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Deschamps, Christine L.

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A COMPARISON OF THE BIOGEOGRAPHICAL AND GEOMORPHOLOGICAL CHARACTERISTICS OF GULLIED AND NON-GULLIED VALLEY HEAD MIRES IN EASTERN LESOTHO

Christine L. Deschamps

School of Geography, Archaeology and Environmental Studies

University of the Witwatersrand, Johannesburg

Dissertation submitted to the Faculty of Science for the degree of Master of Science.

February 2006

Declaration

I declare that this dissertation is my own, unaided work, except where otherwise acknowledged. It is being submitted for the degree of Master of Science in the University of the Witwatersrand, Johannesburg. It has not been submitted before for any degree or examination in any other university.

_____

(Christine Deschamps)

_____ day of _____ 2006

Abstract

Environmental degradation is a serious problem in Lesotho, Africa. The majority of studies dealing with soil loss and gully erosion have mainly focused on the mechanics of the erosion. However, mountain watersheds are sensitive and vulnerable to degradation and in so being, have large inherent environmental implications. Gully erosion diversifies the micro-topographical and hydrological environment. The severe changes and eventual system degradation incurred as a result of gullying has rarely been quantified. This paper specifically highlights mire phytogeographic responses to valley head degradation at selected sites in eastern Lesotho by comparing patterns observed in gullied mires to those seen in other non-gullied mires. Soil and vegetation belt transects are set up across five mires (2 gullied) in the highlands of eastern Lesotho. The response of many of the plant species to the overall environmental gradient was asymmetrical and unimodal in pattern. The spatial distribution of the soil's physical properties, topography and vegetation community patterns were found to reflect the spatial mosaic of the soil moisture gradient. The negative impact that gullying has on the ecohydrological regime of the valley heads is evident and is allowing for shrub encroachment. Both the wetland and dryland vegetation communities correlate strongly with the changes in the surface soil moisture gradient. Gully erosion is clearly a threshold phenomenon. Continued grassland degradation, accelerated soil erosion and subsequent gullying of the wetlands will lead to plant and animal diversity loss, decreased livestock productivity, sediment-laden water and shortened dam life-span.

Preface

Mountain watersheds have always been of global concern because of their important and sensitive function. The impairment of wetlands has considerable implications for surrounding regions (Paudel and Thapa, 2001), however, despite conservation efforts, mountain wetlands remain vulnerable to degradation. In 1971, in the city of Ramsar, Iran, the first international treaty was signed for the protection, conservation and wise-use of wetlands and their resources (Ramsar, 2004). A wetland at Lets'eng-la-Letsie, in the province of Quthing (Lesotho) was added to the list on 1st July 2004. A large number of African wetlands are in good condition, however, as population increases, the pressure to exploit these wetland areas also increases (Schuyt, 2005).

Within this dissertation, the degradation of Lesotho wetlands is discussed with regards to the impact it has hydraulically and economically on the region (Schwabe, 1995; Grab and Morris, 1999; Nüsser and Grab, 2002). The loss of wetlands represents a huge loss in species diversity through habitat destruction, the loss of water regulation, more frequent flash floods, the loss of nutrients, and most importantly, the loss of large tracts of valuable wetland and grassland (Schwabe, 1995). The biogeographical aspects of five valley heads are examined with particular focus on the effect that gully erosion has on wetland characteristics.

This dissertation is divided into seven chapters. **Chapter 1** introduces the basis of this study. A general review of some issues in Lesotho such as, politics, environmental issues and population dynamics are given. The objectives of the study are outlined. **Chapter 2** outlines the environmental setting of the highlands region in Lesotho. **Chapter 3** describes the methods used to collect and interpret the data within this dissertation. **Chapter 4** presents a literature review on the processes and mechanics of gully erosion in general, as well as for two of the five valley heads included in this dissertation. It also summarizes and compares the collected data to past work conducted in the same area. **Chapter 5** presents a literature review on mountain soils, Lesotho soils and wetland soils.

It then describes the soils found within the valley heads studied in this dissertation. **Chapter 6** presents a literature review on the effects of disturbance to vegetation patterning and community type. It then describes the vegetation communities within this dissertation and discusses patterns found and the possible reasons for this occurrence. **Chapter 7** presents a summary of the findings of this research and provides a conclusion to them.

Sections of this work have been presented at the Annual British Ecological Society Meeting in Lancaster, England in 2004 and have been published in the International Journal, Arctic, Antarctic and Alpine Research, 2004.

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Chapter 1

Introduction

1.1 Setting

The small mountainous Kingdom of Lesotho is an enclave within South Africa (Marake *et al*., 1998) and is situated between longitudes 27˚00'E and 29˚30'E and latitudes 28˚30'S and 30˚40'S (Rydgren, 1988) (Figure 1.1). Lesotho is ca. 30 355 km^2 in size and is comprised of four main physiographic regions: the western lowlands, the foothills, the Senqu River Valley and the central and eastern highlands (Figure 1.1). The mountain highlands constitute 80% of the total land area of Lesotho and are the highest in southern Africa, reaching 3 482 m a.s.l. (Bainbridge *et al*., 1991). Two of the four highest mountain ranges, namely the Drakensberg and Maluti, constitute the highlands region of the study area. Together, these two north–south trending mountain ranges represent southern Africa's major drainage divide between the easterly-flowing rivers towards the Indian Ocean (e.g. Tugela) and the westerly-flowing rivers towards the Atlantic Ocean (e.g. Senqu) (Bainbridge *et al*., 1991; Nüsser and Grab, 2002). Deeply dissected basalt valleys, steep slopes, planation surfaces and a dense drainage network characterize the highlands area, with the Great Escarpment acting as the eastern border between Lesotho

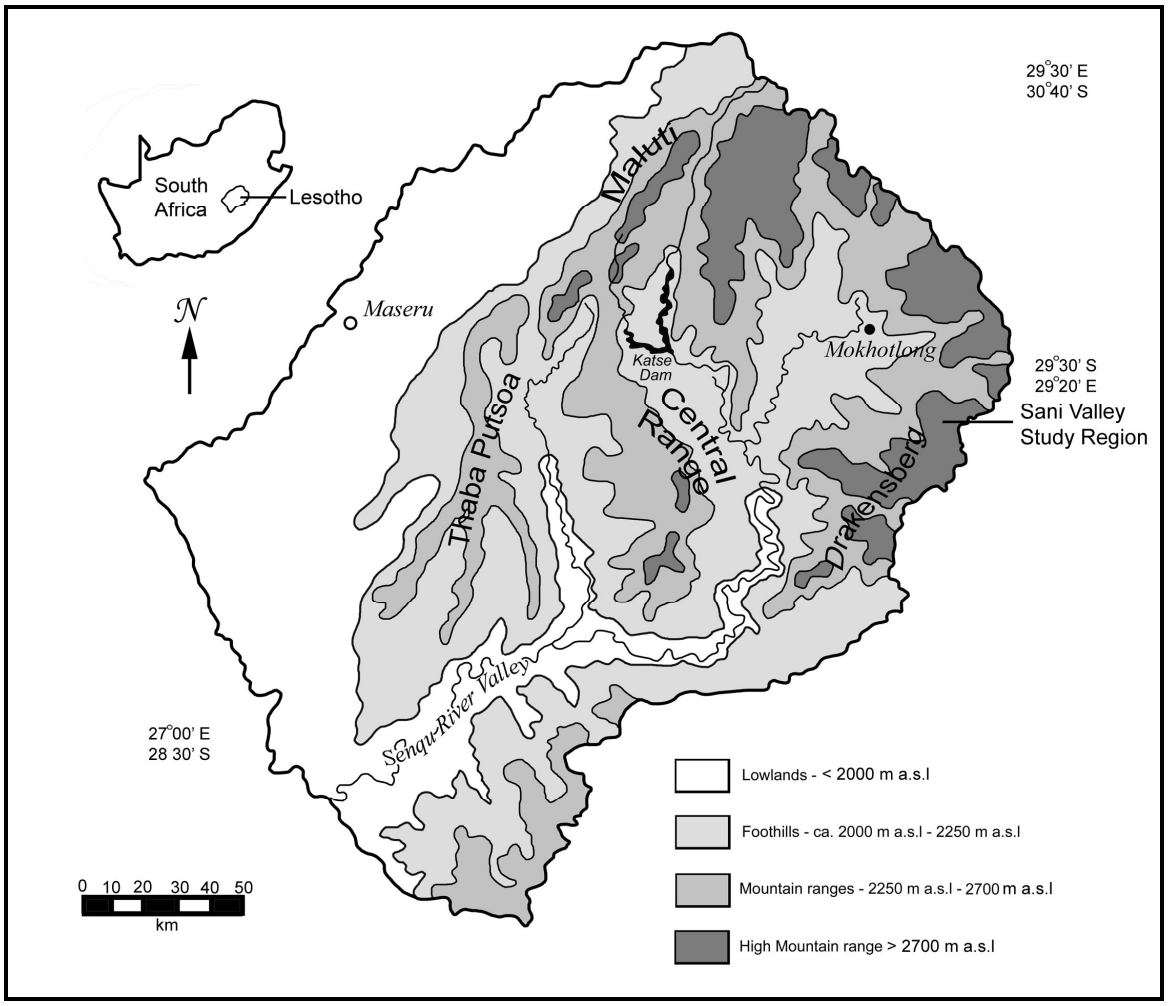


Figure 1.1 Physiographic regions of Lesotho and small-scale location of study site (after Kakonge, 2002, p64).

and South Africa (Jacot-Guillarmod, 1969; van Zinderen Bakker and Werger, 1974; Morris *et al*., 1993).

Lesotho's watersheds are the largest in southern Africa, thus making Lesotho one of the most important sources of fresh water on the African continent (Bainbridge *et al*., 1991; Schwabe, 1995). The water source begins as springs, which are surrounded and filtered by mires. Sadly however, Lesotho is one of the most eroded countries in the world (Calles and Kulander, 1996). The loss of rangeland through soil loss coupled with

population increases and pressures on water resources make the alpine wetlands an increasingly important but vulnerable resource. The wetlands are showing signs of degradation through gully erosion, overgrazing and plant invasions. This study primarily investigates the effects of gully erosion on the geo-ecology of three valley head mires and may thus provide valuable environmental information required for the continuation of the Lesotho Highlands Water Project (LHWP).

1.2 Politics And Demographics

Until independence on October 4th 1966, Lesotho was a British protectorate known as Basutoland. The political history since independence has been unstable, with the Basotho National Party (BNP) being in power until 1986 (Marake *et al*., 1998). In 1986, the BNP was overthrown by the military amidst political pressure from South Africa. In 1993, the Basutoland Congress Party (BCP) was democratically instated and a new constitution was formed, which gave limited power to the king. Despite mediation with the Southern African Development Committee (SADC) there has been instability, attempted coups and successful take-overs and restorations since 1994 (Marake *et al*., 1998).

The population of Lesotho was just over 2 million in 2001 (Nel and Illgner, 2001), which equates to a population density of 66 people per km^2. Over half the population resides in the western lowlands where 80% survive on subsistence agriculture and livestock (Nixon, 1973; Meakins and Duckett, 1993; Makhoalibe, 1999; Nel and Illgner, 2001). The actual population density on available arable land is thus ca. 733 people per km^2 (Marake *et al*., 1998). Since independence, total arable land in Lesotho has decreased from 13% to 8%

(Schmitz and Rooyani, 1987; Marake *et al*., 1998; Nüsser, 2002). The remaining 92% of the country is comprised of mountains, villages, rangeland and barren eroded land (Marake *et al*., 1998).

1.3 Socio-Economic Issues

Although Lesotho's largest source of foreign exchange stems from the Lesotho Highlands Water Project (LHWP) (see section 1.6 for details), agricultural production and livestock rearing are the main sources of employment (Marake *et al*., 1998).

1.3.1 Agriculture

Seventy percent of crops grown in Lesotho are consumed locally. Main crops are grown in both the lowlands and in the mountains, depending on the season and consist of grain (maize, wheat and sorghum), beans and peas (Marake *et al*., 1998). Although 60% – 70% of the workforce is engaged in agriculture, the sector is becoming increasingly dominated by livestock rearing (Marake *et al*., 1998). Agricultural production has declined since 1978 / 79, whilst livestock production has increased over that of agriculture since 1984. Some of the reasons for this decline in agricultural outputs are said to be drought, harsh winters, fertiliser application rates and the loss of land through soil erosion (Marake *et al*., 1998). Despite pressure on existing arable land, large tracts are still left fallow each year (Marake *et al*., 1998), perhaps due to the dominance of the livestock sector.

1.3.2 Livestock Activities - Transhumance

Rangelands comprise 80% of the total land area (Marake *et al*., 1998) and are mainly found in the foothills and subalpine regions. These areas are grazed by cattle, sheep and goat according to the *transhumance* system, which has been an ongoing tradition of livestock rearing since the 19th Century (Quinlan and Morris, 1994; Makhoalibe, 1999; Nüsser, 2002). In late spring (October/November), livestock is taken from villages below 2 400 m a.s.l. and brought to the summer grazing posts in the upper subalpine and alpine grasslands (2 750 m – 3 100 m a.s.l.). In late autumn (May), the livestock is brought back to the village areas to forage on the harvested land and village rangelands. Although Lesotho's monarch allows usufractuary rights, the transhumance grazing practice is controlled in the village areas by the infrequently noted systems of *maboella* and *leboella* (Marake *et al*., 1998). *Maboella* and *leboella* are controlled by the principal chiefs and are systems whereby plots of grazing and rangeland are set aside each summer for rejuvenation. There is also a requirement that each grazing post area have a grazing permit so that livestock numbers and movements can be monitored (Marake *et al*., 1998). However, grassland deterioration remains an ongoing concern in these grazing areas (Quinlan and Morris, 1994; Marake *et al*., 1998). Since the 1970's, the grazing carrying capacity of the village winter grazing areas has decreased due to increases in population and soil erosion. In response to this loss, grazing pressure has increased on the mountain rangelands (Nüsser, 2002). Winter grazing posts are now being established in the lower subalpine regions (2 600 m a.s.l.), above the villages. There is thus an ever-increasing grazing pressure on the grasslands and wetlands of the subalpine zone without an adequate period of vegetation rejuvenation (Quinlan and Morris, 1994).

1.4 Erosion And Controls

Due to perceived ineffective range management by the local chiefs, as well as the fact that traditional conservation controls do not extend into the alpine areas (as these areas were rarely used and more difficult to control) (Marake *et al*., 1998; Nüsser, 2002), the government has introduced various conservation-based legislations and programmes. Systems and laws, such as the Village Development Councils (VDC's), the Land Husbandry Act of 1969, Pasture Management regulations (1973), Grazing Control regulations (1977), Grazing Associations which operate in Range Management Areas (in 1978), livestock import taxes (in 1984) and Livestock Exchange were all introduced in an attempt to create effective management and conservation through the control of livestock numbers and movements (Quinlan and Morris, 1994) (see Table 1.1). In the early 1990's, the area above 2 750 m a.s.l. was categorised as a Managed Resource Area (MRA). This initiative was part of the Drakensberg/Maluti Catchment Conservation Programme (D/MCCP) and was designed to reduce alpine grassland degradation, as well as to determine whether highland erosion could affect the LHWP reservoirs (Quinlan and Morris, 1994). The zonation of the MRA has been criticized for being too high in altitude to efficiently protect the areas that are actually being grazed longer (i.e. the winter grazing areas in the lower subalpine region at ca. 2 600 m a.s.l.).

Soil erosion throughout Lesotho and the consequent grassland and/or wetland degradation is generally believed to have an anthropogenic origin and is often attributed to a lack of local stakeholder (i.e. Basotho) conservation initiatives (Jacot-Guillarmod, 1962, 1963; van Zinderen Bakker & Werger, 1974; Chakela, 1981; Meakins & Duckett,

Table 1.1 Number of livestock in Lesotho over a period of time (after Marake *et al*., 1998, p26).

Year	Cattle	Sheep	Goats	Horses	Donkeys
1985/86	525	1,392	978	110	112
1986/87	627	1,703	1,150	127	136
1987/88	627	1,645	1,121	129	148
1988/89	583	1,501	1,064	116	131
1989/90	523	1,377	995	104	151
1990/91	543	1,481	730	98	110
1991/92	700	1,383	649	89	-
1992/93	658	1,176	811	107	139
1993/94	578	1,277	876	113	140
1994/95	580	1,131	749	100	146

1993; Backéus & Grab, 1995; Schwabe, 1995; Morris & Grab, 1997; Grab & Nüsser, 2001). Apparently, gully erosion did not exist before missionaries first visited the country (Faber and Imeson, 1982) and was unknown before Europeans colonized the area in the 1830's (Showers, 1989). The introduced European farming practices (i.e. use of the plough, keeping cattle in pens, indiscriminate tree-felling and contour plots) differed greatly from the traditional land-use systems. It is worth noting that traditional methods had many conservation elements, such as rotational grazing, shallow hoeing and the keeping of small plots surrounded by grassland (Showers, 1996). By the late 19th Century to early 20th Century, gully erosion had become established in areas with much human activity (e.g. government land and alongside paths and roads). In the first attempt to evaluate the extent of erosion in Lesotho, a non-quantitative government report was written in 1935 (Pim, 1935). Although the report inferred that gully erosion was not a problem on the farmlands (Faber and Imeson, 1982), some gullies were cited as being over 4.5 m deep (Showers, 1996). Ironically, it appears that European land-use practices and soil conservation measures since the late 1930's had contributed to the establishment of gully erosion in parts of Lesotho (Showers, 1996).

1.5 Grasslands And Erosion

The alpine region of eastern Lesotho is within the Austral Afro-alpine belt (van Zinderen Bakker and Werger, 1974; Carbutt and Edwards, 2004). This belt has an unusually large number of endemic and/or threatened species for a high mountain system (Hilliard and Burtt, 1987; Bainbridge *et al*., 1991; Morris *et al*., 1993; Backéus and Grab, 1995; Makhoalibe, 1999). Approximately 3% of the plant species and just over 60 faunal species found in the eastern mountain region are endemic (Bainbridge *et al*., 1991). The main vegetation types within the alpine belt are the grassland communities, *Themeda triandra* ('seboku') and *Festuca caprina* ('letsiri') accompanied by *Merxmuellera spp* (Carbutt and Edwards, 2004). These grassland communities not only provide an essential source of fodder to grazing livestock but, together with the wetland vegetation, they play a key role in influencing the hydrological regime of the region by stabilizing the catchment soils and thereby reducing sheetflow and consequent soil erosion (Grobbelaar and Stegman, 1987; Schwabe and Whyte, 1993). However, grassland vegetation cover is decreasing at high altitudes, especially around permanent grazing posts (Nüsser, 2002). Where grassland cover has decreased, invasion by karroid species occurs (e.g. *Chrysocoma ciliata*) (van Zinderen Bakker and Werger, 1974; Chakela, 1981; Morris *et al.*, 1993; Schwabe and Whyte, 1993; Pooley, 1998; Grab and Morris, 1999). The exact reason(s) for this change in vegetation cover in the various highland regions is difficult to establish, however degradation in close proximity to the grazing posts is evidently owing to over-utilization of the grazing resource around these posts (Nüsser, 2002). As a result of the change from grassland to shrubland, runoff and overland flow velocity increases because the shrubs cannot attenuate water flow as effectively as the grasses (Faulkner,

1990; Abrahams *et al*., 1995; Casermeiro *et al*., 2004). Consequent gullying of the wetlands decreases their ability to attenuate water, and thus generates large sediment loads within fluvial channels and dams (Schwabe and Whyte, 1993; Quinlan and Morris, 1994).

1.6 Hydrology

High altitude wetland systems predominantly located above 2 600 m a.s.l. in eastern Lesotho are areas with high endemism and high biological diversity (Grobbelaar and Stegman, 1987; Bainbridge *et al*., 1991; Backéus and Grab, 1995; Morris and Grab, 1997; Makhoalibe, 1999). These alpine Holocene-aged wetlands (oldest date at 13 490 BP; Marker 1995, 1998) are the only peat producing wetlands (i.e. *mires*) in southern Africa (Bainbridge *et al*., 1991; Backéus and Grab, 1995; Nüsser and Grab, 2002). The mires tend to have a crescentic shape and occur in cirque-shaped valley heads (Jacot-Guillarmod, 1962; Grobbelaar and Stegman, 1987) and at river sources where groundwater surfaces in the form of springs (Jacot-Guillarmod, 1969; Schmitz and Rooyani, 1987). Both minerotrophic fens and ombrotrophic bogs are reported to exist in the alpine areas, with fens occurring on the warmer north-facing slopes and bogs occurring on the steeper, south-facing slopes and at higher altitudes (Schwabe, 1989). These alpine mires are key resources for the economy of Lesotho. Firstly, the mires are important grazing areas and thus contribute to the agricultural revenue. They remain saturated throughout the year, thus allowing for perennially flowing streams that are an especially important water resource during the dry winter months. The wetlands also have a high biomass (3700.6kg/Ha) and a large carrying capacity (2.5 times greater than the surrounding grasslands) (Schwabe and Whyte, 1993; Nüsser, 2002). Thirdly, the

mires have important hydrological functions including organic matter production, soil stabilization, flow reduction, sediment retention, nutrient removal and groundwater storage and recharge (Schwabe, 1995), all of which are essential to the success of the Lesotho Highlands Water Project (LHWP) (FIVAS, 1994).

1.6.1 Lesotho Highlands Water Project (LHWP)

The African continent has an estimated $^{+}$1200 dams, of which more than 60% are located in South Africa and Zimbabwe (Schuyt, 2005). One of these dams occurs within the alpine region of Lesotho on the Senqu River system, which is located in the largest watershed in southern Africa (Schwabe, 1995). The Senqu River drainage basin area within Lesotho is 20 000 km^2 and yields more than 47% of the total water flow in Lesotho (Makhoalibe, 1999). The exploitation of the mountain hydrology is Lesotho's most viable option to improving its economic conditions (Schmitz and Rooyani, 1987; van Rooy and van Schalkwyk, 1993). In order to facilitate the growing need for water in southern Africa, a joint venture was initiated between South Africa and Lesotho. The Malibamatso River, a tributary of the Senqu River, is used for the largest water transfer scheme on the African continent, which has thus far cost $^{+}$8 billion US dollars (FIVAS, 1994; Nel and Illgner, 2001). The aim of the LHWP is to reverse the southerly flow of the targeted water source so that a portion of it is diverted into the Vaal River in neighbouring South Africa (FIVAS, 1994; De Graaf and Bell, 1997). The project officially began in 1986 and consists of dams, transfer tunnels, delivery tunnels and pumping stations, the first phase of which was completed in 1998. Currently, the project is delivering 18 m^3/s of water to the industrial heartland of Gauteng Province, South Africa (Smith, 1999; Nel and Illgner, 2001). The export of this water through the LHWP

generates the largest source of foreign exchange for Lesotho and provides much needed development initiatives (Bainbridge *et al*., 1991). The LHWP also provides Lesotho with its own hydroelectric power supply.

Negative environmental consequences of the dam construction include the destruction of homes, the loss of over 4 000 ha of grazing and agricultural land, degraded downstream riparian land and seismic activity, all of which have negatively affected the functioning of springs in the headwater areas (Nel and Illgner, 2001; Matete and Hassan, 2003). Continued grassland degradation, accelerated soil erosion and subsequent gullying of the wetlands, and the environmental effects of dams can all lead to a loss of plant and animal diversity, a decrease in livestock productivity, sediment-laden water and a shortened dam life-span. These effects will eventually have an impact on the economy of Lesotho if not dealt with adequately in the near future (Grobbelaar and Stegman, 1987; Schwabe and Whyte, 1993; Quinlan and Morris, 1994; Calles and Stålnacke, 2000).

1.6.2 Wetland Degradation

A damaged wetland is one that has a deep water table, areas of exposed soil and evidence of some form of erosion (Schwabe, 1989). This type of wetland may also host plant species that are indicative of disturbance. Since at least 1963, the alpine wetlands of Lesotho have intrigued researchers (Killick, 1963), although many reports remain qualitative to semi-quantitative. All reports have stressed the fragility of the wetland habitats to overgrazing by livestock and most have emphasized the need for conservation

(see Table 1.2). The relatively poor condition of many high altitude wetlands/mires has been attributed to climatic change over the last 4 000 years, past and current land-use

Table 1.2 Research on various aspects of alpine wetlands and associated conservation issues in Lesotho.

Year	Author	Subject
1963	Jacot-Guillarmod	Observation on the bogs of Basutoland Mountains
1969	Jacot-Guillarmod	Effect of land use on alpine wetland vegetation
1974	van Zinderen Bakker and Werger	High-altitude bogs of Lesotho
1987	Grobbelaar and Stegmann	Water quality of high altitude bogs, Maluti Mountains
1987	Schmitz and Rooyani	Lesotho: Geology, geomorphology and soils
1989	Backéus	Flarks in Lesotho
1989	Klug *et al.*	Terrain analysis of the Drakensberg/Maluti
1989	Schwabe	Management of wetlands in the Drakensberg/Maluti
1991	Bainbridge *et al.*	Report on conservation programme, Drakensberg
1993	Meakins and Duckett	Vanishing bogs in the mountains of Lesotho
1993	Morris *et al.*	Classification of alpine vegetation in Lesotho
1993	Schwabe and Whyte	Distribution of wetlands and grasslands
1995	Backéus and Grab	Mires in Lesotho
1995	Schwabe	Alpine mires of Lesotho
1997	Morris and Grab	Threatened wetlands of Lesotho
1999	Grab and Morris	Soil and water issues in alpine wetlands of Lesotho
1999	Makhoalibe	Management of water in the Drakensberg/Maluti
2001	Grab and Nüsser	Case study from Sani plateau on integrated research
2002	Kakonge	Apply chaos theory to environmental decline in Lesotho
2002	Nüsser and Grab	Land degradation and soil erosion, Lesotho highlands
2004	Grab and Deschamps	Controls and processes following gully erosion in mires

practices, natural erosion processes, geomorphic thresholds, or to the synergism of some of these factors (see Schwabe, 1989; Calles and Kulander, 1996; Nüsser and Grab, 2002). Regardless of the cause, 49% - 65% of the wetlands within the Senqu River pilot study were previously classified as being extensively damaged, having low water tables and hosting clayey soils (Schwabe, 1989). These damaged wetlands are still able to filter and attenuate water, but at a reduced capacity. Due to a lowered water table and loss of

vegetation, the soil becomes dry and friable and therefore looses its ability to stabilize the soil. Better information on the status of the alpine wetlands is necessary, as it is evident that Lesotho's water is a valuable asset in need of urgent protection (Grobbelaar and Stegman, 1987).

1.7 Objectives

Thus far, there has been no detailed study on the interrelationship between gully erosion and wetland geomorphology and biology for eastern Lesotho. The objective of this study is thus to provide a detailed and quantitative assessment of the bio-geographical and geomorphological attributes of five alpine wetlands in the Sani Valley region (Figure 1.1). The primary aim is to assess the impact that gully erosion has on specific biophysical characteristics within selected eroded mires. The quantification of the environmental impact that gully erosion has is based on a comparison with adjacent non-gullied mires. The biophysical characteristics are also used to assist with the classification of wetland type, delineation of wetland boundaries and identification of soil type (Carter *et al*., 1988).

The second aim of this study is to discuss possibilities for the continued growth of gully systems within the alpine mires. The level of sensitivity of the landscape will be addressed by examining the environmental factors such as soil characteristics and faunal pressures. Aside from extrinsic factors contributing to the destabilization of valley head equilibrium, intrinsic forces such as drainage basin area and slope gradient are often cited as controlling the spatial occurrence of gullies. Schumm and Hadley's (1957) initial work

on slope thresholds found that oversteepened slopes, in relation to the drainage basin area, were a causative factor for gullying at some sites in the United States. This idea has since been widely accepted, and found to apply to many semi-arid areas of North America, Australia and South Africa (e.g. Patton and Schumm, 1975; Imeson and Kwaad, 1980; Schumm, 1981; Beckhedahl *et al*., 1988; Campbell, 1989; Bocco, 1991; Prosser and Winchester, 1996). This idea is often paired with the hypothesis that headcut migration ceases once at a critical slope in relation to the drainage basin area (Montgomery and Dietrich, 1988).

This study would not be complete without an assessment of the wetland gullies, and thus takes a morphometric approach in describing the wetland gully morphologies, similar to those undertaken by Ebisimiju and Ekiti (1989) and Whitlow (1994a, b).

1.8 Research Hypotheses

Principle Aim: to assess the impact that gully erosion has on specific biophysical characteristics within selected eroded mires.

1. Across the expanses of non-gullied wetlands, soil moisture has a uniform pattern of soil saturation, whilst on gullied wetlands the soil moisture gradient is more mosaic-like and is driest adjacent to the gully.
2. Gullied wetlands are more susceptible to dryland plant invasions than non-gullied wetlands due to disturbances in hydrology and soil mechanics.
3. Gullied wetlands have increased microhabitat diversity and as such, are more species-rich than non-gullied wetlands.

Secondary Aim: to discuss plausible reasons for the perpetuation of gully erosion within the mires.

It has been noted that eroded wetlands in Lesotho have a clayey texture, whilst intact wetlands have a loamy texture (Schwabe, 1989). The present study suggests that there is a connection between soil type and the occurrence of gullied wetlands.

4. Gullied wetlands are more inherently erodible when compared to the soil type of wetlands that have not been gullied. The peat within the gullied wetlands has greater clay content than the non-gullied wetlands.
5. It is expected that gully heads form at locally steepened sections of the valley head floor.

Chapter 2

Environmental Setting

2.1 Study Area

The study area is located within the Drakensberg Mountains of eastern Lesotho (Figure 2.1). This mountain range is the highest of Lesotho's four mountain ranges (max. = 3 482 m a.s.l.) and represents the most prominent section of the Great Escarpment (Backéus and Grab, 1995; Nüsser, 2002). The Great Escarpment forms a natural border between Lesotho and South Africa and is at the eastern edge of the southern African plateau (Partridge and Maud, 1987; Grab, 1996). The core study area is located approximately 3.5 km southwest of the Lesotho border post at Sani Top (2957 m a.s.l., 29˚30'S; 29˚20'E) (Figure 2.1). The north-facing aspect of the Sekhokong Range has been eroded into at least four valley heads that have broad, gentle (0˚-10˚) floors. Low-elevation erosional basalt ridges separate the individual valley heads from each other. The valley heads have been numbered for this study in an east to west direction, with the most eastern valley head (#1) situated near the escarpment edge (Figure 2.1). Valley heads 2, 3 and 4 have previously been labelled as B, C and D respectively, by Marker (1994), and as

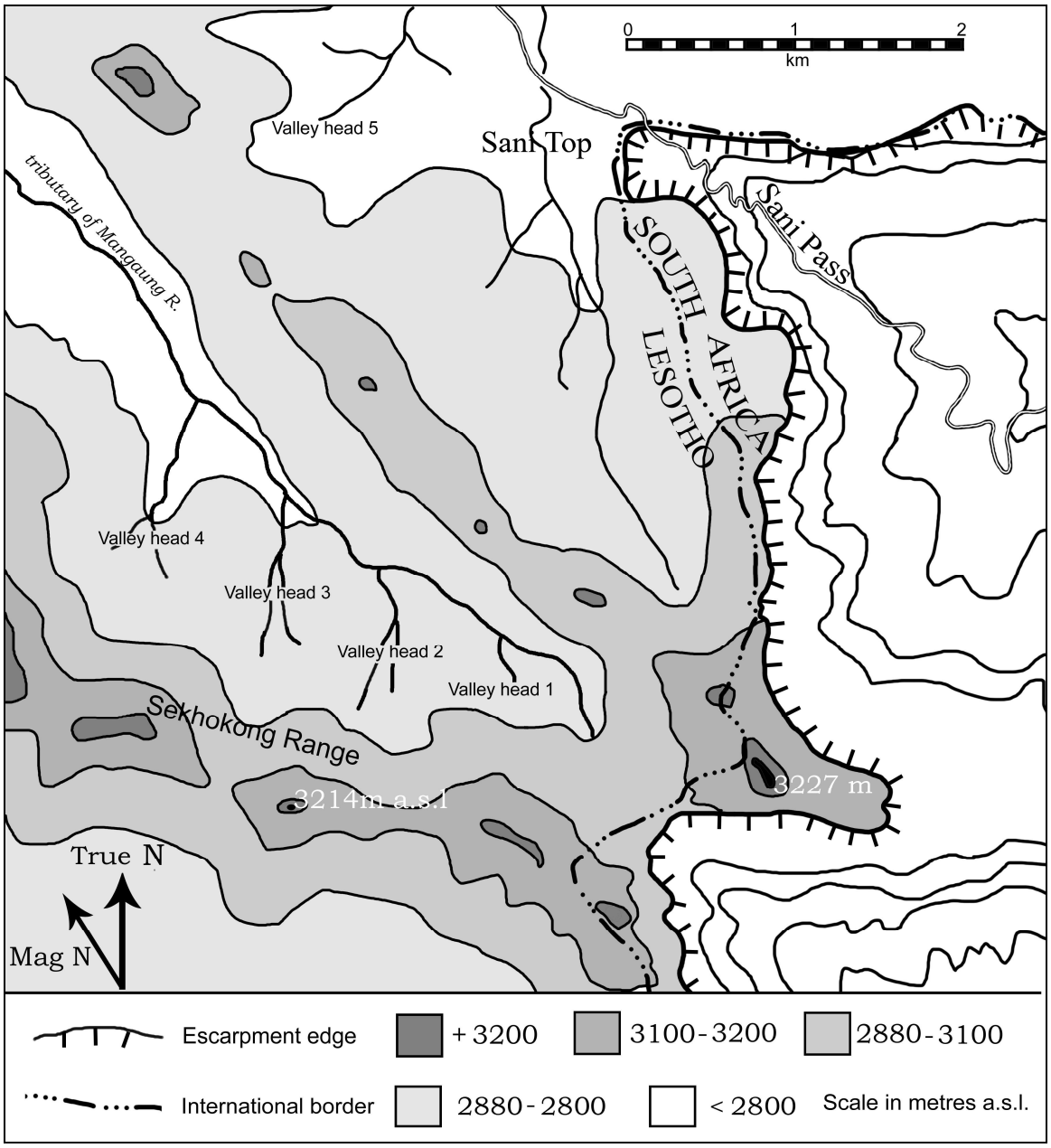


Figure 2.1 Locations of the five study sites within the valley heads, south of Sani Pass, Lesotho. Darkest shade represents peak elevations.

A, B and C respectively, by Marker and Whittington (1971). Valley heads 1 – 4 are north-facing, whilst valley head 5 faces northeast. The floors of the five valley heads are located at similar elevations (between 2900 m a.s.l. and 2970 m a.s.l). The flat floors of the valley heads are occupied by wetlands that vary in size from ca. 1.75 km^2 to 2.4 km^2.

Each of the wetlands within the five valley heads has a tributary stream flowing through it. Each valley head is bounded by rilled slopes and all have at least one valley floor gully cut into the mineral soil and/or peat. All of the studied valley heads are regularly grazed from October to June by horses, cattle, sheep and goats. These gullied and non-gullied valley head wetlands were chosen for comparative geo-ecological and geomorphological analyses, given their similar environmental and anthropogenic settings.

2.2 Climate

2.2.1 Precipitation

The African continent can be divided into eight climatic regions based on temperature and precipitation (Figure 2.2) (Goudie, 1996). The Kingdom of Lesotho lies within the summer rainfall area, of which the Drakensberg is one of its wettest regions (Killick, 1963). The highest precipitation occurs along the northwest face of the Maluti Mountains and along the northeastern to southeastern Drakensberg escarpment (Figure 2.3) (Grobbelaar and Stegman, 1987). Geographically, South Africa and Lesotho lie solely within the subtropical high pressure belt at latitude 30°S (Tyson *et al*., 1976). The belt moves northwards and intensifies in winter, then weakens and moves southwards in summer. A westerly movement of the Indian Ocean high-pressure cell in winter thus results in relatively dry air accompanying the increase in pressure. The dry stable air, together with the winter anticyclone, influences the atmospheric circulation over the subcontinent and results in cloudless skies above 3000 m a.s.l (Tyson *et al*., 1976; Wilken, 1982). In summer, a low pressure cell usually develops over eastern South Africa and the subsidence inversion that rises over the escarpment transports moist air,

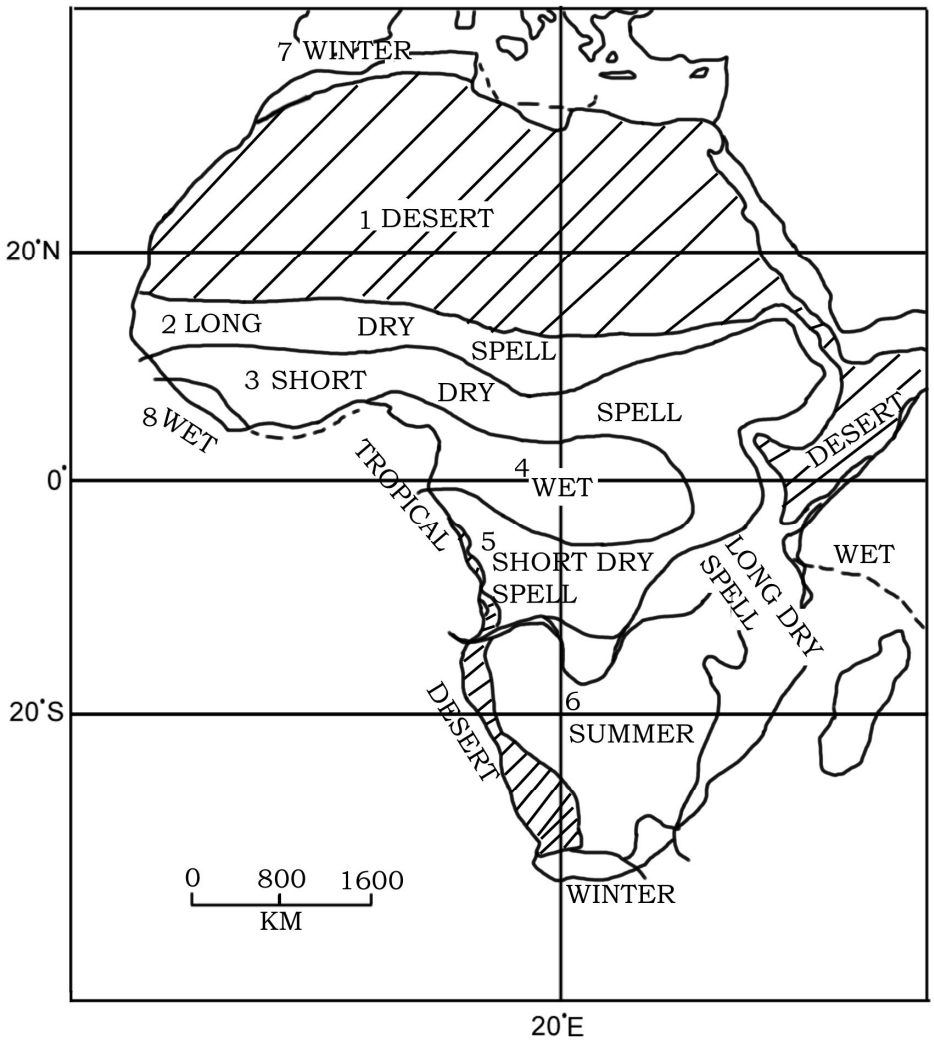


Figure 2.2 Eight climatic regions of Africa (after Goudie, 1996, p37).

thus favouring thunderstorm development (Tyson *et al*., 1976). High intensity thunderstorms can provide 50% of the total precipitation (Preston-Whyte and Tyson, 1988). These thunderstorms originate as eastward moving squall-line storms or as orographically induced thunderstorms (Killick, 1963; Tyson *et al*., 1976).

There is a definite link in Lesotho between variations in mean annual rainfall with elevation. Peak rainfall occurs over the Drakensberg range (Sene *et al*., 1998), with an

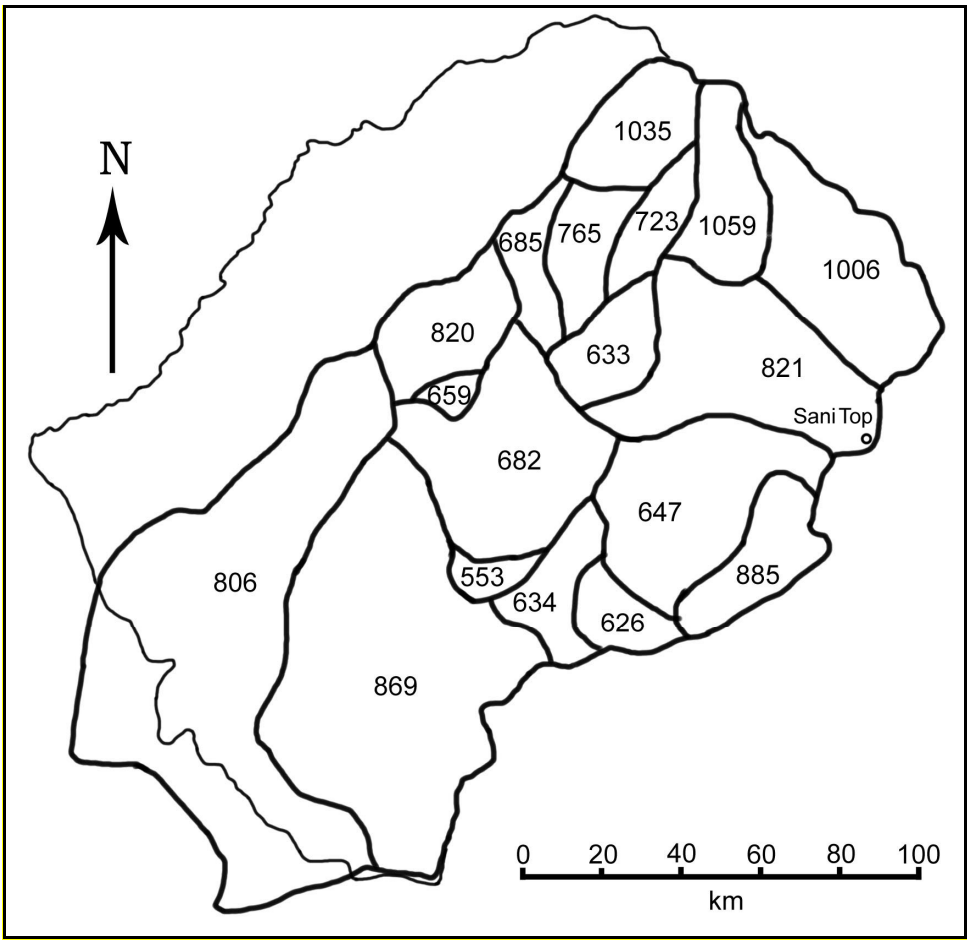


Figure 2.3 Estimated average annual precipitation (in mm) for the years 1975 – 1990 in the highlands region, Lesotho (after Sene *et al.*, 1998, p336).

estimated rainfall average of 1600 mm per annum in the highlands region (Lundén *et al.*, 1990; Nüsser and Grab, 2002); mean runoff coefficients are also greatest in these areas of highest rainfall (Sene *et al.*, 1998). The lowlands region tends to receive an average of 735 mm of precipitation per annum (Hydén and Sekoli, 2000). Within the highlands is a noticeable rainshadow effect particularly over the central region, with east-facing slopes receiving greater amounts of precipitation than slopes facing west (Tyson *et al.*, 1976; Schulze, 1979; Chakela, 1981; Sene *et al.*, 1998). Figure 2.3 illustrates the variation in precipation amounts over the highlands region and indicates that during the period

between 1975 – 1990, the Sani Top area received between 821 – 647 mm; in 2002, Sani Top received an annual rainfall of 742 mm in 2002 (Nel and Sumner, 2005). Approximately 70% - 85% of precipitation falls between October and April, with the wettest months being January, February and March and the driest months being June and July (Tyson *et al*., 1976; Waites, 2000; Nel and Sumner, 2005). Precipitation as snow can occur at any time of the year, but is most common from April to September (Killick, 1963; Jacot-Guillarmod, 1962, 1969; Grab, 2005). It is estimated that snow falls approximately 8 times per annum at high altitudes (Boelhouwers, 1988; Sene *et al*., 1998). In winter, snow may lie on south-facing slopes for several weeks, but usually disappears within a few days on slopes facing north (Mulder and Grab, 2002).

2.2.2 Temperature

The contemporary climate of Lesotho is regionally distinct and has been classified as temperate (Lundén *et al*., 1990). It is characterized by high diurnal and seasonal temperature variations with warm, moist summers and cold, dry winters (Showers, 1989; Boelhouwers and Meiklejohn, 2002; Grab, 2005). However, owing to altitudinal effects, the highlands of Lesotho remain cool to cold throughout the year (van Zinderen Bakker and Werger, 1974; Schmitz and Rooyani, 1987). Frost in the alpine belt is an almost daily occurrence (85%) from June to August, with an average of 180 frost days occurring from April to October (Nixon, 1973; Grab, 1997a; Nüsser and Grab, 2002; Grab, 2005). Mean annual temperature in the lowlands regions is 15ºC (Lundén *et al*., 1990), whereas in the highlands region, the contemporary mean annual temperature is 6ºC with mild, wet summers and cold, dry winters (Mills and Grab, 2005).

2.2.3 Humidity / Evaporation

The atmospheric humidity of Lesotho is generally low (<30%), especially in late winter (Killick, 1963; Nixon, 1973). Low periods of humidity often coincide with the passing of westerly winds, when evaporation is highest (Killick, 1963). Although approximately 735 mm of precipitation falls in the Lesotho lowlands, the evaporation of ca. 1346 mm produces an annual water deficit (Lundén *et al*., 1990).

2.2.4 Wind

The dominant wind direction over the Drakensberg is both diurnally and seasonally variable (Freiman *et al*., 1998). During the day, the mountains act as a heat source; the thermodynamic winds originate from the southeast and are referred to as plain-mountain winds (Tyson, 1968; Tyson and Keen, 1970; Preston-Whyte, 1971). At night, mountain-plain winds blow away from the heat source and down the mountain (Tyson, 1968; Preston-Whyte, 1971); however this statement is contrary to Freiman *et al*. (1998) which states that after midday, regional winds would blow upslope. Over the course of a year, the direction and origin of the winds rotate from being westerly, to easterly and then to southeasterly. In summer (December), the dominant rainbearing winds originate from the east and southeast, whilst in winter (July), the dominant wind is westerly (Killick, 1963). The high velocity westerly winter winds, known as *bergwinds*, often occur from late winter to early spring (Killick, 1963; 1979; Tyson, 1969). Bergwinds are associated with large pre-frontal divergences and often occur as a transition between anticyclonic and cyclonic circulations (Tyson, 1969; Preston-Whyte and Tyson, 1988). These turbulent winds are accompanied by periods of low humidity, minimum levels of soil moisture and

a marked increase in temperature (Killick, 1979; Grab and Simpson, 2000). Bergwinds always result in anomalous winter temperatures that rise at least 5°C and when followed by a cold front, the wind ends abruptly and temperatures drop by as much as 10°C (Tyson, 1969; Grab and Simpson, 2000). In spring (which is the windiest time of the year) and in autumn, the winds blow at varying strengths as they grade from one solstice to the next (Tyson *et al*., 1976).

2.2.5 Insolation

The time of year (season), weather conditions, slope gradient and aspect all affect the amount of solar radiation that a given locality will receive in the high Drakensberg. Apart from daily weather conditions, variation in seasonal and diurnal radiation is completely dependent on slope gradient and aspect (Tyson *et al*., 1976; Granger and Schulze, 1977). As slope gradient increases, the difference between sunshine hours in summer and winter also increases (Tyson *et al*., 1976; Granger and Schulze, 1977). Generally, summer months receive less sunshine than winter months. For example, Killick (1963) recorded an average of 5.5 hours (39%) of possible daily sunshine in December at a site in the 'little Berg' (1817 m a.s.l.) whilst, there were an average of 8.3 hours (82%) of possible daily sunshine in June.

2.2.6 Palaeo-Climate

The climatic history of southern Africa from ca. 35 000 years BP to ca. 4000 years ago consists of several distinct periods. From ca. 35 000 to ca. 22 000 years BP, climatic conditions were mild and moist, especially in the southern Drakensberg (Lewis, 2005).

At this time, known as the Birnam Interstadial, lacustrine deposits and braided streams developed. By ca. 25 000 years BP, climatic conditions became extremely cold and arid at higher altitudes. This period, referred to as the Bottelnek Stadial (Lewis, 2002, 2005) extended from ca. 20 000 – 16 000 years BP and is thought to have been the coldest and driest period in southern Africa over the last 125 000 years (Meadows and Baxter, 1999; Bamford and Grab, 2005). The Last Glacial Maximum (LGM), which occurred ca. 18 000 to 16 000 years BP, was reported as being cool/cold and wet (Lewis, 2002). Data from Cango Cave speleothems indicate a temperature depression of ca. 5°C for the LGM (Tyson, 1987; Meadows and Baxter, 1999). Mean annual rainfall during the LGM was ca. 20% less than contemporary values and resulted in widespread desiccation (Meadows, 2001; Thomas 2004). The lowering of temperatures by ca. 5 to 7°C during the LGM resulted in periglacial and possible glacial conditions at high altitudes in the Drakensberg and Lesotho (Mitchell, 1995; Grab and Simpson, 2000; Lewis, 2005; Mills and Grab, 2005). Although post-glacial conditions remained wet and cool from 14 000 – 12 000 years BP, temperatures began to increase during this time (Tyson, 1987).

