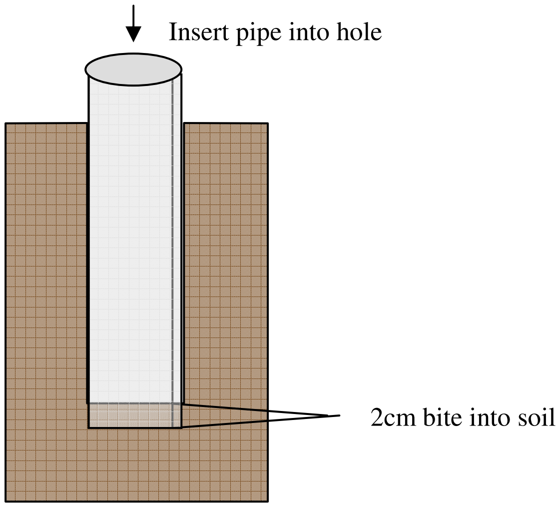
At each location sub-soil K_{sat} is measured using the falling-head lined-borehole technique;

- Based on a assessment with hand classify the soil as either clay, loam or sand
- Determine the moisture condition of the field, wet, medium or dry, based on your prior knowledge or visual assessment.

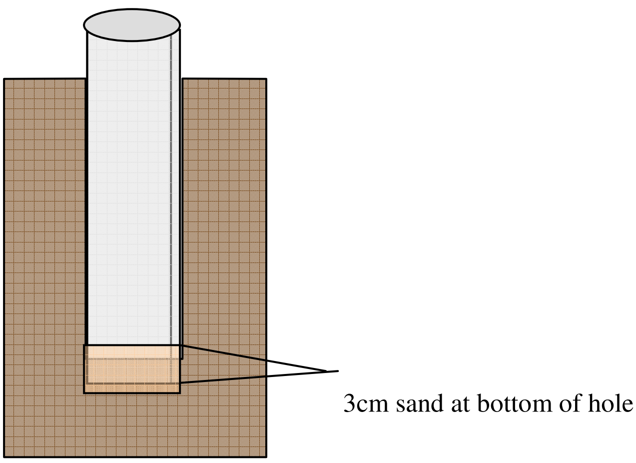
- Holes should be made with a hydraulic ram; this will be the most efficient sampling method. If a hand auger must be used the bottom surface of the hole should be treated an epoxy-resin such as araldite to remove the smeared surface.



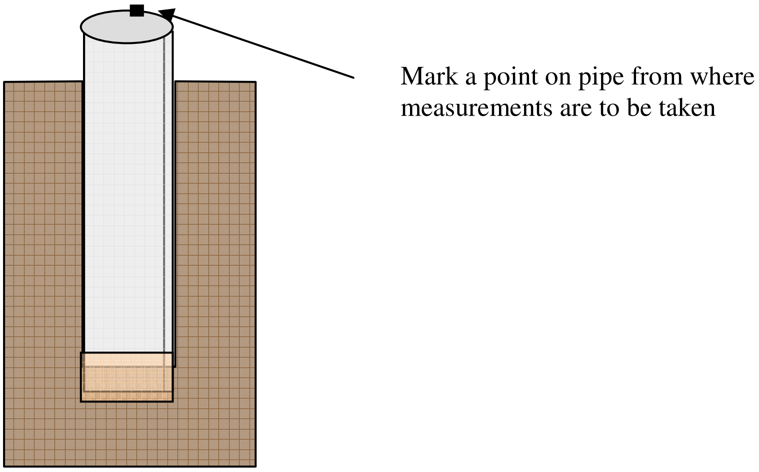
- Holes must be a made to a depth just below the root zone of the crop, or if there is a known constricting layer, measurements should be taken at that depth.
- When inserting the pipe, push it into the soil such that a bite of approximately 2cm is achieved.



