# DISCONTINUOUS GULLY EROSION AS A MECHANISM OF WETLAND FORMATION: A CASE STUDY OF THE KOMPANJIESDRIF BASIN, KROMRIVIER, EASTERN CAPE, SOUTH AFRICA

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## ABSTRACT

The Kompanjiesdrif basin is an unchannelled valley bottom palmiet wetland located near the headwaters of the Kromrivier in the Eastern Cape of South Africa. The wetland itself is underlain by Bokkeveld shales with the bordering mountain ranges comprising more resistant Table Mountain Group quarzitic sandstones. The valley is relatively planar and broad in form over a width of approximately 200 m. None of the existing controls that are considered to lead to valley widening and longitudinal slope reduction are immediately apparent. The basin lies on the Post Africa II erosion surface; with no evidence of a resistant lithology which might act as a local base level, limiting rates of vertical erosion and inducing lateral planing in upstream reaches via a meandering channel. The possible role of sagging of the basin due to long term deep chemical weathering of bedrock is discounted as the lithologies in the basin are sedimentary in origin and thus not susceptible to chemical weathering. The degree to which climate and sea level changes affected rates of incision and subsequent slope reduction is unclear, although their potential influence should be acknowledged.

This study examined the geomorphic dynamics as discerned from the sedimentary record and morphology of the wetland basin, which provide a snapshot into the long-term processes which lowered the longitudinal slope and widened this valley. Coring within the wetland to depths of 1 - 3.3 m revealed that the sedimentary fill generally comprised an upward fining sequence, with sand or fine sand at the base, grading into silt and clay and organic material in the upper sections of cores. Occasional instances of multiple fine sand layers were observed in a few of the cores. An increase in the organic content of material from the north to the south side of the wetland and the occurrence of multiple thin layers of sand in the stratigraphy, highlighted the role of the northern tributary alluvial fans in influencing valley form. Sediment from north bank alluvial fans seem to periodically, partially impound the wetland basin.

Surveyed transects across the wetland basin along with subsurface coring to the depth to refusal, illustrated a localised increase in longitudinal slope downstream of the nodes of tributary alluvial fan deposits, which impinge on the trunk stream basin. Coupled with the presence of deep, drowned, trench-like features (up to 8 m deep) beneath floating mats of palmiet, which were predominantly free of sedimentary fill and found opposite tributary alluvial fans; confirmed that the northern tributaries play a major role in the structure and geomorphic dynamics of the basin. The trench-like features appeared to be remnants of deep,

narrow, discontinuous gullies. Dating of sediment from the base of these features (460-7040 BP) confirmed that they were formed prior to European settlement in the area. Therefore, it is suggested that the localised increase in longitudinal slope, caused by sediment deposition on the alluvial fans, transgresses a geomorphic threshold slope and that gully erosion is thus initiated. The process of repeated gully erosion leads to planing of bedrock and longitudinal slope reduction. Gully erosion forms an integral component of a cycle of deposition and incision referred to as "cut-and-fill". During each iteration of the cycle of cutting and filling, gullies form in novel locations leading to gradual valley widening. Over geological time scales, the planing of bedrock and resultant valley widening creates a broad planar valley with a very low longitudinal slope; producing conditions suitable for unchannelled valley bottom wetland formation.

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# CHAPTER 1 – INTRODUCTION

Wetland science has primarily focused on wide-ranging studies in the northern hemisphere as well as having been mostly restricted to moist and temperate environments (Mitsch & Gosselink, 2007; Tooth & McCarthy, 2007). Earlier publications exhibit a geographical bias with North American and European studies being predominant (Mitsch & Gosselink, 2000).

Along with this earlier geographical bias, research has also exhibited a strong ecological bias which has viewed wetlands as ecological systems (Tooth & McCarthy, 2007), although much work has been done on understanding the hydrological factors which contribute to the functioning of wetlands (Tooth & McCarthy, 2007). The advances made in highlighting the role of wetland plants and their capabilities in terms of nitrogen and carbon sequestration; the ability of wetlands to minimise the impacts of large floods and recognition of the fact that wetlands markedly improve the quality and quantity of potable water available, has been instrumental in driving the need for more research and a greater level of respect for the value that wetlands add or provide in a landscape (Kotze et al., 2005).

Taking into consideration the widespread degradation and destruction of wetlands on a global scale, there has been a concerted effort directed toward understanding and highlighting their provision of ecosystem services; in the hope that this understanding would serve to minimise potential future degradation (Mitsch & Gosselink, 2007; Kotze et al., 2005). This approach has met with some success as there has been an increased level of awareness created, which is evident through the increase in legislation produced around wetland related issues such as the Ramsar Convention (Tooth & McCarthy, 2007).

In addition, recognition of the role wetlands play in ecosystem services, such as flood attenuation, has led to widespread wetland restoration globally (Mitsch & Gosselink, 2007; Kotze et al., 2005). With projects in America including the Kissimmee River (Dahm et al., 1995) and projects in Europe (e.g. Straskrabova & Prach, 1998) taking place quite early in the trajectory of wetland restoration science, South Africa has also been active in the field of wetland restoration, where a statutory agency, Working for Wetlands, has spent approximately R 826 million on rehabilitating over 1000 wetlands across the country since 2004 (DEA, 2017).

However, existing paradigms developed in the northern hemisphere explaining the structure and function of temperate and tropical wetlands, have been found to have little bearing on wetlands which occur in dryland environments (Tooth & McCarthy, 2007). The key difference is that in dryland environments an overall negative water balance is experienced. Therefore, larger wetlands are almost invariably reliant on fluvial inputs for most of their water supply (Tooth & McCarthy, 2007). Consequently, there have been claims that fluvial geomorphology is the most relevant theoretical framework to use as a basis for understanding wetland structure, function and dynamics within dryland environments (McCarthy & Hancox, 2000; Ellery et al., 2009). Increasingly there are a number of studies which examine the role of geomorphology in contributing to the origin and dynamics of wetlands. Examples of geomorphological studies which have served this purpose encompass wetlands that occur across the globe (Adamson et al., 1987; Nanson et al., 1993; Finlayson & Kenyon, 2007; Tooth & McCarthy, 2007). More locally, the use of fluvial geomorphology to uncover the structure and dynamics of wetlands in southern Africa is especially useful in aiding rehabilitation efforts and informing management decisions (Ellery et al., 2009; McCarthy et al., 2010; Tooth et al., 2014).