By ca. 13 500 years BP, climatic conditions were warm and moist (Tyson, 1987; Lewis, 2002). These warmer, wetter conditions led to organic sedimentation, which is observed in valley head 3, and for which an average date of ca. 13 490 ± 130 years BP has been obtained (Marker, 1994, 1995). Drier conditions from ca. 12 200 to ca. 11 000 years BP brought an end to the early Holocene organic accumulation (Tyson, 1987). A gravel horizon visible within the soil profile of valley head 3 coincides with this drier period (Marker, 1994).

A period of very wet and warm conditions known globally as the Altithermal occurred from ca. 8 000 to 6 000 years BP (Jolly *et al*., 1998; Partridge *et al*., 1999; Grab *et al*., 2005a). Temperatures at this time were 1 - 2°C warmer than at present (Partridge *et al*., 1999). After 6 000 years BP, conditions became much drier but remained warm and was represented by a phase of aeolian deposition (Iriondo, 1999; Meadows and Baxter, 1999; Grab *et al*., 2005a). This period, globally referred to as the Hypsithermal, is regarded as the time of maximum warming during the Holocene and persisted until ca. 5 000 years BP (Schmitz and Rooyani, 1987; Tyson, 1987; Schwabe, 1989; Marker, 1995; Meadows and Baxter, 1999; Lewis, 2002). During the late Holocene, a second phase of organic accumulation occurred within the Sani Top valley heads (ca. 5 000 to 1 000 years BP) (Marker, 1994). During the last 1 000 years, the contemporary climate has been described as dry and stable (Schmitz and Rooyani, 1987). In Lesotho, this drier period resulted in a renewal of gravel movement, valley gradient steepening and gully incision, now accentuated by contemporary grazing pressures (Marker 1994, 1995; Grab *et al*., 2005a).

2.3 Geology

The mountain Kingdom of Lesotho is situated within the eastern half of the broad Karoo Basin (Figure 2.4) (Schmitz and Rooyani, 1987; Schwabe and Whyte, 1993). The Karoo Basin covers ca. 700 000 km^2 and has a maximum thickness of ca. 12 km in its southeastern region (Johnson *et al*., 1997). The Karoo Basin, which lies above the stable granitic floor of the Kaapvaal Craton and Namaqua-Natal Metamorphic Belt, is down-tilted towards the west (Schmitz and Rooyani, 1987; Day and King, 1995). The rocks of Lesotho belong to the upper part of the Karoo Supergroup, which range in age from Late

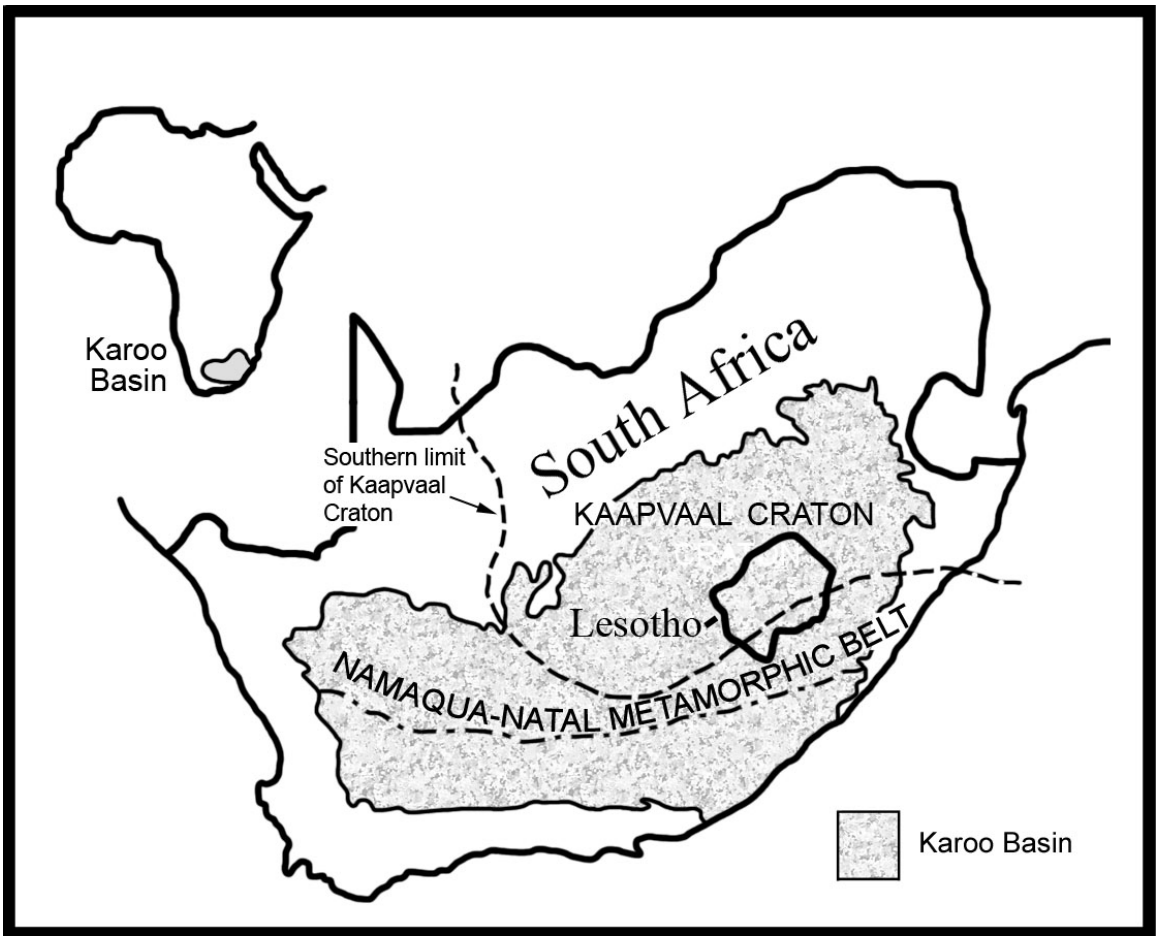


Figure 2.4 General geological structure surrounding Lesotho (Johnson *et al*., 1997, p270).

Carboniferous to Early Jurassic (Nixon, 1973; Johnson *et al*., 1997). The succession includes, from oldest to youngest: the Dwyka Group, Ecca Group, Beaufort Group and the Stormberg Series (including the Molteno, Elliot, Clarens and Drakensberg Formations) (Figure 2.5) (Nixon, 1973; Schmitz and Rooyani, 1987). The rocks of the Stormberg Series underlie much of Lesotho (Low and Robelo, 1996).

The basal part of the succession is the Dwyka Group, comprised of diamictite and other glacially derived rocks (Johnson *et al*., 1997). These rock types were deposited during the Late Carboniferous and Early Permian and consist of six discernible lithofacies. The Ecca

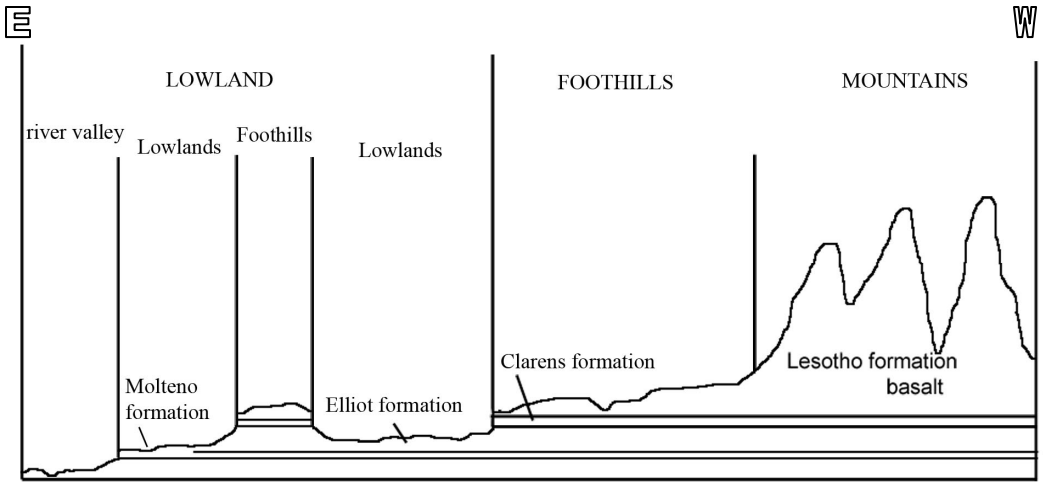


Figure 2.5 Generalized physiogeographic regions of Lesotho (after Schmitz and Rooyani, 1987, p55).

Group consists of mud dominated deposits laid down during the Permian. The 3000 m thick Beaufort Group consists of fossiliferous Mid-Permian to Mid-Triassic aged sandstones and mudrocks (McCarthy and Hancox, 2000; Hancox and Rubidge, 2001).

2.3.1 Stormberg Series

The Stormberg Series formed during the final unloading phase of the Cape Fold Belt during the Late Permian through to the Early Jurassic (Nixon, 1973; Johnson *et al*., 1997; Bordy *et al*., 2004). From oldest to youngest, the series includes the Molteno, Elliot, Clarens and Lesotho Formations. The Molteno Formation has a strongly diachronous contact with the Burgersdorp Formation of the Beaufort Group and a maximum thickness of ca. 600 m in the south and ca.10 m in the north (Nixon, 1973; Schmitz and Rooyani, 1987; Johnson *et al*., 1997; Hancox and Rubidge, 2001). Molteno sandstone originated from the fluvial deposition of boulders, coarse and gritty sandstones, occasional mudstones and rare carbonaceous shales (Nixon, 1973; Johnson *et al*., 1997).

The Elliot Formation, which conforms to the Molteno, ranges in thickness from ca. 500 m in the south to ca. 50 m in the north (Furon, 1963; Nixon, 1973; Johnson *et al*., 1997; Bordy *et al*., 2004). It was formed during the Late Triassic to Early Jurassic and represents the increasingly arid environment that was dominated by aeolian processes. The Elliot Formation is referred to as the Red Beds, as these highly erodible Late Triassic beds consist of alternating sequences of pink, yellow and green sandstones with red, maroon and purple mudstones and shales (King, 1945; Nixon, 1973; SARCCUS, 1982). The soils formed from the Molteno and Elliot Formations include the Duplex soils of the western Lesotho lowlands (Lundén *et al*., 1990).

The Clarens Formation marks the boundary between the foothills and the lowlands of Lesotho, having a gradational contact with the Elliot Formation below and a sharp contact with the Lesotho Formation above (Nixon, 1973). The Clarens Formation, which is of Late Triassic/Early Jurassic age and deposited by aeolian processes before the break-up of Gondwanaland, marks the final stage of Karoo sedimentation (Eriksson, 1986; Johnson *et al*., 1997; Carbutt and Edwards, 2004). The deposited sediments accumulated as aeolian sand along with silt dune and playa lake deposits during a wet desert environment (Eriksson, 1986). The top half of the Clarens Formation is a massive scarp, formed from fine-grained cream-coloured sediment that has a thickness of 70 – 75 m (Nixon, 1973). The lower half of the Clarens Formation is composed of finer-grained red, green and purple sandstones approximately 60 m thick (Nixon, 1973; Schmitz and Rooyani, 1987). Soils formed from the Clarens Formation deposits are generally light to dark in colour and are strongly acidic (Lundén *et al*., 1990).

The Lesotho highlands (> 2000 m a.s.l.) are comprised of Drakensberg flood basalt of the Lesotho Formation (last of the Stormberg Series). These beds are the youngest members of the Karoo sequence and formed ca. 195 to 170 million years ago during the Early Jurassic period (Klug *et al*., 1989; Johnson *et al*., 1997; Carbutt and Edwards, 2004). The Clarens Formation was covered by a series of monotonous lava flows varying from 0.5 m to > 50 m thick, with each flow evenly superimposed upon the next (Nixon, 1973). Each flow layer was composed of a thin basal layer of aphanitic basalt, a layer of pipe amygdales, a layer of weakly vesicular basalt and an upper layer of strongly vesicular basalt (Nixon, 1973; Grab *et al*., 2005b). The Pahoehoe type lavas were laid down in rapid succession (Nixon, 1973; Johnson *et al*., 1997). The speed of the eruptions is evident in the absence of ash lenses and weathered surfaces (Nixon, 1973). The lavas reached the surface by following dolerite dykes, many of which are still apparent in the landscape today (Nixon, 1973; Schmitz and Rooyani, 1987; Johnson *et al*., 1997). The Drakensberg lavas once covered all of Lesotho and much of southern Africa, but today cover only 70% of Lesotho's surface (Schmitz and Rooyani, 1987). Three main tholeiitic lava flows, along with several minor ones, produced a basalt cap that presently has a maximum thickness of ca. 1600 m (Schmitz and Rooyani, 1987; Marake *et al.*, 1998; Grab *et al*., 2005b). These basalt lavas are rich in calcium (10.5%) and high in aluminium (Jacot-Guillarmod, 1969; Klug *et al*., 1989). The basalt becomes slightly less basic with elevation (Nixon, 1973) and tends to weather from its original dark grey, black colour, to a variety of brown, light brown, reddish and purple hues (Klug *et al*., 1989). The neutral soils that are produced from the Drakensberg Formation basalt are dark coloured and are of the Fusi, Popa and Matšana series (Lundén *et al*., 1990).

2.4 Geomorphology

The Lesotho mountains consist of a sequence of horizontal basalt layers, with the terminal layer being a continuous scarp at ca. 3000 m a.s.l. (Boelhouwers and Meiklejohn, 2002). This erosion surface separates Lesotho from the erosion surface of KwaZulu-Natal to the east (at ca. 1200 m a.s.l.), and from the surface of the South African high plateau to the north and west (at ca. 1500 m a.s.l.) (King, 1945). These erosion surfaces are classified as the Basutoland Plateau or peneplane, the Natal Miocene bevel and the High Veld Miocene peneplane, respectively (King, 1945). The Basutoland Plateau, which is in no way connected to faulting, has the appearance of being tilted southward (King, 1945; King and King, 1959). It has been suggested that this impression of tilting is caused by the appearance of a series of horizontal layers that decrease in thickness south/south westwards (Binnie and Partners, 1971). At least five major bevels or peneplains are recognized in Lesotho; their morphology is attributable to erosion cycles, but mostly to structural controls (Nixon, 1973). These surfaces do not occur at the same elevation throughout Lesotho, but all follow the same position relative to the others (Nixon, 1973).

The higher summits in Lesotho are regarded as being cold enough to permit small-scale periglacial processes and associated landform production (e.g. Lewis, 1988; Boelhouwers, 1991; Grab, 1997a). Relict periglacial features and supposed glacial landforms have been identified and used in palaeoenvironmental reconstructions of the alpine environment (Boelhouwers and Meiklejohn, 2002). Some periglacial features and/or processes include, thufur (e.g. Grab 1994, 1997a, 1999, 2005; Backéus and Grab,

1995), flarks (e.g. Backéus, 1989; Backéus and Grab, 1995), needle ice (e.g. Marker and Whittington, 1971; Grab, 1999, 2000), turf exfoliation (e.g. Grab, 2002), frost heaving (e.g. Jacot-Guillarmod, 1969; Grab 2002), soil stripes (e.g. Grab, 1996) and miniature patterned ground (Hanvey and Marker, 1994; Grab, 1997b). The occurrence of localized glacial conditions during the LGM (18 000 – 14 000 year BP) has been suggested (Grab 1996; Mills and Grab, 2005), but literature supporting this theory is "contentious" (Boelhouwers and Meiklejohn, 2002, 51). Most glaciation research reports on marginal and/or limited features (i.e. niche glaciers). Three of the north-facing valley heads in this study (# 2, 3 and 4, see Figure 2.1) for example, have been described as glacial cirques (Marker, 1991, 1994). However, Grab and Hall (1996) have challenged this glacial cirque hypothesis and rather suggest a bog-cirque origin. Because larger-scale Quaternary evidence of glaciation after the LGM is absent, it is accepted by some that glaciation did not occur during this period (Deacon and Lancaster, 1988; Preston-Whyte and Tyson, 2000). However, quantitative work is on-going; for example, Mills and Grab (2005) have recorded radiocarbon dates for a debris ridge at Tsatsa-La-Mangaung (3275 m a.s.l.), which indicate that the ridge was deposited during the Late Pleistocene (21 – 18 kyr) when glaciers were active (Mills and Grab, 2005).

2.5 Soils

The 26 main soil types in Lesotho belong to 6 major soil groups (Table 2.1) (Nixon, 1973). Of these, 80% are located above 2000 m a.s.l. and 6% are highly susceptible to fluvial erosion (Chakela, 1981; Faber and Imeson, 1982; Schmitz and Rooyani, 1987; Rydgren, 1988; Nkalai, 1991; Calles and Kulander, 1996; Rydgren, 1996; Calles and

Table 2.1 Various soil series found in Lesotho (adapted from Marake *et al*., 1998).

Geology	Associated Soil Series
Lesotho Formation	Popa & Matšana
	Fusi & Thabana
	Machache, Nkau,
	Sefikeng, Tumo, Matšaba,
	Seforong, Ralebese, Matela
	Phechela, Khabo, Sofonia,
	Maseru dark
Clarens Formation	Matela, Berea,
	Ntsi, Qalaheng,
	Thoteng, Theko
	Lekhalong, Tšenola, Sani
Molteno & Elliot Formations	Maliehe, Bosiu, Majara
	Moshoeshoe, Tsiki, Sephula,
	Tšakholo, Maseru
	Leribe, Matela, Qalo, Hololo, Rama
	Matšaba, Seforong, Ralebese, Kubu, Khabo
	Khabo, Kubu, Bela, Phechela, Maseru, Maseru dark
	Caledon, Sofonia, Kolonyama
	Phechela
	Ralebese

Stålnacke, 2000). According to the USDA system of soils classification, *Mollisols* and *Alfisols* are the main soil orders identified in Lesotho (Rydgren, 1996). In the mountains, the *Mollisols* and *Lithosols* are the main soils and cover the slopes and valleys (Schmitz and Rooyani, 1987; Schwabe and Whtye, 1993; Rydgren, 1996). The mountain *Mollisols* generally lack a B horizon and tend to be shallow (< 40 cm) (Klug *et al*., 1989). Where deeper *Mollisols* occur, the B horizon tends to be yellowish brown and *cambic* (Bw). Organic matter content is high in *Mollisols* and varies between 10 – 20%, although this amount depends on locality and climatic conditions (Klug *et al*., 1989; pers obs, 2002). The *Alfisols* are well developed in the lowlands and foothills and have an *ochric* epipedon and *argillic* (Bt) horizon (Schmitz and Rooyani, 1987; Schwabe and Whtye, 1993; Rydgren, 1996). The well-known and easily eroded Duplex soils of the lowlands belong

to the class of *Alfisols.* The term *Duplex* refers to the abrupt textural difference between the A and B horizons (Nixon, 1973).

The primary soil producing factors in the mountains of Lesotho are *in situ* basalt weathering, colluviation and eluviation (Schmitz and Rooyani, 1987; Klug *et al*., 1989). Since the parent material of the highland soils is basalt, it tends to have a high calcium (10.5%) and montmorillonite clay content, resulting in a neutral pH with a consequential high base saturation (Schmitz and Rooyani, 1987; Klug *et al*., 1989). As with most alpine soils, the mountain soils of Lesotho generally have higher organic matter content than the lowland soils (Young, 1976; Morris *et al*., 1993). The soils of the mountain area, where frost action and turf exfoliation are active, generally have a udic soil moisture regime and a cryic temperature regime (Mack, 1981; Grab, 1997a, 1999, 2002; Klug *et al*., 1989). Areas with hydromorphic soils at high altitudes (above 2750 m a.s.l.) include valley bottoms, gentle slopes and springs (Jacot-Guillarmod, 1962, 1969; Schwabe, 1989, 1995). These soils are rich in organic matter and are classified as *Histosols* (Backéus, 1988; Schwabe, 1989; Morris *et al*., 1993), *Borosaprists* (Schmitz and Rooyani, 1987) and *Aquic Mollisols* (Klug *et al*., 1989). Depending on the wetland investigated, and the criteria used to define the soil, such terms are often found to describe the same wetland.

2.6 Vegetation

Lesotho is a grassland country in which five grassland types occur within the various vegetation belts (Lundén *et al*., 1990; Marake *et al*., 1998) (Table 2.2). Natural tree growth in Lesotho is almost completely absent (van Zinderen Bakker and Werger, 1974;

Table 2.2 Grassland biomes in Lesotho (adapted from Schmitz and Rooyani, 1987).

Highveld Grassland Types			Area (km2)
	Moist cold highveld	*Cymbopogon-Themeda* Veld Transition; *Aristida unciformis*-Eragrostis *plana* Grassland North-eastern Sandy Highveld; Highland Sweet Grassland; Moist Cool Temperate Grassland; Highland Sourveld to *Cymbopogon-Themeda* Veld Transition	6689
	Moist cool highveld	*Cymbopogon-Themeda* grassland; *Themeda triandra-Eragrostis curvula* grassland	198
	Wet cold highveld	*Rhus dentata-Leucosidea* Thicket; High Cold Sourveld	58
Mountain Grassland Types	Afro mountain	*Themeda-Festuca* Veld; *Monocymbium ceresiiforme-Tristachya* l*eucothrix* grassland	15 489
	Alti mountain	*Merxmuallera-Festuca* grassland;*Themeda-Festuca* Alpine veld; *Erica-Helichrysum* Heath; Erica-*Helichrysum-Eumorphia* Sedge Heath	7118
	Moist upland	*Hyparrhenia hirta* Tall grassland; Highland Sourveld; Dohne Sourveld	3
TOTAL AREA			**29 558**

van Zinderen Bakker, 1981; Schwabe and Whyte, 1993; Backéus and Grab, 1995). In eastern Lesotho, the arrangement and locations of botanical communities tend to be more strongly influenced by altitude and slope aspect, rather than soil type (Nixon, 1973; Carbutt and Edwards, 2004). In addition, many years of overgrazing and uncontrolled fires have influenced the development of present-day plant associations (Nixon, 1973; Killick, 1990). The Drakensberg has been divided into altitudinal zones according to dominant plant communities (Killick, 1963). The montane belt extends from 1280 m – 1830 m a.s.l, the subalpine belt from 1830 m – 2865 m a.s.l and the Afro-alpine belt from 2865 m – 3500 m a.s.l (Killick, 1963) (Figure 2.6).

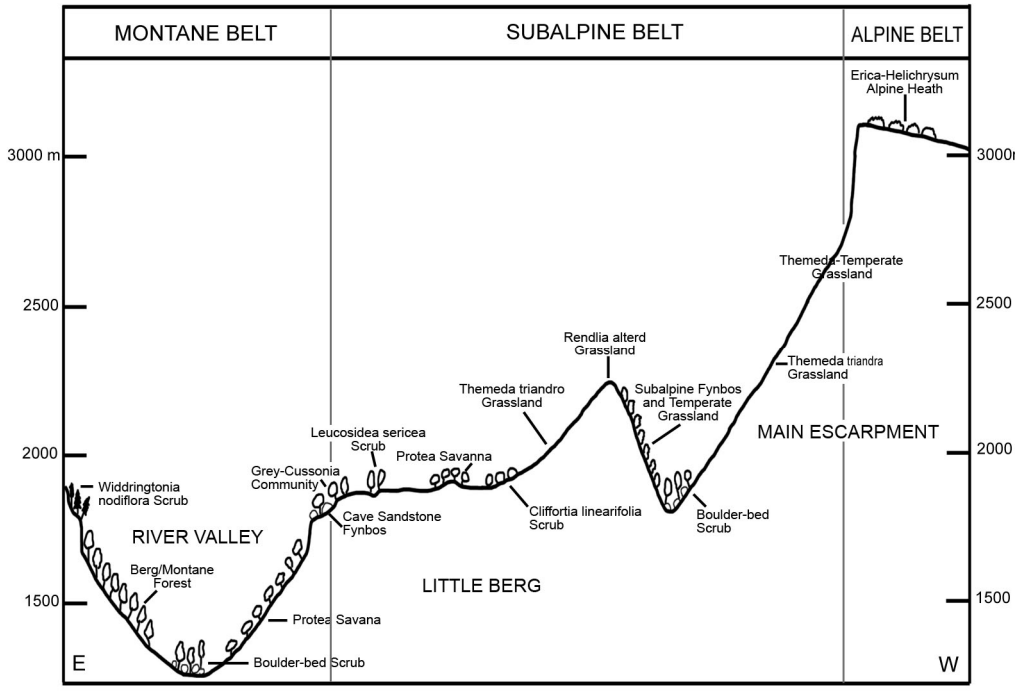


Figure 2.6 Division of the Drakensberg into altitudinal zones according to dominant plant communities (after Killick, 1990, p24).

Under the current practice of veld burning, *Themeda triandra* (grass) has become the dominant botanical species in the montane belt (Killick, 1990). When protected from fires, *T. triandra* becomes replaced by *Hyparrhenia* grassland. The two dominant types of Proteas found in the river valleys of the montane belt are *P. caffra* and *P. roupelliae* (Killick, 1990), but Protea species become increasingly scarce towards the subalpine and alpine belts.

The most common grassland type within the subalpine belt is the *Themeda triandra* grassland (Killick, 1963). *Themeda triandra*, also locally known as Red grass or as 'Seboku', is common to the Little Berg with its associated herbs, including some orchids between 1830 - 2140 m a.s.l. (Killick, 1990). On the gentle north-facing slopes, *T. triandra* can extend to 2600 – 3050 m a.s.l., whilst on the cooler, steeper and moister

south-facing slopes its habitat becomes restricted and is rarely found above 2 750 m a.s.l. (Killick, 1963; Morris, *et al*., 1993). This grassland community is economically valuable to the Basotho people, given its palatability to their livestock (Nixon, 1973; Schwabe, 1989; Schwabe and Whyte, 1993; Nüsser, 2002). Above 2600 m a.s.l, *T. triandra* occurs with the temperate and alpine grassland species *Festuca costata* (Tussock Fescue) and *Festuca caprina* (Goat Beard Grass) (Killick, 1990).

The growing season in the alpine belt is affected by altitude as well as topographic and climatic gradients (i.e. hail, drought and frost), and thus is unpredictable (Carbutt and Edwards, 2004). Plant growth is limited by available moisture and temperature, and thus plant species must be adapted to a wide variety of stressors (Showers, 1989). The alpine belt is a treeless zone consisting of well-suited short grasses, cushion plants, low lying woody heath communities and hydromorphic plant species, all of which display xeromorphic adaptations to the severe alpine conditions (Killick, 1963; Killick, 1990). The climax community of the alpine belt is the Erica – Helichrysum Heath (Jacot-Guillarmod, 1969; Killick, 1990; Morris *et al.*, 1993; Marake *et al*., 1998; Pooley, 1998). Below 3200 m a.s.l., the most extensive community is *Erica dominans* heath, whilst at higher altitudes, this small woody plant with purple flowers commonly occurs with *Helichrysum trilineatum* (Alpine Everlasting) (Killick, 1990). The dominant grasses of the alpine belt include *Merxmuellera disticha*, *Festuca caprina, Pentaschistis oreodoxa* and sometimes *T. triandra* (Killick, 1990). The homogenous grassland species of this altitude have variable heights, are often patchy and irregular and are always interrupted by mud patches and bare stony ground (Killick, 1990; pers. obs.). *F. caprina* is most

often found on south-facing slopes and is most palatable when young and flowering. Together with *Merxmuellera,* the 'Letsiri' grassland is dominant in the landscape and important as fodder grass (Quinlan and Morris, 1994), with *Merxmuellera spp.* and *T. triandra* having the highest carrying capacities of the grasses (Schwabe and Whyte, 1993). However, a combination of fire and overgrazing can lead to damaged and disturbed areas, with consequent encroachment of invasive species such as *Karroochloa purpurea* (Hare Grass) and *Chrysocoma ciliata* (Bitter Bush). The Karroid scrub species, *C. ciliata*, also known as *C. tenufolia*, *C. oblongifolia* and '*Sehalahala'*, is an aggressive and unpalatable shrub estimated to have covered 13% of the mountain area by 1938 (Killick, 1963). By 1998, *C. ciliata* covered 359 680 hectares (16%) of rangeland (Marake *et al*., 1998).

The hydromorphic communities within the alpine belt are located within mires that have originated along seepage areas and in riverheads (Killick, 1990; Marake *et al*., 1998). The wetland vegetation is composed of facultative wetland angiosperms, bryophytes and some algae, all of which are low-growing species rarely exceeding 5 to 6 cm in height (Jacot-Guillarmod, 1969). The locations of these and other wetland herbs, as well as those often found on the 'dry islands' and hummocks within the wetlands, are subject to saturation and disturbance (Hobson *et al*., 1970; Schwabe, 1989, 1995). Among some of the species found within the mire expanses are *Ranunculus meyeri*, *R. multifidus*, *Haplocarpha nervosa*, *Cotula paludosa* and *Trifolium burchellianum* (Killick, 1963, 1990; Schwabe, 1989; Pooley, 1998; Nüsser, 2002). Along the mire margin are species such as *R. meyeri*, *Erica dominans*, *Geranium multisectum* and *Kniphofia caulescens*,

whilst *Merxmuellera drakensbergensis* (Broom Grass) and *K. caulescens* are also common along stream banks (Killick, 1990). Of the ca. 1 750 species of angiosperms that occur in Lesotho, ca. 16% - 30% are endemic to the mountain area and most occur within the highland wetlands (Marake *et al*., 1998; Carbutt and Edwards, 2004). Some of the wetland endemics, not necessarily found near Sani Top, include *Aloe polyphylla*, *Aponogeton ranunculiformis, Kniphofia hirsutas, Haplomitrium hookeri and Leptodontium gemmascens* (Meakins and Duckett, 1993; Marake *et al*., 1998).

The grassland and wetland vegetation within the eastern highlands of Lesotho are an important resource, not only in terms of their intrinsic value and tourist interest, but also for the grazing economy and their hydrological control (Nüsser, 2002). For instance, wetlands provide 2.5 times more forage than the surrounding grasslands and provide a large vegetation biomass of 3700.6kg/TUF/Ha (Schwabe and Whyte, 1993). However, whilst *C. ciliata* encroachment and wetland gullying increase, these important fodder grasses are forced to decrease (van Zinderen Bakker, 1981; Marker, 1994; Schwabe, 1995; Nüsser, 2002; Grab and Deschamps, 2004).

2.7 Wetlands

The 'mokhaobos' or wetlands of the eastern highlands of Lesotho are of post-glacial age, with the oldest date thus far recorded at 13 490 C^{14} years BP (Marker, 1994, 1995). Although it has been suggested that the organic matter content is not high enough to be classified as peat (e.g. Klug *et al*., 1989), it is accepted by others that the wetlands are / were peat forming (Jacot-Guillarmod, 1962; Grobbelaar and Stegmann, 1987; Meakins

and Duckett, 1993; Backéus and Grab, 1995; Morris and Grab, 1997). The water within these peat-forming *mires* is said to originate through springs from kimberlite pipes, from seepage above the impermeable Clarens Sandstone, or exclusively from the accumulation of precipitation (Jacot-Guillarmod, 1969; van Zinderen Bakker and Werger, 1974; Schmitz and Rooyani, 1987; Schwabe, 1989). Most of the mires are said to be 20 to 30 hectares in size and tend to be triangular in shape, with the base upslope (van Zinderen Bakker and Werger, 1974; Meakins and Duckett, 1993; Morris and Grab, 1997). The mires are generally absent below 2 500 m a.s.l. (Grobbelaar and Stegman, 1987), although many authors place that value around 2 750 m a.s.l. (e.g. Schwabe, 1989, 1995; Backéus and Grab, 1995). Peat thickness often ranges from 2 – 3 m, with an average of 0.25 mm of peat accumulating per annum (van Zinderen Bakker and Werger, 1974). The peat has formed solely from the roots of vascular plants and is said to be highly humified and amorphous (Backéus and Grab, 1995). The soil profiles within several mires at Sani Top consist of alternating layers of thick peaty horizons and clay and orange gravels (Marker, 1994, 1995, 1998; pers obs). These alternating horizons represent climatic transitions from cool/humid to cold/dry, respectively (van Zinderen Bakker and Werger, 1974; Hanvey and Marker, 1994; Marker, 1995).

Only a few attempts have been made to classify the wetlands of eastern Lesotho by examining their soil characteristics, pH and/or vegetation type (i.e. Klug *et al*., 1989; Schwabe, 1989). The terms *fen*, *mire*, *bog*, *sponge* and *wetland* seem to be used interchangeably without regard to actual meaning, or to the qualifications required for the terms to be used correctly. Both types of mires, namely bogs and fens, have formed

within this region of Lesotho. However, the literature often uses the term ‘bog’ rather than the more general term ‘mire’ and the term ‘mire’ instead of the more specific terms ‘bog’ and ‘fen’. Further criteria must be defined to separate these terms in order to understand whether the feature is truly a bog or a fen. True bogs are apparently found on high altitude south-facing slopes of Lesotho, have a high water table and high organic matter content. The bogs should also be isolated systems or ombrotrophic, meaning that their water is derived solely from precipitation (Schwabe, 1989). Precipitation has been recorded as having a pH value of 5.2 (Grobbelaar and Stegman, 1987), thus the bogs are species poor or oligotrophic (van Zinderen Bakker and Werger, 1974; Schwabe, 1995; Jonasson and Shaver, 1999). The bogs are covered by dense sedge and grass vegetation (Schwabe, 1989).

Although fens are also covered by sedge and grass vegetation, they are said to have a relatively low water table, less organic matter, high clay content, occur most often on north-facing slopes and have a stream running through them (Schwabe, 1989). The fens have a minerotrophic, soligenous water supply and thus have a circum-neutral pH (Wassen *et al*., 1990; Backéus and Grab, 1995; Schwabe, 1995). The water that exits the mire filtering system is of excellent quality, meaning that it is clear, low in ions and has a pH value between 6.3 and 8 (Jacot-Guillarmod, 1969; van Zinderen Bakker and Werger, 1974; Grobbelaar and Stegman, 1987; pers. obs. 2002). Fens are also generally more damaged by gully erosion than are bogs. The difference in the extent of damage is represented by their slope positions, given that most grazing posts and associated denuded vegetation are located on the warmer, north-facing aspects (Nüsser and Grab,

2002). The surfaces of both bogs and fens are covered in areas by frost-induced hummocks (thufur), with hummock density varying according to mire type, such that bogs generally host the highest densities (Killick, 1963; Schwabe, 1989, 1995). The mounds have an average height of 16 to 30 cm and a diameter of 50 to 70 cm (van Zinderen Bakker and Werger, 1974; Backéus and Grab, 1995, Grab, 1997a). When undisturbed, the hummocks are covered with mire vegetation but may be broken by grazing, needle ice action and solifluction (van Zinderen Bakker and Werger, 1974).

Many of the mires are surrounded by tall tussocks of *Merxmuellera drakensbergensis* and *Kniphofia caulescens* which, when present, help to attenuate surface flow onto the thick wetland soils (Grab and Morris, 1999). However, annual burning of these tussocks is accelerating soil desiccation, runoff and consequential fan development onto the mires (Grab and Morris, 1999). With the aid of GIS (Geographical Information Systems) and LandSat Imagery, all of the wetlands in the Maluti/Drakensberg mountain catchment have had their distributions mapped, carrying capacities calculated and quality assessed (Schwabe, 1989; Schwabe and Whyte, 1993). These studies concluded that the carrying capacity of wetlands in good condition would be ca. 2.5 times greater than that of grasslands (Schwabe and Whyte, 1993). However, the study also revealed that 49% of the surveyed mires hosted plant species indicative of wetland disturbance (Schwabe, 1995). Yet, the report did not include a complete quantitative description of any individual valley head or offer quantitative explanations for the patterns observed either in valley head vegetation distribution, gully characteristics or soil types. It is believed that most of

the ‘bogs’ in Lesotho are degrading (Meakins and Duckett, 1993; Schwabe and Whyte, 1993).

Chapter 3

Methodology

Deep wetland gully erosion occurs within two of the five valley heads included in the study (Figures 3.1 – 3.4). To establish the impacts that gullying has had on the valley head geo-ecology, detailed assessments are made for several key geomorphological and environmental features of the valley heads. To test the hypotheses outlined in Chapter 1, the study incorporates field and laboratory techniques from a variety of disciplines including ecology, pedology and geomorphology.

3.1 Soil Moisture

To measure and quantify the surface soil moisture gradient of gullied mires and compare them to non-gullied mires, basic quantitative field methods were applied. Due to time constraints for field research, soil moisture was only recorded during early and late spring of 2001 and summer of 2002 (Figures 3.1 – 3.4). The line transect method was best suited to depict the hydrological gradient evident across a representative area of the valley floor (see McLean and Cook, 1968). Each transect was clearly marked using wooden pegs and

spray painted rocks to ensure precision during repeat analyses. Other variables that inherently control soil moisture were also recorded along each transect including groundwater seepage, erosion, animal burrows and slope gradient (Taylor & Seastedt, 1994).

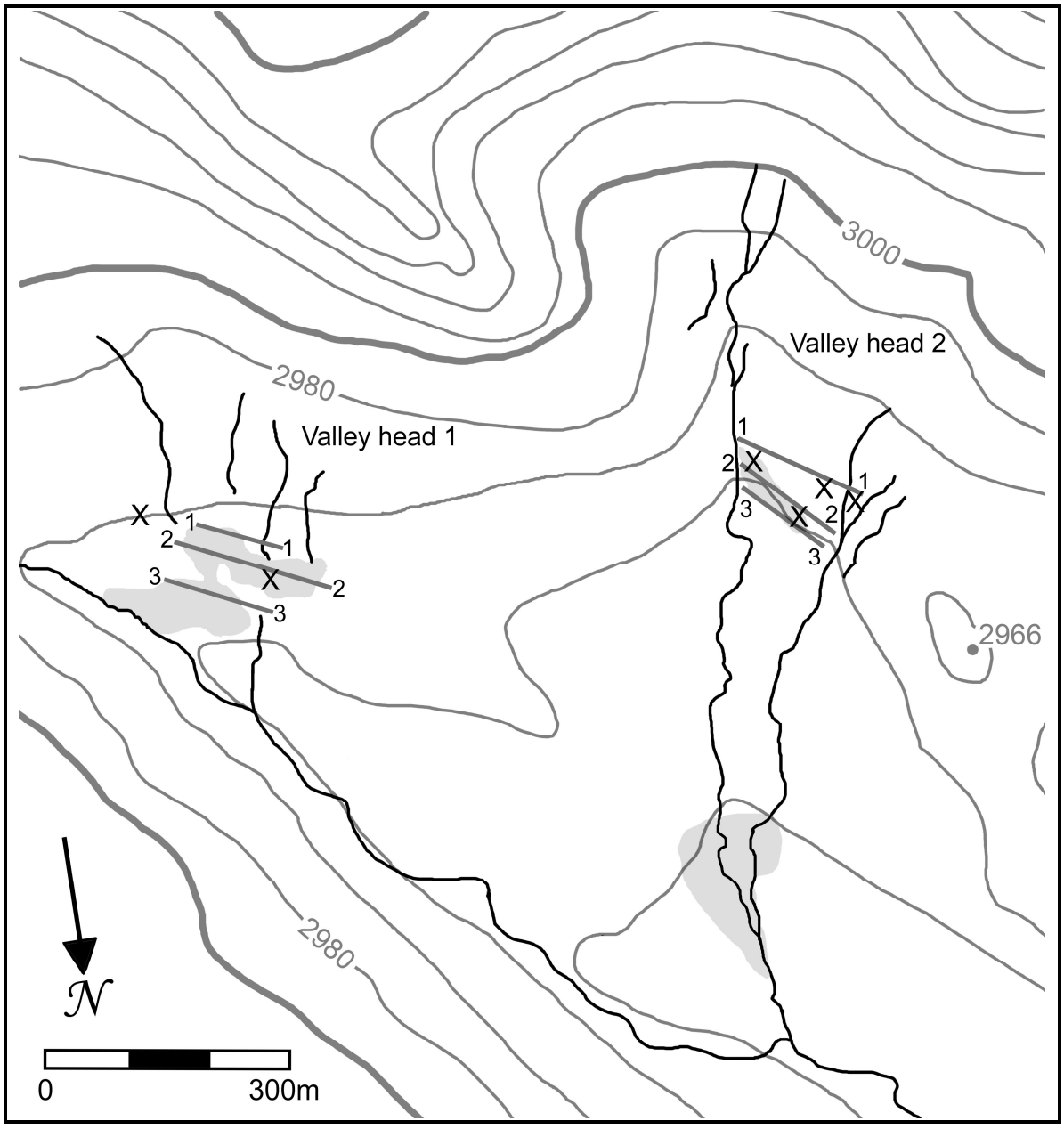


Figure 3.1 Topographic outlines of valley heads 1 (on the left) and 2 (on the right). Shaded areas represent wetland locations. X – approximate locations of soil pits. The horizontal lines represent transects used to measure either soil moisture and/or soil properties. Main transects used are lines 2 in each valley head.

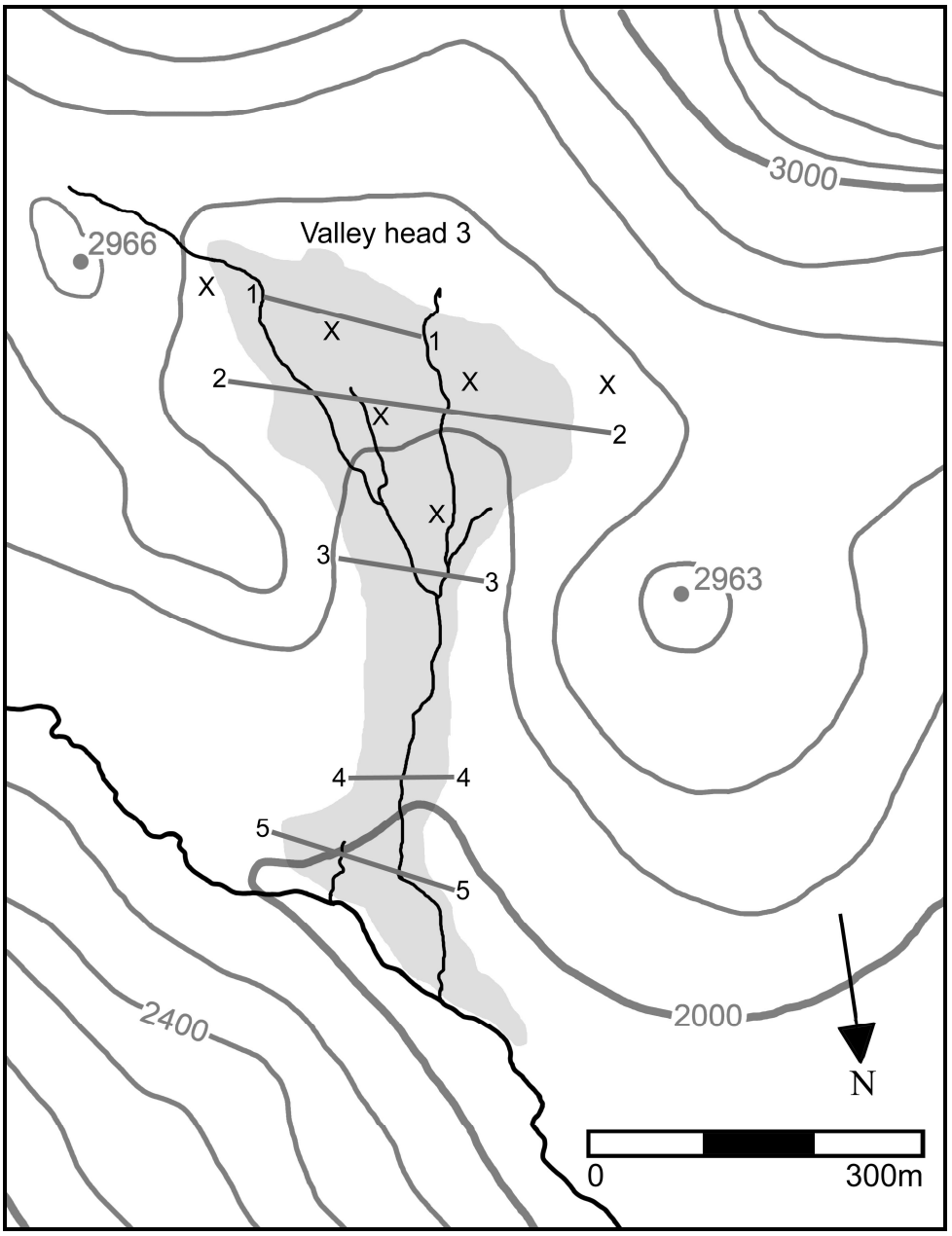


Figure 3.2 Topographic outline of valley head 3. Shaded area represents wetland location. X – approximate locations of soil pits. The horizontal lines represent transects used to measure either soil moisture and/or soil properties. Transect line #1 is the main transect used in vegetation and soil surveys.

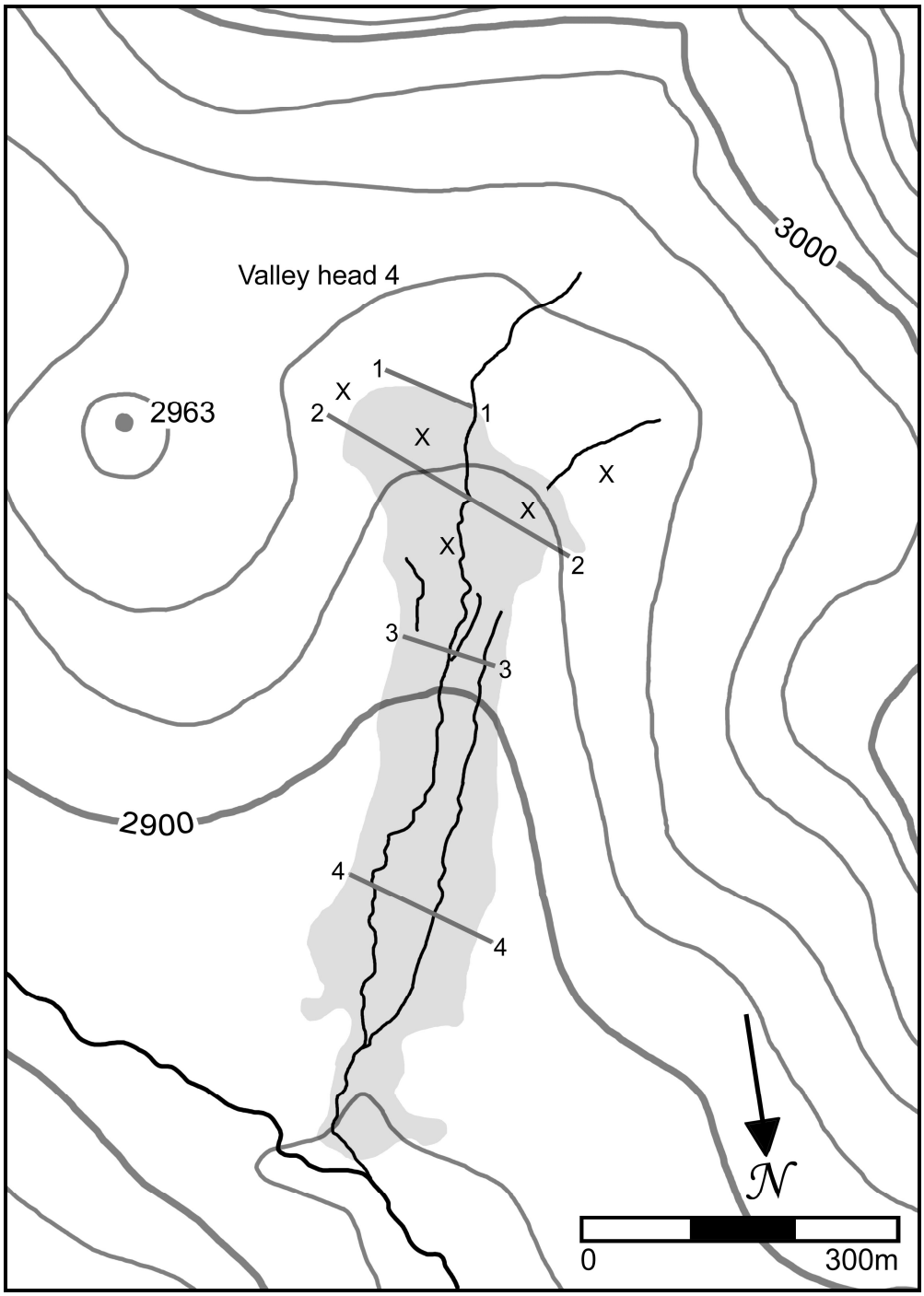


Figure 3.3 topographic outlines of valley head 4. Shaded area represents wetland location. X – approximate locations of soil pits. The horizontal lines represent transects used to measure either soil moisture and/or soil properties. Transect line #2 was used for the main vegetation and soil surveys.

