THE INFLUENCE OF LANDSCAPE DIS-CONNECTIVITY ON THE STRUCTURE AND FUNCTION OF THE KROM RIVER, EASTERN CAPE, SOUTH AFRICA

A thesis submitted in the fulfilment of the requirements for the Degree of

Master of Science,

Rhodes University

Grahamstown, South Africa

By Shaun McNamara

15 March 2018

ABSTRACT

Given that a broad valley and low longitudinal slope are important pre-requisites for wetland formation in dryland environments, it has been proposed that cut-and-fill cycles are largely responsible for the geomorphic evolution of the Krom River valley-bottom wetlands. Research to support this suggestion has focused extensively on the role of phases of incision. As a result, little is known about where sediment mobilised during phases of incision is being deposited (filling phase). This study aimed to address this question to add to the understanding of how cut-and-fill cycles influence the structure and functioning of the Krom River and its wetlands. This was achieved through a reach-scale appraisal of the degree of longitudinal connectivity of the Krom River. The reach used for this appraisal contained an incised section along which the river channel exists as a large gully, and a section immediately downstream of the gully terminus where the Krom River is un-gullied, and flow is diffuse across most of the width of the valley floor.

Quantification of the masses of sediment eroded and deposited within the selected reach of the Krom River during a single recent (2012) flood event revealed that the degree of longitudinal connectivity in the Krom River is generally low. During the flood, much of the sediment mobilised by the cutting of the Krom River channel was deposited immediately downstream of the gully terminus, forming a large floodout feature. Particle size analyses of core samples taken along the floodout feature showed that the coarsest fraction of previously mobilised sediment was deposited at the head of the floodout, while finer sediment fractions were deposited progressively further downstream. Field surveys revealed that the pattern of deposition within the floodout feature led to localised steepening of the studied reach of the Krom River downstream of the gully terminus. Surveys of the recently eroded gully revealed that following incision, the eroded stream bed had a lower longitudinal gradient than both the pre-erosional land surface and the regional slope of the Krom River.

The results of this study suggest that floodout formation downstream of gullies may promote the transgression of geomorphic thresholds for erosion, such that the development of floodout features leads to likely initiation of new cutting phases in novel locations along the course of the Krom River. They further suggest that the Krom River is capable of intrinsic longitudinal self-recovery through ongoing cut-and-fill cycles. Finally, it would appear that the current cutting phases responsible for the “destruction” of wetlands within the system are part of a

cycle that will lead to prolonged geomorphic stability, such that the system is made more suitable for the long-term re-establishment of wetlands.

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ACKNOWLEDGEMENTS

Thank you to my Mother, whose support has never wavered, and whose attitude towards my research has always reinforced the feeling that my work in wetland science will make a difference. Thank you to my Dad for encouraging me to do my master's degree. I am grateful for your investment into my future. Thank you to my siblings, and to Winston and Bailey, for all the light-hearted moments along the way. These were important.

Thank you to Lexi Gagiano. I thought I tested you with endless conversations about the wonders of peat formation during my honours degree dissertation, but that could not prepare you for the influx of thesis related conversation that were to follow during my masters. Thank you for your patience while I rattled on about "wetland stuff". Thank you for your reassurance during the stressful times, and for loving that I am ultimately a wetland nerd. I appreciate you getting your hands dirty during field work. That was fun.

To Fred Ellery, thank you for everything you have helped me with during this thesis, but most importantly, thank you for transferring to me your passion for understanding wetland and river systems. My love of wetlands and research is a direct result of all the (academic and non-academic) conversation and experiences we have shared over the years. Thank you to Simon Pulley, whose enthusiasm and brilliant mind were an important part of this process. Lastly, thank you to Juliette Lagesse, Pippa Schlegel and Matt Hermon for your field work assistance.

CHAPTER 1: **INTRODUCTION**

1.1 Background

Wetland science has generally had a strong ecological focus, with most research relating to wetlands in wet and temperate regions in the northern hemisphere (Tooth & McCarthy, 2007). This focus is problematic when trying to understand the structure and function of wetlands in dryland environments. This is because models explaining temperate and tropical wetland formation may not apply in dryland systems (Tooth & McCarthy, 2007). The fundamental difference is that dryland environments experience overall negative water balance conditions. Therefore, unlike wet and temperate settings, wetland formation in dryland settings is seemingly unlikely.

Large wetlands in water scarce settings are therefore almost always integrated into fluvial networks as river systems provide a reliable and sustainable supply of water (Ellery et al., 2008). As a result, wetland studies in dryland environments use fluvial geomorphology as a basis for understanding wetland formation and dynamics (Tooth & McCarthy, 2007; Ellery et al., 2008).

With localised exceptions, South Africa can be considered a dryland region (Tooth et al., 2007). Despite this, numerous large wetland systems exist within the country's borders. In order to account for this anomaly several models of wetland formation, which focus on geomorphological drivers, have been developed. These models include:

- The promotion of lateral erosion through meander migration in response to a resistant lithology, such that a wide valley with a low longitudinal slope is created (Tooth et al., 2002; Tooth, 2004).
- The blockage of a tributary stream by sedimentation along a trunk stream, which lowers the longitudinal slope of the tributary river, and promotes sedimentation and wetland formation along the toe of the tributary valley (Grenfell et al., 2010).
- The incursion of derived sediment onto the trunk stream valley floor, which blocks the trunk stream and lowers its longitudinal slope upstream of the tributary trunk confluence (McCarthy et al., 2011; Joubert and Ellery, 2013).
- The in situ weathering of bedrock, which leads to sagging and the creation of shallow basins which have the potential to host wetlands (Edwards et al., 2016).

Recent research in the Krom River by Lagesse (2017) has led to the proposition of an additional model of wetland formation that needs to be considered. Lagesse (2017) suggests that discontinuous gullies along the course of the Krom River, which are part of natural cut-and-fill cycles, serve to widen the valley floor and lower the longitudinal slope over geomorphologic timescales, creating an environment suitable for the formation of palmiet wetlands. This proposed model of wetland formation is supported by the findings of Job (2014) and Silbernagl (2014), who both found natural cut-and-fill cycles to contribute to wetland formation in the Goukou River in the Southern Cape, and in the Featherstone Kloof in the Eastern Cape, respectively.

Cut-and-fill systems have been observed in multiple semi-arid settings in different regions of the world (Patton & Schumm, 1975; Grenfell et al., 2012; Burrough et al., 2015). While certain cut-and-fill dynamics, such as the timing and controls of cut-and-fill phases, vary between systems, a basic pattern is applicable to most of these systems. The basic pattern is that cutting phases involve the formation of discontinuous gullies, while the filling phases take longer than cutting phases and are characterised by the backfilling of gullies and sediment deposition immediately downstream of the gully terminus.

In her proposed model of wetland formation in the Krom River, Lagesse (2017) provides an in-depth assessment of the role of cutting via incisional gullies. However, little attention is given to the downstream deposition and storage of previously eroded sediment (the filling phase). Given the notion that sediment storage in fluvial systems has the ability to alter landform morphology, and can promote landscape change through the transgression of slope or runoff thresholds (Patton & Schumm, 1975; Schumm, 1979; Schumm 1981), it is believed that filling phases need to be addressed if wetland formation in the Krom River is to be fully understood.

The concept of landscape dis-connectivity is regularly used to address the dynamics and implications of sediment storage within fluvial systems (Brierley et al. 2006; Hooke 2003; Fryirs 2013; Wohl & Beckman, 2014), and has also been used to explain factors influencing wetland formation in dryland environments (Grenfell et al., 2008; Grenfell et al., 2009; Grenfell et al., 2010 Grenfell et al., 2012; Ellery et al. 2013; Grenfell et al., 2014). Therefore, this concept was selected as a lens to understand filling phases in the Krom River. The concept of landscape dis-connectivity describes the integrated transfer of sediment across all possible sources to all possible sinks within a landscape (Bracken, et al., 2015). This study focuses on the degree of longitudinal connectivity of the Krom River, which includes tributary-trunk and upstream-downstream interaction of water and sediment.

Connectivity has three process-based components; sediment mobilisation (cutting), sediment transport, and sediment deposition (filling; Bracken, et al., 2015). These processes interact in space and time to influence landscape and landform morphology within a catchment system (Schumm, 1979). An understanding of these interactions is an important part of landscape management as it provides information on how environmental change is likely to manifest itself throughout a system (Hooke, 2003; Brierley, et al., 2006). Therefore, to gain a holistic understanding of how cut-and-fill cycles interact to influence landscape and wetland evolution in the Krom River, the filling phases in Krom River cut-and-fill cycles are examined in conjunction with the assessments of cutting phases provided by Lagesse (2017).

Filling phases in cut-and-fill systems involve the deposition and storage of sediment (Antevs, 1952; Schumm, 1981; Fuller & Marden, 2010). This makes the sediment storage features that form as a result of filling an important consideration. Fryirs, et al. (2007a) divides the forms of fluvial sediment storage zones into buffers, barriers and blankets. Buffers are landforms that prevent hillslope sediment from entering a channel network. An example of a buffer feature is an elevated floodplain that limits the transport of hillslope sediment across its surface (Bracken, et al., 2015). Barriers disrupt the movement of sediment along a channel. Over-widened channels can act as barriers if they reduce the transport capacity of the channel, therefore preventing the movement of sediment along the stream course (Fryirs, et al., 2007a; Fryirs, 2013). Blanket features smother channel or floodplain surfaces, reducing the accessibility of these covered surfaces to processes of detachment and transport. Bed armouring is an example of a blanket feature (Fryirs, et al., 2007).

It is believed that the degree of upstream-downstream connectivity in the Krom River is generally low as a result of extended low flow periods. Episodic flood events are believed to mobilise large quantities of sediment, but the rapid return of persistent low flow conditions promotes the deposition and storage of large quantities of sediment at the toe of large gullies. Therefore, wide depositional reaches, known as floodouts, form downstream of gullies. These floodouts cover much of the Krom River valley floor and are thought to act as barrier features, promoting future sediment deposition during extended low flow periods. This processes is comparable to the formation of floodouts downstream of gullies in the semi-arid Karoo (Grenfell et al., 2012). However, due to over steepening as a result of sediment storage, the floodouts along the Krom River are thought to eventually undergo erosion, and the cut-and-fill cycle is repeated.

1.2 Aim and objectives

1.2.1 Aim

The aim of this study is to investigate the degree of longitudinal dis-connectivity of the Krom River by determining where sediment, mobilised by the cutting of gullies, is deposited, in order to add to the understanding of how cut-and-fill cycles influence the structure and functioning of the Krom River and its wetlands.

A reach-scale analysis of the degree of longitudinal dis-connectivity will be undertaken in order to get a better understanding of the processes and dynamics of cutting and filling in the Krom River. The information gained from reach-scale observations will be used to understand the degree of Krom River dis-connectivity in a broader sense.

1.2.2 Objectives

1. Determine the spatial extent of erosion and deposition along a selected reach of the Krom River.
2. Determine the quantity and particle size distribution of sediment eroded and deposited within a selected reach of the study site.
3. Determine how erosional and depositional phases initiate and interact to influence the structure and functioning of the Krom River system over time, and examine the implications of this for future erosional and depositional cycles within wetland management

CHAPTER 2: **LITERATURE REVIEW**

2.1 The anomaly of wetland formation in southern Africa

Wetlands form at the interface between aquatic and terrestrial environments and are areas that are temporarily or permanently inundated with shallow water (Tooth & McCarthy, 2007; Ellery et al., 2008). For a wetland to form, the inundation of the land surface needs to occur for a sufficient period of time that anaerobic soil conditions are created in the rooting zones of herbaceous plants (Ellery et al., 2008). This requires the presence of surplus water, which needs to accumulate and persist on the land surface for a prolonged period of time. The current southern African climatic conditions and the region's recent tectonic history make this unlikely. This is because the climate of southern Africa is characterised by low rainfall and high evapotranspiration rates, while the region's tectonic history has created an elevated landmass that promotes the rapid runoff of water from the land surface to the ocean.

2.2 Recent geological and tectonic history of South Africa

In general, South Africa is a highly-elevated region. This is due to the presence of a broad elevated plateau that occupies the country's interior. The average elevation of this plateau is 1250 m above mean sea level (amsl). The formation of this plateau is a result of two distinct isostatic uplift events. These are thought to have occurred 20 (Ma) and 5 Ma million years ago (Partridge & Maud. 2000; McCarthy & Rubidge, 2005). The subcontinent rises steeply from the coast, and slopes gently downwards towards the west, as the uplift during the two events was greater in the eastern part of the subcontinent than the western part. The first uplift event raised the elevation of the eastern part of the subcontinent by approximately 250 m and the western part by approximately 150 m (Ellery et al., 2008). The subsequent uplift event raised the eastern and western parts of the subcontinent a further 900 m and 100 m respectively (Ellery et al., 2008).

The steep rise of the plateau from the coast of South Africa to the edge of the escarpment has led to the development of many short but steep rivers along the eastern and southern coasts, and has served to increase the gradient of the larger rivers that drain the region's interior. (Ellery et al., 2008). Added to this, the increase in elevation of the interior of the subcontinent during each uplift event served to lower the base level of rivers draining into the ocean. As a result, each uplift event was associated with the initiation of periods of widespread river incision (McCarthy & Rubidge, 2005). The land surface prior to any isostatic uplift activity and subsequent river incision is referred to as the African Erosion Surface (AES). The land surface following river rejuvenation in response to the uplift event 20 Ma is referred to as the Post African I Erosion Surface (PA 1). Further incision by river systems following the most recent uplift event, 5 Ma, formed the land surface known as the Post African II Erosion Surface (PA II).

2.3 Models of fluvially integrated wetland formation in semi-arid settings

Despite being a water scarce region, and despite the region's steep and elevated topography, there is an abundance of wetland systems present in southern Africa. These range from small hillslope seeps to large inland alluvial fans (Tooth, 2004) and extensive floodplain wetland systems (Tooth et al., 2002; McCarthy et al., 2011). A noticeable trend in South Africa is that most large wetland systems in the region are integrated into the fluvial network as rivers provide a reliable source of water in an otherwise water scarce region (Tooth & McCarthy, 2007).

Wetland science in South Africa has therefore started to focus on fluvial geomorphology as a framework for understanding how and why wetlands form in a setting as unlikely as the southern African subcontinent. This has led to the proposition of a number of models of wetland formation that have been developed through the use of South African case studies, and which may be applied to regions with similar climatic and geological settings. These conceptual models make use of geomorphological concepts, with many of them focusing on processes that lead to valley widening and longitudinal slope reduction (Lagesse, 2017). The most well-known and relevant conceptual models will be briefly discussed.

2.3.1 Wetland formation as a result of lateral erosion upstream of a resistant lithology

Tooth et al. (2002) developed a model of wetland formation to explain the presence of floodplain wetlands in semi-arid settings. This model was initially developed to explain the formation of floodplain wetlands along the upper Kliprivier, and was subsequently expanded on by Tooth et al. (2004) using the Schoonspruit and Venterspruit Rivers as additional case studies. All these rivers are located on the South African Highveld, in the eastern Free State. This model acknowledges the influence of catchment geology on alluvial channels, with a specific focus on the effect of resistant outcrops on local base level.

Tooth et al. (2002; 2004) suggest erosion resistant sills and dykes are able to act as local base levels for streams which otherwise exist over a less resistant lithology. The Klip, Schoonspruit and Venterspruit Rivers are all located upon loosely cemented Karoo Supergroup sandstone and shale, but resistant dolerite intrusions (dykes and sills) appear at various locations along the length of each stream (Tooth et al., 2002; 2004). In these systems, vertical fluvial erosion occurs where streams flow directly over sedimentary rocks. The dolerite intrusions however tend to resist fluvial erosion due to their hardness. This allows dolerite outcrops that cross the stream course to act as local base levels for reaches upstream of such intrusions.

Upstream of where dolerite outcrops exist, the vertical erosion of the less resistant sedimentary rocks is hampered as vertical incision in these locations cannot exceed the rate of the channel bed cutting at the location of the resistant dolerite intrusion (Tooth et al., 2002; 2004). Therefore, with dolerite intrusions controlling the rate of vertical erosion, lateral erosion emerges as the dominant process upstream of dolerite intrusions in the short- to medium- term (decades to tens of thousands of years; Tooth et al., 2002). The lateral erosion of the sandstone and shale valleys occurs through the development of dynamic river meanders that constantly rework valley floor sediment, and gradually create a broad valley with a near planar valley floor and a low longitudinal slope (Tooth et al., 2002).

These conditions are ideal settings to host wetlands. This is because the wide flat valley bottom and the gentle longitudinal slope serve to slow the movement of water through the fluvial system. While this conceptual model of wetland formation was originally developed for the Kliprivier, subsequent research has revealed that similar geologic controls and fluvial processes have, to varying degrees, contributed to the development of the Dartmoor Vlei (Edwards et al., 2016), the Stillhurst Vlei (Grenfell et al., 2008), and the Northington wetlands in the foothills of the Drakensburg mountains (Grenfell et al., 2008).

2.3.2 Tributary-trunk interaction related wetland formation

Other conceptual models focus on the interactions of water and sediment between tributary and trunk rivers as a geomorphic control for wetland formation in dryland settings. Grenfell et al. (2010) place tributary-trunk relationships onto a continuum, such that they exist between two extremes. On one end of the continuum are “tributary dominated” systems, and on the other end are “trunk dominated” systems. A tributary dominated system is a fluvial environment in which tributary derived sedimentation on the trunk river valley floor exceeds trunk river transport capacity to the extent that tributary sediment begins to block the flow along the trunk stream (Grenfell et al., 2010). Trunk dominated systems occur when tributary rivers supply a limited amount of sediment to the trunk river valley, while the trunk stream tends to transport and deposit significant quantities of sediment. In this scenario it is possible that the trunk stream sedimentation at the tributary-trunk confluence completely blocks the tributary stream, such that tributary flow does not meet the trunk stream (Grenfell et al., 2010). In South Africa, both these situations have been linked to the development of wetland environments (Joubert and Ellery, 2013, McCarthy et al., 2011).

2.3.2.1 Wetland formation in tributary dominated systems

The geomorphic origin of the Nyl River floodplain wetland, located in the Limpopo Province of South Africa, provides an example of a wetland system that has formed in a tributary dominated fluvial setting. McCarthy et al. (2011) found that the main control on the development of the Nyl River floodplain wetland was the obstruction of flow along the Nyl River by coarse grained tributary alluvial fans. These alluvial fans developed at the distal end of steep tributaries that have their headwaters in the mountainous terrain to the east of the Nyl River Valley (McCarthy et al., 2011). The tributary alluvial fans extend across the Nyl River in the lower reaches of the main areas of wetland.

This conceptual model of wetland formation suggests that the obstruction of the Nyl River leads to back ponding behind the tributary alluvial fans, such that upstream of each fan location the Nyl River forms shallow depression features. The encroaching alluvial fans also lead to a reduction in upstream longitudinal slope as the fans act as an elevated local base level for upstream reaches, and promote long-term valley bottom sediment accumulation (McCarthy et al., 2011).

2.3.2.2 Wetland formation in trunk dominated systems

Where tributary streams have small catchment areas and a low sediment supply, trunk rivers that have a comparatively high sediment supply have the potential to block their tributaries through the deposition of sediment on their floodplain. The more rapid rate of trunk river floodplain aggradation in the region of the tributary-trunk confluence prevents the tributary river from meeting the trunk stream. The elevated floodplain margin acts as an elevated local base level that lowers the longitudinal slope along the tributary river. These conditions promote sustained flooding at the toe of blocked tributary rivers.

Based on a growing number of case studies that link wetland formation to trunk domination in fluvial systems, the blockage of tributary rivers by adjacent trunk streams has increasingly been considered a driver of wetland formation in dryland settings. Examples of wetland formation in trunk dominated systems include Lake Futululu, on the Mfolozi floodplain margin in northern KwaZulu-Natal (Grenfell et al., 2010), Stillhurst Vlei in the foothills of the Drakensburg Mountain in KwaZulu-Natal (Grenfell et al., 2008), the development of the Kap River blocked valley-lake on the floodplain of the Great Fish River in the Eastern Cape (McNamara, 2015), and the development of blocked valley lakes in the Mkuze floodplain in Maputaland, northern KwaZulu-Natal (Ellery et al., 2012).

2.3.3 Wetland formation due to in situ chemical weathering of bedrock

Edwards et al. (2016) have recently proposed that long-term deep chemical weathering can lead to bedrock volume losses, producing sagging. Edwards et al., (2016) attribute the formation of the Dartmoor Vlei, to the extremely low energy conditions along the Mnyamvubu River in the Kwa-Zulu Natal Midlands, to this process.

The Dartmoor Vlei exists on a dolerite sill and ends against a more resistant dolerite dyke. The differences in hardness of the dolerite sill bedrock and the dolerite dyke initially led to the development of a floodplain wetland. This was achieved through the processes described by Tooth et al. (2002; 2004). However, a transition from a floodplain wetland dominated by the presence of a meandering channel, to an unchannelled valley bottom wetland, took place at some point in the wetlands history.

This transition is linked to the long-term inundation of the dolerite sill, which led to the dissolution of metals in the weathering bedrock (Edwards et al., 2016). The volume losses, which are associated with deep chemical weathering were most severe near the head of the wetland, such that the longitudinal slope along Dartmoor Vlei was reduced (Edwards et al., 2016). Land surface sagging has therefore produced the extremely low energy conditions present in the system today, which, in turn, ensures the constant presence of surplus water.

2.3.4 Cut-and-fill cycles as a mechanism of long-term wetland formation

Based on research in the Krom River, Lagesse (2017) has proposed an additional model of wetland formation in dryland settings. This model proposes that repeated cycles of cutting and filling have led to valley widening and longitudinal slope reduction in the Krom River over geological timescales. According to Lagesse (2017) cutting phases, characterised by intense gully erosion along the Krom River channel, are initiated by localised over steepening of the river. This over steepening occurs where tributary alluvial fans encroach onto the trunk stream valley floor. Once a gully is initiated it propagates upstream, eroding to bedrock in the process (Lagesse, 2017). The gullies that develop are discontinuous, and sediment accumulation, or filling occurs at the gully toe (Lagesse, 2017).

Lagesse (2017) suggests that after an alluvial fan has initiated a period of gully erosion, the fan begins to once again block and gradually steepen the slope of the Krom River until gullying is again initiated. Each cutting phase is suggested to erode a different part of the valley floor and, over extended periods of time, this process widens the Krom River valley and lowers its longitudinal slope (Lagesse, 2017). These conditions promote diffuse flow and create an environment that is suitable for the formation of unchanneled valley bottom wetlands.

Whilst cut-and-fill cycles have not been acknowledged as a model of wetland formation prior to the work of Lagesse (2017), repeated cutting and filling have, in a limited number of case studies, been acknowledged as having a role in the formation and dynamics of valley bottom wetlands in South Africa (Job, 2014; Silbernagel, 2014). Given that this model of wetland formation is novel, an increased understanding of the processes associated with it seem appropriate.

2.4 Cut-and-fill cycles

Cut-and-fill cycles describe phases of incision, generally through gully erosion, which are followed by relatively slower filling phases involving gradual sediment accumulation within previously eroded gullies (Nanson & Croke, 1992; Brierley & Fryirs, 1999). Cut-and-fill cycles have been investigated in a wide range of settings including South America (Bekkaddour et al., 2014), South Africa (Botha et al., 1994), Zambia (Burrough et al., 2015), North America (Antevs, 1952; Patton & Schmm, 1975; Womack & Schumm, 1977), and Australia (Brierley & Fryirs, 1999). A common trend in systems undergoing repeated cutting and filling is the presence of a discontinuous stream network.