Geomorphic research on wetlands in southern Africa has employed the fundamental principles of fluvial geomorphology to develop several conceptual models on the origin and dynamics of certain wetlands. These include those authored by Tooth et al., (2002, 2004) in which the origin of large floodplain wetlands on the Highveld of South Africa was explained. Grenfell et al., (2008), Grenfell et al., (2010) and Ellery et al., (2012) summarise the continuum of possible interactions between tributary and trunk streams; from a trunk stream blocking a tributary stream to a trunk being blocked by a tributary. Joubert & Ellery (2013) combined information from the above conceptual models to address the origin of the Wakkerstroom wetland, which possesses three distinct hydro-geomorphic units; with the upper and lower reaches comprising floodplain wetlands and the middle reach comprising a trunk-by-tributary impoundment morphology.

Some conceptual models address the occurrence of peat wetlands in southern Africa (McCarthy et al, 1997; Grenfell et al., 2008; Grenfell et al., 2009a; Grenfell et al., 2009b; Grenfell et al., 2010; McCarthy et al., 2011; Ellery et al., 2012). Nevertheless, as argued by Ellery et al (2012), these models do not address the key controls on peat formation. Given the limited number of conceptual models on wetland and peat formation in southern Africa, and due partly to the highly specialised nature of wetland science; the conceptual models developed thus far do not apply to every peatland in the region (Job, 2014; Silbernagl, 2014). Thus, there is a need to conduct further studies on peat wetlands to add to or improve upon the current understanding of this field of wetland science. When this knowledge gap is considered in conjunction with the uses geomorphic studies serve, particularly in southern Africa where degradation of wetlands via erosion is prevalent, the need for more studies on peatlands increases (McCarthy et al., 2010; Tooth et al., 2013).

The above reasoning is particularly pertinent to palmiet wetlands, which are distributed in the Western and Eastern Cape as well as in limited parts of Kwa-Zulu Natal, South Africa (Job, 2014). None of the existing conceptual models apply to certain palmiet wetland systems; as they occur in broad flat valleys in the Cape Fold Mountain belt, with very shallow longitudinal slopes which are thought to have formed due to valley widening and planing of bedrock producing the lower relative stream power needed for wetland formation (Job, 2014). However, the only known mechanism that exists which produces broad flat valleys in otherwise mountainous catchments is the model proposed by Tooth et al. (2002) and (2004) for floodplain systems on the South African Highveld. This model entails planing of bedrock and valley widening via channel meander migration in a floodplain setting. Yet, palmiet wetlands observed do not have any visible meandering channels present. Moreover, the model proposed by Tooth et al. (2002, 2004) does not explain peat formation in active floodplain systems. Therefore, the mechanism by which valley widening occurs in the presence of concurrent peat formation in these wetlands is unknown.

A number of palmiet wetlands are undergoing severe gully erosion (Rebelo, 2012; Job, 2014). Gully erosion in the palmiet wetlands of the Goukou River situated in the Southern Cape is thought to be part of the natural cycle of this system (Job, 2014). A similar model of natural cycles of cutting and filling has also been developed for the Featherstone Kloof wetland near Grahamstown in the Eastern Cape (Silbernagl, 2014). Cut-and-fill cycles have been observed in a range of semi-arid regions across the globe, while more locally they have been noted in the Karoo (Grenfell et al., 2012; Nelson & Rittenour, 2014). As with the Goukou wetland, some type of cut-and-fill dynamic is thought to be a primary driver of gully erosion in palmiet wetlands distributed along the Kromrivier, also located in the Southern Cape.

All cut-and-fill systems vary slightly in their dynamics, which are dependent on the geomorphic context or setting (Patton & Schumm, 1975; Burrough et al., 2015). Yet the basic principles applicable to most cut-and-fill systems are still relevant (Patton & Schumm, 1981; Fryirs & Brierly, 1998). The cutting phase involves discontinuous gully erosion while the filling phase comprises backfilling of gullies and sediment deposition on gully floodouts (Patton & Schumm, 1981; Grenfell et al., 2012). In cut-and-fill systems where palmiet is present, the plant is thought to play a pivotal role as it is considered to be an ecosystem engineer capable of altering the structure and functioning of its ecosystem (Jones et al., 1994; Sieben, 2012; Job, 2014; Barclay, 2016).

In the Kromrivier context, during longer term periods of lower flows, it is thought that fill dynamics are predominant. During these filling phases palmiet may colonise floodouts and other depositional features along the river channel. Due to its hardy rhizomatous roots and robust stems it is able to resist high flows (Sieben, 2012; Job, 2014). Its tendency to grow in an intertwining network enables palmiet to colonise the entire breadth of a channel reach, which causes diffuse flow (Sieben, 2012; Job, 2014). This, in combination with its roots' effective sediment trapping capabilities, leads to conditions conducive to peat formation i.e. low energy flow and limited clastic sediment in areas distal to sediment supply (Rydin & Jeglum, 2006; Grenfell et al., 2010). Thereafter peat accumulates. Conversely, during periods of higher flow or during isolated flooding events, the cutting phase of the natural cycle predominates and discontinuous gully erosion is initiated in certain reaches of the river channel (Job, 2014; Silbernagl, 2014). Once reduced flows return, the filling phase once again is more prevalent, as palmiet recolonizes the river and peat formation resumes (Job, 2014). During each subsequent cutting episode, different sections of the valley appear to be eroded, such that over time valley widening and planing of bedrock occurs leading to lower relative stream power (Job, 2014).

It is postulated that during the more prolonged filling phases along the Kromrivier valley, wetlands may arise in the wide and relatively planar valleys which result from these cut-and-fill cycles. The occurrence of wetlands in other cut-and-fill systems elsewhere, has been mentioned. However, seldom has the origin of these wetland systems been explicitly linked to the process of cut-and-fill cycles in the southern African sub-region (Grenfell et al., 2009a; Job, 2014).

An alternative and more widely accepted explanation for the gully erosion present in the Kromrivier's palmiet wetlands is that it is caused by anthropogenic impacts that result in the degradation of wetland environments (Bocco, 1991; Poesen, 2003; Valentin, 2005; Rebelo, 2012). Such impacts include increased runoff as a result of catchment overgrazing, frequent burning of the catchment or hardening of surfaces in the catchment. Alternatively, activities in the wetland may increase flow rates locally or across the entire width of the valley, such as by clearing palmiet, cultivation, excavation of artificial drains and road crossings.

Given these contrasting views of erosion and deposition, a greater level of understanding is needed to determine whether the erosion taking place in the Kromrivier's palmiet wetlands is linked to anthropogenic impacts or forms part of the natural dynamics of these complex peat wetlands; and further, whether these cycles play a role in their natural origin and evolution (Tooth et al., 2009; McCarthy et al., 2010).