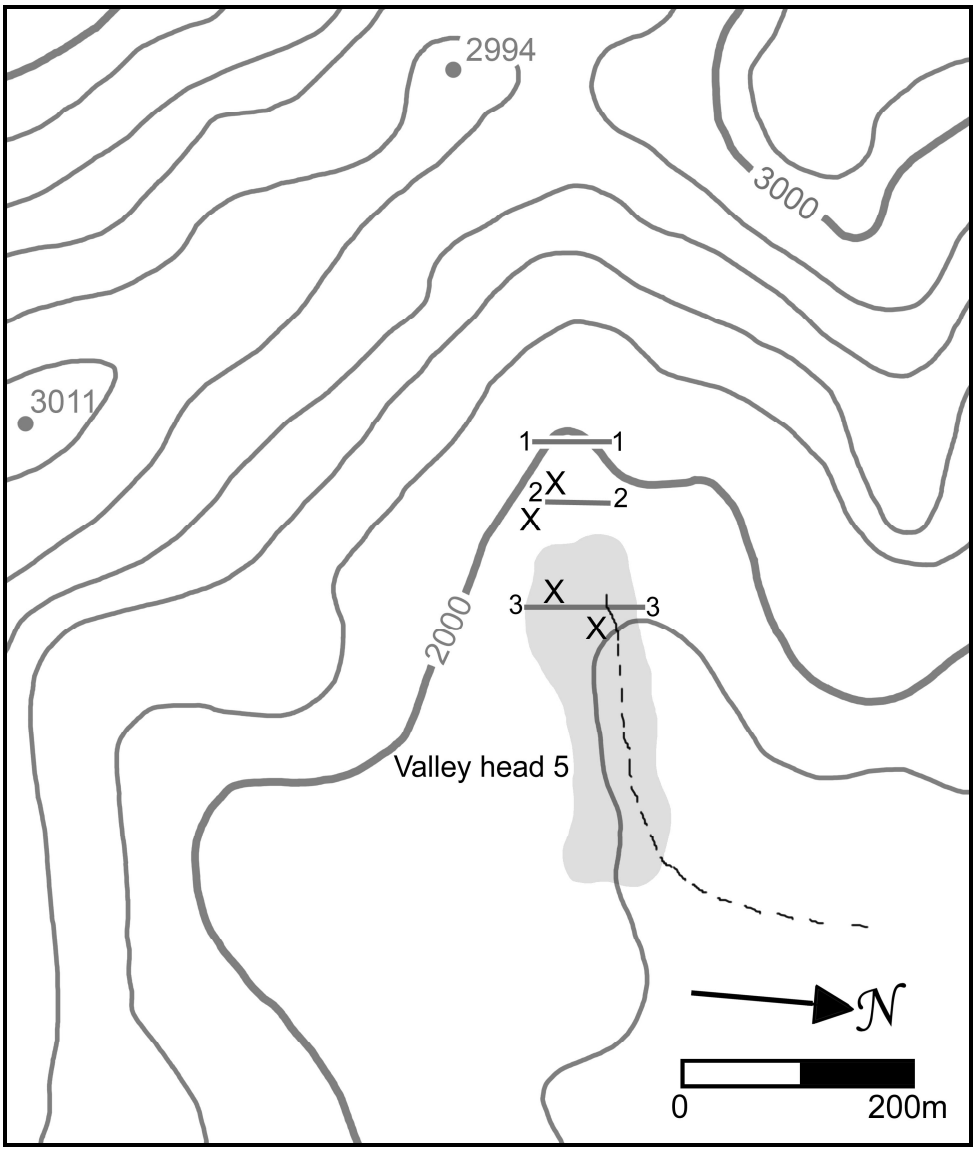


Figure 3.4 Topographic outline of valley head 5. Shaded area represents wetland location. X – approximate locations of soil pits. The horizontal lines represent transects used to measure either soil moisture and/or soil properties. Transect line #3 was used for the main vegetation and soil surveys.

3.1.1 Soil Moisture Belt Transect Procedure

In preparation, a 100 m long builders line was marked with tape at 1 m intervals. The 1 m wide belt transects were laid out perpendicular to various vegetation communities in an east-west direction. This belt transect was divided into 1 m^2 units along the full length of the transect within each valley head (Figure 3.5). Soil moisture was recorded using an

Eijkelkamp soil moisture Theta meter/probe (type HH1). Recordings were taken 5 times within each unit, with one measurement at each corner and one in the centre (Figure 3.5). The probes of the Theta meter were inserted to a depth of 10 cm, which represents the approximate vegetation rooting zone (see Schmitz and Rooyani, 1987). This higher resolution method of soil moisture measurement was undertaken in autumn (April 2002) as this season has relatively low precipitation amounts. This method was performed to collect detailed information about latent soil moisture characteristics and environmental variables that control it. Various shades of blue were used to illustrate the soil moisture gradient within the belt transect and within each $1m^2$. For example, as illustrated in Figure 3.5, the colour shading of the top quadrant represents the average moisture value using the values from corners 1, 2 and 5, whereas the bottom quadrant was calculated using values from corners 3, 5 and 4. From lightest to darkest shade, the moisture gradient was divided into very dry (0 - 2), dry (2 - 3.5), moist (3.5 - 5), wet (5 - <100) and saturated (100).

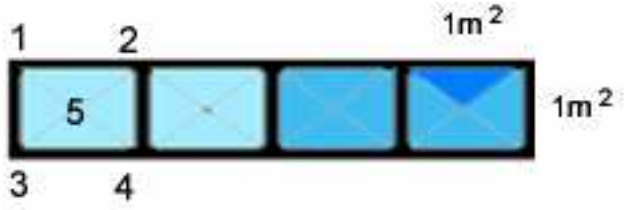


Figure 3.5 Example of the belt transect design used to record detailed soil moisture values.

3.1.2 Soil Moisture Line Transects

To better gauge the overall saturation levels of the wetland, additional line transects were constructed above and below the belt transects. The line transects were separated by at

least 85 m, with an average distance between transects of 130 m. These transects were approximately parallel to one another (Figures 3.1 – 3.4). Soil moisture along the line transects was recorded during excursions in September and November 2001 and in January 2002. During these excursions, the wetland was observed as saturated, thus the intention of these transects was to portray an overall picture of the wetland surface only, as well as to document the general trend. Measurements were taken at 5 m intervals along these line transects.

Transect length and number depended on the width and length of the wetland, with longer wetlands requiring four transects, while shorter ones required only three (see Table 3.1).

Table 3.1 Lengths of the line transects in each valley head; listed from upslope to downslope positions.

Valley 1	190m, 190m, 110m
Valley 2	124m, 140m, 170m
Valley 3	21m, 91m, 175m, 130m
Valley 4	107m, 170m, 300m, 180m
Valley 5	85m, 150m, 150m, 115m

3.1.3 Correlations With Soil Moisture Data

A profile for each transect was individually designed, using transect distance and the soil moisture measurement as the independent and dependent variables, respectively. Spearman's rank correlation coefficient was used to determine the strength, direction and statistical significance of the independent variables that influence the surface soil

moisture along each transect, with particular emphasis on the area adjacent to the gully (i.e. number of burrows, soil characteristics). Emphasis was placed on determining the relationship between the surface soil moisture gradient across the wetland in relation to its relative distance from the gully. Spearman's nonparametric test was preferred over that of Pearson's correlation coefficient, so as to diminish assumptions regarding the frequencies of the variables (Ebdon, 1977).

3.2 Vegetation Analyses

Environmental factors that may have been impacted on by gully erosion, such as soil moisture and soil type, were examined in the context of their impact on vegetation distribution. Other factors that determine the location and/or abundance of vegetation communities, such as slope gradient and various human-induced physical changes, were also examined. Using the belt transect method (see McLean and Cook, 1968; Causton, 1988), the study aimed to determine the existing patterns of various wetland vegetation communities in order to establish factors responsible for changes to the geographical distribution of the various vegetation communities. The belt transect method was chosen because its width (1 m) allowed for a broad representation of the distinct vegetation types found within the valley heads. Transect length was designed to include the representative vegetational patterns that exist across each valley head floor (according to Greig-Smith, 1964). The belt transects were arranged within each valley head so as to traverse perpendicularly through various vegetation communities, at approximately right angles to their boundaries (according to McLean and Cook, 1968). Vegetation transects were also established down relatively undamaged slopes to compare vegetation type, cover and

abundance with other transects that were located across grazing posts and their adjoining grazing areas.

3.2.1 Procedures And Analyses

One vegetation belt transect was placed in each valley head along the same transect lines used for the higher resolution soil moisture recordings (refer to Figures 3.1-3.4). As the plant communities were relatively homogenous, one main belt transect across a typical section of each valley head was representative of the vegetational patterns apparent over the entire valley head floor (see also McLean and Cook, 1968). Each belt transect was 1 m wide, with the soil moisture transect line located on the upslope side of the belt. The belt transect lengths from valley 1 through 5 were: 190 m, 140 m, 175 m, 300 m and 115 m respectively. Additional belt transects were constructed down the slopes of the valley head in November 2002 and in June 2003, so as to examine downslope variations in vegetation patterns. These additional transects were each 65 m in length and were purposely placed through and some distance from a grazing post, so as to capture any differences in the vegetation composition that have possibly been imposed by the inherent activity around a grazing site. These downslope vegetation transects were also used to construct soil transects.

The belt transects were divided into 5 m intervals in order to evaluate gradual changes along it (according to Carter *et al.*, 1988). Geomorphological and vegetational changes along the belt transect and within each grid segment were described and the species density determined within each grid segment (according to Causton, 1988). Where species density was greater than 100 per 1 m^2, a combined determination of abundance

and density was noted (Cain and de Oliveira Castro, 1959; Backéus *et al.*, 1994). Total vegetation cover and maximum height were also measured and/or estimated. A sample of each plant species was collected, pressed and subsequently identified at the University of the Witwatersrand's herbarium.

Plant species have different tolerance coefficients for various environmental gradients. Each ecotone portrays its own distinctive arrangement and composition of species and each species has its own range of tolerance to local environmental conditions (Burrows, 1990). This ecophysiological amplitude is tested in terms of the amplitudinal range of various species. Given that this study is on alpine wetlands, the tolerance of various plant species to changes in the hydrological gradient of the vegetative soil layer was assessed. Tests were also undertaken to determine any relationship between specific plant species and soil pH, and organic matter content.

Spearman's rank correlation coefficient was used to measure the relationship between vegetation communities with soil moisture. The data were converted into ranked form and tested at a 0.05 significance level. The relationship between particular plant species and their proximity to the wetland gully was also tested using Spearman's rank correlation coefficient (King, 1966; Ebdon, 1977).

3.3 Soil Analyses

The aim of this section was to determine how soil conditions differ with distance from the gully. Wetland soil profiles were also analyzed to aid in wetland classification and to

compare the soil type of the gullied wetland with those that have not been gullied. Mineral and wetland soil profiles were examined to establish and compare the developmental history of the valley heads.

Soil samples were collected across the soil transects and from the soil pits, which were either dug into the wetland or obtained from exposed gully sidewalls (see Figures 3.1 – 3.4). All soil samples were placed into sealed plastic bags. All of the soil pits and samples were described using the field methods of Gardiner and Dackombe (1983) and the Field Guide to Forest Ecosystem Classification (1997). Field observations were recorded on a pre-designed recording sheet (Appendix A).

3.3.1 Wetland Soil Sampling

Topsoil samples were collected from the wetland in each valley head along the same transect that was used to measure vegetation patterns. Each sample was measured in the laboratory for pH, conductivity, volumetric water content, % organic matter (section 3.3.5) and hygroscopic moisture content. Additional larger portions of topsoil were collected in each valley head so that particle size distribution and bulk density could be determined. Other samples were also collected from eroded areas and from a gully interfluve in valley head 3. This comparison was conducted to examine if the exposed organic material was physically different from intact organic material.

The gullies in valleys 3 and 4 had incised through the wetland and exposed the adjoining wetland soil profile. Soil horizons were sampled and tested for % organic matter (section 3.3.5) and hygroscopic moisture content (section 3.3.7). The pH of the wetland water and

the degree of humification (Von Post pressing method) of the organic layers were also determined (Denholm and Schut, 1994). The pH value was used to assist in wetland classification and to give an indication of the soil condition in terms of availability of plant-essential elements (Buckman and Brady, 1960; McLean, 1982; Hendershot *et al*., 1993). At each point where a transect intersected a gully, gully profiles were examined. In total, four soil profiles were examined from exposed gully walls in valley 3, and three profiles were examined in valley 4.

3.3.2 Dryland Soil Sampling

Samples from each exposed horizon along the gully sidewalls were collected from each valley head. Soil samples were also collected from alongside, within and below a grazing post in valley head 5 and from a downslope valley transect (refer to Figures 3.1 – 3.4). Bulk density cores were taken from the A-horizons of each mineral soil site. From these soil samples, particle size analysis was performed, organic matter percentage was determined and hygroscopic (Kalra and Maynard, 1991) and volumetric water content measured.

3.3.3 Soil Transects

Along the vegetation belt transects within each valley head, soil samples were collected at 5 m intervals from the top 10 cm of the soil (adapted from Causton, 1988). Sufficient samples (approximately 50 g) were collected to determine % organic matter, hygroscopic moisture, soil pH and electric conductivity (methods described below). This collection method was chosen so as to trace soil changes across the mineral soil of the valley head

floor, across the peaty wetland, and across the wetland gully(ies). The soil properties and possible patterns are then compared to the soil moisture values recorded along the same transects at the same intervals.

3.3.4 Bulk Density

The bulk density sampling technique chosen was the core method (rather than using an in situ technique) (see Culley, 1993). The core method was chosen because it is one of the most commonly used methods to characterize the soil structure; it can also be performed relatively quickly and without much equipment. The bulk density and *not* the mean density was measured to determine the strength, compaction and structure of the *sampled* soil since bulk density usually does not exhibit much spatial variability (according to Culley, 1993). This approach was taken so as to minimise disruption to the wetland areas. The sampling and laboratory procedures were as follows:

- Tin cylinders of approximately 7.25 cm by 10.6 cm were used to collect the samples.
- Each cylinder was weighed in the laboratory and the volume (cm^3) of each cylinder was calculated as:

$$V = \pi d^2 h/4000$$

where d = diameter; h = height (Culley, 1993)

- The top vegetative layer of the soil surface (about 5 cm) was cut back and the cylinder was pressed vertically or horizontally into the levelled soil to fill the cylinder. The cylinder was sealed with aluminium foil, taped, and labelled.
- In the lab the cores were weighed and then placed into an oven at 90˚C. Given that organic matter may decompose above 105˚C (Kalra and Maynard, 1991; Topp, 1993), the samples were set to remain at 90˚C for 70 hours.

- The weight of the cylinders plus dry soil was recorded, and the bulk density (D_b g cm $^{-3}$) calculated:

  D_b = (weight (g) of cylinder plus dried soil – weight of cylinder) / Volume

- Volumetric water content (θ g cm $^{-3}$) was calculated by:

  $\theta = [(w3 - w5)/(w5 - w2)] * D_b$, where

  w3 = weight of cylinder plus moist soil (g)

  w5 = weight of cylinder plus dried soil (g)

  w2 = weight of cylinder (g)

3.3.5 Organic Matter Determination

The dry-ashing method (or loss-on-ignition) was used to determine the degree of both organic and inorganic / ash content. At times, two procedures were used on samples from the same collection so as to test the accuracy of the procedures. The first procedure required the sample to be slowly heated to 375˚C for 16 hours (according to Kalra and Maynard, 1991). The second procedure required the samples to be heated to 375˚C for 1 hour and then be brought to 550˚C for a further16 hours (Karam, 1993).

Both methods required the use of 2 – 5 g of oven-dried soil that had been passed through a 2-mm sieve. Some of the soil used for this procedure came immediately from the oven-dried bulk density experiments, whilst the remaining sediment was first air-dried, then oven-dried and finally ashed. Before weighing the empty crucibles, they were first heated to 105˚C and subsequently cooled (according to Kalra and Maynard, 1991). Approximately 2 g of oven-dried soil was added to the crucibles and then placed into the furnace. The approximate organic matter was calculated as:

$$\% \text{ organic matter} = \frac{\text{oven-dried (g)} - \text{ash (g)}}{\text{oven-dried (g)}} * 100$$

(Kalra and Maynard, 1991).

3.3.6 Soil Reaction (pH)

Soil pH was measured for almost every sample taken from the field. The pH of the soil was required to trace the pH gradient along the wetland, to capture any large changes that may occur adjacent to the gully, to aid in the determination of wetland type, and to assess the condition of the soil in terms of the availability of plant-essential elements (i.e. N, K, Ca, Mg, P, S, Cl, Fe, B, Mn, Zn, Cu, Mo and Ni) (Buckman and Brady, 1960; McLean, 1982; Hendershot *et al*., 1993). A buffer solution of pH 4 and pH 7 were used to calibrate the pH meter. The soil pH was measured in water as this portrays the closest pH to that of the soil solution in the field. $CaCl_2$ was not used, as these soils are not agricultural soils (Hendershot *et al*., 1993).

The pH was measured on air-dried samples that had been passed through a 2-mm sieve. The samples were all either organic or had high organic matter content (> 17%), thus the pH was tested according to the procedure required for organic soils (i.e. 2 g of soil was used in 20 ml of distilled water) (according to Hendershot *et al*., 1993). The soil solution was stirred intermittently for a ½ hour and then the suspension was allowed to settle for a further ½ - 1 hour (Kalra and Maynard, 1991). The probe of the calibrated digital pH meter 766 Calimatic 'knick' was then inserted into the clear supernatant solution, and a reading was recorded once the pH had become constant. The electric conductivity (mV) was also recorded for each pH measured.

3.3.7 Soil Water Content

Although the ratio of the mass of water present in a field-moist sample to the mass of the oven-dried sample (gravimetric) does not reveal field capacity or the water available to plants, it is a good measure for determining differences in soil texture, structure, and drainage across a landscape (Goudie, 1981; Topp, 1993). The volumetric method of determining soil water content was calculated using the bulk density samples taken in March and November 2002 (according to Culley, 1993). The moisture content of the air-dried soil (hygroscopic moisture) was also measured on the soil samples collected along the main transect in each valley head.

3.3.7.1 *Procedure – gravimetric method with oven-drying*

In the field, soil samples were collected and placed directly into sealable plastic bags and labelled (according to Topp, 1993). Once in the laboratory, the moist soil was immediately transferred to weighed and tared aluminium dishes. The samples were oven dried at 90˚C for 72 hours. Samples were then cooled in the desiccator and subsequently weighed to within 0.01 grams. The water content was calculated as a percentage of the dry soil mass by:

$$\text{Water content, \%} = \frac{(\text{moist soil + dish (g)}) - (\text{dry soil + dish (g)})}{\text{dry soil}} * 100$$

(Topp, 1993)

3.3.7.2 *Procedure - hygroscopic moisture (Kalra and Maynard, 1991)*

Approximately 10 g of air-dried soil was passed through a 2-mm sieve. Fractions less than 2 mm were placed onto weighed aluminium dishes and oven-dried at 90˚C for 72 hours. The same procedure as outlined in 3.3.7.1 was followed.

3.3.8 Total Pore Space

Total pore space was calculated from the known bulk density (BD) of the samples and from ash content (according to Carter and Ball, 1993; Parent and Caron, 1993). Total pore space (TPS) was calculated by:

$$PD = (1 + F)/ [(F/1.55) + (1/2.65)], \text{ where}$$

F = % organic matter / % ash

$$TPS = 100(PD – BD) / PD$$

(Parent and Caron, 1993).

3.3.9 Particle Size Distribution

Information on the percentage of each main textural class can be used to predict the hydraulic properties of that soil (Gee and Bauder, 1986). Potential water retention abilities, infiltration rates and productivity of the soil can be inferred once particle-size percentages are known. To assess the texture of the soils sampled, particle-size analysis was performed using the hydrometer method.

The pre-treatment to remove organic matter from the soil samples was first undertaken, as all of the soil samples contained > 5% organic matter.

- 50 g of air-dried soil was passed through a 2-mm mesh sieve so that only the sand, silt, and clay-sized fractions would be included in the experiment. The soil was then added to a 1-litre beaker.
- A separate 10 g of the same sample was air-dried, passed through a 2-mm mesh sieve and oven-dried. The oven-dried weight of the 10 g sample along with its air-dried weight was used to calculate the C_o (oven-dried weight (g/L)) of the air-dried 50 g sample. This was found by cross-multiplication of the known weights.
- Following the pre-treatment procedure outlined in Kalra and Maynard (1991), 50 ml of distilled water was added to the soil in a 1-litre beaker and 50 ml of H_2O_2 (30%) was slowly added. Frothing was observed for 15-20 minutes, sometimes

longer. When the frothing had subsided, the beakers were placed onto a hot plate and heated to about 80˚C. Approximately 10 ml of H_2O_2 was added and again frothing was observed until it had subsided. The addition of H_2O_2 was repeated until frothing had ceased and the supernatant was almost clear.

- Water was then added to the 400 ml mark on the 1-litre beaker. The solution was boiled for 1 hour. Once the soil had cooled and settled, the supernatant liquid was siphoned off using a syringe. 250 ml of distilled water was then added to the sample to re-disperse the soil. Again, the supernatant liquid was siphoned off. This process was repeated a couple of times.
- The method for particle size distribution using the Bouyoucos hydrometer method without pre-treatment was then followed (according to Day, 1965; Gee and Bauder, 1986).

3.3.9.1 *Procedure - Bouyoucos hydrometer method:*

- Distilled water was added to the soil in the 1-litre beaker to produce a volume of 400 ml (according to Day, 1965; Kalra and Maynard, 1991).
- 100 ml of sodium-hexametaphosphate (50 g/litre solution (Calgon)) was added to the beaker and the sample was left to soak overnight (Gee and Bauder, 1986).
- The entire suspension was transferred to a dispersion cup, and was mixed with an electric mixer for 5 minutes. The sample was transferred to a sedimentation cylinder, and room temperature distilled water was added to make 1-litre. Calgon solution in each cylinder is now 5 g/litre. Temperature was recorded once room temperature was reached.
- A control cylinder was prepared using 100 ml of Calgon solution (50 g/litre) and adding distilled water to make 1-litre: Calgon solution was then 5 g/litre. The solution was stirred and the temperature was monitored until room temperature was reached. The Bouyoucos scale (ASTM soil hydrometer 152H) was lowered into the solution and the R_L was read, using the upper edge of the meniscus around the stem.

- A plunger was inserted into the soil solution and moved up and down to mix the contents thoroughly. As soon as mixing had stopped, the time was recorded immediately and the hydrometer was lowered carefully into the solution.
- The hydrometer was read (R) at 30 seconds, 40 seconds, and at one minute. The hydrometer was not removed before the end of one minute to minimize mixing.
- The R_L and temperature of the control cylinder was also read at this time. The control suspension was read after every reading of R.
- Following hydrometer readings were taken at 2, 5, 10, 30, 120, 300, 480, 600, 720 and 1440 minutes (adapted from Day, 1965; Gee and Bauder, 1986; Kalra and Maynard, 1991; Deschamps, 2000).
- The suspension was then passed through a 0.050-mm sieve to capture the sand portion. Once dried, the entire sand fraction was weighed to within 0.1g.

Calculations:

- The concentration of the soil in suspension (C) at each reading was calculated by:

$$C = R - R_L$$

- The summation percentage (P) for each time interval was calculated by:

$$P = C/C_o \times 100,$$

where C_o is the oven-dried weight (g/L) of the soil sample.

- The mean particle diameter (X) in suspension (μm) at time (t) in minutes, was calculated by:

$$X = \theta / (t)^{1/2},$$

where θ is a sedimentation parameter obtained from a table in Day (1965, 564) using R. $(t)^{1/2}$ is the square-root of the time of the reading in minutes.

The θ parameters found in Day (1965) are parameters only applicable at 30˚C, where the viscosity of water (η_{30}) is 0.008007 poise (or 0.8 centipoise). If the temperature recorded during the experiment deviated from 30˚C, then θ must be multiplied by a correction factor of (η/η_{30})½. The viscosity of water with 5 g / litre of Calgon at various

temperatures is given in Gee and Bauder (1986, p394), in Green (1981), Perry and Chilton (1973), and can be calculated as:

$N = -0.025t + 1.51$, where t is any temperature between 15 - 25˚C; the answer is given in centipoises (Schuster, 2003)

Using the hydrometer readings over the time period used, a summation percentage curve (P versus X) was produced. Particle-sizes were plotted as cumulative curves rather than as histograms because cumulative curves are not restricted to displaying the data in terms of the size-class intervals (Gale and Hoare, 1991). From the curves, the percentages of sand, silt, and clay were determined according to the Wentworth scale of classification (Wentworth, 1922). Once particle size percentages were known, the soil textural classes were determined using the USDA classification scheme and textural triangle (from Gee and Bauder, 1986).

3.3.10 Von Post Pressing Technique

Following the guidelines in Table 3.2, the degree of decomposition was assessed for organic layers in the wetland soil profiles (Denholm and Schut, 1994).

3.3.11 Cation Exchange Capacity Estimations

Using percent clay, organic matter and the pH of the substrate, the cation exchange capacity (CEC) was estimated for the topsoil samples. The percentages and pH were placed into the following calculation:

$$CEC = 0.28(clay) + 0.98(OM) + 1.8(pH)$$

(Meyer *et al*., 1994)

Table 3.2 Von Post Pressing Scale (after Denholm and Schut, 1994).

von Post scale of decomposition v - very; mod - moderately		
FIBRIC (Of)	undecomposed	plant structured unaltered; clear water
	almost undecomposed	plant structure distinct; clear water
	v weakly decomposed	plant structure distinct; brown water; no peat passes thrus fingers
	weakly decomposed	turbid water, residue mushy
MESIC (Om)	mod. Decomposed	turbid brown water, some peat escapes
	strongly decomposed	1/3 of peat escapes; strongly mushy
HUMIC (Oh)	strongly decomposed	1/2 peat escapes through fingers
	v strongly decomposed	2/3 peat escapes through fingers
	almost decomposed	nearly all peat escapes through fingers
	completely decomposed	all the peat escapes between fingers

3.4 Morphometric Analysis Of The Wetland Gullies

Gully erosion is indicative that a site-specific landscape is not in equilibrium with its surrounding environment (Bocco, 1991). By taking a morphometric approach to gully analysis (Ebisemiju and Ekiti, 1989), it is possible to gain an insight into its processes. Gully morphology reflects its past and present processes and can also reveal likely future events (Heede, 1976). The plan form of the gully in each valley head was characterized as linear, bulbous, dendritic, trellis, parallel, or compound (according to Bocco, 1991). Gully position within the valley head was classified as either discontinuous or continuous based on defining criteria (see Heede, 1976; Bocco, 1991). Overall gully length (actual and straight line), average depth, channel floor gradient, sinuosity and gully head morphology were measured.

Transect profiles were produced for the valley floor next to the centre line of a wetland gully (in valley heads 3 and 4) and dryland gully (in valley heads 1 and 2). Beginning at

the gully head, slope angle was measured using an abney level every 20 m, or sooner, if there was a break of slope. At each recording point, the gully was entered and the slope of the gully floor and depth of the gully was measured to directly compare to the slope of the valley floor. The transects extended downslope past the lower depositional fans of the gully exit points. The mean slope gradients were converted to radians using the Excel program. For X (horizontal distance), distance travelled in metres was multiplied by the cosine of the radian. For Y (vertical distance), distance travelled in metres was multiplied by the sine of the radian (according to King, 1966). These results were then turned into a cumulative running total and the corresponding graphs were produced (according to Carson and Kirkby, 1972). By comparing the average slope of the valley floor with that of the adjacent slope of the gully channel, gully type (i.e. discontinuous or continuous) was determined. The sinuosity index was calculated by measuring the straight-line distance from the gully head to the gully mouth, and dividing that number against the actual gully length (Ebisemiju and Ekiti, 1989). The relative activity of contemporary gully erosion was established by estimating the percentage of sidewall vegetation at each transect intersection. Sidewall morphology was observed and categorized to help ascertain its possible process mechanisms.

Chapter 4

Gully Erosion

4.1 Introduction

Of all the erosion processes that sculpture the surface of the earth (i.e. all the facets of wind and water erosion), gullying is amongst the most important (Tacconi *et al*., 1982; Billi and Dramis, 2003). This chapter is placed here as it introduces gully erosion mechanics, processes and provides an introduction for Chapters 5 and 6.

Gully erosion is a reaction within the landscape to an instability or exceeded sensitivity caused by an extrinsic (i.e. climate, land use) or intrinsic (i.e. slope aggradation) change in the system (Rowntree, 1988; Allison and Thomas, 1993; Billi and Dramis, 2003). As a response to the imbalance, gullying increases the connectivity of the drainage system and thus increases the drainage density (Leopold *et al*., 1964; Bradford and Piest, 1980; Harvey, 1996; Poesen *et al*., 2003). However, by doing so, drainage basin sediment yields are greatly increased, downstream water quality is reduced, organic matter is oxidized from the topsoil, arable land is decreased and an eyesore is created in the landscape (Rapp, 1975; Bradford and Piest, 1980; Trimble, 1988; Pimentel *et al*., 1995).

Since gully erosion is the direct destruction of an important natural resource (i.e. soil), it has wide and adverse implications for economies (Ebisemiju and Ekiti, 1989; Rienks *et al*., 2000; Billi and Dramis, 2003).

By definition, a gully is a linear incision within unconsolidated material that forms during intense episodes of fluvial erosion (Torri and Borselli, 2003). The formation of an incision or headcut can be initiated and accelerated by the erosion processes that take form as mass failure, seepage, overland flow, rain wash or tunnel erosion (Chakela, 1981; Bettis, 1983; Prosser, 1996). Yet, the exact reasons why and where gullies form, is still poorly understood (Prosser and Winchester, 1996; Billi and Dramis, 2003). Erosion only occurs once the runoff forces exceed soil resistance (Leopold *et al*., 1964).

It is generally agreed that gullies (Rienks *et al*., 2000) have a cross sectional area of at least 929 cm^2 and are impassable by farm machinery (Bradford and Piest, 1980; Campbell, 1989; Bocco, 1991; Vandekerckhove *et al*., 1998; Poesen *et al*., 2003). Gullies are usually lasting features in the landscape that transmit ephemeral flow and characteristically have steep sides, a sharp channel head and, depending on gully type, can end downslope at a mouth (Bradford and Piest, 1980; Eyles, 1980; Ebisemiju and Ekiti, 1989; Prosser and Winchester, 1996). By definition, the channel *head* is the furthest upslope location of the gully that still has well-defined sides (Montgomery and Dietrich, 1988; Bull and Kirkby, 1997), whilst the *mouth* of the gully is at the downstream end where sidewall depth has decreased to zero (Heede, 1976).

All landscapes have key processes that can change in accordance with external/internal forces in order to keep the system in equilibrium. However, when the dominant process changes enough, or when two processes intersect, a threshold is met and the landform responds by becoming qualitatively different (Kirkby, 1980). Sometimes only a small change in certain key factors is required for thresholds to be exceeded and for a major change to occur (Allison and Thomas, 1993). Various key factors within the landscape include, soil, vegetation, valley floor gradient, flow hydraulics, climate and geology (Schumm and Lusby, 1963; Graf, 1979; Vandekerckhove *et al*., 1998; Poesen *et al*., 2003). The threshold that exists for each of these factors, as well as others, varies from region to region. For example, the threshold for vegetation cover varies from as low as 15% in Alberta, Canada to as high as 85% in the Ethiopian highlands (Crouch and Blong, 1989; Evans, 1998).

Depending on the country, approximately 10% – 94% of soil loss is caused by gully erosion (Prosser and Winchester, 1996). Soil erosion generally occurs within the semi-arid Sahel and in the mountain and highland regions at rates 15 times greater than during the average post-Cretaceous (Stocking, 1996). Sheet and gully erosion are often interconnected as gullying results as a mechanism for dealing with the large volumes of runoff and sediment transport derived from sheet erosion (Stocking, 1996). With that in mind, it has been postulated that gully erosion is not to be regarded as indicative of the level or degree of erosion, but instead as one of its forms (Imeson and Kwaad, 1980).

The examination of gully erosion within the Lesotho highlands is important, as there has been little focus on the environmental effects of gullying within Lesotho wetlands. The objective here is to establish the existence of relationships between various geomorphological and biological characteristics and the presence or absence of wetland gullying. These particular valley head wetlands were chosen as they have similar aspects and have developed under similar environmental and anthropogenic conditions, yet do not all have gullied wetland areas. The remainder of this chapter is dedicated to providing a gully erosion literature review and an examination of the Lesotho wetland gully morphologies and gully-forming processes.

4.2 Overgrazing And Thresholds Of Gully Erosion

Gully erosion is certainly a threshold phenomenon; it occurs only once the thresholds of key landscape variables are exceeded (Poesen *et al*., 2003). Although climate change, valley floor gradient and other intrinsic variables are important in the stability of the landscape, extrinsic human actions with regards to their impact on environmental change, is a dominant and driving force for the resulting geomorphic responses (Harvey, 1996; Vanacker *et al*., 2003).

When a landscape is well vegetated, the soil is protected from erosion (Evans, 1980). Protected soils are often porous and well structured, and when not saturated, infiltration is promoted and runoff reduced (Rasiah *et al*., 1992; Guerra, 1994). Soils that are well vegetated thus have a greater surface roughness (> 5-10 mm), undergo less surface crusting, have a higher rainfall storage capacity and are better able to resist the forces of

runoff, sheet flow and sediment detachment, than soils that have a lower vegetation cover (Evans, 1980; Calles and Stålnacke, 2000).

Land use change, in the form of cultivation or from excessive livestock grazing and trampling, can alter landscape heterogeneity and induce structural deterioration of the soil (Imeson *et al*., 1982; Metzger *et al*., 2005). A population explosion of African ice rats (*Otomys sloggetti*) also adds to the grazing pressures in the valley heads (Lynch and Watson, 1992; Hinze, pers. comm 2006). African ice rats, which occur at altitudes above 2000 m a.s.l., graze in a similar manner to that of sheep by tearing out the entire plant/root (Lynch and Watson, 1992). Selective grazing directly affects the cover density and composition of vegetation by decreasing the aboveground biomass through the direct destruction and/or deterioration of the original vegetative cover (Bragg and Tallis, 2001; Metzger *et al*., 2005). In so doing, the soil becomes exposed to the elements and runoff, and becomes susceptible to 'poaching' (i.e. penetration of the soil surface by hooves [Tasker, 1980; Burgess *et al*., 2000]) (Davies, 1985; Rowntree, 1988; Brooks and Stoneman, 1997). *O. sloggetti* also contribute to the pock-marking of the wetland with its extensive burrow systems within the wetland, and consequently the excavated soil then becomes colonized by dwarf sedges (Lynch and Watson, 1992; Richter *et al*., 1997). This deterioration of the vegetative cover promotes the encroachment of invasive woody species, which can establish themselves and thrive in areas where other species are less able to do so (Milton, 1995a; Moleele, 1998; Kakonge, 2002; Visser *et al*., 2004).

Once vegetation has been degraded, the landscape system becomes sensitive to climate changes and anthropogenic impacts (Poesen *et al*., 2003). When the vegetation threshold is exceeded, the soil has a reduced ability to resist rainfall and overland flow erosion forces (Graf, 1979; Vandaele *et al*., 1996). The exposure of the soil thus leads to impaired soil quality (Snyman and du Preez, 2005). Livestock trampling can now further induce structural deterioration and topsoil compaction, resulting in decreased field capacity, percent organic matter, rainfall interception and soil aggregate stability (Guerra, 1994; Morgan and Mngomezulu, 2003; Strunk, 2003; Snyman and du Preez, 2005). The mechanical resistance, aeration and temperature of the soil are ultimately affected as a result (Oztas and Fayetorbay, 2003).

Overgrazing increases the abundance of unpalatable species to livestock and increases bare soil patches, soil detachability, crust formation and the alkalinity/salinity of the soil (Guerra, 1994; Kakonge, 2002; Visser *et al*., 2004; Snyman and du Preez, 2005). Heavy grazing and trampling thus impairs the soil (in relation to its threshold). Trampling results in lost organic matter, degraded aggregate stability and diminished infiltration capacity. The compaction also causes soil pores to be clogged and encourages crust formation, thus increasing runoff that converges in channels downhill (Guerra, 1994; Harvey, 1996; Vandekerckhove *et al*., 1998; Liu *et al*., 2003; Oztas and Fayetorbay, 2003; Strunk, 2003).

Once overland flow begins and converges, the turbulence and velocity of the overland flow increases as it moves downhill (Poesen *et al*., 2003). When passing over a chance

convexity, velocity further accelerates and incision occurs due to soil property thresholds being exceeded (Evans, 1980, 1996, 1998; Guerra, 1994). Even scars produced from sheep or cattle hooves can form a knickpoint that is capable of initiating a gully head (Evans, 1998). The ability for flow to entrain sediment is directly related to the annual precipitation and runoff amounts, once vegetation cover has decreased (Schumm, 1981). The combined potential power of rainfall and overland flow is expressed conceptually as:

$$\text{Erosion} = f\ (\text{erosivity, erodibility}),$$

(Bocco, 1991, p392).

Several publications have suggested that the production of large sediment yields from overland flow is dependent on slope length and shape (Evans, 1980; Lal, 1982; Montgomery and Dietrich, 1988; Bryan and Poesen, 1989; Liggitt and Fincham, 1989). Although gully erosion is still thought of as a threshold phenomenon in these reports, it was found that gully erosion was most severe on gradients between 5–16°. It is suggested that gully erosion does not increase with increasing slope in an indefinite fashion, but that above and below these gradients/threshold values, erosion should generally decrease (Evans, 1980; Liggitt and Fincham, 1989). Relative erosion rates and sediment loads are found to increase rapidly with steepness over convex slopes, increase steadily with length on uniform slopes, decrease towards the bottom on complex slopes (with concave lower portion) and have low peak runoff volumes on concave slopes (Meyer and Kramer, 1973; Harvey, 1996).

Over 34% of global soil degradation and 49% of the soil loss in Africa has been attributed to overgrazing (Evans, 1998). Uncontrolled grazing has been suggested as the

cause for the deterioration and loss of vegetative cover in Russia (Belyaev *et al*., 2004), China (Mahaney and Zhang, 1991), Australia (Friedel *et al*., 2003) and Lesotho (Backéus & Grab, 1995; Schwabe, 1995; Morris & Grab, 1997; Grab & Nüsser, 2001; Kakonge, 2002). In addition, the potential impact of burrowing animals in alpine areas should not be underestimated, as such activity can be substantial (Rapp, 1975; Hall *et al*., 1999). Burrowing animals directly impact the landscape with their burrowing, loading and compaction, digging and trampling. Burrowing provides transportable sediment, alters infiltration rates and disrupts vegetation patterns and abundance (Neave and Abrahams, 2001; Hall and Lamont, 2003). In the alpine areas of Lesotho, African ice rats should be included in any discussion dealing with the effects of grazing animals on the landscape. For example, in a 1 m^2 area, 12 burrow entrances were counted and 60% of the vegetation was destroyed, whilst in other areas within the study region, the impact of the burrowing and sediment removal resulted in the total elimination of vegetation (Grab and Morris, 1999; Hall *et al*., 1999).

The concept of the geomorphic threshold, first developed by Schumm and Hadley (1957), is broadly defined as "one that is inherent in the manner of landform change; it is a threshold that is developed within the geomorphic system by changes in the system itself through time" (Schumm, 1981, p301). With regards to gully erosion, the geomorphic threshold states that for any stable landscape there exists a critical inherent gradient above which the slope angle of the drainage area becomes great enough to produce sufficient runoff for gully incision (Schumm and Hadley, 1957; Schumm, 1963, 1979, 1981; Patton and Schumm, 1975; Montgomery and Dietrich, 1994; Morgan and

Mngomezulu, 2003; Poesen *et al*., 2003). The threshold of the slope angle can be exceeded by changes to key variables such as increased precipitation, increased slope caused by transportation and subsequent deposition, decreased vegetation cover, trampling and compaction. Often, this relationship between thresholds and gully erosion is used to predict areas prone to gullying (Vandekerckhove *et al*., 1998).

The geomorphic threshold suggests that there must be an inverse relationship as well as an inherent threshold between the drainage basin area and the valley slope gradient (Patton and Schumm, 1975; Montgomery and Dietrich, 1994; Morgan and Mngomezulu, 2003; Vandekerckhove *et al*., 2003), so as one decreases in size or angle, the other increases (Tacconi *et al*., 1982). Thus, as the drainage area becomes smaller due to gully erosion expanding the drainage density, slope length becomes smaller and the gradient becomes steeper (Leopold *et al*., 1964; Schumm, 1981; Tacconi *et al*., 1982; Prosser and Winchester, 1996; Vandekerckhove *et al*., 1998; Talling and Sowter, 1999; Martínez-Casasnovas *et al*., 2003; Poesen *et al*., 2003).

Theoretically, gullies should begin at the first position within the drainage area where there is sufficient area and hence runoff to support development. From this, it can then be expected that there would also be a strong inverse correlation between the source area and the gradient above the channel head (Montgomery and Dietrich, 1988; Belyaev *et al*., 2004). In recent models, this relationship has been used to predict the location of headcut initiation (Whitlow, 1994b; Grissinger, 1996; Prosser and Winchester, 1996; Vandekerckhove *et al*., 1998; Belyaev *et al*., 2004). However, this theory is difficult to

apply to older gullies since it assumes that the present gully head has remained at the location of initial incision, when in fact it is unusual for a gully to migrate solely downslope (Nir and Klein, 1974; Rowntree, 1991; Bennett, 1999; Moeyersons, 2003). Instead, this correlation is often used to predict the extent of gully upslope migration. Patton and Schumm (1975) used a less biased approach to measure the upslope gradient and drainage basin area from a locally oversteepened section of the valley floor. Within this and other studies, it was concluded that the discontinuous gullies at these localities had begun at these locally oversteepened reaches of the valley floor (Schumm and Hadley, 1957; Heede, 1976; Patton and Schumm, 1975; Schumm, 1979, 1981; Whitlow, 1992).

Although gully incision requires a minimum runoff value, which is determined by drainage area and is a function of gradient, it must be remembered that other variables play an equally important role in gully incision. Such variables include land-use, vegetation cover, geology, rainfall intensity and soil properties. All of the variables must have their thresholds exceeded by the forces of erosion before gully erosion is initiated (Vandekerckhove *et al*., 1998; Torri and Borselli, 2003).

4.3 The Mechanics Of Gully Erosion

4.3.1 Processes of Gully Incision

The process of headcut migration is complex and involves hydraulic forces (Grissinger, 1996). A knickpoint within the valley floor can develop into a scour hole once overland

flow has cut through the surface seal of the soil. Flow velocity and its erosive power increases as the water flows over and down the knickpoint (Leopold *et al*., 1964; Billi and Dramis, 2003). The tumbling effect of the flow undercuts the vertical drop and the scour hole enlarges and migrates upslope by virtue of collapse and transport (Figure 4.1) (Bennett, 1999). When the back wall of the headcut is vertical, a plunge pool develops and becomes deeper than the gully floor (Leopold *et al*., 1964). Depending on soil stratigraphy, flow capacity and soil characteristics, the gully head lengthens as it migrates upslope by either horizontal retreat (Bocco, 1991) or by vertical incision (Bettis, 1983).

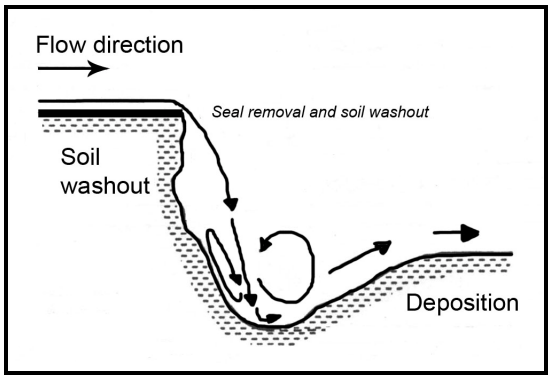


Figure 4.1 The mechanics of creating a plunge pool (after Bennett, 1999, p278).

4.3.2 The Headcut

The morphology of the headcut is diagnostic of the erosional processes and soil characteristics that impact it, but it can also indicate the evolutionary stage of the gully. At least four gully head morphologies exist and have been referred to as *gradual*, *transitional*, *abrupt* and *rill-abrupt* (Figure 4.2) (Oostwoud Wijdenes *et al*., 1999). When a headcut has a vertical wall and is shaped by plunge pool erosion or mass failure, as in the abrupt and rill-abrupt head types, the headcut grows by horizontal retreat in a self-propagating manner. On the other hand, when erosion above the headcut is able to exceed

erosion downstream of the cut, the headcut grows by vertical incision and becomes a rotating or self-degrading formation (Stein *et al.*, 1993; Oostwoud Wijdenes *et al.*, 1999). For gradual or transitional gully head types, the dominant erosion is fluvial rather than mass movement (Oostwoud Wijdenes *et al.*, 1999).

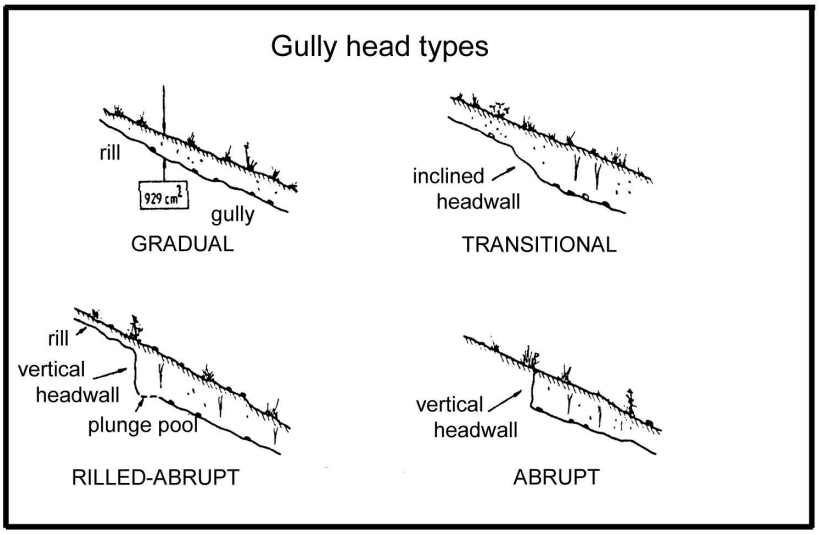


Figure 4.2 Characterizations of various head cut formations (after Oostwoud Wijdenes *et al.*, 1999, p589).

4.3.3 Sidewall Erosion

After gully head retreat, vertical walls of soil are left exposed. Increased pore pressure from subsurface water flowing towards the gully, together with abrasion from water along the base of the gully walls, causes slumping and undercutting of the gully sidewalls (Bettis, 1983; Martínez-Casasnovas *et al.*, 2004). When the basal flow is strong enough, the loosened sediment gets transported downstream and the gully widens. Sidewall erosion is a description of gully wall failure incurred from undercutting, soil dispersion by rainsplash, surface flow, subsurface flow, fluting, frost and/or wind action. Studies have shown that sidewall erosion can contribute greater sediment yields than that

produced from linear incision (i.e. headcutting), and thus contributes more significantly towards the overall gully plan (Blong *et al*., 1982; Bocco, 1991; Whitlow, 1992; Whitlow, 1994b; Martínez-Casasnovas *et al*., 2003). Blong *et al*., (1982) found that 30% of the sediment yield from just one gully sidewall was produced from overland and through-flow alone. These active sidewall processes are generally restricted to the most active area of the gully (Blong *et al*., 1982; Whitlow, 1994a). There are thus a variety of sidewall forms with distance from the gully head; the morphology being shaped by age and position of the area within the gully plan form and by side-wall soil properties (Figure 4.3) (Crouch and Blong, 1989; Crouch, 1992). To locate areas that most urgently require appropriate conservation measures, a scheme of gully sidewall classification was designed by Crouch (1992) and incorporates four levels of description. The sidewall can be described according to 1) degree of activity, 2) by its morphology and 3) by the dominant erosional processes (Table 5.1). A fourth category has been included to account for the description of special features (Crouch, 1992).

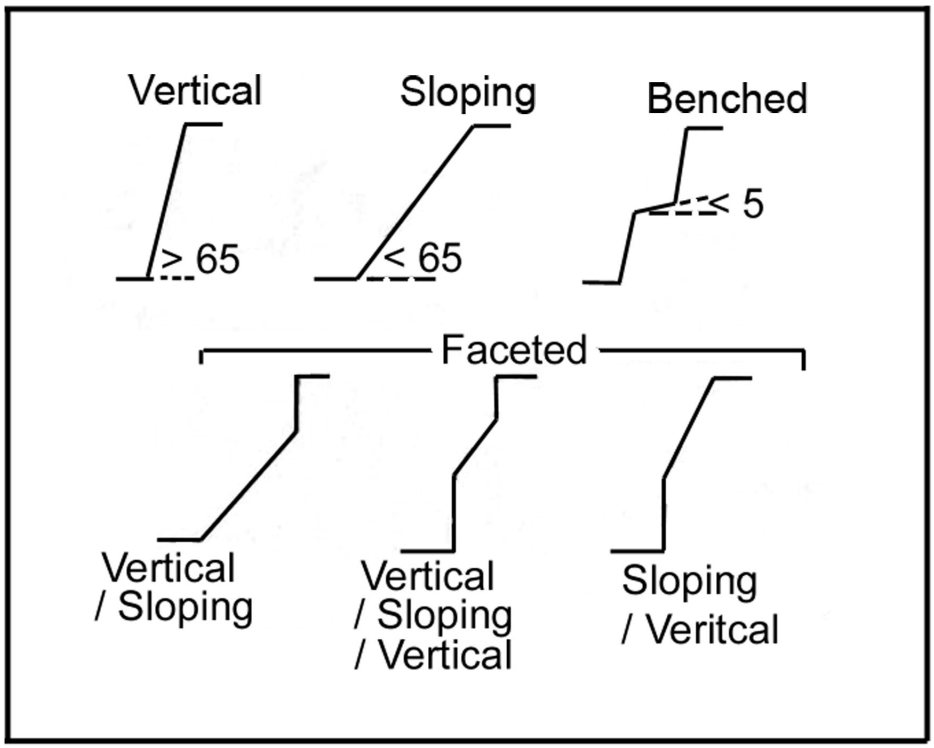


Figure 4.3 Gully sidewall profiles; numbers are angles in degrees (after Crouch and Blong, 1989, p293).