Research into cut-and-fill systems has tended to focus on the factors and conditions that control the transition from a filling phase to a cutting phase. Despite significant research, little consensus exists on the topic. Some believe it is entirely extrinsic factors such as climate, vegetation and anthropogenic activity that control cut-and-fill transitions (Antevs, 1952;

Balling & Wells, 1990; Brierley & Fryirs 1999), while others suggest it is intrinsic threshold related factors that ultimately control these processes (Patton & Schumm, 1975; Womack & Schumm, 1977; Schumm, 1979; Botha et al., 1994).

2.4.1 Extrinsic controls

Antevs (1952) attributed the presence of arroyos (gullies) in the semi-arid southwestern USA to the influence of short-term climate variability on vegetative cover. This research suggests that the lack of soil moisture during drought periods reduces vegetative cover on the land surface such that soils become un-cohesive and are prone to cutting via gully erosion. However, during wetter periods, the abundance of vegetation slows runoff along hillslopes and gullies and therefore promotes sediment deposition. Climate driven controls on cut-and-fill transitions have been more recently identified by Bekkaddour et al. (2014), who linked the Quaternary terrace sequences in the Pisco valley in Peru to orbitally driven climate cycles.

Anthropogenic catchment activities have also been cited as having either partial or complete control over cutting and filling phases. Prosser & Slade (1994) acknowledge the impacts of drain construction and livestock farming on the initiation of valley floor gullies as these processes affect the presence vegetation within the system. However, these authors imply that human activities alone are not sufficient to induce gullying, but rather they interact with different climatic regimes to make valley floor incision a likely outcome. Incisional phases along rivers that drain the central plateau of Zimbabwe have been attributed to unregulated commercial agriculture pre-1950, and to increases in population density in communal lands post-1950 (Whitlow 1988; 1990).

2.4.2 Intrinsic controls

A large body of evidence suggests that the transition from filling phases to cutting phases may be a natural process that is controlled by intrinsic factors. Much of this research links the initiation of cutting phases to the transgression of slope or runoff thresholds, resulting in erosion (Womack & Schumm, 1977; Schumm, 1979; Schumm, 1981). Alternatively, Brierley & Fryirs (1999) suggest that in the Wolumla Creek in New South Wales, Australia, it is the location of tributary streams along the trunk river valley that determined if cutting along tributary valleys would be initiated.

Schumm (1973) found that in semi-arid regions, sediment storage along discontinuous river channels causes a gradual increase in valley bottom longitudinal slope gradients which eventually results in slope failure in the form of gullying. He found this process to be the ultimate control on the initiation of cutting phases, and suggested that over steepened stream reaches that develop in this manner exceed a unique type of intrinsic threshold which he referred to as a geomorphic threshold. A geomorphic threshold can be defined as one that is inherent in the manner of landform change, such that changes occur within the system itself; and through time these changes lead to a condition of emerging instability and eventual failure (Schumm, 1973; Patton & Schumm, 1975).

Research into cut-and-fill cycles reveals that individual stream reaches in northwestern Colorado (Womack & Schumm, 1977; Schumm, 1979), and individual hillslopes in northern KwaZulu-Natal (Botha et al., 1994), reacted independently to extrinsic environmental changes. This suggests that local geomorphic thresholds exert a significant control on the development of cut-and-fill cycles. Supporting the argument that phases of cutting and filling are natural intrinsic processes is the presence of filled gullies in multiple settings which pre-date human settlement (Botha et al., 1994; Grenfell et al. 2012; Lagesse 2017)

2.4.3 The role of cut-and-fill cycles in long-term landscape evolution

Although not extensively recognised, some research acknowledges alternating cutting and filling phases as having an influence on long-term landscape adjustments and landscape evolution. For example Schumm (1973), Patton & Schumm, (1975) and Womack & Schumm (1977) highlight how cutting phases lead to downstream sediment storage and slope steepening, which promotes the transgression of geomorphic thresholds such that incision is again initiated. This process is referred to as a "complex response", and is considered a form of natural, long-term slope adjustment through which a fluvial system can achieve a graded state.

Although it has not been widely studied, cut-and-fill cycles have recently been identified as landscape evolution processes that promote wetland formation in dryland environments. Burrough et al. (2015), for example, found that climatically controlled and wind driven cut-and-fill cycles are an ancient and ongoing process that influences the formation and stability of wetlands within the upper Zambezi Valley. Job (2014), Silbernagl, (2014) and Lagesse (2017) have all found cut-and-fill cycles to contribute to wetland formation in semi-arid South Africa, as the cutting phases were shown to lower the longitudinal slope and widen the valleys of the

Goukou River (Job, 2014), Featherstone Kloof (Silbernagl, 2014), and Krom Rivers (Lagesse, 2017).

2.5 The conceptual framework of landscape dis-connectivity

Fluvial systems convey water and sediment from the land surface towards the ocean. The concept of landscape dis-connectivity describes the efficiency and patterns of sediment transfer through a fluvial system. Early definitions describe dis-connectivity as the movement of sediment through different geomorphic zones in a landscape, from local source areas to local sink areas (Harvey, 2002; Hooke, 2003; Fryirs et al., 2007a). Such definitions are based on the fluvial system model proposed by Schumm (1981), which identifies three main zones in a fluvial system, with a single dominant process occurring in each, as follows:

- The catchment landscape, which supplies stream channels with sediment through processes of hillslope erosion.
- The river system in which transport processes dominate as streams move sediment through the catchment.
- A coastal site of sediment accumulation where processes of deposition dominate (Schumm, 1981).

This model implies that zones of sediment mobilisation through erosion, transport and deposition, exist separately from each other, and that landscape development takes place in three distinct zones down the entire length of the stream, with the upper catchment being erosional (erosion dominated), the middle reaches delivering sediment from the upper catchment to the lower catchment (transport dominated), and the lower catchment being depositional (deposition dominated). In reality, sediment can be mobilised, transported and deposited in any landscape zone in a catchment, and these processes can occur simultaneously in different landscape zones. For this reason, Bracken et al. (2015) described landscape dis-connectivity as the integrated transfer of sediment across all possible sources to all possible sinks within a landscape.

Although it is disputed that a single process is dominant in the individual landscape zones described by Schumm (1981), the division of a landscape into localised zones of sediment erosion, transport and deposition is widely agreed upon (Harvey, 2001; Hooke, 2003; Brierley et al., 2006; Fryirs et al., 2007a; Harvey, 2012; Fryirs, 2013; Bracken et al., 2015). As the erosion, transport and deposition of sediment can take place at any location in a catchment system, the division of a landscape into geomorphic zones can occur at much smaller spatial scales than at the catchment scale, as described by Schumm (1981). Because this project is not concerned with catchment scale landscape function and landscape change, but rather has its focus on the reach and sub-reach interactions of sediment, a geomorphic zone in the context of this project refers to reach and sub-reach zones within a fluvial system in which one of the processes (erosion, transport or deposition) dominate. At this scale of fluvial system analysis, geomorphic zone boundaries include hillslope-channel boundaries, tributary-trunk boundaries, floodplain-channel boundaries, and reach-reach boundaries. Of importance to dis-connectivity theory is how efficiently sediment moves through these local geomorphic zones and across the boundaries between them.

2.6 Component processes of landscape dis-connectivity

2.6.1 Sediment detachment processes

Sediment detachment processes make up an important aspect of landscape dis-connectivity. This is because when sediment is detached from the land surface, it is available for downstream transport. Sediment is detached from the land surface through erosional processes, which can be defined as a physical process that removes soil, weathered rock or previously deposited sediment from the land surface, and transports it to another location (Ellery et al., 2008).

Land degradation associated with soil erosion has resulted in this process being considered a critical worldwide environmental issue (Laker, 2004; Le Roux et al., 2008). Research indicates that at varying intensities, at least 70 % of South Africa's land surface has been affected by soil erosion (Le Roux et al., 2007). According to Le Roux et al. (2008), the Eastern Cape is the South African province most severely affected by soil erosion. One third of the land surface in this province (16 million ha) is classified as having at least moderate soil loss rates (12-25 t/ha/yr.), with parts of the province classified as having extremely high soil loss rates (> 150 t/ha/yr; Le Roux et al., 2008).

Water is considered to be the most important soil erosion agent because of the efficiency with which it detaches sediment from the earth's surface (Laker, 2004; Le Roux et al., 2008). The most common water related soil erosion processes are rain splash, sheet wash, rill erosion, and gully erosion (Le Roux et al., 2007). The presence of incisional gullies indicates that water-related erosional processes are dominant in the Krom River.

Ellery et al. (2008) consider a gully to be a well-defined erosional feature, carved into a landscape by water. Gullying can account for significant amounts of the total erosion in a landscape, and are an efficient sediment detachment and transport process (Poesen et al., 2003; Valentin et al., 2005; Poesen, 2011). The development of gullies is a multi-step process, with different stages being responsible for varying amounts of sediment detachment (Figure 2.1). Sidorchuk (2005) divides gully development into two broad phases. The first phase is gully initiation, which makes up approximately 5% of a gullies life span. According to Sidorchuk, (2005), during this phase a gully can reach 90% of its total length and 60% of its total area. The second phase of gully development described by Sidorchuk (2005) is slow gully deepening in its upper reaches, and aggradation in its lower reaches. Schumm (1994) and Ellery et al. (2008) see gully development as following a similar pattern. During the initiation phase, they describe rapid gully widening and deepening (Figure 2.1, A-C); followed by gully bank slumping, gully bed aggradation, and eventual gully stability (Figure 2.1, D-E).

Bull (1997) notes that increasing streamflow and decreasing material resistance combine to produce entrenchment but typically, neither process is dominant.

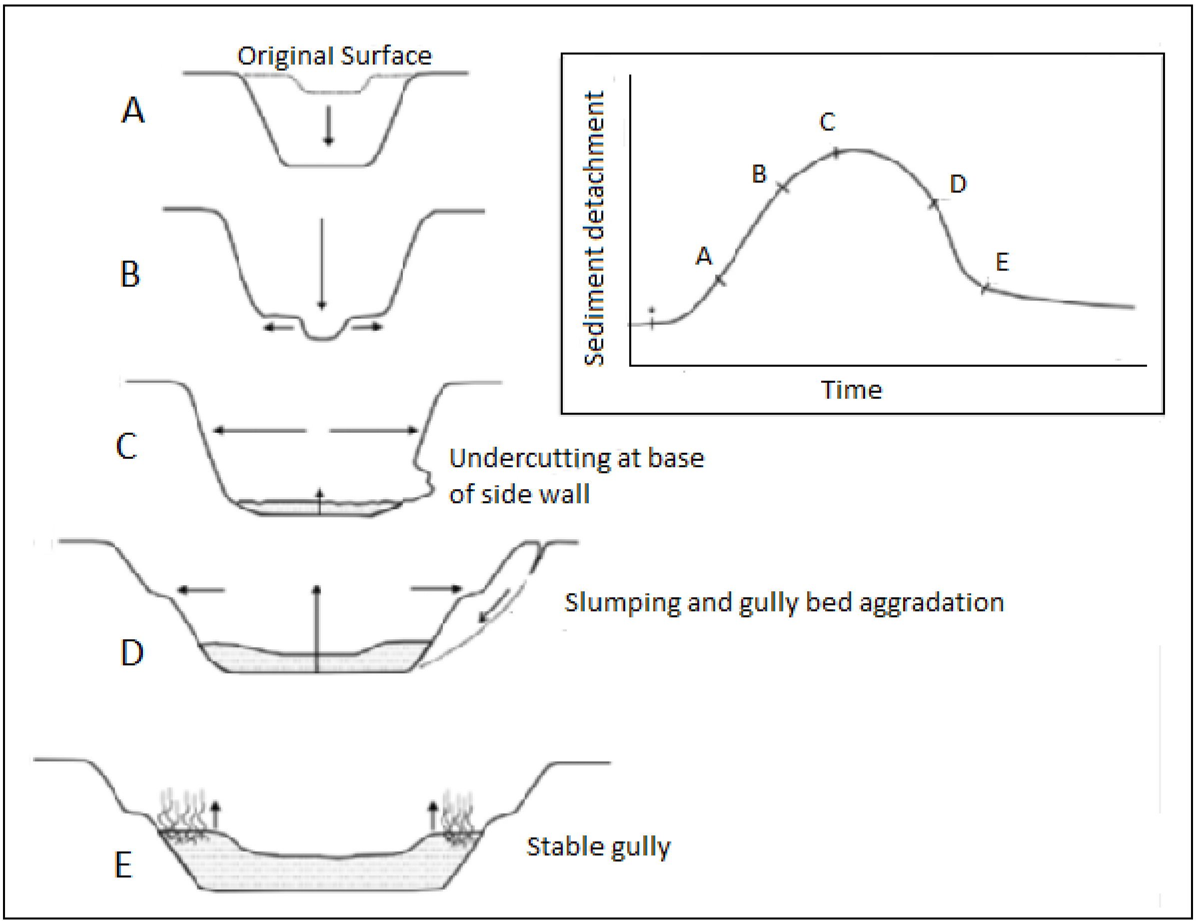

Figure 2.1: *Conceptual model of gully evolution, from initial incision (A-C) to widening (C-D), and eventual gully stability (E). Inset: The curve on the top-right shows changes in sediment detachment during each gully development phase (Adapted from Schumm, 1994 and Ellery et al., 2008)*

Gully initiation processes involve the concentration of runoff. Concentrated runoff flow results in an increase in hydrodynamic power, such that sediment detachment and entrainment occur at accelerated rates (Poesen et al., 2003; Wells et al., 2009). The kinetic energy of moving water initiates erosion and transport processes (Knighton, 1999). Erosion will only take place once a critical level of erosion resistance is reached (Knighton, 1999). Through increases in hydrodynamic power on an over-steepened land surface, or due to the confinement of flow in a gully, it is possible for a head-cut to develop. A head-cut is a steep change in bed surface topography, where localised erosion takes place (Wells et al., 2009). Head-cut erosion occurs in a headward propagating fashion, allowing the gully to extend itself through a landscape while detaching and mobilising significant amounts of sediment (Robinson & Hanson, 2001; Wells et al., 2009). In a fluvial system, where gully erosion has been initiated along a stream, gullies propagate in an upstream direction. Along with the propagating head-cut, bed scour and gully bank collapse contribute to the total amount of sediment detached during the process (Osborn et al., 1989; Robinson & Hanson, 2001; Sidorchuk, 2005; Wells et al., 2009).

In a sediment yield study done on 22 reservoirs in Spain, Verstraeten et al. (2003) showed that, in areas where gullies were rare, the average sediment yield to nearby reservoirs was 0.74 t/ha/yr, while the average sediment yield to reservoirs in areas with many gullies was 9.61 t/ha/yr. Hence, the presence of gullies in a catchment has a clear influence on the amount of sediment transported downstream. In terms of landscape dis-connectivity, gully erosion increases landscape connectivity as they provide a direct link between sources of sediment (head-cuts, bed scour and bank collapse) and the channel network responsible for moving detached sediment further down the system (Poesen et al., 2003; Verstraeten et al., 2003; Wells et al., 2009).

2.6.2 Sediment transport processes

Sediment transport is a size selective process. In her conceptual framework for coarse sediment connectivity in river systems, Hooke (2003) explains that coarse sediment is likely to be transported only during high energy flood periods, and will therefore be transported less frequently, and over shorter distances, compared to fine sediment, which is transported more regularly, under frequently occurring lower magnitude (normal) flow events. Hooke (2003) concludes that differing storage times and transport routes of coarse and fine sediment means that the degree of dis-connectivity in a landscape will vary based on the material available for transport.

2.6.3 Sediment deposition processes

Deposited sediment accumulates in either a sediment store or a sediment sink. A sediment store is a temporary, episodically reworked feature, where the sediment it holds is occasionally re-mobilised for further downstream transport. A sediment sink is a permanent sediment storage feature. The presence of sediment stores and sinks hampers the efficiency of downstream sediment transfer and therefore reduces landscape connectivity. This is because sediment deposited in stores and sinks does not move through the system.

Schumm & Litchty (1965) present three possible timescales for considering the structure and function of geomorphic systems, and explain that the status of certain variables (dependent or independent) changes with the timescale of the investigation. Similarly, when assessing landscape dis-connectivity, the status of a storage feature (store or sink) can depend on the timescale of investigation. During an investigation of landscape dis-connectivity, if a storage feature is not reworked it can incorrectly be considered a sink. However, when assessing landscape dis-connectivity over a longer period, if the same storage feature is reworked it is considered a sediment store. This makes understanding the long-term dynamics of a landscape or fluvial system an important component of effective environmental management.

The breaching capacity of a sediment storage feature describes the threshold conditions required to rework the feature (Fryirs et al., 2007a). The breaching capacity of a storage feature therefore determines whether that feature is a store or sink over a given time period, and influences the timescales of dis-connectivity in a landscape (Harvey, 2002; Fryirs et al., 2007b; Fryirs, 2013; Bracken et al., 2015). The magnitude and frequency of events that exceed the breaching capacity of a sediment store will dictate the timescales of sediment storage and transport, and will therefore influence the timescales of dis-connectivity in a landscape (Harvey, 2002; Fryirs et al., 2007b; Fryirs, 2013; Bracken et al., 2015).

Because rainfall is manifested as flow, the magnitude and frequency of threshold exceeding events is dependent on the amount and timing of precipitation in the catchment being considered (Grenfell et al., 2014). Regional climate therefore influences the patterns of sediment transfer and the length of sediment storage. The geomorphic threshold concept described by Schumm (1973) also exerts an influence on the timescales of sediment mobilisation. This is because steepened slopes make it more likely that flood events will exceed the breaching capacity of a sediment storage feature.

2.7 Definitions and dynamics of landscape linkages

The movement of sediment through a landscape is dependent on the status of longitudinal, lateral and vertical linkages which occur within and between different zones (Brierley et al., 2006; Fryirs et al., 2007a; Fryirs, 2013). These linkages influence the patterns and pathways of sediment movement through a system, and therefore dictate the degree of landscape dis-connectivity.

Longitudinal connections are defined in the context of the channel network and include upstream-downstream and tributary-trunk interactions of water and sediment (Fryirs et al., 2007a). Sediment storage features which inhibit longitudinal connections within a fluvial system are known as barriers (Fryirs et al., 2007a). Barriers often limit the longitudinal interactions of water and sediment through their effect on the local base level of a stream system. The Tooth et al. (2002; 2004) conceptual model of wetland formation provides a useful example of a base level related barrier feature. This is because dolerite intrusions such as sills and dykes, which cross a channel, can act as the local base level for a system, and can inhibit longitudinal linkages of water and sediment. A low slope gradient in the reaches upstream of the local base level is the inevitable result of this channel slope adjustment because a stream naturally adjusts its slope to the local base level in order to achieve a graded state. A low slope gradient results in a loss in stream transport capacity, which could lead to the development of discontinuous channels and within-channel sedimentation (Grenfell et al., 2009; Grenfell et al., 2014). Each of these outcomes limits the movement of sediment along a channel network, thus reducing the degree of longitudinal connectivity in a river system.

However, it is possible for longitudinal connectivity to be affected in situations where local base level adjustments are not the cause of sediment storage. For example, the loss of stream power downstream of gullies, or along wide channels that experience dispersed flow, can lead to sediment deposition and inefficient longitudinal sediment transport. For example, the loss of stream confinement within the Northington Wetland in the KwaZulu-Natal Drakensburg foothills has led to the development of floodout features (Grenfell et al., 2009), which are defined as sand sheet deposits that accumulate on top of an intact valley floor (Brierley & Fryirs, 1999). Floodouts are a common feature of dis-continuous streams in semi-arid settings, and have been shown to influence longitudinal connectivity (Brierley & Fryirs, 1999; Foster et al., 2012; Grenfell et al., 2012; Fryirs, 2013; Grenfell et al., 2014).

Dams and dam-like structures, both natural and artificial, are highly effective barriers. The walls of these structures act as the local base level of system and create backwater areas behind them that reduce flow velocity, such that sediment settles downstream of the dam inlet. Research from various locations all over the world has shown that dams and similar features effectively trap previously mobilised sediment, thus reducing the longitudinal movement of sediment along channel networks (Bednarek, 2001; Butler & Malanson, 2005; Vericat & Batalla, 2006; Fryirs, et al., 2007a; Chen et al., 2008; Xue et al., 2011).

Lateral linkages encompass the relationship between the channel network zones and the wider landscape of a fluvial system (Fryirs et al., 2007a). An example of a lateral linkage is the relationship between a river channel and its floodplain. An example of an enhanced lateral linkage between a channel network and floodplain can be seen in a study done by Wohl &

Beckman (2014) along the North St. Vrain Creek in the Rocky Mountain National Park, Colorado. Here channel spanning logjams have created upstream backwater zones. Deposition of sediment due to reduced transport capacities has resulted in a significant amount of aggradation occurring in these backwater zones. This has raised the elevation of the river bed behind these logjams. The raised river bed has resulted in more regular overbank flow in the reaches of the river behind the logjams, such that water and sediment flow from the channel network onto the floodplain more regularly than they normally would. Such enhanced lateral linkages between two geomorphic zones (channel-floodplain) has led to a reduced linkage between upstream-downstream zones, as the longitudinal transfer of sediment is inhibited by the presence of the channel spanning logjams. Sediment storage features which inhibit lateral linkages within a fluvial system are known as buffers (Fryirs, et al., 2007a). Stable or aggrading alluvial fans are an example of a buffer as they prevent hillslope-derived sediment from reaching the channel network (Harvey, 2001; Harvey, 2002; Harvey, 2012).

Vertical linkages include surface-subsurface interactions of water and sediment (Fryirs et al., 2007a). Any feature disrupting vertical linkages of water and sediment are referred to as blankets. Church et al. (1998) found that the vertical sorting of channel bed sediment, such that coarse sediment is deposited over fine sediment, increased the stability of the Harris Creek in British Columbia. Fine sediment overlain by more coarse sediment is defined as bed armouring. The channel stability in the mentioned example results partly from streamflow not being able overcome the inertia of the coarse surface sediment, therefore it cannot entrain the fine sediment beneath the channel bed surface. Bed armouring is therefore an example of a blanket feature. The presence of coarse sediment on the surface of the channel bed of the Harris Creek, which is acting as a layer of bed armour for the finer sub-surface sediment, is responsible for the disrupted vertical linkages in this system (Church et al., 1998).

2.8 Role of wetland vegetation in landscape dis-connectivity

Wetlands have been shown to have an influence on landscape dis-connectivity. This is because wetlands have a tendency to act as sediment traps, removing suspended sediment from flows as they pass through wetland systems (Chambers et al., 1999; Steiger et al., 2003; Philipp & Field, 2005; Hupp et al., 2008). This is largely due to the high vegetation density in wetlands. Dense wetland vegetation is able to provide resistance to flow, thereby reducing flow energy (Job, 2014). Once sediment has been removed from the streamflow, it can remain stored in wetlands for extended periods of time.

In Jugiong Creek, New South Wales, Australia the development and expansion of in-stream wetlands is considered a natural control on the initiation of filling phases in a system that has experienced intense gully erosion over the past 200 years (Zierholz., 2001). Added to this, the ability of in-stream wetlands in the Jugiong Creek to trap and store sediment, has reduced the degree of longitudinal connectivity in this system, and in doing so, has prevented the discharge of large quantities of sediment to downstream river reaches (Zierholz et al., 2001).

