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4.0 Description of experimental site

4.1 Introduction

A sequence of experiments was carried out to evaluate the effects of grazing on the soil water balance. The surface and subsurface hydraulic properties were determined for two soil profiles, one affected by grazing animals, the other an ungrazed site. In Chapter 5 the effects of grazing on soil hydraulic properties at the soil surface were examined. Differences in water entry through the soil surface, storage and water movement between two grazing treatments were determined. Further hydraulic property measurements were carried out in Chapter 6, including measurement of the subsoil hydraulic properties.

As discussed in Section 3.3, modelling can help to improve one's understanding of water behaviour in the soil. The effects of grazing on soil water can be modelled with the Soil Water Infiltration and Movement (SWIM) model. An evaluation of SWIM's ability to predict subsoil drainage is described in Chapter 7. The drainage component of the water balance plays an important role in pasture growth, with drainage affecting soil aeration and water and nutrient availability to plants.

A number of simulations were carried out using SWIM, in Chapter 8, to compare the effects of changes in rainfall intensity, initial soil potential and depressional storage on the soil water balance of a grazed and ungrazed site.

4.2 Location

The experimental work took place on the CSIRO property, 'Chiswick', situated approximated 15 km south of Armidale on the Northern Tablelands of NSW (Figure 4.1). 'Chiswick' is at an elevation of 1046m, latitude 31°31'S, longitude 150°39'E (Schafer, 1980). The experiments were conducted on the 'Big Ridge 1' site, which is situated in the south-east corner of the property.

4.3 Climate

A summary of meteorological data collected at 'Chiswick' between 1949 and 1976 is presented in Table 4.1. The mean annual rainfall is 866 mm, of which 60 per cent falls between September and February (Figure 4.2). Air temperatures rise in summer to a mean of 24.6°C in January and fall in winter to an average range of -0.6°C to 11.0°C in July, as shown in Figure 4.2. Frosts are common between May and September.

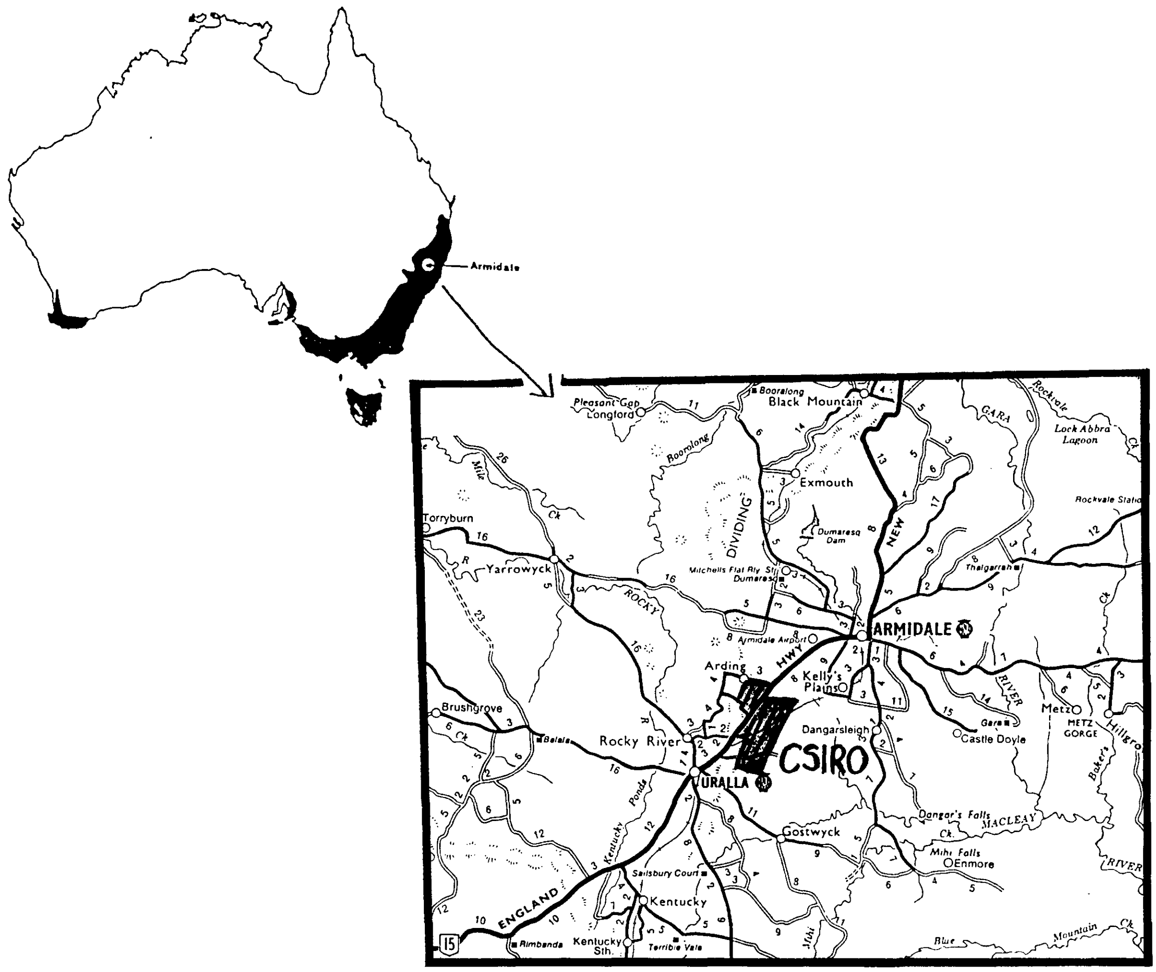
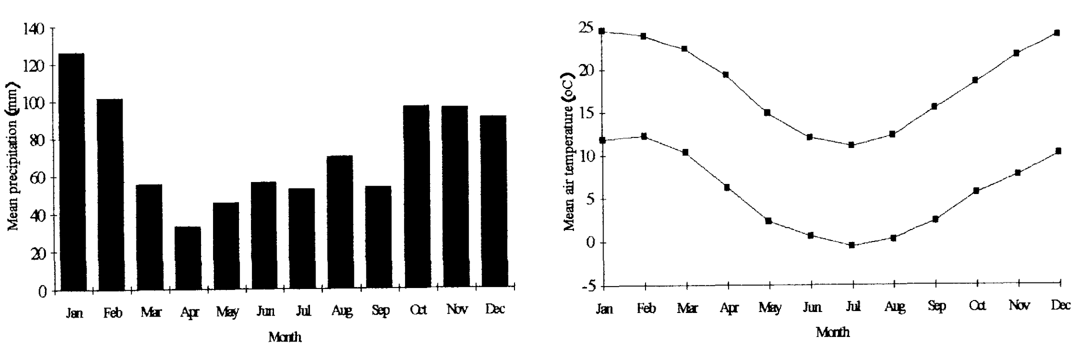


Figure 4.1: Location of 'Chiswick'



(a) (b)

Figure 4.2: (a) Mean monthly rainfall and (b) mean monthly minimum and maximum air temperatures at 'Chiswick' (George *et al.*, 1977)

Table 4.1: Meteorological Data for 'Chiswick', 1949-1976 (George *et al.*, 1977)

Parameter	Lowest value	Highest value	Mean
Sunshine per day (hours)	0.0	13.8	7.4
Maximum air temperature (^{o}C)	0.3	37.7	18.2
Minimum air temperature (^{o}C)	-12.6	23.3	5.7
7.5 cm soil temperature at 9am (^{o}C)	0.3	29.0	13.6
Precipitation per rainy day (mm)	0.1	131.6	8.3
Annual Precipitation (mm)	495.0	1112.0	866.4
Evaporation per day (mm)	-2.8	14.5	3.4
Relative humidity at 9am (%)	2.6	100.0	74.7

4.4 Soils

'Chiswick' covers an area of 1249 ha of undulating country lying within a wide valley. Basalt caps surround the valley, and overlie Tertiary and Palaeozoic sediments, which form the valley floor (Schafer, 1980). These three rock types provide the parent materials from which the soils on 'Chiswick' are derived. A number of soil types are found at 'Chiswick', ranging from black earths, chocolate and prairie soils developed on basalt, to lighter loams, including podzolics and solodics developed on parent material derived from sedimentary rocks. On each parent material a well defined toposequence of soils has been developed (Schafer, 1980).

The three main soil types found at 'Big Ridge' were classified by Schafer (1980) as weisenboden, prairie and gleyed podzolic. The distribution of these soils across the site is shown in Figure 4.3. The experimental plots used in this study are situated on the gleyed podzolic. A soil profile description of this soil is given in Table 4.2. This soil is derived from sedimentary material but is influenced by the basaltic colluvium which has been transported from the upper slopes. It is a duplex soil having a moderately differentiated profile, consisting of a grey-brown loamy A horizon with a bleached A2 horizon overlying a medium to heavy clay B horizon. Mottles are prominent in the B horizon indicating imperfect internal drainage. Particle size analysis details are given in Table 4.3.

