The origin and dynamics of Wakkerstroom Vlei, Mpumalanga Province, South Africa

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As the candidates Supervisor I agree/do not agree to the submission of this dissertation.

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Abstract

The formation and common occurrence of riparian wetlands within the semi-arid Highveld interior of South Africa, a landscape setting undergoing extensive long-term fluvial incision, is an enigma and the underlying controls on the formation and hydro-geomorphological dynamics of these wetlands has not been widely investigated. Wakkerstroom Vlei is one such enigma in that it is a large (~ 1000 ha) Highveld system comprising extensive reaches of unchanneled valley-bottom wetland with considerable (up to 2 m deep) peat deposits. Accommodation space for wetland formation is thought to be controlled by the superimposition of the main (Wakkerstroom/Thaka) river upon an erosion-resistant Karoo dolerite sill at the toe of the system, which forms a stable local base-level along the rivers course. As a result, the river has carved broad (up to 1 300 m), gently sloping (average slope ~ 0.17 %) valleys along softer shale valley reaches upstream of the dolerite barrier.

Examination of the valley fill along these valley-bottom wetland reaches, together with analysis of historic aerial photography, reveals that continuous tracts of meandering river and floodplain wetlands formerly existed, and that the wetland experienced an abrupt shift to valley-bottom wetland conditions where surface flow of water is diffusive. Following the creation of accommodation space along the main river valley, lateral tributary streams began to deposit substantial amounts of course sediment into the main valley via alluvial fans. Several of these fans have coalesced to form multiple coalescing alluvial fan complexes that historically were able to extend far across the floodplain from either side of the valley, resulting in main river valley impoundment. This has promoted flood-out formation, along the main valley which, together with the denser growth of vegetation across the floodplain, has created conditions suitable for organic sedimentation and peat accumulation. The formation and evolution of Wakkerstroom Vlei has thus been controlled by the complex interaction between geological, geomorphological and biotic processes. Understanding the role of these factors in shaping both the short- and long-term hydro-geomorphic dynamics of the system is essential in implementing effective management and conservation strategies

both within Wakkerstroom Vlei and other large valley-bottom wetlands within the South African Highveld interior.

Preface

The work described in this dissertation was carried out in the School of Environmental Science, Department of Geography, University of KwaZulu-Natal, Durban, from January 2006 to March 2009, under the supervision of Dr. Serban Proches (supervisor), University of KwaZulu-Natal, Durban, and Professor William Ellery (co-supervisor), Rhodes University, Grahamstown.

This study represents original work by the author and has not been otherwise submitted in any form for any degree or diploma to any tertiary institution. Where use has been made of the work of other authors it is duly acknowledged in the text.

Plagiarism Declaration

I, **Rebecca Joubert** declare that,

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Chapter 1

Introduction

1.1. Background

South Africa hosts a diverse array of wetland types (Ewart-Smith *et al.*, 2006); however, many of these systems have already been lost and are currently at risk of being lost from the South African landscape at a relatively rapid rate (Begg, 1986; Kotze *et al.*, 1995). Despite this, there have been relatively few attempts to develop systematic knowledge of the fundamental controlling factors on the formation and dynamics of southern African wetlands (McCarthy and Hancox, 2000; Ellery *et al.*, 2004), in view of effectively managing and conserving these systems. Most wetland studies in southern Africa have focused on short-term processes and dynamics and there is great need to study the longer-term geomorphological processes that form underlying controls on the development and long-term existence of wetlands in the landscape (Ellery *et al.*, 2004; Tooth and McCarthy, 2007). This situation is mirrored internationally, and knowledge of wetland formation is limited (Rogers, 1995; Tooth, 2000). Most wetlands in the Northern Hemisphere have formed within depressions created by glaciation during the last Ice Age that are not linked to the drainage network (Mitsch and Gosselink, 2000). These wetlands are maintained hydrologically by rainfall alone, due to high rainfall and low evaporation rates. In contrast to the situation in the Northern Hemisphere, many wetlands in southern Africa experience a strong seasonal deficit in water as a result of the semi-arid climate characterised by potential evaporation rates that substantially exceed precipitation rates (Rogers, 1995; McCarthy and Hancox, 2000; Ellery *et al.*, 2004). This means that most southern African wetlands are sustained hydrologically by water inputs (surface and/or sub-surface flows) from the surrounding catchment and are thus commonly linked with drainage lines. Most southern African rivers are in a long-term state of incision as a result of multiple tectonic uplift events of the sub-continent following the rifting of Gondwana, which has resulted in an interior plateau of unusually high mean elevation (> 1000 m amsl, Cowan, 1995) characterised by a dense, well integrated drainage network (Partridge and Maud, 2000). The Highveld interior is thus undergoing erosion on a large scale and its high elevation reduces the likelihood of water accumulating within the landscape. These factors compromise the formation and long-term existence of wetlands within the South African landscape and one would expect a lack of wetlands particularly within the Highveld interior. However, this is not the case and most catchments within South Africa host wetlands (Cowan, 1995) and within the South African Highveld interior large floodplain systems are common features of the landscape (Tooth *et al.,* 2004).

This raises the question of which factors are promoting the formation and existence of wetlands within the southern African landscape, and especially within the Highveld interior. The unique geological history of southern Africa, which has been largely shaped by its association, and subsequent break-away from the supercontinent Gondwana, means that the landscape comprises a diverse array of rock types of varying age and resistance to erosion (King, 1963). Many rivers within the Highveld interior are situated on or close to bedrock (Tooth *et al*., 2004) due to long-term landscape incision and are thus subject to strong geological controls which include variations in rock structure and lithology (King, 1963; Tooth *et al*., 2002a, 2004). Hence, research on the controls on wetland formation and dynamics has thus far been focused around geological controls and has been limited to a few large inland floodplain wetland systems situated within the eastern South African Highveld interior (e.g. Tooth *et al.,* 2002a, 2004; Tooth and McCarthy, 2007). Within these systems, resistant Karoo dolerite dykes and sills have strongly influenced alluvial river behavior and concomitant floodplain wetland formation, where they have become exposed along drainage lines. These dolerite dykes and sills are more resistant to erosion than surrounding softer Karoo sedimentary rocks, and thus form stable local base-levels along the course of a river. Rivers respond upstream of these dolerite barriers by laterally planing the softer Karoo sedimentary rocks and have thus been able to carve broad (between 1 and 1.5 km wide), near-planar valleys (e.g. Tooth *et al*., 2002a, 2004). The floodplain wetlands that form within the accommodated space are characterised by an extensively meandering river and a diverse array of floodplain features such as oxbow lakes, abandoned meander belts, point bar deposits and flooded backswamps (Tooth and McCarthy, 2007). Alluvial valley fill is maintained as a relatively thin layer (< 4 m) over the bedrock by the reworking of floodplain sediments by a continually shifting river channel (Tooth *et al*., 2002a, 2004).

Consequently, these systems are characterised by low organic content and minimal peat accumulation. Contrastingly, within the downstream dolerite valley, lateral erosion is limited and the river follows a much straighter course through a narrow valley (< 400 m) of limited floodplain width (Tooth *et al*., 2002a, 2004). Over time, breaching of dolerite barriers results in incision and headcut gully formation along upstream reaches, and marks the onset of floodplain erosion and desiccation (Tooth *et al*., 2004).

Wakkerstroom Vlei is situated with the eastern South African Highveld interior and is a large (~ 1000 ha) wetland system dominated by unchanneled valley-bottom wetland and has been described as a wetland of high regional significance (Begg, 1989; Birdlife South Africa, 2006). Wakkerstroom Vlei lies within a broad (> 1 km), near-planar Karoo shale valley, and is confined within the head and toe regions by Karoo dolerite sills. Along the upper reaches of the wetland, where dolerite outcrops along the valley, the main (Wakkerstroom) river is well defined and meanders through a relatively narrow floodplain for a short distance before disappearing where it enters broader, shale valley reaches. Flow becomes largely diffusive along these reaches and is widely dispersed by the dense growth of *Phragmites australis*. Towards the toe of the system the main (Thaka) river reforms and once again meanders through a relatively narrow floodplain shortly before exiting the wetland and flowing through the downstream dolerite valley. Wakkerstroom Vlei contrasts to the formerly described floodplain systems, in being characterised by the absence of a river channel and floodplain features over most of its reaches, and although alluvial valley fill is thin (< 4.5 m), extensive peat deposits are common along the unchanneled valley-bottom wetland reaches. Several tributaries enter the Wakkerstroom valley and disappear at the wetland margins as alluvial fans. The hydro-geomorphological differences between Wakkerstroom Vlei and those wetlands described by Tooth *et al.* (2002a; 2004) are sufficient to warrant a study of the factors that have controlled the formation and hydro-geomorphological dynamics of Wakkerstroom Vlei.

1.2. Aim and objectives

This study aims to investigate the factors that have influenced the formation and hydro-geomorphological dynamics of Wakkerstroom Vlei in an attempt to expand the conceptual framework of the formation of large inland valley-bottom wetland systems in South Africa. In order to achieve this aim the following objectives have been identified:

- To describe the geomorphological characteristics and processes together with the controls thereon, within Wakkerstroom Vlei and the immediate surrounding catchment;
- To describe the extent, depth and physical properties of the wetland sedimentary fill;
- To develop a conceptual model of the origin and development of Wakkerstroom Vlei.

Chapter 2

Study Area

2.1. Location, physiography and geology

Wakkerstroom Vlei is situated in the uppermost reaches of the Tugela catchment (Figure 2.1), and is a large wetland system (~ 1000 ha) comprising mostly unchanneled valley-bottom wetland. The wetland lies immediately west of the town Wakkerstroom and falls within both the provinces of Mpumalanga and KwaZulu-Natal in the South African Highveld (Figure, 2.1).

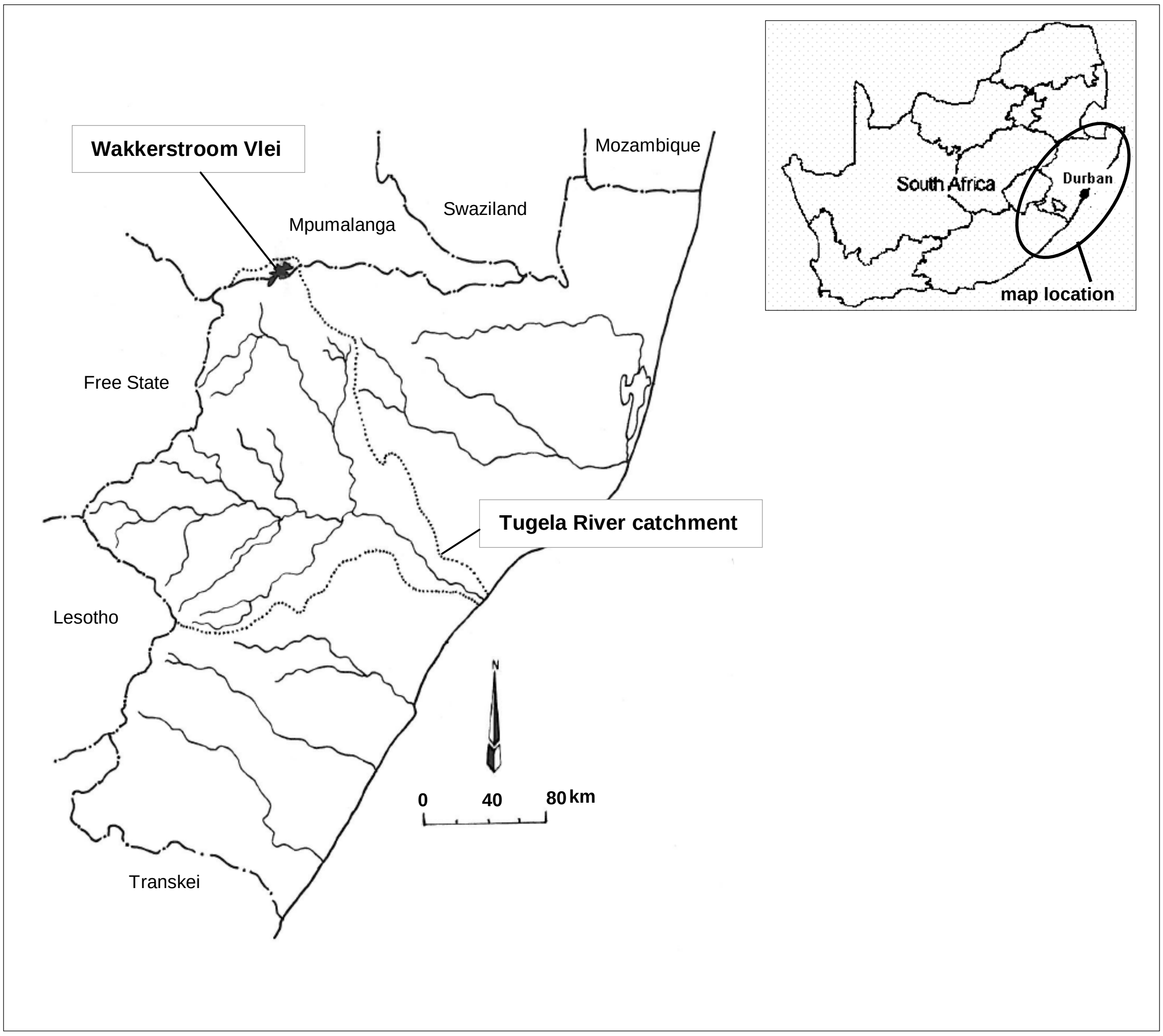


Figure 2.1: The location of Wakkerstroom Vlei (adapted from Begg, 1989).

According to Begg (1989), the wetland occupies a relatively small percentage (~ 4.8 %) of its ~ 207 km^2 catchment (quaternary sub-catchment V051, Figure 2.2), which is estimated to have a high mean annual runoff of approximately 30 x 10^6 m^3.

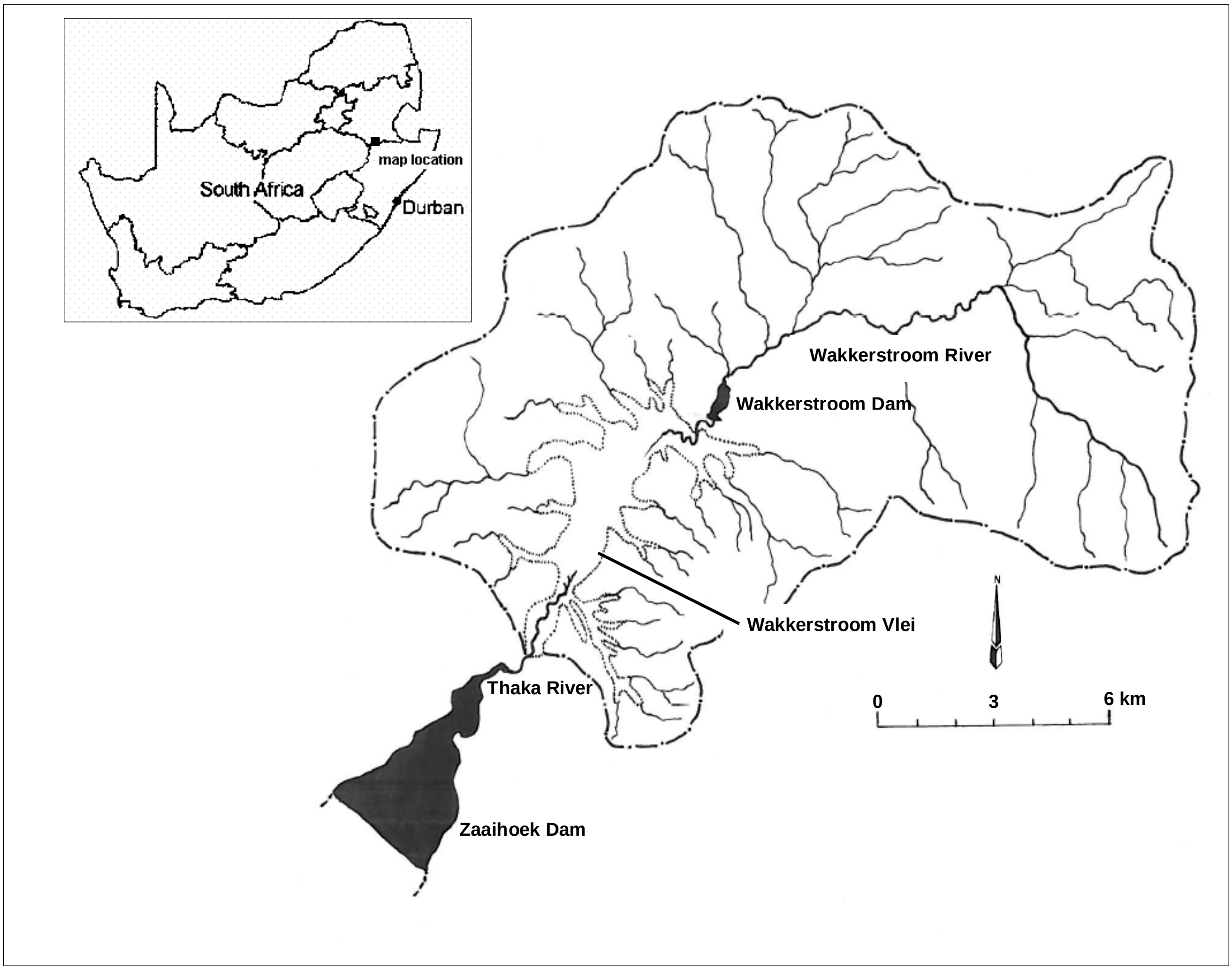


Figure 2.2: Wakkerstroom Vlei and surrounding catchment (adapted from Begg, 1989).

Main surface water inputs occur via the Wakkerstroom River, which was dammed (Wakkerstroom Dam) shortly upstream of the wetland prior to 1953, as well as via several relatively short, steep tributary streams, that rise within the higher lying hills surrounding the wetland (Figure 2.2). The Wakkerstroom River disappears along approximately 5 km of the main body of the wetland but reforms once again within the lower reaches of the system as the Thaka River, which was dammed (Zaaihoek Dam) shortly below the wetland in 1987 (Begg, 1989).

Most of the wetland is underlain by Ecca shale (Volksrust formation) however, Karoo dolerite outcrops at the head and toe of the system (Figure 2.3). Karoo sandstone (Escourt formation) together with Karoo dolerite form higher lying hills surrounding the wetland (Figure 2.3).

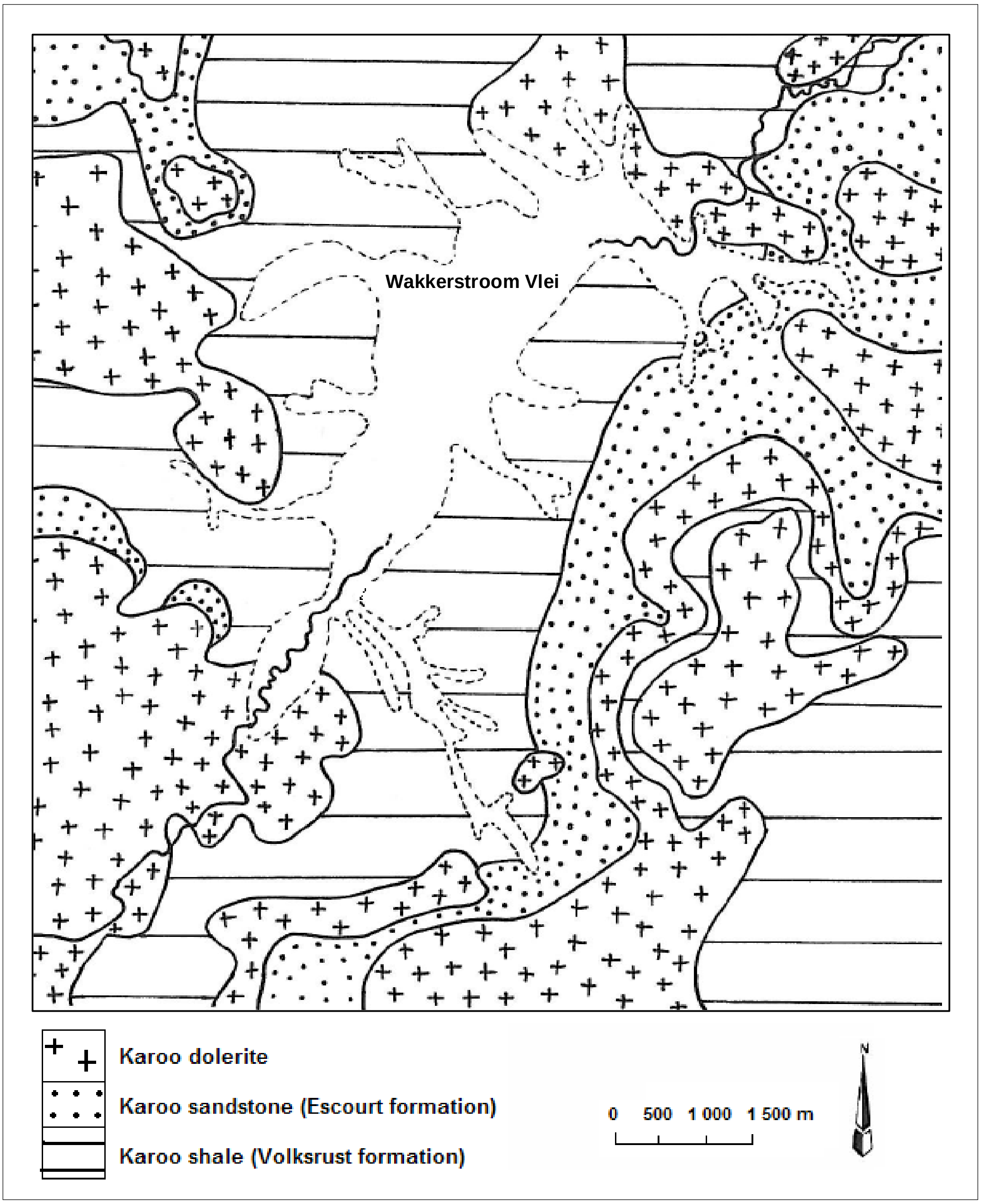


Figure 2.3: The geology of Wakkerstroom Vlei and the immediate surrounding catchment (adapted from Begg, 1989).