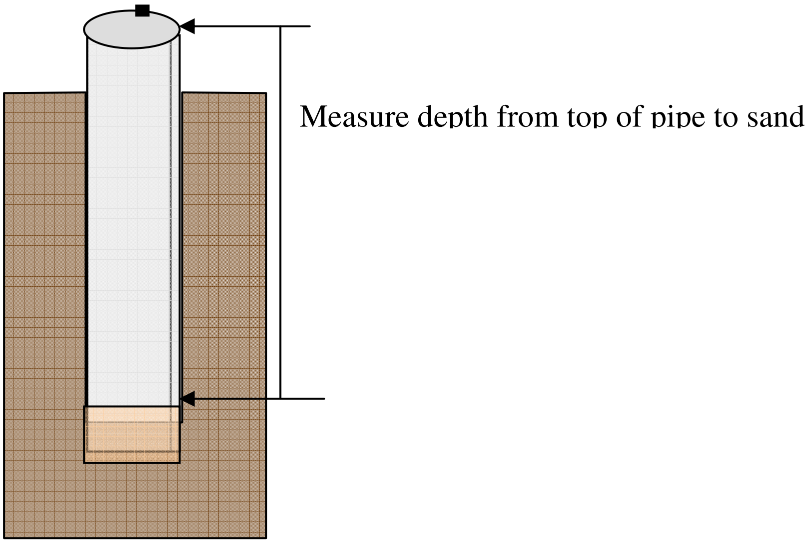
- Add one third of a cup of sand, to the hole, this will prevent destruction of the soil surface when water is poured into the hole.



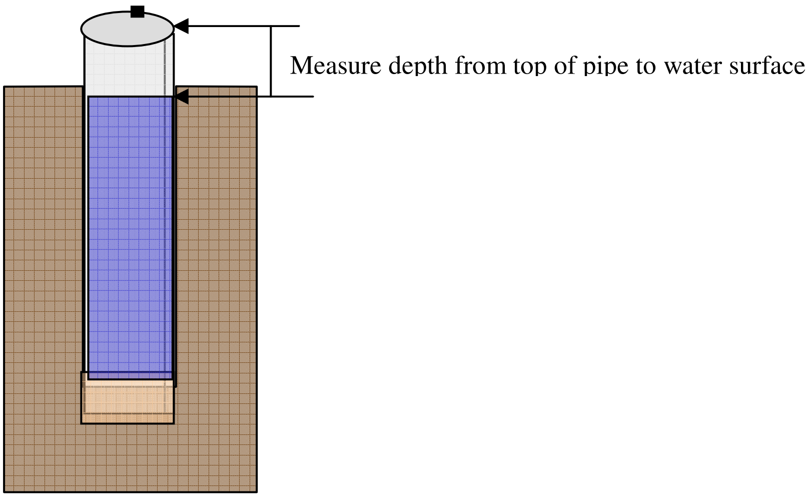
- After pipe insertion, mark a reference point on pipe circumference and always take measurement from this point to ensure consistency of measurement.



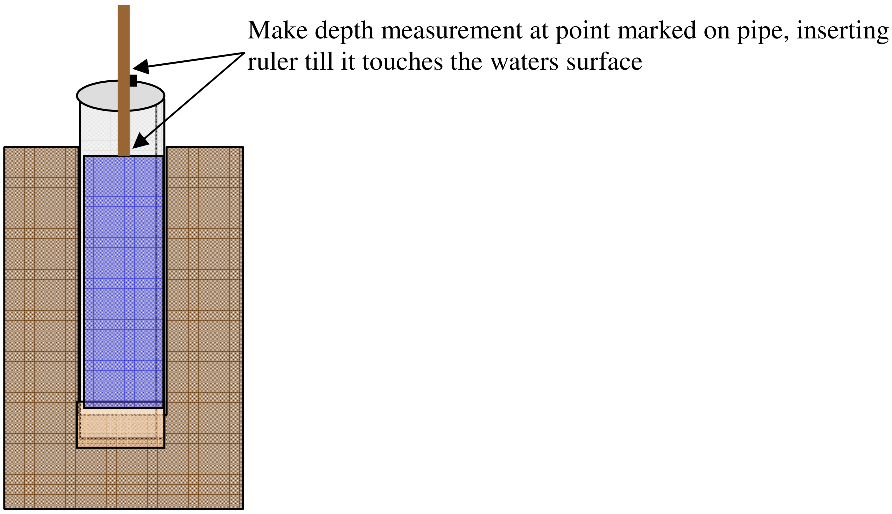
- Measure the depth from surface of the sand to the top of the pipe.



- Fill the hole with water to approximately 10cm below the top of the pipe, measuring the initial depth of water below the top of the pipe.



- The change in water depth over time must then be recorded. For the first 2 days 3 measurements should be made each day, approximately 3 hours apart. By this time the area of infiltration should be saturated. On the last day 5 measurements, approximately 2 hours apart should be made.
- Measurements are made by inserting a stiff graduated ruler into the hole, at the point where the pipe has been marked until the bottom of the ruler just touched the surface of the water.