4.5 History of the 'Big Ridge' site

'Big Ridge' was set up in 1958 as a long term grazing trial. A mixture of phalaris (*Phalaris aquatica*) and white clover (*Trifolium repens*) was sown into a prepared seedbed and 250 kg of superphosphate per hectare was applied annually. Between 1959 and 1963 it was lightly grazed enabling the pasture species to establish. In 1963 the site was divided into sixteen 0.405 ha plots and stocked with Merino wethers at 10, 20, 30 and 40 Dry Sheep Equivalents (DSE) per ha, with four replications of each stocking rate. From 1963, 125 kg per ha of potassium chloride was applied annually in addition to superphosphate. In 1968 after severe drought conditions it was evident that the 40 DSE per ha treatment could no longer be sustained. The four plots under this treatment were destocked and became ungrazed treatments. In 1978, six plots were subdivided into two plots of equal area and the annual fertiliser applications were discontinued on one side (K. Hutchinson, *pers. comm.*), as shown in Figure 4.3. The stocking rates changed in 1983 to 0, 10, 15 and 20 DSE per ha, each replicated in a Latin square design (Figure 4.3). The latin square design, by having each treatment occurring once in every row and once in each column, takes into account variation due to soil type.

The two plots to be used in this study have a stocking rate of 0 and 10 DSE per ha respectively. They are located on the gleyed podzolic soil and marked plot number 5 and 1 as shown in Figure 4.3. A stocking rate of 10 DSE per ha is a typical commercial stocking rate on the Northern Tablelands. Previous research has also found that the greatest differences in soil properties between stocking rates on Big Ridge 1 were between the 0 and 10 DSE per ha (Lemin, 1992 and Whitbread, 1992).

Table 4.2: Soil profile description - Gleyed Podzolic, Big Ridge 1, 'Chiswick' (Schafer, 1980 and MacKenzie, 1993)

Horizon	Soil Property	
A1	Colour	Light grey (10YR 7/2 dry) to dark greyish brown (10YR 4/2 moist)
(0-11cm)	Texture	Sandy clay loam
	Structure	Weak 5-10 mm polyhedral peds
	Fabric	Earthy
	Consistence	Weak
	Segregations	Few fine ferromanganiferous concretions
	Field pH	6.0
Boundary		Clear and smooth
A2	Colour	Light grey (10YR 7/2 dry) to light brownish grey (10YR 6/2 moist)
(11-31cm)	Texture	Sandy clay loam
	Structure	Massive
	Fabric	Earthy
	Consistence	Firm
	Segregations	Many medium sized ferromanganiferous concretions
	Field pH	6.5
Boundary		Sharp and wavy
B1	Colour	Pale brown (10 YR 6/3 dry) to greyish brown (10YR 5/3 moist)
(31-42 cm)	Texture	Medium clay
	Structure	Strong 50-100 mm prismatic peds
	Fabric	Rough ped
	Consistence	Very strong
	Segregations	Common medium ferromanganiferous nodules
	Field pH	5.8
Boundary		Diffuse and smooth
B21	Colour	Greyish brown (10YR 5/2 dry) to brown (10YR 5/3 moist)
(42-101 cm)	Texture	Medium heavy clay
	Structure	Strong 50-100 mm prismatic peds
	Fabric	Smooth ped
	Consistence	Very strong
	Segregations	Common medium ferromanganiferous nodules
	Field pH	6.0
Boundary		Diffuse and smooth
B22	Colour	Brown (10 YR 5/3 dry and moist)
(101-150+ cm)	Texture	Medium heavy clay
	Structure	Strong 50-100 mm prismatic peds
	Fabric	Smooth ped
	Consistence	Very strong
	Segregations	Common medium ferromanganiferous nodules
	Field pH	6.0

Table 4.3: Particle size analysis (Schafer, 1980)

Depth (cm)	Coarse Sand %	Fine sand %	Silt %	Clay %
0-3	22.4	29.6	28.0	21.4
15-30	29.8	29.4	24.0	19.4
30-51	8.0	15.0	16.0	59.4
76-91	9.2	13.2	10.2	65.4
107-114	10.8	14.1	11.8	62.0

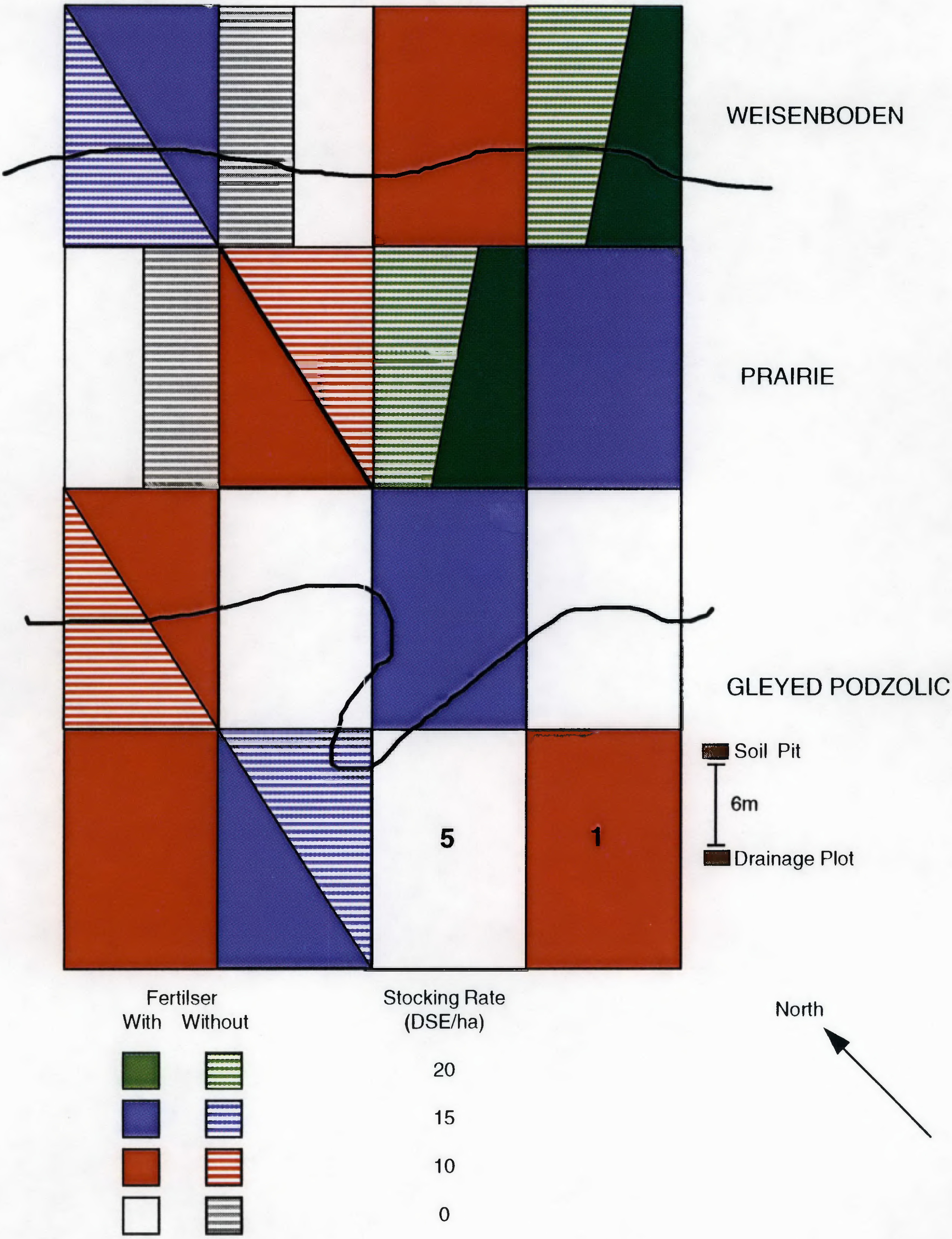


Figure 4.3: Big Ridge 1 plot design

5.0 A comparison of surface hydraulic properties between two grazing treatments

5.1 Introduction

There is increasing concern that trampling by grazing animals breaks down soil aggregates and compacts the soil surface under pasture, thus reducing the productivity of the pasture (Section 3.2). Grazing animals indirectly affect pastures by changing soil physical properties such as bulk density, porosity and aggregate stability (Witschi and Michalk, 1979; Willatt and Pullar, 1983; Kelly, 1985). These properties affect plant growth through their effect on aeration, strength, temperature and most importantly available soil water.

The principles governing water entry, storage and movement can generally be derived from a knowledge of two important soil hydraulic properties, namely the moisture characteristic $\psi(\theta)$, and the hydraulic conductivity function $K(\theta)$ (Greacen and Williams, 1983). In the present study these hydraulic properties were measured to investigate the effect of animal treading on soil surface hydraulic properties. The hydraulic properties were then used as inputs into the Soil Water Infiltration and Movement (SWIM) model to assess changes in the water balance due to grazing (Chapter 8.0).

5.2 Materials and Methods

5.2.1 Soil bulk density and porosity

The bulk density was measured on each of the undisturbed cores taken for the moisture characteristic using the core method outlined by Blake and Hartge (1986). Total porosity was calculated from the bulk density using the following equation:

$$P = 1 - (\rho_b / \rho_d) \qquad [5.1]$$

where, P is total porosity, ρ_b is the bulk density and ρ_d is particle density, assumed to be equal to 2650 kg m^{-3}.

5.2.2 Moisture characteristic

Undisturbed soil cores were taken at three depths, 5-9 cm, 20-24 cm and 30-34 cm, at six sites in each plot using steel cores with an internal diameter of 73 mm and a depth of 40 mm. These depths were chosen as they are, respectively, above, within and below the compacted zone found in the 10 DSE per ha treatment by Lemin (1992).