Table 4.1 Gully Sidewall Classification (adapted from Crouch and Blong, 1989, p293-294).

Erosional activity	active; < 20% cover semi-active; 20-70% cover stable; > 70% cover
Morphology * see Figure 4.3	vertical sloping benched faceted
Dominant processes	fluted; crenelation wall failure; slip/topple/fall seepage; diffuse/concentrated overfalls; overhang/undermine/scour/cave

4.4 Gully Types

A gully can be classified and categorized either by the erosion process responsible for its initiation, its sidewall morphology, position in the landscape and/or by its plan form

(Bradford and Piest, 1980; Campbell, 1989). To illustrate this, gullies have been described as being discontinuous, ephemeral or continuous, U-shaped or V-shaped, valley-floor, valley-side or valley head gullies, linear, bulbous, dendritic, etc, or by a combination of these terms (Leopold and Miller, 1956; Bocco, 1991; Rowntree, 1991; Grissinger, 1996). Gullies can erode by upstream extension into an existing channel (i.e. entrenchment; Rowntree, 1991), or initiate a channel where no drainage paths had previously existed (Bocco, 1991; Prosser and Winchester, 1996). The former example usually applies to valley floor gullies, whose U-shaped morphology generally forms through the processes of channel entrenchment, undercutting and overall instability (Imeson and Kwaad, 1980; Grissinger, 1996). The latter example is more applicable to valley side gullies, which generally form on a substrate conducive to concentrated overland flow with a V-shaped morphology (Rowntree, 1988, 1991; Prosser, 1996). By detecting certain patterns in the morphological characteristics and initiating processes of various gully types, Imeson and Kwaad (1980) have isolated four gully forms (Table 4.2). Type 1 gullies develop from rills or whenever overland flow becomes concentrated (i.e. ephemeral); these gullies often have a V-shaped cross-section and do not form on the valley floor. Type 2 gullies develop from the upstream migration of plunge pools and often acquire a deep U-shape with steep walls. Type 3 is similar to type 2 except that this type is associated more with subsurface water and piping, and can often lead to the formation of badlands. Type 4 gullies develop within the alluvial sediments of the valley floor by headcutting, are U-shaped and are often characteristic of the discontinuous gullies that Schumm (1979) identified and has connected to the concept of geomorphic thresholds. Since the morphology of the gully cross-section strongly depends on the

erosion process (Bocco, 1991; Whitlow, 1994a), there is no steadfast rule as to what shape a gully should have in accordance to its position in the landscape. In addition, but also adhering to the four main types of gullies identified by Imeson and Kwaad (1980) and Imeson *et al*. (1982), are ephemeral, continuous and discontinuous gullies.

Table 4.2 Characteristics of particular gully types (adapted from Imeson and Kwaad, 1980, p432).

Gully type	Gully cross-section	Position in landscape	Source of runoff
Type 1	V-shaped	Anywhere, but valley bottom	Overland flow
Type 2	U-shaped	Anywhere, but valley bottom	Overland flow with subsurface water
Type 3	U-shaped	Anywhere, usually on lower slopes	Subsurface flow
Type 4	U-shaped	Valley bottoms	Overland flow from tributary gullies

4.4.1 Ephemeral Gullies

Ephemeral gullies are linear erosion features (Grissinger, 1996), with a minimum cross-section of 930 cm^2 (Vandekerckhove *et al*., 1998). They are temporary forms that may reoccur annually, but are easily removed from the landscape by machinery; thus the term 'ephemeral' refers to the permanence of their structure and not to their flow (Bull and Kirkby, 1997). Ephemeral gullies are formed by the processes of overland flow and begin as microrills, which can enlarge into gullies downstream (Oostwoud Wijdenes *et al*., 1999). Although they can be continuous or discontinuous in nature and U-shaped or V-shaped in morphology, they are always topographically controlled (Grissinger, 1996; Vandaele *et al*., 1996; Bull and Kirkby, 1997; Vandekerckhove *et al*., 1998; Nachtergaele and Poesen, 1999; Poesen *et al*., 2003).

4.4.2 Continuous and Discontinuous Gullies

The concept of continuous and discontinuous gullies (developed by Leopold and Miller, 1956) is based on their dissimilar evolutions. Continuous gullies are initiated where ground cover has become depleted, provided that all other factors are conducive to erosion (Heede, 1976). After this initial incision, the main drainage lines become gullied by headward elongation and tributaries are added later. Usually, the gullies begin as finger-like rills in headwater areas. Continuous gullies therefore characteristically form a network pattern that has a central V-shaped gully whose depth increases in a downstream direction, ending in an abrupt gully mouth (Imeson and Kwaad, 1980). Discontinuous gullies, on the other hand, can begin at any position in the landscape and occur singly or as a series of chains one after the other (Heede, 1976). They characteristically have a vertical headcut and a bed gradient less than that of the valley floor, so that when and where the gully bed intersects with the valley floor, a small depositional fan is produced (Leopold *et al*., 1964; Blong, 1970; Heede, 1976; Schumm, 1981; de Oliveira, 1989; Billi and Dramis, 2003). As the discontinuous gullies migrate upstream, it is possible for them to coalesce (Bocco, 1991). This gully fusion represents the early stages of the transformation of a discontinuous gully into a continuous gully. The wetland gullies in valleys 3 and 4 in the study area, fit the description of discontinuous gully systems.

The above morphologies are not necessarily true to all situations (Ebisemiju and Ekiti, 1989; Whitlow, 1994b). For example, the typical morphology produced by mass failure and pipe collapse is a U-shape (Blong, 1970; Imeson and Kwaad, 1980), however pipe collapse in the Eastern Cape Province of South Africa is also known to have produced V-shaped discontinuous gullies (Beckedahl and Dardis, 1988). In addition, a study

conducted in Colorado U.S.A. describes discontinuous gullies with floors that do not intersect with that of the valley (Mosley, 1972).

4.5 Gully Erosion In Lesotho

There are some areas where the chances of natural restabilization of bare, exposed soil are extremely rare. Some of these areas include places that have peat soils, duplex soils, soils on steep slopes or are in areas that have harsh climates (Rapp, 1975; Trimble, 1988). Notably, Lesotho has all of the above-mentioned landscape components. It is thus not surprising that Lesotho is considered as one of the most eroded countries in the world (Faber and Imeson, 1982; Dregne, 1990; Rydgren, 1996).

In terms of both size and area, the extent of degradation across the border region between Lesotho and South Africa is visible from satellite images, with Lesotho clearly depicting the most extensive degradation (Rydgren, 1988). Land use patterns, usually associated with degradation caused by poor landuse and increasing populations, appear to be the main variable that differs between South Africa and Lesotho (Rydgren, 1988; Showers, 1989; Strömquist, 1990). It has been suggested that extrinsic factors are primarily responsible for the initiation of the gully erosion, whereas intrinsic changes or responses activate further development and growth of the gullies (Watson and Ramokgopa, 1997).

Due to the increasing loss of arable land in the lowlands, a system of transhumance began in the 1880's (Grab and Morris, 1999). Severe signs of erosion in the lowlands were recorded around the time of the arrival of missionaries (Showers, 1989, 1996), and in the

alpine belt deep channels were reported within the wetlands by at least the mid 1970's (Rapp, 1975). The ongoing degradation and soil erosion has been linked to continuous anthropozoogenic impacts (Grab and Morris, 1999; Grab and Deschamps, 2004) (Plate 4.1). The alpine grasslands are used for pasture and are subjected to frequent winter burning, trampling, needle ice, deflation, burrowing and runoff (Grab and Morris, 1999; Nüsser and Grab, 2002). The alpine areas that appear to be most affected by erosion are the drainage lines, rather than the valley slopes, as has also been observed in the Pennines (Bower, 1962; Schwabe, 1995). As the slopes become denuded and the vegetation threshold (70%) required on a mountain range is exceeded (Skovlin, 1984), overland flow and sheet wash progress downslope and deposit large volumes of water and sediment onto the valley floor wetlands (Schwabe and Whyte, 1993; Schwabe, 1995). The wetlands below the north-facing slopes, which tend to have a gentle gradient, appear to be more susceptible to gullying (Nüsser and Grab, 2002).

4.6 Gully Morphology And Principal Processes In The Sani Valley Region

Valley heads 3, 4 and 5 exhibit alpine mires that have been affected by gully erosion. The morphologies of these wetland gullies were examined to document their present state as well as to describe contemporary processes.



Plate 4.1 Valley heads are actively grazed and gullies are degraded partly by anthropozoogenic impacts (November 2002).

4.6.1 Valley 3

The wetland area within valley head 3 had been cut into by a valley-bottom gully system. The gully system was palmate dendritic in form, with dryland plant species and large areas of dry, exposed peat dominating the periphery of the gully system, as well as across the concave gully interfluves. Several ice-rat burrows (+20) were observed within these areas. The valley floor outside of the gully system (to the west and east of it) consisted of wetland plant species on a 'wet' peaty soil.

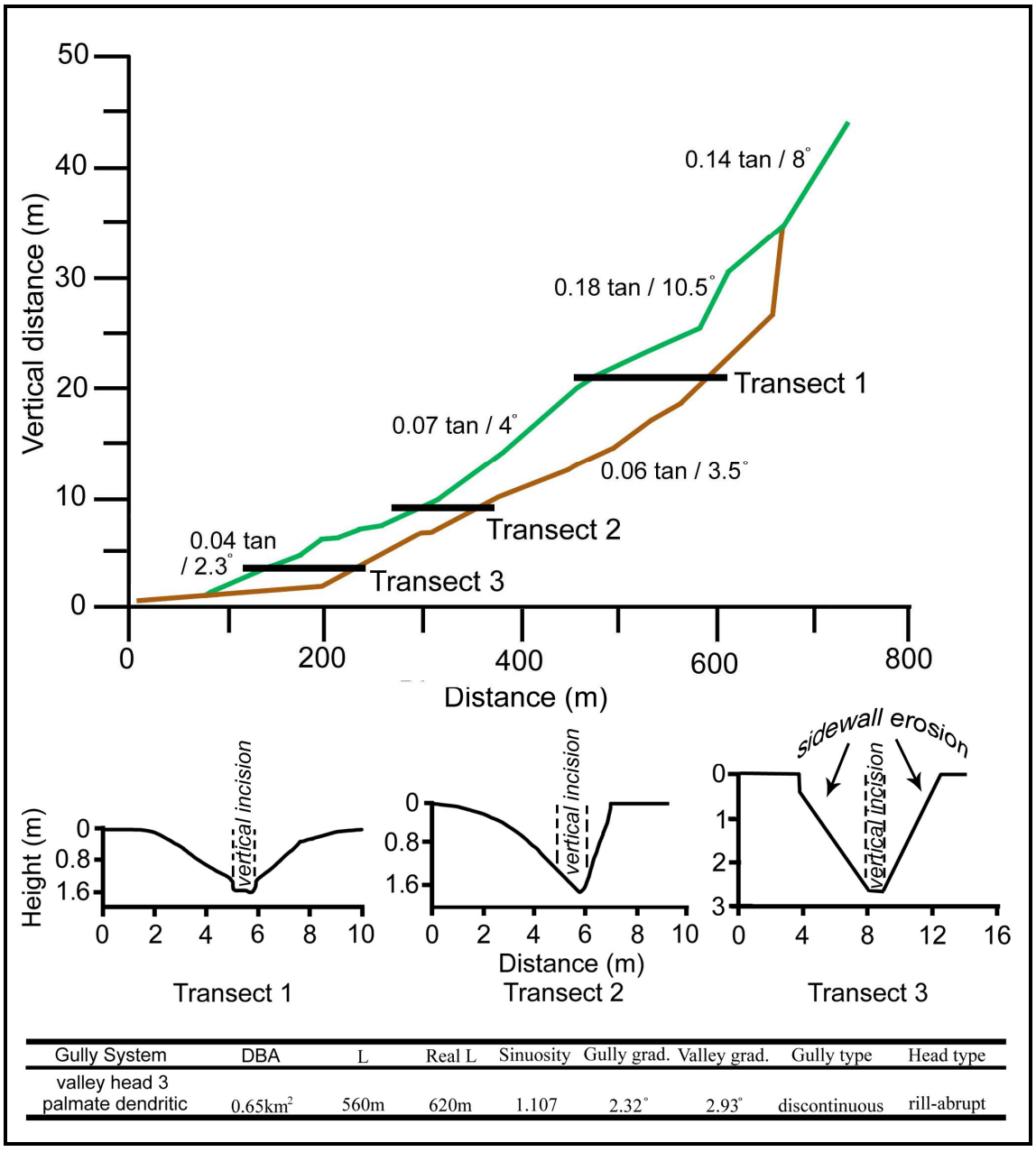
The headcut of the gully system's main branch was rill-abrupt and bedrock controlled, although without a plunge pool. The headcut was 4.1 m wide and 2.7 m deep (Plate 4.2). The gradient of the valley head floor immediately above the gully head was 8˚, whilst the gradient of the channel immediately below the head was 4˚ (Figure 4.4). The average gradient of the gully channel was 2.3˚ (tan 0.04) whilst the adjacent valley floor was

steeper, with an average gradient of 2.9˚ (tan 0.05). Gully depth then became shallower downstream, intersecting the valley floor, at which point the sediment is deposited to produce a fan (S29˚36'22.1" E29˚15'56.1" [2925 m a.s.l]). The actual length of the gully was 620 m, with sinuosity being 1.11. Gully width was generally uniform, with the deepest and widest point located at its mid-section. A meandering drainage line over a slope of 0.5˚- 3˚ transfers water from the alluvial fan downstream to the main river (a tributary of the Manguang River). The drainage basin area above the gully system was approximately 0.350 km^2.



Plate 4.2 Rill-abrupt headcut in valley head 3 (June 2003).

Cross-sectional profiles of the gully at three points along its length revealed similar results. The highest cross-section has exposed a dryland soil profile, while the other two cross-sections show exposed wetland soil (Figure 4.4). All profiles have an approximate V-shape, active sidewalls (i.e. no vegetation) and are sloping and steep ($^{+}$30˚). From



Gully System	DBA	L	Real L	Sinuosity	Gully grad.	Valley grad.	Gully type	Head type
valley head 3 palmate dendritic	0.65km^2	560m	620m	1.107	2.32°	2.93°	discontinuous	rill-abrupt

Figure 4.4 Longitudinal profiles of valley head 3; where green is the valley slope and brown is the gully slope. The cross-sections are shown from each transect intersection and the vertical incision versus sidewall erosion is highlighted.

highest to lowest, the morphologies of the cross-sections reveal that they were most affected by the processes of wall failure, undercutting and soil fall. The dominance of wall failure rather than vertical incision within this gully was evident from the cross-sectional area. The quantity of soil eroded outside of the reach of the actual watercourse,

as well as the tension cracks, support this contention (c.f. Martínez-Casasnovas *et al*., 2004) (Figure 4.4). The cross-sections show the discontinuous nature of the gully as cross-sectional depth decreases with elevation.

4.6.2 Valley 4

The drainage system in valley head 4 was generally trellis shaped, with all headcuts occurring at similar approximate elevations. Gullies within this valley head may exist entirely within the dryland or wetland, or in some instances have developed through both hydrological regimes. A central gully has cut through the wetland centre and extended upslope into the dryland portions of the valley side. The wetland area has conceivably covered the whole of the valley floor but is, at the time of writing, divided into four sections by the gully system and its draining effects. These dry areas in the centre of the wetland are slightly convex or raised, especially towards the gully terminus where sediment would have been more recently deposited.

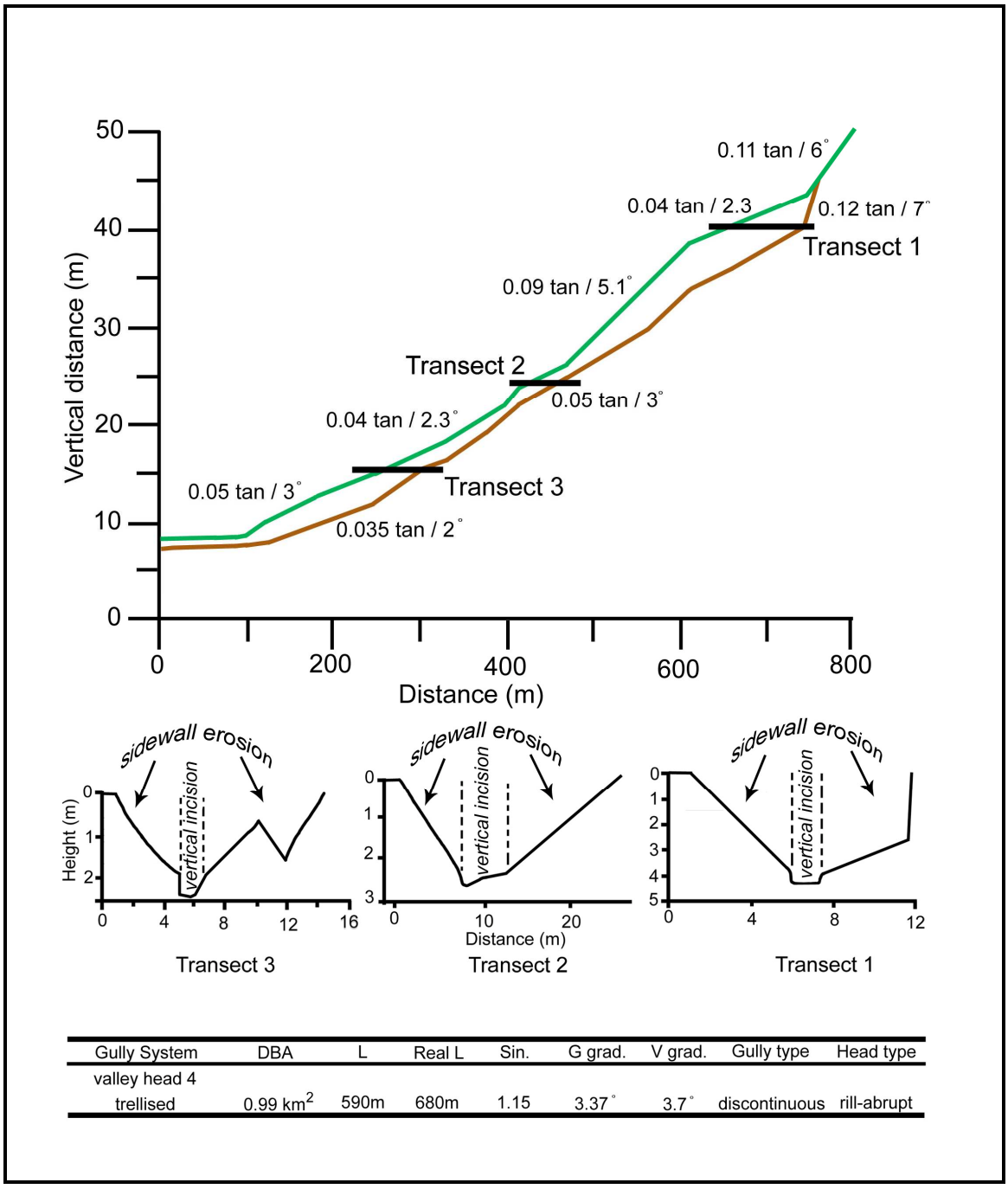
The headcut of the central gully was rill-abrupt, 5.3 m wide and 2.98 m deep and was located at S29˚ 36’26” E29˚15’27.1” (Plate 4.3). The gradient of the valley floor immediately above the headcut was 6˚, whilst immediately below the headcut the gully channel had a gradient of 2.5˚. The average gradient of the gully channel floor (3.37˚ [0.06 tan]) was less than that of the adjacent average valley floor gradient of 3.7˚ (0.07 tan) (Figure 4.5). Gully depth had thus decreased gradually downslope, eventually meeting the valley floor (S29˚36’0.94” E29˚15’32.3”), at which point a narrow shallow stream acutely meandered to the main river. The overall gully plan shape was widest in



Plate 4.3 Rill-abrupt headcut of the main gully in valley head 4 (June 2003).

the upper middle section (± 25 m), but tapered towards the head and terminus. Actual gully length was 680 m and sinuosity is 1.15. The drainage area above the gully system was approximately 0.998 km^2 (Figure 4.5).

Three cross-sectional gully profiles were examined at various distances along the main gully (Figure 4.5). Channel depth was recorded as decreasing in depth with decreasing elevation, although channel width was variable along the course. The channel profiles had varying morphologies with some walls being more concave than others, although most were somewhat V-shaped and sloping. All of the sidewalls appeared to be active as denoted by the absence of vegetation; the cross-sectional area outside of the reach of the watercourse establishes that sidewall erosion is a dominant force.



Gully System	DBA	L	Real L	Sin.	G grad.	V grad.	Gully type	Head type
valley head 4 trellised	0.99 km^2	590m	680m	1.15	3.37°	3.7°	discontinuous	rill-abrupt

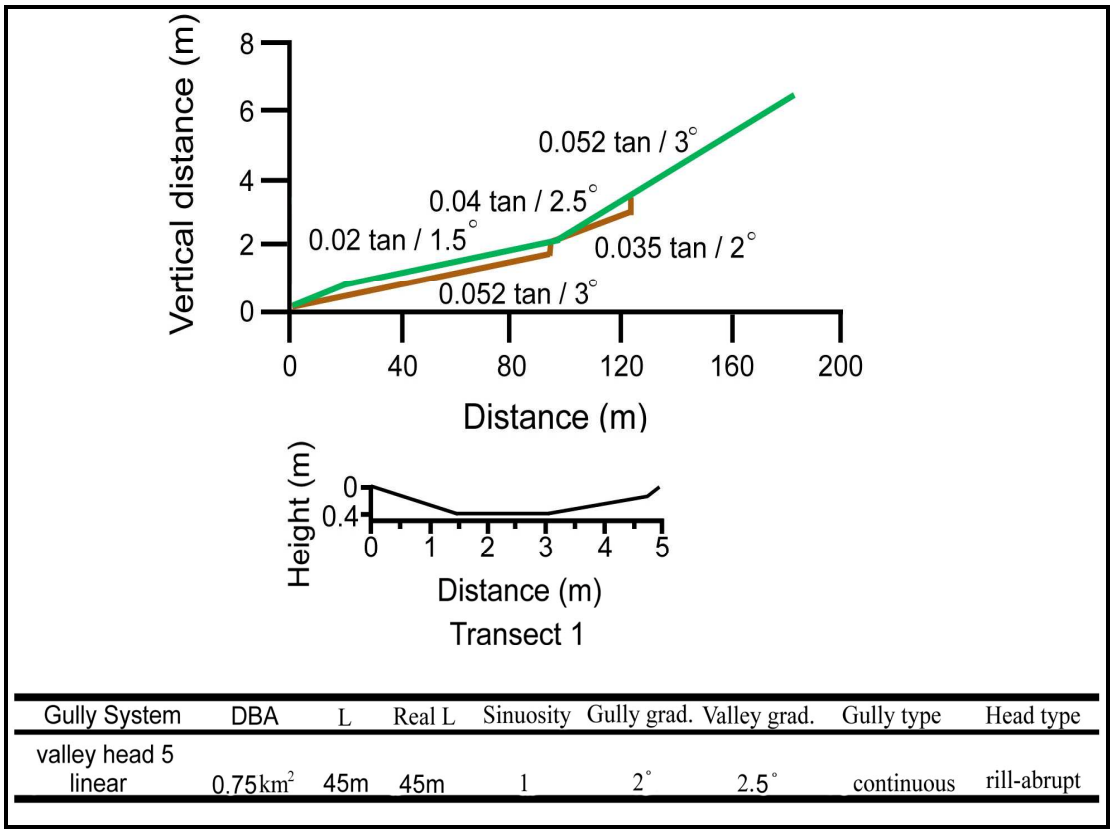
Figure 4.5 Longitudinal profiles of valley head 4; where green is the valley slope and brown is the gully slope. The cross-sections are shown from each transect intersection and the vertical incision versus sidewall erosion are highlighted.

4.6.3 Valley 5

A small developing gully was actively cutting into the wetland area within valley head 5. In June 2003, water was seen flowing down the centre of the wetland towards this drainage system. The system consisted of a rill connected to a small gully. One large wetland rill (30 - 50 cm deep) has occupied the lower half of the valley floor since at least February 2001. This rill was checkmark-shaped with an average channel gradient of 2˚. The adjacent valley floor had an average gradient of 2.5˚. Above the rill headcut (S29°35'06.5" E029°15'59.4" [2900 m a.s.l.]), which had a maximum depth of 30 cm, the average gradient of the valley floor was 3˚. The U-shaped gully that was connected to the rill had a maximum depth of 50 cm and a downstream gradient of 3˚ for 33 m, after which the gradient decreased to 0.5˚ for a further 41 m (Figure 4.6). The average gradient of the adjacent valley floor was 1˚, whilst downslope of the gully terminus, the average gradient of the valley floor was 2˚.

4.6.4 Discussion

The wetland gullies studied in valley heads 3 and 4 both have intermediate morphologies, with the cross-sectional profiles of valley head 4 being the more variable of the two. Both gully systems had rill-abrupt gully heads, with weakly developed plunge pools. The presence of rill-abrupt gully heads suggests that the gullies have migrated upstream following an existing drainage path. This is shown by the connection of drainage lines across the wetland surfaces of valley heads 3, 4 and 5. The two non-gullied wetlands (valley heads 1 and 2) do not host such drainage lines. The processes most often noted in upstream extending valley floor gullies are entrenchment, undercutting and



Gully System	DBA	L	Real L	Sinuosity	Gully grad.	Valley grad.	Gully type	Head type
valley head 5 linear	0.75km^2	45m	45m	1	2°	2.5°	continuous	rill-abrupt

Figure 4.6 Longitudinal profiles of valley head 5; where green is the valley slope and brown is the gully slope. The cross-sections are shown from the transect intersections.

consequent general mass instability. Such upstream gullies also usually have a deep U-shape morphology (as with Type 2 gullies). The wetland gullies of valley heads 3 and 4 are wide and moderately deep, with morphologies tending towards a U-shape. However, some V-shaped sections were observed in valley head 3. The absence of plunge-pools beneath the gully heads indicates that the heads are migrating more by vertical incision rather than horizontal retreat and that the erosion has been occurring more quickly above the head than below it. These gullies are actively eroding and are contemporary features within the valley heads. The steepness of the walls, the presence of running water through the gullies, and the almost complete lack of vegetation on the sidewalls suggest contemporary gully development. Sidewall erosion was more prevalent within each

cross-sectional profile than vertical incision, with the most obvious erosional process being soil fall (Figures 4.4 and 4.5).

In all cases, the slope gradient above the gully heads was higher than the gradient of the channel floor immediately below the gully head. This pattern continues for the length of both gullies, where the average channel gradient was less than that for the adjacent valley floor gradient. The fact that the channel floor intersects the valley floor at the gully terminus and has an associated downslope depositional fan, indicates that the gully was of a discontinuous nature. Slope gradients of around tan 0.06 and 0.03, representing 'oversteepened slopes', has been reported by Patton and Schumm (1975). They further suggested that these slope values were great enough to result in gully initiation. The tan values at the bases and headcuts of the Sani gullies are all much higher (i.e. 0.05 – 0.11) than the valley floor gradients measured by Patton and Schumm (1975). Following Patton and Schumm's theory, the most downstream portions of the gullies in valley heads 3 and 4 would have been initiated where the gully terminus occurs at present. Given the gentle slope gradients of the slope in close proximity to the gully terminus, it is possible that the gullies have migrated upslope from this location (c.f Evans, 1980; Liggitt and Fincham, 1989). However, other locations of gully initiation upstream, and therefore areas where they may have coalesced, are difficult to identify. In addition, no pattern could be detected as to whether the upstream reaches/gully heads were related to an inverse relationship with the drainage basin area above the head (i.e. had either begun or had ceased migrating at a critical upslope position, as discussed by Montgomery and Dietrich (1988) (see Table 4.3). This could imply that either the drainage area was too small (> 1 km^2) for this relationship to exist, as was also found by Patton and Schumm

(1975) for their threshold theory, or alternatively, that these gullies had not yet reached their final upslope location given that the upslope gradient was often slightly greater than the gradient directly above the head.

Table 4.3 Gully (G) and valley (V) gradients within each valley head.

	Valley 1	Valley 2	Valley 3	Valley 4	Valley 5
Area (km^2)	0.400	0.340	0.650	0.998	0.750
G. gradient above head	14	8.5	8	6	3
G. gradient of head	10	5	4	2.5	2
V. gradient above head	14	9	8	7	3

Chapter 5

Valley Head Soils

5.1 Introduction

This chapter is designed to briefly introduce the soils of Lesotho, the general characteristics of wetland and mineral soils, and the affects that human-induced changes may have on them. The remainder of the chapter focuses on the results of the soil transects at the study site. Soil characteristics (i.e. pH, percent organic matter and soil moisture) are examined across the mire expanse, with particular focus on the area adjacent to the gullies (in valley heads 3, 4 and in 5). By comparing the data sets collected within the gullied valley heads with those of non-gullied valley heads, it is possible to determine differences in soil characteristics, should they occur. Soil profiles in valley heads 3 and 4 are examined and their depositional history is described and compared to previous profiles produced for the same valley heads by Marker and Whittington (1971), Hanvey and Marker (1994) and Marker (1994, 1998).

5.2 Classification Of Lesotho Soils

Soil is a product of various soil forming factors and reflects the varying influence of each of these factors (Schmitz and Rooyani, 1987). Although each soil type consists of solids, liquids and gases, including flora and fauna, every type is influenced differently by a combination of soil forming factors; namely parent material, climate, vegetation, topography, time and biological activity (Buckman and Brady, 1960; Brady and Weil, 2002). The complex and diverse soils found in Lesotho today have been influenced by a mixture of the six major soil forming factors mentioned above, which consequently impacts on soil forming processes (i.e. weathering, leaching, translocation, gleying, bio-mixing, deposition, erosion and organic matter accumulation) (Pearsall, 1950; Buckman and Brady, 1960; Brady and Weil, 2002).

5.2.1 Dryland / Mineral Soil

The majority of Lesotho soils are mineral soils that have organic matter ranging from less than 1% to ca. 10% (Schmitz and Rooyani, 1987). The source of the soil mineral component is derived from sedimentary rocks in the lowlands and igneous rocks in the mountains (Rydgren, 1990). The organic matter component is formed from the abundant grasses (Schmitz and Rooyani, 1987). Of the eleven soil orders that exist globally (using USDA Soil Taxonomy), six occur in Lesotho, of which, three are predominantly zonal soils (i.e. the Mollisols, Alfisols and Vertisols) (Table 5.1) (Van der Merwe *et al*., 2002).

Table 5.1 Description of the main mineral soil orders of Lesotho; Histosols are the sixth soil order occurring in Lesotho (after Nixon, 1973; Schmitz and Rooyani, 1987; Rydgren 1990).

	Description
Mollisol	Black soil rich in organic matter under grass vegetation. Has a mollic epipedon.
Alfisol	A loamy sand soil with an eluviated surface horizon and an argillic endopedon.
Vertisol	Clay soil with deep cracks and a vertic epipedon. 30% clay in all horizons above 100cm.
Entisol	Weakly developed horizons with ochric epipedon.
Inceptisol	A young soil, but older than Entisol. With a cambic horizon.

5.2.1.1 *Montane Soils*

The most influential soil forming factors in the lowland region are topography, parent material and palaeoclimatic conditions. The montane soils are characterized by a ustic moisture regime (limited moisture), a thermic temperature regime (MAAT: 15 - 22°C) and usually have less than 2% organic matter content (Chakela *et al*., 1986). Since the Montane soils are derived from sedimentary rock (i.e. the Molteno, Elliot and Clarens Formations), they are dominant in minerals such as mica, potassium feldspar and illite, and are generally sandy, acidic soils that can either be light or dark in colour (Lundén *et al*., 1990).

The dominant soil orders found in the lowland and foothill regions of Lesotho are the Alfisols, Entisols and Inceptisols (in the USDA Soil Taxonomy) (Table 5.2 for listing of series). Alfisols, which are similar to Luvisols (in the FAO soil map of the world), cover approximately 3000 km^2 of Lesotho (Schmitz and Rooyani, 1987) and are characterized

Table 5.2 General list of some soil series found in Lesotho and their description (modified from Nixon, 1973 and Schmitz and Rooyani, 1987).

Soil Order	Soil Series	Characteristics
Alfisol	Leribe/Qalo/Rama	Of Elliot Formation
	Machache	On basalt
	Maseru/Sephula	Duplex
Entisol	Caledon	Very young, alluvium
	Majara	Buried Mollisol; acidic
	Thoteng	Young, on Clarens
	Ntsi	Young, on Clarens
Inceptisol	Berea/Qalaheng	Young, on Clarens
	Matela	Young, on Clarens and basalt
Mollisol	Moroke/Popa	On steep slopes, shallow soil
	Fusi	Mountain valleys, overthickened epipedon
	Nkau/Seforong	Very deep soil on basalt. Reddish brown loam surface
	Soloja/Thamathu	Clay illuviation, on basalt
	Sofonia/Kolonyama	On basalt, alluvial, deep soil, stratified
	Sani	Aquic, clay loam epipedon, moderately deep
	Matšana	Moderately deep loam soil on basalt saprolite
Vertisol	Thabana	Clayey soil on basalt saprolite, poor permeability
	Phechela	Clayey on basalt colluvium in bottom land areas

by having an ochric epipedon (light coloured) overlying an argillic (clay) or natric (clay and sodium) endopedon (Driessen and Decker, 2001). The Alfisols are highly erodible and are often cut into by U-shaped gullies (Schmitz and Rooyani, 1987). Two major groups of Alfisols are found in Lesotho, which include at least six soil series. One of these groups includes the lowland ‘red soils’. This soil type has an uncharacteristic low base saturation and was formed from the weathering of the Elliot Formation Sandstones. The second major group of Alfisols includes the duplex soils, which have an aquic moisture regime and best reflect the processes of eluviation and illuviation seen in the abrupt textural and structural change between the A and B horizons (Schmitz and

Rooyani, 1987; Rienks *et al*., 2000). The Entisols and Inceptisols are described in section 5.2.1.2.

5.2.1.2 *Subalpine Soils*

The soils that have formed within the subalpine region of Lesotho are derived from basaltic colluvium and alluvium. These soils are generally shallow, brown to red brown in colour, and have a udic moisture regime (moist) and a mesic temperature regime (MAAT: 8 – 15°C). Inceptisols, Entisols and Mollisols predominate over the foothills, whilst Vertisols typically occupy depressions in the landscape (Table 5.2; Figure 5.1) (Rydgren, 1990; Van der Merwe *et al*., 2002). Inceptisols and Entisols (USDA), similar to Brunisols and Regosols (in FAO) respectively, are recent but poorly developed soils mainly occupying steep slopes. Entisols are the lesser developed of the two and occupy ca. 1280 km^2 in Lesotho, whilst Inceptisols are more acidic and wet, occupying ca. 1850 km^2 (Ntokoane and Nthebe, 1998). Vertisols are basic, clayey soils ($30\%^+$ clay within the top 100 cm) with 2:1 smectite clays, and are recognized by their shrinking, swelling and cracking characteristics (Juma, 1998; Ntokoane and Nthebe, 1998; Driessen and Deckers, 2001). Vertisols normally develop under grassland vegetation, are high in exchangeable bases (Ca and Mg) and have a near neutral pH (Strahler, 1975). Mollisols, which are similar to Chernozems (CSSC and FAO), are more strongly influenced by the tall grassland vegetation in which they are found and are therefore one of the most fertile soil types in the world (Strahler, 1975; Driessen and Deckers, 2001). By definition, Mollisols have a mollic epipedon that is at least 20 cm thick with 1% to 18% organic matter and have less than 30% clay (Driessen and Deckers, 2001) (Figure 5.2). The mollic epipedon

must have a dry colour value darker than 5.5 and a moist chroma less than 3.5 (Strahler, 1975; Schmitz and Rooyani, 1987). The thick, organic rich surface horizon that is most noticeable in Mollisols is a result of the grassland vegetation, whilst the characteristic high base status (> 50%) and dominance of 2:1 clays (i.e. montmorillonite) reflects the qualities of the basaltic parent material (Schmitz and Rooyani, 1987).

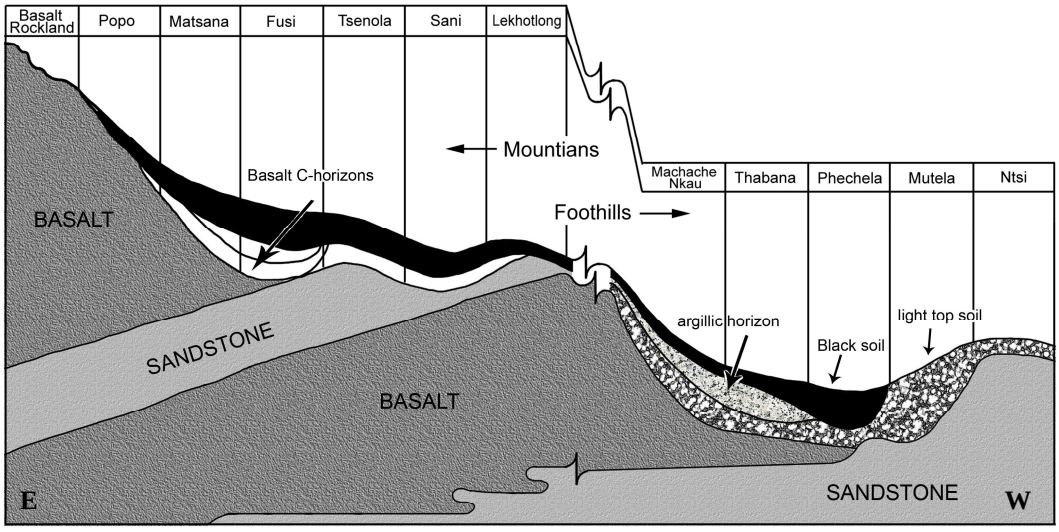


Figure 5.1 Toposequence of the major soil types found in the montane and sub-alpine regions of Lesotho, showing east to west (after Schmitz and Rooyani, 1987, p143).

5.2.1.3 *Alpine soils*

The mountain mineral soils of Lesotho have a higher organic matter content than the montane mineral soils in the lowlands. The relatively high organic matter content (1% to 10%), compared to that in the lowlands (< 2%), is typical of most alpine soils (Leopold *et al*., 1964) and is attributable to the soil forming factors of time, topography and climate (Young, 1976). Basalt weathering of the Lesotho Formation, which has a mineralogy consisting of plagioclase, sodium/calcium aluminosilicate, iron, magnesium silicate and

olivine, releases products such as silica, calcium, magnesium, iron and aluminium into the developing soil (Schmitz and Rooyani, 1987). As altitude increases, the extent of weathering and the development of the B horizon decrease, as does clay content, iron and acidity. However, with an increase in altitude and associated decrease in temperature, the percentage organic matter increases (Munn and Spackman, 1990; Miller and Birkeland, 1992). Since the weathering and leaching of the basalt saprolite is neither intensive nor rapid in the Drakensberg, the dominant clay mineral of the alpine soils is montmorillonite (Schmitz and Rooyani, 1987). Montmorillonite is a highly cohesive 2:1 clay that has a high net negative charge and is capable of swelling (Table 5.3) (Juma, 1998). The type of clay within these alpine soils, as well as their high amounts of organic matter, is partly responsible for the neutral to basic pH values and high base saturation. Organic matter content and clay percentage are important features of all soils, due to their high negative charges, and thus have the largest capacities to adsorb cations (Buckman and Brady, 1960; Brady and Weil, 2002).

Organic matter within a soil serves many important functions: it helps to promote good soil structure, improves water retention and infiltration, buffers against acidity, helps to regulate the soil microclimate, affects microflora and microfaunal activities and increases the cation exchange capacity, and thus determines the nutritional health of the soil (Brady and Weil, 2002; Snyman and du Preez, 2005). Together with clay, organic matter has the ability to adsorb essential nutrients (cations) such as: N, P, K, Ca, Mg, S and CHO (Brady and Weil, 2002). Organic matter is present within a soil as either humic (dark brown) or fulvic (yellow brown) acids, which create mor, moder or mull (Buckman and Brady, 1960; Brady and Weir, 2002). Mor is a low activity, acidic organic matter type

that usually develops in coniferous areas. Moder (aka *duff*) tends to be a thicker accumulation of organic matter (2 – 3 cm) and is characteristic of mountain grasslands, whilst mull is a biologically active, well-humified, neutral organic matter found in grassland areas (Buckman and Brady, 1960).

Table 5.3 Comparative properties of common silicate clay minerals (adapted from Juma, 1998; Brady and Weil, 2002).

Property	Montmorillonite	Illite	Kaolinite
Size (μm)	0.01-1.0	0.02-2.0	0.5-5.0
Shape	Flakes	Irregular flakes	Hexagonal crystals
External surface area (m^2/g)	70-120	70-100	10-30
Internal surface area (m^2/g)	550-650	-	-
Plasticity	High	Medium	Low
Cohesiveness	High	Medium	Low
Swelling capacity	High	Low to none	Low
Unit-layer charge	0.5-0.9	1.0-1.5	0
Interlayer spacing (nm)	1.0-2.0	1.0	0.7
Bonding	Van der Waal's bonds (weak attractive force)	Potassium ions	Hydrogen
Net negative charge ($cmol_c/kg$)	80-120	15-40	2-5

Mountain soil orders

The dominant soil orders of the Lesotho mountain area include Mollisols, Vertisols and Histosols (Schwabe, 1989; Rydgren, 1996) (Figure 5.2). These basalt-derived soils tend to be dark coloured, well structured, have a neutral pH and are classified as having a udic moisture regime (with the exception of the Histosols) and a cryic temperature regime (Schmitz and Rooyani, 1987; Lundén *et al*., 1990). Soils with an aquic moisture regime are often formed in depressions (Lundén *et al*., 1990).

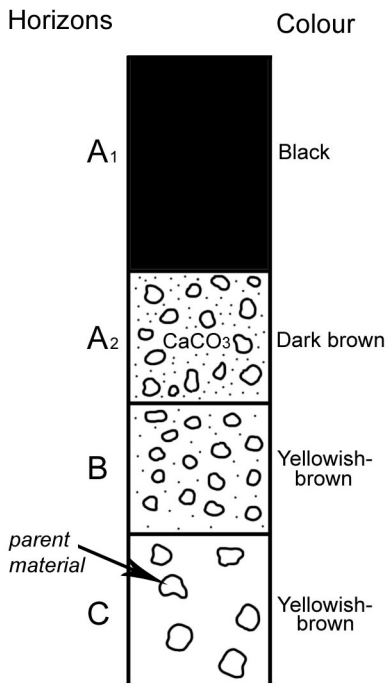


Figure 5.2 Typical profile of a Mollisol with upward weathering of the parent material; $CaCo_3$ is found in the A_2 layer (after Burrows, 1990, p57).

Mollisols are the dominant soil order on mountain slopes and within valleys, and cover approximately 1.5 million hectares in Lesotho (Schmitz and Rooyani, 1987). These Mollisols have a thick (>18 cm), dark humus-rich mollic epipedon that develop under grassland vegetation (Klug *et al*., 1989). The Mollisol soil series that is located near Sani Pass includes the *Matšana*, *Popa*, *Fusi* and *Sani* (Lundén *et al*., 1990) (Table 5.2). The *Matšana* soil series belongs to the subgroup Typic Hapludoll and is a moderately deep well drained gravely soil. The term Hapludoll denotes that the soil is a Mollisol with minimum horizonation in a udic moisture regime (Schmitz and Rooyani, 1987). The *Popa* soil series is classified as a Lithic Hapludoll (shallow soil) and covers approximately 30% of Lesotho. It is usually located on steep slopes and is a shallow, well-drained soil. *Fusi* soils are Cumulic Hapludolls and therefore have overlying thick epipedons, which are high in organic matter and in base saturation. *Fusi* soils are well

drained and are normally found in concavities, whereas *Popa* soils are skeletal soils found in convex areas. Finally, *Sani* soils have an aquic moisture regime and are classified as Typic Natraqualfs (Schmitz and Rooyani, 1987), meaning that they have a natric horizon, have a thin O horizon and mottled subsoil. The main Vertisolic soil series that occurs in the alpine region is *Thabana*, which is a black clayey soil that has 3% to 7% organic matter content in its upper layers. *Thabana* soils typically have a neutral pH, a high cation exchange capacity, poor drainage and are dominated by montmorillonite clay. The Histosols of the mountain areas (which are not mineral soils) include the alpine mires, which have been classified as well-decomposed *Borosaprists* (Schmitz and Rooyani, 1987).

5.2.2 Wetland Soil

The generic term 'wetland' describes a habitat that is characterized by the presence of water, anaerobic soil conditions (in most cases) (Tiner, 1999) and biota (i.e. hydrophytes) that have adaptations to anoxia (Mitsch and Gosselink, 1986; Charman, 2002). Wetlands develop in positions on the landscape where water inflow exceeds rates of outflow and evapotranspiration (i.e. in depressions, floodplains, below ground water discharge, etc.) (Tiner, 1999; Charman, 2002) (Figure 5.3). Thus, the water that is contained within a wetland comes directly from precipitation, overland runoff flow, groundwater sources, rivers, or from a combination of these sources (Haslam, 2003).

Wetlands act as sinks, sources and transformers of nutrients, and thus affect the quality of water (Lewis *et al*., 1995). Some of the functions that wetlands provide are chemical and

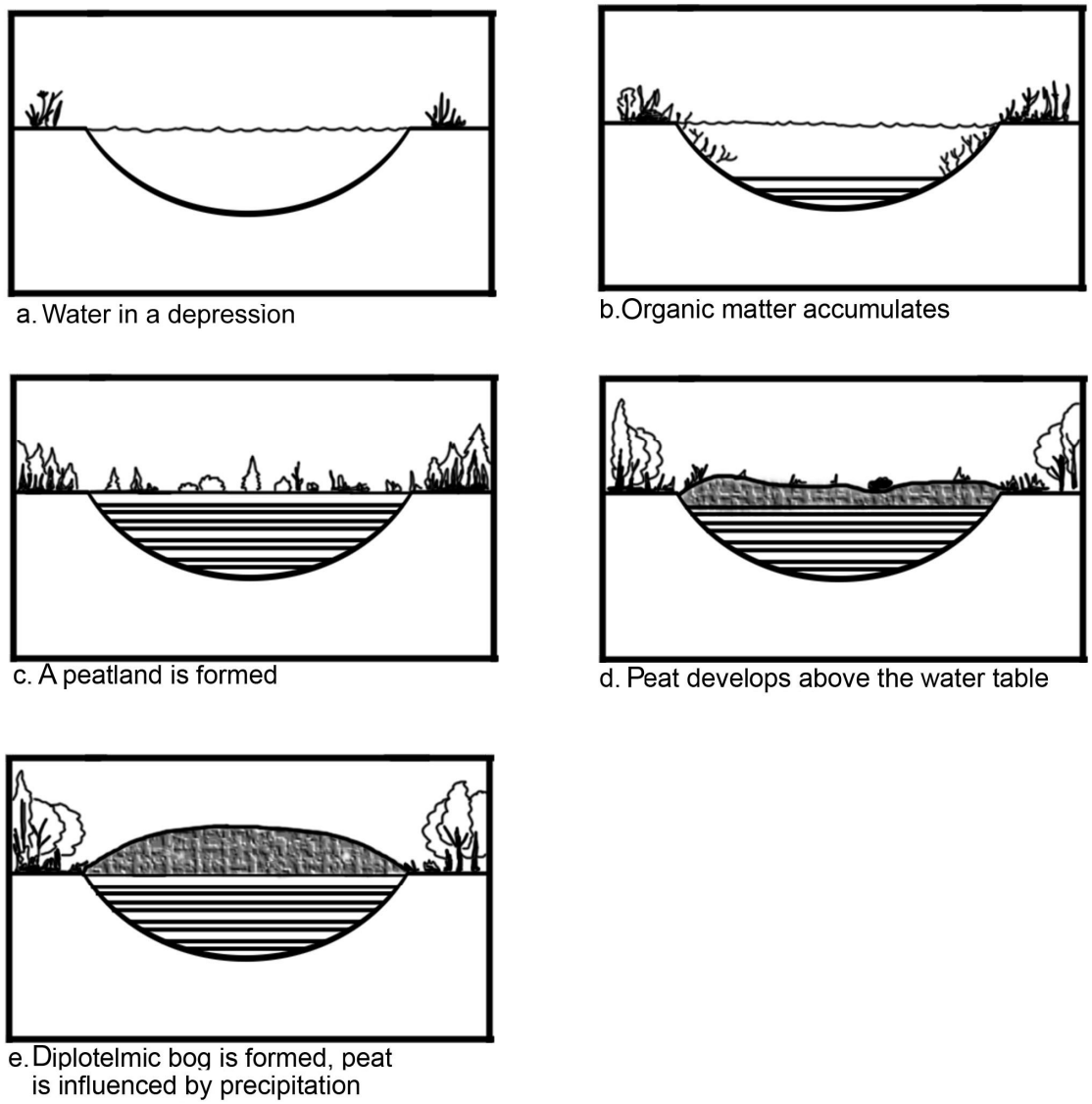


Figure 5.3 Development of a wetland and its eventual transformation into a raised bog (after Brooks and Stoneman, 1997, p31).

biochemical, biological (wetlands are capable of achieving high levels of biodiversity), hydrological and/or physical (i.e. groundwater recharge, soil stabilization, and flood protection). In addition, wetlands have an economic value and are perceived as being intrinsically important (Lewis *et al*., 1995; Haslam, 2003; Nakamura *et al*., 2004). Most wetlands are highly productive and extremely sensitive ecosystems (Mitsch and

Gosselink, 1986; Charman, 2002), which cover ca. 6.4% of the world (Table 5.4) (Mitsch and Gosselink, 1986; Haslam, 2003).

