1.0 Introduction

An important element of sustainable pasture management is the conservation of the biophysical soil resource (Hutchinson, 1992). The soil resource is the basis for plant growth, providing water, nutrients and support to the plant. Unfortunately many Australian farmers are forced to place little emphasis on protecting this important asset, due to pressing financial commitments. In terms of management, soils are often given lowest priority in the short term as it is the animal and cropping enterprises that provide the financial returns. However, the emphasis given to these short term priorities and the neglect of soil conservation issues will inevitably lead in the long term to soil degradation and a decline in pasture productivity.

Soil degradation is affecting the productivity and long-term viability of many Australian soils. Although degradation is generally perceived as salinity and erosion, other forms of soil degradation include nutrient depletion, acidification and soil structural decline.

Soil structure refers to the arrangement of soil particles and the air spaces in between them. A well structured soil allows air, water and nutrients to move freely through the pore space within and between soil aggregates. A poorly structured soil, therefore, adversely affects several factors essential to plant growth including available water, aeration and strength. Soil structural decline refers to the undesirable physical changes in soil structure that result from various land use practices. These physical changes stem from both internal and external forces applied to a soil causing a breakdown in its structure. Internal forces involve the stability of soil aggregates. For example, the presence of water can disrupt a soil aggregate by moving in between soil particles, and pushing out trapped air, resulting in a breakdown of soil structure. External forces are those from machinery, livestock trampling and raindrop impact.

Compaction is an important process leading to soil structural decline. Although it is usually associated with heavy farm machinery, trampling from grazing animals can also be a source of compaction. The ground pressures exerted by livestock are comparable to those exerted by agricultural machinery (Packer, 1988). Soil compaction involves the compression of a mass of soil into a smaller volume when a pressure is applied to the soil (Harris, 1971). The change in volume may be due to the following: 1) a compression of the solid particles, 2) a compression of the liquid and gas within pore spaces and 3) a rearrangement of the soil particles. The severity of compaction depends on the state of the soil and the magnitude of the load applied (Willatt and Pullar, 1983).

The soil water balance describes the additions of water to the soil, storage within the soil and water losses through runoff, deep drainage and evapotranspiration.

Treading by grazing animals can change soil physical properties, including bulk density and porosity (Willatt and Pullar, 1983), strength (Weigel *et al.*, 1990) and aggregate stability (Packer, 1988), which affect the soil water balance. Grazing indirectly affects the water balance by changing the amount and composition of vegetation on the soil surface.

A most important aspect of treading by grazing animals affecting the soil water balance is a reduction in macropores. Macropores play an important role in water infiltration, aeration and rapid drainage. A loss of macropores will reduce infiltration leading to increased runoff. Soil aeration is reduced because of slower drainage of water through the soil. Increases in soil bulk density can restrict root penetration, decreasing the volume of soil which can supply water and nutrients to the plant. Therefore, compaction reduces the amount of water that is potentially available to pasture plants.

A greater understanding of the soil-plant system will lead to improved pasture management practices and increased production. A greater understanding of the processes involved in this system can be gained through the use of modelling. Models can be used to determine the effect of certain parameters on a process or they may provide estimates of actual quantities. For example, the Soil Water Infiltration and Movement (SWIM) model has been developed to quantify the soil water balance. SWIM is a water balance model based on the soil hydraulic properties that can simulate water movement through a given soil.

Little research has been conducted on the effects of grazing on the soil water balance. SWIM may provide a means of investigating these effects.

The overall object of the research described in this thesis was to address some aspects of these issues. The specific aims were:

(1) to determine the surface hydraulic properties of a gleyed podzolic soil and investigate changes in these properties as a result of animal grazing;

(2) to determine the hydraulic properties of the gleyed podzolic soil in order to evaluate the SWIM model for its prediction of drainage;

(3) to examine the consequences of changed hydrological properties under grazing on the soil water balance, using SWIM;

(4) to determine the sensitivity of SWIM output to changes in input parameters.

Part I of this study comprises three chapters that review literature dealing with soil water. Chapter 2 reviews literature associated with the principles and measurement of soil water content, potential and water movement in the soil. Literature associated with the soil water balance, the effects of grazing animals on the soil water balance and water balance modelling are reviewed in Chapter 3.

Part II is an experimental section consisting of a sequence of experiments carried out to examine the effects of grazing on the soil water balance. Chapter 4 provides a detailed description of the experimental site. Soil hydraulic properties are measured in Chapters 5 and 6, providing essential data to examine the differences in hydraulic properties under two grazing treatments. An evaluation of SWIM for its prediction of drainage was carried out in Chapter 7. Chapter 8 examines the consequences of changed hydraulic properties due to grazing on the soil water balance through use of a water balance model, SWIM.

Literature Review - Table of contents

Figures

Tables:

2.0 Principles and measurement of soil water content, potential and movement in the soil

2.1 Soil water content

2.1.1 Principles

The amount of water in a soil directly affects plant growth. It also influences soil aeration, temperature and soil strength.

The amount of water held in a given volume of soil depends on the amount of pore space (porosity) and pore size distribution. Water may occur in pores within and between structural units (peds). Pores of different sizes have different functions as shown on Table 2.1. Macropores are those pores greater than 30 μm. De Leenheer (1977) divided the macropores into aeration pores those >300 μm and transmission pores those between 300 and 30 μm. When the pore size is reduced to 30 μm hydraulic conductivity is low, drainage becomes very slow and water is stored in the soil. When pore size is less than 0.02 μm the water is held strongly by the fine soil particles and is no longer available for plant uptake.

Table 2.1: Pores size groups their functions and equivalent matric potential (De Leenheer, 1977, Craze and Hamilton, 1991)

Pore size diameter (μm)	Function	Matric potential required to drain these pore sizes (kPa)
> 300 (Macropores)	Aeration	> -1
300 - 30 (Macropores)	Rapid water transmission	-1 to -10
< 30 (Micropores)	Slow water transmission	< -10
0.02 - 30	Storage of plant available water	-10 to -1500
< 0.02	Residual water (unavailable)	< -1500

The soil matrix holds water in two ways: adsorbed onto the surfaces of soil particles and held as capillary water in the micro and macropores (Figure 2.1). Colloidal particles such as clay or humus have a net negative charge so that polar molecules of water are adsorbed onto the surfaces forming a film of water around the particles. Clay particles have a large surface area, so that water adsorption is the predominant force in clay soils. When a soil is saturated, the pores are full of water. The smaller the diameter of the soil pore the stronger the capillary force. As the soil dries water is initially lost from the macropores in which the capillary forces are relatively weak. As a soil dries the film decreases in size and water is held more tightly.

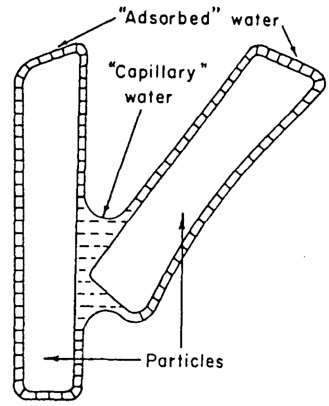


Figure 2.1: Two ways in which water is held in soil (Hillel, 1980)

2.1.2 Measurement

There are several field and laboratory methods for measuring water content. Soil water content may be expressed as mass of water per unit mass of soil (gravimetric moisture) or as a volume of water per unit volume of soil (volumetric moisture). Several authors give detailed descriptions of soil water measurement including Gardner (1986), Gardner *et al.* (1991), and Topp (1993). The more commonly used methods used are outlined below.

2.1.2.1 Gravimetric method

Gravimetric water content (θ_m) is measured by weighing a moist soil sample, then oven drying it at 105°C, reweighing to obtain its dry weight. Water content is then calculated from the following formula:

$$\theta_m = M_w/M_s \quad (kg_{water} / kg_{soil}) \qquad [2.1]$$

where M_w is the mass of water present in the soil sample and M_s is the mass of dried soil. This method is simple and does not require much equipment. However the disadvantage of this method is destructive sampling, which causes damage to the sampling site. Also, the measurements are not made *in situ* so that error may be introduced during transport and handling.

Volumetric water content (θ_v) can be calculated from gravimetric moisture when the bulk density of the soil is known:

$$(\theta_v) = \theta_m * (\rho_b/\rho_w) \quad (m^3{}_{water} / m^3{}_{soil}) \qquad [2.2]$$

where, ρ_b is bulk density and ρ_w is the density of water.

2.1.2.2 Neutron Method

The neutron method of measuring soil water content is a non-destructive field method based on the slowing-down by water of fast neutrons emitted by a radioactive source (Greacen, 1981). A neutron water meter contains both a radioactive source and a slow neutron detector which is lowered into an aluminium tube, known as an access tube, installed vertically into the soil (Figure 2.2). Fast neutrons are emitted from the radioactive source which are slowed (or thermalised) by hydrogen. Given that most of the hydrogen in the soil is water, the number of thermalised neutrons (count rate) is proportional to the soil water content. The count rate is converted to water content by use of a soil-specific calibration (Gardner *et al.*, 1991).

The volume of soil sampled by the detector depends on the soil water content. The 'sphere of measurement' is the area around the source which has the greatest influence on the count rate (Figure 2.2). The radius of the sphere varies from about 0.15 m for a saturated soil to about 0.70 m for a dry soil (Topp, 1993).

The neutron method is widely accepted both theoretically and practically. Once access tubes are installed, measurements can be made periodically and an accurate measurement of soil water content over time can be obtained. The main disadvantages are the cost of equipment, the difficulty in satisfactorily installing the access tubes in some soil types and the danger associated with the radioactive source. Also, the neutron water meter does not accurately measure water content at the soil surface.

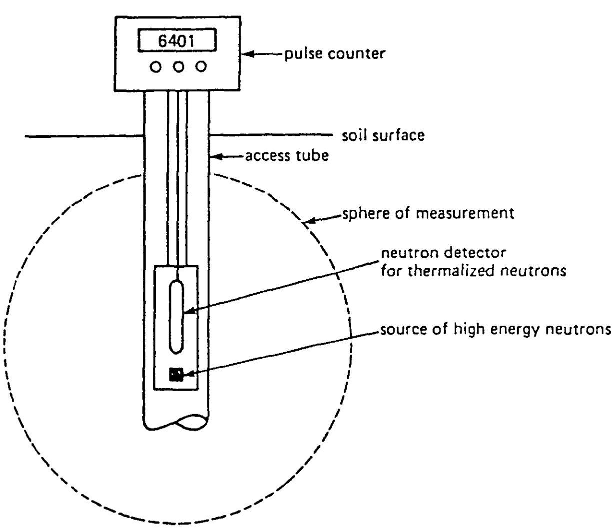


Figure 2.2: Neutron water meter (Topp, 1993)

2.1.2.3 Time-Domain Reflectometry

Time-domain reflectometry (TDR) makes use of the electrical properties of the water molecule to determine the soil water content (Topp, 1993). It measures the transit time of an electrical signal along metallic probes. This time is related to the dielectric constant of the material surrounding the probe. The dielectric constant of water is around 80 and for most soil types it is between 3 and 5, therefore the dielectric constant can be used as a measure of soil wetness. The TDR is particularly useful for taking water content measurements at the soil surface. TDR does not require calibration for different soil types as does the neutron water meter. There is some difficulty in installing the TDR probes so that there is no air gap between the probes and soil. This is particularly so in stony soils.