Sampling sites were chosen by dividing each plot into three strata of equal area and within each stratum two sampling points were chosen using random co-ordinates. Atypical sites, such as sheep camps and the zone within one metre of fence lines, were avoided during sampling. In order to reduce the effects of soil variation, cores were taken directly beneath the surfaces within the profile at which unsaturated hydraulic conductivity had been measured (see following section).

The amount of water held by the cores at matric potentials of -5, -10, -30, -100 and -300 kPa was determined after equilibration on pressure plates (Reeve and Carter, 1991).

Disturbed soil samples were also taken at each depth. They were oven dried at 40°C, sieved through a 2 mm sieve then sampled to determine water retention at matric potentials of -1000 and -1500 kPa using a Decegon SC-10A Thermocouple Sample Changer Psychrometer, as described by Rawlins and Campbell (1986).

The moisture retention data were fitted to Campbell's (1974) power function describing the relationship between water content and matric potential in the form:

$$\psi = \psi e\left(\frac{\theta}{\theta s}\right)^{-b} \qquad [5.2]$$

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

To determine the air-entry potential and the *b* value a linear regression was carried out in the form:

$$\ln\theta = A + B\ln\psi \qquad [5.3]$$

where A = ln θs + 1/*b* lnψe

B = - 1/*b*

Saturated water content (θs) occurs when all pores are filled with water, and was taken to be equal to total porosity (*P*).

5.2.3 Unsaturated hydraulic conductivity

Unsaturated hydraulic conductivity was measured in each plot using a negative head disc permeameter, as described by Perroux and White (1988). Measurements were taken at the same six sites used for determining moisture characteristic. Measurements were taken at three depths: at the soil surface, within the compacted layer (20 cm) and below the compacted layer (30 cm). In order to take measurements below the soil surface, the overlying soil was excavated to the required depth by hand.

Unsaturated hydraulic conductivity was determined by measuring unsaturated infiltration measurements at four tensions: 10 mm, 20 mm, 30 mm and 40 mm. The steady state flow at each tension was then used to calculate unsaturated hydraulic conductivity using the following equations of Ankeny *et al.* (1991), which are based on Wooding's (1968) algebraic description of three-dimensional steady state infiltration into the soil from a circular source:

$$Q(\psi_1) = [\pi r^2 + 4r / A]K(\psi_1) \qquad [5.4]$$

$$Q(\psi_2) = [\pi r^2 + 4r / A]K(\psi_2) \qquad [5.5]$$

where $Q(\psi_1)$ and $Q(\psi_2)$ are the steady state infiltration rates at potentials of ψ_1 and $\psi_2)$
$K(\psi_1)$ and $K(\psi_2)$ are the hydraulic conductivities at these two potentials
r is the radius of the disc permeameter
A is a constant.

Equations 5.4 and 5.5 contain three unknowns. In order to solve for these Ankeny *et al.* (1991) provides another equation developed from the relationship between hydraulic conductivity and matric potential:

$$[K(\psi_1) - K(\psi_2)] / A = \Delta\psi [K(\psi_1) + K(\psi_2)] / 2 \qquad [5.6]$$

Equations [5.4], [5.5] and [5.6] were solved simultaneously for pairs of infiltration rates at different tensions as follows:

$$K(\psi_1) = \frac{Q(\psi_1)}{(\pi r^2 + \frac{4r}{A})} \qquad [5.7]$$

$$K(\psi_2) = \frac{Q(\psi_2)}{(\pi r^2 + \frac{4r}{A})} \qquad [5.8]$$

$$A = \frac{2[Q(\psi_1) - Q(\psi_2)]}{\Delta\psi[Q(\psi_1) + Q(\psi_2))} \qquad [5.9]$$

There is no single value of A for each individual tension, therefore hydraulic conductivity is calculated from two A values. For example, the hydraulic conductivity at -30 mm tension is equal to the average of two K values each calculated with a different A, i.e. A for -20 mm to -30 mm and A for -30 mm to -40 mm tension.

5.2.4 Unsaturated hydraulic conductivity function [K(θ)]

Campbell's (1974) K(θ) function was used to predict the unsaturated hydraulic conductivity function from water retention data:

$$K(\theta) = K_s(\frac{\theta}{\theta s})^m \qquad [5.10]$$

where K(θ) = hydraulic conductivity function

K_S = saturated hydraulic conductivity

θ = soil water content

θs = saturated soil water content

m = 2b + 3, where b is the best fit line relating θ to ψ on a log-log scale (see equation 5.2).

Saturated hydraulic conductivity (K_S) was estimated from K measured at -10 mm tension using the disc permeameter, θs was calculated from bulk density data and b was calculated from the moisture characteristic.

5.2.5 Statistical analysis

Analysis of variance was used to compare the properties of the treatments. Differences between means were assessed by the least significant difference test. A two-way analysis of variance was carried out on moisture characteristic and unsaturated hydraulic conductivity data, to examine the interaction between depth and stocking rate on these two factors.

5.3 Results

5.3.1 Bulk density and porosity

Bulk density increases with depth, but there is no significant differences between treatments at any depth, as shown in Tables 5.1. Likewise there are no significant differences in porosity between the two grazing treatments (Table 5.2). Lemin (1992) did however, find a trend for bulk density to be highest within the compacted zone.

Table 5.1: Mean bulk density at three depths in each grazing treatment

	Bulk Density (kg m^{-3})	
Depth	Zero graze	10 DSE/ha
5 - 9 cm	1371^a	1367^a
20 - 24 cm	1512^a	1544^a
30 - 34 cm	1522^a	1490^a

Data within a row followed with the same letter are not significantly different at P=5%

Table 5.2: Mean total porosity at three depths in each grazing treatment

	Total porosity (m^3 m^{-3})	
Depth	Zero graze	10 DSE/ha
5 - 9 cm	0.483^a	0.484^a
20 - 24 cm	0.430^a	0.417^a
30 - 34 cm	0.426^a	0.438^a

Data within a row followed with the same letter are not significantly different at P=5%

5.3.2 Soil moisture characteristic

Significant differences in the amount of water held in the soil at different potentials were found between stocking rate and soil depth (Table 5.3). Stocking rate has a significant effect on the moisture characteristic. Significant interactions between stocking rate and soil depth were found at the higher potentials of -5, -10 and -30 kPa.

The water retention data were fitted to Campbell's water release curve (equation 5.1). The values required for fitting the data are given in Table 5.4. The high R^2 values indicate a good fit of the data by the Campbell model. The moisture characteristics presented in Figures 5.1 to 5.3 show that the greatest difference between treatments occurs at 30 cm depth. An analysis of variance carried out on the regression lines indicated that there was no significant differences between lines at 5 or 20 cm depth.

However, there was a significant difference in both the intercept and slope of the 30 cm moisture characteristics.

Table 5.3: Effects of stocking rate and soil depth on moisture retained at different matric potentials

	Matric potential (kPa)						
	-5	-10	-30	-100	-300	-1000	-1500
Stocking rate (SR)	*	*	**	***	**	**	**
Depth	***	*				**	***
SR x depth	**	**	**	-	-		

*** P = 0.1%, ** P = 1%, * P = 5% and - P=10%

Table 5.4: Parameters required for calculating Campbell's water release curve: A and B are regression coefficients, R^2 = coefficient of determination , θs = saturated water content, b = the slope of the moisture characteristic, and ψ_e = the air-entry potential

	5 cm		20 cm		30 cm	
Parameter	**zero graze**	**10 DSE/ha**	**zero graze**	**10 DSE/ha**	**zero graze**	**10 DSE/ha**
A	4.61	4.42	4.35	4.21	4.10	4.13
B	-0.24	-0.20	-0.21	-0.17	-0.17	-0.13
R^2	0.93	0.94	0.97	0.94	0.99	0.99
θs (m3 m^{-3})	0.48	0.48	0.43	0.42	0.43	0.44
b	4.22	5.04	4.81	5.76	5.93	7.78
ψ_e (cm water)	-22.02	-15.44	-17.10	-16.20	-7.87	-14.72

