THE ORIGIN AND EVOLUTION OF DARTMOOR VLEI IN THE KWAZULU-NATAL MIDLANDS, SOUTH AFRICA

By

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As the candidate's supervisor I have/~~have not~~ approved this thesis/dissertation for submission.

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ABSTRACT

Dartmoor Vlei is a 42ha un-channelled valley-bottom wetland system located in the headwaters of the Myamvubu River in the KwaZulu-Natal Midlands, South Africa. The wetland and its catchment are entirely underlain by a large dolerite sill that forms the Karkloof escarpment and plateau and the wetland terminates against a dolerite dyke ridge that has intruded into the sill. Wetlands on the Highveld and KwaZulu-Natal Midlands are thought to arise due to the lateral erosion of valleys upstream of resistant lithologies that impede vertical erosion. This is typical of valleys where Karoo sediments occur upstream of resistant dolerite dykes. Such valley widening by lateral planing is typically associated with actively migrating meanders. As a result, wetlands found upstream of dolerite intrusions are generally located on floodplains characterised by actively migrating meanders, extensive backswamps, ox bow lakes, alluvial ridges and clastic alluvial fill. However, in contrast to these floodplain wetlands, Dartmoor Vlei has evolved into an un-channelled valley-bottom wetland characterised by diffuse flow conditions, minimal channelled flow, extensive peat deposits and a general lack of floodplain features.

Coring within the wetland has established that the sedimentary fill of the wetland generally comprises upward fining sequences of sediment characterised by sands and gravels near the valley floor that grade into fine organic-rich silt sediments and peat at the surface. These findings confirmed that the wetland has evolved from a floodplain wetland characterised by laterally migrating meanders to a valley-bottom wetland characterised by discontinuous streams and peat accumulation. Coring also established that the wetland is predominantly underlain by residual saprolite that extends to depths in excess of 7m. The occurrence of a large discontinuity between the residual saprolite and fresh dolerite surfaces underlying the toe of the wetland indicated that the residual saprolite surface has sagged relative to the fresh dolerite and dolerite dyke at the toe of the wetland over time. Chemical and mineralogical analyses of fresh dolerite and residual clay within the valley confirmed that the chemical transformation of the dolerite bedrock into residual clay has resulted in both volume and thickness losses in the weathered dolerite sill mass. This has in turn resulted in the sagging of the valley floor and the wetland surface over time. These findings provide an explanation for the extremely low energy conditions of Dartmoor Vlei and explain why the wetland did not evolve in the same fashion as other wetlands in a similar geological and geomorphological setting. The extensive chemical weathering of the dolerite sill underlying the wetland has been attributed to the extremely long-time period that the soils within the wetland have been saturated. The long-term saturation of soils within Dartmoor Vlei has been facilitated by the formation and preservation of the African Erosion Surface on which Dartmoor Vlei is located.

PREFACE

The experimental work described in this dissertation was carried out in the School of Environmental Sciences, University of Natal, Durban, from January 2006 to July 2009, under the supervision of Professor Willliam N. Ellery and Professor Roseanne Diab, and co-supervision of Professor John Dunlevey.

These studies represent original work by the author and have not otherwise been submitted in any form for any degree or diploma to any tertiary institution. Where use has been made of the work of others it is duly acknowledged in the text.

DECLARATION 1 - PLAGIARISM

I, Ryan Edwards declare that

1. The research reported in this thesis, except where otherwise indicated, is my original research.

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3. This thesis does not contain other persons' data, pictures, graphs or other information, unless specifically acknowledged as being sourced from other persons.

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DETAILS OF CONTRIBUTION TO PUBLICATIONS that form part and/or include research presented in this thesis (include publications in preparation, submitted, *in press* and published and give details of the contributions of each author to the experimental work and writing of each publication)

Publication 1: n/a

Signed: 

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CHAPTER 1 - INTRODUCTION

Southern Africa hosts a diverse range of wetland habitats despite the semi-arid to arid climate of the subcontinent (McCarthy and Hancox, 2000; Tooth and McCarthy, 2007; Ellery *et al*, 2009). These include floodplain wetlands, valley-bottom wetlands that may be un-channelled or channelled, hillslope seepage wetlands and pans (Kotze *et al*, 2005). With the exception of pans, wetlands occur within the well developed drainage network. As a result, most wetlands are the product of fluvial geomorphic processes that create conditions in the landscape conducive to valley widening rather than valley deepening (Tooth and McCarthy, 2007). Despite the strong link between wetland formation and fluvial geomorphology, geomorphological research on southern African wetlands is limited as biologists and hydrologists have dominated wetland research (McCarthy and Hancox, 2000; Ellery *et al*, 2009). Thus, there are currently large gaps in understanding how wetlands originate and evolve such that they continue to remain poorly understood features of many landscapes (McCarthy and Hancox, 2000; Tooth and McCarthy, 2007). This deficiency has negative implications for the effectiveness and success of wetland management and rehabilitation interventions in southern Africa because of conflict between rehabilitation actions and natural processes (McCarthy and Hancox, 2000; Ellery *et al*, 2009).

At present, human-induced erosion brought about by the alteration of wetlands and their catchments is the biggest threat to the existence of wetlands in South Africa. Understandably, the prime focus of wetland rehabilitation interventions and strategies is thus on halting erosion and restoring natural hydrological conditions within wetlands. However, owing to the fluvial geomorphic evolution of southern Africa, wetlands in the region are generally transient features in the landscape over long timescales and in some cases erosion is a natural process in wetland evolution (Ellery *et al*, 2009). Thus, with a lot of resources being injected into wetland rehabilitation in South Africa, it is becoming increasingly important for wetland practitioners to understand the origin and evolution of wetlands and the possible underlying causes of wetland degradation before implementing rehabilitation and management strategies (Grenfell, 2007; Ellery *et al*, 2009). Such knowledge would enable wetland practitioners to effectively diagnose the causes of wetland degradation and better predict the effects of interventions in light of the natural evolution of wetlands (Grenfell, 2007; Ellery *et al*, 2009). Recent studies on wetland geomorphology in southern Africa have indicated that understanding the hydrogeomorphic controls on the formation and evolution of wetlands is invaluable to both the design of effective rehabilitation interventions and the prioritization of rehabilitation resources (McCarthy and

Hancox, 2000; Grenfell, 2007). Ignorance of these factors can cause wastage of resources if interventions work against natural processes that operate within wetlands, or it can cause loss of wetlands in the long term (Ellery *et al*, 2009). Within the context of South Africa being a developing country this is problematic (Ellery *et al*, 2009).

Although biologists and hydrologists have dominated wetland research in South Africa to date (McCarthy and Hancox, 2000), the importance of wetland geomorphology is increasingly being recognised as an integral part of wetland science and management (Tooth and McCarthy, 2007). As a result, a number of studies on the geomorphology of important South African wetlands have recently been undertaken and conceptual models of wetland evolution are currently being developed and tested (Tooth and McCarthy, 2007). To date, the most well known model of wetland formation in southern Africa is that of Tooth *et al* (2002a; 2004). Their research on the geomorphology of large floodplains located above resistant dolerite outcrops in the eastern interior of South Africa has established that resistant dolerite dykes and sills outcropping along river courses with their beds on the softer lithologies of the Karoo sedimentary rocks form stable local base levels along river courses such that floodplain wetland formation is initiated in the less resistant Karoo sedimentary valleys upstream (Tooth *et al*, 2002a; 2004). Due to the resistant nature of dolerite outcrops relative to the Karoo sedimentary rocks, vertical erosion in the sedimentary valleys upstream of dolerite intrusions is impeded by the slow erosion of the dolerite outcrops, and rivers adjust by using excess energy to laterally plane wide gently sloping valleys (Tooth *et al*, 2002a; 2004). The widening and planing of a valley over time by laterally migrating meandering streams creates accommodation space for lateral and low level vertical accretion and associated floodplain wetland environments (Tooth *et al*, 2004; Tooth and McCarthy, 2007).

Like the floodplain wetlands that have been studied by Tooth *et al* (2002a; 2004) on the Highveld of the Free-State Province and by Grenfell (2007) in the foothills of the KwaZulu-Natal Drakensberg, Dartmoor Vlei, a 42ha valley-bottom wetland in the KwaZulu-Natal Midlands, is located in a broad gently sloping valley immediately upstream of a dolerite dyke. However, in contrast to the floodplains that have been studied, Dartmoor Vlei is characterised by:

- small discontinuous meandering channels;
- a lack of classic floodplain features like backswamps, alluvial ridges, oxbow lakes and levees;

- the predominance of diffuse flow and permanently wet conditions throughout the wetland; and
- peat deposits.

In addition, the wetland is entirely underlain by a large dolerite sill that forms the Karkloof Escarpment and plateau, which contrasts with the Karoo sedimentary rocks that underlie the floodplains that have been previously studied. In light of these anomalous characteristics and the general deficiency in studies on wetland geomorphology in South Africa, this study aims to determine why Dartmoor Vlei does not conform to the current conceptual model of dolerite controlled floodplain wetland formation proposed by Tooth *et al* (2002a; 2004) by identifying the controls on the origin and evolution of the wetland and how these controls have influenced the wetland's evolution.

The objectives to achieve this aim are to describe the current hydrology and geomorphological structure of Dartmoor Vlei, to determine the nature of the wetland sedimentary fill and long-term history of sedimentation, to broadly determine the effect of climate on natural wetland dynamics and evolution, and to develop a conceptual model of the origin and evolution of Dartmoor Vlei and integrate it with current understanding of other South African wetlands.

CHAPTER 2 - THEORETICAL FRAMEWORK

2.1 Introduction

Wetlands occupy positions in the landscape that are transitional between purely aquatic and terrestrial environments and are defined as areas where the water table is usually at or near the surface, or where the surface of the land is covered by shallow water, such that anaerobic conditions develop within the rooting zone and vegetation is predominantly hydrophilic (Breen and Begg, 1989; Mitsch and Gosselink, 1993; Tooth and McCarthy, 2007; Ellery *et al*, 2009). The climate of a region is generally the major driving force behind wetland formation as wetlands are characterised by a surplus of water near the land surface (Mitsch and Gosselink, 1993; Ellery *et al*, 2009). However, climate is not the only factor responsible for wetland formation (Mitsch and Gosselink, 1993; Ellery *et al*, 2009). The availability of water at or near the surface is also strongly controlled by the geomorphology and geology of a region that influence and control the movement of water within the landscape (Brinson, 1993; McCarthy and Hancox, 2000; Tooth and McCarthy, 2007). In South Africa in particular, geomorphological and geological processes are integral in creating adequate accommodation space in the landscape within which water can accumulate, as well as creating features in the landscape that impede or restrict the movement of water and promote the shallow saturation of soils (Mitsch and Gosselink, 1993; McCarthy and Hancox, 2000; Tooth and McCarthy, 2007; Ellery *et al*, 2009).

This chapter reviews literature on the geological and geomorphological controls responsible for the formation and evolution of freshwater inland wetlands along southern African rivers as well as important concepts from the disciplines of hydrology, fluvial geomorphology and sedimentology relevant to an understanding of wetland origin and evolution.

2.2 Wetland hydrology

Before reviewing literature on the geological and geomorphlogical factors responsible for the occurrence of wetlands in southern Africa, it is necessary to understand the hydrology of wetlands that differentiates them from purely aquatic systems like rivers and lakes. Wetland hydrology is characterised by a surplus water supply at or near the soil surface for prolonged periods of time (Ellery *et al*, 2009). The prolonged shallow flooding and saturation of soils characteristic of wetland environments are generally characterized by the slow movement of water, long residence times, and low energy conditions, which combine to create the physiochemical conditions that make wetland ecosystems different from fast flowing high energy rivers and deeply flooded lakes (Mitsch and Gosselink, 1993; Ellery *et al*, 2009). Under natural conditions, the hydrology of a wetland is the product of climate, which determines the availability of water within the landscape, and the topography and geology of the landscape which determine the movement and distribution of water within the landscape (Mitsch and Gosselink, 1993; Ellery *et al*, 2009).

An important concept in understanding the nature of wetland hydrology is the ‘wetland hydroperiod’ which is the seasonal pattern of water distribution and level, which defines the rise and fall of a wetland’s surface and subsurface water (Mitsch and Gosselink, 1993). The hydroperiod is an integration of all inflows and outflows of water, but it is also influenced by physical features of the terrain and by proximity to other bodies of water (Mitsch and Gosselink, 1993). The hydroperiod can be summarized as being the result of the following factors (Mitsch and Gosselink, 1993):

- The balance between the inflows and outflows of water
- The surface contours of the landscape
- Subsurface soil, geology and groundwater conditions

The first condition defines the water budget of the wetland, whereas the second and third define the capacity of the wetland to store water (Mitsch and Gosselink, 1993). The general balance between water storage and inflows and outflows is expressed as an equation that states that the change in the volume of water storage in a wetland is equal to the balance of the inflows and the outflows (Mitsch and Gosselink, 1993; Ellery *et al*, 2009). The equation is expressed as follows (Mitsch and Gosselink, 1993):

$$\Delta V/t = Pn + Si + Gi - ET - So - Go$$

where:

V = volume of water storage in a wetland

t = time

$\Delta V/t$ = change in volume of water storage in wetlands over time

Pn = net precipitation that reaches the surface

Si = surface inflows via streams or overland flow

Gi = groundwater inflows

ET = evapotranspiration

So = surface outflows

Go = groundwater outflows

This equation provides a useful summary of the major hydrologic components of any wetland water budget (Mitsch and Gosselink, 1993). The potential sources of water supplying a wetland include rainfall, surface runoff, river or stream inflow, and groundwater inflow (Mitsch and Gosselink, 1993; Ellery *et al*, 2009). In each case, the relative contribution of each input source may vary with time, as may the flow characteristics of each source (Ellery *et al*, 2009). The outflows or losses from a wetland include evapotranspiration, groundwater outflows, surface outflows via streams or rivers, or a combination of any or all of these outflows (Mitsch and Gosselink, 1993; Ellery *et al*, 2009).

Wetland water budgets are characterized by a positive storage component where the sum of the input components is greater than the sum of the output components for some time during the year (Mitsch and Gosselink, 1993; Ellery *et al*, 2009). As a result, processes that add water rapidly to a site will increase the likelihood of wetland formation, as will any factors that inhibit the rapid water movement away from a site and increase water storage at a site (Mitsch and Gosselink, 1993; Ellery *et al*, 2009). Processes that that add water rapidly to a site include rainfall and surface inflow via rivers; processes that inhibit the movement of water away from a site include geological and geomorphological barriers to streamflow, dense stands of vegetation, and gentle topographical gradients. In contrast groundwater inflow adds water slowly to a site while surface outflows in the form of confined steeply sloping stream and river channels remove water rapidly from a site (McCarthy and Hancox, 2000; Tooth and McCarthy, 2007; Ellery *et al*, 2009). As a result, wetlands are expected to occur most extensively in settings where there is high rainfall, where there are large rivers that discharge into shallow gently sloping basins, or where surface outflows are absent or restricted (Tooth and McCarthy, 2007; Ellery *et al*, 2009). In general, wetland formation in open drainage basins is less likely due to the rapid delivery of water away from a site via channelled flow (Ellery *et al*, 2009). However, flow within rivers and streams is often restricted or impeded by natural barriers, as described above, which results in limited surface outflows from a site (Ellery *et al*, 2009).

2.2.1 Surface flow

Wetlands can be both the receiving systems and source areas of surface water flows (Mitsch and Gosselink, 1993). Wetlands may be subject to surface inflows of several types namely, hillslope runoff, direct streamflow or indirect streamflow through the overtopping of channel banks during floods. Within fluvial settings, wetlands generally occur in wide shallow basins in and around river channels like floodplains and broad valley-bottom areas and they either receive flow directly from a stream channel or indirectly through the seasonal overtopping of channel banks during flood discharges (Mitsch and Gosselink, 1993). Wetlands occurring on floodplains have been called riparian and riverine wetlands and the flooding of these wetlands varies in intensity, duration, and number of floods from year to year (Mitsch and Gosselink, 1993; Rogers, 1997; Ewart-Smith *et al*, 2006).

The properties of channelled flow play in integral part in the hydrology of wetlands located within a drainage basin and it is important to understand the basic concepts of channelled flow. The amount or volume of water within a cross sectional unit of channel is referred to as the discharge of the channel and is described simply as the product of the cross sectional area of a stream channel and the average velocity, which together are measured in m^3/sec (Mitsch and Gosselink, 1993). When river flow fills the entire channel the flow is referred to as bank-full discharge and when river flow begins to overflow onto the floodplain, the flow is referred to as flood discharge (Schumm, 2005).

The velocity of channelled flow can be described by the Manning equation if the slope of the stream and a description of the surface roughness are known (Mitsch and Gosselink, 1993). The velocity of flowing water within a channel is a function of the gradient of a stream, the volume of water flowing within the stream cross-section, the viscosity of water, and the characteristics of the channel cross-section and bed (Morisawa, 1968). This relationship is expressed by Manning's equation: $V=R^{2/3}.S^{½}/n$ where V is the mean velocity, R is the hydraulic radius of the channel, S is the longitudinal gradient of the channel, and n is the Manning's roughness coefficient indicating the roughness of the channel (Ellery *et al*, 2009). The hydraulic radius is a measure of the relative friction imposed on the sides and bed of the channel and is calculated as the cross-sectional area (m^2) divided by the wetted perimeter (m) (Morisawa, 1968; Ellery *et al*, 2009). The roughness factor has to be determined empirically and varies not only for different streams but also for the same stream under different conditions and at different times (Morisawa, 1968). Factors which affect the roughness of a channel include the size and shape of the grains on the bed, sinuosity, obstructions in the channel such as vegetation, logs and sandbars, any irregularity within the channel cross-section, and the depth and volume of water flowing within a channel described by discharge (Morisawa, 1968).

Assuming all other variables constant, this equation shows that velocity is:

- proportional to hydraulic radius such that for a given roughness and slope, wide shallow channels have a lower velocity than channels with a narrower, deeper cross section;
- proportional to slope such that for a given hydraulic radius and roughness, channels with a steep slope have a higher velocity than channels with a shallow slope; and
- inversely proportional to bed roughness, such that for a given hydraulic radius and slope, streams with boulder strewn or vegetated beds have a lower velocity than unobstructed sandy or unvegetated beds (Ellery *et al*, 2009).

It is also important to note that velocity increases with the depth of the channel for a constant slope and bed roughness (Leopold and Maddock, 1953; Ellery *et al*, 2009).

2.2.2 Groundwater flow

In this section, details on the nature and importance of groundwater flow within wetlands has been sourced from Mitch and Gosselink (1993). The recharge-discharge function of wetlands on groundwater resources has often been cited as one of the most important attributes of wetlands, but it does not hold for all wetland types Groundwater inflows occur when the surface water level of a wetland is at a lower elevation than the water table of the surrounding landscape. In this situation wetlands are referred to as discharge wetlands as groundwater discharges into the wetland. Another type of discharge wetland is called a spring or a seep wetland and is often found at the base of slopes where the groundwater surface intercepts the land surface.

When the water level in a wetland is higher than the water table of its surroundings, groundwater will flow out of the wetland. This occurs because surface flow down a valley is more efficient than subsurface flow. In this situation, wetlands are referred to as recharge wetlands as the wetland recharges the groundwater. When a wetland is well above the groundwater of an area, the wetland is referred to as being perched. Many wetlands also occur where soils have poor permeability. In these type of wetlands the major source of water can be restricted to surface water runoff, with losses occurring only through evapotranspiration and other surface outflows. This type of wetland often has fluctuating hydroperiods and intermittent flooding and standing water is dependent on seasonal surface inflows.

The flow of groundwater into and out of a wetland is often described by Darcy’s Law. The law states that flow of groundwater is proportional to the slope of the water table, or the hydraulic slope, and the hydraulic conductivity or permeability of the geological material through which the water flows. In equation form Darcy’s Law is given as:

G = kas

Where,

G = flow rate of groundwater measured as a volume per unit time (m^3/sec)

k = hydraulic conductivity or permeability measured in length per unit time (m/sec)

a = groundwater cross-sectional area perpendicular to the direction of flow (m^2)

s = hydraulic gradient or slope of the water table (no unit as slope is a fraction of $\Delta y/\Delta x$)

2.2.3 Wetland hydrology in southern Africa

Owing to the semi-arid to arid climate and negative atmospheric water balance of southern Africa, rainfall is generally insufficient to directly sustain wetlands over most of the country (McCarthy and Hancox, 2000; Tooth and McCarthy, 2007; Ellery *et al*, 2009). Surface water within the country is thus predominantly located within perennial, seasonal and ephemeral rivers and streams that originate in the Drakensberg Mountains and related mountain ranges where precipitation is the highest (McCarthy and Hancox, 2000; McCarthy and Rubidge, 2005; Tooth and McCarthy, 2007; Ellery *et al*, 2009). As a result, most wetlands within southern Africa are fed by surface flows of rivers and streams within the well developed drainage network of the subcontinent (McCarthy and Hancox, 2000; Tooth and McCarthy, 2007). Within the South African drainage network, wetland hydrology is predominantly controlled by the geomorphology and geology of the valleys within the drainage network (Tooth and McCarthy, 2007; Ellery *et al*, 2009). Due to the strong association between geomorphic setting and wetland hydrology in South Africa, a classification of South African wetlands has been developed that describes wetlands in terms of hydrogeomorphic setting (Kotze *et al*, 2005; Ewart-Smith *et al*, 2006). This classification system for South African wetlands developed by Kotze *et al* (2005) has identified six broad hydrogeomorphic wetland types, namely floodplain wetlands, valley-bottom wetlands with a channel, valley-bottom wetlands without a channel, hillslope seepage wetlands feeding a watercourse, hillslope seepage wetlands not feeding a watercourse, and depressions (Kotze *et al*, 2005). The hydrogeomorphic types are defined based on geomorphic setting, water source and how water flows through the wetland (Kotze *et al*, 2005). Each one is briefly described below.

Floodplains are relatively flat gently sloping valley-bottom areas bordering a stream that are fed by the seasonal overtopping of the channel banks (Wolman and Leopold, 1957; Kotze *et al*, 2005; Ewart-Smith *et al*, 2006). Floodplains are characterized by a well defined stream channel that is often actively migrating and by the alluvial transport and deposition of sediment, usually leading to the net accumulation of sediment and the formation of sedimentary structures like channel bars, levees, alluvial ridges, scroll bars and oxbow lakes (Bridge, 2003; Kotze *et al*, 2005). Depending on the specific context, wetlands may occur on all or few parts of the floodplain (Nichols, 1999; Slingerland and Smith, 2004). In general, most of the wetland on floodplains is generally located within the seasonally and permanently inundated backswamp areas, with some wetland areas often occurring in oxbow and abandoned channel depressions (Tooth *et al*, 2002a; Tooth and McCarthy, 2007). Water inputs into floodplain wetlands are mainly from the main channel during seasonal floods, with smaller contributions from lateral seepage (Kotze *et al*, 2005). Due to the gentle floodplain gradients and the sealing off of the floodplain surface by fine clays, sheet and diffuse flow generated during seasonal floods usually accumulates on the floodplain surface for prolonged periods of time (Tooth *et al*, 2002b). Water loss from these wetlands generally occurs via evapotranspiration, groundwater discharge, and channelled outflow if wetlands are in the vicinity of the floodplain channel (Ellery *et al*, 2009).

A channelled valley-bottom wetland is a valley-bottom area that has a well defined stream but lacks the characteristic floodplain features described above (Kotze *et al*, 2005; Ewart-Smith *et al*, 2006). These wetlands may be gently sloping and characterised by the net accumulation of alluvial deposits, or they may be relatively steeply sloping characterised by the net loss of sediment (Kotze *et al*, 2005). Water inputs are mainly from adjacent slopes with limited inputs from the main channel when flow overtops the channel banks during exceptional floods (Kotze *et al*, 2005; Ellery *et al*, 2009). Flow within the valley-bottom areas elevated above the channel is predominantly diffuse as hillslope seepage generally drains toward the main channel (Kotze *et al*, 2005). The main channel represents the main reason for the loss of water from these wetlands, with smaller losses occurring due to evapotranspiration (Ellery *et al*, 2009).

An un-channelled valley-bottom wetland is a gently sloping valley-bottom area of low relief with no clearly defined stream channel (Kotze *et al*, 2005; Ewart-Smith *et al*, 2006). Within these wetlands, incoming surface flow usually via one or more influent streams, usually spreads out across the valley-bottom as sheet flow and/or subsurface and surface diffuse flow (Kotze *et al*, 2005; Ewart-Smith *et al*, 2006). Hillslope seepage from adjacent slopes also provides some input into these wetlands (Kotze *et al*, 2005; Ellery *et al*, 2009). The often abrupt reduction in stream power and definition associated with these wetlands leads to the regular deposition of sediment within the valley and aggradation of the valley surface over time (Kotze *et al*, 2005; Ewart-Smith *et al*, 2006). The slow diffuse nature of the flow of water through the wetland in conjunction with large inputs associated with the spreading out of incoming stream flow, usually promotes shallow seasonal and permanent water tables, which in turn promotes localised organic sedimentation (Ellery *et al*, 2009). Losses from these wetlands are generally via channelled outflow as the valley becomes confined, and evapotranspiration is also an important means of water loss (Kotze *et al*, 2005; Ewart-Smith *et al*, 2006; Ellery *et al*, 2009).

Hillslope seepage wetlands not feeding a watercourse are typically concave slopes on hillsides that are characterised by the slow movement of subsurface flow at or near the surface (Kotze *et al*, 2005; Ellery *et al*, 2009). Brinson (1993) identifies two processes that give rise to hillslope seepage wetlands. The first involves the intersection of the ground surface by the water table as a result of an abrupt break in slope or the occurrence of impermeable strata that direct groundwater to the surface (Ellery *et al*, 2009). The second involves the upward movement of groundwater towards the surface as a result of hydraulic forces (Brinson, 1993; Ellery *et al*, 2009). Within these wetlands, water inputs are mainly from upslope groundwater flow and subsurface seepage (Kotze *et al*, 2005; Ellery *et al*, 2009). Flow through these wetlands is predominantly via diffuse subsurface and surface flow as groundwater moves into and through the wetland under the force of gravity (Kotze *et al*, 2005; Ellery *et al*, 2009). Outflows also comprise diffuse subsurface flow from the wetland under the force of gravity, with smaller losses occurring through evapotranspiration (Kotze *et al*, 2005; Ellery *et al*, 2009).

Hillslope seepage wetlands feeding a watercourse also typically occur on concave slopes of hillsides for the same abovementioned reasons. However, outflow is via a well defined stream channel connected to a watercourse (Kotze *et al*, 2005; Ewart-Smith *et al*, 2006).

Depression wetlands are generally basin-shaped areas with closed elevation contours (Kotze *et al*, 2005). Ewart-Smith *et al* (2006) identify three types of depression wetlands that include depressions linked to a channel with channelled outflow, depressions linked to a channel but without appreciable channelled outflow, and isolated depressions. Depressions linked to a channel with channelled outflow may receive surface water from a channel and have an outflow of surface water via a channel, while depressions linked to a channel without channelled outflow receive water from a channel but do not have a visible outflow of surface water via a channel (Ewart-Smith *et al*, 2006). An isolated depression is a basin-shaped area with a closed elevation contour that allows for the accumulation of water via subsurface inputs and/or rainfall, but is not connected via a surface inlet or outlet to the drainage network (Ewart-Smith *et al*, 2006).

From this classification it is evident that the majority of wetlands within southern Africa generally occur in settings within the drainage network where stream channels are generally alluvial, longitudinal gradients are gentle, flow is often unconfined and diffuse, the velocity of flowing water is low, and deposition is an important fluvial process with little or no erosion (Ellery *et al*, 2009). However, most of the southern African drainage network is currently in a state of long term incision due to two episodes of epeirogenic uplift that occurred most extensively along the south-eastern continental margin of the subcontinent during the Miocene (20Ma BP) and Pliocene (5Ma BP) (Dardis *et al*, 1988; McCarthy and Hancox, 2000; Tooth *et al*, 2002b; McCarthy and Rubidge, 2005). As result, the hydrogeomorphic wetland types and associated low energy settings described by Kotze *et al* (2005) and Ewart-Smith *et al* (2006) are expected to be rare unless drainage is impeded or restricted (McCarthy and Hancox, 2000; Ellery *et al*, 2009). The fact that these hydrogeomorphic wetland types exist within the drainage network of southern Africa indicates that the prevailing evolution of the drainage network has been controlled by factors besides uplift and warping (McCarthy and Hancox, 2000; Ellery *et al*, 2009). Such factors have temporarily impeded incision in a setting where incision should be widespread (McCarthy and Hancox, 2000). Recent research on a number of wetlands within the interior of the subcontinent by Tooth *et al* (2002a; 2002b; 2004) and Grenfell (2007) has revealed that structural, lithological and geomorphological barriers to incision and flow occur extensively throughout the interior of the subcontinent and that these controls have significantly influenced the prevailing evolution of parts of the drainage network, leading to wetland formation. Before reviewing these controls on wetland formation it is necessary to review both the basic concepts of fluvial geomorphology and the geomorphological and sedimentological characteristics of wetlands. This will enable a better understanding of how these controls influence the evolution of the South African drainage network.

2.3 Fluvial geomorphology

Fluvial geomorphology is the study of how flowing water within streams and rivers shape the surface of the earth (Leopold *et al*, 1964). Rivers are networks of channels that collect and convey runoff and sediment that is generated by rainfall (Allen, 1970). Rivers drain water from the continents to the ocean and the power of running water makes it possible for rivers to erode and transport sediment and shape landscapes (Leopold *et al*, 1964). In accomplishing this transfer, the water flowing off the land towards the sea forms and maintains a highly organised system of physical and hydraulic features (Leopold *et al*, 1964).

2.3.1 Fluvial processes

Rivers shape landscapes through the processes of erosion, sediment transport and deposition (Ritter *et al*, 2002). The ability of river channels to erode and transport water and sediment and shape landscapes is determined by the driving and resisting forces built into the system (Ritter *et al*, 2002). A river will characteristically erode, transport or deposit sediment depending on the driving energy given to the water by velocity, depth and slope, and on the energy consumed by resistance to flow offered by elements such as channel shape, channel configuration, particle size, sediment concentration and the total volume and physical properties of water (Morisawa, 1968; Ritter *et al*, 2002).

Erosion and Transportation

The velocity and discharge of channelled flow is considered a key property in determining the ability of running water to erode and transport sediment as it generally represents the amount of energy being supplied to the water to do the work of erosion and transportation (Ritter *et al*, 2002).

The amount of sediment transported within a river channel is also dependent on the size of the particles that are available for transportation (Ritter *et al*, 2002). The load carried by natural streams can be separated into two components: dissolved load and clastic load (Knighton, 1998). Clastic load in turn can be divided into suspended and bed load (Knighton, 1998). Dissolved load comprises all the material transported in solution by rivers (Knighton, 1998). The dissolved load varies in magnitude and composition according to the dominant sources, the rates of solute mobilization and the hydrological pathways operating within the catchment (Knighton, 1998). Suspended load comprises all clastic sediment transported in suspension and generally consists of silt and clay-sized particles (Morisawa, 1968). Suspended load is the finest grain fraction of the total sediment load, consisting of particles whose settling velocities are so low that they are transported in suspension at approximately the same velocity as the flow and settle out when velocity of flow is much reduced (Knighton, 1998). Coarse particles (sand and gravel) may also travel in suspension but they are likely to be deposited more quickly and stored in channel banks (Ritter *et al*, 2002). Except for short periods of suspension, coarse sediment, generally consisting of sand, gravel and boulder sized particles, is usually transported as bed load (Ritter *et al*, 2002). Bed load refers to sediment transported close to or at the channel bottom by rolling, sliding or bouncing (Ritter *et al*, 2002). The sediment load transported by streams is supplied by the weathering of exposed rocks and by the down-slope movement of loose material (Bridge, 2003). The rate of weathering and the texture and composition of weathered material is controlled by the nature of the exposed rocks, the amount of precipitation, temperature variations and the presence of vegetation (Bridge, 2003; Schumm, 2005). These are in turn controlled by topography and climate (Bridge, 2003). For a given parent material, relatively coarse weathered material tends to be most common in cold climates and where there are steep slopes, whereas clays tend to dominate in warm humid climates or regions of low slope (Bridge, 2003; Schumm, 2005).

The ability of a river to transport sediment can be described by the concepts of competence and capacity (Morisawa, 1968; Ellery *et al*, 2009). The largest grain a stream can move along its bed under any given set of hydraulic conditions is referred to as its competence (Morisawa, 1968). The competence of a river changes greatly, not only from the head of a stream to its mouth, but also from time to time at the same spot (Morisawa, 1968). The sediment transport rate or capacity of a stream is the maximum amount of sediment of a given size that a stream can carry (Morisawa, 1968; Ellery *et al*, 2009). Both these properties determine the nature of erosion and deposition (Morisawa, 1968; Bridge, 2003). Where the capacity of a stream is greater than the sediment load, erosion will occur but where the capacity of a stream is less than that required to transport the available load, deposition should occur (Ritter *et al*, 2002; Ellery *et al*, 2009).

2.3.2 The fluvial system

In order to simplify the discussion of the complex assemblage of landforms that make up a fluvial system, an ideal fluvial system can be divided into three zones: the eroding headwaters, the water-sediment transport-transfer system and the depositional lowlands (Schumm, 2005). These three simplified compartments of a river system basically show that rivers shape the landscape either by vertical erosion, lateral erosion, or both. The eroding headwaters comprise of a network of streams where the downstream gradient is steep and velocities are high (Bridge, 2003; Schumm, 2005). In this region rivers are eroding into the landscape, are deeply incised, have steep sided valleys with narrow V-shaped cross sections, and have small catchment areas and small discharges (Schumm, 1977; Ritter *et al*, 2002). The dominant fluvial activity is erosion which takes place by corrasion, hydraulic action and attrition (Knighton, 1998). In the middle reaches of a drainage basin, the dominant fluvial activity is sediment transportation (Schumm, 2005). Here river gradients and flow velocities generally decrease and valleys become more U-shaped (Ellery *et al*, 2009). Despite the reduction in downstream gradient, velocities can still remain moderate to high and capacity and competence are sufficient to transport the load (Ritter *et al*, 2002). This is because rivers in the middle reaches have larger catchment sizes than their upstream counterparts, resulting in larger discharges and wider and deeper channels, compensating for the decrease in gradient (Ritter *et al*, 2002). In the lower reaches of the drainage basin, deposition is the dominant fluvial activity (Ritter *et al*, 2002). The lower reaches of rivers typically have a shallow gradient, low flow velocity and U-shaped valleys that are shallow in cross section (Ellery *et al*, 2009). However, like the middle reaches of rivers, rivers in the lower reaches have large catchment sizes, resulting in high discharges (Ritter *et al*, 2002).

Base level

The lowest elevation below which a river cannot incise is referred to as the base level and occurs where rivers flow into the sea or at confluences with the trunk river (Leopold and Bull, 1979). Fluvial processes cease where a river flows into a lake, a larger trunk stream or the ocean, because the hydraulic gradient is reduced to zero or close to zero (Leopold and Bull, 1979). The ocean is regarded as the ultimate base level and trunk river beds are regarded as local base levels for incoming tributary rivers and streams (Leopold and Bull, 1979). In reality, however, a river's catchment is highly varied along the length of a stream in terms of geology and geomorphology, and local base level formation is a common occurrence between the headwaters of a stream and the ocean (Leopold and Bull, 1979; Ellery *et al*, 2009). Local base levels are areas along rivers and streams where incision is temporarily halted due to the river being unable to erode below a certain elevation as a result of fixed local controls like impoundments or resistant lithologies (Schumm, 2005). For example, a fixed local control like a resistant lithology forms a local base level because the stream is unable to incise the bed through the resistant lithology as easily as through the materials upstream and downstream of it (Leopold and Bull, 1979; Tooth *et al*, 2002a). Thus, the elevation of the stream bed as the river flows over the outcrop represents the elevation below which the river cannot incise. In addition to the formation of local base levels due to the occurrence of fixed local controls, a number of different upstream, downstream and local controls on river behaviour and form also occur between headwaters of a stream and the ocean or its confluence with a tributary base level (Leopold and Bull, 1979; Schumm, 2005). As a result of local factors that influence river behaviour and form, the longitudinal profile of rivers and their drainage basins are never smooth and generally contain convex reaches (Knighton, 1998).

The longitudinal profile

The longitudinal profile is a plot of channel bed elevation over channel distance from an upstream reference point to a downstream reference point (Philips and Lutz, 2008). The longitudinal profile is a fundamental measure in understanding fluvial geomorphology because it reflects the slope and energy gradients of a stream along its length (Philips and Lutz, 2008).

The longitudinal profile has been widely used as a diagnostic indicator of the stage of evolution of a landscape, tectonic uplift or subsidence, variations in lithology, base level changes and the effects of climate on landscapes (Philips and Lutz, 2008).

A more or less uniformly variant concave upward longitudinal profile has long been considered the ideal or attractor state for channel evolution and an indicator of steady state, where a stream grade approximates an equilibrium state (Ritter *et al*, 2002; Philips and Lutz, 2008). The strong association of smooth concave upward longitudinal profiles with steady state, grade and equilibrium is based on the notion that, as discharge increases downstream, the gradient necessary to transport sediment load decreases (Ritter *et al*, 2002; Philips and Lutz, 2008). River systems characteristically achieve a concave upward longitudinal profile due to downstream increases in discharge and channel widths and depths (Ritter *et al*, 2002). The increasing discharge and resultant channel widths and depths imply an ability to transport the same load over progressively lower slopes (Ritter *et al*, 2002). However, a river can accommodate increasing discharge by varying current velocities, depth and width, and different combinations of the adjustable variables may be associated with different profile forms (Knighton, 1998). Increasing discharge and decreasing material size provide only a general explanation of profile concavity (Ritter *et al*, 2002).

Traditionally geologists and geographers have assumed that a concave upwards longitudinal profile is the channel form assumed by graded rivers or rivers in equilibrium (Ritter *et al*, 2002). This is because channel longitudinal slope has always been recognised as the prime adjustable property of rivers in its effect on current velocity (Knighton, 1998). However, changes in channel gradient have been found to account for a small part of the adjustment required to achieve a balance between the input and output of sediment, most of the adjustment being accomplished by changes in other components of flow and channel geometry, which respond more rapidly (Leopold and Bull, 1979; Knighton, 1998). It is now recognised that, in response to a given set of external upstream, local and downstream controls, each river strives to develop a particular combination of gradient, hydraulic geometry and pattern that allows it to accommodate available discharge and transport available sediment most efficiently (Knighton, 1998).

Grade and equilibrium

Leopold and Bull (1979: 195) define a graded stream as “one in which, over a period of years, slope, velocity, depth, width, roughness, pattern and channel morphology delicately and mutually adjust to provide the power and the efficiency necessary to transport the load supplied from the drainage basin without aggradation or degradation of the channels". In the graded condition, river and stream channels reach a state of stability or dynamic equilibrium where mass entering the system is approximately equal to mass leaving the system, resulting in the efficient transport of water and sediment (Morisawa, 1968). Equilibrium is not a purely static condition and true stability never exists in natural rivers, which frequently change their position and which must continue to pass a range of discharge and sediment loads (Knighton, 1998). In reality natural rivers are always eroding and depositing along their channels as discharge and sediment change daily, seasonally and annually (Ritter *et al*, 2002). However, viewed over a long period of time with no major climatic and tectonic changes, natural rivers generally develop characteristic equilibrium channel and morphological forms that most efficiently transport the prevailing sediment load in a given hydrogeomorphic setting and in a given time period (Morisawa, 1968; Knighton, 1998).