Table 5.4 Estimated global coverage of wetlands (after Mitsch and Gosselink, 1986).

Zone	Total Land Area %
Polar	2.5
Boreal	11.0
Sub-Boreal	13.4
Subtropical	29.3
Tropical	10.9

Hydrology is the key component to any wetland and it is the quantity and quality of water that differentiates one wetland from another (Haslam, 2003). Generally, wetlands in arid environments are typically represented as salt flats, whilst in humid, cool regions they form peatlands. In more temperate, subtropical and tropical areas, wetlands form marshes and swamps. Perhaps it is due to their ability to develop in a variety of climatic regions and topographic settings that there is no single, universally recognized definition of the term 'wetland' (Mitsch and Gosselink, 1986; Tiner, 1999; Haslam, 2003). Of the definitions used to help define the physical boundaries of a wetland, one that was developed by the United States Fish and Wildlife Service and has been widely adopted, states that a wetland is

> " …land where an excess of water is the dominant factor determining the nature of soil development and the types of animals and plant communities living at the soil surface. It spans a continuum of environments where terrestrial and aquatic systems intergrade."
>
> (Cowardin *et al*., 1979, p4).

The Ramsar Convention defined a wetland as being

> "…areas of marsh, fen, peatland or water, whether natural or artificial, permanent or temporary, with water that is static or flowing, fresh, brackish or salt, including areas of marine water the depth of which at low tide does not exceed six metres."
>
> (Hughes, 1996, p267).

According to the South African Water Act, a wetland is

> "land which is transitional between terrestrial and aquatic systems, where the water table is usually at or near the surface or the land is periodically covered with shallow water, and which land in normal circumstances supports or would support vegetation typically adapted to life in saturated soil".
>
> (Dini *et al*., 1998, p4).

Because wetlands are found to exist as an ecotone between two major ecological communities, the entire wetland area is viewed as an environmental gradient. As with many gradients, the official boundary between the wetland and the upland area, or between the wetland and the lower aquatic system (i.e. lake/river/ocean), can be extremely difficult to locate (Mulamoottil *et al*., 1994; Wetlands Pro, 2001). The United States Army Corp of Engineers, who can professionally determine if an area is a jurisdictional wetland, examine along a transect for the required presence of hydrophytic vegetation, a hydric soil and wetland hydrology (Wetlands Pro, 2001). The wetland boundary is found only once all three criteria are present at the same location (Charman, 2002). Some popular wetland types, which meet such requirements, include *swamp*, *marsh*, *bog*, *fen* and *wet meadow* (Mitsch and Gosselink, 1986) (Figure 5.4).

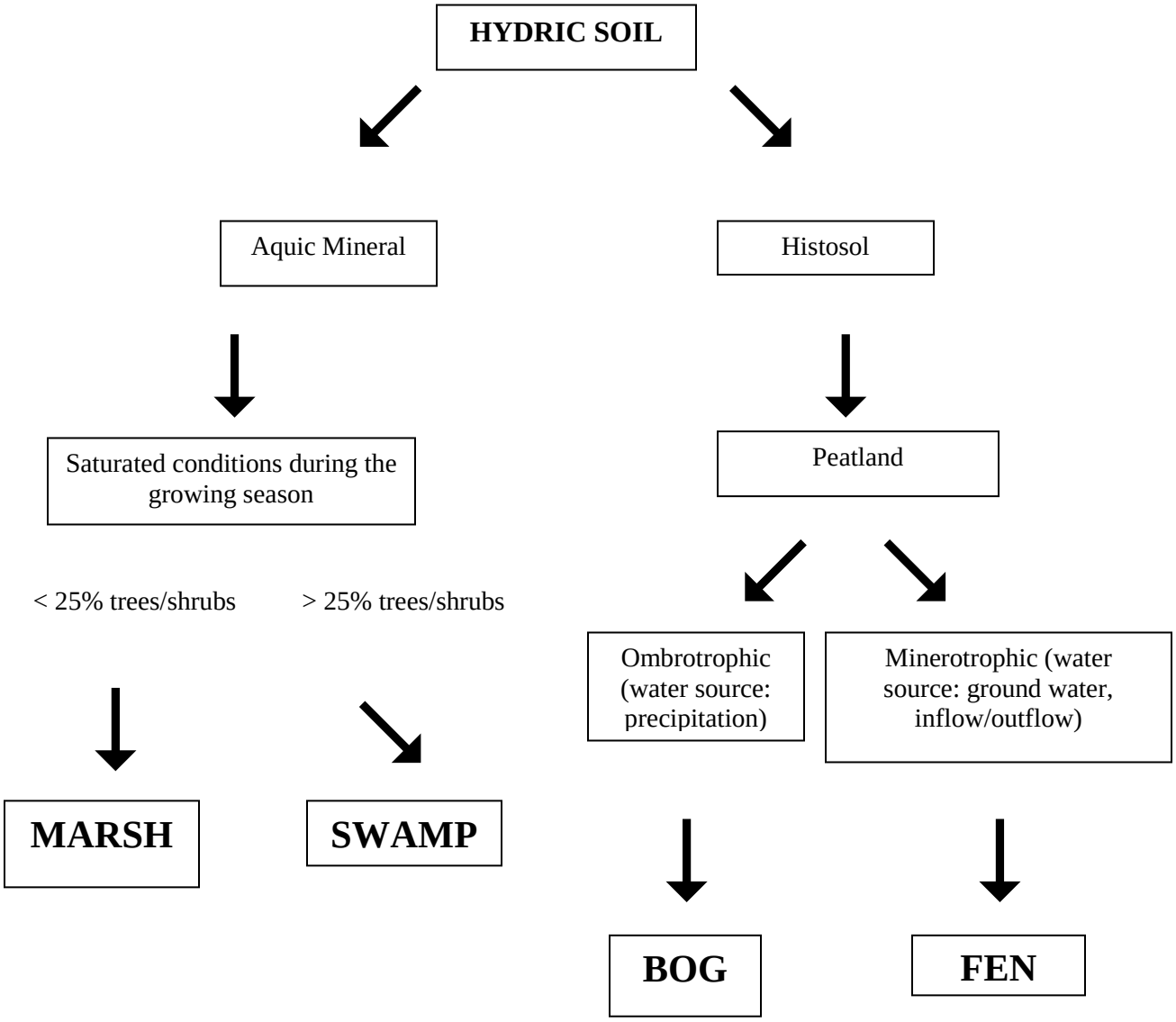


Figure 5.4 Flow chart illustrating the differences between various wetland types.

5.2.2.1 *Hydric soils*

A wetland soil that has anaerobic conditions and supports hydrophytic vegetation is known as a *hydric* soil (Mitsch and Gosselink, 1986). All wetlands accumulate soil (Haslam, 2003), either through the deposition of sediment or as undecomposed, humified plant remains that produce peat. These two main types of hydric soils develop in

conditions of varying wetness, hydrological conditions and climatic regimes. The main similarities between these two broad types of soil are the water content and chemical transformations that occur within them due to depleted oxygen amounts.

Upon saturation, the diffusion of oxygen is stopped and the available oxygen within the soil ultimately becomes consumed by the respiration of aerobic organisms and is completely depleted within a few hours (Mohanty and Dash, 1982; Tiner, 1999). The diffusion of oxygen becomes drastically reduced upon becoming waterlogged because of a decrease in the effective area coupled with an increase in diffusion length (Mohanty and Dash, 1982). The rate of diffusion also depends on whether the water is lentic or lotic; in flowing water the diffusion of oxygen is 10 000 times faster than in standing water (Mohanty and Dash, 1982). The only free oxygen present under these waterlogged circumstances is found within a thin layer of oxidized soil at the soil-water interface, which comes from the surface air as well as from photosynthesizing plants (Sparling, 1966; Tiner, 1999). The diffusion of oxygen within this layer, as well as throughout the entire profile, is not uniform, nor can it move as freely as in an aerobic soil. When the redox potential, which can range from 700 to –300mV in a saturated soil, decreases to at least 350 to 300mV, anaerobic conditions result whereby facultative and obligate anaerobes increase in number (Tiner, 1999). These organisms promote denitrification and reduce otherwise non-toxic forms of iron, manganese, nitrogen and sulphate to toxic forms (Tiner, 1999; Charman, 2002). Therefore, hydrogen sulfide, methane, organic carbon and nitrogen (as ammonium through mineralisation) accumulate in these anaerobic substrates (Shotyk, 1988; Charman, 2002; Holden *et al*., 2004).

5.2.2.2 *Aquic mineral soils*

The wetland soils on the drier end of the wetland gradient are called mineral soils and belong to an aquic suborder and/or subgroup (Lewis *et al*., 1995). This type of hydric soil profile generally consists of mineral horizons. The top surface of a mineral wetland soil is usually not saturated and will contain oxidative compounds of iron and other essential elements, as well as organic matter, which all contribute to the rich brown colour of the layer (Tiner, 1999). The subsurface mineral layers however, are saturated and therefore have undergone reducing physical and chemical behavioural changes, which produce metabolically anaerobic bluish grey coloured horizons known as *gley*. Gleyed soil is a characteristic redoximorphic feature of an aquic mineral soil and is formed by the reduction of iron into iron oxide or ferrous iron (Fe^{2+}) (Lewis *et al*., 1995). With periods of long saturation, the ferrous iron becomes mobile and leaves a grey, gleyed soil. During its movement, the reduced iron may become aerated and therefore oxidized to ferric iron (Fe^{3+}) (Tiner, 1999). The aeration can result from either water level fluctuations or from the presence of oxygen around plant roots; the oxidized iron precipitates as red or orange mottles and is visible within a matrix of reduced, gleyed mineral soil. A typical mineral soil that has developed anaerobic features from being saturated most of the year is the aquic Mollisol (Schmitz and Rooyani, 1987). Aquic Mollisols may develop a histic epipedon over a mollic epipedon and are recognized by their redox features within 20 cm of the surface (Mitsch and Gosselink, 1986). Marshes are also formed on waterlogged mineral soils, but are periodically inundated with standing fresh, salt or brackish water (Burrows, 1990). If the entire wetland were to begin accumulating thick deposits of peat, the wetland would eventually become a peatland (Brooks and Stoneman, 1997).

5.2.2.3 *Mires / peatlands*

The second type of hydric soil is the organic or peaty soil, known as the *Histosol* (Mitsch and Gosselink, 1986; Driessen and Deckers, 2001). Histosols are formed when the soil is permanently or nearly saturated and are distinguished by the presence of thick deposits of peat (Tiner, 1999). Organic soils are different from mineral soils in many respects, including lower bulk density and porosity, higher organic carbon levels, greater conductivity, higher organic matter content and more nutrient availability (Buckman and Brady, 1960; Mitsch and Gosselink, 1986). Since mineral wetland soils also tend to accumulate peat, the general rule used to distinguish between a mineral/aquic hydric soil and a Histosol is that the 'peat' within a mineral hydric soil will be less than 30 cm in depth and will have less than 20% to 35% organic matter (dry weight) (Holden *et al*., 2004) (see Figure 5.5). As a rule, a Histosol will have a soil profile that consists of peat for at least more than half of its upper 80 cm. This peat will contain 18% or more organic matter if there is greater than 60% clay, have at least 12% organic matter if there is no clay, or will have between 12% to 18% organic matter if the clay fraction is less than 60% but still greater than zero (Strahler, 1975; Mitsch and Gosselink, 1986) (Figure 5.5). The rate at which the peat is able to accumulate depends on the rate of decay. The rate of decay is dependent on the water content, the temperature of the soil, the biomass of the vegetation, the pace at which the material takes to fall beneath the water surface, and the microbial populations (Warner, 1996; Charman, 2002). As the hydrophytic plants die within the saturated soil, their remains fall into the water. Once submerged, the rate of decomposition becomes drastically reduced since the anaerobic conditions of the soil inhibit the respiration and decay processes of the decomposing agents (Warner, 1996;

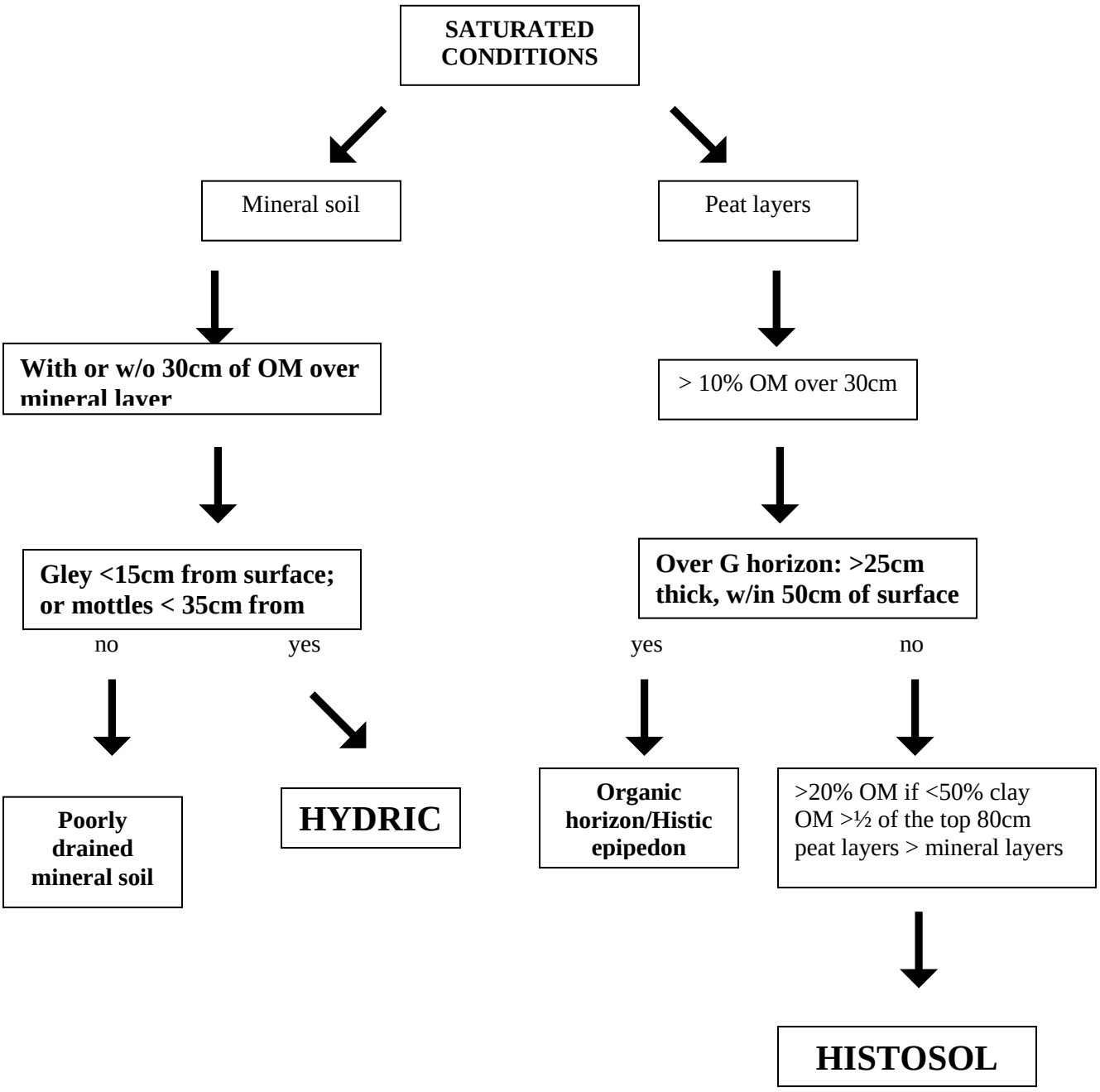


Figure 5.5 Flow chart describing the differences between a hydric soil and a Histosol, where OM = organic matter; w/o = 'without' (information from Mitsch & Gosselink, 1986).

Bozkurt *et al*., 2001; Bragg and Tallis, 2001; Charman, 2002). Mineralisation of certain compounds does occur quickly, however the main material of the plant and animal matter becomes humin, which is the bulk of the peat mass (Charman, 2002). Peat is therefore the accumulation of physically and chemically transformed organic residues in varying stages of decomposition formed in cold and waterlogged environments (Bozkurt *et al*., 2001; Holden *et al*., 2004). Peat is typically light brown to black in colour; although colour changes with the passing of time and increasing decomposition. The relative proportions of carbon, hydrogen and oxygen along with other nutrients within the peat depend on the botanical composition and state of decomposition (Mitsch and Gosselink, 1986; Shotyk, 1988). During decomposition there is a subsequent loss of organic matter, an increase in bulk density and a decrease in hydraulic conductivity (Mitsch and Gosselink, 1986; Bozkurt *et al*., 2001).

According to the relative proximity to the water table, three main peat types exist and all can potentially occur one on top of the other. These are known as primary, secondary and/or tertiary peat types (Charman, 2002), or as limnic, telmatic and terrestric peats, respectively (Moore and Bellamy, 1973). Primary/limnic peats develop in depressions and below the water table and are therefore minerotrophic; secondary/telmatic peats tend to develop around the primary peat in a periodically saturated zone and can therefore be minerotrophic or ombrotrophic, while tertiary/terrestric peats develop above the influence of ground water and hence are humified and oxidized, as well as ombrotrophic in nature (Charman, 2002). Peat can be further classified according to its plant forming communities (over 61 types of peat have been classified this way), by the degree of

decomposition (ranging from fibrous and highly porous to amorphous and highly humified (see Table 3.2, Chapter 3) and/or by its topographic position (Moore and Bellamy, 1973; Pears, 1985; Bozkurt *et al*., 2001).

Areas that have attained at least a peat thickness of 30 cm are known as peatlands and/or mires (Shotyk, 1988). There is estimated to be ca. 42 to 1500 billion tonnes (10^{15}g) of peat globally (Charman, 2002) and therefore a net pool containing ca. 570 x 10^{15}g of fixed carbon (Bozkurt *et al*., 2001; Sjors, 1982; Shotyk, 1988; Williams, 1988); although this estimation is dependent on vegetation type and depth. Currently, peatlands cover approximately 4% (500 million hectares) of the total global land area with the most extensive coverage being within the Boreal forest ecosystems (Sjors, 1982). It is estimated that the initiation of most peatlands began around the time of the Pleistocene deglaciation, ca. 8000 – 12 000 years BP (Marker, 1994; Bozkurt *et al*., 2001). In Africa in particular, two main periods of peat initiation are apparent during phases of increased moist conditions; one during the Late-glacial period and one ca. 5000 – 3500 years BP (Meadows, 2001; Charman, 2002).

The average accumulation rate of peat within peatlands varies considerably from one continent to another. Since peat accumulates during specific environmental conditions (i.e. cool and wet), its existence is a testament to the climatic and vegetational conditions at the time of peat formation. For example, in North America, peat developed at an average rate of 100–200 cm/1000 yrs, whilst in Europe it progressed more slowly, at an average rate of 20–80 cm/1000 yrs (Bozkurt *et al*., 2001). As peat develops, its vertical

gradient becomes composed of alternating layers of organic matter, sands, silts and clays that all reflect the alternation of wet and dry climatic phases, whereby times of organic matter accumulation represent wet phases and times of mineral sedimentation represent drier periods (Moore and Bellamy, 1973; Burrows, 1990). More specifically, in some peatlands, laminations, which are a reflection of the cycle of the seasons, may develop in calm conditions at the mud-water interface. The autochthonous mud layer would represent the organic sedimentation while the mineral layers would represent allochthonous inwash (Goudie, 1981). Preserved peat layers, consisting of the various host plant materials (e.g. fibre, pollen and spore) thus become a historical dialogue of past events. The defining factors used to differentiate one peatland type from another are usually the water source and the source of nutrients (Moore and Bellamy, 1973; Holden *et al*., 2004). A single type of peatland can exist within a specific landform, or various types of peatlands can occur along a horizontal gradient within a landscape. Based on water source and nutrient content respectively, a peatland is typically minerotrophic or eutrophic, ombrotrophic or oligotrophic, or as mesotrophic if it falls between these classifications (Moore and Bellamy, 1973; Pears, 1985; Malmer, 1986).

Fens

When groundwater or mineral soil water from runoff is involved in present peat formation, the peatland is classified as minerotrophic and can be soligenous and/or topogenous. Since fens develop in the landscape where minerotrophic water accumulates, they are classic minerotrophic peatland examples, rich in dissolved minerals (Malmer, 1986; Tiner, 1999; Charman, 2002; Haslam, 2003; Holden *et al*., 2004). Due to the

neutralizing effect of the base-rich minerals upon any acids that enter the water, fens typically have a pH value around 5.5 to 8 (Shotyk, 1988). When a fen has an abundance of dissolved minerals, either from its groundwater source or during the early stages of its development, the fen is given the term, *eutrophic* (Brooks and Stoneman, 1997). Eutrophic fens are floristically rich and have a high base status (Bozkurt *et al*., 2001). However, as the amount of dissolved minerals becomes less, or the water becomes more dilute, the fen becomes *oligotrophic* (Burrows, 1990; Shotyk, 1988). Oligotrophic fens are less basic and so tend to be bog-like (Shotyk, 1988). This type of fen is sometimes described as being a poor-fen due to its chemical status as well as being vegetatively poor, in contrast to eutrophic rich-fens (Moore and Bellamy, 1973; Malmer, 1986). The fen morphology tends to be flat and is normally dominated by bryophytes, sedges and grasses (Haslam, 2003). Fens are commonly distinguished from bogs by the presence of flowing water that is either entering or exiting the mire (Moore and Bellamy, 1973).

Bogs

The major difference between minerotrophic (i.e. fens) and ombrotrophic (i.e. bogs) peatlands is that bogs form in the absence of mineral water (i.e. bogs are currently growing under the sole influence of precipitation) (Moore and Bellamy, 1973; Brooks and Stoneman, 1997; Bragg and Tallis, 2001) (Figure 5.4). The term bog therefore applies to the accumulation of peat above the level of groundwater (Malmer, 1986). Bog peat can either form directly on top of the mineral surface due to the paludification of an area, by the terrestrialisation of a water body or when terrestric peat develops within a fen (Warner, 1996; Brooks and Stoneman, 1997). Since bogs are not influenced by mineral

water and are also non-flowing, there is little input of bases. This depletion of nutrients together with the increased capacity to adsorb and exchange cations, as well as the inability to neutralize acids produced from decomposition (i.e. CO^2 and organic acids), all contribute to the development of characteristically acidic bog waters (Shotyk, 1988; Burrows, 1990; Jonasson and Shaver, 1999). Bogs typically have acidic surface waters with a pH of ca. 4, especially when inhabited by Sphagnum moss (Sparling, 1966; Moore and Bellamy, 1973; Malmer, 1986; Shotyk, 1988). During the progression towards independence from surface water supplies and a dependence on precipitation as the sole provider of nutrients such as potassium, magnesium and calcium, plant communities are forced to adjust to the harsher habitat and as a consequence, bogs tend to be more species-poor than fens (Smith, 1966; Shotyk, 1988). Although there are various types of bogs, all bogs have similar cushion-like vegetation and are formed from water that has been cut-off from geogenous sources (Smith, 1966; Haslam, 2003).

Bogs are often described as being *diplotelmic* because they are essentially composed of two to three separate layers of peat (Malmer, 1986) (see Figure 5.6). The thin layer closest to the surface is known as the euphotic layer. The euphotic layer is the main growing portion of the peat and also has the lowest bulk density within the profile (i.e. 100 – 200 g cm^{-3}). Just below the euphotic layer, within the upper 3-5 cm of the bog, is the *acrotelm*. The acrotelm is chemically active, aerobic and also growing (Haslam, 2003). The acrotelm experiences the fluctuation of the water table and characteristically has a high hydraulic conductivity and consists of well-decomposed porous peat (Bozkurt *et al*., 2001; Bragg and Tallis, 2001). Rapid decomposition, death and growth all occur

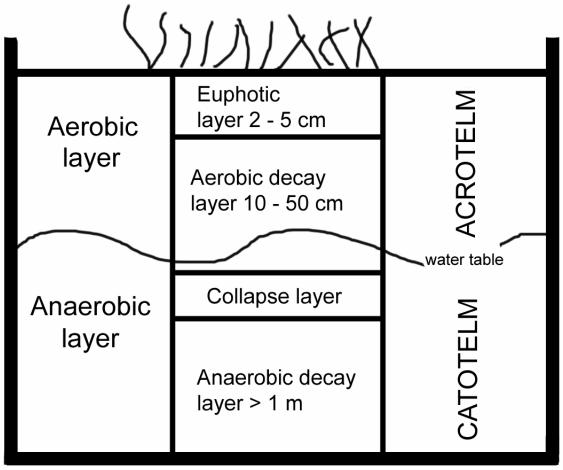


Figure 5.6 Structural layers of a peatland (after Bozkurt *et al*., 2001, p110).

within this layer, with 90% – 97% of the fixed carbon being lost to the acrotelm. The rate of aerobic decomposition is dependent on the moisture content of the peatland, and thus the acrotelm is thickest when the peatland is drier (Bragg and Tallis, 2001). Depending on the rate of decomposition within the acrotelm, the remaining carbon becomes stored in the lower peat layer, known as the *catotelm*. This lowest peat layer is compact, anaerobic, constantly waterlogged (95% -97% water content) and has a high bulk density (>1000 g cm^{-3}), low hydraulic conductivity and is highly humified (Ingram, 1978; Bozkurt *et al*., 2001; Bragg and Tallis, 2001). The location of the production of the various gases produced within the peat profiles is coincident and dependent on these functional peat zones (i.e. acrotelm and catotelm). Within the euphotic functional zone, or in the case of gas production, the aerobic zone, respiration occurs and carbon dioxide is produced. Beneath this lies the euphotic zone within the acrotelm, where maximum carbon dioxide production occurs (Bozkurt *et al*., 2001). Between the euphotic and catotelm layers, at about the level of the water table, is the transition zone where both aerobic and anaerobic conditions occur. Therefore both carbon dioxide and methanogenesis takes place within

this area (Bozkurt *et al*., 2001). Within the permanently anaerobic catotelm, methane and carbon dioxide are produced due to the slow rate of organic matter decay (Charman, 2002).

There are two main types of bogs: blanket bogs and raised bogs. Blanket bogs are a rare occurrence of which the majority (13%) occur in the wet and cool regions of Britain (Brooks and Stoneman, 1997; Charman, 2002). They can form directly on top of the mineral surface due to paludification or can form by terrestrialisation of a water body (Brooks and Stoneman, 1997). Conversely, blanket bogs do not rely on mineral ground water for their existence and so may extend upslope to ca. 35˚ (Mitsch and Gosselink, 1986; Burrows, 1990).

5.3 Sani Top Soil Characteristics

During field observations in June, September and November 2001, January, February and April 2002, and in June 2003, all of the mires examined had 'wet' to 'saturated' top soils. The mires are mostly covered with cropped wetland vegetation and have varying densities of *Otomys sloggetti* (ice rat) burrows (Plate 5.1). It was further observed that during November 2001 and June 2003, the gullies within the mires contained clear, neutral (6.89) pH values for flowing water.

A single transect was laid across a representative cross-section of the valley floor within each valley head, beginning and ending on mineral soil, and crossing the vegetation boundaries perpendicular to the slope (following Meadows and Dewey, 1986). Detailed



Plate 5.1. *Otomys sloggetti* (Ice Rat) outside of a burrow (September 2001).

soil moisture along one transect within each valley head was recorded at $1m^2$ intervals, as outlined in Chapter 3. This allowed for detailed observations of the soil moisture gradient. Soil moisture was also recorded along these transects and others, once in summer, autumn and in spring, so as to gain a better overall impression of the soil moisture gradients during different seasons.

To help delineate the wetland boundaries and to reveal relationships that may exist between the soil properties themselves, topsoil collections were taken along the transects at 5 m intervals, as described in Chapter 3 (Table 5.5). These samples were collected across representative areas of the valley floor. Although they cannot be regarded as being representative of the valley head as a whole, it will be demonstrated that these

results provide a good indication of the general trend of topsoil characteristics in the valley heads.

Table 5.5 Number of soil samples taken along each transect at 5 m intervals.

	Soil Samples ca. 250 g	Length of Transect
Valley 1	39	190 m
Valley 2	29	145 m
Valley 3	36	180 m
Valley 4	57	300 m
Valley 5	25	120 m

Soil horizons within the soil profiles were examined to further understand the depositional history of the valley heads and samples were collected and examined to aid in the classification of wetland type as well as to provide information on soil characteristics and how they pertain to gully initiation and erosion.

The following results are grouped so that the non-gullied wetlands transects are shown separate from the gullied wetlands. This is done so that comparisons can be more easily made within each group and between each group. Since the valley heads were formed through similar depositional conditions, the descriptions of the soil types and profiles are grouped according to type (i.e. mineral and wetland) rather than by geographic position.

As outlined in Chapter 3, the soil moisture values are grouped along a gradient. The term 'very dry' is used to describe values between 0 – 0.2 $m^3.m^{-3}$, 'dry' describes values between 0.21 – 0.35 $m^3.m^{-3}$, 'moist' refers to values between 0.351 – 0.5 $m^3.m^{-3}$, 'wet'

refers to values between 0.51 - <1 $m^3.m^{-3}$, and 'saturated' is used for soil moisture values of 1 $m^3.m^{-3}$.

5.3.1 Results from Soil Analyses From Valley Heads 1 And 2 (Non-Gullied Wetlands):

Soil transect results: soil moisture and soil properties

The surface soil moisture recorded in November 2001, January 2002 and April 2002 across the representative soil transects in valley heads 1 and 2 were consistently saturated; Figures 5.7 and 5.8 highlight the wetland's uniform soil saturation in April 2002 across valley heads 1 and 2. However, in September 2001, which preceded 5 relatively dry winter months (< 133 mm precipitation) in which the probe could not penetrate the soil, the soil moisture data from the mire expanse in valley 1 revealed two localized areas that were below saturation (Figure 5.9). At the same time, in valley head 2, the surface water gradient of the mire was not saturated across a width of 55 m, as for the other months, but instead was saturated across only 30 m of the wetland. The isolated soil moisture decreases within the wetland of valley head 1, from 'saturated' to 'wet' conditions, were each associated with livestock damage and/or burrows. Conversely, the soil moisture content of the mineral soil along the valley head floor was erratic in each month and shifted from being 'dry' to 'very dry'. A clear distinction between the wetland and dryland areas of the valley floor is evident (Figures 5.7 and 5.8).

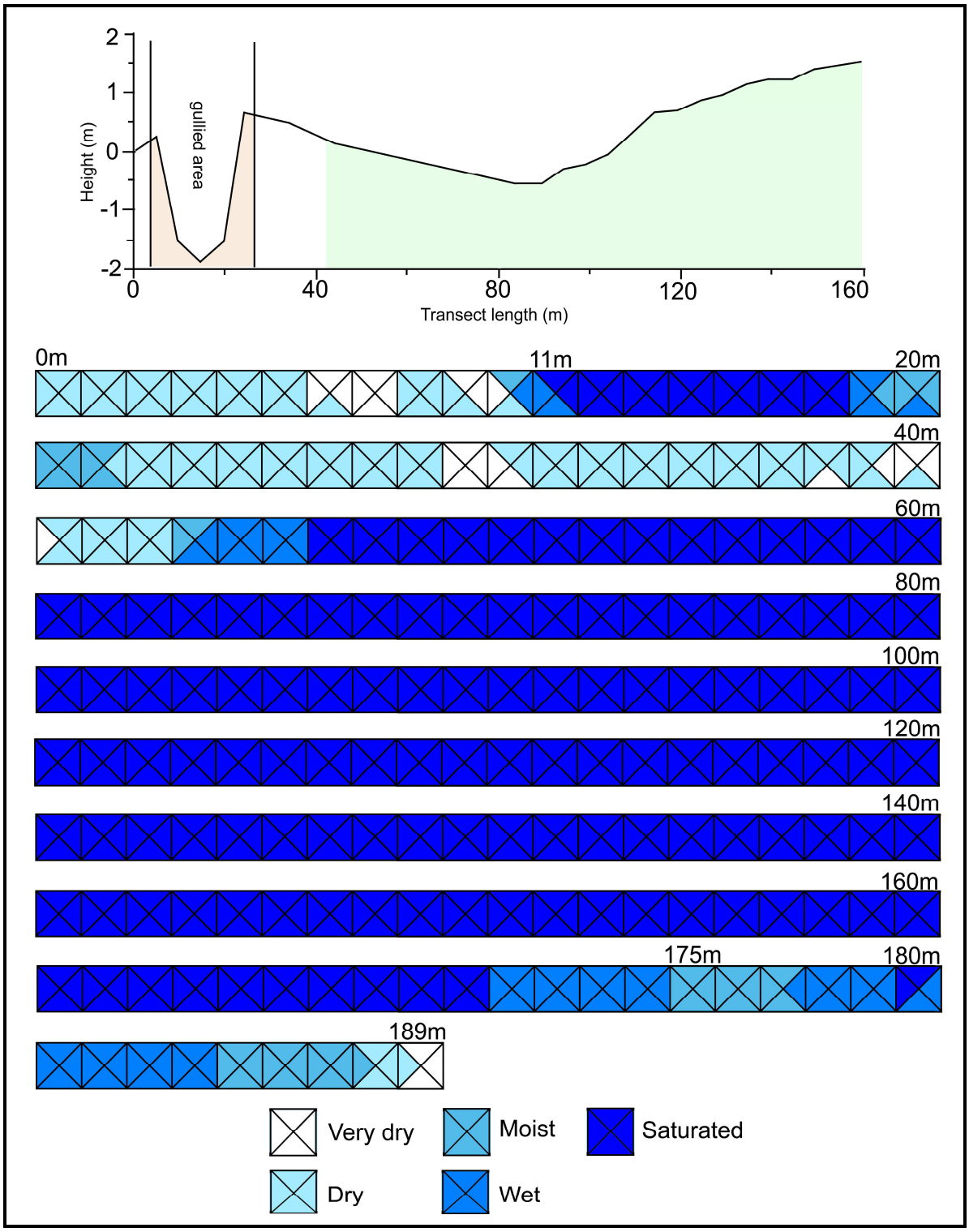


Figure 5.7 Detailed soil moisture grid within valley head 1. The top profile illustrates the slope along transect #2, highlighting the wetland area (in green) and the gullied area (in brown). The areas not shaded represent mineral dryland soil.

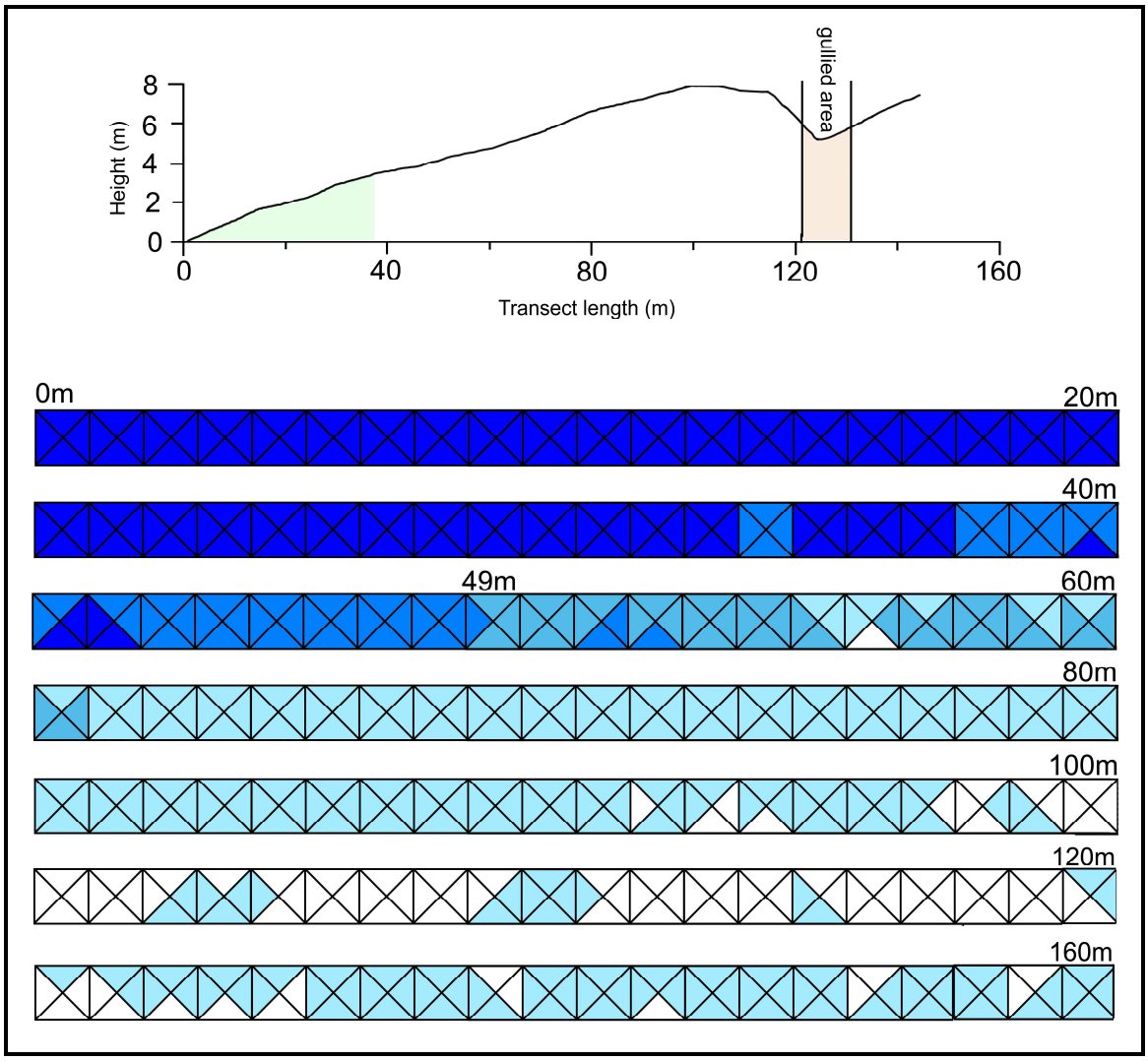


Figure 5.8 Detailed soil moisture grid within valley head 2. The top graph illustrates the slope profile of transect #2, highlighting the wetland area (in green) and the gullied area (in brown). The non-shaded area represents the mineral valley head floor.

It appears that the mire margin fluctuates slightly depending on season and moisture availability, however there is a stable, visible vegetation community change in this fringe zone (described in Chapter 6). The mire margin is also marked by a marked decrease in organic matter from the mire expanse. For example, when passing from the mire expanse onto the mire margin in valley head 1 (ca. 10 m band), soil moisture content decreased,

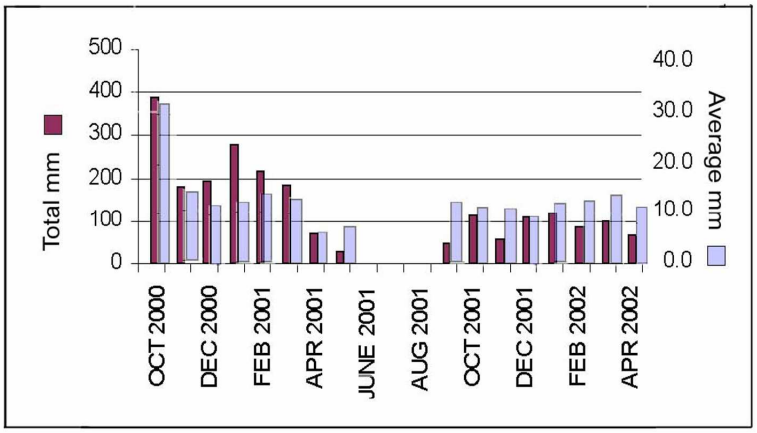


Figure 5.9 Rain data from October 2000 to April 2002. The light purple bar illustrates the average amount of rain received in each month. The burgundy bars represent the total amount of precipitation received in the respective months.

the vegetation community changed to dryland herbaceous species and organic matter content decreased from 24% to 16%. In valley head 2, when passing from the expanse into the margin, organic matter content decreased from 30% to 15% over a distance of only 10 m.

Transect data revealed definite trends between the soil moisture, pH and organic matter content collected across the two non-gullied wetlands (Figure 5.10). The highest pH value appears to coincide with the lowest organic matter content within the gullied areas where soil moisture is low. Inversely, the lowest pH value coincides with the highest organic matter content within the mire expanse, where the soil is saturated. These apparent relationships were rank correlated and the results from each transect were compared and contrasted (Table 5.6). The pH of the surface soil provided a weak correlation with soil moisture in valley heads 1 and 2, whereas in valley head 1, pH and organic matter shared a strong, negative Spearman's Rank correlation. Organic matter and the surface soil moisture along each transect revealed a strong, positive relationship.

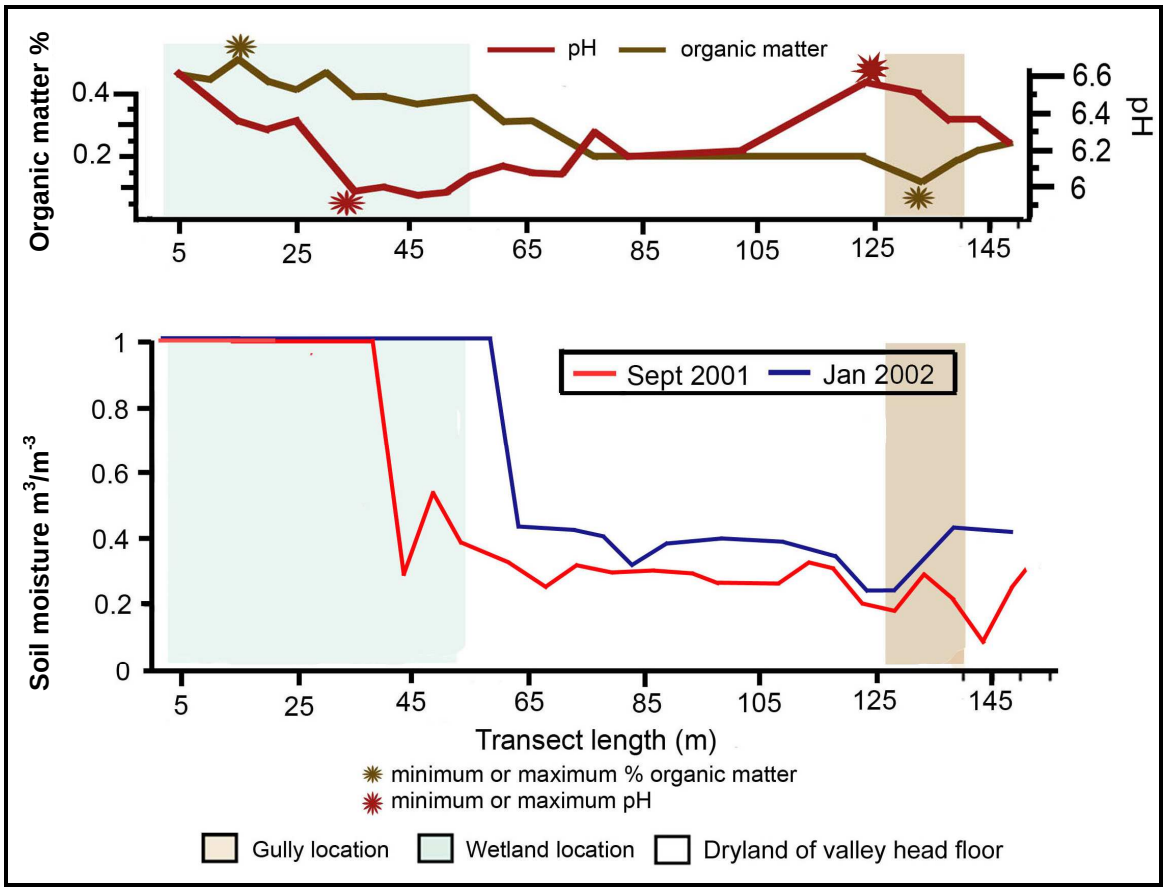


Figure 5.10 Example of the correlation between soil moisture (m^3/m^{-3}) content with organic matter % and pH values recorded along the non-gullied soil transect in valley head 2. Soil moisture is shown from the driest month (i.e. September 2001) and the wettest month (January 2002) for comparison.

Table 5.6 Spearman's rank correlation coefficients for various parameters measured along the entire soil transect in each valley head. ✓ = the null hypothesis was rejected, the sample is statistically significant. ✕ = the null hypothesis is not rejected; 'wet' = wetland; 'dry' = dryland; OM = % organic matter; SM = soil moisture.

	V1 n = 39		0.05	**V2** n = 22		0.05	**V3** n = 29		0.05	**V4** n = 57		0.05	**V5** n = 25		0.05
OM/SM	0.5		✓	0.9		✓	0.6		✓	0.4		✓	0.5		✓
OM/pH	-0.7		✓	-0.1		✕	0		✕	-0.3		✓	-0.3		✕
SM/pH	-0.1		✕	-0.3		✕	0.2		✕	0.3		✓	0		✕
	ca.	min	max	ca.	min	max	ca.	Min	max	ca.	min	max	ca.	min	max
Wet pH	6.15	5.5	6.7	6.14	5.9	6.5	6.33	6.1	6.6	6.33	4.8	7.1	6.5	6.1	7.1
Dry pH	6.34	5.9	6.9	6.19	6.0	6.5	6.26	5.8	7.1	6.11	5.2	7.1	6.6	6.3	6.7
Wet OM	39	14	72	32	28	38	28	22	31	34	14	62	39	18	49
Dry OM	15	5	20	17	9	23	20	10	27	19	8	31	23	17	30

Surface soil properties adjacent to the gully edges

Since valley head wetlands 1 and 2 are not gullied, the surface soil properties adjacent to the dryland gullies, each crossed by transect #2, were examined instead (labels shown in Figure 3.1 – 3.5). In valley head 1, this particular section of the gully had been reclaimed (the bottom of the gully was saturated and inhabited by wetland vegetation) and was a few metres upstream of the gully mouth. The dryland areas adjacent to the gully and the gully sidewalls had relatively high pH values (average 6.64) and a low organic matter content (< 16%). The soil adjacent to the dryland gully in valley head 2 had some of the lowest recorded organic matter contents (14% and 15%) and also had a high maximum pH value (pH 6.19) relative to the overall pH values for the transect (6.45) (Figure 5.10). Interestingly, the lowest organic matter percentages and the highest pH values from both valleys 1 and 2 were sampled from the dryland gully walls/edges (Table 5.6). Soil moisture was also driest adjacent to the gully edges and became progressively less so towards the wetland, away from the gully edge. During the study, recorded soil moisture values on either side of the gully walls was never above 'moist' (0.35 – 0.5 $m^3.m^{-3}$), and was usually 'dry' (0.2 – 0.35 $m^3.m^{-3}$) to 'very dry' (0 – 0.2 $m^3.m^{-3}$).

5.3.2 Valley Head 5

Soil transect results: soil moisture and soil properties

In January 2002 and in April 2002, surface soil moisture was recorded along transect #3 across valley head 5. Surface soil moisture trends collected in these two months varied considerably (Figure 5.11). The mire expanse was saturated in January, yet in April the soil moisture gradient had become more varied. Adjacent to and around the small gully,

the soil moisture gradient was irregular and dry (Figure 5.12). From approximately 15 m away from the gully (from 50 m to 105 m along the transect), soil moisture increased but did not become saturated. Similarly, percent organic matter (collected in November) also began to gradually decrease at the 50 m transect position.

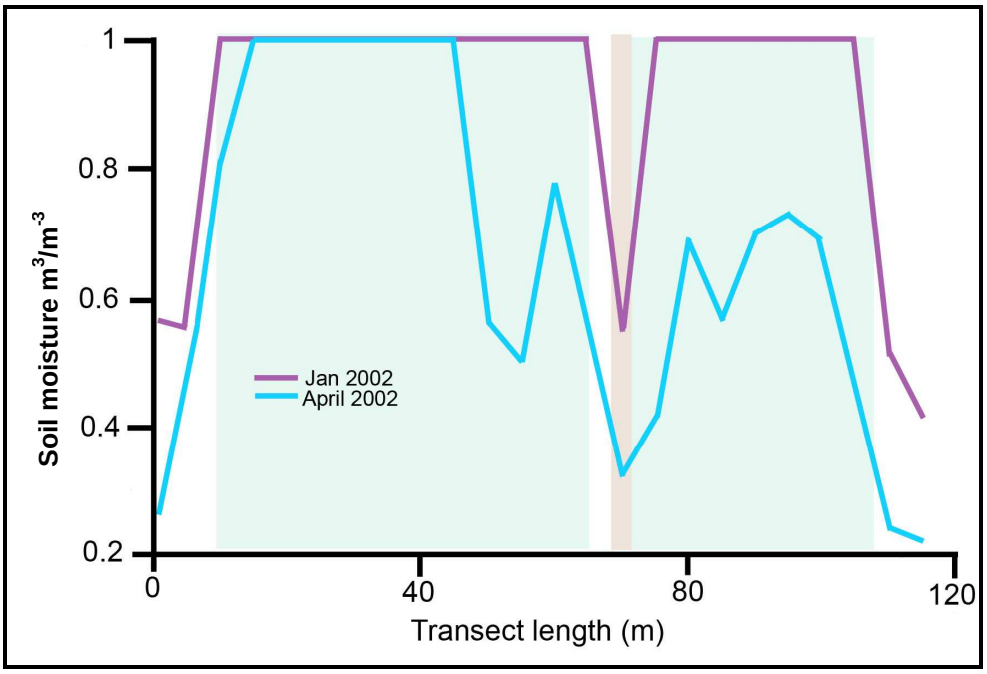


Figure 5.11 Soil moisture (m^3/m^{-3}) recorded along transect #3 in valley head 5 on two separate occasions. The wetland area is highlighted in green, whereas the gullied area is highlighted in brown. Non-shaded area represents dryland valley floor or degraded wetland.