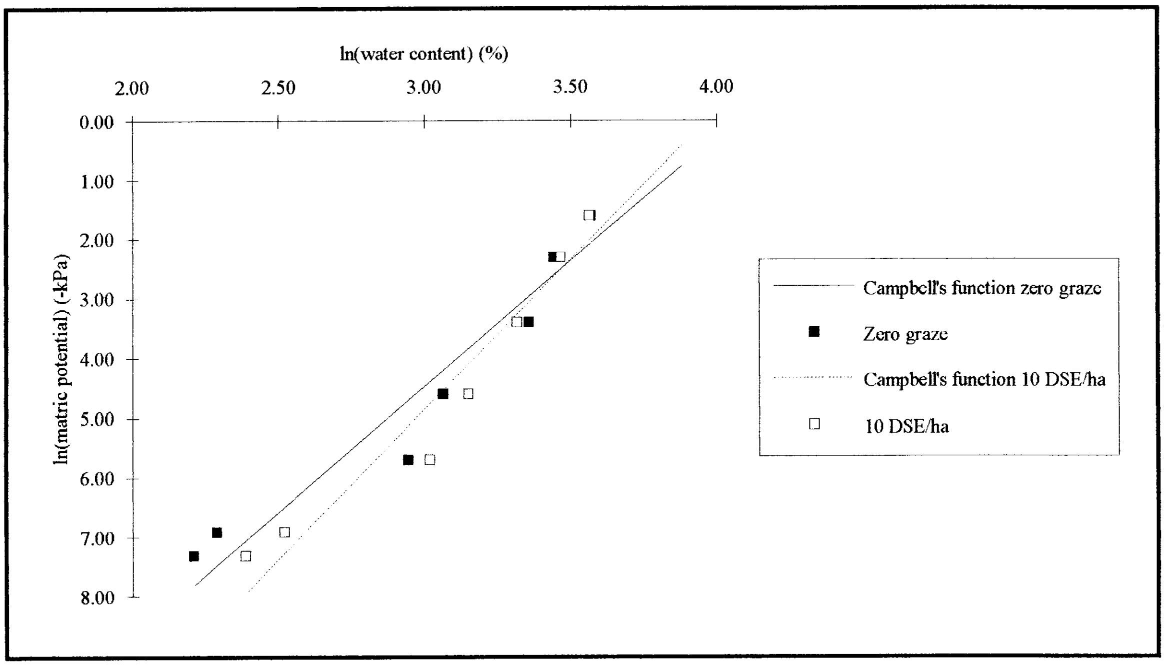


Figure 5.1: Moisture characteristic at 5 cm depth

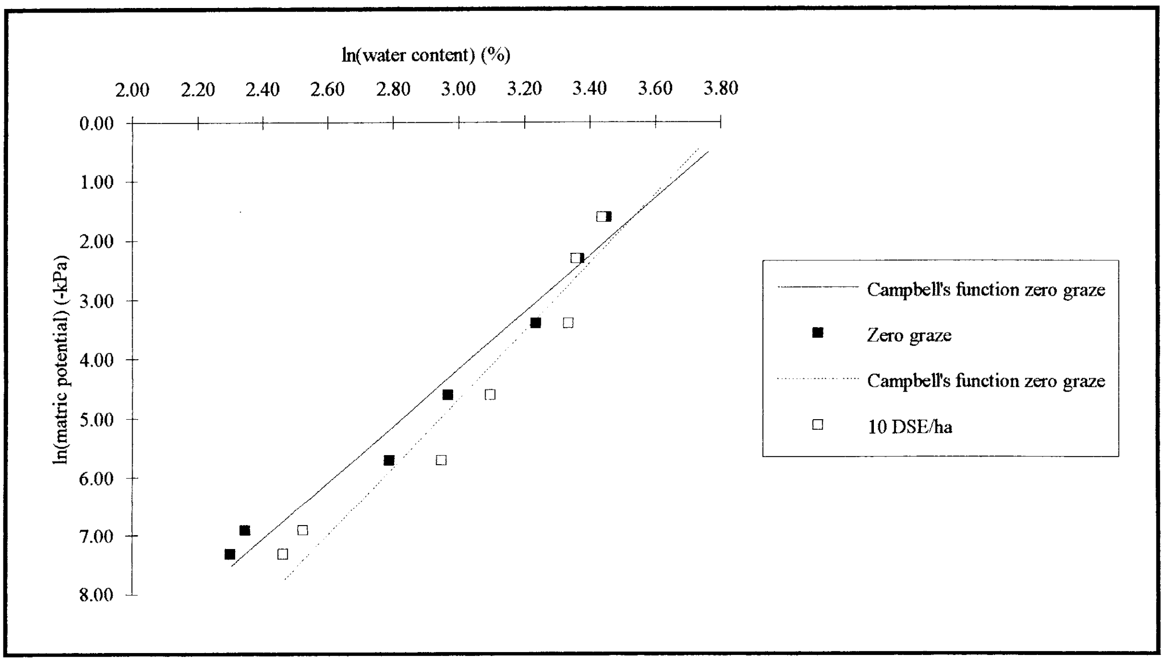


Figure 5.2: Moisture characteristic at 20 cm depth

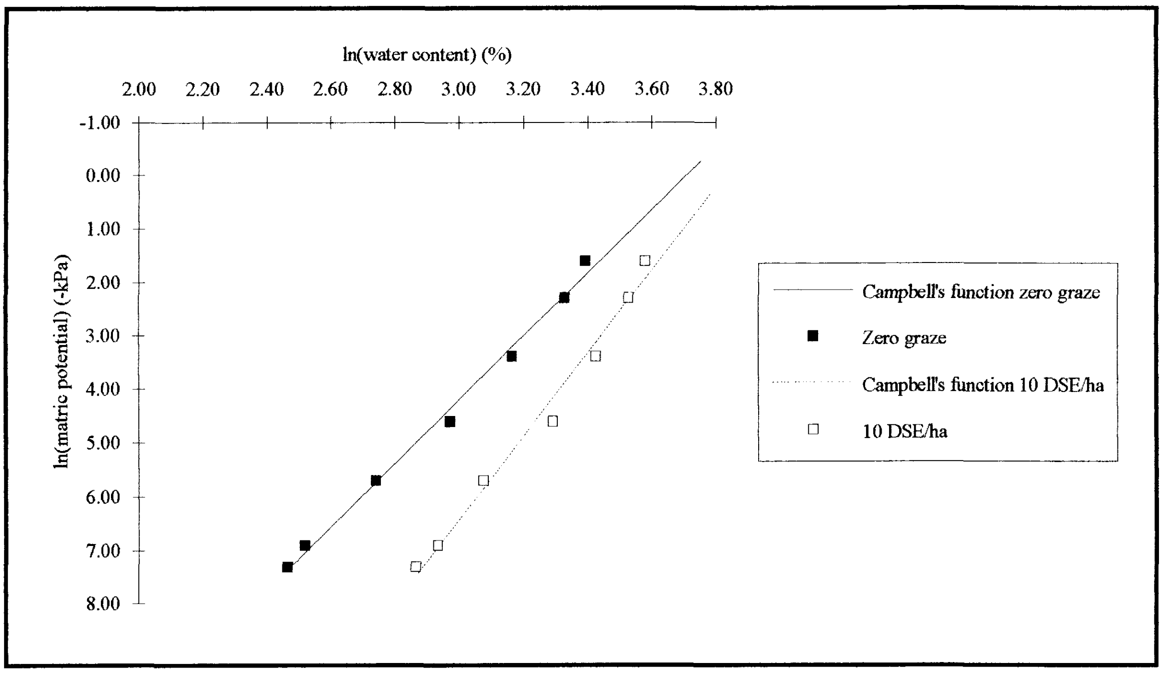


Figure 5.3: Moisture characteristic at 30 cm depth

5.3.3 Unsaturated hydraulic conductivity

Significant differences in unsaturated hydraulic conductivity between stocking rate and depth are shown in Table 5.5. At 10 mm and 20 mm tensions there was a significant interaction between stocking rate and depth.

Table 5.5: Effects of stocking rate and depth on unsaturated hydraulic conductivity

	Tension (mm)			
	-10	-20	-30	-40
Stocking rate (SR)	-		*	
Depth (D)	**	*		-
SR x D	**	*		

*** P = 0.1%, ** P = 1%, * P = 5% and - P=10%

The mean values of unsaturated hydraulic conductivity for each treatment at different depths are presented in Figure 5.4.

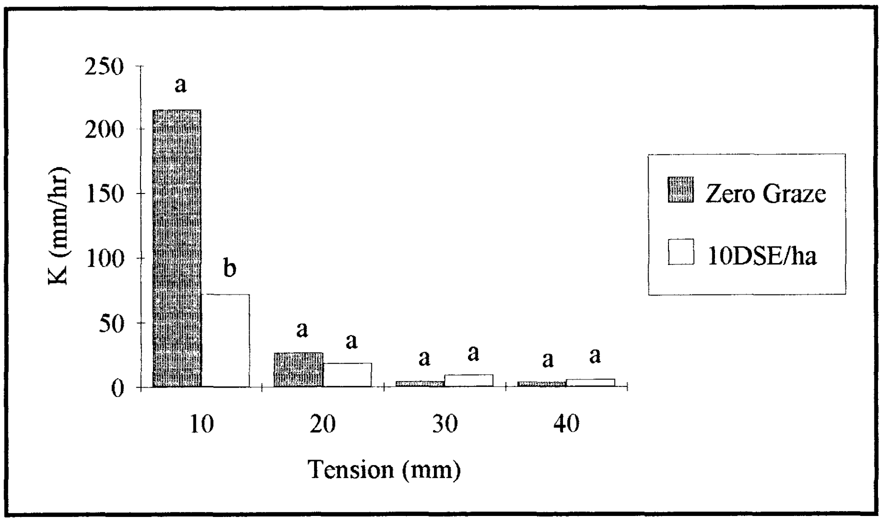
5.3.4 Predicted unsaturated hydraulic conductivity function [K(θ)]

The relationships between hydraulic conductivity and water content at different depths are illustrated in Figure 5.5. The parameters used for calculating the K(θ) function are given in Table 5.6. At the soil surface the hydraulic conductivity at any given water content is greater for the zero graze than the 10 DSE/ha treatment.