The Kompanjiesdrif basin is viewed as a relatively pristine "basin" of the Kromrivier wetland (Haigh et al., 2002), in that it is dominated by extensive stands of palmiet and shows no visible signs of gullies (presently or historically). However, two large erosion-control structures have been installed at the toe of the basin to halt the progressive headward erosion of a large gully that was viewed as threatening the basin. The Kompanjiesdrif basin was therefore chosen as the site of investigation in this study.

The focus of this research is to shed light on the origin and geomorphic dynamics of a single palmiet wetland basin that falls along a reach of the Kromrivier in the Eastern Cape of South Africa. The Kromrivier is subject to extensive reaches of gully erosion and the Kompanjiesdrif basin is one of the few remaining intact palmiet wetland basins along its length. Using this basin as a case study the aim is to determine the mechanism by which valley widening and longitudinal slope reduction has occurred. In this way the study aims to also verify whether gully erosion is a natural phenomenon that has taken place across the valley over extended periods, perhaps more recently influenced by anthropogenic activities; or whether it is likely to be solely a consequence of human activities in the study area. The objectives of the study are to:

- Develop a conceptual model to explain wetland origin and dynamics in this setting (Objective 1)
- Determine the topography and cross-sectional depth to bedrock beneath sedimentary fill in the Kompanjiesdrif basin, and describe the characteristics of sedimentary fill across sections of the wetland (Objective 2)
- Determine the ages of sediment across selected sections of the Kompanjiesdrif basin (Objective 3)

- Determine variation in the width and longitudinal slope of the entire Kromrivier as a point of reference against which the width and local gradient of the Kompanjiesdrif basin may be compared (Objective 4)

# CHAPTER 2 – LITERATURE REVIEW

## *2.1 Erosion and wetlands*

Wetlands around the world have been subjected to varying degrees of anthropogenic impacts and subsequent degradation, which has led in some cases to their complete destruction (Davis & Froend, 1999; Finlayson & Rea, 1999; Acreman et al., 2000; Mitsch & Gosselink, 2007; Zhou et al., 2009). Examples of impacts that may negatively affect wetlands include expansion of agriculture or urban development, both of which can lead to the draining of portions of or entire wetland complexes (Van Asselen et al., 2013). Associated with this development is an increase in the amount of infrastructure such as roads or dams, which are known to have potentially severe effects on the health and integrity of wetlands in their vicinity (Van Asselen et al., 2013). In many regions the proportion of wetlands damaged by human activities is greater than 50% (Mitsch & Gosselink, 2007; Van Asselen et al., 2013).

A wetland can be defined as an environment in which protracted shallow inundation typically occurs for a sufficient period of time per annum for anaerobic soil conditions to arise such that hydrophytes are the dominant plants (Ellery et al., 2009). Wetlands occupy the interface between aquatic and terrestrial environments and are viewed as depositional environments that possess low relative stream power as a consequence of their near-planar cross-sections and very gentle longitudinal slopes of less than 1 %, with small hillslope seepage wetlands being an exception (Ellery et al., 2009). Hence, soil erosion in a wetland is generally associated with wetland degradation.

Soil erosion is a process whereby sediment is detached and transported either via wind or water (le Roux, 2007; Hermon, 2016). In the majority of cases the mechanism which induces soil erosion is water (le Roux, 2007; Hermon, 2016). Soil erosion is a natural process which can be accelerated and thus exceed normal rates, through the effects of anthropogenic drivers. The resultant landforms which arise through the process of soil erosion are as variable in size, form and occurrence as the landscapes they occupy (Hermon, 2016). These include micro rills, rills, mega-rills, ephemeral gullies and gullies (Poesen, 2003).

The southern African landscape is known to be highly elevated and therefore incised in comparison to other landmasses that have not experienced recent mountain building events (McCarthy & Rubidge, 2005; Ellery et al., 2009). Yet, despite this, southern Africa plays host to a wide range of wetlands in a variety of landscape settings (Ellery et al., 2009; Job, 2014).

In order to fully appreciate how wetlands persist under what appear to be adverse conditions, a sound understanding of the geological and tectonic history of the region is a prerequisite (Ellery et al., 2009).

*2.2 The recent geological and tectonic history of southern Africa*

The study site is in the Cape Fold Mountains that extend from the Western Cape to the Eastern Cape, with the mountains comprising highly resistant quartzite of the Cape Supergroup. Folding of the sediments that make up the mountain range took place as a consequence of the collision of Antarctica with the African continent about 300 million years ago, long before the break-up of Gondwanaland (McCarthy & Rubidge, 2005). Before the break-up of Gondwanaland the elevation of the subcontinent ranged between 1000 and 2000 m above sea level (Partridge & Maud, 2000; McCarthy & Rubidge, 2005)

The break-up of Gondwanaland and the rifting that occurred contributed further to an elevated southern African subcontinent and the formation of a marginal escarpment (Partridge & Maud, 2000; McCarthy & Rubidge, 2005). In the more recent geological past, the southern and eastern African regions experienced two isostatic uplift events, the first of which took place about 20 million years ago, while the second event took place about 5 million years ago (Partridge & Maud, 2000; McCarthy & Rubidge, 2005). The cause of these uplift events is generally poorly understood as they are unique to the African continent (McCarthy & Rubidge, 2005). The uplift associated with both events was greater in the east than the west of southern Africa, causing the subcontinent to slope gently downwards from east to west, but more specifically in the southern Cape (the region of the present study), uplift amounted to approximately 200 m for each event (Partridge & Maud, 2000; McCarthy & Rubidge, 2005).

These increases in elevation resulted in a decrease in the base levels of fluvial systems draining into the ocean, causing river incision to occur (McCarthy & Rubidge, 2005). The original land surface that was present before uplift took place is referred to as the African Erosion Surface (AES). Two phases of erosion were associated with these uplift events and have formed two additional erosion surfaces, the Post Africa I (PA I) and Post Africa II (PA II) surfaces that represent stable land surfaces created over the period 20 to 5 million years ago (PA I), and less than 5 million years ago (PA II) respectively (Partridge & Maud, 2000;

McCarthy & Rubidge, 2005). Therefore, these events are largely responsible for the incised topography of southern Africa.

Mechanisms of landscape evolution which have led to the current topography and distribution of wetlands in southern Africa are not only linked to the tectonics and geology of the region, but also to climate (McCarthy & Hancox, 2000). Soil erosion is prevalent across most of southern Africa; and water is viewed as the primary process through which erosion takes place in the region (le Roux et al., 2007). Thus, the current and past climates of southern Africa have played a significant role in producing the current landscape, by either slowing rates of erosion during drier climatic phases or catalysing periods of rapid landscape denudation during more humid climatic phases (McCarthy et al., 2011; Job, 2014; Silbernagl, 2014). Therefore, a sound understanding of present and past climates has an important bearing on understanding the geomorphic dynamics that have led to wetland formation.