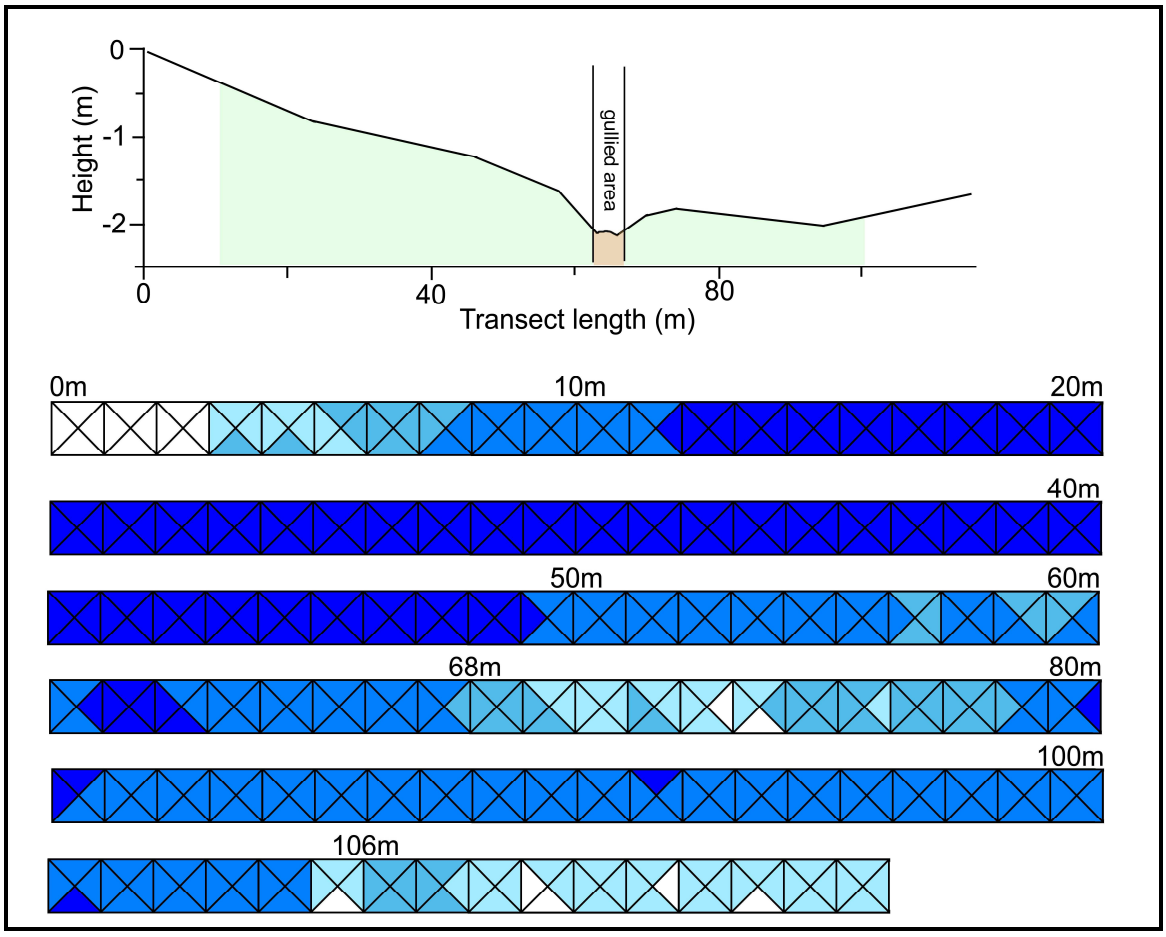


Figure 5.12 Detailed soil moisture grid within valley head 5. The top graph illustrates the slope profile of transect #3, highlighting the wetland area (in green) and the gullied area (in brown). The non-shaded area represents the mineral valley head floor.

The transitional or fringe area of the mire was defined by lessened soil moisture, small dryland herbs and a lower average percentage organic matter. The quantity of moisture within this area varied according to season; from 'moist' in January to 'dry' in April. As one moves outwards from the mire expanse towards the surrounding dryland area, abrupt changes in vegetation composition is recorded, and is also associated with an increase in pH and a decrease in organic matter content. On the north end of the transect, when moving just 1 m from the mire expanse into the transitional area, organic matter content decreased from 39% to 23%. On the south end of the transect, organic matter content

decreased from 49% in the wetland to 30% on the dryland (Figure 5.13). The maximum dryland organic matter content and minimum pH value recorded along the transect were measured along the mire margin (i.e. at the 10 m transect position) (Table 5.6 and Figure 5.13).

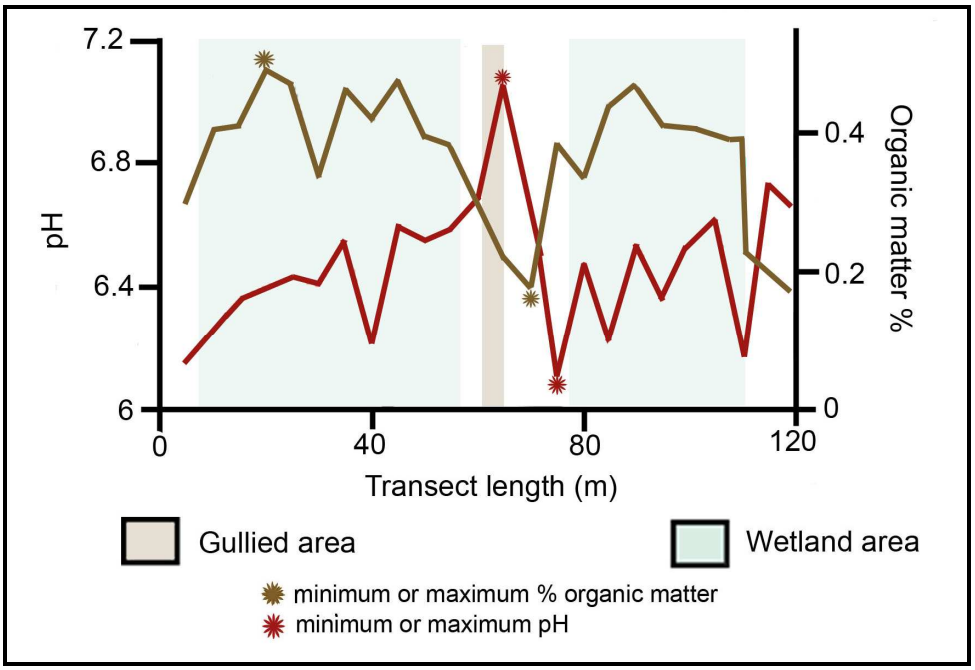


Figure 5.13 Organic matter % contrasted against pH values across the soil transect (#3) in valley head 5.

Surface soil properties adjacent to the gully

Valley head 5 has a small gully with a rill-abrupt morphology in the approximate centre of the wetland that is crossed by transect #3. Adjacent to the gully, from the 75 - 80 m transect position, the soil surface consists of exposed dry gravels and denuded wetland vegetation. The lowest organic matter content and maximum pH value were sampled from this damaged area in valley 5 (Figure 5.13). Although the surface of the wetland was saturated in January, it appears that the changes incurred to the peat as a result of the

wetland erosion (i.e. decreases to organic matter content and an increased pH) are only being masked by the seasonal saturated conditions. During the drier months, the effect of the gully on soil moisture becomes more evident. The decreasing trend in soil moisture content in April revealed that although the surface soil moisture fluctuates seasonally, the organic matter content portrays the more usual saturation state since its accumulation is dependent on more consistent conditions. As expected, organic matter (collected in November) was found to be dependent on, and to share, a significant and positive rank correlation with the surface soil moisture (0.5 where $P < 0.05$) (recorded in January) (Table 5.6). No significant relationship was found between the pH and soil moisture or organic matter content.

5.3.3 Valley Heads 3 And 4 (Gullied Wetlands)

Soil transect results: soil moisture and soil properties

Transect #'s 1 and 2 in valley heads 3 and 4 respectively, illustrate that the soil moisture content across the surface of the wetland was erratic within all recorded seasons; Figure 5.14 is an example of the soil moisture gradient across valley head 3 in April 2002. Across all of the transects, there were localized areas which are classified as 'dry' to 'very dry'. The surface soil moisture of the dryland areas was consistently 'dry' to 'moist' in all months recorded.

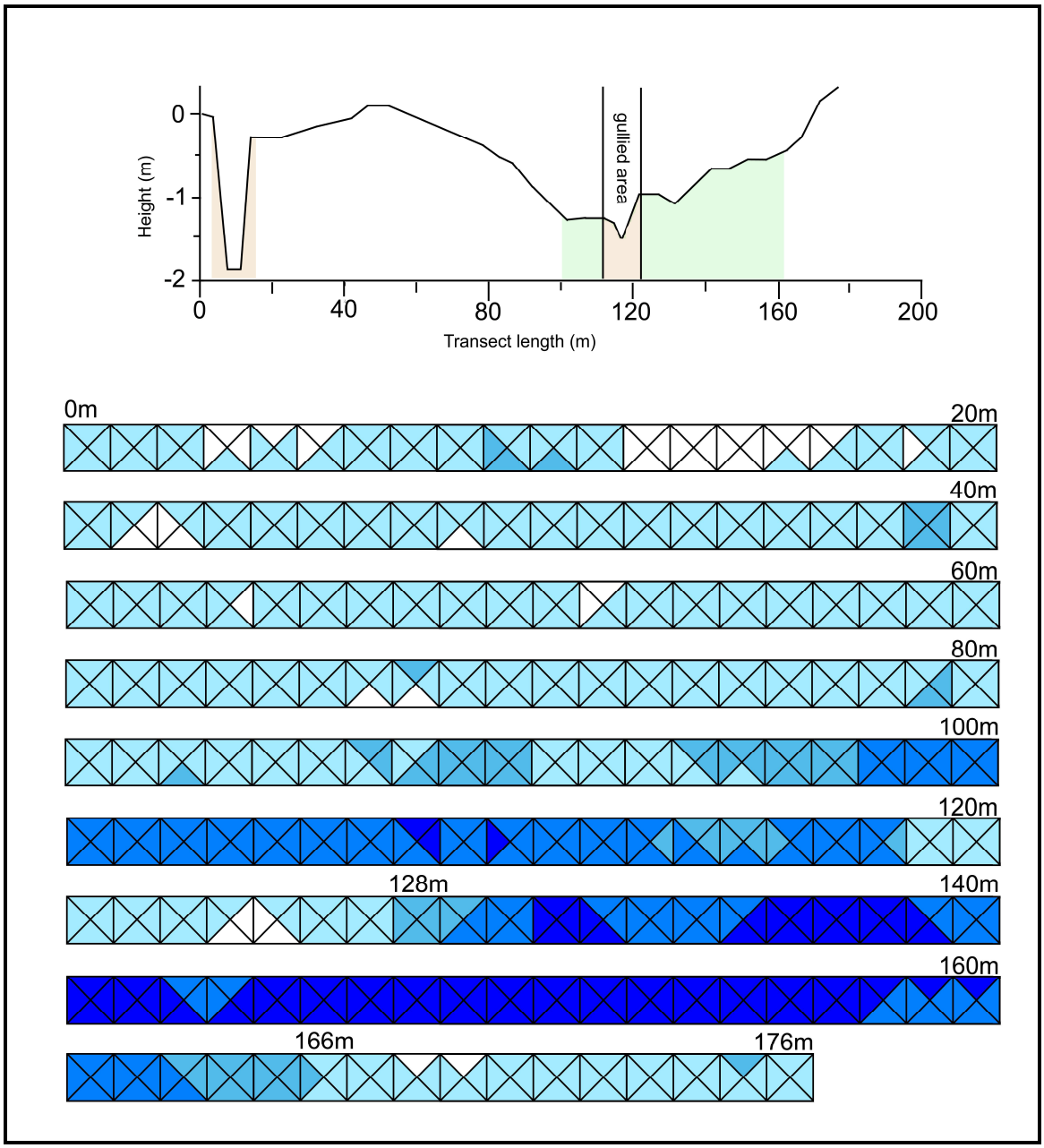


Figure 5.14 Detailed soil moisture grid within valley head 3. The top graph illustrates the slope profile of transect #1, highlighting the wetland area (in green) and the gullied area (in brown). The non-shaded area represents the mineral valley head floor.

Valley Head 3

According to the soil moisture gradient in January 2002, the width of the wetland expanse along transect #1 was 65 m (extending from 95 to 160 m along the transect).

However, in September 2001, November 2001 and April 2002, the surface soil moisture gradient did not extend as far into the dryland, nor was the mire expanse as uniformly wet (i.e. soil moisture only became saturated at 100 m). In September 2001, the entire transected mire expanse never reached a wetness category higher than 'moist'. The highest recorded organic matter content and lowest pH value were recorded within the undamaged area of the mire (Figure 5.15). The mire margins appeared to be well defined along the transect, with vegetation community changes and increases in organic matter content occurring from the dryland towards the wetland expanse.

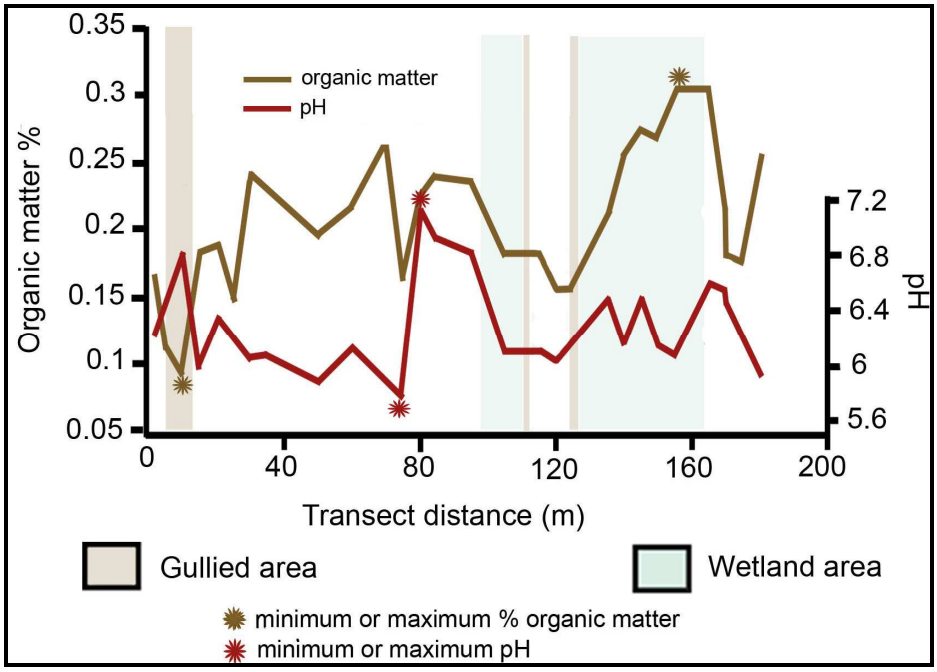


Figure 5.15 Organic matter % contrasted against pH values across the soil transect in valley head 3.

Surface soil properties adjacent to the gully

Visible along transect #1 in valley head 3 were the affects of the small rills and gullies from 110 – 125 m and of the deep gully at the eastern end of the transect (Figure 5.14).

The former erosional features were shallow (< 60 cm deep), however the adjacent soil moisture was never recorded as 'saturated'. In September 2001, the mire expanse to the west of these features (from 125 m – 160 m along the transect) was never saturated. Organic matter content for the western portion of the mire expanse is not as high as that for the eastern portion. Where soil moisture decreased adjacent to the gully edges, organic matter contents also decreased (i.e. from ca. 24% to 16%) (compare Figure 5.14 to 5.15). The increase in soil moisture and organic matter content with increasing distance from the gully provided a Spearman's rank correlation of 0.53 and 0.39 respectively, where $n = 29$, $P < 0.05$) (Table 5.6). The maximum pH and lowest organic matter and soil moisture contents were recorded adjacent to the deep dryland gully (Figure 5.15). No significant rank correlation was recorded for soil pH with distance from the gully edge (Table 5.6).

Valley Head 4

Soil transect results: soil moisture and soil properties

In valley head 4, the wetland surface was not consistently saturated during any month; Figure 5.16 is an example of the erratic soil moisture gradient recorded in April 2002. The wetland expanse is essentially divided into two portions by a large gully (Plate 5.2). The western side of the mire has extensively burrowed areas, which coincide with localized areas of dry soil and low organic matter content (Figure 5.16). Generally, the eastern portion of the wetland had higher soil moisture values, greater organic matter content and lower pH values (Figure 5.16).

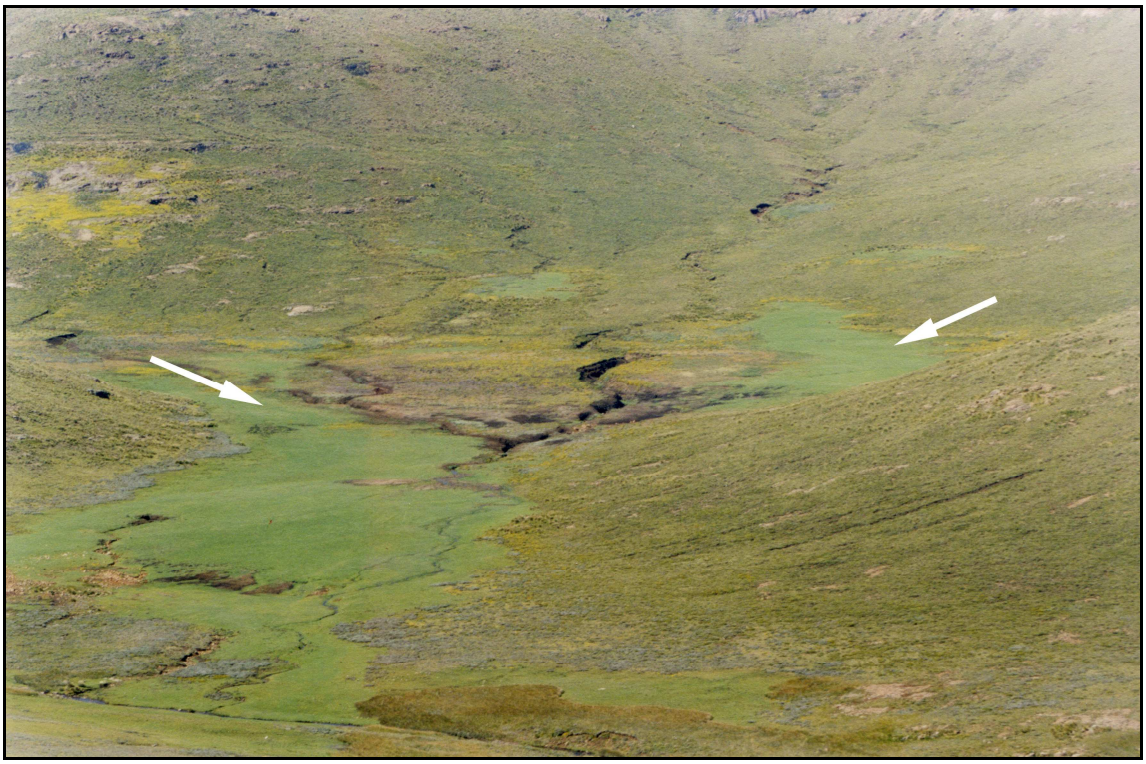


Plate 5.2 View of valley head 4 taken from the north; the centre gully is clearly seen to have divided the top half of the wetland into two; arrows highlight the two large wetland areas (November 2001).

The eastern mire fringe area is distinct and coincides with high soil moisture values (saturated) and increased organic matter contents (27% to 41%). This transitional area has many *Otomys* burrows (> 2 m^2) and is associated with isolated lessened soil moisture levels and isolated decreases in organic matter content (Figure 5.16). The western mire margin was less distinct and even less uniformly saturated than the eastern side.

Surface soil properties adjacent to the gully

For at least 25 m on either side of the large wetland gully, surface soil moisture values are highly variable (Figure 5.17). Coinciding with the decreasing soil moisture the organic matter content also began to decrease substantially (62% to 36%). A relatively strong correlation value of 0.61 (where $n = 53$; $P < 0.05$) is recorded for decreasing soil moisture

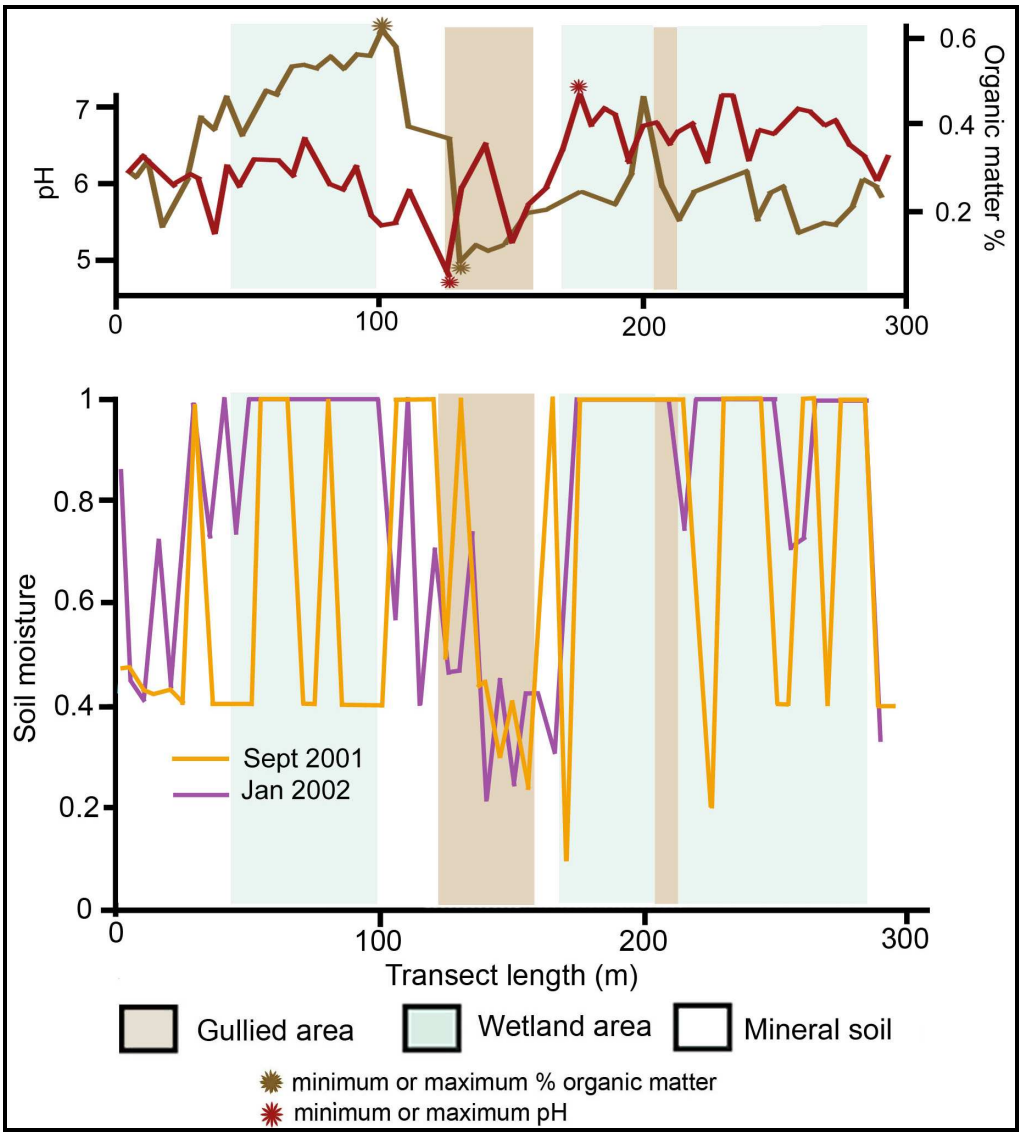


Figure 5.16 Example of the correlation between soil moisture content with organic matter % and pH values recorded along the non-gullied soil transect in valley head 4. Soil moisture (m^3/m^{-3}) is shown from the driest month (i.e. September 2001) and the wettest month (January 2002) for comparison.

with distance from the gully edge (Table 5.6). A relatively strong correlation of 0.4 moisture (where n = 53; $P < 0.05$) was also recorded for organic matter content with soil moisture. Although the soil profile at the gully edge consists of thick organic profiles,

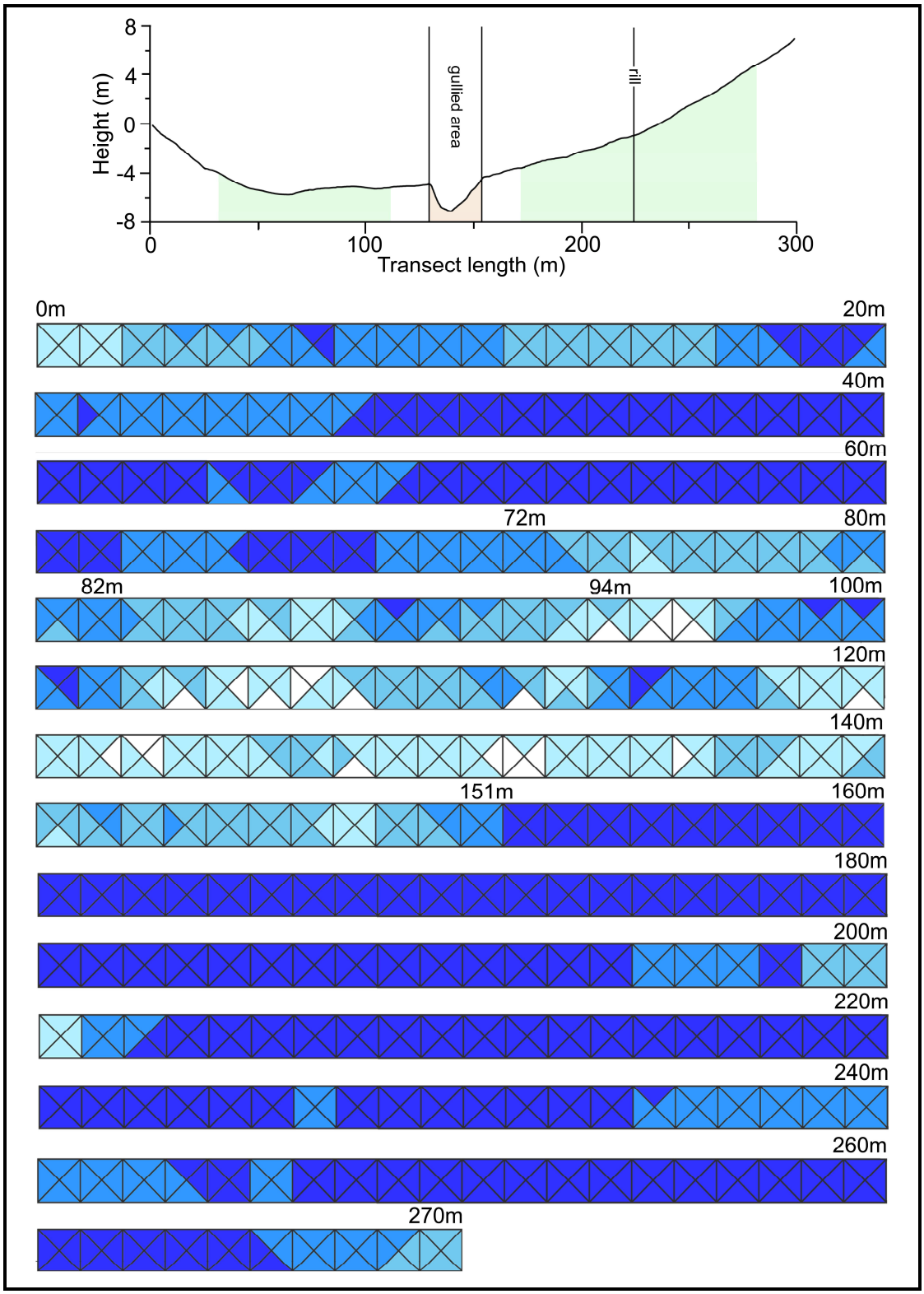


Figure 5.17 Detailed soil moisture grid within valley head 4. The top graph illustrates the slope profile of transect #2, highlighting the wetland area (in green) and the gullied area (in brown). The non-shaded area represents the mineral valley head floor.

organic matter content is insufficient to classify such sediments as peat. This affirms the assumption that the gully did indeed cut through the mire and that the exposed peat at the gully edge had been oxidized. The lowest organic matter content was recorded from the gully floor and the maximum pH value was recorded from the desiccated gully edge (Figure 5.16). The topsoil pH was also rank correlated with increasing distance from the gully (0.33, $n = 49$, $P < 0.05$). There was no significant relationship between soil organic matter content and distance from the gully.

5.3.4 Summary Of The Surface Soil Results

In summary, the non-gullied wetlands were almost consistently saturated during the field recordings in summer, spring and autumn. All of the mire expanses demonstrated a trend of decreasing soil moisture towards the mire margins.

The surface soil moisture of the gullied mire expanses was not consistently saturated. The surface soil moisture of the wetlands in valley heads 3 and 4 could not mask the effects of burrows, rills and deep gullies during any of the months investigated. The average organic matter contents decreased by ca. 10% adjacent to the gullies (within 15 m from the gullies). Similarly, soil moisture is rapidly reduced adjacent to the gullies and significant rank correlations were calculated for both valley heads. The lowest organic matter content, lowest soil moisture readings and maximum pH values were all recorded in proximity to the wetland gullies in valley heads 1, 2 and 5.

5.3.5 Description And Classification Of The Alpine Mineral Soils And Histosols

Depositional history of the gullied wetlands

To aid in the classification of the wetlands and to understand the developmental histories of the valley heads, soil profiles were examined from wetland and dryland locations within each valley head. Emphasis is placed on valley heads 3 and 4 because of the deep soil profiles that are exposed by gully erosion (Figures 5.18 and 5.19). The profiles were then compared to the profiles analyzed and dated by Marker and Whittington (1971) and Marker (1994, 1998).

The valley floors consist of an accumulation of two main types of sediment, namely clastic (diamictons and orange gravels) and organics. Clastic sediments generally record fluctuating depositional energies of the period (Thomas, 2004). In the high Drakensberg, the diamictons are a product of cold periods and cryogenically-induced processes. The orange gravels, which obtained their colour from the oxidation of the basalt-derived iron, are actually derived from the diamicton material and have accumulated through weathering and transportation from upslope (Hanvey and Marker, 1994).

The organic sediment (< 80% organic matter) represent climatic conditions that allowed for long periods of saturation and is thus peat forming (Marker, 1995). The differing horizon thicknesses throughout the valley profiles suggest varying local depositional conditions, whilst the profile sequences represent climatic transitions (Marker, 1995).

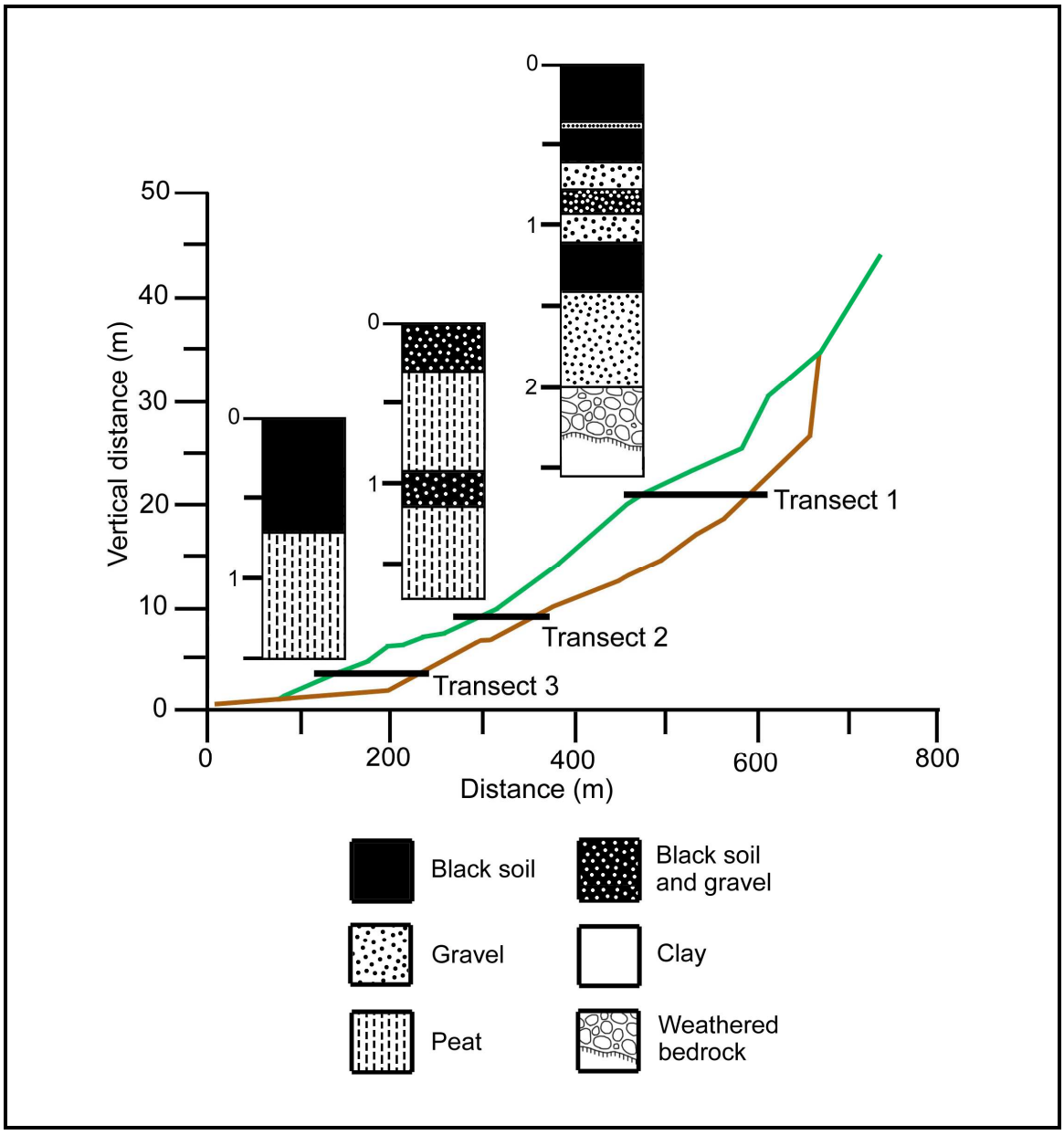


Figure 5.18 Longitudinal profile of the gully (brown) /valley (green) slope within valley head 3 with transect positions. The soil profiles of these intersections are shown and described.

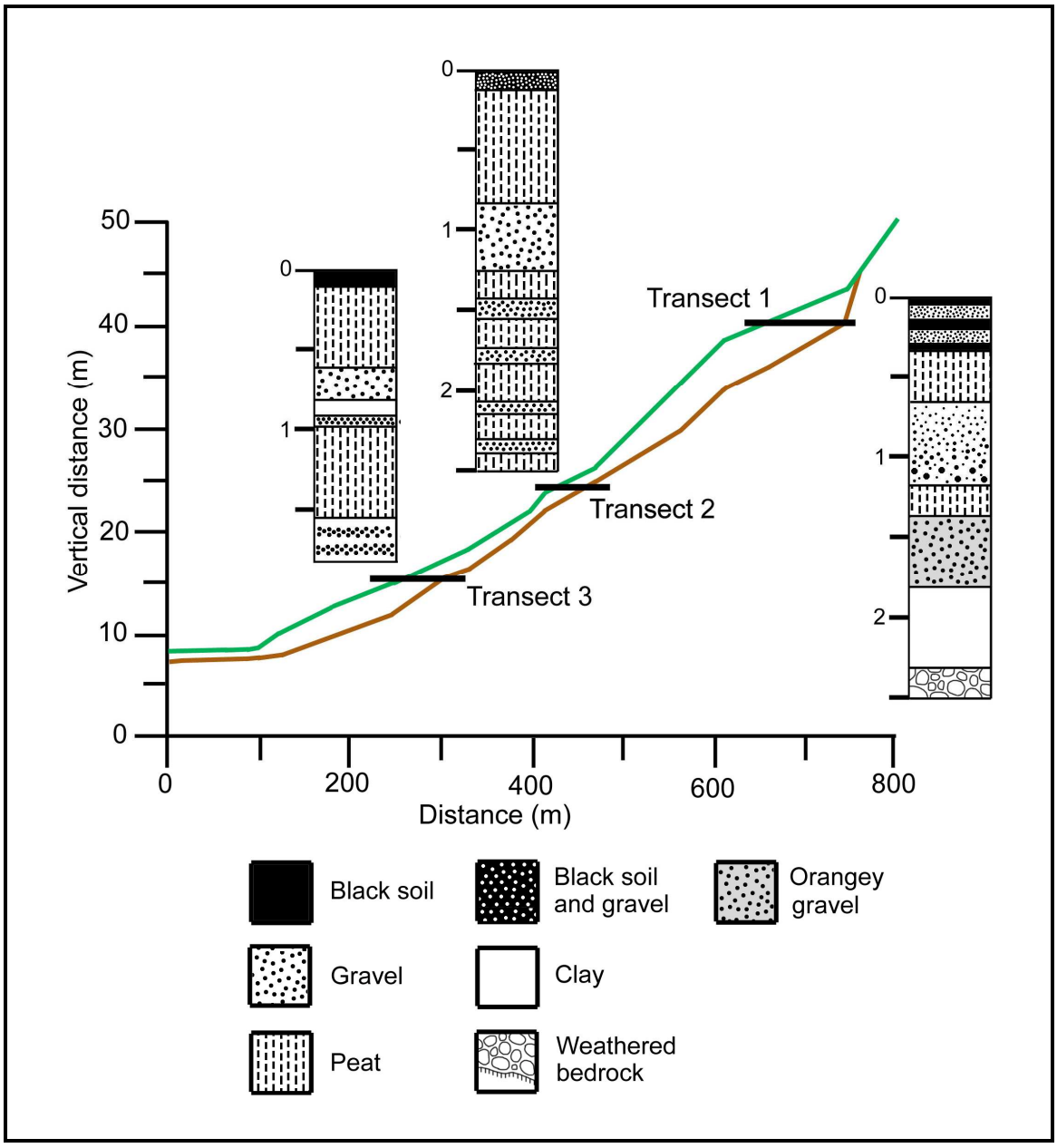


Figure 5.19 Longitudinal profile of the gully (brown) /valley (green) slope within valley head 4 with transect positions. The soil profiles of these intersections are shown and described.

The organic horizons are always found above gravels, which suggests that periods of cold, dry conditions with reduced vegetation cover and associated increased runoff, were replaced by warmer and wetter conditions that would have favoured continuous cover and organic accumulation in wet areas, and clay accumulation in drier areas (Marker

1994, 1995). Narrow, varve-like layers, which are most evident in the gully sidewalls within valley 4, could represent seasonal fluctuations rather than climatic changes to the hydrological conditions of the area (Marker and Whittington, 1971; Marker, 1994, 1995). As is typical of alpine soils, the fines (sands, silts and clays) are most dominant towards the top of the profiles (Munn and Spackman, 1990; Hanvey and Marker, 1994). The present climate of Lesotho has formed the dark, loamy cap of organic matter and/or mollic epipedons (Hanvey and Marker, 1994) Within many of the profiles, two main phases of organic accretion are evident (Plate 5.3). These organic horizons imply that two distinct periods of enhanced accumulation occurred (see also Marker, 1995). Marker (1994) dated two large organic horizons in valley 3, similar to the ones observed in this



Plate 5.3 Two large peat deposits clearly evident in this soil profile, valley head 3 (June 2003).

study. The deepest horizon provided a radiocarbon date of 13 490 ± 130 BP (Marker, 1994), which places the beginning of the organic accumulation at the onset of global warming succeeding the Last Glacial Maximum (Grab *et al*., 2005a; Lewis, 2005). The date from the basal area of the second large peat accumulation (4 740 ± 60 BP to 1 910 ± 50 BP) indicates renewed phase of warmer and wetter conditions (Marker, 1994, 1998). The time frame provided for this phase correlates well with the warmer periods that are said to have occurred in southern Africa (Partridge *et al*., 1990; Mitchell, 1995; Meadows and Baxter, 1999).

5.4 Discussion

It has been demonstrated that the affect of gully erosion on surface soils is both direct and indirect. As illustrated with the soil moisture transects in each valley head, gully incision has caused the adjacent soil to desiccate, as has also been reported by Hanvey and Marker (1994). It is further demonstrated that all surface soil moisture decreases coincide with areas of erosion or surface cover change (i.e. burrows, rills and/or gullies). Decreases in soil moisture correlate well with distance from the gully. Peat desiccation enhances oxidation and consequent deflation by wind (de Mars *et al*., 1996; Evans and Warbuton, 2001). The desiccation of sediment within the mire provides a more suitable habitat for *Otomys sloggetti* (Grobbelaar and Stegman, 1987; Hanvey and Marker, 1994; Morris and Grab, 1997; Hall *et al*., 1999).

5.4.1 Dryland Soils

Most of the mineral soil types examined within this study are classified here as Mollisols, since they have dark brown to brown black (chroma value ≤ 4) mollic epipedons thicker than 18 cm, and are rich in organic matter (11% – 16%). However, depending on the topographic position, a variety of Mollisols were encountered within each valley head. Most soil profiles examined lacked a well-developed B horizon (Plate 5.4). Instead, these Mollisols have a paralithic or lithic contact onto the basaltic saprolite or basalt, respectively.



Plate 5.4 Dryland gully wall with limited soil development (June 2003).

The mineral Mollisols have a udic moisture regime, meaning that they are not dry for as long as 45 consecutive days during the growing season. These soils also reflect the

properties of the basaltic Lesotho Formation parent material (i.e. high levels of basic cations associated with calcium [10.5% CaO] and magnesium) and have a clay mineralogy typified by montmorillonite (Young, 1976; Schmitz and Rooyani, 1987; Klug *et al*., 1989; Kakonge, 2002). On average, the dryland soils have a relatively neutral reaction, high percent clay (especially in valley 5), low bulk densities and high total pore space percentages (Table 5.9) (Figure 5.20). Due to the high net negative charge and large surface area associated with montmorillonite clay, as well as the high percentage of it, these results suggest that the mineral soils are efficient at adsorbing and exchanging cations with those in the soil solution (Juma, 1998; Snyman and du Preez, 2005).

Other Lesotho alpine soil studies have also found high cation exchange capacities, similar pH vales (ca. 6), high base saturation, significant percentage of organic matter and high proportions of clay (Schmitz and Rooyani, 1987; Klug *et al*. 1989; Nkalai, 1991; Kakonge, 2002). The results correspond well with cited characteristics for dry, upland soils (i.e. upland mineral soils are expected have bulk density values from 1000 – 2000 gcm^{-3}) and total pore space, regardless of clay content (Mitsch and Gosselink, 1986) (Table 5.8). Results also correspond well with organic matter contents typical for Popa and Fusi Mollisols (i.e. >10%) (Schmitz and Rooyani, 1987).

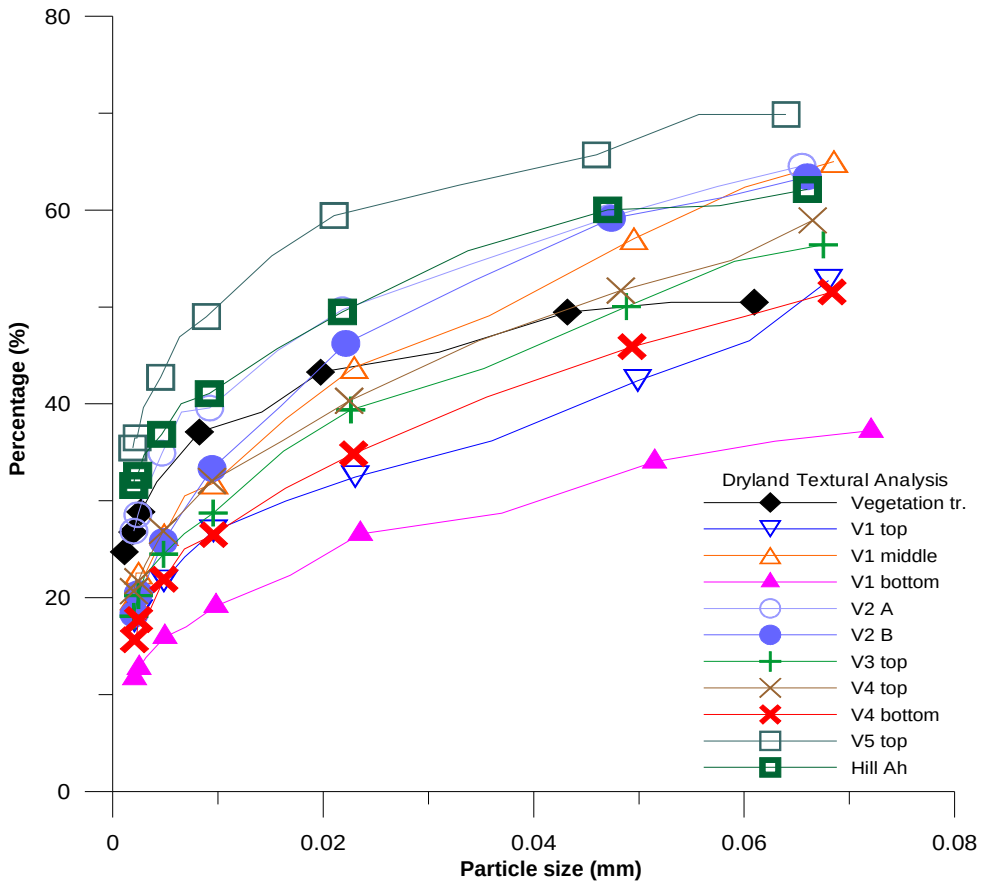


Figure 5.20 Particle-size graph of the dryland minerals collected from all valley heads, with accompanying table, depicting percentiles and class types.

Table 5.7 Table depicts percentiles and class types for Figure 5.20 above.

	>0.06mm	>0.002mm	<0.002mm	Class
V1 top	53%	29%	18%	Sandy loam
V1 *clay* middle	38%	42%	20%	Loam
V1 bottom	64%	24%	12%	Sandy loam
V2 top	38%	35%	27%	Loam/CL
V2 bottom	38%	44%	18%	Loam
V3 top	45%	37%	18%	Loam
V4 top	45%	34%	21%	Loam
V4 *clay* bottom	51%	33%	16%	Loam
V5 top	30%	34%	36%	Clay loam
Hill top	40%	27%	33%	Clay loam
Veg'n transect	50%	23%	27%	Sandy clay loam

Table 5.8 Chart summary of the mineral soil A horizon characteristics.

	pH	OM	Bd	Clay:Sand:Silt	Class	CEC	TPS	Colour	Type
VALLEY 1									
A	6.47	11.25	1003.4	18:53:29	L/SL	35	58.9	Brn-blk	1) *Fusi* 2) *Popa/Moroke*
VALLEY 2									
A	6.45	12	801.04	27:38:35	L/CL	43	67.1	Brn-blk	*Matšana* w/o A
VALLEY 3									
A	6.2	18	703.2	18:45:37	L	32.1	89	Grey Y-brn	Similar to *Sofonia*
VALLEY 4									
A	6.32	19	705.04	21:45:34	L	40.2	70.6	Brn-blk	Mollisol, buried histic
VALLEY 5									
A	6.65	23	826.58	36:30:34	CL	50.8	65.2	Black	*Phecela*, Vertisol

These values are from dryland bulk density samples or from dryland profiles from each valley head.
OM = Organic matter %; Bd = Bulk density (gcm-3); CEC = Cation Exchange Capacity;
TPS = total pore space. 1) = A valley bottom profile. 2) = A valley side profile.