Certain wetland plant species have been shown to be particularly efficient sediment traps. Palmiet (Prionium serratum) is one such wetland plant species (Figure 2.2; Sieben, 2012; Barclay 2016). Valley bottom wetlands dominated by palmiet have been shown to greatly decrease longitudinal connectivity in a number of settings in South Africa, and have been considered a vital agent controlling the structure and function of certain systems (Sieben, 2012; Job; 2014; Lagesse, 2017). Given that the focus of this study is on the filling phase of cut-and-fill cycles in the Krom River, the influence of palmiet (a common wetland plant within the system) on the degree of longitudinal connectivity in the system is an important consideration.

Figure 2.2: *Palmiet plants along the course of the Krom River (Photo: Pippa Schlegal).*

A number of factors influence the efficiency with which palmiet wetlands trap sediment. Firstly, each individual stand of palmiet has web-like leaves that surround the base of the stem.

These web-like leave bases act as a net and efficiently trap sediment being transported through the wetland (Figure 2.3; Barclay, 2016). Secondly, palmiet plants have deep penetrating root systems and very thick (10 cm diameter) palm-like stems that allow them to withstand periods of high flow (Figure 2.3; Sieben, 2012; Job, 2014). They are therefore able to serve as sediment traps during flood events, when sediment loads are generally elevated. Lastly, palmiet wetlands have a tendency to densely colonize open water channels, decreasing flow velocity such that sediment deposition and storage is a likely outcome (Job, 2014).

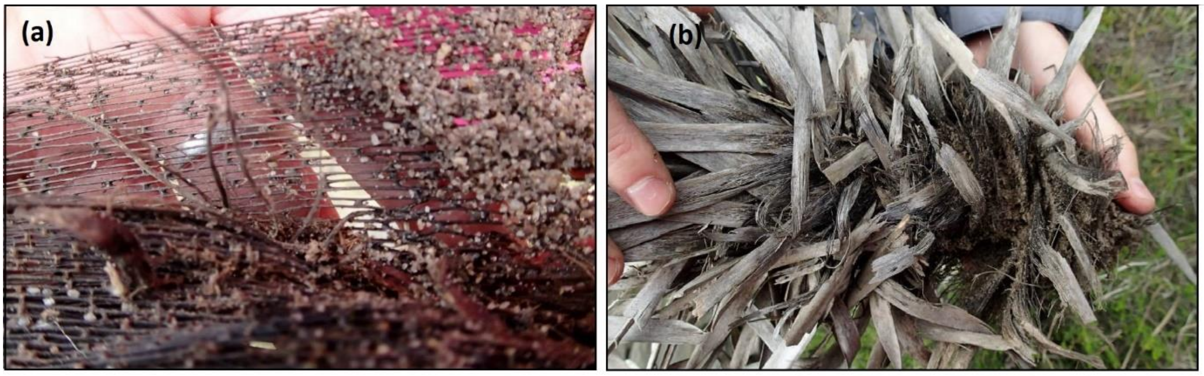

Figure 2.3: *Sediment trapped in web-like leaves that surround the stem of the palmiet plant (a) and an intact palmiet plant stem (b)*

CHAPTER 3: STUDY AREA

3.1 Location

The Krom River is situated in the southwestern part of the Eastern Cape province of South Africa. Its catchment is comprised of quaternary catchments K90A-K90E (Figure 3.1). The study site for this project is along the upper Krom River, within quaternary catchment K90B.

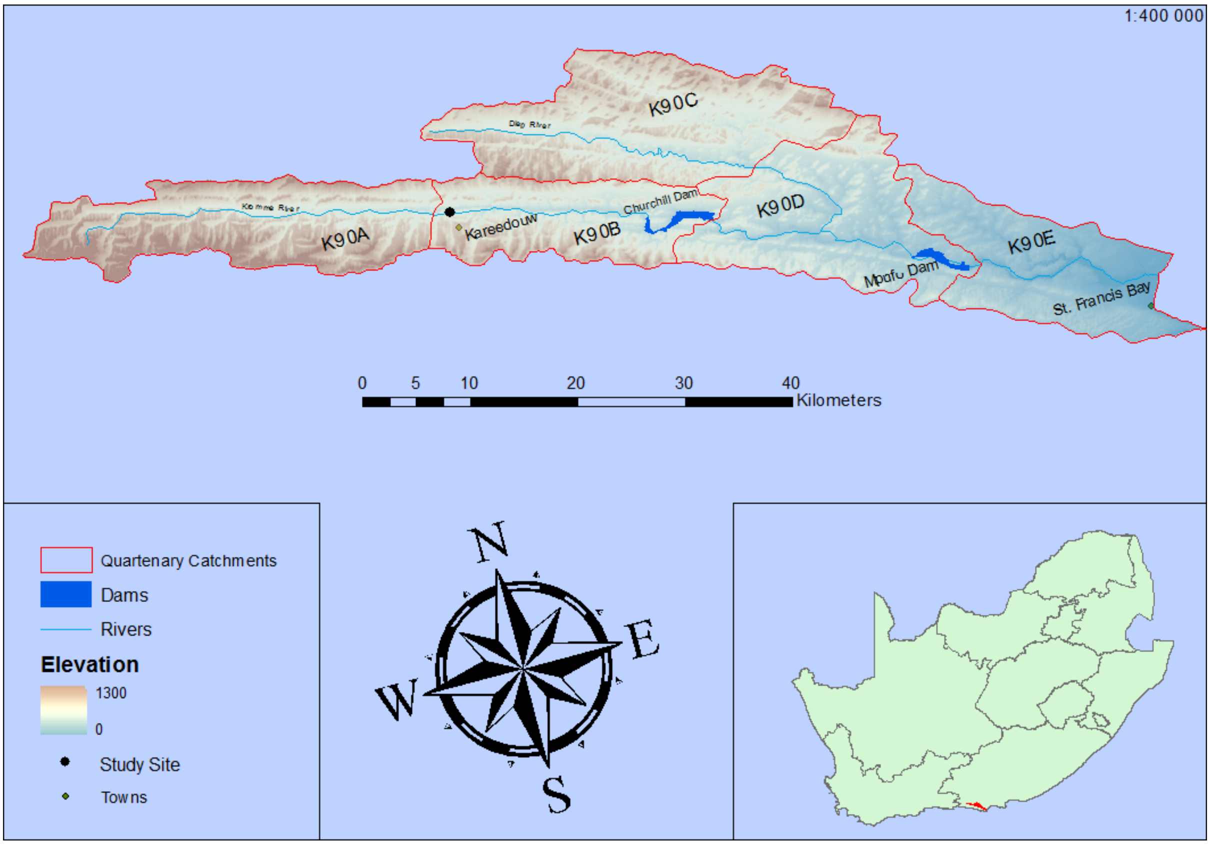

Figure 3.1: *Map of quaternary catchments K90A-K90E. Inset: Location of quaternary catchments K90A-K90E in South Africa.*

3.2 Geology and topography

The Krom River flows in an easterly direction and empties into the Indian Ocean at St Francis Bay (Figure 3.1). Its headwaters lie within the Formosa Nature Reserve located in the Tsitsakamma Mountain range (Rebelo et al., 2013). The course of the Krom River covers a distance of approximately 95 km from its headwaters to its estuary, and drains a catchment area of approximately 112 500 ha (Dennis & Wentzel, 2007). The Suuranys Mountains, with an elevation of approximately 1050 m amsl, border the Krom River to the north (Rebelo et al., 2013). To the south the Krom River is bordered by the Tsitsakamma Mountain range, which has a maximum elevation of 1251 m amsl (Haigh et al., 2008). The presence of these mountain ranges causes the Krom River catchment to slope steeply onto the valley floor, especially in its upper reaches (Haigh et al., 2008).

The geology of the upper Krom River catchment is made up of sandstone and shale of the Cape Supergroup, with the Peninsula, Nardouw, Goudini, Skurweberg and Baviaanskloof Formations being predominantly sandstone and the Cedarberg, Gydo and Ceres Formations being predominantly shale (Haigh et al., 2008). Quartzitic sandstone makes up the ridges and slopes of the catchment while softer sandstone and shale are present within the valley-bottom (Figure 3.2; Rebelo et al., 2015).

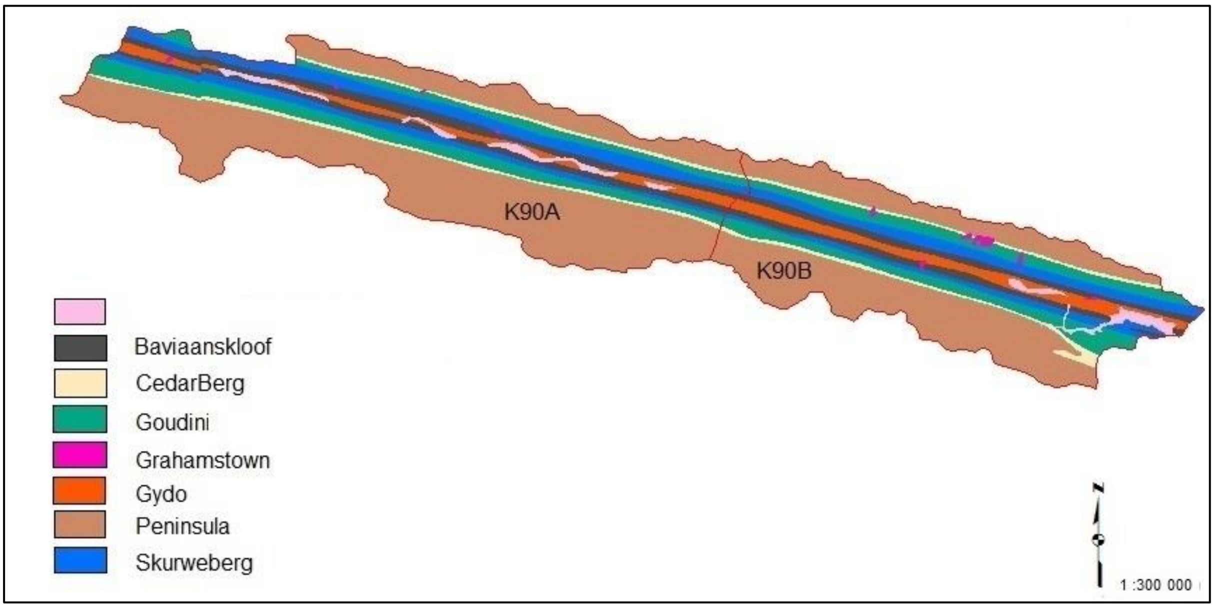

Figure 3.2: *Geology map of quaternary catchments K90A and K90B*

3.3 Climate and hydrology

A bimodal pattern of rainfall exists within the Krom River catchment, with maximum rainfall occurring in autumn and spring from February to April, and August to October, respectively (Nsor & Gambiza, 2013). The mean annual precipitation (MAP) for the entire Krom River catchment from 1950 to 2000 is 614 mm (Rebelo et al., 2015).

Spatially, the amount of rainfall in the catchment varies considerably, with some areas having average rainfall as high as 800 mm while others have average values as low as 400 mm (Rebelo et al., 2013). Typically, the south and southwest areas of the catchment in the vicinity of the Tsitsakamma Mountains, receive more rainfall than the north and northwest areas in the Suuranys Mountain region (Haigh et al., 2008). The MAP measured at the weather station in the town of Kareedouw is 716 mm (Haigh et al., 2008). Kareedouw is located in the quaternary catchment K90B, within close proximity to the selected study site, making this MAP figure relatively accurate for the study site.

The mean annual water loss through evapotranspiration for the upper Krom River catchment is 1300-1500 mm (Middleton & Bailey, 2008). The mean annual runoff (MAR) for quaternary catchments K90A and K90B is 100-200 mm (Middleton & Bailey, 2008).

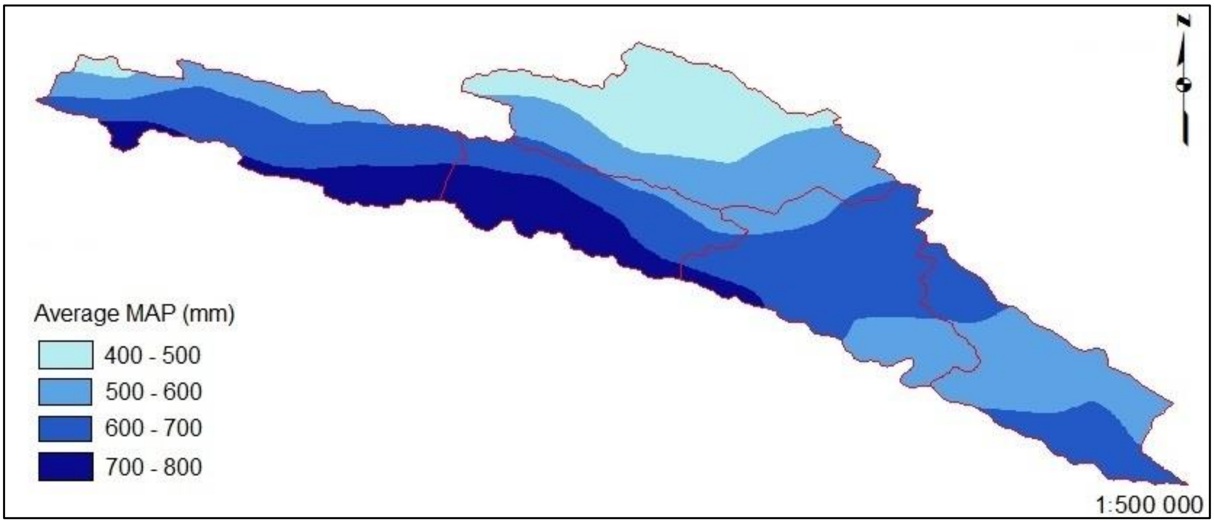

Figure 3.3: *Spatial variation in mean annual precipitation (MAP) within the Krom River catchment*

The Krom River catchment experiences occasional floods of varying magnitudes. In recent history, notable floods have occurred in 1931, 1965, 1981, 1996, 2001, 2006, 2007 and 2012. Flood return intervals and their associated discharge values for the Krom River catchment are presented in Table 3.1.

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Table 3.1: *Return intervals of floods in the Krom River based on data from 1955-2006. The data used was obtained from a gauging weir downstream of Churchill Dam, in quaternary catchment K90B (Adapted from Haigh et al., 2008).*

Return Interval (years)	Discharge (cumecs)
2	2.26
5	25.17
10	93.18
25	389.73
50	999.06
100	2366
200	5264.15

3.4 Vegetation

Fynbos is the dominant vegetation cover within the Krom River catchment. Various classes of fynbos cover approximately 46 % of the catchment area, while grasslands cover 12.5 %, thicket covers 9.8 %, renosterveld occupies 9.2 %, and forest cover 2.2 % (Mucina & Rutherford, 2006; Mander et al., 2010). Mander et al. (2010) highlight that 10.8 % of the catchment is degraded vegetation, with 0.28 % of natural vegetation being degraded due to the encroachment of invasive alien plant species.

Within the upper Krom River catchment region the dominant vegetative cover is indigenous Fynbos (Mucina & Rutherford, 2006; Nsor & Gambiza, 2013). The most dominant Fynbos classes in this region are the Tsitsikamma Mountain Fynbos and the grassier Humansdorp Fynbos (Rebelo et al., 2013). Also abundantly present within the upper Krom River catchment is Renosterveld (Mucina & Rutherford, 2006).

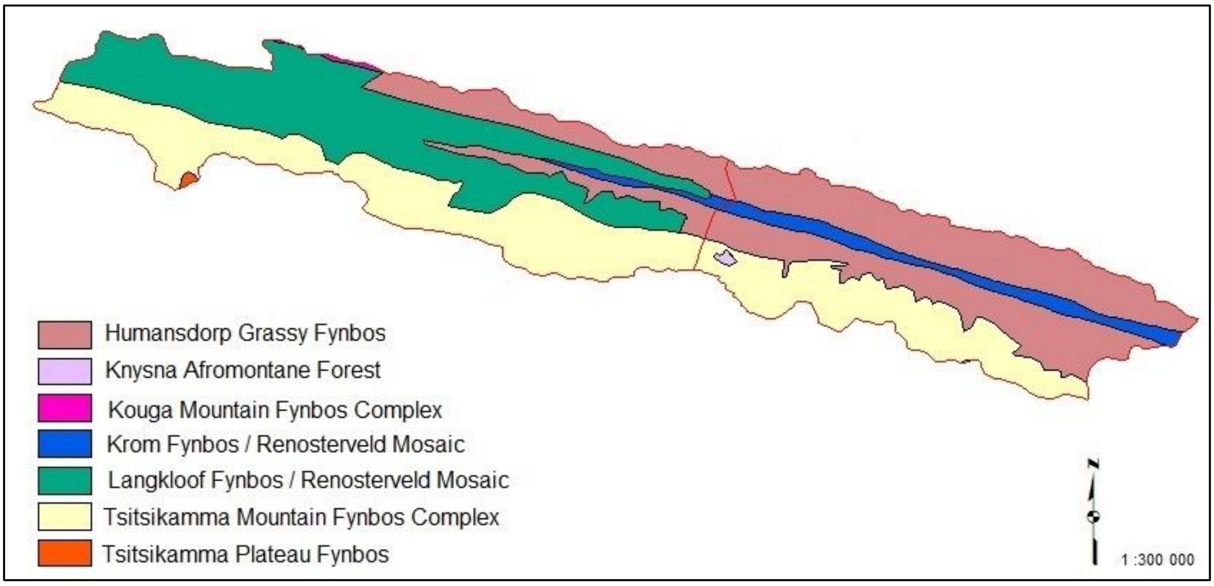


Figure 3.4: *Vegetation map of quaternary catchments K90A and K90B.*

Valley bottom wetlands occur extensively along the Krom River in quaternary catchments K90A and K90B. Palmiet, which forms monospecific stands and is an efficient peat forming wetland plant species, is the dominant vegetative cover in these areas (Barclay, 2016). The extensive presence of palmiet in these wetlands has led to the accumulation of deep peat deposits within the wetland systems (Haigh et al., 2008).

Smaller areas of grasses, sedges, ferns and reeds are also present in the wetlands, contributing to a mosaic of wetland plant communities (Haigh et al., 2008; Nsor & Gambiza, 2013). At various locations within the upper Krom River Catchment, black wattle trees (*Acacia mearnsii*) have encroached on valley-bottom wetland areas (Dennis & Wentzel, 2007).

3.5 Hydro-geomorphology

A trellis network drainage pattern is present in quaternary catchments K90A and K90B, with multiple large and minor tributaries joining the trunk river from the wet Tsitsakamma Mountains to the south, and multiple large, but mostly ephemeral tributaries joining the Krom River from the less moist Suuranys Mountains to the north (Haigh et al., 2008). Alluvial fans have formed at the distal ends of many of the tributaries in the upper Krom River catchment. To varying degrees, these alluvial fans have restricted the width of the valley floor, and have encroached on the Krom River channel and the palmiet wetlands (Haigh et al., 2008).

In certain situations the alluvial fans extending across the Krom River valley bottom have locally increased the longitudinal gradient of the trunk stream (Haigh et al., 2008; Lagesse, 2017). This is believed to promote channel incision and increases the likelihood of the formation of upstream propagating head-cuts which can lead to gully development along the Krom River channel (Haigh et al., 2008; Schlegel 2017).

3.6 Historical land use

The upper Krom River catchment area was first occupied by settlers in 1775 (Haigh et al., 2008). The development of Mosselbaai and Plettenberg Bay harbours in 1787 and 1788, respectively, led to increased harvesting of indigenous trees and intensified farming in the upper Krom river region (Haigh et al., 2008). Established in 1905, Kareedouw is the only town in quaternary catchment K90B. In 1906 a railway line linking the area to Port Elizabeth was completed leading to even more intensified farming in the upper Krom River catchment (Haigh et al., 2008). From the beginning of the 20th century until the 1940’s soft fruit orchards and grazing were the most common land uses in the upper Krom region (Haigh et al., 2008). At this time, orchards were established on the fertile Krom River floodplains. After 1942 fruit and vegetables were the main farming products in the region, with dairy and sheep farming also making up a significant part of the regional economy (Haigh et al., 2008). Much of the road and bridge construction in the Krom River catchment took place between 1950 and 1970 (Haigh et al., 2008).

3.7 Current land use

Fruit, vegetable and livestock farming remain common land uses in the upper Krom River region, with farmland currently occupying 23.1 % of the area of the Krom River catchment (Mander et al., 2010). A small number of nature reserves, game farms and holiday farms are present in the upper regions of the catchment (Haigh et al., 2008; Mander et al., 2010). Very little urban development has occurred within the upper Krom River catchment area, with the population of Kareedouw being less than 5 000 (Rebelo et al., 2013). Urban and peri-urban land cover in the Krom River catchment totals approximately 3.3% of the total catchment area.

The Krom River serves as a vital water resource for the Nelson Mandela Bay Metropole, which includes the Port Elizabeth, one of South Africa's largest cities (Haigh et al., 2008; Rebelo et al., 2015). Two large dams exist along the Krom River course. The Churchill Dam, with a capacity of 35 710 106 m^3, is located in quaternary catchment K90B, southeast of Kareedouw (Figure 3.1; Haigh et al., 2008). The Mpofu Dam, with a capacity of 10 706 106 m^3, is located in quaternary catchment K90D (Figure 3.1; Haigh et al., 2008). These dams together supply

Port Elizabeth with approximately 40% of its water, with 24% being supplied by the Churchill Dam which is fed by the upper Krom River alone (Haigh et al., 2008; Rebelo et al., 2015).

The palmiet wetlands along the upper Krom River are a vital natural resource because they purify the water of the Krom River, and thereby improve the quality of the water supplied to the Nelson Mandela Metropole (Haigh et al., 2008). The presence of erosional gullies along the river course are believed to be threatening the integrity of the wetlands, and therefore threatening the supply of good quality water (Haigh et al., 2008). This has led Working for Wetlands, a national agency tasked with the rehabilitation and conservation of South African wetlands, to invest considerable amounts of time and money into wetland rehabilitation in the catchment (DEA, 2017).

This rehabilitation has been in the form of a number of erosion control structures that have been built along the course of the upper Krom River. Construction of the first rehabilitation structures began in 2000 (Haigh et al., 2008). These structures were designed to stabilise the head-cuts of gullies present along the course of the Krom River to prevent them propagating through the palmiet wetlands, and to improve their hydrological functioning. The intention of these interventions is to ensure a sustained supply of good quality water to the Churchill and Mpofu Dams (Haigh et al., 2008). Added to this, these structures act as sediment traps, reducing the rate of sedimentation in the Churchill and Mpofu reservoirs.

3.8 Study site location

The study site for this project is a 3 km long reach of the Krom River. The upper boundary of this selected reach is 12.5 km upstream of the town of Kareedouw (Figure 3.5). This reach of the Krom River was chosen as a study site as large scale valley bottom deposition appeared to be occurring immediately downstream of an actively eroding reach. This made the site potentially useful one for understanding sediment processes and dynamics along the courseof the Krom River.