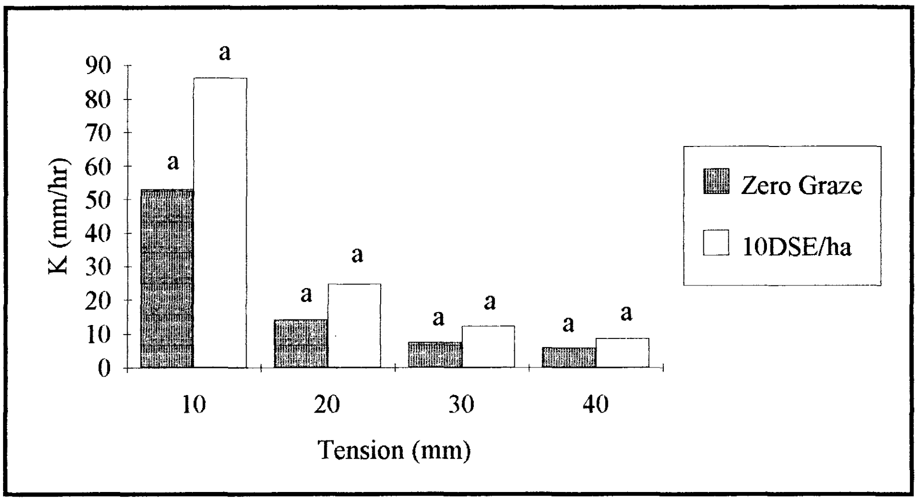
Table 5.6: Parameters required for calculation of K(θ)

	Surface		20 cm		30 cm	
Parameter	**Zero graze**	**10 DSE/ha**	**Zero graze**	**10 DSE/ha**	**Zero graze**	**10 DSE/ha**
Ks (mm/hr)	214.81	71.87	53.22	86.16	36.81	34.56
θs (m3/m3)	0.48	0.48	0.43	0.42	0.43	0.44
m*	11.44	13.08	12.61	14.53	14.85	18.56

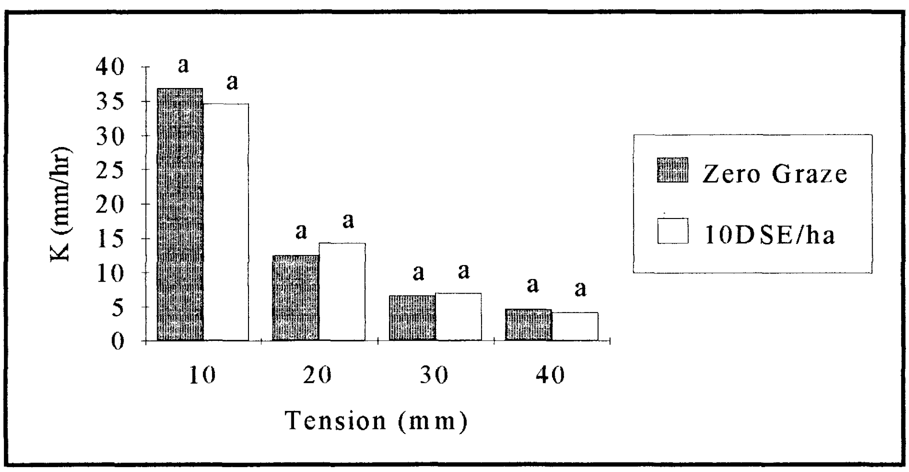
* m = 2*b*+3, where *b* is the slope of the moisture characteristic



(a) soil surface



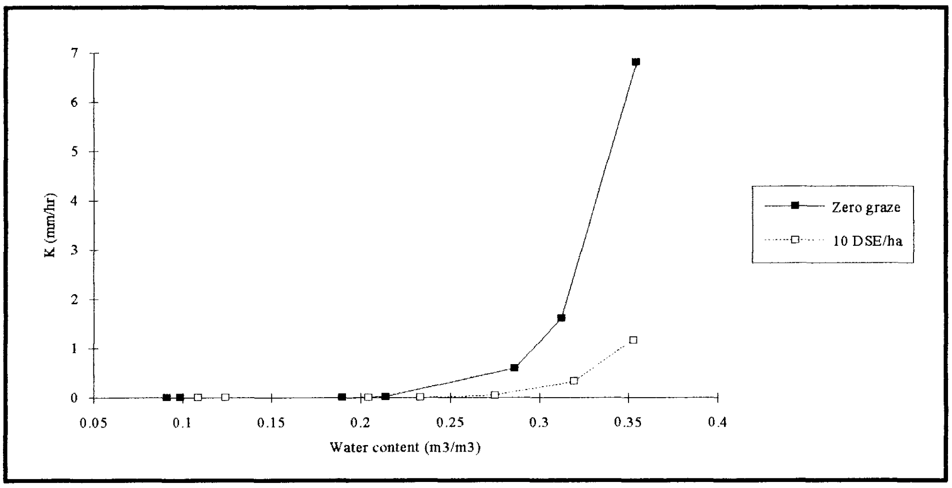
(b) 20 cm



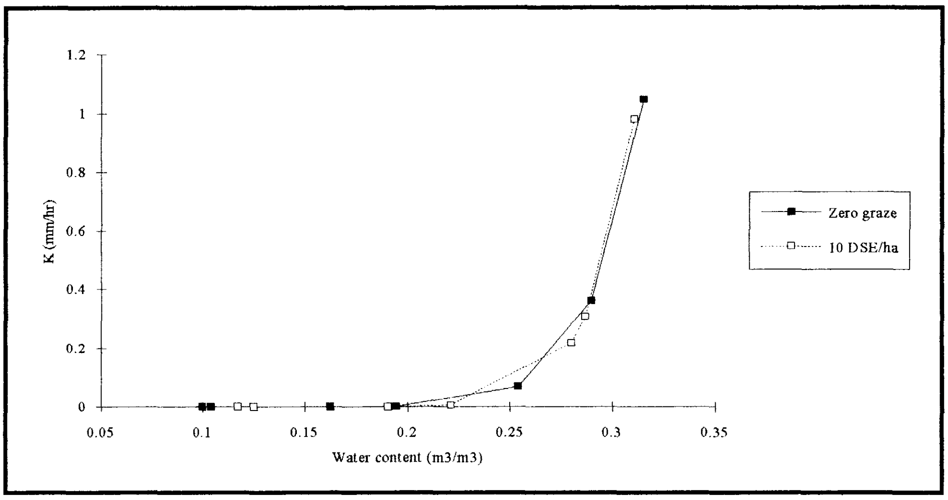
(c) 30 cm

Figure 5.4: Unsaturated hydraulic conductivity at four tensions and three depths, (a) surface, (b) 20 cm and (c) 30 cm*

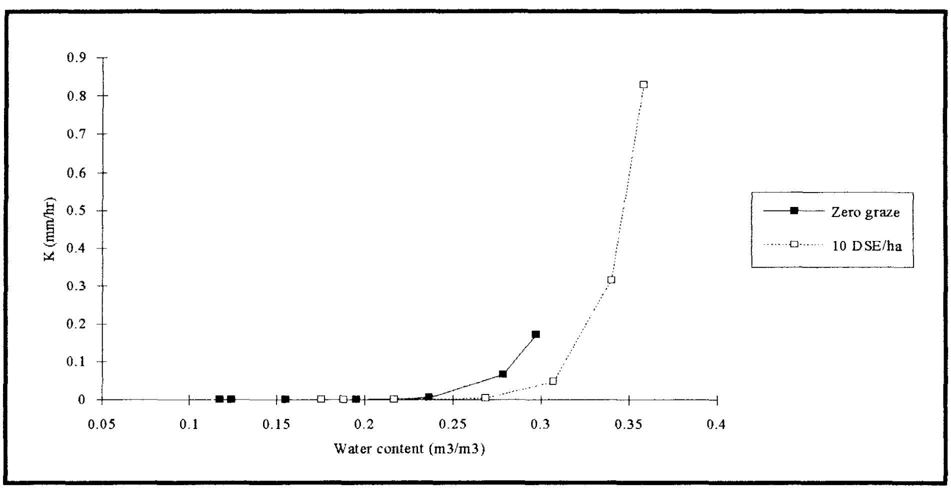
***For each tension, values shown with the same letter are not significantly different**



(a)



(b)



(c)

Figure 5.5: K(θ) relationship at three depths,(a) surface, (b) 20 cm and (c) 30 cm

5.4 Discussion

5.4.1 Bulk density

Grazing did not have a significant effect on bulk density at the depths sampled in this study. Alderfer and Robinson (1947) found that for soils ranging in texture from clay loams to sandy loams, differences in bulk density between grazed and ungrazed sites occurred only in the top 2.5 cm. On the other hand, McCarty and Mazurak (1976) found significantly higher bulk densities for a clay loam soil that had been grazed for 25 years compared to ungrazed sites. Average bulk densities in the top 7.6 cm were 1.22 g cm^{-3} for the continuously grazed compared to 1.04 g cm^{-3} for the ungrazed plots. Witschi and Michalk (1979) measured bulk density in the 0 to 4 cm layer of an irrigated clay soil which was trampled by sheep 24 hours after irrigation ceased. Bulk density increased with increasing stocking rate, with the greatest increase in bulk density occurring in the wettest plot. From a review of the effects of grazing on bulk density, Packer (1988) concluded that the effects of grazing occurred only in the top 10 cm of soil for medium to heavy textured soils.