Begg (1989) describes Wakkerstroom Vlei as a priority wetland within the upper Tugela catchment as its large area, relatively low slope (~ 0.27 %, Kotze *et al.,* 1994) and high surface roughness (provided by *Phragmites australis*) means that it has high water purification, water storage and flood attenuation value and is thus important to downstream water users and those supplied with water from Zaaihoek Dam. These physical characteristics mean that the wetland is also important in providing nesting and foraging habitat for many globally and locally threatened bird species, some of which include: the White-winged Flufftail, which is one of the rarest bird species in Africa, the Grey Crowned Crane and the African Marsh Harrier (Kotze *et al.,* 1994; Birdlife South Africa, 2006). Wakkerstroom Vlei is thus a birding hotspot within KwaZulu-Natal/Mpumalanga as it hosts a rich diversity of bird species (Birdlife South Africa, 2006).

2.2. Climate and vegetation

Mean annual rainfall for the region is relatively high (~ 800 mm/annum), however annual potential evapotranspiration rates are more than double this (~ 1 650 mm/annum) (Phillips, 1973, cited in Kotze *et al.,* 1994). Strong seasonal differences in temperature are experienced over the study area with maximum temperatures of 35º C in summer and minimum temperatures of - 11º C in winter.

The rainfall over the study area, as indicated by rainfall data from two nearby rainfall stations at Volksrust (Figure 2.4, Station VA and VB), is variable over both short- (year-to-year) and longer-term (several years) time periods. Over longer time periods, there is clear evidence for cyclic shifts between relatively wet rainfall periods (Figure 2.4, ‘wet cycle’) of above-average annual rainfall (623 mm/annum, Figure 2.4, ‘Average annual for period of record’) and relatively dry rainfall periods (Figure 2.4, ‘dry cycle’) of below-average annual rainfall that last for periods of between 4 to 10 years.

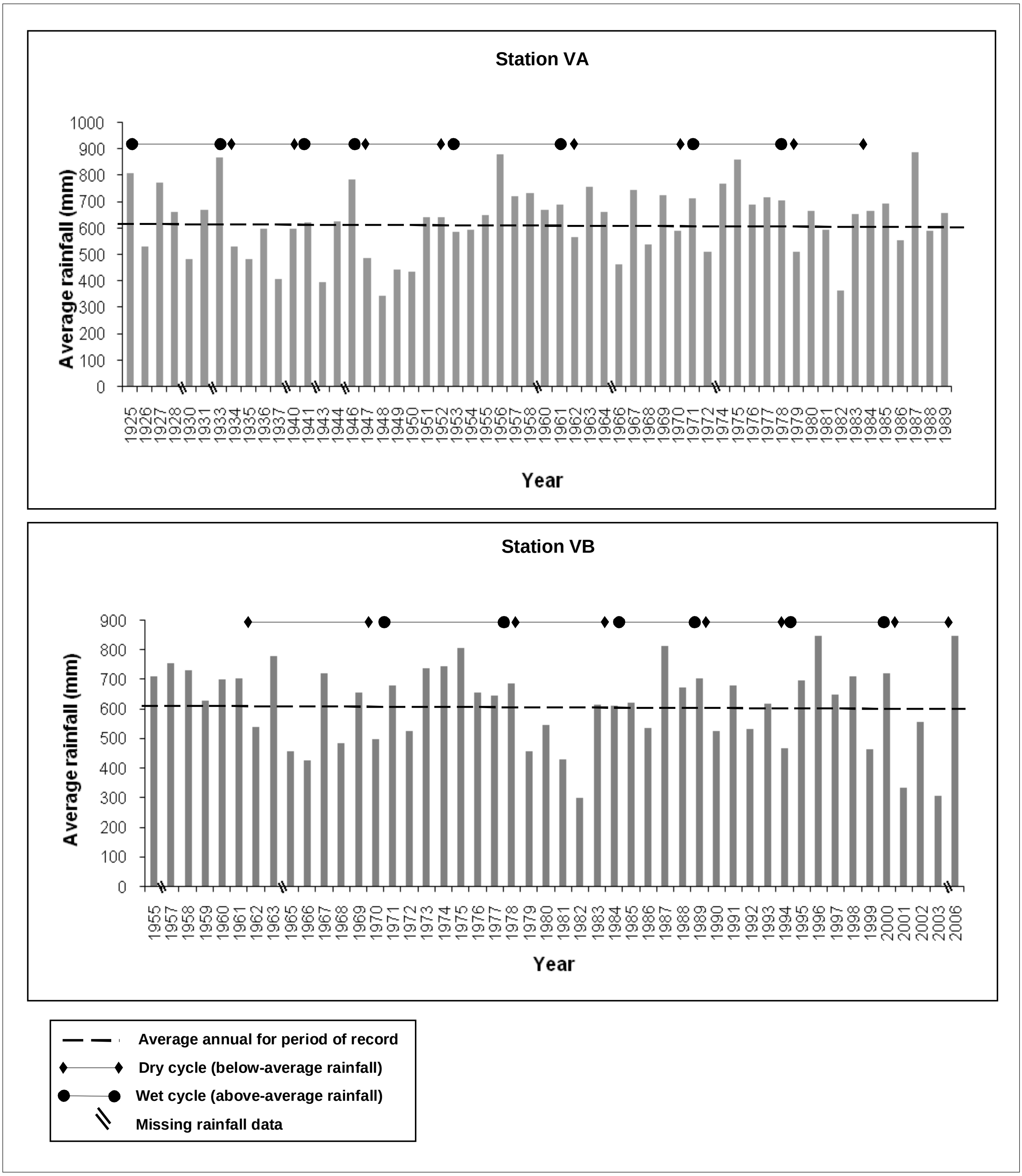


Figure 2.4: Annual rainfall for the periods 1925-1989 and 1955-2006 using available rainfall data from two rainfall stations in Volksrust (Station VA and Station VB respectively), approximately 30 km north-west of Wakkerstroom Vlei.

Several vegetation communities are distributed throughout the wetland (Figure 2.5). The unchanneled main body comprises reed marsh, dominated by the tall reed *Phragmites australis* which forms a dense colony across the wetland (Kotze *et al.*, 1994). The uppermost and lower most floodplain regions are characterised by short growing wet grassland, on the higher lying floodplain areas, and sedge/bulrush marsh on the lower lying floodplain areas. Wetland sidearms are characterised by short growing wet grassland, sedge meadow and sedge/bulrush marsh (Kotze *et al.,* 1994). The reeds *P. australis* and *Carex acutiformis* are the two major peat forming species within the system (MC Grenfell, 2006, pers. comm.).

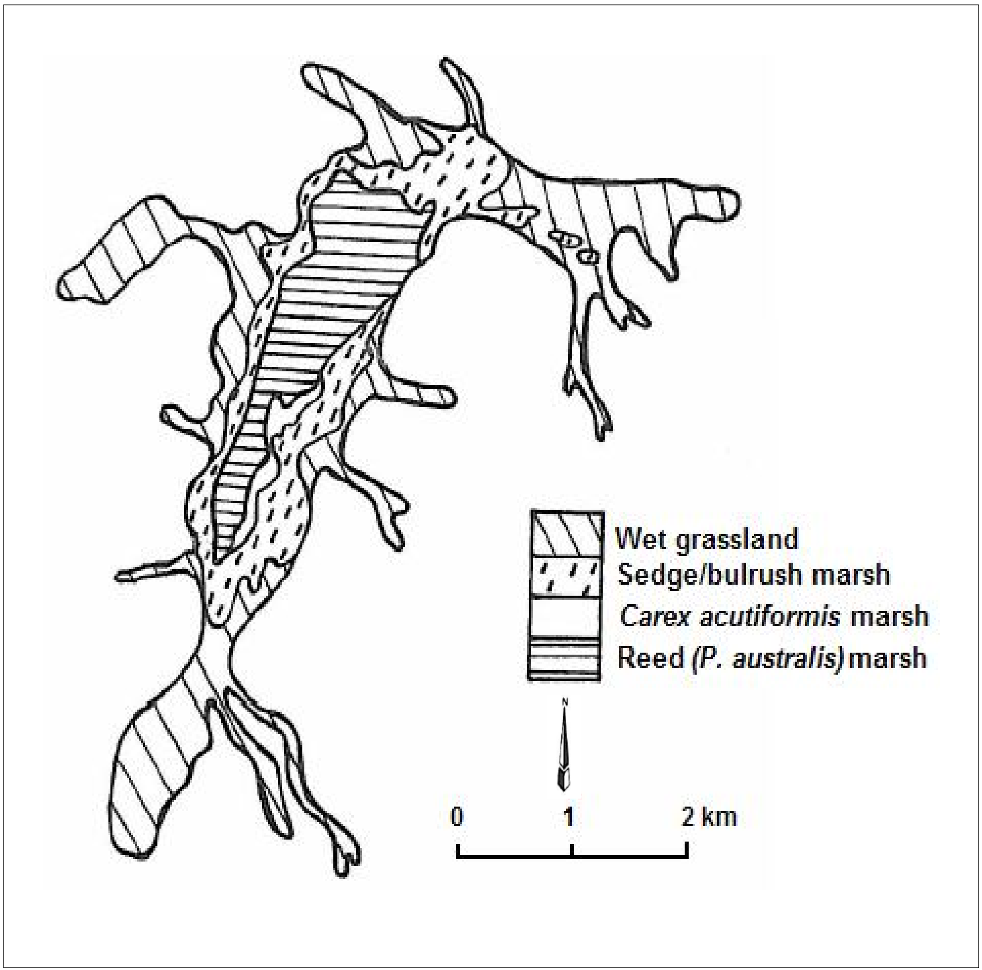


Figure 2.5: General vegetation characteristics of Wakkerstroom Vlei (from Kotze *et al.,* 1994).

2.3. Socio-economic characteristics

Wakkerstroom Vlei is one of the few wetlands still owned by a town municipality (Kotze *et al.,* 1994). Land-use in the wetland catchment is dominated by cattle and sheep farming which use natural veld grazing. Small amounts of crop and pasture production also take place within the catchment. The wetland is used by both local residents and commercial farmers to graze cattle and this forms the dominant land-use within the wetland, especially during drier years with up to 2 500 head of cattle being reported (Begg, 1989). Thus far this activity has promoted slight degradation of wet grassland communities and slight gully erosion, however these gullies have remained stable (Kotze *et al.,* 1994). The wetland burns annually in winter and this does not appear to significantly affect hydrological values or organic matter content (Kotze *et al*., 1994). Other small-scale activities within the wetland include mowing of grass for hay by local farmers (Begg, 1989) and harvesting of reeds and grasses by local residents (Kotze *et al.,* 1994).

Chapter 3

Literature Review

3.1. Introduction

Understanding the controlling factors in wetland formation and dynamics involves understanding the interaction between hydrological, geological and geomorphological processes (McCarthy and Hancox, 2000). Wetlands may be broadly defined from a hydrological basis by soils that are saturated at or close to the land surface for prolonged time periods (> several weeks) (McCarthy and Hancox, 2000). This unique characteristic is ultimately controlled by the way in which geological and geomorphological factors interact with local hydro-climatic factors to promote accumulation of water within the landscape. Wetlands thus occupy specific hydro-geomorphological positions within the landscape and this requires an understanding of the way in which historical and current geological, climatic and geomorphological processes have combined in shaping the landscape. These fundamental processes are dynamic in time and space and govern the long-term existence of wetlands. The southern African landscape has been shaped by geological and climatic events associated with the fragmentation of Gondwana and subsequent geological and geomorphological processes. The current physiography and climate of southern Africa thus owes its origin to this geological and geomorphological history of the sub-content. Knowledge of the role of these historical processes in creating particular hydro-geomorphological settings conducive to wetland formation is important in understanding the formation and long-term existence of wetlands within the southern African landscape.

3.2. Broad-scale physiography and climate of southern Africa

The southern African landscape is ancient with no recent (last 100 000 years) significant tectonic events, mountain building episodes or glaciation (Moon and Dardis, 1988). The interior Highveld plateau has an unusual hypo-symmetry and is bordered by a considerable marginal escarpment both of which are drained by a well integrated drainage network (Partridge and Maud, 2000). The modern climate is semi-arid over most of southern Africa excluding areas along the coastal margin, and it is characterised by mean annual potential evaporation rates (1100 - 3000 mm) that are generally more than twice the mean annual rainfall (< 490 mm) (Tyson and Preston-Whyte, 2000). A distinct east-west rainfall gradient exists with eastern and south eastern coastal regions receiving considerably more rainfall than the interior and western regions of the sub-continent. These modern characteristics of the southern African environment have been inherited from unique geological and climatic events associated with the inclusion and subsequent break-away of the African continent from the supercontinent Gondwana.

3.3. The pre-Cenozoic environment

The geological, geomorphological and climatic history of the southern African landscape is summarized in Table 3.1. Between approximately 550 Ma and 182 Ma the African continent formed part of Gondwana and the southern African landscape was largely aggradational (Partridge and Maud, 2000). Tectonic activity around the southern margins of Gondwana resulted in an extensive inland basin within which sediments of the Karoo Supergroup were deposited between approximately 300 – 180 Ma (Smith *et al.,* 1993; Catuneanu *et al.,* 2005). The rocks of this Supergroup are the most extensive in South Africa and currently cover around two thirds of southern Africa (Smith, 1990), totaling 8 km in thickness in some areas (Smith *et al.,* 1993). The first Karoo Sediments were glacially derived and deposited at the onset of the Dwyka glaciation (~ 310 Ma) during which southern Africa was covered by a continental ice sheet as it was located approximately at the South Pole (McCarthy and Rubidge, 2005). The general southward movement of this ice sheet resulted in ice scoured pavements, especially over the southern region of southern Africa (King, 1963). These glacial pavements were covered during subsequent Karoo sedimentation. The Karoo Supergroup has not been appreciably faulted or folded, and it is responsible for much of the flat topography over the interior of South Africa (King, 1963).

The period of Karoo sedimentation ended around 180 Ma with extensive basaltic lava outflows associated with the breakup of Gondwana (Smith *et al.,* 1993; Partridge and Maud, 2000). Some basic magma cooled below-surface as intrusive dolerite sills and dykes that are widespread in the rocks of the Karoo Supergroup. The mantle plume associated with this event resulted in extensive and considerable upwarping and uplift of the sub-continent (Segev, 2002), during which the interior of the sub-continent was considerably elevated to a mean altitude of around 2000 m amsl (McCarthy and Rubidge, 2005), at the same time exposing an escarpment along the continental margin.

Table 3.1: Summary of the major geological, geomorphological and climatic events in the history of the southern African landscape.

Geological time period	Geological event	Geomorphological process	Climatic event
Late Jurassic – early Cretaceous (180 – 140 Ma)	Sub-continental uplift associated with mantle lava plume; extensive lava outpourings and intrusion by Karoo dolerite dykes and sills (Smith *et al*. 1993)	Macro-scale drainage development (Partridge and Maud, 2000)	
Early Cretaceous – late Cretaceous (140 – 65 Ma)	Opening of Proto-Indian, Atlantic and Southern Oceans (Partridge and Maud, 2000)	African erosion cycle (King, 1963)	Shift to relatively warm, moist conditions (Tyson and Preston-Whyte, 1988)
Late Cretaceous (65 Ma)	Global catastrophe (assumed meteorite impact); (McCarthy and Rubidge, 2005)		Sharp shift to relatively cold, arid conditions (Tyson and Preston-Whyte, 1988)
Late Cretaceous to early Miocene (65 – 24 Ma)		Widespread deposition of fluvial sediments and aeolian sands (King, 1963)	
Early – mid-Miocene (24 – 14 Ma)			Shift to relatively warm, moist conditions (Tyson, 1986)
End of early Miocene (20 Ma)	Moderate asymmetric sub-continental uplift (up to 250 m); (Partridge and Maud, 1987)		
Early Miocene to early Pliocene (20 – 5 Ma)		Post-African I erosion cycle associated with Miocene uplift and river rejuvenation (King, 1963)	
Early Pliocene (5 Ma)	Major asymmetric sub-continental uplift (up to 900 m); (Partridge and Maud, 1987)		
Late Pliocene - present		Post-African II erosion cycle associated with Pliocene uplift and river rejuvenation (King, 1963)	

Uplift associated with the rifting of Gondwana was significant in initiating macro-scale drainage development over southern Africa, which existed in the form of three principle ancient drainage systems, namely the Kalahari, Karoo and paleo-Limpopo systems (McCarthy and Rubidge, 2005). These systems drained most of central and south western southern Africa with several smaller systems draining the marginal escarpment and this, together with the elevated nature of the sub-continent, exposed the southern African landscape to extensive and relatively rapid fluvial erosion. As the continent of Gondwana began to break-up, the sub-continent of southern Africa was surrounded with warm oceans that resulted in an associated climatic shift to relatively warm, moist conditions (Tyson and Preston-Whyte, 1988; McCarthy and Rubidge, 2005). This enhanced the rapid erosion of the interior and marginal escarpment associated with sub-continental uplift, and as a result large amounts of the Karoo sedimentary rocks were removed from the interior during the Cretaceous (~ 142 - 65 Ma) exposing a variety of underlying lithologies - especially the Karoo dolerite dykes and sills that are relatively more resistant to erosion than surrounding softer Karoo rocks (de Wit *et al.,* 2000). These dolerite dykes and sills commonly outcrop across the southern African landscape where they form erosion resistant features in the landscape (King, 1963). This major cycle of erosion in the history of the southern African landscape has been termed the African erosion cycle (King, 1963) and was the first of three major erosional cycles of the southern African landscape.

The African Erosion Cycle was halted by a major shift in climatic conditions at the end of the Cretaceous (~ 64 Ma) during a severe global shift towards aridification and decline in temperatures (Tyson and Preston-Whyte, 1988). Several causes for this global climatic shift have been proposed the most widely accepted of which has been the influence of a large-scale meteorite impact (Ward *et al.,* 2000). Fluvial activity in southern Africa was strongly affected by this event as flow declined in rivers and ceased in many smaller drainage basins (Partridge and Maud, 2000), and mass floral extinctions associated with this climatic shift resulted in basin-wide changes in fluvial form from meandering rivers to braided rivers as rivers responded to a marked increase in sediment yield (Ward *et al.,* 2000). Pronounced cooling of the Southern Ocean at around 38 Ma resulted in further declines in temperature and aridity over southern Africa (Tyson, 1986). The southern African landscape thus underwent little change during the Palaeogene (~ 65 - 24 Ma). Major shifts in drainage however occurred over the interior in association with downwarping and formation of the Kalahari Basin (Partridge and Maud, 2000; Haddon and McCarthy, 2005). During this event the Limpopo River was cut from its upper tributaries forming the ancestral Lake Makgadikgadi and the Kalahari River captured the Karoo River to form the present day Orange River which forms one of the largest drainage systems covering most of the South African interior (McCarthy and Rubidge, 2005).

The southern African landscape experienced two major uplift events during the Miocene (~ 20 Ma) and the Pliocene (~ 5 Ma) during which the sub-continent was substantially elevated and warped along its margins (Partridge and Maud, 2000). These events were significant in reinitiating widespread erosion and degradation of the southern African landscape by the newly developed drainage networks and were key events in shaping the modern climate and landscape. Uplift was preferential along the southern and eastern coastal margins with uplift values of around 250 m in the east and 150 m in the west during the first uplift event (Miocene uplift), and around 900 m in the east and 100 m in the west during the second uplift event (Pliocene uplift) (Partridge and Maud, 1987). Drainage lines, particularly along the eastern and southern escarpment were considerably steepened resulting in rejuvenation and renewed incision along river networks. At the same time, fluvial erosion was enhanced by the relatively warm and moist climatic conditions which prevailed over the sub-continent during the first half of the Miocene (Tyson, 1986; Tyson and Partridge, 2000). As a result the southern African interior was extensively and rapidly eroded and large quantities of basalt and Karoo rocks were removed following these two successive erosional cycles related to the Miocene and Pliocene uplifts and these have been termed the Post African I and Post African II erosion cycles (King, 1963). This further contributed to extensive exposure of underlying more resistant lithologies, most commonly Karoo dolerite dykes and sills.

The modern climate of southern Africa was also shaped during the Miocene and Pliocene uplift events as well as by the evolution of ocean currents during the Miocene. Asymmetric uplift of the sub-continent isolated the interior and western regions of the landscape from rainfall, thereby establishing the pronounced east-west rainfall gradient. This was further enhanced by the establishment of the cold Benguela ocean current during the late Miocene (~ 14 Ma ago) (Tyson and Preston-Whyte, 1988). Thus during the Miocene and Pliocene substantial amounts of basalt and Karoo sediments were removed from the southern African interior and rivers became superimposed upon underlying Karoo dolerite dykes and sills, which exert extensive structural and lithological controls on alluvial river behaviour within the southern African region (Partridge and Maud, 1987; de Wit *et al.*, 2000).

3.4. The Cenozoic environment

Over the past 5 million years the southern African climate has been highly variable and has been characterised by variations between relatively cool and dry climatic periods followed by relatively warm and moist climatic periods over varying time-scales. Tyson and Partridge (2000) suggest that these fluctuations have occurred around a semi-arid climatic mean. Associated with these climatic shifts have been shifts in fluvial activity between periods of fluvial erosion (degradational regimes) and fluvial deposition (aggradational regimes) as a result of changes in river flow regimes and vegetation cover within the catchment (e.g. Ward *et al.*, 2000; Williams *et al.*, 2001).