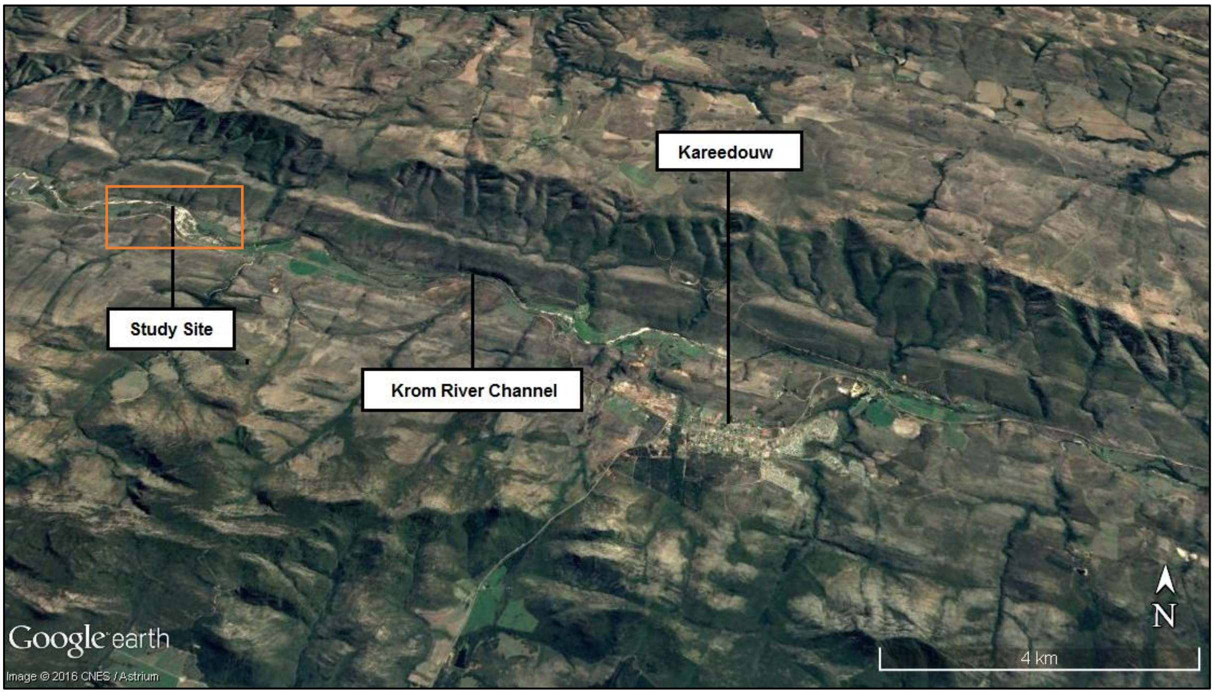


Figure 3.5: *Google Earth image of part of the Krom River showing the location of the study site in relation to the downstream town of Kareedouw.*

3.9 Study site characteristics

The upper 1.4 km of the selected reach exists in the form of a confined channel (Figure 3.6). Within this confined reach erosion through channel incision appears to be the dominant fluvial process. Downstream of the 1.4 km long confined erosional reach, the Krom River loses confinement (Figure 3.6). Downstream from this location the Krom River exists as multiple channels that run through a 1.6 km long unconfined (wide) channel that spans most of the width of the valley floor (Figure 3.6). Steep sided river banks are absent throughout this unconfined reach, and significant amounts of recently deposited alluvial sediment are present. This indicates that depositional processes have been dominant through this reach in recent times. The downstream boundary of the study site is marked by the presence of an erosion control structure.

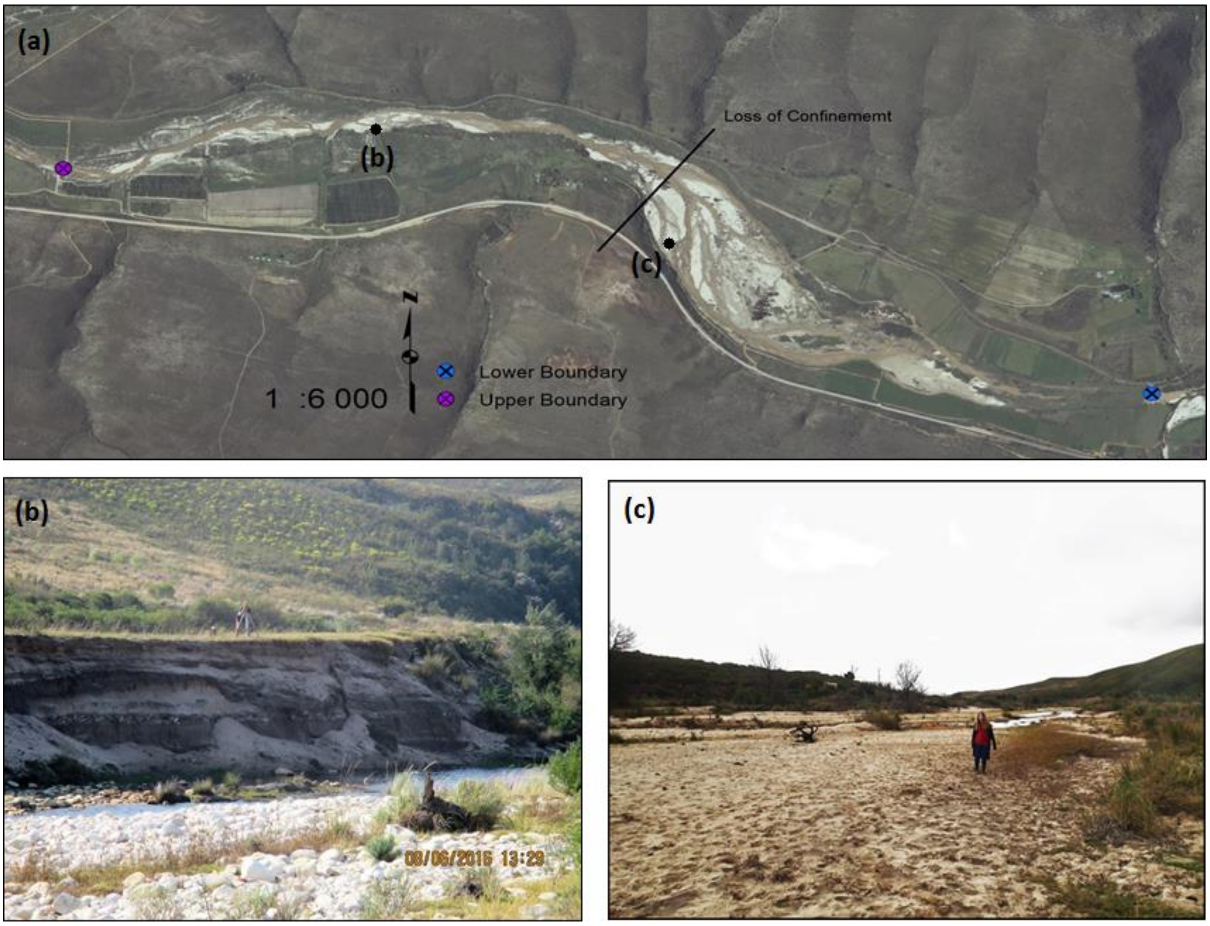


Figure 3.6: *Aerial photograph of the selected reach of the Krom River in 2012, after the occurrence of a large flood event (a), with photographs of the channel along the confined and incised reach (b), and along the unconfined depositional reach (c).*

CHAPTER 4: **METHODS**

Topographic surveys, coring, particle size analysis and an analysis of selected aerial photographs were used to achieve the objectives of this project. Table 4.1 provides an explanation of how these techniques were used to achieve each objective.

Table 4.1: *Table showing each objective of this project, and a summary of the methods used to achieve them.*

Objective	**Method**
1. Determine the spatial extent of erosion and deposition along a selected reach of the Krom River.	Cross-section d-GPS surveys of the selected reach of the Krom River in the chosen study site.
2. Determine the quantity and particle size distribution of eroded and deposited sediment within the study site.	Eroded quantity: calculated based on the dimensions (cross-section area and length) of the actively eroding channel present within the study site, and on the average packing density of the eroded sediment. Deposited quantity: Sediment deposited during the most recent large flood in 2012 was used as a basis for understanding the system's sediment dynamics. This was calculated using the dimensions (cross-sectional area and length) of the depositional feature that formed during the large flood in 2012, and its average packing density Particle size distribution of eroded and deposited sediment: Particle size distribution using dry sieving of sediment samples collected using a barrel corer
3. Determine how erosional and depositional phases initiate and interact to influence the structure and functioning of the Krom River system over time, and examine the implications of this for future erosional and depositional cycles.	Analysis of historical and current aerial photographs, and the construction of longitudinal profiles of the study site.

4.1 Spatial extent of eroded and deposited sediment

4.1.1 Field work

Objective 1 required that the spatial extent of eroded and deposited sediment be determined for the selected reach of the Krom River. A differential Global Positioning System (d-GPS) was used to topographically survey a total of 13 transects along the river course (Figure 4.1). The accuracy of d-GPS surveys is within 0.2 m in x, y and z dimensions, particularly when undertaken at a small scale such as in this study. These surveys allowed for measurements of the cross-sectional morphology of the Krom River at different locations throughout the selected reach. This information, combined with field observations, made it possible to delineate the extent of erosion and deposition within the study site.

Transects 1 to 8 were completed within the confined reach of the study site where erosion was the dominant process (Figure 4.1). These surveys extended across parts of the floodplain of the Krom River, and included survey points in the channel. Transects 9 to 13 were completed at selected locations within the unconfined reach of the study site where deposition was the dominant process (Figure 4.1). Again, these surveys included detailed measurements of the channel, and extended across the floodplain.

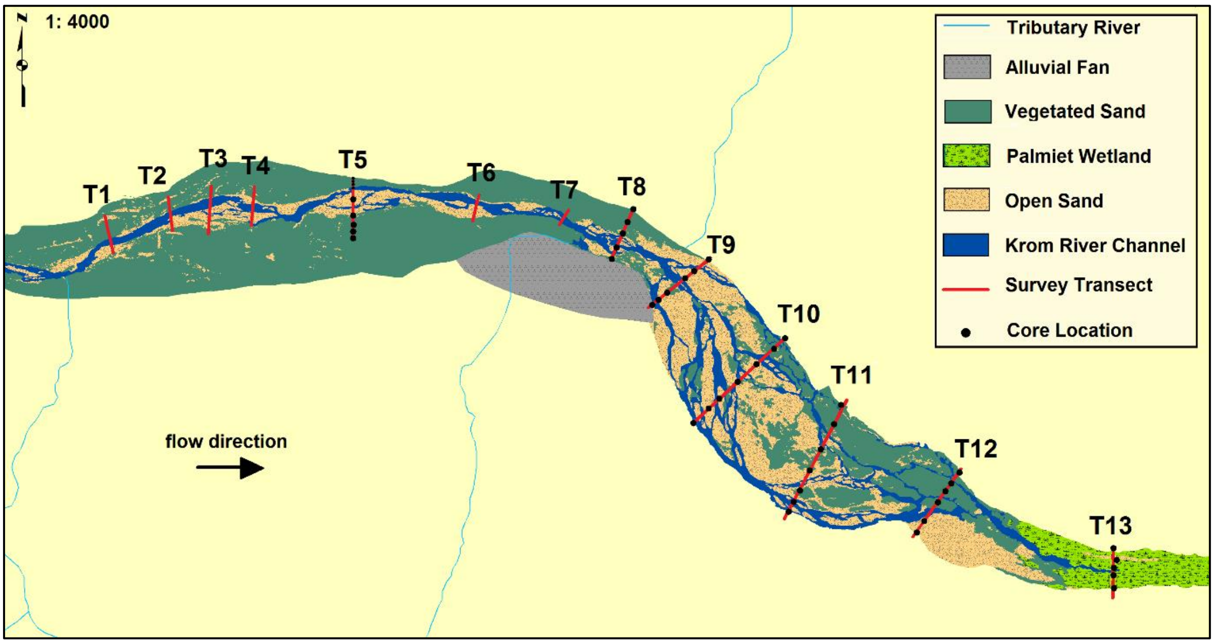


Figure 4.1: *The locations of survey transect 1 to 13 in the study area, including the locations of the cores taken along transect 5, and transects 8 to 13.*

4.2 Quantification of eroded and deposited sediment

Sediment mass is commonly used measure to assess sediment dynamics within a fluvial system (Fryirs et al., 2007). It was therefore decided that the mass of eroded sediment within the confined reach of the study site (where erosion is the dominant process), and the mass of sediment deposited in the unconfined reach of the study site (where deposition is the dominant process) during the 2012 flood event, would be determined (Objective 2). This required the determination of the volume of eroded and deposited sediment, which would be converted to mass figures based on the bulk density determined from the core samples.

4.2.1 Volume determinations

The total volume of sediment from the incised channel was calculated by multiplying the interpolated bankfull cross-sectional area (assuming a horizontal surface between the top of existing banks) by the channel length.

The volume of sediment deposited during the 2012 flood was determined by coring through the recent sedimentary deposits to the pre-2012 flood surface, and recording the depth of recent sediment. The pre-2012 surface was easily identifiable due to the abrupt transition from off-white sand deposits to a black organic-rich topsoil at the contact between flood and former valley-fill deposits, respectively (Figure 4.2). The elevation of the surface at each core location was surveyed using d-GPS such that the cross-section of the Krom River floodplain surface prior to the deposition of sediment during the 2012 flood could be determined. This information was then used to determine the volume of flood-related sediment deposition.

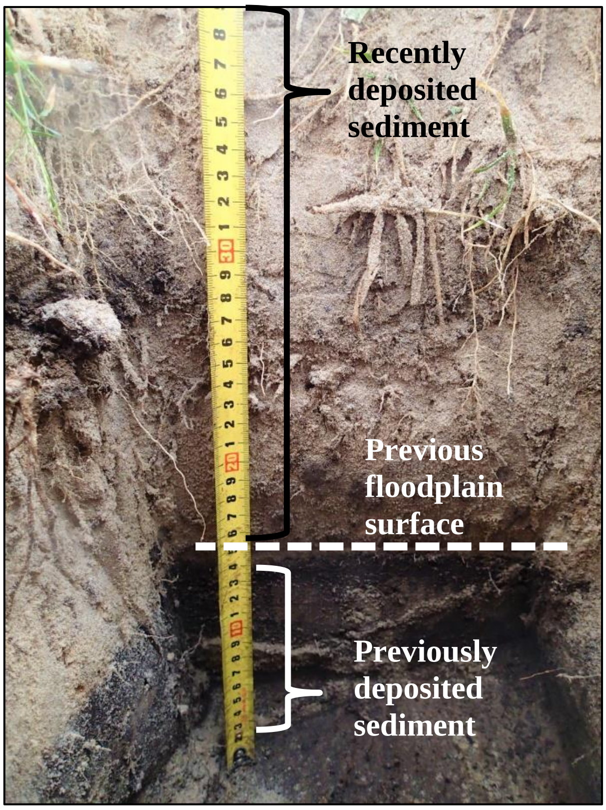


Figure 4.2: *The distinction between recently and previously deposited sediment*

Mass calculations

To convert the volumes of eroded and deposited sediment to mass figures, the average bulk density of eroded and deposited sediment was determined. This was done by collecting 10 samples of floodplain sediment (eroded sediment) from the floodplain adjacent to the erosional gully within the confined reach of the study site, and 10 samples of deposited sediment from the unconfined reach of the study site. These samples were all collected in a vessel of a known volume. They were then dried and weighed to determine their masses.

4.3 Particle size distribution of eroded and deposited sediment

The particle size distribution of eroded and deposited sediment within the study site (Objective 2) was determined using a set of stacked and graded sieves. Samples of eroded sediment were collected from the floodplain of the confined reach, whereas samples of deposited sediment were collected from the unconfined depositional reach of the study site.

To determine the particle size distribution of the sediment eroded within the confined reach of the study site, a total of 8 surface samples of sediment from the floodplain were collected using a trowel. These samples were collected at a depth of 0.1 to 0.2 m along T5, with 4 samples being collected from each bank (Figure 4.1).

To determine the particle size distribution of the sediment deposited within the unconfined reach of the study site, cores were taken through the recently deposited sediment, with each core being sampled at a range of depths. Through the unconfined reach of the study site cores were taken along surveyed transects T9 to T13 (Figure 4.1), with each core being included in the topographic survey.

Cores were retrieved using a barrel auger and, owing to the unconsolidated nature of the deposited sediment, coring sleeves were used to maintain the integrity of the core holes. Each core was sampled at 20 cm depth intervals and at every noticeable change in stratigraphy.

4.3.1 Laboratory analysis

Prior to sieving, each sample was oven-dried at 100 ºC for 48 hours and disaggregated using a pestle and mortar. Sediment samples were dry-sieved using stacked woven wire mesh sieves and a sieve shaker. Each sample was sieved using an automatic shaker for 10 minutes. The selected sieve sizes included 1000 µm, 500 µm, 250 µm, 125 µm, and 63 µm, with sediment particles < 63 µm being collected in the pan at the base of the sieve stack. These sieves were selected to correspond with the Udden-Wentworth grain-size classes). After each sample had been sieved for 10 minutes the accumulated sediment in each sieve was weighed using an electronic balance.

Table 4.2: *Udden-Wentworth grain-size classification scheme (Wentworth, 1922).*

Particle Size (mm)	**Particle Size (µm)**	**Sediment Class**
0.5-1	500-1000	Coarse Sand
0.25-0.5	250-500	Medium Sand
0.125-0.25	125-250	Fine Sand
0.063-0.125	63-125	Very Fine Sand
< 0.063	< 63	Silt and Clay

Most of the samples predominantly contained sand sized material; hence, it was considered unnecessary to fractionate particles below 63 µm.

4.4 Initiation and interaction of erosion and deposition

4.4.1 Historical aerial photography

To gain an understanding of how sediment erosion and deposition have alternated and interacted within the study area (Objective 3), and to gain an understanding of the extent of landscape change(s) that these processes caused over time, an analysis of historical and recent aerial photographs was undertaken. The analysis of maps and aerial imagery has long been considered a useful method for understanding patterns and processes of landscape change (Gautam et al., 2003; Turner et al., 1989). It is an analysis technique that has been effectively applied to both current and historical aerial imagery (Gautam et al., 2003), making it a useful tool for gaining an understanding of how the Krom River system has been influenced by erosional and depositional process over both long and short periods of time.

Aerial images from 1969, 2009 and 2012 were selected for analysis. The aerial photograph of the study site from 1969 was selected for analysis as it provided a high-quality image that allowed for an effective understanding of the historical planform morphology of the Krom River that could be compared with the present-day situation. The image from 2009 was selected for analysis as it was the most recent aerial photograph of the study site that pre-dated the large flood that occurred along the Krom River in 2012. The 2012 aerial image was captured after the flood. This provided an opportunity for a useful comparison of the structure of the selected reach in pre- and post- flood conditions, making it possible to isolate any landscape change(s) that may have been due to gully erosion and sediment deposition initiated because of this flood event

Each of the aerial photographs was obtained from the Department of Rural Development and Land Reform (DRDLR), through the National Geo-Spatial Information (NGI) agency. The aerial photograph of the selected reach from 1969 was captured at a photo-scale of 1:36 000 using a wide-angle camera. The images from 2009 and 2012 were captured using an Intergraph digital mapping camera and have a ground sample distance (GSD) of 0.5 m. These images were available at a scale of 1:10 000.

The use of geographic information systems (GIS) software for the analysis of aerial photographs has increased the ability and effectiveness of this technique for assessing landscape change. GIS has been regularly and successfully used to map land cover to asses landscape change (Cousins, 2014; Gautam et al., 2003; Haile, 2014). Arc Map 10.3.1 was used for the analysis of the selected aerial images. For each photograph, the main land cover classes making up the Krom River valley bottom area were mapped and digitised through the creation of polygons. Descriptions of each valley bottom land cover class are provided in Table 4.3. For the 1969 image this digitisation process was completed at a scale of 1:1000 or at a higher scale of resolution when increased detail was required. The increased quality of the 2009 and 2012 images allowed for the digitisation process to be completed at a finer scale; therefore, at least 1:500 was used throughout.

Table 4.3: *Descriptions of the land cover classes used for the analysis of the selected aerial photographs.*

Land cover class	**Description**
Channel	Open water areas of the Krom River channel.
Palmiet wetland	Areas of intact wetland dominated by the presence of palmiet (*Prionium serratum*).
Bare sand	Recently deposited alluvial sediment deposits that have not yet been colonised by vegetation.
Vegetated sand	Areas of previously deposited alluvial sediment that have since been vegetated by either terrestrial or wetland plants.

4.4.2 Longitudinal profiles

To understand how and why erosional and depositional phases initiate, and to better understand the implications of these processes (Objective 3) two longitudinal profiles of the study site were created. Key elevation data, from each transect survey were used to produce these longitudinal profiles of the Krom River, to identify the thalweg, the top of the left bank (erosional reach), and the central floodplain, and floodplain margin elevations (depositional reach; Figure 4.3).

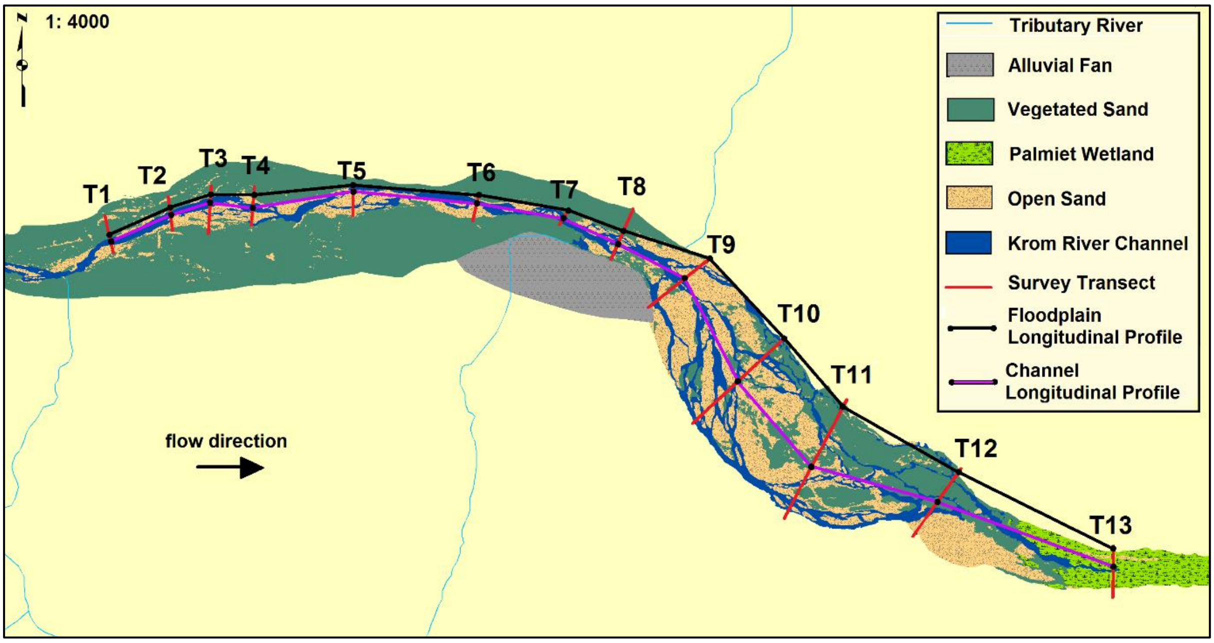


Figure 4.3: *The locations of survey readings from each cross-section to construct the longitudinal profiles of the Krom River.*

4.5 Data quality and uncertainty

In this study the mass of eroded and deposited sediment was calculated and then used to make inferences about the degree of dis-connectivity of the Krom River. The sample collection and measurement processes associated with the above mentioned data are subject to varying degrees of uncertainty (Horowitz, 2017). Uncertainty, often due to sample variability, is mostly a result of the instruments used and/or human error. Given that the data produced from samples and measurements are used to make inferences about natural processes and systems it is important to be critical about the degree and source of uncertainty associated with data.