Proffitt *et al.* (1993) found no significant differences between grazing treatments in the bulk density of the surface 0-4 cm for a sandy loam soil. They found that the depth of hoof prints were only 2 cm or less. They suggested that differences in bulk density probably occurred in the upper soil surface, but were masked by the deeper volumes of soil sampled. This could also account for no differences being found between treatments in this present study, since bulk density was not measured above 5 cm. Lemin (1992) measured bulk density on the same gleyed podzolic soil on Big Ridge 1. She took undisturbed soil cores at 0 to 8 cm and 8 to 16 cm depth, and divided each core into 1 cm slices. Bulk density was significantly higher in the grazed treatment only at 1 to 2 cm and 6 to 7 cm. There was no differences in bulk density below 7 cm. The depth of sampling used in this present study may have masked the differences observed by Lemin (1992).

5.4.2 Total porosity

As total porosity is calculated from bulk density, no significant differences were found between the grazed and ungrazed plots. Again, the depth of coring used may have masked differences.

Several researchers have reported no changes in total porosity with grazing, but have found significant differences in porosity at certain tensions, indicating differences in the volume of different pore sizes. Generally grazing reduced the volume of larger pores (those drained at up to 10 kPa tension). Alderfer and Robinson (1947) compared capillary and non-capillary porosity in the surface 2.5 cm layer of grazed and ungrazed sites. The grazed site had a much lower non-capillary porosity, indicating a reduction in macropores (> 0.03 mm).

Reed (1957) found no differences in total porosity between trampled and ungrazed treatments, but porosity measured at 60 cm suction (6 kPa) was significantly lower in the grazed area. Gradwell (1968) measured porosity at 50 cm water suction (5 kPa) and found a decrease in the volume of large pores with increased stocking rate.

Whitbread (1992) measured total porosity and air-filled porosity at -10 kPa, at 0-4 cm and 8-12 cm depths on the gleyed podzolic soil at Big Ridge. He found no significant differences in total porosity between the zero graze and 10 DSE per ha treatment at 0-4 cm. However, total porosity was significantly higher for the 10 DSE per ha at the 8-12 cm depth. Porosity measured at -10 kPa, was not significantly different between treatments at either depth. Porosity at this tension was, however, significantly lower for a stocking rate of 15 DSE per ha compared with 0 and 10 DSE per ha at both 0-4 cm and 8-12 cm.

5.4.3 Moisture characteristic

Compaction by grazing animals reduces the macroporosity of a soil. The suction required to drain macropores is low, at around 10 kPa. The effects of compaction are exhibited at the wet end of the moisture characteristic (0 to -100 kPa). Compaction may reduce total porosity and therefore the water content at saturation should be greater in the uncompacted soil. However, compaction may decrease the proportion of large transmission pores so that, as potential decreases, the uncompacted soil would initially lose water more quickly than the compacted soil. The compacted soil could hold more water due to the larger volume of smaller sized pores (Warkentin, 1971). At the dry end of the moisture characteristic compaction does not greatly affect the micropores so that little difference between compacted and uncompacted soils would be expected.

Table 5.3 indicates that there is a significant interaction between depth and stocking rate at matric potentials of -5, -10 and -30 kPa. Figures 5.1 to 5.3 show that the interaction is due mainly to the difference between the ungrazed and grazed plots at 30 cm depth. There is no difference between the 5 cm and 20 cm depths.

The high values of the coefficients of determination (R^2) shown in Table 5.4 suggest that Campbell's (1974) function describes the water retention data well. Campbell's function best fits soil layers with a high clay content, hence narrower pore size distribution (van

Genuchten and Neilsen, 1985) As shown in Table 5.4, Campbell's function provides the best fit at the 30 cm depth, with an R^2 of 0.99 for both treatments. The *b* values increase with depth for both treatments. At each depth the moisture characteristic is steeper for the 10 DSE per ha than the zero graze treatment. The steeper curves mean that for a given change in matric potential, there is a smaller change in water content.

The moisture characteristic at the 5 cm depth (above the compacted zone), presented in Figure 5.1, shows that there is no significant difference in the amount of water held at any given potential between the two treatments. Likewise, at the 20 cm depth (within the compacted zone) there is no significant difference in the moisture characteristic between the two grazing treatments (Figure 5.2).

At the 30 cm depth, which according to Lemin's (1992) data for soil strength is believed to be below the compacted zone, the 10 DSE per ha treatment holds significantly more water than the zero graze treatment at all potentials. Given that the total porosity for the two treatments is similar (Table 5.2) the differences in water retention at 30 cm are possibly due to textural differences. All cores were taken at a fixed depth of 5 to 9 cm, 20 to 24 cm and 30 to 34 cm. The depth to the top of the B horizon varies and hence the cores taken at 30 to 34 cm could have contained varying amounts of B horizon material. The texture of the sampled soil could thus vary, resulting in the observed difference in soil moisture characteristic.

Whitbread (1992) measured the moisture characteristic on a gleyed podzolic soil type on Big Ridge 1 at 'Chiswick' at 0-4 cm and 8-12 cm. At 0-4 cm he found no significant differences between the zero graze and 10 DSE per ha; however, the higher stocked treatments, 15 DSE per ha and 20 DSE per ha, held significantly more water than the zero-graze at -5 kPa. This was attributed to the decrease in transmission pores and hence less loss of water by drainage. At 8-12 cm, no significant differences between treatments were found. These results are in agreement with the lack of a significant effect of grazing for the 5 cm and 20 cm depths.

5.4.4 Unsaturated hydraulic conductivity

Hydraulic conductivity is affected by pore size distribution and pore continuity. There is a significant difference in unsaturated hydraulic conductivity with depth (Table 5.5). Unsaturated hydraulic conductivity is significantly higher at 10 and 20 mm of tension at the soil surface compared to the 20 and 30 cm depths. Table 5.5 also shows a significant interaction between stocking rate and depth for these tensions. The zero graze treatment has a significantly higher infiltration rate than the 10 DSE per ha treatment at the soil surface as shown in Figure 5.4.

The pore size that is conducting water varies with the tension applied. At 20 mm tension, pores with a diameter greater than 1.5 mm cannot exert enough suction to move water and will remain dry, but at a lower tension of 10 mm, pores up to 3 mm in diameter will conduct water. The treatment difference found in the surface soil at 10 mm tension but not at 20 mm tension means that the ungrazed soil is conducting more water through pores greater than 1.5 mm in diameter compared to the grazed soil. There may be a greater number of pores of that size in the zero graze treatment and/or the pores may be better connected resulting in more continuous pores.

In reviews by Packer (1988) and Gifford and Hawkins (1978), there is considerable evidence that infiltration is reduced with increased stocking. The main cause is thought to be a reduction in the number of macropores, particularly near the soil surface. The soil moisture characteristics did not indicate significant differences in macroporosity at 5 cm depth, but differences could be present above 5 cm. Furthermore, soil retention data do not provide information on the continuity of pores.

Whitbread (1992) measured hydraulic conductivity at 13 mm and 36 mm tensions on the gleyed podzolic soil. Hydraulic conductivity at both tensions decreased as the stocking rate increased, with the ungrazed treatment having a significantly higher infiltration rate at 13 mm tension than the 15 and 20 DSE per ha stocking rates. There was, however, no difference between the 0 and 10 DSE per ha stocking rates. He found no significant differences in hydraulic conductivity at 36 mm tension.

Willatt and Pullar (1983) measured hydraulic conductivity on a silty loam soil from a grazing trial where sheep were stocked at 0, 10, 15, 19 and 22 per hectare. They found hydraulic conductivity was reduced by an increase in stocking rate. Significant differences between the ungrazed and grazed treatments only occurred at stocking rates of 15, 19 and 22 sheep per ha.

Recently Proffitt *et al.* (1993) examined the effects of grazing on water infiltration into a red-brown earth. They found infiltration rates, in the plots where grazing was deferred for 6-11 weeks after opening winter rainfall, to be significantly higher than the plots were grazing commenced from the start of the wet season The ungrazed plot had a significantly higher infiltration rate than both grazing treatments.

Reduced infiltration has important consequences for runoff and soil water storage. Neath *et al.* (1991) attributed reduced soil water under grazing to low infiltration rates. Likewise, Proffitt *et al.* (1993) found water content to be higher after rain throughout the soil profile in the ungrazed treatment compared to grazed treatments as a result of greater infiltration.

5.4.5 Hydraulic conductivity function [K(θ)]

Figure 5.5 shows that the K(θ) curves are much steeper at high water contents, with a small change in water content leading to a large change in hydraulic conductivity. At water contents less than 0.25 m^3 m^{-3} there is very little change in conductivity for both treatments at all depths.

At a given water content the zero graze treatment has a higher hydraulic conductivity at each depth compared to the 10 DSE per ha treatment, implying larger more continuous pores.

Williams (1983) stated that at a given water content, conductivity can be reduced by about three orders of magnitude as the field texture increases from a sandy loam to a clay. Schafer (1980) reported the clay content decreases from 21.4 percent in the 0 to 3 cm layer to 19.4 in the 15 to 30 cm layer. At 30 to 51 cm, clay content increases to 59.4 per cent. A decrease in hydraulic conductivity with depth due to increasing clay content is evident in Figure 5.5.

5.5 Conclusions

There were no differences in bulk density or porosity at 5 to 9 cm, 20 to 24 cm or 30 to 34 cm. However, past research at 'Chiswick' suggests that changes in soil physical properties due to grazing occur in the top 7 cm. Such changes may have been masked by taking cores only from 5 to 9 cm.