5.4.2 Wetland Soils

Using percentage organic matter as a classification factor, the wetland soils of eastern Lesotho have previously been classified as Histosols (peat-forming) (Schmitz and Rooyani, 1987; Schwabe and Whyte, 1993) and as aquic Mollisols (wet mineral soil) (Klug *et al*., 1989). However, this study has found that these wetland soils are not typical peatland examples as defined in section 5.2.2.3. Organic matter content is not greater than 80% and not all peat layers are thicker than 40 cm (Warner, 1996; Bozkurt *et al*., 2001). However, the definition of *peat* is simply that the material must be composed of plant matter with a minimum of 20% organic matter if clay content is less than 50%, or of 30% organic matter if clay content is greater than 50% (Mitsch and Gosselink, 1986; Tiner, 1999). The saturated soils within these valley heads have organic-rich layers with surface and subsurface horizons having greater than 20% organic matter (Table 5.9 and 5.10) and less than 50% clay (Figure 5.21). Furthermore, there is no gleyed horizon

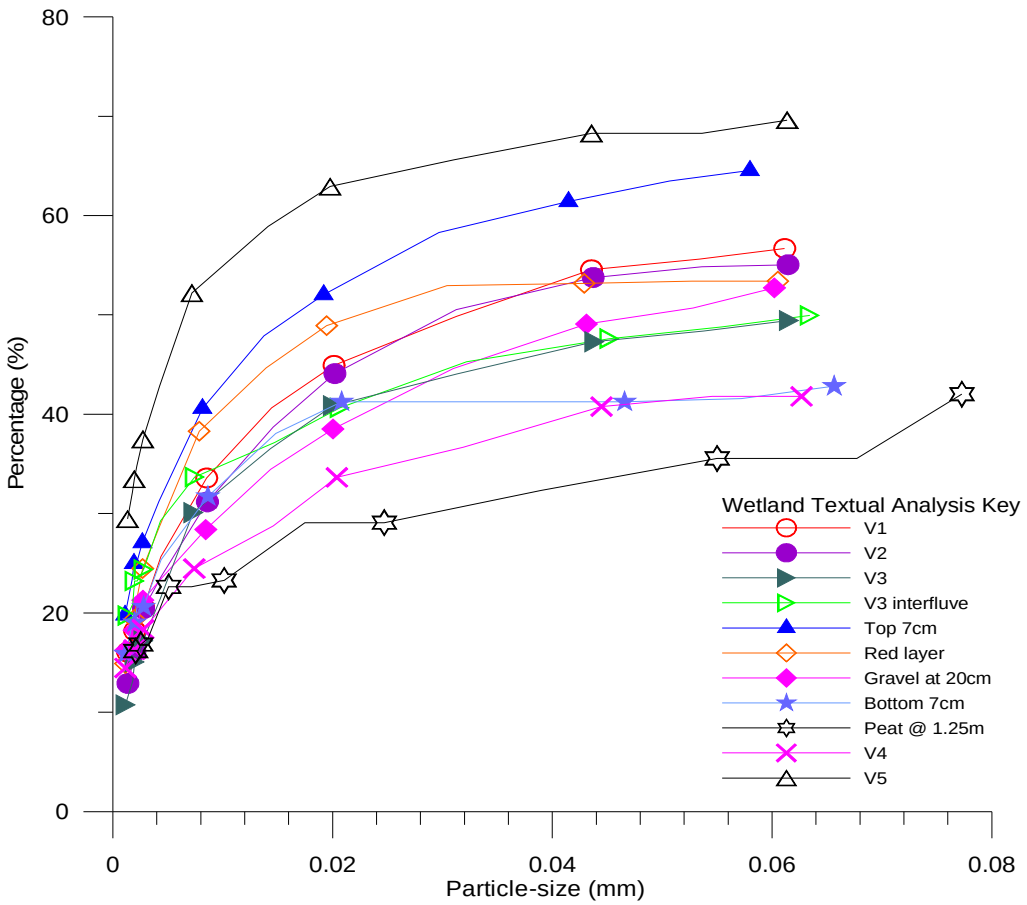


Figure 5.21 Particle-size distribution of the wetland soils collected from all valley heads, with accompanying table, depicting percentiles and class types.

Table 5.9 Table depicts percentiles and class types for Figure 5.21 above.

	>0.06mm	>0.002mm	<0.002mm	Class
V1	43%	39%	18%	Loam
V2	45%	39%	16%	Loam
V3	51%	34%	15%	Loam
V3 interfluve	50%	27%	23%	Sandy clay loam
Top 7cm	*36%*	*39%*	*25%*	*Loam*
Red layer	*47%*	*33%*	*20%*	*Loam*
Bottom 7cm	*57%*	*24%*	*19%*	*Sandy loam*
Gravels @ 20cm	*48%*	*34%*	*18.3%*	*Loam*
Peat @ 1.25m	*65%*	*19%*	*16%*	*Sandy loam*
V4	58%	26%	16%	Sandy loam
V5	30%	35%	35%	Clay loam

Table 5.10 Average values/results from topsoil wetland samples from each valley head. OM = organic matter %; Bd = bulk density; θ = volumetric water content; TPS = total pore space; CEC = cation exchange capacity; used the pH and OM values from the bulk density sample. L = loam; SL = sandy loam; CL = clay loam.

	pH	OM	Bd$_{(gcm^{-3})}$	θ(gcm^{-3})	Clay:Sand:Silt	Class	CEC	TPS
VALLEY 1	6.28	32	452	670	18:44:38	L	48	79
VALLEY 2	6.94	31	526	671	16:45:39	L	46	76
VALLEY 3	5.85	31	534	427	15:50:35	L	47	75
VALLEY 4	6.08	30	371	660	16:58:26	SL	63	81
VALLEY 5	6.45	39	445	656	34:31:35	CL	59	79

Table 5.11 Checklist and comparison of the valley head wetland soils to determine if they meet the requirements for an aquic Mollisol.

AQUIC MOLLISOL	V1	V2	V3(4)	V4	V5
Histic epipedon over Mollic epipedon	✗	✗	✗	✗	✗
Redox features in 100cm of mineral	✓	✗	✗	✗	✗
Grey or mottles in/under Mollic epipedon	✗	✗	✗	✗	✗
High organic matter content (>0.58%)	✓	✓	✓	✓	✓
Base saturation > 50%	✓	✓	✓	✓	✓

Table 5.12 Checklist and comparison of the valley head wetland soils to determine if they meet the requirements for a Histosol.

HISTOSOL	V1	V2	V3 (4)	V4	V5
>20% OM if clay <50%	✓	✓	✓	✓	✓
Organic soil for > ½ of 1st 80cm	✓	?	✓	✓	?
Parent material composed of plants	✓	✓	✓	✓	✓
Developed under saturated conditions	✓	✓	✓	✓	✓

underlying any of the histic epipedons, suggesting that the surface peat horizon is not just a diagnostic Organic (O) horizon. The study therefore classifies these wetlands as mires and the soil profiles as Histosols (see Tables 5.11 and 5.12). More specifically, because these peats are dominantly fibric, tending towards a neutral reaction and have a contemporary and past MAAT of <8˚C, this study classifies these Histosols as

Borofibrists in the *Euic* reaction class (pH > 4.5). The mires themselves are to be classified here as minerotrophic. Specifically, they will be described as eutrophic fens, at the lower limit, since the average pH values for all of the mires was recorded as being above 5.5 but less than 7 (Malmer, 1986; Boeye *et al*., 1996). The presence of flowing water in the drainage channels suggests that these fens are soligenous, as well as geogenous in nature (Wassen *et al*., 1996), which confirms suggestions by van Zinderen Bakker and Werger (1974) and Backéus and Grab (1995).

Although no comparisons of the peat humification and level of compaction can be made between the valley heads due to the limited number of samples, the average bulk density values (between 370 – 530 gcm^{-3}) indicate that the sampled surface peats are well decomposed and/or had been compacted. In comparison to light, newly forming peat has an average bulk density of ca. 40 – 80 gcm^{-3} (according to Mitsch and Gosselink, 1986; Bozkurt *et al*., 2001). This of course is contrary to the initial Von Post field assessments in which the peat samples were given humification values ca. 1 – 3 (i.e. virtually unhumified). However, the peat samples may reflect compaction through livestock trampling for grazing and the loss of organic matter, rather than the actual degree of humification.

Chapter 6

Vegetation Patterns

6.1 Introduction

The science of recognizing that landscapes are composed of separate and natural units of plant species adapted to similar local conditions is known as *phytosociology* (Kent *et al*., 1997; Miles *et al*., 2001). Phytosociology recognizes that although plants respond individually to a simultaneous multitude of gradients according to species type, individual plant species can often share similar and coinciding spatial distributions (Zimmerman and Thom, 1982; Morris *et al*., 1993; Dale, 1999). The manner and tolerance (i.e. their ecophysiological amplitude) with which species respond to a combination of environmental gradients produces the distinctive floral patterns known as a plant community (Burrows, 1990; Rezaei and Gilkes, 2005). Examples of environmental gradients include temperature, pH, aspect, elevation, water table, water chemistry, and nutrient type and availability (Slack *et al*., 1980; Boyer and Wheeler, 1989; Wassen *et al*., 1990; Tickner *et al*., 2001). Other environmental gradients that affect large-scale vegetation patterning are the effects of disturbance including fire, trampling and grazing (Dale, 1999). Although often difficult to detect, changes in plant

community composition represents an area of ecological change and indicates that gradient thresholds have been exceeded (Miles *et al*., 2001).

The zone of change between two or more communities is referred to as a transitional area and can be viewed at a regional, community or quadrat scale (Kent *et al*., 1997). This study focuses on assessing the transitional areas between the upland and wetland communities at the community scale and defining various plant communities within the valley heads. Plant species distributions are examined at the scale of the landform (i.e. physiographic plant geography) (Zimmerman and Thom, 1982).

6.2 Effect Of Grazing On Vegetation Community Patterns

Vegetation cover, abundance and community heterogeneity reflects the micro-environment within the larger landscape, and is thus representative of the climate, parent material, slope, aspect, nutrient availability, altitude, etc. (Milne & Hartley, 2001). When the affects of grazing occur (i.e. compaction, trampling and vegetation loss), the benefits that vegetation cover provides become reduced (Boardman *et al*., 2003; Visser *et al*., 2004; Abule *et al*., 2005). The degradation of vegetation by livestock and/or natural grazing fauna decreases the infiltration capacity of the soil/peat, creates surface crusting, lowers the nutrient status of the soil, can result in shorter root systems, lowers the soil-water content and directly changes the structure, productivity and biodiversity of the vegetation community (Rowntree, 1988; Moleele, 1998; Boardman *et al*., 2003; Rey, 2003; Casermeiro *et al*., 2004; Metzger *et al*., 2005; Snyman and du Preez, 2005). Vegetation community change that inevitably occurs as a consequence of grazing and

trampling is unpredictable and dependent on the resilience and resistance of vegetation, and other factors such as, fluctuations in precipitation, grazing patterns and fire (Rowntree, 1988; Wiegand and Milton, 1996; Milne and Hartley, 2001). Where it occurs however, overgrazing is regarded as the most important cause of landscape degradation (Morgan and Mngomezulu, 2003; Snyman and du Preez, 2005).

Selective grazing and the creation of bare areas caused by compaction and overgrazing can allow for a decrease in palatable herbs or grasses that is often seen to coincide with an increase in shrubs or less palatable species (Thornes, 1985; Rowntree, 1988; Moleele, 1998; Boardman *et al*., 2003; Visser *et al*., 2004; Abule *et al*., 2005; Metzger *et al*., 2005). This change and/or loss of vegetation composition and/or cover allows for accelerated runoff and concentrated overland flow due to the wide spacing of the shrubs and loss of natural protection from raindrop impact (Faulkner, 1990; Abrahams *et al*., 1995; Casermeiro *et al*., 2004).

The invasion of woody species into grass-dominated hillsides and subsequent erosion has been recorded in semi-arid areas in the United States (Abrahams *et al*., 1995; Parsons *et al*., 1996), on Table Mountain and in the Karoo, South Africa (Rossouw, 1997; Boardman *et al*., 2003; Visser *et al*., 2004), as well as in Ethiopia (Abule *et al*., 2005), to name a few. The reversal of this pattern is difficult to achieve since shrub encroachment creates a positive feedback in which the propagation of other vegetation simply becomes too difficult in areas around shrub-islands. This difficulty may be attributed to the higher organic matter, soil moisture and infiltration capacity beneath the shrubs, than between them (Abrahams *et al*., 1995; Abule *et al*., 2005).

6.3 Vegetation Change In The Lesotho Highlands

Alpine areas are particularly sensitive to their environment and are susceptible to the effects of stress due to damage and/or change (Slack *et al*., 1980; Williams, 1988; Hall *et al*., 1999; Milne and Hartley, 2001). The grazing that occurs on the alpine mires in Lesotho initiates similar ecohydrological consequences that would be expected in any grazed area; the vegetation communities experience change in response to species removal, trampling and a lowering of the water table (Brooks and Stoneman, 1997). The removal and/or loss of wetland plant species opens new niches and allows for the encroachment of invader species and exposes peat to erosion (Cronk and Fuller, 1995; Nakamura *et al*., 2004). Vegetation change has been ongoing in eastern Lesotho. This change has often been attributed to climate change, land use and overgrazing (Jacot-Guillarmod, 1969; Van Zinderen Bakker, 1981; Morris *et al*., 1993). According to Van Zinderen Bakker (1981), more than 50% of the alpine area has been occupied by invading Karoo shrubs (i.e. *Chrysocoma ciliata*) and between 49% - 65% of the alpine mires in the Senqu basin have species indicative of disruption (Schwabe, 1989). These values could possibly be higher at present, given population and livestock increases during the last 15 years. As the encroachment of Karoo species such as *C. ciliata* continues, the carrying capacity of the local vegetation will further decrease (Kakonge, 2002). Apparently, *C. ciliata* provides much less biomass (280.1 kg/TUF/Ha) than the local grasses (+1875.7 kg/TUF/Ha) (Schwabe and Whyte, 1993).

The bright green foliage of the wetland areas clearly stands out from the surrounding valley floors, which is primarily composed of low-lying angiosperms, bryophytes and

some algae (Jacot-Guillarmod, 1969; Killick, 1990; Pooley, 1998). Mires are extremely rare to Africa (Hughes, 1996), so the structure and composition of those in the alpine regions of Lesotho are unique (Jacot-Guillarmod, 1969; Backéus, 1988; Schwabe *et al*., 1989; Morris *et al*., 1993). Aside from the endemic angiosperms (30%) that occur in the region (Marake, 1998), many of the non-endemic wetland species are restricted to the alpine zone of the southern African mountains (e.g. *H. nervosa*), or are only found in southern Africa (e.g. *T. burchellianum*) (van Zinderen Bakker and Werger, 1974).

Vegetation patterns reveal the strong impacts of topographically induced local conditions affecting soil moisture, insolation and temperature. As many of the plants have similar adaptive characteristics to cope with the cold conditions (Williams, 1988), community structure tends to be simplistic. At least three main communities are distinct within the valley heads, with the term community being used to describe groups of "species with a similar distribution" (Morris *et al*., 1993, p48).

The wetlands within each valley head are grazed by cattle, horse, sheep, goats and native fauna (e.g. ice rats). Regardless of the incline of the wetland, only the gullied wetlands have rills and drainage lines across the wetlands. Only the gullied wetlands have 'dry islands' and only the gullied wetlands have a grazing post directly above them (Plate 6.1). By examining the vegetation of the valley heads and the patterns that are associated with their communities, it is hoped that the differences between gullied and non-gullied wetlands will be determined

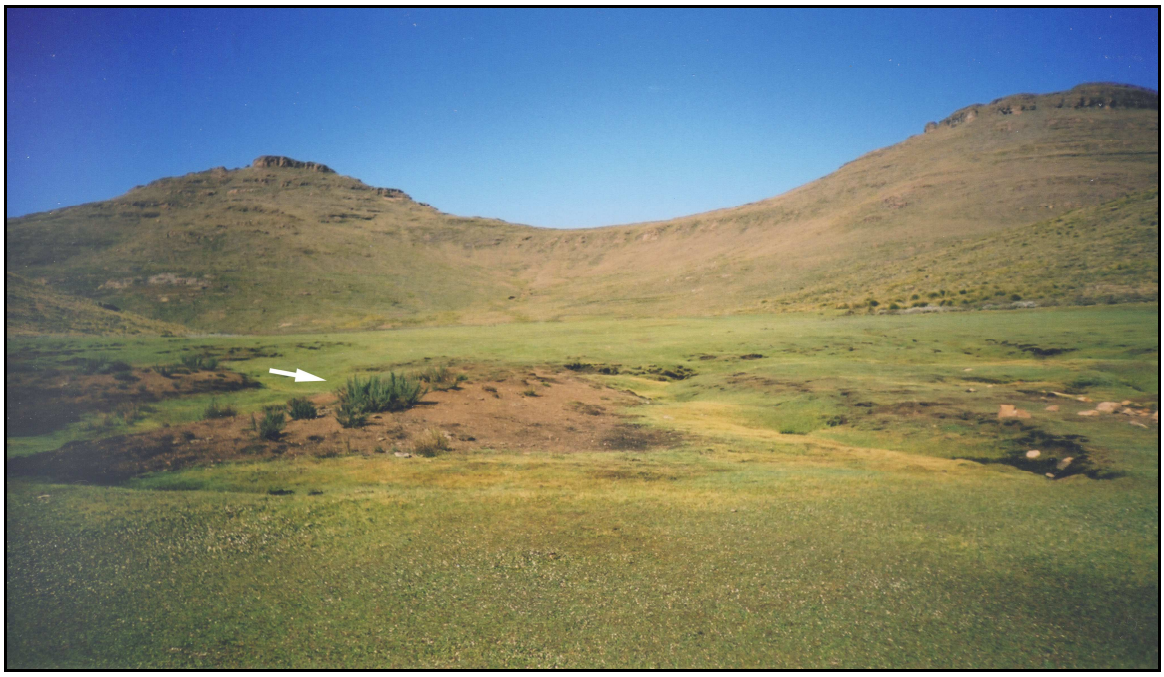


Plate 6.1 Dry island in the centre of the mire (shown by arrow). Shrubs next to arrow are ca. 0.5m tall.

6.4 Patterns Along The Vegetation Transects

Thirty-four plant species were collected and identified in total along the belt transects from each valley head (Table 6.1). The transect in valley head 1 contained 24 species, valley head 2 had 21 species, valley head 3 had 16 species, valley head 4 had 22 species and valley head 5 had 13 species.

The majority of the belt transect data were collected in late November 2001. The remaining data were collected in January 2002, when some of the previously identified plant species were no longer in flower (e.g. *Haplocarpha nervosa*). Yet, other species such as *Cotula paludosa* were observed in flower for the first time. It was observed that the location of plant communities along each transect is primarily influenced by the soil moisture gradient.

Table 6.1 Plant species noted along the belt transects (gullied wetlands are indicated in bold).

	Valley 1	Valley 2	**Valley 3**	**Valley 4**	Valley 5
Wetland species					
Alepidea ± pusilla	✓	✓			
Aponogetum sp.	✓				
Athospermum basutocum		✓	✓	✓	
Cotula paludosa				✓	✓
Haplocarpha nervosa	✓	✓	✓	✓	✓
Limosella major	✓	✓	✓	✓	✓
Lobelia ± filiformis	✓	✓	✓	✓	✓
Oxalis obliquifolia					✓
Polygala amatymbica		✓		✓	
Ranunculus meyeri	✓	✓	✓	✓	✓
Ranunculus multifidus		✓	✓	✓	
Rhodohypoxis rubella	✓			✓	✓
Saniella verna	✓	✓			
Sebaea filiformis	✓				
Trifolium burchellianum	✓	✓		✓	✓
Wurmbea elatior	✓	✓			
Dryland herbaceous species					
Alchemilla woodii	✓	✓			✓
Centella asiatica			✓	✓	
Crassula sp.	✓				
Delosperma sp.	✓				
Erica dominans	✓	✓	✓	✓	
Geranium multisectum	✓	✓	✓	✓	✓
Helichrysum aureum var. aureum	✓				
Helichrysum flanaganii	✓	✓	✓	✓	
Helichrysum milfordiae	✓				
Helichrysum sessiliodes	✓	✓			
Helichrysum subglomeratum	✓	✓	✓	✓	✓
Shrubs and Grasses					
Senecio inaequadens				✓	✓
Chrysocoma ciliata			✓	✓	
Helichrysum trilineatum	✓	✓	✓	✓	
Selago flanaganii		✓	✓	✓	✓
Eumorphia sericea	✓	✓	✓	✓	
Kniphofia caulescens			✓	✓	
Merxmuellera drakensbergensis	✓	✓		✓	

Three graminoid (gc) and one robust emergent (re) dominant plant community types were noticeable along the belt transects (Table 6.2). Plant location and community change were concurrent with changes to the surface soil moisture gradient (Figures 6.1 – 6.5).

Table 6.2 The four general plant communities, following the surface soil moisture gradient.

Vegetation communities	Surface soil condition	Theta meter reading
Mire	wet - saturated	0.5 - 1 $m^3.m^{-3}$
Mire margin / dryland herbs	moist	0.35 - 0.5 $m^3.m^{-3}$
K. caulescens	dry - moist	0.2 - 0.5 $m^3.m^{-3}$
Shrubs & herb species	very dry - dry	0 – 0.2 $m^3.m^{-3}$

Figures 6.1 – 6.5 indicate that there is a clear separation along the soil moisture gradient into these four broad, partially overlapping, plant communities (i.e. wetland, dryland, shrubs and *K. caulescens*).

6.4.1 Upland / Dryland Vegetation Communities

The *dryland community* (a term used by Schwabe, 1989) consists of grasses, shrubs and some low growing perennial herbs primarily on the mineral soil, and is divided into three groups that loosely follow a topographically controlled divide. These groups occasionally overlap and include: 1) those species that are dominant at higher altitudes as well as along the valley sides; 2) the shrubs and dryland herbs on the main, dry areas of the valley head floors, and 3) the dryland species along the valley floor mire margins.

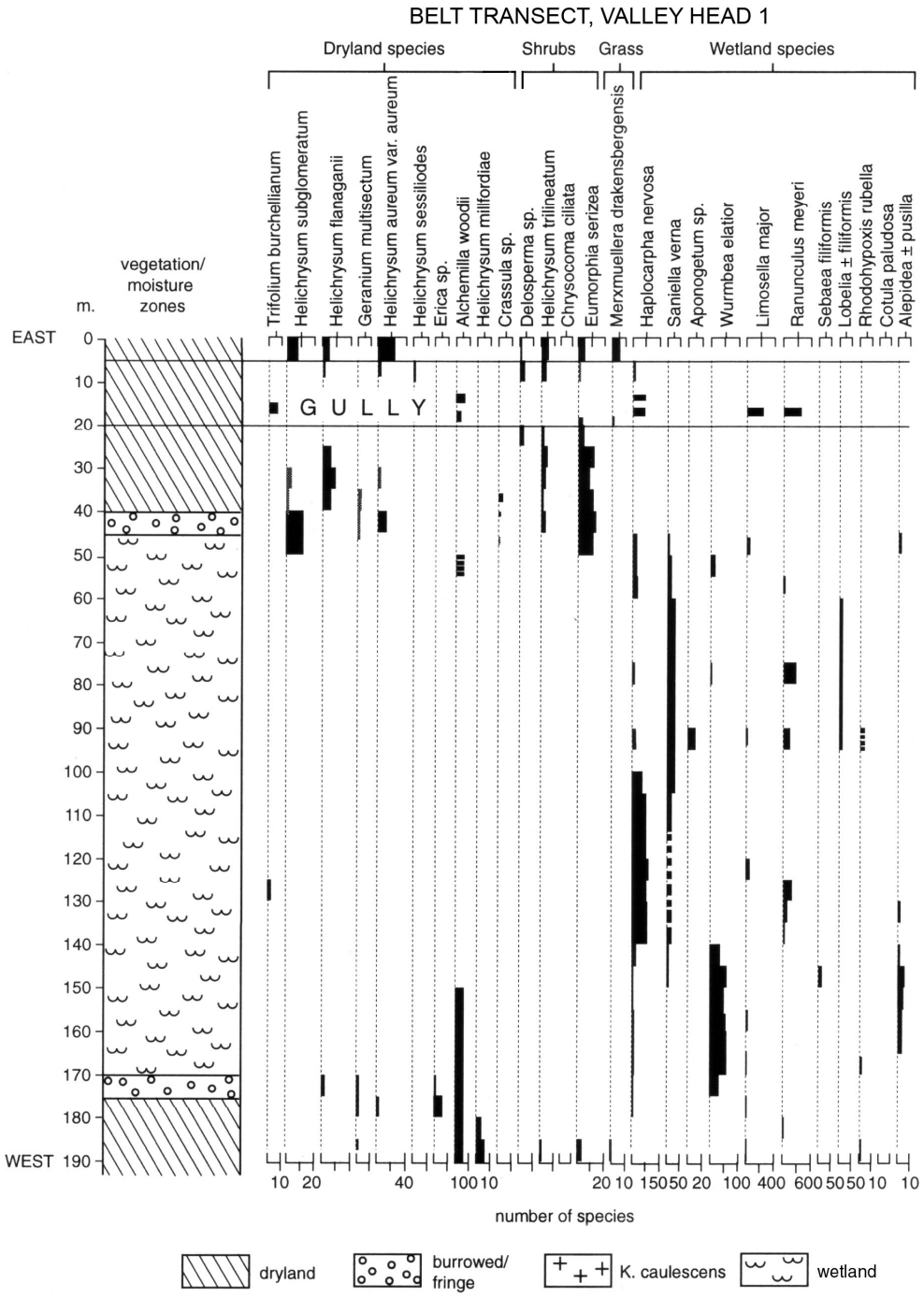


Figure 6.1 Vegetation belt transect within valley head 1.

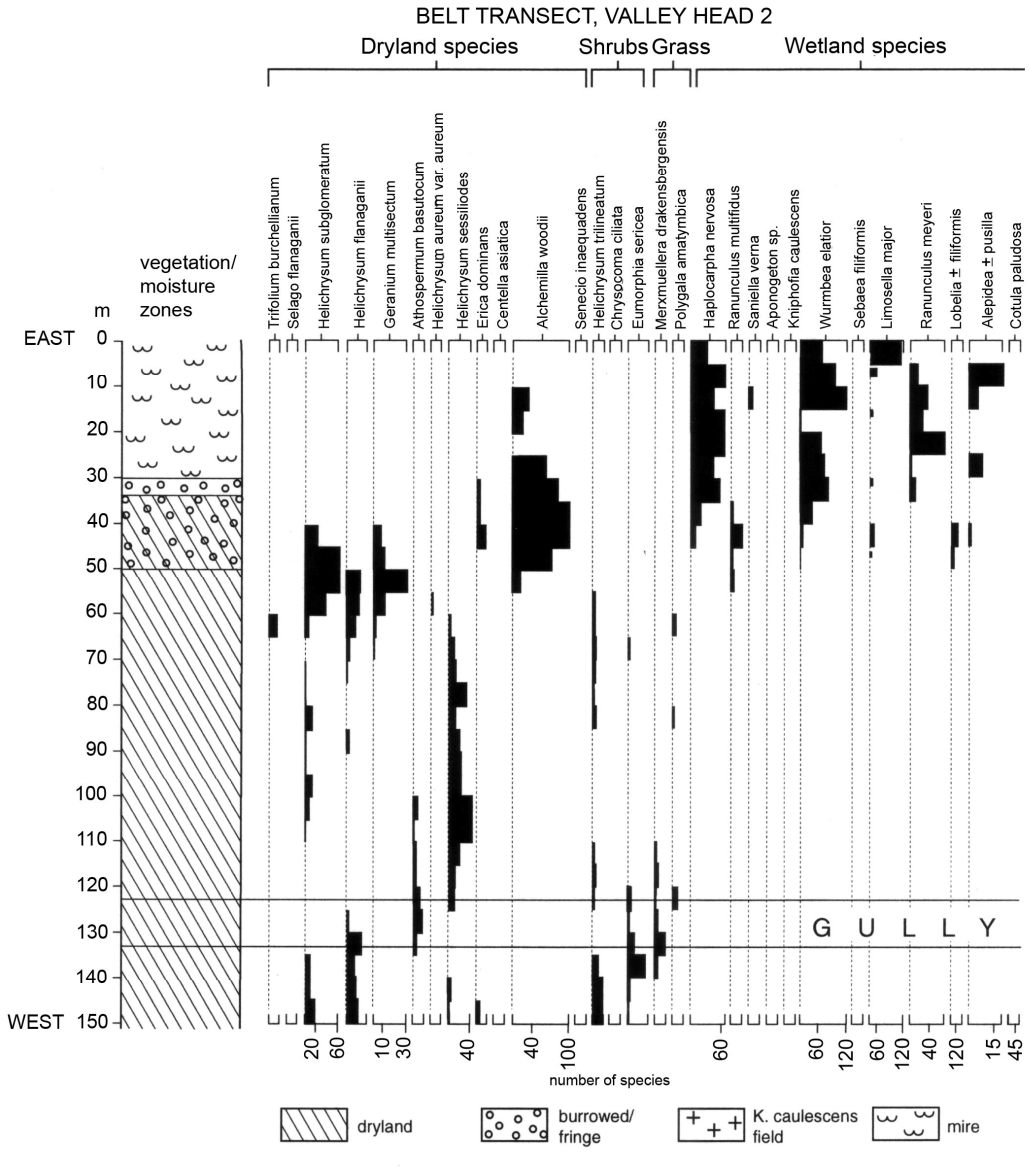


Figure 6.2 Vegetation belt transect within valley head 2.

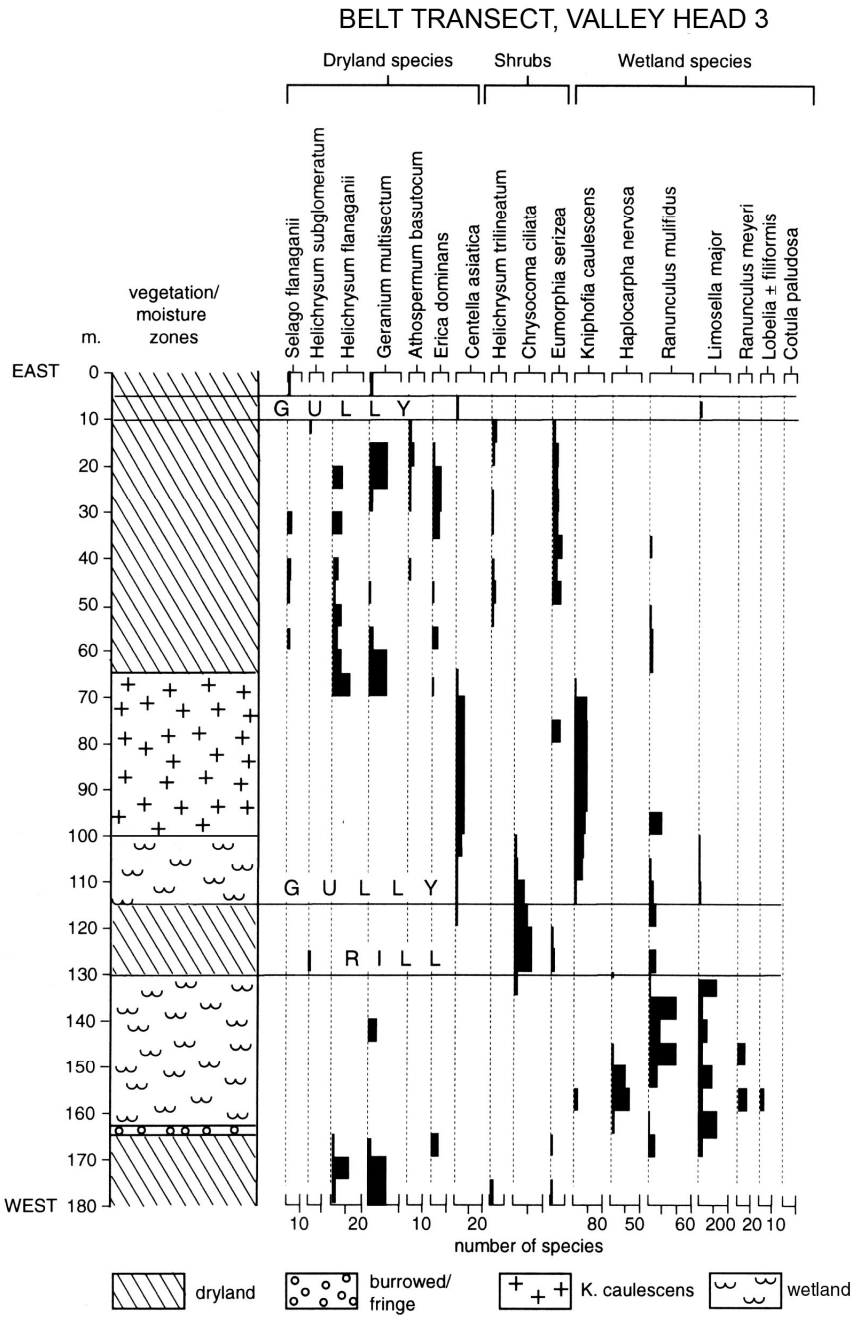


Figure 6.3 Vegetation belt transect within valley head 3.

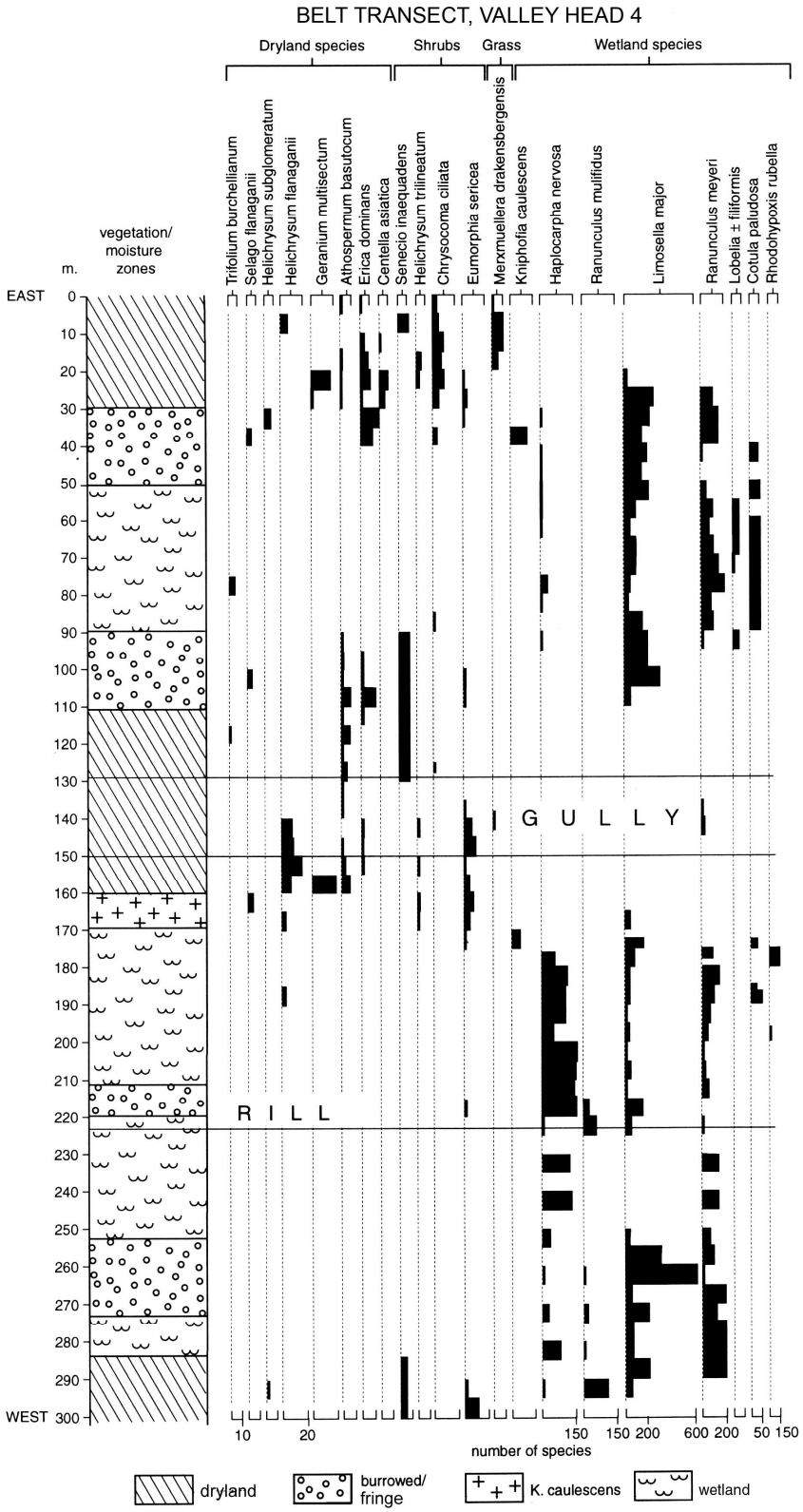


Figure 6.4 Vegetation belt transect within valley head 4.

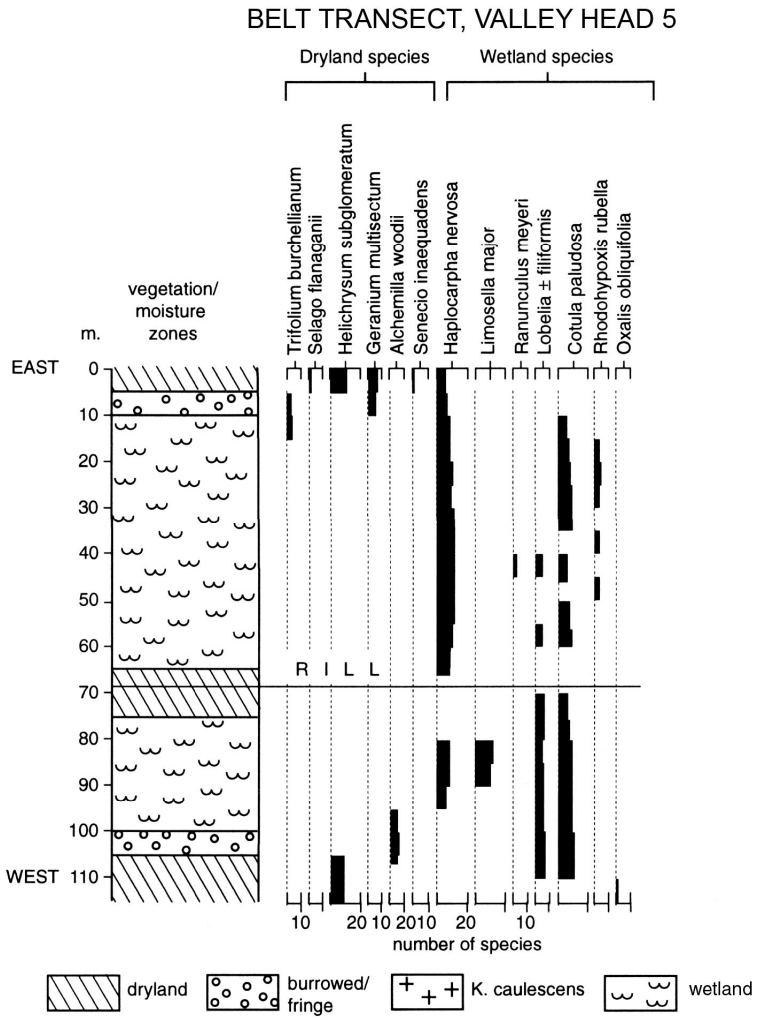


Figure 6.5 Vegetation belt transect within valley head 5.

6.4.1.1 Downslope Vegetation Transects

Vegetation surveys along the mountain plateaus as well as down the valley sides (visible in longitudinal belt transects conducted down valley head 5 in November 2002 and June 2003) reveal that only small herbs and weeds (i.e. *Berkheya multijuga*) grow within these rock-dominated areas (Figure 6.6).

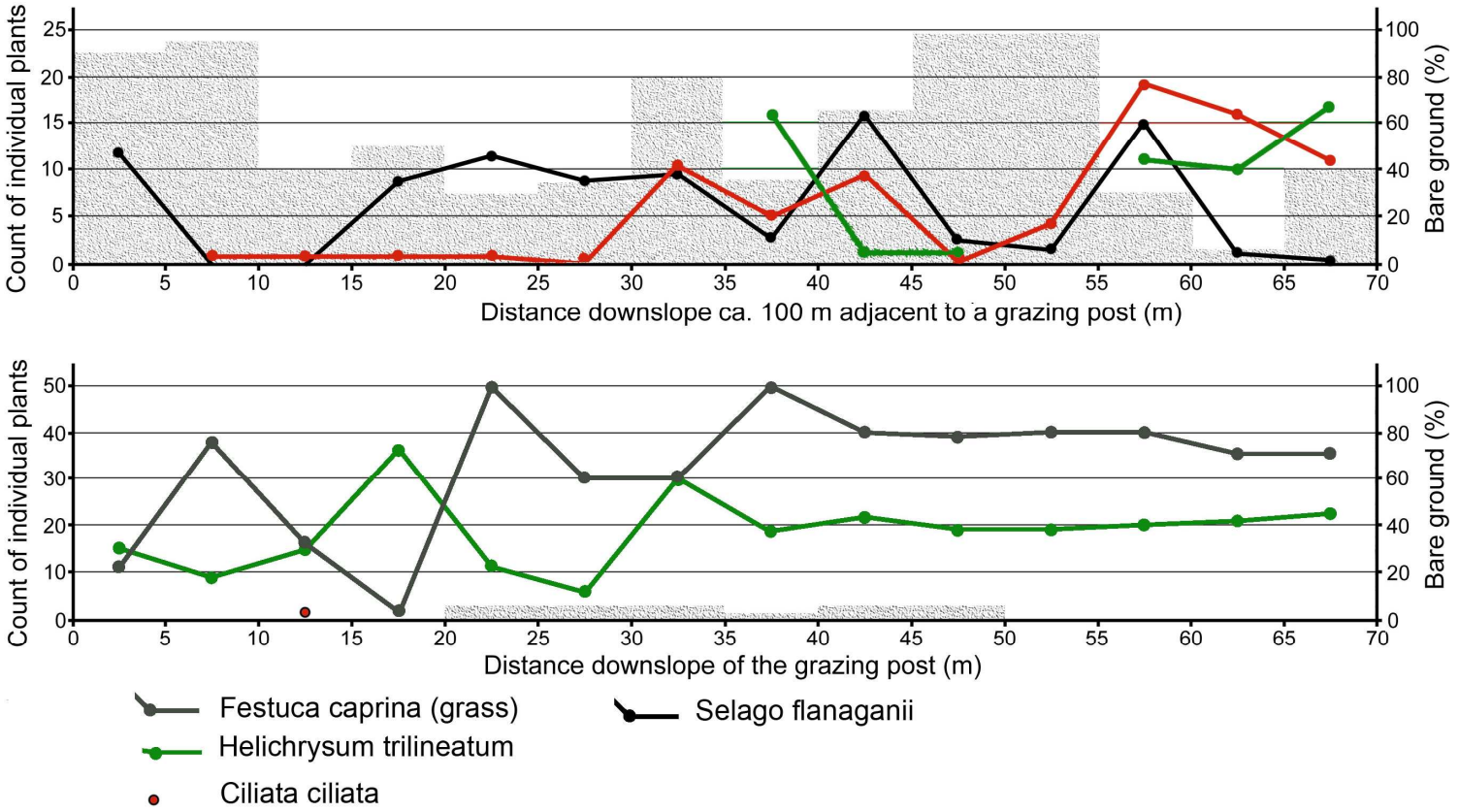


Figure 6.6 The top graph indicates a transect below a grazing post, whilst the lower graph indicates vegetation cover and composition ca. 100 m adjacent to a grazing post. The shaded background represents the percentage of bare ground.

With a decrease in elevation and an increase in protection, plant species such as *Selago flanaganii* and *Helichrysum trilineatum* occur more frequently and are also more robust. There is also a considerable difference in ground cover characteristics below grazing posts and on adjacent slopes next to such 'motebos' (Figure 6.6). Below the grazing post, the transect indicates a much greater area of exposed rock and/or bare soil as opposed to adjacent slopes (Plate 6.2). Vegetation cover was less than 30% on the steeper upslope portions of the valley head wall, in line with the grazing post transects. The invader species, *Chrysocoma ciliata*, is also present at sites surrounding grazing posts but absent on the adjacent slopes. Plate 6.3 demonstrates the dark green foliage of *C. ciliata* dominating the landscape around a grazing post located to the south of the study area. Adjacent slopes away from grazing posts host an almost 100% grass cover, especially the short curly grass, *Festuca caprina*. Shrubby vegetation is also more robust and abundant at higher elevations above the grazing posts. Generally, the lower slopes consist of a mix of grasses (*Merxmuellera drakensbergensis*), small herbs (*Helichrysum flanaganii*) and shrubs (*Helichrysum trilineatum*, *Eumorphia sericea* and *Chrysocoma ciliata*) (Plate 6.4). These shrubs were also observed on the valley head floor as well as within the gullied mires on isolated, gravely, dry islands and adjacent to the wetland gully edges (Figure 6.1 – 6.5).

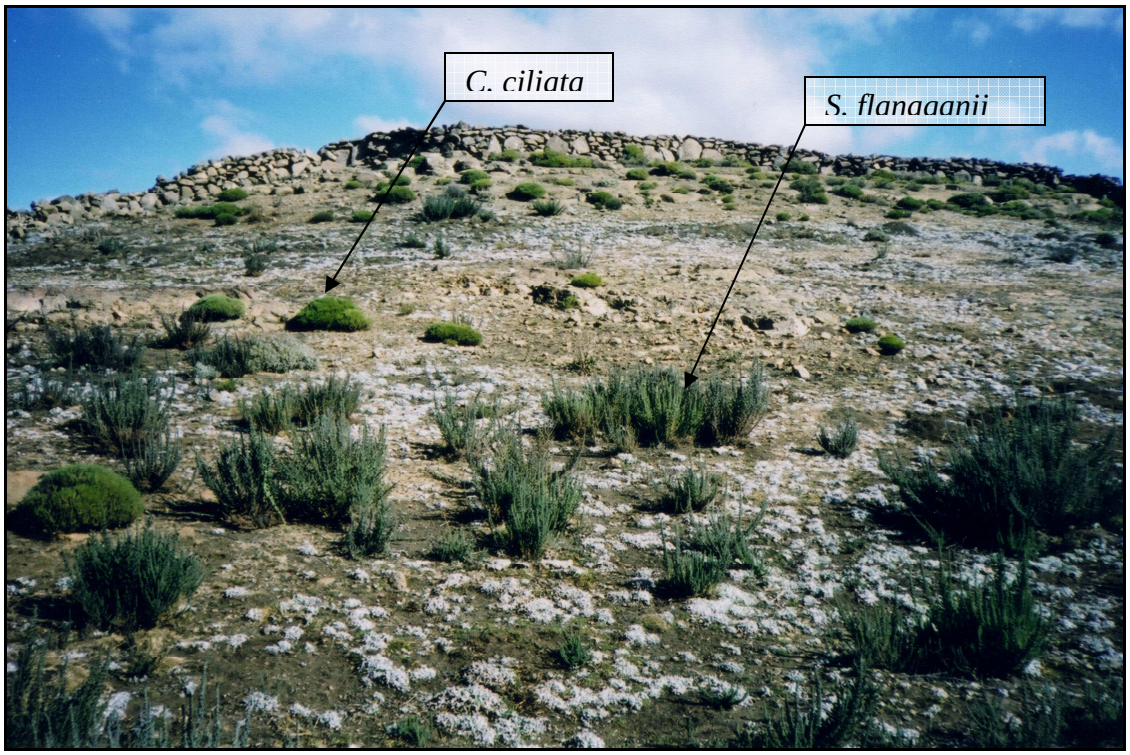


Plate 6.2 Looking upslope towards an abandoned grazing post. Notice the sparse vegetation cover and prevalence of *S. flanaganii* and *C. ciliata* (Bitter Bush) (November 2002).

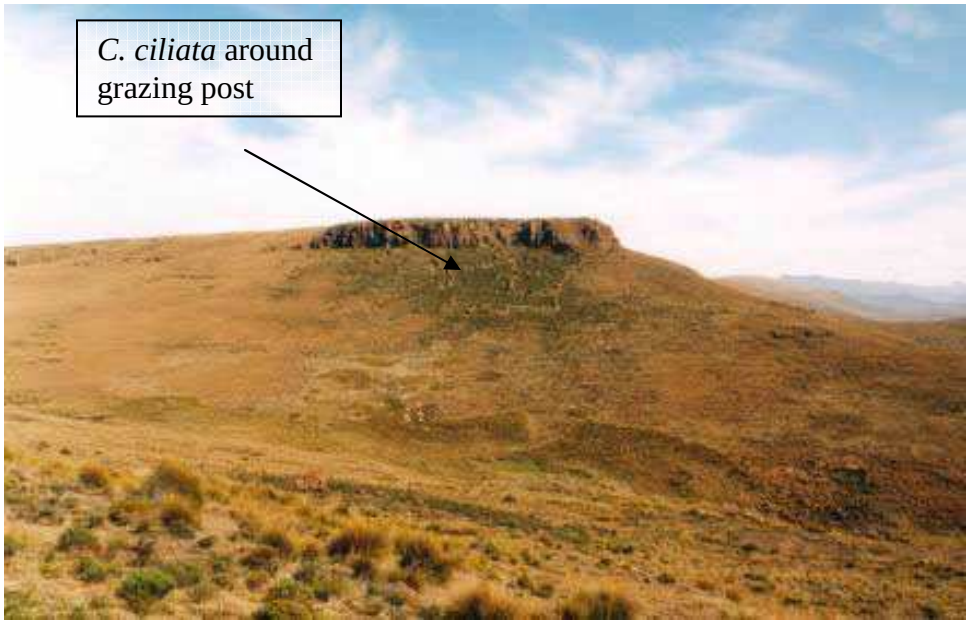


Plate 6.3 **N**otice the dark green hue created by *C. ciliata* around an abandoned grazing post below the basalt scarp (November 2001).