Although inherent instrument inaccuracies are unavoidable, all field surveys in this study were completed on the same day using a d-GPS that was operated by the same people throughout the process. Added to this the study site for this project is relatively small, such that the sample variability and error associated with the survey measurements are likely to be minimal. The distinct contrast in sediment colour between recently deposited sediment and the pre-2012 flood surface made the measurement of the depth of deposited sediment at the selected sample locations an accurate and easily repeatable process. Sample variability was therefore largely reduced for this measurement process.

Average bulk density and volume figures were used to determine the mass of eroded and deposited sediment within the study site. Any calculations that involve average values will be associated with uncertainty as an average represents a range of values in reality. Added to this, recently deposited sediment was present on the bed of the confined reach of the study site when this study was completed. It was therefore not possible to survey the bed of the gully prior to the 2012 flood. Instead the current gully bed was surveyed and this survey data was used for subsequent calculations and interpretations. As a result, the calculated volume of the gully along the eroded reach, and the calculated mass of eroded sediment along this reach, are likely to be conservative estimates.

Therefore, the interpretations made in this study are subject to the cumulative uncertainty of the measurement and calculation process. However, it is believed that the efforts made to reduce sample variability during the collection and measurement process of this study means the interpretations made based on collected and calculated data are valid.

CHAPTER 5: **RESULTS**

5.1 Landscape change

The aerial photograph analyses highlight the dynamic nature of the Krom River system. Through processes of erosion and deposition significant landscape changes have taken place along the studied reach of the Krom River. Mapping the patterns of landscape change in the study site over time provides insights into how erosion and deposition interact within this system (Objective 3).

In 1969 palmiet wetlands were the dominant valley floor land cover for the reach of the study site that is downstream of the large impinging alluvial fan that originates in the Tsitsikamma Mountain Range to the south of the Krom River (Figure 5.1, a). Except for two small, isolated, areas of open sand, the study site downstream of the alluvial fan was free of recently deposited alluvial sand in 1969. However, upstream of the point where intact palmiet wetlands ceased, vegetated sand and open, recently deposited alluvial sand were the dominant land covers. Through the study site in 1969, the Krom River largely existed as a single but discontinuous channel. Because of the tendency of palmiet to colonise open water channels it is likely that some of the flow of the Krom River was not visible in the aerial photographs in the reaches where this vegetative cover was thickest.

In 2009 vegetated sand was the most abundant land cover class through much of the selected reach of the Krom River. Small amounts of open, non-vegetated sand were present near the head of the selected reach. This open sand did not extend across the valley bottom, but was generally within close proximity to the Krom River channel (Figure 5.1, b). Compared to the 1969 land cover map, the presence of intact palmiet wetland had been substantially reduced in 2009 such that between 1969 and 2009 extensive areas of palmiet wetland were replaced by areas of vegetated sand. The Krom River channel underwent noticeable change from 1969 to 2009, such that it appeared to increase in width and become more continuous. However, downstream of the southern tributary alluvial fan, the stream became increasingly divided over time, such that it formed multiple discontinuous branches in the 2009 imagery.

Between 2009 and 2012 the Krom River valley bottom areas again experienced significant changes. The presence of palmiet wetland was further reduced and open sand extended across most of the valley bottom area downstream of the south tributary alluvial fan, but it was not confined to areas near to the Krom River channel (Figure 5.1, c). The reach of the study site upstream of the alluvial fan also contained significant amounts of open sand. The open sand in this upper reach was, however, generally restricted to areas within close proximity to the Krom River channel. In 2012, upstream of the alluvial fan the Krom River channel was continuous. However, immediately downstream of the impinging alluvial fan to the toe of the study site, the Krom River existed as a braided stream network.

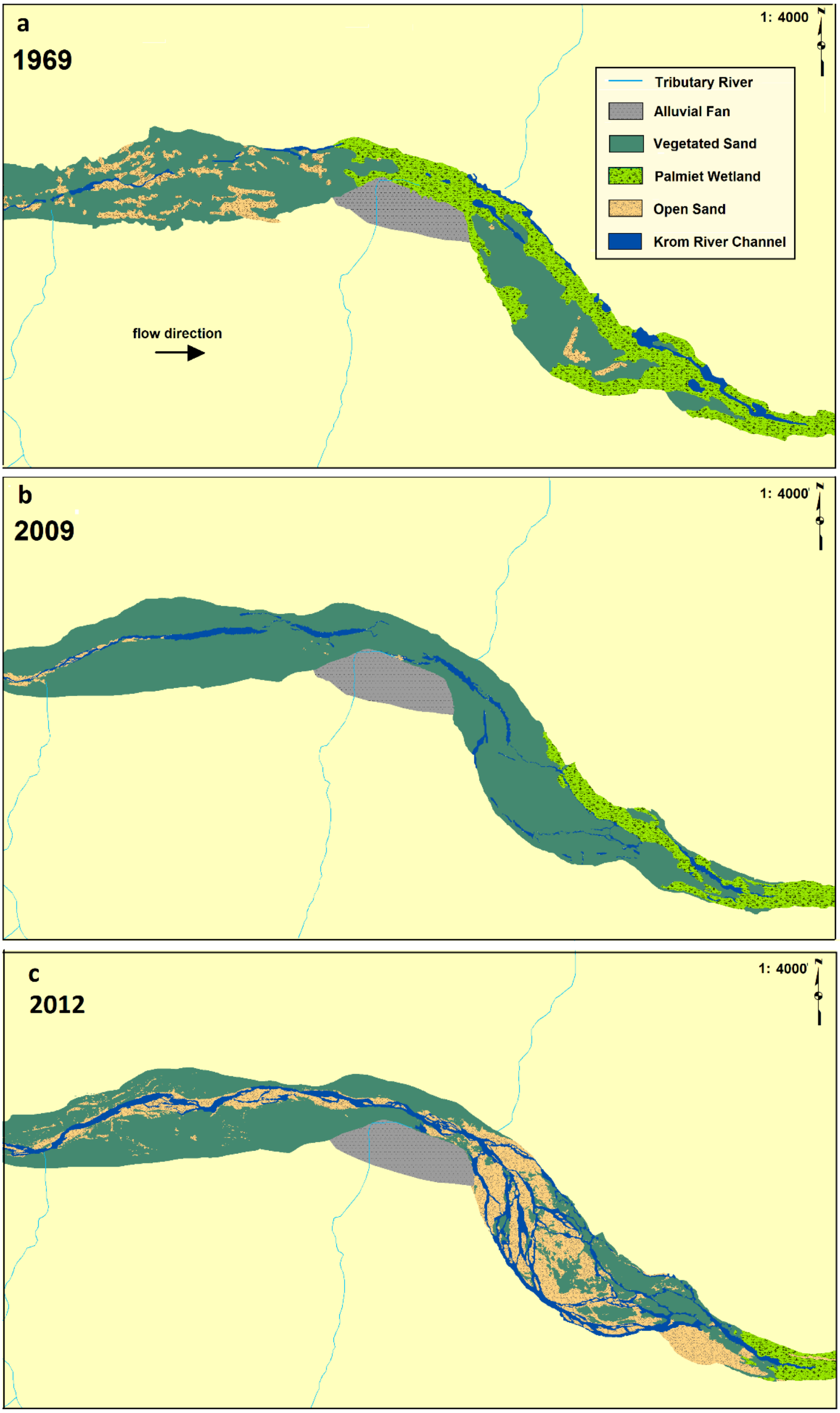


Figure 5.1: *Land cover maps of the selected reach of the Krom River from 1969, 2009 and 2012, produced from aerial photography.*

5.2 *Spatial extent of eroded sediment*

Along the upper 1.4 km of the study site, the Krom River exists as a confined and deeply incised channel. Gully erosion has been the dominant process through this confined section. However, some recently deposited alluvial sediment is present on the channel bed. Despite considerable variability, the width of the Krom River channel shows an overall general decrease with distance along this confined reach, decreasing from a width of 56 m at the head of the study site (Figure 5.2, T1) to a width of 30 m immediately upstream of where the Krom River channel loses confinement (Figure 5.2, T8). Within this confined reach, the channel reaches a maximum width of 111 m (Figure 5.2, T3). Similarly, and despite variation downstream, the depth of the incised channel decreases with distance along the confined section of the study site. At the head of the study site the Krom River channel has a thalweg depth of 5.8 m (Figure 5.2, T1). Immediately upstream of where the Krom River loses confinement the channel has a thalweg depth of 2.4 m (Figure 5.2, T8).

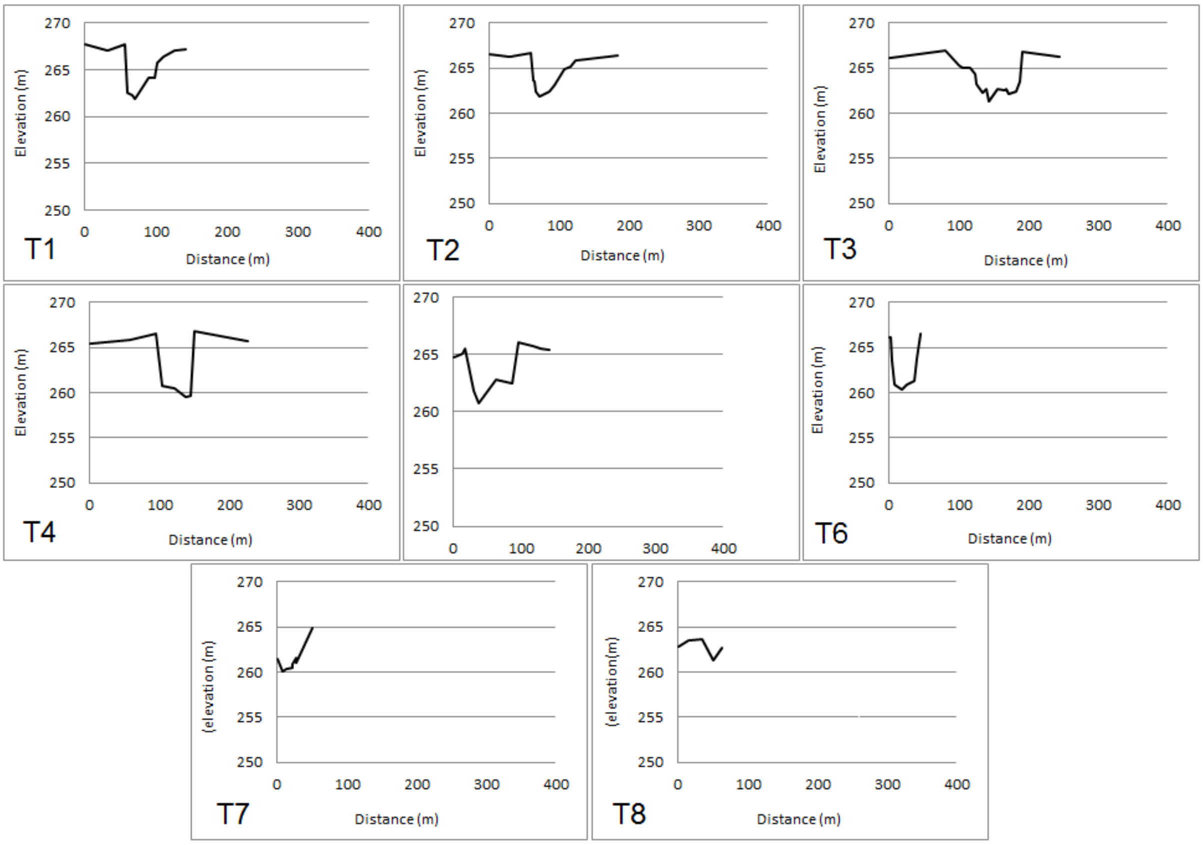


Figure 5.2: *Channel cross-sections through the confined reach of the study site, starting from the upper boundary of the study site (T1), extending downstream, to the point where the Krom River loses confinement (T8).*

5.3 Spatial extent of deposited sediment

Immediately downstream of T8 (Figure 5.2) the Krom River channel experiences a loss of confinement, with the channel no longer having steep sided and eroded river banks. Noticeable after the loss of channel confinement is a significant amount of recently deposited alluvial sediment that covers much of the valley bottom area.

At T9, where the loss of confinement is first noticeable, the depositional area has a width of 180 m (Figure 5.3). At this location, the deposited sediment has formed an elevated mound. At its crest, this mound is 1.4 m higher than the land surface adjacent to the channel. The elevated depositional feature is still evident at T10 (Figure 5.3) but here the mound is less distinct, rising 1 m above the adjacent land surface at its crest. Downstream of the initial loss of confinement the depositional area gets progressively wider, reaching a maximum width at T11. Here the unconfined depositional area has a width of 310 m (Figure 5.3). Downstream of T11 the depositional area begins to narrow, eventually reaching a width of 107 m near the toe of the study site (Figure 5.3, T13). At the toe of the study site is a rock gabion erosion control structure, immediately downstream of which the Krom River is a confined and incised channel once again.

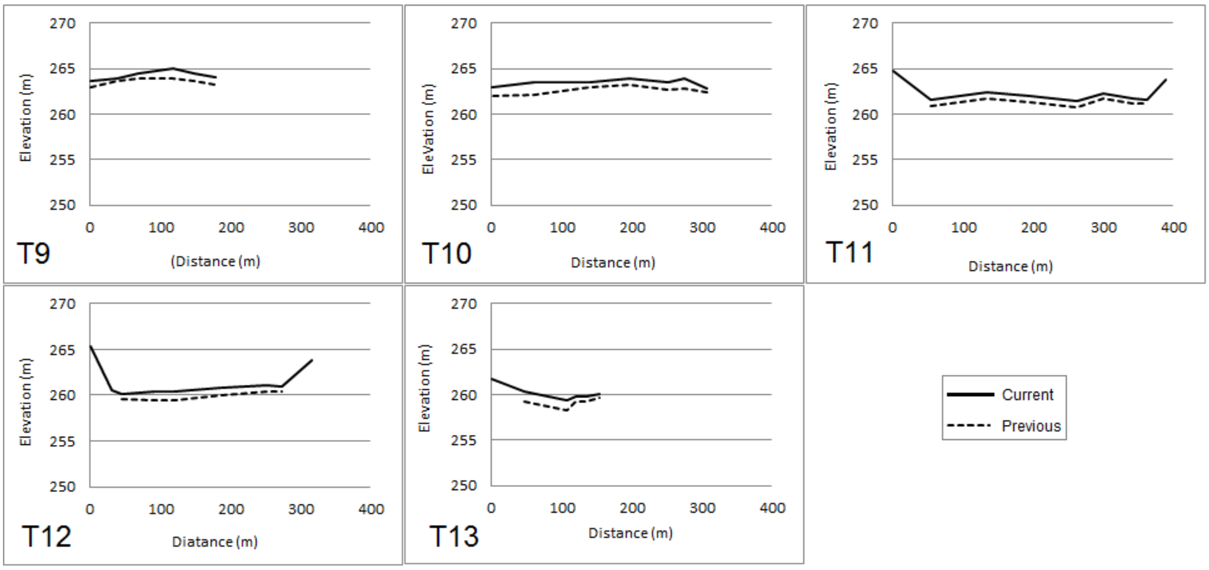


Figure 5.3: C*ross-sections through the unconfined reach of the study site, starting at T9, and extending to the lower boundary of the study site (T13).*

5.4 Mass of eroded and deposited sediment

5.4.1 Eroded sediment

While evidence of this erosion is prominent throughout the incised reach, the greatest concentration of sediment generated from erosion occurs between T2 and T5 (Table 5.1) The severity of erosion decreases with distance from the head of the study site, such that the last 198 m (14.1 % of the length of the eroded reach) accounts for approximately 4 % of the total mass of eroded sediment along the reach (Figure 6.4). The approximate mass of sediment eroded over time, through channel incision processes along the confined reach of the study site is estimated to be 328 002 – 404 799 metric tonnes (t).

Table 5.1: *Mass of sediment eroded per meter of channel through the confined section of the study site.*

Reach	Length (m)	Eroded sediment per meter of channel (t/m)	Total mass per section (t)
T1-T2	212	202 – 250	42 880 – 52 920
T2-T3	133	302 – 373	40 105 – 49 495
T3-T4	130	393 – 485	51 026 – 62 – 974
T4-T5	277	317 – 391	87 909 – 108 491
T5-T6	297	248 – 306	73 585 – 90 815
T6-T7	161	127 – 156	20 411 – 25 189
T7-T8	198	61 – 75	11 996 – 14 804
TOTAL			328 001 – 404 799

5.4.2 Deposited sediment

The greatest concentration of deposited sediment occurs along the 650 m long stretch of the study site between T9 and T11, which comprises 40 % of the length of the depositional reach but where 53 % of the total mass of deposited sediment is stored. The concentration of deposited sediment decreases with distance downstream of the loss of confinement such that the last 432 m (27 % of the length of the depositional reach) stores 17 % of the total mass of deposited sediment (Table 5.2). The total mass of sediment deposited within the unconfined reach of the study site during the flood event in 2012 is calculated at 436 000 – 451 000 t (Table 5.2).

Table 5.2: *Mass of sediment deposited per meter of channel through the unconfined section of the study site.*

River Reach	**Length (m)**	**Deposited sediment per meter of channel (t/m)**	**Total mass per section (t)**
T8-T9	211	250 - 258	52 768 – 54 632
T9-T10	257	381 - 395	98 166 – 101 634
T10-T11	389	338 - 350	131 675 – 136 325
T11-T12	329	244 - 253	80 282 – 83 118
T12-T13	432	169 - 175	73 109 – 75 691
TOTAL			436 000 – 451 000

5.5 Particle size of eroded sediment

The sediment from the floodplain adjacent to the incised Krom River channel was considered representative of the sediment that was mobilised by the incision of the Krom River channel. This sediment contained a low percentage of coarse sand (~5.5 %). The most abundant particle size classes along the Krom River floodplain are medium sand (~29 %) and fine sand (~25 %). Very fine sand and silt each make up about 20 % of the eroded sediment (Figure 5.4).

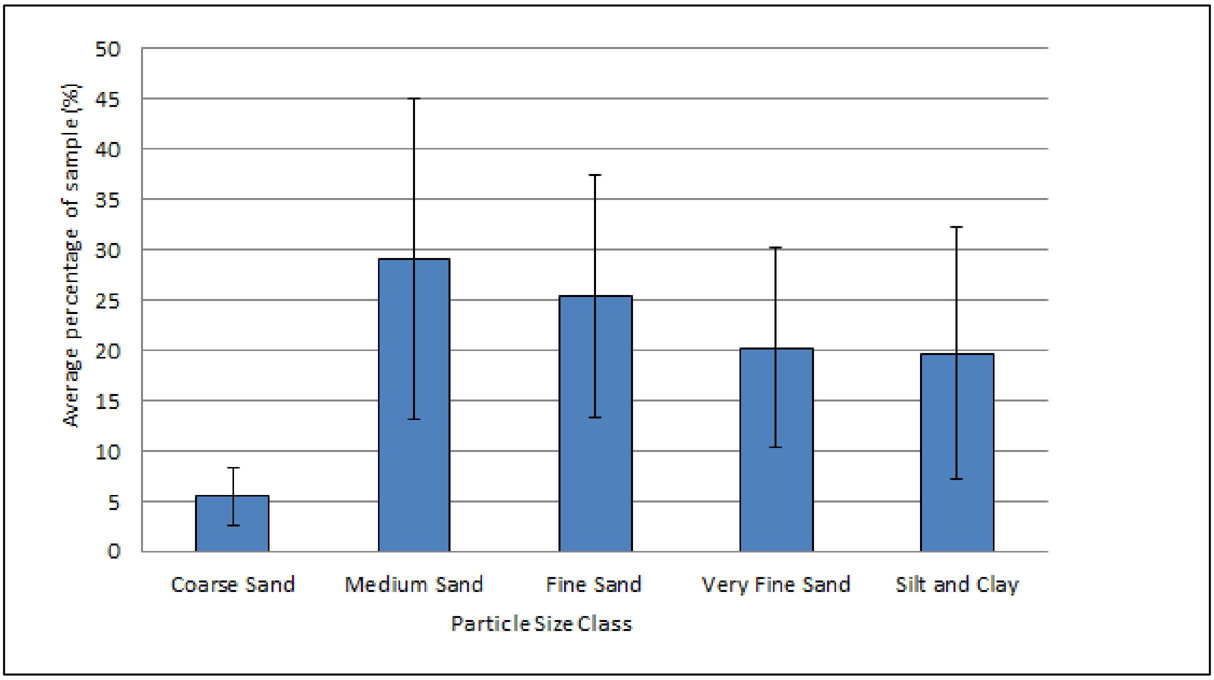


Figure 5.4: *The average percentage of sediment from each sediment size class for the samples taken from the floodplain at T5.*

5.6 Particle size of deposited sediment

The cores taken along T9, T11 and T13 were used to the particle size distribution of sediment deposited at the head, near the middle, and at the toe of the unconfined section of the study site respectively. The cores along T9 show the nature of the sediment that is deposited within the elevated depositional mound that exists immediately downstream of where the Krom River loses confinement. A total of 6 cores were taken across the 180 m wide depositional area at T9. These cores show an increase in the presence of coarse sand and a decrease in the presence of fine material (fine sand, very fine sand and silt) with increasing depth, suggesting an overall upward fining sequence throughout (Figure 5.5).

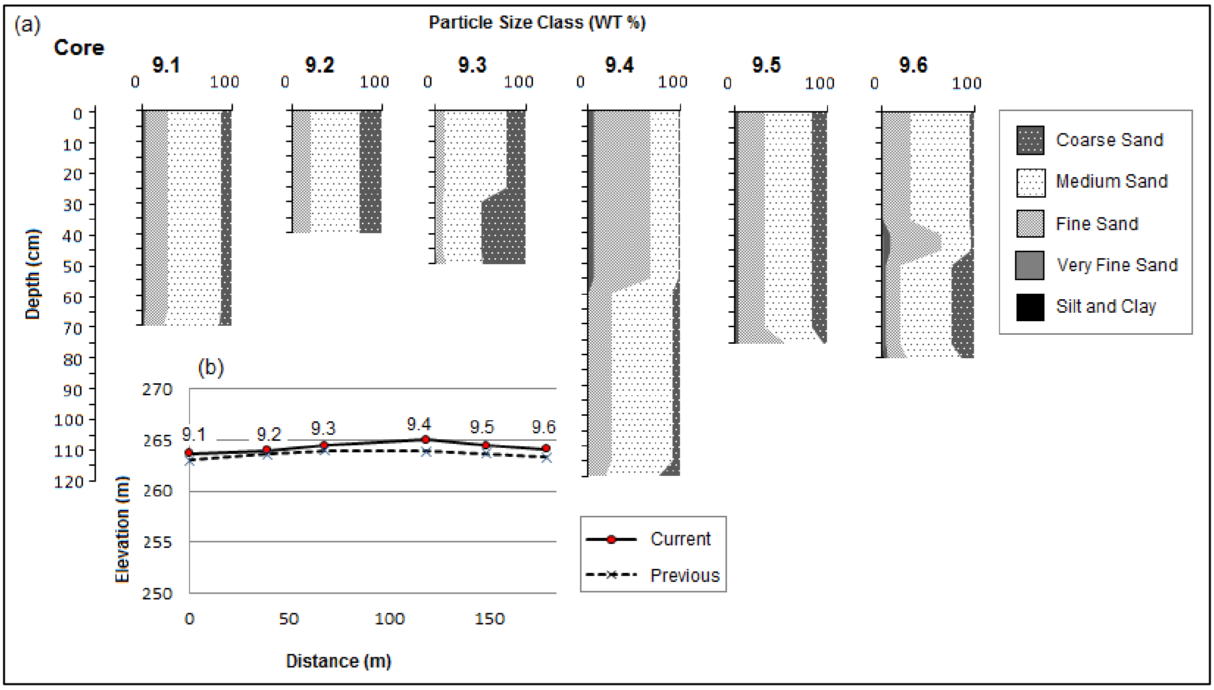


Figure 5.5: *Particle size distribution of Cores 9.1 to 9.6 (a) and cross-section profile of T9 showing the location of cores (b).*

T11 was 650 m downstream of T9, near the middle of the unconfined section of the study site. 7 cores were taken across the width (310 m) of the depositional reach. Medium sand, ranging between 27 % and 73 % was the most abundant sediment size class along this transect (Figure 5.6). Small amounts of coarse sand (< 5 %) were present in the cores taken near the middle of the valley (Cores 11.2, 11.3, 11.4 and 11.5). Within these cores the presence of fine sand and very fine sand was elevated compared to the cores taken near the edges of the depositional area (Cores 11.1, 11.6 and 11.7). The cores near the left and right edges of the floodplain area have elevated amounts of coarse sand compared to the more central cores. A general upward fining pattern is again evident for the cores along T11.

