DEVELOPING HYDROLOGICAL RESPONSE MODELS FOR SELECTED SOILSCAPES IN THE WEATHERLEY CATCHMENT, EASTERN CAPE PROVINCE

By

Darren Bouwer

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Supervisor: Prof P.A.L. le Roux

Co-supervisor: Dr J.J. van Tol

Dedicated to Hendrik Bouwer,

For all the support and opportunities, you are deeply missed.

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DECLARATION

I hereby declare that this dissertation submitted for the Magister Scientiae Agriculturae degree at the University of the Free Sate, is my own work and has not been submitted to any other University.

I also agree that the University of the Free State has the sole right to publication of this dissertation.

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Darren Bouwer

ABSTRACT

Soil acts as a first order control and determines the flowpaths of water exiting a catchment. The long term interactive relationship between water and parent material has resulted in present soil morphology. The soil chemical properties react faster than morphological properties to a change in pedological processes predominantly controlled by the water regime in soil. Flowpath characteristics, from very fast to very slow, determines the resultant soil chemistry. Soils can act as flowpath or as a storage mechanism of water in the soilscape.

Soils were grouped according to their hydrological response. Recharge soils are soils, which serve as conduits of infiltrated water and recharge underlying fractured bedrock. Interflow soils were divided into deep interflow and shallow-interflow responsive depending if the lateral movement of water was in the deep subsoil or in an E horizon close to the surface. Responsive soils, due to the saturated nature, will saturate quickly after infiltration inducing overland flow.

The soil chemical properties for representative profiles of a soilscape were described, characterised and interpreted. Pedological processes were inferred from the soil morphology, chemical properties and water regime measurements in soil profiles. Chemical properties were used to verify if the morphological properties are in phase with the current water regime. Annual duration of saturation ($ADs_{>0.7}$), which is defined as 70% saturation of porosity was also used to support the chemical property deductions.

A conceptual hydrological response model was constructed using soil morphology as an ancient indicator of flowpaths, which was improved using chemical properties as recent indicators of the current water regime. The current water regime was verified with real time snapshots of hydrology using hydrometric data.

Keywords: Morphology, Chemical properties, Conceptual hydrological model, Flowpath, Storage mechanism.

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LIST OF ABBREVIATIONS

$ADs_{>0.7}$ -	Annual duration of saturation above porosity of 0.7
Av	Avalon
Bd -	Bloemdal
BS -	Base saturation
DUL -	Drained upper limit
ET -	Evapotranspiration
gh -	G horizon
gs -	E horizon
Gf -	Griffin
Hu -	Hutton
Ka -	Katspruit
Kd -	Kroonstad
K_{sat} -	Saturated hydraulic conductivity
M.A.P. -	Mean annual precipitation
Ms -	Mispah
Ne -	Neocutanic B
Lo -	Longlands
on -	Unspecified with signs of wetness
ot -	Orthic A
P201 ect -	Refers to profiles of Weatherley catchment
Pn -	Pinedene
re -	Red apedal B
SLF-	Subsurface lateral flow
so -	Saprolite
sp -	Soft plinthic
TMU -	Terrain morphological unit
Tu -	Tukulu
ye -	Yellow brown apedal

CHAPTER 1

INTRODUCTION

1.1. BACKGROUND

Soil is a natural entity. Its properties developed slowly over a long time equilibrating with environmental impact (Dorronsoro, n. d.). The dominant environmental impact, namely the interaction between water and parent material, manipulated by terrain, continue in soil. Hydropedology as described by Lin (2003) incorporates this interactive relationship between these factors in the unsaturated zone. Hydropedology has the ability to link hydrology and pedology (Pachepsky *et al.,* 2006). Lin (2010) states that hydropedology is an important contributor to critical zone studies. This zone, with high bio-chemical activity, is of importance in soil science, hydrology and ecology, especially with rising levels of pollution in soils. The interaction of water between soil and fractured rock plays a significant role in the behaviour of catchments (Van Tol *et al.*, 2010; Kuenene *et al.*, 2011; Lorentz, 2004)

Hydrological studies are typically researched on catchment scale. Catchment response is determined by the response of individual hillslope (Sivapalan, 2003). Hillslope response is mainly controlled by soil distribution patterns (Soulsby *et al.*, 2006), and its interaction with underlying rock, and is therefore a first order control of water movement within the earth (Park *et al.*, 2001, Soulsby *et al.*, 2006). The soil distribution pattern controls the hydrological processes such as flowpaths, residence times and storage mechanism (Soulsby *et al.*, 2006) which influence the quantity and chemical composition of the water exiting the soilscape (Jacks & Norrström, 2004). Soils have specific hydrological functions (Van Tol *et al.*, 2010, Kuenene *et al.*, 2011) and they can be divided in horizons with specific hydrological functions (Kuenene *et al.*, 2011, Van Tol *et al.*, 2011).

Soil morphology is a visible indicator of the interaction between water and parent material and the resultant variation indicates soil water regimes. There are many factors like hydraulic conductivity, bulk densities, preferential flow paths, different soils which make it very difficult to classify a hillslope and it is even suggested by Bevan (2000), that all hillslopes differ and that hillslopes should be viewed individually. Soil morphology has the ability to decipher these differences and produce conceptual hydrological response models (Fritsch & Fitzpatrick, 1994, Soulsby *et al.*, 2006, Ticehurst *et al.,* 2007, van Tol *et al.,* 2010, van Tol *et al.,* 2011 and Kuenene *et al.*, 2011).

Although soil physical data usually reported in soil surveys are originally used to design pedotransfer functions (Bouma 1989), soil morphology, as reported in soil surveys, can also serve as a pedotransfer function to infer the hydrological response of soils (Fritsch & Fitzpatrick, 1994). Soil morphological data is more accessible than soil physical and hydrometric data commonly used to determine hydrological processes within a soilscape. Gathering soil physical and hydrometric data are tedious, time consuming and often unreliable (Park and Burt, 1999). Soil morphology can play an important role in conceptualising soilscape hydrology and assist in assigning the correct model structure in soilscape hydrological models (Lorentz *et al.*, 2007; Van Tol *et al.*, 2011).

The soil water regime correlates with the morphology of soils (Van Huyssteen *et al.*, 2005). For instance soils with red apedal B (*re*) horizons are well drained and responded with short periods of near saturation contrary to soils with grey (*gh*) horizons which are saturated for long periods. The genetic *re* horizons respond as recharge hydrological units compared to grey *gh* horizons serving as storage mechanisms. Recharge soils are defined as soils that rapidly transmit water vertically in the profile. Interflow soils, also known as throughflow, are soils were water is diverted laterally down the soilscape by an impeding layer. The impeding layer has a lower saturated hydraulic conductivity (K_s). Responsive soils are saturated for long periods of time due to saturation excess with water or low K_s (Le Roux *et al.*, 2011; Van Tol *et al.*, 2012). The yellow brown apedal B (*ye*) horizon is saturated slightly longer than the *re* horizon (Van Huyssteen *et al.*, 2005). However the *ye* horizon of the *Gf* soil is interpreted as recharge. This contradicts the results of Van Huyssteen *et al.* (2005). The yellow-brown colour of the *Gf* soil is related to soil mineralogy (Fey, 1983). However the colour variation is rather soil redox chemistry related and interpreted as being due to quick reduction due to higher OC contents in the topsoil rather than longer saturation typical of other *ye* subsoils. The hydrological relationships of horizons, pedons and soilscapes discussed above, contributed to the improvement of a hydrological model by accommodating soil interflow (Lorentz *et al.*, 2007).

The process of soil formation is relatively long (10^2-10^4 years) whereas the process of pH change is shorter (10^0 - 10^3 years) at catchment scale (MacEwan, 1997). Properties relating to soil morphology; cutans, drainage, which is largely related to structure, soil colour (hue), and the delineation of master horizons in most soils, except for acid sulfate soils with sulfidic material is unlikely to change in a human lifetime whereas soil chemical properties are likely to change (MacEwan & Fitzpatrick, 1996).

Although physical weathering in soil is recognised by rock fragments in the soil, it is chemical weathering that changes rock into soil with specific features. Chemical weathering can't take place without the presence of water and water is the mediator in further chemical reactions in the soil (Essington, 2004). Hydrology therefore plays an important role in the resultant soil chemistry.

Geochemical indicators of hydrological processes have been found by several workers (e.g. McDaniel *et al.*, 1992 and Park & Burt, 1999) and could transform the procedure of soilscape modelling by reducing the time factor of physical measurements and improve the quality as hydrometric and isotope data, which, without the support of other data could produce erroneous predictions (McDonnell *et al.,* 2007).

Modelling soil chemistry using water chemistry, at soilscape scale, is very difficult and consequently most soil chemistry is modelled using water chemistry at catchment scale (Burns *et al.,* 1998). Numerous one dimensional studies have been done on the effects of hydrology on soil chemistry focussing on a single soil attribute or at plot scale. However soilscapes should be studied three dimensionally, instead of at point observations, in order to gain a holistic understanding of the complexity of soilscape hydrological behaviour (Tromp van Meerveld & McDonnell 2006), in other words in a soilscape context.

Studies have been done on the effects of water on soil chemical properties. The Fe-rich and Mn-rich concretion contents are commonly highest in horizons with fluctuating water tables rather than horizons that are more permanently saturated (D'Amore *et al.*, 2004; le Roux, 1996). According to Park and Burt (1999) and McDaniel *et al.* (1992) Fe and Mn can be used as a pedochemical indictors and identification of through-flow in soils. Ferrrous iron (Fe^{2+}) can be absorbed on the exchange sites thereby freeing Ca and Mg and increasing the Fe content (Phillips & Greenway, 1998).

Developing a concept is the first step in scientific research and that is also true of hydrology. Soilscape hydrology can be conceptualised using signatures related to soil/water interaction. Soil morphology is effectively applied as "signatures of soil features". This emphasises the role of a soil profile description as a pedotransfer functions. However, the relatively slow reaction of soil morphology to changes in soil water regime and the fact that some morphological features are irreversible, questions its universal application. Hydrometrics is applied as current (real time) indicators of response but its application is limited by its snapshot nature. Soil chemistry can fill this gap as it can connect point data in a soilscape and connect soil horizon and soil pedon response at soilscape scale.

The term soilscape replaces hillslope in this discussion. Hydrological soilscapes are related to the pedological catena. Milne (1936) first referred to a catena as the typical distribution of soils in a landscape and was later modified to toposequence (Bushell, 1942). Toposequence found in Fritsch & Fitzpatrick, (1994) and Van Tol *et al.* (2010) of red freely drained soils on the crest and hydromorphic soils close to the stream is typical example of a sequence of soils that increase in wetness down slope relating the distribution of soils to the landscape. Soilscape refers to this orderly distribution of soils in the landscape (Landis 2013). Soilscape is the merging of the terms soil and landscape (Blume *et al.*, 2002), therefore it doesn't detract anything from previous understanding of a hillslope but emphasises the third dimension of width to the distribution of soils in the catchment.

1.2. MOTIVATION and HYPOTHESIS

1.2.1. MOTIVATION

The Weatherley catchment is an ideal location for data mining due to the extensive pedological and hydrological research conducted in the catchment. Soil chemical properties are a result of pedological processes, controlled by soil forming factors. Therefore an understanding of the soil forming factors (most importantly climate and topography) and correlating these factors with soil morphology and chemistry. Soil chemistry will increase the hydropedological interpretations based on morphology, and the interaction with current soil morphology will enable the explanation of current soil chemical properties and the extrapolation of these trends in similar morphological, geological, topographic and climatic settings.

1.2.2. HYPOTHESIS

A conceptual hydrological response model of a soilscape can be developed from soil morphology, an ancient signature of hydrology on pedogenesis, and improved using soil chemistry, a recent indicator of pedogenesis, to represent current hydrology.

CHAPTER 2

LITERATURE REVIEW

2.1. CHEMICAL PROPERTIES

2.1.1. pH

Buol *et al.* (1980) regards pH as the most important chemical measurement that can be made in soil. Sumner (2000) defines pH, [-log (H^+)], as the negative logarithm, base 10, of H^+ activity. Buol et al. (1980) states that by specifying activity instead of concentration, it is recognised that there are other hydrogen ions in the soil but that only the H^+ in solution is measured.

Lin *et al.* (2005) found that the wetter regions in the study area had a lower pH due to leaching than in drier areas. The summit and the backslope were usually a drier area and had the highest pH as this is where the least leaching took place. Low pH can cause metal mobilization in the deeper soil horizons (Abesser *et al.*, 2006). Soils having a pH dependant charge will increase in CEC with an increase in pH and a decrease in CEC with a decrease in pH (Phillips and Greenway, 1998). A general classification of pH ranges are found in Table 2.1.

Table 2.1 General interpretation of pH ranges (Bruce & Raymond, 1982)

pH	Rating
>9	Very strongly alkaline
9 - 8.5	Strongly alkaline
8.4 - 7.9	Moderately alkaline
7.8 - 7.4	Mildly alkaline
7.3 - 6.6	Neutral
6.5 - 6.1	Slightly acid
6 - 5.6	Moderately acidic
5.5 - 5.1	Strongly acidic
5 - 4.5	Very strongly

2.1.2. BASE SATURATION

It is defined by the following formula

$$\text{Percentage Base Saturation} = \frac{\sum(\text{exchangable Ca, Mg, Na, K}) \times 100}{\text{CEC at pH 7 or 8.2}} \quad \text{[Eq 1]}$$

Buol *et al.* (1985) states that soils with a low percentage base saturation (BS) were considered to be dominated by kaolinitic clay and hydrous oxides, whereas soils with a high BS were considered to be dominated by 2:1 clay minerals. A general range of BS and the rating with regards to leaching in soils is represented in Table 2.2.

Table 2.2 Ratings of base saturation in soil (Metson, 1961)

Range (%)	Rating
70 - 100	Very weakly leached
50 - 70	Weakly leached
30 - 50	Moderately leached
15 - 30	Strongly leached
0 - 15	Very strongly leached

2.1.3. IRON

2.1.3.1. Origin, Formation and Distribution

Iron is the most abundant micronutrient in soil (Sumner, 2000) and found in most igneous rocks and in many minerals (Fitzpatrick, 1980). The iron is then released from the minerals through hydrolysis or oxidation of Fe^{2+} of parent material into the soil (Bohn, 1985) and is represented in Eq 2

$$\text{Fe (II)-O-Si} + H_2O \rightarrow \text{Fe (III)} + OH + \text{-Si-OH} + e^- \quad \text{[Eq 2}\}$$

Iron is found in Fe^{2+} and Fe^{3+} forms as iron oxides, silicates, carbonates and sulphates (Sumner, 2000). Goethite [FeOOH] is the most common iron oxide and found in almost all soil types and climate zones although it is more abundant in cool wet climates (Essington, 2004). Sumner (2000) differs and states that hematite [Fe_2O_3] is the most abundant but that it is closely associated with goethite. They both commonly occur as thin coatings on the soil particles (Bohn, 1985). Iron is reduced at pH levels lower than 3 and hydrolysis occurs when the pH level is higher than 3 (van Huyssteen *et al.,* 2005). Every unit of increase of pH, the solubility of Fe decreases a thousand fold (Sumner, 2000)

2.1.3.2. Aerobic Conditions

Iron is relatively evenly distributed and results in a homogeneous colour through the profile (Schwertmann, 1993). The high affinity of Fe^{3+} for OH ligand results in the formation of Fe oxides in soil and therefore hydrolysis proceeds (van Huyssteen *et al.,* 2005) as described above [Eq 2] . A low soil pH is needed for protolysis to occur and since soil pH levels are seldom low enough rarely will iron oxides dissolute.

2.1.3.3. Anaerobic Conditions

In Table 2.3 the typical reduction reactions and redox potentials. The reactions are in sequence, therefore after the oxygen and nitrogen is reduced, manganese and then iron is reduced.

Table 2.3 Reduction reactions and redox potentials in soil (Bohn et al., 1985)

Half reaction	Redox potential measured in soil (mV)
$O_2 + 4\ e^- + 4\ H^+ \rightarrow 2\ H_2O$	600 - 400
$NO_3 + 2\ e^- + 2\ H^+ \rightarrow NO_2^- + H_2O$	500 - 200
$MnO_2 + 2\ e^- + 4\ H^+ \rightarrow Mn^{2+} + 2\ H_20$	400 - 200
$FeOOH + e^- + 3\ H^+ \rightarrow Fe^{2+} + 2\ H_2O$	300 - 100
$SO_42+ + 6\ e^- + 9\ H^+ \rightarrow HS^- + 4\ H_2O$	0 - -150
$2\ H+ + 2\ e^- \rightarrow H_2$	-150 - -220
$2\ CH_2O \rightarrow CO_2 + CH_4$	-150 - -220

In anaerobic conditions there is a heterogeneous distribution because Fe oxides are reduced and unevenly distributed (Schwertmann, 1993). Under these conditions iron is reduced, therefore Fe $^{3+}$ is reduced to Fe^{2+} (Eq 3).

$$FeOOH + e^- + 3H^+ \text{-----} Fe^{2+} + 2H_2O \qquad \text{[Eq 3\}}$$

Van Huyssteen *et al.* (2005) suggested that reduction takes occurs before 100% saturation and assumed reduction at 70% saturation. Smith & Van Huyssteen (2011) found different amounts of Fe^{3+} reduced to Fe^{2+} at different saturation levels for 121 days, the results are presented in Figure 2.1. All the measurements at 60% saturation were not reduced, half the measurements at 70% saturation were reduced, and most the Fe was reduced at 80% saturation and all the Fe was reduced at 90% saturation.

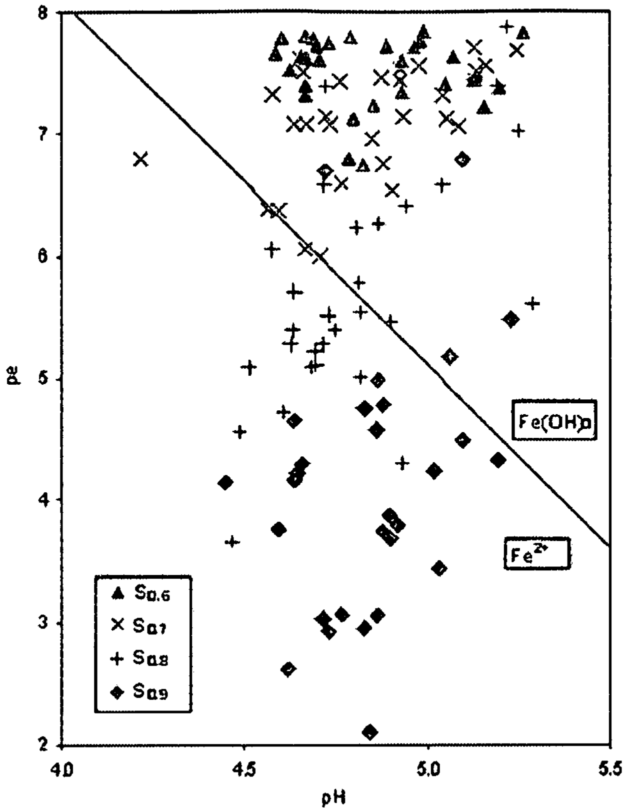

Figure 2.1 Stability diagram of Fe at different degrees of saturation (S0.6=60%, S0.7=70%, S0.8=80%, S0.9=90%) (from Smith & Van Huyssteen, 2011).

2.1.4. MANGANESE

Manganese is a transitional metal very similar to iron in chemistry and is only second to iron in concentration. It is mostly found in mafic rocks rather than silica rich rocks (Essington, 2004). (Table 2.4)

Table 2.4 General concentration of Mn in different rock types (Heal, 2001)

Rock type	Concentration (mg kg^{-1})
	Igneous
Basalt	1300
Andesite	1200
Gabbro	1200
Diorite	1400
	Metamorphic
Gneiss	600
Granulite	800
	Sedimentary
Greywacke	700
Sandstone	170
Shale	600
Limestone	550

Manganese oxides are particularly found in redoximorphic conditions (e.g. Essington, 2004). Mn is generally very mobile in soils and is found in three ionic forms Mn^{2+}, Mn^{3+} and Mn^{4+} of which the divalent form is the most common and is stable in reducing environments (Sumner, 2000). Chemistry is the most important control on Mn mobilisation with Mn^{2+} and Mn^{4+} being the most common forms in soil (Heal, 2001). There is also a decrease in solubility as with iron in an increase with pH but to a lesser extent (Sumner, 2000). The relative concentration of complex Mn ions is determined by pH (Lindsay, 1979). The Oxidation on Mn can be abiotic or microbial as illustrated in Figure 2.2.

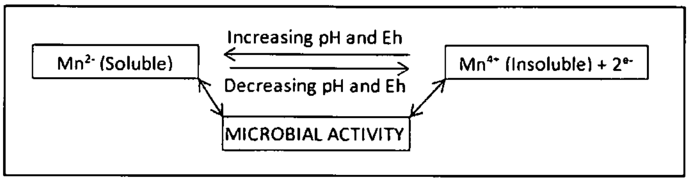

Figure 2.2 Representation of Mn redox reaction (Heal, 2001).

2.1.5. CARBONATES

Distribution of carbonate associated cations can be used as a paleoclimatic and paleoecologic indicator (Cerling, 1984). Their distribution is related to leaching resulting in more abundance in dry climates and less in humid climates.

Lin *et al.* (2005) found an increase to mean depth to $CaCO_3$ layer which correlated with an increase in the precipitation and leaching gradient. They found that the depth to the $CaCO_3$ was less on the summit and backslope than in the depression in the drier area as is seen in Figure 2.3. There was a negative correlation between pH and depth to $CaCO_3$.The opposite is true of the wetter region were the depth to $CaCO_3$ was found to be the greatest on the summit and backslope and shallower in the depression. PiPujol & Buurman (1997) suggests that calcium carbonate, like iron, reacts to water regimes.

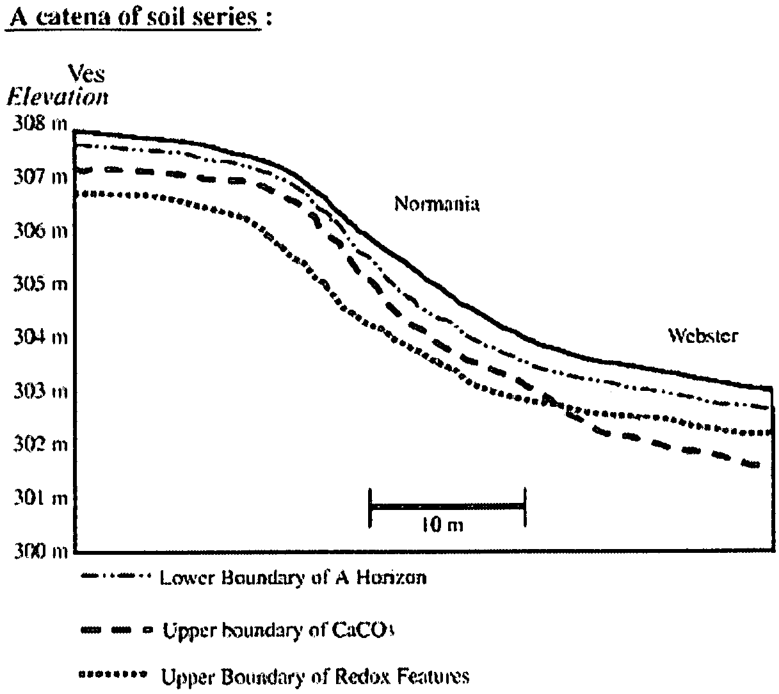

Figure 2.3 Depth to CaCO3 layer in dry climate (from Lin et al., 2005).

Decalcification is defined by Buol *et al.* (1980) as the eluviation of specifically carbonates and an associated process calcification as the accumulation of carbonates. Calcium and magnesium are the primary minerals that bind with carbonates in soil. Common reaction by which $CaCO_3$ is formed in soil can be described by the following equation:

$$Ca^{2+} + CO_2 + H_2O \rightarrow CaCO_3 + H_2 \qquad [Eq\ 4]$$

2.1.5.1. Calcium and Magnesium

The chemical behaviour of calcium and magnesium in soil is very similar and can be discussed together (Sumner, 2000). Ca is most abundant in calcareous sedimentary rocks, than in basic igneous rocks and also found in acidic igneous rocks (Jenny, 1941). Mg is mostly found in basic igneous rocks than acidic rocks and least of all in sedimentary rocks (Metson, 1974). Table 2.5 shows different concentrations of Ca and Mg in different soils and climates.

Table 2.5 Ca and Mg in soil (Brady, 1974)

	Ca (g kg^{-1})	Mg (g kg^{-1})
Temperate soils	0.7 - 36	1.2 - 15
Humid region soils	4	3
Arid region soils	10	6
Peat soils	1.1- 48.3	

Essington (2004) states that the accumulation of Ca is due to poor drainage therefore a lack of leaching is present were carbonates accumulate. The removal of calcium from soil reduces the base saturation and pH. The removal of calcium can be either from leaching or uptake of plants (Brady and Weil, 1996)

Hazelton & Murphy (2007) give a general indication of the different levels of Ca and Mg in soils (Table 2.6)

Table 2.6 Different concentrations of Ca and Mg in soil (Metson, 1961)

Cation	Very low	Low	Moderate	High	Very high
Ca (cmol(+) kg^{-1})	0-2	2-5	5-10	10-20	>20
Mg(cmol(+) kg^{-1})	0-0.3	0.3-1	1-3	3-8	>8

2.1.6. EFFECT OF HYDROLOGY ON SOIL CHEMICAL PROPERTIES

Smith & Van Huyssteen (2011) could not find significant correlation between pH and increased degree and duration of saturation but found fluctuations of pH during saturation (Figure2.3). Phillips and Greenway (1998) found that oxidation of previously reduced soils did not have a significant effect on the pH.

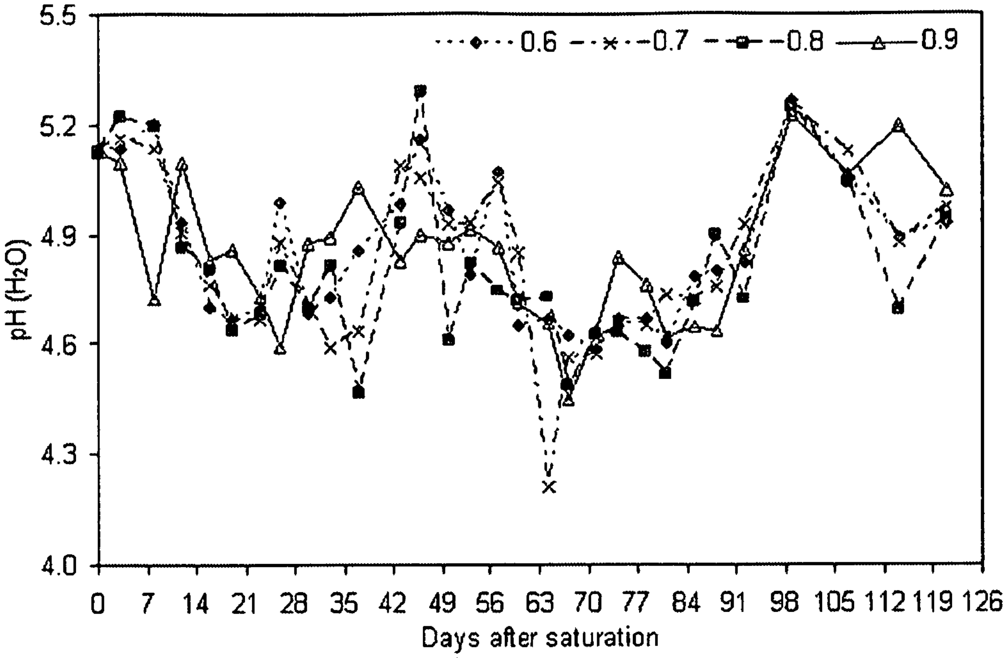

Figure 2.4 pH for different degrees of saturation (0.6=60%, 0.7= 70%, 0.8=80% and 0.9=90%) (Smith & Van Huyssteen, 2011).

Base cations in the soil solution increased with an increase in saturation due to the competition for negative exchange sites with the increased Fe^{2+} and Mn^{2+} caused by reduction (Larson *et al.*, 1991). Therefore losses of cations can occur when soils are saturated and the solution drains. (Phillips and Greenway, 1998) also found increased amounts of Ca, Mg, K and Na in waterlogged soil solution attributed to increased solubility of organic carbon and elevated Fe and Mn levels. Reoxidation of waterlogged soil does not have an effect on the base cations.

An increase in degree of saturation will cause an increase in the mean Fe^{2+} and Mn^{2+}in soil and is represented in Figure 2.5 & 2.6 (Smith & Van Huyssteen, 2011). Increased saturation will result in more Fe and Mn being reduced and potentially be removed from the profile.

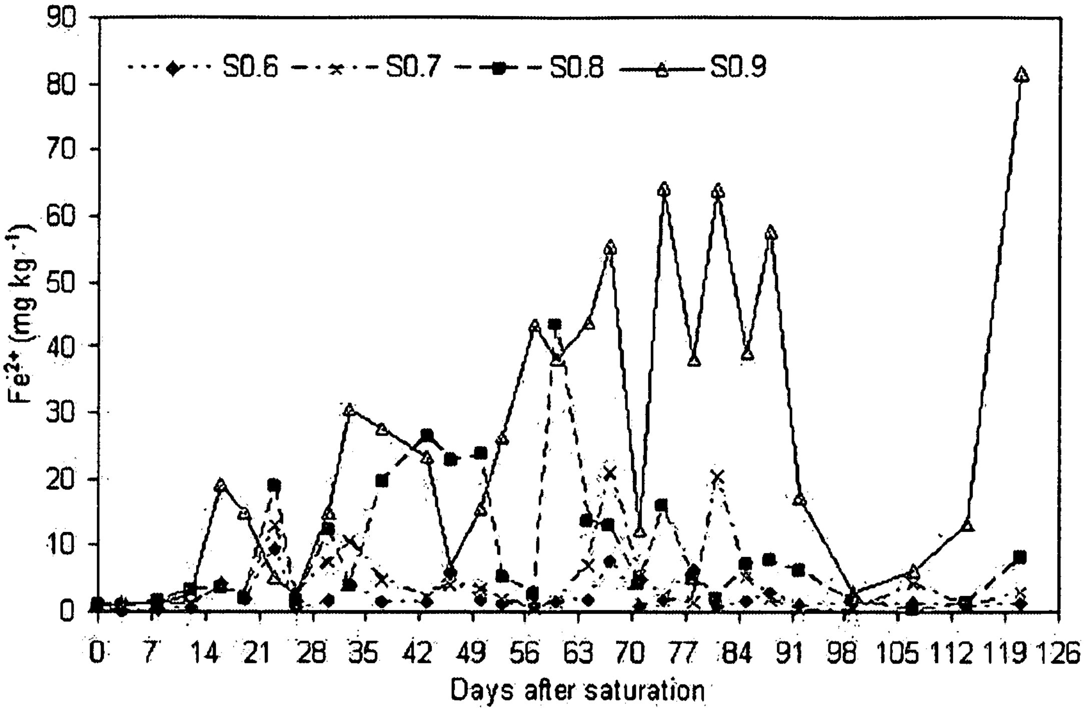

Figure 2.5 Fe^{2+} concentrations for different degrees of saturation (0.6=60%, 0.7= 70%, 0.8=80% and 0.9=90%) (Smith & Van Huyssteen, 2011).