There were no significant differences in the moisture characteristic between grazing treatments at 5 to 9 cm or 20 to 24 cm. Again, it is possible that the effects of compaction are occurring closer to the soil surface. The 10 DSE per ha treatment held significantly more water at all potentials than the zero graze treatment at 30 cm depth. This is attributed to textural differences due to variation in horizon depth over the plots.

Differences in macroporosity and continuity of pores at the soil surface would account for the higher infiltration rates at the soil surface under the zero graze treatment.

6.0 Measurement of the hydraulic properties of a gleyed podzolic soil

6.1 Introduction

A knowledge of the soil moisture characteristic and the hydraulic conductivity function is essential to the understanding of the soil-water system. The soil moisture characteristic describes the relationship between soil water potential and water content and indicates the ability of the soil to store water. Hydraulic conductivity is a measure of a soil's ability to transmit water. It is affected by pore size and shape and decreases with a reduction in soil water content. Saturated hydraulic conductivity describes the hydraulic conductivity of a saturated soil. It is constant for a given soil, but is affected by soil texture and structure.

The Soil Water Infiltration and Movement (SWIM) model uses Campbell's (1974, 1985) equations to define soil hydraulic properties. The inputs used to solve Campbell's equations are saturated water content, air-entry potential, *b* (the slope of the best fit line relating water content to matric potential) and saturated hydraulic conductivity.

The aim of this experiment was to measure the hydraulic properties used as inputs into SWIM which are later used in SWIM simulations (Chapters 7 and 8). Differences in the moisture characteristic and saturated hydraulic conductivity at different depths were examined in a gleyed podzolic soil.

6.2 Materials and methods

6.2.1 Moisture characteristic

Undisturbed soil cores were taken from the A1, A2, B1 and the B2 horizons, from a soil pit located outside the boundary of plot 1 (Figure 4.3). The middle of the core was located at depths of 5, 25, 40 and 70 cm. Steel cores with an internal diameter of 73 mm and a depth of 40 mm were used, and six replicates were taken at each depth. The soil profile had similar morphology to the gleyed podzolic soil used to assess the effects of grazing on soil hydrology, as described in Chapter 5. The area outside the experimental plots was grazed periodically.

Moisture retention at matric potentials of -5, -10, -30, -100 and -300 kPa was determined after equilibration on pressure plates (Reeve and Carter, 1991). A Decegon SC-10A Thermocouple Sample Changer Psychrometer, as described by Rawlins and

Campbell (1986), was used to determine moisture retention at lower matric potentials (-1000 and -1500 kPa).

Bulk density was measured on three extra undisturbed cores that were taken at each depth. The average bulk density of the three cores was used to convert gravimetric to volumetric water content. At the 40 cm depth (B1 horizon) the water content at saturation, as calculated from bulk density, was found to be lower than the water content at potentials of -5, -10, -30, -100 and -300 kPa for three replicates. This anomaly is attributed to soil variability and the data from these three cores was excluded from further analysis. Therefore, the moisture characteristic for the B1 horizon was obtained using only three replicates. At the 70 cm depth (B2 horizon), the water content measured at -5 kPa for two cores was greater than saturated water content measured from the average bulk density. These two cores were excluded from the experiment. As the bulk density cores were taken within 1 m of the moisture characteristic cores, it suggests great soil variability. Spatial differences in the depth to the B horizon is an important source of variation. Given this variation, it would have been better to obtain a bulk density of each individual core which was taken for moisture characteristic determination.

The moisture characteristic data were fitted to Campbell's (1974) power function describing the relationship between water content and matric potential, as described in Section 5.2.2.

Subsequent analysis of the moisture characteristic showed that water content of the cores subject to 100 and 300 kPa pressures did not relate well to the other points on the moisture characteristic curve. The reason was most probably being the age of the plates. Over time the pores in the pressure plates become blocked with fine soil particles and precipitated salts so that water movement through the plate becomes restricted. The cores subject to -100 and -300 kPa were therefore omitted and Campbell's (1974) function was calculated for each horizon from the water contents corresponding to matric potentials of -5, -10, -30, -1000 and -1500 kPa.

6.2.2 Saturated hydraulic conductivity

Saturated hydraulic conductivity (K_S) was measured at depth in the field using a well permeameter. A well permeameter measures steady state infiltration rate when a head of water is ponded to a constant depth. The well permeameter consists of a water reservoir, which has an air inlet tube on the inside (Figure 6.1). The water in the reservoir fills an augured test hole to a predetermined depth. The air inlet tube is then opened allowing air to enter at the top, and its lower opening coincides with the water level in the hole. A 0.01 M solution of calcium chloride was used in the reservoir to avoid dispersion occurring in the test hole due to the rapid influx of water into the hole. As water in the test hole drops, air is admitted through the air inlet tube releasing water from the reservoir into the test hole to maintain a constant head. The water level in the reservoir was observed over time and the rate of outflow determined, from which saturated hydraulic conductivity was calculated.

The measurements were taken at four depths: 20-50 cm, 60-90 cm, 90-120 cm and 120-150 cm. The K_S measurement reflects an integration of different values of K_S over the height of ponded water (30 cm). Each measurement was replicated eight times. Saturated hydraulic conductivity was calculated using the following equation of Elrick and Reynolds (1992):

$$K_{fs} = CQ / (2\pi H^2 + \pi r^2 C + 2\pi H / \alpha^*) \qquad [6.1]$$

where K_{fs} is the field-saturated hydraulic conductivity (cm sec^{-1})

C is a dimensionless shape factor which depends on the H/r ratio ($C = 2.3$)

Q is the steady state infiltration rate (cm^3 sec^{-1})

H is the height of ponded water ($H = 30$ cm)

r is the radius of the auger hole ($r = 3.25$ cm)

α^* is equal to saturated hydraulic conductivity divided by the matric flux potential ($K_S/\phi m$) ($\alpha^* = 0.12$)

C was estimated using the relationship between H and r described by Reynolds (1993). Elrick and Reynolds (1992) provide estimates of α^* based on soil textural and structural considerations.

In equation [6.1], the terms in the denominator on the right-hand-side of the equation describe the contribution of hydrostatic pressure, gravity and capillarity, respectively, to the total flow of water out of the well (Elrick and Reynolds 1992).

6.2.3 Statistical analysis

Statistical analysis was performed on all data using analysis of variance and the significance of differences between means was tested by Duncan's multiple range test (DMRT).

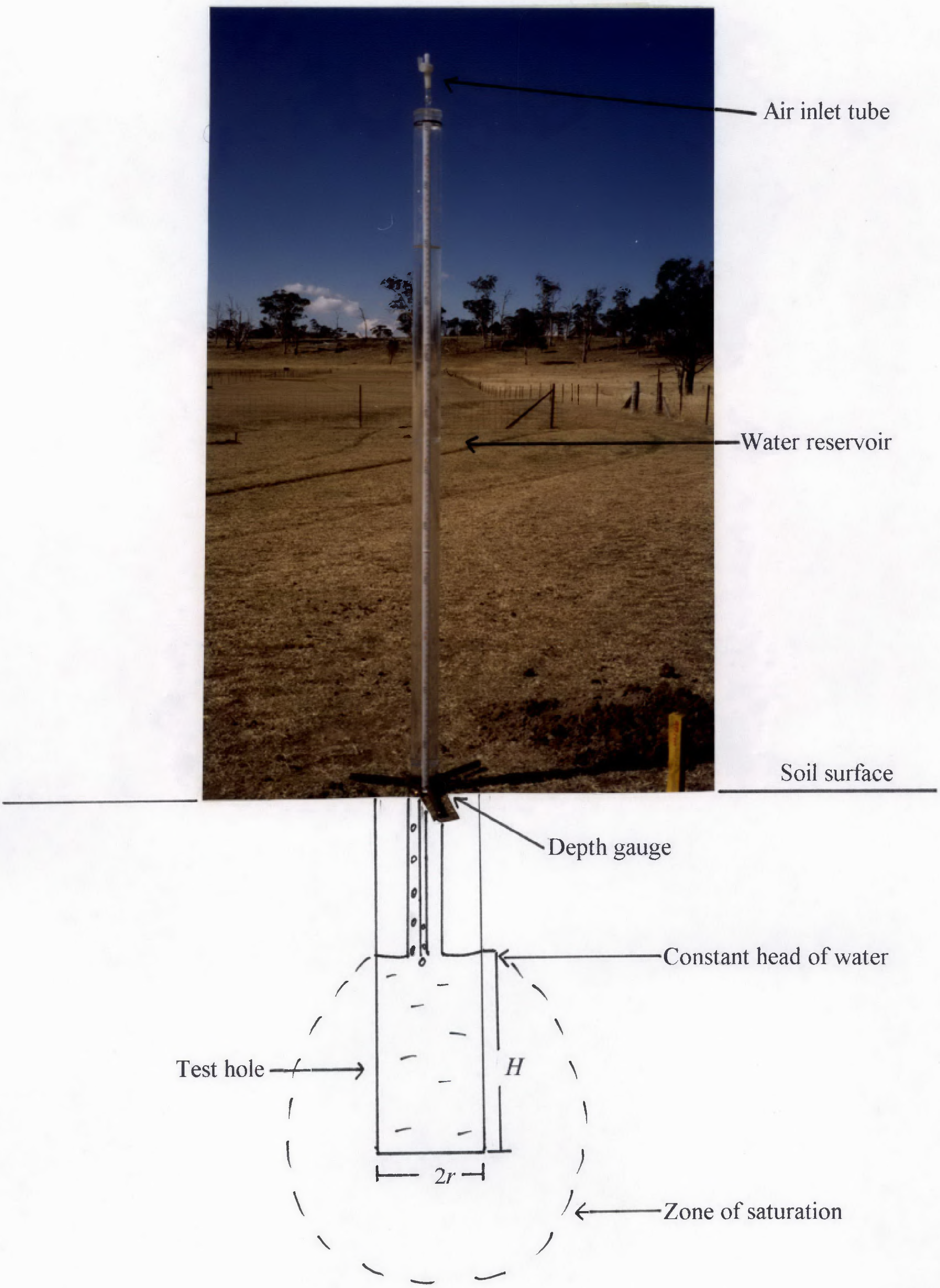


Figure 6.1: A well permeameter used to measured K_S at depth in the soil profile

6.3 Results and discussion

6.3.1 Laboratory determined soil moisture characteristic

The moisture characteristic data determined in the laboratory were fitted to Campbell's (1974) water release curve using the parameters listed in Table 6.1. The moisture characteristics are shown in Figure 6.2 for each horizon. Coefficients of determination values (R^2) of 0.99 for the A1 and A2 horizon and 0.98 and 0.96 for the B1 and B2 horizons indicate that Campbell's function fitted the data well.