Evidence suggests that over prolonged time-scales (hundreds of thousands of years) the global climate has shifted between relatively cool and dry conditions (glacial periods) and relatively warm and moist conditions (interglacial periods) as indicated by the waxing and waning of northern and southern hemisphere ice sheets. Eustatic shifts in sea-level associated with these climatic variations were most significant in promoting cycles of erosion along coastal drainage networks, with rivers becoming more active during falling sea-levels associated with glacial periods. The last major eustatic shift occurred around the time of the last glacial maximum approximately 20 000 – 12 000 years before present (Tyson and Partridge, 2000) when global sea levels fell to around 120 m below present levels (Peltier, 2002). In contrast, rivers of the Highveld interior entered erosional cycles and became more active during moister interglacial periods (King, 1963), related to increased flow regimes as well as decreased sediment supply within the catchment as a result of stabilization of hill-slopes by increased vegetation cover (Maddy, 2002). The most recent large-scale shift in climate occurred after the last glacial maximum at around 12 000 years ago with a global shift to relatively warm, moist conditions (Tyson and Partridge, 2000).

Shorter-term variations in rainfall have been superimposed upon these longer-term climatic trends and are a unique feature of the modern southern African climate (Tyson *et al.,* 2000; Tyson and Preston-Whyte, 2000). These variations range from millennial to multi-decadal shifts between relatively warm and wet and relatively cool and dry conditions (Scott, 1989; Talma and Vogel, 1992; Tyson *et al.,* 2002; Holmgren *et al.,* 2003). In particular, there is evidence for distinct quasi-cyclic oscillations in rainfall ranging for periods of between ~500-800 years, ~ 80 years and ~ 18 years (Tyson *et al.,* 2002). The quasi-18 year oscillations are described as the most pronounced feature of the southern African climate and since the early 1900's southern Africa has experienced 8 quasi-cyclic oscillations (Tyson and Preston-Whyte, 2000, Table 3.2).

Table 3.2: Time intervals for wet and dry quasi-cyclic oscillations in rainfall over South Africa for the period 1905 - 1981 (Tyson and Preston-Whyte, 2000).

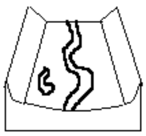
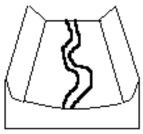
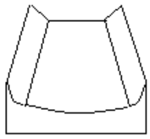
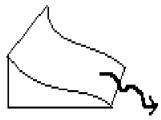
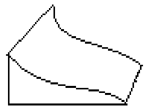
Relatively wet	Relatively dry
1916 – 1925	1905 – 1916
1933 – 1944	1925 – 1933
1953 – 1962	1944 – 1953
1971 – 1981	1962 - 1971

The last peak of a relatively wet rainfall cycle occurred around 1983/84 (Tyson, 1986), followed by a shift toward relatively dry conditions during the 1990's. At present South Africa is experiencing relatively warm, wet climatic conditions with the approach toward another peak within a relatively wet rainfall cycle of the quasi-18 year rainfall oscillations. In general, these climatic variations resulted in enhanced erosion during relatively warm and moist climatic periods, particularly over the southern African Highveld interior, and this further contributed to the removal of Karoo sedimentary rocks and exposure of underlying more resistant Karoo dolerite dykes and sills.

3.5. Accomodation for wetland formation in southern Africa

Hydrology (water regime) and geomorphology (landform) are recognized as the two fundamental determinants of the existence of wetlands within the landscape (Mitsch and Gosselink, 2000). These factors are not mutually exclusive and where they interact in a unique way are able to form the fundamental underlying controls on wetland formation despite characteristics of climate, soil type, vegetation or geological/geomorphological origin (Ewart-Smith *et al.,* 2006). Wetlands are hydrologically defined by a positive water balance over long enough time periods for the development of anaerobic soil conditions that support hydrophytic plants adapted to living in these conditions (DWAF, 1998). This condition requires that water inputs (surface, subsurface, precipitation) exceed water outputs (surface, subsurface, evapotranspiration) from a particular landscape setting over prolonged time periods (> several weeks). Local geomorphological characteristics of the landscape can thus strongly control wetland formation by limiting the rate of surface and subsurface water losses relative to the rate of surface and subsurface water inputs from a particular landscape setting. This is particularly important where local climatic conditions do not promote a positive water balance as is the case in southern Africa on the Highveld, where the semi-arid climate means that most wetlands are associated with low energy fluvial depositional environments (Ellery *et al.*, 2004). Alluvial depositional environments should be lacking considering the geological and geomorphological history of the sub-continent, as rivers are in a long-term state of incision. However, the South African interior hosts a wide range of hydro-geomorphic wetland types associated most commonly with fluvial landforms (Rogers, 1995; Kotze *et al.,* 2005). Kotze *et al.* (2005) define six hydro-geomorphic wetland types, based upon geomorphological (landscape) setting, the main water inputs to the system and how water flows through and leaves the wetland system, and these include: floodplain, valley-bottom with a channel, valley-bottom without a channel, hillslope seepage feeding a stream; hillslope seepage not feeding a stream and depression (including pans) wetlands (Table 3.3). This means that understanding the controls on alluvial river behaviour and accommodation for wetland formation, through valley widening and sediment deposition, is essential in understanding wetland formation and dynamics in southern Africa.

Table 3.3: Wetland hydro-geomorphic types defined for inland wetlands in South Africa (from Kotze *et al*., 2005).

Hydro-geomorphic type	**1.** Floodplain 	**2.** Valley-bottom with a channel 	**3.** Valley-bottom without channel 	**4.** Hillslope seepage feeding a stream 	**5.** Hillslope seepage not feeding stream 	**6.** Depression (including Pans) 
Main water inputs/outputs	Mainly surface	Mainly surface	Mainly surface	Mainly subsurface inputs and surface outputs	Mainly subsurface	Either surface /subsurface or precipitation/ evaporation

The behaviour of rivers in the landscape is governed by the concept of river equilibrium, wherein rivers seek to achieve minimum energy expenditure in transporting the available sediment load (Schumm, 1977). This is achieved by balancing discharge and channel slope, which define the transport capacity of a river, with the available sediment load. Over long time-scales a river attempts to achieve equilibrium over its entire longitudinal profile by lowering longitudinal slope to the lowest position within the landscape which is global base-level (sea-level), thereby achieving no net erosion or deposition along its length (Schumm, 1977).

The feature that limits the depth to which a stream can lower its bed is known as the base-level. Ultimately, the base-level for streams is sea-level, since a stream flowing into the sea cannot erode its bed below this level. Once a river has eroded a slope such that no net erosion or deposition along its length is possible, it is referred to as a graded stream (Schumm, 1977). Such an equilibrium stream profile is characteristically concave upwards in shape (Knighton, 1984). Rivers also grade in response to local base-levels along the drainage line which may exist for example in the form of resistant rock outcrops or man made dams where these are relatively stable over long time periods of several thousand years. Given local base-levels along the course of a stream, a stream may achieve a stepped longitudinal profile (Figure 3.1).

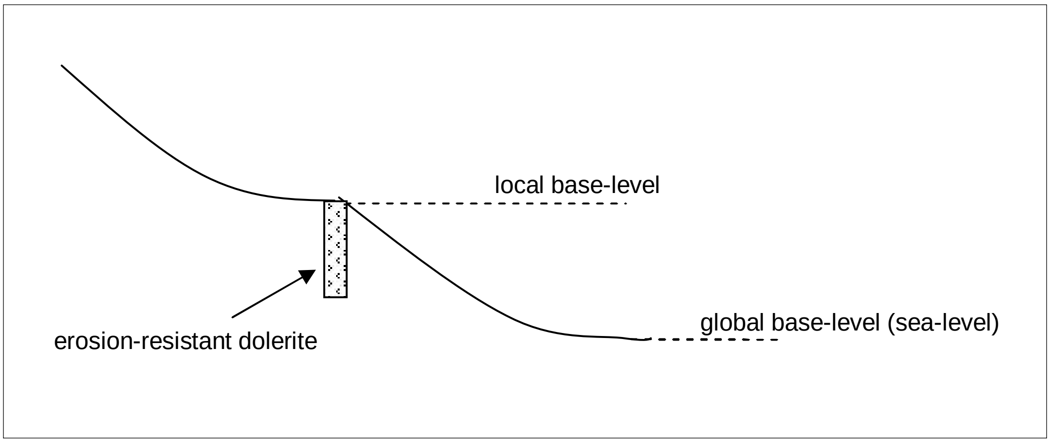


Figure 3.1: Schematic diagram of a stepped river longitudinal profile achieved through river grade according to local and global base-levels.

Equilibrium thresholds may be crossed through changes in internal stream variables as the river naturally works towards its grade, or through changes in external river variables such as variation in the hardness of local lithology and geological structure, climatic variability, neotectonics, and variation in the elevation of base-level through variation in sea-level for example (Bridge, 2005). These are not mutually exclusive factors and may combine to influence river equilibrium.

Rivers respond to disequilibrium through processes of erosion or deposition combined with associated channel morphologic changes (geometry and pattern) as the river works to attain an equilibrium condition. The equilibrium channel morphology (channel geometry and channel pattern) that a river adopts may be considerably influenced by local environmental factors which include local geology, climate, soils and vegetation (Bridge, 2005). Rivers respond over different spatial and temporal scales according to the nature and magnitude of change in the external variable/s (Dollar, 2002). Rivers are thus dynamic in time and space with behavioral changes shaped by the integrated influence of external variable conditions on internal river equilibrium conditions.

In southern Africa several principle factors are thought to control river behaviour and thus accomodation space for fluvial deposition and concomitant wetland formation, including, neotectonic activity, climatic and eustatic sea-level variations and lithological and structural geological controls along drainage lines (McCarthy and Hancox, 2000). In all cases climatic variation has featured as a combining variable with other controls on accommodation space and wetland formation in southern Africa. Coastal rivers have been primarily influence by eustatic shifts in sea-level in combination with neotectonics (McCarthy and Hancox, 2000). In contrast, rivers of the Highveld have been influenced primarily by lithological and structural geological controls along drainage lines (King, 1963; Moon 1988; Partridge and Maud, 2000). As a result extensive alluvial depositional environments hosting large floodplain wetlands are common within the Highveld interior (Tooth *et al.*, 2002a, 2004).

3.6. Current concepts of wetland formation in southern Africa

Geological control on inland wetland formation was first suggested by Tooth *et al.* (2002a, 2004) for large floodplain wetland systems along the Klip, Schoonspruit and Venterspruit Rivers within the semi-arid Highveld of South Africa. This is currently the most widely accepted model of wetland formation in South Africa and forms the basis of understanding the geological origin of most inland floodplain systems. The floodplain systems described by Tooth *et al.* (2002a, 2004) drain relatively large catchments and their formation and dynamics are strongly controlled by the influence of variations in the hardness of different lithologies that straddle alluvial rivers and influence their behaviour over short to medium time-scales (centuries to tens of thousands of years). Where drainage systems on Karoo sediments have become superimposed upon Karoo dolerite dykes and sills, these resistant lithologies (dykes and sills) act as stable local base-levels that limit downward erosion of softer lithologies. Upstream of these dolerite barriers, vertical erosion is thus limited by the dolerite barrier and as a result the river has excess erosional energy which it uses by laterally eroding and planing the softer Karoo rocks (Figure 3.2). Along these valley reaches lateral planation is enhanced through horizontal bedding of the Karoo sedimentary rock strata. As a result the river carves out a broad valley of low longitudinal slope through the process of lateral-migration, thus creating accommodation space for fluvial deposition and floodplain wetland formation.

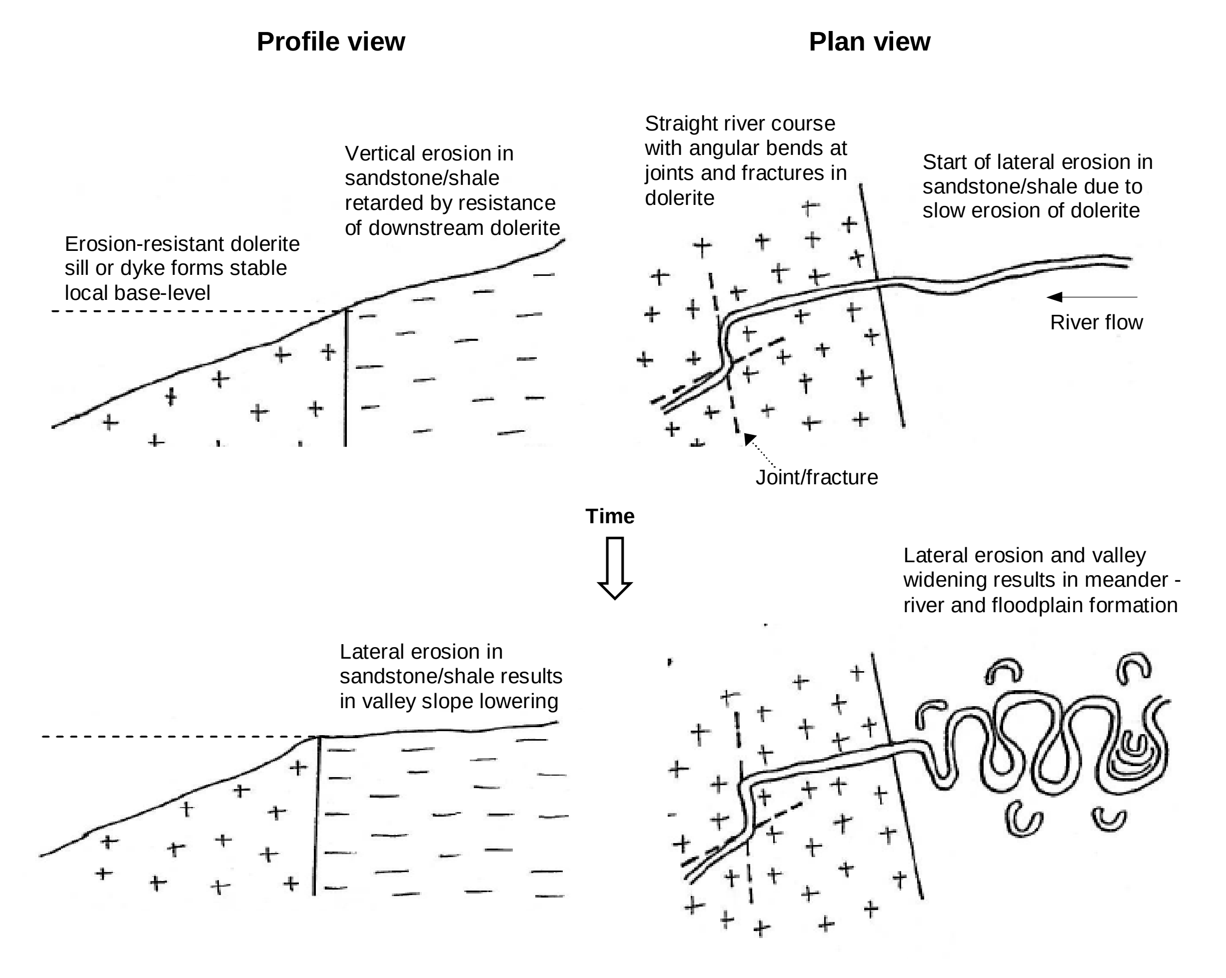


Figure 3.2: The current model of inland floodplain wetland formation illustrating control by resistant Karoo dolerite dykes and sills on alluvial river behaviour (from Tooth *et al*., 2002a).

As a result of these processes described by Tooth *et al.* (2002a, 2004) within these systems, Karoo dolerite valleys contrast sharply in geomorphological characteristics with Karoo sandstone/shale valleys as depicted in Table 3.4. The sandstone/shale valleys are characterised by an extensively meandering river channel and floodplain environment situated within a broad, gently sloping valley (Table 3.4). Common floodplain features along these broad valley reaches include numerous oxbows, meander scars, alluvial ridges, and flooded backswamps. As a result, floodplain alluvium is continually reworked and valley fill is characteristically thin and dominated by clastic sediments. Most tributaries entering the valley disappear at the margins of the floodplain wetlands with only few crossing the floodplain and joining the trunk stream. The upstream and downstream dolerite valleys are considerably narrower and steeper in gradient and the river follows a much straighter course as erosion is restricted to vertical lines of weakness such as joints and fractures (Table 3.4). As a result lateral erosion and floodplain width is limited within the dolerite valleys and alluvial fill is also characteristically thin. These catchments are in a long-term (hundreds of thousands of years) state of landscape degradation. However, over short to medium time-scales (centuries to tens of thousands of years) the interaction between the erosional processes within the dolerite and sandstone/shale valleys means that lateral erosion is able to dominate over vertical erosion within the shale/sandstone valleys, thus accommodating floodplain wetland formation (Tooth *et al.*, 2002a, 2004).

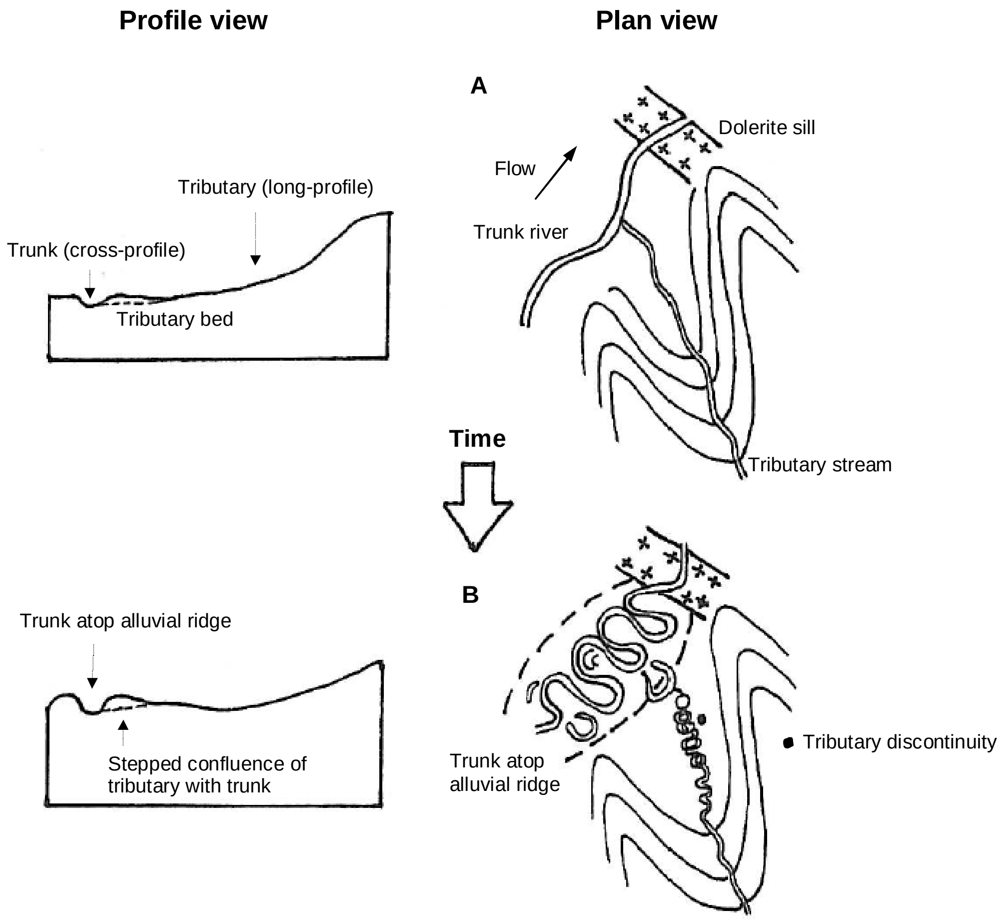
Table 3.4: Major differences in channel and floodplain geomorphological characteristics between floodplains that occur in resistant Karoo dolerite valleys and floodplains that occur in less resistant sandstone or shale valleys upstream of dolerite sills/dykes (Tooth *et al.,* 2002a, 2004).

	Dolerite valley	**Sandstone/shale valley**
Floodplain width	< 200 m	> 1000 m
Channel gradient	> 0.002	< 0.0011
Channel sinuosity	< 1.45	> 1.64
Depth of floodplain alluvial fill	< 4 m	< 4 m
Floodplain features	Levees, adjacent swales	Levees, oxbow lakes, scroll bars, alluvial ridges, backswamps, abandoned meander belts

The influence of the dolerite barriers in controlling the relative rate of vertical over lateral erosion is important in defining the long-term existence of these wetlands. Over longer time-scales (> tens of thousands of years) where these dolerite barriers gradually become eroded from the landscape, vertical erosion within the upstream floodplain reaches is reinitiated through gradual lowering and disappearance of the local base-level control (Tooth *et al.*, 2004). As a consequence, floodplain degradation occurs and is marked by numerous gullies that erode headward into the alluvial valley fill and soft bedrock lithologies, leading to wetland loss (Tooth *et al.*, 2004).

In this setting the relative resistance of different lithologies to erosion is important in influencing the relative rates of vertical and lateral erosion, and the time-scale over which these erosional processes occur. Areas of considerable variability in resistance to erosion will allow more time for the creation of broad floodplain wetlands.