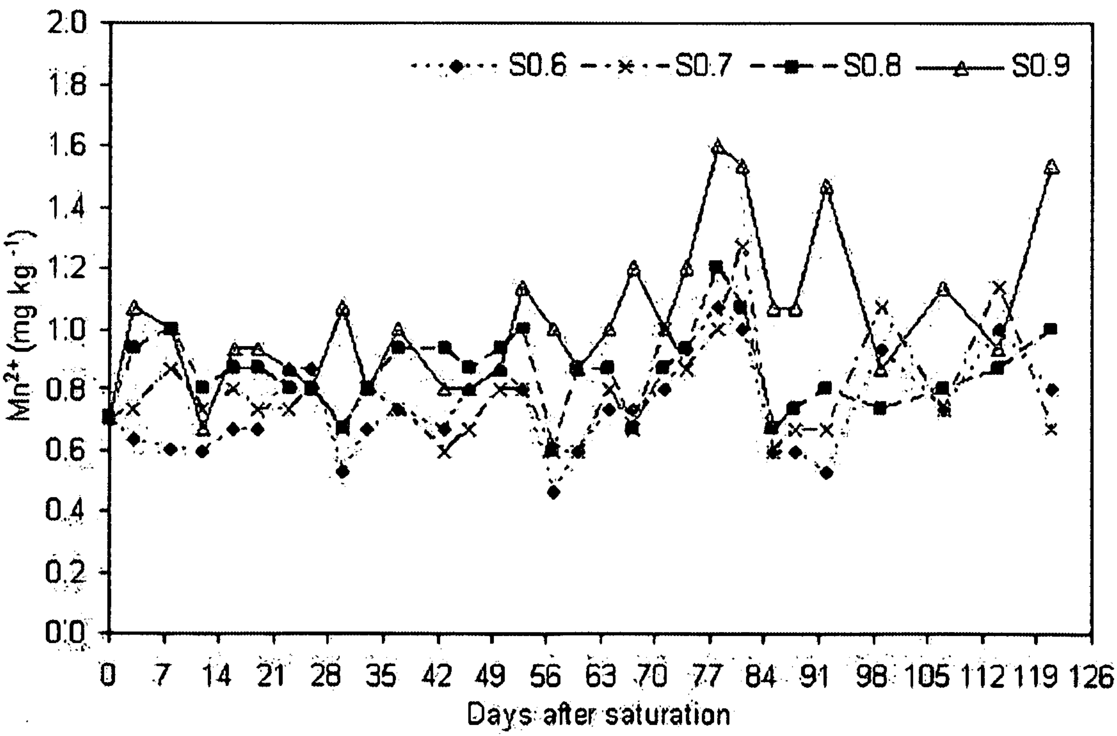

Figure 2.6 Mn^{2+} concentrations for different degrees of saturation (0.6=60%, 0.7= 70%, 0.8=80% and 0.9=90%) (Smith & Van Huyssteen, 2011).

Fe^{2+} and Mn^{2+} are soluble and can be removed if the soils are saturated and Fe and Mn are reduced. Note that during a redox potential of between 200Mv and 300Mv (Table 2.3) both Fe and Mn will be reduced. Sumner (2000) states that the reduction of Fe and Mn has direct impacts on soil chemistry, 1- increased water soluble Fe^{2+} and Mn^{2+}, 2- increase in pH, 3- breakdown of organic matter and nutrient release, 4- formation of new minerals (Gambrell, 1994), 5- displacement of cations from soil complex to soilwater and 6-increased solubility of P and Si. Ticehurst *et al.* (2007) affirms that the redistribution of soluble compounds reflects the movement of water.

Fe^{2+} redistributed either lower in the soil profile or even found in the stream completely out of the soil (Abesser *et al.*, 2006). Reuter & Bell (2003) revealed that the Fe decreased from summit to wetland and attributed the Fe loss to increases in water saturation and therefore reduction of Fe. Heal (2001) and Abesser *et al.* (2006) found that in a big riparian zone that Fe and Mn were being flushed into the streams during storm events that were dominated by soilwater. Therefore Fe and Mn were being leached out of the soil in saturated conditions. Phillips and Greenway (1998) found that reoxidation did not have an effect on exchangeable Fe and Mn.

Fe and Mn are also responsible for the colour in the soil (Bohn *et al.,* 1985). The soils where Fe and Mn have been removed were very gleyed and lacked the red and orange colours associated with these elements. Van Huyssteen *et al.* (2005) states that the reduction and redistribution of these minerals can be associated with the water regime. Smith & Van Huyssteen (2011) found a high positive correlation between Mn^{2+} and Fe^{2+} concentration increase and increased degree of water saturation.

Lin *et al.* (2005) suggests that local topography is very significant in the formation of $CaCO_3$ and is also found where local variations occur within climatic regions (i.e. implying $CaCO_3$ formation is subject to the presence of hydrological processes). The difference in the two regions is due to more leaching in the depressions of drier areas where water accumulates and that there is water saturation and watelogging in wetter areas where $CaCO_3$ can accumulate.

2.2. HYDROLOGICAL PROCESSES

2.2.1. PATHWAYS

Hydrological pathways are important as they determine how and when precipitation reaches the stream. Figure 2.7 shows the general processes involved in the hydrological cycle and all processes shown are relevant to hydropedology. McDonell (2003), entitles an article "where does the water go

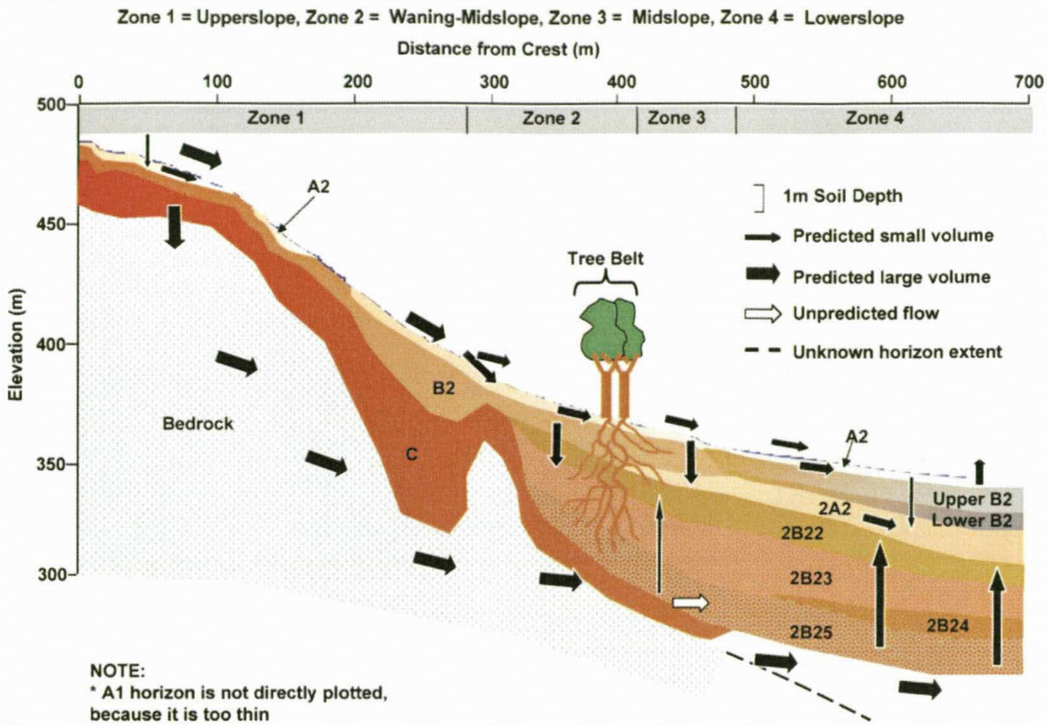

Figure 2.8 Hydrological response model of a hillslope in Australia (Ticehurst *et al.*, 2007).

Overland flow can occur as saturation excess or infiltration excess. Where plants and shrubs are spars overland flow is more likely than where there is dense vegetation (Kirby, 1978). Ticehurst *et al.* (2007) found both types of overland flow. Saturation excess occurred when a low intensity storm occurred with the soil having very high antecedent moisture content and infiltration excess occurs during a high intensity storm with low antecedent moisture content. Two important factors considering overland flow is antecedent moisture content and rainfall intensity.

Topography is also a factor controlling overland flow. Ticehurst *et al.* (2007) found that where the gradient is less steep there is more infiltration and therefore less overland flow. Most overland flow was found at the break of slope.

Three dominant conditions are necessary for SLF to occur. Firstly a development of saturation at the soil-bedrock interface is necessary (McGlynn *et al.,* 2002). Secondly, a rise in the water table from material with low hydraulic conductivity into a material with a higher hydraulic conductivity (Kendall *et al.,* 1999). A deflection of water moving vertically by impeding layers to move horizontally (Ticehurst *et al.*, 2007).

Ticehurst *et al.* (2007) identified four SLF as can be seen in Figure 2.8. Lin *et al.* (2006) also had similar findings and are represented in Figure 2.9, the different moisture contents of the profiles in different position on the hillslope are also shown in Figure 2.9. The flowpaths are subsurface when it rains" and this is a rather simple answer given in Figure 2.7 in the different pathways that may be present in soils.

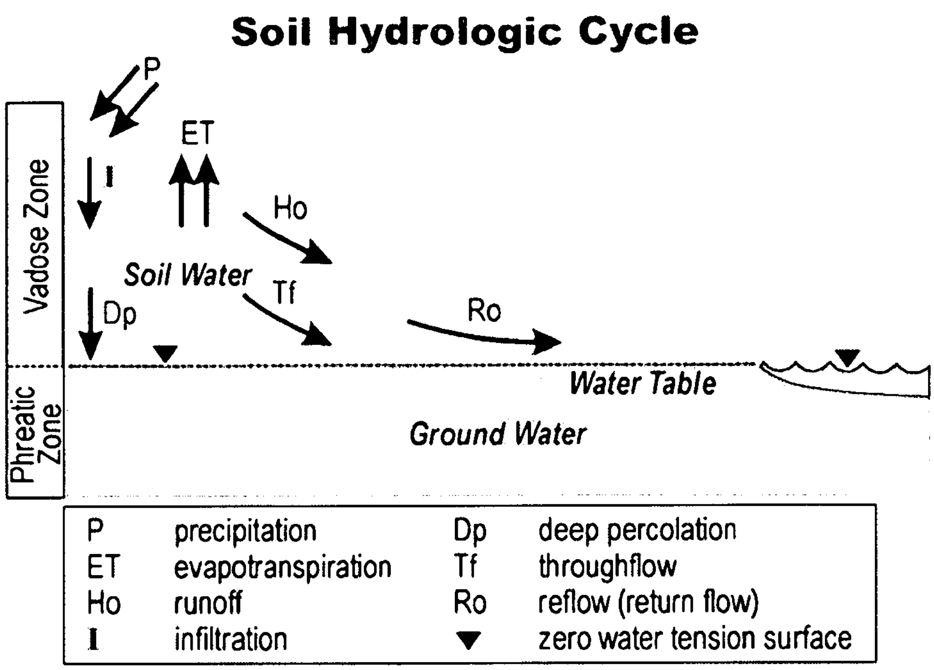

Figure 2.7 Soil hydrologic cycle (from Schoeneberger & Wysocki, 2005).

Ticehurst *et al.* (2007) indentifies three lateral flow paths as are shown in Figure 2.8. The pathways are named as: overland flow, subsurface lateral flow (SLF) and bedrock interflow pathways that are present in a soilscape. These pathways can be further subdivided. The flowpaths are more dominant on different parts of the hillslope and they are not mutually exclusive from each other, the different pathways are also dominant at different saturation periods (Ticehurst *et al.*, 2007). Ticehurst *et al.* (2007) also found that a slope of 10% to 15% was enough for SLF to occur which is lateral flow within the solum of the soil (Figure 2.8).

seepage trough macropore networks in subsoil, lateral flow through the A/B interface, return flow at the footslope and toeslope and flow on the soil-bedrock interface.

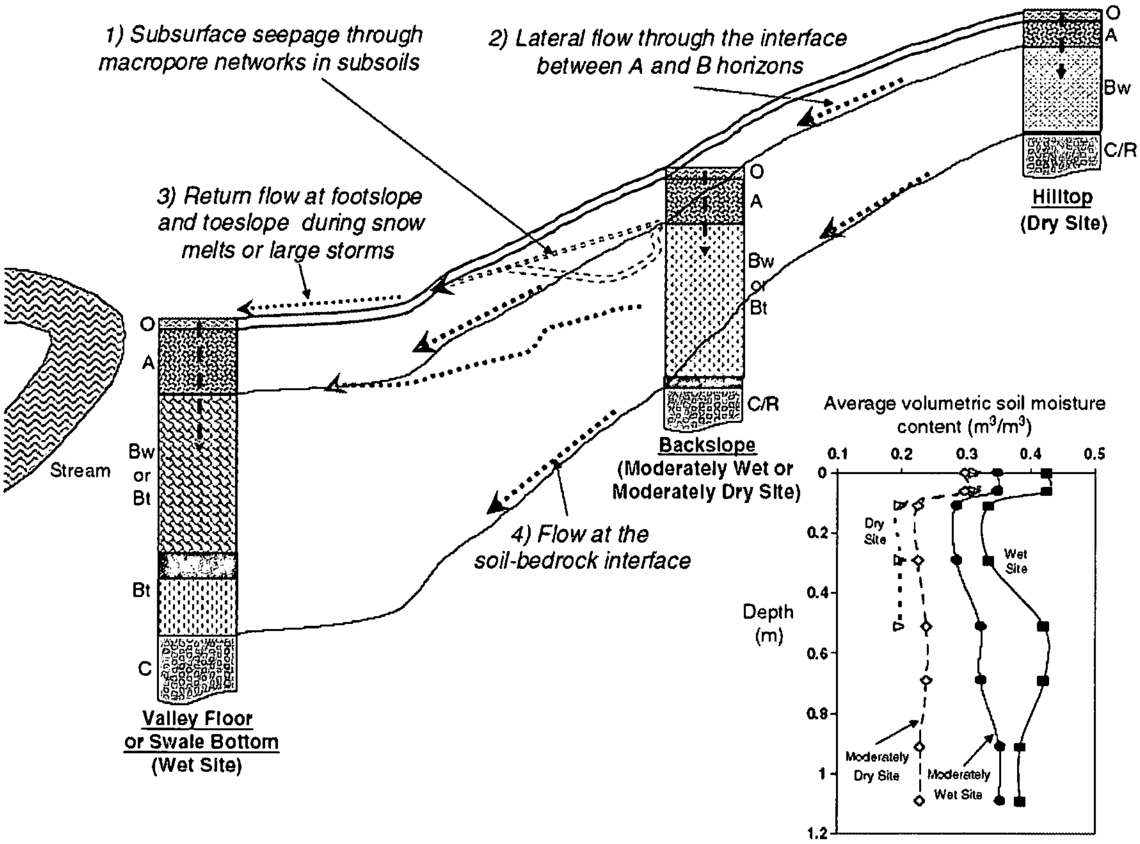

Figure 2.9 Hydrological response of a conceptual hillslope and soil moisture content of three profiles in the hillslope (from Lin *et al.* 2006).

Macropores are defined as having a diameter greater than 75 μm and can be caused by root channels or animal activity. The macropores can be interconnected and respond differently at different antecedent moisture content. Lin *et al.* (2006) found that some subsoils responded very quickly to rainfall events and attributed this to the macropore flow conducting water quickly to the horizons. Soils dominated by 2:1 clay minerals will shrink in dry periods and therefore crack. These cracks are able to conduct water very quickly until the soils starts to swell and the cracks disappear. These soils will conduct water slowly during wet periods and quickly during dry periods.

Ticehurst *et al.* (2007) and Lin *et al.* (2006) found movement of water within the soil matrix. Lin *et al.* (2006) found it to be more specifically on the interface between the A and B horizons. This was due to different structures, density and hydraulic conductivity of the horizons. This was visible and the B horizon had very low soil moisture content.

Ticehurst *et al.* (2007) found return flow at the toe of the hillslope which correlates with Lin *et al.* (2006). In both cases the return flow is diverted to overland flow due to saturation excess. It then follows the paths described for overland flow.

Tromp-van Meerveld and McDonnell (2006b) show how subsurface topography can control the flow of water in a hillslope. The pathway on top of the bedrock found in a few studies (Lin *et al.*, 2006, Ticehurst *et al.*, 2007 and Asano *et al.*, 2002) is due to the same reason as the flow at the A and B interface, the rock has different hydraulic properties to soil. This would mean that there was no lateral flow within the soil as water would have to drain vertically onto the bedrock and conditions necessary for flow would have to occur. Tromp-van Meerveld and McDonnell (2006a) found in their study that there is a threshold value in storm size that increases the stormflow in the soil. They found that if a storm event of more than 55mm occurred the stormflow in the soil would increase more than fivefold. Tromp-van Meerveld and McDonnell (2006b) ascribe this to their "fill and spill" hypothesis which is that depressions in the hillslope first need to be filled before water will run out of it. This hypothesis is shown in Figure 2.10, here the rise in water level in a hollow over time is shown in a hillslope for two different storms.

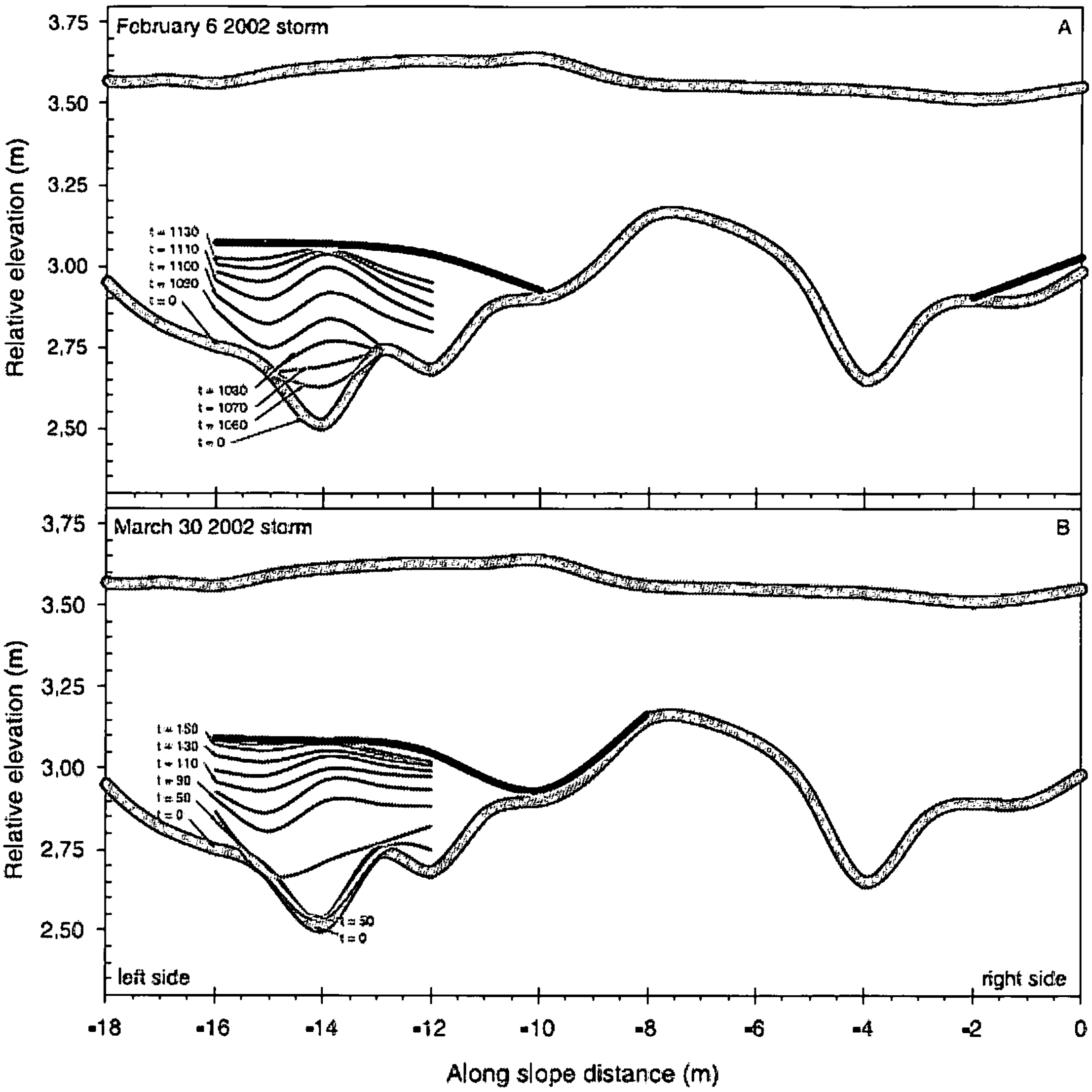

Figure 2.10 Graph showing "fill and spill" hypothesis (Tromp-van Meerveld and McDonnell 2006b).

If there are cracks in the bedrock a flowpath through the rock to bottom of the slope is found by Ticehurst *et al.* (2007) if the macropores are connected and this is seen in Figure 2.8, also found by Asano *et al.* (2002).

2.2.1.1. Resident times

Resident time implies the time that water will stay in a specific system until it leaves the system into another system. Another word synonymous with resident time is transit time which is more specifically the time taken from when water enters a system to when it exist the system.

Resident times can reveal a lot about the storage mechanism, pathways and the source of the water (McGuire & McDonnell, 2006). Most chemical reactions are dependent on the environment and time, hence the importance of residence time of water in a hillslope as this will determine the environment and the time for a chemical reaction.

Asano *et al.* (2002) found that soil depth also effects the resident time and is shown in Figure 2.11. The vertical drainage was found to be more important factor than the lateral drainage.

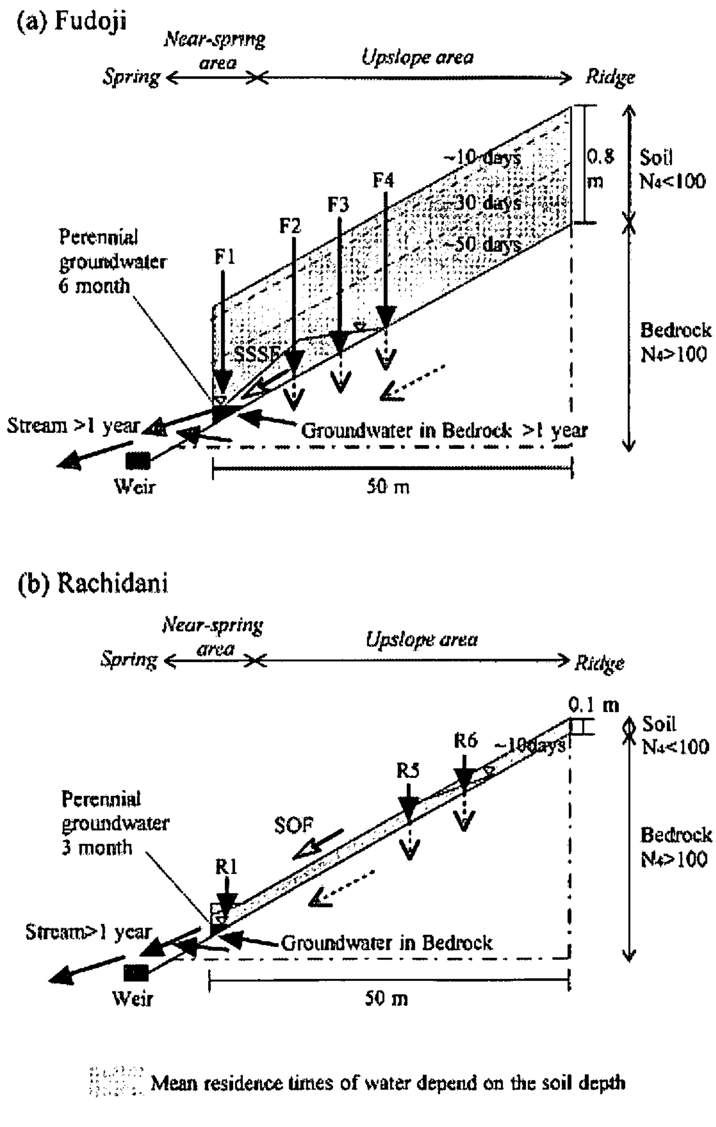

Figure 2.11 Conceptual model illustrating the influence of soil depth on resident time in two catchments in Japan (Asano *et al.*, 2002).

McGuire *et al.* (2005) found that residence time didn't correlate to catchment area but rather the internal form and structure of a catchment. It was found that residence time was correlated to simple topographic features such as median flow length and gradient of slopes.

2.2.2. SOIL MORPHOLOGY

Lin *et al.* (2006) states that topography plays an important role during wet periods but that soil characteristics play a dominate role in the drier periods.

Ticehurst *et al.* (2007) found that soil colour was the most useful soil indicator of flowpaths in the soil and that morphology was the reflection of the most dominant hydrological processes. Red colour found in soil due to the presence of hematite (Fe_2O_3) is an indication a well drained soil as the Fe is not reduced. Soils that are more poorly drained than red soils have a yellow colour dominating due to the presence of goethite (FeOOH). In very poorly drained soils the Fe is reduced and removed from the soil resulting in grey, low chroma colours dominating (van Huyssteen *et al.,* 2005).

Redoximorphism is the reduction of Fe (III) contained in hematite and goethite, as discussed in previous sections. Ticehurst *et al.* (2007) also found that Fe and Mn concretions and mottles were a sign of periodic saturation and that an increase in size and abundance corresponded to a longer period of saturation. These features indicate the presence of stagnant water in the landscape. D'Amore *et al.* (2004) found that the amount of mottles did not indicate duration of saturation as much as a fluctuating watertable.

Buol *et al.* (1980) defines eluviation as the movement of material out of a portion of a soil profile and the process whereby the material accumulates as illuviation. Ticehurst *et al.* (2007) uses an increase in clay content downslope to indicate subsurface lateral flow down the hillslope.

2.2.3. HYDROLOGICAL SOIL TYPES

Three main soil types are found in literature regarding their hydrological response (Asano *et al.*, 2002; Soulsby *et al.*, 2006; Ticehurst *et al.*, 2007) and are defined by Van Tol *et al.* (2011) as recharge, interflow and responsive.

Red soil horizons with a massive structure and coarse texture are soils, which rapidly transmit water (Fritsch and Fitzpatrick, 1994). These soils are termed recharge soils, as the water will move through the soil and into saprolite were it can recharge the groundwater or flow through fractured rock and feed soils at the bottom of the hillslope. Hutton or Glenrosa soil forms are examples of recharge soils.

Interflow soils are soils were water infiltrates the soil and then is diverted laterally down the hillslope. This diversion could be caused by a difference is hydraulic properties of the soil. These soils are also known as throughflow soils. A distinction can be made as to the depth of the horizontal movement of water. Shallow interflow is expected in the E horizon (*gs*) close to the surface due to an impeding underlying horizon caused by the presence of higher clay content or long periods of water saturation (Kroonstad or Cartref soil). Deep interflow occurs in soils which have a B horizon which promotes vertical flow but due to reduced permeability water ponds in the C horizon and laterally flow occurs in the C horizon (Bloemdal or Avalon soil).

Responsive soils are soils that are saturated quickly and therefore produce overland flow. Shallow soils are classified as responsive due to the low waterholding capacity and therefore saturate quickly (Mispah soil). Soils that have a low K_{sat} or are saturated for long periods are also responsive as they will cause overland flow due to saturation excess (Katsptuit).

Figure 2.12 is a representation of the different hydrological soil types and the correlation with position in the landscape. Lin *et al.* (2006) found that wet soils were found at the bottom of the hillslope were water accumulated.

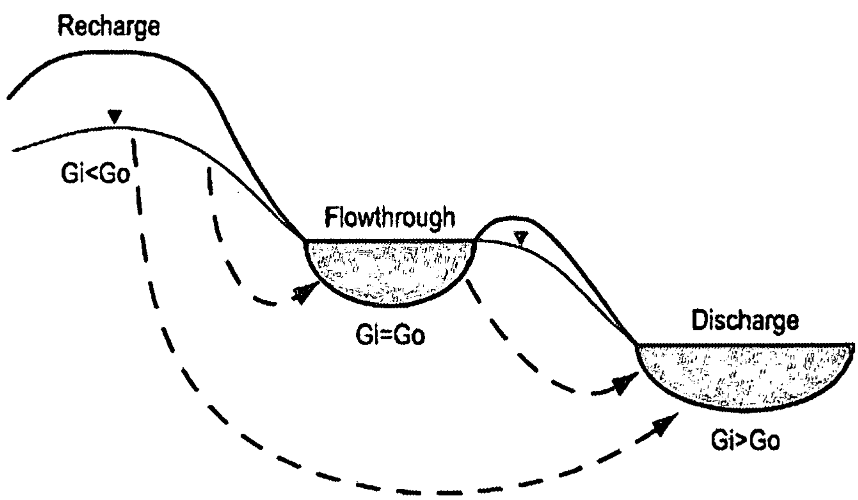

Figure 2.12 Position of different hydrological soil types in a hillslope (from Vepraskas *et al.*, 2006).

2.3. WEATHERLEY CATCHMENT

2.3.1. SITE DESCRIPTION

The Weatherley catchment is situated south West of Maclear in the Eastern Cape (Figure 2.13). The catchment is 160ha (Lorentz 2001) and forms part of the Mooi River catchment, which is a quaternary catchment of the Umzimvubu basin. It is at an altitude of approximately 1300m above sea level.