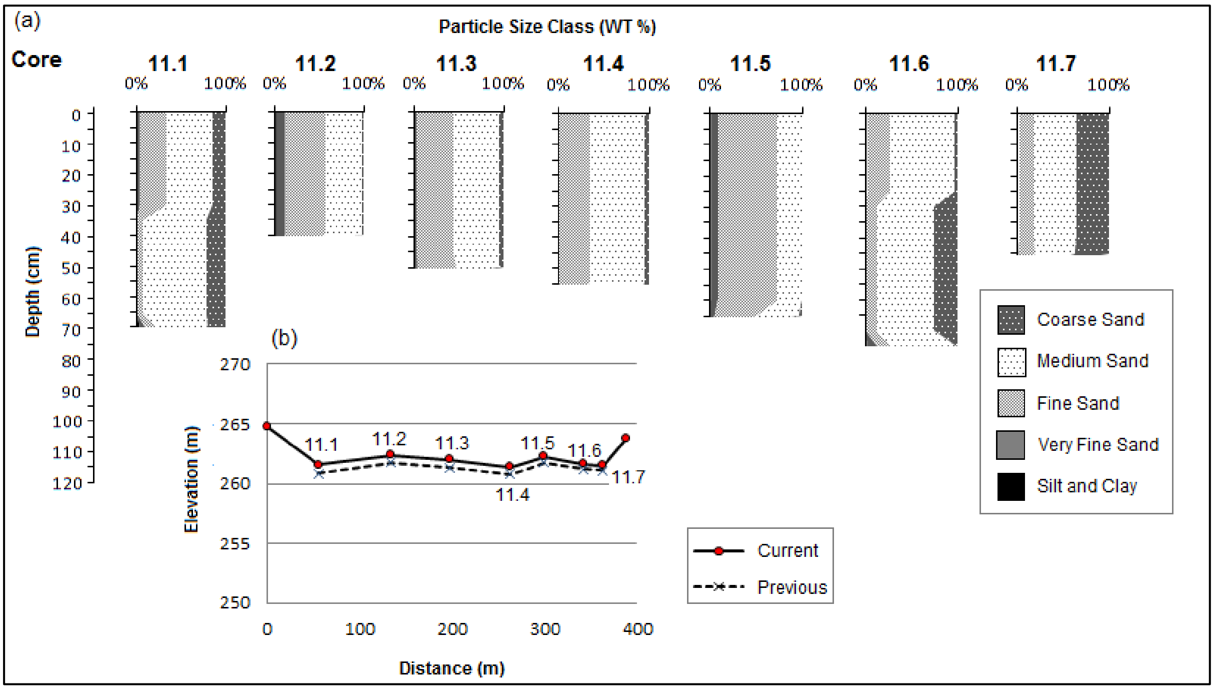


Figure 5.6: *Particle size distribution of Cores 11.1 to 11.7 (a) and cross-section profile of T11 showing the location of cores (b).*

T13 was taken at the toe of the study site, 760 m downstream of T11 (Figure 4.1). A total of 5 cores were taken along the 107 m depositional area. Fine sand (19 % to 50 %) and medium sand (2 % to 73 %) are the dominant sediment classes for these cores (Figure 5.7). The presence of coarse sand is limited in each core along this transect, and again experiences its highest percentages near the base of each core. Very fine sand represents < 2 % of the material in each core, except for Core 13.3, where it ranges between ~ 13 % (near the surface) and ~ 2 % (at its base). Again, the cores were consistently upward fining.

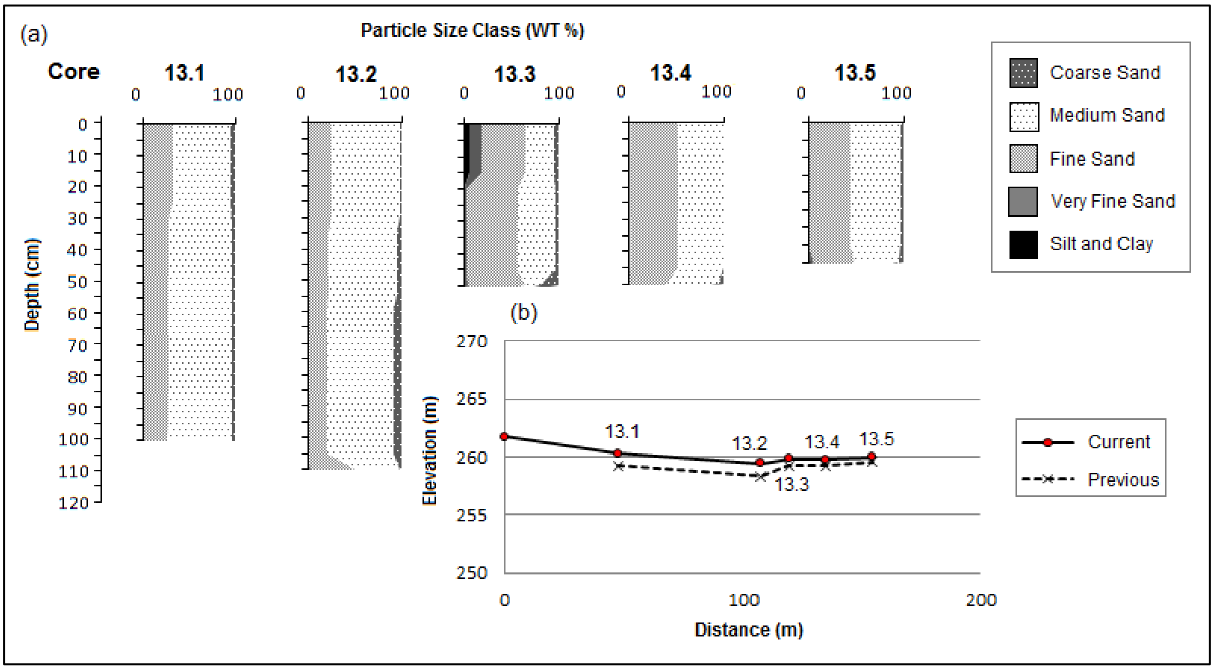


Figure 5.7: *Particle size distribution of Cores 13.1 to 13.5 (a) and cross-section profile of T13 showing the location of cores (b).*

5.7 Longitudinal depth of deposits and patterns of sediment particle size

The depth of sediment deposited in the centre of the Krom River channel increases from T5 to T8, and then from T8 to T9, which was taken immediately downstream of the point where the Krom River loses channel confinement (Figure 5.8). The depth of sediment deposited in the middle of the channel at T9 is 120 cm, making this the deepest recorded deposit in the study site. Downstream of T9, the depth of sediment deposited at the centre of the channel quickly decreases with distance away from the point where confinement is lost, such that the sediment deposit from T9 was deeper than the deposit from T10, while the sediment deposit from T10 was deeper than that from T11. The depth of centrally deposited sediment increases from T11 to T12 (95 cm), before decreasing at T13, where the lowest depth of centrally deposited sediment exists (50 cm). It is important to note that each core was abandoned as soon as the layer of previously deposited sediment was reached. It is therefore likely that the thickness of each recorded sediment deposit is likely to represent a minimum depth.

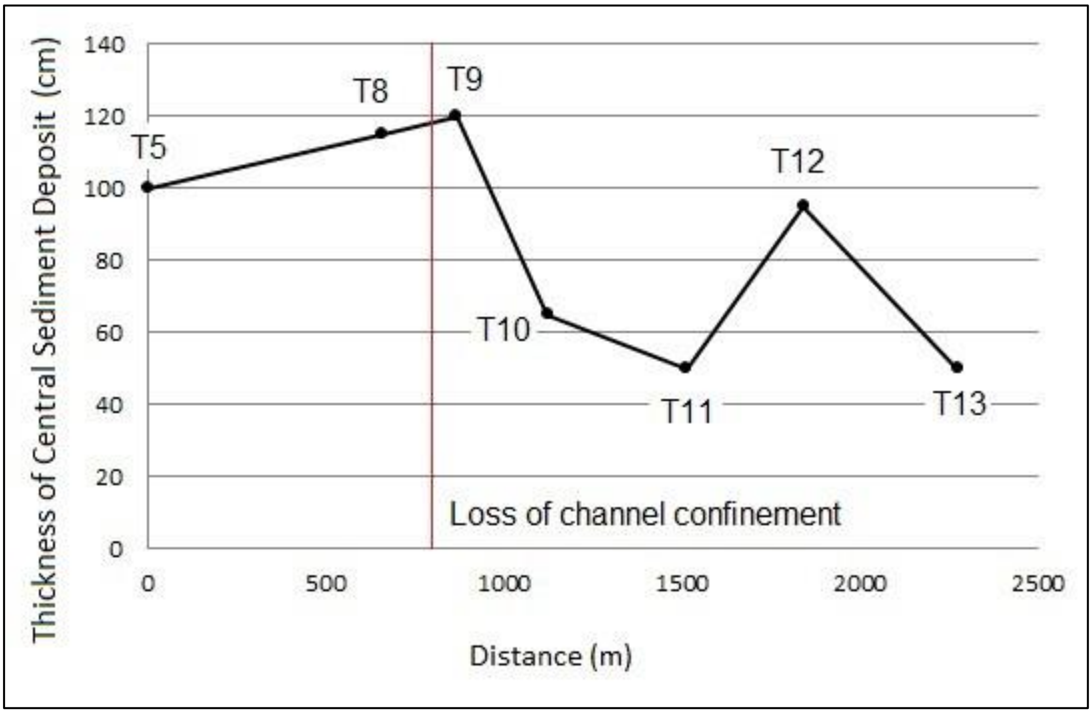
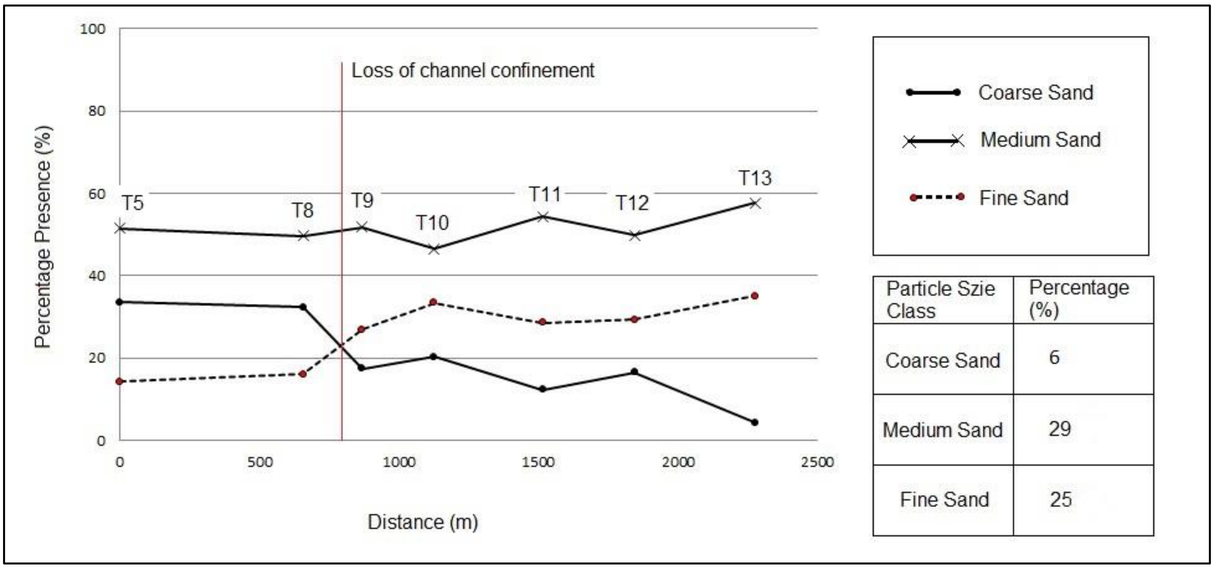


Figure 5.8: *Depth of sediment deposited near the centre of the Krom River during the 2012 flood at T5 and T8 within the confined section, and T9 to T13 within the unconfined section.*

The distinct longitudinal sedimentation patterns associated with the 2012 flood are made clear when looking at the average percentage of each sediment class at different distances along the study site (Figure 5.9 and Figure 5.10). Medium sand remains consistently high throughout the study site, and ranges between 46 % and 58 %. The presence of coarse sand decreases with distance downstream, decreasing from ~ 33 % at T5 to ~ 4 % at T13 (Figure 5.9). A sudden decrease in the presence of coarse sand is evident at T9, immediately downstream of where the Krom River loses confinement. Despite the general decrease in coarse sand along the length of the study site, the presence of coarse sand throughout is consistently higher than the average presence of coarse sand in the eroded sediment (Figure 5.9). The presence of fine sand increases with distance along the study site, from ~14 % at T5 to ~35 % at T13, with a noticeable increase in the presence of fine sand occurring at T9.



Particle Szie Class	Percentage (%)
Coarse Sand	6
Medium Sand	29
Fine Sand	25

Figure 5.9: *Average presence of coarse sand, medium sand and fine sand at T5 and T8 within the confined section, and T9 to T13 within the unconfined section. Inset: Table showing the average presence of coarse, medium and fine sand for the floodplain (eroded) sediment through the confined reach of the study site*

The presence of very fine sand at T5 and T8 is 0.60 % and 1.40 % respectively (Figure 5.10). Downstream of the loss of channel confinement the presence of very fine sand increases, reaching a peak of 4.40 % at T10. Downstream of T10 the average presence of very fine sand progressively decreases, although never to less than 2 %. The average presence of silt increases from T5 to T12, before decreasing slightly at T13 (0.80 %). Compared to the average presence of very fine sand and silt in the eroded sediment, the average presence of these sediment classes in the deposited sediment is consistently very low throughout the length of the study site.

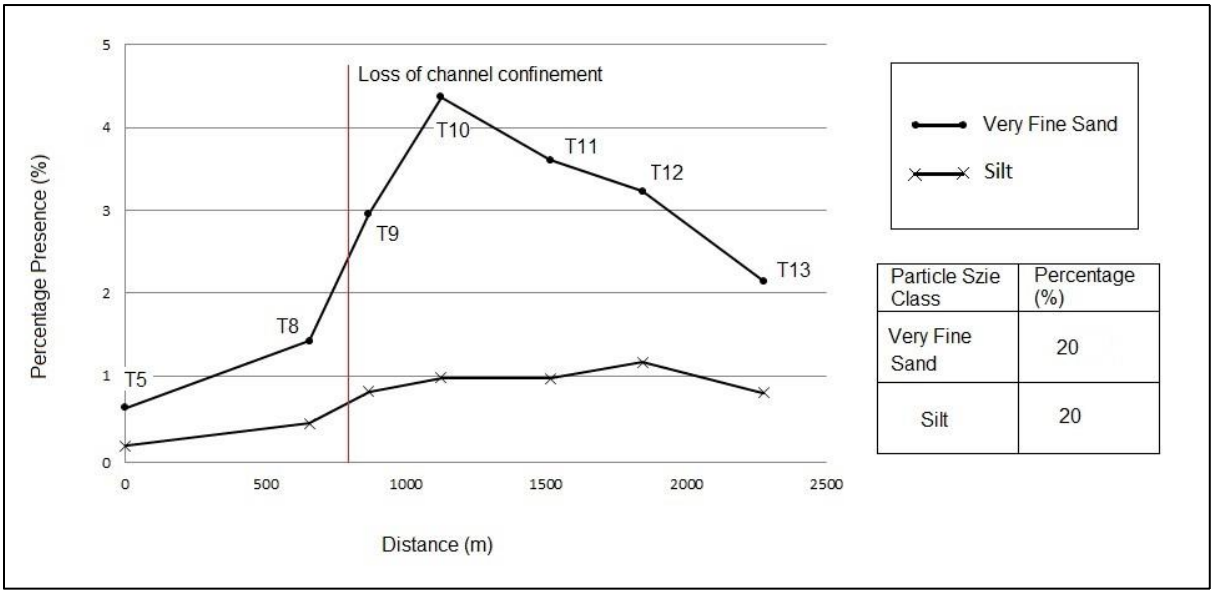


Figure 5.10: *Average presence of very fine sand and silt for T5 and T8 within the confined section, and T9 to T13 within the unconfined section. Inset: Table showing the average presence of very fine sand and silt for the floodplain (eroded) sediment through the confined reach of the study site.*

5.8 Longitudinal slope variation

The longitudinal profile of the Krom River floodplain represents the approximate profile of the study site prior to being influenced by any erosional and depositional processes. The longitudinal profile of the current bed of the channel shows how processes of erosion and deposition have altered the longitudinal structure of the valley.

Through the confined section of the Krom River, the floodplain has a slope gradient of 0.0031 m/m (0.31 %; Figure 5.11). This is less steep than the~0.0035 m/m (0.35 %) regional slope gradient of the Krom River. Approximately 1000 m downstream of the upper boundary of the study site a large right bank alluvial fan is impinging on the Krom River valley bottom. The 570 m stretch of the floodplain that is adjacent to this alluvial fan has a slope gradient of 0.0066 m/m (0.66 %). The slope gradient of the floodplain of the Krom River, downstream of the point where the Krom loses confinement is 0.0028 m/m (0.28 %), also making this stretch of floodplain less steep than the regional slope.

The current channel bed of the confined section of the study site is an average of 4.56 m lower than the elevation of the floodplain adjacent to the incised channel. The bed of the Krom River has a gradient of 0.001 m/m (0.1 %) through the confined section of the study site, such that it is approximately a third of the steepness of the floodplain along this reach of the study (0.33 %). It is also less steep than the regional slope of the Krom River (0.35 %). The sediment deposits downstream of where confinement is lost are thickest at the head of the unconfined reach, and thin towards the toe the study site. As a result, the slope gradient through the unconfined reach of the Krom River is approximately 0.003+7 m/m (0.37 %), which is steeper than the slope of the adjacent floodplain through this same reach (0.28 %), and steeper than the regional slope of the Krom River.

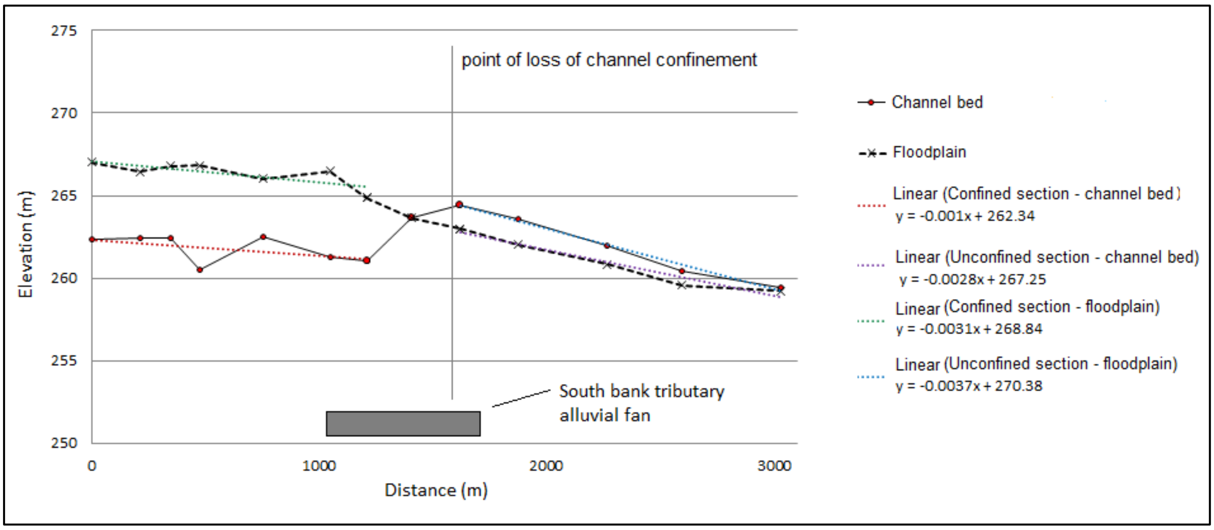


Figure 5.11: *Longitudinal profile of Krom River channel and floodplain for the reach that makes up the study site.*

CHAPTER 6: **DISCUSSION**

6.1 Longitudinal connectivity of the Krom River System

Within a fluvial system longitudinal connectivity describes tributary-trunk as well as upstream-downstream interactions of water and sediment (Brierley et al., 2006; Fryirs et al., 2007a). The degree of tributary-trunk and upstream-downstream connectivity in the Krom River both interactions influence the structure and functioning of the Krom River and its wetlands

6.1.1 Tributary-trunk interactions in the Krom River

Multiple steeply sloped north and south bank tributaries feed into the Krom River at various locations along its length. The combination of the loss of confinement at the tributary-trunk confluence and the comparatively low longitudinal slope of the Krom River compared to its tributaries leads to the deposition of tributary sediment onto the Krom River valley floor. The normal flow regime of the Krom River is dominated by lengthy low flow periods. This limits the trunk stream's ability to mobilise and transport tributary derived sediment downstream of the tributary-trunk confluence. Similar tributary-trunk interactions have been observed in the Nyl River in the Limpopo Province of South Africa where sediment has accumulated at the distal ends of several tributaries as a consequence of low trunk river flows and the coarse calibre of the tributary derived sediment (McCarthy et al., 2011).

The inability of the Krom River to rework and transport tributary derived sediment along its course has led to the gradual storage of tributary sediment in tributary alluvial fans. Over time these fans partially encroach onto the Krom River valley floor (Haigh et al. 2008; Lagesse, 2017). Lagesse (2017) predicted that the degree to which tributary alluvial fans are encroaching on the valley floor of the Krom River is positively correlated to tributary catchment size. McCarthy et al. (2011) found the volume, timing and calibre of tributary derived sediment, as well as the tributaries discharge regime, to influence the extent to which tributary alluvial fans encroach on trunk streams.

Sediment storage within tributary alluvial fan features forms "buffers" (Fryirs et al., 2007a), and is a common fluvial phenomenon in both semi-arid and humid environments. The presence of tributary alluvial fans has been shown to cause a general reduction in system connectivity over varying timescales, depending on individual system characteristics (Carson et al., 1989; Harvey, 2002; Grenfell et al., 2009; Harvey, 2012; Bracken et al., 2015). Alluvial fans in the

Krom River are having a similar effect on tributary-trunk connectivity. The storage of tributary derived sediment in encroaching alluvial fans limits the efficiency with which sediment moves from tributary rivers to the trunk river, and how tributary sediment is transported along the trunk river. Therefore, the degree of tributary-trunk connectivity within the Krom River system can be considered to be low.