It is important to note that because the equilibrium concept is most frequently used qualitatively rather than quantitatively in geomorphology, there is, as yet, no universally accepted set of criteria for determining whether all or part of a river system is in equilibrium (Thorn and Welford, 1994; Knighton, 1998). Despite this, it is still evident that most natural rivers achieve a state of stability where channel form is adjusted to the average discharge and sediment load, and where there is no obvious rapid degradation or aggradation over the short term (Knighton, 1998).

Channel pattern is also recognized as a mode of channel form adjustment in the horizontal plane, which is additional to but nevertheless linked with cross-sectional and longitudinal modes of adjustment towards equilibrium (Knighton, 1998; Ritter *et al*, 2002). Channel pattern influences resistance to flow and can be regarded as an alternative to slope adjustment when valley slope is fixed in the short- and medium-term (Knighton, 1998). Channel patterns are usually classified as straight, meandering or braided (Ritter *et al*, 2002). The distinction between straight and meandering channels is based on a property called sinuosity, which is the ratio of stream length to valley length (Ritter *et al*, 2002). The transition from a straight to meandering channel is usually placed at a sinuosity value of 1.5 (Ritter *et al*, 2002). Most streams do not have straight banks for any significant distance, making the straight pattern uncommon in the landscape (Ritter *et al*, 2002). "The initiation of meanders requires localised bank retreat which, if a series of bends develop, must alternate from one side of the channel to another in a more less regular fashion" (Knighton, 1998: 220). The periodic deformation of the channel bed and the development of a sinuous thalweg are regarded by many as necessary precursors to erosion of the channel bank (Knighton, 1998). However, there is no completely satisfactory explanation of how or why meanders develop (Knighton, 1998. Most explanations fall into two categories, regarding oscillation as either an inherent property of turbulent flow or as a result of the interaction between the flow and a mobile channel bed (Knighton, 1998).

Other explanations have been less concerned with specific processes and rather more with why streams meander in the broader context of channel form adjustment (Knighton, 1998). In this regard, meandering is considered one of the ways whereby a river can adjust its rate of energy expenditure and ability to transport sediment so that a form of equilibrium is reached (Knighton, 1998). In line with the minimum stream power concept, Chang (1980) argued that meander development is explained by a river's tendency to establish a minimum channel slope for given input conditions (Knighton, 1998). He proposed that along a stretch of river with uniform water and sediment discharges but variable valley slope, a river can maintain a relatively constant gradient by lengthening its course where the valley slope is too steep (Knighton, 1998). Thus, meanders are effective at using up excess energy within the river system so that prevailing discharge can most efficiently transport the available load (Knighton, 1998). Nevertheless, it is still important to note that meandering streams are sites of net aggradation (Bridge, 2003).

2.3.3 General controls on the form and behaviour of channels

Schumm (2005) has identified a number upstream, downstream and fixed local and variable local controls on river morphology and behaviour, which ultimately determine the nature of landscapes shaped by fluvial activity. The majority of this section has been sourced from Schumm (2005).

Upstream controls include landscape history, tectonic events and altered relief, catchment geology, climate and climate change, and human activities; downstream controls include base level and channel length; fixed local controls include bedrock outcrops and their characteristics, interactions of the trunk stream with tributary streams, active local tectonic activity and valley morphology; and variable local controls include floods, vegetation and episodic events . As the downstream controls of base level have already been reviewed, only upstream and local controls relevant to the study are reviewed below.

Landscape history

Understandably, the landscape inherited from the past, formed under different climatic and tectonic conditions, has a large influence on the dynamics and morphology of most contemporary rivers. For example, many rivers in the northern hemisphere flow in broad gently sloping valleys that were carved out by retreating glaciers during the last ice age. As result, many contemporary rivers are forced to migrate within these broad valleys which exert a strong control on channel form. Similarly, ancient landscapes like those found in central southern Africa have been planed by past erosion cycles and existing rivers and streams are heavily influenced by the gentle topographic gradients inherited from the landscape (McCarthy and Hancox, 2000).

Tectonic activity and relief

Tectonic activity generates the relief responsible for the rejuvenation of energy gradients and the initiation of cycles of erosion (Tooth *et al*, 2004). These erosional periods during and succeeding periods of uplift are responsible for the initiation of steep headwater areas characterised by the delivery of vast amounts of sediment to downstream areas. As a result, tectonics and relief play an integral role in determining the quantity and type of sediment load provided to rivers and hence influence the form and behaviour of rivers downstream. For example, high relief mountains in cooler climates that are tectonically active often produce large quantities of coarse sediment, which in turn produce wide, high width-depth ratio, braided bedload rivers. In contrast, low relief stable mountains, especially in humid regions, deliver finer, more cohesive, sediment, which produces low width-depth ratio, sinuous or straight channels. It is important to note that the effect of tectonics and relief on sediment yields and hence upon the type of river found downstream is always modified by other upstream variables like catchment lithology and climate.

In addition to the upstream controls on river type created by uplift, rivers also follow structural lows or major fault systems. Melton (1959) cited in Schumm (2005) estimated that between 25 and 75% of all continental drainage in un-glaciated regions is influenced and/or controlled by tectonic processes and associated land movements responsible for earthquakes that deliver increased sediment yields to rivers downstream.

Catchment lithology

The type of bedrock exposed in the headwaters greatly influences the quantity and type of sediment delivered to channels downstream. Catchment lithology directly affects runoff and sediment yield such that for a given relief, sound rock will yield limited quantities of sediment and shattered rock large quantities of sediment. The type and amount of sediment delivered to a valley is a significant determinant of the type of channel found there as well as the morphology and behaviour of the channel. Thus, lithology exerts a strong control on the nature and amount of the sediment available for transportation and deposition (Knighton, 1998). Understandably rivers draining shale or siltstone will have a high percentage of silt and clay in their banks and beds and thus, should be narrow, deep and sinuous. Cohesive sediments have long been associated with channel stability and the presence of meander bends (Schumm, 1977). In contrast, rivers draining sandstone and gravel areas will have a high percentage of gravel and sand within their banks and beds and should be wider, shallower and straighter than their suspended sediment counterparts.

Rock resistance to erosion is also an important factor determining sediment yield and the type of sediment delivered to and controlling stream morphology. Resistant rocks generally produce less sediment than softer rocks.

Climate

Climate is extremely important in determining river type because it establishes the hydrologic character of the drainage basins, the vegetation which stabilises hillslopes and channel banks, and the quantity of sediment delivered to active channels. Firstly, climate is a major determinant of the river discharge, which plays an important role in determining channel form. The flow and sedimentary processes that operate during seasonal flood discharges generally have a large effect on the geometry of alluvial rivers (Bridge, 2003). Increases in discharge and sediment transport rate during exceptional floods may even change the normal channel pattern, which may be reinstated during subsequent flood periods (Bridge, 2003). On the other hand, during seasonally low flows, sediment transport capacity is low and modification of the channel pattern is considered to be minimal (Bridge, 2003).

Secondly, climate exerts a strong control on the type of vegetation present in the landscape, which in turn exerts a strong control on sediment yield. The relationship between natural rates of erosion and precipitation for continental climates reveals that natural maximum erosion rates generally occur in semi-arid regions. Even though humid regions produce the greatest discharges, the moist conditions are conducive to dense vegetation cover that protects the soils. Conversely, arid regions have little vegetation cover to protect soil but have limited discharge to remove the exposed soils. Thus, at an intermediate rate of precipitation vegetation protection is relatively low and there is sufficient runoff to remove sediment out of the drainage basin. Thus, like upstream lithology, climate exerts a strong control on the amount and type of sediment delivered to downstream river systems, and hence the type and form of the channels.

Bedrock and resistant alluvium

Rivers are strongly influenced locally by the presence of bedrock, resistant alluvium in river terraces, the presence of laterally impinging alluvial fans, colluvium and glacial outwash, that operate in various ways to form a local base level along portions of the trunk stream, which initiate large changes in river type and form (Tooth *et al*, 2002a). Bedrock resistance also determines valley width, channel width and the ability of the channel to shift laterally.

Floodplain and bank sediments also exert strong controls on channel behaviour. For example, it is obvious that sandy banks will erode more rapidly than clayey banks because of the cohesiveness of silt and clay banks and beds. Studies have shown that clay deposits and clay plugs influence meander migration and meander shape. When an alluvial meander encounters resistant sediments or bedrock, the downstream limb of the meander will be fixed in position and the upstream limb will continue to migrate, thereby deforming the meander. A river entering a region of resistant floodplain sediments such as backswamp deposits will develop stable bends of relatively high sinuosity.

Tributary inputs

The impact of small tributaries on the trunk river valley is usually small, perhaps causing minor valley widening and deepening. However, if the tributary is steep and large it can have a number of effects on the trunk river ranging from valley widening and deepening to a compete change of channel character. In particular, steep sediment-laden tributaries can have substantial impacts on the form and behaviour of trunk channels. For example, the introduction of high quantities of suspended load from incoming tributaries will likely alter the nature and cohesiveness of the prevailing bank and bed material of the trunk stream and can change a high width-depth channel to a narrower and deeper channel. In other examples, coarse alluvial fans deposited in the main river valleys by steep tributary valleys have been found to impound trunk streams and form local base levels that initiate significant changes in alluvial channel morphology (Tooth and McCarthy, 2007).

2.4 Geomorphological and sedimentological characteristics of inland wetlands

2.4.1 Floodplain Wetlands

Due to the climate and geomorphology of the country, South African floodplains are generally characterised by a single meandering or sinuous channel (Tooth and McCarthy, 2007). Meandering and sinuous channelled floodplains generally comprise an alluvial ridge on which the channel is located, and the floodbasin or backswamp, which is the lowest lying area on the floodplain that is presently isolated from channel activity (Bridge, 2003). Typical floodplain landforms associated with alluvial ridges include the meandering river channel, point bars on the inside and cut banks on the outside of the channel bends, oxbow lakes, meander scroll bars, levees, crevasse channels and crevasse splays (Wolman and Leopold, 1957; Leopold *et al*, 1964; Bridge, 2003). All floodplain landforms and their associated deposits can be generally categorised as in-channel, overbank and avulsion forms and deposits (Wolman and Leopold, 1953; Bridge, 2003; Slingerland and Smith, 2004).

In-channel forms and deposits

In-channel deposits include all sediment that is deposited within the confines of the channel and can be further subdivided into point bar and channel bed or lag deposits (Wolman and Leopold, 1957; Bridge, 2003). Channel bed deposits comprise coarse gravel and sand that is deposited on the channel bed during low flows (Nichols, 1999; Bridge, 2003). During high energy flows, these sand and gravel deposits are entrained and transported along the channel bed by rolling, saltation and/or suspension (Ritter, 2002; Bridge, 2003). Point bars comprise sediment deposited on the inside banks of meandering and sinuous channels (Nichols, 1999). A vertical cross section through a point bar generally exhibits a gradation from coarser material at the base to finer at the top due to the decreasing energy of the flow up to the surface of the point bar that is generally capped by fine overbank deposits (Nichols, 1999). South African floodplains are generally characterised by laterally dynamic meandering channels that are associated with the deposition of laterally extensive channel bed deposits and local point bar deposits (Nichols, 1999). Thus, laterally extensive channel deposits are generally a diagnostic property of floodplain conditions in South Africa (Tooth *et al*, 2002a; Grenfell, 2007).

Overbank forms and deposits

Overbank deposits consist of sediment deposited outside of the channel by flood discharges that overtop the banks of the river (Bridge, 2003). Overbank sediments generally comprise suspended loads of silt and clay, with limited sand sized material (Nichols, 1999). As water leaves the confines of the channel during floods, it spreads out and its velocity drops very quickly, resulting in the deposition of the suspended load, leaving only clay material in suspension (Nichols, 1999). The thickness of sediment deposited by major seasonal floods averages in the order of millimetres to centimetres (Bridge, 2003).

Sheets of sand and silt deposited during floods are thickest near the channel bank because of the abrupt reduction in sediment transport capacity of water leaving the channel, and clay usually accumulates as a more continuous blanket (Nichols, 1999; Bridge, 2003). Repeated deposition of silt and sand close to the channel edge leads to the formation of levees (Nichols, 1999), which are discontinuous, wedge shaped ridges of elevated ground that border channels and vary in height above the adjacent flood basin (Bridge, 2003). Levee deposits tend to be similar in facies to material at the top of the channel-bar deposits (Bridge, 2003). Often levees are breached during floods by ephemeral channels called crevasse channels (Nichols, 1999). These channels split downslope into smaller distributaries and deposit fan or lobe shaped mounds of sediment called crevasse splays (Bridge, 2003), which tend to be coarser grained and thicker than levee deposits (Bridge, 2003).

Flood basins or backswamps are the lowest lying areas on the floodplain and are located outside of and adjacent to the active meander belt (Bridge, 2003). Although most of the coarser overbank sediments are deposited on the alluvial ridge during floods, suspended sediment transported via overbank sheetflow and/or crevasse channels is often deposited within flood basin areas (Slingerland and Smith, 2004). The nature of flood basin environments is strongly controlled by climate and hydrogeomorphic setting: in wet climates there may be permanent lakes and marshes within the flood basins, whereas in semi-arid and arid climates floodbains are often characterised by ephemeral and seasonal inundation (Bridge, 2003). In addition, lakes and marshes are also often present in flood basins directly fed by springs and/or where the water table is shallow (Slingerland and Smith, 2004). As a result, backswamp sediments often comprise organic sediments and organic-rich clays and silts that reflect past and/or present palustrine marsh conditions (Rogers, 1997).

Avulsion

It has long been held that aggradation of floodplains is the product of overbank deposition during flood events, but it is clear that most of the aggradation on floodplains associated with overbank deposition is restricted to areas close to the active channel and that areas outside of the channel belt generally receive limited amounts of overbank sediments (Bridge, 2003; Slingerland and Smith, 2004). The continued increase in gradient between the channel belt and the surrounding floodbasins often results in a geomorphic instability that is periodically relieved by channel avulsion, leading to the deposition of sediment within the former lower lying flood basin areas (Slingerland and Smith, 2004). As a result, channel avulsion is now recognised as a dominant mechanism in basin filling and a key process in the vertical aggradation of floodplains (Collinson, 1996; Slingerland and Smith, 2004; Jones and Hajek, 2007).

“Avulsion is a process whereby flow is diverted out of an established river channel into a new course on the adjacent floodplain” (Slingerland and Smith, 2004: 259). According to Slingerland and Smith (2004) channel avulsion is a feature of aggrading floodplains that occur as a result of the lateral steepening of gradients between the alluvial ridge and floodbasins, which are necessary for the channel to shift. However, Tooth *et al* (2008) investigated channel avulsions on long-term non-aggrading floodplains in South Africa, and found that this is not always the case. Within these floodplains animal tracks result in the migration of knickpoints into the floodbasins that ultimately form new channels capable of diverting flow and sediment, thus resulting in an avulsion (Tooth *et al*, 2008).

Avulsion deposits

Avulsion generally results in the repositioning of the channel bed and associated deposits within floodbasins that were previously isolated from channel activity (Nichols, 1999; Slingerland and Smith, 2004). Frequent avulsion results in the occurrence of narrow channel deposits within the floodplain fill as there is less time for lateral migration to occur (Nichols, 1999).

Flow and sediment load are largely confined within the newly formed or reoccupied channels that arise by avulsion such that early post-deposition avulsion generally occurs within the channel areas (Slingerland and Smith, 2004). However, channelled flow will likely be associated with overbank flow such that overbank sedimentation will also occur in areas previously isolated from the channel itself (Slingerland and Smith, 2004).

Alluvial fans

Floodplains are also often characterised by alluvial fans deposited on the floodplain surface by incoming tributaries that abruptly lose confinement as they flow onto the floodplain (Grenfell, 2007). Alluvial fans are relatively gently sloping cone-shaped landforms that generally occur where valleys abruptly lose confinement and decrease their slope (Nichols, 1999). Alluvial fans are characterised by the abrupt deposition of sediment and spreading out of the incoming flow, usually as sheet flow (Nichols, 1999). Coarser material is deposited first and suspended load is carried to the distal reaches of the fan (Nichols, 1999). The sediments of alluvial fans are generally poorly sorted with a chaotic arrangement of clasts (Nichols, 1999).

Within the South African drainage network, alluvial fans have been found to have substantial effects on the processes and forms of floodplains as they often result in the impoundment and confinement of flow (Smith, 2006; Tooth and McCarthy, 2007; Joubert, 2009).

2.4.2 Valley-Bottom Wetlands

To date there has been limited research on the processes, forms and sedimentology of large valley-bottom wetlands in South Africa, particularly of those that are located within trunk river valleys (Tooth and McCarthy, 2007). South African valley-bottom wetlands are generally characterised by permanently and seasonally inundated marshes, with channels that often become discontinuous (Grenfell, 2007; Grenfell *et al*, 2008; Joubert, 2009). Common features of valley-bottom wetlands include variable degrees of organic sedimentation as a result of shallow water tables and the predominance of diffuse flow as well as the occurrence of ponds and depressions (Grenfell 2007; Ellery *et al*, 2009). The marshes of valley-bottom wetlands are similar in most respects with the various palustrine wetland types identified in the wetland classification systems developed by the United States Fish and Wildlife Service (USFWS) and the Ramsar Convention Bureau (Ewart-Smith *et al*, 2006). These include freshwater marshes, shrub-dominated swamps, peat swamps and swamp forests (Ewart-Smith *et al*, 2006). These wetland environments are generally characterised by subsurface and diffuse surface flow, low unit stream powers and sediment deposition leading to aggradation of the wetland surface over time (Ewart-Smith *et al*, 2006; Grenfell, 2007). As a result, valley-bottom wetland fill comprises mainly clay and fine silt, and often it comprises peat (Grenfell, 2007). However, limited coarse material, which is generally introduced into the marsh by sheetflow, may occur if a stream is present (Rogers, 1997; Grenfell, 2007), (Grenfell, 2007).

Channel forms and deposits

Valley-bottom wetlands hosting palustrine marshes and peat swamps are often characterised by small sinuous and straight channels that are often discontinuous (Grenfell, 2007; Watters and Stanley, 2007). Jurmu and Andrle (1997) have found that streams passing through palustrine marsh valley-bottom wetlands are often structurally distinct from alluvial channels because their bed and banks comprise mainly organic material. Whereas alluvial channels adjust geometry to efficiently transmit sediment and water, the form of ‘wetland’ channels have been found to be predominantly controlled by bank stabilising vegetation and shallow water tables (Watters and Stanley, 2007). The limited research on the geomorphology of wetland channels indicates that these channels are characterised by low width/depth ratios, rectangular cross sections and irregular planform geometries (Watters and Stanley, 2007). Though wetland channels are not as laterally dynamic as alluvial channels, they are still often able to transport and erode clastic and organic sediments (Watters and Stanley, 2007).

Like floodplain wetlands, valley-bottom wetlands are also often characterised by the presence of alluvial fans that either occur where there is a loss of valley confinement within the upper reaches of the wetland or where incoming tributary valleys lose confinement as they flow onto the valley-bottom surface (Grenfell, 2007).

2.5 Inland wetland formation and evolution in Southern Africa

Recent research has shown that the depositional settings that host wetlands along southern African rivers need to be understood within the context of the broad geological and geomorphological evolution of southern Africa, which has resulted in the creation of a set of geological and geomorphological controls that impede prevailing incision and drainage development of rivers (Tooth and McCarthy, 2007; Ellery *et al*, 2009). These local controls occur in the form of resistant lithologies that form stable local base levels along river courses that modify river morphology, valley planform geometry and longitudinal slope in a way that promotes wetland formation (Tooth *et al*, 2002a; 2004; Ellery *et al*, 2009). Despite recent research and increasing interest in wetland origin and evolution, geomorphological research on southern African wetlands still remains limited and relatively marginalized from a predominantly inventory and classification based approach to understand wetland dynamics and process (McCarthy and Hancox, 2000; Tooth and McCarthy, 2007). As a result, much wetland geomorphological theory is based on relatively few case studies (McCarthy and Hancox, 2000; Tooth and McCarthy, 2007).

Recent case studies on the geomorphology of important southern African wetlands have to date identified four unique local base level forming controls that initiate wetland formation and control wetland evolution:

- resistant dolerite dykes and sills have been found to impede vertical incision and initiate floodplain wetland formation in valleys upstream of the dolerite intrusions that are made up of softer sedimentary (Karoo Supergroup) rocks (Tooth *et al*, 2002a; 2004). This model applies along the Klip River on the South African Highveld (Tooth *et al*, 2002a), the Nsonge and Mooi Rivers in the foothills of the Drakensberg in KwaZulu-Natal (Grenfell, 2007), and the Blood River in north-western KwaZulu-Natal (Tooth and McCarthy, 2007);
- the deposition of alluvial ridges on floodplains have been found to impound incoming tributary streams and initiate floodplain-abutting valley-bottom wetlands adjacent to the

floodplains of the Mooi and Nsonge Rivers in the foothills of the Drakensberg in KwaZulu-Natal (Grenfell, 2007) and the Mfolozi Floodplain in Northern KwaZulu-Natal;

- alluvial fans and deposits from incoming tributaries have been found to form alluvial barriers that impound trunk river floodplains and initiate valley-bottom wetland formation along the floodplains of the Nyl River in Mpumalanga Province (Tooth *et al*, 2002b), the Blood River in northern KwaZulu-Natal (Tooth and McCarthy, 2007), and Wakkerstroom Vlei in the Mpumalanga Province (Joubert, 2009); and
- large scale subsidence associated with continental-scale fault systems associated with rifting in southern Africa, which have been found to negate the effects of uplift and create gentle basins within which extensive wetlands have formed along the Okavango Rivers (Haddon and McCarthy, 2005).

2.5.1 Resistant dolerite intrusions

During the break-up of the Gondwana supercontinent, the Karoo sedimentary strata that existed on the surface of South Africa were intruded extensively by dolerite sills and dykes, with extensive flood basalts forming over most of the South African land surface (Tooth *et al*, 2002a; Tooth *et al*, 2004). Following this event, landscape development has essentially involved downcutting by rivers and the backwearing of slopes as a result of the elevated topography inherited from the breakup of Gondwana and two epeirogenic uplifts that occurred 20Ma and 5Ma BP, which rejuvenated the drainage network of South Africa and initiated pulses of erosion and fluvial incision (Tooth *et al*, 2002a; Tooth *et al*, 2004; McCarthy and Rubidge, 2005). The consequent erosion over time has resulted in the removal of most of the flood basalts and a substantial amount of underlying Karoo sedimentary rocks, with the exception of the Lesotho Highlands (Tooth *et al*, 2002a; Tooth *et al*, 2004; McCarthy and Rubidge, 2005). This has led to the widespread superimposition of many South African rivers onto underlying dolerite dykes and sills, which crop out sporadically over an area covering nearly two-thirds of South Africa (Tooth *et al*, 2002a; Tooth *et al*, 2004). Owing to the persistence of the energy gradients created during the two uplift events, many rivers in the eastern interior of South Africa continue in a long-term state of incision, with their beds positioned on or close to bedrock (McCarthy and Hancox, 2000; McCarthy and Rubidge, 2005). This applies especially to the short steep river systems of KwaZulu-Natal that drain the Drakensberg and associated foothills (McCarthy and Hancox, 2000; McCarthy and Rubidge, 2005). Thus, lithological and structural controls are expected to greatly influence the character and behaviour of fluvial systems and regional drainage (Tooth *et al*, 2002a; Tooth *et al*, 2004).

Tooth *et al* (2002a; 2004) have developed a conceptual model that describes how resistant dolerite outcrops exposed along river courses influence river behaviour and control the formation and evolution of floodplain wetlands in the interior of southern Africa underlain by Karoo Supergroup strata. According to this model, a resistant dolerite outcrop exposed along a river course impedes vertical erosion in the less resistant Karoo sedimentary valley upstream and forms a stable local base level (Tooth *et al*, 2002a; Tooth *et al*, 2004). Because of this resistant lithology, the river upstream of the dolerite intrusion has excess energy to incise (Tooth *et al*, 2002a). Thus, with vertical erosion impeded upstream of the dolerite outcrop, the river adjusts by expending this excess energy by laterally eroding its valley (Tooth *et al*, 2002a; Tooth *et al*, 2004). Lateral erosion is accomplished by the formation of laterally migrating meanders, which both widen the valley and reduce the valley gradient through planing the bedrock valley floor over time (Tooth *et al*, 2002a; 2004). The widening and planing of the valley over time creates accommodation space for lateral and low level vertical accretion and associated floodplain wetland formation (Tooth *et al*, 2004; Tooth and McCarthy, 2007). In the long term, net vertical accretion on the floodplain is effectively negative due the gradual lowering of the valley floor and the reworking of floodplain alluvium over time (Tooth *et al*, 2002a; Grenfell, 2007).

Wetlands that are a product of this control are generally seasonal in nature and characterised by large backswamp areas, oxbow depressions and abandoned meander belts that are usually at a lower elevation on the floodplain than the meandering channel (Tooth *et al*, 2002a). These floodplain wetlands are predominantly fed by the overtopping of the main meandering channel during flood discharges, and the high clay content of overbank sedimentation in the backswamp areas effectively seals the backswamp surfaces, thus inhibiting infiltration and promoting the accumulation of standing water (Grenfell, 2007).

The channel beds on these floodplains are generally positioned on or just above bedrock and continue to slowly plane their bedrock floors over time (Tooth *et al*, 2002a; Tooth *et al*, 2004; Grenfell, 2007). As a result, the bedrock floor of these floodplains is relatively planar (Tooth *et al*, 2002a; Tooth *et al*, 2004). According to Tooth *et al* (2002a), meander formation in this setting has two main effects:

- it increases sinuosity, which reduces the erosive power of the river; and
- the lateral migration of the river results in the dissipation of excess energy by frictional resistance around bends, by lateral erosion of the bedrock channel floor, and by reworking of floodplain alluvium (Tooth *et al*, 2002).

Thus, meandering in this setting provides a way of ‘marking time’ while vertical erosion is impeded by the dolerite barrier (Tooth and McCarthy, 2007). Meanders in these settings are generally dynamic, with processes such as point bar deposition, cut-off formation and channel avulsion, giving rise to a diverse range of fluvial landforms (Tooth and McCarthy, 2007).

Sedimentation in dolerite controlled floodplains that have been studied by Tooth *et al* (2002a) and Grenfell (2007) is mainly clastic, with floodplain sediments generally fining upwards from basal gravels and sands above bedrock to fine organic-rich clays near the surface (Tooth *et al*, 2002a; 2004; Tooth and McCarthy, 2007; Grenfell, 2007). This fining upwards succession of the sedimentary fill is a result of the reworking of the floodplain alluvium and the lateral accretion of the point bar deposits as the migrating channel planes the valley floor over time (Tooth *et al*, 2002a; Tooth *et al*, 2004; Grenfell *et al*, 2008). Channel abandonment as a result of avulsion and/or meander cut-off maintains the lateral migratory dynamic of the channel and the reworking of floodplain alluvium (Grenfell, 2007; Tooth *et al*, 2008).

In the wetter oxbow and backswamp areas, sediments have a higher organic content near the surface owing to a low clastic sediment input, greater plant growth, and reduced rates of decomposition (Tooth and McCarthy, 2007). Peat formation within these floodplain wetlands is rare due to the active nature of the meandering channels, the regular input of fine clastic sediments across the floodplain, and seasonally wet conditions on the floodplain (Tooth *et al*, 2002; Grenfell, 2007).

2.5.2 Tributary impoundment by trunk stream alluvial ridges

A tributary stream cannot erode lower than the bed of the trunk river into which it drains (Leopold and Bull, 1979). The level of the trunk bed thus acts as the local base level of the tributary stream (Leopold and Bull, 1979). Studies on tributary-trunk interactions in the Wolumla catchment in New South Wales, Australia by Brierly and Fryirs (1999) have shown that in some cases the drainage of tributary streams and their associated wetlands is disconnected from the drainage of trunk streams due to the position of the trunk stream channel at tributary confluences. This suggests that in some cases the local base level of incoming tributaries may be elevated or lowered depending on geomorphic processes taking place within the trunk river (Brierly and Fryirs, 1999).

Using these ideas, Grenfell (2007) has established that many floodplain-abutting valley-bottom wetlands in the interior of KwaZulu-Natal in South Africa may be the product of the impingement of alluvial ridges on trunk river floodplains that serve to disconnect the drainage of tributary valleys from the aggrading, meandering trunk stream. In this section, details on floodplain abutting valley-bottom wetlands controlled by alluvial ridges are sourced mainly from Grenfell (2007).

Research on two floodplain-abutting tributary valley-bottom wetlands in the foothills of the Drakensberg by Grenfell (2007) has established that alluvial ridges on these large floodplains have impounded the drainage of incoming tributary valleys, raising the base levels of these tributary valleys. The formation of a new base level at a higher elevation by floodplain overbank accretion impounds the flow of tributary streams and initiates headward aggradation (backfilling) of the tributary valleys as the streams lose transport capacity and deposit their load. Headward aggradation gradually results in the reduction of tributary stream gradients and the formation of channelled and un-channelled valley-bottom wetlands upstream of the impoundment.

The wetlands that have been studied in these settings are generally located in the lower reaches of the tributary valleys extending from the permanently wet regions behind the alluvial ridges to a break in slope upstream, marking the extent of headward aggradation. These floodplain-abutting valley-bottom wetlands are predominantly fed by flow originating in the incoming or through-flowing tributary streams. Flow within the upper portions of the tributary wetlands is largely confined within a channel. In the lower reaches flow becomes diffuse as the tributary channel disappears and loses definition. Where streamflow is defined in the upper reaches of the tributary valley-bottom wetlands, these wetlands are generally fed by overbank flooding with limited groundwater and hillslope seepage inputs. Where the tributary streams are discontinuous and undefined in the middle reaches, the valley-bottom wetland is fed by unconfined sheetflow that spreads out across the wetland surface. Un-channelled valley-bottom segments in the lower reaches are generally fed by surface and subsurface diffuse flow from the upper reaches.

In the case of Stillerust Vlei, the tributary channel was found to undergo several changes through the length of the wetland (Grenfell *et al*, 2008). The tributary stream is initially a single-thread straight channel, which changes to a laterally stable single thread sinuous channel in the slightly less steep middle reaches (Grenfell *et al*, 2008). The channel then splits to become a multiple thread channel comprising many small intertwined channels as the channel bed gradient continues to decrease (Grenfell *et al*, 2008). As the channel bed decreases at the break in slope associated with the stream entering the floodplain-valley, the multiple-thread sinuous channel terminates in a series of small flooded depressions or ponds (Grenfell *et al*, 2008).

Little to no vertical or lateral erosion occurs within these tributary wetlands due to rapid reductions in stream energy and stream definition in the downstream direction, and, as a result, these wetland systems are generally depositional environments. Unlike dolerite controlled floodplain wetlands, the floodplain-abutting valley-bottom wetland channels are relatively inactive and are purely alluvial channels with no contact with the bedrock floor.

Sedimentation throughout the two floodplain-abutting valley-bottom wetlands studied in the Drakensberg is dominantly clastic with little evidence of peat formation. However, limited peat formation has been observed in the permanently flooded depressional settings that exist in the region of the floodplain-tributary valley-bottom transition behind the alluvial impoundment. These regions are characterised by permanently wet conditions and an absence of clastic sediment input, making peat accumulation inevitable. The majority of the valley fill in areas not originally carved out by laterally migrating meanders comprises uniform accumulations of clay above sedimentary rock, punctuated by channel bed deposits at variable depths (Grenfell *et al*, 2008). This sequence of sediments reflects the downstream loss of tributary stream power associated with stream impoundment due to aggradation associated with the trunk stream, which has resulted in the filling of the valley with fine overbank and sheetflow sediments (Grenfell *et al*, 2008).

2.5.3 Neo-tectonic subsidence associated with faulting

Apart from the two uplift events at 20 and 5Ma BP, the southern African subcontinent is tectonically stable and much of the interior has been subjected to prolonged planation (McCarthy and Hancox, 2000; Haddon and McCarthy, 2005). As a result, land surfaces in the interior are relatively old and mature with low topographic gradients (McCarthy and Hancox, 2000). Owing to the propagation of ancient rift and continental scale fault systems into southern Africa, some of these areas are experiencing neo-tectonic activity and have been characterised by neo-tectonic subsidence associated with various types of faults (McCarthy and Hancox, 2000). Where these fault systems have crossed river courses such as the Okavango River, the subsidence of the valley has created basins conducive to wetland formation (McCarthy and Hancox, 2000; Tooth *et al,* 2002b; Tooth and McCarthy, 2007).

In the case of the Okavango Fan, a graben fault system running perpendicular to the Okavango River has impounded the flow of the Okavango River (Haddon and McCarthy, 2005; Tooth and

McCarthy, 2007). Within the graben, subsidence has outstripped sedimentation and a large low gradient alluvial fan that hosts the extensive Okavango Fan has developed (Haddon and McCarthy, 2005; Tooth and McCarthy, 2007). The entire wetland system comprises a narrow low gradient entry corridor known as the Panhandle that enters a graben depression in which a large, broad alluvial fan has developed (Tooth and McCarthy, 2007). Owing to the abrupt loss of confinement of the Okavango River on entering the graben, the transport capacity of the stream abruptly decreases and the clastic sediment load of the Okavango River is deposited within the proximal reaches of the graben (Haddon and McCarthy, 2005). As a result, organic sedimentation dominates the portions of the graben where sediment free water accumulates (McCarthy *et al*, 1997). Up to 300m of sediment has accumulated in the depression but local topographic relief on the fan is generally less that 1.5m (Tooth and McCarthy, 2007). In the Panhandle the Okavango is an actively meandering, largely single thread river (Tooth and McCarthy, 2007). However, beyond the Gomare Fault it divides into a number of straighter more laterally stable sinuous distributary channels that disperse across the fan (Tooth and McCarthy, 2007). In both the Panhandle and the alluvial fan the channels are typically well defined with the channel banks made of peat with low amounts of clay (Tooth and McCarthy, 2007). Due to the permeability of the channel banks and the low topographic relief, water leaks from the channels and is widely dispersed throughout the fan, helping to promote and sustain the extensive permanent and seasonal wetlands (Tooth and McCarthy, 2007).

In contrast to the graben system associated with the Okavango Fan, a study on the geomorphology of the Nyl River floodplain by Tooth *et al* (2002b) concludes that the floodplain has formed within a basinal to synclinal structure on the downthrown side of the Zebediela fault within the Springbok flats (Tooth *et al*, 2002b). Instead of the deposition of large alluvial fans like that of the Okavango Fan, a large broad floodplain underlain by extensive alluvial deposits has formed within the depression (Tooth *et al*, 2002b). However, the structural control on the formation of the floodplain has recently come under scrutiny by Ellery *et al* (2009). An analysis of the longitudinal profile by Ellery *et al* (2009) indicates that the termination of the Nyl floodplain is not linked to the presence of geological structure or lithology. Rather the break in slope is associated with the location of two incoming tributaries (Ellery *et al*, 2009). Thus, Ellery *et al* (2009) conclude that the Nyl River floodplain is the product of the impoundment of the valley by tributary alluvial deposits, which has led to headward aggradation along the valley, giving rise to the present floodplain conditions (Ellery *et al*, 2009). In light of these findings, Tooth and McCarthy (2007) now attribute the origin and evolution of the Nyl River floodplain to a combination of both fault movement and partial blockage of the valley by tributary deposits.

2.5.4 Trunk River Impoundment by Tributary Alluvial Fans/Deposits

Studies on the geomorphology of the Nyl River and Blood River floodplains indicate that alluvial fans and coarse deposits introduced by tributary streams onto trunk rivers and their floodplains, often form alluvial barriers that impound the flow of the trunk channel, reduce upstream gradients, and initiate the creation of broad low gradient valley-bottom surfaces that host seasonal floodplains (Nyl River floodplain) and large un-channelled floodouts (Blood River floodplain) (Tooth, 2000a; Tooth *et al*, 2002b; Tooth and McCarthy, 2007). In arid central Australia, where many ephemeral streams decrease in size and eventually disappear, the alluvial surface beyond the channel termini are termed ‘floodouts’ (Tooth, 2000a; Tooth *et al*, 2002b). These floodouts are predominantly fed by seasonal floods that spill across the low gradient alluvial surface as shallow creeping sheetflow (Tooth *et al*, 2002b).

In the case of the Nyl River floodplain, two tributaries, the Rooisloot River and the Dorps River, drain large and elevated catchments to the east of the wetland (Ellery *et al*, 2009). A field investigation confirmed that the termination of the wetland is associated with the deposition of coarse deposits on the floodplain surface by two incoming tributary streams (Ellery *et al*, 2009). Over time these alluvial fans deposited on the floodplain by tributary streams have created barriers to flow resulting in the current floodplain-floodout system that exists today (Tooth and McCarthy, 2007; Ellery *et al*, 2009). However, Tooth and McCarthy (2007) maintain that subsidence associated with fault movement is also responsible for the current floodplain system and that this interpretation should not be overlooked.

Unlike the floodplains studied by Tooth *et al* (2002a; 2004), the Nyl River floodplain is an aggrading floodplain characterised by a sinuous alluvial channel that terminates in a broad floodout in the middle reaches of the floodplain (Tooth *et al*, 2002b).

The Blood River floodplain and associated wetlands are located in a low gradient sandstone/shale valley upstream of a dolerite sill that has impeded vertical erosion in the upstream valley (Tooth and McCarthy, 2007). Thus, the origin and evolution of the floodplain generally conforms to the model developed by Tooth *et al* (2002a; 2004). However, like the Nyl River floodplain, the main channel of the Blood River floodplain terminates in a large un-channelled floodout in its upper reaches. Like the Klip River floodplains, the Blood River floodplain is initially characterised by well defined levees, alluvial ridges and backswamps.

However, the channel rapidly declines in size downstream leading to an abrupt decrease in unit stream power and the resultant deposition of bedload sand and minor gravel (Tooth and McCarthy, 2007).

Both the sedimentation and the stratigraphy of the well defined meandering channel belts of the Blood River floodplains resemble those of the Klip River floodplains (Tooth and McCarthy, 2007). The floodouts on the other hand are underlain by clastic deposits with organic deposits dominating surface layers (Tooth and McCarthy, 2007). This is in contrast to the drier conditions of the Nyl River floodplain floodout that is characterised by seasonal inundation (Tooth *et al*, 2002b). Despite the differences, the Blood River floodplain floodout, like that of the Nyl River, is also characterised by sheetflows that carry silt and clays across the floodout (Tooth and McCarthy, 2007).

2.5 Conclusion

To date, research on the geomorphology of large floodplain wetlands within southern Africa indicates that the vertical and lateral accommodation space within the steep and incising drainage network of southern Africa required for the formation of floodplain wetlands is generally the product of either resistant dolerite intrusions which initiate the lateral planning of the less resistant valleys upstream as described in the conceptual model developed by Tooth *et al* (2002a; 2004), neo-tectonic subsidence associated with large fault systems that create basins within which sediment can be deposited and on which large rivers can meander, and alluvial barriers to trunk river flow due to deposition of sediment by incoming tributaries that initiate valley filling and the reduction of upstream gradients, which is conducive to the formation of aggrading floodplains and permanently wet floodouts. In addition, the downstream decrease in discharge associated with floodplains within the drier regions of southern Africa has also resulted in the formation of floodouts characterised by seasonal inundation.