A recent study by Grenfell *et al*. (2008) on the formation and dynamics of two wetland systems within the foothills of the KwaZulu-Natal Drakensberg has led to expansion of the Tooth *et al.* (2002a) model. Grenfell *et al*. (2008) studied Hlatikulu Vlei and Stillerust Vlei wetland systems associated with the Nsonge and Mooi Rivers respectively. Hlatikulu Vlei is a relatively small floodplain wetland system in comparison to the floodplain systems described by Tooth *et al.* (2002a) and lies upstream of a dolerite barrier within an unconfined valley of Karoo shale. Characteristic floodplain features and a thin alluvial valley fill (< 3 m) are maintained in accordance with the Tooth *et al.* (2002a) model of floodplain wetland formation. However, the findings for Stillerust Vlei deviate somewhat from the Tooth *et al.* (2002a, 2004) floodplain systems. Stillerust Vlei is a relatively small system comprised of two hydro-geomorphic wetland components, a floodplain component along the Mooi (trunk) River which lies upstream of a dolerite sill, and an abutting tributary valley-bottom component, the formation of which has been controlled by floodplain dynamics along the trunk river (Figure 3.3). In this system, sedimentation in the form of lateral and low vertical accretion along the trunk river floodplain resulted in the formation of an alluvial ridge which had the combined effect of blocking the mouth of the tributary stream valley and raising the base-level from that of the trunk river bed to the level of the alluvial ridge (Grenfell *et al*., 2008). The tributary stream responded by becoming more sinuous and consequently deposited sediment in a headward direction up-valley. This resulted in the formation of a valley-bottom wetland with a multiple thread stream channel plan-form as a consequence of impoundment and loss in stream energy (Figure 3.3) (Grenfell *et al*., 2008). These trunk-tributary interactions are ultimately controlled by the continued existence and influence of the dolerite sill that acts as a local base-level on the floodplain of the trunk river. Grenfell *et al*. (2008) noted that in this setting it is important that lateral accretion along the trunk river floodplain occurs at a faster rate than within the tributary valley floodplain to maintain geomorphological control on the existence of the tributary valley-bottom wetland. Stillerust Vlei thus represents a form of geomorphological control on inland wetland formation which has been born out of the underlying control of dolerite local base-levels on inland floodplain dynamics.



A: Trunk superimposed upon dolerite sill starts to laterally erode along upstream reaches. Base-level for tributary is trunk bed.

B: River meanders and floodplain wetland form. Lateral accretion along trunk valley results in tributary valley impoundment, tributary enters trunk via oxbows and has deferred junction with trunk river.

Figure 3.3: Model of trunk-tributary relations and inland valley-bottom wetland formation for Stillerust Vlei (from Grenfell *et al*., 2008).

Chapter 4

Methods

4.1. Division of the wetland into hydro-geomorphic units

The wetland system was divided into hydro-geomorphic components based on the classification of Kotze *et al.* (2005) (see Table 3.3, p.20) which aided in better understanding the hydro-geomorphological dynamics and interactions within Wakkerstroom Vlei. Hydro-geomorphic characteristics such as geomorphological setting together with the general nature of water flows into, through and out of the wetland were assessed from aerial photographic analysis and field survey evidence.

4.2. Current and historic geomorphology of the wetland and surrounding catchment

Several cross-valley profiles, perpendicular to the axis of the wetland were surveyed using dumpy level and staff and plotted as relative elevations. Profiles were positioned to capture downstream changes in main river (Wakkerstroom/Thaka River) channel and valley geomorphology from the dolerite valley shortly above the wetland to the dolerite valley shortly below the system (Figure 4.1). Cores to bedrock were taken at regular intervals along each profile within the wetland to determine wetland basin morphology. Cross-valley surveys were linked to one another by surveying stakes, positioned at the start of each profile, along the length of the valley using dumpy level and staff. In this way, relative valley-long bedrock, thalweg (both channeled and unchanneled) and wetland surface elevations could be plotted on a longitudinal profile to display downstream variation in the slope of bedrock, trunk channel bed and wetland alluvial fill.

Recent geomorphological change within the wetland and surrounding catchment was determined by analyzing a time sequence of aerial photography dated 1938, 1953, 1978, and 2003. This sequence included the oldest and most recent aerial photography for the study area. The morphology of the main river, tributary streams, geomorphological features (such as oxbow, meander cutoffs) and man-made structures, together with major changes in these features were mapped. This was done by digitizing rectified and georeferenced 2003 aerial photography of the study area using ArcGIS version 3.3, to produce a base-map upon which historic geomorphologic changes within the wetland and immediate surrounding catchment could be digitized. Each time phase of aerial photography was visually analyzed and compared using stereoscopic vision. Major geomorphologic changes were then digitized and mapped upon the 2003 base-map to produce a time sequence of maps displaying major changes in man-made features and geomorphology of the wetland and immediate surrounding catchment between 1938 and 2003.

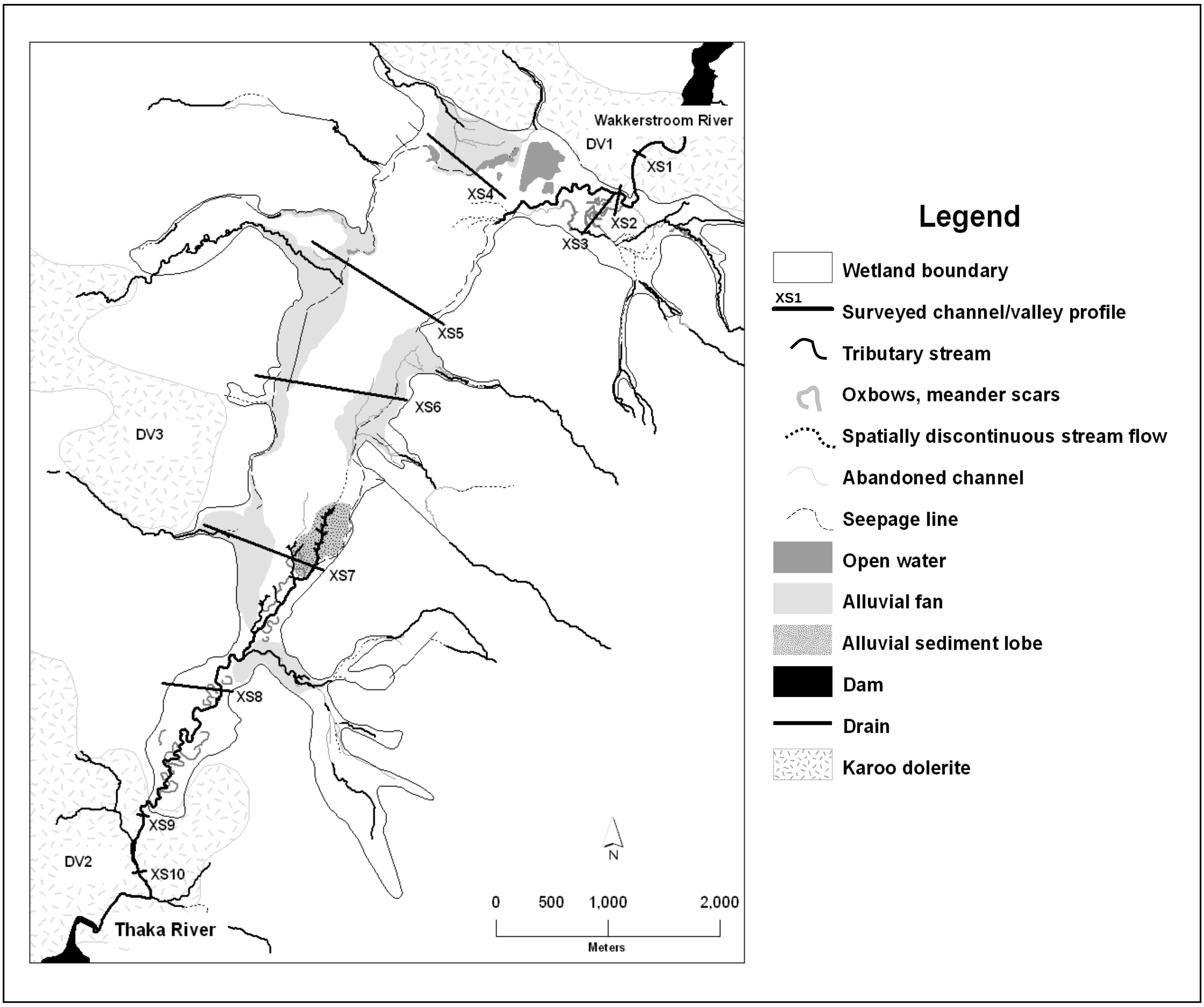


Figure 4.1: Position of surveyed cross-valley profiles within Wakkerstroom Vlei

4.3. Characteristics of wetland sedimentary fill

Wetland sediments were sampled at each core to bedrock along the cross-valley profiles using a gouge corer, which provides a relatively undisturbed core of sediment displaying detailed sediment stratigraphy. This allowed for accurate sampling of sediment at 0.5 m depth intervals from the wetland surface down to bedrock, including where noticeable morphologic and textural changes in sediment occurred. Where it became physically impossible to use the gouge corer method (where very ‘sticky’ clay layers of sediment were encountered), an auger was used to sample sediments. This method does not allow as much detailed and precise sampling of sediments as the gouge corer method however, in view of the aims of this study this method provided the necessary sedimentary data for determination of general grain size distribution (% gravel, sand, silt and clay) of the wetland sediments.

Each soil sample was analyzed for soil particle size distribution (% gravel, sand, silt and clay) and soil organic matter content. Samples were prepared for analysis by oven drying each air dried sample at 105°C for 1 hour. Samples were then cooled in a desiccator to prevent absorption of atmospheric moisture. Approximately 40 g of sediment used to analyze particle size distribution. The gravel fraction (% by mass) was determined by sieving each 40 g sample through a 2 mm diameter sieve and weighing the retained fraction of sediment. The % sand, silt and clay (% by mass) of the remaining sediments was determined using the pipette method described by Briggs (1977). Samples were classified (e.g. clay loam, sandy clay loam) according to the USDA soil texture triangle and named accordingly (Gee and Bauder, 1986). Soil organic matter content (% by mass) was determined according to the loss on ignition method described by Heiri *et al.* (2001).

Chapter 5

Results

5.1. General geological and hydro-geomorphologic characteristics of Wakkerstroom Vlei and the immediate surrounding catchment

Wakkerstroom Vlei lies within a broad valley of Karoo shales stretching over a distance of approximately 7.5 km, and is bounded at its head (Figure 5.1, ‘D1’) and its toe (Figure 5.1, ‘D2’) between two relatively narrow Karoo dolerite sills. The dolerite sill at the toe of the system is considerably broader than the sill at the head of the system. Dolerite is not only common within the head and toe regions, but also outcrops for a short distance along the western side of the valley towards the lower-middle reaches of the wetland (Figure 5.1, ‘D3’). Two distinctive hydro-geomorphic wetland types comprise the system: *floodplain* wetland, which forms two relatively small hydro-geomorphic components (< 2 km long each) confined to the upper (Figure 5.1, ‘UF’) and lower (Figure 5.1, ‘LF’) reaches of the system where dolerite outcrops along the valley, and *unchanneled valley-bottom* wetland (Figure 5.1, ‘VB’) which stretches for 5 km along the main body of the system where shale outcrops along both sides of the valley. Wetland tributaries exist where tributary streams enter the wetland from the east and west at several places along the wetland valley, varying in size from small (< 0.5 km long) to large (> 2 km long) tributary streams.

5.2. Downvalley variation in wetland longitudinal and cross-sectional morphology

The Wakkerstroom River arises at an elevation of ~ 1900 masl within the Mpumalanga Highveld and flows southwest for a short distance (~ 18 km) before reaching the wetland (~ 1760 m amsl). The river shows pronounced and abrupt downstream changes in channel form and behaviour that corresponds with downstream variation in lithology. Through D1 and D2, where dolerite outcrops along both sides of the valley, the trunk river is restricted to vertical incision as indicated by the relatively narrow, sub-rectangular to rectangular shape of the channel (Figure 5.2, XS1, XS9 and XS10); as a result the river is confined within a narrow valley (base width ~ 200 m) largely absent of a floodplain and is relatively straight (sinuosity (P), ~ 1.48 and ~ 1.29 respectively).

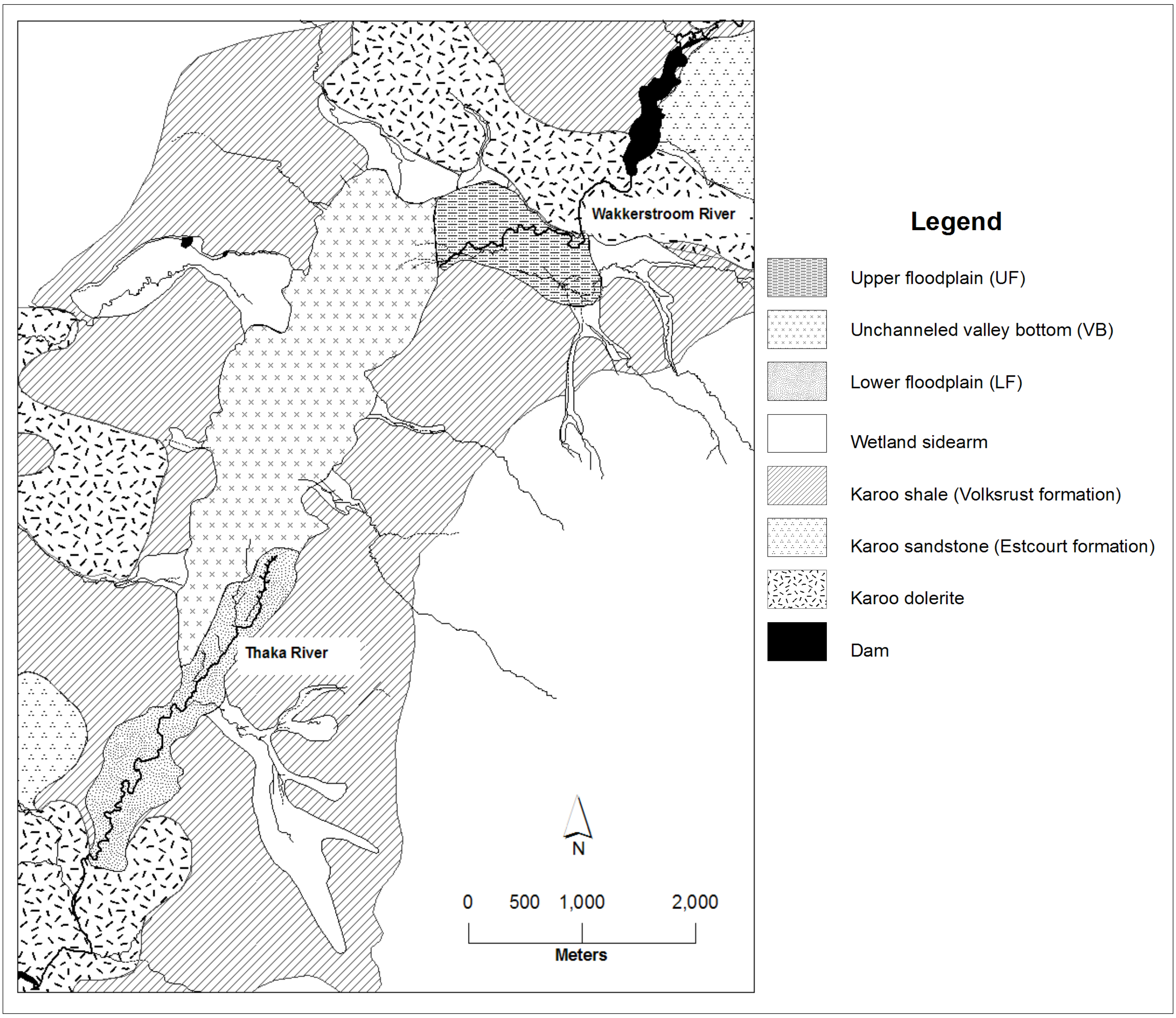


Figure 5.1: General geology (adapted from Begg, 1989) and hydro-geomorphology of Wakkerstroom Vlei showing the location of major Karoo dolerite sills at the head (D1) and toe (D2) regions of the wetland through which the trunk river (Wakkerstroom/Thaka) flows.

Contrastingly, within upstream and downstream valley reaches (UF and LF) where the trunk river becomes superimposed upon softer shale and where dolerite outcrops occur only along one valley side-wall, the valley broadens considerably (base width between ~ 500 – 650 m) and the trunk channel loses confinement, widening considerably (Figure 5.2, XS2). At the same time the river becomes more sinuous (P ~ 1.5), meandering through a floodplain marked with characteristic floodplain features such as oxbow lakes, meander cutoffs and point bars.

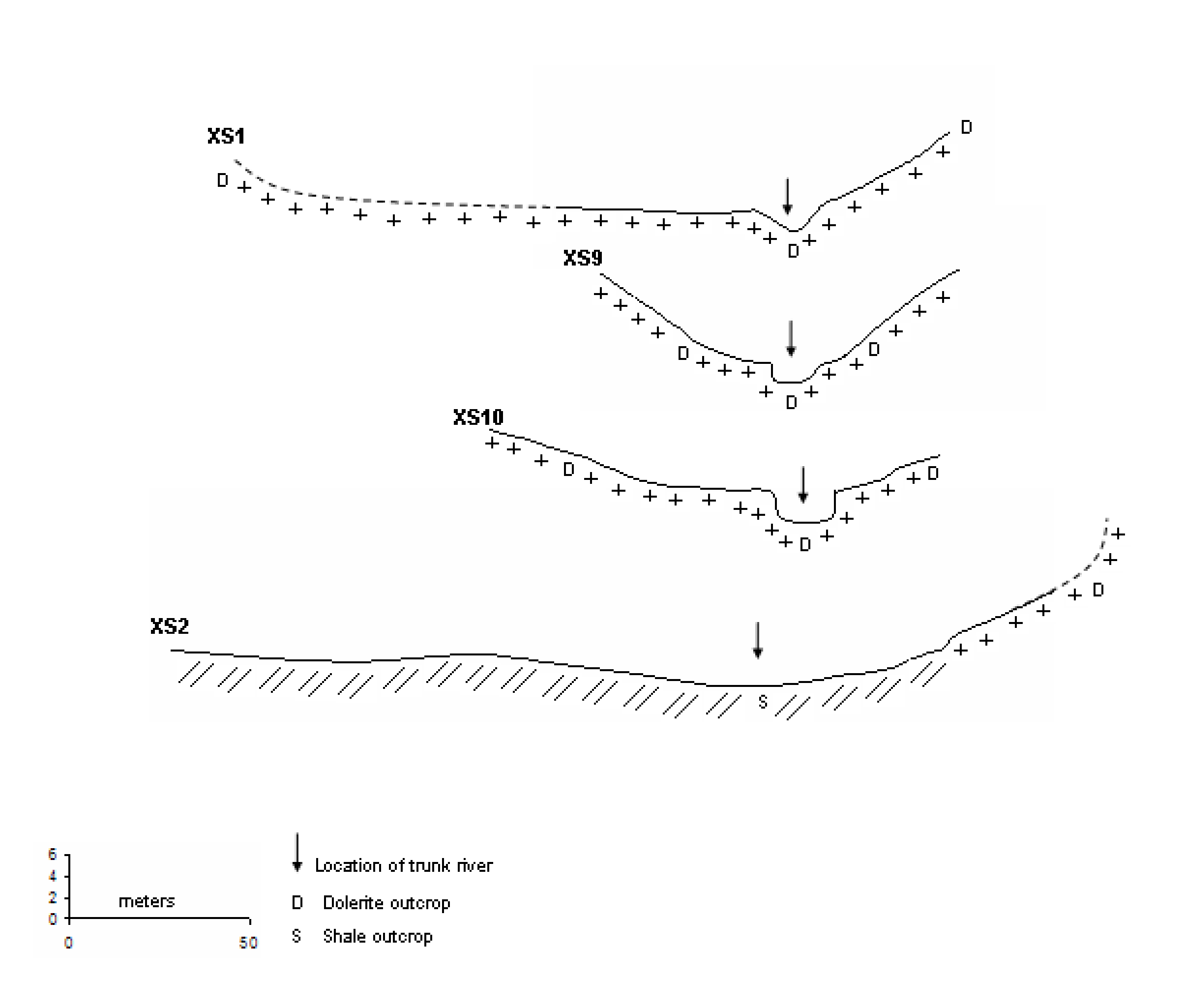
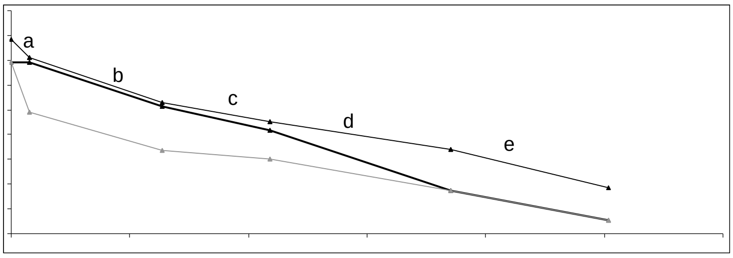


Figure 5.2: Surveyed cross-sectional profiles, viewed looking downstream, of the trunk river (Wakkerstroom and Thaka) through D1 (XS1), D2 (XS9 and 10) and the UF (XS2) of Wakkerstroom Vlei (refer to Fig. 4.1 for location of transects). Underlying lithology is shown as shale (S) and dolerite (D) after Begg (1989).