6.1.2 Upstream-downstream interactions

Although the Krom River is characterised by lengthy low flow periods, episodic but short-lived high discharge events are a feature of the systems hydrological regime. Added to this, the nature of the Krom River valley bottom sediment has a large proportion of fairly coarse sediment (sand), which the stream is unable to mobilise and transport during periods of normal flow. Therefore, the transport capacity of the Krom River during normal flow conditions is limited. Rather than being regularly and efficiently transported downstream, sediment is stored in depositional features along the course of the Krom River. This further contributes to the low upstream-downstream connectivity of the river.

The increase in the discharge of the Krom River during temporary periods of high flow, or during episodically experienced flood events, leads to an increase in the stream's capacity. As a result, stored sediment is mobilised and transported downstream with a greater level of efficiency compared to during sustained low flow conditions. The short-lived nature of high discharge events means that the low discharge conditions quickly return.

The reduced transport capacity of the Krom River during normal flow means that following a high discharge event, large quantities of sediment are deposited a short distance downstream from where the sediment was originally mobilised. Sediment is therefore irregularly mobilised and transported along the course of the trunk stream of the Krom River, with transport being dependent on the occurrence of isolated flood events. As a result, sediment moves along the trunk stream of the Krom River in an extremely dis-connected manner, such that the movement of sediment through the system can be likened to a “jerky conveyor belt” (Ferguson, 1981). A possible implication of the high degree of longitudinal dis-connectivity of the Krom River is that the Churchill Dam, located 25 km downstream of the study site, will not fill with sediment as quickly as previously predicted (Haigh et al., 2008).

6.2 Erosion and deposition as agents of longitudinal slope adjustment

Erosion has been identified as a slope lowering process, with the bed of eroded gullies typically having a lower longitudinal slope gradient than the pre-erosional land surface (Patton & Schumm, 1975; Ellery et al., 2008). Deposition results in aggradation, which leads to longitudinal slope steepening (Patton & Schumm, 1975; Ellery et al., 2008).

While research has commonly investigated the causes of gully erosion, much less effort has been expended in determining/evaluating how gully erosion and subsequent deposition may be influencing the geomorphic structure and evolution of a landscape. Exceptions include Grenfell et al. (2012), who investigated the influence of depositional floodouts on the morphology of the semi-arid Karoo, and Lagesse (2017), who recently proposed that cut-and-fill cycles are a major geomorphic control on the presence of wetlands within the Krom River system. Like Patton & Schumm (1975) and Schumm (1979), Lagesse (2017) viewed the lowering of the longitudinal slope of the Krom River through gully erosion as an ongoing and intrinsic catchment process.

Based on her research in the Krom River, Lagesse (2017) proposed a novel conceptual "cut-and-fill model" of wetland formation in semi-arid environments. Lagesse (2017) focused extensively on the influence and implications of the presence of discontinuous gully erosion (cutting) in longitudinal slope adjustment and valley widening. Despite her focus on the cutting phase, Lagesse (2017) acknowledged the importance of the filling phases in the evolution of the Krom River and provided a limited but incomplete explanation of the controls and implications of the filling phases. Therefore, to add to the understanding of how cut-and-fill cycles are interacting to influence the structure and function of the Krom River and its wetlands, an adapted version of the cut-and-fill conceptual model of wetland formation proposed by Lagesse (2017) is presented (Figure 6.1) More extensive explanations of the depositional processes following gully initiation are also provided.

As shown by Lagesse (2017), and by the longitudinal surveys of the study site of this project, the incursion of tributary sediment onto the Krom River valley floor locally alters the streams longitudinal slope; lowering the slope in an upstream direction and steepening it in a downstream direction (Figure 6.1, a). The localised alteration of the slope of the Krom River greatly increases the likelihood that erosion thresholds will be exceeded during flood events (Schlegel, 2017). This is because the over steepened longitudinal slope, downstream of the locus of tributary sediment deposition, is too steep for the increase in discharge during flood events; hence, erosion occurs reducing the slope (Schlegel, 2017). Therefore, during high discharge events, gully erosion is initiated where the Krom River is locally steepened by encroaching tributary alluvial fans (Figure 6.1, b). As the gully propagates upstream, it mobilises significant amounts of sediment, and erodes to bedrock, gradually lowering the longitudinal slope of the Krom River, such that it has a lower slope gradient than the pre-erosional land surface (Figure 6.1, b). The slope gradient of the bed of the gully that exists within the upstream reach of the study site for this project has a lower slope gradient (0.1 %) than the floodplain along the same length of river (0.33 %). This illustrates the slope reducing effect of gully erosion in the Krom River.

The dis-connected transfer of sediment along the course of the Krom River and the loss of confinement experienced at the toe of gullies means that the sediment mobilised by upstream propagating head-cuts is deposited in large quantities at the toe of gullies (Figure 6.1, b). This recently deposited sediment forms wide depositional reaches downstream of discontinuous gullies , which are termed "floodouts" (Tooth & McCarthy, 2007). The wide and unconfined depositional reach present within the study site is an example of a modern floodout feature of this nature. Within the Kompanjiesdrif basin about 4 km upstream of the present study site, Lagesse (2017) found a layer of sandy clastic sediment across the valley (~ 200 m wide) and along most of the length (~ 1.5 km) of the Komanjiesdrif basin. This sandy layer was buried by 1 to 3.5 m of organic sediment. The carbon date of the sediment immediately below this sand layer was 7000 (+/- 30) BP (Lagesse, 2017). It appeared as though this buried layer of clastic sediment covered much of the valley floor of the Kompanjiesdrif basin, suggesting that it is the remains of an ancient floodout that would have developed at the distal end of a large discontinuous gully. This is evidence that the deposition of significant amounts of eroded sediment at the toe of gullies is an ancient and ongoing process in the Krom River system.

The sediment deposits within floodouts in the Krom River are thickest immediately downstream of the loss of gully confinement, and thin progressively with distance downstream from the gully toe (Figure 6.1, c). This pattern of deposition locally steepens the longitudinal slope of the Krom River. Along with the influence of encroaching alluvial fans on local longitudinal slope, the formation of over steepened floodouts along certain reaches of the Krom River is an additional localised slope steepening mechanism that promotes likelihood of erosion within this system. The results of this study suggest that this is an important control on the initiation of cut-and-fill cycles in the Krom River, and should be viewed in conjunction with the role of tributary alluvial fans (Lagesse, 2017, Schlegel, 2017) if the structure and functioning of the Krom River and its wetlands are to be fully understood.

According to the conceptual model proposed by Lagesse (2017), after a cycle of erosion and deposition is completed, tributary sediment accumulation on the trunk valley floor once again begins to locally steepen the longitudinal slope of the Krom River, and eventually the formation of a new gully, in a novel location, is initiated (Figure 6.1, d). The results of this project suggest that locally steepened floodouts will have a similar influence on the initiation of future cycles of cutting and filling. Thus, the cut-and-fill process is repeated (Figure 6.1, a-d). The initiation of gullies in novel locations serves to widen the Krom River valley floor over extended periods of time (Lagesse, 2017). Therefore, over time scales of thousands to tens-of-thousands of years, erosion and the dis-connected transfer of sediment through the Krom River system, interact to lower the longitudinal slope of the Krom River, and to widen the trunk stream valley (Lagesse, 2017).

A reduction in longitudinal slope steepness and a wide valley bottom are conditions that slow the movement of water through a fluvial system, and have been considered vital prerequisites for the formation of wetlands in semi-arid environments (Tooth et al., 2002; Tooth et al., 2007; Ellery et al., 2008; McCarthy et al., 2011). Therefore, cut-and-fill cycles along the course of the Krom River appear to create an environment suitable for the establishment of palmiet wetlands (Figure 6.1, e).

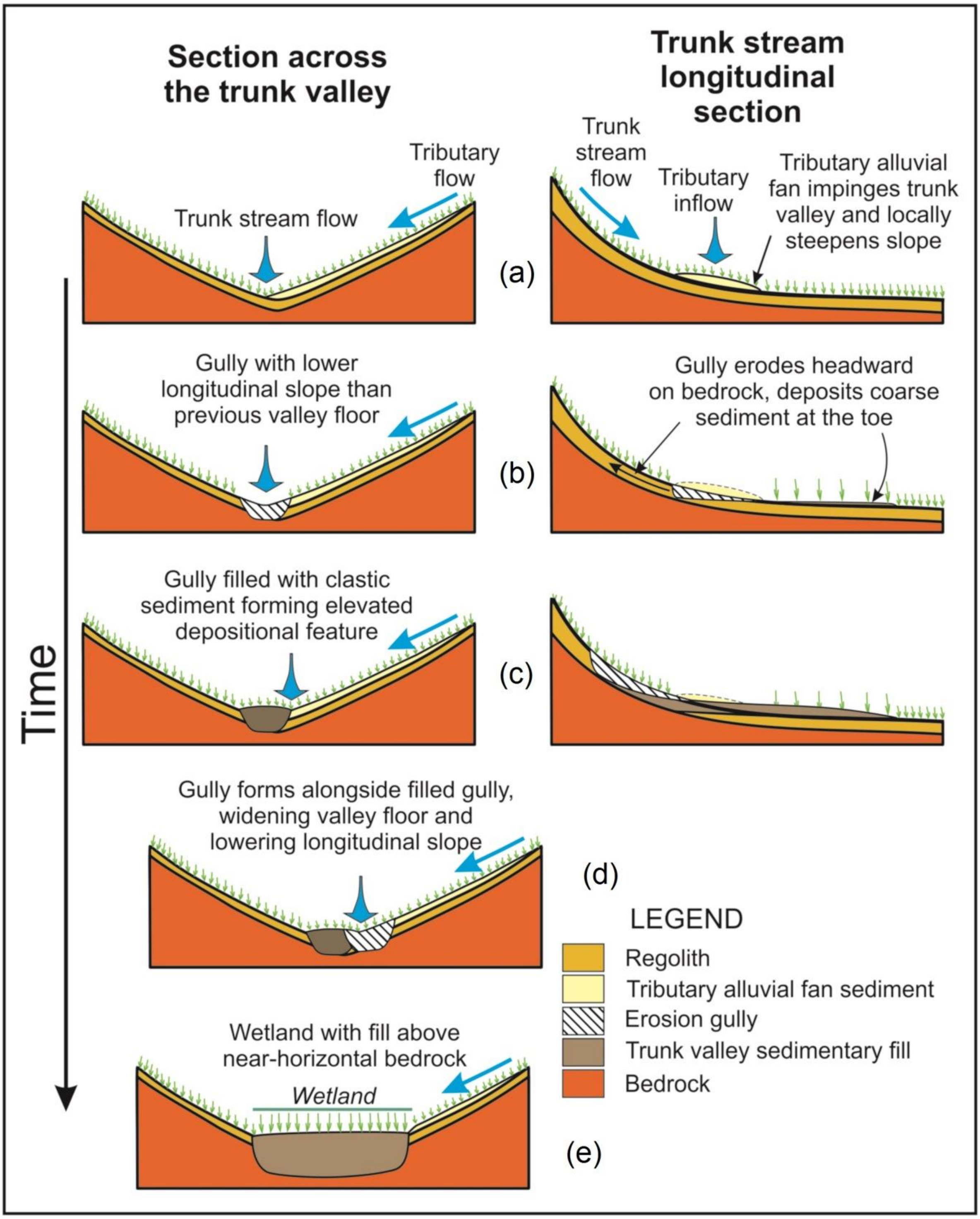


Figure 6.1: *A simplified diagram of how erosion and deposition lead to longitudinal slope adjustment, valley widening and wetland formation in the Krom River (adapted from Lagesse, 2017).*

6.3 The role of floodout formation in long-term landscape evolution in the Krom River

Floodout development in the Krom River, as a result of inefficient upstream-downstream sediment transport appears to be an important process in the long-term evolution and adjustment of the Krom River system. This process should be considered in conjunction with the role that tributary alluvial fans play in initiating cut-and-fill cycles in the Krom River if landscape change and wetland formation are to be fully understood. A conceptual model of the influence of floodout formation on long-term landscape evolution in the Krom River was developed (Figure 6.2) which incorporates the role of valley floor deposition on slope steepening and gully initiation.

The depositional phase of this conceptual model illustrates the implications of the dis-connected transfer of sediment along the course of the Krom River on the systems longitudinal slope. The sudden decrease in discharge associated with the loss of gully confinement, and the selective nature of sediment transport by streams (Carson et al., 1989; Wathen et al., 1997; Bracken et al., 2015) leads to an accumulation of material at the toe of large gullies. This creates expansive floodout features that cover the Krom River valley floor with stored clastic sediment. The sediment deposits along the floodout features are thickest near the toe of the gully and thin progressively with distance downstream from the loss of gully confinement. Therefore, as sediment accumulates within floodout features the longitudinal slope along the floodout increases over time (Figure 6.2).

Local slope steepening through floodout formation is clear in the unconfined reach of the study site such that a large quantity of the sediment, mobilised during the 2012 flood event was deposited at the toe of an actively eroding gully. As a result, the longitudinal slope of the floodout was steepened from 0.28 % to 0.37 %. The sediment derived from the 2012 flood event would have been deposited over older flood deposits. These older deposits are likely to have locally raised the longitudinal slope of the floodout from a lower slope gradient to a gradient of 0.28 %.

Patton & Schumm (1975) and Schumm (1979) suggest that valley floor stability is strongly influenced by sediment storage due to its role in altering longitudinal slope steepness. Stability in this context refers to the likelihood of a geomorphic threshold being transgressed (Patton & Schumm, 1975). Essentially, steep slopes lead to unstable landscapes and landforms which are inherently prone to erosion. Erosion is the system's way of lowering its slope to an appropriate gradient for the available discharge (Ellery et al., 2008). Therefore, according to Patton &

Schumm (1975) and Schumm (1979), as well as other research that has attributed valley floor instability to slope steepness (Schumm 1981; Schumm 1994; Phillips 2003; Fuller & Marden 2010), sediment deposition onto floodout features locally increases valley floor instability in the Krom River, leaving it prone to erosion as the system attempts to achieve a graded state. The blue downward facing vertical arrow in Figure 6.2, b represents the transgression of a geomorphic threshold for erosion due to an over steepened slope which has formed in response to the deposition of sediment onto a floodout feature.

The erosion phase of the conceptual model shows the effect of the formation of a gully along a floodout feature in response to local over steepening. When a critical slope value is reached along over steepened floodouts, the magnitude of flood events generally experienced in the Krom River will be sufficient to support erosion. In this state, floodouts downstream of discontinuous gullies in the Krom are event sensitive and are “primed for change” (or "unstable"; Fryirs et al., 2007a). When the Krom River experiences a high enough flow, gully erosion will be initiated along the unstable but previously un-gullied floodout features (Figure 6.2, b). The response of an unstable fluvial system to a single large flood event can therefore have a significant effect on stream structure (Schumm, 1979).

After a gully is initiated it rapidly propagates upstream along over steepened floodouts. The ancient gullies Lagesse (2017) noted within the Kompanjiesdrif basin each contained a small amount of clastic sediment that was deposited immediately over bedrock. This implies that the gullies in the Krom River erode through the regolith and reach bedrock before flood derived sediment covers the exposed gully floor. This process serves to gradually lower the longitudinal slope of the bedrock of the Krom River (Figure 6.2). Reaches of the Krom River that have recently had their longitudinal slopes lowered because of gullying will be more stable than before gullying occurred (Figure 6.2, b).

Throughout time certain reaches of the Krom River will be actively eroding, with the sediment generated from this erosion accumulating in new floodout features at the toe of the gullies. This downstream sediment accumulation causes slope steepening in a new reach. Therefore, in response to one reach of the Krom River being made stable through longitudinal slope lowering processes, another reach is made increasingly unstable in response to the downstream storage of previously mobilised sediment. Such reaches will experience increased instability following each flood event until failure occurs. Therefore, the cut-and-fill process is repeated along new reaches of the Krom River over time (Figure 6.2). In this manner, the dis-connected longitudinal transfer of sediment along the course of the Krom River, and the associated periods of valley floor instability, contribute to a long-term reduction in the longitudinal slope of the Krom River. The occurrence of gullying therefore appears to be a necessary process that facilitates longitudinal self-recovery within the Krom River system. This longitudinal self-recovery ultimately makes the Krom River a more stable fluvial environment that is suited to the long-term formation of wetlands.

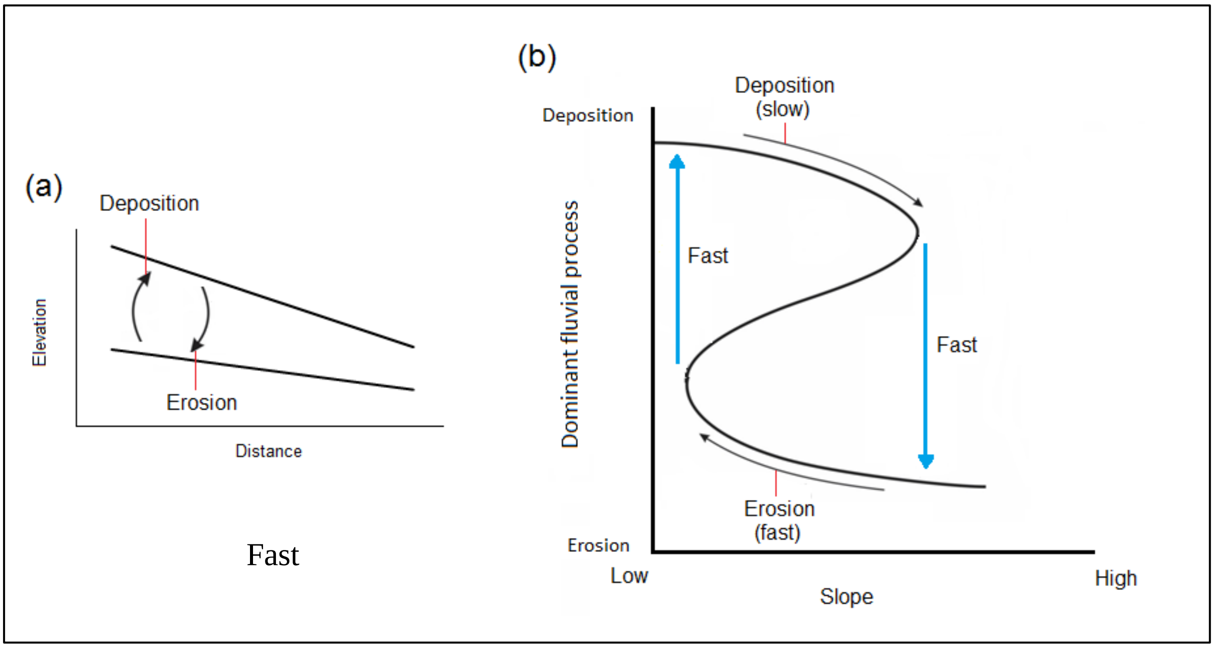


Figure 6.2: *Conceptual model of the influence of floodout formation on the long-term evolution of the Krom River, including (a) showing variation in longitudinal slope in response to erosion and deposition and (b) showing a threshold catastrophe curve leading to sudden slope reduction during flood events.*

6.4 Timescales of cutting and filling in the Krom River

Figure 6.2 does not consider the timescales over which slope steepening and slope reduction take place within the Krom River. In the Kompanjiesdrif basin, Lagesse (2017) dated sediment from the bases of a series of ancient gullies, which ranged in age from 460 BP to 7040 BP, and used the time gaps between each dated gully to estimate that new cutting phases are initiated every 100 to 1000 years. This estimation represents the rate at which sediment deposition increases the longitudinal slope along floodout features, such that gully erosion is initiated. It is however acknowledged in her research that this is a very tentative estimation of the timescales of a cut-and-fill cycle in the Krom as climate variability and a lack of agreement between sedimentary facies that act as historical records make it difficult to interpret cut-and-fill timescales (Lagesse, 2017). Cut-and-fill cycles taking place over similar timescales to those in the Krom River were noted in the Kanab Creek in Northern Colorado by Nelson & Rittenour (2014), while Patton & Schumm (1975) suggest that cut-and-fill cycles in in Piceance Creek occur of over much shorter timescales.

Research on cut-and-fill systems has suggested that gully initiation and propagation is a relatively rapid process (Patton & Schumm, 1975; Patton & Schumm, 1981; Nelson & Rittenour, 2014). The rapid nature of gully initiation and propagation suggests that variability in the rate at which cut-and-fill cycles take place is dependent on the factors controlling the longer-term filling phases. In the Krom River sediment accumulation and slope steepening of floodouts is dependent on the occurrence of episodic high discharge events such that filling phases take place over extended periods of time. As the gullies present within the Kompanjiesdrif basin pre-date human settlement in the area (Lagesse, 2017) it would appear that fluctuations in valley floor stability are intrinsic to the Krom River, and that gully erosion is a natural response to the periodically unstable Krom River valley floor.

6.5 Palmiet as an agent of dis-connectivity: Implications for longitudinal slope adjustment

Floodout features downstream of discontinuous gullies provide a suitable environment for the medium-term establishment of wetlands along the Krom River. The presence of wetland vegetation within the unconfined reach of the present study site provides evidence of the establishment of zones of intact wetland along reaches of the Krom River that are dominated by recently deposited flood derived-sediment.

Due to the tendency of wetlands to act as fluvial sediment traps (Phillips 1989; Zierholz et al., 2001; Ellery et al. 2008) the presence of wetlands along floodouts in the Krom River promotes increased sediment storage (Barclay, 2016). Floodout wetlands along the course of the Krom River therefore influence the degree of longitudinal connectivity in the system. Sediment storage within wetlands along the Krom River is enhanced by the abundant presence of palmiet. This is because palmiet wetlands are known for having remarkable sediment trapping capabilities (Sieben, 2012; Nsor & Gambiza, 2013; Job, 2014; Barclay; 2016). Due to their tendency to promote sediment storage along already dis-connected reaches of the Krom River, floodout wetlands contribute to the long-term longitudinal slope adjustment that is occurring along the course of the Krom River (Figure 6.2).

Ultimately, the presence of palmiet wetlands increases sedimentation downstream of gullies, which increases the rate of longitudinal slope steepening of the floodouts. Once the longitudinal slope along the floodouts exceeds a critical slope steepness value, and gullying occurs, these wetlands will be destroyed, as longitudinal slope lowering through channel incision occurs. Therefore, despite being prone to destruction by gully erosion in the medium-term, these wetlands assist in the longitudinal self-recovery of the Krom River, such that it is a more stable system over time, and ultimately suited to the long-term re-establishment of valley-bottom wetlands within the system.

6.6 Erosion control structures in the Krom River

Gully erosion within wetland and river environments is commonly considered a consequence of human land use practices within the catchment (Sidorchuk, 1999; Nyssen et al., 2002; Poesen et al., 2002). At this point, the recognition of erosion as a precursor to wetland formation is a relatively new concept that is limited to only a few case studies (Grenfell et al., 2009; Job, 2014; Silbernagel, 2014; Lagesse, 2017). Furthermore, there is a strong emphasis on the negative consequences of gully erosion in wetland management tools produced for South Africa, such as “The Wetland Management Series” (Dada et al, 2007). One of the handbooks in this series, known as “WET-Health: A technique for rapidly assessing wetland health”, (Macfarlane et al., 2008), focuses particularly strongly on the negative consequences of wetland erosion. This combination of factors has led to an uncompromising view by wetland managers that gully erosion is the principle cause of permanent wetland degradation in South Africa. Given this narrative, the inevitable response of wetland managers is to attempt to preserve wetland integrity by preventing the initiation and propagation of gullies. To date, Working for Wetlands has spent about ZAR 1 billion on wetland protection projects, often involving interventions designed to nullify erosion impacts (DEA, 2017).