In contrast, research on valley-bottom wetlands as defined by Kotze *et al* (2005) and Ewart-Smith *et al* (2006) is limited and to date only two floodplain-abutting valley-bottom wetlands have been studied from a geomorphological perspective. This limited research indicates that valley-bottom wetlands are the product of alluvial impoundments which initiate the downstream reduction in channel size and stream power, and the associated headward aggradation of the impounded valleys. However, it is important to note that the broad, low gradient valleys in which valley-bottom wetlands are located are almost always the product of laterally migrating meanders controlled by dolerite intrusions or neo-tectonic subsidence associated with large fault systems. Thus, valley-bottom wetlands usually represent an additional stage in the evolution of floodplain wetlands that are impounded by alluvial barriers. Alternatively, tributary valley-bottom wetlands may occur where tributaries are impounded by trunk stream floodplains, as described in the conceptual model of floodplain-abutting valley-bottom wetlands developed by Grenfell (2007). Thus, the origin and evolution of valley-bottom wetlands is generally more complex than floodplain wetlands, and is often the product of multiple controls operating at different temporal and spatial scales.

CHAPTER 3 - STUDY AREA

3.1 Locality

Dartmoor Vlei is a 42ha valley-bottom wetland system located on the Karkloof Plateau (Figure 3.1) in the KwaZulu-Natal Midlands, approximately 95km northwest of Durban and 25km east of Mooi River. The wetland is situated in the headwaters of the Myamvubu River, a right bank tributary of the Mooi River. The Myamvubu River rises on the highest portions of the Karkloof Plateau at an elevation of 1650m above sea level and flows in a north, north-easterly direction away from the Karkloof escarpment and plateau across the undulating KwaZulu-Natal midlands to its junction with the Mooi River at an elevation of 1100m. The head of the wetland is located approximately 1.5km to the north of the Karkloof escarpment.

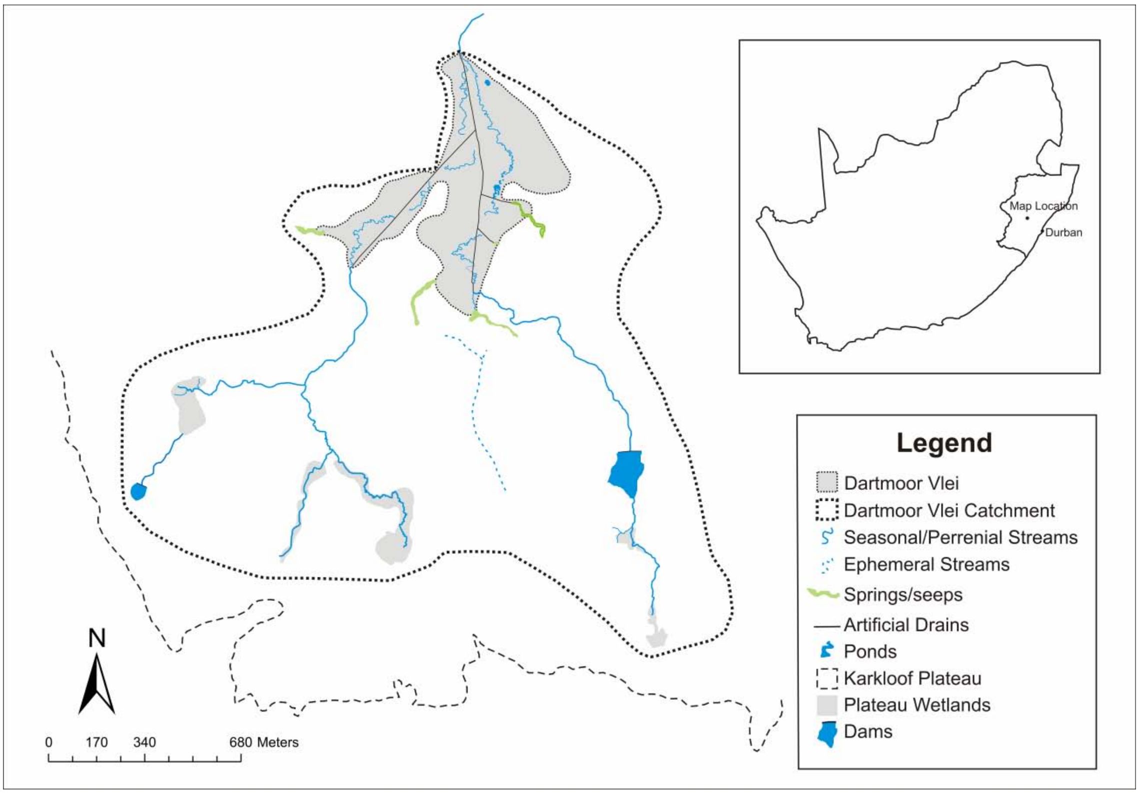


Figure 3.1: Dartmoor Vlei and surrounding catchment.

3.2. Climate

Dartmoor Vlei occupies approximately 11% of its 385ha catchment, which is located within quaternary catchment V20F. This quaternary catchment has an estimated median annual runoff of 112.4 million m^3, a mean annual precipitation of 868.4mm and a mean annual potential evaporation (A-pan equivalent) of 1649.1mm (Schulze, 1997). The area is a summer rainfall region with most precipitation occurring between October and March (Mucina and Rutherford, 2006). Heavy frost occurs in winter and occasional falls of snow have been recorded in the Karkloof plateau area (Begg, 1989). A mean annual temperature of 14.6ºC and 26 frost days per year are indicative of a cooler submontane form of warm-temperate climate (Mucina and Rutherford, 2006).

Climate data specific to the Karkloof plateau area indicate that the catchment of Dartmoor Vlei receives over 1000mm of rainfall on average every year. The majority of this rainfall occurs during summer and early autumn (October to March) as indicated in Table 5.1.

Table 3.1: Rainfall data for the Karkloof Plateau area (adapted from www.agis.agric.za)

Rainfall Period	Amount (mm)	
	Lower	Upper
Mean annual	1000	>1000
January	150	175
February	125	150
March	125	150
April	50	75
May	25	50
June	10	25
July	10	25
August	25	50
September	50	75
October	75	100
November	100	125
December	125	150

3.3 Topography and drainage

KwaZulu-Natal is bordered by the Drakensberg escarpment in the west and the Indian Ocean in the east. The province descends from the Drakensberg to the sea in a series of scarps and terraces such that the Karkloof escarpment and plateau form one of these scarps and terraces respectively (Bredell *et al*, 1980). The Karkloof Range forms the drainage divide between the Mgeni and Tugela catchments (Begg, 1989).

Topographically the Karkloof Plateau can be divided into the high lying western and eastern portions characterized by gently sloping ridges and hilltops and incised valleys, and the lower lying central portions characterized by broad gently sloping valleys separated by distinct dolerite ridges (Bredell *et al*, 1980). Dartmoor Vlei is located within the central topographical region, being situated in a broad gently sloping valley that terminates against a dolerite dyke ridge extending across the toe of the wetland in the north. Much of the plateau is characterised by broad terrestrial planation surfaces separated by dolerite ridge outcrops and depressions hosting wetlands. Bredell *et al* (1980), identify the Karkloof Plateau as one of the last remaining portions of the African Erosion Surface in KwaZulu-Natal that was created during the Cretaceous, some 144 to 66.4 million years BP.

Two short steep-sided mountain streams, rising on the highest portions of the Karkloof Plateau, drain into Dartmoor Vlei and abruptly lose confinement as they enter the broad valley hosting the wetland. At present the flow of both incoming streams has been diverted into artificial drains that have been dug along the entire length of the wetland as indicated in Figure 3.1. The eastern stream rises at an elevation of 1600m and flows over the plateau for approximately 1km before entering a steep and narrow dolerite boulder strewn valley that hosts the stream for approximately 700m. On entering the wetland the stream attains 2nd order. The western stream is the product of two small first order streams that rise at an elevation 1600m. These streams coalesce to form a single second order stream immediately before entering asteep-sided dolerite boulder strewn valley. Like the eastern stream, the western stream abruptly loses confinement as it enters Dartmoor Vlei. The eastern and western streams have 72ha and 129ha size catchments respectively. A third steep valley is evident between the western and eastern valleys, but flow within this valley is very limited for unknown reasons. In addition, a number of springs occur at the base of the valley sides that drain as diffuse flow into the wetland.

Rehabilitation interventions to counteract the effects of the artificial drains were undertaken by Eastern Wetlands on behalf of Working for Wetlands in 2004 and early 2005 (Kotze *et al*, 2009). The rehabilitation interventions consisted primarily of the establishment of nine concrete weirs and associated spreader canals down the length of the artificial drains in order to plug the drains, raise the water table, spread-out flow and promote diffuse flow (Kotze *et al,* 2009). Following the evaluation of the effectiveness of the rehabilitation structures in Dartmoor Vlei by Kotze *et al* (2009), it was found that the impact of the drains on the functioning of the wetland was moderate and that the wetland was likely not eroding prior to the establishment of the weirs and spreader canal (Kotze *et al*, 2009). Nevertheless, the weirs and spreader canals were found to be reducing water loss from the wetland and raising the water tables in the vicinity of the drains, thereby improving the health of the wetland (Kotze *et al*, 2009).

3.4 Geology

The Karkloof Plateau is underlain by the shale, siltstones and sandstones of the Estcourt Formation (Beaufort Group) and the shales and sandstones of the Volksrust Formation (Ecca Group) that have been extensively intruded by Jurassic dolerite sills and dykes (Bredell *et al*, 1980; Begg, 1989) (Figure 3.2). The headwaters of the Myamvubu River are almost entirely underlain by a large thick dolerite sill that forms the high lying topography of the Karkloof escarpment and plateau (Figure 3.2). The sill itself has been extensively intruded by a series of dolerite dykes that outcrop in the form of large dolerite ridges (Figure 3.2). The plateau capping dolerite sill has been subject to an extensive amount of chemical weathering with some lower lying valley-bottom basins comprising residual saprolite to depths in excess of 7m. The dolerite dykes, on the other hand are highly resistant and show limited amounts of chemical weathering. The dolerite is predominantly fine grained and well jointed (Bredell *et al*, 1980).

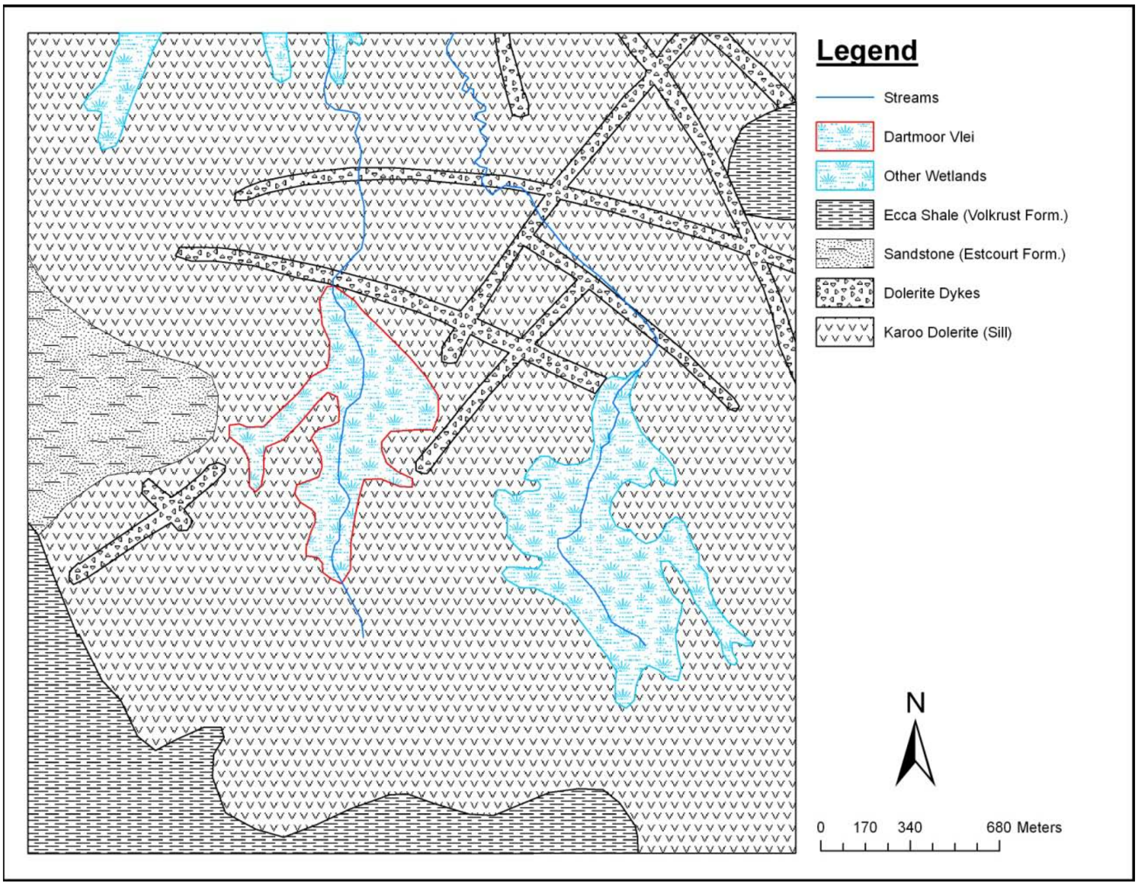


Figure 3.2: The geology of Dartmoor Vlei and the surrounding area (after Begg, 1989).

3.5 Vegetation cover

According to Mucina and Ruherford (2006), the Karkloof plateau area falls within the Sub-Escarpment Grassland Bioregion Unit Gs10, the Drakensberg Foothill Moist Grassland. This vegetation unit occurs across a broad arc of Drakensberg piedmonts covering the surrounds of Bergville in the north, Nottingham Road, Impendle and Bulwer in the north-east, and Kokstad, Mount Currie, Underberg and the surrounds of Mt Fletcher, Ugie, Maclear and Elliot in the south-west (Mucina and Rutherford, 2006). The region is characterised by forb-rich grassland dominated by short bunch grasses including *Themeda triandra* and *Tristachya leucothrix* (Mucina and Rutherford, 2006). The wetland falls within the Ezemvelo KwaZulu-Natal Wildlife (EKZNW) administered Karkloof Nature Reserve, on Dartmoor Farm, and the wetland is composed mostly of short grasses and sedges (Begg, 1989). There is a distinct zonation of plants in Dartmoor Vlei with the most hygrophilous species occurring in the moist, central parts and more mesic species occurring on the drier margins (Begg, 1989). The fringing hygrophilous communities appear to be composed mainly of grasses (*Setaria obscura, Aristida junciformis, Monocymbium sp*.) and various herbs, whilst in the central areas the vegetation comprises sedge-meadows that comprise *Cyperus sp., Carex cognata, Pycreus oakfortensis, Kyllinga melanosperma, Mariscus sp,* and *Juncus lomatophyllus* (Begg, 1989).

3.6 Land Use

Although the wetland and its catchment falls within a provincial nature reserve administered by Ezemvelo KwaZulu-Natal Wildlife, with the objective of protecting and managing for the purposes of biodiversity conservation, small numbers of domestic livestock occasionally forage in the catchment and upslope of the wetland. Nevertheless, almost the entire catchment of Dartmoor Vlei is under natural vegetation in a moderate to good condition (Kotze *et al*, 2009).

CHAPTER 4 - METHODS

4.1 Field work

4.1.1 Sediment sampling

Four transects were undertaken across the width of the wetland at intervals along the length of the wetland as indicated in Figure 4.1. Sedimentary fill sampling was undertaken systematically at points located along each transect in order to determine the nature and depth of the wetland fill, the depth and slope of the water table and the depth to bedrock. At each sample site the sedimentary fill down to bedrock was sampled using a gouge corer and an auger. Sediment samples were taken at 0.5m intervals and where a change in the texture and colour of the sedimentary fill was encountered. Each core was logged qualitatively in the field during the extraction of each core and the presence of peat was identified in the field by observation of fibrous material and by squeezing the material in the hand to determine the behaviour of the soil between the fingers and the colour of the water released. In addition, completely weathered residual saprolite underlying the sedimentary fill was also sampled until refusal. The sediment and residual saprolite samples were placed in plastic bags, sealed, labelled and transported back to the laboratory for analysis. The depth of the water table relative to the land surface was recorded using a tape measure and sounding device. Each sample point was then geo-referenced using a Trimble GPS with a remote base station (accuracy 1m in x, y and z fields).

Due to the highly variable and weathered geology underlying the wetland, samples of fresh dolerite outcropping within the wetland and on the lower slopes of the wetland valley were sampled systematically using a geological pick and a hammer. The sample locations are indicated with the prefix D in Figure 4.1. These samples were placed in plastic bags, sealed, labelled and transported back to the laboratory for analysis.

4.1.2 Cross-sectional surveys

In total, four cross-sectional surveys were undertaken within the wetland using a dumpy level (Figure 4.1). The sediment sampling sites were surveyed into each transect so that the subsurface characteristics of the wetland could be related to the surface characteristics. The relative elevations and distances of the survey points and the depths to the water table and bedrock were determined and plotted using MS Excel to show wetland cross-sectional and longitudinal morphology, slope and depth of the water table, and the morphology of and depth to the bedrock surface.

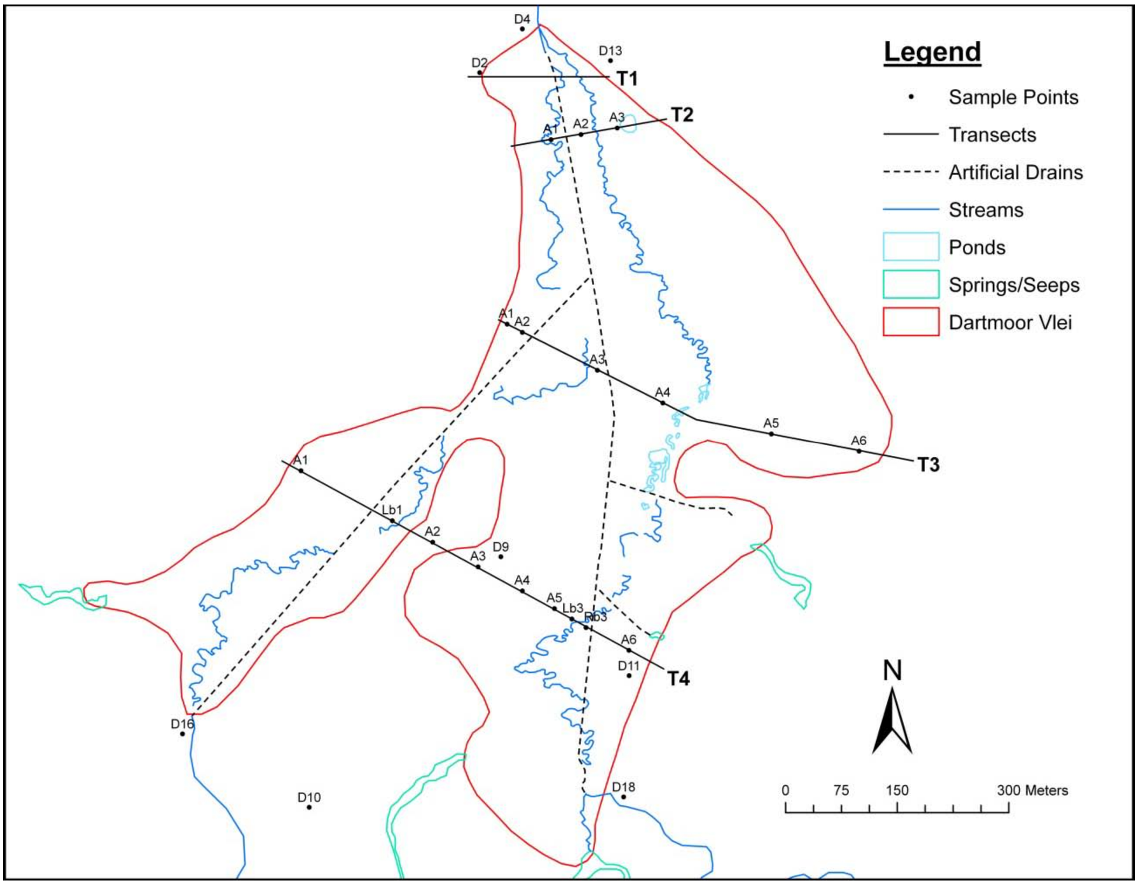


Figure 4.1: Location of surveyed transects (‘T’), coring locations (‘A’) and fresh dolerite sample points (‘D’).

4.1.3 Channel cross-sectional and longitudinal surveys

The cross-sectional geometry of the eastern stream channel was surveyed using a dumpy level. The channel cross-sections were systematically measured down the length of the stream at points between channel bends where the stream was relatively straight and where changes in channel geometry were obvious (Figure 4.2). The relative elevations and distances of the surveyed points were determined and plotted using MS Excel.

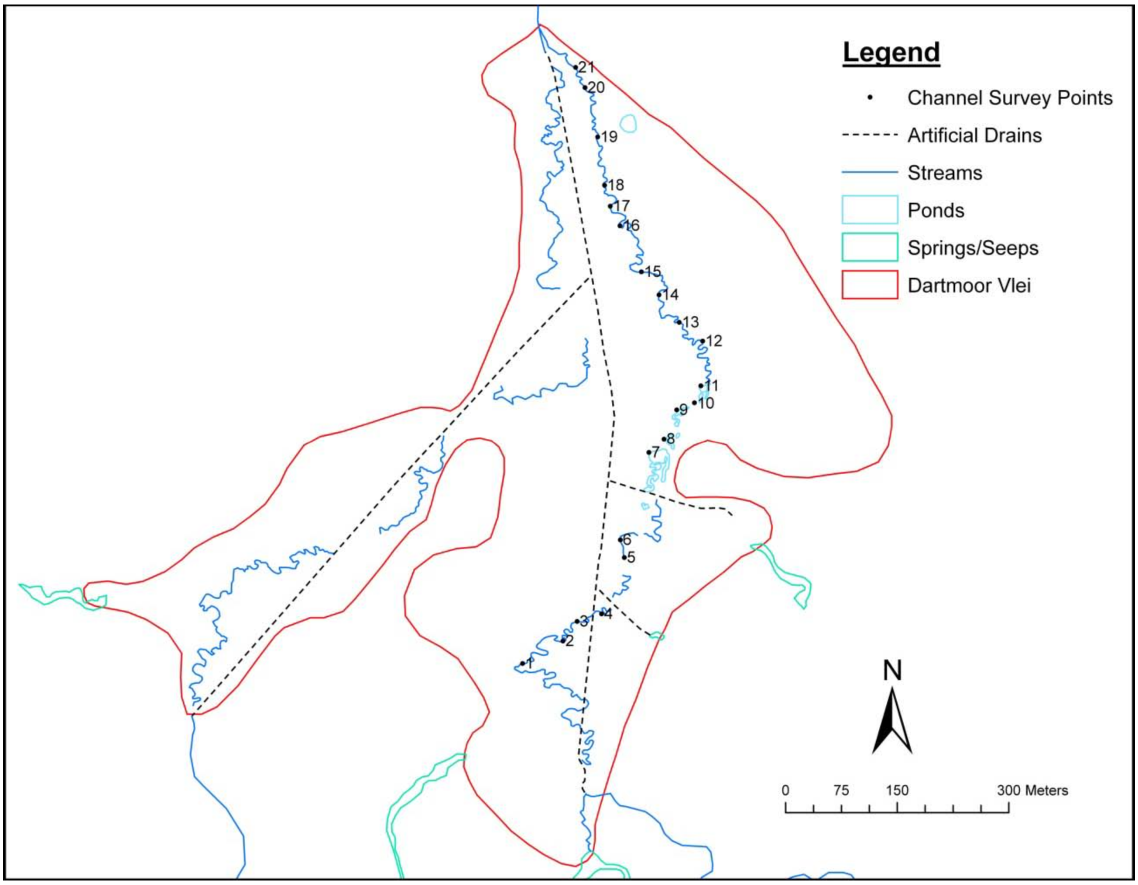


Figure 4.2: Location of individual cross sectional surveys of the eastern stream channel.

The longitudinal profile of the eastern stream bed was surveyed using a dumpy level. At each survey point the depth to bedrock relative to the stream bed was determined using a gouge corer, an auger and a measuring tape. The relative elevations and distances of the surveyed points were plotted using MS Excel. Once the longitudinal profile was constructed, geological features were superimposed on the longitudinal profile in order to establish relationships between changes in channel slope and geology.

4.1.4 Digital elevation mapping

Further detail on the morphology of the wetland surface was determined using differential GPS and a local base station (Myeza, 2007). The x, y and z data, which were accurate to approximately 10cm, were entered into the software package ‘Surfer’ and a digital elevation model was compiled.

4.1.5 Wetland hydrology

A broad understanding of the hydrology of the wetland was determined by collecting qualitative data on the following:

- water inputs into the wetland and their relative contributions in qualitative terms
- the nature of water movement through the wetland

- the volume of the inflow compared to the volume of the outflow in qualitative terms
- relationship between groundwater and surface water

Data on the abovementioned hydrological parameters was collected using a combination of field observation, aerial photographic analysis on ArcView GIS 9.2, cross-sectional water table characteristics and laboratory results of particle size analysis and measurement of loss-on-ignition at 450^{0}C. The field observation period extended from June 2006 to November 2008.

4.2 Laboratory analyses

4.2.1 Soil preparation

Each sediment sample was air dried for a week. Thereafter, each sample was crushed using a pestle and a mortar and placed in a labelled paper bag.

4.2.2 Organic matter content: Loss-on-ignition (LOI)

Approximately 40g of each sample was placed in a glass beaker and oven dried at a temperature of 105^{0}C for 4 hours to remove any residual moisture. The samples were then re-weighed and placed in a furnace at a temperature of 450^{0} C for 12 hours. After 12 hours, the furnace was turned off and the samples were left to cool down for an hour. Thereafter, the samples were placed in a glass desiccator. The samples were then re-weighed and the loss of mass was calculated as a percentage of the original mass of the oven dried sample to give the organic matter content of the sample. Percentage LOI with depth was plotted for each core using MS Excel.

Following a review of various peat classification schemes for temperate and tropical regions included in Wust *et al* (2003), and the soil classification system for South Africa developed by MacVicar *et al* (1977), a broad sediment classification scheme in terms of percentage organic matter content was developed to aid the description of the sediment samples. Due to the sub humid climate of the KwaZulu-Natal Midlands, peat soils for this study were generally classified as material comprising 30-100% organic matter content; peaty soils 15-30%; organic-rich soils 5-15%; and mineral soils 0-5%.

4.2.3 Particle size analysis

Due to the fine nature of the wetland sedimentary fill, particle size was determined using the Malvern Mastersizer 2000 laser based particle analyser. However, due to the high cost of using the Malvern Mastersizer, particle size was determined in only seven cores located throughout the wetland. These included cores A1 and A3 of Transect 2, Cores A3 and A4 of Transect 3 and

Cores A1, A2 and Rb3 of Transect 4. The choice of these specific cores for particle size analysis was based on their central location within the wetland as well as their distribution across different areas of the wetland considered to provide a representative sample of the different sequences of alluvial and organic fill that exist within the wetland.

In preparing samples for particle size analysis, organic matter was removed from the sediment samples using hydrogen peroxide. Approximately 20g of each sediment sample was placed in a 1000ml glass breaker and 20ml of hydrogen peroxide was added. The samples were then left to stand overnight at room temperature. The following day the sediment and hydrogen peroxide solution was placed on a hotplate to enhance the reaction. When the reaction ceased with the addition of further dilute hydrogen peroxide the beaker was removed and left to cool down for particle size analysis with the Malvern Mastersizer 2000.

Particle size distribution data for each sample obtained from the Malvern Mastersizer 2000 was used to calculate mean particle size, skewness and sorting values. Each sample was then classified in terms of mean particle size using the Udden-Wentworth scale and associated nomenclature (Gordon *et al*, 1992).

4.2.4 Chemical and mineral composition of the dolerite and residual saprolite

Due to the highly weathered nature of geology underlying the wetland, the chemical and mineralogical composition of the residual clay and fresh dolerite was determined in order to confirm the origin of the residual saprolite. The chemical and mineralogical composition of the fresh dolerite and residual saprolite samples was determined using X-ray fluorescence (XRF) and X-ray diffraction (XRD). Eight fresh dolerite samples, and thirteen residual saprolite samples from cores A3 of Transect 2, A4 of Transect 3 and A3 of Transect 4, were identified for XRF and XRD analysis. The location of the intact dolerite sample points is shown in Figure 4.1.

The CIPW method for calculating the mineral composition of rock samples from XRF data in Hutchison (1974) was used to determine the mineralogical composition for each fresh rock sample in terms of weight percentage, and a method based on molar concentrations and idealised chemical formulas was used to calculate the mass of kaolinite, quartz, feldspar and iron oxides that would constitute the residual clay samples. As the chemical composition of the fresh dolerite samples were very similar, an average chemical composition was calculated and used to determine the average mineralogical composition using the CIPW method. The weight percentage of each mineral for the averaged fresh dolerite samples and each residual saprolite sample calculated in terms of kaolinite, quartz and hematite was tabulated for comparative purposes.

The mineralogical composition of the fresh dolerite and residual saprolite was confirmed using XRD. An XRD pattern was obtained for each sample and the mineral identification was undertaken by Prof. J. N. Dunlevey from the School of Geological Sciences of the University of KwaZulu-Natal.

CHAPTER 5 - RESULTS

5.1 Valley morphology

Dartmoor Vlei is located within a broad, irregular valley that drains two short, steep headwater streams that enter the wetland from the south. Within the upper reaches of Dartmoor Vlei, the wetland comprises two arms that are located in two broad shallow valleys (Figures 5.1 and 5.2). These valley-bottom arms coalesce in the middle reaches of the wetland, downstream of which the valley gradually narrows towards the toe of the wetland (Figures 5.1 and 5.2). The narrowing of the valley correlates with a change in the geology of the eastern valley slopes, where a large dolerite dyke forms a ridge along the eastern valley side, extending from the middle to the lower reaches of the wetland. The western valley sides are steep in the upper and middle reaches and abruptly decrease in steepness and size in the lower reaches (Figures 5.1 and 5.2). The abrupt reduction in the height and slope of the western valley sides near the head and the toe of the wetland correlate with the occurrence of a broad gently sloping valley that enters the existing wetland valley from the west (Figures 5.1 and 5.2; T1 and T2). This broad valley lies approximately 1-2m above the Dartmoor Vlei surface (Figure 5.2; T1 and T2). The valley entering the wetland from the west in the lower reaches forms part of an extensive terrestrial planation surface that is located behind another dolerite dyke north of the dolerite dyke ridge that cuts across the toe of Dartmoor Vlei. This planation surface is largely terrestrial and the present drainage is largely confined to the lower lying valley-bottom wetlands like Dartmoor Vlei. Beyond this point, the confinement of the valley and the termination of Dartmoor Vlei correlate strongly with the occurrence of a dolerite dyke ridge that forms the eastern slopes of the wetland and cuts across its toe to the north of the wetland.

5.2 Wetland morphology

5.2.1 Upper reaches

In the upper reaches (Figure 5.2; T4) there are three distinct valley-bottom surfaces comprising a western valley-bottom arm hosting permanent wetland, an elevated ridge hosting terrestrial and hygrophilous grassland, and an eastern valley-bottom arm at a lower elevation than the western arm, hosting permanent wetland. The western valley-bottom arm is characterised by two relatively flat surfaces (A; Figure 5.2; T4) separated by a low ridge (B; Figure 5.2; T4). Within this arm the stream is located on the eastern flank against the ridge (C; Figure 5.2; T4) that separates the two arms, and an artificial drain has been created west of the stream. The elevated ridge is approximately 0.5 to 1m higher in elevation than the wetland arms. The eastern valley-bottom arm is relatively flat with the exception of a small terrace-like feature (D; Figure 5.2; T4) to the east of the stream channel. Conditions on top of this terrace feature are substantially drier than the lower lying permanently wet areas, and vegetation cover predominantly comprises *Aristida junciformis*. An artificial drain has been established immediately to the east of the channel and the western bank of the channel comprises a low, rounded ridge (E; Figure 5.2; T4). Both the terrestrial elevated ridge and terrace-like feature within the upper reaches of the wetland were observed to be characterised by localised dolerite outcrops.

5.2.2 Middle reaches

In the middle reaches (Figure 5.2; T3) the elevated ridge disappears and the two distinct valley-bottom arms become a single valley-bottom surface hosting extensive permanent wetland and two discontinuous meandering channels that comprise a chain of ponds. Two artificial drains are also present. The valley-bottom surface in the middle reaches comprises an extremely flat surface within the western and central portions of the reach, to the east of which is a gently sloping west-facing slope (G; Figure 5.2; T3) that is morphologically distinct from the flat western portion of the wetland. This gently-sloping surface reflects the steepest gradient within the valley-bottom wetland and hosts both permanent and seasonal wetland. The only departure from the prevailing flat morphology is the occurrence of a small depression located immediately to the east of the eastern stream channel (H; Figure 5.2; T3) and an elevated spur that extends into the wetland from the east (Figure 5.1).

5.2.3 Lower reaches

In the lower reaches the gently sloping valley-bottom surface in the east disappears and there is a single relatively flat valley-bottom surface characterised by depressions along both the western and eastern flanks of the wetland surface (I; Figure 5.3; T2). The western stream is located within the western depression and converges with the eastern stream channel near the toe of the wetland (Figure 5.2; T1). The western stream increases in width and depth towards the toe (Figure 5.3; T1). Like the western stream channel in the middle reaches, the eastern stream in the lower reaches has an elevated right bank (Figure 5.2; T2). However, further downstream the difference in the elevation of the channel banks is not evident.

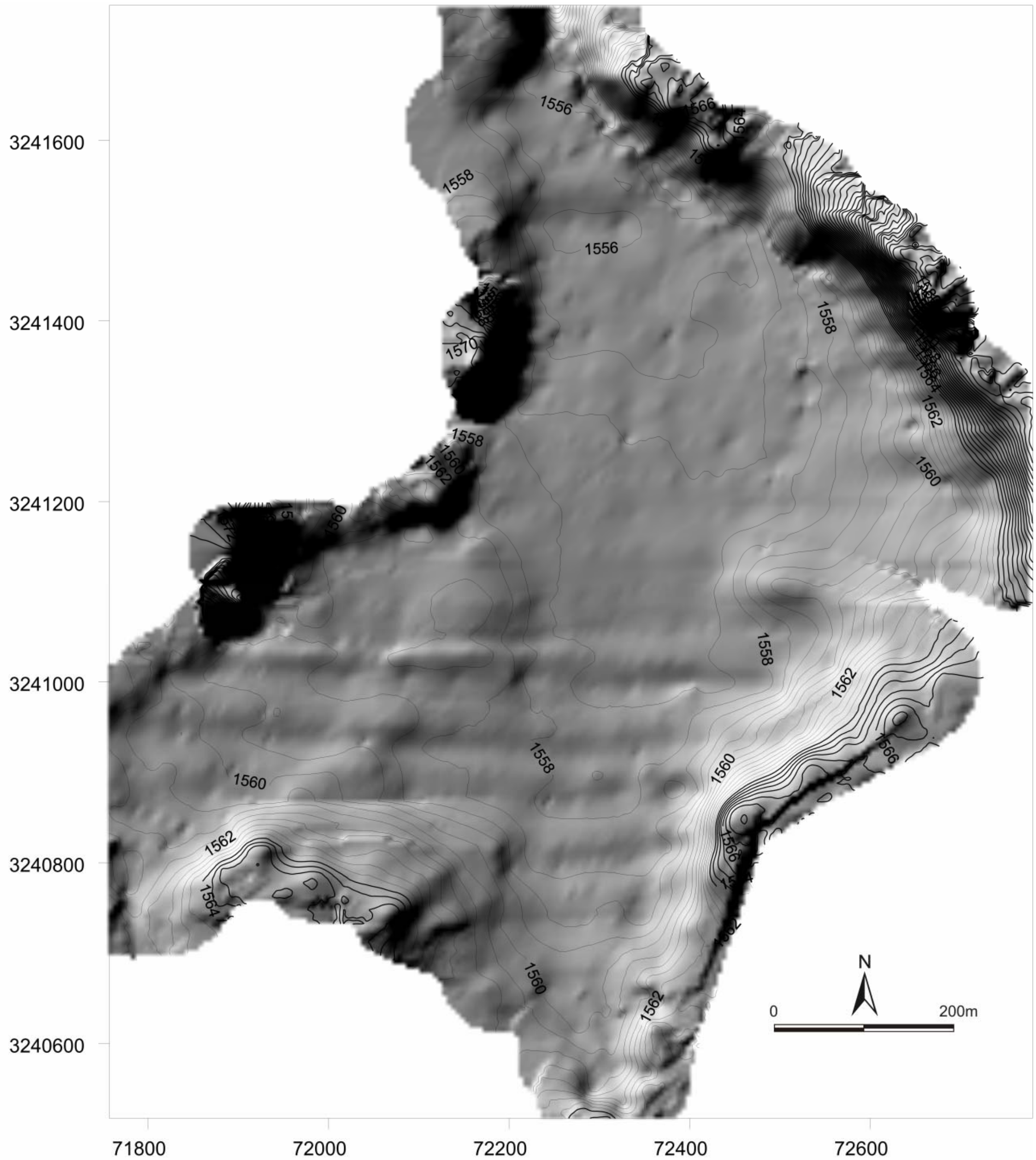


Figure 5.1: Digital elevation model of Dartmoor Vlei.

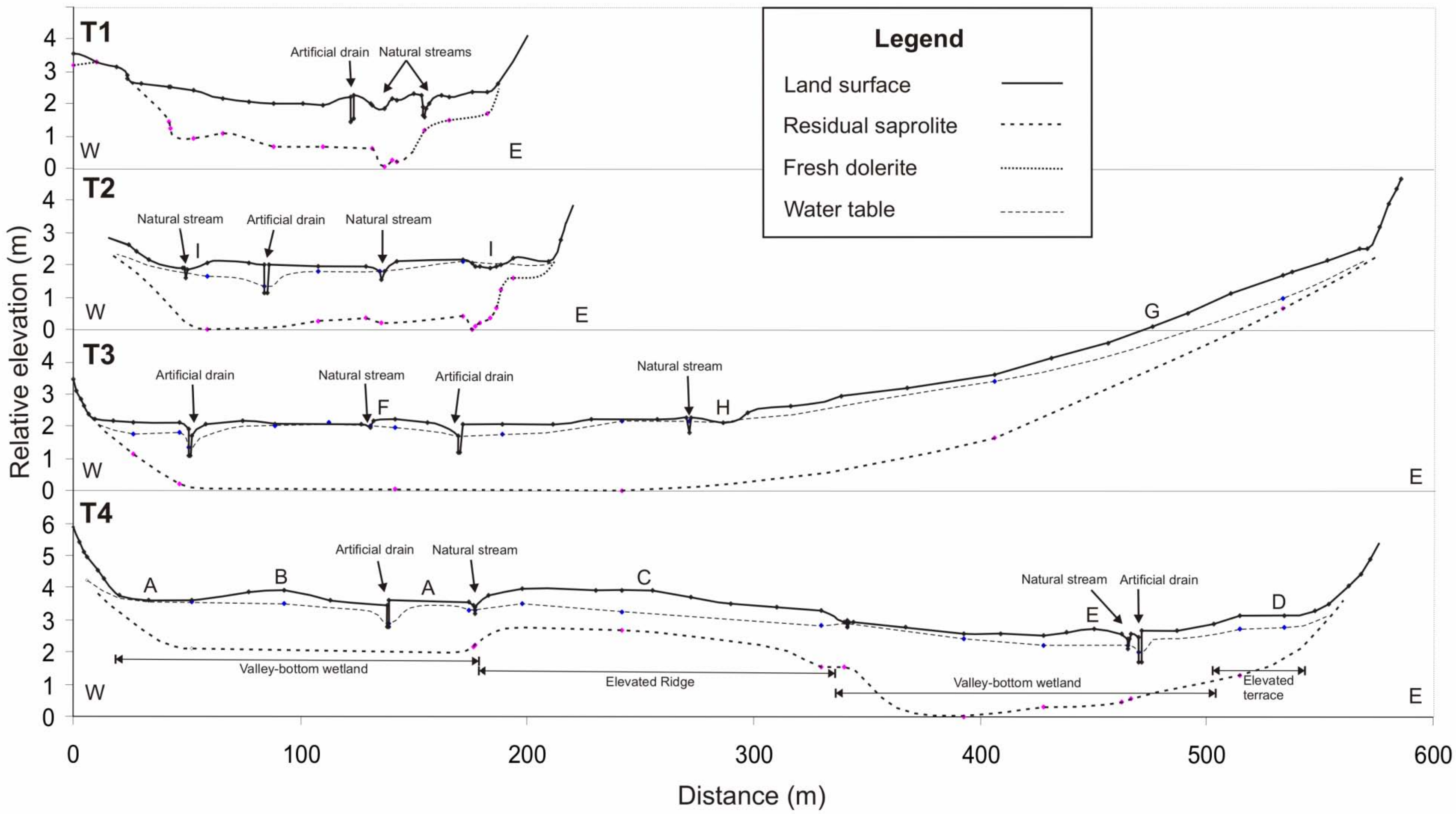


Figure 5.2: Cross-sectional profiles of the wetland, water table and bedrock basin surface

5.3 Wetland hydrology

5.3.1 Water inputs and outputs

Dartmoor Vlei is predominantly fed by two incoming streams that enter from the south-west and south-east and by a number of springs that occur along the lower slopes of the valley in the upper and middle reaches. Based on personal observations, these streams are seasonal with high discharges during the summer rainfall period, while during the winter months flows within the steep valleys above Dartmoor Vlei are reduced to a gentle trickle.