### *2.3 The evolution of the current climate of southern Africa*

Our current climatic regime has its origins in a number of events that took place between approximately 35 and 2 million years ago and is closely linked to the geological history of the landmass (McCarthy & Rubidge, 2005). The break-up of Gondwanaland into its component parts formed shallow warm oceans which progressively became colder and deepened due to the continental drift that took place (McCarthy & Rubidge, 2005). The separation of the fragments of Gondwanaland was responsible for thermally distancing Antarctica from the other landmasses positioned on the surface of the globe and led to the creation of circum – Antarctic circulation 35 Ma (McCarthy & Rubidge, 2005). Movement of the continents closer to their current positions led to the expansion of the Southern Ocean and a reduction in temperature on Antarctica, as well as the formation and expansion of ice sheets (Deacon & Lancaster, 1988; McCarthy & Rubidge, 2005). By 23 Ma ice had reached the margins of the continent, which resulted in the cooling of the Southern Ocean and subsequent reductions in temperatures were felt in the Atlantic approximately 14 Ma (Partridge & Maud, 2000; McCarthy & Rubidge, 2005).

Prior to the cooling of the Atlantic, both the Atlantic and Indian Oceans were responsible for supplying warm moist air to southern Africa, creating a warm humid climatic regime over the landmass (Partridge & Maud, 2000; McCarthy & Rubidge, 2005). Reduction in temperatures of the Atlantic led to the generation of the Benguela current along the western periphery of the continent (McCarthy & Rubidge, 2005). The upwelling of cold water from this newly generated current led to the propagation of westerly winds and a supply of cold dry air interacting with the already existent tropical easterlies that supplied warm moist air (McCarthy & Rubidge, 2005). Subsequently, the western side of southern Africa developed a more arid environment and the east coast maintained a warmer more humid regime that led to a rainfall gradient across South Africa (Partridge & Maud, 2000; McCarthy & Rubidge, 2005). Over the last 2 Ma the climate has generally been cooler than the current regime and has undergone both a glacial and interglacial period with the earth currently in an interglacial phase (Deacon & Lancaster, 1988; McCarthy & Rubidge, 2005). Our climatic regime today retains the west to east rainfall gradient which has been established and enhanced over the history of the continent (Tyson, 1986; Deacon & Lancaster, 1988; McCarthy & Rubidge, 2005).

### *2.4 Wetlands in drylands and the fluvial network*

Southern Africa is defined as a dryland environment as mean annual evapotranspiration is greater than the mean annual rainfall (Tyson, 1986; Ellery et al., 2009, Tooth & McCarthy 2007). This has resulted in a net negative water balance (Tooth & McCarthy, 2007; Ellery et al., 2009). Consequently, in a highly incised landscape that exhibits a negative water balance, wetlands tend to be highly integrated with the fluvial network (Tooth & McCarthy, 2007; Ellery et al., 2009). With the exception of pans and depressions, groundwater and precipitation may act as supplementary water sources, but they are unlikely to be primary sources of water for wetlands in the region (Tooth et al., 2002; Ellery et al., 2009; Job, 2014). Due to the close association wetlands have with fluvial systems in the region, fluvial geomorphology has become an important theoretical framework for understanding the origin, dynamics and collapse of many wetlands. Studies that have employed this approach include: Tooth et al (2002, 2004), Tooth & McCarthy, (2007), Grenfell et al., (2008, 2009a & b, 2010) and Ellery et al., (2009, 2013). Fluvial geomorphology is a field of study that attempts to understand how flowing water in streams and rivers alters and affects the evolution of landscapes that make up the surface of the earth (King, 1963, Leopold et al., 1964; Edwards et al., 2009).

Therefore, it is possible to concede that the geomorphic dynamics of wetlands in southern Africa, including erosion and deposition, behave in a similar manner to those reflected in rivers and streams around the world (Ellery et al., 2009). Several fundamental paradigms have been developed in this field which are integral to understanding rivers and wetlands in southern Africa (Ellery et al., 2009). Thus, reviewing essential concepts central to fluvial geomorphology will aid in understanding the evolution of many wetlands in the region.

### *2.4.1 Dynamic equilibrium in fluvial systems*

One of the fundamental concepts utilised to understand changes in the dynamics of a stream or a wetland, is that of an integrated system in dynamic equilibrium (Leopold, 1964; Schumm, 1977; Ellery et al., 2009). Streams react in a cybernetic manner with feedbacks such that if one reach of a stream system is altered naturally or artificially, it will in turn influence other reaches of the system (Schumm, 1977). Important variables that play a role in shaping a stream system at the local scale include variables that form part of the stream power equation ($\Omega = p.g.Q.S$; where p is the density of water, g is acceleration due to gravity, Q is discharge and S is slope) and the discharge equation ($Q = V.A$; where V is velocity and A is cross-sectional area) other factors that are important include base level and sediment load (Leopold & Maddock, 1953; Schumm, 1977; Phillips, 2010). Therefore, a change in one of these variables, such as discharge, will lead to alteration of one or more of the other variables (such as width, depth or velocity) in order for the stream to adjust to the new state. In order to adjust to an increased discharge for example, stream power will increase and thereafter the stream will reduce its slope through erosion of the stream bed. The reduction in slope will cause a concurrent decrease in stream power (Leopold & Maddock, 1953; Phillips, 2010). Erosion (leading to slope reduction) and deposition (leading to slope increase) are two of the most important mechanisms by which streams maximise energy dispersion along their course for a given flow regime (Ellery et al 2009).

The tendency towards either deposition, which is a process associated with slope-steepening, or erosion, which leads to a decrease in slope (Leopold & Bull, 1979), is dependent upon, among other factors, the interactions between the capacity of a stream and the quantity of sediment supply (Ellery et al., 2009). Capacity is defined as the amount of sediment which can be transported by a stream such that if the capacity of a stream is greater than the load (amount of sediment being transported) the stream will erode in order to increase its sediment load sufficiently to maintain uniformity of flow along its length (Ellery et al., 2009). Conversely, if the amount of sediment (sediment load) which is being transported by a stream begins to exceed its capacity, then sediment deposition takes place (Ellery et al., 2009).

Given the fact that streams are integrated systems which make adjustments that aim to disperse their energy along the path of least resistance, and that discharge typically increases downstream, a stream may achieve a continuously declining longitudinal slope from its head to its mouth (Leopold & Bull, 1979; Philips, 2010). Where this is achieved, the stream is referred to as a graded stream (Davis, 1902). Such streams display certain characteristics along their course from their source to their mouth, which either systematically increase (such as discharge or stream power) or decrease (such as gradient or mean particle size being transported) downstream (Leopold & Bull, 1979; Ellery et al., 2009).