The importance of the Krom River valley-bottom wetlands for the water supply of the Port Elizabeth metropole, and the destruction of large areas of palmiet wetland in the system by extensive gully erosion, prompted Working for Wetlands to intervene in the system to protect the remaining Krom River wetlands (DEA, 2017). To this end, between 2000 and 2007 Working for Wetlands constructed a number of erosion control structures along the course of the Krom River (Haigh et al., 2008; DEA, 2017). These structures are in the form of rock gabions, often constructed in combination with concrete, built across the Krom River at selected locations. The flooded areas behind these structures serve to nullify the propagating head-cuts of selected gullies, which were considered threatening to upstream palmiet wetlands. Despite the structures along the Krom River experiencing regular and costly flood damage (Haigh et al., 2008), these structures have been successful in stabilising gullies and preventing the initiation of new gullies along certain reaches of the Krom River (DEA, 2017).

The conceptual longitudinal slope adjustment model presented in this study suggests that the Krom River is a complex adaptive system that is inherently capable of longitudinal self-recovery. Self-recovery in this context refers to a river's ability to respond to changing inputs of water and sediment, which allow river systems to enter periods of relative geomorphological stability (Powell, 2016). Ongoing cut-and-fill cycles are the Krom Rivers' natural response to variation in water and sediment supply, and are therefore important longitudinal self-recovery processes within the system. Given that erosion control structures in the Krom River prevent the initiation of periods of cutting, these structures are likely interfering with the system's ability to recover and achieve a state of relative geomorphic stability.

Inherent self-recovery via erosion and deposition is not a new concept in fluvial geomorphology (Schumm & Litchy, 1965; Schumm, 1973; Schumm, 1981). Powell (2016) recently added to the literature dealing with self-recovery in fluvial geomorphology through the novel use of a modified version of the concept of adaptive cycles. Although the concept of an adaptive cycle was originally developed to explain complex interactions within social and natural systems (Gunderson & Holling, 2001), Powell (2016) used this concept to explain the interaction of fluvial form and processes in the Baviaanskloof River floodplain. It is believed the concept of an adaptive cycle used by Powell (2016) can similarly be used to explain reach scale longitudinal self-recovery via cut-and-fill cycles in the Krom River; and in doing so, will provide insight into the effects of human intervention on the long-term health of the Krom River and its wetlands.

6.6.1 Adaptive cycles in the Krom River

An adaptive cycle is considered in the context of two axes, the connectedness of internal controlling variables and processes (the x-axis) and the internal potential for change (the y-axis). In the context of an adaptive cycle the term "connectedness" refers to the interaction between process and variables that result in the ability of a system to self-regulate its internal condition based on negative feedback (Gunderson & Holling, 2001; Powell 2016). The internal potential for change towards another internal future stable state relates to the ways that systems can be drawn into single or alternative stable states once natural or artificial thresholds are crossed. On the basis of these 2 axes, Gunderson (2008) describes four phases which make up the adaptive cycle (Figure 6.3). These are:

1. The exploitation (or growth) phase. In this phase, the system experiences rapid change in a predictable manner following system collapse and reorganisation (r).

2. The conservation phase, where structural complexity is high and the system's variables and processes become highly connected and tend to operate slowly (K).

3. The destruction (or release) phase, which is initiated when a threshold is breached such that the system experiences a collapse, and accumulated resources are released from storage, and system change is inevitable (Ω).

4. The reorganisation phase when the system adjusts to its new conditions by reorganising its available resources for the next exploitation phase (α).

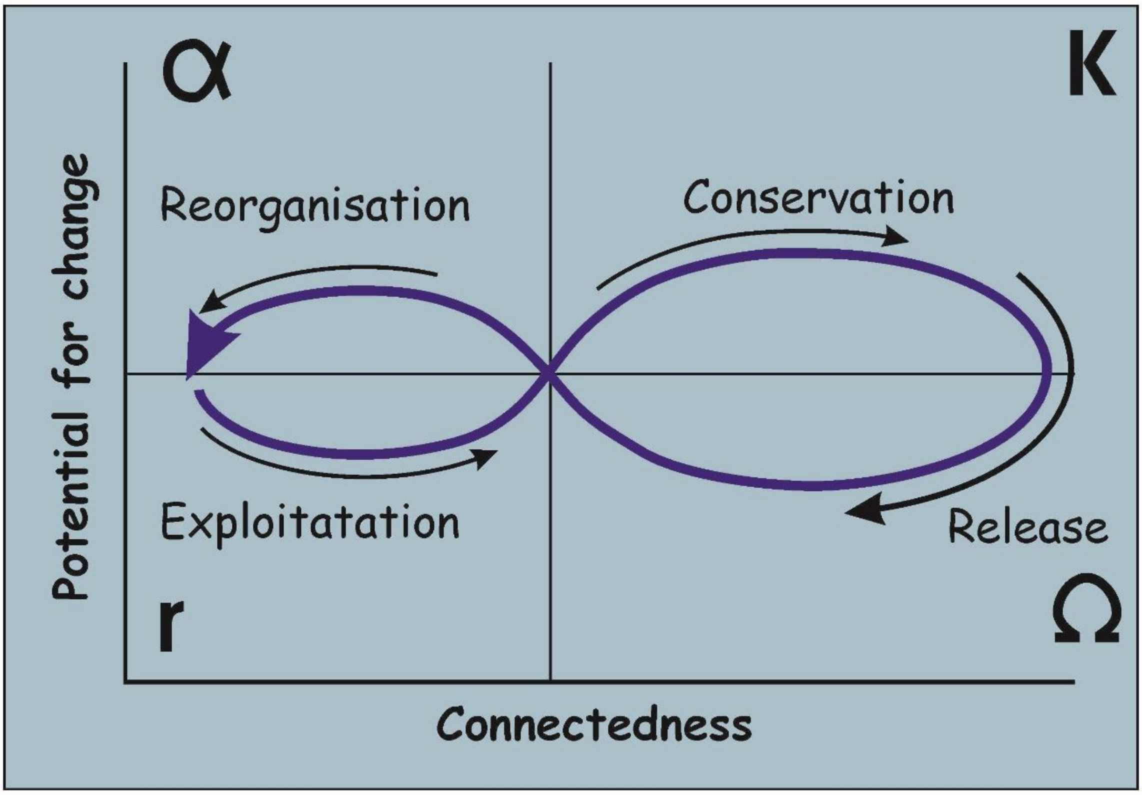


Figure 6.3: *The adaptive cycle that is used to describe phases of local stability and instability in natural systems that might contribute to overall stability in the long term (adapted from Gunderson and Holling, 2001).*

Reach scale longitudinal self-recovery via cut-and-fill cycles along the Krom River can be explained through the four phases of the adaptive cycle mentioned above. It is useful to consider a reach of the Krom River without an erosion control structure along its course, and that has recently experienced intense channel incision. Following incision, the Krom River channel enters a reorganisation phase (α), which is characterised by geomorphological restructuring of the channel. This restructuring is associated with the redistribution of sediment mobilised during the cutting process. Following reorganisation, the reach enters the growth phase (r), which is characterised by the formation and growth of in-channel depositional floodout features. These features form due to the predominantly low flow conditions of the Krom River, the loss of confinement at the toe of the recently incised channel, and the lowered longitudinal slope of the reach following cutting. During this phase the channel is stable, and erosion is unlikely to occur again, thus, indicating a shift towards the conservation (K) phase (Powell, 2016). During the conservation phase sediment transport is an important process. The floodout features form barriers, such that the downstream transport of sediment is limited, and gradual reach scale aggradation (filling) becomes the dominant process.

The filling process occurs over long-periods of time. Therefore, the channel stability associated with the growth (r) and conservation phase (K) is maintained for extended periods. It is during these phases that palmiet wetlands are able to re-establish in the medium-term, after experiencing destruction during the previous release phase.

The pattern of gradual filling during the conservation phase leads to an increase in channel slope gradient. As a result, slope thresholds for erosion are gradually approached, and the Krom River channel becomes increasingly unstable. Eventually a large enough flood will breach the slope threshold for erosion and the reach in question will enter the destruction/release phase (Ω). This phase involves the rapid structural breakdown of the channel via erosion. The cutting process associated with the release phase in the Krom River serves to lower the longitudinal slope of the reach, and contributes to the gradual widening of the valley (Lagesse, 2017). Following cutting the reach in question is more geomorphologically stable, and a new phase of reorganisation (α) is initiated (Powell 2016).

6.6.2 Influence of erosional control structures on longitudinal self-recovery in the Krom River

Erosion control structures along the course of the Krom River act as permanent barriers that limit headward propagation of erosional nick points and also trap sediment generated by upstream erosion. The effect of permanent sediment trapping for reaches downstream of the erosion control structures is that gully filling via floodout formation is inhibited. The presence of exposed bedrock on the bed of gullies downstream of erosion control structures provides evidence that the structures are limiting gully filling in the Krom River (Hermon, 2016). Minimal filling downstream of gullies means that natural slope steepening, an important part of the adaptive cycle of the Krom River, is limited along certain reaches.

Upstream of erosion control structures trapped sediment leads to gradual increases in channel gradient. As with the growth of floodout features (in reaches uninfluenced by erosion control structures) during the conservation (K) phase of the adaptive cycle, the channel upstream of erosion control structures gradually approaches threshold gradients for erosion. However, in reaches where erosion control structures exist, the release (Ω) phase is prevented, and cutting cannot take place. Instead of experiencing collapse via cutting, which leads to the continuation of the adaptive cycle, the self-recovery of the Krom River is paused in time as the channel upstream of the erosion control structures is prevented from entering the reorganisation (α), growth (r) and conservation (K) phases of the adaptive cycle.

Therefore, erosion control structures in the Krom River are preventing the loss of existing palmiet wetlands but are playing an active role in limiting periods of cutting (upstream of structure) and filling (downstream of structures) that ultimately lead to reductions in longitudinal slope steepness and to valley bottom widening; conditions that are likely to be associated with the long-term re-establishment of palmiet wetlands in a more geomorphologically stable setting.

CHAPTER 7: **CONCLUSIONS**

This project aimed to investigate the degree of longitudinal dis-connectivity of the Krom River by determining where sediment, mobilised by the cutting of gullies, was deposited, in order to add to the understanding of how cut-and-fill cycles influence the structure and functioning of the Krom River and its wetlands. It is believed that this aim was well met. This is because this project led to a sound understanding of the relationship between cut-and-fill cycles and landscape dis-connectivity, culminating in the proposal of a new conceptual model displaying the importance of floodouts on the systems structure and functioning.

Table 3.1: *Table showing each objective and the Tables/Figures that contributed towards achieving them*

Objective	Figure/Table
1. Determine the spatial extent of erosion and deposition along a selected reach of the Krom River.	Figure 5.2, Figure 5.3
2. Determine the quantity and particle size distribution of sediment eroded and deposited within a selected reach of the study site.	Table 5.1, Table, 5.2, Figure 5.4, Figure 5.5, Figure 5.6, Figure 5.7, Figure 5.8, Figure 5.9, Figure 5.10
3. Determine how erosional and depositional phases initiate and interact to influence the structure and functioning of the Krom River system over time, and examine the implications of this for future erosional and depositional cycles within wetland management	Figure 5.11, Figure 6.1, Figure 6.2, Figure 6.3

Through a reach scale analysis of the Krom River's depositional processes, and an understanding of how erosion and deposition interact within the system, it can be concluded that emerging valley floor instability in the Krom River develops within the geomorphic system through intrinsic landform changes that are a result of the dis-connected longitudinal transfer of previously eroded sediment. Current and historic gully erosion experienced by the Krom River are natural and necessary responses to the transgression of geomorphic thresholds. Filling phases can therefore be considered a mechanism that promotes the initiation of new cut-and-fill cycles in novel locations, such that ultimately, longitudinal slope reduction and valley widening gradually occurs over geologic timescales throughout the Krom River system. The slope adjustment model presented in this study suggests that the Krom River is most capable of longitudinal self-recovery when erosion and deposition are allowed to naturally occur within the system.

A limitation of this study is that a reach scale appraisal of longitudinal connectivity was used to gain an understanding of the dynamics of the filling phases, and the results of this focused study were used to make broad inferences about the role of filling phases on the structure and function of the Krom River. Although the inferences made from this study are supported by data presented by Lagesse (2017), further research is needed to investigate whether this proposed mechanism of long-term slope reduction is applicable to the entire Krom River, and to other systems, with similar tectonic and geologic settings.

Given this understanding, the long-term influence of erosion control structures in the Krom River should be reconsidered. The structures along the Krom River have proven to be useful for the short-term conservation and preservation of existing palmiet wetlands. However, this study suggests that their implementation likely limits the system’s ability for long-term longitudinal self-recovery. The use of erosion control structures in the Krom River appears to be a result of what Schumm (1994) considers to be two common misconceptions relating to fluvial systems; (1) fluvial system are inherently stable, and any fluctuation in stability is unnatural, and (2) when unstable, the occurrence of any landscape change is catastrophic, and will not cease without intervention. The models proposed in this study suggest that these misconceptions could easily be avoided through a long-term understanding of the system's dynamics.

A dilemma therefore exists within the Krom River, and in wetland conservation in general. Landscape manager’s need to decide if it is best to protect and conserve wetlands over timescales relevant to the life span of current generations, or if it is more responsible to allow wetland degradation to occur within our lifetimes, as part of natural long-term fluvial processes that promote wetland formation in more stable setting.

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APPENDIX

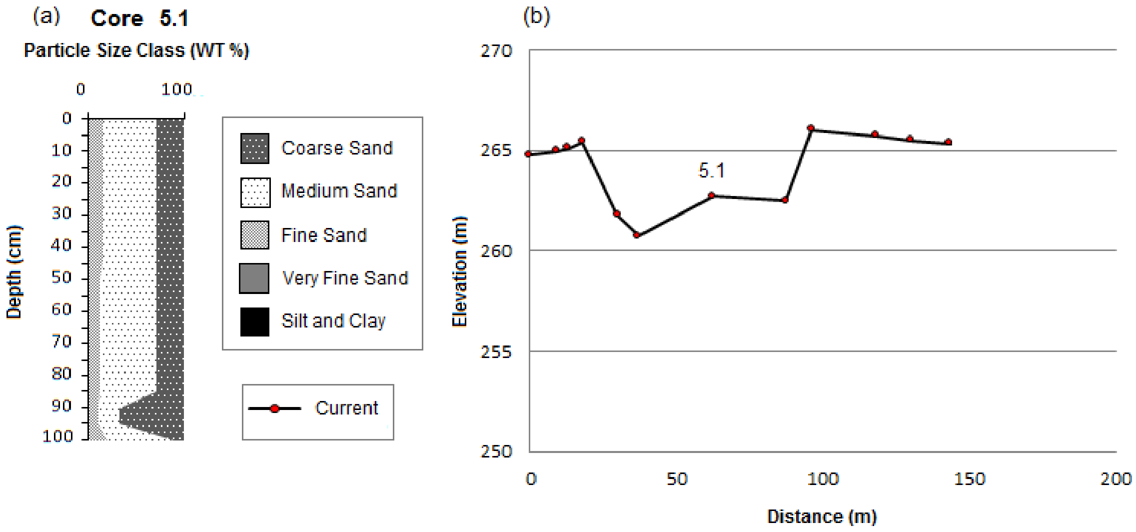

Figure 8.1: *Particle size distribution of Cores 5.1 (a) and cross section profile of T5 showing the location of this core (b).*

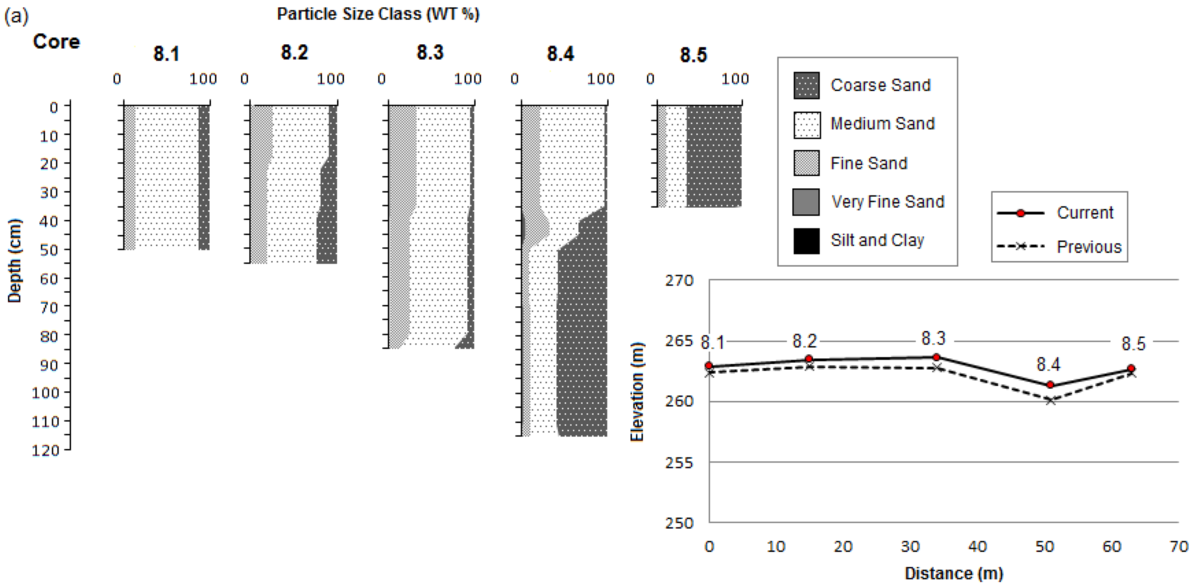

Figure 8.2: *Particle size distribution of Cores 8.1 to 8.5 (a) and cross section profile of T8 showing the location of these cores (b).*

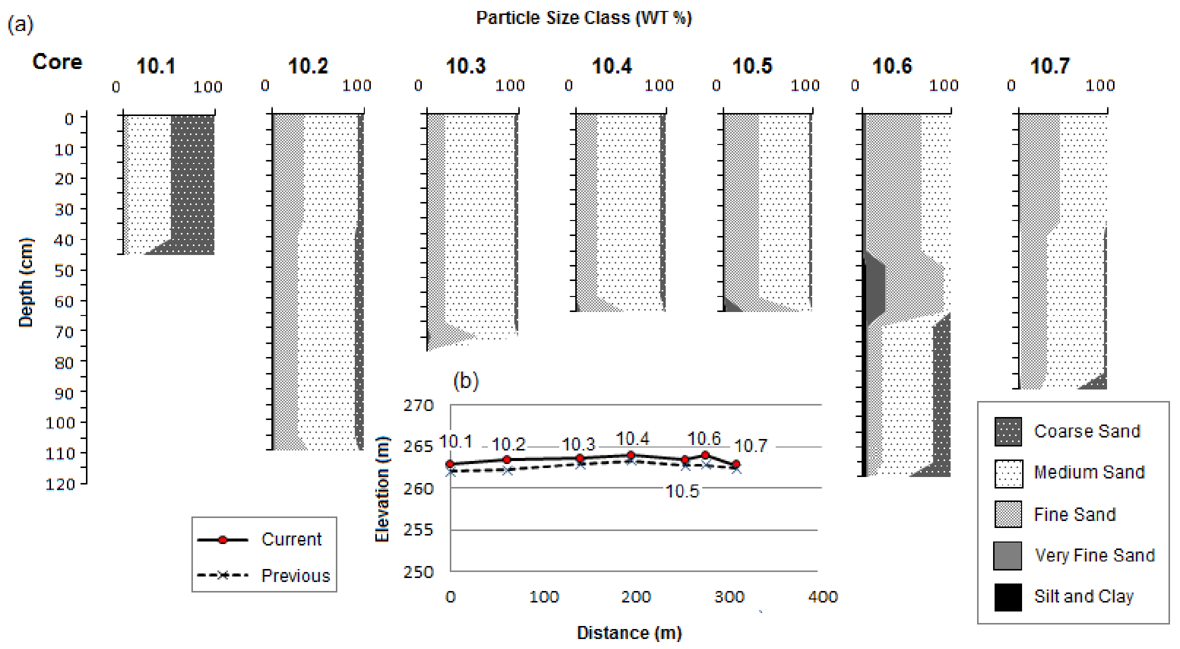

Figure 8.3: *Particle size distribution of Cores 10.1 to 10.7 (a) and cross section profile of T10 showing the location of these cores (b).*

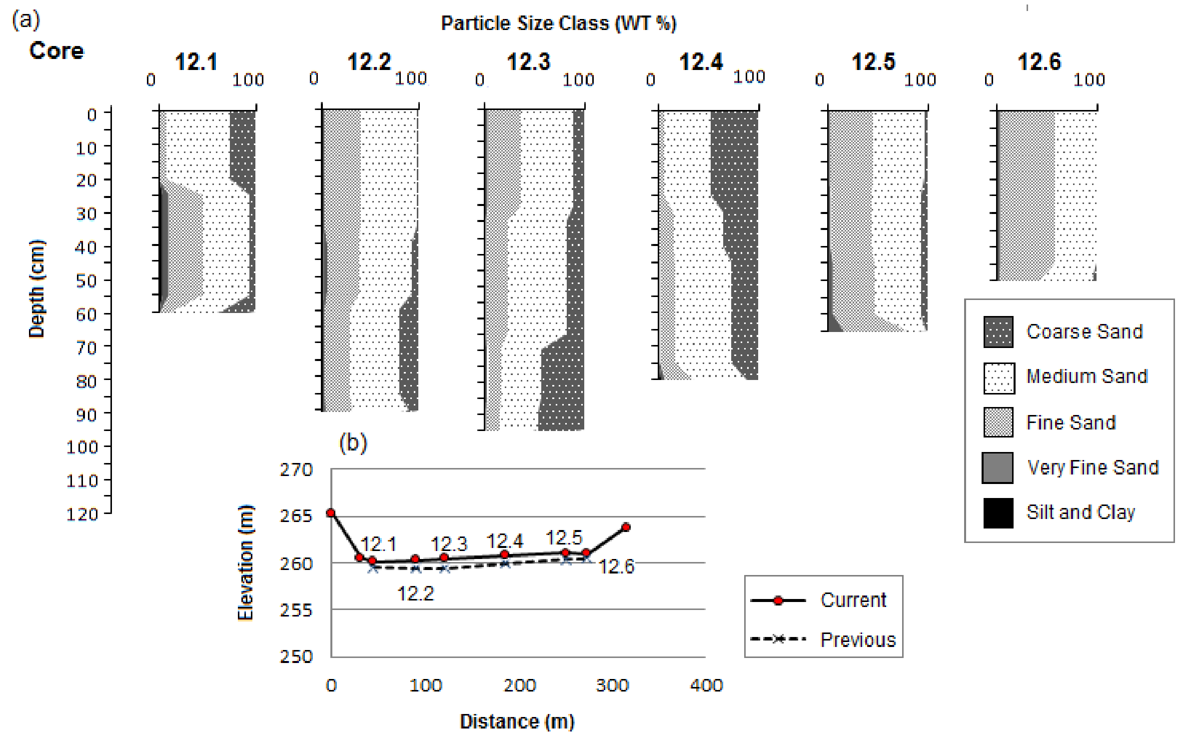

Figure 8.4: *Particle size distribution of Cores 12.1 to 12.6 (a) and cross section profile of T5 showing the location of these cores (b).*