Outflows from the wetland are predominantly via streamflow at the termination of the wetland, and there is likely to be considerable water loss from the wetland due to evapotranspiration. At present a dam wall has been constructed across the toe of the wetland, which is reducing the amount of water lost to the stream channels below the dam. Outflows from the wetland are most substantial in summer, with low flows being maintained through winter.

5.3.2 Water movement within the wetland

Upper reaches:

Flow within the western arm is predominantly concentrated within the lower lying areas on either side of the ridge and within the artificial drain. Despite the existence of the drain, the water table of the wetland arm during the summer period of 2006 was relatively high and surface water was present. To the west of the ridge the water table was at the surface and to the east of the ridge the water table was slightly lower. The stream channel located against the ridge feature had a minimal effect on the water table. Field visits confirmed that flow within the western arm is strongly diffuse while flow within the stream was minimal. The extremely low flows within the natural channels are partly the result of the diversion of streamflow into the artificial drain that cuts across the natural stream course. As a result, the stream channels are now predominantly fed by the local water table rather than from influent streamflow from the catchment. Currently, the major input into the western arm is from the spring located at the intersection of the wetland with the east-west aligned valley entering the upper reaches of Dartmoor Vlei.

Within the western flank of the eastern arm the water table during the summer period of 2007 was at the surface and gently sloped towards the artificial drain that is currently drawing down the water table within the arm. Like the western arm, flow within the eastern arm is strongly diffuse with limited flow in the natural channel. This is partly because the artificial drain intersects the natural channel and has diverted flow away from it. As a result, the channel is now fed by the local water table rather than from channelled flow from upstream. In addition, a small channel has also formed along the western flank of the valley-bottom arm against the elevated ridge between the western and eastern valley-bottom arms. Flow within the channel is limited and the water level within the channel reflects the elevation of the water table. At present, the eastern arm is predominantly fed by a number of springs that discharge into the wetland in the upper reaches. Incoming streamflow is largely concentrated within the artificial drains and does not contribute to the water level in the upper reaches.

Middle reaches:

The water table was generally located at or in the vicinity of the land surface except in the vicinity of the artificial drains, which have lowered the local water table. Within the sloping portion of the wetland, the water table is located just below the surface and increases in depth eastwards as the elevation of the wetland increases. Flow within the middle reaches is strongly diffuse and stagnant in places where a series of ponds occur. Currently, flow within the natural streams is limited and both stream channels are discontinuous. The water level within the channels generally reflects the depth of the local water table. The lower lying areas of the wetland are predominantly fed by diffuse flow from the upper reaches and by seepage from a spring that occurs immediately south of the elevated spur that extends into the wetland from the east. The lack of influence of the streams on the movement of water in the middle reaches is illustrated by the fact that the water table elevation remains uniform in the vicinity of the channel.

The lack of flow within the western channel is partly the result of the diversion of channelled flow into the artificial drain that repeatedly cuts across the natural channel in the middle reaches. However, the greatly reduced depth and width of the western channel indicates that natural processes are also responsible for the reduction of flow within the channel. The series of ponds correlate with the occurrence of the small elevated spur that extends into the wetland from the east, and the occurrence of a spring that currently feeds the ponded area. The higher lying eastern slopes of the valley-bottom surface are fed by hillslope seepage from upslope that is concentrated near the surface.

Lower reaches:

Figure 5.2 shows that the artificial drain is currently drawing down the water table and draining water from the wetland because the elevation of the water table is generally lowest in the vicinity of the drains. The water table recorded in the winter of 2006 was relatively uniform in the vicinity of the natural streams and flow within this reach was predominantly diffuse with a small amount of channelled flow occurring within the eastern stream. Within the eastern flank of the wetland, the water table was within 0-10cm of the surface and surface water was observed within the eastern depression of the wetland. At present, a dam wall has been constructed across the toe of the wetland and there is a moderate amount of ponding behind the dam wall. The dam wall was constructed to reduce the loss of water diverted within the artificial drains.

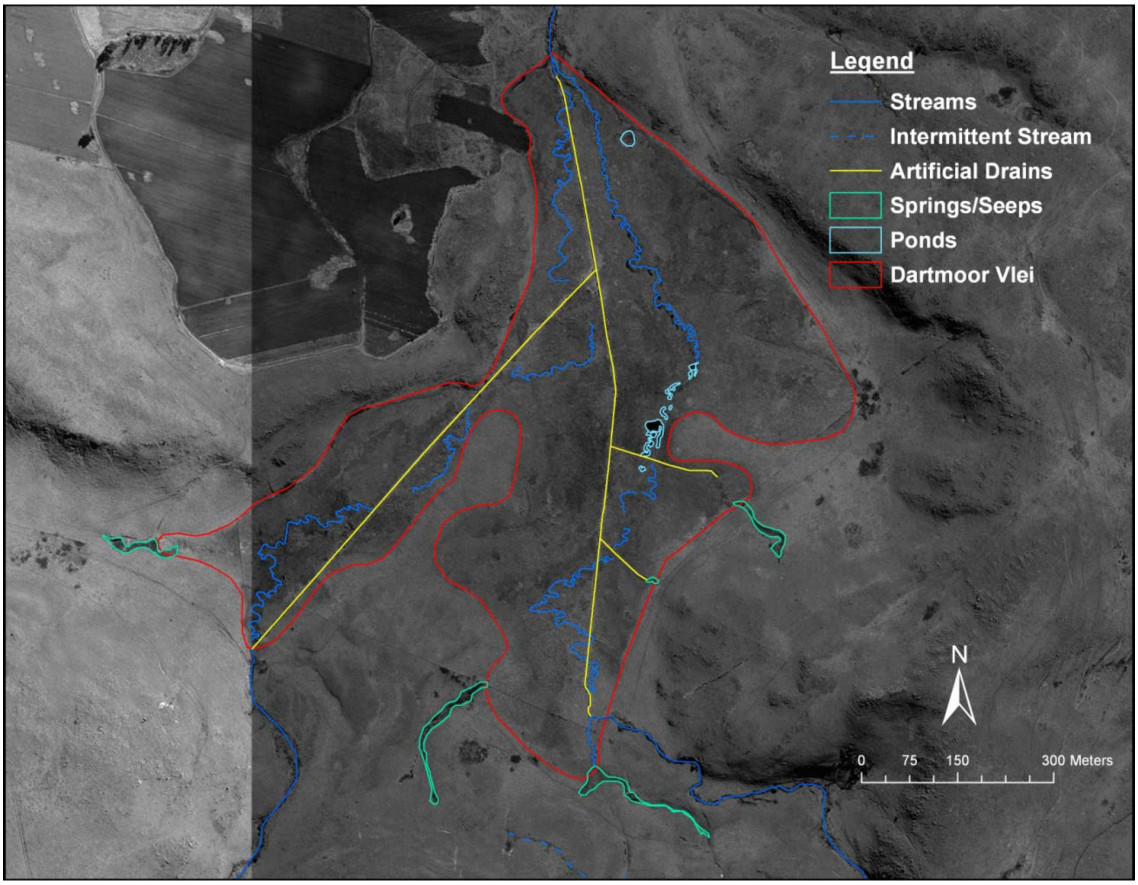


Figure 5.3: Aerial photograph of Dartmoor Vlei indicating hydrological features observed during field visits.

5.4 Downstream Variations in the Eastern Stream Channel Geometry

At present, stream flow from the incoming eastern stream valley has been diverted into a large longitudinal drain that has been dug down the length of the wetland. This drain repeatedly cuts across the path of the natural sinuous and irregular meandering eastern stream within the upper reaches of the wetland and has had a significant effect on the flow within the stream. Nevertheless the existing eastern channel within undisturbed areas of the wetland is more or less distinct and still retains its original morphology.

5.4.1 Cross-sectional and plan form channel characteristics

The channel undergoes a wide range of changes from upstream to downstream (Figure 5.4). The upper reaches of the eastern valley-bottom arm are characterised by a small, well defined parallel-sided channel with an irregular meandering plan form. The channel is roughly 1m wide with a 0.8m high vertical right bank and a 0.3m high vertical left bank from the channel bed. Over the next 100m the channel gradually becomes shallower and narrower and eventually becomes discontinuous and choked with vegetation. Within this portion of the stream course, the channel repeatedly loses and regains definition, alternating between a wide basin characterised by surface and subsurface diffuse flow and a more defined, shallow, vegetation-choked stream. This discontinuous flow path terminates in a series of ponds that are separated by un-channelled surfaces characterised by diffuse flow. Firstly a large, circular pond, approximately 24m in diameter and 1m deep occurs followed by alternating diffuse and ponded areas. The series of ponds extends for approximately 200m downstream before a discontinuous meandering channel starts to re-form. Beyond the series of ponds, the re-forming meandering channel is initially poorly defined and is more of a drainage path roughly 5m wide and shallow. This channel extends for approximately 200m downstream before splitting into two narrow defined channels. In the lower reaches, the channel gradually narrows and deepens becoming more defined. The channel eventually becomes U-shaped and roughly 1.5m wide and 1m deep, indicating an increase in discharge within the channel in the lower reaches that coincides with the narrowing of the valley.

The cross-sectional channel profiles indicate that there are roughly five distinct fluvial zones within the wetland. These include a zone of small well defined meandering channels occurring in the upper reaches, a zone of discontinuous meandering channels characterised by diffuse and sheet flow occurring in the middle upper reaches, a series of ponds occurring within the middle reaches, a zone of discontinuous meanders occurring within a diffuse flow path depression in the middle to lower reaches, and another zone of defined meandering channels in the lower reaches. These zones are referred to as Zone 1 to 5 respectively (Figure 5.5).

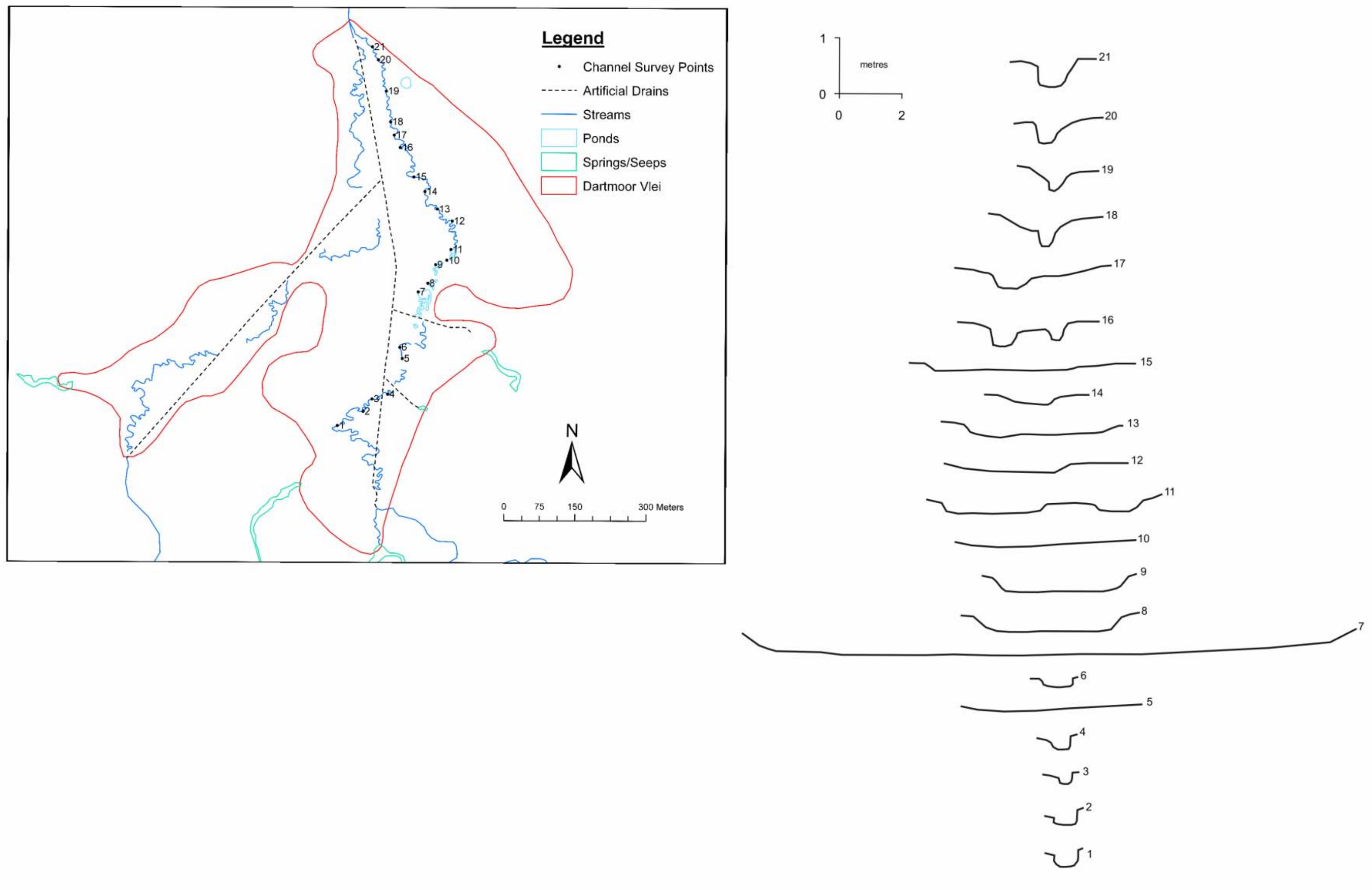


Figure 5.4: Cross-sectional profiles of the eastern stream channel

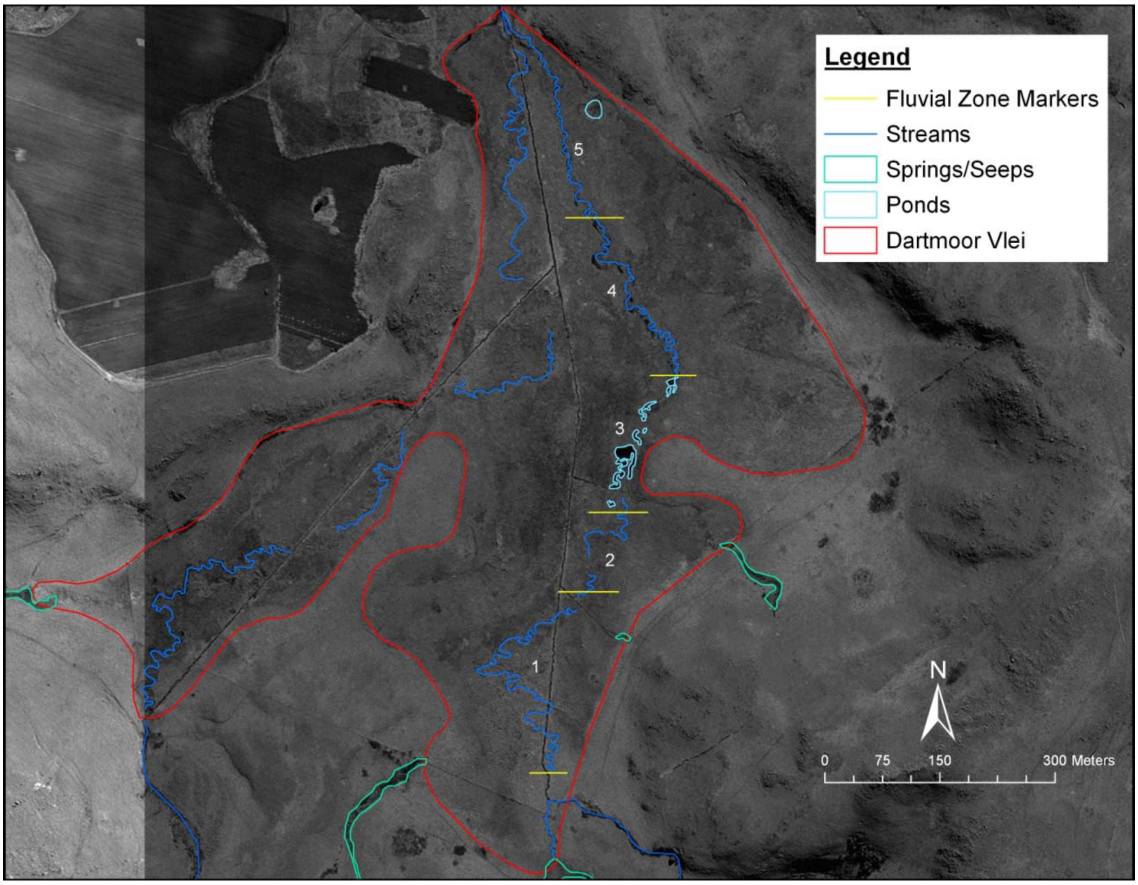
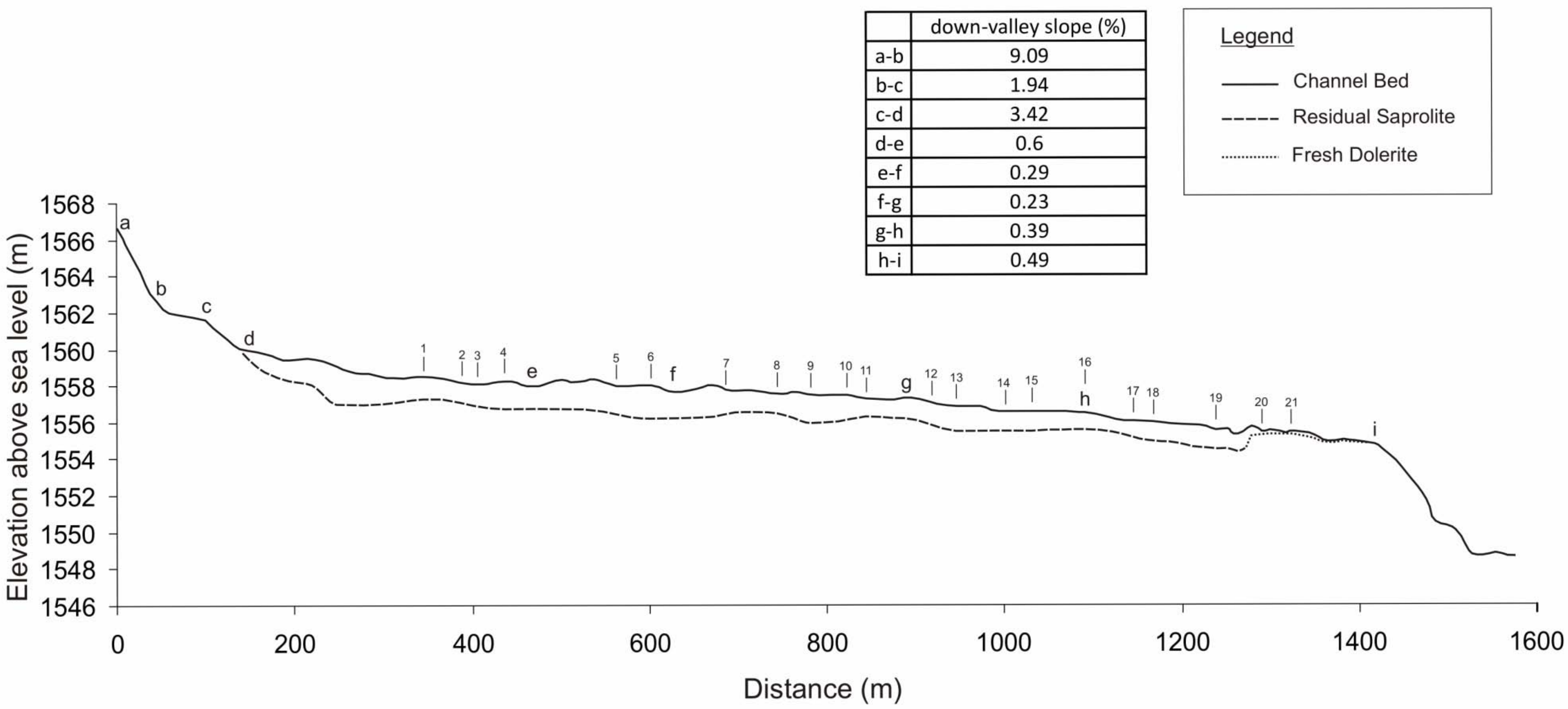


Figure 5.5: Aerial photograph of Dartmoor Vlei indicating approximate fluvial zones.

5.4.2 Longitudinal Stream Bed Profile

The stream bed slopes steeply down from the higher lying areas of the Karkloof plateau, where it is confined in a steep sided dolerite boulder strewn valley, the bottom of which is reflected in Figure 5.6. A break in slope occurs at approximately 50m along the longitudinal profile associated with the widening of the valley and a loss of confinement of the stream. Here the stream gradient changes from 9.09% to 1.94% as indicated by the slope b-c in Figure 5.6. Thereafter, the stream bed gradient steepens for a short distance to 3.42% (slope c-d) and then flattens out abruptly as the stream enters the broad valley-bottom wetland. On entering the wetland the stream changes from a relatively straight bedrock stream to a meandering stream with its bed elevated 1-2m above the underlying residual saprolite surface.

The longitudinal profile of the stream bed within the wetland indicates that there are roughly five sloping surfaces that correlate well with the five fluvial environments mentioned earlier. In the upper reaches the average slope of the stream bed is 0.6% extending from the head of the wetland at 150m along the longitudinal profile to the upper-middle reaches at 460m along the longitudinal profile (slope d-e). This corresponds with the region where the channel is small and well defined in the upper region of the wetland, as indicated by cross-sections 1 to 4 in Figure 5.4. At 460m along the profile the stream bed gradient decreases further to a gradient of 0.29% (slope e-f), corresponding with the change in the stream morphology from well defined to shallow and discontinuous as indicated by cross-sections 5 and 6 in Figure 5.4. At approximately 610m along the longitudinal profile the gradient of the stream bed decreases slightly to 0.23% correlating with the occurrence of a series of ponds that occur from 610m to 840m along the profile as indicated by cross-sections 7 to 11 in Figure 5.4 and by slope f-g in Figure 5.6. At approximately 840m along the longitudinal profile the stream bed gradient increases to 0.39% (slope g-h) correlating with the disappearance of the ponds and the reformation of a shallow discontinuous meandering channel that alternates between a channel and a shallow diffuse flow path. Cross-sections 12 to 15 in Figure 5.4 illustrate this variation. At approximately 1100m along the longitudinal profile the stream bed gradient again steepens to a gradient of 0.49% (slope h-i) correlating with the reformation of a well defined meandering channel as illustrated by cross-sections 16 to 21 in Figure 5.4. This stream bed gradient is maintained until the confinement of the stream in the steep-sided dolerite dyke valley, where the wetland disappears and stream gradient abruptly increases. The average gradient of the stream bed throughout the wetland is 0.4% as illustrated by slope d-i in Figure 5.6.



	down-valley slope (%)
a-b	9.09
b-c	1.94
c-d	3.42
d-e	0.6
e-f	0.29
f-g	0.23
g-h	0.39
h-i	0.49

Figure 5.6: Longitudinal profile of the eastern stream channel bed with underlying geology

5.5 Bedrock Basin Morphology

5.5.1 Cross-sectional Bedrock Basin Characteristics

The wetland is predominantly underlain by residual saprolite that is a product of the deep weathering of the dolerite sill valley, with the toe of the wetland being underlain by relatively fresh dolerite bedrock. Like the valley-bottom surface morphology, there is variation in the bedrock (residual saprolite) morphology from the upper to lower reaches of the wetland (Figures 5.2 and 5.6). Depth to residual saprolite is typically about 1.5 to 2m, which declines greatly at the toe of the wetland, such that there is a near-vertical discontinuity in the depth to bedrock at the toe of the wetland. Depth to the dolerite bedrock underlying the toe of the wetland is typically about 0.5 to 0.8m.

The cross-sectional profile of the residual saprolite surface underlying the wetland in the upper reaches indicates that the two valley-bottom wetland arms in the upper reaches have formed within distinct valleys separated by a ridge of bedrock (residual saprolite) (Figure 5.2). The residual saprolite surface is near-planar within the western valley and is asymmetrical in the eastern valley arm, sloping gently upward toward the east. The ridge that separates the two valley-bottom arm basins generally follows the profile of the ridge on the surface of the wetland.

Like the valley-bottom surface profiles, the elevated ridge that separates the two valley basins disappears at the joining of the two arms. The residual saprolite surface in the middle reaches is near planar in the western and central-western portions, with a large, gently upward sloping surface towards the east that reflects the eastern slopes of the valley-bottom surface (Figure 5.2; T3).

The cross-sectional profiles of the bedrock surface in the lower reaches (T1 and T2; Figure 5.2) clearly indicate that there is a large discontinuity at the contact between the residual saprolite and fresh dolerite bedrock along the eastern margin of the wetland. The residual saprolite surface is located approximately 1m below the dolerite bedrock surface and there is an abrupt near-vertical rise in the dolerite bedrock surface at the contact. The residual saprolite surface is characterised by 3-4m wide and 0.5m deep channel-like depressions adjacent to the contact zone, and to the west of the channel-like depression the residual sparolite surface has a relatively uniform elevation with small-scale ridges and depressions. The dolerite bedrock surface underlying the toe of the wetland is near-planar and extends across the toe of the wetland at roughly the same angle as the dolerite dyke ridge that cuts across the toe of the wetland. (T1 and T2; Figure 5.2 and Figure 5.7).

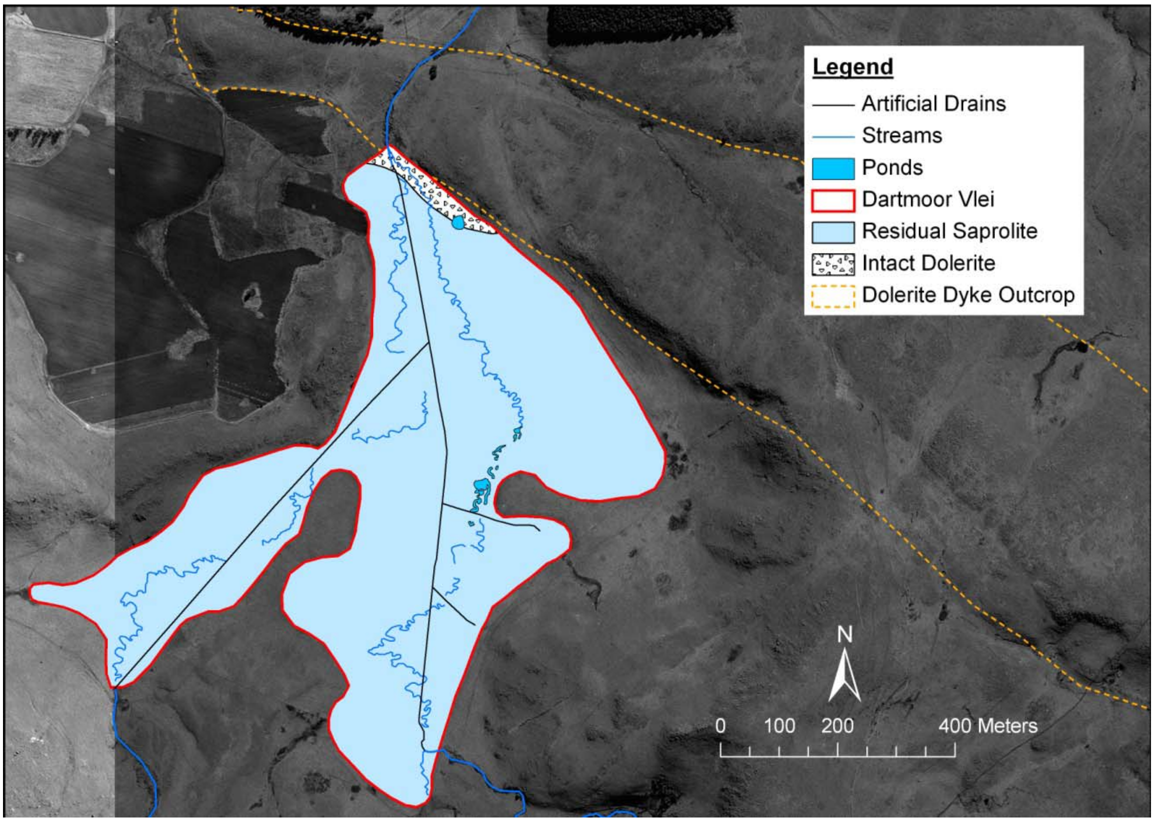


Figure 5.7: Geology underlying Dartmoor Vlei derived from coring.

5.5.2 Longitudinal Bedrock Basin Profile

It is evident that the longitudinal bedrock profile underlying the wetland is somewhat irregular and that it forms a depression behind the fresh dolerite surface that lies at a higher elevation than the residual saprolite surface at the toe of the wetland (Figure 5.6).

Upstream of the wetland, the stream flows over dolerite bedrock and the bedrock profile reflects the stream bed profile. As the stream flows into the wetland the stream bed gradient abruptly flattens out and the dolerite bedrock changes to a residual saprolite surface that slopes at a gradient of 3.42% in the uppermost region of the wetland (Figure 5.6). An abrupt break in slope in the residual saprolite surface occurs at about 250m along the longitudinal profile, beyond which the gradient of the residual saprolite surface slopes gently upwards for about 100m before steepening at about 350m along the profile (Figure 5.6). This surface terminates at a break in slope at about 600m along the profile, below which the saprolite surface is somewhat irregular characterised by two ridges extending from about 600m to about 950m along the profile. From about 950m to 1100m, the gradient of the saprolite surface again flattens out. At 1100m along the profile, the saprolite surface steepens to form a uniformly sloping surface. This surface terminates at a large step that occurs at the contact between the residual saprolite and dolerite bedrock. At the contact between the two lithologies, the dolerite bedrock surface is elevated approximately 1m above the residual saprolite surface. The dolerite bedrock surface has a fairly uniform gentle slope that is maintained until the stream is confined in the steep-sided dolerite dyke valley, where the wetland disappears and the gradient of the stream bed and bedrock surface abruptly increase.

5.6 Sedimentary fill

5.6.1 Transect 4 (Upper reaches)

Analysis of core A1 of Transect 4 (Figure 5.8) indicates that the alluvial fill in the western arm is characterised by a 20-30cm basal layer of fine peaty silt overlain by a 10-15cm organic-rich gravel layer, a 50cm sequence of upward coarsening organic-rich to peaty silt and a 30-40cm surface layer of peat that was fibrous within the top 10cm and amorphous between 10-40cm. The gravel layer comprises medium gravel in a matrix of fine sand and extends across the whole western arm at a depth ranging between 90 and 120cm below the surface. The gravel layers comprised angular fragments of altered sandstone and shale with a few rounded sandstone pebbles. The residual saprolite surface underlying the western arm comprises soft homogeneous blue clay. The residual clay in the vicinity of core A1 was observed to extend to a depth of 550cm below the ground surface before the clay became too hard to auger. The clay was initially a pale blue to dark blue-grey colour that graded into an orange-brown colour with depth. Evidence of fresh dolerite at the refusal depth was absent.

Organic matter content within the western arm wetland fill generally decreases with depth, increasing slightly at the base of the alluvial fill and reaching the lowest values in the vicinity of the gravel layer. Peat deposits were most substantial in the western portions of the western arm occurring in a gentle depression behind the small elevated ridge.

The elevated ridge that separates the two wetland arms comprises a 10cm basal layer of fine gravel overlain by a 1m sequence of upward fining organic-rich and peaty silt and by a 20-30cm decomposed amorphous peat surface layer. Organic matter content generally decreases with depth and the valley floor comprises soft white residual clay.

Like the western arm, the alluvial fill of the eastern arm is characterised by a 10-50cm fine organic-rich silt basal layer directly overlain by a 10-30cm organic-rich gravel layer, with the exception of the basal layers in the vicinity of core LB3, where there is a 30cm gravel basal layer. Above the basal layer the alluvial fill generally comprises sequences of organic-rich and peaty silt punctuated by one or more gravel layers and topped by a 30-40cm surface layer of peat. Like the western arm, the eastern arm sedimentary fill is characterised by a gravel layer near the base of the fill that extends laterally across the eastern arm basin. This gravel layer comprises medium gravel in a matrix of medium organic-rich silt. In the western portions of the wetland, the sedimentary fill overlying the gravel layer generally comprises an upward fining sequence of organic-rich to peaty silt overlain by a 30-40cm layer of peat. The residual clay in the vicinity of core A3 was observed extend to a depth of 750cm below the ground surface before the clay became too hard to auger. Evidence of fresh dolerite at the refusal depth was absent. Like the residual saprolite underlying the western arm, the residual saprolite surface underlying the western arm comprises soft homogeneous blue clay. The clay was initially a pale blue colour that graded into a bright orange colour with depth. Cores Lb3, Rb3 and A6 indicate that the sedimentary fill in the eastern portion of the eastern arm is characterised by a second thinner gravel layer that lies approximately 40-50cm above the lower gravel layer. The second gravel layer comprises fine gravel in a matrix of fine peaty silt and extends laterally across the eastern portion of the arm at a roughly uniform elevation. As a result, the wetland fill profile in the eastern portions of the arm generally comprises two upward fining sequences of organic-rich and peaty silt, with the lower sequence comprising an upward fining sequence from medium gravel to fine peaty silt and the upper sequence comprising an upward fining sequence from fine gravel to medium peaty silt. The gravel layers generally comprised angular fragments of altered sandstone and shale with a few rounded sandstone pebbles.

Organic matter content within the eastern arm wetland fill generally decreases with depth, increasing slightly at the base of the alluvial fill and reaching the lowest values in the vicinity of the main gravel layer. Peat deposits are most substantial in the western portions of the eastern arm occurring in the lower elevations of the arm.

Unlike the western arm and the elevated ridge, the bedrock underlying the eastern arm is highly variable. The bedrock surface underlying the western edge of the eastern arm (Core LB2) comprises a 5cm hard pan layer of weathered dolerite that sits on top of a light blue, soft residual saprolite. This bedrock surface then changes to a green coloured soft residual saprolite with no evidence of a hard pan layer as indicated in Core A4. However, in the middle of the eastern arm as indicated in Core A5, the bedrock surface again changes to a hard pan layer sitting atop the residual saprolite. From Core Lb3 to just past Core Rb3 the bedrock surface changes back to a soft residual saprolite similar to that underlying the western arm. As one moves further east along the cross-sectional profile, the residual saprolite surface changes from a homogeneous blue clay to a bright orange speckled clay.

Analysis of the excavated faces of the artificial drains at the head of the wetland reveal the presence of a relatively thick sequence of coarse to medium gravel in a matrix of organic-rich silt occurring at depths between 0.5 and 1.5m below the surface. The thickness and extent of this gravel deposit was not encountered in any of the cores analysed in transect 4 which indicates that there is an abrupt decrease in the size and extent of gravel deposits from the head of the wetland to Transect 4.

In terms of sorting, the fill underlying the wetland in the upper reaches was moderately poorly to very poorly sorted. The general trend showed that the fine silt layers were generally moderately poorly to poorly sorted, the coarse silt layers poorly sorted, and the gravel layers very poorly sorted.

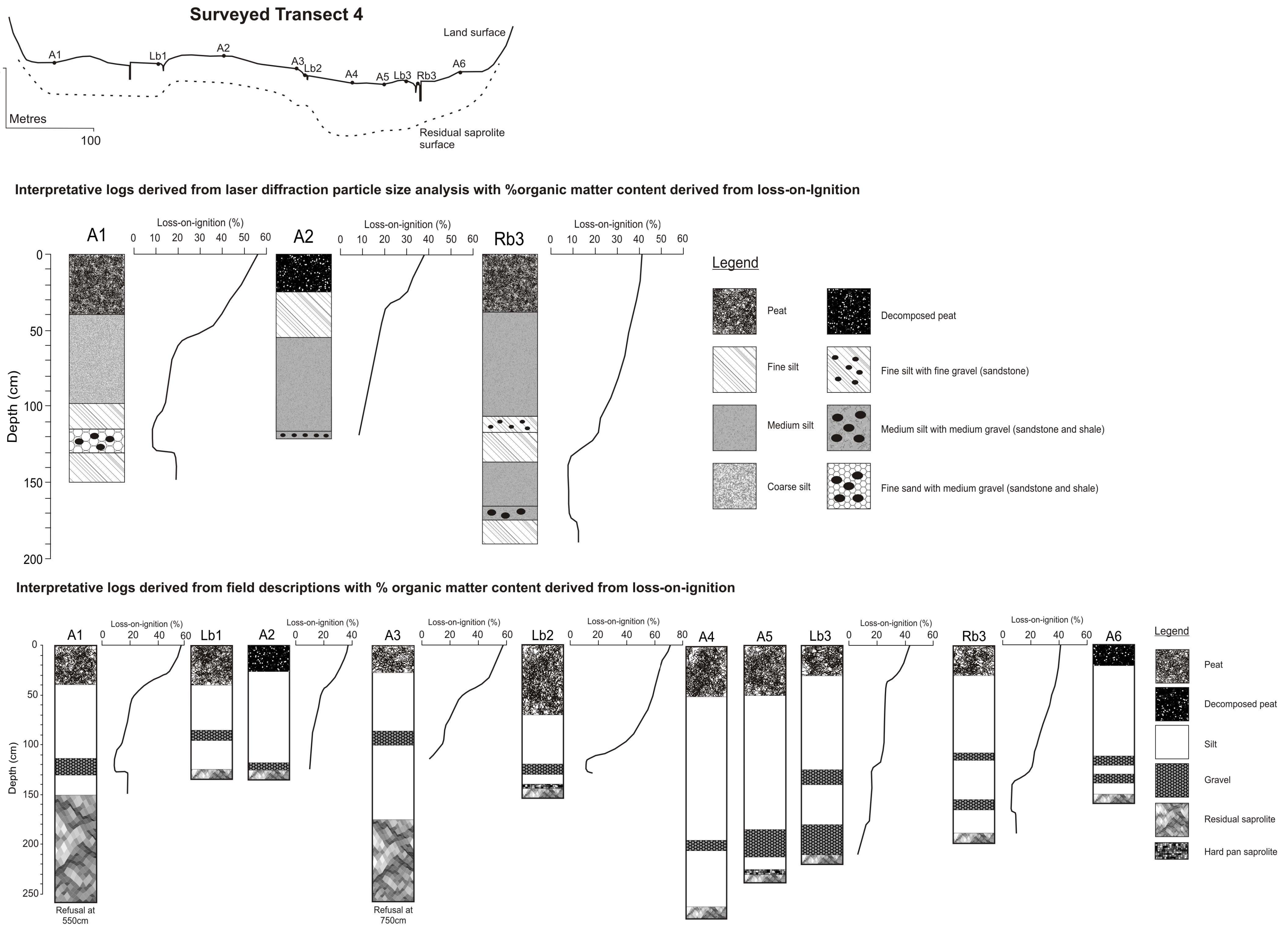


Figure 5.8: Sedimentary cores taken along transect 4

5.6.2 Transect 3 (Middle reaches)

The fill in the western portion of the valley comprises a basal gravel layer overlain by a 1-1.5m sequence of fine organic-rich to peaty silt and a 30cm layer of fibrous and amorphous peat at the surface. The gravel comprises fine stones directly overlying and meshing with soft residual saprolite. The residual saprolite surface underlying the western portions of the wetland comprises soft homogeneous pale blue clay that grades into orange clay with depth. Organic matter content generally decreases with depth (Figure 5.9).