- Measurements must be taken until steady-state infiltration has been achieved, this time will vary depending upon soil type and initial moisture content, for vertosols in NW NSW 3 days is sufficient.
- These measuring specifications are a rough guide, if water infiltration is occurring at a rapid rate, then the hole should be refilled, and the last 5 measurements, made when soil saturation has occurred may need to be only half to one hour apart.
- To remove pipes it may be easiest to drill two holes on opposite sides of the pipe, secure a nut and bolt through these holes, and pull out, some pipes may require the hydraulic ram for removal.

Sub-soil saturation duration must be determined; this requires soil profile moisture content data for the duration of the irrigation period, it is preferred for this data to be collected in-field, however this may not be practical, and having data from the same soil type is sufficient. From this data the length of time for which the sub-soil is saturated after an irrigation event can be approximated. If the data is collected in field the total time period for which the sub-soil was saturated during the season can be determined. The effect of heavy rain should be considered, when calculating the total time period of sub-soil saturation.

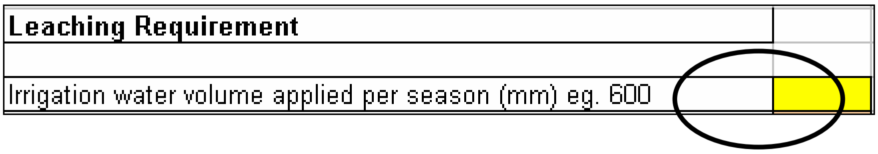
Once all of this data is collected and entered into the spreadsheet, information relating to the efficiency of the irrigation schedule in place for the crops being grown will be output.

The Interface – A User's Guide

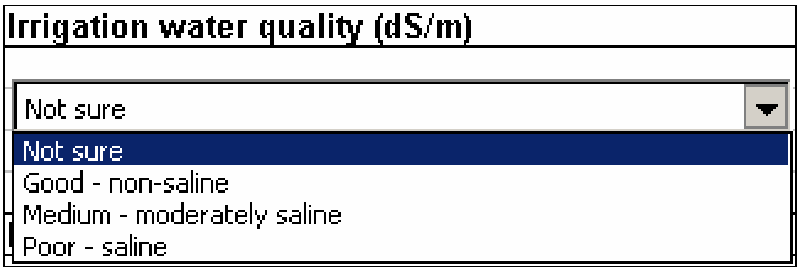
A simple colour scheme has been incorporated into this interface for ease of use. Any box coloured yellow must be filled in for correct function of the model, any box coloured blue may be filled in as necessary, a pink box has already been set to a default value, but should be changed as necessary. Orange boxes require a selection to be made from a drop down menu and green boxes display a result, which may be important for making irrigation decisions.

Data pertaining to the irrigation water and the field under observation must be entered into the model. So you need to know:

- How much irrigation water, altogether, is applied over the season, in mm, enter under the section titled leaching requirement



- How saline is this applied water. If you do not know the exact EC rough selection of its salinity is required; good (eg. non-saline river water), medium, or poor (eg. borewater).Select from a drop down menu.



- What is the salinity tolerance you wish to maintain; this will be for the least tolerant crop in the rotation, which can be selected from a drop down menu.

Least salt tolerant crop
Cotton
Cotton Wheat Sorghum Dolichos Field pea Faba bean

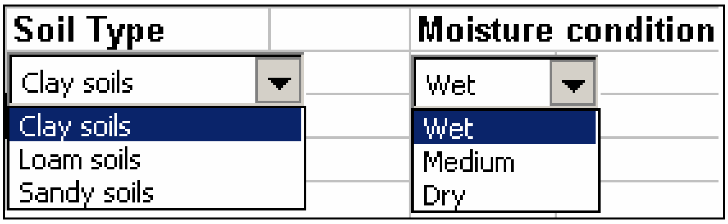
- You must determine the number of days for which the sub-soil is saturated over the growing season, as outlined above.

Potential Deep Drainage	
Sub-soil saturation duration for the irrigation period (days)	

- The depth from the sand to the top of the pipe must be measured, and infiltration measurements using the falling-head lined-borehole technique, as described above, must be taken.

Hole	Time (minutes)	Depth (cm)
1		
Depth from sand to the top of pipe (cm)		
Initial depth to waters surface (cm)		

- For each site where measurements are made the soil type, clay, loam or sand must be selected. Also the initial moisture condition of the soil, wet, medium or dry must be chosen.



Each spreadsheet is sufficient analyse one class, having six sample sites with five holes at each site.

Once all of this information has been entered, the output sheet will provide information about how much leaching is required and how much potential deep drainage is occurring and if this is an efficient management scheme.

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