2.2 Soil water potential

2.2.1 Principles

Water content does not indicate how much water is available for plant growth unless one knows how tightly the water is being held in the soil. The amount of plant-available water in a soil is dependent on the soil water potential.

Soil water, like any other body in nature, contains energy. Whereas kinetic energy is negligible, potential energy of soil water, resulting from position or internal condition is important in determining the state and movement of soil water (Ward and Robinson, 1990). Soil water potential refers to the energy with which water is retained by the soil and consequently to the energy necessary for water to be removed from the soil by plants.

Soil water potential is defined as the amount of work that must be done to transfer a unit quantity of water from the reference state to the equilibrium soil water system where the reference state is a hypothetical reservoir of pure free water at atmospheric pressure, at the same temperature as soil water and at a given elevation (Hillel, 1982). Soil water potential differs from the potential of pure, free water because soil water has several other forces acting upon it, such as those resulting from the attraction of the soil matrix for water, presence of solutes and the action of external gas pressures and gravitation. Total soil water potential therefore consists of a number of different components as shown in equation 2.3.

$$\psi_t = \psi_m + \psi_s + \psi_p + \psi_z \qquad [2.3]$$

where ψ_t is total water potential, ψ_m is matric potential, ψ_s is solute potential, ψ_p is pressure potential and ψ_z is gravitational potential.

2.2.1.1 Gravitational potential (ψ_z)

Soil water is subject to the force of gravity and the gravitation potential is therefore associated with the position of the soil water in the earth's gravitational field. It is calculated as follows:

$$\psi_z = \rho_W.g.(z\text{-}z_O) \qquad [2.4]$$

where ρ_W.is the density of water, g is the acceleration due to gravity and $(z\text{-}z_O)$ is the height above the reference level z_O. Gravitational potential increases with elevation and in the absence of strong retention forces water will drain downwards from higher to lower elevations.

2.2.1.2 Matric potential (ψ_m)

Matric potential is a negative potential resulting from adsorptive and capillary forces due to the soil matrix. Unsaturated soil spontaneously absorbs free water due to the adsorption of water molecules onto soil particle surfaces and the capillary attraction of soil pores for water (section 2.1.1). Matric potential is an important soil factor determining the availability of water to plants. Differences in matric potential within a soil also provide the driving force for unsaturated water flow once any differences in elevation have been allowed for (Mullins, 1991).

Matric potential is determined by both soil texture and structure and is calculated using the following equation:

$$\psi_m = -2T/r \qquad [2.5]$$

where T is the surface tension of water and r is the largest radius of pores occupied by water.

2.2.1.3 Pressure potential (ψ_p)

Pressure potential is an important component of total water potential below the water table. It is the potential associated with soil water being submerged and is caused by the hydrostatic pressure head. It is a positive potential when a soil is saturated and is zero in an unsaturated soil. It is calculated from:

$$\psi_p = \rho_w . g . d \qquad [2.6]$$

where d is the depth of submergence.

2.2.1.4 Solute potential (ψ_s)

Solute potential arises from the presence of solutes in the soil solution. The solutes attract water molecules creating an osmotic pressure of the solution. If the soil solution has a high salt content the osmotic pressure increases and therefore plants have difficulty in taking up water.

Solute potential only makes a contribution to differences in total potential if there is water movement across a semi-permeable membrane. It does not affect water flow within the soil.

2.2.2 Measurement

Studies that investigate water transport and storage in soils and soil-water-plant relationships require the energy status of the soil water to be known. Total soil water potential is often thought of as the sum of matric and solute potentials and this is a useful index for characterising the energy status of soil water with respect to plant uptake (Hillel, 1982). The sum of the matric and gravitational potentials is called the hydraulic potential and is useful in evaluating the directions and magnitudes of the water-moving forces throughout the soil profile. Methods are available for measuring matric potential and total soil moisture potential, separately or together.

2.2.2.1 Tensiometers

Tensiometers are used for the measurement of *in situ* matric potential, hydraulic potential and hydraulic gradients. Tensiometers consist of a plastic tube fitted with a porous ceramic cup at one end, and a removable air-tight cap at the other as shown in Figure 2.3. When the porous cup is in contact with the soil, water moves into or out of the cup until the potential inside the cup is equal to that of the water in the soil surrounding the cup. As soil water content decreases, the potential of the soil water decreases relative to that of the water in the tensiometer cup. Water will therefore move out of the tensiometer through the pores in the ceramic cup and into the soil until the energy level of the water in the tensiometer is in equilibrium with the soil water.

There are different devices used to measure the pressure (negative) of the water in the tensiometer, including a mercury manometer, pressure transducer and Bourdon-vacuum gauge. The amount of water that must move for a given change in potential (known as the 'gauge sensitivity') differs between the different devices. Mercury manometers and Bourdon-vacuum gauges are much less sensitive than pressure transducers (Mullins, 1991). A mercury manometer gives a more precise pressure measurement than a Bourdon gauge but setting it up is much more rigorous. Where several tensiometers are set up in the same vicinity for hydraulic gradient measurements a mercury manometer is often preferred.

A portable transducer system is described by Marthaler *et al.* (1983). A syringe needle attached to a pressure transducer is inserted through a septum stopper at the top of the tensiometer. The pressure in the air below the septum stopper, which is in equilibrium with the water pressure, is determined by the transducer. Portable pressure transducers are easy to use and cost effective when monitoring large numbers of tensiometers (Cresswell, 1993).

Most tensiometers give accurate readings within the range of 0 to -100 kPa (Cassel and Klute, 1986). The lowest potential depends on the air-entry potential of the ceramic cup. At potentials less than this air enters the porous cup.

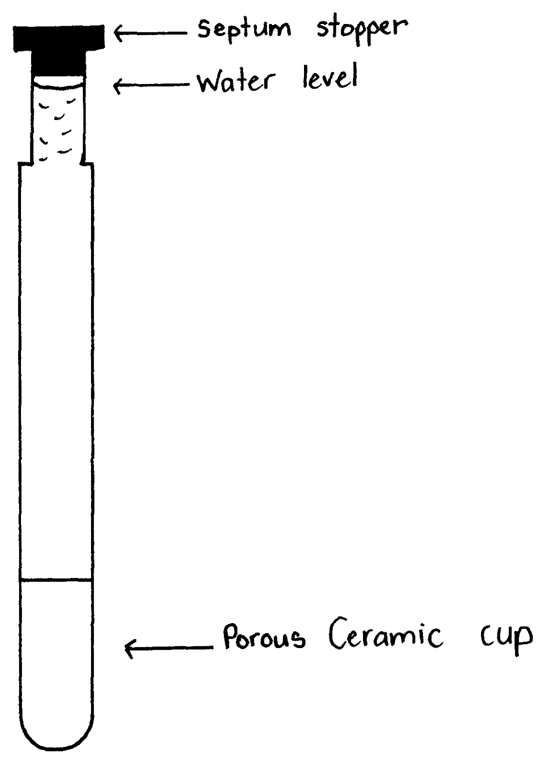


Figure 2.3: A tensiometer

2.2.2.2 Thermocouple Psychrometer

A thermocouple psychrometer measures matric plus solute potential. It provides accurate readings of water potential from around -80 kPa to less than -1500 kPa (Mullins, 1991). Rawlins and Campbell (1986) give a detailed description of psychrometers and their use. Water potential is determined by measurement of relative vapour pressure of air that is in equilibrium with the soil pores (Livingston, 1993). A small ceramic bead is dipped in water, from which water evaporates at a rate controlled by its temperature and relative humidity of the surrounding air. The evaporation causes the temperature of the bead to drop. Eventually a steady rate of evaporation is reached and the bead has a constant temperature difference to its surrounding. This difference in temperature is measured as a voltage which is related to water potential. The main factors that influence the accuracy of psychrometer results are associated with temperature, including (i) temperature gradients, (ii) temperature fluctuations over time and (iii) vapour pressure gradients. All of these factors are described in detail by Rawlins and Campbell (1986).

2.2.2.3 Filter Paper Technique

The filter paper method is a cheap and simple method for measuring matric potential. It involves placing filter paper in contact with a soil sample in a sealed tin, at constant temperature until an equilibrium is reached (5-7 days). The filter paper is then carefully removed, brushing off any adhering soil and the paper is quickly weighed before oven drying to determine gravimetric water content. Gravimetric water content is converted to matric potential using a calibration curve. The calibration curve is derived by measuring water content of the filter papers at defined matric potentials using suction plates, pressure plates or a psychrometer (Mullins, 1991). It is important to weigh the filter paper, both wet and dry very accurately for reliable results.

2.3 Moisture characteristic

2.3.1 Principles

Different soils at the same water content have different water potentials. The relationship between soil water potential and water content is known as the moisture characteristic (Childs, 1940) and is illustrated in Figure 2.4. A soil which is saturated has a matric potential equal to zero as the soil water is in equilibrium with free water. As the soil dries, matric potential decreases and the largest water-filled pores will begin to empty once a certain critical value of potential is reached. This critical value is the air-entry potential at which the largest water filled pores just drain. Further decreases in water potential will correspond to further decreases in water content as smaller pores drain. This is described by the capillary function:

$$\psi m = -2T / r \qquad [2.7]$$

where T is the surface tension of water and r is the radius of capillary tube representing a soil pore of this diameter.

Further decreases in matric potential result in the emptying of smaller pores.

The moisture characteristic can be described by a power function of the form:

$$\psi = a\theta^{-b} \qquad [2.8]$$

that may be written as

$$\ln\psi = \ln a - b\ln\theta \qquad [2.9]$$

where ψ = matric potential

θ = water content

a = intercept

b = slope

Campbell (1985) described the moisture characteristic by a power function of the form:

$$\psi = \psi e\left(\frac{\theta}{\theta s}\right)^{-b} \qquad [2.10]$$

where ψ = matric potential

ψe = air-entry potential

θ = water content

θs = saturated water content

b = slope of the best fit line relating θ to ψ on a log-log scale

which may be written as

$$\ln\psi = \ln\psi e - b\ln\theta + \ln\theta s \qquad [2.11]$$

The model parameters, ψ_e and b are found by regressing ψ on θ/θ_s.

Since most water retention data is collected by applying a defined ψ to a soil then determining the water content, θ may be regressed against ψ, so that equation [2.9] becomes:

$$\ln\theta = A + B\ln\psi \qquad [2.12]$$

which in terms of equation [2.11] is

$$\ln\theta = \ln\theta s + 1/b\ln\psi e - 1/b\ln\psi \qquad [2.13]$$

From equations [2.12] and [2.13]

$$A = \ln(\theta s) + 1/b\ln(\psi e) \qquad [2.14]$$

and

$$B = -1/b \qquad [2.15]$$

Therefore from equation [2.14]

$$\ln(\psi e) = \frac{(A - \ln(\theta s))}{(1/b)} \qquad [2.16]$$

and from equation [2.15]

$$b = -1/B \qquad [2.17]$$

Soil water content as a function of the matric potential may also be described in terms of van Genuchten's (1980) water retention function:

$$\theta = \theta r + \frac{\theta s - \theta r}{(1 + (\alpha\psi)^n)^m} \qquad [2.18]$$

where θ = water content

θs = saturated water content

θr = residual water

ψ = matric potential (cm)

α, n are constants

m = 1-1/n

Equation [2.18] contains four independent parameters θr, θs, α and n that are estimated from observed soil retention data. The residual water is defined as the water content for which the gradient ($d\theta/d\psi$) becomes zero.