The fill in the vicinity of core A3 near the middle of the valley comprises a basal layer of upward fining organic-rich silt overlain by an organic-rich fine gravel layer, an upward fining sequence of silt that grades from coarse organic-rich silt to fine peaty silt, and a 30cm fibrous and amorphous peat surface layer. The upper portion of the fine peaty silt layer has a 10cm peat layer at a depth of 0.5m. The residual clay in the vicinity of core A3 was observed extend to a depth of 350cm below the ground surface before the clay became too hard to auger. Evidence of fresh dolerite at the refusal depth was absent. The residual saprolite surface underlying the wetland in the vicinity of A3 comprises soft homogeneous white clay that grades into orange clay with depth.

In the vicinity of core A4, the alluvial fill comprises a basal organic-rich gravel layer overlain by a 30cm upward fining sequence of organic-rich to peaty silt, a 20cm fibrous peat layer, two upward fining sequences of peaty silt, and a 50cm fibrous and amorphous peat surface layer. The residual clay in the vicinity of core A4 was observed extend to a depth of 515cm below the ground surface before the clay became too hard to auger. Evidence of fresh dolerite at the refusal depth was absent. The residual saprolite surface underlying the wetland in the vicinity of A4 comprises soft homogeneous light grey-green clay that grades into orange-brown clay with depth.

In the middle portion of the valley, organic matter content generally decreases with depth with the exception of the fill in the vicinity of core A4 where localised organic layers comprising undecomposed fibrous material occur at intervals between the depths of 1.1 and 1.6m below the surface.

Like the sedimentary fill in the upper reaches, the fill underlying the wetland in the middle reaches was moderately poorly to very poorly sorted. The general trend showed that the fine silt layers were generally moderately poorly to poorly sorted, the coarse silt layers were poorly sorted, and the gravel layers very poorly sorted.

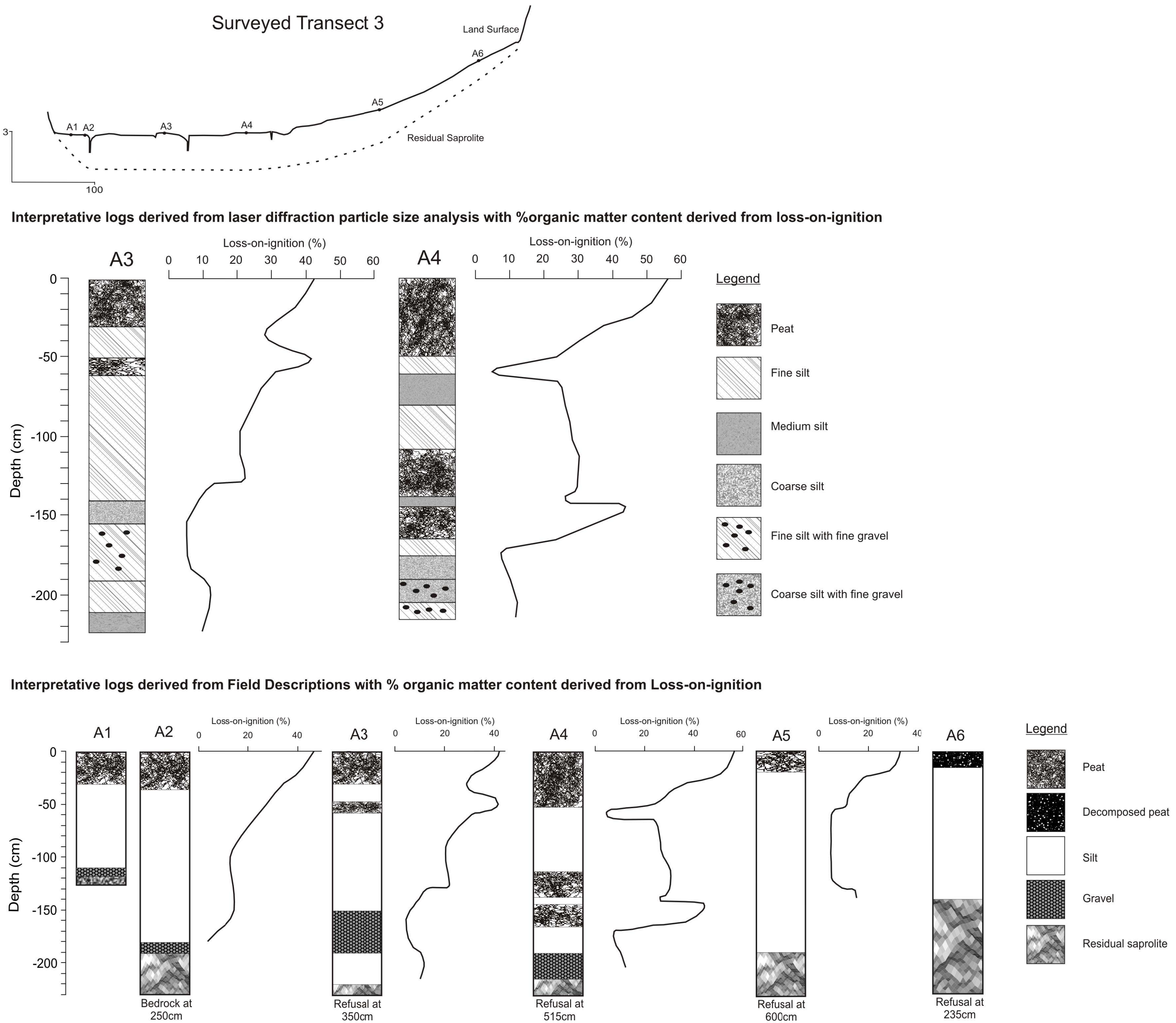


Figure 5.9: Sedimentary cores taken along transect 3

5.6.3 Transect 2 (Lower reaches)

The fill in the vicinity of core A1 comprises a well defined basal gravel layer with sharp horizontal boundaries overlain by a 1.5m sequence of alternating fine and medium peaty silt and a 40cm peat surface layer. The well defined fine gravel layer extends laterally across the western portions of the residual saprolite basin surface and disappears in the vicinity of core A2. Much of the gravel comprised soft weathered sandstone fragments with little evidence of dolerite. The residual clay in the vicinity of core A1 was observed extend to a depth of 480cm below the ground surface before the clay became too hard to auger. Evidence of fresh dolerite at the refusal depth was absent. The residual saprolite surface underlying the wetland in the vicinity of A1 comprises soft homogeneous pale blue clay that grades into orange-brown clay with depth. Organic matter content generally decreases relatively uniformly with depth. With the exception of the basal sand and gravel layer, the majority of the core comprises peaty material.

In terms of sorting, the organic-rich silt layers vary between 1.61 to 1.92 phi, indicating that the silt sediments are generally moderately poorly to poorly sorted. The basal sand and gravel layer has a sorting value of 2.8 phi indicating that it is very poorly sorted. There appear to be no trends between mean particle size and sorting. However, the coarsest silt layer does have the highest sorting value of all the silt layers.

In the vicinity of core A3, the alluvial fill comprises a basal layer of medium organic-rich silt overlain by two upward fining sequences of sediment. The lower upward fining sequence consists of coarse gravel in a matrix of medium organic-rich silt that grades upwards into medium organic-rich silt. Above this sequence there is a well defined fine gravel layer overlain by an upward fining sequence of organic-rich and peaty silt consisting of medium organic-rich silt that grades upwards into fine peaty silt. The well defined fine gravel layer identified within cores A2 and A3 extends laterally across the middle and eastern portions of the wetland fill at roughly the same elevation. The coarse gravel layer identified in core A3 is not laterally extensive and was only found between the eastern channel and the eastern depression. The gravel layers comprised angular fragments of altered sandstone and shale with a few rounded sandstone pebbles. The residual clay in the vicinity of core A3 was observed extend to a depth of 570cm below the ground surface before the clay became too hard to auger. Evidence of fresh dolerite at the refusal depth was absent. Like the residual saprolite surface underlying the wetland in the vicinity of A1, the residual saprolite comprises soft homogeneous pale blue clay that grades into orange-brown clay with depth. The organic matter content initially decreases rather rapidly with depth as the core grades from peat to organic-rich silt. After this point, the decrease in organic matter content becomes more gradual, reaching its lowest point at 115cm, which corresponds with the fine gravel layer. From this point, organic matter content gradually increases with depth.

Organic matter content is also variable in the lateral direction. Core A1 has substantially higher organic matter content than A3 and it (core A1) does not have a basal organic-rich silt layer.

In terms of sorting, the silt layers vary between 1.76 to 1.95 phi indicating that the silt layers are generally moderately poorly to poorly sorted. Like core A1, the coarser gravel layers have higher sorting values ranging from 2 to 2.12 phi, indicating that they are poorly sorted.

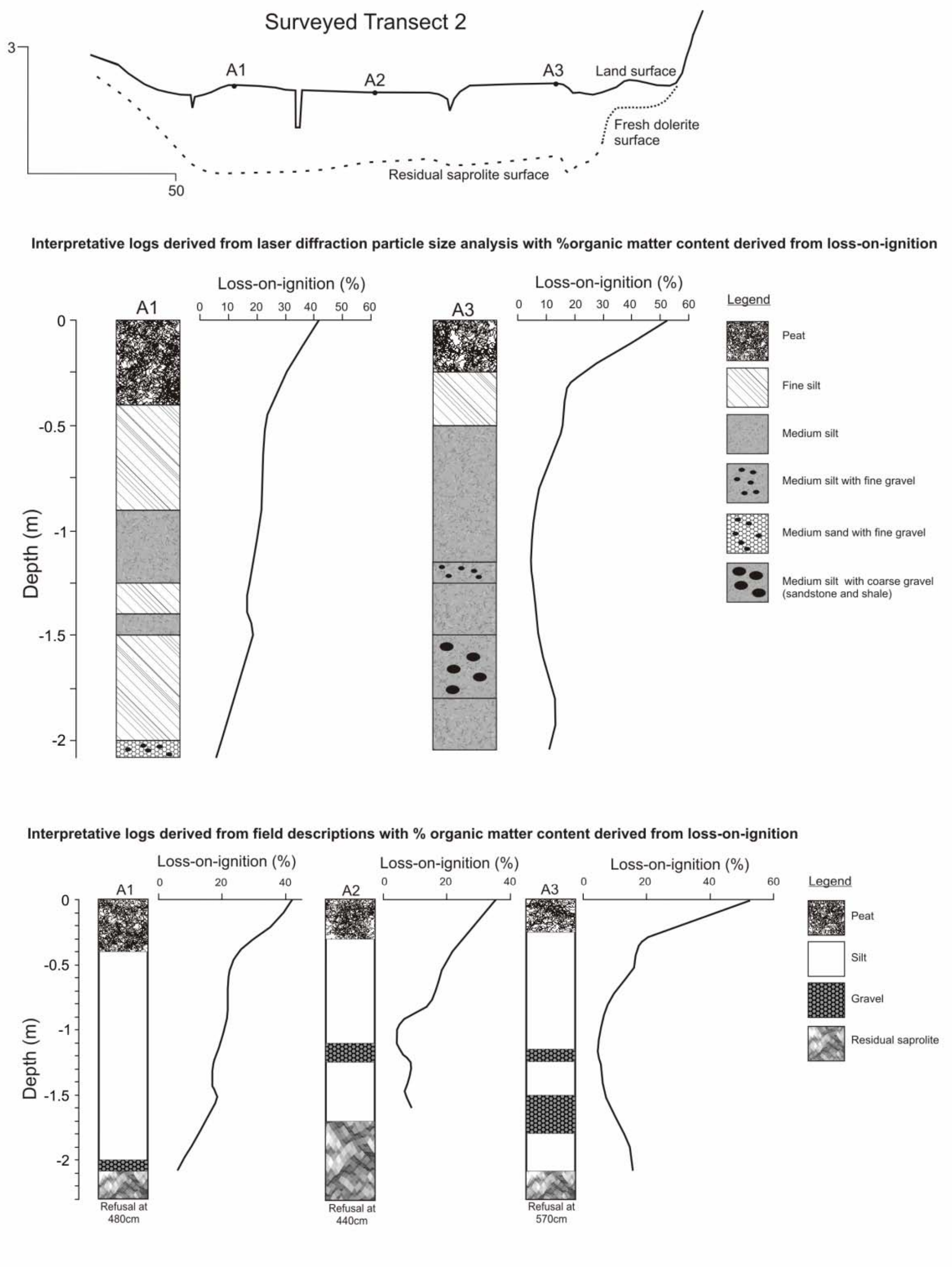


Figure 5.10: Sedimentary cores taken along transect 2

5.7 Chemical and mineralogical properties of the fresh dolerite and residual saprolite

5.7.1 Chemical properties of the fresh dolerite and residual saprolite

Table 5.1 indicates that the fresh dolerite samples are predominantly composed of SiO_2 and Al_2O_3 with notable amounts of FeO, MgO and CaO. SiO_2 generally accounts for 51.08% of the dolerites with Al_2O_3 accounting for 14.08%. Compared to dolerites in the region, Fe_2O_3 was very low averaging 1.46%. No significant differences occurred in the chemical composition of the dolerites, even in the oxides CaO, MgO, Na_2O and K_2O which are strongly influenced by magmatic differentiation and crustal contamination as well as being considered the most mobile during weathering.

Table 5.1: Chemical composition (weight percent) of the fresh dolerite samples within the Dartmoor Vlei valley as determined by XRF

	D2	D4	D9	D10	D11	D13	D16	D18	Average
SiO_2	51.37	51.81	51.26	50.30	50.75	51.37	51.26	50.50	51.08
TiO_2	1.06	1.24	1.04	1.25	1.07	1.27	1.14	1.04	1.14
Al_2O_3	14.38	13.33	14.53	14.06	15.51	12.93	13.34	14.54	14.08
Fe_2O_3	1.30	1.60	1.31	1.65	1.32	1.63	1.54	1.31	1.46
FeO	12.65	11.73	12.79	12.37	13.65	11.38	11.74	12.80	12.39
MnO	0.17	0.21	0.17	0.21	0.19	0.22	0.21	0.18	0.20
MgO	6.72	6.45	6.64	6.24	6.42	6.39	6.77	7.39	6.63
CaO	11.49	10.38	11.48	10.20	11.50	10.33	10.78	11.12	10.91
Na_2O	2.62	2.55	2.69	2.66	2.60	2.68	2.49	2.67	2.62
K_2O	0.63	0.69	0.62	0.72	0.65	0.70	0.63	0.63	0.66
P_2O_5	0.21	0.19	0.19	0.19	0.21	0.21	0.16	0.20	0.20
Cr_2O_3	0.04	0.03	0.04	0.03	0.05	0.04	0.05	0.07	0.04
NiO	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01
LOI	0.51	0.05	0.38	0.32	0.35	0.19	0.33	0.61	0.34
Total	103.16	100.27	103.15	100.21	104.28	99.35	100.45	103.07	

Like the fresh dolerite samples, Table 5.2 indicates that the residual clay underlying the wetland in the lower reaches is predominantly composed of SiO_2 and Al_2O_3. SiO_2 generally accounts for 59.3 to 62.18% of the residual saprolite mass between 200 and 500cm depth with Al_2O_3 varying between 16.57 and 22.41% and K_2O between 2.35 and 3.25%. With the exception of Fe_2O_3, which increased variably with depth, the chemical composition of the residual saprolite mass remained relatively constant with depth. The relatively high weight percentage attributed to loss-on-ignition indicates that the residual saprolite contains appreciable amounts of water.

Table 5.2: Chemical composition (weight percent) of residual saprolite clay samples within core T2A3 (lower reaches) as determined by XRF

	198-208cm	295-305cm	400-410cm	495-505cm
SiO_2	62.18	60.12	59.30	61.62
TiO_2	0.77	0.71	0.77	0.62
Al_2O_3	22.41	16.57	18.95	19.32
Fe_2O_3	1.61	11.64	7.77	6.23
MnO	0.02	0.03	0.06	0.16
MgO	0.67	0.56	0.78	0.64
CaO	0.00	0.00	0.02	0.03
Na_2O	0.00	0.06	0.00	0.00
K_2O	2.65	2.49	3.25	2.35
P_2O_5	0.05	0.13	0.12	0.16
Cr_2O_3	0.00	0.01	0.01	0.01
NiO	0.00	0.00	0.00	0.00
LOI	9.44	8.00	8.75	8.84
Total	99.80	100.32	99.78	99.98

Like the residual saprolite underlying the lower reaches, Table 5.3 indicates that the residual saprolite underlying the middle reaches is predominantly composed of SiO_2 and Al_2O_3. SiO_2 generally accounts for 31.22 to 59.01% of the residual saprolite mass between 200 and 430cm depth with Al_2O_3 varying between 15.35 and 25.43%, Fe_2O_3 varying between 5.88 and 24.49% and K_2O varying between 0.25 and 2.44%. The percentage Al_2O_3 and LOI decreased with depth and the percentage of Fe_2O_3 varied considerably with depth, increasing initially and then decreasing at depths of 300cm and lower. Besides these compounds, the chemical composition of the residual saprolite mass remained relatively constant with depth. The percentage Fe_2O_3 and LOI in the residual saprolite generally increases from the lower to middle reaches.

Table 5.3: Chemical composition (weight percent) of residual saprolite clay samples within core T3A4 (middle reaches) along as determined by XRF

	215-225cm	235-245cm	300-310cm	420-430cm
SiO_2	45.26	31.22	58.29	59.01
TiO_2	2.16	1.67	0.67	0.62
Al_2O_3	25.43	20.18	15.35	15.50
Fe_2O3	5.88	24.49	13.61	11.85
MnO	0.05	0.06	0.03	0.08
MgO	0.48	0.56	1.13	0.95
CaO	0.02	0.07	0.05	0.07
Na_2O	0.00	0.00	0.00	0.00
K_2O	0.25	0.15	2.44	2.23
P_2O_5	0.10	0.49	0.16	0.15
Cr_2O_3	0.02	0.02	0.01	0.01
NiO	0.02	0.02	0.00	0.00
LOI	20.82	20.39	8.53	9.37

Total	100.49	99.32	100.27	99.84

Table 5.4 indicates that the residual saprolite underlying the wetland in the upper reaches is predominantly composed of SiO_2, Al_2O_3 and Fe_2O_3. SiO_2 generally accounts for 34.9 to 44.52% of the residual saprolite mass between 125 and 730cm depth with Al_2O_3 varying between 19.66 and 30.03%, Fe_2O_3 varying between 4.33 and 23.43%, TiO_2 between 1.4 and 2.24% and MgO between 0.73 and 1.48%. There is a slight decrease in the percentage of SiO_2 and an increase in the percentage of Fe2O3 within the residual saprolite in the upper reaches with depth. In contrast to the residual saprolite in the middle reaches, the percentage LOI within the clays generally remained constant with depth. In general, Tables 5.2, 5.3 and 5.4 indicate that the percentage Fe_2O_3 within the residual saprolite increases towards the head of the wetland and the percentage SiO_2 decreases towards the head of the wetland.

Table 5.4: Chemical composition (weight percent) of residual saprolite clay samples within core T4A3 (upper reaches) as determined by XRF

	125-135cm	260-270cm	420-430cm	530-540cm	635-645cm	730-740cm
SiO_2	44.22	44.19	44.52	35.28	36.14	34.90
TiO_2	1.86	2.19	2.24	1.40	1.56	1.42
Al_2O_3	30.03	28.92	27.85	20.61	21.86	19.66
Fe_2O_3	4.33	5.69	5.68	22.00	19.95	23.43
MnO	0.06	0.07	0.06	0.06	0.07	0.07
MgO	0.73	0.88	1.07	1.19	1.25	1.48
CaO	0.00	0.01	0.06	0.15	0.15	0.20
Na_2O	0.00	0.00	0.01	0.02	0.00	0.00
K_2O	0.07	0.08	0.09	0.36	0.37	0.40
P_2O_5	0.02	0.05	0.05	0.38	0.33	0.31
Cr_2O_3	0.05	0.05	0.06	0.04	0.05	0.03
NiO	0.01	0.01	0.02	0.01	0.02	0.01
LOI	18.91	18.12	18.35	18.69	18.75	18.41
Total	100.29	100.26	100.06	100.19	100.50	100.32

5.7.2 Comparative analysis of the chemistry of the fresh dolerite and residual saprolite

As the fresh dolerite rock has been transformed into completely weathered residual saprolite, concentrations of SiO_2 and TiO_2 have remained relatively constant showing slight decreases in concentration within the middle and upper reaches, Al_2O_3 has increased in concentration by between 5 and 10%, Fe_2O_3 has increased, and there has been significant losses in the concentrations of FeO, MgO, CaO and Na_2O. In general there has been a general loss of Fe ions and substantial losses of Mg, Ca and Na ions as the intact dolerite has weathered to residual saprolite. The variability in iron loss can be ascribed to inconsistencies in oxidation at different sites with higher ion losses in more reducing conditions that generated more mobile Fe 2+ ions.

5.7.3 Mineralogical properties of the fresh dolerite and residual saprolite

The mineral composition of the fresh dolerite samples reflects that of a typical dolerite, and is dominated by pyroxene and plagioclase feldspar. On average, the fresh dolerite samples are composed of approximately 65% feldspar (albite + anorthite + orthoclase), 24% pyroxene (enstatite + wollastonite + ferrosilite), 8% quartz, and 3% illmenite with very low concentrations of magnetite, apatite and chromite. Olivine (fayalite, fosterite) was not present within any of the samples (it is not mineralogically compatible with quartz) (Hutchison, 1974).

Table 5.5: Averaged mineralogical composition of the fresh dolerite samples in terms of weight percentage as determined by the CIPW method in Hutchison (1974)

Apatite	0.54
Chromite	0.08
Ilmenite	2.75
Orthoclase	4.94
Albite	28.13
Anorthite	31.36
Magnetite	0.32
Wollastonite	14.90
Enstatite	8.03
Ferrosilite	0.86
Fayalite	0.00
Fosterite	0.00
Quartz	8.10

Tables 5.6, 5.7 and 5.8 show that the residual clay samples predominantly comprise quartz, kaolin and orthoclase, of which kaolin is the most dominant within the upper and middle reaches and quartz within the lower reaches of the wetland. In the lower reaches, the percentage orthoclase within the residual saprolite is variable with depth, the percentage kaolin increases slightly with depth, the percentage quartz decreases slightly with depth and the percentage iron oxides is variable with depth. In the middle reaches, the percentage orthoclase within the residual saprolite increases with depth, the percentage kaolin decreases with depth, the percentage quartz increases slightly with depth and iron oxides is variable with depth, increasing initially and then decreasing at depths of 300 to 400cm. In the upper reaches, the percentage orthoclase increases steadily with depth, the percentage kaolin decreases with depth, quartz remains relatively constant with depth and percentage iron oxides increases with depth. Overall, the percentage orthoclase decreases considerably towards the head of the wetland, the percentage kaolin remains relatively constant towards the head, of the wetland, the percentage quartz decreases slightly towards the head of the wetland and the percentage iron oxides increase towards the head of the wetland.

Table 5.6: Averaged mineralogical composition of the residual clay samples within core T2A3 (lower reaches) in terms of weight percentage as determined by mineralogical calculations based on the XRF data

	198-208cm	295-305cm	400-410cm	495-505cm
Orthoclase	17.03	13.65	17.08	12.82
Kaolin	25.47	32.89	36.06	38.16
Quartz	54.20	43.30	40.37	43.62
Iron Oxides	3.29	10.15	6.50	5.41

Table 5.7: Averaged mineralogical composition of the residual clay samples within core T3A4 (middle reaches) in terms of weight percentage as determined by mineralogical calculations based on the XRF data

	215-225cm	235-245cm	300-310cm	420-430cm
Orthoclase	1.58	1.00	13.67	12.73
Kaolin	58.30	48.82	31.13	32.05
Quartz	34.19	24.15	43.07	44.45
Iron Oxides	5.92	26.03	12.13	10.77

Table 5.8: Averaged mineralogical composition of the residual clay samples within core T4A3 (upper reaches) in terms of weight percentage as determined by mineralogical calculations based on the XRF data

	125-135cm	260-270cm	420-430cm	530-540cm	635-645cm	730-740cm
Orthoclase	0.42	0.48	0.55	2.32	2.36	2.60
Kaolin	65.15	63.24	61.98	48.15	50.40	46.20
Quartz	30.30	30.80	31.91	26.95	27.04	27.01
Iron Oxides	4.13	5.47	5.55	22.59	20.21	24.19

A comparison of the mineralogy of the fresh dolerite and residual saprolite samples indicates that the dominant pyroxene and feldspar group minerals have largely been transformed into clay minerals of the kaolinite family. The transformation of the dominant pyroxene and the feldspar group minerals into clay minerals is generally confirmed by the XRD patterns of the fresh dolerite and residual clay samples as shown in Figures 5.11 and 5.12. Figure 5.11 is characterised by sharp intense peaks indicating that the minerals analysed have strong crystalline structure. The d spacings of the sharp peaks are indicative of the minerals pyroxene, quartz and plagioclase feldspar that typically make up a tholeiitic dolerite rock. The small broad peak at approximately 7^0 2 theta is considered to be due to a trace of poorly crystalline chlorite formed in the very last stages of crystallisation. Figure 5.12 is generally characterised by broad, low intensity peaks with the exception of the isolated sharp, intense peaks of quartz. The broad low intensity peaks indicate poor crystal structure that is indicative of clay minerals of the kaolinite family.

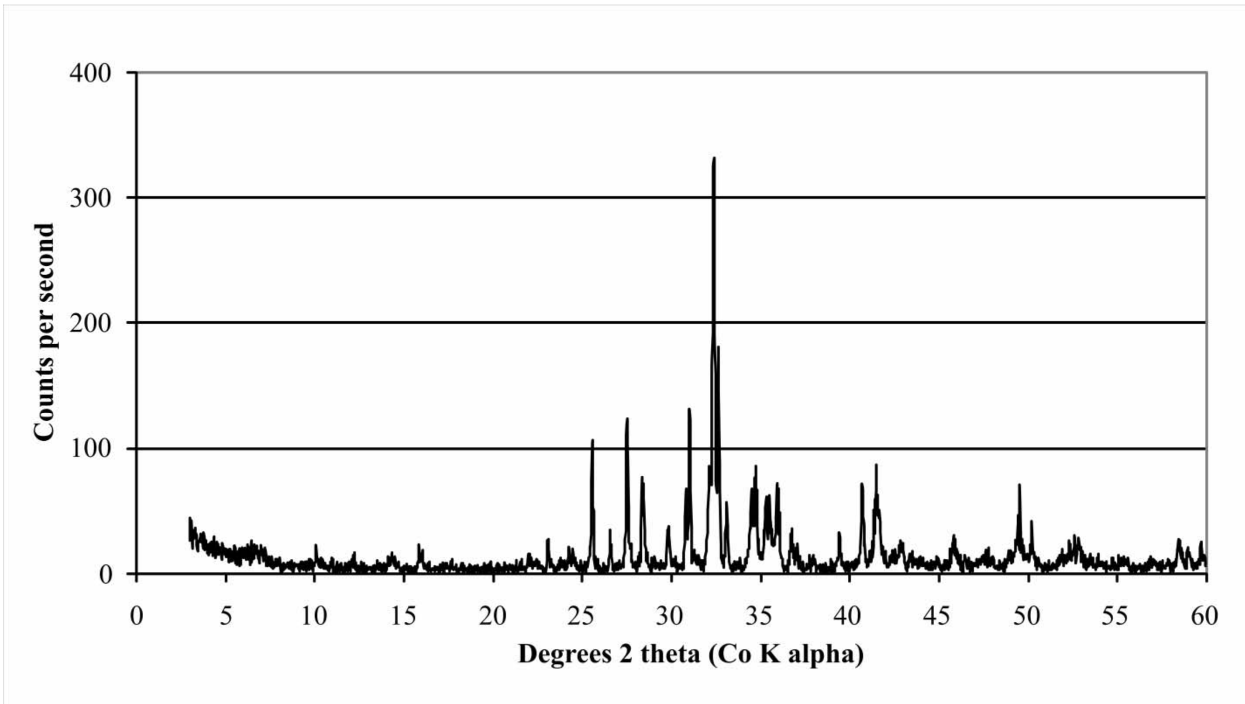


Figure 5.11: XRD pattern of fresh dolerite sample D4.

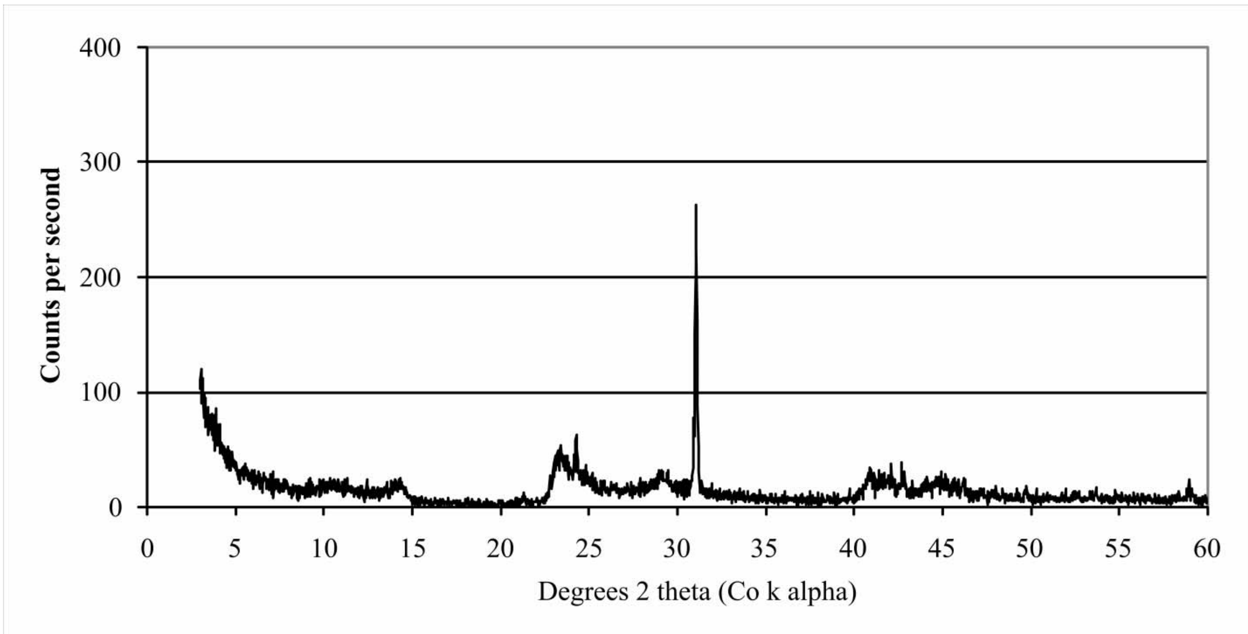


Figure 5.12: XRD pattern of residual saprolite sampled at a depth of 215-225cm.within core T3A4

CHAPTER 6 - DISCUSSION

6.1 Wetland hydrology

Dartmoor Vlei is characterised by a high water table all year around, slow diffuse water movement and a high volume of water storage despite the relatively permeable peat surface layers and the artificial drains. These characteristics are largely the result of a large amount of water inputs relative to outputs and the retardation of water movement down the wetland due to the gentle longitudinal gradient of the wetland surface, local depressions and discontinuous streams with low unit stream powers. The fact that the wetland still remains permanently wet in parts despite streamflow being diverted into drains indicates that the numerous springs that occur around the fringes of the wetland are the dominant water inputs at present and have likely been a major input throughout the evolution of the wetland. The high organic content of the majority of the wetland fill confirms this. In addition, much of the peat surface layers were highly amorphous and had a high water holding capacity. As a result, the amorphous peat layers were also playing an important role in impeding the draining of the wetland by the artificial drains. The seasonal nature of rainfall indicates that the incoming streams and springs are seasonal and that the wetland receives most water during the summer season. However, hydrological inputs into the wetland are maintained to a lesser extent during winter as a result of the high occurrence of mist and the occurrence of snow within the region during winter.

The maintenance of high water tables during the low rainfall periods confirms that the residence time of water in the wetland is likely to be long, and that a large volume of water is stored within the wetland during the winter months as a result of the slow movement of water through the wetland. In addition, the gradual change in the underlying bedrock from jointed and fractured dolerite to relatively impermeable clay would have greatly reduced transmission losses to groundwater and provided a relatively impermeable lining to the basin. Under natural conditions, the stream channels would have also been a major water input into the wetland via overtopping of the channel banks in certain reaches, diffuse leakage through the organic stream banks, and sheetflow along discontinuous reaches. The presence of a 30-50cm peat surface layer throughout the wetland indicates prolonged wetness and extremely low rates of clastic sedimentation. This is confirmed by the lack of clastic sediment found on or within the beds and banks of the wetland streams. This in turn confirms that at present, sediment carried into the wetland from the two incoming mountain streams is largely deposited at the head of the wetland and that limited clastic sediment is carried by the wetland streams.

6.2 Wetland geomorphology

The permanently wet valley-bottom wetland conditions described in section 6.1 is strongly controlled by the geomorphology of the wetland, which is characterised by a gentle longitudinal average gradient of 0.4% and the occurrence of localised depressions. The gentle longitudinal gradient of the valley-bottom wetland is in turn the product of a number of geomorphological controls.

6.2.1 Lithological controls on wetland geomorphology, stream behaviour and sedimentation

The longitudinal profile of the eastern stream bed through and downstream of the wetland demonstrates that the steepening of the valley and stream bed that is co-incident with the termination of the wetland is linked to the change in lithology from dolerite sill to dolerite dyke. The dolerite dyke at the toe of the wetland is clearly more resistant to weathering than the dolerite sill underlying the body of the wetland as the sill is deeply weathered and exists as soft residual clay and saprolite (Figure 5.3). Despite its extremely soft and weathered nature, it has a gently sloping near-planar cross-sectional morphology which indicates that the body of the wetland (valley floor) has been subject to lateral erosion and planation as described for wetlands on the Highveld of South Africa (Tooth *et al*, 2002a; 2004). The preservation of dolerite dyke ridges throughout the plateau area downstream (north) and north-east of the wetland, and the occurrence of broad flat valleys and terrestrial planation surfaces underlain by dolerite sills adjacent to (west of) the wetland indicate that the dolerite dykes in the area are more resistant to erosion than the dolerite sills into which they have intruded. The high resistance of the dolerite dyke relative to the dolerite sill confirms that the dolerite dyke immediately below the wetland is currently impeding the incision of the dolerite sill valley and that it is acting as a local base level for the wetland valley and streams. Thus, the formation of the broad low gradient valley that hosts Dartmoor Vlei is largely the product of the lateral planation of the valley by the incoming streams as described in the conceptual model developed by Tooth *et al* (2002a).

The presence of laterally extensive gravel channel-bed deposits at and in the vicinity of the residual saprolite and dolerite bedrock surfaces and the occurrence of 30-60cm thick lateral accretion deposits evidenced by the upward fining sequences of silt above many of these gravel channel-bed deposits, indicate that at least two distinct phases of laterally migrating mixed-load streams were responsible for the lateral planation of the residual saprolite bedrock basin. However, in contrast to the sedimentary deposits of the laterally planed floodplains studied by Tooth *et al* (2002a; 2004) and Grenfell (2007), much of the residual saprolite basin, specifically in the lower and upper reaches, is mantled by a thin basal layer of fine organic-rich silt instead of a thin basal gravel layer. The weak evidence of thin laterally extensive basal gravel deposits is likely to be due to the nature of the parent source material of the gravel, which is the same as the completely weathered rock mass underlying the wetland. Chemical weathering would weather the gravel and incorporate it into the saprolite. This is confirmed by the limited presence of dolerite stones and pebbles within the gravel layers, which generally comprise sandstone and shale material. Furthermore, the fine organic-rich basal layers are also not typical of the lateral accretion deposits that are expected to occur immediately above planed valley basins, as described by Tooth *et al* (2002a) and Grenfell (2007). The organic-rich basal deposits in Dartmoor Vlei are probably the remnants of earlier lateral accretion deposits that were reworked by laterally migrating streams and replaced and overlain with a gravel layer and more recent sequences of lateral accretion deposits. The occurrence of more than one gravel layer within the valley fill sedimentary sequence above deeply weathered bedrock also suggests that this particular landscape setting of Dartmoor Vlei is aggradational and not degradational as is the case for the wetlands described by Tooth *et al* (2002a) and Grenfell (2007). The silty lateral accretion deposits indicate that the sediment load of the early floodplain streams predominantly comprised fine to coarse silt with sand and gravel bed loads. These coarse bed load deposits seem unlikely to be due to the present streams due to the small catchments and the low unit stream powers of the two streams that currently drain into the wetland.

6.2.2 The role of landscape history

The presence of an abandoned laterally planed valley-bottom network that intersects Dartmoor Vlei in the lower reaches of the wetland from the west, indicates that landscape history has likely had a strong influence on the origin and evolution of the Dartmoor Valley, particularly the extent of chemical weathering of the dolerite sill underlying Dartmoor Vlei. The geomorphological link between the Dartmoor Valley and the abandoned valley-bottom network to the west indicates that flow within the wetland likely drained into the westward oriented valley in the past before being diverted through the dolerite dyke that cuts across the toe of the current wetland in the north. The reasons for the capture of flow along the present flow path through the dolerite dyke and the abandonment of the broad planed valley-bottom network is unknown. However, the current elevation of the dolerite dyke local base level and the dolerite bedrock underlying the toe of the wetland at approximately 1-2m below the terrestrial planation surface indicate that the dolerite dyke base level and valley floor of Dartmoor Vlei have been lowered relative to the planation surface of the valley-bottom network to the west.

In a reconnaissance survey of bauxitic material in the KwaZulu-Natal Midlands, Bredell *et al* (1980) identified the Karkloof Plateau as one of many high-lying dolerite capped plateaus in the region which are remnants of the African Erosion Surface. In addition, Partridge *et al* (2006) broadly identify the KwaZulu-Natal Midlands as a region that likely has remnants of the African Erosion Surface. The African Erosion Surface was a widespread low relief planation surface that existed below and above the Great Escarpment at the end of the Cretaceous (144 to 66.4 million years BP) (Partridge and Maud, 2000a; Partridge *et al*, 2006). There is general consensus that climates during much of the Cretaceous were warm and humid, which facilitated both rapid erosion and extensive chemical weathering of specific lithologies, resulting in substantial in-situ residual clay and saprolite formation (Partridge and Maud, 2000a; Partridge *et al*, 2006). This provides an explanation for the origin of the extensive terrestrial planation surface adjacent to Dartmoor Vlei as well as the occurrence of extensive in-situ chemical weathering of the dolerite sill underlying Dartmoor Vlei.