Plate 6.4 Two of the shrub species present within the gullied transects. *C. ciliata* is the darker green shrub to the left and top left; *Helichrysum trilineatum* is the lighter green shrub on the right. The inserted scale is in decimetres: the above shrubs are approximately 30 cm wide and are 50 cm in height (November 2002).

6.4.1.2 *Valley Floor Vegetation Communities*

The second group of dryland vegetation is that growing on the drier valley head floors, and is comprised mainly of the taller herbs, *Senecio inaequidens* and *Selago flanaganii*, the curly grass, *Festuca caprina*, and the small mat-forming herbs, *Helichrysum subglomeratum*, *H. flanaganii* and *H. aureum var aureum* (Plate 6.5). A slight overlap is noticeable between this community and those found on the slopes and wetland fringes.

6.4.1.3 *Mire Margin Vegetation Communities*

The mire margin is determined by decreasing soil moisture and organic matter percentage, and consists of small herbaceous species such as *Trifolium burchellianum* (wild clover), *Geranium multisectum*, *Erica dominans* and *Alchemilla woodii*. These



Plate 6.5 The dryland valley floor vegetation with the yellow-topped herb, *Helichrysum subglomeratum*, and the whiter herb, *H. flanaganii*. Use the inserted scale for approximate size of the herb immediately above it; scale bar is broken into two-centimetre segments (January 2002).



Plate 6.6 The transition from mire to dryland is distinct, with a transitional area evident by the presence of burrows and smaller herbaceous plants (January 2002).

species appear to indicate a boundary between the mire expanse (where soil is saturated) and the dryer mineral soil (Plate 6.6).

6.4.2 Wetland Communities

Within the non-gullied wetlands, the vegetation forms a continuous wet green turf or lawn that highlights the wetland area unmistakeably. None of the species growing within the wetland are listed as obligate hydrophytes (according to Killick, 1990; Pooley, 1998). However, most of the wetland species appear to be obligate to their surroundings. For instance, it is extremely rare to observe species such as *Saniella verna*, *Wurmbea elatior* and/or *Aponogetum sp.* growing beyond the wetland. *Ranunculus meyeri*, *Limosella major* and *Haplocarpha nervosa* are found in high cover-abundance (e.g. > 100/m^2) where surface water is present (Plate 6.7). However, *H. nervosa* is also recorded in areas that are wet to moist (3.5 – 5 $m^3.m^{-3}$ x 0.01). These findings are in contrast to published reports stating that the species prefers semi-aquatic conditions (Hobson *et al*., 1970). Wetland areas with slightly less moisture (e.g. on hummock apexes, edges of small rills and alongside burrow openings) frequently host *Lobelia filiformis*, *Ranunculus multifidus* and *R. meyeri* (+25%). The tiny pink *Rhodohypoxis rubella*, along with the purple and white petalled *Lobelia filiformis*, is also found in large numbers on hummock apexes and within undamaged wetland areas (see also Schwabe, 1989). In disturbed wetland areas, *Ranunculus multifidus* is dominant, as has also been observed by Schwabe (1989). *Haplocarpha nervosa* and *Limosella major* are the two most frequently recorded plants in the valley heads (ca. total of 8368 individuals along the transects). However, in terms of



Plate 6.7 View of the mire surface, which presents itself as a flowering lawn. The white flowers are *L. major*, and the yellow flowers are *R. multifidus* (November 2001).

percentage cover and visibility, the most visible wetland species are *Haplocarpha nervosa* and *Cotula paludosa* (80% - 100% cover or partial overlap within the saturated mire portion of transect).

It should be noted that many of the plant species are not found within each valley head and are not always found within their expected habitats. In particular, the small yellow flower, *Cotula paludosa* and the dark green shrub, *Chrysocoma ciliata*, are sometimes completely absent from one valley head floor, yet abundant in another.

6.4.3 Relationship Between Soil Moisture And Vegetation

By observing the average surface soil moisture readings along each belt transect (Figures 6.1 – 6.5), it is evident that there is an association between the location of plant species and/or community types with particular soil moisture regimes. These relationships were rank correlated in order to summarise the data obtained. Wetland plant species correlate

positively and significantly with soil moisture over the entire transect (Spearman's rank correlation of 0.64 to 0.68; $P < 0.05$); meaning that as the mire surface changes from moist → wet → saturated, the number of recorded wetland species increases (Table 6.3). Specifically, *H. nervosa*, a dominant wetland species, is positively correlated (0.57 to 0.72, $P < 0.05$) with soil moisture along each vegetation transect (Table 6.3). Dryland species in most of the valley heads, except valley 2, correlate negatively with the soil moisture across the transect (-0.42 to -0.93, $P < 0.05$), whilst in valley 2 the relationship was slightly negative (-0.11, $P < 0.05$). In valley 2, a negative correlation for the dryland plant community did become more pronounced after small herbs were omitted from the calculation, which may reflect a strong preponderance of small dryland herbs within the wetland area. Within each valley head (except valley head 2), shrub species, dryland herb species and dryland plant communities had significant negative relationships with the soil moisture gradient along the entire transects (see Table 6.3).

Table 6.3 Calculated values for correlations of individual plant species and plant communities with surface soil moisture along the entire transect. * = Not statistically significant at the 0.05 level.

	Valley 1 N = 38	Valley 2 N = 30	Valley 3 N = 37	Valley 4 N =60
H. nervosa	0.67	0.72	0.57	0.66
H. trilineatum	-0.98	-0.45	-0.54	-0.39
E. sericea	-0.89	-0.5	-0.21*	-0.68
C. ciliata	n/a	n/a	-0.42	-0.4
Only shrubs	-0.88	-0.74	-0.43	-0.62
Only herbs	-0.87	-0.2*	-0.31*	-0.57
Wetland species	0.64	0.83	0.62	0.68
Dryland	-0.93	-0.11*	-0.42	-0.73
L. major & *H. nervosa*	0.04*	0.58	0.68	0.28

6.4.4 Vegetation Range Overlap

The response of many of the plant species to the overall environmental gradient was asymmetrical and unimodal in pattern (Figures 6.1 – 6.5). Apparently, such a skewed density curve for a single species is common (Dale, 1999). However, some species (e.g. *Eumorphia sericea* in valley head 1) had bimodal responses. In order to examine the causal factors determining the composition of the vegetational communities along the vegetation transects, it was useful to study their arrangement along a direct gradient (i.e. soil moisture) (Dale, 1999). Since a strong rank correlation only exists between the various vegetation communities in relation to soil moisture, and not with either soil pH or organic matter content, the gradient referred to in this context is thus of surface soil moisture. By examining the arrangement of the species within the specified ranges (i.e. wetland versus dryland etc.), community range dynamics can be assessed by specifically measuring species overlap and the placement of boundaries. Qualitatively, the communities found along each vegetation transect tend to replace each other with no intermediate zone of coexistence. However, mire or fringe communities often have competing individual species that coexist until completely replaced by another community. For instance the fringe species, *A. woodii* in valley head 1, extends from the fringe community and overlaps into the wetland community by ca. 10 m. Apart from community on community dynamics, there is also a spatial pattern of inter-community competition, where species either completely or partially cohabitate with all or some of the other competing species of the same community (Figures 6.1 – 6.5). The individual species within the wetland communities of valley heads 2 to 5 all partially overlap with one another, whilst the mire species in valley head 1 do not coexist, but rather occupy

discrete patches. The arrangement of each species and the way it coexists with other species could be quantitatively compared to a null hypothesis. Calculations were prepared by counting the number of pairs of each species that fell within each of three designated classes: 0) no overlap, 1) partial overlap and 2) complete overlap (according to Dale, 1999). From this calculation, the actual pairings are compared to the expected number of pairs stated in the null hypotheses, which in this context assumes that the order of events, and beginnings and endings of each species range is random (Dale, 1999) (Table 6.4). Based on the null hypothesis, the results for each transect reveal that the actual number of pairs of ranges that partially overlap, is surprisingly lower than expected for most valley heads (except for valley head 2). The actual number of pairs with no overlap was calculated as being higher than expected for each valley head, except for valley head 3 (Table 6.4). For all of these cases, the null hypothesis can be rejected.

Table 6.4 The calculated means of pairs of species ranges in each category of overlap for each valley head is found in the second column. The first number = the expected number of non-overlapping pairs, the second = the expected number of partially overlapping pairs and the third = the expected number of pairs whose ranges completely overlap. The actual numbers of occurrences are in the third column.

	Hypothesis	Actual
Valley 1	92 / 92 / 92	167/ 36.5 / 72
Valley 2	70 / 70 / 70	106 / 48 / 56
Valley 3	40 / 40 / 40	65 / 27.5 / 27.5
Valley 4	70 / 70 / 70	126 / 31 / 53
Valley 5	26 / 26 / 26	45 / 6 / 27

6.4.5 Vegetation Patterns Within The Gullied Wetlands

Gully erosion through the central portion of a mire in valley head 4 provides an opportunity to establish the possible impacts of gully erosion on the surrounding vegetation. The transect crosses the gully perpendicularly and thus captures gradient

changes moving towards and away from the gully. The location of shrubs and dryland herb species in the fringe areas and within the wetland area, is associated with erosion in each case (Figures 6.3 and 6.4). The disturbance is either in the form of burrows (ca. > $2/m^2$) or the existence of a gully and/or rill that has cut through the mire. As indicated in Table 6.1, the gullied valley heads have a greater number of shrub species than the non-gullied valley heads. In particular, *C. ciliata* is only present along the transects in the gullied valley heads. Shrubs, exposed soil and herbaceous plants are dominant for a distance of at least ca. 15 m on either side of the wetland gullies. Soil moisture content was also the lowest adjacent to the gully edges (i.e. 'dry' to 'very dry'). With increasing distance from the gully, soil moisture content increased to 'moist' and 'wet' levels. However, within the non-gullied mire expanses, the soil was saturated across most parts of the wetland. Along the vegetation and soil moisture transects in valley head 4, rank correlation indicates that the two variables share a significant relationship. Thus, with increasing distance from the gully edge, as the surface soil moisture increases, so the wetland vegetation cover increases (0.61, n = 237; 0.45, n = 53; $P<0.05$) (Table 6.5).

Table 6.5 Spearman's rank correlations: various variables with distance from the gully edge, valley head 4. SM = soil moisture.

Mire in valley head 4	Correlations with distance from the gully
Dryland species	-0.48 (n = 53)
Wetland species	0.45 (n = 53)
Shrubs	-0.21 (n = 53)
Soil moisture	0.61 (n = 237)
Burrows	-0.07 (n = 237)
Burrows & SM	-0.39 (n = 237)

Within the wetland, dryland plant species and wetland plant species have inverse relationships with distance from the wetland gully edge (between –0.48 and +0.45) (as

shown in Table 6.5). The results indicate that gully erosion is affecting the soil moisture across the wetland to some extent, which consequently influences the location and occurrence of wetland vegetation and the invasion of dryland herbs. However, it should be recognized that soil moisture and the establishment of invader species within the mire are also affected by other environmental variables such as ice rat burrowing and micro-topography, hence not permitting a stronger correlation with distance from the gully.

6.5 Discussion And Summary

The relationship between the vegetation of a region and the soil it inhabits is a complex system of interrelated variables. Vegetation is dependent on soil and soil is dependent on vegetation type and density (Rowntree, 1988). In the Sani study, the variables that impact the health and location of the plant species (i.e. climate, insolation and aspect) are similar for all the valley heads due to their proximity to each other. Thus it is reasonable to assume that the valley heads have undergone parallel developmental histories. Following this, description of vegetation type and location within the valley heads from various locations within the Drakensberg and Maluti mountains is consistent amongst authors (e.g. van Zinderen Bakker and Werger, 1974; van Zinderen Bakker, 1981; Morris *et al*., 1993; Pooley, 1998).

Vegetation patterns were recorded across the valley head floors, the mire expanses, and along slope profiles to the mountain summits. Species such as, *R. meyeri*, *L. major* and *H. nervosa* closely reflect changes in soil moisture, soil type and position across the mire. *R. meyeri* and *L. major* grow in zones of runoff or around pools, whilst *H. nervosa* is most

dominant in the undamaged, wetter areas of the wetlands. Fringe species and dwarf shrubs and grasses such as *H. flanaganii*, *H. trilineatum*, *E. sericea* and *S. inaequadens*, were recorded along the mire margins. These shrubs and grasses have an approximate aerial cover of 30%. On the slopes, the coverage of dwarf shrubs and grasses increases to 75% -90%. Once on the plateau however, shrub coverage decreases to ca. 30% (see also van Zinderen Bakker and Werger, 1974) and low herbaceous species such as *Helichrysum subglomeratum* increase in number. Below mountain summits, species most associated with the mountaintops often overlap with species common to the slopes, such as *Festuca sp*., *H. trilineatum* and *Delosperma sp*.

For all the valley heads investigated, the wetland vegetation has an ecotonal transition with the dryland communities (c.f. Kent *et al*., 1997). However, within the gullied mires, the wetland vegetation becomes degraded due to a lowering of the water table (see also Thornes, 1985). Slack *et al*. (1980) demonstrated that the location of wetland species in particular reflect the importance of the water level gradient in fens of western Alberta, Canada. Even though the water was constricted by slope gradient, wetland surfaces with zero to minimal destruction of the underlying peat had a water table just below the surface (Slack *et al*., 1980).

The impact that gullying has had on the ecohydrological regime of the valley heads is evident and portrayed by the vegetation types and densities found along the gully edges and interfluves (Figure 6.1– 6.5). In a wetland landscape, water level and supply are the main determining factors for the location and survival of plant species (Wassen *et al*.,

1996; Hughes, 1997). In valley heads 3 and 4, which both have disturbed and damaged wetland areas, the disturbance is illustrated by the uncharacteristic location of some plant species, particularly when compared with the plant composition in valley heads 1, 2 and 5. Species such as *Alchemilla woodii*, *Geranium multisectum* and *Helichrysum subglomeratum* usually occupy the periphery of an undamaged wetland, as recorded in valley heads 1 and 2. However, within the wetland expanse of valley heads 3 and 4, these 'fringe species', together with species indicative of disturbance (e.g. *S. inaequadens*, *C. ciliata* and *Rumex*), are found around locally dryer areas such as at burrow exits, along rill edges and within desiccated inter-drainage areas. The spatial patterning of these plants reflects the greater mosaic of the soil moisture gradients in these damaged mires.

6.5.1 Summary

Non-gullied valley head wetlands have a continuous green lawn consisting of typical wetland plant species. Other than the occasional small herbaceous species in areas with a lowered water table, there is no encroachment by species normally associated with disturbance. The mire expanse is also completely saturated and has relatively few *Otomys* burrows. In contrast, the gullied mires have a dry to very dry soil moisture regime adjacent to the gully edges, whilst the mire expanse is encroached by dryland species and shrubs. Areas between gully arms (inter-drainage areas) are usually bare and dry. Gully erosion thus reduces the soil moisture available to the mire and is a primary factor contributing to the reduction of wetland plant species abundance across the mire surface.

Chapter 7

Discussion and Conclusion

7.1 The State Of Wetlands

Since the 1900's, more than half of the total global wetlands have been destroyed (Schuyt, 2005) or altered (Kennedy *et al*., 2003). As outlined in previous chapters, research has identified that wetlands have important functions such as soil stabilization, water filtration, flood reduction, organic carbon retention, groundwater recharge and regulating discharge (Schwabe, 1995; Middleton, 2002; Nel, 2003). Wetlands are also valuable for agriculture, recreation, scientific research, and are important ecosystems with tremendous biodiversity (Mitsch and Gosselink, 1986; Mulamoottil *et al*., 1994; Nel, 2003). Despite the attention wetlands have received, communities have become endangered, whilst much faunal extinction has resulted from wetland habitat destruction (Mulamoottil *et al*., 1994; Kennedy *et al*., 2003).

In 1971, in the city of Ramsar, Iran, the first international treaty was signed for the protection, conservation and wise-use of wetlands and their resources (Ramsar, 2004). Before being added onto the list of internationally important wetlands, each wetland site must meet at least one of the criteria set forth by the convention, which is based on whether it is a rare wetland type, or

whether the wetland supports valuable, threatened or unique species of ecological communities (Ramsar, 2004). By December 2003, 138 countries had listed a total of 1328 wetlands on the Ramsar List of Wetlands of International Importance (Ramsar, 2004). Presently, Africa has 74 wetland sites listed, of which 17 are located in South Africa. A wetland at Lets'eng-la-Letsie, in the province of Quthing (Lesotho) was added to the list on 1st July 2004.

A large number of African wetlands are in good condition compared to those in North America and Europe. However, as population increases, the pressure to exploit these wetland areas also increases (Schuyt, 2005). Wet soils are particularly vulnerable to trampling and compaction, especially if they are peaty (Bower, 1962; Trimble, 1988). The key component to the type and process formation of a particular wetland is its hydrology (Mulamoottil *et al*., 1994), therefore disturbances or changes to hydrology can lead to biological, physical and/or chemical changes that can negatively impact downstream processes and hydrological conditions (Haslam, 2003). The effect of grazing animals is often considered to be a major cause for vegetation cover change, the consequent deterioration of topsoil and disturbance to wetland hydrology. Trampling by grazing animals also directly provides a ready source of transportable material to wind and water action (Evans, 1998). As livestock graze, the shortened turf becomes weakened and susceptible to 'poaching' by hooves (Tasker, 1980; Davies, 1985; Brooks and Stoneman, 1997) and further damage by rodent burrows (Evans, 1998).

External disturbances to sensitive ecosystems, such as alpine mires, can result in severe changes and eventual system degradation (Hüttle and Gerwin, 2004; Rodríguez *et al*., 2005). This type of externally induced degradation directly decreases soil quality, and in terms of succession, it can consequently lead to ecological regression (Rodríguez *et al*., 2005). The

indirect and direct effects that gully erosion has on the adjacent soil properties, as well as on the vegetation community dynamics within these valley heads, has been highlighted in this study. Once the wetland surface has been degraded by soil erosion and the peat exposed through cracks (hags), such surfaces become more susceptible to the forces of wind, water and frost heave during winter (Bower, 1962; Selkirk and Saffigna, 1999; Grab and Deschamps, 2004). Consequently, the exposed peat desiccates and becomes oxidized and ever more susceptible to erosion (Schwabe, 1995). Eroded mires are more easily drained and more likely to undergo vegetation changes than intact mires (Bower, 1962). Some authors believe that wetlands and peat are able to self-heal, which is demonstrated by bands of silt and peat within soil profiles (McFarlane and Whitlow, 1990). It has been postulated that peat may become unstable at a particular thickness, since gullies tend to form where the wetland is deepest (Bower, 1962). The current study is not about normal rates of erosion, but rather is focused on accelerated erosion, which occurs in response to anthropogenic impacts. Biotic and human induced changes can trigger dramatic changes in vegetation type and cover, cause peat erosion and initiate a loss of sediment and wetland area (Bower, 1962; Smith, 1982; Skovlin, 1984). Fauna (ice rats and moles) and flora (Karoo shrubs) then invade the desiccated wetland areas and perpetuate the cycle (Schwabe, 1995; Grab and Morris, 1999).

Mountain watersheds have always been of global concern because of their important and sensitive function. Despite conservation efforts, mountain wetlands remain vulnerable to degradation; this thus holds considerable environmental implications for the surrounding region (Paudel and Thapa, 2001). Since the Lesotho highlands area is hydraulically and economically important, the decrease in wetland area and change in wetland functioning should raise much concern (Schwabe, 1995; Grab and Morris, 1999; Nüsser and Grab, 2002).

The loss of wetlands represents a huge loss in species diversity through habitat destruction, the loss of water regulation, more frequent flash floods, the loss of nutrients, and most importantly, the loss of large tracts of valuable wetland and grassland (Schwabe, 1995). The larger mires with a more complete vegetation cover are better able to attenuate flow, prevent soil erosion and filter impurities. Yet, even in some of the larger wetlands, desiccation of peat results in the channelling of flow out of the basin, resulting in a large loss of sediment and water (Schwabe, 1995; Grab and Morris, 1999).

7.2 The Effect Of Gully Erosion On Specific Biophysical Characteristics Within Selected Eroded Mires

Two of the five alpine mires investigated in this study have been incised by deep discontinuous gullies (i.e. valley heads 3 and 4), whilst a third alpine mire has been eroded by a short, shallow gully system ca. 50 cm deep (i.e. in valley head 5). These gullies tend to have U-shaped morphologies with active sidewall erosion and rill-abrupt channel heads. Wetland gully sidewalls within valley heads 3 and 4 indicate at least two pronounced phases of peat accumulation, alternating with thinner horizons of clastic sediment. The two thickest peat horizons in valley head 3 have been dated at ca. 4740 ± 60 BP (Marker, 1994, 1998) and ca. 13 490 ± 130 BP (Marker, 1994).

Gully erosion is a well-studied subject, with researchers having examined the causes, consequences and management of this widespread geomorphic phenomenon (e.g. Heede, 1976; Ebisemiju and Ekiti, 1989; de Oliveira, 1997; Talling and Sowter, 1999). Research has shown that gully erosion is a reaction within the landscape to instability or exceeded sensitivity caused

by an extrinsic (i.e. climate, land use) or intrinsic (i.e. slope aggradation) change in the system (Rowntree, 1988; Allison and Thomas, 1993; Billi and Dramis, 2003). However, the actual effect of gully erosion on the natural landscape is still a somewhat understudied component of gully erosion research. A consequence of gully erosion is the increased connectivity of the drainage system and a corresponding increase in drainage density (Leopold *et al*., 1964; Bradford and Piest, 1980; Harvey, 1996; Poesen *et al*., 2003). It is also accepted that gully erosion involves the direct destruction and removal of soil, which is a valuable natural resource (Ebisemiju and Ekiti, 1989; Rienks *et al*., 2000; Billi and Dramis, 2003).

An objective of this study was to quantify the environmental effects of gully erosion on several alpine mire ecosystems within the Sani Valley region of eastern Lesotho. Three hypotheses were presented in section 1.8. These hypotheses are based on the central assumption that gully erosion degrades mire ecosystems in the eastern Lesotho Highlands (Schwabe, 1989; Grab and Deschamps, 2004). Below, are summaries based on conclusions formed in chapters 5 and 6, of the effect of gully erosion on three main valley head bio-ecological components, i.e. soil moisture, vegetation composition and soil properties.

7.2.1 Summary: Gully Effects On Soil Moisture

Gully erosion impacts the adjacent soil through sediment removal and the draining and desiccation of at least the upper soil horizons (Grab and Deschamps, 2004). The loss of topsoil moisture within these mires, as a consequence of gully erosion, diversifies the micro-topographical and hydrological environment and creates a semi-arid microhabitat suitable for dryland shrubs (Grab *et al*., 2005a). The present study has demonstrated that gully erosion

affects the soil moisture gradient for several metres adjacent to the gullies. In fact, the soil moisture transects revealed that the soil was driest at gully edges and that moisture content increased with distance away from the gully edge, yet not necessarily in a linear direction (Figures 5.14 – 5.17). The soil moisture in valley heads 3 and 4 indicated a significant correlation with distance from the gully (0.53 and 0.61, $P<0.05$, respectively). The study has demonstrated through a series of comparative soil moisture transects that non-gullied mires have consistently saturated moisture levels across their surfaces, whilst gullied mires have irregular and lower soil moisture values (Figures 5.7 – 5.17), as was also observed in Lesotho by Hanvey and Marker (1994). This pattern of non-uniformity across the mire expanse, particularly areas that are relatively 'dry', always coincide with areas of erosion and/or cover change (i.e. burrows, rills and/or gullies) (Grab and Deschamps, 2004; Grab *et al*., 2005b).

7.2.2 Summary: Gully Effects On Vegetation Distribution And Composition

Soil moisture is considered to be the primary controlling factor affecting plant growth, and the deciding variable for vegetation community patterns (Hall and Lamont, 2003). In fact, the spatial mosaic of soil moisture reflects the spatial distribution of the soil physical properties and other variables such as micro-topography (Fitzjohn *et al*., 1998). The lower soil moisture levels adjacent to the gullies thus affect vegetation community patterns. Unlike the gullied mires, the non-gullied mire expanses are not occupied by shrubs and large numbers of dryland herbaceous species, and do not have exposed areas of soil. The environmentally intact parts of the mires are dominated by the occurrence of *Ranunculus meyeri*, *Limosella major*, *Haplocarpha nervosa* and *Cotula paludosa*. However, the gullied and eroded mires host dryland shrubs and have large areas of exposed soil. *Senecio inaequadens*, *Rumex* spp. and *C.*

ciliata, which are species indicative of disturbance, frequent the mire expanses in valley heads 3 and 4, yet are not recorded within the mires of valley heads 1, 2 and 5. The location of all plant species/communities relative to distance from the gully edges is found to be significant (ca. 0.45 for wetland plant species and –0.48 for dryland species, $P<0.05$). Thus, within the mire, dryland plant species are dominant towards the gullies whilst wetland plants occur in greater number with distance from the gullies. Strong relationships also occur between vegetation type and soil moisture across the valley head floors. Positive correlations are recorded between wetland plant species and soil moisture (r value = 0.64 to 0.83, $P<0.05$) whilst negative correlations occur between dryland plants and soil moisture (r value = -0.42 to –0.93, $P<0.05$) (Table 6.3).

It is important to consider the possibility that variables other than gullying had created favourable conditions for the dominance of dryland shrubs within the gullied mires where it is envisaged that more palatable wetland species had once occurred. A discussion on how this happens and the effects of this cycle are outlined in section 7.3. Given that the non-gullied wetlands do not show a trend of dry soil conditions and the presence of dryland shrubs, it is envisaged that the dryland vegetation within the gullied mires has established itself due to external forces favouring its growth (e.g. from a decrease in soil moisture associated with gullying). Table 7.1 illustrates that the two gullied mires have the highest number of dryland plants within the mires.

Due to the favourable conditions for the invasion of dryland species adjacent to the gullies, it is suggested that the gullied wetlands should be more species rich than non-gullied wetlands. Similarly, Taddese *et al*., (2002) found a greater species richness for grazed areas than non-

grazed areas. Hence, where livestock grazing is involved, as is the case for the valley heads investigated, it is most likely to have an influence on vegetation changes. Although species richness within each wetland is similar (Table 7.1), the gullies and their adjacent inter-drainage areas have induced a habitat that is greatly fragmented, thus permitting the diversification of plant communities. These new habitats are less homogenous and there is a greater ratio of fringe area to mire expanse.

Table 7.1 Number of plant species per valley head. Wet sp.% – fractional percentage of wetland plant species; dry sp.% – fractional percentage of dryland plant species. Gullied valley head wetlands are indicated in bold.

	Valley 1	Valley 2	**Valley 3**	**Valley 4**	Valley 5
Total species #	27	29	17	24	13
species # on wetland	14 (51%)	7 (24%)	9 (53%)	11 (46%)	8 (62%)
wet sp% : dry sp%	71:27	86:14	56:44	63:36	75:25

7.2.3 Summary: Gully Effects On Soil Properties

The presence of gullies within valley heads 3 and 4 has also indirectly impacted the adjacent organic matter content and to a lesser extent the pH of the topsoil. Organic matter is an important soil component as it increases runoff infiltration, aggregate stability and soil fertility (Snyman and du Preez, 2005). Shifts in species composition from grassland to shrubs, as recorded in the gullied valley heads, may result in changes to the soil organic matter content (Rezaei and Gilkes, 2005). The soil transects, the soil moisture transects, as well as upslope valley soil profiles reveal thick deposits of organic accumulation within each valley head. The average organic matter content is generally high within the mires (> 28%), and results from the transects indicate that in almost all situations, the samples are identified as being peat (Table 7.2). It should therefore be assumed that without the presence of gully erosion through the centre of the mires in valley heads 3 and 4, that all five mire expanses would be similar, in

terms of having deep, uninterrupted peat accumulations. The gullies have affected the adjacent soil chemical properties of the mire, either through decreasing soil moisture content and/or through vegetation change (Snyman and du Preez, 2005). A significant correlation is found between the percent organic matter and soil moisture within each valley head, thus as soil moisture decreases so too does organic matter content (Table 7.2). Although not always significant, the topsoil pH correlates negatively with organic matter content. Similarly, a study on rangelands in Iran also found a negative correlation between pH and organic matter content (Rezaei and Gilkes, 2005). The Iranian work also found no significant relationship between pH and other soil chemical variables, which could be due to the fact that pH is mainly affected by parent material (Rezaei and Gilkes, 2005). Although soil moisture correlates with vegetation type and organic matter content, vegetation type does not correlate with either pH or organic matter content. Thus, despite organic content often being cited as the best predictor of soil quality, the Sani 'wetland' plant species are not dependent on the wetland being peat producing, but rather are dependent on the moisture gradient for determining vegetation patterns and growth (c.f. Hall and Lamont, 2003).

Table 7.2 Spearman's rank correlation coefficients for various parameters measured along the entire soil transect in each valley head. ✓ = the null hypothesis was rejected, the sample is statistically significant. ✕ = the null hypothesis is not rejected; 'wet' = wetland; 'dry' = dryland; OM = % organic matter; SM = soil moisture.

	V1 n = 39		0.05	**V2** n = 22		0.05	**V3** n = 29		0.05	**V4** n = 57		0.05	**V5** n = 25		0.05
OM/SM	0.5		✓	0.9		✓	0.6		✓	0.4		✓	0.5		✓
OM/pH	-0.7		✓	-0.1		✕	0		✕	-0.3		✓	-0.3		✕
SM/pH	-0.1		✕	-0.3		✕	0.2		✕	0.3		✓	0		✕
	ca.	min	max	ca.	min	max	ca.	Min	max	ca.	min	max	ca.	min	max
Wet pH	6.15	5.5	6.7	6.14	5.9	6.5	6.33	6.1	6.6	6.33	4.8	7.1	6.5	6.1	7.1
Dry pH	6.34	5.9	6.9	6.19	6.0	6.5	6.26	5.8	7.1	6.11	5.2	7.1	6.6	6.3	6.7
Wet OM	39	14	72	32	28	38	28	22	31	34	14	62	39	18	49
Dry OM	15	5	20	17	9	23	20	10	27	19	8	31	23	17	30

7.3 Environmental Factors Contributing Towards Wetland Degradation And Erosion

Lesotho is regarded as one of the most eroded countries in the world and is faced with serious environmental degradation (Dregne, 1990; Showers, 1996; Grab *et al*., 2005a). The rapid rate of degradation is said to be the consequence of a complex set of short and long-term regional and site-specific factors (Figure 7.1) (Kakonge, 2002). The main factors are apparently overgrazing, the loss of soil fertility in agricultural areas and the loss of over 0.7 million tonnes/ha of soil per year through gully erosion (Kakonge, 2002).

As discussed in Chapter 4, the effects of livestock grazing are numerous and likely to be irreversible (Milton, 1995a). Overgrazing alters landscape heterogeneity through trampling and compaction of the soil and/or vegetation cover. Degradation of the soil leads to decreased infiltration, clogged soil pores, decreased aggregate stability, reduced fertility and a decrease in biomass production (Neave and Abrahams, 2001; Strunk, 2003; Metzger *et al*., 2005; Snyman and du Preez, 2005). Direct effects of overgrazing on vegetation type includes the replacement of palatable plants by those less palatable, due mainly to the loss of the entire plant, buds and/or seeds. In addition, overgrazing is known to cause changes to the soil structure and modify the microhabitat of the plant-growing environment (Milton, 1995a, 1995b; Taddese *et al*., 2002).

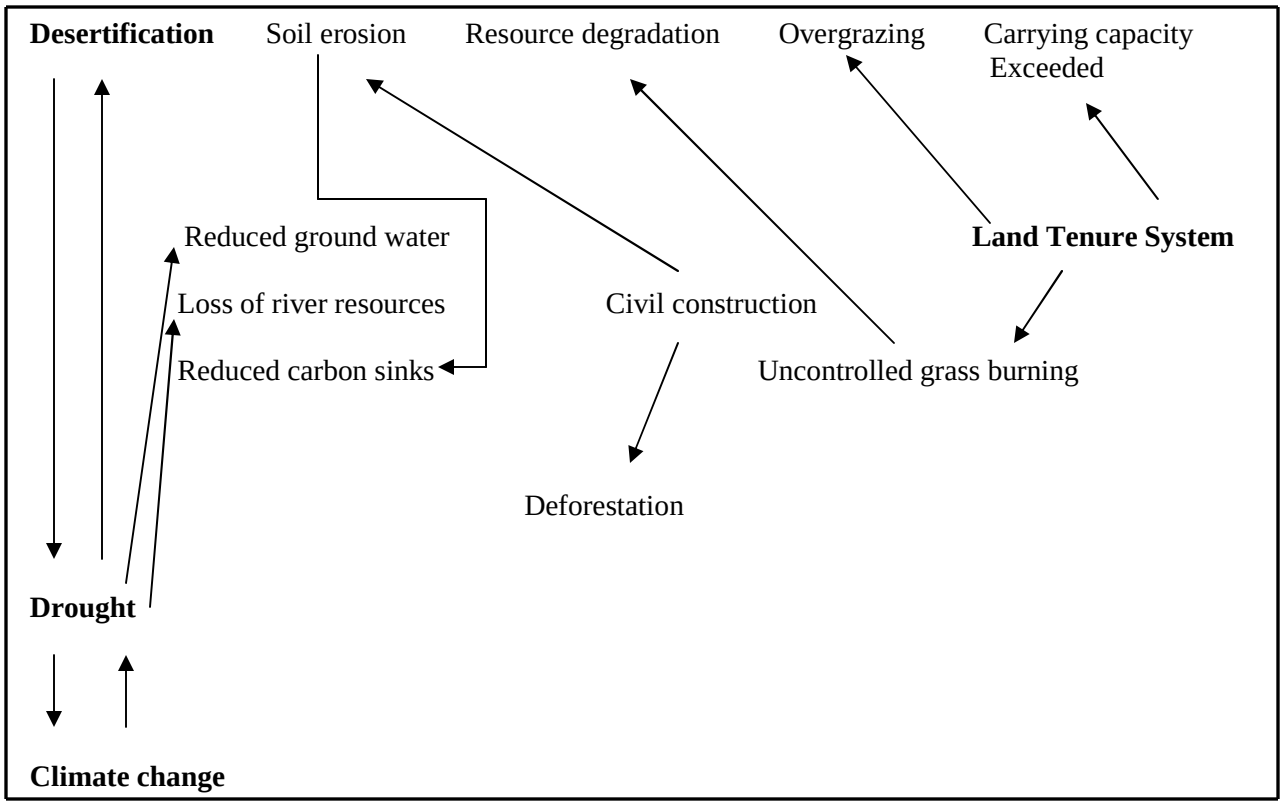


Figure 7.1 Environmental problems facing Lesotho (after Kakonge, 2002, p74).

Gully erosion essentially occurs because the velocity and force of the flow from runoff and raindrop splash has overcome soil particle resistance to erosion. Many thresholds are involved and must be overcome in order for this to occur (i.e. vegetation cover, soil type, rain intensity, anthropogenic factors, slope and drainage area, etc.).

7.3.1 Vegetation Thresholds

Raindrop impact and rain splash dislodge the soil particles most efficiently on a slope and/or in a strong wind, as such factors enhance the directional force. However, slopes that are well vegetated are better able to resist the forces of runoff and sediment detachment (Evans, 1980; Rowntree, 1988; Poesen *et al*., 2003). The threshold for vegetation cover varies from region to

region, ranging from as low as 15% in Alberta, Canada, to as high as 85% in the Ethiopian highlands (Crouch and Blong, 1989; Evans, 1998).

Once the vegetation threshold has been exceeded, the resistance of the soil to rainfall and overland flow erosion is decreased (Graf, 1979; Vandaele *et al*., 1996) and sediment yield increases following the Langbein-Schumm curve (Schumm and Lusby, 1963, 1981; Trimble, 1988). This curve indicates that once vegetation cover has decreased, sediment yield is directly related to the annual precipitation and runoff amounts (Schumm, 1981). The combined potential power of rainfall and overland flow is expressed conceptually as:

$$\text{Erosion} = f\ (\text{erosivity, erodibility}),$$

(Bocco, 1991, p392).

7.3.2 Rainfall Thresholds

In southern Africa, intensities of rain splash less than 30 ∞ 25 mm/hour are considered non-erosive, which is thus used as the erosive rain threshold value (Calles and Kulander, 1996). Calles and Kulander (1996) concluded that the high intensity, low frequency rainfall events required to exceed the intensity threshold value was not frequent enough to be responsible for the severe gully erosion in parts of Lesotho. Over 2200 (32%) gullies are estimated to occur in the lowlands and foothills region of Lesotho (Kakonge, 2002). However, a separate study has found that the low intensity, high frequency Lesotho rainstorms were the cause of severe gully erosion in grazing areas (Rydgren, 1988). Since much grazing (as opposed to cultivation) takes place in the foothills and mountain areas, perhaps rainfall could be considered as a major factor for the status of the soil erosion in the alpine areas. After only 20 minutes of rain, water from

approximately 150 m upslope can converge at a gully head to contribute to the total runoff amount (Faber and Imeson, 1982).

In Lesotho, the time period between 1979 – 1996 is reported to have experienced the highest incidence of droughts in two centuries (Kakonge, 2002). Land misuse during drought years rapidly declines organic matter content and soil porosity within the rooting zone. High intensity rainfall on air-dried soil (i.e. exposed, unvegetated soil), as opposed to soil with antecedent moisture content, leads to aggregate breakdown and surface seal formation (Imeson *et al*., 1982; Lado *et al*., 2004). A consequence of surface seal formation is a reduction in infiltration rates and increase of runoff rates (Morin *et al*., 1981).

7.3.3 Soil Thresholds

Soils that have large amounts of organic matter content (> 3%) and/or have less than 60% silt and/or fine sand are generally most water-stable, most porous and well structured (Evans, 1980). Soils with high clay contents have a high surface roughness (> 5-10 mm), low surface crusting and high rainfall storage capacity and thus encounter less runoff and sheetflow (Calles and Stålnacke, 2000). Inversely, soils with a low clay content (<100 g kg/sup-1) also have high infiltration rates and less runoff because the percentage of clay is insufficient to clog the soil pores and form a seal. A study on soil wetting and the effects of clay found that the most susceptible soil to form a seal was one that has an intermediate amount of clay (200 g kg^sup-1^) (Loch, 1994; Lado *et al*., 2004). Evans (1980) plotted more than 56 eroded soils from various regions (eg. Canada, India and England) onto a textural triangle, and the resulting pattern demonstrates that soils with greater than 35% clay are stable and more resistant to

raindrop and splash erosion (Figure 7.2) (Lado *et al*., 2004). The study also concluded that the most erodible soils contain an intermediate amount of clay (between 9% – 35%) (Evans, 1980). Soils with intermediate amounts of clay become highly erodible when wet, such that a rainfall of only 10 mm on a saturated soil is sufficient to generate runoff and sediment entrainment (Faber and Imeson, 1982; Evans, 1996). It should also be noted that the effects of grazing and agriculture on the soil compaction, loss of bulk density and surface roughness are also increased substantially on wet soils (Evans, 1980, 1996, 1998; Skovlin, 1984; Davies, 1985; Trimble, 1988).

Clay content of the peat taken from each valley head is similar between sites except for valley head 5, which has a much greater amount (Table 4.8). The clay content from valley head 5 is ca. 34%, which classifies it as a clay loam. According to Evans (1980) and Lado *et al*. (2004), soils are most erodible with clay contents between 9% and 35%, in which case, all of the valley head soils fall within the category of 'high erosion risk' (Figure 7.2).

7.3.4 Slope Thresholds

Even if all of the afore mentioned thresholds are surpassed, gully erosion cannot be initiated without confined, concentrated overland flow (Beckedahl *et al*., 1988; Hall and Lamont, 2003). Animal tracks, grazing routes and man-made drainage lines are often suitable structures promoting such concentrated flow. However, the erosive ability of channelized flow is related to slope angle and drainage basin area (Horton, 1945; Montgomery and Dietrich, 1988). Gully

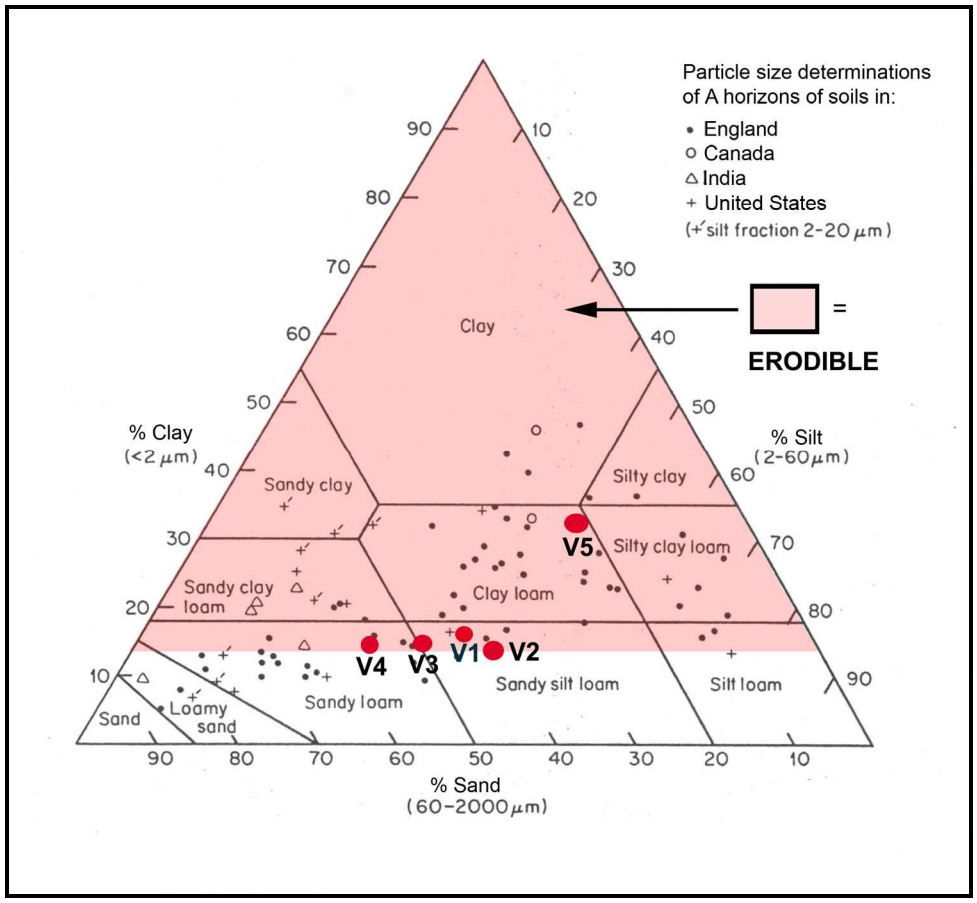


Figure 7.2 Soil textural triangle showing the results from previous soil tests summarized by Evans (1980, p117), and the results from the five valley heads from this study.

erosion is therefore not simply a function of overgrazing and poor land management. As discussed in chapter 4, the geomorphic threshold implies that for any stable landscape there exists a critical inherent gradient above which the slope angle of the drainage area becomes great enough to produce sufficient runoff to initiate gully incision (Schumm and Hadley, 1957; Patton and Schumm, 1975; Schumm, 1979, 1981; Montgomery and Dietrich, 1988; Morgan and Mngomezulu, 2003; Poesen *et al*., 2003). The present locations of the gully mouths in valley heads 3 and 4 have gentle slopes (<7°) (i.e. 'most erosive', as described in Chapter 4, section 4.6.4); it is possible that the gullies have migrated upslope from this location (c.f. Evans, 1980; Liggitt and Fincham, 1989; Boardman *et al*., 2003). However, no relationship

was observed between drainage basin area above the head and the position of gully incision. As mentioned in chapter 4 (section 4.6.4), this could simply imply that the drainage area was insufficient, or that the gullies had not yet reached their final upslope location (c.f. Patton and Schumm, 1975). Valley heads 3 and 4 have evident grazing routes to the east and west of the wetlands and these valley heads have active grazing posts from which channelized flow paths could be observed. In contrast, valley heads 1, 2 and 5 did not have active grazing posts within the valley heads, yet the slopes are grazed in summer.

7.4 Concluding Summary

Over 88% of the soil in Lesotho is in poor condition (Marake, 1998), possibly because the carrying capacity has been exceeded by over three times the acceptable limit (286 000 animal units) (Kakonge, 2002). Despite the lower population density and lower grazing intensities than in the lowland areas, ca. 9.7% of the highland area has been degraded by gully erosion (Kakonge, 2002) and half of the mires in the Senqu River valley have been damaged to some extent (Schwabe, 1995).

There is evidence of reduced vegetation cover, vegetation change, multiple track development and gully extension on the slopes surrounding the grazing posts. Reduced infiltration rates, decreased vegetation cover and increased runoff, channelize the surface flow through grazing routes and drainage lines towards the centre of the valley heads where the wetlands are located. Drought, uncontrolled grazing, an apparent population explosion of *O. sloggetti*, and an overall lack of awareness of the environmental degradation, has led to increased pressures on the valley head systems and peatlands (Figure 7.3). A survey in 1989 found that only 12% of the

current rangelands were in excellent condition (Kakonge, 2002). In another study conducted in 1989, 49% to 65% of the mires in the Senqu River Valley of eastern Lesotho had been damaged, indicating lowered water tables, increased scrub invasion and vegetation characteristics typical of dryland areas (Schwabe, 1989; Brooks and Stoneman, 1997). The high altitude mires of Lesotho are subjected to environmental pressures including, annual burning, grazing, trampling and water extraction. These factors are known to cause a possible loss of nutrients (notably K, Ca, Ph and N), loss of palatable vegetation, increase species invasion, contribute to vegetation loss and promote soil / peat erosion (Moore and Bellamy, 1973; Blackburn, 1984; Brooks and Stoneman, 1997; Evans and Warbuton, 2001; Visser *et al*., 2004).

Even though the Lesotho alpine wetlands occur at higher elevations than the surrounding protected South African wetlands, which also provide a habitat for unique and endemic plant species, it appears that the topographical remoteness and a lack of local conservation enforcement are placing the Lesotho alpine wetlands at a higher risk of becoming further damaged and degraded.

On the African continent, where approximately 21 million km^2 (or 75%) of its surface area is classified as dryland, existing wetlands are of great value to populations that depend on regular sources of water (Hughes, 1996; Schuyt, 2005). In 1997, wetlands of KwaZulu-Natal Drakensberg Park in South Africa (with ca. 2428 km^2 area) were added onto the Ramsar List of Wetlands of International Importance. This park has an abundance of high altitude wetlands and provides habitat for at least 36 endemic plants (Ramsar, 2004). More recently, the entire South African Drakensberg was proclaimed a World Heritage Site.

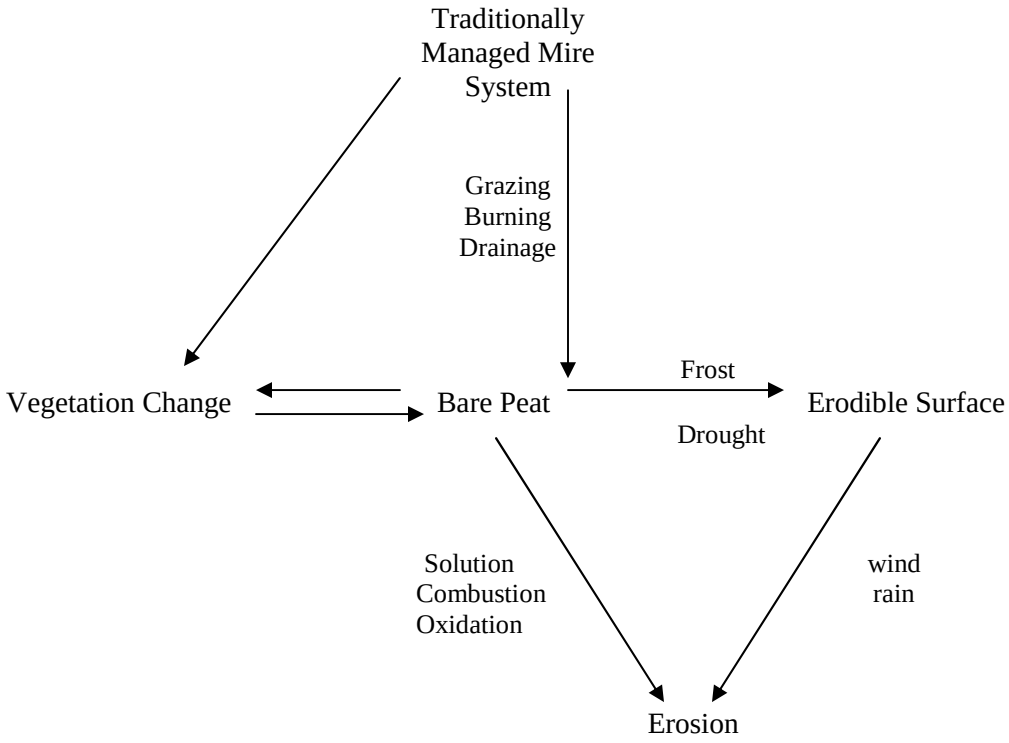


Figure 7.3 Sensitivity of the mire system in Lesotho. The system is therefore equally susceptible to both vegetation change and erosion (after Bragg and Tallis, 2001, p353).

This thesis has highlighted the causes and consequences of gully erosion within alpine wetlands of eastern Lesotho. Thus, given the current World Heritage Site status and trans-boundary initiatives, it is hoped that this work will be of value to the various stakeholders involved in the future management of such programmes.

Appendix A

Appendix A. Example of the form used for the collection of field data when conducting soil pit analyses.

References

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