Given that each matric potential corresponds with the emptying of pores of a certain diameter, the amount of water held at any given matric potential is therefore affected by soil texture and structure as shown on Figure 2.4.

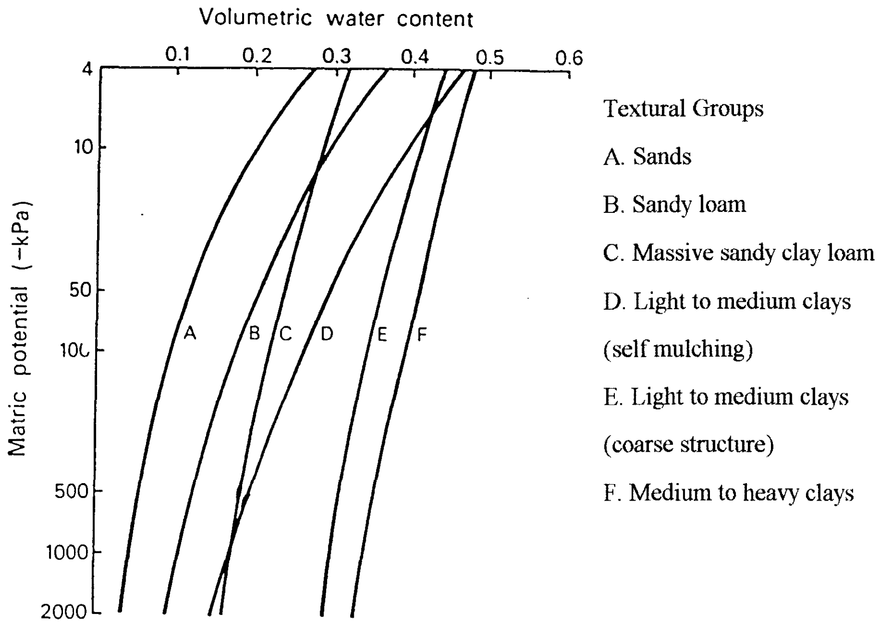


Figure 2.4: The moisture characteristic (Williams, 1983)

Reeve and Carter (1991) proposed that the amount of water held at potentials between 0 and -100 kPa is dependent on the capillary effect and on pore size distribution, and is therefore strongly affected by soil structure. At potentials less than -100 kPa, water retention is due more to adsorption forces and therefore the amount of water held is affected more by soil texture.

Sandy soils have a large number of macropores resulting in the majority of water being released at high potentials. A clay soil will have a higher water content at any given water potential than a sandy soil because of the strong adsorption of water by the clay particles and smaller pore size. A clay soil requires much lower potentials for water to be released. Figure 2.4 shows the moisture characteristic curves for six different textural groups. It is seen that an increase in clay content displaces the moisture characteristic to the right.

Soil structure plays an important role in determining the shape of the moisture characteristic, particularly in the high potential range. Compaction results in a reduction in the number of macropores and the number of intermediate sized pores is increased. From Figure 2.5 it can be seen that the loss of macropores from compaction results in less water being held at high potentials. A decreasing potential (more negative) corresponds to a larger suction. As matric potential increases the compacted soil will retain more water due to an increase in the volume of intermediate sized pores. Micropores are not affected by compaction so the amount of water held at low potentials (high suction) is similar to the non-compacted soil.

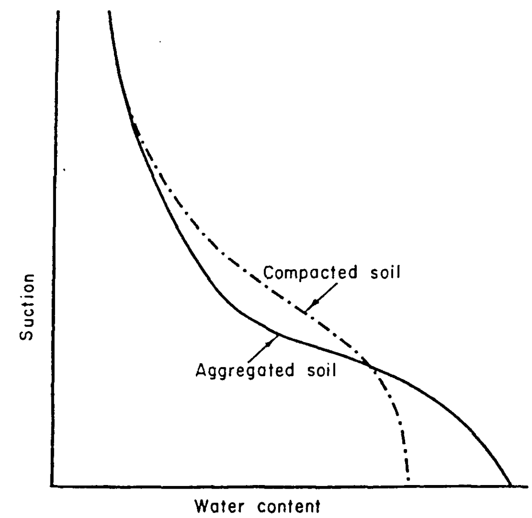


Figure 2.5: The effect of compaction on the soil moisture characteristic (Hillel, 1980)

The moisture characteristic can indicate the ability of a soil to store water for plants. The water retained by the soil at matric potentials between -10 kPa and -1500 kPa gives an approximation of the available water content (Williams, 1983). Table 2.2 shows the differences in available water between different textured soil with good and bad structure.

Although sandy soils have a high proportion of their water content available to plants, water drains readily through the profile resulting in a low water content at field capacity compared to clay soils. Clay soils have a high water holding capacity, but, strong adhesive forces result in a low available water content. The importance of structure in determining the available water content is shown by the poorly structured soils having a substantially decreased available water content. A well structured loam enables a reasonable amount of water to be stored for plant growth and has a sufficient number of macropores to allow the soil to lose water at low potentials, reducing the chance of waterlogging.

Table 2.2: Influence of texture and structure on water retained between -10 and -1500 kPa for a range of Australian soils (Williams, 1983)

	Water retained between -10 kPa and -1500 kPa (mm /dm)		
Field Texture Class	**Structured Soil**	**Structureless soil**	**All soils in group**
	Sands		**13.6**
Sand		14.1	
Fine sands		15.3	
Loamy sands	15.5	12.3	
Clayey sands	14.0		
	Sandy loams		**15.5**
Sandy loams		11.9	
Fine sandy loams	26.5	14.6	
	Loams		15.8
Loam	24.3	13.6	
Sandy clay loam		12.7	
	Clay loams		**16.9**
Clay loam	17.7	18.5	
Silty clay loam	12.7	12.7	
	Light clays		**13.8**
Sandy clays		16.7	
Silty clays		9.2	
Light clays	11.5	14.0	
	Medium to heavy clays		**11.5**
Medium clays	12.2		
Heavy clays	11.5		
	Self-mulching clays		**21.4**
	21.4		

2.3.1.1 Hysteresis

The shape and position of the moisture characteristic is not only dependent on soil properties, but also on the wetting and drying history of the soil (Reeve and Carter, 1991). A soil can have a different water content at the same matric potential due to the phenomena known as hysteresis. As shown in Figure 2.6, at a given matric suction more water is held when the soil is desorbing.

As a wet soil begins to drain or a 'drying' soil is re-wet the relationship between water content and matric potential follows the 'scanning curves' as the relationship shifts between the two main curves. The scanning curves are the intermediate loops in between the sorption and desorption curve, indicating transitions between these two main curves.

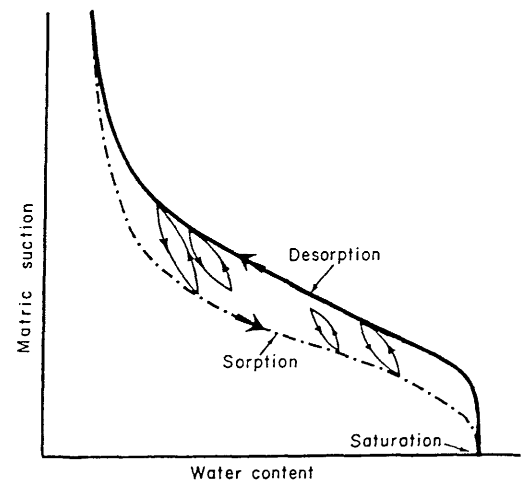


Figure 2.6: Hysteresis in the relationship between water content and suction during the wetting (sorption) and drying (desorption) of a soil (Hillel, 1980)

Hillel (1982) describes the various reasons for hysteresis as follows:

1) Differences in pore shape result in the 'ink-bottle' effect. If a soil pore is irregularly shaped and full of water (Figure 2.7), water will not drain until the matric potential is less than that corresponding to the narrowest radius, r_1 of the pore (ie $\psi_m = -2T/r_1$). Thereafter all the water would drain as the rest of the pore has a larger diameter than r_1. For this pore to re-wet the potential must increase above the potential corresponding to the entry point of narrowest radius r_1. Then for water to move further into the pore a matric potential's corresponding to the increasing pore diameter are required. Once the matric potential exceeds $\psi_m = -2T/r_2$ the pore will abruptly fill. Therefore desorption depends on the narrow radii of the connecting pores, whereas sorption depends on the maximum diameter of the large pores resulting in different wetting and drying paths for the same pore.

2) Contact-angle refers to the angle of contact between the water and solid walls of pores. This tends to be greater in an advancing meniscus than in a receding one.

3) Air trapped in pore space will decrease the water content of a soil that is wetting up.

4) Swelling, shrinking and aging result in changes to soil structure which can affect the moisture characteristic in wetting and drying systems.

The moisture characteristic is usually measured on a drying soil so that the desorption curve represents the moisture characteristic.

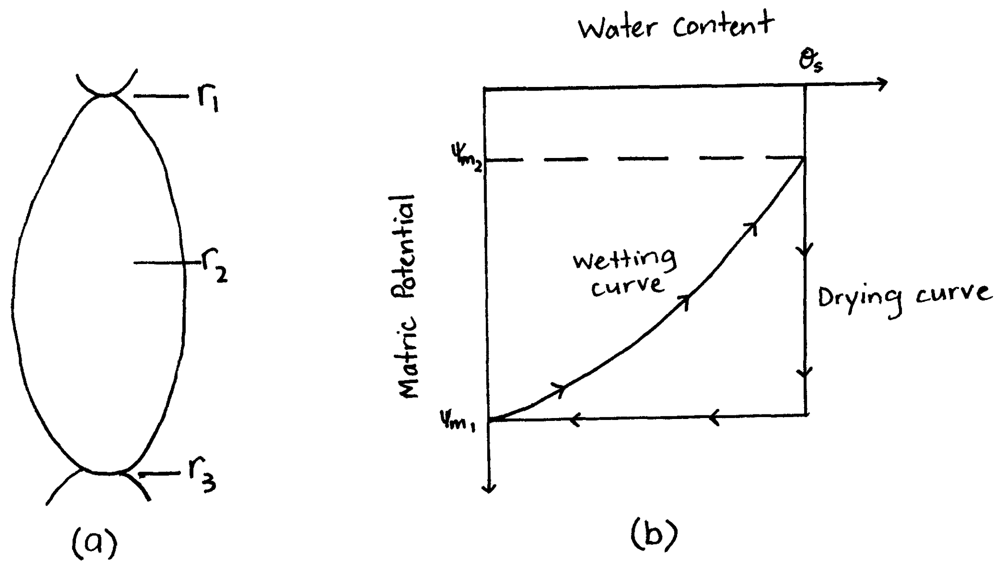


Figure 2.7: A soil pore which is irregularly shaped (a), having a maximum radius, r_2, and is connected to other pores of radius r_1 and r_3. The moisture characteristic for the pore is shown in (b), illustrating the different wetting and drying paths. ψ_{m1} = -2T/r_1, ψ_{m2} = -2T/r_2, and θs = saturated water content (adapted from Hanks, 1992)

2.3.2 Measurement

There are two ways in which to determine the moisture characteristic of a soil. The first is to equilibrate the soil at a chosen range of potentials using suction tables or pressure plates and to determine moisture content at each potential. The second way is to measure moisture potential and content at a number of times as the soil sample dries down.