### *2.4.2 Base level and accommodation space*

One of the reference conditions used for assessing stream characteristics is that of a logarithmic profile. This reference condition is attributed in part to the effect that sea level has on how streams develop their longitudinal form, given that streams always maintain a downward slope towards the ocean (Ellery et al., 2009). This is simply related to the fact that as a stream in the vicinity of the ocean erodes its bed, longitudinal slope is reduced to a point where velocity is so low that erosion can no longer occur, and therefore slope can no longer be reduced. A physiographic feature (in this case sea level) that controls the depth to which a stream is able to erode its bed, is defined as a base level (Summerfield, 1991). While sea level is the ultimate base level for all streams, there may be physiographic features along the course of a stream that limit the depth to which a stream is able to erode its bed in an upstream direction, which are referred to as local base levels (Davis, 1902; Leopold & Bull, 1979). For example, the elevation of the bed of the trunk stream acts as a local base level below which a tributary cannot erode (Grenfell et al., 2010). Another example of a local base level is a resistant lithology along the course of a stream in a catchment of otherwise easily weathered and eroded bedrock, such as a resistant dolerite dyke, oriented across a stream in a catchment of easily weathered and eroded shale (Tooth et al., 2002). A local base level may also be created by tectonic processes like faulting, where a valley becomes blocked by local uplift across a fault line (McCarthy et al., 2002). Interactions between trunk and tributary streams, geological factors, tectonic activity and natural or artificial impoundments may act as local base levels (McCarthy et al., 2002; Tooth et al., 2002; Ellery et al., 2009; Grenfell et al., 2010).

Accommodation space in a fluvial setting is a zone along a stream, upstream of a local base level, where, erosion has momentarily ceased and the potential for sediment accumulation is likely (Job, 2014). Processes leading to the creation of accommodation space include tectonic subsidence (McCarthy & Hancox, 2000), deposition along a trunk stream that blocks a tributary stream (Ellery et al., 2012), or even deposition by a tributary that reduces the longitudinal slope along the trunk stream (Joubert & Ellery, 2013). Local base levels on a river's longitudinal profile that are associated with variation in the resistance of different lithologies to weathering and erosion, can be identified by distinctive reaches with a near planar cross-section and gentle longitudinal slope, succeeded by a step which is often associated with valley narrowing and stream straightening (Ellery et al., 2009). Accommodation space, along with near planar cross sections and gentle longitudinal slopes, which in combination lead to lower relative stream power, are all factors which may contribute to wetland formation.

### *2.4.3 The origin of fluvially integrated wetlands*

It is important to note that although wetlands in southern Africa are generally integrated with the fluvial network, the hydrological characteristics of wetlands differ from those of rivers and streams (Ellery et al., 2009; Edwards, 2009). Wetlands are environments characterised by prolonged periods of inundation but with slow flow, low energy conditions and long residence times, which contrasts to the faster flow and higher energy regimes of streams (Ellery et al., 2009; Edwards, 2009). Instead of being a product of a pre-existing inherited landscape structure, they arise where fluvial processes contribute to longitudinal slope reduction of the valley in which they occur, through incision, and where lateral planing of the valley has come to dominate over vertical erosion (Tooth et al., 2002; 2004; Ellery et al., 2009; Edwards, 2009). Therefore, the production of broad, flat valleys with a gentle longitudinal slope happens as a consequence of the work of streams to create environments which play host to wetland systems (Ellery et al., 2009; Edwards, 2009).

Resultant investigations into the origin and dynamics of southern African wetlands have focused primarily on how mechanisms of valley widening or the presence of local base levels have contributed to their creation (Ellery et al., 2009; Grenfell 2010). Geomorphological studies have culminated in the creation of conceptual models for certain types of wetlands in southern Africa (Ellery et al., 2009). A conceptual model of how a wetland forms and changes over time provides insight into potential future trajectories. Most existing conceptual models employ geomorphic concepts to provide the rationale which attributes wetland formation to various controls or factors that either influence wetlands independently or in combination (Tooth et al., 2002; Tooth et al., 2004; Grenfell et al., 2008; Grenfell et al., 2010; Joubert & Ellery, 2013; Edwards et al., 2016).

### *2.4.4 Existing conceptual models of fluvially integrated wetland origins*

Although thought has been given to the origin of wetlands for some time, initially by considering the effect of a rifted graben which hosts the Okavango Delta (McCarthy & Ellery 1998), as well as the effects of sea-level rise on coastal wetlands on the coastal plain of northern KwaZulu-Natal (McCarthy & Hancox, 2000), the role of more passive factors such as long-term fluvial processes (tens of thousands of years) in relation to local base levels has received greater attention more recently.

#### *2.4.4.1 Floodplain wetlands in South Africa*

One of the best-known models of wetland formation which has been developed was conceived by Tooth et al. (2002) and was expanded upon by Tooth et al. (2004). This involves a geological source of wetland formation where the occurrence of floodplain wetlands on the Highveld in South Africa has been explained (Tooth et al., 2002; Tooth et al., 2004). Typical features of these floodplain wetlands include oxbow lakes, back swamps and elevated alluvial ridges, within a context of limited sedimentary fill above bedrock (Tooth et al., 2002; Tooth et al., 2004). The presence of these wetlands is linked to a specific combination of underlying geology; which leads to the creation of wide valleys accommodating floodplains with a sinuous meandering channel reach, upstream of deep, narrow valleys with relatively straight channels (Tooth et al., 2002; Tooth et al., 2004).

One of the lithologies involved is a resistant one, normally dolerite (Tooth et al., 2002; Tooth et al., 2004). Dolerite is an igneous rock which arises in the form of sills (near horizontal features of varying thickness that are oriented parallel to pre-existing sedimentary strata) or dykes (generally near-vertical features of varying width that are oriented across the bedding of pre-existing sedimentary strata; McCarthy & Rubidge, 2005; Marshak, 2011; Christiansen & Hamblin, 2014). As a consequence of its resistance to erosion, incision within this lithology takes place over very long term time scales and usually along the near-vertical fractures and joints in the rock (Tooth et al., 2002; Tooth et al., 2004). This produces a deep fairly straight stream course in a deep narrow valley (Tooth et al., 2002; Tooth et al., 2004).