The UF is orientated with its longest axis from north-west to south-east although the Wakkerstroom (trunk) River crosses it from east to west (Figure 5.1). A steep dolerite bluff extends along the northern margin of the valley. Valley bedrock slope is steep at the head of the wetland at 2.57% (Figure 5.3, reach 'a') through the UF and the trunk river meanders extensively through a relatively narrow strip of floodplain over a distance of about 450 m. The slope on the stream thalweg is close to horizontal at 0.006% at the head of the wetland (Figure 5.3). Within the upper reaches of the floodplain the trunk river is narrow and deep with its bed positioned upon bedrock where it is aligned initially on the northern side-wall of the wetland against the dolerite bluff (Figure 5.4, XS3). The Wakkerstroom River changes from being relatively straight to sinuous as it leaves dolerite (D1) and becomes superimposed on shale (Figure 5.5). To its south the channel has a levee that is elevated approximately 1 m above the floodplain (Figure 5.4, XS3) which has enhanced the inundation of the floodplain backswamp to the south (Figure 5.5). The dynamic nature of the trunk river is indicated by numerous oxbow lakes and abandoned channel sections which characterize the backswamp (Figure 5.5).

		downvalley slope (%)			
	hydro-geomorphic component	**wetland surface**	**thalweg**	**bedrock**	**sinuosity**
a	UF	0.97	0.006	2.57	**1.51**
b	VB	0.32	0.32	0.27	no channel; diffusive flow
c	VB	0.17	0.21	0.08	no channel; diffusive flow
d	VB	0.15	0.32	0.17	mainly diffusive flow; start of headcut gullies
e	upper half of LF	0.24	0.18	0.18	**1.21**



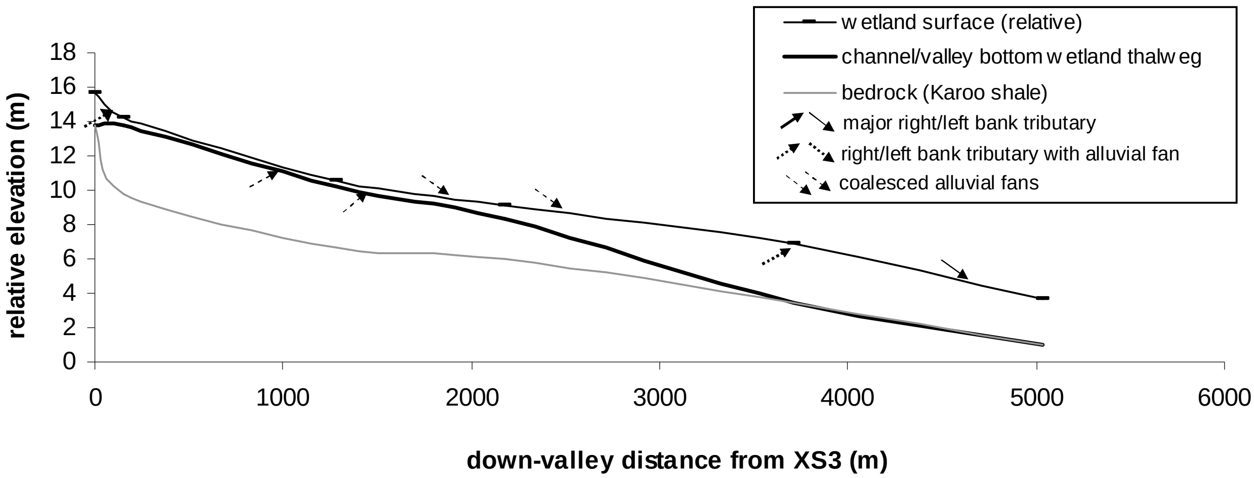


Figure 5.3: Longitudinal profile of Wakkerstroom Vlei between XS3 and XS8, showing variation in slope of the wetland, thalweg and bedrock as well as channel sinuosity where the stream is present.

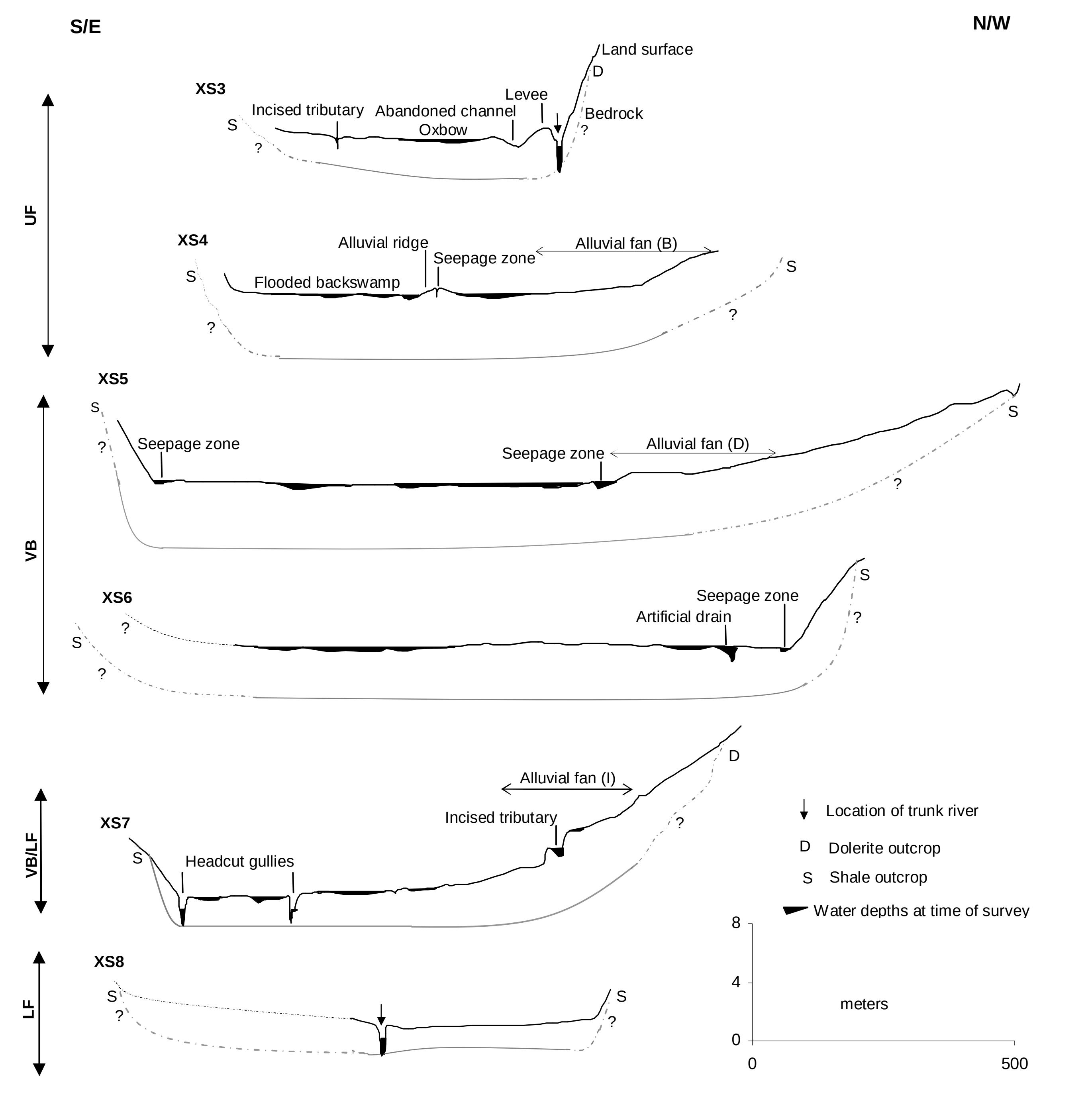


Figure 5.4: Surveyed cross-sectional profiles of Wakkerstroom Vlei arranged downvalley from XS3 to XS8, showing geomorphological detail.

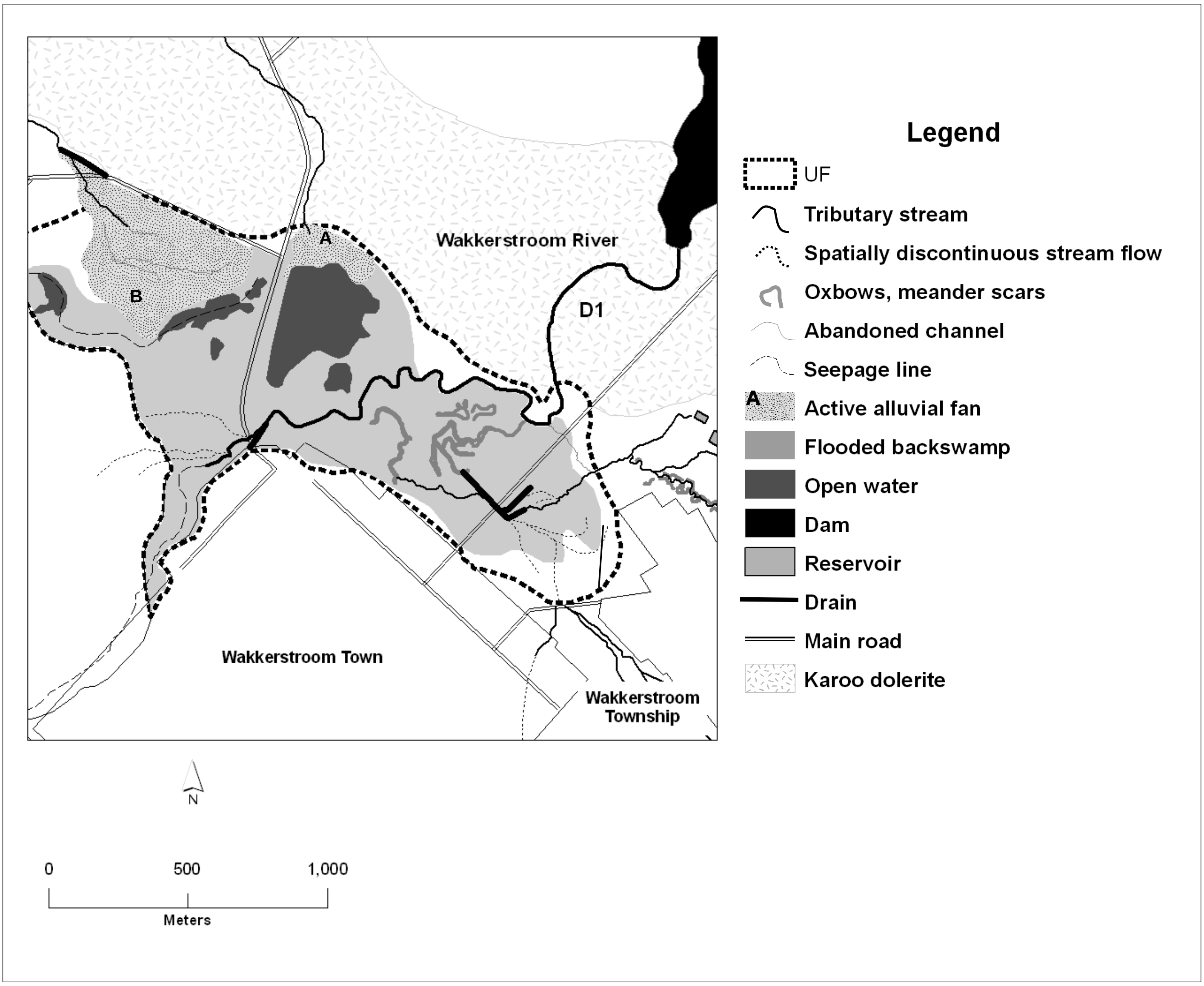


Figure 5.5: Geomorphological characteristics, roads and artificial drains of the upper floodplain hydro-geomorphic component (UF) of Wakkerstroom Vlei.

The strong westward alignment of the trunk channel has been promoted by the presence of a tributary stream which enters from the east side of the valley. There are clear indications that this stream meanders before straightening as it enters the floodplain from the east. However, this stream has been radically altered as a result of impedence of stream flow by a road that crosses the eastern-most side of the floodplain (Figure 5.5). Water has ponded above the road, which is fed back into the floodplain via a drain, but a small stream channel backswamp has cut away from the drain (to its south) thus enhancing perennial inundation of the floodplain (Figure 5.5). Similarly, within the lower reaches of the UF, another road crossing has promoted ponding of a small tributary entering the UF from the north. Across the UF the valley widens (~ 850 m base width) and the trunk channel diverges away from the dolerite bluff (Figure 5.4, XS4). Two alluvial fans have formed within this region where tributary streams lose confinement as they enter the floodplain (Figure 5.5, ‘A’ and ‘B’). A small channel has formed along the distal end of alluvial fan B, which rests atop an alluvial ridge elevated approximately 0.5 m above the floodplain (Figure 5.4, XS4).

Immediately downstream of the UF valley bedrock slope sharply declines to 0.27% (Figure 5.3, reach ‘b’) and the valley widens considerably (~ 1 300 m base width). Along most of the VB shale forms the side-walls on both sides of the valley (Figure 5.4, XS5 and XS6). Valley bedrock slope is gentle averaging 0.17% along the length of the VB, and the trunk river disappears where it enters the VB (Figure 5.6). Diffuse water flow within these reaches is enhanced by the growth of dense stands of the reed *Phragmites australis* and the wetland is extensively inundated throughout the year. Along the middle reaches of the VB where the valley becomes widest (Figure 5.4, XS5), valley bedrock slope becomes near-flat at 0.08% (Figure 5.3, reach ‘c’) however, the wetland slope increases considerably relative to bedrock at 0.17% along these reaches. Several alluvial fans have formed where tributary streams lose confinement as they enter the wetland (Figure 5.6, ‘C’, ‘D’, ‘E’, ‘F’, ‘G’, ‘H’ and ‘I’). These fans extend up to 300 m across the wetland surface and many have coalesced with neighbouring fans. In particular, alluvial fans D, E, and F have coalesced along the western side of the wetland and approximately opposite to alluvial fans G and H, which have coalesced along the eastern side of the wetland. These two alluvial fan complexes form an extensive gently sloping alluvial slope that stretches almost the entire length of the VB. Preferential drainage down the valley occurs along the distal reaches of these alluvial fans resulting in lateral seepage zones on either side of the wetland (Figure 5.4, XS5; Figure 5.6). An artificial drain which extends along the western side of the wetland between alluvial fans E and F has concentrated seepage water and promoted slight incision of wetland fill (Figure 5.4, XS6; Figure 5.6).

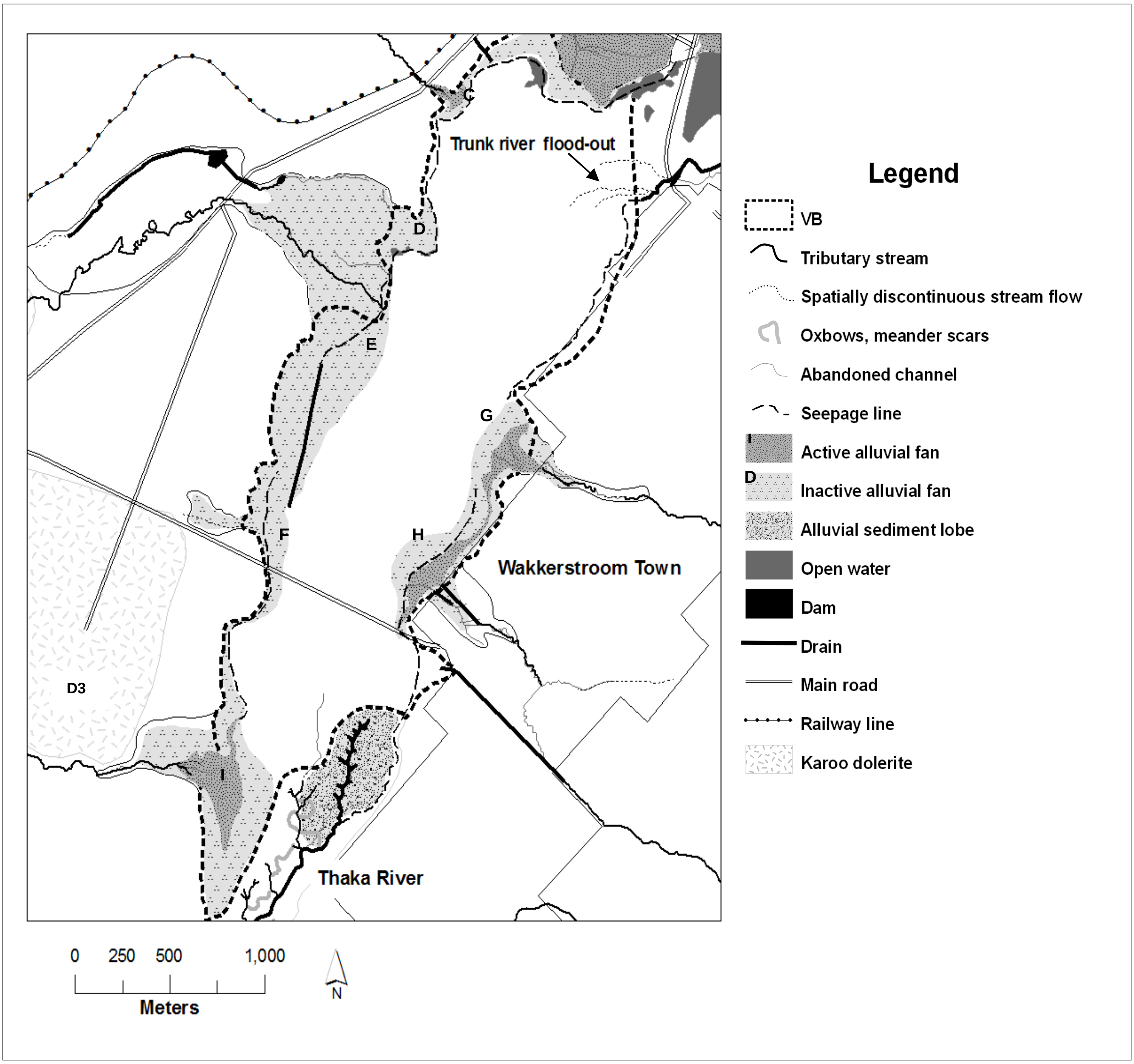


Figure 5.6: Geomorphological characteristics of the middle VB reaches of Wakkerstroom Vlei.

Towards the lower reaches of the VB the wetland valley gets progressively narrower (~ 800 m base-width), particularly where dolerite (Figure 5.6, 'D3') outcrops for a short distance along the western side of the valley (Figure 5.4, XS7). There are no tributaries or alluvial fans that enter the valley from the west where dolerite outcrops in this region of the valley. The wetland thalweg and bedrock slope steepen along this reach but the slope of the wetland remains relatively gentle (Figure 5.3, reach 'd'). Channel formation is reinitiated shortly downstream, along the upper reaches of the LF, where the slope of the wetland steepens considerably (Figure 5.3, reach ‘e’) within the vicinity of a large alluvial sediment lobe (Figure 5.6). Gullies have formed at the head of this sediment lobe and preferentially drain along the eastern side of the wetland opposite alluvial fan I (Figure 5.6). These gullies have incised through wetland fill (Figure 5.4, XS7) and join shortly downstream to form a larger trunk channel, the Thaka River which forms an integral part of the LF (Figure 5.7).

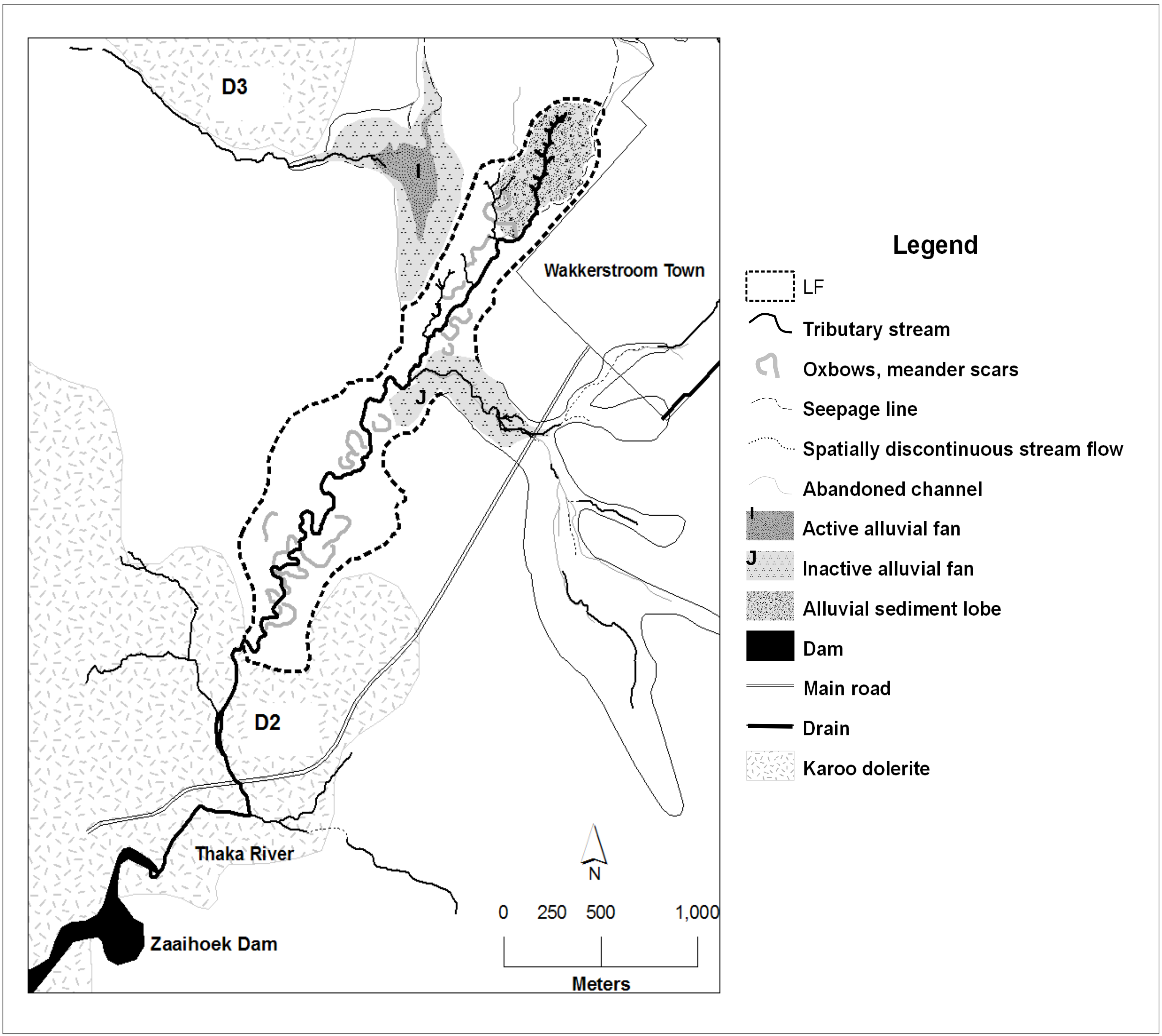


Figure 5.7: Geomorphological characteristics of the LF reaches of Wakkerstroom Vlei.