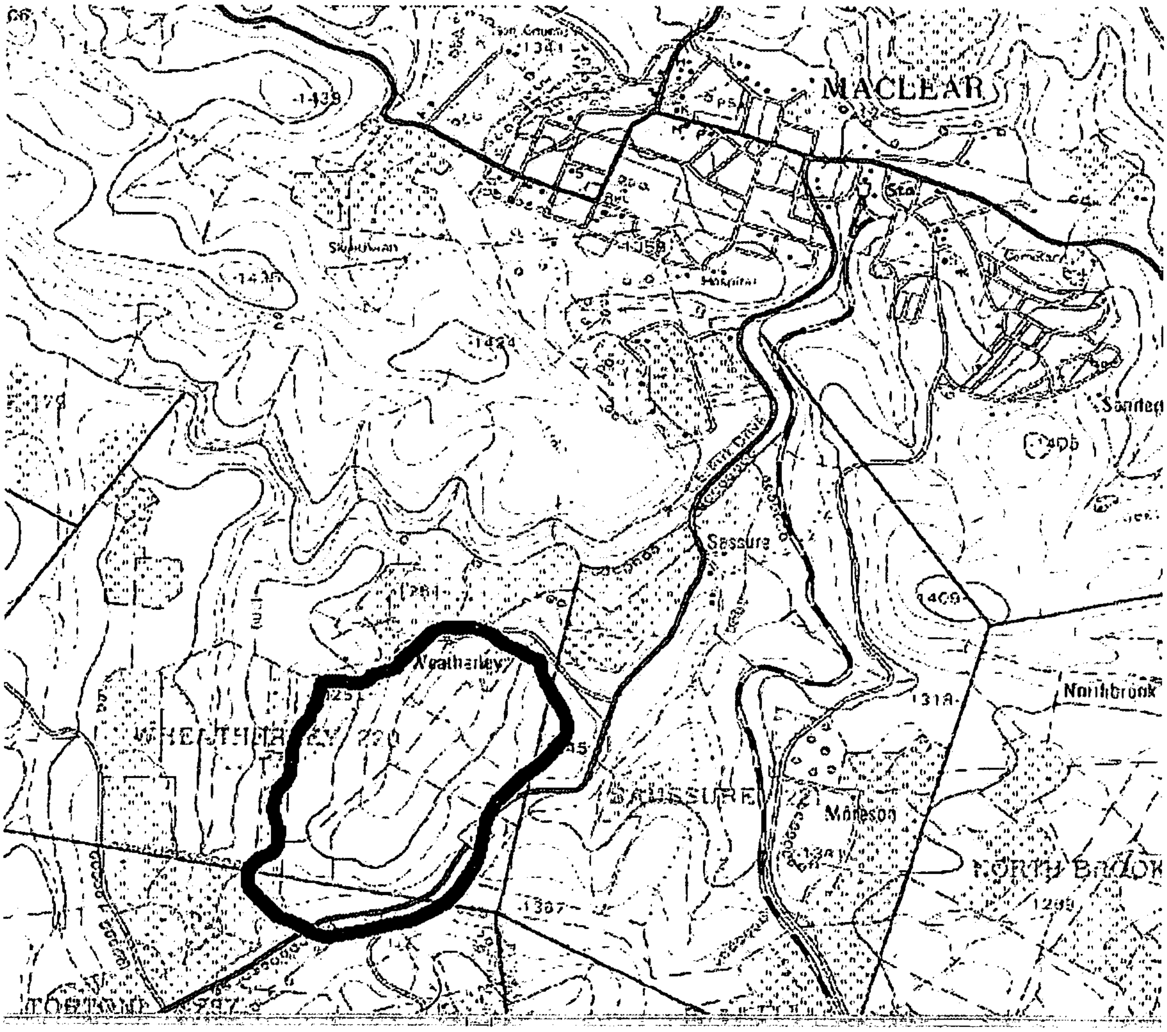

Figure 2.13 Location of the Weatherley catchment.

2.3.2. CLIMATE

The average daily temperature ranges from 11^0C in winter to 20^0C in summer, winters are very cold with the minimum daily average of 4^0 C, frost and snow are common. The catchment has a mean annual precipitation (MAP) of 1000mm and a mean annual evaporation (MAE) of 1488 (BEEH, 2003).

2.3.3. TOPOGRAPHY AND LAND USE

The catchment drains in a northerly direction towards the Mooi River (Van Tol, 2008). The eastern and southern slope have an average of 12% and the steeper western slope has an 18% slope (Lorentz *et al.*, 2004). The catchment is naturally covered by Highveld sour grass (Acock's, 1975) with a basal cover of 50 - 75% on the hillslopes (Esprey, 1997). An area of 5ha on the western slope previously used as agricultural lands (Lorentz, 2001) and there has been some afforestation on the slopes since 2002 (Van Tol *et al.*, 2010). The catchment topography can be seen in Figure 2.14.

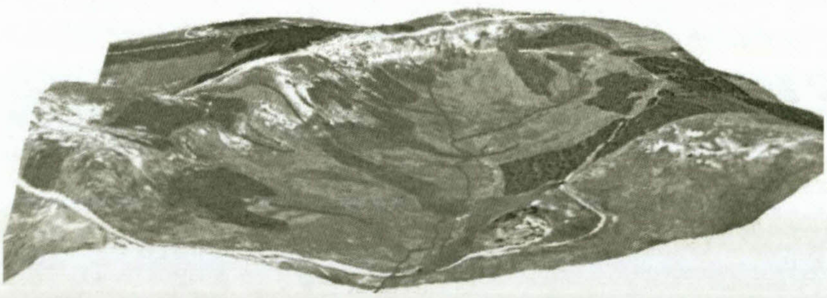

Figure 2.14 3D view of the Weatherley catchment.

2.3.4. GEOLOGY

The catchment consists of mudstone and sandstone of the Elliot formation above 1 320 m a.m.s.l. and Molteno formation below this altitude (De Decker, 1981). The Molteno formation forms a prominent shelf, which plays a major role in the hydrology of the catchment (Van Tol, 2008). There are two dolerite dykes in the catchment which can be seen in Figure 2.15.

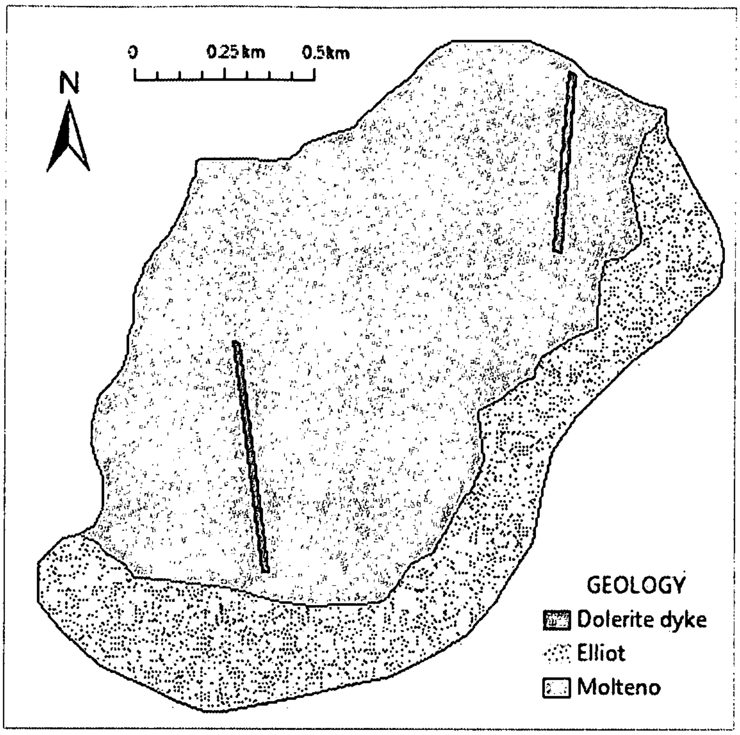

Figure 2.15 Geology of the Weatherley catchment (from De Decker, 1981).

2.3.5. SOILS

The soils were classified by Roberts *et al.* (1996) using the Soil Classification- A Taxonomic system for South Africa (Soil Classification Working Group, 1991) and Figure 2.16 is the soil map produced. Due to the high rainfall the soils are leached causing a low pH in the soils that are freely drained. The freely drained soils consisting of the Hutton soils are found on the terrain morphological unit (TMU) 1 and 2 positions. The TMU 3 positions are consist interflow soils (Bloemdal, Tukulu, Pinedene and Sepane) but due to the high rainfall responsive soils are also present on this position (Katspruit). The presence of flucuating watertables in the catchment also allow the formation of plinthic soils (Avelon, Westleigh and Longlands).

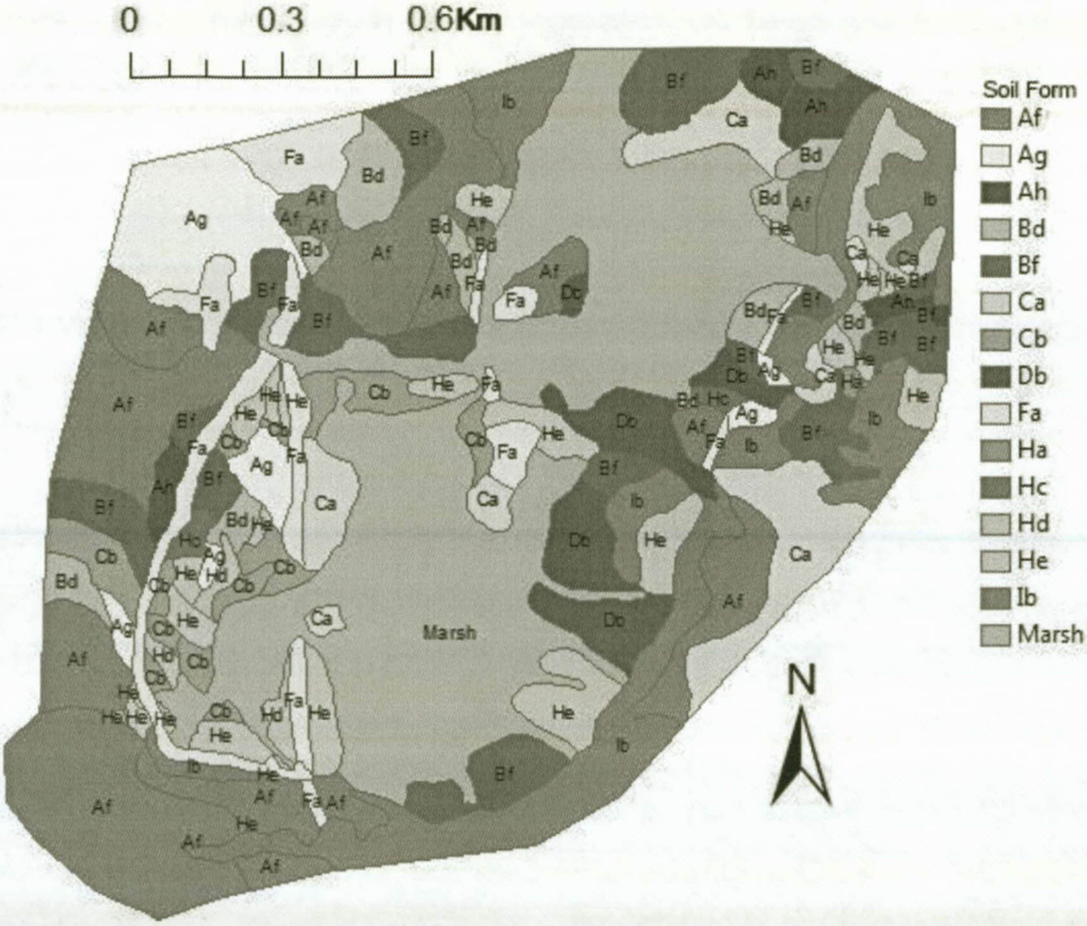

Figure 2.16 Soil map of Weatherley (from Roberts *et al.*, 1996).

2.4. SUMMARY OF PREVIOUS STUDIES ON WEATHERLEY CATCHMENT

The many studies done in the Weatherley catchment have all contributed to the understanding of the hydrological nature of the catchment and complement and enhance the development of conceptual models and ideas. Figure 2.17 is a map of Weatherley and shows the location of the nests, which consist of Piezometer and tensiometers. Other measuring apparatuses are also shown like neutron water meter access tubes and crump weirs. The data collected at these sites have contributed to many of studies conducted in the catchment.

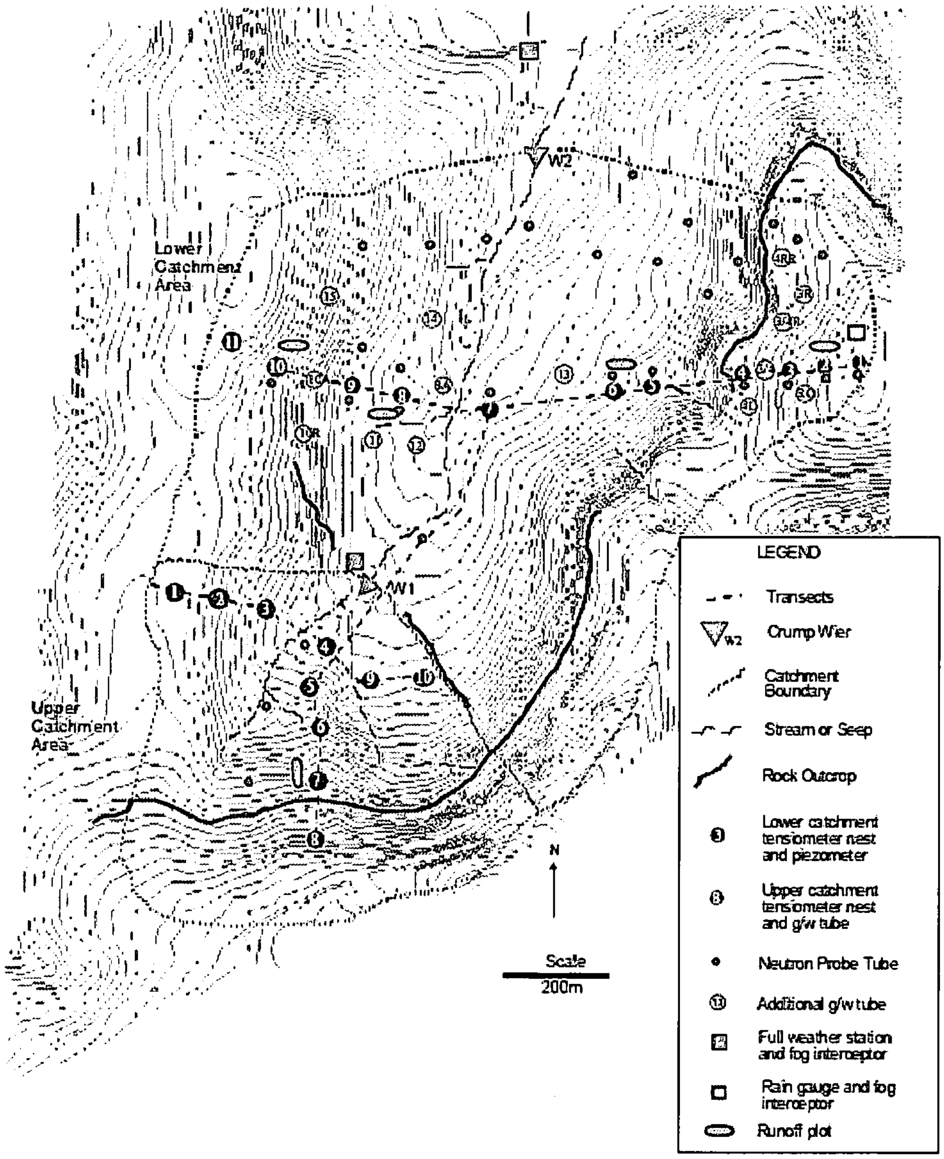

Figure 2.17 Instrumentation network of Weatherley (Lorentz *et al.*, 2004).

Esprey (1997) study focused on the transect 1 which incorporates nest 1 to nest 10 but more specifically on nest 1 to nest 4. Concluded that luviation of clay was present both down the profile and down the hillslope. A low bulk density indicated the presence of macropores on the crest and the toe slope. There is a decrease in hydraulic conductivity with depth, however there are variations within certain profiles on the midslope that have deep well drained soil. Poorly drained soils are found on the crest and toe slope. It was concluded that water drains vertically through the profile till it reaches the bedrock, were it forms a water table that begins to move laterally through the hillslope.

Lorentz (2001) findings are schematically represented by Figure 2.18 and the different flow paths and mechanisms can be seen. The study concluded that in the hillslopes above the Molteno shelf, rapid near surface macropore flow in larger rain events and that perched watertables would occur during these events that would flow downslope and dry up. There is a rapid increase in groundwater levels after rain events indicating fissures that transmit the water. In the upper slope there is a continuous accumulation at the toe even long after a rainfall event. It is found that long duration high volume rainfalls contribute more to the groundwater than high intensity rainfall events.

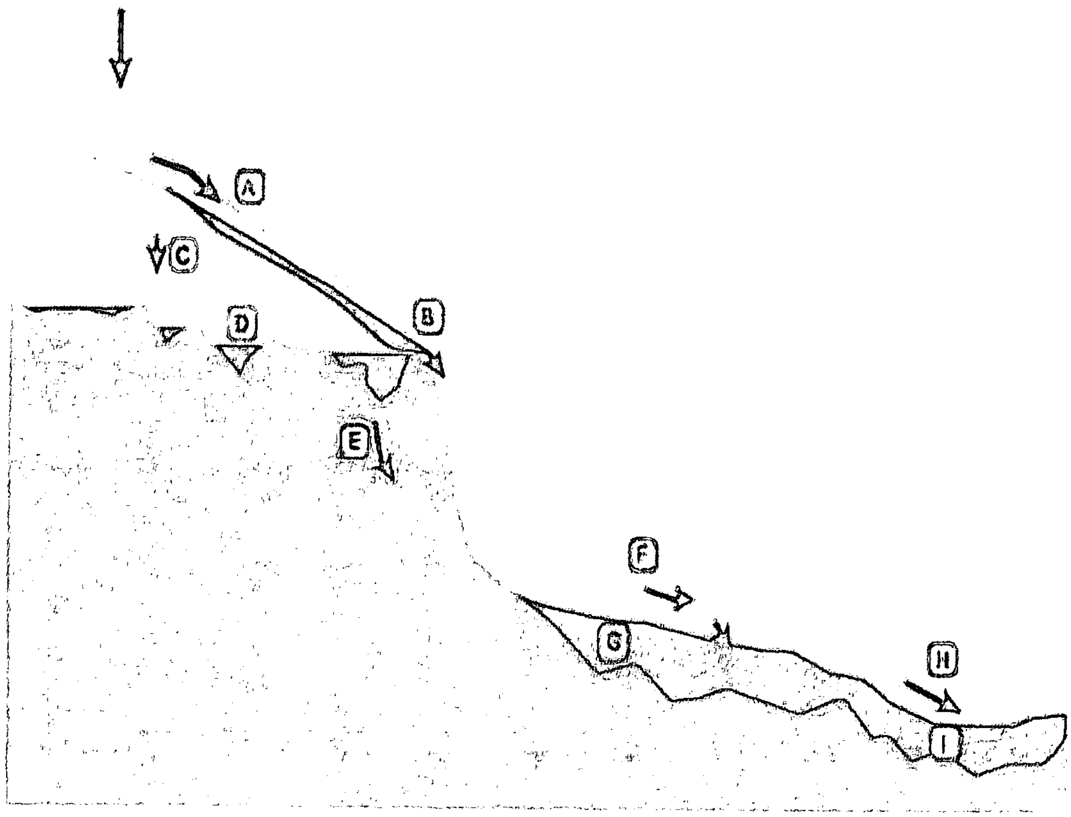

Figure 2.18 Conceptual model of flow mechanism in the Weatherley catchment (Lorentz, 2001).

Lorentz *et al.* (2004) quantified the three main streamflow processes. These are perched groundwater response, surface runoff-response and near-surface macropore flow. It was found that the perched watertable contributed steadily and constantly (0.1cu.m/s) for long durations (1 hour) which is a sharp contrast to the surface runoff-response which contributed 0.9cu.m/s in 2.5 minutes and stopped contributing. The near-surface macropore response was the intermediate response which peak at 0.9cu.m/s but tapered off to 0.1 in 15 minutes. This all occurred in a rain event of 31.4mm with a duration of 20 minutes.

Uhlenbrook *et al.* (2005) found perched watertable above the bedrock that seeps out the toe of the slope but also recharges the sandstone aquifer below the watertable through fissures. This was validated by soil moisture contents and the use of electrical resistivity tomography (ERT). The results from the ERT can be seen in Figure 2.19, it was found that ERT readings were useful to gain further insights into the hydrological processes when correlated with other measurements. The study was done on nests 1 to 4.

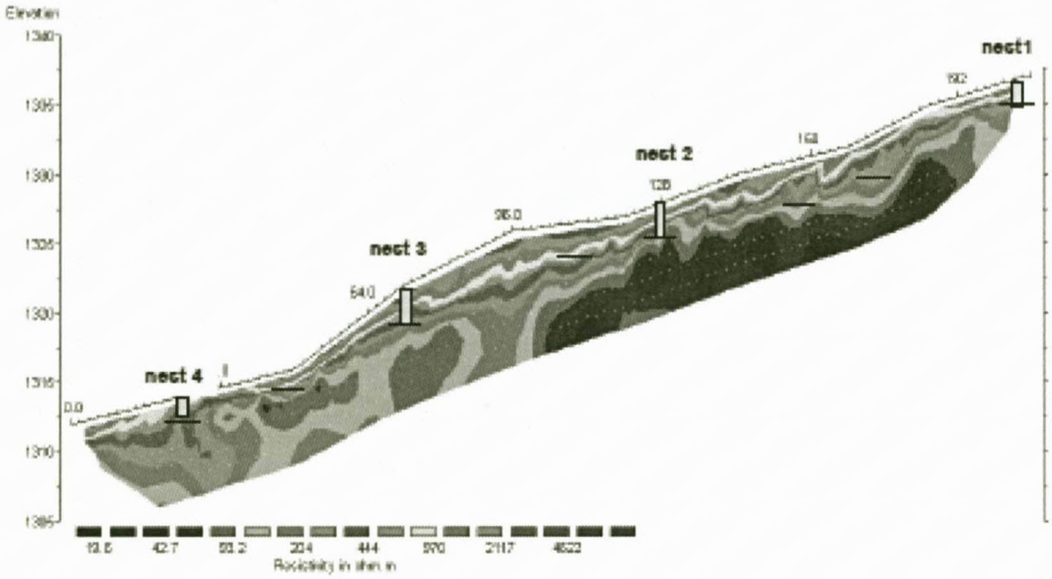

Figure 2.19 Results of the 2D electrical imaging using ERT from nest 1 to nest 4 (Uhlenbrook *et al.*, 2005).

Van Huyssteen *et al.* (2005) found that *ot* horizon could be subdivided using the nature of the underlying material. These classes are well drained soils, moderately drained soils and poorly drained soils. The well drained soils consisted of red apedal B, yellow-brown apedal B and neocutanic horizons. Moderately drained soils had few to many mottles and was found overlying a *gs* or *gh* horizon. Poorly drained soils similarly overly a *gh* or *gs* horizon and soft plinthic but have many mottles.

A mean for selected properties is found in Table 2.7. The properties are group in soil horizons.

Table 2.7 Means of selected soil properties group in diagnostic horizons in Weatherley (Van Huyssteen *et al.*, 2005)

Hor	Clay	S	CEC_{soil}	CEC_{clay}	BS	OC	N	pH_{water}	pH_{KCl}	Fe	Mn	Mean $AD_{S>0.7}$
	(%)	($cmol_c\ kg^{-1}$)			(%)		($mg\ kg^{-1}$)			($mg\ kg^{-1}$)		($days\ year^{-1}$)
ne	18.6	2.0	4.6	20.1	45.1	0.35	343	5.22	4.31	9741	155.3	37
ye	13.6	1.2	4.4	28.0	28.3	0.22	187	5.18	3.99	5152	25.2	75
re	21.4	1.6	4.5	16.3	38.5	0.35	259	5.36	4.17	10338	144.6	80
ot	14.8	3.2	6.8	27.0	46.0	0.93	645	5.33	4.36	7586	46.6	148
sp	17.6	3.2	6.0	32.7	52.5	0.19	249	5.79	4.31	6835	64.5	182
gs	13.0	1.9	6.8	44.5	34.3	0.45	374	5.44	4.29	7924	16.8	202
on	23.4	4.3	7.0	28.1	58.2	0.13	208	5.59	4.27	8669	115.4	248
gh	28.2	6.9	10.5	35.3	64.2	0.20	306	6.15	4.44	9227	33.0	331

Figure 2.20 shows the drainage curve at two depths, 250mm and 750mm, at P204 during the period of 1 May 2001 to 12 May 2001. It can be seen that the *ot* horizon drains fairly quickly while the little change in the C horizon is attributed to ET by Van Huyssteen *et al.* (2005).

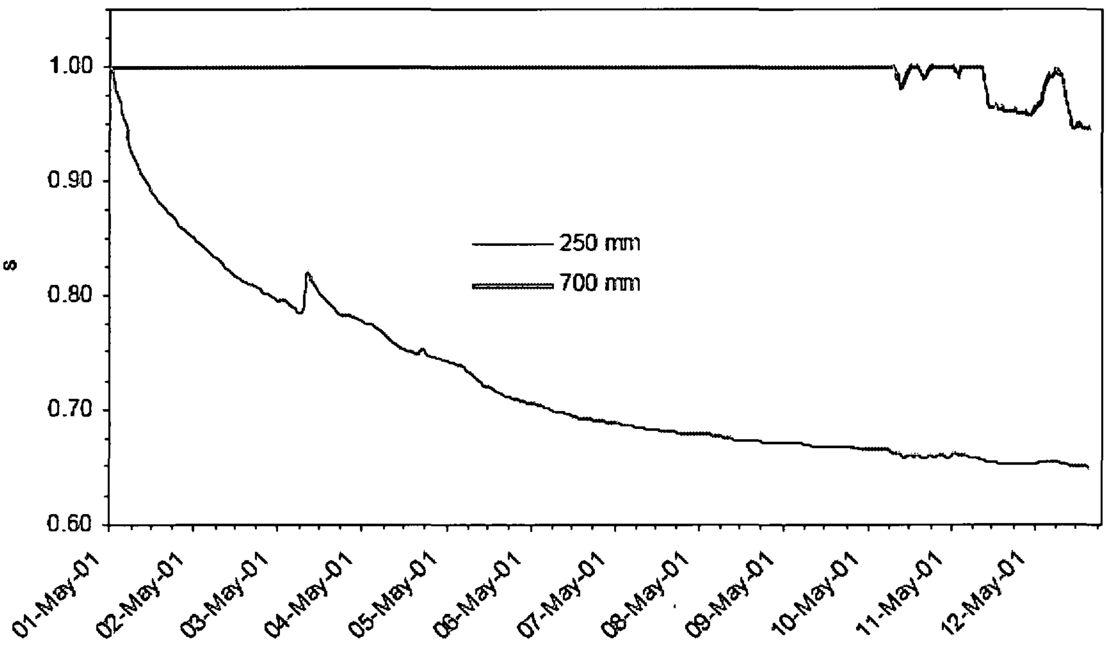

Figure 2.20 Changes in degree of saturation (s) at P204 (Van Huyssteen *et al.*, 2005).

Van Huyssteen *et al.* (2005) calculated the water balance of the Upper Eastern Catchment (*UEC*) of Weatherley and the results are found in Table 2.8. The surface runoff and vertical drainage below the on were considered negligible.

Table 2.8 Water Balance of the Eastern Upper Catchment (Van Huyssteen *et al.*, 2005)

Parameter	Value	
Estimated area of UEC (m^2)		36 530
Width of water conducting zone at P204 (m)		100
	mm	m^3
Total P	104	3 792
Total ET	102	3 741
P-ET		*51*
Water loss out of 0-470 mm depth:		
During constant rate period (38 days)	2 884	288
During decreasing rate period (12 days)	66	7
Total		*295*
Water loss out of 470-770 mm depth:		
During constant rate period (48 days)	2 088	209
During decreasing rate period (2 days)	5	0.5
Total		*209*
Estimated total water loss		*504*
Estimated total outflow (corrected for P and ET)		**453**

Van Huyssteen *et al.* (2005) found that the Molteno shelf, large area of the marsh and the difference in topography of the east and the west soilscapes made it difficult to correlate soil characteristics with hydrological characteristics. The interflow portion of the hydrograph was found to be closely correlated with the distribution pattern of the soils. The apedal soils which constitute 55% of the total area contributed about 70% of the *DWC* in the *UEC*. The deepest soils had the highest *DWC* values.

Van Huyssteen *et al.* (2005) found that that *ot*, *ye* and *gs* on the TMU 3 reacted rapidly to rainfall events while *gs* on TMU 5 remain saturated for long periods because of water being supplied by the higher lying well drained soils.

Le Roux *et al.* (2005) conducted a study on the organic matter and vegetation baseline. The main findings where that the amount and distribution of *OM* was similar in similar hydrological soils with the freely drained soils having the highest amounts of *OM*. A linear decrease in *OM* to 600mm was found in all soils irrespective of the hydrological response.

Van Tol *et al.* (2010) study was based on a hillslope in the upper catchment above W1. A description and class of hydrological soil type of all the soils in the study area can be seen in Table 2.9.