If most of the Karkloof Plateau is a remnant of the African Erosion Surface, the initial planation of the valley that hosts Dartmoor Vlei took place during the Cretaceous period. Subsequent to the Cretaceous, Dartmoor Vlei seems to have been predominantly aggradational, which is in stark contrast to all of the floodplains that have been studied by Tooth *et al* (2002a; 2004) and Grenfell (2007), which were laterally planed during the Miocene and Pliocene eras and are still being laterally planed at present as a result of the persistence of the energy gradients created by the most recent Pliocene uplift (McCarthy and Hancox, 2000; Partridge and Maud, 2000a). Thus, in contrast to most South African wetlands that have been studied, the formation of Dartmoor Vlei has largely been controlled by the tectonic and climatic conditions prevalent during the Cretaceous.

During the initial planing of the valley during the Cretaceous, the streams responsible for the lateral planation of the valley floor would have been substantially larger than those that currently feed Dartmoor Vlei. This is because the extent of the stream catchments have been greatly reduced by the northwards retreat of the Karkloof escarpment as a result of erosion linked to the Mgeni River and its catchment. The presence of larger stream catchments and associated discharges during the initial planation of the valley provides a possible explanation for the occurrence of the extensive abandoned valley-bottom network linked to Dartmoor Vlei, which could only have been laterally planed by streams larger than those that exist today. In addition, the larger stream discharges in the past provide an explanation for the occurrence of relatively coarse sand and gravel channel bed deposits within the sedimentary fill of Dartmoor Vlei. Thus, the evolution of Dartmoor Vlei has likely been characterised by the gradual reduction in stream discharges as the streams' catchments have been cut back by erosion within the Mgeni River catchment.

At the end of the Cretaceous, the energy gradients responsible for the rapid erosion of KwaZulu-Natal had been substantially reduced by the lateral planation of the province, and drainage was largely sluggish (Partridge and Maud, 2000a; Partridge *et al*, 2006). As a result, the stream power of the laterally migrating meandering streams within the Dartmoor Valley at the end of the Cretaceous would have been greatly reduced encouraging the formation of low energy palustrine marshes. In addition to the decline in erosion rate during the latter part of the Cretaceous, there was also major cooling and drying of climates which further slowed landscape denudation (Tyson and Partridge, 2000; Partridge and Maud, 2000a).

6.2.3 Chemical weathering and wetland geomorphology

Although the processes of wetland formation described in the conceptual model developed by Tooth *et al* (2002a) provides an adequate explanation for the general geomorphology of the broad gently-sloping valley in which Dartmoor Vlei has formed, the conceptual model cannot explain the current hydrology and geomorphology of the diffuse peat-dominated valley-bottom wetland and the discontinuous nature of the wetland streams. Neither can it explain the occurrence of a number of gravel channel-bed deposits above the weathered bedrock (saprolite) surface, which indicates that aggradation of the stream channel beds has occurred. The discontinuous stream styles, strongly diffuse valley-bottom conditions, and the predominance of organic sedimentation within Dartmoor Vlei, are not usually associated with dolerite-controlled floodplains like those that have been studied along the Klip, Mooi and Nsonge Rivers (Tooth *et al*, 2002a; Tooth *et al*, 2004; Grenfell, 2007). Rather, such characteristics as observed in Dartmoor Vlei have been found to be largely the product of alluvial impoundments that impede flow within broad valleys, or of neo-tectonic subsidence associated with continental scale fault systems (McCarthy and Hancox, 2000; Tooth and McCarthy, 2007; Grenfell, 2007; Joubert, 2009). In the case of Dartmoor Vlei, the longitudinal profile of the residual saprolite and bedrock surface underlying the wetland strongly suggests that flow retardation, the creation of surface depressions, and the reduction in the longitudinal gradient of the wetland surface, can be attributed to the sagging of the residual saprolite surface underlying the wetland – as indicated by the occurrence of a large discontinuity between the residual saprolite surface and the fresh dolerite surface at the toe of the wetland. The 1m difference in elevation between the residual saprolite and the fresh dolerite surfaces at the contact between the two lithologies, indicates that the residual saprolite and wetland surface within the vicinity of the contact zone has sagged by approximately 1 metre.

The mineralogy of fresh dolerite adjacent to, and the residual saprolite underlying Dartmoor Vlei, indicates that the pyroxene and plagioclase dominated (with a lesser amount of quartz) mineralogy of the fresh dolerite has been converted to kaolinite and micro crystalline quartz by in-situ chemical weathering. In addition, the relative enrichment of the highly resistant quartz and orthoclase minerals confirms that there has been a general loss of material mass as the fresh dolerite has been transformed into residual saprolite. The loss and transformation of minerals and the general loss of material mass indicates that the volume of the dolerite sill underlying the wetland has likely been reduced as the fresh dolerite has been transformed into residual saprolite. Simply put, volume equals mass divided by density. As the mass and density of each of the minerals within each of the samples is known, the volume of each of the dolerite and residual samples could be calculated and the percentage loss in volume determined (Table 6.1).

From the calculated mineral concentrations, the average density of the fresh dolerite within the wetland valley was calculated to be 3.2 g/ml. The average density of the residual clay samples was approximately 2.5g/ml, which is substantially lower than the density of the fresh dolerite. However, despite the decrease in density during the alteration of the parent dolerite to residual clay, there is a significant volume reduction that relates to the leaching of water soluble ions (notably sodium, magnesium and calcium) during the process of in-situ chemical weathering. Table 6.1 indicates that the general 0.7g/ml reduction in the density of the dolerite mass as it has been transformed from fresh dolerite to residual clay, translates into an approximate 2.9% volume loss. As the cross-sectional and longitudinal profiles of the underlying dolerite basin indicate that the residual saprolite surface has sagged relative to the fresh dolerite at the toe of the wetland, the percentage volume loss was translated into an approximate percentage thickness loss to give an indication of the possible reduction in the thickness of the weathered dolerite mass as it has been chemically altered. The average thickness loss that has occurred within the weathered dolerite mass underlying the wetland was calculated to be approximately 8.5%. Table 6.1 also shows that the potential percentage volume and thickness losses within the residual saprolite increases towards the head of the wetland. This is paralleled by an increase in the percentage kaolin and a decrease in the percentage quartz and orthoclase in the residual saprolite towards the head of the wetland. This correlation indicates that a greater amount of weathering has likely occurred within the dolerite underlying the upper reaches of the wetland.

Table 6.1: Calculated density of the residual saprolite samples and the calculated % volume loss and % thickness loss that has taken place as each sample has weathered from fresh dolerite to residual saprolite

Core T2A3 (Lower reaches)							
Depth (cm)	198-208	295-305	400-410	495-505			Average
Density (g/ml)	2.54	2.51	2.50	2.49			2.51
% Volume Loss	2.55	2.67	2.76	2.78			2.69
% Thickness Loss	7.47	7.80	8.04	8.12			7.86
Core T3A4 (Middle reaches)							
Depth (cm)	215-225	235-245	300-310	420-430			Average
Density (g/ml)	2.42	2.47	2.52	2.52			2.48
% Volume Loss	3.15	2.90	2.63	2.65			2.83
% Thickness Loss	9.16	8.45	7.68	7.73			8.26
T4A3 (Upper reaches)							
Depth (cm)	125-135	260-270	420-430	530-540	635-645	730-740	Average
Density (g/ml)	2.39	2.40	2.40	2.47	2.46	2.48	2.43
% Volume Loss	3.30	3.26	3.23	2.90	2.95	2.86	3.08
% Thickness Loss	9.58	9.45	9.38	8.45	8.60	8.33	8.96

The approximate 2.9% volume loss and potential 8.5% thickness loss in the weathered dolerite sill mass as it has been transformed from fresh dolerite into residual saprolite, confirms that the sagging of the residual saprolite surface is a plausible explanation for the discontinuity in the longitudinal basin profile, and that chemical weathering processes are likely responsible for the sagging of the residual clay basin. The occurrence of residual saprolite at depths in excess of 7m, with little evidence of rock material, suggests that the weathering level is significantly deeper than 7m. If, for example, the residual saprolite extends to depths in the vicinity of 10-12m below the ground surface, the residual saprolite mass would be approximately 8-10m thick. Thus, an average thickness loss of 8.5% associated with the chemical transformation of fresh dolerite into residual saprolite is expected to result in an approximate 0.85m loss in thickness. This projected thickness loss would be generally consistent with the difference in elevation that currently exists between the residual saprolite and fresh dolerite surfaces in the lower reaches of the wetland.

The gradual sagging of the wetland surface associated with the mass loss that results from the deep weathering of the dolerite sill provides an adequate explanation for the presence of low unit stream powers, ‘wetland streams’ as defined by Jurmu and Andrle (1997), diffuse valley-bottom wetland conditions, and the predominance of organic sedimentation – despite the wetland being controlled geomorphologically by a dolerite dyke. The gradual sagging of the wetland surface upstream of the dolerite dyke, associated with chemical transformation of the underlying dolerite, is expected to reduce the gradient of the streams and the wetland surface over time. The decrease in the slope of the wetland surface and channels gradually decreases the transport capacity of the streams and results in the deposition of sediment that the stream channels can no longer transport. Over time the stream beds would thus gradually aggrade as the streams adjusted to the gradient change and the vertical accommodation space created by sagging would be slowly filled with fine sediment. Ultimately, continued sagging would reduce the ability of the channels to carry sediment and water, resulting in the predominance of diffuse flow and the termination of clastic sedimentation. Under these conditions organic sedimentation would replace clastic sedimentation and the channels would either disappear or become largely vegetation controlled – much like the palustrine wetland and peatland channels that have been studied by Jurmu and Andrle (1997) and Watters and Stanley (2007).

The sedimentary fill of the wetland generally confirms that the wetland has evolved from a floodplain wetland characterised by mixed-load laterally migrating streams and organic-rich clastic sedimentation to the current valley-bottom wetland that is characterised by diffuse flow and organic sedimentation. There is a clear gradation from organic-rich silt sequences punctuated by gravel channel bed deposits into more peaty and 'mucky' deposits, and ultimately into peat deposits near the surface indicating that organic sedimentation has gradually replaced clastic sedimentation over time and that organic sedimentation is the dominant mode of sedimentation at present.

The predominance of organic sedimentation over the most recent period of wetland evolution indicates that sagging of the wetland basin has occurred more rapidly than the aggradation of the basin by accumulation of clastic sediment. However, the maintenance of relatively defined stream channel segments and an average positive longitudinal gradient at present indicates that organic sedimentation has kept pace with the sagging of the basin and flow is not being significantly impounded at present.

As the thickness loss of the weathered dolerite sill mass has likely not been uniform throughout the wetland, differential sagging is likely responsible for the occurrence of many of the surface features of the wetland. These include the depressions in the lower reaches, elevated ridges in the upper and middle reaches, the series of ponds in the middle reaches, and ultimately the form and behaviour of the discontinuous meandering streams.

The reasons for differential sagging are numerous. However, generally differences in weathering are related to differences in the structure of the rock being weathered. This is evident in the upper reaches where the elevated surfaces within the wetland ridge correlate strongly with the occurrence of dolerite outcrops and hard-pan residual saprolite layers, indicating that the dolerite in this region is compositionally different from the dolerite underlying the rest of the basin. The presence of dolerite outcrops on top of the ridges, and hard-pan saprolite layers underlying the ridges, indicates that the dolerite underlying the ridge is likely more resistant to weathering than the completely weathered homogeneous residual saprolite underlying the valley-bottom arms. This provides an explanation for why the residual saprolite surface underlying the current valley-bottom arms has sagged more than the residual saprolite underlying the more elevated surfaces.

6.2.4 Controls on the chemical weathering of the dolerite sill underlying Dartmoor Vlei

The present reduced blue colours of the majority of the residual saprolite underlying the wetland in combination with the high organic matter content of the alluvial fill and a valley-bottom setting conducive to the permanent saturation of the valley, suggest that the process of hydrolysis is largely responsible for extensive weathering of the dolerite underlying the wetland. The process of hydrolysis requires a continuous supply of water to a site over a long period of time which would ultimately flush out the ions and solutes from the site, thereby promoting the hydrolysis reactions.

The extremely long timescales of valley-bottom saturation required to produce the extensive weathering profiles that currently exist underneath the wetland has been facilitated by the preservation of the African Erosion Surface weathering profiles on the Karkloof Plateau and continued saturation thereafter. Owing to the Miocene and Pliocene uplifts, much of the deeply weathered saprolite profiles present at the end of the Cretaceous were removed by erosion (Partridge and Maud, 2000a; Partridge *et al*, 2006). As a result, most of the wetlands that have been studied in South Africa occur on or below the Post African I and II Erosion Surfaces that have been subject to minimal weathering due to the shorter time period of saturation, and which have thus not weathered to a substantial depth. In addition, subsequent to the Cretaceous, continued chemical weathering of the dolerite underlying Dartmoor Vlei has been facilitated by the increasingly permanently wet conditions that have characterised the long-term evolution of Dartmoor Vlei, as indicated by the high organic matter content of the majority of the sedimentary fill of Dartmoor Vlei.

An exception to the deep weathering of the dolerite sill underlying the wetland is the preservation of a small portion of laterally planed fresh dolerite underlying the toe of the wetland. However, owing to the close association of this portion of fresh dolerite with the dolerite dyke that cuts across the toe of the wetland, it is likely that this portion of fresh dolerite has been altered by the process of contact metamorphism associated with the intrusion of the dolerite dyke, which has increased its hardness and resistance to the weathering.

6.3 A conceptual model of the origin and evolution of Dartmoor Vlei

Figure 6.1 gives a schematic illustration of the three broad stages in the evolution of Dartmoor Vlei. Firstly, the broad valley that hosts Dartmoor Vlei was initially laterally planed during the Cretaceous by laterally migrating meandering streams in a manner that is consistent with the conceptual model of floodplain formation developed by Tooth *et al* (2002a). During this time, the Dartmoor Valley and the streams planing the valley were larger than those that currently exist due to the larger stream catchments that existed prior to the Miocene and Pliocene uplifts. The end-product of the phase of lateral planation in terms of the longitudinal valley gradient is illustrated in profile A of Figure 6.1. During this time the humid conditions and saturation of the soils in places within the floodplain facilitated the start of the chemical weathering of the dolerite sill underlying the initial floodplain.

Nearing the end of the Cretaceous, the lateral planation of the Dartmoor Valley gradually terminated due to the reduction in the longitudinal gradients and discharges of the streams feeding the Dartmoor Valley associated with the occurrence of the African Erosion Surface. This resulted in the gradual aggradation of the stream channels, the reduction in the longitudinal gradients of the wetland and streams, and the increased saturation of floodplain soils which facilitated further chemical weathering of the dolerite sill underlying the floodplain. By the end of the Cretaceous, limited sagging of the valley had likely started to occur as a result of the substantial chemical weathering of the dolerite sill during the Cretaceous as indicated in Profile B of Figure 6.1. The sagging resulted in the creation of accommodation space for clastic deposition and the further reduction in the longitudinal gradients of the wetland and stream channels (Figure 6.1; B).

Since the inception of sagging, the residual saprolite basin has sagged more rapidly than the dolerite dyke local base level has been lowered and eroded, which has resulted in the gradual impoundment of flow, the reduction in the longitudinal stream gradients, loss of stream definition, and the predominance of diffuse flow within the wetland. Ultimately, continued sagging has resulted in the reduction in the ability of the channels to carry sediment and water and the termination of clastic sedimentation. As a result, the recent aggradation of the valley has generally involved peat accumulation within the accommodation space created by the continued sagging of the basin (Figure 6.1; C).

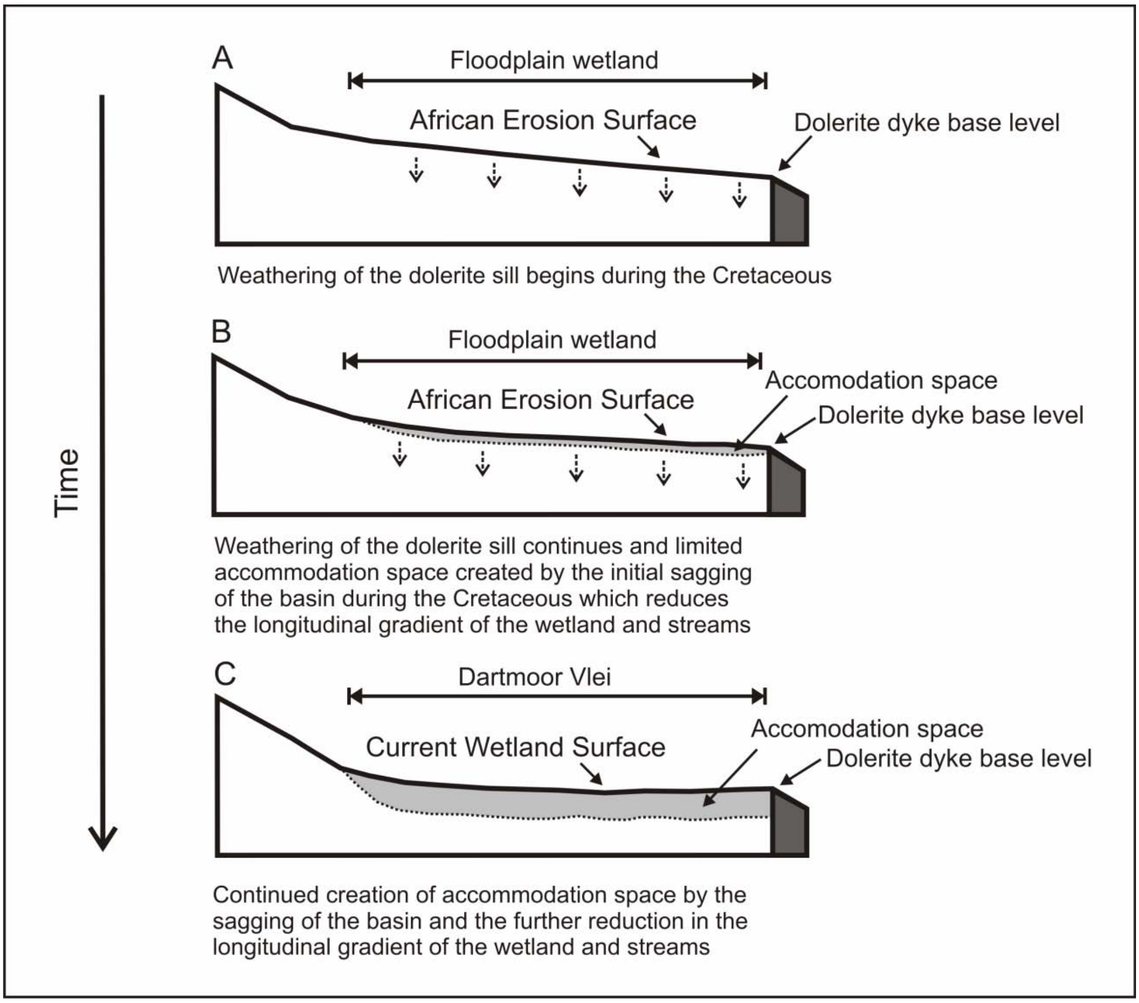


Figure 6.1: Schematic illustration showing the different phases in the evolution of Dartmoor Vlei.

CHAPTER 7 - CONCLUSION

At present, most of the floodplain and valley-bottom wetlands that have been studied in South Africa have formed under the climatic and tectonic conditions that have prevailed since the Pliocene, such that they are located within an incising drainage network below the African Erosion Surface. However, the identification of Karkloof Plateau as the remains of the African Erosion Surface indicates that Dartmoor Vlei has been the product of the geomorphic processes that were operative under the climatic and tectonic conditions of the Cretaceous, and that Dartmoor Vlei has been marginally affected by the Miocene and Pliocene uplift events. The humid climates of the Cretaceous together with the long-term saturation of the soils within the Dartmoor Valley have facilitated the extensive in-situ chemical weathering of the dolerite sill underlying the wetland into residual saprolite extending to depths in excess of 7m below the ground surface. The long-term saturation of the soils within the Dartmoor Valley was made possible by the lateral planation and widening of the valley during the Cretaceous by laterally migrating streams consistent with the processes described in the conceptual model of floodplain wetland formation developed by Tooth *et al* (2002a). The gentle longitudinal gradients of the early floodplain environments would have encouraged the spreading out of flow and the increased saturation of soils conducive to the start of in-situ chemical weathering of the dolerite sill. The gradual reduction in stream power associated with extensive planing of the KwaZulu-Natal Province nearing the end of the Cretaceous would have further increased the saturation of the floodplain soils and enhanced the rate of in-situ chemical weathering.

The occurrence of a near-vertical discontinuity at the contact between the residual saprolite and fresh dolerite bedrock underlying the toe of the wetland indicates that the residual saprolite basin has sagged relative to the dolerite dyke local base level. Chemical and mineralogical analysis of the fresh dolerite and residual saprolite underlying the wetland have confirmed that the sagging of the residual saprolite surface has been the product of volume and thickness losses in the weathered dolerite sill mass that have taken place during the in-situ chemical weathering of the dolerite sill. The process of basin sagging and resultant creation of accommodation space within the Dartmoor Valley provides an adequate explanation for the occurrence of diffuse valley-bottom wetland conditions, low unit stream powers, discontinuous streams, and peat formation – despite the wetland being controlled geomorphologically by a dolerite dyke.

This is the first time that the process of basin sagging, resulting from the long-term in-situ chemical weathering of igneous rocks, has been documented as a major control on the origin and evolution of a valley-bottom wetland in South Africa. This has implications for all wetlands currently located on the remains of the African Erosion surface as well as for the future rehabilitation of these valley-bottom wetland systems.

Although basin sagging associated with the transformation of fresh dolerite into residual clay provides an explanation for the current hydro-geomorphology of Dartmoor Vlei, there is still a need to undertake studies on similar depressional wetlands located on the remains of the African Erosion Surface to confirm that basin sagging is a major control within wetlands underlain by lithologies susceptible to chemical weathering on the African Erosion Surface.

Furthermore, it is evident from this study that South African wetlands on the remains of the African Erosion Surface are the product of the fluvial geomorphic processes operative under the climatic and tectonic conditions of the Cretaceous. In light of the poorly constrained ages of the sedimentary fill described in this study, the dating of these sediments would be invaluable in constructing a more detailed picture of the evolution of Dartmoor Vlei during the Cretaceous and thereafter.

REFERENCES

Allen, J. R. (1970). *Physical Processes of Sedimentation: An Introduction, Earth Science Series*. George Allen and Unwin Ltd, London.

Begg, G. W. (1989). *The Wetlands of Natal (Part 3): The location, status and function of the priority wetlands of Natal*. Natal Town and Regional Planning Commission, Pietermaritzburg. Report 73.

Bredell, J. H., Barnardo, D. J. and Horn G. F. J. (1980). *A Reconnaissance Survey of Bauxitic Material in Southern Natal – Report 1: The Area 2930 A – North of Howick*. Council for Geoscience, Pretoria.

Breen, C. M. and Begg, G. W. (1989). Conservation status of southern African wetlands. *In*: Huntley, B. J. (Ed.), *Biotic diversity in southern Africa*. Oxford University Press, Cape Town, 254-264.

Bridge, J. S. (2003). *Rivers and Floodplains: Forms, Processes, and Sedimentary Record*. Blackwell Science, London.

Brierly, G. and Fryirs, K. (1999). Tributary-trunk stream relations in a cut-and-fill landscape: a case study from Wolumla Catchment, New South Wales, Australia. *Geomorphology* **28**, 61-73.

Brinson, M. M. (1993). *A Hydrogeomorphic Classification for Wetlands*. US Army Corps of Engineers, Washington DC. Technical Report WRP-DE-4.

Chang, H. H. (1980). Stable Alluvial Canal Design. *Journal of the Hydraulics Division* **106** (HY5), 873-891. American Society of Civil Engineers.

Collinson, J. D. (1996). Alluvial Rivers. *In*: Reading, H. D. (Ed.), *Sedimentary Environments: Processes, Facies and Stratigraphy, Third Edition*. Blackwell Science, Oxford, 37-82.

Dardis, G. F., Beckedahl, H. R. and Stone, A. W. (1988). Fluvial Systems. *In*: Moon, B. P. and Dardis, G. F. (Eds.), *The Geomorphology of Southern Africa*. Southern Book Publishers, Johannesburg, 31-48.

Ellery, W. N., Grenfell, M. C., Grenfell, S. E., Kotze, D. C., McCarthy, T. S., Tooth, S., Grundling, P-L., Beckedahl, H., le Maitre, D. and Ramsay, L. (2009). *WET-Origins: Controls on the distribution and dynamics of wetlands in South Africa*. Water Research Commission, Pretoria. WRC Report No. TT 334/08.

Ewart-Smith, J. L., Ollis, D. J., Day, J. A. and Malan, H. L. (2006). *National Wetland Inventory: Development of a Wetland Classification System for South Africa*. Prepared for the Water Research Commission (WRC) and South African National Biodiversity Institute (SANBI). WRC Report: KV 174/06.

Gordon, N. D., McMahon, T. A. and Finlayson, B. L. (1992). *Stream Hydrology: An Introduction for Ecologists*. Wiley, Chichester.

Grenfell, M. C (2007). *The geomorphic origin and evolution in the KwaZulu-Natal Drakensberg foothills: Implications for rehabilitation*. M.Sc. Thesis, University of KwaZulu-Natal, Durban.

Grenfell, M. C., Ellery, W. N. and Grenfell, S. E. (2008). Tributary valley impoundment by trunk river floodplain development: a case study from the KwaZulu-Natal Drakensberg foothills, eastern South Africa. *Earth Surface Processes and Landforms* **33**, 2029-2044.

Haddon, I. G. and McCarthy, T. S. (2005). The Mesozoic-Cenozoic interior sag basins of Central Africa: The Late-Cretaceous-Cenozoic Kalahari and Okavango basins. *Journal of African Earth Sciences* **43**, 316-333.

Hutchison, C. S. (1974). Laboratory Handbook of Petrographic Techniques. Wiley-Interscience, Toronto.

Jones, H. L. and Hajek, E. A. (2007). Characterizing avulsion stratigraphy in ancient alluvial deposits. *Sedimentary Geology* **202**: 124-137.

Joubert. R. (2009). *The origin of Wakkerstroom Vlei, Mpumalanga Province, South Africa*. M.Sc. Thesis, University of KwaZulu-Natal, Durban.

Jurmu, M. C. and Andrle, R. (1997). Morphology of a Wetland Stream. *Environmental Management* **21**(6), 921-941.

Knighton, D. (1998). *Fluvial forms and processes: a new perspective*. Arnold Publishers Ltd, London.

Kotze, D. C., Ellery, W. N., Patrick, B. and Bambus, O (2009) An assessment of the effect of rehabilitation interventions on the health of Dartmoor Vlei, KwaZulu-Natal Midlands. *In*: *WET-Outcome Evaluate: An evaluation of the rehabilitation outcomes at six wetland sites in South Africa*. Water Research Commission, Pretoria. WRC Report No. TT 343/08.

Kotze, D. C., Marneweck, G. C., Batchelor, A. L., Lindley, D. S. and Collins, N. B. (2005). WET-EcoServices: a technique for rapidly assessing ecosystem services supplied by wetlands. Report submitted to the Water Research Commission, Pretoria.

Leopold, L. B. and Maddock, T. J. (1953). *The Hydraulic Geometry of Stream Channels and Some Physiographic Implications*. U. S. Geological Survey Professional Paper 252.

Leopold, L. B., Wolman, M. G. and Miller, J. P. (1964). *Fluvial Processes in Geomorphology*. W. H. Freeman and Company, San Francisco.

Leopold, L. B. and Bull, W. B. (1979). Base Level, Aggradation, and Grade. *Proceedings of the American Philosophical Society* **123**(3), 186-202.

MacVicar, C. N., De Villiers, J. M., Loxton, R. F., Verster, E., Lambrechts, J. J. N., Merry-Weather, F. R., Le Roux, J., Van Rooyen, T. H. and Harmse, H. J. von. M. (1977). *Soil classification: a binomial system for South Africa*. Department of Agriculture and Technical Services, Pretoria, Science Bulletin 390.

McCarthy, T. C., Barry, M., Bloem, A., Ellery, W. N., Heister, H., Merry, C. L., Ruther, H. and Sternberg, H. (1997). The gradient of the Okavango fan, Botswana, and its sedimentological and tectonic implications. *Journal of African Earth Sciences* **24**(1/2), 65-78.

McCarthy, T. S. and Hancox, P. J. (2000). Wetlands. *In*: Partridge, T. C. and Maud, R. R. (Eds.), *The Cenozoic of Southern Africa: Oxford Monographs on Geology and Geophysics*. Oxford University Press, Oxford, 218-235.

McCarthy, T. S. and Rubidge, B. (2005). *The Story of Earth and Life: A southern African perspective on a 4.6-billion-year journey*. Struik Publishers, Cape Town.

Mitsch, W. J. and Gosselink, J. G. (1993). *Wetlands. 2nd ed*. Van Nostrand Reinhold, New York.

Morisawa, M. (1968). *Streams: their dynamics and morphology*. McGraw-Hill, New York.

Mucina, L. and Rutherford, M. C. (Eds.) (2006). The Vegetation of South Africa, Lesotho and Swaziland. Strelitzia 19. South African National Biodiversity Institute, Pretoria.

Myeza, M. (2007). Using kinematic method for contouring a wetland area. Undergraduate Research project, School of Civil Engineering, Surveying and Construction, University of KwaZulu-Natal, Durban.

Nichols, G. (1999). *Sedimentology and Stratigraphy*. Blackwell Science, Oxford.

Partridge, T. C. and Maud, R. R. (2000a). Macro-scale geomorphic evolution of southern Africa. *In*: Patridge, T. C. and Maud, R. R. (Eds.), *The Cenozoic of Southern Africa*. Oxford University Press, New York, 145-161.

Partridge, T. C., Botha, G. A. and Haddon, I. G. (2006). Cenozoic Deposits of the Interior. In: Johnson, M. R., Anhaeusser, C. R. and Thomas, R. J. (Eds.), *The Geology of South Africa*. Geological Society of South Africa, Johannesburg/Council for Geoscience, Pretoria, 585-604.

Philips, J. D. and Lutz, J. D. (2008). Profile convexities in bedrock and alluvial streams. Geomorphology **102**: 554-566.

Ritter, D.F., Kochel, C.R. and Miller, J.R. (2002). *Process Geomorphology, 4th Edition*. McGraw Hill, New York.

Rogers, K. H. (1997). Freshwater Wetlands. *In*: Cowling, R. M., Richardson, D. M. and Pierce, S. M. (Eds.), *Vegetation of southern Africa*. Cambridge University Press, 322-347.

Schulze, R. E. (1997). South African Atlas of Agrohydrology and Climatology. Water Research Commission, Pretoria, WRC Report TT82/96.

Schumm, S. A. (1977). *The Fluvial System*. John Wiley & Sons, New York.

Schumm, S. A. (2005). *River Variability and Complexity*. Cambridge University Press, New York.

Slingerland, R. and Smith, N. D. (2004). River avulsions and their deposits. *Annual Review of Earth and Planetary Sciences*. **32**, 257-285.

Smith, D. (2006). *Geomorphological Evidence for the Origin and Evolution of a Wetland: Case Study of Kruisfontein Wetland, Mooi River, KwaZulu-Natal*. B.Sc. Hons. Thesis, University of KwaZulu-Natal, Durban.

Thorn, C. E. and Welford, M. R. (1994). The Equilibrium Concept in Geomorphology. *Annals of the Association of American Geographers* **84**(4), 666-696).

Tooth, S. (2000a). Process, form and change in dryland rivers: a review of recent research. *Earth-Science Reviews* **51**, 61-107.

Tooth, S., McCarthy, T. S., Brandt, D., Hancox, P. J. and Morris, R. (2002a). Geological controls on the formation of alluvial meanders and floodplain wetlands: the example of the Klip River, Eastern Free State, South Africa. *Earth Surface Processes and Landforms* **27**, 797-815.

Tooth, S., McCarthy, T. S., Hancox, P. J., Brandt, D., Buckley, K., Nortje, E. and McQuade, S. (2002b). The geomorphology of the Nyl River and floodplain in the semi-arid Northern Province, South Africa. *South African Geographical Journal* **84** (2), 226-237.

Tooth, S., Brandt, D., Hancox, P. J. and McCarthy, T. S. (2004). Geological controls on alluvial river behaviour: a comparative study of three rivers on the South African Highveld. *Journal of African Earth Sciences* **38**, 79-97.

Tooth, S. and McCarthy, T. S. (2007). Wetlands in drylands: geomorphological and sedimentological characteristics, with emphasis on examples from southern Africa. *Progress in Physical Geography* **31** (1), 3-41.

Tooth, S., Rodnight, H., McCarthy, T. S., Duller, G. A. T. and Grundling, A. T. (2008). Late Quaternary dynamics of a South African floodplain wetland and the implications for assessing recent human impacts. *Geomorphology* **106**, 278-291.

Tyson, P. D. and Partridge, T. C. (2000). Evolution of Cenozoic Climates. *In*: Patridge, T. C. and Maud, R. R. (Eds.), *The Cenozoic of Southern Africa*. Oxford University Press, New York, 371-386.

Watters, J. R. and Stanley, E. H. (2007). Stream channels in peatlands: The role of biological processes in controlling channel form. *Geomorphology* **89**, 97-110.

Wolman, M. G. and Leopold, L. B. (1957). *River Flood Plains: Some Observations On Their Formation*. U. S. Geological Survey Professional Paper 282-C.

Wust, R. A. J., Bustin, R. M. and Lavkulich, L. M. (2003). New classification systems for tropical organic-rich deposits based on studies of the Tasek Bera Basin, Malaysia. *Catena* **53**, 133-163.

http://www.agis.agric.za/agismap_atlas/AtlasViewer.jsp?MapService=agis_atlas2006&ProjectId=5&LId=0&OId=0&LayerIdVisList=none.