Within the upper reaches of the LF the floodplain is narrow (width of ~ 200 m) where two alluvial fans (I to the west and J to the east) occur in sub-opposite positions, contributing to constriction of the LF in this region (Figure 5.7). The trunk channel is straight (P ~ 1.15) through the reach of the floodplain where it is relatively confined between these two alluvial deposits, but downstream where shale once again outcrops on both sides of the valley, the floodplain widens (~ 600 m) and the trunk river becomes more sinuous (P ~ 1.71). In this region the dynamic nature of the meandering river is indicated by the density of abandoned meander belts (Figure 5.7). The river has deeply incised the floodplain fill through these reaches and is positioned on bedrock (Figure 5.4, XS8). This together with the lack of adjacent channel levees suggests that the trunk river has recently begun to incise through the LF, the floodplain is thus rarely inundated by overbank flows. The trunk river becomes increasingly more tortuous in meander style towards its junction with D2, where it abruptly straightens (P ~ 1.29) as it becomes confined within a deep (3 – 4 m) narrow (< 100 m wide) valley where dolerite forms both valley side-walls.

5.3. Recent (last 70 years) changes in the geomorphological characteristics of Wakkerstroom Vlei

Analysis of aerial photographic sequences from 1938 - 2003 reveals that the trunk river has remained stable over the last 70 years with no major channel relocations or adjustments in the upper and lower hydro-geomorphic components (UF and LF respectively). Within the UF alluvial fan B maintained much the same extent despite damming of its tributary stream after 1978 (Figure 5.8).

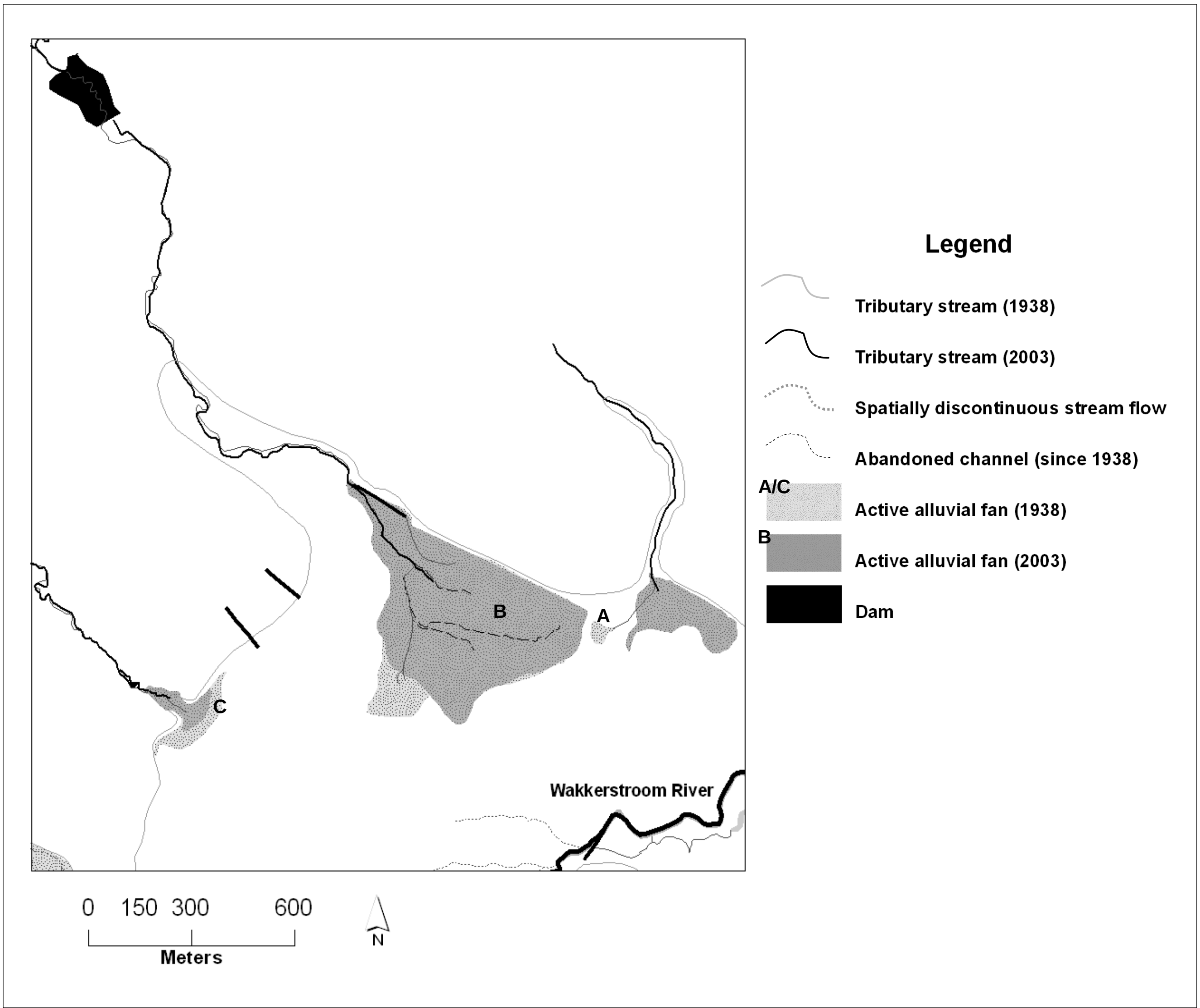


Figure 5.8: Recent (1938 - 2003) geomorphological change within the upper reaches of Wakkerstroom Vlei as observed from aerial photography.

The tributary stream of this alluvial fan has remained highly unstable avulsing several times eastwards across the fan between 1938 and 2003, leaving a series of abandoned channels on the fan surface. Considerable changes in alluvial fan activity have occurred along the VB where coalescent fans D, E and F have ceased activity and activity on alluvial fans G, H and I has decreased considerably in overall extent (Figure 5.9). Most of these changes were accompanied by decreased channel activity across the surface of alluvial fans after 1978 following the damming and draining of a number of tributary streams shortly upstream of the wetland (Figure 5.9). The cessation of alluvial fan E however seems to be linked rather with the natural entrenchment of the tributary stream across the fan surface prior to 1978.

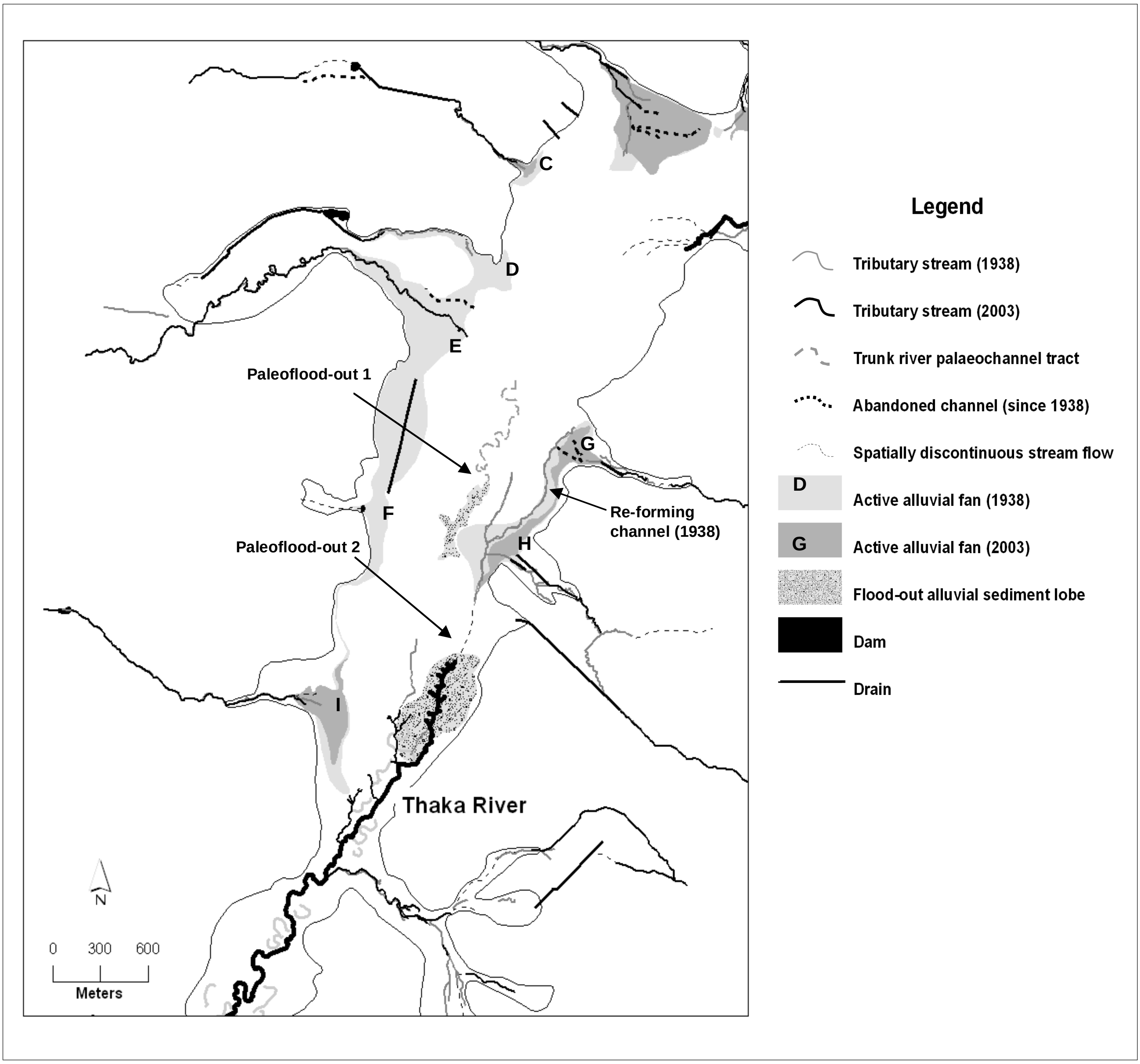


Figure 5.9: Recent (1938-2003) changes in the geomorphological characteristics of Wakkerstroom Vlei along the middle VB reaches of the wetland as observed from aerial photography.

The 1938 aerial photography reveals an abandoned tract of the trunk river (Figure 5.9, 'trunk river paleochannel tract') that terminates in a flood-out abruptly upstream of its confluence with coalescent fans G and H (Figure 5.9, 'paleoflood-out 1'). Although this flood-out is clearly depicted within the 1938 aerial photography, it has subsequently been covered over by the dense growth of *P. australis* along the VB reaches of the wetland. Also evident in 1938 aerial photography is the existence of a re-forming channel tract (Figure 5.9, 're-forming channel') along the flood-out zone which flowed for a short distance before flooding out once again and forming the large alluvial sediment lobe evident along the upper reaches of the LF (Figure 5.9, 'paleoflood-out 2'). This second flood-out sediment lobe is considerably larger than that of paleoflood-out 1 due to the absence of the confining influence of impinging alluvial fans along the lower reaches of the VB component of the wetland. Several headcut gullies have formed within the vicinity of this second paleoflood-out sediment lobe and have maintained much the same form, showing no overall advance or retreat between 1938 and 2003 (Figure 5.9).

5.4. Characteristics of wetland sedimentary fill

Cores of valley fill sediments taken to bedrock within the UF shows that wetland fill is shallow (< 2.5 m) and consists largely of upwards fining deposits which overlie a basal layer of bedrock scour deposits consisting of shale fragments with some coarse sand (Figure 5.10). These sediments typically comprise sandy clay loam overlying the gravel and sand on bedrock, and grade upwards to clay and organic clay loam. Such sequences are indicative of a meandering river floodplain backswamp setting. In oxbow lake settings the stratigraphy is more complex, with alternating lenses of material of varying grade, with the uppermost layer being fine-grained (Figure 5.10, cores 2 and 3). Such deposits indicate alternating high and low energy flow regimes that would typify overbank deposits in close proximity to a meandering river channel. Throughout the cores there is an upwards increase in organic content with typical values of between 5 and 10 % organic matter towards the wetland surface. Bedrock is fairly flat varying by about 0.5 m over a distance of approximately 400 m, rising steeply to the adjacent upland dolerite terrain. Cores taken within the LF reveal that once again bedrock is planar and covered by a thin basal layer of shale bedrock fragments of gravel grade (Figure 5.11). Once again upward-fining backswamp deposits are present, as are deposits typical of meander channel belts (Figure 5.11), although the latter are confined to a relatively narrow belt along the trunk channel. Organic matter content in general displays a similar increasing upwards trend. However, a localized sharp peak in organic content occurs shortly west of the trunk channel at a depth of approximately 0.1 – 0.3 m (Figure 5.11, core 2). This accumulation has taken place within a permanently flooded former oxbow lake (refer to Figure 4.1 for location of XS8, LF).

Valley fill along the VB of Wakkerstroom Vlei thickens considerably (up to 4.5 m) and overlies a broad near-planar bedrock surface covered by a thin, basal layer made up of shale clasts of gravel grade material (Figures 5.12, 5.13, 5.14, 5.15). Cores taken within the upper, middle and lower reaches of the VB reveal general upward fining sedimentary sequences consisting mostly of thick, uniform clay deposits with or without organic material present, which indicates low energy wetland environments characteristic of diffusive flow conditions. These fine-grained deposits are extensive in the central, lowest lying regions of the wetland where they commonly overlie basal upwards fining sequences comprising coarser, sandy loam to sandy clay loam deposits. In contrast, towards the edge of the wetland, particularly the western edge, fine-grained valley bottom sequences alternate with sequences comprising thin layers of material of varying grade (Figures 5.12 – 5.15). These deposits typically lie on alluvial fans that slope into the wetland from the west, representing alluvial fan deposits. These alluvial fan deposits often extend further across the wetland valley than the extent of current/recently active alluvial fans and in some cases underlie buried meander channel belt and floodplain sedimentary sequences (e.g. Figure 5.12, core 3). Organic matter content increases towards the surface of the wetland and is highest along the broader middle reaches of the VB, where peat deposits can reach a thickness of up to 2 m (Figures 5.13 – 5.15). This indicates a progressive decline in flow energy as well as deeper, permanent flooding of the wetland, allowing for the accumulation of organic matter and the formation of peat.

In general, the bedrock slopes away from areas where alluvial fans impinge laterally on the valley-bottom wetland. This is especially true for the larger alluvial fans (Figure 5.12, Alluvial fan B; Figure 5.13, Alluvial fan D; Figure 5.15, Alluvial fan I). This suggests that bedrock may have been protected to a large degree from bedrock scour and planation along the trunk river valley by the sediments of the alluvial fans.

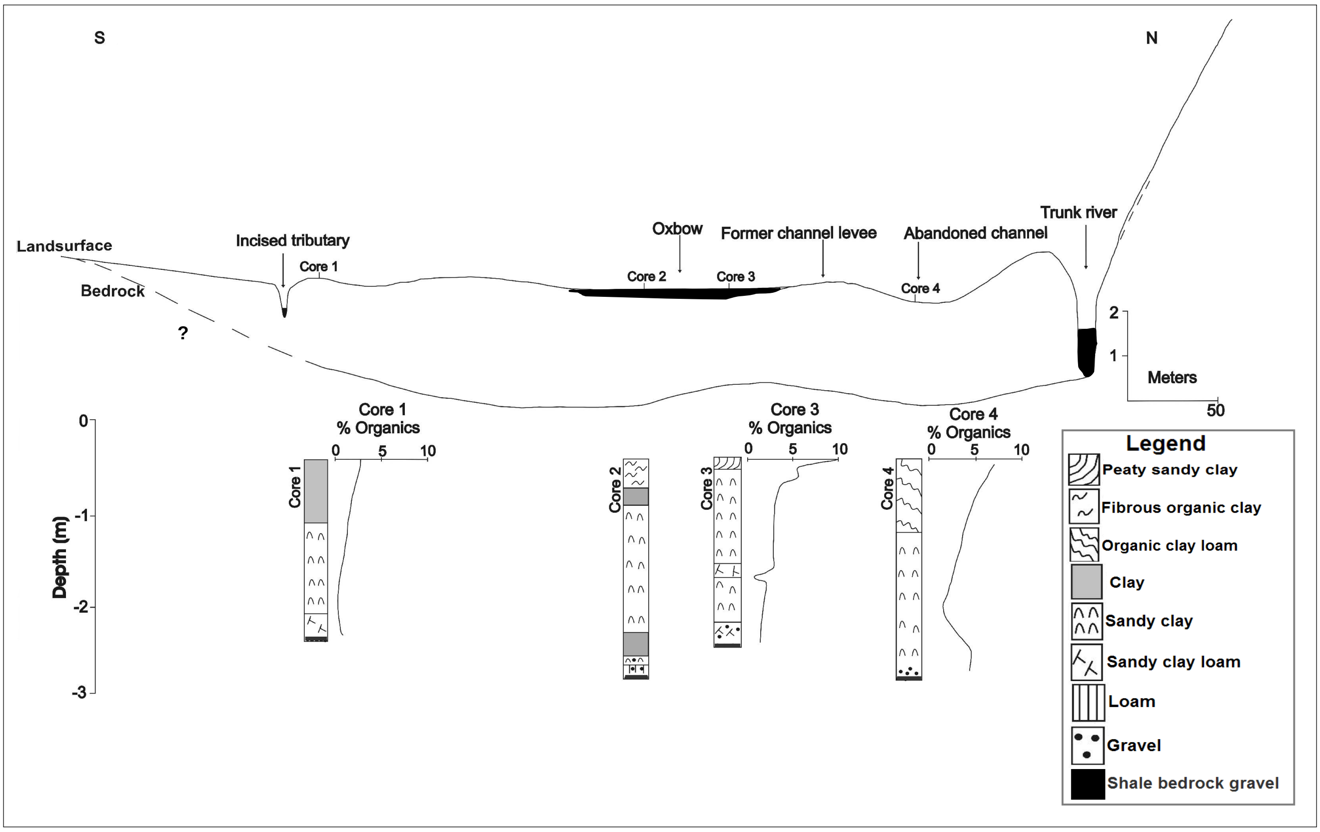


Figure 5.10: Sediment cores taken in the UF of Wakkerstroom Vlei along surveyed cross-valley profile XS3 indicating stratigraphy of the wetland valley fill.

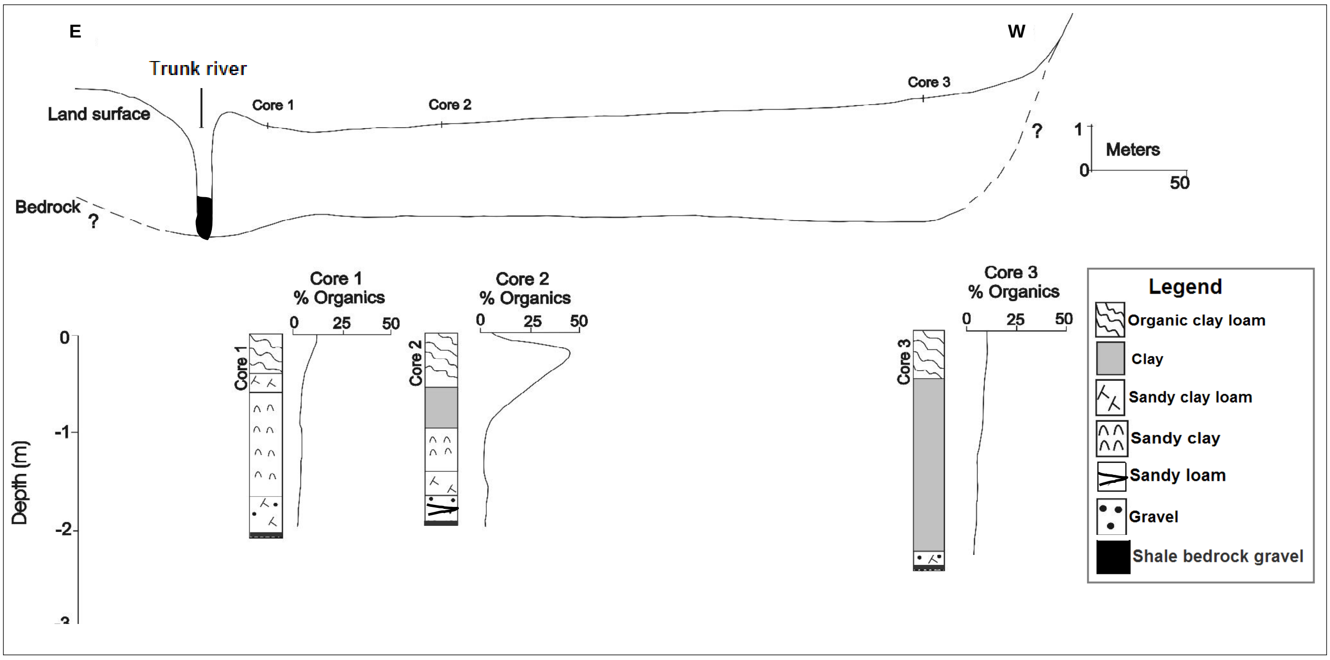


Figure 5.11: Sediment cores taken within the LF of Wakkerstroom Vlei along surveyed cross-valley profile XS8 indicating stratigraphy of the wetland valley fill.