Table 6.1: Parameters required for calculating Campbell's water release curve: A and B = regression coefficients, R^2 = coefficient of determination, θ_s = saturated water content, b = the slope of the moisture characteristic, and ψ_e = air-entry potential

	A1(5 cm)	A2(25 cm)	B1(40 cm)	B2(70 cm)
A	4.41	4.19	4.16	4.24
B	-0.19	-0.16	-0.07	-0.09
R^2	0.996	0.992	0.981	0.960
θs (m^3 m^{-3})	0.47	0.41	0.52	0.50
b	5.17	6.13	13.56	11.56
ψe (cm water)	-18.09	-19.12	-19.11	-46.97

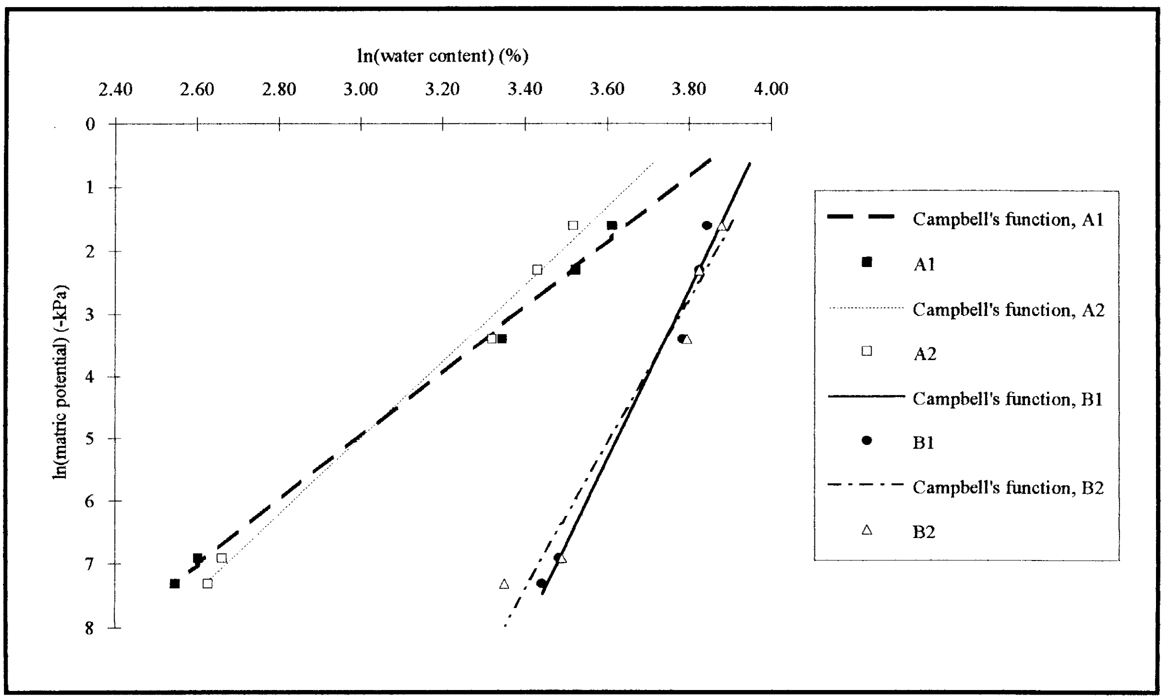


Figure 6.2: Laboratory determined moisture characteristic for different horizons of a gleyed podzolic soil

Williams (1983) examined the moisture characteristics of different textured soils, ranging from sands to loams to medium clays and found that the higher the clay content, the more water the soil held at all potentials from -4 to -1500 kPa. The moisture characteristics presented in Figure 6.2 show that the B1 and B2 horizons hold more water over the range of the potentials measured compared to the A1 and A2 horizons. This is attributed to the higher clay content of these lower horizons (Section 4.4, Table 4.3). The clay particles and small pores hold onto water tightly, requiring much lower potentials than the topsoil for water to move. This is supported by the low air-entry potential of -46.97 cm H_20 for the B2 horizon. Campbell (1985) stated that as clay soils have a smaller mean particle diameter, they have lower (more negative) air-entry potentials and larger *b* values than coarse-textured soils. However, the air-entry potential for the B1 horizon is only slightly lower than for the overlying A1 and A2 horizons despite its higher clay content. This is probably a result of soil variability. The air-entry potential is also determined by soil structure and a weakly structured soil may have a higher air-entry potential compared to a well structured soil despite texture differences.

The moisture characteristic curves for the B1 and B2 horizon have a higher *b* value, hence a steeper curve than the A1 and A2 horizons. Therefore, a given change in water potential results in a smaller change in water content in the subsoil compared to the topsoil. A high *b* value also indicates a narrow pore size distribution, which is expected in the clay B horizons.

Spatial variability of soil characteristics has been well documented (Petersen and Calvin, 1986; Kutilek and Nielsen, 1994). The differences in bulk density measurements for the B1 and B2 horizons (Section 6.2.1) are an example of this variability. Depth to the B horizon is a major source of variation in the gleyed podzolic at Big Ridge 1. To improve bulk density measurements it would have been better to have taken bulk density measurements of each individual soil core. Another improvement of technique would be, instead of using different cores for each point on the moisture characteristic curve, to use the same core to determine the moisture characteristic. The water content of each core would be measured at -5, -10, -30, -100 and -300 kPa, starting at the highest potential. Once the soil had equilibrated the core would be weighed then placed back onto the pressure plate and equilibrated at a lower potential. This process would be repeated at each potential and then finally the core would be oven dried to obtain the moisture content at each potential. Using a single core for each moisture characteristic should give a better fit of data points because variability between cores would be removed. However, variation between cores would still result in differences between curves.

6.3.2 Saturated hydraulic conductivity

The saturated hydraulic conductivities at 20-50, 60-90, 90-120 and 120-150 cm depths are presented in Table 6.2. Saturated hydraulic conductivity (K_S) decreased with depth. K_S is significantly greater at 50 cm compared to the depths below. From the classification of K_S given by Marshall (1969), K_S at 50 cm is moderately slow, K_S at 90 cm and 120 cm are slow and K_S at 150 cm is very slow. These values suggest that water movement is retarded in the clay rich subsoil leading to poor drainage.

Table 6.2: Mean saturated hydraulic conductivity for various depths in the soil profile

	Saturated hydraulic conductivity (mm/hr)			
Depth	**20-50 cm**	**60-90 cm**	**90-120 cm**	**120-150 cm**
Mean Ks (mm/hr)	13.16^{a}	2.25^{b}	1.88^{b}	0.50^{b}
Standard error	2.20	0.49	0.55	0.09

Means within a row followed by the same letter are not significantly different at P=0.05 according to DMRT

McKeague *et al.* (1982) stated that the major factors contributing to high K_S values were biopores, textures coarser than loamy fine sand, and strong fine to medium blocky structure. The subsoil of the gleyed podzolic had a strong structure consisting of 50 to 100 mm long prismatic peds. However, the soil was dense, and showed signs of periodic waterlogging, such as mottles and gleying. No cracking was observed at depth when the soil pit was initially excavated and, when pits were left open, cracks did not develop until the soil had dried to almost wilting point (D.A. MacLeod, *pers. comm.*). The subsoil had very few biopores.

6.4 Conclusion

The hydraulic properties of the A horizon differed quite markedly from those of the B horizon due to the duplex nature of the gleyed podzolic soil. The B horizon being rich in clay holds more water than the A horizon over the potential range -5 kPa to -1500 kPa. The B horizon has a lower air-entry potential and steeper moisture characteristic than the A horizon. Saturated hydraulic conductivity decreased with depth, again as a result of increasing clay content.

The hydraulic properties of the gleyed podzolic indicate that water movement is restricted in the B horizon due to the predominance of micropores, leading to poor drainage.

Spatial variability in soil properties led to some measurement problems. When determining the moisture characteristic on a soil which exhibits much variation, the moisture characteristic should be determined for individual cores so any outlying results due to spatial variability can be identified.