Table 2.9 Description of soils found in study area (Van Tol *et al.*, 2010)

Observation	Terain morpho-logical unit (TMU)	Soil form	Soil family	Horizons	Depth (mm)	Clay (%)	Moist colour	Hydrological behaviour [1]*	Hydrological soil type
P211 Uc4	5	Katspruit (Ka)	Ka1000	orthic A (*ot*) G-horizon (*gh*)	150 1 200	32 38	10YR4/2 10YR5/1	Macropore flow Waterlogged	Responsive
354	5	Ka	Ka1000	*ot* *gh*	350 1 500	25 40	2.5YR4/2 2.5YR6/0	Macropore flow Waterlogged	Responsive
Uc5	5	Ka	Ka1000	*ot* *gh*	400 1 500	15 50	7.5YR4/6 7.5YR5/2	Macropore flow Waterlogged	Responsive
353	5	Ka	Ka1000	*ot* *gh*	300 800	10 40	10YR4/2 10YR5/3	Macropore flow Waterlogged	Responsive
P213	5	Ka	Ka1000	*ot* *gh*	500 1 500	27 30	10YR3/2 10YR4/1	Macropore flow Waterlogged	Responsive
Uc6	4	Kroonstad (Kd)	Kd1000	*ot* E -horizon (*gs*) *gh*	300 1 000 1 500	30 30 55	7.5YR3/2 10YR5/2 10YR6/1	Macropore flow Lateral flow Waterlogged	Interflow
358	3	Tukulu (Tu)	Tu2220	*ot* neocutanic B (*ne*) unspecified material with signs of wetness (*on*)	550 1 450 1 510	18 25 30	10YR4/3 7.5YR4/4 5YR4/4	Vertical drainage Vertical drainage Lateral flow	Interflow
357	3	Tu	Tu2110	*ot* *ne* *on*	400 900 1 510	8 10 10	10YR4/1 10YR5/2 10YR5/3	Vertical drainage Vertical drainage Lateral flow	Interflow
180	3	Tu	Tu2210	*ot* *ne* *on*	400 900 1 500	15 15 20	10YR3/4 5YR5/6 ND [2]*	Vertical drainage Vertical drainage Lateral flow	Interflow
P212 Uc7	3	Tu	Tu2220	*ot* *ne* *on*	300 1 300 1 500	10 12 17	10YR3/2 7.5YR4/4 7.5YR4/4	Vertical drainage Vertical drainage Lateral flow	Interflow
188	3	Kd	Kd2000	*ot* *gs* *gh*	200 600 900	20 16 30	ND ND ND	Macropore flow Lateral flow Waterlogged	Interflow
Uc8	2	Longlands (Lo)	Lo1000	*ot* *gs* soft plinthic (*sp*)	300 700 900	12 25 25	7.5YR3/2 10YR3/2 7.5YR5/6	Macropore flow Lateral flow Periodic saturation	Interflow
240	1	Hutton (Hu)	Hu2100	*ot* red apedal B (*re*)	380 1 500+	12 15	5YR3/2 10R3/4	Vertical drainage Vertical drainage	Recharge
Uc1	1/3	Bloemdal (Bd)	Bd1100	*ot* *re* *on*	500 1 200 1 500+	15 20 30	5YR2.5/1 5YR4/6 2.5YR5/6	Vertical drainage Vertical drainage Lateral flow	Interflow

[1] The expected dominant hydrological behaviour of various horizons based on morphological properties [2]*ND = not determined*

Figure 2.21 is a conceptual model developed by Van Tol *et al.* (2010) by interpreting the soil properties of the soil. The different processes are indicated by arrow.

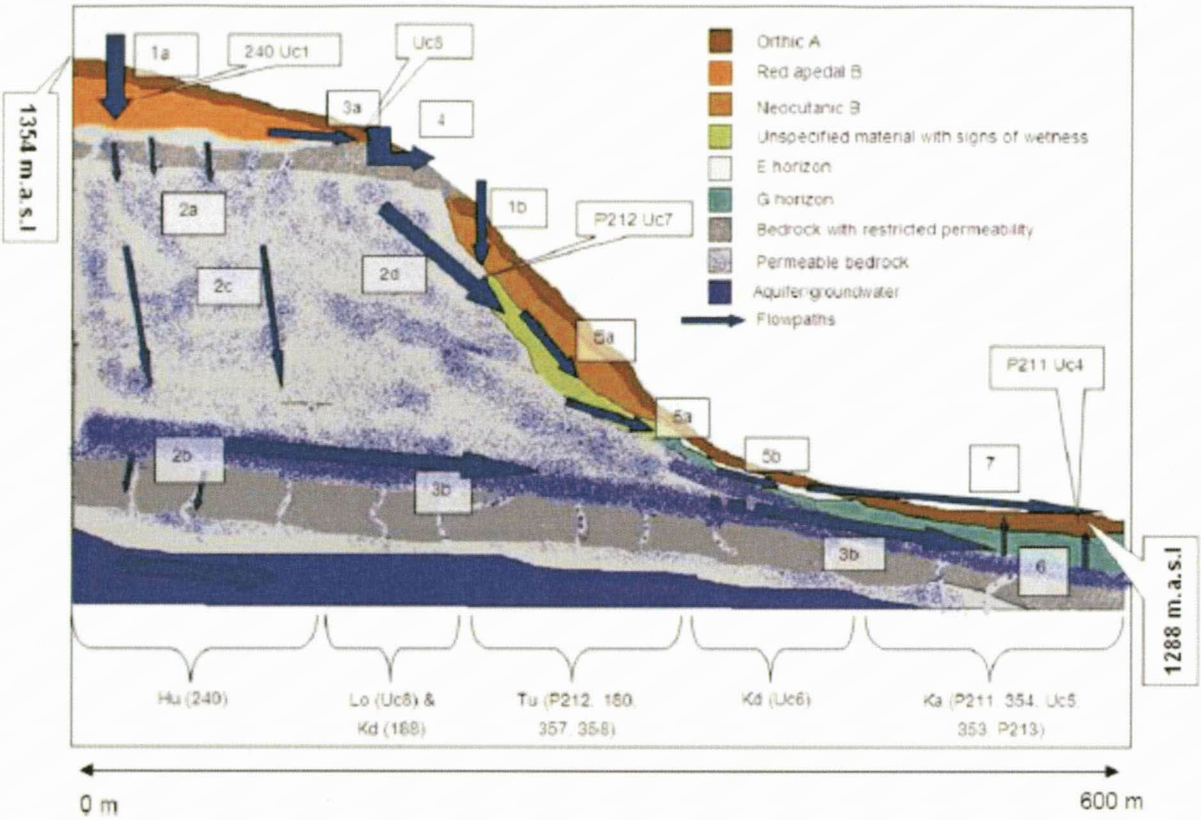

Figure 2.21 Conceptual model of study area (from Van Tol *et al.*, 2010).

Van Tol *et al.* (2010) used tensiometer and hydrograph data and found that the data supported the conceptual model. It is therefore possible to use soil morphological and chemical properties in the Weatherley catchment to hypothesis hydrological processes of soilscapes.

CHAPTER 3

HYDROPEDOLOGICAL INTERPRETATION OF SOIL CHEMICAL PROPERTIES

3.1 INTRODUCTION

Soil properties are generally perceived as being in equilibrium with the environment and that is expected to hold for all parameters derived from them. The irreversible nature of some soil properties like soil texture, and the lack of morphological expression in immature soils, questions the correctness of interpretations derived from morphological soil properties. Historical and current climate change, and annual, and event variation make the system more complex. Hydrology commonly revert to the use of soil physical parameters controlling the hydrological response of soilscapes e.g. soil texture, bulk density, hydraulic conductivity, etc., and real time hydrometry *e.g.* water level, water tension, etc. as snapshots of sections of the hydrological response of a transect, representative of a soilscape, which is extrapolated further to represent a catchment, and suitable to model hydrological response. These methods parameterise hydrological models *e.g.* with flow rates in unknown directions assumed to be vertical or lateral, lacking representivity of vertical horizonisation of soils and lateral distribution of soil horizons and the hydrological role of each.

The hypothesis is that the trends in chemical soil parameters relate to different hydropedological response types making it useful for pedotransfer function of hydrological response from soil data reported in conventional soil surveys.

3.2 METHODOLOGY

The soils of the Weatherley catchment are represented by 28 soil profiles. These profiles were grouped hydrologicaly according to their conceptual hydrological response using the classification developed by Le Roux *et al.* (2011). A soil was selected to represent each of the recharge, deep interflow, shallow interflow responsive (dry and wet examples) and responsive hydrological groups (Figure 3.1). Hutton (*Hu*) soils are grouped as recharge soils and P221 was selected as representative. The Avalon (*Av*), Pinedene (*Pn*), Tukulu (*Tu*) and Bloemdal (*Bd*) soils are grouped as deep interflow soils. A Bd soil (P220) was selected as a representative profile. Kroonstad (*Kd*), Westleigh (*We*) and Longlands (*Lo*) soils are grouped as shallow-interflow responsive soils. A Lo soil (P201) was selected to represent the drier shallow interflow responsive soils and a Kd soil (P205) to represents the wetter members. Katspruit (*Ka*) soils are grouped as responsive (saturation excess) soils and P218 was selected to represent the group.

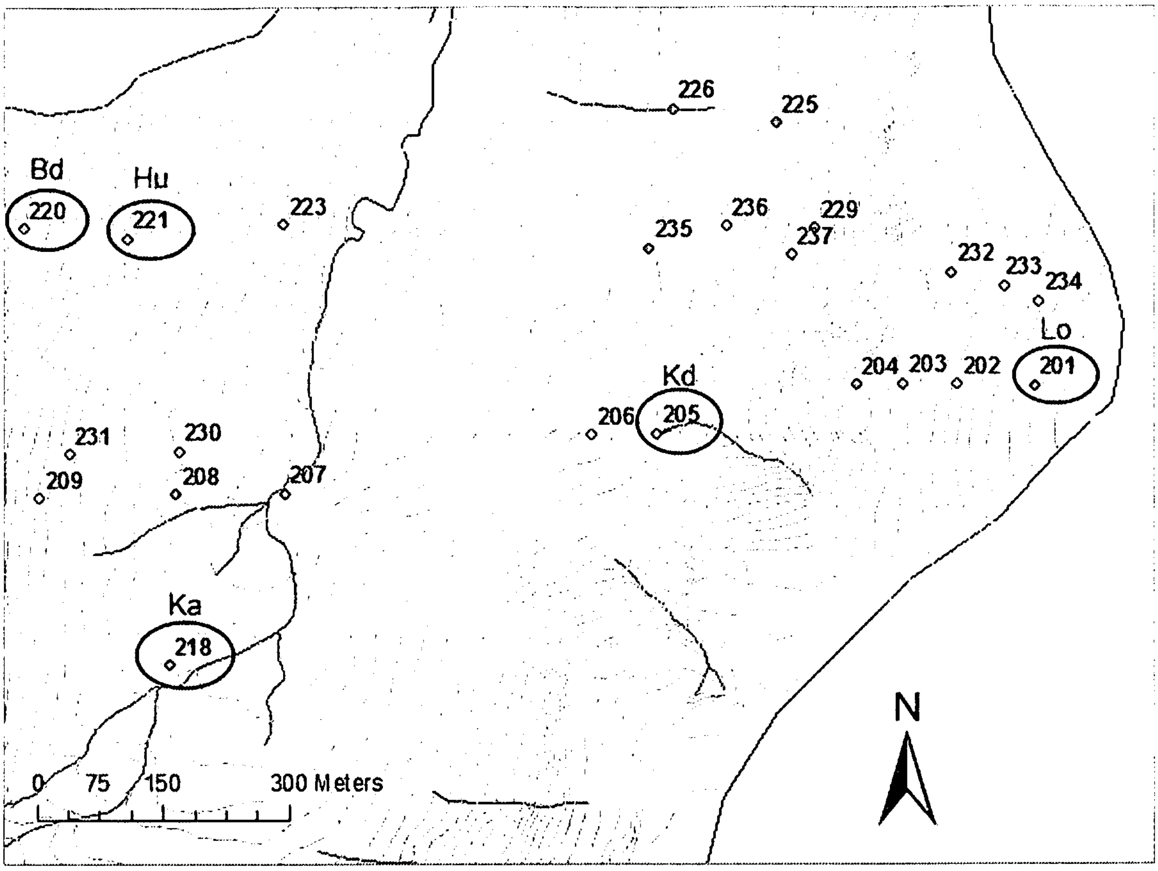

Figure 3.1 Position of the selected representative profiles (P201, P205, P218, P220 and P221).

The chemical properties of profiles (Van Huyssteen *et al.,* 2005) of different hydrological response groups were described and related to the morphology and water regime of the horizons and profiles. Exchangeable Ca, Mg and S-value (sum of base cations); pH_{water}, CBD extractable Fe and Mn were the chemical properties used for discussion. The water regime is presented as drainable water, quantified as the degree of saturation exceeding 70% of porosity, in days per year ($ADs_{>0.7}$), from long term soil water content measurements (Van Huyssteen *et al.,* 2005). The Neutron water meter which was used to measure water contents measured at 300mm intervals and therefore the horizon in which the $ADs_{>0.7}$ is represented occupied the relative measured area. The degree of saturation selected (s>0.7), represent the condition where soil pores are filled to the point where oxygen exchange are slowed down, reduction is expected to occur and significant amounts of drainable water is present in the soil.

3.3 RESULTS AND DISCUSSION

The chemical properties in the Weatherley catchment vary (Table 4.1). The Ca and Mg contents range from very low to high (Metson, 1961). Generally the responsive soils have the highest concentrations and the interflow soils have the lowest concentrations. The base saturation of the soils in the catchment indicate that soils range from very weakly leached in the responsive soils to very strongly leached in the interflow soils (Metson, 1961). The pH ranges from strongly alkaline in the responsive soils to strongly acid in the interflow soils (Bruce & Raymond, 1982). The Fe and Mn both have the minimum and maximum values in the responsive soils.

Table 3. 1 The ranges of chemical properties in the Weatherley catchment and the hydrological response soil type

	Ca	Mg	S	Base sat	pH	Fe	Mn
				Catchment			
Min	0.22	0.10	0.51	8.00	3.63	390	1.0
Max	17.35	7.86	20.93	153.40	8.90	44940	720.0
Mean	2.35	1.43	4.11	50.61	5.60	7023	73.0
STD	2.40	1.34	3.70	27.41	0.62	4734	104.8
				Responsive soils			
Min	0.26	0.21	0.63	13.00	4.87	390	1.0
Max	17.35	7.86	20.93	153.40	8.90	44940	329.5
Mean	3.46	2.21	6.06	64.51	5.84	7464	52.4
STD	2.65	1.64	4.11	29.13	0.56	5925	64.3
				Interflow soils			
Min	0.22	0.10	0.51	8.00	3.63	1145	1.5
Max	9.95	6.51	14.94	128.40	7.13	22050	447.0
Mean	1.90	1.12	3.32	43.16	5.49	6284	57.0
STD	2.14	1.11	3.29	24.90	0.66	3566	73.8
				Recharge soils			
Min	0.77	0.60	1.61	38.40	5.17	4980	94.5
Max	6.52	3.67	10.67	93.70	5.97	16150	720.0
Mean	2.22	1.44	4.01	58.56	5.52	10096	286.1
STD	1.63	0.83	2.43	14.92	0.24	3556	162.8

3.3.1 RECHARGE SOILS

Hutton soil was selected to represent the recharge soils. The *Hu* soil consists of an orthic A (*ot*) horizon (0 - 250mm), *re1* horizon (250mm - 600mm), *re2* horizon (600mm - 1200mm) and unspecified without signs of wetness (*od*) horizon (1200mm - 1400+mm). The *Hu* soil is classified as a recharge soil due to the presence of a *re* horizon as the first subsoil, related to unrestricted vertical flow and absence of redox and reduction morphology as signs of periodic or permanent saturation.

The lack of redox morphology at the deep saprolitic subsoil horizon indicates that the water that infiltrates the soil moves through the *re* horizon will not stagnate for prolonged periods but will infiltrate the underlying rock recharging laterally movement in rock fractures and/or the regional groundwater. The bleached colour of the *ot* horizon indicates short intervals of reduction, which could be enhanced by the higher absolute and easily oxidisable OC content of the topsoil. It supports the pedogenetic yellowing of the B1 horizon of the Griffin soil observed in transect 210 (see Chapter 4). Yellow-brown apedal B (*ye*) horizons overlying *re* horizons also occur in humic soils related to the topsoil. The *re1* is a transitional horizon between the *ot* and *re2* horizons. The morphology of the *re* horizon indicates that internal drainage exceeds infiltration rate and saturated hydraulic conductivity of the *ot* horizon creating an oxidising environment.

The drainable water in the *ot* and *ot/re* transitional horizon ($ADs_{>0.7}$ of 115 days of the year) can be explained by near surface macropore flow recorded by Lorentz (2007) using tensiometry. It may be related to the bleached colour of the *ot* horizon. Similar results were recorded in *ot* horizons overlying *re* horizons *e.g.* the *Bd* soil (P210). The upper *re* horizon has drainable water for short periods increasing downwards ($ADs_{>0.7}$ of 4, 8 and 50 days per year respectively) and a similar duration in the *od* horizon indicating a fractured rock water table and lateral water flow in the shallow fractured rock.

The Ca, Mg and S-value profiles support these interpretations (Figure 3.2). The high values at the surface, decreasing in the *ot* horizon, is indicative of biocycling. The increasing values of these properties in the *re* horizon and increase in the *od* horizon, indicate redistribution of bases downwards by leaching and weathering of saprolite.

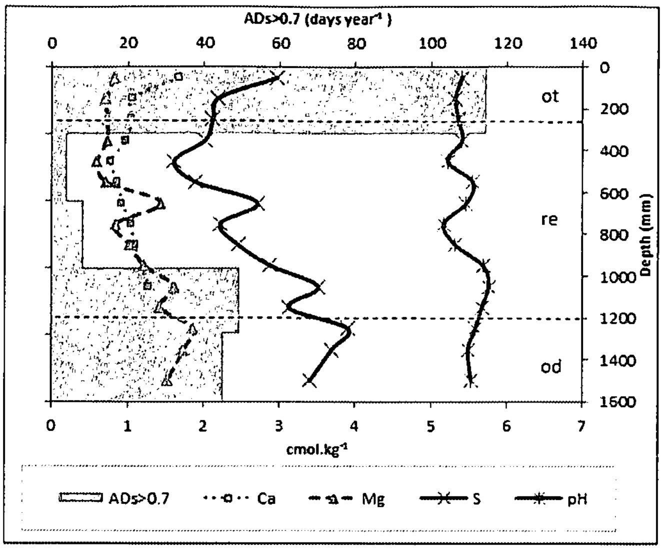

Figure 3. 2 ADs$_{>0.7}$, Ca, Mg, S value and pH for the *Hu* soil (Adapted from Van Huyssteen *et al.*, 2005).

The erratic crude relationship of Fe Mn distribution with horizonation cannot be explained (Figure 3.3).

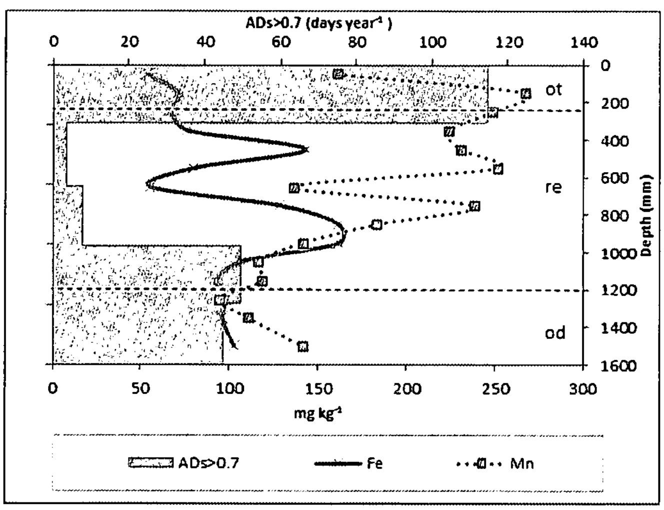

Figure 3. 3 ADs$_{>0.7}$, Fe/100 and Mn in the *Hu* soil (Adapted from Van Huyssteen *et al.*, 2005).

3.3.2 INTERFLOW SOILS

3.3.2.1 Deep Interflow

A *Bd* soil was selected to represent the deep interflow soils The *Bd* soil consists of an *ot* horizon (0 - 540mm), *re* horizon (540mm - 880mm) and unspecified material with signs of wetness (*on*) horizon

(880mm - 1500mm). It classifies as a deep interflow soil. The surface and first subsoil are oxidic and respond as recharge horizons. The *on* horizon has redox morphology, indicating an extended duration of drainable water expected to contribute to lateral flow. The chemical property profiles indicate that several processes are active in pedogenesis.

The *re* horizon is never saturated and the *on* horizon is saturated for more than 300 days a year indicating a lateral water source and interflow.

The Ca, Mg and S-value profiles support these interpretations (Figure 3.4). The high values at the surface, decreasing in the *ot* horizon, is indicative of prominent biocycling activity. The consistent low values of these properties in the *re* horizon and increase in the *on* horizon, indicate redistribution of bases downwards by leaching. The increase in bases in the *on* horizon, although still low according to Hazelton & Murphy (2007), and when compared to the mean values of Ca and Mg (2.35 $cmol_c$ kg^{-1} & 1.43 $cmol_c$ kg^{-1} respectively) in the catchment, suggests that the cations have been translocated to the *on* horizon by vertical and interflow drainage. The enrichment is related to a degree of stagnation, slow flow rate, long flowpath and relative long residence time. The chemical processes are horizon related except at the transition of the *re* and *on* horizons. Either the *on* horizon is migrating up or different degrees of saturation in the *on* horizon have created different chemical environments.

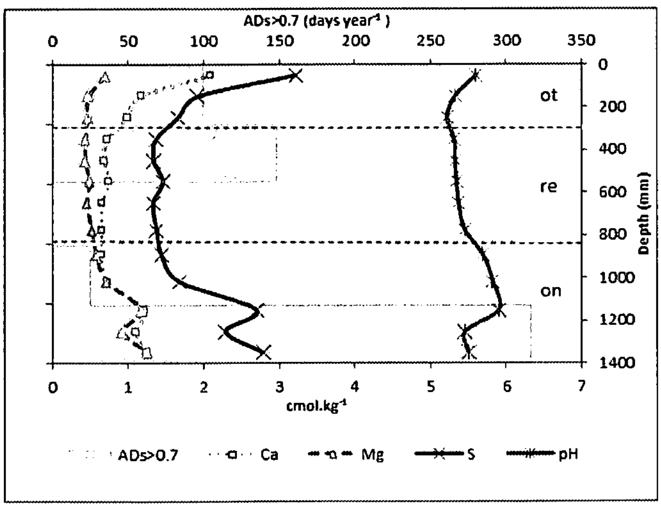

Figure 3. 4 $ADs_{>0.7}$, Ca, Mg and S value for the *Bd* soil (Adapted from Van Huyssteen *et al.*, 2005).

Base saturation decreases sharply from the surface and continues to decrease in the *ot* horizon (Figure 3.5). There is an increase in base saturation with depth in the *re* horizon towards the *on* horizon. In the *on* horizon there is a decrease in base saturation. The increase in base saturation does not correspond with the Ca and Mg profiles.

Fe and Mn (Figure 3.5) both decrease from the surface in the *ot* horizon. Contrary to the bases, Fe is not sensitive to leaching at this pH and Mn is less sensitive for redistributed by leaching, supporting a biocycling process. The Mn concentration profile is systematic and its relationship with genetic soil horizons indicates that the redistribution of Mn is controlled by the same mechanisms controlling horizonisation. The low Mn concentration in the *re* horizon is an indication that solubility at low pH and leaching may the controlling processes of Mn in this horizon. The Mn concentrations in the profile are below the mean concentrations of interflow soils (447 mg kg^{-1}) in the catchment. The high Mn concentration at the *re/on* transition and systematic increase in the *re* horizon and decrease in the *on* horizon, is indicating redox redistribution supported by diffusion and capillary rise. Diffusion is typically expected to contribute to concentration variation over short distances, as in mottle formation, while capillary rise is controlled by root uptake of water and creation of a water pressure gradient and unsaturated flow.

The Fe concentration is inconsistent and not related to soil morphology. It may be related to mineral weathering. A general increasing trend in the lower *ot* and *re* horizons and some consistency in the *on* horizon is related to the distribution of Mn and the processes responsible for it. The chemical properties confirm the difference in morphology of the *re1* and *re2* horizons and that the *re1* horizon reacts similarly to the *ot* horizon.

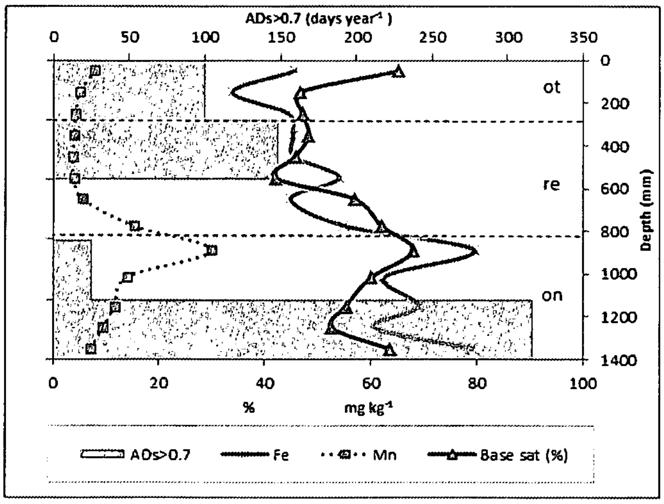

Figure 3. 5 ADs$_{>0.7}$, Fe/100, Mn and base saturation for the *Bd* soil (Adapted from Van Huyssteen *et al.*, 2005).

The contribution of recharge and deep interflow soils to recharge of groundwater is uncertain. The fact that the crest and upper midslope have drainable water that feeds lower lying soils should be an indication that it could make a continuous contribution to groundwater over time.

3.3.2.2 Shallow-Interflow Responsive

A *Lo* soil was selected to represent the dry end members of the shallow-interflow responsive soils. The Lo soil consists of an *ot* (0 - 430mm), *gs* (430mm - 730mm) and soft plinthic B (*sp*) (730mm - 980mm) horizons. It classifies as a shallow-interflow responsive soil due to the presence of *gs* and *sp* horizons related to interflow. Mottling in the *sp* horizon is associated with a "fluctuating" watertable indicating a more stagnant condition. The dominant pedological processes are ferrolysis, which breaks down silicate clay and releases Fe and Mn, and redistribution of Fe and Mn in the *gs* and *sp* horizons. The relative uniform grey, sandy morphology of *gs* horizon indicates ferrolysis under conditions resulting in a relative homogeneous reducing condition. That implies that either the depth associated with high OC content creates an intense reduction or a relative homogeneous reducing environment created with a degree of flow that results in miscible displacement.

The duration of drainable water as indicated by $ADs_{>0.7}$ values, systematically increase down the profile (Figure 3.6). The $ADs_{>0.7}$ value for the *ot* horizon is 17.67 days of the year. $ADs_{>0.7}$ in the upper *gs* horizon is 40.83 days of the year and the $ADs_{>0.7}$ in the *sp* horizon is 93.83 days of the year. The systematic increase is attributed to one sustainable lateral water source systematically varying in level. The flow rate is expected to be slower and deeper in the profile. Flow rate is controlled by the varied water table levels as to produce the variation typical of plinthic soils. The systematic increase in $ADs_{>0.7}$ profiles of all plinthic soils of the catchment is similar and in contrast with the other $ADs_{>0.7}$ profiles *e.g.* in the *Bd* soil the increase to the *on* is indicating a very stable water table level implying a large buffer action as event and seasonal impacts are buffered.

A change in chemical properties correlates with a change of horizon in the profile. The Ca, Mg and S-value profiles are similar and support the hydrological deductions from morphology (Figure 3.6). Enrichment of the *ot* horizon is indicative of bioclycling. The elements then stabilize in the *gs* horizon and further down the profile. The depletion of base cations in the *gs* and *sp* horizons are a result of ferrolysis due to Fe and Mn competing for exchange sites with the base cations confirming the morphological interpretations. The *sp* horizon has close to the lowest Ca and Mg concentrations in the catchment. The changes in base cations are correlated with changes in horizons indicating that the processes attributed to horizon formation are controlling the distribution of base cations. The pH decreases in the *ot* horizon and decreases in the *gs* horizon (Figure 3.6). The decrease in pH especially in the *gs* horizon indicates prominent ferrolysis. There is a slight increase in pH in the soft plinthic horizon indicating that the strongest ferrolysis is present in the upper *sp* horizon or there is a variation caused inheritantly by the parent material.

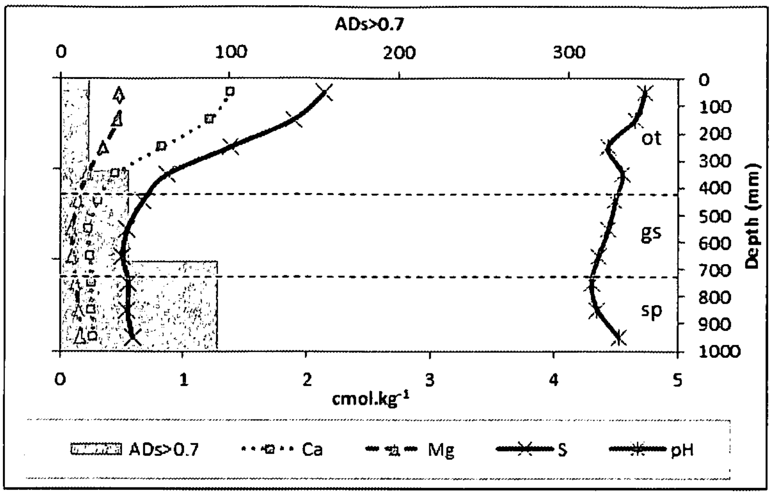

Figure 3. 6 ADs$_{>0.7}$, Ca, Mg and S value for the *Lo* soil (Adapted from Van Huyssteen *et al.,* 2005).

The Fe content increases with depth and there is a peak at the transition of *gs* horizon and the *sp* horizon (Figure 3.7). The peak at the *gs/sp* transition can be due to more oxidising conditions present and Fe can precipitate due to a lower ADs$_{>0.7}$. The Fe contents are also low indicating that Fe has been depleted and ferrolysis is active in the *gs* and *sp* horizons. The Mn content decreases in the *ot* horizon and stabilizes in the *gs* horizon only to increase towards the transition with the *sp* horizon (Figure 3.7). The decreasing trend in Mn suggest that the reduction potential is more consistently higher than 300mV as more Mn is being reduced and removed from the profile than Fe. These properties not only suggest that the horizons undergo reduction as indicated by the ADs$_{>0.7}$ in the *gs* horizon and *sp* horizons but that water movement is present to leach the cations from the soil solution. Indicating that the classification based on morphology as an interflow soil is correct and that *Lo* soil is still phase with the morphology.