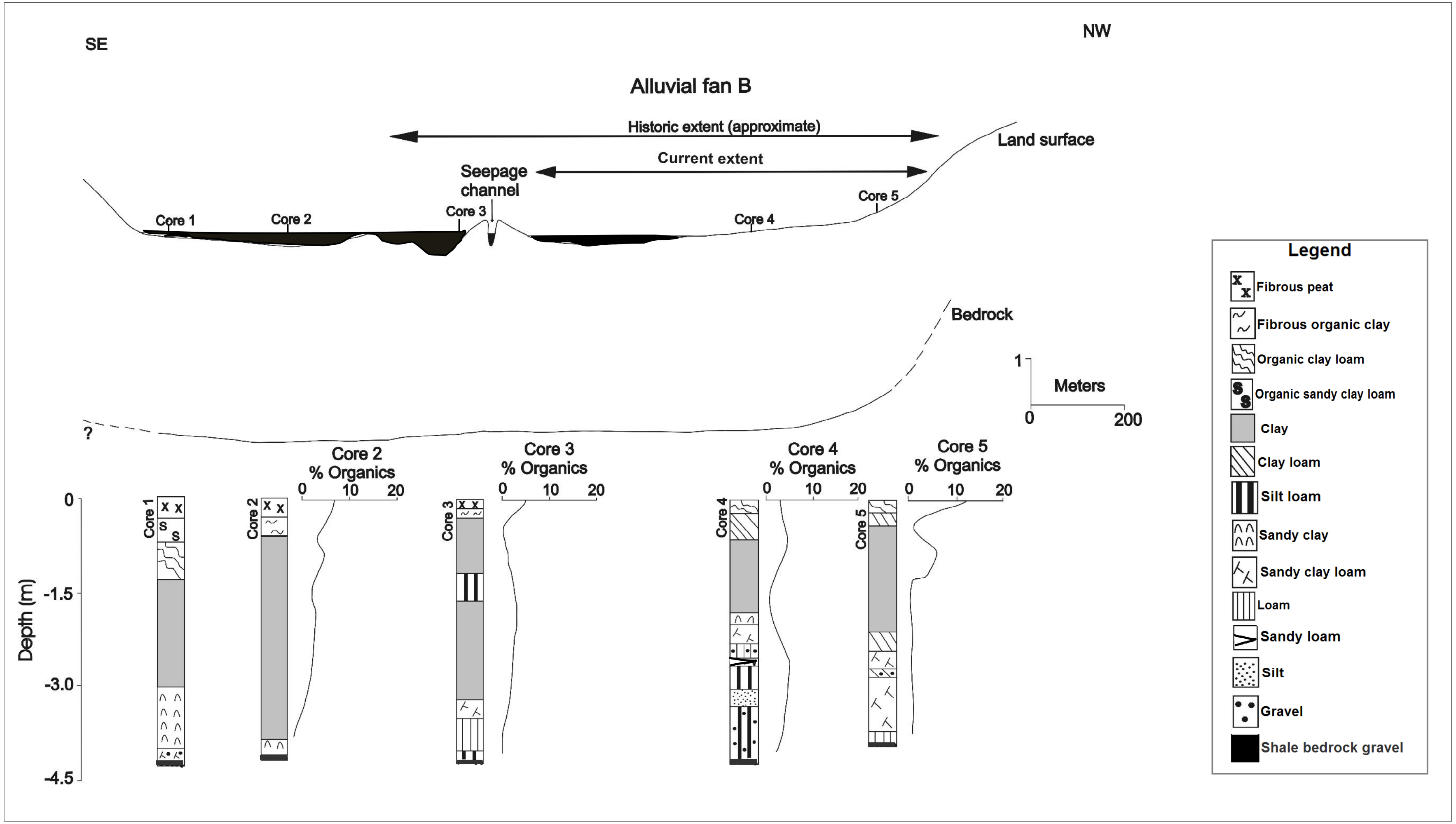


Figure 5.12: Sediment cores taken within the VB of Wakkerstroom Vlei along surveyed cross-valley profile XS4 indicating stratigraphy of the wetland valley fill.

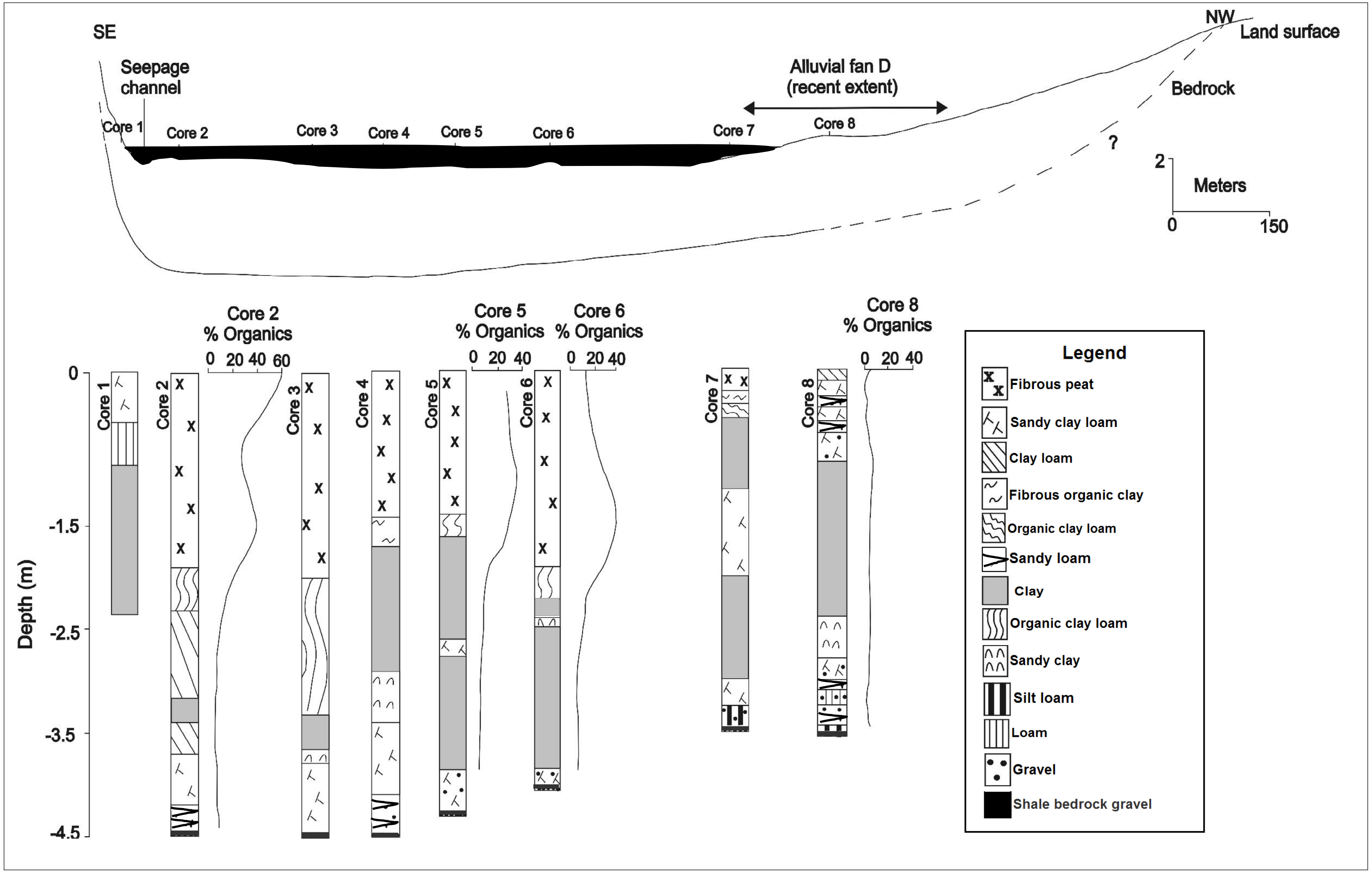


Figure 5.13: Sediment cores taken within the VB of Wakkerstroom Vlei along surveyed cross-valley profile XS5 indicating stratigraphy of the wetland valley fill.

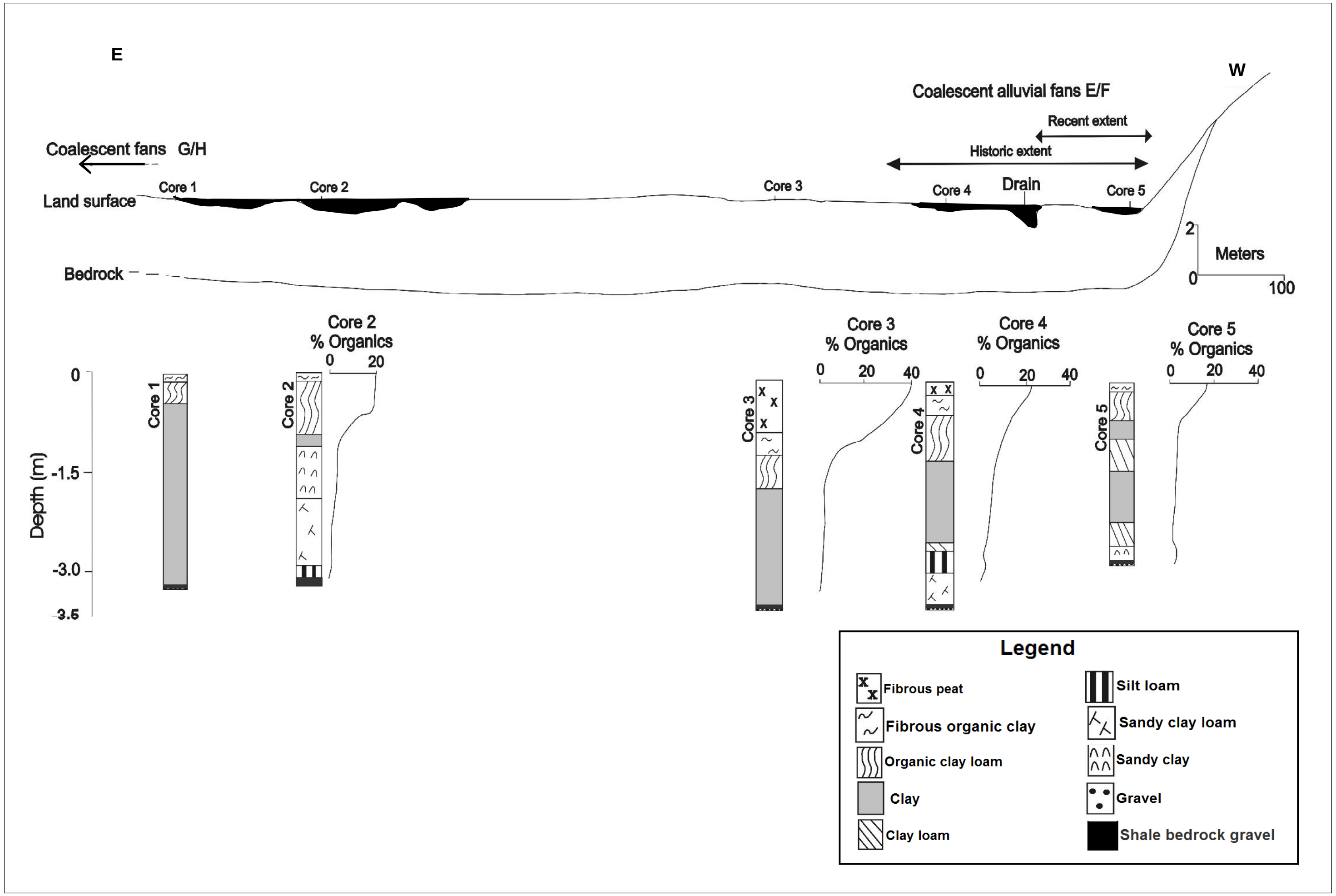


Figure 5.14: Sediment cores taken along the lower-middle reaches of the VB of Wakkerstroom Vlei along surveyed cross-valley profile XS6 indicating stratigraphy of the wetland valley fill.

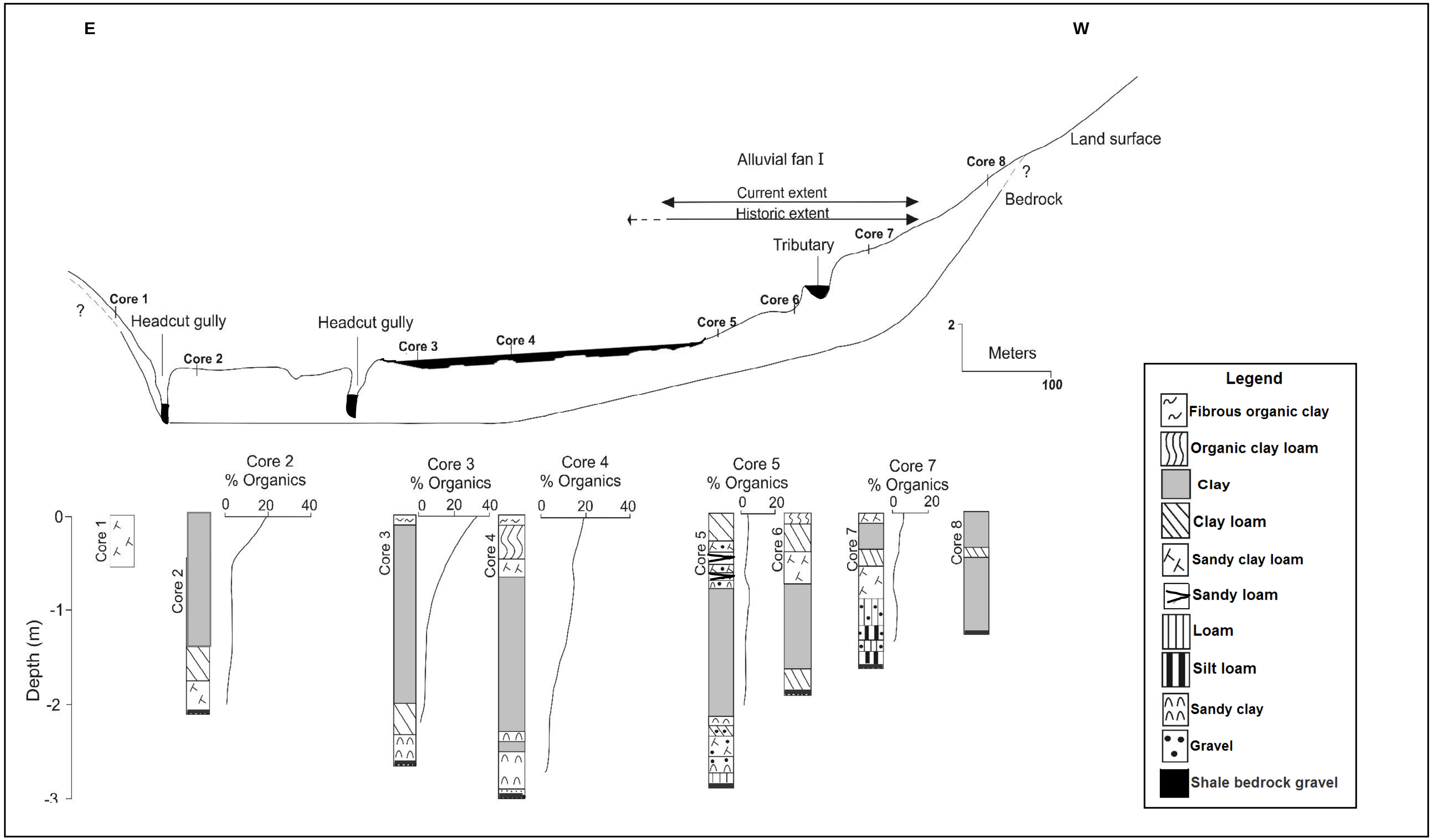


Figure 5.15: Sediment cores taken at the toe of the VB of Wakkerstroom Vlei along surveyed cross-valley profile XS7 indicating stratigraphy of the wetland valley fill.

Chapter 6

Discussion

6.1. Geological controls on trunk river (Wakkerstroom/Thaka) behaviour

Wakkerstroom Vlei displays evidence of the impact of dolerite control on wetland development similar to the dolerite controlled floodplain wetland systems described by Tooth *et al.* (2002a, 2004). The long-term inherited incisional energy of a South African Highveld river from Miocene and Pliocene uplift events has resulted in superimposition of the trunk river (Wakkerstroom/Thaka) upon Karoo sediments and a dolerite sill at the head and the toe of the wetland. The erosional resistance of dolerite over long time-scales (tens of thousands of years, Tooth *et al.,* 2002a), means that it is able to exert control on the behaviour of the trunk river by limiting the rate of incision along the valley where soft, easily weathering and eroding Karoo rocks are extensive.

Where the river flows over resistant dolerite it is straight and valleys are narrow. In these situations wetlands are absent, as is the case where the trunk river (Wakkerstroom/Thaka) flows through D1 and D2 respectively. Contrastingly, where the river flows over softer shale, the valley widens and the stream becomes more sinuous as it meanders through wider and gently sloping valley reaches as is evident along upper floodplain and lower floodplain hydro-geomorphic components of the wetland. The lateral-migration dynamic of the river along these valley reaches has been indirectly controlled by impeded incision of the trunk river by the downstream dolerite sill, which acts as a local base-level, limiting vertical erosion upstream. The river upstream of the dolerite sill thus achieves grade and expends excess energy by adjusting to a more sinuous channel form and laterally eroding the valley, on softer sediment creating conditions suitable for floodplain formation in which a meandering river is well developed (Figure 6.1).

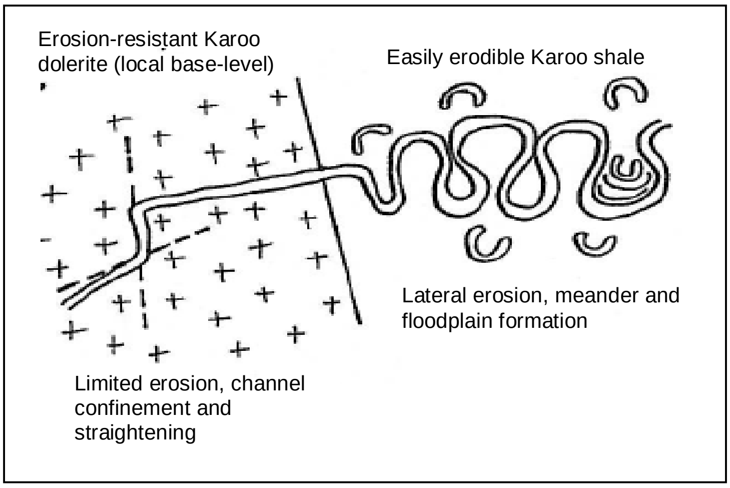


Figure 6.1: The model of dolerite control on inland floodplain wetland formation showing the control of an erosion resistant dolerite local base-level on the formation of alluvial meanders along upstream softer shale valley reaches (after Tooth *et al.,* 2002a).

This process thus represents a river equilibrium response which is currently evident along the upper floodplain and lower floodplain hydro-geomorphic components of Wakkerstroom Vlei: the meandering trunk channel is positioned upon bedrock and has left a thin basal layer of gravel consisting of shale bedrock fragments where it has shifted across the valley floor. Valley fill has been maintained as a thin layer over bedrock, comprising mostly lateral accretion floodplain deposits resulting from reworking of floodplain alluvium by a frequently migrating river channel. Similarly, along the valley-bottom hydro-geomorphic component reaches of the system, the widespread occurrence of a basal layer of shale fragments provides evidence of historic bedrock planation by a laterally migrating river channel.

Dolerite is however able to directly influence this vertical and lateral-migration dynamic where it crosses a valley in an area where softer rock extends across the valley upstream. As a result along D1 and D2 the river has created a narrower and steeper valley than along the shale reaches of the system which are similar in morphology to the shale/sandstone valley reaches described elsewhere on the Highveld by Tooth *et al.* (2002a, 2004) (Table 6.1).

Table 6.1: Morphologic comparisons between Wakkerstroom Vlei and the floodplain systems along easily erodible shale/sandstone valley reaches described for wetlands elsewhere on the Highveld by Tooth *et al.* (2002a, 2004).

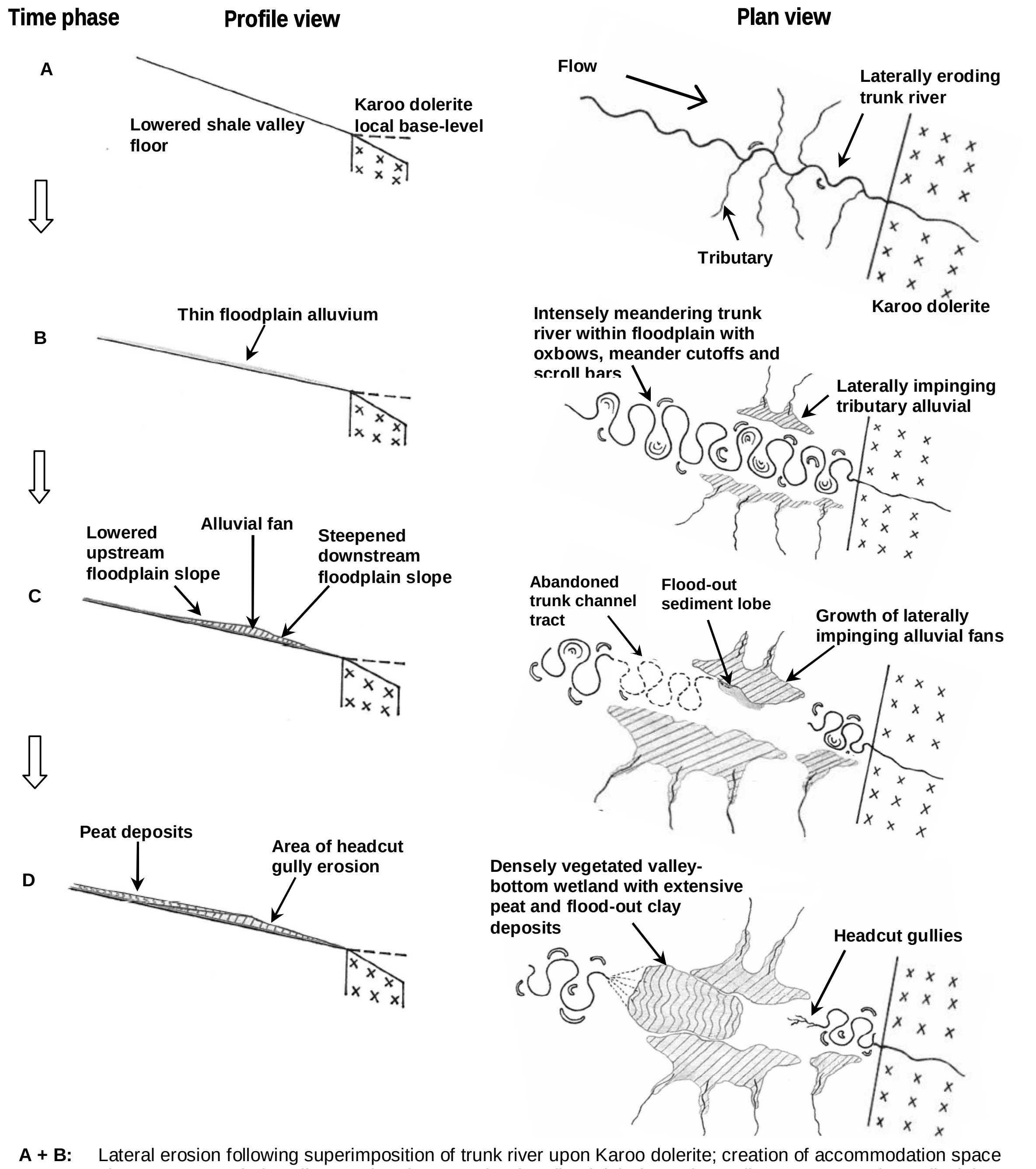
	Wakkerstroom Vlei	**Klip, Venterspruit, Schoonspruit**
Valley width	up to ~ 1 300 m	up to ~ 1 500 m
Average valley slope	< 0.17 %	< 0.11 %
Depth of alluvial fill	< 4.5 m	< 4 m

However, in contrast to the extensive floodplain wetlands and meandering channel tracts that characterize the broad, gently sloping shale/sandstone reaches of the wetland systems described by Tooth *et al.* (2002a, 2004), within Wakkerstroom Vlei the trunk river floods out along the valley-bottom hydro-geomorphic component. This results in an extensive diffusive and deeply inundated valley-bottom wetland environment characterised by deep and extensive peat accumulation. Evidence of continuous tracts of meandering river and floodplain environment along the valley-bottom hydro-geomorphic component is apparent in the wetland alluvial fill by the common occurrence of buried meander channel belt and flooded backswamp floodplain sedimentary sequences comprising upwards fining sequences of primarily sandy loam deposits (Figures 5.12 – 5.15). The high sinuosity of the trunk river paleochannel tract (Figure 5.9, 'trunk river paleochannel tract') evident along these valley reaches in aerial photography further suggests the existence of a meandering river floodplain environment. Buried alluvial fan deposits which underlie some of these buried floodplain deposits (Figures 5.12 and 5.15) indicate that tributary streams deposited sediment over extensive areas of the valley floor. Furthermore, it seems that tributary streams and associated alluvial fans have extended much further across the floodplain than currently active alluvial fans such that they impinged upon the behaviour of the trunk river to a greater extent than at present as suggested by alluvial fan sequences below valley-bottom wetland sequences (Figures 5.12 – 5.15). Wakkerstroom Vlei thus shows considerable hydro-geomorphic divergence from the dolerite controlled wetland systems described elsewhere on the Highveld by Tooth *et al.* (2002a, 2004).