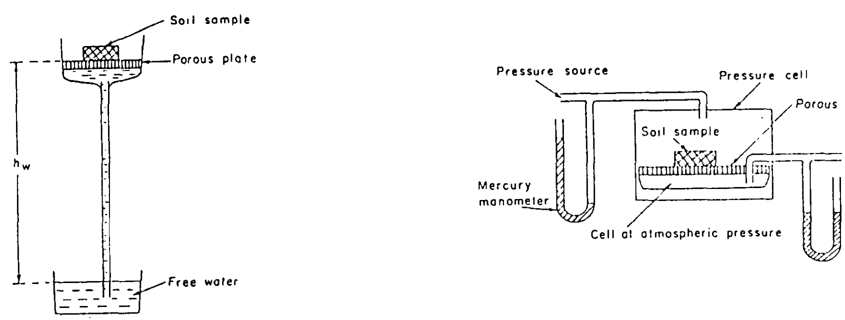
2.3.2.1 Suction (tension) plate

As the soil moisture characteristic is affected by structure, particularly at high potentials (0 to -100 kPa), measurement must be made on undisturbed soil cores. The soil core is wet up then placed onto a porous suction plate. A known suction is applied to the core by a hanging water column as shown in Figure 2.8a. Once the water has drained from the sample and the sample has reached equilibrium it is weighed and returned to the suction plate. The matric potential is adjusted to a lower value and the procedure repeated for a range of potentials. Finally the soil cores are dried in an oven to obtain a dry weight in order to calculate water content at each potential. The maximum suction value obtainable by porous suction plates is often limited by the height available below the plate for a hanging water column. For potentials lower than -10 kPa, a complex sequence of bubbling towers or a vacuum system is required (Reeve and Carter, 1991).

2.3.2.2 Pressure plate

The pressure plate method is similar to the suction plate method in that a soil core is placed onto a porous ceramic plate, but the plates are placed inside pressure chambers and a positive pressure is applied to the soil pushing water out of the soil sample (Figure 2.8b). The maximum suction value obtainable by porous pressure plates depends on the pressures that the chambers can withstand and the maximum pressure the porous plates can bear before allowing air into its pores. Pressure plates are usually used for a range of water potentials between -5 and -1500 kPa.

A disadvantage of using these desorption methods is the time required for samples to equilibrate. At potentials greater than -30 kPa samples will equilibrate in less than a week. However, at potentials between -100 and -300 kPa at least 30 days is required and over six weeks at potentials less than -1000 kPa.



a) **b)**

Figure 2.8: Two methods for measuring the moisture characteristic a) suction plate, b) pressure plate (Hillel, 1980)

***2.3.2.3 In situ* moisture characteristic**

The moisture characteristic may be measured directly in the field by taking simultaneous measurements of matric potential using tensiometers and moisture content determined gravimetrically or by a neutron probe. Tensiometers should be placed around a neutron probe access tube with measurements being taken as the soil dries. Tensiometers are unable to measure matric potential at potentials less than -100 kPa, therefore other methods would need to be used at lower potentials.

The filter paper technique described in Section 2.2.2.3, can also be used to measure the moisture characteristic of a soil *in situ*. To obtain the moisture characteristic soil samples dried to a range of moisture contents are required. These can be obtained by successive sampling of field soils as they dry out (Reeve and Carter, 1991).

2.4 Soil water movement

2.4.1 Principles

Water flows through soil pores and also in the water films that surround the soil particles. When a soil is saturated all the soil pores are filled with water and all are conducting. Pore continuity is at a maximum so the ability of the soil to conduct water is also at its greatest. Water in a saturated soil is moved by gravity and pressure, hence the driving force is a gradient of positive potential.

As the soil dries, the largest pores become air-filled, obstructing flow of water and reducing the conductivity of the soil because the soil pores are no longer continuous. Water begins to flow through the water films and the smaller sized pores that remain water-filled. Water movement then becomes dominated by matric potential differences (negative potential) arising from differences in water content in the soil.

Flow of soil water is described by Darcy's Law (equation 2.19), which states that the flow of water through the soil is in the direction of, and at a rate proportional to (i) the hydraulic gradient, $\partial\psi_h/\partial x$ which is the driving force acting on the water, and (ii) the hydraulic conductivity, $K(\theta)$, which is a function of water content that indicates the ease with which water is transmitted through a soil. Flow of water can be expressed as:

$$q = -K(\theta).\frac{\partial\psi_h}{\partial x} \qquad [2.19]$$

where q is the flux density of water, i.e. the volume of water flowing per unit time per unit area

$\psi_h/\partial x$ is the hydraulic gradient. ψ_h is hydraulic potential which is the sum of matric and gravitational potential. $\psi h = \psi m + \psi z$ and ∂x is the change in distance.

Equation [2.19] is applicable to one-dimensional steady state flow where the flux remains constant along the conducting system (Hillel, 1980).

Soil water is rarely in steady state due to intermittent rainfall and evapotranspiration which cause the soil water content to fluctuate with time. Soil water flow is then described by combining Darcy's equation with the continuity equation to form Richards' equation. The continuity equation [2.20] is a mathematical expression for the conservation of mass law, which states that water is not lost or destroyed, but that water that flows in is either stored or flows out (Hanks, 1992).

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial x} \qquad [2.20]$$

where θ is the volumetric water content at time t. If equation [2.19] is substituted into equation [2.20] it forms Richards' equation which describes one dimensional water flow in a vertical direction:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} K(\frac{\partial \psi m}{\partial x} + \frac{\partial \psi z}{\partial x}) \qquad [2.21]$$

For horizontal flow there is no gravity gradient influencing flow, therefore $\partial\psi z/\partial x$ would be zero and flow would be described by the following equation:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} K(\frac{\partial \psi m}{\partial x}) \qquad [2.22]$$

2.4.1.1 Hydraulic conductivity

Whether water movement is under saturated or unsaturated conditions, it is dependent on hydraulic conductivity. Hydraulic conductivity (K) is a measure of a soil's ability to transmit water and is defined by Darcy's equation (equation 2.19). The conductivity of a soil is affected by the size and shape of pores, determined by soil texture and structure. The larger the pore, the better the conductor. As a soil dries and the largest pores stop conducting water, the tortuosity of flow paths increase resulting in a decrease in conductivity with decreasing water content (Klute and Dirksen, 1986). Hydraulic conductivity is therefore a function of water content, $K(\theta)$. It is also a function of matric potential, $K(\psi)$, given the relationship between water content and matric potential. These relationships are illustrated in Figures 2.9 and 2.10. Rose *et al.* (1965) calculated K at different depths for a loam soil. The surface soil was a fine sandy loam texture overlying a clay B horizon, with the clay becoming less porous with depth. Figure 2.9 shows the decrease in conductivity at a given moisture content as clay content increases and the decrease in conductivity as moisture content decreases. The slope of the latter relationship is steeper for the lighter textured soils.

Figure 2.10 shows that hydraulic conductivity at a given potential differs greatly between texture classes. K decreases much more rapidly for a coarse textured soil than a finer clay soil as the matric potential decreases, due to the higher percentage of larger pores.

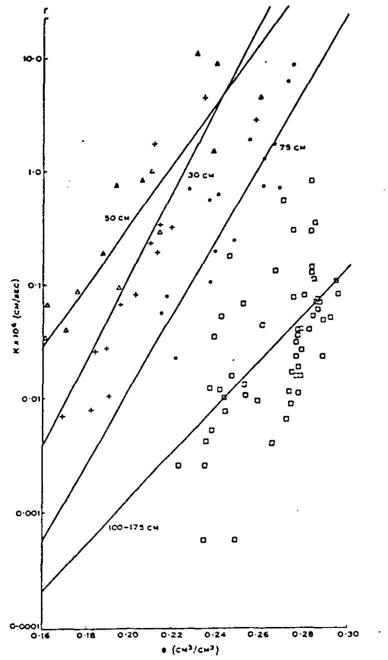


Figure 2.9: Relationship between hydraulic conductivity and volumetric water content at different depths. + 30cm, Δ 50cm, • 75cm, □ 100, 125, 150 and 175cm (Rose *et al.*, 1965)

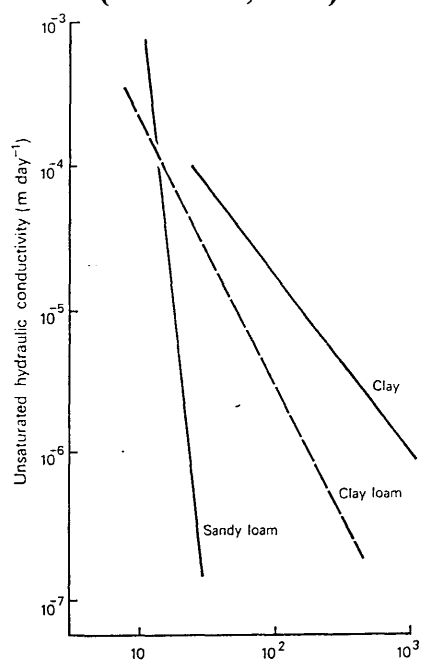


Figure 2.10: The relationship between hydraulic conductivity and matric potential (Williams *et al.*, 1983)

The value of K between saturation and -10 kPa is determined mainly by structure (Williams, 1983). This is illustrated in Figure 2.11, which compares $K(\psi)$ for two well structured and two poorly structured soils of different texture. In each case the well structured soils have a higher K at any given potential.

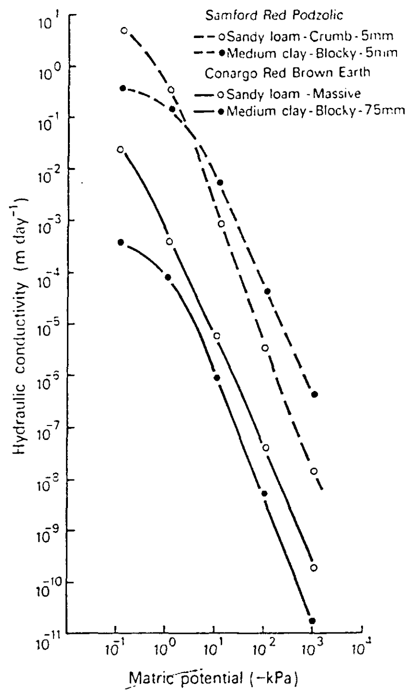


Figure 2.11: The hydraulic conductivity-matric potential relationship as affected by structure (Williams, 1983)

Saturated hydraulic conductivity, K_S, is the conductivity of a saturated soil, i.e. $K_S = K(\theta_S)$. It is constant for a given soil, but varies with texture and structure. Saturated hydraulic conductivity is much greater in coarse textured than in fine textured soil due to the greater number of larger sized pores. Structure also affects K_S due to reduced porosity in poorly structured soils.