APPENDIX A – SURVEY DATA

Transect 1	T	M	B	Bearing	Bedrock	Dist	Rel. Dist	Rel. Ht	Bedrock
1(T4A1)	7.3	4.4	2	0	0.37	53	0	-0.44	-0.81
2	9.4	7.2	5.1	0	0	43	10	-0.72	-0.72
3	10.1	8.4	6.7	0		34	19	-0.84	
4	12.4	11	9.45	0		29.5	23.5	-1.1	
5	13.5	12.1	10.6	0		29	24	-1.21	
6	14.7	13.5	12.4	0		23	30	-1.35	
7(T4A2)	15	14.9	13.9	0	1.08	11	42	-1.49	-2.57
8(T4A3)	15.5	14.9	14.45	0	1.25	10.5	42.5	-1.49	-2.74
9(T4A4)		15.85		0	1.5	0	53	-1.585	-3.085
10(T4A5)	19.05	18.48	17.75	180	1.08	13	66	-1.848	-2.928
11	20.5	19.2	18.05	180		24.5	77.5	-1.92	
12(T4A6)	21.5	19.75	18	180	1.35	35	88	-1.975	-3.325
13	22.5	20.1	17.7	180		48	101	-2.01	
14(T4A7)	23.2	20.35	17.5	180	1.3	57	110	-2.035	-3.335
15 (furrow bank)	21.45	18	14.55	180		69	122	-1.8	
back on p15	18	16.6	15.25	0		27.5			
16 (furrow bed)	18	24.4	15.25	0		27.5	122	-2.58	
18 (furrow bed)	17.45	23.3	14.75	0		27	123	-2.47	
17 (furrow bank)	17.45	16.05	14.75	0		27	123	-1.745	
19 (channel 2 bank)	19.2	18.25	17.35	0		18.5	131	-1.965	
20(T4A8)	19.9	19.05	18.1	0	1.36	18	131.5	-2.045	-3.405
21(T4A9)	20.5	19.85	19.25	0	1.8	12.5	137	-2.125	-3.925
22(T4A10)	17.6	17.1	16.7	0	1.9	9	140.5	-1.85	-3.75
23(T4A11)	17.65	17.3	16.95	0	1.91	7	142.5	-1.87	-3.78
24		15.6				0	149.5	-1.7	
25	16.19	16.01	15.85	180		3.4	152.9	-1.741	
26 (channel 1 bank)	19.5	19.3	19.1	180		4	153.5	-2.07	
27 (channel 1 bed)	19.5	21.9	19.1	180		4	153.5	-2.33	
28(T4A12)	22.5	22.35	22	180	0.43	5	154.5	-2.375	-2.805
29 (channel 1 bank)	20.4	20.15	19.85	180		5.5	155	-2.155	
30	18.75	18.4	18.05	180		7	156.5	-1.98	
31	16.75	16.1	15.5	180		12.5	162	-1.75	
32(T4A13)	17	16.2	15.45	180	0.73	15.5	165	-1.76	-2.49
33	16	14.7	13.4	180		26	175.5	-1.61	
34(T4A14)	16.4	14.7	13.1	180	0.71	33	182.5	-1.61	-2.32
35	14.05	12.2	10.3	180		37.5	187	-1.36	

Transect 2												
	T	M	B	Bearing	Dist	Rel. Dist(m)	Rel. Ht(m)	Soft Bedrock	Water table	G1	G2	Peat
Pt 1 Stake	23.5	22.25	21.05	103	24.5	24.5	-2.225					
Pt 2	25.5	24.1	22.7	105	28	28	-2.41					
Pt 3	28.5	26.85	25.2	108	33	33	-2.685					
Pt 4	31.9	29.5	27.1	109	48	48	-2.95					
Pt 5 (Channel Bank)	31.9	29.45	27	108	49	49	-2.945					
Pt 6 (Channel Bed)	34.85	32.3	29.8	108	50.5	49.5	-3.23					
Pt 7 (Channel Bank)	32.2	29.6	27	108	52	50	-2.96					
Pt 8 (Auger 1)	30.6	27.8	24.7	109	59	59	-2.78	-4.84	-3.18	-4.78		-3.18
Pt 9	31.5	27.7	23.8	109	77	77	-2.77					
Pt 10 (Furrow Bank)	32.65	28.45	24.25	108	84	84	-2.845					
Pt 11 (Water Level)	39.3	35.1	30.8	107	85	84	-3.7		-3.51			
Pt 12 (Channel Bottom)	41.3	37	32.8	106.5	85	85	-3.7					
Pt 13 (Furrow Bank)	32.6	28.2	23.9	107	87	86	-2.82					
Back on Pt 13	18.7	15.8	12.4	291	63							
Pt 14 (Auger 2)	18.4	16.5	14.2	291	42	108	-2.89	-4.59	-3.02	-3.99		-3.19
Pt 15 Channel Bank	17.25	16.2	15.1	288	21.5	128.5	-2.86	-4.46		-4.06		
Pt 16 (Water level)	18.65	17.85	17.05	287	16	135	-3.025		-3.025			
Pt 17 (Channel Bottom)	21.25	20.5	19.8	284	14.5	135.5	-3.29	-4.64				
Pt 18 Channel bank	18.5	17.85	17.2	284	13	137	-3.025					
Pt 19	15.4	15.05	14.65	285	7.5	142.5	-2.745					
Pt 20 (Auger 3)	15.6	14.55	13.5	97	21	171	-2.695	-4.445	-2.745	-3.845	-4.195	-2.945
back on Pt 20	17.1	15.8	14.8		23							
Pt 21 (A4)	17.5	16.6	15.65		18.5	175.5	-2.775	-4.835				
Pt 22 (A5)	18.55	17.7	16.85		17	177	-2.885	-4.725				
Pt 23 (A6)	18.3	17.55	16.8		15	179	-2.87	-4.61				
Pt 24 (A7)	18.7	18.15	17.65		10.5	183.5	-2.93	-4.46				
Pt 25 (A8)	18.15	17.85	17.35		8	186	-2.9	-4.17				
Pt 26 (A9)	17.45	17.15	16.85		6	188	-2.83	-3.61				
Pt 27 (A10)		15.3			0	194	-2.645	-3.245				
Pt 28	16.75	16.3	15.85		9	209	-2.745					
Pt 29	9.9	9.15	8.45		14.5	214.5	-2.03					

Transect 3								
	T	M	B	Bearing	Rel Ht	Rel Dist	Corr. Ht	Soft Bedrock
Pt1 (Stake 1)	4.85	4.5	4.2	310	4.5	0	-0.45	
Pt2	7.95	7.7	7.45	311	7.7	1.5	-0.77	
Pt3	10.6	10.45	10.32	311	10.45	3.7	-1.045	
Pt4	12.85	12.75	12.65	309	12.75	4.5	-1.275	
Ht of Dumpy		15.15			15.15	6.5	-1.515	
Pt5	16.72	16.57	16.42	140	16.57	9.5	-1.657	
Pt6	17.6	17.05	16.5	139	17.05	17.5	-1.705	
Pt 7 Auger 1	18.6	17.6	16.6	137	17.6	26.5	-1.76	-2.76
Pt8 Auger 2	19.7	17.7	15.7	141	17.7	46.5	-1.77	-3.67
Pt 9 Furrow bank	21.9	19.7	17.5	139	19.7	50.5	-1.97	
Pt10 furrow bed	30.5	28.2	26	139	28.2	50.5	-2.82	
						51.5	-2.82	
Pt11 furrow bank	24.1	21.8	19.5	139	21.8	52	-2.18	
Pt12	20.6	18	15.4	139	18	58.5	-1.8	
Pt13	20.6	17.2	13.8	140	17.2	74.5	-1.72	
Pt14	22.2	18.1	14	137	18.1	88.5	-1.81	
Pt15	22.85	17.5	12.25	142	17.5	112.5	-1.75	
Pt16	24	18	12	143	18	126.5	-1.8	
Pt17 Small meandering channel bed	25.7	19.5	13.3	143	19.5	130.5	-1.95	
Pt18	20.5	15.2	9.9	311	17.3	132	-1.73	
Pt19 Auger 3	19.3	14.4	9.6	308	16.5	141.5	-1.65	-3.85
Pt20	19.8	15.7	11.5	307	17.8	155.5	-1.78	
Pt21 furrow bank	23.1	19.6	16.1	307	21.7	169	-2.17	
Pt22 furrow bed	28.5	25.1	21.6	307	27.2	169	-2.72	
						170	-2.72	
Pt23 furrow bank	19.3	16	12.5	306	18.1	171	-1.81	
Pt24	18.6	16.1	13.6	305	18.2	189	-1.82	
Pt25	17.8	16.3	14.85	288	18.4	211.5	-1.84	
Pt26	15.6	14.7	13.8	256	16.8	228.5	-1.68	
Pt27 Auger 4	16.2	15.3	14.4	211	17.4	242	-1.74	-3.89
Pt28	15.9	14.75	13.6	169.5	16.85	257	-1.685	
Pt29 channel bank	15.6	13.9	12.2	153	16	270.5	-1.6	
Pt30 channel bed	20.6	18.8	17	152	20.9	271.5	-2.09	
Pt31 channel bank	15.85	14	12.2	151	16.1	272.5	-1.61	
Pt32	17.2	15.8	12.3	145	17.9	286.5	-1.79	
Pt33	15.4	12.4	9.4	142	14.5	297.5	-1.45	
Pt34	14.2	10.3	6.4	137	12.4	316	-1.24	
Pt35	12.6	7.6	2.6	134	9.7	338.5	-0.97	
Pt36	33.9	29.5	25	301	6.8	367.5	-0.68	
Pt37 Auger 5	27.85	25.3	22.8	296	2.6	406	-0.26	-2.26
Pt38	21.8	20.5	19.3	295	-2.2	431.5	0.22	
Ht of Dumpy		15.5			-7.2	456.5	0.72	
Pt 39	11.7	10.7	9.75	113	-12	476	1.2	
Pt40	8.35	6.6	4.85	123	-16.1	491.5	1.61	
Pt41	46.8	43.2	39.8	303	-22.4	510.5	2.24	
Pt42 Auger 6	40	37.6	35.3	306	-28	533.5	2.8	1.8
Pt43 fence	38.4	36.3	34.1	306	-29.3	537.5	2.93	
Pt44	34.1	32.7	31.4	305	-32.9	553.5	3.29	
Pt45	29.8	29.2	28.5	305	-36.4	567.5	3.64	
Pt46	29.4	29	28.45	304	-36.6	571	3.66	
Pt47	22.9	22.7	22.45	306	-42.9	576	4.29	
Ht of Dumpy		15.4			-50.2	580.5	5.02	
Pt 48	11	10.8	10.65	129	-54.8	584	5.48	
Pt49	7.8	7.52	7.27	131	-58.08	585.8	5.808	

Transect 4										
	T	M	B	Bearing	Rel. Ht(dm)	Rel.Dist(m)	Rel. Ht (m)	Corrected Rel. Ht	Bedrock	WT
Ht of Dumpy		15.9			15.9	0	1.59	-1.59		
Pt1	20.5	20.35	20.25	147	20.35	2.5	2.035	-2.035		
Pt2	23.65	23.45	23.2	144	23.45	4.5	2.345	-2.345		
Pt3 Stake	25.2	24.9	24.6	142	24.9	6	2.49	-2.49		-3.19
Pt4 Fence	29.9	29.3	28.8	140	29.3	11	2.93	-2.93		
Pt5 Wetland Edge	32.4	31.7	31.05	143	31.7	13.5	3.17	-3.17		
Pt6	37.8	36.8	35.8	142	36.8	20	3.68	-3.68		
Pt7	40.1	38.4	36.8	142	38.4	33	3.84	-3.84		
Pt8 Auger1	41.1	38.5	35.9	140	38.5	52	3.85	-3.85	-5.35	-3.9
Pt9	39.9	36	32.2	142	36	77	3.6	-3.6		
Pt10	40	35.3	30.7	143	35.3	93	3.53	-3.53		-3.93
Pt11	44.1	38.5	32.8	144	38.5	113	3.85	-3.85		
Pt12 Furrow Bank	47	40.1	33.2	143	40.1	138	4.01	-4.01		
						138		-4.66		
Pt13 Furow Bed	32.9	27.1	21.3	322	46.6	139	4.66	-4.66		-4.56
Pt14 Furrow Bank	24.8	19	13.3	322	38.5	139	3.85	-3.85		
Pt15	23.3	19.3	15.2	321	38.8	174	3.88	-3.88		-4.13
Pt16 Channel Bank 1	24.5	20.6	16.6	321	40.1	176	4.01	-4.01	-5.26	
Pt17 Channel Bottom 1	26.7	22.8	18.9	321	42.3	176.5	4.23	-4.23	-5.23	-4.13
Pt18 Channel bank 1	25	21.1	17.3	321	40.6	177	4.06	-4.06	-5.21	
Pt19	20.6	17.1	13.4	321	36.6	183	3.66	-3.66		
Pt20	17.8	15	12.1	322	34.5	198	3.45	-3.45		-3.95
Pt21	17.05	15.8	14.6	321	35.3	230.5	3.53	-3.53		
Pt22 Auger 2	16.3	15.6	15	324	35.1	242	3.51	-3.51	-4.76	-4.21
Ht of Dumpy		15.9			35.4	255	3.54	-3.54		
Pt23	18.65	17.8	16.9	137	37.3	272.5	3.73	-3.73		
Pt24	21.5	19.8	18	135	39.3	290	3.93	-3.93		
Pt25	23.7	20.9	18.2	134	40.4	310	4.04	-4.04		
Pt28 Auger3	25.7	22	18.2	132	41.5	330	4.15	-4.15	-5.9	-4.6
Pt 29 Left Bank 2	18.7	13.3	7.7	323	45.1	340	4.51	-4.51	-5.91	
Pt 30 Channel Bank 2	18.2	12.8	7.3	323	44.6	341	4.46	-4.46		
Pt 31 Channel Bottom 2	20.1	14.8	9.3	323	46.6	341.5	4.66	-4.66		-4.54
Pt 32 Channel Bank 2	18.6	13.3	7.8	323	45.1	342	4.51	-4.51		
Pt 33 Right Bank 2	18.3	13.1	7.7	323	44.9	344	4.49	-4.49		
Pt 34	19.1	15	10.8	323	46.8	367	4.68	-4.68		
Pt 35 Auger 4	19.5	16.6	13.8	324	48.4	393	4.84	-4.84	-7.44	-5.04
Pt 36	18.8	16.8	14.7	324	48.6	409	4.86	-4.86		
Pt 37 Auger 5	18.2	17.1	16	329	48.9	428	4.89	-4.89	-7.14	-5.24
Pt 38	16.7	16.1	15.5	327	47.9	438	4.79	-4.79		
Ht of Dumpy		15.1			46.9	450	4.69	-4.69		
Pt 39 Left Bank 3	17.6	16.95	16.35	167	48.75	462.5	4.875	-4.875	-6.975	
Pt40 Channel Bank 3	19.2	18.45	17.7	164	50.25	465	5.025	-5.025		
Pt41 Channel Bottom 3	22.2	21.4	20.65	164.5	53.2	465.5	5.32	-5.32		-5.22
Pt42 Channel Bank 3	19.2	18.4	17.65	165	50.2	466	5.02	-5.02		
Pt43 Right Bank 3	17.65	16.8	16	165.5	48.6	466.5	4.86	-4.86	-6.86	
Pt44 Furrow Bank	18.8	17.8	16.8	157	49.6	470	4.96	-4.96		
Pt45 Furrow Bed	26.65	25.65	24.65	154	57.45	470	5.745	-5.745		-5.445
						471		-5.745		
Pt46 Furrow Bank	16.95	15.9	14.85	154	47.7	471	4.77	-4.77		
Pt47	17.8	16	14.1	146	47.8	487	4.78	-4.78		
Pt48	16.2	13.6	10.9	143	45.4	503	4.54	-4.54		
Pt 49 Auger 6	14.3	11.1	7.8	140	42.9	515	4.29	-4.29	-6.14	-4.7
Pt50	15.2	11.1	6.8	140	42.9	534	4.29	-4.29		-4.66
Pt 51	14.7	9.8	4.9	140	41.6	548	4.16	-4.16		
Pt52	12.7	7.5	2.3	140	39.3	554	3.93	-3.93		
Pt 53	23.8	23.3	22.8	318	33.65	562.5	3.365	-3.365		
Pt 54	19.8	19.55	19.35	316	29.9	568	2.99	-2.99		
Ht of Dumpy		15.2			25.55	572.5	2.555	-2.555		

Long Profile	T	M	B	Bear	Bedrock	Dist	Dist with bearing	Rel. Dist	Rel. Ht	
1	7.8	6.45	5.1	96		27	0	0	-0.645	
2	15.8	15.2	14.7	86		11	11	11	-1.52	
3	38.5	37.8	37.05	334		14.5	21	32	-3.78	
4	47.65	46.25	44.85	340		28	13.5	45.5	-4.625	
back on p4	17.4	15.6	13.8	142		36		45.5		
5	24.2	23.5	22.9	116		13	23	68.5	-5.415	
6	27.2	26.4	25.5	336		17	31	99.5	-5.705	
7	32	30.65	29.35	322		26.5	10	109.5	-6.13	
8	39.2	37	35.2	328		40	15	124.5	-6.765	
9	45.7	42.8	39.9	319		58	23	147.5	-7.345	
back on p9	24.45	23.4	22.4			20.5		147.5		
10 (L1A1)	27.4	25.9	24.5	25	100	29	20.5	168	-7.595	-8.595
11	31.4	28.9	26.6	28		48	19	187	-7.895	
12 (L1A2)	32.2	28.6	24.8	28	135	74	26	213	-7.865	-9.215
back on p12	16.3	12.8	9.3	173		70		213		
13 (L1A3)	18.65	16.85	15.05	182	210	36	35	248	-8.27	-10.37
14	20.65	20.25	19.8	175		8.5	27	275	-8.61	
15	24.7	22.7	20.7	317		40	47	322	-8.855	
16 (L1A4)	25.3	22.2	19	323	130	63	24	346	-8.805	-10.105
17	29.1	26.1	23.1	4		60	42	388	-9.195	
18	29.7	26.2	22.6	17		71	19	407	-9.205	
19 (L1A5)	29.1	24.7	20.4	35	162	87	30	437	-9.055	-10.675
back on p19	21.5	17.5	13.5	221		80		437		
20	22.5	19.8	17.3	211		52	30	467	-9.285	
21 (L1A6)	17.9	17.1	16.3	212	168	16	35.5	502.5	-9.015	-10.695
22	18.65	18.4	18.15	204		5	11	513.5	-9.145	
23	17.75	16.9	16.05	1		17	23	536.5	-8.995	
24	21.85	19.8	17.7	332		41.5	27	563.5	-9.285	
25 (L1A7)	23.85	20.5	17.1	3	179	67.5	38.5	602	-9.355	-11.145
26	28.45	23.8	19.25	5		92	24.5	626.5	-9.685	
back on p26	24.4	21	17.55	212		68.5		626.5		
27 (L1A8)	19.6	17.65	16.65	243	176	29.5	45.5	672	-9.35	-11.11
28	20.95	20.05	19.3	256		16.5	14	686	-9.59	
29 (L1A9)	21.35	20.4	19.45	333	119	19	22	708	-9.625	-10.815
30 (L1A10)	24.4	21.7	19.1	355	118	53	36	744	-9.755	-10.935
31	24.25	21.1	18.1	9		61.5	16	760	-9.695	
back on p31	20	17.4	14.8	246		52		760		
32 (L1A11)	22.1	19.45	17.05	270	159	50.5	22	782	-9.9	-11.49
33	20.15	19.2	18.3	321		18.5	41	823	-9.875	
34 (L1A12)	22.25	20.5	18.85	0	105	34	22	845	-10.005	-11.055
35	24.25	21.45	18.7	6		55.5	22	867	-10.1	
36 (L1A13)	24.7	20.85	17.2	8	115	75	20	887	-10.04	-11.19
back on p36	17.9	14.4	10.85	172		70.5		887		
37 (L1A14)	19.2	18.6	18	185	140	12	58.5	945.5	-10.46	-11.86
38	20.5	19.1	18.35	293		21.5	27.5	973	-10.51	
39	23	21.6	20.2	344		28	22	995	-10.76	
40	23.3	21.3	19.35	357		39.5	14	1009	-10.73	
41 (L1A15)	24.05	21.1	18.05	343	115	60	23	1032	-10.71	-11.86
back on p41	20.8	17.45	14.1	192		67		1032		
42 (L1A16)	19.25	18.7	18.15	223	100	11	58	1090	-10.835	-11.835
43 (L1A17)	25.1	22.7	20.3	346	104	48	55	1145	-11.235	-12.275
44 (L1A18)	29.2	24.7	20.4	356	108	88	42	1187	-11.435	-12.515
back on p44	20.2	16.7	13.2	196		70		1187		
45 (L1A19)	19.7	17.9	16.1	204	126	36	34	1221	-11.555	-12.815
46	21.6	20.5	19.4	222		22	16.5	1237.5	-11.815	
47 (L1A20)	19.8	19.3	18.9	220	115	9	13	1250.5	-11.695	-12.845
48	23.15	22.7	22.25	269		9	7	1257.5	-12.035	
49 (L1A21)	21.5	20.9	20.3	333	114	12	11.5	1269	-11.855	-12.995
50 (L1A22)	19.8	19.1	18.4	351	104	14	4.5	1273.5	-11.675	-12.715
51 (L1A23)	19.7	19.05	18.45	17	52	12.5	8	1281.5	-11.67	-12.19
52	22.3	21.2	20.05	354		22.5	9	1290.5	-11.885	
53 (L1A24)	22.3	20.8	19.2	358	20	31	9	1299.5	-11.845	-12.045
54	24.5	22	19.4	352		51	17	1316.5	-11.965	
55 (L1A25)	23.65	21.25	18.85	357	22	48	5.5	1322	-11.89	-12.11
56 (L1A26)	26.45	22.45	19.55	356	22	69	21	1343	-12.01	-12.23
back on p56	20.8	19.5	18.2	171		26		1343		
57 (L1A27)	24.2	23.6	23	224	10	12	21	1364	-12.42	-12.52
58 (L1A28)	23.4	22.6	21.7	301	10	17	19	1383	-12.32	-12.42
59 (L1A29)	24.7	23.4	22.1	323	10	26	12	1395	-12.4	-12.5
60	28.7	26.3	23.9	348		48	27	1422	-12.69	
61	30.2	27.7	25.2	352		50	4	1426	-12.83	
back on p61	24.7	22.5	20.45	225		42.5		1426		
62	35.6	34.4	33.2	252		24	23.5	1449.5	-14.02	
63	39.55	38.6	37.65	266		19	8	1457.5	-14.44	
64	47.35	46.25	45.15	301		22	13	1470.5	-15.205	
back on p64	13.4	12.1	10.75	244		26.5		1470.5		
65	29.35	28.55	27.8	272		15.5	15	1485.5	-16.85	
66	31.65	31.05	30.35	351		13	17.5	1503	-17.1	
67	37.65	36.6	35.6	23		20.5	12	1515	-17.655	
68	45.95	44.55	43.25	34		27	8.5	1523.5	-18.45	
69	48.55	46.85	45.1	41		34.5	8.5	1532	-18.68	
70	48.25	45.6	42.75	47		55	21	1553	-18.555	
71	50.25	46.8	43.45	48		68	13	1566	-18.675	

Channel Cross-sections									
	T	M	B	Dist	Rel. Ht(m)	Dist	Ht	Fdist	FHt
a (CS1)	15.7	15.4	15	7	-1.54	7	0.66	7	13.66
b	17	16.7	16.3	7	-1.67	7	0.53	7	13.53
c	18.6	18.2	17.8	8	-1.82	8	0.38	7.5	13.38
d	17.1	16.7	16.2	9	-1.67	9	0.53	8	13.53
e	16.4	15.9	15.5	9	-1.59	9	0.61	8	13.61
f	14.4	13.9	13.5	9	-1.39	9	0.81	8	13.81
a (CS2)	17.9	17.5	17.2	7	-1.75	16	0.45	7	12.45
b	18.6	18.3	17.9	7	-1.83	16	0.37	7	12.37
c	19.3	18.9	18.5	8	-1.89	17	0.31	7.5	12.31
d	18.9	18.5	18.1	8	-1.85	17	0.35	7.5	12.35
e	18.4	18	17.5	9	-1.8	18	0.4	8	12.4
f	15.9	15.4	15	9	-1.54	18	0.66	8	12.66
a (CS3)	16.7	16.3	16	7	-1.63	25	0.57	7	11.57
b	17.6	17.2	16.8	8	-1.72	26	0.48	7.5	11.48
c	18.6	18.2	17.8	8	-1.82	26	0.38	7.5	11.38
d	17.5	17	16.6	9	-1.7	27	0.5	8	11.5
e	15.8	15.4	14.9	9	-1.54	27	0.66	8	11.66
a (CS4)	19.1	18.7	18.4	7	-1.87	34	0.33	7	10.33
b	21.6	21.2	20.8	8	-2.12	35	0.08	7.5	10.08
c	20.6	20.2	19.7	9	-2.02	36	0.18	8	10.18
d	18.4	17.9	17.5	9	-1.79	36	0.41	8	10.41
a (CS5)	14.4	14	13.6	8	-1.4	44	0.8	8	9.8
b	16.5	16	15.6	9	-1.6	45	0.6	8.5	9.6
c	16.6	16.1	15.6	10	-1.61	46	0.59	9	9.59
d	15.6	15	14.5	11	-1.5	47	0.7	9.5	9.7
e	14.8	14.2	13.7	11	-1.42	47	0.78	9.5	9.78
a (CS6)	15.1	14.7	14.3	8	-1.47	55	0.73	8	8.73
b	17.1	16.6	16.2	9	-1.66	56	0.54	8.5	8.54
c	18	17.5	16.9	11	-1.75	58	0.45	9.5	8.45
d	17.4	16.7	16.1	13	-1.67	60	0.53	10.5	8.53
e	17.6	16.8	16	16	-1.68	63	0.52	12	8.52
f	17.4	16.5	15.6	18	-1.65	65	0.55	13	8.55
g	16.4	15.4	14.3	21	-1.54	68	0.66	14.5	8.66
a (CS7)	16.2	15.9	15.5	7	-1.59	75	0.61	7	7.61
b	17.1	16.7	16.3	8	-1.67	76	0.53	7.5	7.53
c	18.3	17.8	17.3	10	-1.78	78	0.42	8.5	7.42
d	18.5	18	17.4	11	-1.8	79	0.4	9	7.4
e	17.3	16.7	16.1	12	-1.67	80	0.53	9.5	7.53
f	16.3	15.6	14.9	14	-1.56	82	0.64	10.5	7.64
a (CS8)	16.4	15.25	14.15	22.5	-1.525	104.5	0.675	13.5	6.675
b	17	16.1	15.2	18	-1.61	100	0.59	11.25	6.59
c	17.7	17.05	16.3	14	-1.705	96	0.495	9.25	6.495
d	17.4	16.9	16.4	10	-1.69	92	0.51	7.25	6.51
e	17.5	17.1	16.7	8	-1.71	90	0.49	6.25	6.49
f	17.4	17.1	16.8	6	-1.71	88	0.49	5.25	6.49
g	15.7	15.5	15.25	4.5	-1.55	86.5	0.65	4.5	6.65
a (CS9)	15.25	14.9	14.55	7	-1.49	112	0.71	7	5.71
b	17.15	16.7	16.25	9	-1.67	114	0.53	8	5.53
c	21.65	21.15	20.65	10	-2.115	115	0.085	8.5	5.085
d	17.8	17.2	16.6	12	-1.72	117	0.48	9.5	5.48
e	17.7	17	16.3	14	-1.7	119	0.5	10.5	5.5
f	20.3	19.5	18.8	15	-1.95	120	0.25	11	5.25
g	16.1	15.3	14.5	16	-1.53	121	0.67	11.5	5.67
h	15.8	14.9	14	18	-1.49	123	0.71	12.5	5.71
a (CS10)	15.5	15.2	14.9	6	-1.52	129	0.68	6	4.68
b	17.25	16.8	16.4	8.5	-1.68	131.5	0.52	7.25	4.52
c	20.6	20.1	19.6	10	-2.01	133	0.19	8	4.19
d	18.2	17.6	17	12	-1.76	135	0.44	9	4.44
e	17.3	16.6	15.8	15	-1.66	138	0.54	10.5	4.54
f	15.7	14.8	13.9	18	-1.48	141	0.72	12	4.72
a (CS11)	15.6	14.9	14.2	14	-1.49	155	0.71	9.75	3.71
b	16.1	15.5	14.9	12	-1.55	153	0.65	8.75	3.65
c	18.9	18.35	17.83	10.7	-1.835	151.7	0.365	8.1	3.365
d	22.5	22	21.5	10	-2.2	151	0	7.75	3
e	18.8	18.35	17.9	9	-1.835	150	0.365	7.25	3.365
f	17.45	17.1	16.7	7.5	-1.71	148.5	0.49	6.5	3.49
g	14.65	14.35	14.1	5.5	-1.435	146.5	0.765	5.5	3.765
a (CS12)	14.25	14	13.7	5.5	-1.4	157.5	0.8	5.5	2.8
b	16.3	16	15.65	6.5	-1.6	158.5	0.6	6	2.6
c	18.45	18.05	17.7	7.5	-1.805	159.5	0.395	6.5	2.395
d	20.1	19.7	19.35	7.5	-1.97	159.5	0.23	6.5	2.23
e	16.8	16.4	15.9	9	-1.64	161	0.56	7.25	2.56
f	15.55	15	14.45	11	-1.5	163	0.7	8.25	2.7
a (CS13)	15.15	14.85	14.6	5.5	-1.485	168.5	0.715	5.5	1.715
b	15.15	14.8	14.5	6.5	-1.48	169.5	0.72	6	1.72
c	19.1	18.75	18.4	7	-1.875	170	0.325	6.25	1.325
d	20.4	20	19.65	7.5	-2	170.5	0.2	6.5	1.2
e	16.83	16.3	15.83	10	-1.63	173	0.57	7.25	1.57
f	14.25	13.6	13	12.5	-1.36	175.5	0.84	8.5	1.84
a (CS14)	15.1	14.85	14.6	5	-1.485	180	0.715	5	0.715
b	16.75	16.45	16.1	6.5	-1.645	181.5	0.555	5.75	0.555
c	20.3	20	19.65	6.5	-2	181.5	0.2	5.75	0.2
d	21.4	21	20.65	7.5	-2.1	182.5	0.1	6.25	0.1
e	19	18.55	18.1	9	-1.855	184	0.345	7	0.345
f	15	14.5	14	10	-1.45	185	0.75	7.5	0.75

APPENDIX B – PARTICLE SIZE ANALYSIS

Raw Data																
Sample Name	d95 (um)	d84 (um)	d50 (um)	d16 (um)	d5 (um)	d95 (phi)	d84 (phi)	d50 (phi)	d16 (phi)	d5 (phi)	Median (um)	Mean (um)	Median (phi)	Mean (phi)	Sorting (phi)	Skewness
4	2.594	11.785	90.443	714.994	1213.965	8.59	6.41	3.47	0.48	-0.28	90.44	272.41	3.47	1.88	2.82	0.07
4	2.592	11.845	92.247	670.953	1115.53	8.59	6.40	3.44	0.58	-0.16	92.25	258.35	3.44	1.95	2.78	0.10
4	1.959	7.296	52.227	177.278	365.773	9.00	7.10	4.26	2.50	1.45	52.23	78.93	4.26	3.66	2.29	0.24
13	2.545	10.048	83.91	1121.196	1443.324	8.62	6.64	3.58	-0.17	-0.53	83.91	405.05	3.58	1.30	3.09	0.00
13	1.59	5.008	33.204	102.284	203.927	9.30	7.64	4.91	3.29	2.29	33.20	46.83	4.91	4.42	2.15	0.25
13	1.578	5.021	33.445	104.001	212.846	9.31	7.64	4.90	3.27	2.23	33.45	47.49	4.90	4.40	2.17	0.25
20	1.695	5.598	30.163	77.387	128.913	9.20	7.48	5.05	3.69	2.96	30.16	37.72	5.05	4.73	1.89	0.31
20	1.676	5.492	29.55	75.733	127.028	9.22	7.51	5.08	3.72	2.98	29.55	36.93	5.08	4.76	1.89	0.30
20	1.675	5.466	29.225	76.082	129.268	9.22	7.52	5.10	3.72	2.95	29.23	36.92	5.10	4.76	1.90	0.29
31	1.691	4.325	22.345	60.347	112.27	9.21	7.85	5.48	4.05	3.15	22.35	29.01	5.48	5.11	1.87	0.24
31	1.664	4.193	21.382	56.677	95.603	9.23	7.90	5.55	4.14	3.39	21.38	27.42	5.55	5.19	1.82	0.26
31	1.664	4.2	21.441	57.117	98.494	9.23	7.90	5.54	4.13	3.34	21.44	27.59	5.54	5.18	1.83	0.25
23	2.142	8.195	40.455	96.568	137.61	8.87	6.93	4.63	3.37	2.86	40.46	48.41	4.63	4.37	1.80	0.35
23	2.167	8.326	41.523	102.597	153.629	8.85	6.91	4.59	3.28	2.70	41.52	50.82	4.59	4.30	1.84	0.33
23	2.145	8.189	40.589	96.687	136.86	8.86	6.93	4.62	3.37	2.87	40.59	48.49	4.62	4.37	1.80	0.36
8	1.627	4.157	22.747	66.334	114.372	9.26	7.91	5.46	3.91	3.13	22.75	31.08	5.46	5.01	1.93	0.23
8	1.64	4.227	23.257	69.817	129.61	9.25	7.89	5.43	3.84	2.95	23.26	32.43	5.43	4.95	1.97	0.21
8	1.63	4.175	22.838	68.32	123.053	9.26	7.90	5.45	3.87	3.02	22.84	31.78	5.45	4.98	1.95	0.22
16	1.438	3.928	21.023	54.645	87.977	9.44	7.99	5.57	4.19	3.51	21.02	26.53	5.57	5.24	1.85	0.29
16	1.63	4.09	20.991	54.041	85.885	9.26	7.93	5.57	4.21	3.54	20.99	26.37	5.57	5.24	1.80	0.28
16	1.434	3.889	20.648	53.939	86.519	9.45	8.01	5.60	4.21	3.53	20.65	26.16	5.60	5.26	1.84	0.29
7	1.445	3.056	16.484	61.245	174.38	9.43	8.35	5.92	4.03	2.52	16.48	26.93	5.92	5.21	2.13	0.07
7	1.441	3.027	16.101	59.182	170.257	9.44	8.37	5.96	4.08	2.55	16.10	26.10	5.96	5.26	2.12	0.07
7	1.541	3.233	15.956	51.08	101.413	9.34	8.27	5.97	4.29	3.30	15.96	23.42	5.97	5.42	1.91	0.14
9	1.542	3.808	22.229	67.042	133.465	9.34	8.04	5.49	3.90	2.91	22.23	31.03	5.49	5.01	2.01	0.21
9	1.536	3.768	21.833	65.414	127.195	9.35	8.05	5.52	3.93	2.97	21.83	30.34	5.52	5.04	1.99	0.22
9	1.578	3.766	20.833	59.643	109.524	9.31	8.05	5.58	4.07	3.19	20.83	28.08	5.58	5.15	1.92	0.23
17	1.165	3.043	20.146	80.779	434.234	9.75	8.36	5.63	3.63	1.20	20.15	34.66	5.63	4.85	2.48	0.06
17	1.102	2.685	15.999	49.886	91.64	9.83	8.54	5.97	4.33	3.45	16.00	22.86	5.97	5.45	2.02	0.22
17	1.178	3.091	17.848	51.673	101.676	9.73	8.34	5.81	4.27	3.30	17.85	24.20	5.81	5.37	1.99	0.23
6	1.247	2.99	16.148	52.864	97.936	9.65	8.39	5.95	4.24	3.35	16.15	24.00	5.95	5.38	1.99	0.17
6	1.348	3.345	17.632	54.25	93.508	9.53	8.22	5.83	4.20	3.42	17.63	25.08	5.83	5.32	1.93	0.20
6	1.36	3.353	17.531	54.344	92.801	9.52	8.22	5.83	4.20	3.43	17.53	25.08	5.83	5.32	1.93	0.20
25	1.668	6.745	44.773	340.049	536.524	9.23	7.21	4.48	1.56	0.90	44.77	130.52	4.48	2.94	2.68	0.05
25	1.359	5.103	31.352	113.236	321.961	9.52	7.61	5.00	3.14	1.64	31.35	49.90	5.00	4.32	2.31	0.16
25	1.707	7.143	49.197	447.904	650.714	9.19	7.13	4.35	1.16	0.62	49.20	168.08	4.35	2.57	2.79	0.03
5	1.391	2.787	12.033	39.65	80.962	9.49	8.49	6.38	4.66	3.63	12.03	18.16	6.38	5.78	1.85	0.08
5	1.376	2.751	11.809	38.329	73.976	9.51	8.51	6.40	4.71	3.76	11.81	17.63	6.40	5.83	1.82	0.09
5	1.389	2.766	11.9	39.245	80.226	9.49	8.50	6.39	4.67	3.64	11.90	17.97	6.39	5.80	1.84	0.08
3	1.349	2.808	9.482	28.43	59.682	9.53	8.48	6.72	5.14	4.07	9.48	13.57	6.72	6.20	1.66	0.04
3	1.341	2.789	9.37	27.599	54.54	9.54	8.49	6.74	5.18	4.20	9.37	13.25	6.74	6.24	1.64	0.05
3	1.355	2.82	9.522	28.684	65.492	9.53	8.47	6.71	5.12	3.93	9.52	13.68	6.71	6.19	1.68	0.03
2	1.495	3.311	13.833	48.957	113.767	9.39	8.24	6.18	4.35	3.14	13.83	22.03	6.18	5.50	1.92	0.04
2	1.474	3.287	13.778	48.872	116.464	9.41	8.25	6.18	4.35	3.10	13.78	21.98	6.18	5.51	1.93	0.04
2	1.49	3.37	14.358	58.853	143.918	9.39	8.21	6.12	4.09	2.80	14.36	25.53	6.12	5.29	2.03	0.00

27	1.132	2.54	10.904	29.578	49.782	9.79	8.62	6.52	5.08	4.33	10.90	14.34	6.52	6.12	1.71	0.19
27	1.127	2.53	10.908	29.67	50.402	9.79	8.63	6.52	5.07	4.31	10.91	14.37	6.52	6.12	1.72	0.19
27	1.121	2.519	10.942	30.027	51.783	9.80	8.63	6.51	5.06	4.27	10.94	14.50	6.51	6.11	1.73	0.19
11	1.182	2.827	11.496	33.43	72.412	9.72	8.47	6.44	4.90	3.79	11.50	15.92	6.44	5.97	1.79	0.12
11	1.18	2.823	11.485	33.785	75.631	9.73	8.47	6.44	4.89	3.72	11.49	16.03	6.44	5.96	1.80	0.11
11	1.188	2.82	11.434	33.612	71.645	9.72	8.47	6.45	4.89	3.80	11.43	15.96	6.45	5.97	1.79	0.12
18	1.446	3.768	18.139	45.524	85.552	9.43	8.05	5.78	4.46	3.55	18.14	22.48	5.78	5.48	1.79	0.25
18	1.405	3.695	17.907	45.005	84.512	9.48	8.08	5.80	4.47	3.56	17.91	22.20	5.80	5.49	1.80	0.25
18	1.39	3.679	17.896	44.967	83.768	9.49	8.09	5.80	4.47	3.58	17.90	22.18	5.80	5.49	1.80	0.26
15	1.34	2.498	9.325	30.137	59.716	9.54	8.65	6.74	5.05	4.07	9.33	13.99	6.74	6.16	1.73	0.04
15	1.337	2.478	9.183	29.096	55.98	9.55	8.66	6.77	5.10	4.16	9.18	13.59	6.77	6.20	1.70	0.05
15	1.318	2.478	9.186	29.393	57.997	9.57	8.66	6.77	5.09	4.11	9.19	13.69	6.77	6.19	1.72	0.04
24	1.203	2.131	8.852	25.913	43.308	9.70	8.87	6.82	5.27	4.53	8.85	12.30	6.82	6.35	1.68	0.13
24	1.189	2.115	8.802	25.696	42.974	9.72	8.89	6.83	5.28	4.54	8.80	12.20	6.83	6.36	1.68	0.13
24	0.92	1.982	8.528	25.627	43.169	10.09	8.98	6.87	5.29	4.53	8.53	12.05	6.87	6.38	1.76	0.15
21	1.53	3.481	21.027	84.114	142.376	9.35	8.17	5.57	3.57	2.81	21.03	36.21	5.57	4.79	2.14	0.14
21	1.509	3.393	20.074	81.034	137.263	9.37	8.20	5.64	3.63	2.86	20.07	34.83	5.64	4.84	2.13	0.13
21	1.647	3.807	21.587	80.824	131.581	9.25	8.04	5.53	3.63	2.93	21.59	35.41	5.53	4.82	2.06	0.16
32	1.267	2.784	11.22	31.769	63.618	9.62	8.49	6.48	4.98	3.97	11.22	15.26	6.48	6.03	1.73	0.13
32	1.258	2.759	11.069	31.076	60.76	9.63	8.50	6.50	5.01	4.04	11.07	14.97	6.50	6.06	1.72	0.13
32	1.254	2.754	11.067	30.997	60.728	9.64	8.50	6.50	5.01	4.04	11.07	14.94	6.50	6.06	1.72	0.14
14	1.152	2.393	9.637	35.262	69.381	9.76	8.71	6.70	4.83	3.85	9.64	15.76	6.70	5.99	1.87	0.04
14	1.145	2.377	9.535	34.759	68.293	9.77	8.72	6.71	4.85	3.87	9.54	15.56	6.71	6.01	1.86	0.04
14	1.139	2.355	9.363	33.514	65.323	9.78	8.73	6.74	4.90	3.94	9.36	15.08	6.74	6.05	1.84	0.04
1	1.404	2.881	9.717	26.846	55.185	9.48	8.44	6.69	5.22	4.18	9.72	13.15	6.69	6.25	1.61	0.07
1	1.372	2.857	9.621	26.631	55.019	9.51	8.45	6.70	5.23	4.18	9.62	13.04	6.70	6.26	1.61	0.07
1	1.357	2.836	9.557	26.286	54.198	9.53	8.46	6.71	5.25	4.21	9.56	12.89	6.71	6.28	1.61	0.07
19	1.453	2.858	10.34	33.866	84.496	9.43	8.45	6.60	4.88	3.56	10.34	15.69	6.60	5.99	1.78	0.00
19	1.44	2.842	10.247	33.306	78.857	9.44	8.46	6.61	4.91	3.66	10.25	15.47	6.61	6.01	1.76	0.01
19	1.447	2.859	10.391	34.679	87.418	9.43	8.45	6.59	4.85	3.52	10.39	15.98	6.59	5.97	1.80	0.00
22	1.422	2.762	10.387	32.048	59.292	9.46	8.50	6.59	4.96	4.08	10.39	15.07	6.59	6.05	1.70	0.07
22	1.413	2.748	10.327	31.784	59.228	9.47	8.51	6.60	4.98	4.08	10.33	14.95	6.60	6.06	1.70	0.07
22	1.414	2.751	10.333	31.857	59.061	9.47	8.51	6.60	4.97	4.08	10.33	14.98	6.60	6.06	1.70	0.07
26	1.336	2.957	11.481	29.725	49.004	9.55	8.40	6.44	5.07	4.35	11.48	14.72	6.44	6.09	1.62	0.18
26	1.337	2.962	11.551	29.964	50.507	9.55	8.40	6.44	5.06	4.31	11.55	14.83	6.44	6.08	1.63	0.18
26	1.411	2.992	11.625	30.225	51.267	9.47	8.38	6.43	5.05	4.29	11.63	14.95	6.43	6.06	1.62	0.17
28	1.151	2.633	12.707	35.006	58.032	9.76	8.57	6.30	4.84	4.11	12.71	16.78	6.30	5.90	1.79	0.22
28	1.151	2.63	12.694	35.163	58.827	9.76	8.57	6.30	4.83	4.09	12.69	16.83	6.30	5.89	1.80	0.22
28	1.147	2.62	12.652	34.982	58.31	9.77	8.58	6.30	4.84	4.10	12.65	16.75	6.30	5.90	1.79	0.22
12	1.337	2.78	9.815	33.33	69.33	9.55	8.49	6.67	4.91	3.85	9.82	15.31	6.67	6.03	1.76	0.01
12	1.332	2.768	9.741	33.182	70.415	9.55	8.50	6.68	4.91	3.83	9.74	15.23	6.68	6.04	1.76	0.01
12	1.331	2.763	9.684	32.603	68.565	9.55	8.50	6.69	4.94	3.87	9.68	15.02	6.69	6.06	1.75	0.01
30	1.309	2.754	10.826	29.685	49.645	9.58	8.50	6.53	5.07	4.33	10.83	14.42	6.53	6.12	1.65	0.16
30	1.303	2.738	10.768	29.275	49.115	9.58	8.51	6.54	5.09	4.35	10.77	14.26	6.54	6.13	1.65	0.16
30	1.302	2.733	10.722	29.239	49.17	9.59	8.52	6.54	5.10	4.35	10.72	14.23	6.54	6.13	1.65	0.16
29	1.119	3.258	27.274	63.559	95.735	9.80	8.26	5.20	3.98	3.38	27.27	31.36	5.20	4.99	2.04	0.43
29	1.115	3.245	27.187	63.131	94.473	9.81	8.27	5.20	3.99	3.40	27.19	31.19	5.20	5.00	2.04	0.44