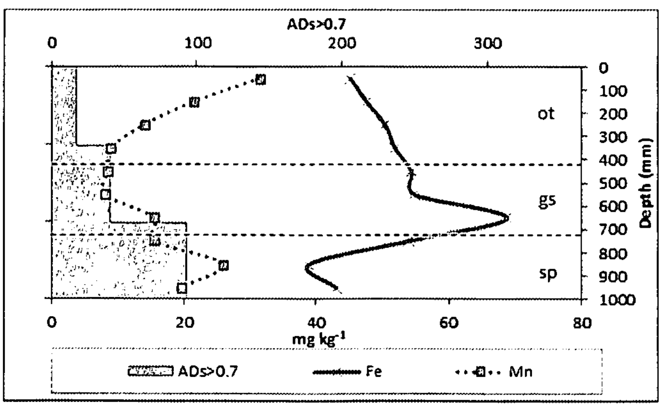

Figure 3. 7 ADs$_{>0.7}$, Fe/100 and Mn for the *Lo* soil (Adapted from Van Huyssteen *et al.*, 2005).

The *Kd* soil was selected as representative of the wet end members of the shallow-interflow responsive soils. The profile consists of an *ot* horizon (0 - 220mm), *gs*1 horizon (220mm - 460mm), *gs2* (460mm - 660mm) horizon and G horizon (*gh*) horizon (660mm - 1400mm). The *Kd* soil is classified a shallow-interflow responsive soil as interflow is expected in *gs* horizon as the first subsoil. Contrary to the Lo soil the interflow is controlled by the *gh* horizon as a third horizon not a fourth horizon. The bleached colour of the *ot* horizon indicates short intervals of reduction which could be enhanced by the higher OC content of the topsoil. The *gs1* horizon lacks colour compared to the *gs2* horizon, which has Fe accumulation, indicating that reduction is more intense in the *gs1* horizon . The *gs2* horizon is likely a younger horizon with Fe not completely removed from the horizon. The reduced nature of the *gh* horizon indicates prolonged periods of reduction caused by saturation and associated with stagnation and storage mechanism.

The $ADs_{>0.7}$ value of the *ot* horizon is 56.83 days of the year. The $ADs_{>0.7}$ at the *gs1/gs2* transition is 84.83 days per year. At the *gs/ gh* transition the $ADs_{>0.7}$ was 258.17 days per year and increase to 354.86 days per year in the *gh* horizon and saturated for the entire year at the bottom of the *gh* horizon. The *Kd* soil is saturated for longer periods than the *Lo* soil.

Calcuim, Mg, S-value and pH of *Kd* soil support the morphological interpretations (Figure 3.8). The Ca, Mg and therefore S value all react similarly, decreasing in the *ot* horizon from the surface indicating biocycling in the *ot* horizon. The consistently low values in the *gs* horizon suggest that ferrolysis is active in the *gs* horizons due to the leaching of cations. The properties initially increase in the *gh* horizon but then decreases with an increase peak at the bottom. The general enrichment of cation the *gh* horizon is an indication of water stagnation functioning as a storage mechanism. The pH is very consistent in the profile; the only noteworthy change is half a unit increase from the *gs* horizon to the *gh* horizon, indicating that the ferrolysis in the gs horizon is not present in the *gh* horizon which is the difference when compared to the sp horizon of the Lo soil. The rise in pH in the *gh* horizon also indicates that no leaching is present. In Figure 3.8 it is evident that the chemical properties are associated with soil horizons but even more so with $ADs_{>0.7}$. The enrichment in the upper 200mm of the *gh* horizon is in contrast with the lower *gs2* horizon with the same $ADs_{>0.7}$. The lower OC content at the depth could cause a decrease in reduction of Fe and Mn, therefore increase in clay and cations. The $ADs_{>0.7}$ could also be below the threshold at the deeper depths and the increase in $ADs_{>0.7}$ in the *gh* horizons increases reduction. The *gs* horizon could be migrating into the *gh* horizon. The increase at the bottom of the *gh* horizon can be caused by the less weathered transition to the parent material.

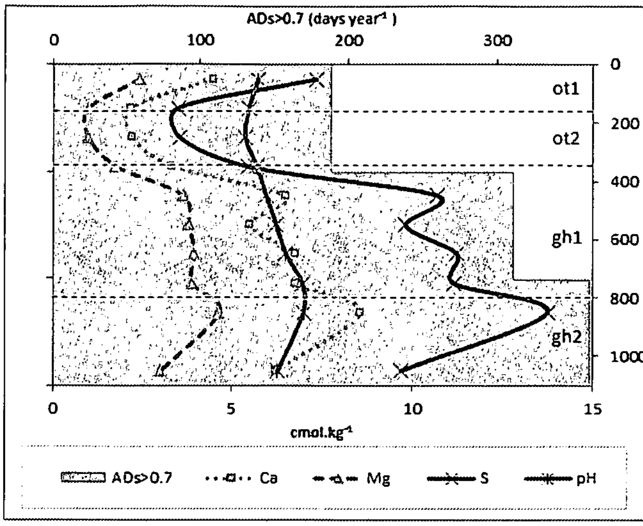

Figure 3. 8 ADs$_{>0.7}$, Ca, Mg, S value and pH for the *Kd* soil (Adapted from Van Huyssteen *et al.*, 2005).

Iron and Mn concentrations are inconsistent in the profile (Figure 3.9). There is a general increasing trend from the surface to the transition of the *gs2* and *gh* horizons, the increase in Fe corresponds to increase in colour of the *gs2* horizon. Then changes to a decreasing trend in the *gh* horizon, indicating that the reduction in the *ot* and *gs* horizon is less prominent as in the *gh* horizon were Fe is reduced and translocated from the *gh* horizon. The Mn concentrations change with a change in horizon. Mn is low in the *ot* horizon, increases in the *gs2* horizon and decreases towards the *gh* horizon. The increase in Mn in the *gs2* correlates with the increase in Fe and the increased colour of the *gs2* horizon. The more mobile Mn is most abundant in the *gs2* horizon confirming the immaturity of the horizon. There is a decrease in Mn and Fe is low indicating that Fe and Mn are not competing for exchange sites but the opposite is true in the lower *gh* horizon where there is an increase in Mn and a decrease in basic cation. The concentrations of basic cations have a negative correlation with Mn in the *gh* horizon.

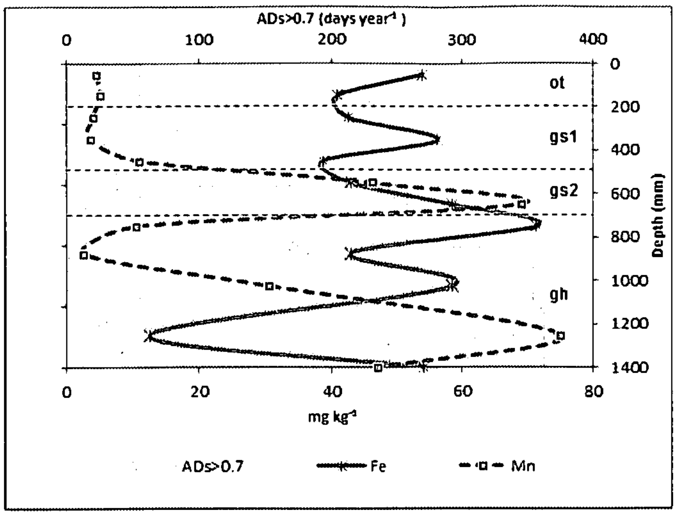

Figure 3. 9 $ADs_{>0.7}$, Fe/100 and Mn for the *Kd* soil (Adapted from Van Huyssteen *et al.*, 2005).

3.3.3 RESPONSIVE SOILS

The *Ka* soil consists of an *ot1* horizon (0 - 180mm), *ot2* horizon (180mm - 350mm), *gh* horizon (350mm - 900+mm). *Ka* soil is classified a responsive soil due to the saturated nature of the *gh* horizon, any infiltration will saturated the soil quickly inducing overland flow. Reduction is expected in the *gh1* and *gh2* horizons due to water saturation for long periods as indicated by the gleyed morphology. The *ot1* and *ot2* are very similar in colour and structure. The reddish colour at the *ot2/gh* transitions due to reduced Fe being mobilized up the profile and then oxidizing in the O_2 rich environment. The abundance of mottles in the *gh* horizon indicates reducing conditions which the Fe is reduced and brief oxidizing conditions for the Fe mottles to form. The gley morphology is synonymous with saturated conditions and suggests that the *gh* horizon functions as a storage mechanism of water in the landscape.

The $ADs_{>0.7}$ value of the *ot* horizon for 189.5 days of the year. The transition of the *ot2* and most of *gh1* horizons has an $ADs_{>0.7}$ value of 312.17 days per year. The *gh2* horizon has an $ADs_{>0.7}$ of 363.17 days per year.

There is a decrease in all chemical properties in the *ot1* horizon due to surface enrichment by biocycling and an increase in the *ot2* horizon (Figure 3.10). Ca becomes consistent in the *gh1* horizon with a slight decrease in the *gh2* horizon. Mg and S value, due to the irregular nature of the Mg, generally has an increasing trend in the *gh1* horizon. The enrichment of the *gh1* horizon indicates water stagnation and base cations are deposited in the horizon due to ET losses of water. Cations decrease in the *gh2* horizon. There is a large progressive increase in pH from the *ot2* horizon to the bottom of the *gh1* horizon and a decrease in the *gh2* horizon. The increase in pH down the profile suggests a stagnation of water movement and that no leaching takes place and hence the accumulation of Ca and Mg in the *gh1* horizon. The decrease in all properties in the *gh2* horizons suggests a change in processes from the *gh1* horizon. Ferrolysis will cause a decrease in cations and therefore a decrease in pH, water movement instead of stagnation will leach cations from horizons and also cause a decrease in pH. The trends in the chemical properties are confined to horizons and transitions of horizons. The trends of the *ot2* are migrating into the *gh1* horizon.

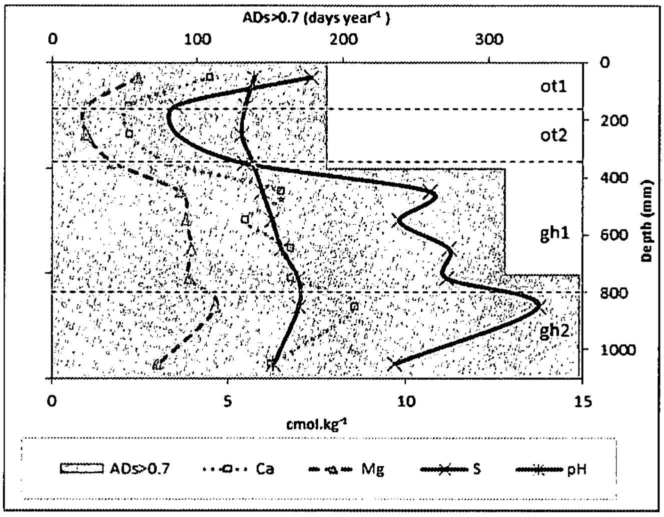

Figure 3. 10 ADs$_{>0.7}$, Ca, Mg, S value and pH for the *Ka* soil (Adapted from Van Huyssteen *et al.*, 2005).

Fe and Mn for *Ka* soil decreases in the *ot1* horizon and stabilize in the *ot2* horizon (Figure 3.11). There is a large increase in both Fe and Mn to the transition of the *gh1* horizon and a decrease in the *gh1* horizon. Fe is constant to the bottom of the profile but Mn increases towards the transition with *gh2* horizon and decreases in the *gh2* horizon. The accumulation of Fe and Mn at the transition of the *ot2* and *gh1* horizons corresponds to the reddish colour in the morphology and due to Fe^{2+} and Mn^{2+} being translocated upwards through capillary rise and due to more oxidizing conditions stabilize. The Fe and Mn are also reduced in the *ot2* horizons and could have illuviated to the transition. The decrease and stabilization at low concentrations of Fe in the *gh1* and *gh2* horizons confirmed that Fe has been reduced and removed from the horizons. The change in properties corresponds with a change in horizons suggest that the horizon is in phase with the morphology. The *ot2* horizons could be an immature *gs* horizon.

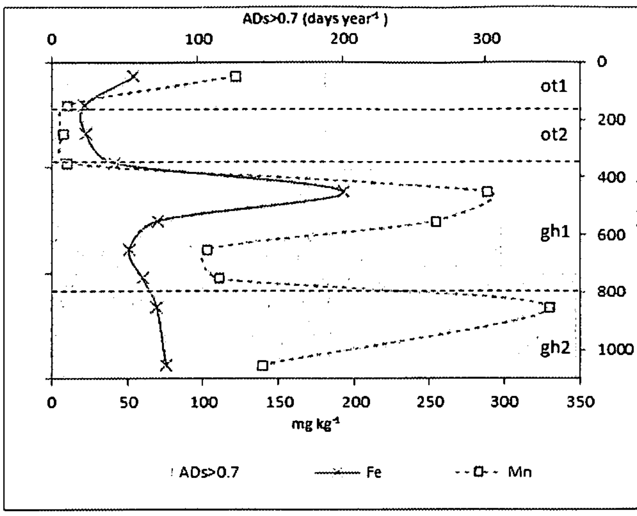

Figure 3. 11 ADs$_{>0.7}$, Fe/100 and Mn for *Ka* soil (Adapted from Van Huyssteen *et al.*, 2005).

3.4 CONCLUSION

Soil chemical parameters largely support the interpretations from the relationships between soil morphology and duration of drainable water typically used by hydropedologists to design conceptual hydrological response models. Therefore, current hydrological responses of soilscapes are expected to behave according to morphological indicators and chemical parameters commonly reported in soil surveys. Soilscapes and soil chemical parameters can function as entities to transfer soilscape hydrological information. Horizonisation, commonly considered the result of pedogenesis, vertically group soil properties and soil forming processes related to hydrology, controlling the presence and expression of soil morphological features and chemical parameters.

Both soil morphology and chemical parameters indicate that soil fertility and hydrological responses focus in the surface horizons and subsoil horizons of soilscapes of the Weatherley catchment, respectively. Bio-activity, biocycling and bioturbation can be responsible for accumulation of humus (darkening), cations, Fe and Mn in the *ot* horizons, decreasing from the surface to the transition with the B1 horizon. In recharge and deep-interflow soils free vertical flow dominates in the B1 horizons. Properties inherited from the parent material and lateral flow respectively dominates in the deep subsoils. In shallow-interflow responsive and responsive soils the morphology and chemistry are controlled by secondary chemistry related to water logging and residence time. Minor peaks in chemical parameters in genetic horizons indicate the presence of thin immature flowlines that is not old enough or intense enough to leave morphological signatures.

However, the variation in chemical parameters that could not be explained emphasizes our limited perception on the temporal and spatial variation of the processes driving these chemical parameters.

CHAPTER 4*

USING ANCIENT AND RECENT SOIL PROPERTIES TO DESIGN A CONCEPTUAL HYDROLOGICAL SOILSCAPE RESPONSE MODEL

*Chapter 4 is in final preparation to be submitted as an article and as a consequence written as an independent chapter with references included.

4.1 INTRODUCTION

Soils are natural entities. In general, their properties have developed slowly over a long time equilibrating with environmental drivers (Dorronsoro, n. d.). The dominant environmental impact, namely the chemical interaction between water and parent material, manipulated by terrain, continues in soil. The interactive relationship between soil and hydrology has led to the development of a new field of study called hydropedology (Lin, 2010). The interaction of water, soil and fractured rock plays a significant role in the behaviour of catchments (Van Tol *et al.*, 2010; Kuenene *et al.*, 2011; Lorentz, 2004). This zone of activity has a large impact on the biosphere and is therefore called the critical zone Lin (2010). The critical zone, with high bio-chemical activity, is of importance in soil science, hydrology and ecology, especially with rising levels of pollution in soil and water. Hydropedology has the ability to bring hydrology and pedology together and broaden knowledge in water movement (Pachepsky *et al.*, 2006).

Hydrological studies are typically researched on catchment scale. Catchment response in turn, is determined by the response of individual soilscapes (Sivapalan, 2003). Soilscape response is mainly controlled by soil distribution patterns (Soulsby *et al.*, 2006), and its interaction with underlying rock, and is therefore a first order control of water movement within the critical zone (Park *et al.*, 2001, Soulsby *et al.*, 2006). The soil distribution pattern controls hydrological processes such as flowpaths, residence times and storage mechanism (Soulsby *et al.*, 2006) which influence the quantity and chemical composition of the water exiting the soilscape (Jacks and Norrström, 2004).

The term soilscape replaces hillslope in this discussion for the following reasons: hydrological soilscapes are related to the pedological catena. Milne (1936) first referred to a catena as the typical distribution of soils in a landscape and was later modified to toposequence (Bushell, 1942). Toposequences found in Fritsch & Fitzpatrick, (1994) and van Tol *et al.* (2010) of red freely drained soils on the crest and hydromorphic soils close to the stream is a typical example of a sequence of soils. Soilscape refers to this orderly distribution of soils in landscapes (Landis, 2013). Soilscape is the merging of the terms soil and landscape (Blume *et al.*, 2002), therefore it does not detract anything from the previous understanding of a landscape/hillslope but emphasises the third dimension of width to the distribution of soils in the catchment.

Although soil physical data usually reported in soil surveys are originally used to design pedotransfer functions (Bouma, 1989), soil morphology, as reported in soil surveys, can also serve as a pedotransfer function to infer the hydrological response of soils (Fritsch & Fritzpatrick, 1994). Soil morphology data are more accessible than soil physical and hydrometric data commonly used to determine hydrological processes within a soilscape. Gathering soil physical and hydrometric data is tedious, time consuming and often unreliable (Park and Burt, 1999). Soil morphology can play an important role in conceptualising soilscape hydrology and assist in assigning the correct model structure in soilscape hydrological models (Lorentz *et al.*, 2007; Van Tol *et al.*, 2011). Soil morphology is a visible indicator of the interaction between water and parent material and the resultant variation indicates soil water regimes. Soil morphology was successfully applied by Fritsch & Fitzpatrick (1994), Soulsby *et al.* (2006), Ticehurst *et al.* (2007), Van Tol *et al.* (2010) and Kuenene *et al.* (2011) to develop conceptual hydrological response models of soilscapes and subsequently catchments.

The soil water regime correlates with the morphology of soils (Van Huyssteen *et al.*, 2005). Soils with *re* (red apedal B) horizons are well drained and respond with short periods of near saturation contrary to soils with grey (*gh*) horizons which are saturated for long periods. These phenomena allows for the distinction of three dominant soil water responses. The genetic *re* horizons respond as recharge hydrological units compared to grey *gh* horizons serving as storage mechanisms. Recharge soils are defined as soils that rapidly transmit water. Interflow soils, also defined as throughflow, are soils were water is diverted laterally down the soilscape by an impeding layer. The impeding layer has a lower saturated hydraulic conductivity (K_s). Responsive soils are saturated for long periods of time due to saturation excess with water (Le Roux *et al.*, 2011; Van Tol *et al.*, 2012). The yellow brown apedal B (*ye*) horizon is saturated slightly longer than the *re* horizon (Van Huyssteen *et al.*, 2005). However the *ye* horizon of the Griffen (*Gf*) soil is interpreted as recharge. This contradicts the results of Van Huyssteen *et al.* (2005). This is because the yellow-brown colour of the *Gf* soil is related to soil mineralogy (Fey, 1983). However the colour variation is rather soil redox chemistry related and interpreted as arising from quick reduction due to high OC contents in the topsoil rather than longer saturation typical of other *ye* horizons. The hydrological relationships of horizons, pedons and soilscapes discussed above, contributed to the improvement of a hydrological model (Lorentz *et al.*, 2007).

The process of soil formation is relatively long (10^2-10^4 years) whereas when compared to the process of pH change which is shorter (10^0 - 10^3 years) at catchment scale (MacEwan, 1997). Properties relating to soil morphology; cutans, drainage which is largely related to structure, soil colour (hue), and the delineation of master horizons is unlikely to change in a lifetime whereas soil chemical properties are likely to change (MacEwan & Fitzpatrick, 1996), confirming that soil chemistry is more sensitive to change than soil morphology.

Chemical weathering can't take place without water and water is the mediator in further chemical reactions in the soil (Essington, 2004). Hydrology therefore plays an important role in resultant soil chemistry. Geochemical indicators are related to hydrological processes (McDaniel *et al.*, 1992 and Park & Burt, 1999) and could transform the procedure of soilscape modelling by reducing the time factor related to physical measurements and improve the quality of hydrometric and isotope data, which, without the support of other data could produce erroneous predictions (McDonnell *et al.,* 2007).

Modelling soil chemistry using water chemistry confined to a soilscape, is very difficult and consequently most soil chemistry is modelled using water chemistry at catchment scale (Burns et al., 1998). Numerous one dimensional studies have been done on the effects of hydrology on soil chemistry focussing on a single soil attribute or at plot scale. Soilscapes should be studied three dimensionally, instead of point observations, in order to gain a holistic understanding of the complexity of soilscape hydrological behaviour (Tromp van Meerveld & McDonnell, 2006), in other words in a soilscape context.

Studies have been done on the effects of water on soil chemistry. The Fe and Mn concretion contents are highest in horizons with fluctuating water tables rather than horizons that are more permanently saturated D'Amore *et al.* (2004) and le Roux (1996). According to Park and Burt (1999) and McDaniel *et al.* (1992) Fe and Mn can be used as pedochemical indictor and identification of throughflow in soils. Fe^{2+} can be absorbed on the exchange cites thereby freeing Ca and Mg and increasing the Fe content (Phillips & Greenway, 1998).

Many studies in the Weatherley catchment contribute to the understanding of the conceptual model. Lorentz & Esprey (1998) found that water accumulation at the toe slope was dependant on soil properties, surface and bedrock topography. A large component of quick flow was ascribed to near-surface macropore flow implying soil contact only (Lorentz *et al.,* 2004). The Molteno formation forms a prominent shelf which plays an important role in the hydrology of the catchment (Van Tol *et al.,* 2010). A comprehensive hydropedological study was done by Van Huyssteen *ET al. (2005)* from which some data is extracted for this paper. The study revealed that similar hydrological responses from soils exhibited similar amounts and distribution of organic carbon in the profile and there is a linear decrease from the surface to 600mm irrespective of the different B horizons (Le Roux *et al.*, 2005).

Developing a concept is the first step in scientific research and that is also true of hydropedology. Soilscape hydrology can be conceptualised using signatures related to the interaction of water with soil and fractured rock. Soil morphology is effectively applied as signatures of this interaction. That emphasises the role of a soil profile description as pedotransfer function. However, the relative slow reaction of soil morphology to changes in soil water regime and the fact that some morphological features are irreversible, bring into question its universal application. Hydrometrics is applied as current (real time) indicators of response but its application is limited by its snapshot nature. Soil chemistry data, as sensitive, equilibrated products of water/soil interaction can fill this gap as it can connect point data in a soilscape and connect soil horizon and soil pedon response at soilscape scale.

The hypothesis is that soil chemical properties are likely to be a recent indicator of hydrological processes, and soil morphology an indicator of ancient waterflow processes

4.2 METHODOLOGY

4.2.1 SITE DESCRIPTION

The Weatherley catchment is situated southwest of Maclear in North Eastern Cape. The catchment is 160ha (Lorentz, 2001) and forms part of the Mooi River catchment, which is a quaternary catchment of the Umzimvubu basin. Elevation ranges from 1254 to 1352 meters above sea level (Figure 4.1). It has a mean annual precipitation (MAP) of 1000mm and mean annual potential evapotranspiration (PET) of 1480 mm (BEEH, 2003). The soilscape was initially covered by Highveld sour grass (Acocks', 1975) and has a basal cover of 50 - 70% (Esprey, 1997). Parts of the soilscape have been afforested since 2002. The catchment consists of mudstone and sandstone of the Elliot formation above 1320m a.m.s.l. and Molteno formation below (De Decker, 1981).

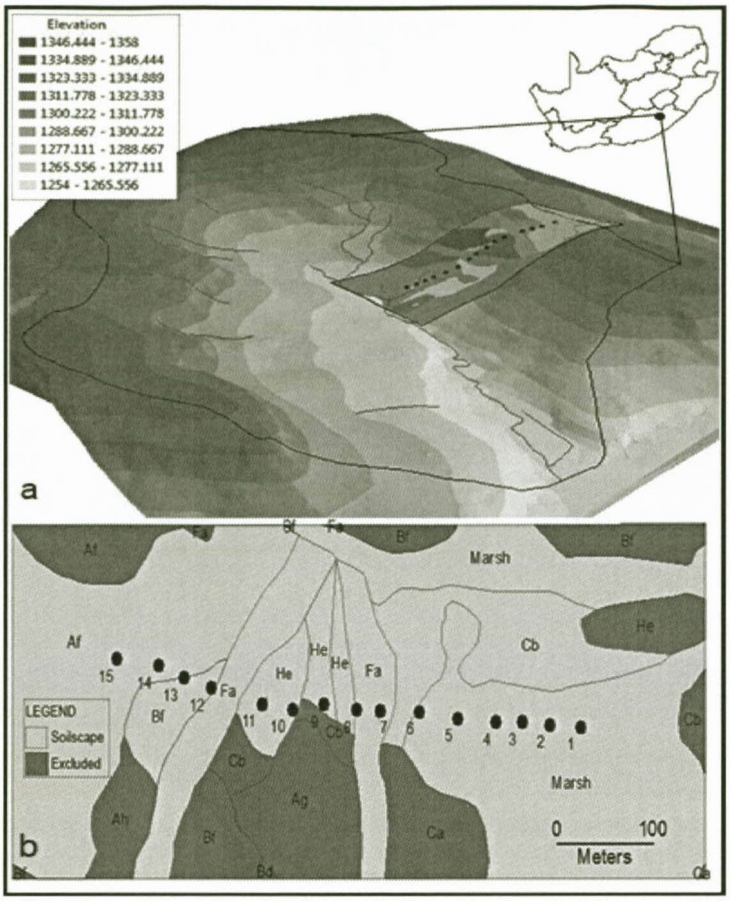

Figure 4.1 A) Location of the Weatherley catchment in South Africa and B) transect of observations.

Profile descriptions and chemical analyses of three profiles (Van Huyssteen et al., 2005) were used to establish the relationship between soil morphology and soil chemistry. Soil water contents and tension were measured for more than ten years using a neutron water meter (Van Huyssteen *et al.*, 2005) and tensiometers (BEEH, 2003). Soil auger observations were made every 20m and samples taken at 200mm intervals. At each auger observation the soil morphology was described in detail and classified according to Soil Classification Working Group (1991). Samples were dried, crushed and analysed using a Mid inferred spectrometer (MIR). Soils were divided into three main soil types related to their hydrological response as defined by Van Tol *et al.* (2011).

Soil tension and soil water content data were used to verify the deductions made in the conceptual hydrological response model. In 2001 the catchment experienced less rain than usual and the month of March is characterized by a few days with no rainfall and then a few days with rain (Figure 4. 9). There is a large rainfall event on the 11^{th} of 47.3mm. This pattern allows a wetting up sequence and a drying sequence which gives a good indication of the water regimes. Annual duration of saturation

($ADs_{>0.7}$), which is defined as 70% saturation of porosity and expresses in days $year^{-1}$ as proposed by Van Huyssteen *et al.* (2005) to be the beginning of reduction.

Three steps were used to identify the factors and processes controlling the soilscape hydrology to develop the hydrological response model (i) Morphology was used to design the conceptual hydrological model; (ii) interpretation of the chemical data was used to improve the model (iii) long term soil water content and tension data were finally used to verify the interpretations.

4.3 RESULTS and DISCUSSION

4.3.1 MORPHOLOGY

The maturity of the soils is visible in morphology which changes downwards and features as well developed horizons. This is an indication that vertical water movement down the profile contributes to soil formation in all the soils of this soilscape. The morphology of soils also changes down slope, especially in the subsoil, with an increase in redox morphology, indicating a strong interflow component (Table 4.1). Downslope the redox morphology changes to gley morphological features indicating saturation due to increased accumulation of interflow water. Gley morphology of near surface horizons of the Katspruit and Kroonstad soil indicates saturated conditions.