McKeague *et al.* (1982) investigated the possibility of making estimates of K_S from observation of soil morphology. They found that macroporosity and structure strongly affected K_S. High K_S values were found in soils with a texture coarser than fine sandy loam, and also in clayey soils which had a lot of biopores or a strong blocky structure. Massive, compressed soils and clay soils with few macropores had low K_S values. Most of these soils were collected from plough-pans and paddocks which had been cultivated when the soil was too wet. McKeague *et al.* (1982) concluded that K_S was not closely related to texture, and that macroporosity and structure were the main factors controlling K_S in many soils. Tillage practice and current land-use has a major effect on soil structure, porosity and therefore K_S.

Marshall (1959) grouped K_S values into different classes, which illustrates the large differences in K_S between different soil types. For a well structured Krasnozem or a coarse sand K_S was 10-50 m per day, whereas for the B horizon of a red-brown earth or a cracking grey clay K_S may be as low as 10^{-4} m per day.

2.4.2 Measurement

2.4.2.1 Saturated hydraulic conductivity (K_S)

The are several laboratory and field methods developed for measuring saturated hydraulic conductivity. Laboratory techniques are based on Darcy's equation. Undisturbed soil cores are taken from the field. They are placed onto a permeable base, saturated and water is ponded on the surface so that water moves through the column of soil under a hydraulic head and the rate of flow of water is measured. The hydraulic conductivity is equal to the rate of flow per unit cross-sectional area per unit hydraulic gradient. There are several methods based on these general principles, including the constant head and the falling head method which are described in detail by Klute and Dirksen (1986).

Like the laboratory methods, field methods for determining saturated hydraulic conductivity are also calculated from Darcy's law after measuring soil water flux and hydraulic gradient. Methods for measuring K_S vary according to whether the soil has a shallow or deep water table. A commonly used field method for K_S measurements at sites below the water table is the auger-hole method (Amoozegar and Warrick, 1986).

The methods used to determine K_S above the water table are often more complex and time consuming than those used below the water table (Amoozegar and Warrick, 1986). These techniques include the constant head well permeameter (Reynolds and Elrick, 1985), also known as the Guelph permeameter, and the ring infiltrometer methods (Reynolds and Elrick, 1990).

Given the sensitivity of K_S to soil structure, when taking undisturbed cores for laboratory methods great care must be taken to avoid changing structure during sampling. If there is some difficulty in obtaining undisturbed soil cores then *in situ* methods would be preferable. The main disadvantage of measuring K_S in the field is the spatial variability of soil hydraulic conductivity (Nielsen *et al.*, 1973). The presence of macropores such as pores formed by soil fauna or plant roots and cracks and fissures will result in rapid water movement, much faster than the K_S of the soil surrounding theses pores, leading to unrealistic K_S values in the field. Care must also be taken with field methods which require augered test-holes. A moist soil may develop smeared or compacted walls during augering, which result in underestimates of K_S (Talsma, 1987; Reynolds, 1993). Spatial variability is also a problem with undisturbed cores, and many samples may be needed to obtain a true estimate.

Several studies have compared K_S values measured using different techniques and have found large differences in K_S measured by different methods (Gupta *et al.*, 1993, Paige and Hillel, 1993 and Mohanty *et al.*, 1994,). The choice of methods often relates to the intended purpose for which the measurements were made, along with the availability of equipment, nature of the soil and skills and knowledge of the operator (Dirksen, 1991).

2.4.2.2 Unsaturated hydraulic conductivity (K)

Darcy's equation is used for calculation of K in unsaturated soils by assuming that K is a function of water content (θ). Therefore, calculation of K requires measurement of water content or matric potential to derive the following relationships, $K(\theta)$ and $K(\psi)$. There are several laboratory methods used to calculate unsaturated hydraulic conductivity (Klute and Dirksen, 1986 and Dirksen, 1991). Conductivity is usually measured by applying a constant hydraulic head difference across a soil core and measuring the steady flux of water. Measurements are made at different matric potentials and water contents. Field techniques include the sprinkling infiltration method, use of disc permeameters and the instantaneous profile method (Green *et al.*, 1986).

The sprinkling infiltration method involves supplying a constant supply of water to the soil, at a rate lower than the hydraulic conductivity. The flux of water entering the soil will reach steady state, at which stage the hydraulic gradient should be close to zero so that the hydraulic conductivity is equal to the flux. The water content and matric potential are measured so as to derive $K(\theta)$ and $K(\psi)$. A limitation of this method is the difficulty of applying water at the low fluxes associated with low water contents and the assumption that water flow is one-dimensional. An advantage is the large sample area relative to soil cores so that more representative results are obtained (White *et al.*, 1992).

Disc permeameters supply water to the soil surface through a disc at a constant supply pressure or under tension. Water flows away from the source in three dimensions and over time the rate of water flow from the source becomes steady. This steady state rate is used to calculate hydraulic conductivity using Wooding's (1968) equation for steady state unconfined infiltration. Ankeny's *et al.* (1991) method for determining unsaturated hydraulic conductivity in the field involves taking infiltration measurements at several different tensions on the same infiltration surface using a modified disc permeameter.

The instantaneous profile method is described by Dirksen (1991) and its practical application by Kablan *et al.* (1989). Advantages of this method are that a large volume of soil is sampled, $K(\psi)$ and $\theta(\psi)$ are measured simultaneously at different depths in the profile, and $K(\psi)$ and $\theta(\psi)$ can generally be measured over a reasonable range of tensions (Reynolds, 1993). This method does, however, require considerable time to obtain a set of results, a lot of water is needed to saturate the soil and it is expensive to set up. The method is described in detail in Chapter 7.

2.4.2.3 Hydraulic conductivity function determined from water retention data

Despite there being many different field and laboratory techniques available for measuring unsaturated hydraulic conductivity, methods have also been developed to estimate the unsaturated hydraulic conductivity function from empirical equations (Clapp and Hornberger, 1978) or from models which predict $K(\theta)$ from soil moisture data (Campbell, 1974, van Genuchten, 1980).

Campbell (1974) combined the empirical equation relating water content to matric potential, (equation [2.10]) with the capillary rise equation to derive a relationship for $K(\theta)$.

$$K(\theta) = K_s\left(\frac{\theta}{\theta s}\right)^m \qquad [2.23]$$

where $K(\theta)$ = hydraulic conductivity function

K_S = saturated hydraulic conductivity

θ = soil water content

θs = saturated soil water content

m = $2b + 3$, b is the best fit line relating θ to ψ on a log-log scale.

Van Genuchten (1980) combined the water retention function with the model proposed by Mualem (1976) that relates the water retention curve to hydraulic conductivity to obtain the following equation:

$$\frac{K(\theta)}{K_s} = (\frac{\theta - \theta r}{\phi - \theta r})^n * (1 - ((\frac{\theta - \theta r}{\phi - \theta r})^{1/m})^m)^2 \qquad [2.24]$$

where θ = soil water content

θr = residual water

ϕ = porosity

ψ = potential (cm)

α, n and m are constants

All parameters needed for both equation [2.23] and [2.24] are available from the moisture characteristic curves, except for saturated hydraulic conductivity.

3.0 The hydrological cycle

The hydrological cycle involves the continuous movement of water between the earth's surface and the atmosphere (Ward and Robinson, 1990). The cycle is illustrated in Figure 3.1. Water vapour condenses in the atmosphere to gives rise to precipitation. The precipitation that reaches the ground surface may infiltrate into the soil, pond on the surface, depending on slope and microrelief, or runoff. The water that infiltrates into the soil has three fates:

1) it is removed from the soil by evaporation off the soil surface or by transpiration from plants (evapotranspiration)

2) some is stored in the soil

3) the excess water drains laterally or downward to the ground water table. The underground water component is eventually removed by upward capillary movement to the soil surface, or to the root zone where it is taken up by plants or lost by ground water seepage and flows into streams and into the ocean.

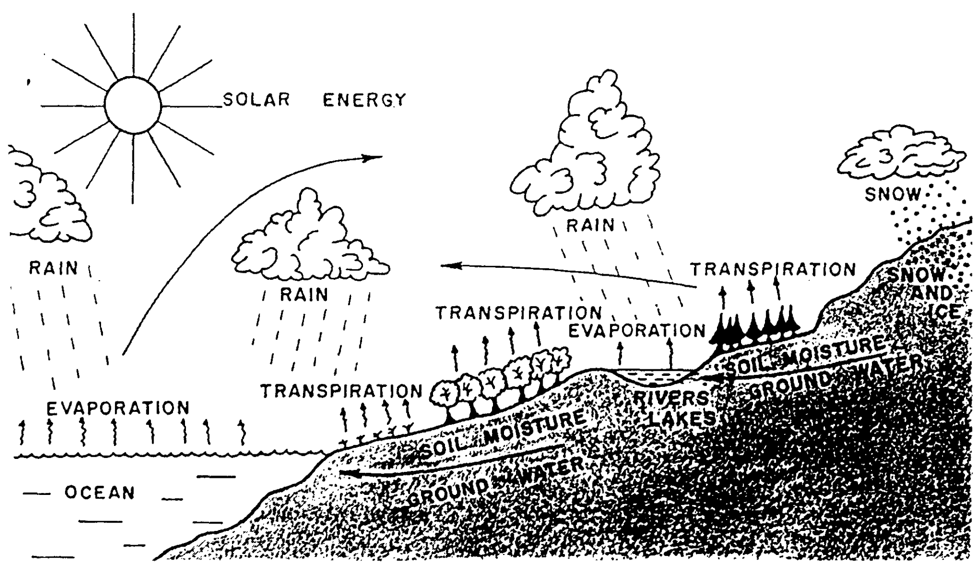


Figure 3.1: A schematic description of the hydrological cycle (Bertrand, 1965)

The hydrological processes which take place in the hydrological cycle can be investigated over a very wide range of spatial and temporal scales. At a macroscale, the hydrological cycle may be examined at a regional or even global level, whereas at a microscale, just the soil hydrological cycle may be explored.

3.1 The soil water balance

From the global hydrological cycle, the processes which take place in the soil are infiltration, water redistribution, drainage to a water table, evaporation from a bare soil and transpiration from plants (Kutilek and Nielsen, 1994). Except for infiltration, all the above processes cause water loss from the soil profile.

The soil hydrological cycle is described by the water balance equation (equation 3.1), which simply states that the difference between the amount of water added ($P+I$) and the amount of water removed ($R+Et+D$) during a certain time period is equal to the change in water content in that volume of soil during that time.

$$\Delta S = (P+I) - (R + Et + D) \qquad [3.1]$$

where: ΔS = change in soil water storage
P = precipitation
I = irrigation
R = runoff
Et = evapotranspiration
D = deep drainage

3.1.1 Infiltration

Infiltration is the entry of water into the soil through the soil surface. It divides the precipitation that reaches the soil surface into two parts. One part replenishes the soil water store that supplies water to plants and the excess recharges ground water. The other part that does not enter the soil is responsible for surface runoff.

Movement of water under non-steady state conditions, into and through the soil, is described by Richards' equation (see Section 2.4.1). This general flow equation is used to describe water movement during infiltration, redistribution and drainage.

Infiltration of water into initially dry soil is controlled mainly by the matric potential gradients and the nature of the surface pores. As the process continues and the wetting front moves deeper below the soil surface, the matric potential gradients decrease and gravity begins to play a larger role. The infiltration rate therefore decreases with time and with the depth of the wetting front. Eventually the rate of flow approaches the saturated hydraulic conductivity once the soil is wetted to depth and the hydraulic gradient is simply gravitational, i.e. the hydraulic gradient, $\partial H/\partial x$, equals one.