29	1.111	3.225	27.095	63.345	94.938	9.81	8.28	5.21	3.98	3.40	27.10	31.22	5.21	5.00	2.05	0.43
10	1.284	2.885	14.225	38.016	56.509	9.61	8.44	6.14	4.72	4.15	14.23	18.38	6.14	5.77	1.76	0.25
10	1.273	2.853	14.073	37.612	55.776	9.62	8.45	6.15	4.73	4.16	14.07	18.18	6.15	5.78	1.76	0.25
10	1.465	2.966	14.362	38.05	56.544	9.41	8.40	6.12	4.72	4.14	14.36	18.46	6.12	5.76	1.72	0.24
4_av	2.593	11.815	91.33	691.292	1166.498	8.59	6.40	3.45	0.53	-0.22	91.33	264.81	3.45	1.92	2.80	0.09
13_av	1.584	5.014	33.324	103.14	208.29	9.30	7.64	4.91	3.28	2.26	33.32	47.16	4.91	4.41	2.16	0.25
20_av	1.682	5.518	29.644	76.407	128.402	9.22	7.50	5.08	3.71	2.96	29.64	37.19	5.08	4.75	1.90	0.30
31_av	1.664	4.196	21.411	56.894	97.003	9.23	7.90	5.55	4.14	3.37	21.41	27.50	5.55	5.18	1.83	0.25
23_av	2.144	8.192	40.521	96.628	137.233	8.87	6.93	4.63	3.37	2.87	40.52	48.45	4.63	4.37	1.80	0.35
8_av	1.632	4.186	22.946	68.113	121.932	9.26	7.90	5.45	3.88	3.04	22.95	31.75	5.45	4.98	1.95	0.22
16_av	1.436	3.908	20.835	54.289	87.244	9.44	8.00	5.58	4.20	3.52	20.84	26.34	5.58	5.25	1.85	0.29
7_av	1.443	3.042	16.291	60.206	172.309	9.44	8.36	5.94	4.05	2.54	16.29	26.51	5.94	5.24	2.12	0.07
9_av	1.539	3.788	22.03	66.215	130.332	9.34	8.04	5.50	3.92	2.94	22.03	30.68	5.50	5.03	2.00	0.21
17_av	1.178	3.091	17.848	51.673	101.676	9.73	8.34	5.81	4.27	3.30	17.85	24.20	5.81	5.37	1.99	0.23
6_av	1.354	3.349	17.581	54.297	93.151	9.53	8.22	5.83	4.20	3.42	17.58	25.08	5.83	5.32	1.93	0.20
25_av	1.668	6.745	44.773	340.049	536.524	9.23	7.21	4.48	1.56	0.90	44.77	130.52	4.48	2.94	2.68	0.05
5_av	1.385	2.768	11.913	39.064	78.214	9.50	8.50	6.39	4.68	3.68	11.91	17.92	6.39	5.80	1.84	0.08
3_av	1.349	2.808	9.482	28.43	59.682	9.53	8.48	6.72	5.14	4.07	9.48	13.57	6.72	6.20	1.66	0.04
2_av	1.484	3.299	13.806	48.915	115.111	9.40	8.24	6.18	4.35	3.12	13.81	22.01	6.18	5.51	1.92	0.04
27_av	1.126	2.53	10.918	29.757	50.632	9.79	8.63	6.52	5.07	4.30	10.92	14.40	6.52	6.12	1.72	0.19
11_av	1.183	2.823	11.472	33.606	73.151	9.72	8.47	6.45	4.90	3.77	11.47	15.97	6.45	5.97	1.79	0.12
18_av	1.414	3.714	17.98	45.165	84.608	9.47	8.07	5.80	4.47	3.56	17.98	22.29	5.80	5.49	1.80	0.25
15_av	1.332	2.485	9.231	29.534	57.875	9.55	8.65	6.76	5.08	4.11	9.23	13.75	6.76	6.18	1.72	0.04
24_av	1.196	2.123	8.827	25.805	43.14	9.71	8.88	6.82	5.28	4.53	8.83	12.25	6.82	6.35	1.68	0.13
21_av	1.586	3.639	21.311	82.41	136.844	9.30	8.10	5.55	3.60	2.87	21.31	35.79	5.55	4.80	2.10	0.15
32_av	1.26	2.766	11.118	31.278	61.676	9.63	8.50	6.49	5.00	4.02	11.12	15.05	6.49	6.05	1.73	0.13
14_av	1.146	2.375	9.511	34.503	67.678	9.77	8.72	6.72	4.86	3.89	9.51	15.46	6.72	6.02	1.86	0.04
1_av	1.378	2.858	9.631	26.587	54.807	9.50	8.45	6.70	5.23	4.19	9.63	13.03	6.70	6.26	1.61	0.07
19_av	1.447	2.853	10.326	33.94	83.52	9.43	8.45	6.60	4.88	3.58	10.33	15.71	6.60	5.99	1.78	0.00
22_av	1.416	2.754	10.349	31.896	59.194	9.46	8.50	6.59	4.97	4.08	10.35	15.00	6.59	6.06	1.70	0.07
26_av	1.361	2.97	11.552	29.97	50.224	9.52	8.40	6.44	5.06	4.32	11.55	14.83	6.44	6.08	1.62	0.18
28_av	1.149	2.627	12.684	35.05	58.383	9.77	8.57	6.30	4.83	4.10	12.68	16.79	6.30	5.90	1.79	0.22
12_av	1.333	2.771	9.746	33.037	69.428	9.55	8.50	6.68	4.92	3.85	9.75	15.18	6.68	6.04	1.76	0.01
30_av	1.302	2.736	10.745	29.257	49.143	9.59	8.51	6.54	5.10	4.35	10.75	14.25	6.54	6.13	1.65	0.16
29_av	1.115	3.243	27.185	63.344	95.048	9.81	8.27	5.20	3.98	3.40	27.19	31.26	5.20	5.00	2.04	0.43
10_av	1.279	2.869	14.149	37.814	56.145	9.61	8.45	6.14	4.72	4.15	14.15	18.28	6.14	5.77	1.76	0.25

Mean Particle Size Classification													
Sample	Mean (um)	Class um	Mean (phi)	Class phi	Sorting (phi)	Skewness	Clay %	Silt %	Sand	Clay+Silt %	Sample	Depth	LOI
4_av	264.81	Med sand	1.92	Med sand	2.80	0.09	3.753	37.406	58.84	41.16	T2A1	200-208	7.36
13_av	47.16	Coarse silt	4.41	Coarse silt	2.16	0.25	6.599	63.336	30.065	69.935	T3A3	140-150	9.74
20_av	37.19	Coarse silt	4.75	Coarse silt	1.90	0.30	6.141	71.766	22.093	77.907	T3A4	175-185	7.39
31_av	27.50	Med silt	5.18	Med silt	1.83	0.25	6.885	79.994	13.121	86.879	T4RB3	137-147	7.41
23_av	48.45	Coarse silt	4.37	Coarse silt	1.80	0.35	4.578	62.234	33.189	66.811	T4A1	60-70	19.96
8_av	31.75	Med silt	4.98	Coarse silt	1.95	0.22	7.154	74.744	18.102	81.898	T2A3	125-135	5.38
16_av	26.34	Med silt	5.25	Med silt	1.85	0.29	7.936	80.444	11.62	88.38	T3A3	215-225	10.25
7_av	26.51	Med silt	5.24	Med silt	2.12	0.07	9.504	75.395	15.101	84.899	T2A3	115-125	4.55
9_av	30.68	Med silt	5.03	Med silt	2.00	0.21	7.991	74.7	17.309	82.691	T2A3	150-160	7.12
17_av	24.20	Med silt	5.37	Med silt	1.99	0.23	10.111	78.626	11.263	88.737	T3A4	60-70	4.71
6_av	25.08	Med silt	5.32	Med silt	1.93	0.20	8.901	78.824	12.275	87.725	T2A3	80-70	7.55
25_av	130.52	Fine sand	2.94	Fine sand	2.68	0.05	6.002	52.768	41.231	58.769	T4A1	115-125	8.61
5_av	17.92	Med silt	5.80	Med silt	1.84	0.08	10.372	81.923	7.705	92.295	T3A3	50-60	15.88
3_av	13.57	Fine silt	6.20	Fine silt	1.66	0.04	9.711	85.705	4.585	95.415	T2A1	150-160	18.49
2_av	22.01	Med silt	5.51	Med silt	1.92	0.04	8.499	79.4	12.101	87.899	T2A1	140-150	16.83
27_av	14.40	Fine silt	6.12	Fine silt	1.72	0.19	11.912	85.202	2.886	97.114	T4A2	45-55	19.86
11_av	15.97	Fine silt	5.97	Med silt	1.79	0.12	10.538	83.132	6.331	93.669	T3A3	97-107	20.48
18_av	22.29	Med silt	5.49	Med silt	1.80	0.25	8.16	82.967	8.873	91.127	T3A4	65-75	23.7
15_av	13.75	Fine silt	6.18	Fine silt	1.72	0.04	11.666	84.178	4.155	95.845	T3A3	200-210	11.79
24_av	12.25	Fine silt	6.35	Fine silt	1.68	0.13	14.694	84.027	1.28	98.72	T4A1	98-108	14.04
21_av	35.79	Coarse silt	4.80	Coarse silt	2.10	0.15	7.838	68.601	23.56	76.44	T3A4	190-200	10.16
32_av	15.05	Fine silt	6.05	Fine silt	1.73	0.13	10.36	84.819	4.821	95.179	T4RB3	175-185	11.36
14_av	15.46	fine silt	6.02	Fine silt	1.86	0.04	12.823	81.367	5.811	94.189	T3A3	190-200	10.37
1_av	13.03	Fine silt	6.26	Fine silt	1.61	0.07	9.573	86.478	3.949	96.051	T2A1	125-135	16.96
19_av	15.71	Fine silt	5.99	Med silt	1.78	0.00	9.672	82.995	7.334	92.666	T3A4	165-175	23.52
22_av	15.00	Fine silt	6.06	Fine silt	1.70	0.07	10.182	85.518	4.3	95.7	T3A4	205-215	12.35
26_av	14.83	Fine silt	6.08	Fine silt	1.62	0.18	9.568	87.655	2.777	97.223	T4A1	130-140	18.05
28_av	16.79	Med silt	5.90	Med silt	1.79	0.22	11.771	84.202	4.027	95.973	T4A2	55-65	17.64
12_av	15.18	Fine silt	6.04	Fine silt	1.76	0.01	10.133	83.82	6.046	93.954	T3A3	130-140	12.87
30_av	14.25	Fine silt	6.13	Fine silt	1.65	0.16	10.53	86.899	2.57	97.43	T4RB3	117-127	21.7
29_av	31.26	Med silt	5.00	Med silt	2.04	0.43	10.753	73.053	16.194	83.806	T4A2	95-105	11.33
10_av	18.28	Med silt	5.77	Med silt	1.76	0.25	10.318	86.69	2.992	97.008	T2A3	180-190	13.14

APPENDIX C – LOSS-ON-IGNITION

T2A1				
Depth	W1	W2	W1-W2	%LOI
0	18.52	10.77	7.75	41.87
45	19.17	14.65	4.52	23.59
90	39.39	30.92	8.47	21.50
125	39.93	33.16	6.77	16.96
140	39.40	32.77	6.63	16.83
150	40.60	33.09	7.51	18.49
200	20.51	19.00	1.51	7.36
208	34.89	33.35	1.54	4.41

T2A2				
Depth	W1	W2	W1-W2	%LOI
0	22.06	14.23	7.83	35.49
40	40.73	31.96	8.77	21.54
55	21.22	17.36	3.86	18.18
70	40.04	33.47	6.57	16.42
82	41.38	35.79	5.59	13.52
92	40.98	38.34	2.64	6.44
100	42.09	40.27	1.83	4.34
110	41.37	39.59	1.77	4.28
125	41.20	37.63	3.57	8.67
147	41.11	38.38	2.73	6.65
160	41.12	37.43	3.68	8.96
170	40.11	38.95	1.16	2.90

T2A3				
Depth	W1	W2	W1-W2	%LOI
0	19.73	9.38	10.35	52.48
30	21.04	17.12	3.92	18.62
50	20.56	17.30	3.27	15.88
80	23.31	21.55	1.76	7.55
115	45.03	42.98	2.05	4.55
125	42.41	40.13	2.28	5.38
150	41.96	38.97	2.99	7.12
180	40.48	35.16	5.32	13.14
198	40.96	40.46	0.50	1.23
260	40.97	40.75	0.22	0.53

T2LB2				
Depth	W1	W2	W1-W2	%LOI
0	40.03	30.70	9.33	23.31
20	39.81	32.14	7.67	19.27
30	39.96	32.84	7.12	17.82
50	38.91	31.83	7.09	18.21
70	39.31	31.99	7.32	18.62
80	39.90	33.51	6.39	16.02
100	40.09	37.33	2.76	6.88
110	41.02	38.96	2.06	5.01
120	40.34	37.22	3.12	7.74
130	40.78	38.55	2.23	5.47

T3A2				
Depth	W1	W2	W1-W2	%LOI
0	19.92	10.65	9.27	46.54
30	38.39	25.16	13.22	34.45
50	40.53	29.45	11.08	27.35
80	40.35	33.36	6.98	17.31
100	40.87	35.63	5.24	12.83
150	40.35	34.52	5.83	14.45
170	41.84	38.69	3.15	7.52
180	42.12	40.60	1.52	3.60

T3A3				
Depth	W1	W2	W1-W2	%LOI
0	19.00	10.99	8.01	42.16
30	20.89	14.86	6.04	28.90
50	20.58	12.02	8.57	41.61
60	40.25	27.76	12.48	31.02
97	41.37	32.89	8.47	20.48
125	15.56	12.15	3.41	21.92
130	40.21	35.04	5.18	12.87
140	41.62	37.56	4.05	9.74
155	41.46	39.33	2.14	5.15
190	41.31	37.03	4.28	10.37
200	39.64	34.96	4.67	11.79
215	20.70	18.58	2.12	10.25

T3A4				
Depth	W1	W2	W1-W2	%LOI
0	18.73	8.19	10.54	56.27
30	20.21	12.66	7.55	37.36
50	19.86	15.04	4.82	24.27
60	41.80	39.83	1.97	4.71
65	20.00	15.26	4.74	23.70
80	19.41	14.35	5.06	26.07
113	41.10	28.60	12.50	30.41
138	19.98	14.73	5.25	26.28
144	20.76	11.67	9.09	43.79
165	21.34	16.32	5.02	23.52
175	40.72	37.71	3.01	7.39
190	42.02	37.75	4.27	10.16
205	41.85	36.68	5.17	12.35

T3A5				
Depth	W1	W2	W1-W2	%LOI
0	40.82	27.45	13.37	32.75
25				18.00
45	42.44	37.49	4.95	11.66
50	39.78	35.27	4.51	11.34
55	40.89	38.00	2.89	7.07
65	40.92	38.90	2.02	4.94
120	41.55	39.41	2.14	5.15
130	40.73	35.30	5.43	13.33

T4A1				
Depth	W1	W2	W1-W2	%LOI
0	20.31	8.90	11.41	56.18
30	39.18	22.13	17.06	43.53
60	39.89	31.92	7.96	19.96
98	40.37	34.70	5.67	14.04
115	41.55	37.97	3.58	8.61
130	39.80	32.62	7.18	18.05

T4A2				
Depth	W1	W2	W1-W2	%LOI
0	20.74	12.87	7.868	37.94
45	40.88	32.76	8.118	19.86
55	40.28	33.18	7.108	17.64
95	40.46	35.88	4.583	11.33

T4A3				
Depth	W1	W2	W1-W2	%LOI
0	19.93	8.397	11.529	57.86
45	38.98	25.73	13.249	33.99
55	39.53	29.14	10.386	26.28
85	40.69	34.04	6.65	16.34
115	40.19	38.12	2.072	5.155

T4LB2				
Depth	W1	W2	W1-W2	%LOI
0	16.89	4.857	12.03	71.24
20	17.79	6.334	11.451	64.39
50	20.06	8.317	11.746	58.55
120	41.54	36.98	4.56	10.98
125	40.4	34.11	6.284	15.56

T4LB3				
Depth	W1	W2	W1-W2	%LOI
0	40.23	22.95	17.278	42.95
36	42.04	30.56	11.48	27.31
96	40.51	30.48	10.028	24.76
143	42.07	35.39	6.68	15.88

T4RB3				
Depth	W1	W2	W1-W2	%LOI
0	21.85	12.85	8.999	41.18
50	40.11	26.13	13.98	34.86
107	40.19	31.13	9.053	22.53
117	39.84	31.19	8.645	21.7
137	39.75	36.8	2.947	7.414
165				
175	22.38	19.84	2.543	11.36

APPENDIX D – X-RAY FLUORESCENCE

XRF data

Dolerite

	D2	D4	D9	D10	D11	D13	D16	D18	Average	Average	Max	Min	SD	COV	Mineral	Wt%
SiO2	51.37	51.81	51.26	50.30	50.75	51.37	51.26	50.50	51.08	51.08	51.81	50.30	0.510	0.998	Ap =	0.54
TiO2	1.06	1.24	1.04	1.25	1.07	1.27	1.14	1.04	1.14	1.14	1.27	1.04	0.100	8.800	Cr =	0.08
Al2O3	14.38	13.33	14.53	14.06	15.51	12.93	13.34	14.54	14.08	14.08	15.51	12.93	0.844	5.993	Ilm =	2.75
Fe2O3	1.30	1.60	1.31	1.65	1.32	1.63	1.54	1.31	1.46	1.46	1.65	1.30	0.160	10.982	Or =	4.94
FeO	12.65	11.73	12.79	12.37	13.65	11.38	11.74	12.80	12.39	12.39	13.65	11.38	0.742	5.993	Ab =	28.13
MnO	0.17	0.21	0.17	0.21	0.19	0.22	0.21	0.18	0.20	0.20	0.22	0.17	0.020	10.256	An =	31.36
MgO	6.72	6.45	6.64	6.24	6.42	6.39	6.77	7.39	6.63	6.63	7.39	6.24	0.357	5.385	Mt =	0.32
CaO	11.49	10.38	11.48	10.20	11.50	10.33	10.78	11.12	10.91	10.91	11.50	10.20	0.560	5.129	Wo =	14.90
Na2O	2.62	2.55	2.69	2.66	2.60	2.68	2.49	2.67	2.62	2.62	2.69	2.49	0.071	2.691	En =	8.03
K2O	0.63	0.69	0.62	0.72	0.65	0.70	0.63	0.63	0.66	0.66	0.72	0.62	0.039	5.876	Fs =	0.86
P2O5	0.21	0.19	0.19	0.19	0.21	0.21	0.16	0.20	0.20	0.20	0.21	0.16	0.017	8.668	Fa =	0.00
Cr2O3	0.04	0.03	0.04	0.03	0.05	0.04	0.05	0.07	0.04	0.04	0.07	0.03	0.013	29.771	Fo =	0.00
NiO	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.02	0.01	0.004	31.427	Qz =	8.10
LOI	0.51	0.05	0.38	0.32	0.35	0.19	0.33	0.61	0.34	0.34	0.61	0.05	0.1734	50.637		
Total	103.16	100.27	103.15	100.21	104.28	99.35	100.45	103.07								

Calculated density = 3.20 g/ml

Clays	T2A3				T3A4				T4A3						
	198-208	295-305	400-410	495-505	215-225	235-245	300-310	420-430	125-135	260-270	420-430	530-540	635-645	730-740	
SiO2	62.18	60.12	59.30	61.62	45.26	31.22	58.29	59.01	44.22	44.19	44.52	35.28	36.14	34.90	
TiO2	0.77	0.71	0.77	0.62	2.16	1.67	0.67	0.62	1.86	2.19	2.24	1.40	1.56	1.42	
Al2O3	22.41	16.57	18.95	19.32	25.43	20.18	15.35	15.50	30.03	28.92	27.85	20.61	21.86	19.66	
Fe2O3	1.61	11.64	7.77	6.23	5.88	24.49	13.61	11.85	4.33	5.69	5.68	22.00	19.95	23.43	
MnO	0.02	0.03	0.06	0.16	0.05	0.06	0.03	0.08	0.06	0.07	0.06	0.06	0.07	0.07	
MgO	0.67	0.56	0.78	0.64	0.48	0.56	1.13	0.95	0.73	0.88	1.07	1.19	1.25	1.48	
CaO	0.00	0.00	0.02	0.03	0.02	0.07	0.05	0.07	0.00	0.01	0.06	0.15	0.15	0.20	
Na2O	0.00	0.06	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.02	0.00	0.00	
K2O	2.65	2.49	3.25	2.35	0.25	0.15	2.44	2.23	0.07	0.08	0.09	0.36	0.37	0.40	
P2O5	0.05	0.13	0.12	0.16	0.10	0.49	0.16	0.15	0.02	0.05	0.05	0.38	0.33	0.31	
Cr2O3	0.00	0.01	0.01	0.01	0.02	0.02	0.01	0.01	0.05	0.05	0.06	0.04	0.05	0.03	
NiO	0.00	0.00	0.00	0.00	0.02	0.02	0.00	0.00	0.01	0.01	0.02	0.01	0.02	0.01	
LOI	9.44	8.00	8.75	8.84	20.82	20.39	8.53	9.37	18.91	18.12	18.35	18.69	18.75	18.41	
Total	99.80	100.32	99.78	99.98	100.49	99.32	100.27	99.84	100.29	100.26	100.06	100.19	100.50	100.32	
Si/(Si+Al)	0.82	0.86	0.84	0.84	0.75	0.72	0.87	0.87	0.71	0.72	0.73	0.74	0.74	0.75	
Al/Al dol*	1.59	1.18	1.35	1.37	1.81	1.43	1.09	1.10	2.13	2.05	1.98	1.46	1.55	1.40	
Mols															
Or	0.0563	0.0529	0.0690	0.0499	0.0053	0.0032	0.0518	0.0473	0.0015	0.0017	0.0019	0.0076	0.0079	0.0085	
Kl	0.0537	0.0813	0.0929	0.0948	0.1247	0.0990	0.0753	0.0760	0.1473	0.1418	0.1366	0.1011	0.1072	0.0964	
Qz	0.9811	0.9192	0.8939	0.9307	0.6285	0.4206	0.8948	0.9060	0.5886	0.5935	0.6043	0.4860	0.4942	0.4844	
Fe OxHy	0.0202	0.0729	0.0487	0.0390	0.0368	0.1534	0.0852	0.0742	0.0271	0.0356	0.0356	0.1378	0.1249	0.1467	
Wt															
Or	18.53	17.41	22.72	16.43	1.75	1.05	17.06	15.59	0.49	0.56	0.63	2.52	2.59	2.80	
Kl	27.70	41.96	47.98	48.92	64.39	51.10	38.87	39.25	76.04	73.23	70.52	52.19	55.35	49.78	
Qz	58.96	55.24	53.71	55.93	37.76	25.27	53.77	54.44	35.37	35.67	36.31	29.21	29.70	29.11	
Fe Ox	3.58	12.95	8.65	6.93	6.54	27.25	15.14	13.19	4.82	6.33	6.32	24.48	22.20	26.07	
sum	108.76	127.55	133.07	128.21	110.45	104.67	124.84	122.47	116.71	115.78	113.78	108.39	109.83	107.75	
Wt%															
Or	17.03	13.65	17.08	12.82	1.58	1.00	13.67	12.73	0.42	0.48	0.55	2.32	2.36	2.60	
Kl	25.47	32.89	36.06	38.16	58.30	48.82	31.13	32.05	65.15	63.24	61.98	48.15	50.40	46.20	
Qz	54.20	43.30	40.37	43.62	34.19	24.15	43.07	44.45	30.30	30.80	31.91	26.95	27.04	27.01	
Fe Ox	3.29	10.15	6.50	5.41	5.92	26.03	12.13	10.77	4.13	5.47	5.55	22.59	20.21	24.19	
density	2.54	2.51	2.50	2.49	2.42	2.47	2.52	2.52	2.39	2.40	2.40	2.47	2.46	2.48	2.47
% vol loss	2.55	2.67	2.76	2.78	3.15	2.90	2.63	2.65	3.30	3.26	3.23	2.90	2.95	2.86	2.90
% th loss	7.47	7.80	8.04	8.12	9.16	8.45	7.68	7.73	9.58	9.45	9.38	8.45	8.60	8.33	8.45
Average	7.86				8.26				8.96						

APPENDIX E – X-RAY DIFFRACTION

Description:
D4 RYAN

Original scan: d4ryan Date: 12/10/2007 09:55
Description of scan:
D4 RYAN

Used wavelength:	K-Alpha
K-Alpha1 wavelength (Å):	1.78897
K-Alpha2 wavelength (Å):	1.79285
K-Alpha2/K-Alpha1 intensity ratio :	0.50000
K-Alpha wavelength (Å):	1.78897
K-Beta wavelength (Å):	

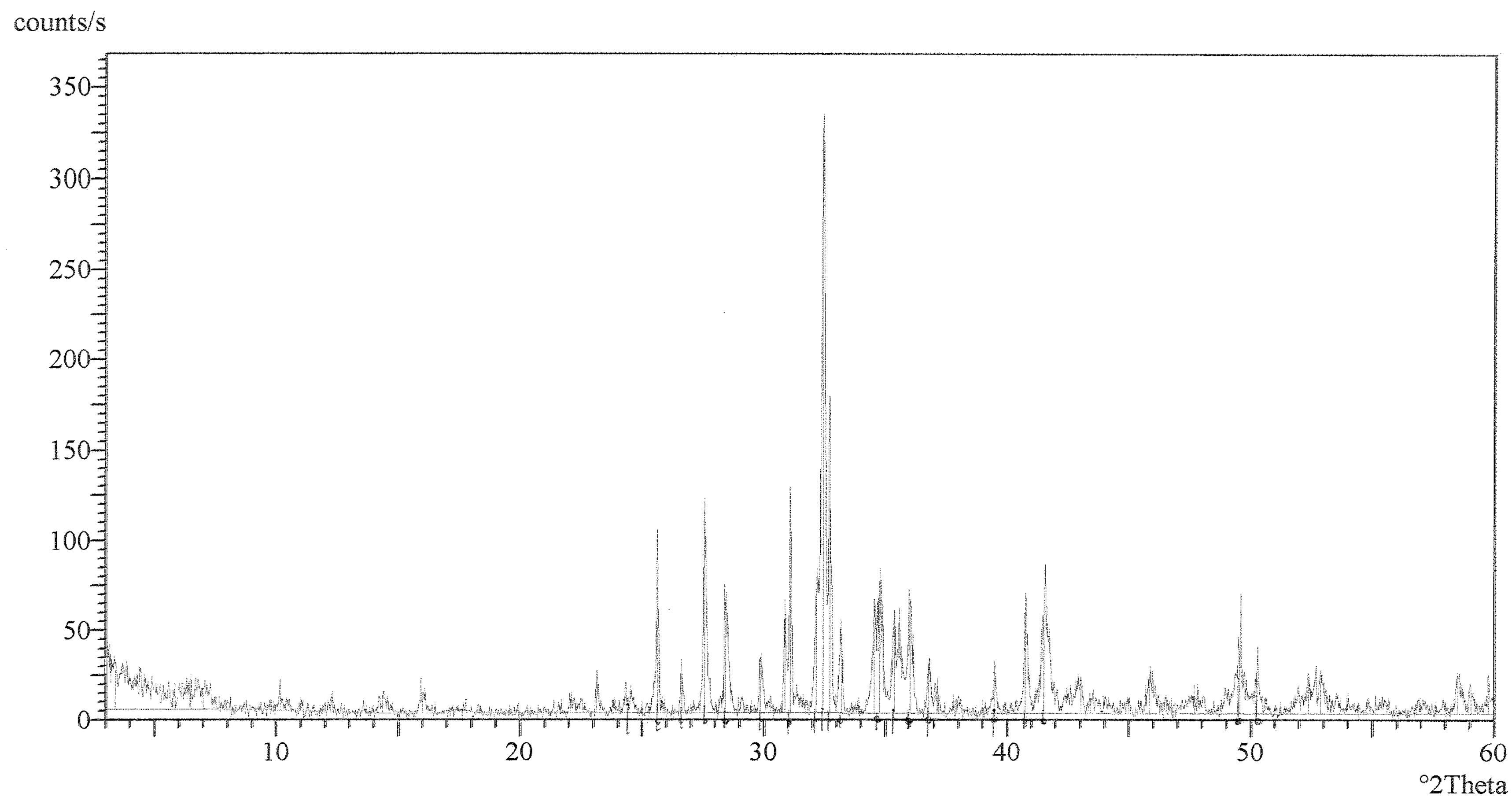
Peak search parameter set:	**As Measured Intensities**
Set created:	04/28/1999 12:06
Peak positions defined by:	Minimum of 2nd derivative
Minimum peak tip width (°2Theta):	0.00
Minimum peak tip width (°2Theta):	1.00
Peak base width (°2Theta):	2.00
Minimum significance:	0.60

d-spacing (Å)	Relative Intensity (%)	Angle (°2Theta)	Peak Height (counts/s)	Background (counts/s)	Tip Width (°2Theta)	Significance
30.62502	6.69	3.34742	22.17	5.39	0.60000	1.03
17.82459	2.04	5.75291	6.77	5.39	0.12500	0.72
14.72531	3.15	6.96510	10.45	5.39	1.00000	1.78
10.12940	2.43	10.13227	8.05	4.89	0.15000	0.77
9.36957	0.90	10.95638	2.98	4.46	0.10000	0.76
8.43064	1.57	12.18098	5.20	3.88	0.30000	0.92
7.17898	2.16	14.31503	7.15	3.33	0.40000	0.97
6.44956	2.35	15.94398	7.80	3.27	0.30000	1.81
4.68592	2.46	22.00912	8.15	3.52	0.15000	0.71
4.47104	7.24	23.08110	24.00	3.50	0.10000	0.97
4.22804	1.86	24.42754	6.17	3.48	0.40000	1.12
4.03838	22.32	25.59377	73.95	3.46	0.10000	2.02
3.88930	4.70	26.59241	15.58	3.45	0.15000	1.97
3.75949	35.72	27.52838	118.35	3.43	0.12500	3.59
3.65000	17.70	28.37119	58.66	3.42	0.10000	1.48
3.47802	9.44	29.80580	31.29	3.40	0.12500	1.49
3.36724	14.72	30.81031	48.77	3.38	0.07500	3.11
3.34227	36.41	31.04627	120.63	3.38	0.10000	2.46
3.23321	20.93	32.12125	69.35	3.36	0.10000	0.65
3.20667	100.00	32.39438	331.33	3.36	0.12500	4.44
3.18148	52.70	32.65810	174.61	3.35	0.10000	2.06
3.13618	11.43	33.14330	37.88	3.34	0.15000	2.50
3.01767	15.46	34.48476	51.23	3.32	0.15000	1.11

d-spacing (Å)	Relative Intensity (%)	Angle (°2Theta)	Peak Height (counts/s)	Background (counts/s)	Tip Width (°2Theta)	Significance
2.99722	20.45	34.72751	67.76	3.32	0.07500	0.92
2.94885	16.83	35.31581	55.76	3.31	0.07500	1.21
2.89685	17.94	35.97110	59.46	3.30	0.12500	0.90
2.83823	7.54	36.74035	24.98	3.29	0.15000	0.83
2.81419	3.24	37.06563	10.75	3.28	0.20000	0.64
2.75576	1.41	37.88105	4.69	3.27	0.30000	1.32
2.65133	8.97	39.43361	29.72	3.26	0.07500	1.09
2.57228	20.08	40.69816	66.54	3.26	0.10000	1.22
2.52639	17.95	41.47118	59.47	3.26	0.30000	3.65
2.44128	5.15	42.98733	17.08	3.26	0.25000	1.27
2.41995	2.11	43.38540	7.00	3.26	0.25000	0.73
2.33964	2.09	44.95447	6.91	3.26	0.15000	0.61
2.29905	6.73	45.79280	22.30	3.26	0.25000	1.33
2.26605	2.67	46.49842	8.86	3.26	0.15000	0.84
2.21418	1.93	47.65428	6.40	3.26	0.60000	0.61
2.13652	9.72	49.50037	32.19	3.26	0.25000	1.62
2.10797	7.80	50.21689	25.83	3.26	0.15000	2.32
2.02846	6.35	52.33102	21.04	3.26	0.07500	0.60
2.01090	7.13	52.82340	23.61	3.26	0.12500	0.84
1.91978	1.34	55.54073	4.43	3.26	0.50000	2.91
1.87511	1.30	56.98325	4.31	3.26	0.50000	0.90
1.83217	6.09	58.44594	20.18	3.26	0.20000	0.82
1.81637	3.61	59.00444	11.95	3.26	0.15000	1.16
1.79671	5.72	59.71497	18.95	3.26	0.07500	1.05

X'Pert Graphics & Identify
Graph: d4ryan

User-1
12/11/2007 11:49



Pyrox.
Plag

- Typical dolerite Φ.

Description:
~~T2A4~~/215-225 RYAN
T3A4

Original scan: t2a42152 Date: 10/18/2007 12:33
Description of scan:
~~T2A4~~/215-225 RYAN
T3A4.

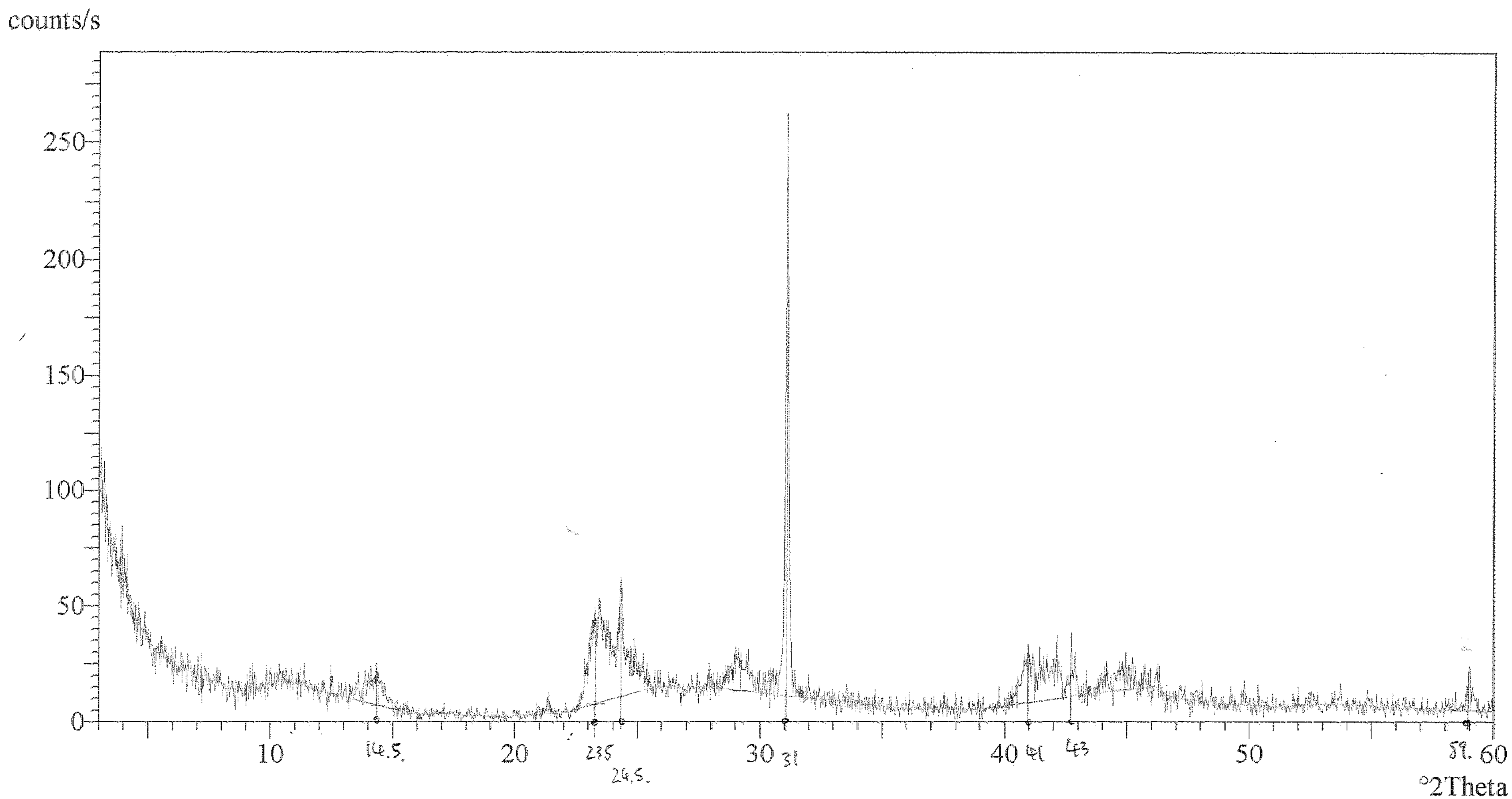
Used wavelength:	K-Alpha
K-Alpha1 wavelength (Å):	1.78897
K-Alpha2 wavelength (Å):	1.79285
K-Alpha2/K-Alpha1 intensity ratio :	0.50000
K-Alpha wavelength (Å):	1.78897
K-Beta wavelength (Å):	

Peak search parameter set:	**As Measured Intensities**
Set created:	04/28/1999 12:06
Peak positions defined by:	Minimum of 2nd derivative
Minimum peak tip width (°2Theta):	0.00
Minimum peak tip width (°2Theta):	1.00
Peak base width (°2Theta):	2.00
Minimum significance:	0.60

d-spacing (Å)	Relative Intensity (%)	Angle (°2Theta)	Peak Height (counts/s)	Background (counts/s)	Tip Width (°2Theta)	Significance
8.23037	2.53	12.47854	6.10	11.55	0.15000	0.73
7.14520	5.18	14.38306	12.50	6.74	0.60000	1.39
4.83137	2.29	21.33867	5.52	2.99	0.20000	0.65
4.43421	14.20	23.27546	34.24	7.33	0.80000	5.58
4.24885	21.12	24.30605	50.94	10.49	0.07500	0.92
3.55131	5.17	29.17683	12.46	13.05	0.60000	0.70
3.34024	100.00	31.06562	241.18	10.89	0.10000	2.71
2.55949	8.61	40.91059	20.78	7.95	0.35000	1.54
2.44690	5.42	42.88386	13.06	10.71	0.30000	1.39
2.34327	3.20	44.88104	7.71	13.50	0.80000	0.80
2.12717	2.64	49.73287	6.37	7.15	0.20000	0.68
2.10290	2.74	50.34658	6.61	6.84	0.07500	0.67
1.81853	6.47	58.92711	15.60	4.67	0.15000	1.20

X'Pert Graphics & Identify
Graph: ~~t2a42~~152
t3a4

User-1
10/19/2007 08:40



Typical residual clay.