Table 4.1 Down slope arranged data of observation number (Obs), slope shape, soil, hydrological soil response type and TMU

Obs.	Slope shape*	Soil Form	Diagnostic horizons	Depth (mm)	Munsell colour	Hydrological response	TMU**
15	xl	Hutton (*Hu*)	Orthic A horizon	400	7.5YR 4/3	Recharge	1
			Red Apedal B horizon	2500	5YR 5/8		
			Rock	2500+	N/A		
14	xl	Hutton (*Hu*)	Orthic A horizon	300	10YR 4/4	Recharge	
			Red Apedal B horizon	1200	5YR 7/6		
			Saprolite	1700	N/A		
13	xl	Griffin (*Gf*)	Orthic A horizon	400	10YR 4/4	Recharge	3
			Yellow Brown B horizon	900	7.5YR 5/8		
			Red Apedal B horizon	1200	7.5YR 8/4		
12	ll	Mispah (*Ms*)	Orthic A horizon	300	10YR 5/3	Shallow responsive	
			Rock		N/A		

11	II	Hutton (*Hu*)	Orthic A horizon	300	10YR 4/3	Deep interflow	
			Red Apedal B horizon	1200	7.5YR 6/6		
			Saprolite	1400+	N/A		
P210	II	Bloemdal (*Bd*)	Orthic A horizon	530	10YR 4/3	Deep interflow	
			Red apedal B horizon	1200	7.5YR 4/6		
			Unspecified material with signs of wetness	1580	10YR 6/4		
			Saprolite	1750	2.5YR 2/6		
10	II	Tukulu (*Tu*)	Orthic A horizon	200	10YR 5/4	Deep interflow	
			Neocutanic B horizon	800	7.5YR 5/6		
			Unspecified material with signs of wetness	1400	10YR 6/3		
9	Ix	Mispah (*Ms*)	Orthic A horizon	400	2.5Y 4/3	Shallow responsive	
			Rock		N/A		
P209	Ix	Katspruit (*Ka*)	Orthic A horizon	450	10YR 5/2	Responsive	
			G Horizon	1100	10YR 6/1		
			Saprolite	1400	10YR 6/1		
8	Ix	Katspruit (*Ka*)	Orthic A horizon	300	2.5Y 5/3	Responsive	
			G Horizon	1300	2.5Y 6/3		
			Saprolite	1700	N/A		
7	Ix	Katspruit (*Ka*)	Orthic A horizon	300	2.5Y 6/2	Responsive	
			G Horizon	1200	2.5Y 7/4		
6	Ix	Katspruit (*Ka*)	Orthic A horizon	300	2.5Y 7/2	Responsive	
			G Horizon	1500	2.5Y 7/6		
5	Ix	Katspruit (*Ka*)	Orthic A horizon	300	2.5Y 5/3	Responsive	
			G Horizon	800	2.5Y 7/3		
4	Ix	Katspruit (*Ka*)	Orthic A horizon	300	10YR 5/3	Responsive	4
			G Horizon	1800	2.5Y 7/3		
P208	II	Kroonstad (*Kd*)	Orthic A horizon	350	10YR 6/1	Shallow-interflow responsive	
			E Horizon	550	10YR 6/2		
			G Horizon	1700	10YR 7/1		
3	II	Kroonstad (*Kd*)	Orthic A horizon	300	10YR 6/4	Shallow-interflow	

			E Horizon	600	7.5YR 7/6	responsive
			G Horizon	1700	2.5Y 7/4	
2	ly	Kroonstad (*Kd*)	Orthic A horizon	300	10YR 6/3	Shallow-interflow responsive
			E Horizon	600	2.5Y 6/4	
			G Horizon	1700	2.5Y 6/6	
1	yy	Kroonstad (*Kd*)	Orthic A horizon	300	5Y 5/2	Shallow-interflow responsive
			E Horizon	400	2.5Y 7/4	
			G Horizon	600	2.5Y 5/2	

*Slope shape: l = linear; x = convex, y= concave

**TMU (Terrain morphological unit) 1 = crest, 3 = midslope, 4 = footslope

The high chroma soils above the Molteno shelf (Obs 15,14,13) indicates an oxidising condition. The colour of the top soil in this zone is consistent with very little change down slope in value and chroma. The surface colour is generally a hue different from the subsoil but from 200mm downwards the colours are uniform. The subsoils change slightly in matrix colour becoming redder in colour. The *re* horizon of the *Hu* and *Gf* soils is uniform in colour with black mottles occurring in the bottom 50mm just above the hard rock, indicating that water is only stagnant for brief periods and that the underlying Molteno rock is permeable. Down slope the hard rock changes locally to saprolite (observation 14) and back to hard rock (observations 13 & 12). The *ye* horizon, occurring in the third observation as the second horizon in a *Gf* soil (observation 13), is also uniform. The chromas of the *ot* horizons are similar to that of the subsoils. Total soil depth varies but generally decreases down slope. Immediately upslope of the Molteno sandstone shelf it is shallow and a *Ms* soil on sandstone hard rock occurs. The well expressed shelf probably lies underneath the crest for most of the divide. Lack of redox morphology in the subsoils indicate that the shelf is fractured, probably associated with the intrusion of the nearby dolerite dyke. The soils lack prominent gley and redox morphological features, which were previously reported in nearby soilscapes in the catchment (Van Tol *et al.*, 2010).

The redox morphology in the C horizons indicates an interflow section below the Molteno sandstone shelf and a second sequence of soils. The *re* horizon of the *Hu* soil (observation 11) is on chemically weathered saprolite that changes downslope to a *Bd* soil with redox morphology at 800mm depth stretching to 1400mm depth on saprolite. The subsoil disappears downslope to form another *Ms* soil on dolerite. The *Bd* soil profile (P210) is representative of the interflow soils (Table 4.2, Figure 4.2). It is situated on the middle midslope just below the Molteno sandstone shelf. It has high chroma morphology with reduction morphology in the deep subsoil. This morphology is representative of the transitional zone between the higher lying oxidic red soils and lower lying reduced grey soils. The *re* horizon of the *Bd* soil (5YR 4/6) is sandwiched between a bleached *ot* horizon (10YR 3/2) and grey *on* (unconsolidated horizon with signs of wetness) horizon with no mottles.

Table 4.2 Bd (P210) profile description (Van Huyssteen *et al.*, 2005)

Horizon	Depth(mm)	Description	Diagnostic horizons
A1	0 - 260	Moisture status: dry; dry colour: 10YR4/2 (100 %); moist colour: 10YR3/2 (100 %); 19.0 % clay; clay loam; no mottles; moderate fine subangular blocky; slightly hard, loose, non-sticky, non-plastic; many normal fine and very fine pores; few normal medium and coarse pores; water absorption 2 second(s); many normal roots; gradual smooth transition;	Orthic A horizon
A2	260 - 530	Moisture status: dry; dry colour: 10YR4/3 (100 %); moist colour: 7.5YR3/2 (100 %); 16.7 % clay; loam; no mottles; moderate fine subangular blocky; soft, loose, non-sticky, non-plastic; many normal fine and very fine pores; few normal medium and coarse pores; water absorption 1 second(s); many normal roots; gradual smooth transition; Remark: This horizon forms the transition between orthic A horizon andre horizon.	Orthic A horizon
B1	530 - 810	Moisture status: dry; dry colour: 7.5YR4/6 (90 %), 7.5YR4/2 (10 %); moist colour: 5YR4/6 (90 %), 7.5YR3/2 (10 %); 23.9 % clay; loam; many medium distinct 7.5YR4/2 dry, 7.5YR3/2 moist, humus mottles; weak fine subangular blocky; slightly hard, loose, non-sticky, non-plastic; many normal fine and very fine pores; few normal medium and coarse pores; common 2-6mm round stones; very few 6-25mm round biocasts; water absorption 2 second(s); many normal roots; gradual smooth transition; Remark: Gravel layer at 830mm.	Red apedal B horizon
B2	810 - 1200	Moisture status: moist; dry colour: 7.5YR6/6 (90 %), 5YR4/6 (10 %); moist colour: 5YR4/4 (90 %), 5YR4/4 (10 %); 18.0 % clay; loam; many fine distinct 7.5YR4/2 dry, 7.5YR3/2 moist, humus mottles; apedal massive; hard, friable, non-sticky, non-plastic; many normal fine and very fine pores; few normal medium and coarse pores; common 2-6mm round stones; very few 6-25mm round biocasts; water absorption 3 second(s); common normal roots; gradual smooth transition; Remark: humus in channels between 600 and 1200mm.	Red apedal B horizon
C1	1200 - 1580	Moisture status: moist; dry colour: 10YR6/4 (100 %); moist colour: 7.5YR3/4 (100 %); 17.3 % clay; coarse sandy loam; no mottles; apedal massive; hard, friable, non-sticky, non-plastic; many normal fine and very fine pores; few normal medium and coarse pores; very few 2-6mm round stones; water absorption 2 second(s); few normal roots; clear smooth transition;	Unspecified material with signs of wetness
C2	1580 - 1750	Moisture status: moist; dry colour: 2.5YR3/6 (100 %); moist colour: 5YR6/2 (100 %); 20.1 % clay; loam; no mottles; strong coarse prismatic; hard, firm, sticky, plastic; few normal fine and very fine pores; few normal medium and coarse pores; common silica cutans; water absorption 7 second(s); no roots; transition not reached;	Saprolite

The *Ka* soil with a gleyed subsoil starts the third sequence of soils (observations 8 to 4). Gleyed horizons are associated with saturation for long periods and therefore react as responsive soils. Redoximorphic mottles qualify as abundant in the matrix of the *gh* horizon of the higher lying *Ka* soils on the midslope but vary from many to few, in the lower lying *Ka* soils. The *Ka* profile (P209) is representative of these observations (Figure 4.2; Table 4.3). It is situated below the dolerite dyke on a lower TMU 3 on mudstone underlying material. The transition between the *ot* and *gh* horizons is abrupt in terms of colour and texture (Figure 4.2). Cracks, slickenslides and cutans occur in the *gh* horizon. The clay content increases down the profile. $CEC_{clay,}$ within the *gh* horizon, increases downwards from 21.6 $cmol_c\ kg^{-1}$ to 40.3$cmol_c\ kg^{-1}$ indicating a shift in mineralogy from 1:1 towards 2:1 silicate clays. The *gh* horizon has a stagnic morphology (dark root channels in a grey matrix) in contrast to the gleyic morphology (bleached root channels in a higher chroma matrix) of the saprolite.

Table 4.3 Ka (P209) profile description (Van Huyssteen *et al.*, 2005)

Horizon	Depth(mm)	Description	Diagnostic horizons
A	0 - 450	Moisture status: moist; dry colour: 10YR5/2 (90 %), 7.5YR4/6 (10 %); moist colour: 10YR3/2 (90 %), 7.5YR3/4 (10 %); 16.3 % clay; loam; many fine distinct 7.5YR4/6 dry, 7.5YR3/4 moist, Fe-oxide mottles; moderate coarse granular; hard, slightly firm, non-sticky, non-plastic; many bleached and rusty fine and very fine pores; many bleached and rusty medium and coarse pores; fine cracks; very few 6-25mm mixed shaped stones; water absorption 2 second(s); many bleached roots; clear smooth transition; Remark: ∀70 m to dolerite outcrop. Rocks in 2 layers at 700 and 900mm.	Orthic A horizon
G	450 - 1100	Moisture status: moist; dry colour: 10YR6/1 (90 %), 7.5YR6/8 (10 %); moist colour: 10YR5/1 (90 %), 7.5YR5/8 (10 %); 35.7 % clay; clay loam; many fine prominent 10YR6/8 dry, 7.5YR5/8 moist, Fe-oxide mottles; strong coarse prismatic; very hard, very firm, very sticky, plastic; common bleached fine and very fine pores; few bleached and rusty medium and coarse pores; many slickensides; medium cracks; very many clay cutans; common 75-250mm mixed shaped stones; very few 2-6mm round Mn concretions; water absorption 6 second(s); few bleached roots; clear smooth transition; Remark: root channels are always dark gray (10YR4/1). Vertical streaking light to dark gray.	G horizon
C	1100 - 1400	Moisture status: moist; dry colour: 10YR6/1 (90 %), 7.5YR5/8 (10 %); moist colour: 10YR5/1 (90 %), 7.5YR4/6 (10 %); 35.7 % clay; clay loam; many fine distinct 7.5YR5/8 dry, 7.5YR4/6 moist, Fe-oxide mottles; strong coarse prismatic; very hard, very firm, very sticky, plastic; few bleached fine and very fine pores; few bleached and rusty medium and coarse pores; many slickensides; medium cracks; very many silica & clay cutans; many 75-250mm mixed shaped stones; water absorption 6 second(s); common bleached roots; transition not reached; Remark: C horizon is horizontally layered mudstone.	Saprolite

E (gs) horizons overly *gh* horizons in soils on the toeslopes of a *Kd* soil (Figure 4.2) (observations 1-3). A *Kd* soil (P208) is representative profile of this zone. Morphologically the soil is quite uniform and the only prominent morphological feature is Fe mottles found in the *ot* and *gs* horizons. There is an increase in the bulk density from 1.62 $Mg\ m^{-3}$ in the *gs* to 1.76 $Mg\ m^{-3}$ *gh* horizon.

The increased indication of saturation in the subsoils down slope is questioned by the abrupt transition of the Ka soil. This duplex morphology continues down slope in the Kd soils. Duplex character typically indicates a strong vertical flow component. However, the abruptness may rather be an indicator of a dominant lateral flowpath in the topsoil starting in soils as near surface macro pore flow as reported by Van Tol et al. (2012).

Table 4.4 Kd (P208) profile description (Van Huyssteen *et al.*, 2005)

Horizon	Depth(mm)	Description	Diagnostic horizons
A1	0 - 350	Moisture status: moist; dry colour: 10YR6/1 (100 %); moist colour: 10YR4/1 (100 %); 13.9 % clay; silty loam; common fine distinct 10YR5/8 dry, 10YR4/6 moist, Fe-oxide mottles; weak coarse granular; hard, friable, non-sticky, non-plastic; many rusty fine and very fine pores; few rusty medium and coarse pores; water absorption 1 second(s); many bleached roots; gradual smooth transition;	Orthic A horizon
E	350 - 530	Moisture status: moist; dry colour: 10YR6/2 (90 %); 10YR5/8 (10 %); moist colour: 10YR4/1 (90 %), 10YR4/6 (10 %); 13.1 % clay; loam; many medium distinct 10YR5/8 dry; 10YR4/6 moist, Fe-oxide mottles; weak medium subangular blocky; hard, friable, non-sticky; non-plastic; many rusty fine and very fine pores; few rusty medium and coarse pores; few clay cutans; water absorption 1 second(s); many bleached roots; gradual smooth transition;	E horizon
G1	530 - 800	Moisture status: moist; dry colour: 10YR7/1 (80 %), 10YR6/6 (20 %); moist colour: 10YR5/2 (80 %), 10YR4/6 (20 %); 23.7 % clay; loam; common medium distinct 10YR6/6 dry, 10YR4/6 moist, Fe-oxide mottles; apedal massive; very hard, firm, sticky, slightly plastic; many bleached fine and very fine pores; few dark medium and coarse pores; few clay cutans; water absorption 1 second(s); common bleached roots; clear smooth transition; Remark: This horizon forms the transition between B and G	G horizon
G2	800 - 1700	Moisture status: moist; dry colour: 10YR6/2 (70 %), 10YR5/8 (20 %), 10YR8/1 (10 %); moist colour: 10YR5/2 (70 %), 10YR4/6 (20 %), 10YR7/1 (10 %); 25.1 % clay; loam; many medium distinct 10YR5/8 dry, 10YR5/6 moist, Fe-oxide mottles; common medium distinct 10YR4/3 dry, 10YR4/2 moist, humus mottles; few fine prominent 10YR2/1 dry, 10YR2/1 moist, Mn mottles; strong coarse prismatic; very hard, very firm, very sticky, plastic; common normal fine and very fine pores; common dark medium and coarse pores; many slickensides; very many clay cutans; water absorption 10 second(s); few bleached roots; transition not reached;	G horizon

The *Bd* (P210) soil has a relatively uniform texture profile (Figure 4.2). The clay content varies between 12.5% and 23.5%. It decreases in the gleyed *C1* and increases in the *C2*. The clay in the *ot* horizon of the *Ka* (P209) soil (Figure 4.2) is uniform and there is a sharp increase in clay % from the *ot* to the *gh* and a steady increase in clay down the profile. In the *Kd* (P208) soil there is an initial increase in clay (Figure 4.2) from the surface in the *ot* with a decrease at the bottom of the *ot* horizon towards the middle of the *gs* horizon and a steady increase of clay from 350mm down the profile to 900mm. At 900mm there is a decreasing trend.

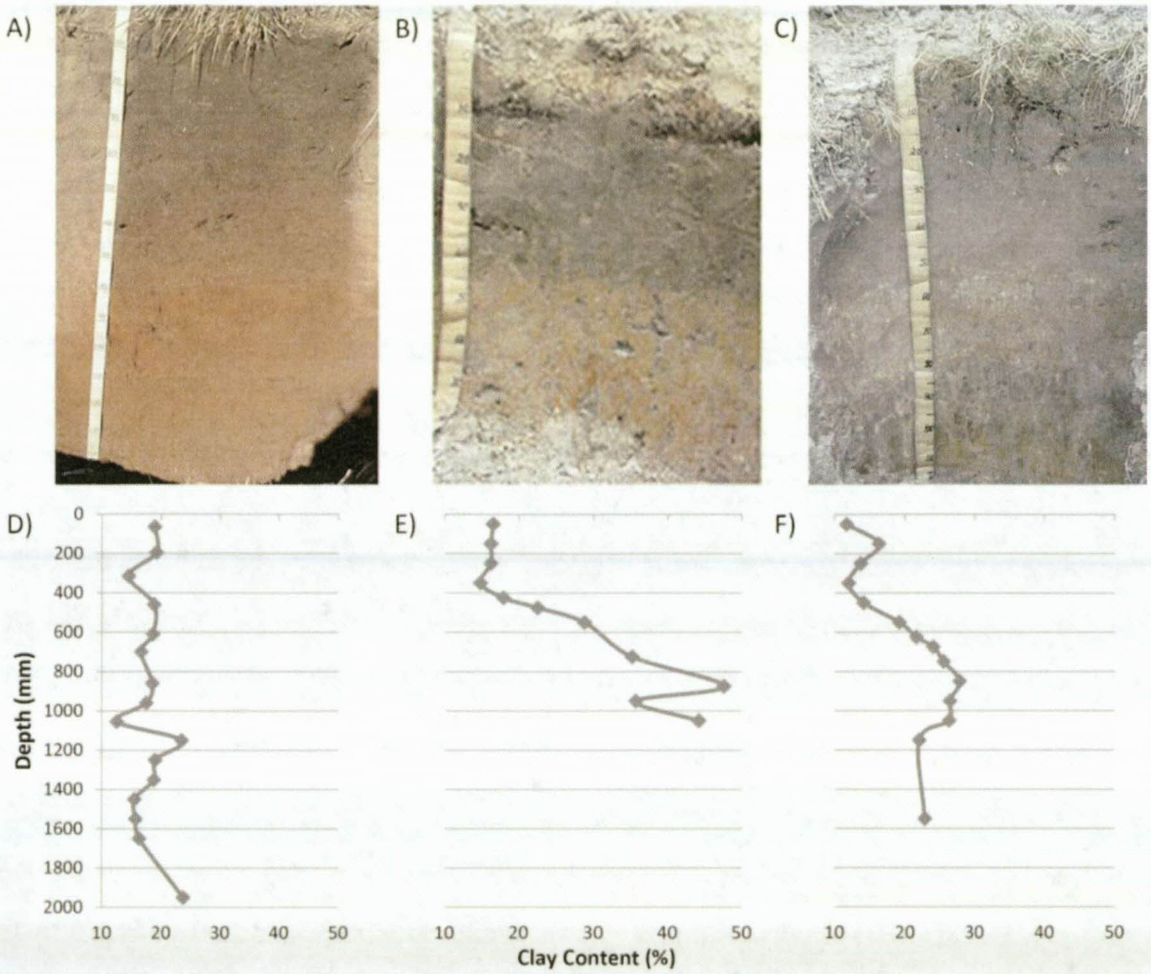

Figure 4.2 Morphology of A) Bloemdal soil (210), B) Katspruit (209), C) Kroonstad (208) and Clay Content of D) Bloemdal soil (210), E) Katspruit (209) and F) Kroonstad (208) (Van Huyssteen et al., 2005).

The oxidic morphology of the *Hu* and *Gf* soil forms indicates that both serve as recharge soils. They serve as conduits of infiltrated water and recharge underlying fractured bedrock. The hard rock/saprolite implies variation in degree of weathering in the rock and may be attributed to local variation in composition in the rock or local variation in water regime or both as the water regime is controlled by the fracture pattern of the rock.

The texture of the *Hu* soil show little pedogenetic differences and is therefore probably experiencing a uniform, draining pedogenesis of the sandy material inherited from the quartzitic Molteno sandstone underlying material (low potential to form clay). Contrary to other *ye* horizons in the catchment, with a wetter soil water regime than *re* horizons, overlying gleyed horizons, with lower permeability the *ye* horizon of the *Gf* soil can be due to extended saturation and/or increased reduction rate due to the higher OC content of the topsoil. The duration of water contents at near saturation for several days was measured in topsoils in other relevant research. However, quick reduction in the topsoil during prolonged soft rain as described by Fey (1985) could hydrate the Fe-oxides to form the yellow-brown colour. The implication is that the *re* horizons are not a limitation to vertical flow.

Water that infiltrates the soil causing water contents to exceed drained upper limit (*DUL*), will move through the soil, and infiltrate the bedrock. Via bedrock flowpaths this water can either recharge regional groundwater aquifers (vertical) or return to the solum downslope through lateral flow on bedding planes.

The morphology of the *Bd* (P210) soil (Figure 4.2) indicates that it serves as a deep interflow soil. This is the highest position in the landscape where subsoil saturation was detected and implies interflow originating from higher lying recharge soils returning from the bedrock to the deep subsoil. It is the transition zone from the recharge soils to the responsive soils. Water infiltrating this soil will move vertically in the *re* horizon which serves as a conduit until it reaches the impeding layer with lower K_s or a periodic water saturation, from where water will be diverted laterally. The increase in slope (from 3% to 8%) promotes interflow. The redox morphology of the *on* horizon is an indication of the increased duration of drainable water. Increased water contents in the *on* horizon are probably return flow from the higher lying recharge zone. By implication the impeding layer may be a perched water table. The bleached *ot* horizon may be related to the same pedogenetic processes as the *ye* horizon of the *Gf*. The lack of mottles in the *ot* horizon can be related to water movement downslope during reduction as near surface macro pore flow.

The dolerite dyke probably acts as a wall controlling interflow increasing lower vadose zone storage. The high storage capacity supply water to the deep interflow zone and soils of the responsive zone downslope.

The mottled, concretionary morphology of the *Ka* (P209) soil is indicative of periodic saturation to the surface during the peak rain season, implying that, after a short wetting up phase, it functions as a responsive soil. After rain events the profile is saturated and the topsoil drains laterally and dries out by ET, creating an alternating reduction/oxidation cycle and short wetting up phases to saturate to the surface hence the gleyed morphology and mottles. The abrupt A/G horizon transition is indicative of distinctly different soil formation in the *ot* and *gs* horizons which may be ascribed to luviation and therefore significant vertical water flow expected during periods of unsaturated conditions, ferrolysis acting under alternating redox conditions implying the soil water regime explained above and several other soil forming processes. The abundance of mottles indicates that the water table fluctuates in the *ot* horizon, with associated varying redox conditions. The more uniform matrix colour with less mottles indicates longer periods of saturation in the *gh* horizon.

The terrain of the *Ka* (P209) soil (TMU 3 and a steep slope of 13%) increases the possibility of a steady supply of water from a large aquifer. It is possibly related to the recharge zone and the dolerite dyke, about 50 m upslope, storing water. Water movement is more likely in the saprolite than in the *gh* horizon, the presence of bleached root channels in the saprolite is signs of water movement compared to dark root channels in the *gh* horizon. The *ot* horizon of the *Ka* soil is thicker other *ot* horizon in the soilscape (450mm) as the soil will not receive as much water on TMU 3 as it has a smaller source area when compared to the soils lower down in the landscape.

In the *Kd* (P208) soil, water that has infiltrated the soil is expected to move largely vertically through the profile till it reaches an impeding layer. The restriction of water movement would be caused by the decrease in K_s due to the increase of clay in the *gh* horizon or water saturation. The impediment can cause water to move laterally above the *gh* horizon in the *gs* horizon resulting in breakdown (ferrolysis) and eluviation of clays in the the *gs* horizon. The accumulation of water from upslope can result in excessive saturation in the *gh* horizon diverting water into the *gs* horizon and thereby promoting lateral and overland flow. Return flow to the soil implies that the water moves upwards under saturated conditions (besides vertical extraction by capillary rise). Water may exit the *gh* horizon and drain down slope in the sandier *gs* horizon, increasing the redox variation.

A conceptual hydrological response model of the soilscape constructed from deductions made from morphological data (Table 4.1) is presented in Figure 4.3. In the recharge zone on the TMU 1, which consists of *Hu* and *Gf* soil forms, water is expected to move vertically through the *re* and *ye* horizons. The prominent shelf of the Molteno formation plays an important role in the hydrology of the catchment but not in this soilscape. The abrupt transition to an interflow zone is rather related to topography created by the shelf on the TMU 3 position. There is an increase in the slope gradient below the shelf. The deep interflow zone of the soilscape immediately below the shelf consists of *Bd*, *Tu* and *Hu* soils, with redox morphology in the deep subsoils of the *Bd* and *Tu* soils and chemically weathered saprolite in the *Hu* soil. The deep interflow soils cover the area down to the dolerite dyke, which seems to influence the hydrology of the lower sections of hydromorphic soils. This water from the recharge zone that moves through the fractures and the subsoil of the deep interflow soils and keep the lower lying hydromorphic soils saturated creating the gleyed horizons found on the lower midslope and the bottom of the soilscape. These horizons will act more as a storage mechanism which is in contrast to the latter which are flowpaths. The shallow-interflow responsive *Kd* soil will have interflow in the *gs* horizon. Interflow in the *gs* horizon may be increased from the soils higher up. The environmental setting suggests that returnflow that keep the *gh* horizon wet may enter the *gs* horizon and contribute to the interflow. By implication the *gs* horizon controls the water level in the soil as it has a high saturated hydraulic conductivity compared to the *gh* horizon (Van Tol *et al.*, 2012).

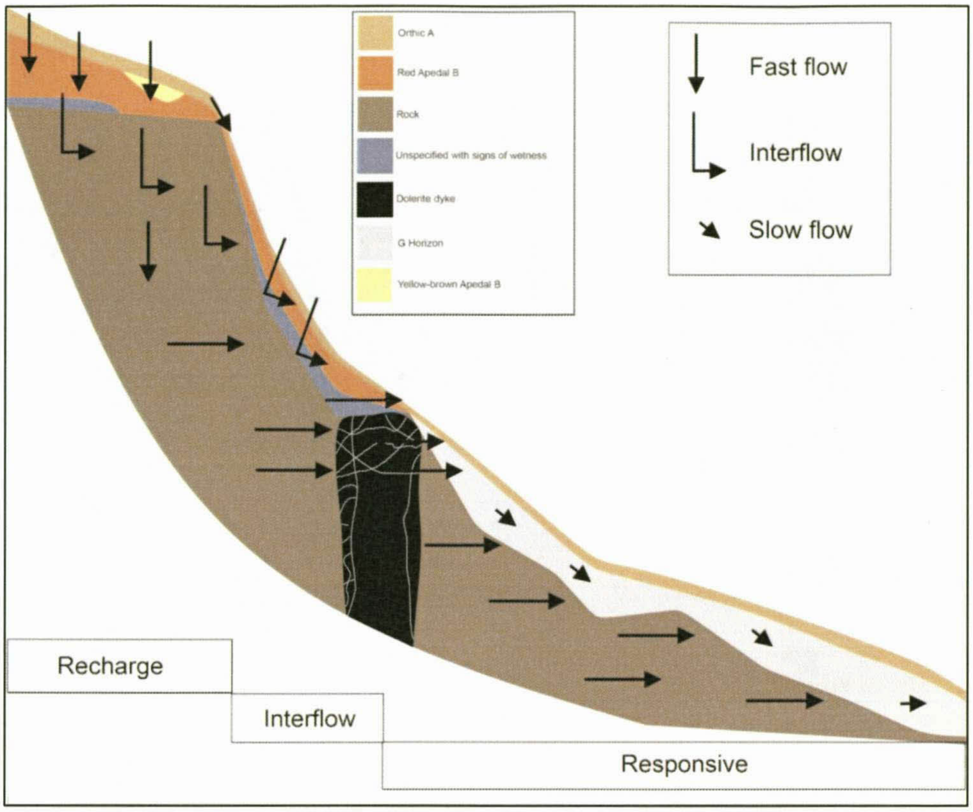

Figure 4.3 Conceptual hydrological response model of soilscape based on morphology.

4.3.2 CHEMISTRY

The chemical parameters largely relate to the morphology of the genetic horizons. Some variation cannot be explained and is assumed to be inherited from the parent material. The base saturation is from the 15 observations made and the profiles from Van Huyssteen *et al.* (2005).

Base saturation profiles and distribution down slope shows that the recharge zone is a leaching environment and basic cations are leached from the surface horizons and increase concentration with depth, indicating that there is vertical movement of water transporting the cations (Figure 4.4). In the interflow section there is removal of basic cations in the top horizons or the entire profile, indicating an element of recharge, and an increase of bases in the subsoil that supports the deduction of horizontal movement and arrival of enriched water. The responsive soils typically have higher base saturation which infers that enriched water arrives in this zone and the water is stagnant and enriched further by ET when water is removed and base cations are deposited into the soil. Cations accumulate in these soils.

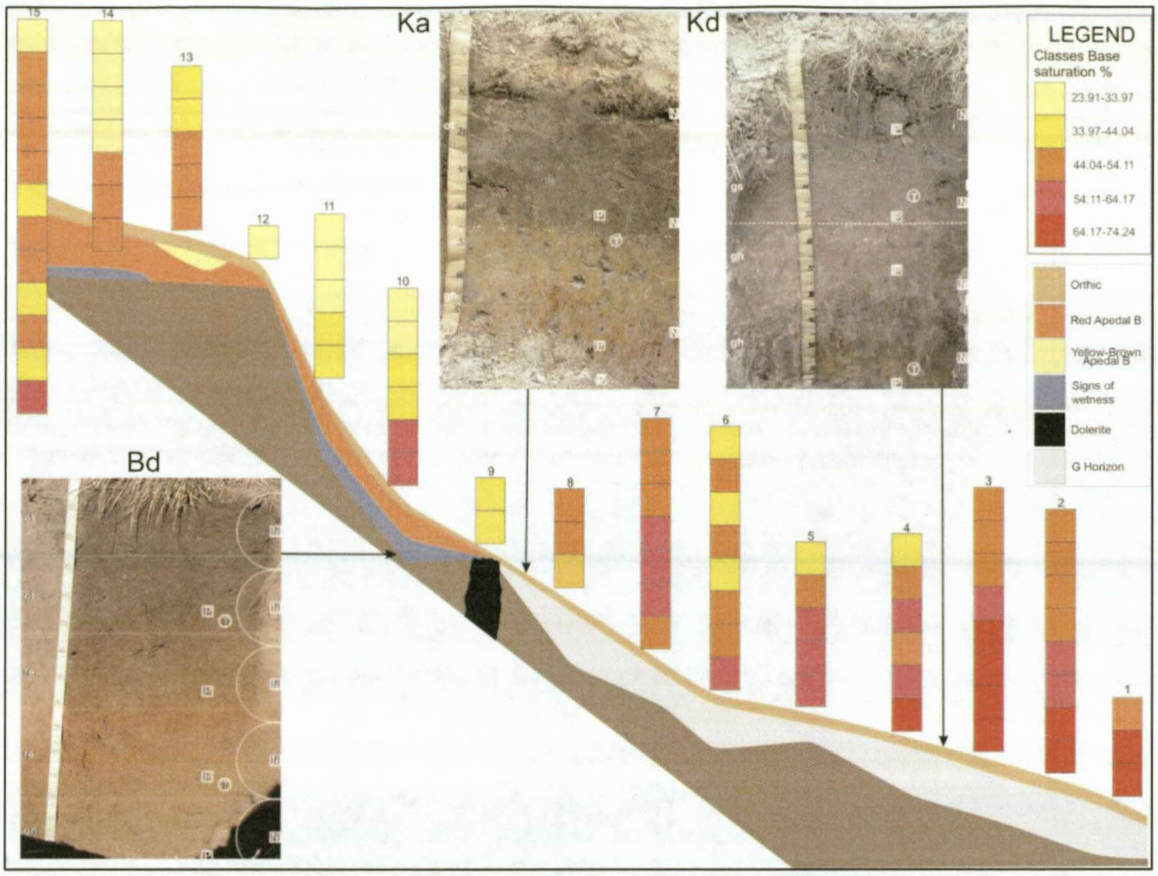

Figure 4.4 Base saturation distributions within profiles and in relation to soilscape position.