6.2. The origin and development of Wakkerstroom Vlei

The origin and development of Wakkerstroom Vlei has been strongly controlled by the slow weathering rate of the dolerite sill at the toe of the wetland which has thus formed a local base-level along the course of the trunk river, such that the origin and early development of Wakkerstroom Vlei would have closely followed the model of floodplain development described by Tooth *et al*. (2002a). As the trunk river became superimposed upon Karoo dolerite lateral erosion was initiated upstream on softer shale valley reaches (Figure 6.2, Time phase 'A'). Over time this lateral-migration dynamic resulted in the creation of wide, gently sloping valley reaches upstream of the dolerite sill, thus providing accommodation for alluvial deposition. As the trunk river began to meander an extensive floodplain wetland would have formed, marked with characteristic floodplain features such as alluvial levees, scroll bars, oxbows and a flooded backswamp environment. Within this environment floodplain alluvium would have been maintained as a relatively thin layer over the valley floor by the continual shifting of the river channel and reworking of floodplain sediments. At the same time, tributary stream drainage would have been essential in influencing the lateral-migration dynamic of the trunk river. In this regard, the power of the trunk river and thus its ability to laterally migrate across the valley would have been greatly enhanced along the middle reaches of the system by inputs of water and sediment from several tributary streams adjoining the trunk channel from either side of the valley. As a result meander migration and bedrock bevelling would have been intensified along these reaches of the valley as indicated by the valley widening considerably and bedrock longitudinal slope becomes near-flat towards the middle reaches of the valley-bottom hydro-geomorphic component. This created a broad, basinal-type setting upon which these relatively short and steep tributary streams carrying relatively high sediment load deposited their load in the form of alluvial fans across the floodplain surface (Figure 6.2, Time phase 'B'). The deposition of sediment in these alluvial fans resulted in disconnection of the tributary streams from the trunk channel.

With time alluvial fans began to prograde across and down the floodplain, coalescing with adjacent downstream alluvial fans and thus began impounding the trunk river valley, reducing upstream wetland longitudinal slope and thus further promoting losses in stream power along the broad floodplain valley reaches (Figure 6.2, Time phase ‘C’).



A + B: Lateral erosion following superimposition of trunk river upon Karoo dolerite; creation of accommodation space along upstream shale valley reaches for meander river floodplain formation; Tributary streams form alluvial fans upon planed valley and floodplain surface.

C + D: Trunk river valley impoundment and flood-out formation; shift to diffusive valley-bottom wetland environment with extensive peat deposits.

Figure 6.2: Schematic illustration of the origin and sequential development of Wakkerstroom Vlei.

This meant that the trunk river was not powerful enough to respond to slope reduction caused by impinging larger coalescent alluvial fans and was thus forced to flood-out and deposit its load headwards along the channel forming a sediment lobe and channel plug, further promoting upstream decreases in channel-longitudinal slope and losses in stream power. As a result, the trunk river lost channel confinement along the broad, gently sloping floodplain valley reaches and maintained flow only along narrower and steeper dolerite influenced valley reaches within the uppermost and lowermost regions (upper floodplain and lower floodplain hydro-geomorphic components respectively) of the system (Figure 6.2, Time phase 'C').

Downstream steepening along this main node of sediment deposition resulting from lateral alluvial fan deposition and trunk channel aggradation resulted in headcut gullies and a tract of reforming trunk channel which are evident in 1938 aerial photography. This reforming channel tract forms a second much larger flood-out sediment lobe (paleoflood-out 2) shortly downstream where the confining influence of laterally impinging alluvial fans is lost. Similarly, along this second flood-out zone, downstream steepening has promoted the formation of gullies which drain into the Thaka River along the lower floodplain hydro-geomorphic component of the wetland. As indicated by examination of historical aerial photographs, gully form along the flood-out zones of the wetland waxes and wanes. These changes in gully form however, are not considerable and in general do not appear to be in response to short-term (several years) variation in rainfall. For example, gully form appears to be waxing in 1938 aerial photography (refer to Figure 5.9) during a relatively dry rainfall period over the study area (refer to Figure 2.4).

Vegetation would have begun to grow in greater profusion above the main paleotrunk river flood-out zone gradually shifting to more robust reed species such as *Phragmites australis*. This would have promoted rapid and widespread dispersion of water from the trunk river where it enters along the broad, flood-out wetland reaches, promoting increased inundation of the floodplain. Furthermore, widespread diffuse flow of this water across the floodplain would have induced deposition of suspended load fine silts and clays and promoted deeper and more prolonged flooding of the floodplain. This, together with dense vegetation growth created accommodation space for organic sedimentation and peat formation upstream of and along the flood-out reaches of the system. The paleochannel flood-out and reforming channel tract and flood-out were thus covered by clastic and organic sediment.

Although the timing of the trunk river paleoflood-out event and the overall age of Wakkerstroom Vlei cannot be alluded to from the findings of this study, it is clear that trunk river flood-out occurred relatively abruptly within the history of the wetland, and that sedimentation shifted abruptly from lateral and low vertical accumulation (Tooth *et al.*, 2002a) to predominantly vertical accretion. This is suggested by the fact that valley-bottom wetland clay and organic clay deposits along the flood-out reaches of the wetland abruptly overlie floodplain sedimentary sequences. This suggests a relatively abrupt switch from floodplain to diffuse-flow valley-bottom wetland conditions. Furthermore, the overall depth of alluvial fill along the valley-bottom hydro-geomorphic component of the system is approximately 4.0 - 4.5 m, which is greater than expected for a laterally planed and filled floodplain system, such as those systems described by Tooth *et al.* (2002a, 2004) and Grenfell *et al.* (2008, 2009), where alluvial fill is typically less than 3 m. It is suggested here that this is the result of aggradation along the trunk stream upstream of laterally impinging alluvial fans.

6.3. Flood-out formation and hydro-geomorphic variation

Flood-outs have been commonly described along episodic rivers situated within arid to semi-arid regions of Australia (e.g. Tooth, 1999; Tooth, 2000; Gore *et al.*, 2000) and were first described by Tooth (1999) for rivers along which the trunk channel disappears and deposits its load across the floodplain, forming an alluvial surface over which floodwaters spill as sheetflows. These flood-outs have commonly been associated with downstream decreases in discharge, through transmission losses (infiltration and evaporation) and diminished tributary stream inputs along the river valley, together with abrupt decreases in downstream channel bed gradients. The combination of these factors essentially promotes considerable losses in stream power along the trunk channel thus promoting channel break-down and flood-out formation (Tooth, 1999). In South Africa, flood-outs have not been commonly described but have similarly been associated with downstream decreases in discharge (transmission losses) and channel bed gradients. This has been described for the Nyl River floodplain (Tooth *et al.*, 2002b; McCarthy *et al.*, 2004) and for the Blood River floodplain (Tooth and McCarthy, 2007), situated within the semi-arid Highveld and sub-humid eastern coastal foothills of South Africa respectively.

The Blood River floodplain, described by Tooth and McCarthy (2007) (Figure 6.3), closely resembles Wakkerstroom Vlei and possesses the following similar characteristics:

- both meandering channel and flood-out floodplain reaches occur upstream of a Karoo dolerite sill barrier;
- the main (Blood) river floods-out where it enters broad, gently sloping Karoo shale and sandstone valley reaches at the head of the system, together with tributary streams which disappear along the margins of the floodplain, forming in some cases alluvial fans;
- the flood-out zone is also densely vegetated (sedges and grasses) and this has promoted localised clastic and organic sedimentation and steepening in wetland longitudinal slope. This has initiated downstream gully formation and as a result the river reforms and flows for a short distance before flooding out once again, forming a second flood-out sediment lobe;
- within the flood-out zone, abandoned tracts of meandering river provide evidence of the former existence of continuous tracts of meander river floodplain wetland.

Tooth and McCarthy (2007) suggested that the formation of the Blood River floodplain has been fundamentally controlled by the dolerite local base-level at the toe of the system and that following long-term blockage of the Blood River valley by tributary stream sediment inputs together with associated downstream decreases in discharge, flood-out formation within the upstream parts of the floodplain and breakdown of the former through-going channel resulted.

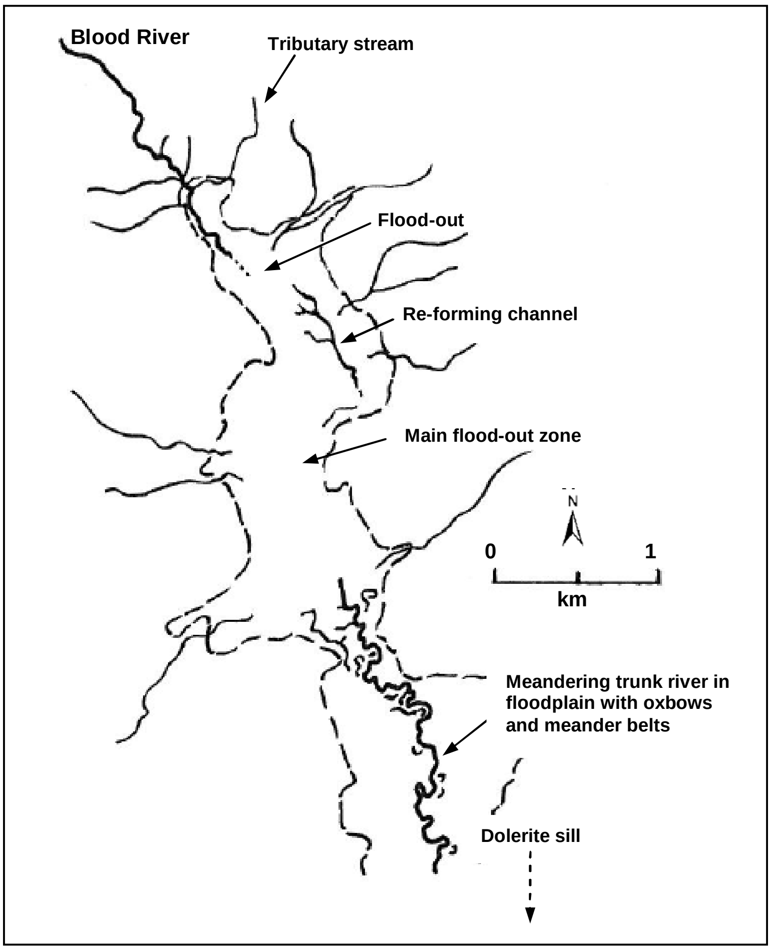


Figure 6.3: The Blood River floodplain within the upper flood-out floodplain reaches of the wetland (from Tooth and McCarthy, 2007).

Within Wakkerstroom Vlei this has largely been the case and tributary-trunk interactions have been key in driving flood-out formation and hydro-geomophic change within the wetland. The formation of alluvial fans and consequent trunk valley impoundment have principally been controlled by dolerite local base-level control on the behaviour of the trunk river. A similar but contrasting observation was made by Grenfell *et al.* (2008) in his study on Stillerust Vlei in the KwaZulu-Natal Drakensberg foothills. Within Stillerust Vlei, dolerite controlled lateral and low vertical accretion along the trunk river floodplain resulted in tributary stream valley impoundment and a consequent hydro-geomorphic shift from meander floodplain wetland to unchanneled valley-bottom wetland (refer to Figure 3.3, p.26). The development of this trunk-tributary dynamic is largely dependent upon faster sedimentation along the trunk river valley than along the tributary stream valley (Grenfell *et al.*, 2008). In the case of Wakkerstroom Vlei the opposite dynamic to

Stillerust Vlei is displayed such that, rather than tributary stream valley impoundment by sedimentation along the trunk river valley, within Wakkerstroom Vlei, sedimentation by laterally impinging tributary streams has led to impoundment of the trunk river valley. This has resulted in a tributary-by-trunk impoundment scenario whereby the tributary streams modify the behaviour of the trunk river by lowering the longitudinal slope in an upstream direction, drowning the trunk river and enhancing peat formation through development of diffusive and prolonged flooding conditions. The establishment of this relationship, although initially influenced by dolerite local base-level control on lateral-migration and valley widening along the trunk river valley, is dependent upon relative erosion rates along the trunk river and tributary stream valleys and sediment input onto the valley floor from tributary streams. When tributary streams entering the floodplain at regular intervals from either side of the valley contribute significant amounts of sediment into the trunk valley, the lateral-migration dynamic of the trunk river along the middle reaches of the wetland valley is reduced by trunk river valley impoundment.

The question of what determines the dominance of the trunk or tributary in wetland evolution is unclear. Within Wakkerstroom Vlei, climate is likely to be an important overriding factor in creating this tributary-trunk erosion/sedimentation dynamic. Variation in rainfall over lengthy time-scales of centuries to millennia means that the wetland and surrounding catchment have undergone shifts between phases when catchment degradation and sedimentation along tributary stream alluvial fans dominates wetland evolution (at present), and phases when diminished stream activity and trunk river sedimentation (through meander migration and bedrock planation) dominates wetland evolution. These shifts are displayed in the wetland alluvial fill along the valley-bottom hydro-geomorphic component of Wakkerstroom Vlei where historic alluvial fan sequences associated with catchment degradation inter-finger with finer, valley-bottom wetland sediments associated with trunk river valley sedimentation. During relatively wet rainfall periods, the laterally impinging tributary streams become degradational and supply large amounts of coarse sediment into the wetland valley. Historically this was important in contributing to trunk valley impoundment where alluvial fans were able to contribute significant amounts of sediment into the trunk river valley. Within these longer-term rainfall cycles, shorter-term (decadal) cycles in rainfall over the study area may also have been important in promoting abrupt plugging and flooding-out of the trunk river and shift to diffusive valley-bottom wetland conditions where alluvial fans were able to prograde relatively fast across the floodplain and extend as far as the trunk channel during relatively wet rainfall cycles.

Vegetation and in particular the dense growth of *Phragmites australis* across the flood-out reaches of the system has also been key in promoting and stabilizing valley-bottom wetland conditions where flow is diffusive. This reed has a rigid and upright growth form and is highly competitive, spreading rapidly through thick underground stems (stolons) and is thus able to form dense monotypic colonies (Struyf *et al.*, 2007). It is thus able to form an impeding barrier to water flow, promoting rapid and widespread dispersion of water and inducing widespread sediment deposition. Following flood-out and abandonment of the trunk river, the dense and extensive growth of this plant species across the valley-bottom hydro-geomorphic component has not only promoted the formation of diffusive-flow valley-bottom wetland conditions, but it has also been a key role player in stabilizing this environment by binding freshly deposited sediments. Its extensive root mat system and tall, robust growth prevents the development of erosional channels within the wetland by dispersing water flows during episodic flood-events, promoting sedimentation and binding sediments such that entrainment is difficult. This enhances the creation of a low energy, permanently flooded environment characterised by low clastic sediment inputs, especially towards the middle flood-out reaches of the valley-bottom hydro-geomorphic component, where the valley widens considerably and where valley slope becomes near-flat. This, together with the high above ground biomass of *P. australis* (Struyf *et al.*, 2007) means that considerable accommodation space has been created for organic sedimentation and peat accumulation across the valley-bottom hydro-geomorphic component. This has led to aggradation along the flood-out valley-bottom wetland reaches of Wakkerstroom Vlei which has further promoted downstream slope steepening along the wetland and this has been important in ensuring continued gully erosion and channelized flow along the lower floodplain reaches of the wetland.

6.4. Longer-term wetland dynamics

The long-term (several thousand years) existence of Wakkerstroom Vlei within the landscape is fundamentally dependent upon the persistence of the downstream dolerite sill as an impeding barrier to erosion along the trunk river. Tooth *et al.* (2004) showed how breaching of this barrier initiates knickpoint retreat and incision along the trunk channel leading to floodplain abandonment and desiccation through the development of headcut gullies and extensive erosional dongas within the upstream floodplain wetland reaches. Within Wakkerstroom Vlei, the downstream dolerite barrier appears to be largely intact. Channel straightening and headcut gully formation along the upper reaches of the lower floodplain hydro-geomorphic component are rather associated with steepening along the wetland downstream of the reforming trunk channel flood-out sediment lobe. Along the lower reaches of the lower floodplain hydro-geomorphic component, gullies are absent and the trunk river has retained a meandering channel form. However, gullies are present in the upper part of the lower floodplain hydro-geomorphic component and their headward retreat may be limited by the low slope along the upstream valley-bottom hydro-geomorphic component reaches of the system. In this regard, the relative rates of aggradation and headward erosion within the valley-bottom and lower floodplain hydro-geomorphic components respectively may affect the upstream advancement of erosional gullies in the short-term but aggradation along the valley-bottom hydro-geomorphic component wetland reaches of the system will steepen slope in a downstream direction such that headward erosion is inevitable within the future of the wetland.

However, this short-term erosion/sedimentation dynamic along the lower floodplain and valley-bottom hydro-geomorphic components respectively has been significantly compromised by the abandonment and cessation of activity of many of the alluvial fans along the valley-bottom hydro-geomorphic component of the system due to damming and draining of tributary streams, which has reduced sediment yields along tributary streams and promoted entrenchment of alluvial fans. This will lead to erosion of alluvial fans and with time may induce erosion of the wetland where these incising tributary streams are able to cut across the wetland and promote reformation of a trunk channel along the valley-bottom hydro-geomorphic component reaches. The construction of dams and drains should therefore be discouraged on tributary streams associated with Wakkerstroom Vlei in order to maintain the natural flow and geomorphological regimes along the tributary streams and the main trunk river valley.

6.5. Conclusion

The findings in this study enhance the understanding of the origin and dynamics of inland valley-bottom wetland systems within the South African landscape and reveal the importance of understanding the underlying controls on the formation and dynamics of wetlands in view of making effective wetland conservation and management decisions. Although large inland floodplain wetlands have been described as common features of the South African landscape (Tooth *et al.*, 2002a, 2004), large unchanneled valley-bottom wetlands characterised by extensive and deep organic and peat accumulations are rare features in the landscape and have not been well studied. Headcut gully formation and donga erosion is common to many inland riparian wetlands in South Africa and has thus far been attributed to poor catchment land-use and management. In view of preventing wetland degradation various rehabilitation strategies focused at halting gully/donga erosion have been implemented, especially along the larger, more pristine wetland systems. However, without understanding the underlying geological, geomorphological, climatic and biotic processes that govern the formation and dynamics of these systems and the complex interaction between these various processes, rehabilitation strategies will continue to work against rather than alongside the natural hydro-geomorphologic dynamics of the system. Furthermore, rehabilitation activities implemented within tributary catchments have the potential to induce erosion and degradation along the trunk stream due to reduced sediment input into the trunk stream.

Within Wakkerstroom Vlei headcut gully erosion has thus far been attributed to poor wetland and catchment land-use and management (Kotze *et al.* 1994). As a result several rehabilitation strategies that focus on preventing further incision and advancement of headcut gullies have been implemented within the wetland. These include primarily the construction of gabion weirs across the trunk stream within the lowermost reaches of the wetland as well as erosion control work in eroding tributary alluvial fans. In view of the findings of this study it would be useful to review future planned rehabilitation activities in the Wakkerstroom Vlei, and possibly for other large inland valley-bottom wetland systems in South Africa that have a similar origin, to ensure effective conservation of these rare and valuable wetland systems.

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