Philip (1969) provided an analytic solution to the general flow equation used to describe the infiltration process into a homogenous soil with water ponded on the surface. The solution is in the form of a power series, $t^{1/2}$:

$$I(t) = st^{1/2} + A \qquad [3.2]$$

where I is the cumulative infiltration, s is sorptivity, t is time and A is a transmission factor which approaches K_S as the soil becomes saturated. Sorptivity describes the tendency of soil to absorb water by matric forces in the early stages of infiltration.

The rate of infiltration (i) is described by the following equation:

$$i = 0.5st^{-0.5} + At \qquad [3.3]$$

3.1.2 Runoff

When the rate of water supply exceeds the rate of infiltration, water will pond on the soil surface. The amount of water stored on the surface (surface storage capacity) depends on its topography (microrelief). The soil surface will have different sized depressions depending on the land use, which are capable of holding a certain amount of water. Once all the depressions are full, water will begin to overflow and runoff will occur.

3.1.3 Water redistribution and drainage

Downward movement of water under the influences of gravity and matric forces continues after infiltration at the surface stops. During redistribution, water moves out of the wet upper layers in the soil profile to deeper drier layers. This process controls the quantity of water retained in the plant root zone, the available air-filled porosity for subsequent storage of water and the recharge to the ground water (Ward and Robinson, 1990). Drainage of water out of the root zone is also responsible for the removal of plant nutrients, which are transported to ground water.

Redistribution occurs because the topsoil within the wetted zone after infiltration has a large soil water potential, and the soil below the wetted zone generally has a low water potential resulting in the spontaneous downward movement of water in order to reach an equilibrium. Water redistribution within a soil profile is composed of both drainage and wetting, the water draining from the upper horizons so increasing the water content of deeper horizons.

Redistribution of water slows down with time as the hydraulic conductivity in the former wetted zone is reduced with decreasing water content and also because the potential gradients are less as the moisture content becomes more uniform.

Evaporation and capillary movement of ground water interrupts the process of redistribution. Figure 3.2 illustrates water redistribution in a soil profile under three conditions: 1) redistribution without evaporation, 2) simultaneous evaporation and redistribution and 3) evaporation only. Evaporation dries the top soil; however, the lower parts of the curve indicate that evaporation has little effect on the shape and rate of advance of the wetting front (compare Figure 2.13a and b). Gardner *et al.* (1970) found that evaporation reduced drainage by only about 10 per cent.

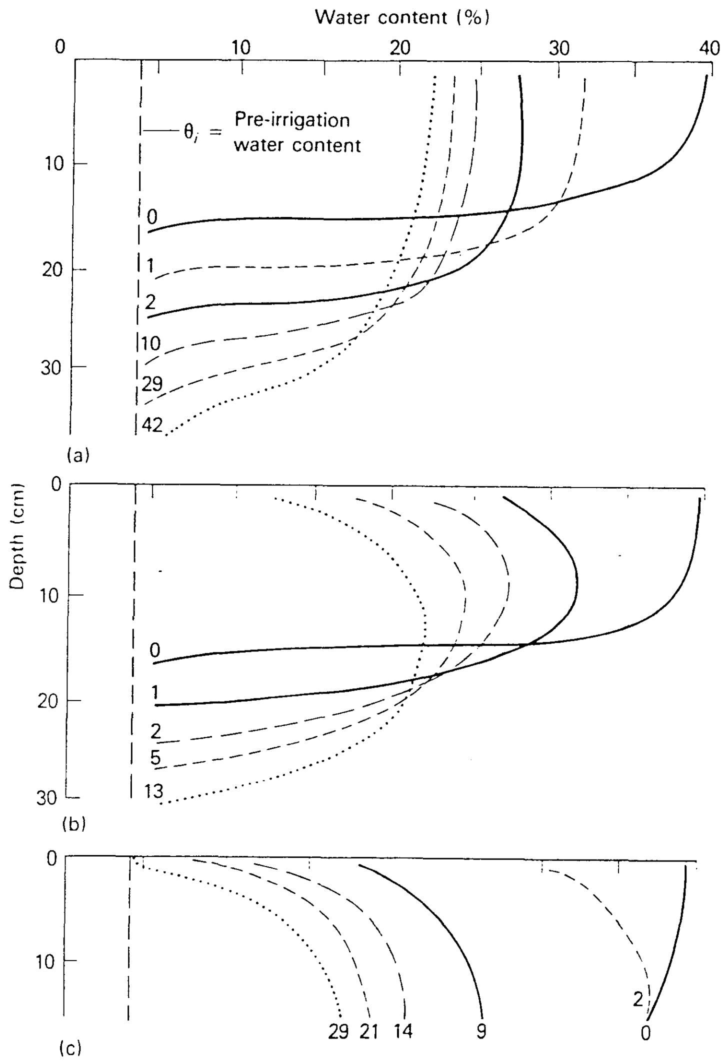


Figure 3.2: Successive soil water profiles in soil columns following an irrigation of 50 mm: a) redistribution with no evaporation, b) redistribution and evaporation, c) evaporation only. The values on each curve indicate the time in days since irrigation ended (Gardner *et al.*, 1970)

Movement of excess water out of the root zone is known as deep percolation (or deep drainage). This water may eventually end up in the ground water table. If a soil has a shallow water table, the wetting front will be close to the ground water, resulting in excess water draining directly from the topsoil to the ground water. When the soil profile is wet throughout, the potential gradients are close to zero and water movement will be due to gravity alone. If this is the case, the downward movement should be equal to the saturated hydraulic conductivity and therefore decrease as water content decreases.

3.1.4 Evapotranspiration

Evapotranspiration is the combined loss of water from the soil surface and by transpiration from plants (Kutilek and Nielsen, 1994). Evaporation from the soil surface may be a continual process where the ground water table is close to the soil surface, resulting in an almost steady flow of water without changing soil water content. Where the water table is not close to the surface, loss of water from the surface will result in an upward flow of water in the profile causing the soil moisture content to decrease. In the absence of a water table, three stages of evaporation exist (Hillel, 1982):

1) Constant rate stage: Early in the process when the soil is wet with a corresponding high conductivity, evaporation will be equal to the atmospheric evaporative demand. The evaporation rate remains constant, as although the hydraulic conductivity decreases at the surface the hydraulic gradient increases enough to compensate for this decrease.

2) Falling rate stage: There is a gradual decrease in evaporation rate with time. The rate of evaporation depends upon the rate of transport of water from deeper parts of the profile to the soil surface. As the soil moisture content decreases with time, hydraulic conductivity continues to decrease and the increase in hydraulic gradients get smaller, thus reducing the amount of water moved to the soil surface.

3) Slow rate stage: At this stage the only water movement to the soil surface is through vapour diffusion.

The rate of evapotranspiration is ultimately controlled by the evaporative demand of the atmosphere, which is a result of temperature, sunlight, humidity and wind speed. Soil characteristics influence water movement in the soil, both movement of water to the soil surface and movement towards plants roots. Transpiration is also affected by plant factors, mainly albedo, stomatal control and root water uptake (Ward and Robinson, 1990).

3.2 The effect of grazing livestock on the soil water balance

The soil water balance is constantly being modified by human activities. Agricultural practices can affect different components of the water balance by changing soil properties. The ideal soil structure for plant growth would require sufficient macroporosity to allow water and air movement into and through the soil, good water holding capacity, little mechanical impedance to root growth and stability of the structure when wetted. Agricultural practices can lead to an improvement in soil physical condition. However, some practices can cause soil degradation, manifest as erosion, salinity, a decline in soil fertility and soil structural decline.

Soil structural decline has a significant influence on the soil water balance. It affects infiltration and hence runoff, water and air movement in the soil, soil strength, aeration and drainage. Soil compaction is an important process leading to soil structural decline.

Compaction is usually associated with frequent trafficking by heavy farm machinery. Animal traffic also causes compaction, especially when animal grazing takes place after rains when the soil is wet and compactable. Although the effect of grazing on soils has been recognised as a problem in the USA, (Alderfer and Robinson, 1947; McCarty and Muzurak, 1976) since the 1930's, the UK (Mullen *et al.*, 1974) and New Zealand (Edmond, 1958; Gradwell, 1968), there has been comparatively little research conducted in Australia.

Studies in Australia have shown that trampling by stock reduces the productivity of pasture. Grazing animals directly affect pasture by reducing ground cover and changing the botanical composition (Witschi and Michalk, 1979; Willatt and Pullar, 1983). Indirectly grazing animals affect pastures by changing soil physical properties (Witschi and Michalk, 1979; Willatt and Pullar, 1983; Kelly, 1985). Proffitt *et al.* (1993) believe that without strategic grazing management, trampling by grazing animals will breakdown soil aggregates and compact the soil surface under pasture.

The effects of grazing on soil and pasture are illustrated in Figure 3.3. Soil compaction by grazing animals will reduce water infiltration, resulting in an increase in runoff and a decrease in the amount of water stored in the soil. High bulk density and soil strength restrict root penetration through the compacted layers, limiting the ability of the root system to supply adequate moisture and nutrients to the plant. Therefore, grazing reduces the efficiency of precipitation usage.