The base saturation of the representative profiles (Figure 4.5b) of the soilscape vary between 13% in the *re* horizon of the *Bd* soil and 130% in the *gh* horizon of the *Kd* soil on the toeslope. There is generally an increase in base saturation down the profile and increases of base saturation in the profiles with a distance from the crest with the exception of the *gs* horizon in the *Kd*. The soils on the crest of the soilscape have a lower average base saturation (20.68%) than the soils in the toeslope of the soilscape (69.92%). In the *Bd* soil the base saturation at the surface is low (22.8%) (Hazelton and Murphy, 2007) and decrease with depth to 500mm ($y = -0.0258x + 24.395$) were it becomes constant and there is a slight increase at 800mm and then relatively constant till 1010mm. At 1010mm there is a spike in the concentration of base cations just above *on* horizon. From 1010mm till 2000mm the base saturation is inconsistent but a general increase with depth ($y = 0.0049x + 18.377$). The surface at *Ka* soil has a moderate saturation (52.7%) (Hazelton and Murphy, 2007). At 450mm there is increase in base saturation to 900mm ($y = 15.8x + 59.667$). There is a sharp decrease the last 200mm of the profile. In the *ot* and *gs* horizons of the *Kd* soil at the surface there is a low Base saturation (26.6% - 43.4%) (Hazelton and Murphy), and steadily increases down the profile ($y = 0.0445x + 34.09$).

The pH of the soil in the soilscape varies between 5.09 and 7.13 (Figure 4.5a). There is generally an increase of pH with depth and an increase downslope. The interflow soils (*Bd*) have lower pH than the responsive soils (*Ka* and *Kd*). The *Bd* soil has a comparatively low, slightly increasing, pH throughout. The pH in the *Ka* soil increases the most of the profiles and the increase is systematic (Figure 4.4). There is a slight deviation at the *ot/gh* transition. The pH of the *Kd* soil is generally less acidic (Figure 4.4). The pH decreases in the *ot* horizon to reach a minimum at transition to the *gs* horizon. It increases sharply in the morphological *gs* and *gh1* horizons. At 800mm the pH increases slower down to 1550mm. This trend increases then decreases in the same area above the gh as the pH in the profile.

Calcium and Mg trends (Figure 4.5c&d) in the *Bd, Ka* and *Kd* soil are very similar and can be described together. The concentrations are lowest in the well drained horizons and the horizons with a low pH. The *Bd* soil there is a drop in the concentration from the surface and it stabilizes in the *re* horizon but spikes in the transition of the *re* and *on* horizons. There is an increasing trend in the *on* horizon towards the bottom of the profile. The Ca in *Ka* soil is very constant in the *ot* horizon but there is a sharp decrease in transition to the *gh* horizon. The concentration increases with depth to 900mm and then decreases to 2.83 $cmol_c kg^{-1}$ at the bottom of the profile. The Mg is similar but the concentration is constant in the whole *ot* horizon. In the *gh* horizon it increases to 900mm. There is initially a decrease in Mg similar to Ca but then spikes again at the bottom of the profile. In *Kd* soil there is a decrease in Ca and Mg from the surface to 350mm in the *gs* horizon. This is the same as the pH and base saturation found at these depths. The concentrations then increase to 1000mm and stabilize.

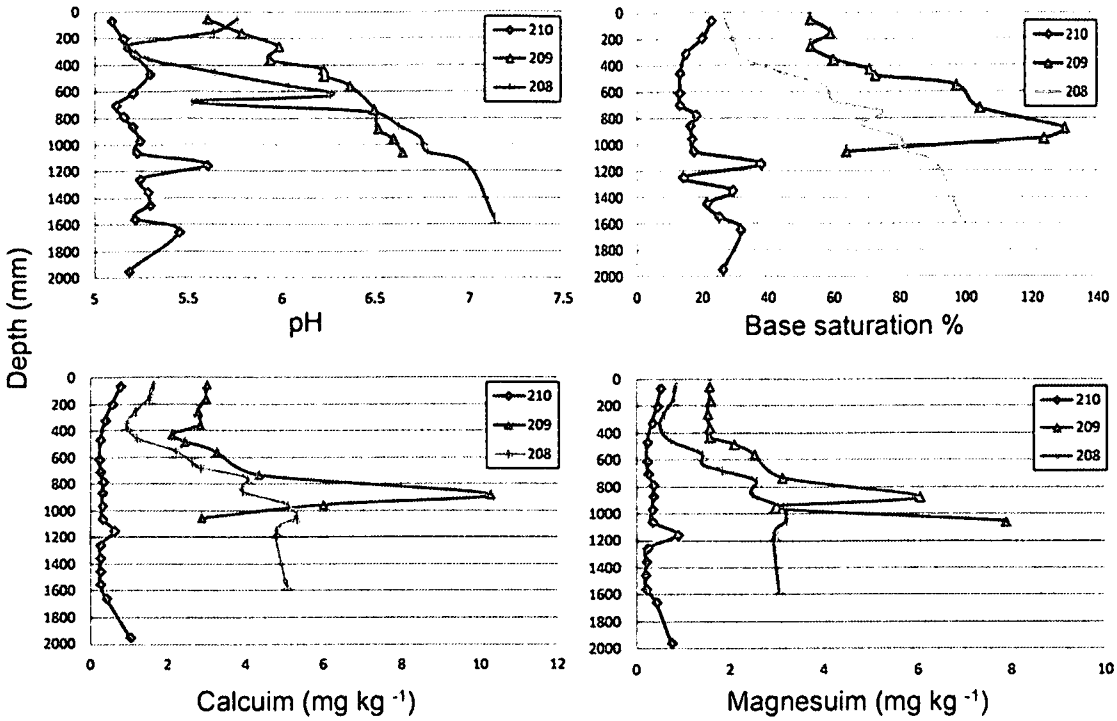

Figure 4.5- a) pH, b) base saturation, c) Ca d) Mg contents of Bl (P210), Ka (P209) and Kd (P208) soils (Adapted from Van Huyssteen *et al.*, 2005).

The general trend of vertical leaching in the topsoils decrease down slope as indicated by all parameters. This is an indication that the whole profile of shallow-interflow responsive and responsive soils are influenced by enriched water. The increase in all parameters with depth is an indication that some rainwater may enter the profile in the wetting up phase at the beginning of the rainy season and during a draining phase at the end of the rainy season when these soils are not saturated.

The values for all parameters are higher in the *Ka* soil than the Kd soil. It is an indication that the water impacting on the Ka soil is more enriched. As more enrichment is expected in a soil flowpath, compared to a fractured rock flowpath, it may indicate that the source for the Kd soil is dominantly the lower vadose zone contrary to the shallow-interflow responsive Kd soil that is the outflow of water from a source interacting with the soil.

The spike in the Bd soil at 1100mm depth is expressed in all parameters (Figure 4.5). It is associated with the reduction morphology of the underlying horizon (Table 4.2).

The low pH of the *ot* and *gs* horizons of the *Kd* soil is related to acidification by ferrolysis linked to redox conditions. These processes play a role in the *ot* of the *Ka* soil explaining the low values of all the parameters (Figure 4.5) and the low clay content (Figure 4.2e).

The pH is related to the soil morphology. The low pH in leached horizons (*re* & *gs*) and high in gh horizons.

The pH trends in the *Bd* soil indicate a general acidification and leaching in the *ot* horizon attributed to acid weathering. It is an indication that the soil is seldom water logged (Phillips & Greenway, 1998). The water, by implication, doesn't arrive enriched or stagnate for long enough to be enriched but flows through the soil. The extreme variation in pH gradients in the transition between *ot* and *re* horizons at 400mm to 800mm cannot be explained. A continuation of the trend in the *ot* horizon is expected in the *re* horizon. The bleached character of the *ot* horizon may be related to redox conditions supporting these pH gradients.

The systematic increase in pH in *Ka* soil (P209) (Figure 4.6) is correlated (R^2 = 0.7443) with an increase in clay. It can be related to water logged condition and water saturation which is expected to increase with depth.

The pH in the *Kd* soil (P208) is consistent with water table fluctuations predicted by the morphology. The major systematic deviation from the trend in the *gs* and *gh1* horizons is due to ferrolysis caused by fluctuating redox conditions mostly in the *gs* horizon but also in the *gh1* horizon. The decrease in pH and gradient of decrease in the surface to 300mm depth cannot be explained by acid weathering but relates to the process of ferrolysis caused by intermitted redox. A peak minimum at 300mm is an indication of ferrolysis activity peaking where the water table often meet oxygen supply. Acidification and ferrolysis is less down the profile which could be expected as *gh2* horizon will be saturated more consistently.

The low base status in the *ot* and *re* horizons of the *Bd* soil is caused by a leaching environment. The difference in trend cannot be explained. The accumulation of cations just above the *on* horizon indicating a section in the profile were the leached base cations accumulate by capillary rise and root extraction in the profile. A degree of water stagnation and arrival of enriched water is expected at this transition but still considered a faster flowpath than the *gh* horizons.

The base saturation profile of the *Ka* soil support the interpretations of the pH profile. The variation in the *ot* horizon may relate to the interaction between acid weathering and ferrolysis interaction driven by rain water and soilscape water entering the soil. The sharp decrease the last 200mm of the profile; this can be due to (contradictory as it has high clay (44.25), CEC soil (17.835) and CEC clay (40.3) the parent material which is Molteno mudstone which is low in basic cations.

The base saturation profile of the *Kd* soil is indicative of a leaching process (Figure 4.6).

Calcium and the Mg are correlated in all three representative profiles (r^2= 0.6347) implying similar behaviour. They largely follow base saturation as expected. In the *Bd* soil the data support the small accumulation at the surface typical of biocycling and the peak at 1150mm depth attributed to a water table. The lower re horizon may experience interaction between vertical leaching with relative "young" water and lateral flow with older water.

In the *Ka* and *Kd* soils the Ca and Mg profiles support the base saturation distribution and interpretation.

Iron values are high and variable in all profiles. There is an increasing trend from the surface to the bottom of the *re* horizon. The spike at 1250mm is just above the spike in bases (Figure 5). There is a decrease in the *on* horizon with an increase in the saprolite.

In *Ka* soil there is a systematic decrease in Fe content from the surface to 250mm where there is a sharp increase in the lower half of the *ot* horizon (Figure 4.6). The Fe in the *gh* horizon in *Ka* soil is relatively constant and very low with a sharp increase in the last sample.

The Fe content in *Kd* soils' *ot* and *gs* horizons is low and decreases systematically. Of all horizons in the soilscape only the *gh* horizon of the *Ka* soil has a lower Fe concentration (Figure 4.6). There is a relative systematic increase in Fe in the *gh* horizon till 1000mm. There is a sharp decrease and then increase to the highest concentration in the profile (359.5 mg kg^{-1}) and then decrease again at the bottom of the profile.

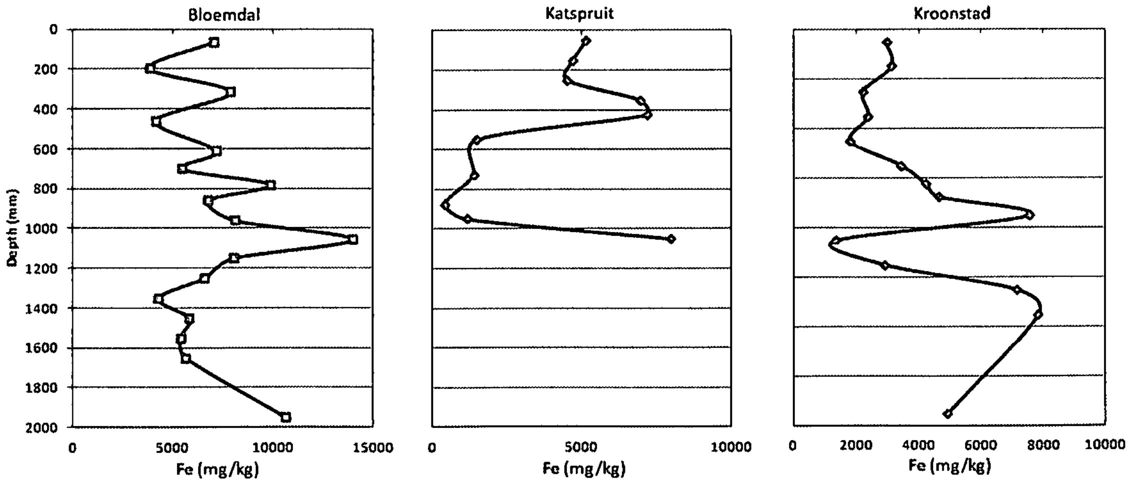

Figure 4.6 Iron distribution of Bd (P210), Ka (P209) and Kd (P208) soil (Adapted from Van Huyssteen *et al.*, 2005).

The Fe profiles are supporting the pedogenetic processes indicated by morphology. This data also emphasises the hydrological differences between diagnostic horizons.

The results indicate that acid weathering under oxidising conditions is responsible for the Fe accumulation in the *Bd* soil. The peaks in Fe concentrations at the transition between a reduced *on* horizon and an oxidised *re* horizon are signatures of redistribution of free Fe.

The Fe profile of the *Ka* soil indicates that redistribution of Fe occurs by being driven by diffusion and capillary rise. Reducing conditions in the *gh* horizon is related to stagnation, interflow of soil water in contact with soil reducing conditions higher up or reduced old water residing in the soilscape, or a combination of them, which could be responsible for dissolving Fe. Diffusion is rather active over short distances and could be responsible for transporting of Fe and precipitation in the bottom of the *ot* horizon under oxidising conditions. Capillary rise could be responsible for transporting the dissolved Fe to the *ot* horizon. The high Fe concentration in the deepest sample is an indication of Fe release by mineral weathering of saprolite. This is supported by the drop in the adjacent sample indicating a transition of less than 200mm between mineral weathering in the saprolite and *gh* horizon depletion of Fe processes.

The Fe concentration profiles of the *Kd* soil indicates slight accumulations in the *ot* horizon compared to the *gs* horizon which indicate capillary rise in Fe during the reduction stage of the *gh* horizon and oxidising state of the *ot* and *gs* horizons.

The Mn contents (Figure 4.7) show that the more freely drained horizons are more predictable and less erratic than the gleyed horizons. The *Bd* soil initially has a general increase of Mn ($y = 0.1178x - 8.9095$) in the *ot* and *re* horizons (Figure 4.6). Mn is relatively consistent in the *on* horizon ($y = -0.085x + 216.5$), concentration then decreases in the saprolite which starts at 1580mm and increases to its highest concentration (145.5 mg kg^{-1}).

The *ot* horizon of *Ka* soil has a decreasing trend of Mn ($y = -0.0696x + 156.96$) (Figure 4.6) in which is contrary to the other profiles. The concentrations of Mn in the *gh* horizon are all relatively low and inconsistent ranging from 95.5 mg kg^{-1} to 5.5 mg kg^{-1}.

In the *Kd* Mn is almost completely removed from the *ot* and *gs* horizons but then accumulates in the *gh* horizon with a sharp decrease at the bottom of the *Kd* (Figure 4.6). The drastic change from lowest concentration of Mn (1.5 mg kg^{-1}) to the highest measured in the three profiles at 1100mm (380 mg kg^{1}).

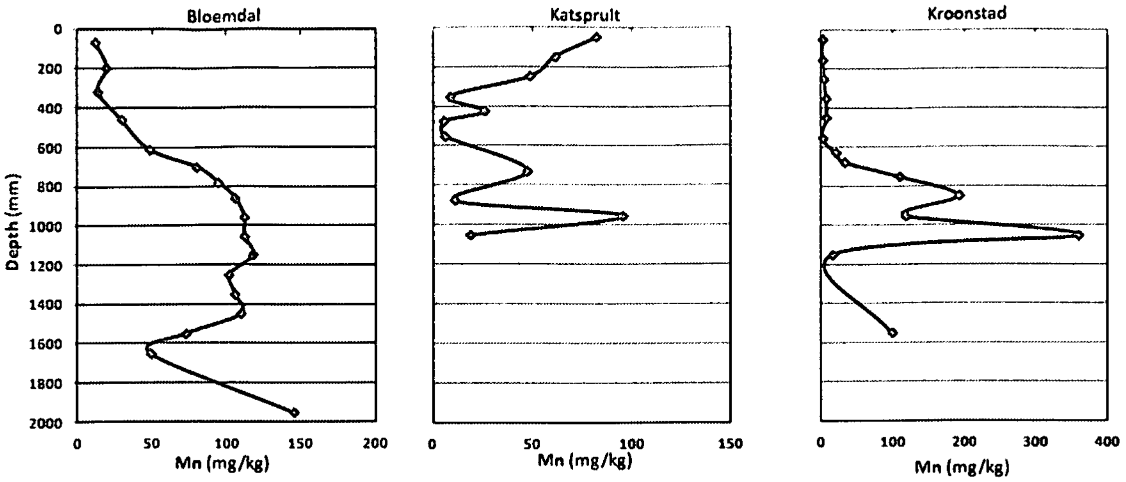

Figure 4.7 Manganese distribution of Bd (P210), Ka (P209) and Kd (P208) soils (Adapted from Van Huyssteen *et al.*, 2005).

In the *Bd* soil the Mn is low at the surface and increases systematically to the *on* horizon which is an indication of vertical movement of water and Mn being released by acid weathering. Manganese which is reduced during brief periods of saturation can be translocated downwards in the profile and then accumulating below drained upper limit were oxidising conditions are prevailing. In *Kd* soil it is possible that the Mn on the surface is organically bound and therefore as carbon decreases so does Mn. If reducing conditions are present as expected in a *gs* horizon and water movement is present there will be removal of Mn from that horizon. There is less Mn in the *gh* horizon compared to the *gh* horizon of the *Ka* soil which would indicate longer periods of saturation. The Mn concentrations fluctuate therefore indicating water movement in the profile were Mn is translocated and not leached, or if it is leached it is replaced by an upslope source.

The *Bd* soil have low Mn/Fe ratios (Figure 4.8) throughout the profile with a very slight increasing trend downwards (y = 7E-06x + 0.0042). The *Ka* soil has a decreasing trend in the *ot* horizon (y = -4E-05x + 0.0183) and then is inconsistent in the *gh* horizon with a few peaks. The Mn/Fe ratio in the *Kd* soil is very low in the *ot* and *gs* horizons, and increases dramatically in the *gh* horizon (y = 7E-05x - 0.0378) to 800mm and decrease in the *gh* horizon (y = -0.0001x + 0.1953).

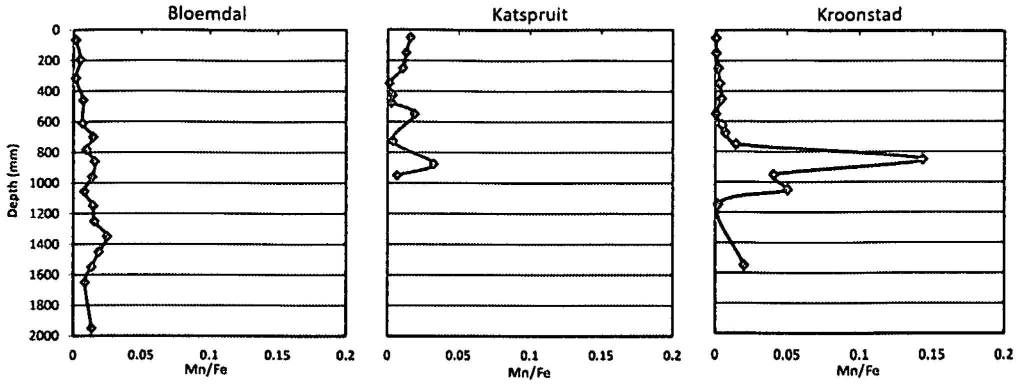

Figure 4.8 Mn/Fe ratios of Bd(P210), Ka(P209) and Kd (P208) soils (Adapted from Van Huyssteen *et al.*, 2005).

In the *Bd* soil the Mn/Fe ratio indicates a homogeneous pedogenetic impact on the parent material expected from oxidising weathering conditions (Figure 4.8). The systematic decrease in the ratio at 1350mm to 1650mm is indicative of a change in redox conditions to more reducing conditions, which could be related to the variation in water table depth during the season and/or a more reduced condition deeper in the water table. The more reduced condition is related to aging due to time spent draining from the recharge zone down slope.

The systematic decrease of the Mn/Fe ratio in the top 300mm of the *ot* horizon of the *Ka* soil can be related to a redox gradient supporting Mn mobilisation. The low values of the next 300mm are due to both depletion of Mn (Figure 4.7) and accumulation of Fe (Figure 4.6). Depletion of Mn varies slightly in these three depths and Fe accumulation is very high in the first two samples and much lower in the third relating to Fe accumulation in the bottom of the ot horizon and initial plinthite mottle formation indicating a minor flowpath on top of the *gh* Horizon. The inconsistency in the *gh* horizon could be due to both elements being mobile under saturated conditions, which is expected in the *gh* horizon and therefore Fe and Mn will be depleted in some areas and deposited in others.

The ratio in the *Kd* soil confirms the conclusions that were deducted by the morphology that water movement will occur in the *ot* and *gs* horizon and that the *gs* horizon will be saturated for long periods. This is supported by the low Mn/Fe ratio, which indicates that the horizons have been saturated and that the Fe and Mn have been reduced and removed from the horizons.

4.3.3 HYDROMETRICS

In the *Bd* (P210) soil (Figure 4.9) the sensor the *ot* (400mm) dried out during the first 10 days of the study period. The small rain events on 8, 9 and 10 March caused a slight decrease in the matric pressure but it was only after the large event of 47mm that the matric pressure in this horizon dropped significantly. The matric pressure increased in the following 5 days. The pressure dropped following rain events on the 17th and 18th of March and again on the 24th and 25th of March. In between these events there was a steady increase (drying out) in the pressure due to ET and vertical drainage.

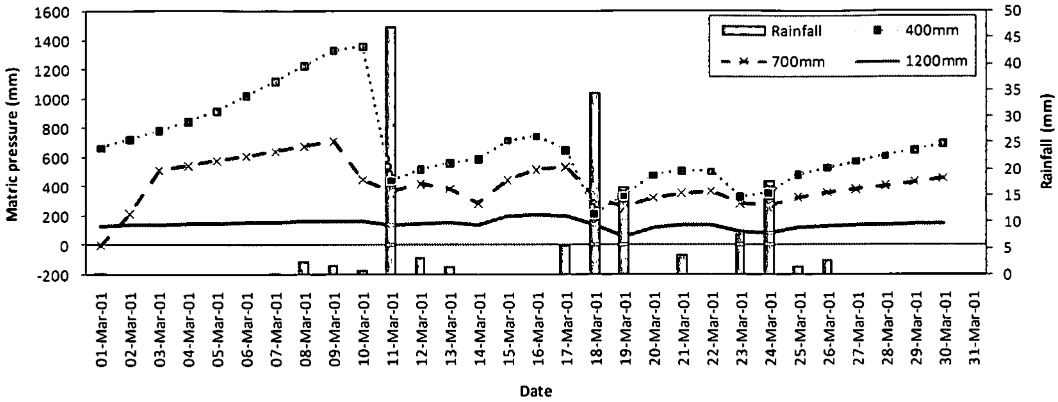

Figure 4.9 Rainfall and matric pressure measured in the *Bd* (P210) soil during March 2001.

The sensor at 700mm in the *re* horizon reacted similar to the sensor in the *ot* horizon but the gradients in increasing and decreasing pressures are less, probably due to the absence of evaporation in this horizon. The sensor also shows a lag time of approximately 3 days in its response to the large event on the 11th of March. The sensor at 1200mm is at the transition to from the *re* horizon to the *on* horizon. This sensor differs greatly from the above two sensors in that this sensor is relatively constant and close to 200mm in matric pressure throughout the study period. There is a slight decrease in pressure after with the event on 11th and the pressure increases from the 14th to 16th but decrease with the events on the 17th, 18th and 19th. The pressure then returns to the constant pressure observed in the beginning of the month.

The absence of saturation (matric pressure < 0mm), suggests that this profile is freely drained to a depth of 1 200mm. Even with large rain events (11th of March) the *ot* and *re* horizons are capable of transmitting this water to deeper horizons without reaching saturation, thereby supporting the morphological interpretations regarding these horizons.

In the *Ka* (P209) soil (Figure 4.10) the sensor in the *ot* horizon (300mm) presents an initial increase in pressure following a rain free period. It decreased after the series of small rain events from 7 to 10 March and reaches saturation after the large event on the 11th of March. This horizon remains close or at saturation for the remainder of the study period.

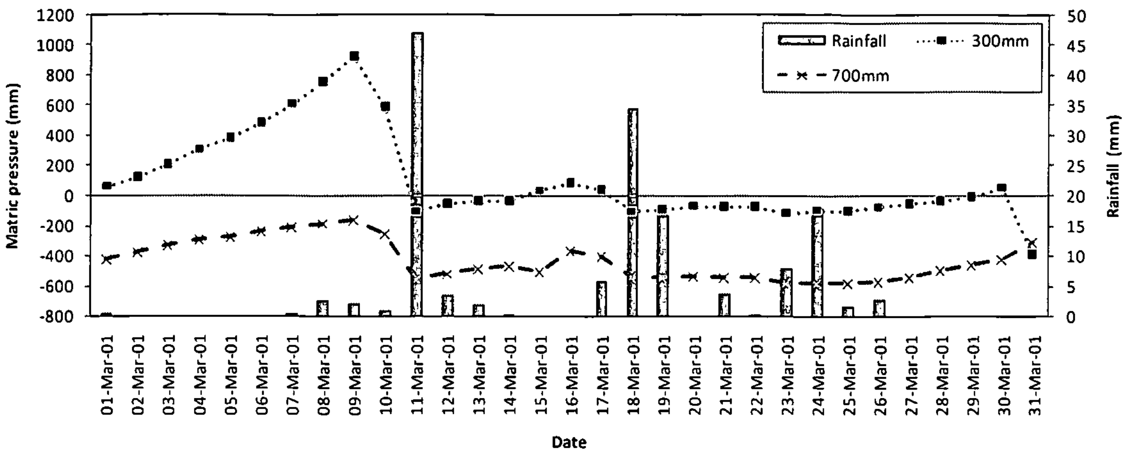

Figure 4.10 Rainfall and matric pressure measured in the *Ka* (P209) soil during March 2001.

The sensor at 700mm in the *gh* horizon remains negative throughout the study period. Slight variation is noted after the large rainfall event (11th of March), where after this horizon remains at approximately – 500mm of matric pressure. The saturated state of both the *ot* and the *gh* horizon during the majority of the study period imply that additional rain will not be able to infiltrate, overland flow will therefore be generated due to saturation excess. Near surface macropore flow can also be generated in the *ot* horizon as a result of a rising water table.

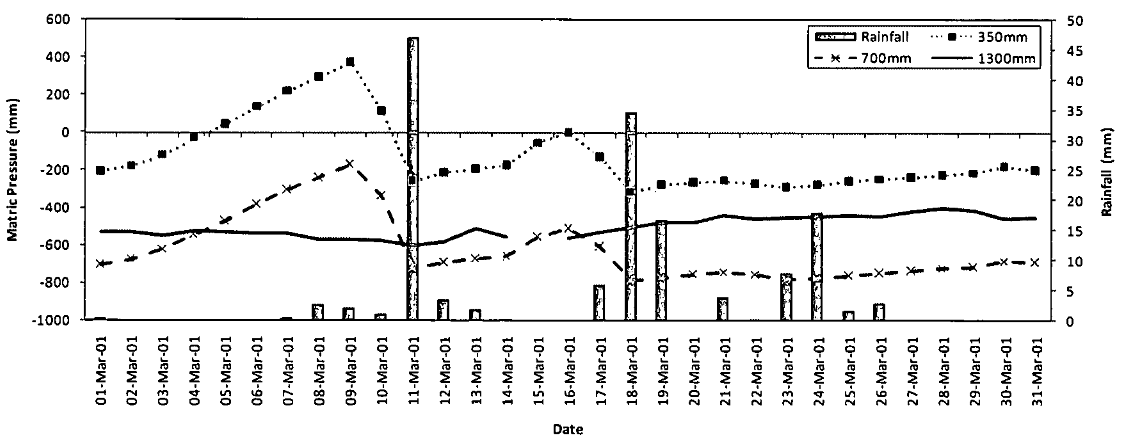

Figure 4.11 Rainfall and matric pressure measured in the *Kd* (P208) soil during March 2001.

In the *Kd* (P208) soil (Figure 4.11), the *ot* horizon (350mm) was saturated in the beginning of the study period (matric pressure < 0mm) but dried out due to evapotranspiration. It responded to the relatively small rain events on the 9th and 10th of March and reached saturation between the 10th and 11th of March. The *ot* horizon remained saturated throughout the rest of the study period. The sensor at 700mm representing the upper part of the *gh* horizon responded very similar to the *ot* horizon, implying that: 1) there was rapid vertical delivery of water through the *ot* to the *gh* horizon and/or 2) similar lateral flow processes control the water contents of these two horizons. Towards the end of the study period these horizons did not show the same drying slopes as in the beginning of the event. It is postulated that lateral contribution from upslope kept these horizons wet during this time. The deeper parts of the *gh* horizon remained saturated throughout the study period with very little fluctuation, suggesting that there is a constant feed of water from higher lying terrain positions.

In the *Bd* (P210) soil (Figure 4.11) shows that the *ot* horizon is saturated for brief periods and the *re* horizon to 1000mm is only saturated for 20% of the year at 70% saturation. At the bottom of the *re* horizon and the transition to the *on* horizon there is an increase in $ADs_{>0.7}$ and the horizon is saturated for most of the year. At this depth in *re* horizon that indicated an increase in pH, base saturation, Ca, Mg, Fe and Mn. This is an indication that the *re* horizon is out of phase with the morphology. The *on* horizon is also saturated for more than 6 months during the year.