The other lithology involved in the formation of such wetlands is one that is much more easily weathered and eroded than the resistant lithology. Karoo sedimentary rocks, such as sandstone and shale, are more easily weathered and eroded than dolerite and normally erode vertically and laterally over medium to short term timescales (Tooth et al., 2002; Tooth et al., 2004). However, when such less resistant lithologies are found upstream of dolerite, dolerite acts as a local base level such that vertical incision dominates upstream of the resistant lithology, until the stream upstream of this local base level has reached an appropriate gradient for its discharge and sediment load (Tooth et al., 2002; Tooth et al., 2004). Thus, the longitudinal slope is lowered to a very low slope e.g. less than 0.01 % for the Kliprivier floodplain (Tooth et al., 2002). Once the appropriate gradient has been reached, the more easily eroded and extensive lithology is eroded laterally, as vertical erosion is restricted to the rate at which vertical erosion takes place in the dolerite sill or dyke downstream (Tooth et al., 2002; Tooth et al., 2004). The resultant dynamic is that of predominantly lateral erosion in the Karoo sedimentary rocks, leading to valley widening and the creation of a broad valley with a near-horizontal cross section (Tooth et al., 2002). Lateral erosion occurs via a meandering channel which migrates across the valley floor constantly reworking sediment (Tooth et al., 2002; Tooth et al., 2004). The eventual outcome is a wide flat valley that hosts a floodplain wetland upstream of a narrow, straighter river channel with a steeper longitudinal slope (Tooth et al., 2002; Tooth et al., 2004). Rivers in southern Africa which play host to these types of floodplain wetlands include the Kliprivier in the eastern Free State Province, the Stillerust Vlei and the Northington and Dartmoor floodplains in the foothills of the Kwa-Zulu Natal Drakensberg (Grenfell et al., 2008; Tooth et al., 2009; Edwards et al. 2016).

*2.4.4.2 Trunk and tributary related models of wetland formation*

Other conceptual models of wetland origin stem from the fundamental geomorphic controls on these systems involving trunk and tributary streams (Grenfell et al., 2008; Grenfell et al., 2010; McCarthy et al., 2011; Ellery et al., 2012). For example, trunk tributary relationships are important within the context of a fluvial network and in some cases can contribute to the formation of blocked tributary valley lakes or wetland systems (Grenfell et al., 2008; Grenfell et al., 2010; McCarthy et al., 2011; Ellery et al., 2012). This all depends on the relative levels of sediment input and transport capacity of a trunk stream in comparison to one of its tributaries (Grenfell et al., 2010).

On one end of the spectrum it is possible to encounter a steep, well connected tributary that is supplied by a large and/or eroding catchment. If its associated trunk stream has a lower capacity and rate of sedimentation, the tributary is likely to dominate their interaction and reduce the slope of the trunk by creating topographic relief across the trunk stream (McCarthy et al., 2011; Joubert & Ellery, 2013). This results in aggradation taking place at a more rapid pace as a result of the input of sediment from the tributary across the trunk stream valley (McCarthy et al., 2011; Joubert & Ellery, 2013). Such impoundment, whether partial or complete, acts as a local base level along the trunk stream (McCarthy et al., 2011). Behind the local base level accommodation space is created, flow becomes more diffuse and sediment deposition occurs (McCarthy et al., 2011). Complete impoundment of the trunk stream is more likely to occur if the confluence is situated in a narrow valley (McCarthy et al., 2011).

In comparison to complete impoundment, partial impoundment results in a different regime upstream of the local base level created by tributary sedimentation (McCarthy et al., 2011). Partial blockage creates backwater lakes and flood-outs while complete impoundment forms lakes across the trunk valley (Grenfell et al., 2010; McCarthy et al., 2011). The conditions present in these lakes or flood-outs may promote sufficiently prolonged inundation to facilitate peat formation (Grenfell et al., 2010; McCarthy et al., 2011). In the southern African context these conditions include the creation of accommodation space upstream of a local base level such that prolonged periods of inundation and peat formation may occur, provided there is limited input of clastic sediment upstream of the elevated local base level (Grenfell et al., 2010; McCarthy et al., 2011).

On the other end of the spectrum of possibilities, the rate of sedimentation in a trunk stream may be far greater than one or more of its tributaries (Grenfell et al., 2008; Grenfell et al., 2010). This occurs when a tributary has a small catchment as well as a relatively low sediment supply (Grenfell et al., 2008; Grenfell et al., 2010). As a consequence, more rapid aggradation of the trunk stream either leads to the formation of a blocked valley lake (Grenfell et al., 2010), or in less severe cases of dis-connectivity, a valley bottom wetland or tributary discontinuity (Grenfell et al., 2008). In this scenario the level of deposition in the trunk stream elevates the tributary local base level and causes lowering of the tributary's longitudinal slope in an upstream direction of the confluence (Grenfell et al., 2008; Grenfell et al., 2010), decreasing the competence of the tributary stream, which results in a decline in clastic sediment supply to its lower reaches and possibly to peat formation (Grenfell et al., 2008; Grenfell et al., 2010).

Intermediate scenarios which occur along the spectrum of possible trunk-tributary interactions are of a more fluvial nature (Grenfell et al., 2010). These encompass more marginal differences in sediment input and transport capacities related to slope and discharge, which may result in the adjustment of both streams to maintain connectivity (Grenfell et al., 2010). Adjustments may manifest as alterations to the width:depth ratio of the trunk or changes in channel patterns (Grenfell et al., 2010). In other cases no adjustments are necessary and trunk and tributary streams remain integrated (Grenfell et al., 2010). Intermediate states thus do not usually lead to wetland formation (Grenfell et al., 2010).

#### *2.4.4.3 Chemical weathering of bedrock and long term sagging of the landscape*

More recently a new model of wetland formation has been proposed by Edwards et al., (2016) which outlines initially geological and subsequently geomorphic controls on wetland formation. This model describes long term chemical weathering of bedrock and subsequent sagging of the landscape, which has led to the occurrence of peat wetlands on the ancient African erosion surface. What is intriguing about this model of wetland formation is the extremely extended period of time and combination of conditions which have persisted for these distinctive wetlands to form.

From their investigations, Edwards et al. (2016) came to the conclusion that Dartmoor Vlei, located in the Drakensberg region of Kwa-Zulu Natal, was originally a floodplain wetland; that developed initially according to the Tooth et al. (2002, 2004) model of floodplain development upstream of a resistant fine-grained dolerite dyke. Once erosion had reduced the longitudinal slope and the valley had been widened through meander migration upstream of the resistant dolerite dyke, prolonged inundation of the valley floor contributed to deep weathering of the coarser-grained dolerite sill upstream of the dolerite dyke at the toe of the wetland (Edwards et al., 2016). Long term inundation of the dolerite sill led to gradual weathering of this bedrock, which was associated with dissolution of metals in the bedrock (Ca, Fe, Mg, Na), which resulted in mass and volume loss associated with sagging, thereby decreasing the longitudinal slope of the wetland further (Edwards et al., 2016). The diffuse flow and accommodation space created behind the dolerite dyke promoted peat accumulation (Edwards et al., 2016).