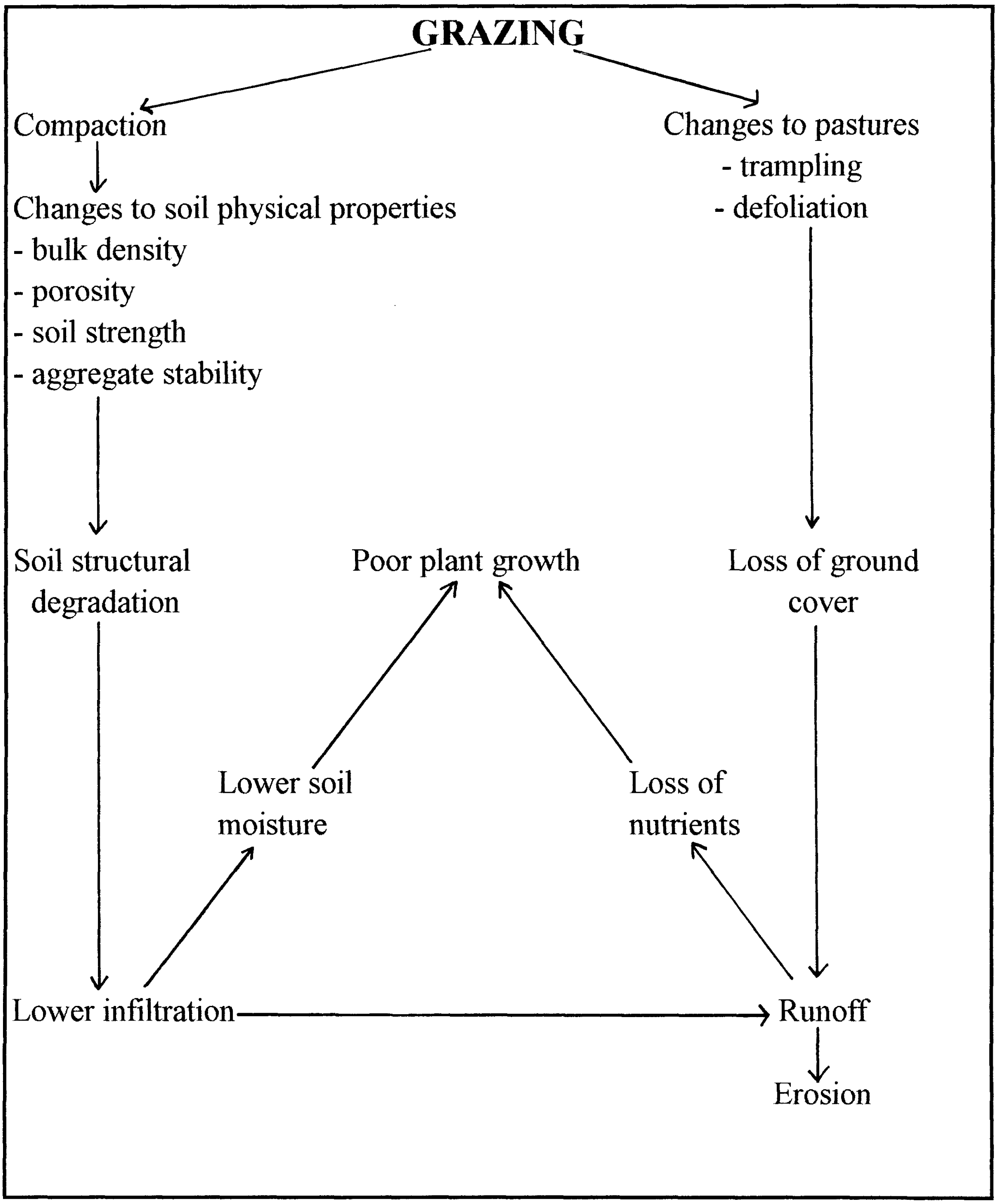


Figure 3.3: The effects of grazing on soil and pasture (Packer, 1988)

3.2.1 The process of compaction

Soil compaction involves the reduction in volume of a given mass of soil when a pressure is applied to the soil (Harris, 1971). The severity of compaction depends on the state of the soil and the magnitude of the load applied (Willatt and Pullar, 1983). The ground pressures exerted by livestock are relatively high and comparable to the pressure exerted by agricultural vehicles. Table 3.1 summarises the potential static loads of various hoofed animals and the ground pressures of agricultural vehicles. The livestock values are likely to be underestimated since they are calculated only on the basis of weight per projected unit area of contact. A walking animal may place their entire weight on one foot at a time, therefore, increasing the static pressure four times.

Table 3.1: Static pressures exerted by livestock and agricultural vehicles (adapted from Packer, 1988; Kirby and Blunden, 1992)

Livestock class	Ground Pressure (kPa)	
Cattle	160-190	
Sheep	63-83	
Goats	60	
Horses	200-390	
Vehicle	**Measured at 10cm depth**	**Estimated at surface**
Toyota Landcruiser	125	250
Tractor (John Deer 4650)	250	180
D8 for chiselling etc		45-90
rig up	160	
rig down	145	
Cat. Challenger 65		
no rig	50	40-75

An animal will tread up to 0.01 ha per day (Packer, 1988). Therefore, treading by stock over a number of years, exerting pressures comparable to agricultural vehicles, is likely to cause problems of soil compaction and decrease pasture productivity. Animals do not graze pastures uniformly due to behavioural characteristics, pasture variation and shelter. Certain areas such as sheep camps, watering points and tracks become more compacted. This variation in trampling increases the variability of soil hydrologic characteristics under grazing regimes, particularly water infiltration (Neath *et al.*, 1991).

3.2.2 The effect of grazing on soil physical properties

Treading by grazing animals affects soil physical properties such as bulk density, porosity, strength and aggregate stability. These changes influence the water balance by affecting infiltration, runoff, ground water recharge and evapotranspiration.

3.2.2.1 Bulk density

Compacted soils have higher bulk densities and a lower proportion of pore space to solids due to a reduction in macropores compared to uncompacted soils. Several studies have found bulk density to increase with grazing compared to ungrazed sites (Alderfer and Robinson, 1947; Gradwell, 1968; Mullen *et al.,* 1974; Witschi and Michalk, 1979;

Willatt and Pullar, 1983). Alderfer and Robinson reported that heavy grazing by cattle increased bulk density in the top 2.5cm of soils that ranged in texture from clay loams to sandy loams, whereas Edmond (1964) and Gradwell (1968) found the depth of consolidation to be around 5-6 cm from the surface in silty loam soils. Packer (1988) concluded that soil texture affected the depth of compaction. Lighter textured soils may compact to a depth of 60 cm whereas in heavier textured soils compaction is confined to the top 10 cm.

Mullen *et al.* (1974) examined the effects of cattle treading (0, 2 and 6.2 animals per ha) on a clay loam soil. The intensively trodden (6.2 animal per ha) plots had a much higher bulk density than lightly trodden or untrodden plots. These results agree with those of Edmond (1958, 1964) and Gradwell (1968) who investigated the effects of treading on a silty loam soil in New Zealand. Witschi and Michalk (1979) found that the bulk density of a clay soil increased from 1350 kg m^{-3} on an ungrazed site to 1880 kg m^{-3} on the heavily grazed site (39.2 DSE/ha). The maximum bulk densities that permit root penetration are in the range 1500-1800 kg m^{-3}, with root penetration severely restricted above a level of 1900 kg m^{-3} (Harte, 1992). Willatt and Pullar (1983) also found significant increases in bulk density in silty loam soils with increasing stocking rates up to 25 DSE/ha.

3.2.2.2 Soil porosity

An increase in bulk density from compaction results in a loss of pore space. Macropores are highly susceptible to compaction resulting in their loss when a soil is compressed.

Several studies have reported a loss of macroporosity from animal treading (Alderfer and Robinson, 1947; Tanner and Mamarill, 1959; Gradwell 1968; Willatt and Pullar 1983). Loss of macroporosity reduces infiltration leading to runoff and erosion risks. Soil aeration also declines and may lead to prolonged periods of deficient aeration after rain or irrigation due to waterlogging.

3.2.2.3 Soil strength

The ability of a soil to support a load is determined by its strength. Soil strength is defined as the maximum stress that can be induced in a soil body without causing the body to fail. As bulk density increases, soil shear strength increases because of an increase in the interlocking of soil particles (Hamblin, 1987).

Bryant *et al.* (1972) found that increasing grazing intensity increased penetrometer resistance. The soil depth at which maximum resistance was encountered was closer to the surface with increasing trampling pressure. Mullen *et al.* (1974) reported significant increases in soil strength with increased stocking rate on a medium textured clay loam soil. Weigel *et al.* (1990) examined the effects of short duration grazing on soil strength. Soil strength was measured before and after each grazing cycle over two years. They found soil strength in the grazed treatment was higher than the ungrazed areas. Strength measurements taken directly under the hoof print were even higher, reaching a maximum of 1.54 MPa after heavy rain during the previous grazing cycle.

3.2.2.4 Aggregate stability

Aggregate stability refers to the ability of a soil to resist disruptive forces. Slaking and dispersion result from unstable soil aggregates. Slaking is the partial breakdown of soil aggregates in water due to the swelling of clay and the expulsion of air from pore spaces. Aggregates are broken down to microaggregates that are as small as $2\mu m$ in diameter resulting in poor infiltration and seedling emergence. If the bonding forces between clay plates become weak once the soil is wetted dispersion of clay particles ($<2\mu m$) will occur.

Aggregate stability is affected by the treading of animals. The animal hoof has a direct impact causing disintegration of soil aggregates by compactive and abrasive forces. Grazing may cause a decrease in pasture production lowering soil organic matter content further. Lemin (1992) found that under grazing, the number of water-stable aggregates of a gleyed podzolic soil declined along with reduced organic matter content. A reduction in ground cover also exposes unstable soil surfaces to the impact of raindrops that may cause soil surface crusting.

3.2.3 The effect of grazing on components of the water balance

The soil water balance is directly affected by grazing animals. Compaction changes surface hydraulic properties resulting in reduced infiltration, increased runoff and lower available water. Indirectly the soil water balance is affected by grazing through changes in the vegetation on the soil surface. The amount of ground cover along with its composition is influenced by grazing. Vegetative cover affects infiltration, runoff and evapotranspiration.

3.2.3.1 Infiltration

Grazing influences infiltration by its effect on soil physical properties, ground cover and pasture composition. Several studies have found infiltration to be reduced with increased stocking (Rauzi and Hanson, 1966; Dadkhah and Gifford, 1980; Willatt and Pullar, 1983; Proffitt *et al.*, 1993).

The amount of water that can infiltrate into a soil is directly related to the number, size and distribution of soil pores and the initial water content. As discussed above, grazing results in a reduction in the number of large pores particularly near the soil surface that play an important role in water entry and rapid water movement through the soil profile.

The importance of plant and mulch cover in reducing puddling, surface sealing and increasing infiltration is mainly due to absorbing raindrop impact (Packer, 1988). Animals will reduce the amount of plant material on the soil surface by defoliation and treading effects. Johnston (1962) showed that infiltration rates increased with increasing amounts of standing vegetation and natural mulch on the soil surface.

Karl Wood *et al.* (1986) examined the effects of both stocking rate and ground cover on infiltration. A high stocked plot was fertilised so as to maintain pasture production at a similar level to the low graze plot. It was found that increased stocking rate did not affect infiltration where pasture production was enough to maintain ground cover.

3.2.3.2 ***Runoff***

Compaction by grazing animals reduces soil porosity at the soil surface, reducing infiltration and resulting in increased runoff. Runoff occurs when rainfall exceeds infiltration rate. Plant cover and microrelief influence the amount of runoff, both of which are altered by grazing.

Several studies have examined the effect of ground cover on runoff. Branson and Owen (1970) found runoff to be much greater on bare soil than where a plant cover was present. Lang and McCaffrey (1984) also found ground cover to affect both the occurrence and amount of runoff on a duplex chocolate soil with a 12 per cent slope. They concluded that a 75 per cent ground cover was necessary to reduce runoff. Costin (1980) found runoff to increase greatly once ground cover fell below 70 per cent on podzolic soils with a 3.7 per cent slope. Grazing influences ground cover through its effect on plant growth.

Grazing also affects runoff through its effect on micro-relief. During a wet season, hoofed animals are likely to increase depressional storage due to soil pugging. However, over a long dry season continual treading may flatten out the surface.

3.2.3.3 ***Soil water***

When macropores are compressed, the pore size diameter is reduced. A compacted soil will therefore have a greater volume of smaller sized pores. Small pores require a much lower potential than macropores for water to drain out of them. A compacted soil will consequently retain more water at low potentials and less at higher potentials compared to a well structured soil (Warkentin, 1971). However, in a compacted soil the increased proportion of small pores reduces the ability for plants to extract water from these smaller pores, thereby reducing the available water content. Proffitt *et al.* (1993) found soil water content was generally higher throughout the soil profile (to a depth of 1.1m) in ungrazed plots compared to grazed plots. They attributed this to higher infiltration in the ungrazed plots.

Although soil moisture is reduced by an increase in evaporation as grazing reduces the amount of litter on the soil surface (Neath *et al.*, 1991), the dry topsoil will favour greater capacity for infiltration at the next rainfall or irrigation event. This effect may, however, be counteracted by decreased macroporosity at the surface.