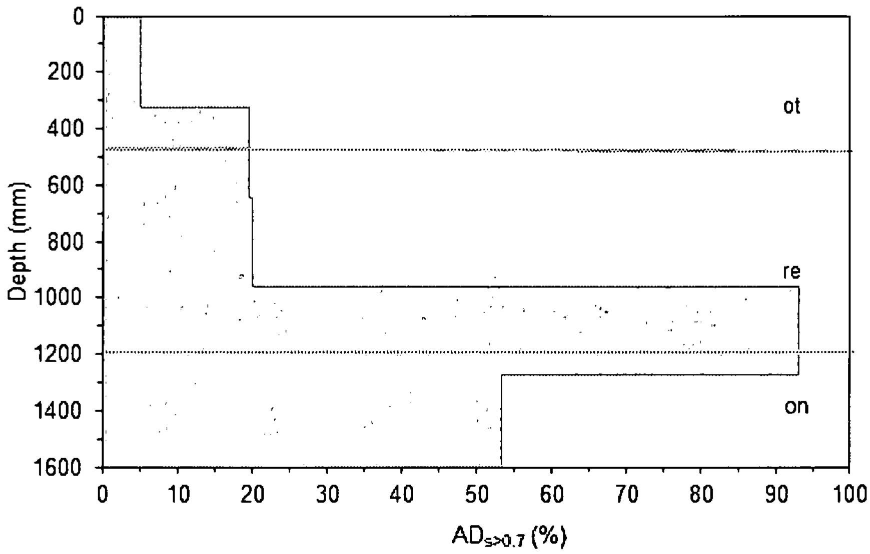

Figure 4.12 $AD_{s>0.7}$ of Bloemdal (Bd) (Adapted from Van Huyssteen *et al.*, 2005).

The *Ka* (P209) soil (Figure 4.12) is saturated for more than 6 months to 300mm and then saturated for nearly the entire year from the bottom of the *ot* horizon and the transition to the *gh* horizon. Due to the slope (8%) if the *gh* horizon is saturated water will move laterally on the *ot/gh* interface. This is a response expected in a *Kd* soil form with a gs horizon. This response was not notable in the morphology but was deduced from the chemical properties and now supported water contents.

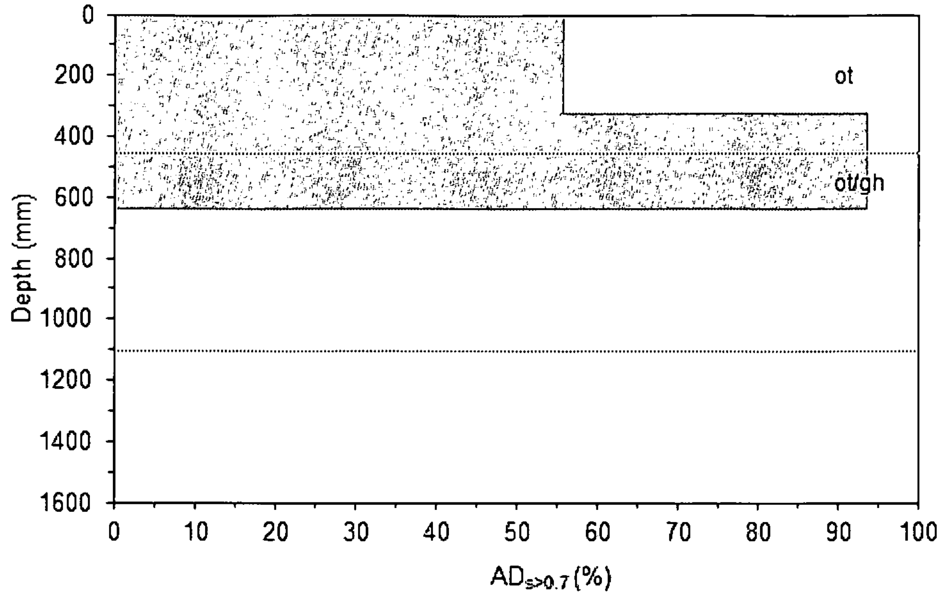

Figure 4.13 $AD_{s>0.7}$ of Katspruit (Ka) (Adapted from Van Huyssteen *et al.*, 2005).

The *Kd* (P208) soil form (Figure 4.13) is has very high $ADs_{>0.7}$ values for all horizons. The *ot* horizon is saturated for almost 6 months a year with increases in the *gs* horizon and the transition to the *gh* horizon. The *gh1* horizon is saturated considerable less than the *gh2* horizon which is saturated for the entire year. The high saturation is expected due to the toeslope position of the soil.

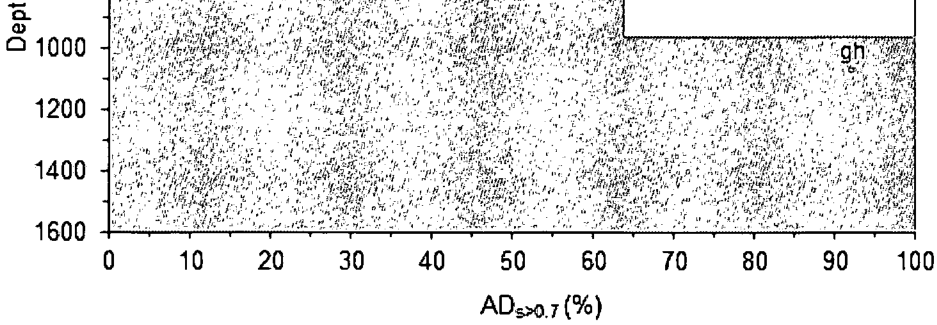

Figure 4.14 $AD_{s>0.7}$ of Kroonstad (Kd) (Adapted from Van Huyssteen *et al.*, 2005).

4.4 CONCLUSIONS

Soil morphology was used to successfully describe the hydrological response of a soilscape in the Weatherley catchment. Soil chemical properties confirmed the hydropedological processes derived from the observed morphology and were validated by hydrometrics. Soil chemical properties are therefore an important tool for hydropedology in the future and can help validate conceptual models based on morphology relatively quickly and inexpensively using conventional soil survey data.

Base cations and pH are very important indicators of pedological processes and corresponded very well with the morphology. In the studied soilscape three distinct zones can be identified based on the chemistry namely recharge zone characterised by low pH and leaching of base cations, interflow zone characterised by low pH and the removal of base cations from specific horizons and responsive zone characterised by higher pH and accumulation of base cations. Hydrological processes can therefore be deduced from the chemical properties. The chemical properties at profile scale also corresponded to the observed morphology.

Identification of point locations for hydrometrical instrumentation can be assisted by interpretation of soil morphology supported by soil chemistry. Soil morphology and chemistry can act as a transfer vehicle of point-scale data obtained from soil moisture content measurements. This is especially true for the extrapolation of point data, vertically in the horizons of the soilscape and also horizontally between different soil types. It can also be used to indentify representative soilscapes in a larger catchment considering the same soil forming factors by indentifying similar trends in the same soil types.

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GENERAL CONCLUSIONS

Soil properties reported in conventional soil surveys are useful for application of pedotransfer functions defining the physical parameters of hydrological models. The strong relationship between soil morphology and soil water regime has been used to transfer pedological processes to develop hydrological response models of soilscapes. Time as a factor of soil formation implies continuous development and change. The morphology could be out of phase with the recent impact of the current climate due to natural buffering questioning its validity for modelling.

The soil chemical parameters are sensitive to pedological processes controlled by current hydrological responses. The secondary processes *e.g.* reduction, ferrolysis, etc., also react to pedological conditions created by different soil water regimes. Where soil chemistry is in phase with soil morphology a relative stable hydrology is expected. Differences between them can distinguish between soils, which are in phase with the current climatic and soilscape conditions and soils in a process of change. Chemical parameters and soil morphology, are products of related pedological processes and therefore generally in phase. Differences between these soil properties may therefore indicate a change in horizonation in soils due to a changing water regime.

Chemical properties related to processes of soil formation confirmed water regimes indicating the rate and direction of flow associated with flowpaths or storage mechanisms. Biocycling was a process dominantly active in recharge and deep interflow soils and must be considered when inferring hydrological response.

The impact of the soil organic matter distribution down the profile and the impact of alternating redox conditions on chemical weathering and acidification could be distinguished. Ferrolysis and the impact on soil chemical properties must be assessed with care in conditions of high soil organic matter typical of topsoils in general and closer to the surface within topsoils. In the deep interflow soils there is an accumulation of cations at the transition of an oxidic and redox horizon due to leaching and capillary rise of the redox horizon.

A conceptual hydrological response model of a soilscape was constructed using soil morphology. Soil chemical parameters were used to verify if the soil profiles were still in phase with the current morphology. Chemistry confirmed that most the interpretations made were correct and minor improvements were made. The conceptual hydrological response model was validated by tension data for event response and long term water contents.

Soil morphology has the ability to provide dimensions to point data and therefore allow more accurate extrapolations. Soil chemical parameters, which are more easily attainable than hydrometric data, can be used to identify soil bodies with similar hydrological response and distinguish between different hydrological responses in similar soil bodies. These trends in the chemical parameters can therefore be extrapolated in morphologically and climatically similar soils to help deduce hydrologically similar soils.

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Appendix

Table 1 Profile description of Longlands (P201) (Van Huyssteen et al., 2005)

Profile No: 201
Map/photo: 3128AB
Latitude & Longitude: 31° 06' 06.0" / 28° 20' 18.0"
Surface stoniness: None
Altitude: 1 337 m
Terrain unit: Upper midslope
Slope: 4 %
Slope shape: Convex
Aspect: West
Microrelief: None
Parent material solum: Origin single, solid rock
Underlying material: Feldspathic sandstone
Geological Group: Elliot

Soil form: Longlands
Soil family: 2000
Surface rockiness: None
Occurrence of flooding: None
Wind erosion: None
water erosion: None
Vegetation / Land use: Grassveld, closed
Water table: None
Described by: C.W. van Huyssteen & W. Boshoff
Date described: 19/06/2001
Weathering of underlying material: Weak physical, weak chemical
Alteration of underlying material: Ferruginised

Horizon	Depth(mm)	Description	Diagnostic horizons / material
A	0 – 430	**Moisture status:** dry; **dry colour:** 10YR4/4 (80 %), 10YR4/2 (20 %); **moist colour:** 10YR3/3 (80 %), 10YR3/2 (20 %); 8.9 % clay; coarse sandy loam; few fine faint 10YR4/2 dry, 10YR3/2 moist, geogenic and biological mottles; apedal massive; soft, friable, non-sticky, non-plastic; no fine and very fine pores; common normal medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 1 second(s); many normal roots; gradual smooth transition	orthic A horizon
E	430 – 730	**Moisture status:** moist; **dry colour:** 7.5YR6/4 (95 %), 10YR4/4 (5 %); **moist colour:** 10YR4/4 (95 %), 10YR4/4 (5 %); 11.9 % clay; coarse sandy loam; common medium distinct 10YR4/4 dry, 10YR3/3 moist, Illuvial humus mottles; common fine distinct 2.5YR4/8 dry, 10R4/8 moist, iron oxide mottles; apedal massive; soft, friable, non-sticky, non-plastic; no fine and very fine pores; few normal medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 1 second(s); many normal roots; gradual smooth transition	E horizon
B	730 – 980	**Moisture status:** moist; **dry colour:** 7.5YR6/4 (100 %); **moist colour:** 10YR4/4 (100 %); 12.6 % clay; coarse sandy loam; many fine prominent 2.5YR4/8 dry, 10R4/8 moist, iron oxide mottles; few fine distinct 2.5YR3/0 dry, 2.5YR2.5/0 moist, manganese mottles; apedal massive; soft, friable, non-sticky, non-plastic; few normal fine and very fine pores; no medium and coarse pores; no slickensides; no cracks; no cutans; very few 6-25 mm mixed shaped stones; water absorption 1 second(s); many normal roots; gradual tonguing transition; Remark: B2 horizon has tongues of sandstone saprolite, with manganese concretions and red surfaces, surrounded by grey sillans.	soft plinthic B horizon
C	980 – 1210	**Moisture status:** moist; **dry colour:** 7.5YR7/2 (40 %), 10YR4/8 (10 %), 10YR8/4 (50 %); **moist colour:** 7.5YR4/4 (40 %), 10R4/6 (10 %), 7.5YR6/8 (50 %); 11.4 % clay; coarse sandy loam; no mottles; apedal massive; soft, friable, non-sticky, non-plastic; no fine and very fine pores; no medium and coarse pores; many slickensides; no cracks; no cutans; no coarse fragments; water absorption 1 second(s); no normal roots; transition not reached; Remark: 870 to 1000 mm is wavy transition between the soft plinthic B horizon and rock horizons. Cores are gray and yellow with red mottles.	unspecified material with signs of wetness

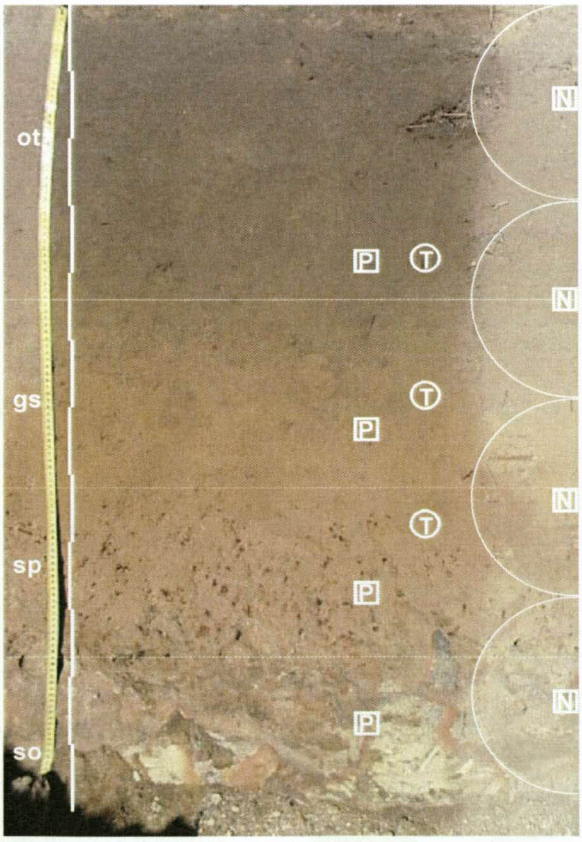

Figure 1 Photograph of Longlands (P201) (Van Huyssteen et al., 2005)

Table 2 Profile description of Kroonstad (P205) (Van Huyssteen et al., 2005)

Profile No: 205
Map/photo: 3128AB
Latitude & Longitude: 31° 06' 07.9" / 28° 20' 03.4"
Surface stoniness: None
Altitude: 1 286 m
Terrain unit: Lower midslope
Slope: 7 %
Slope shape: Concave
Aspect: North-west
Microrelief: None
Parent material solum: Origin single, solid rock
Underlying material: Mudstone
Geological Group: Molteno

Soil form: Kroonstad
Soil family: 2000
Surface rockiness: None
Occurrence of flooding: None
Wind erosion: None
Water erosion: None
Vegetation / Land use: Grassveld, closed
Water table: None
Described by: D. Scholtz & T.B. Zere
Date described: 19/06/2001
Weathering of underlying material: Moderate physical, strong chemical
Alteration of underlying material: Ferruginised
WRB Classification: Gleyic Albic Planosol (hyperdistric)
Soil Taxonomy: Aeric Albaqualfs

Horizon	Depth(mm)	Description	Diagnostic horizons / material
A	0 - 220	Moisture status: moist; dry colour: 2.5Y5/2 (100 %); moist colour: 2.5Y3/2 (100 %); 13.1 % clay; coarse sandy loam; no mottles; weak fine subangular blocky; hard, friable, non-sticky, non-plastic; many rusty fine and very fine pores; few bleached medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 1 second(s); many normal roots; diffuse smooth transition;	orthic A horizon
E1	220 - 460	**Moisture status:** moist; **dry colour:** 2.5Y6/1 (70 %), 10YR6/6 (30 %); **moist colour:** 10YR3/3 (70 %), 10YR4/6 (30 %); 11.0 % clay; coarse sandy loam; many medium distinct 10YR6/6 dry, 10YR4/6 moist, iron oxide mottles; common medium distinct 2.5Y4/2 dry, 2.5Y3/2 moist, humus mottles; weak fine granular; hard, friable, non-sticky, non-plastic; many rusty fine and very fine pores; no medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 2 second(s); many normal roots; clear smooth transition;	E horizon
E2	460 - 660	**Moisture status:** moist; **dry colour:** 10YR6/2 (80 %), 10YR5/8 (10 %), 10YR5/2 (10 %); **moist colour:** 10YR5/2 (80 %), 7.5YR5/8 (10 %), 10YR4/3 (10 %); 11.0 % clay; coarse sandy loam; common medium distinct 10YR6/8 dry, 7.5YR5/8 moist, iron oxide mottles; common medium prominent 2.5Y4/2 dry, 2.5Y3/2 moist, humus mottles; weak fine granular; hard, friable, slightly sticky, slightly plastic; many rusty fine and very fine pores; no medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 4 second(s); common normal roots; abrupt smooth transition;	E horizon
G	660 - 1400	**Moisture status:** moist; **dry colour:** 2.5Y7/1 (70 %), 10YR4/8 (30 %); **moist colour:** 2.5Y6/2 (70 %), 10YR4/8 (30 %); 28.0 % clay; coarse sandy loam; common fine prominent 2.5YR4/8 dry, 10R4/8 moist, iron oxide mottles; common medium prominent 10YR6/8 dry, 7.5YR4/8 moist, iron oxide mottles; strong coarse angular blocky; very hard, very firm, sticky, plastic; common normal fine and very fine pores; no medium and coarse pores; many slickensides; fine cracks; common silica cutans; few 2-6 mm round stones; water absorption 8 second(s); few normal roots; transition not reached;	G horizon

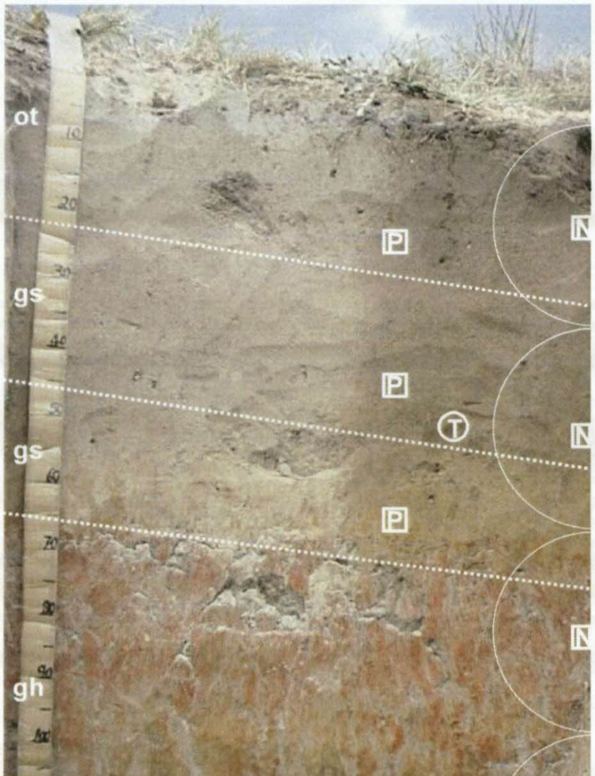

Figure 2 Photograph of Kroonstad (P205) (Van Huyssteen et al., 2005)

Table 3 Profile description of Katspruit (P218) (Van Huyssteen et al., 2005)

Profile No: 218
Map/photo: 3128AB
Latitude & Longitude: 31° 06' 18.8" / 28° 19' 46.2"
Surface stoniness: None
Altitude: 1 276 m
Terrain unit: Valley bottom
Slope: 0 %
Slope shape: Concave
Aspect: Level
Microrelief: 20 % mole mounds, 0.3 m high, profile between features
Parent material solum: Origin single, alluvium
Underlying material: Feldspathic sandstone
Geological Group: Molteno

Soil form: Katspruit
Soil family: 1000
Surface rockiness: None
Occurrence of flooding: Frequent
Wind erosion: None
Water erosion: None
Vegetation / Land use: Grassveld, closed
Water table: 400 mm
Described by: C.W. van Huyssteen
Date described: 19/06/2001
Weathering of underlying material: Advanced physical, strong chemical
Alteration of underlying material: Ferruginised

Horizon	Depth(mm)	Description	Diagnostic horizons / material
A1	0 - 160	**Moisture status:** moist; **dry colour:** 10YR7/1 (90 %), 10YR4/4 (10 %); **moist colour:** 7.5YR4/1 (90 %), 10YR4/4 (10 %); 18.6 % clay; loam; few fine faint 10YR5/8 dry, 10YR4/4 moist, iron oxide mottles; few fine faint 10YR7/1 dry, 10YR5/1 moist, reduced iron mottles; moderate fine subangular blocky; slightly hard, slightly firm, non-sticky, slightly plastic; many bleached fine and very fine pores; few normal medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 1 second(s); many bleached roots; gradual smooth transition; Remark: A1 possible organic horizon.	orthic A horizon
A2	160 - 350	**Moisture status:** moist; **dry colour:** 10YR7/1 (90 %), 10YR4/4 (10 %); **moist colour:** 7.5YR4/2 (90 %), 10YR4/4 (10 %); 18.0 % clay; loam; few fine faint 10YR5/8 dry, 10YR4/4 moist, iron oxide mottles; few fine faint 10YR7/1 dry, 10YR5/1 moist, reduced iron mottles; weak fine subangular blocky; slightly hard, friable, slightly sticky, slightly plastic; common bleached fine and very fine pores; few normal medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 1 second(s); common bleached roots; gradual smooth transition; Remark: A2 possible E horizon.	orthic A horizon
G	350 - 750	**Moisture status:** Wet; **dry colour:** 10YR5/6 (80 %), 10YR5/1 (20 %); **moist colour:** 7.5YR5/6 (80 %), 10YR7/1 (20 %); 30.1 % clay; clay loam; many coarse prominent 10YR7/1 dry, 10YR5/1 moist, reduced iron mottles; apedal massive; hard, very firm, sticky, very plastic; few rusty fine and very fine pores; few rusty medium and coarse pores; no slickensides; fine cracks; very many silica cutans; no coarse fragments; water absorption 10 second(s); common bleached & rusty roots; transition not reached;	G horizon

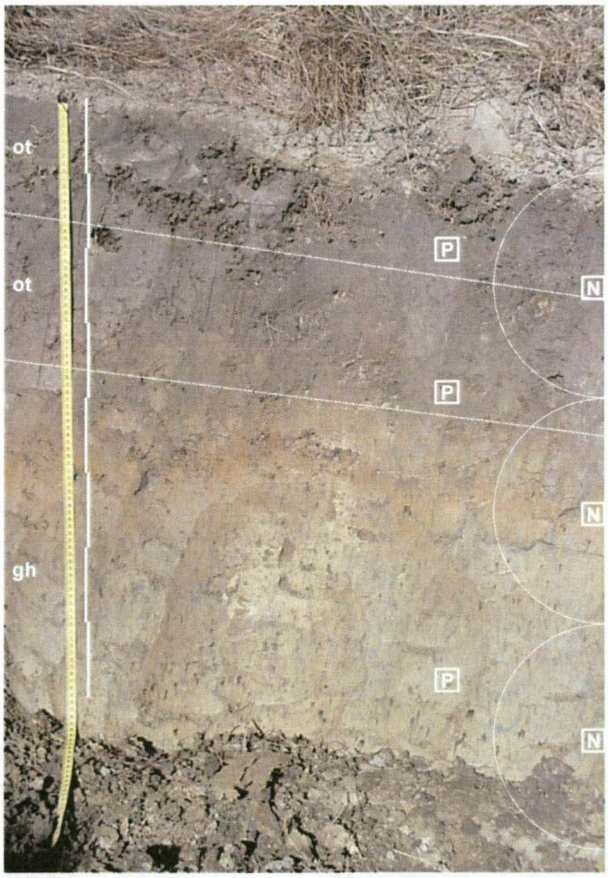

Figure 3 Photograph of Katspruit (P218) (Van Huyssteen et al., 2005)

Table 4 Profile description of Bloemdal (P220) (Van Huyssteen et al., 2005)

Profile No: 220
Map/photo: 3128AB
Latitude & Longitude: 31° 06' 59.9" / 28° 19' 38.9"
Surface stoniness: None
Altitude: 1 306 m
Terrain unit: Upper midslope
Slope: 16 %
Slope shape: Straight
Aspect: East
Microrelief: None
Parent material solum: Origin binary, local colluvium, solid rock
Underlying material: Feldspathic sandstone
Geological Group: Molteno

Soil form: Bloemdal
Soil family: 2100
Surface rockiness: None
Occurrence of flooding: None
Wind erosion: None
Water erosion: None
Vegetation / Land use: Grassveld, closed
Water table: None
Described by: P.A.L. le Roux & M. Smit
Date described: 19/06/2001
Weathering of underlying material: Strong physical, moderate chemical
Alteration of underlying material: Ferruginised

Horizon	Depth(mm)	Description	Diagnostic horizons / material
A	0 - 540	**Moisture status:** moist; **dry colour:** 10YR4/3 (90 %), 7.5YR4/6 (10 %); **moist colour:** 7.5YR3/6 (10 %), 7.5YR3/2 (90 %); 13.6 % clay; sandy loam; no mottles; moderate medium subangular blocky; very hard, friable, non-sticky, non-plastic; many normal fine and very fine pores; common normal medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 2 second(s); many normal roots; gradual smooth transition; Remark: Red and grey colour intensity increases from 540 to 1500 mm.	orthic A horizon
B	540 - 880	**Moisture status:** moist; **dry colour:** 5YR5/4 (90 %), 5YR6/8 (10 %); **moist colour:** 5YR4/6 (90 %), 5YR3/6 (10 %); 15.0 % clay; sandy loam; few fine prominent 5YR6/8 dry, 5YR3/2 moist, humus mottles; weak medium subangular blocky; very hard, friable, slightly sticky, non-plastic; many normal fine and very fine pores; few normal medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 2 second(s); many bleached roots; clear smooth transition;	red apedal B horizon
C1	880 - 1080	**Moisture status:** moist; **dry colour:** 5YR4/6 (70 %), 5YR5/6 (20 %), 5YR7/2 (10 %); **moist colour:** 5YR3/4 (70 %), 5YR4/6 (20 %), 5YR5/2 (10 %); 10.8 % clay; sandy loam; common medium prominent 5YR5/6 dry, 5YR3/2 moist, humus mottles; many coarse prominent 5YR7/2 dry, 5YR5/2 moist, bleached mottles; apedal massive; very hard, firm, slightly sticky, slightly plastic; many normal fine and very fine pores; few bleached medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 2 second(s); many normal roots; clear smooth transition;	unspecified material with signs of wetness
C2	1080 - 1500	**Moisture status:** moist; **dry colour:** 5YR5/4 8(5 %), 5YR4/6 1(5 %); **moist colour:** 5YR4/6 1(5 %), 5YR4/4 8(5 %); 11.2 % clay; sandy loam; many fine distinct 5YR4/6 dry, 10YR5/3 moist, iron oxide mottles; many coarse prominent 7.5YR5/4 dry, 10YR5/3 moist, bleached mottles; apedal massive; very hard, firm, sticky, plastic; few normal fine and very fine pores; few normal medium and coarse pores; no slickensides; no cracks; no cutans; no coarse fragments; water absorption 3 second(s); few normal roots; transition not reached;	unspecified material with signs of wetness

Figure 4 Photograph of Bloemdal (P220) (Van Huyssteen et al., 2005)

Table 5 Profile description of Hutton (P221) (Van Huyssteen et al., 2005)

Profile No: 221
Map/photo: 3128AB
Latitude & Longitude: 31° 06' 00.3" / 28° 19' 42.9"
Surface stoniness: None
Altitude: 1 289 m
Terrain unit: Upper footslope
Slope: 8 %
Slope shape: Straight
Aspect: East
Microrelief: None
Parent material solum: Origin single, local colluvium
Underlying material: Mudstone
Geological Group: Molteno

Soil form: Hutton
Soil family: 2100
Surface rockiness: None
Occurrence of flooding: None
Wind erosion: None
Water erosion: None
Vegetation / Land use: Grassveld, closed
Water table: None
Described by: M.L. Smit & T.B. Zere
Date described: 19/06/2001
Weathering of underlying material: Strong physical, strong chemical
Alteration of underlying material: Ferruginised

Horizon	Depth(mm)	Description	Diagnostic horizons / material
A	0 - 250	**Moisture status:** dry; **dry colour:** 2.5YR4/6 (100 %); **moist colour:** 2.5YR3/4 (100 %); 16.9 % clay loam; no mottles; weak fine granular; slightly hard, friable, slightly sticky, slightly plastic; few normal fine and very fine pores; no medium and coarse pores; no slickensides; fine cracks; no cutans; no coarse fragments; water absorption 1 second(s); many normal roots; abrupt smooth transition;	orthic A horizon
A/B	250 - 580	**Moisture status:** moist; **dry colour:** 7.5YR3/4 (90 %), 2.5YR4/6 (10 %); **moist colour:** 7.5YR3/2 (90 %), 2.5YR3/4 (10 %); 18.7 % clay loam; no mottles; moderate medium subangular blocky; slightly hard, firm, very sticky, plastic; common normal fine and very fine pores; few normal medium and coarse pores; no slickensides; fine cracks; no cutans; no coarse fragments; water absorption 1 second(s); many normal roots; gradual smooth transition;	orthic A / red apedal B horizon
B	580 - 1250	**Moisture status:** moist; **dry colour:** 2.5YR4/8 (100 %); **moist colour:** 2.5YR3/4 (100 %); 31.3 % clay sandy clay loam; no mottles; weak fine crumb; slightly hard, slightly firm, sticky, plastic; common normal fine and very fine pores; common normal medium and coarse pores; no slickensides; fine cracks; no cutans; no coarse fragments; water absorption 2 second(s); many normal roots; diffuse smooth transition;	red apedal B horizon
C	1250 - 1800	**Moisture status:** moist; **dry colour:** 10R4/8 (100 %); **moist colour:** 2.5YR3/4 (100 %); 33.3 % clay sandy clay loam; few fine prominent 2.5YR4/4 dry, 2.5YR3/4 moist, geogenic iron mottles; weak fine subangular blocky; slightly hard, slightly firm, sticky, plastic; common normal fine and very fine pores; common rusty medium and coarse pores; no slickensides; fine cracks; common clay cutans; no coarse fragments; water absorption 3 second(s); few normal roots; transition not reached;	unspecified material with signs of wetness

Figure 5 Photograph of Hutton (P221) (Van Huyssteen et al., 2005)