### *2.4.4.4 The role of sea level and tectonic activity in forming fluvially integrated wetlands*

Another model which does not involve a geomorphic or geological control yet leads to wetland formation includes sea level rise (Grenfell et al., 2009b). Sea level rise triggers a rise in base level and a phase of deposition along coastal plains such as in northern KwaZulu-Natal (Grenfell et al., 2009b). After the decline in sea level of greater than 120 m during the last glacial period, streams eroded their beds and formed relatively incised valleys up to 45 m deeper than their current stream beds (Ellery et al., 2012). Following the rise in sea level after the last glacial maximum, such valleys were drowned and filled with sediment, forming near-horizontal floodplains tens of kilometres long and several kilometres wide (Grenfell et al., 2009b).

Aside from changes in sea level (the ultimate base level) and other local base level controls, a significant influence on the formation of wetlands can be tectonic activity (McCarthy et al., 1997). An example in which ongoing rifting associated with faulting has led to the creation of large basin like depressions is the Okavango Delta, Botswana, which is a large alluvial fan and wetland system associated with the East Africa Rift system (McCarthy et al., 1997).

### *2.4.4.5 Multiple controls leading to wetland formation*

While some wetlands have one overriding primary control that leads to wetland formation, wetland complexes may exhibit a variety of different geological, tectonic and geomorphic controls on their origin simultaneously (Grenfell et al., 2009b). This is best exemplified in the Mfolozi River and its associated tributaries (Grenfell et al., 2009b). For example, the Msunduze River, one of the Mfolozi's tributaries, and also the formation of Lake Teza which is situated in a tributary valley, were caused by the tributary-by-trunk impoundment dynamic (Grenfell et al., 2009b). However, the Mfolozi lower floodplain and central floodplain are affected by variation in sea level as well as by faulting respectively, leading to a variety of hydro-geomorphic units forming this wetland complex (Grenfell et al., 2009b). Other examples in southern Africa where different controls interact include Stillerust and Wakkerstroom Vlei wetlands that involve bedrock planing and trunk-tributary interactions (Grenfell et al., 2008; Joubert & Ellery, 2013).

As shown from the existing paradigms of wetland formation, there are a number of models that have been developed to explain the formation of wetlands. In addition to these models, which have highlighted the potential controls that tectonics, geology, geomorphology, sea level and climate can have in wetland formation and dynamics, there are also biological factors that have been shown to influence wetland systems (Ellery et al., 1995; Van Breemen, 1995; Wright et al., 2002; Mitsch & Gosselink, 2007; Sieben, 2012; Job, 2014). One of the more common influences biological controls can have is on the process of peat formation in wetlands.

### *2.5 Peat formation in fluvially integrated wetlands*

Peat comprises the organic remains of plants and animals that accumulate due to a reduced rate of decomposition caused by anoxic conditions associated with permanent or semi-permanent inundation (Rydin & Jeglum, 2006; Belyea & Clymo, 2001; Job, 2014). Peat is known to form in low flow conditions according to Rydin and Jeglum (2006), as this reduces the amount of clastic sediment input and can promote anoxic conditions, which are both key factors contributing to peat formation.

The amount of peat that forms will fall along a continuum depending on a number of variables including the degree of aeration, the amount of clastic sediment present, the flow regime, as well as many secondary factors, such as temperature, which affects microbial activity, and geology which affects water retention and the chemical regime present (Rydin & Jeglum, 2006; Yeloff et al., 2006; Belyea & Clymo, 2001; Ellery et al., 2009; Job, 2014).

Often, periodic or permanent inundation of the soil is associated with high plant productivity, which creates a positive feedback whereby the amount of undecomposed organic matter increases as the rate of plant productivity becomes greater than the rate of decomposition, leading to even more stagnant anoxic conditions (Rydin & Jeglum, 2006, Sieben, 2012; Job, 2014).

### *2.6 Ecosystem engineers*

Where biota are viewed as controlling the structure and functioning of an ecosystem other than via trophic interactions associated with energy and nutrient flow, they are referred to as ecosystem engineers (Jones et al., 1994). A number of plants and animals which occur in riverine environments have been shown to promote peat formation and in this way act as a biological control which contributes to a given wetland's structure and function (Barclay, 2016). There are several plants and animals that promote peat and wetland formation, which have been categorised as ecosystem engineers. These include plants such as *Cyperus papyrus* in the Okavango Delta, as well as sphagnum moss in the northern hemisphere (Ellery et al., 1995; Van Breemen, 1995; Mitsch & Gosselink, 2007). Another example is that of the beaver which builds dams that block streams and fundamentally control their structure and function (Wright et al., 2002).

### *2.6.1 Palmiet as an ecosystem engineer*

The plant palmiet (*Prionium serratum*) appears to be the most common and ubiquitous plant dominating the wetland basins of the southern and eastern Cape Fold Mountains, including the Kromrivier (Rebelo, 2012; Nsor & Gambiza, 2013). The common name of this plant was derived from its robust appearance, which early Dutch settlers of the South Western Cape likened to be a combination of a palm and a reed (riet), hence the name "palmiet". Palmiet is thought to promote the presence or continuation of wetlands along the Kromrivier by facilitating peat formation. It is for this reason that palmiet has been considered an ecosystem engineer by Sieben (2012) and Job (2014).

Palmiet has remarkable biological characteristics which contribute to its capabilities as an ecosystem engineer (Sieben, 2012; Job, 2014). It grows to a height of approximately 3 m and is a perennial plant that comprises thick woody stems with a diameter of 100 mm or more, as well as a system of thick adventitious roots (Job, 2014). One of its most advantageous features is its robust nature as the plant's thick stems, which are lined with the leaf bases of dead leaves, make it well adapted to survive fires, after which it recovers relatively quickly with vigorous new growth (Boucher & Withers, 2004; Job, 2014). Any old leaves tend to become wrapped around the stem so that during floods the plant is protected to some degree from loose debris that flows down the valley (Boucher & Withers, 2004; Job, 2014). Aside from its robust nature, the clonal manner in which palmiet grows, leads to an intertwining network of thick stems that allow it to grow across a river channel (Sieben, 2012; Job, 2014). Its tall and robust nature also results in its ability to increase the roughness coefficient of the channel reach, which, combined with its ability to colonise and form a dense mat, leads to more diffuse flow across a channel reach (Boucher & Withers, 2004). The thick network of stems is also capable of trapping sediment fairly effectively in areas proximal to sediment supply sources, while creating relatively sediment free conditions in areas distal to supply sources, allowing palmiet to control sediment dispersal (Sieben, 2012; Job, 2014). Thus, palmiet promotes conditions conducive to peat formation in areas more distal to sediment sources i.e. an environment with limited clastic sediment, diffuse flow and generally low energy flow conditions (Sieben, 2012; Job, 2014).