3.2.3.4 Evapotranspiration

Grazing influences evapotranspiration by reducing plant cover. Although transpiration is reduced with a decrease in plant cover, an increase in the amount of bare ground may increase the amount of evaporation.

3.3 Water balance modelling

The recent severe drought in Eastern Australia has highlighted the importance of water in agricultural systems. Water plays a major role in determining agricultural productivity and there is an increasing need for more efficient use of water resources. Models can be used to increase our understanding of the behaviour of water in the soil-plant-atmosphere system. They provide a means of simulating hydrological process and predicting subsequent outcomes and thus they can be used as a tool in increasing water use efficiency. Much progress has been made towards understanding soil water processes and describing them as analytic mathematical models (Allison *et al.*, 1983).

Mathematical models describing the soil-water regime are classified as empirical, mechanistic, stochastic or deterministic (Kutilek and Nielsen, 1994). Empirical models are basically direct descriptions of observational data. Early attempts to model the soil water balance were based on observed empirical relationships between variables. With improved understanding of the processes underlying the observed data, models are able to integrate these underlying mechanisms. A mechanistic model incorporates mathematically the mechanisms relating the variables (Thornley and Johnson, 1990), contributing to further understanding of the observed data.

When mathematical formulations of hydrologic processes depend on chance or random variables the model is a stochastic model (Allison *et al.*, 1983). For example, when the study area is much larger than that of a pedon, such as a regional catchment, the hydrologic processes will differ over the study area. The hydrological processes are then a function of some random variable.

Deterministic models adhere strictly to a mathematical formulation of each of the hydrological processes (Kutilek and Nielsen, 1994). However, their outcomes are based on the assumption that predictions concerning hydrological processes are only possible if previous events are considered.

3.3.1 Modelling the soil water balance

Williams *et al.* (1991) suggest that soil water balance models can be of two forms: the storage overflow model or the solution of Richards' equation by numerical methods.

Storage overflow models are the simplest models, in which the soil profile is described as a bucket of water into which water flows until it is full. Water then overflows as runoff or drainage into the next soil layer. However, in practice, water does not behave in this way and outflows can occur before the bucket (or horizon) is full (Nott, 1992).

The numerical solution of Richards' equation is becoming the commonly accepted basis for detailed studies of soil water movement (Ross, 1990b). The equation is the mathematical expression of Darcy's flow law combined with the principle of conservation of mass (Ross, 1990b), as described in Section 2.4.1. To solve Richards' equation, both the soil moisture characteristic and the hydraulic conductivity function for each soil horizon or layer must be known. SWIM is an example of a water balance model developed using Richards' equation. The major advantage of these models over storage overflow models is that they include a more accurate description of soil physical processes.

3.3.2 Soil Water Infiltration and Movement model (SWIM)

SWIM is a computer software package developed by CSIRO, Division of Soils, to simulate aspects of the soil water balance. It models soil water infiltration, runoff, surface soil movement, evapotranspiration and deep drainage. SWIM also calculates soil moisture content profiles over the simulation time.

SWIM deals with a one-dimensional vertical profile. The soil is horizontally uniform, but vertically soil properties can change between different layers as illustrated in Figure 3.4.

SWIM uses Campbell's (1974, 1985) equations to define soil hydraulic properties (Section 2.3.1 and 2.4.1). The inputs required to solve Campbell's water retention and hydraulic conductivity function are field saturated water content, air-entry potential, the slope of the best fit line relating water content to matric potential on a log-log scale (b), and saturated hydraulic conductivity.

The SWIM model calculates surface soil detention (the amount of water stored in the depressions before runoff occurs) and its prediction of runoff incorporates this factor. The effects of surface crusting can also be simulated by SWIM.

Actual evaporation is calculated using an equation of Campbell (1985):

$$E = Ep(hs - ha) / (1 - ha) \tag{3.4}$$

where potential evaporation (Ep) is known, hs is the humidity at the soil surface and ha is the atmospheric humidity. This method gives reasonable estimates during the first stage of evaporation, when most of the water is lost, however, is less accurate during the third stage (Campbell, 1985).

3.3.2.1 Assumptions

Ross (1990a) outlines several assumptions inherent in SWIM. Firstly, SWIM assumes a rigid soil matrix, therefore cannot be used with swelling soils. It ignores vapour flow, temperature effects on water movement and hysteresis in the soil moisture characteristic and hydraulic conductivity function.

Given that the model assumes the soil is horizontally uniform and deals with one dimensional vertical flow, lateral water movement is not accounted for. However, lateral water movement could occur between horizons within a catchment that have different hydraulic properties (Cresswell *et al.*, 1992) indicating an obvious weakness in this assumption.

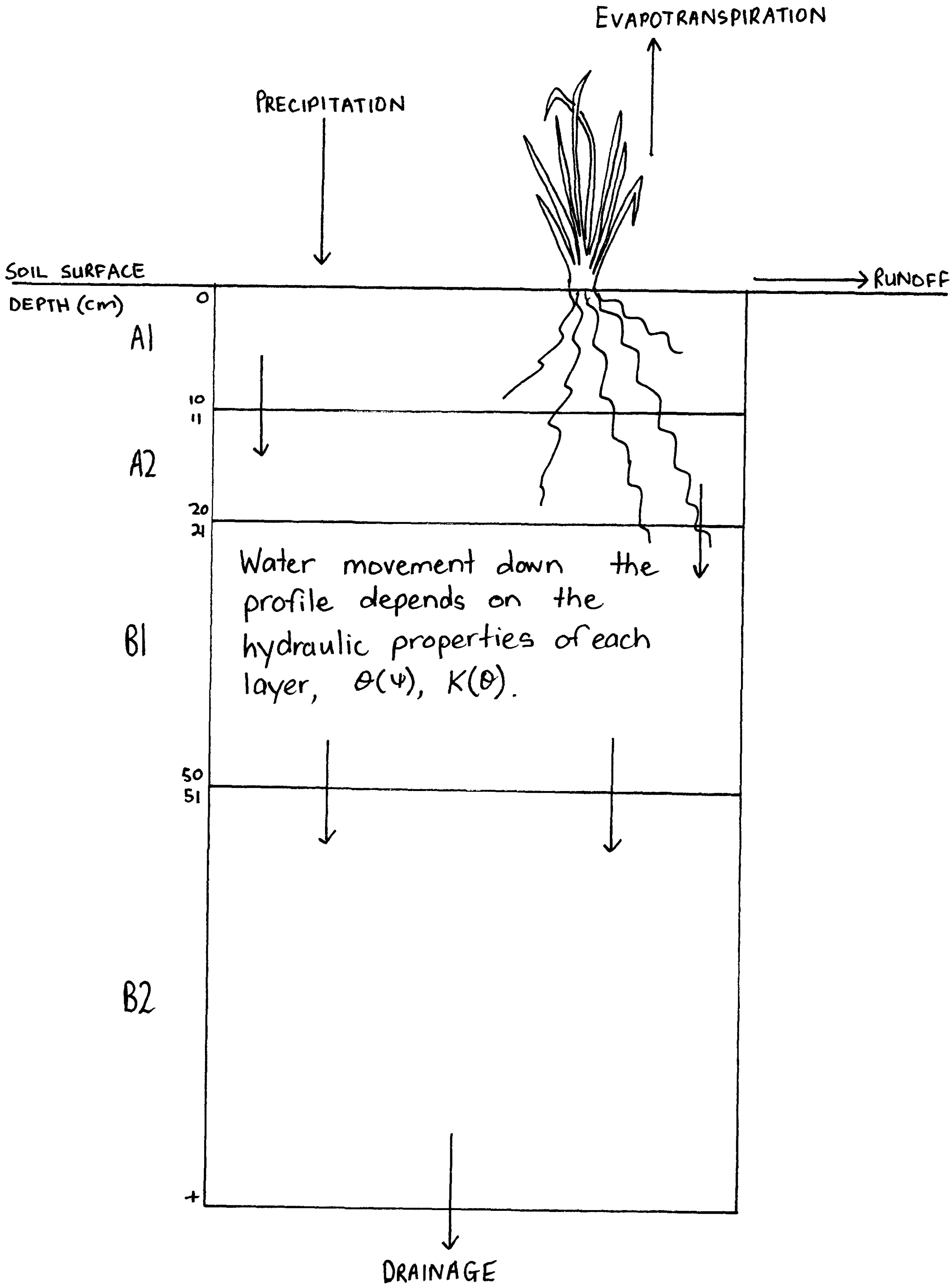


Figure 3.4: A soil profile used in SWIM for calculating the soil water balance

3.3.2.2 Inputs required to run swim

The inputs required for SWIM are divided into seven parameter menus:

1. Simulation times, describe the starting and finishing points of the simulation of the soil water balance.

2. Vegetation. The model allows up to four vegetation types, having certain characteristics that determine its water extraction pattern. The inclusion of vegetation in a simulation is optional.

3. Conductance, describes the flux of water through the very surface of the soil and how it is affected by surface seals.

4. Runoff, occurs when the surface water depth is greater than the surface storage as defined by inputs that relate to surface roughness, conductance and precipitation energy.

5. Soil hydraulic properties. The equations of Campbell (1974, 1985) are used to define saturated water content (θs), field saturated hydraulic conductivity (Ks), matric potential (ψm), air-entry potential (ψe) and *b*.

6. Precipitation data, an optional parameter describing cumulative rainfall over the simulation time.

7. Evaporation data, an optional parameter describing cumulative evaporation over the simulation time.

3.3.2.3 Applications of the SWIM model

SWIM was used by Cresswell *et al.* (1992) to illustrate the effects of changes in soil surface structure on the soil water balance. They investigated changes in structure due to different soil preparation techniques, namely conventional tillage and direct drilling. The effects of plough pans, surface crusts and decreasing surface detention were also examined. Infiltration and water content were measured in the field. Other inputs not measured were estimated using published data or derived using appropriate mathematical equations.

The SWIM simulations indicated that the water balance of a bare soil is significantly affected by major changes in soil structure. Soil water content and runoff did not differ between the direct drill and conventional tillage treatment. There were differences in the amount of evaporation from the bare soil surface from these treatments. A plough pan resulted in increased runoff. For 90 per cent or more of the rainfall to infiltrate, the saturated hydraulic conductivity of the plough pan needed to be greater than 2.5 mm h^{-1}. Reduced hydraulic conductivity due to soil crusting increased runoff. The importance of surface detention was shown, with small increases is surface detention resulting in large decreases in runoff.

Nott (1992) applied SWIM to investigate the effect of vegetation on the soil water balance in a catchment affected by dryland salinity. Six vegetation types were examined: wheat, 1 year old lucerne, 2 year old lucerne, native forest, one year bare fallow and rotation fallow following wheat. Infiltration was the only input measured in the field. All other inputs were estimated from the work of other researchers. SWIM simulations found soil profile water to be higher under short season crops indicating inefficient water-use compared to the perennial systems.

Nott (1992) also examined the effect of land-use history (cereal cropping, perennial lucerne under heavy grazing and native pasture under light grazing) on the water balance. The SWIM simulations showed no major differences between the treatments.

Williams *et al.* (1994) investigated the effect of tree clearing, introduction of different pasture species, and use of native grasses on the soil water balance of a red earth. SWIM was used to analyse the experimental data and provide predictions of evaporation, runoff, deep drainage and profile water storage under different land management.