### *2.7 An alternative mechanism of valley widening and slope reduction*

The presence of peat in the Kromrivier wetlands may be attributed in part to palmiet. However, an alternative mechanism to Tooth's model of valley widening (Tooth et al., 2002) in the context of peat filled wetlands is relatively rare (Grenfell et al., 2009a; Job, 2014; Sibernagl, 2014), particularly if tectonics, variation in sea level and sagging due to deep weathering are discounted (McCarthy et al., 1997; Tooth et al., 2002, 2004; Grenfell et al., 2009; Edwards et al., 2016). The wetland basins located along the Kromrivier appear to be peat dominated systems and there is no evidence of meander channel migration. Thus, the Tooth et al. (2002) model of wetland formation via lateral erosion and meander channel migration does not appear to be a possibility given the fact that systems examined by them are dominated by clastic sediment rather than organic sediment. Moreover although Edwards et al. (2016) have demonstrated how a floodplain wetland over long time scales may evolve into a peatland system, this model does not reflect the dynamics along the Kromrivier for two reasons. Firstly, the Cape Fold Mountains in the study area of the Kromrivier are known to be free of dolerite as their strength and density precluded the intrusion of dolerite (Job, 2014). Secondly, weathering of igneous rocks such as dolerite differs to weathering of sedimentary rocks such as sandstone and shale of the Cape Supergroup because of differences in the complexity of their chemistry and mineralogy. Basic igneous rocks have a complex chemistry and mineralogy such that weathering leads to considerable loss of mass in the form of metals that go into solution during weathering, a process that may lead to volume reduction. However, sedimentary rocks are unlikely to weather chemically to yield large solute loads given their more simple mineralogy, particularly since quartzite is the predominant lithology of the Cape Supergroup. Therefore, mass and volume losses of previously weathered and eroded sedimentary rocks is unlikely to be as great as for basic igneous rocks (Alistoun, 2014; Edwards 2016).

The distinction between these different types of rocks relates to how they were formed (King, 1963, McCarthy & Rubidge, 2005). Basic igneous rocks such as dolerite are formed through volcanic processes which introduce complex minerals from depth within the mantle, to the crust. However, sedimentary rocks form from material which has already been exposed to a cycle of weathering and erosion, and thus comprise minerals that have already undergone mineralogical simplification and are more stable and thus more resistant to chemical action (King, 1963; McCarthy & Rubidge, 2005).

### *2.8 The conundrum of a wetland in the Kompanjiesdrift basin of the Kromrivier*

Given all of this background, the flat and wide nature of the valley of the Kompanjiesdrif basin, located along the Kromrivier, is unlikely to be attributed to the existing mechanisms outlined in the hypotheses above. That is to say, the existing mechanisms proposed to widen valleys and lead to longitudinal slope reduction as proposed by Tooth et al., (2002, 2004) or the model of wetland origin proposed by Edwards et al., (2016). Thus, an alternative explanation for the broad flat valley floor in this otherwise long, steep-sided valley may be applicable and will be considered alongside the other hypotheses. The occurrence of gullies in the Krom presents a potential alternative mechanism for valley widening and slope reduction (Fryirs & Brierly, 1998; Brierly & Fryirs, 1999). Consequently, in the following section a review of gully erosion as a natural or human induced phenomenon is considered, and the ability of this form of erosion to act as a mechanism of landscape evolution that leads to valley widening and slope reduction is explored (Fryirs & Brierly, 1998; Brierly & Fryirs, 1999).

### *2.9. Natural versus anthropogenic gully erosion*

Gullies in a wetland context have been closely linked to deterioration of wetland health (Riddell et al., 2010). The process of gully erosion involves runoff eroding a narrow area that rapidly propagates headward as "headward erosion", such that a deep, narrow channel-like feature develops over a short time period (Poesen, 2003). It is important to clarify that an attempt to classify erosion processes into rill erosion, ephemeral gully erosion, gully erosion and river channel erosion is somewhat subjective as these processes fall along a continuum rather than functioning as discrete entities (Poesen, 2003). Thus, what distinguishes a permanent gully from an ephemeral gully and a gully from a stream, varies according to the parameters chosen (Poesen, 2003). Although ephemeral gullies are more transient in nature and are known to have smaller dimensions than permanent gullies, the details are often debatable (Poesen, 2003). Similarly, the upper limit of permanent gully dimensions which separates a feature from being classified as a stream channel remains relatively indeterminate (Poesen, 2003). For the purpose of this review unless stated otherwise, the term gully erosion encompasses both permanent and ephemeral gullies and is associated with the generation (as opposed to just "transport") of sediment from a wetland to downstream areas.

In addition to the negative connotations of gully formation in wetland settings, the process of gully erosion is a pervasive phenomenon that is viewed as having negatively affected a wide range of environments (Bocco, 1991; Poesen, 2003; Valentin et al., 2005). This in turn has increasingly highlighted the need to address the subject through research (Bocco, 1991; Poesen, 2003; Valentin et al., 2005). Numerous studies on gully erosion, point time and time again, to anthropogenic impacts, which range from installing roads, to deforestation or cultivation, to overgrazing of natural vegetation as one of the primary causes and drivers of gully initiation and propagation (Bocco, 1991; Poesen, 2003; Valentin et al., 2005). This trend in the literature has become well established and widely accepted as the most accurate narrative, to the extent that it has almost become entrenched as dogma (Bocco, 1991; Poesen, 2003; Valentin et al., 2005).

Nevertheless, the alternative, although less frequently considered narrative, is that natural factors may lead to gully initiation and propagation (Patton & Schumm, 1975; Job, 2014; Nelson & Rittenour, 2014). The most prominent natural factors include: climatic variability, vegetation degradation, geomorphic thresholds and highly erodible soil type, among others (Patton & Schumm, 1975; Patton & Schumm, 1981; Prosser & Slade, 1994; Fryirs & Brierly, 1998). When considering the cause of gully erosion, studies have demonstrated that a number of contributing factors are involved and work in combination (Brierly & Fryirs, 1999; Valentin et al., 2005). This poses a challenge in pin-pointing the primary driver of their initiation (Brierly & Fryirs, 1999; Valentin et al., 2005).

Geomorphic studies on gully erosion in conjunction with quantifying chronological controls, often aid in determining whether gully erosion is human-induced or driven by natural factors (Fryirs & Brierly, 1998). Dating of sediment samples has revealed instances of gully erosion that occurred prior to human settlement in various parts of the world (Botha & Partridge, 1988; Dardis, 1989; Fryirs & Brierly, 1998, Avni et al. 2010). These include gullies dated in KwaZulu-Natal, South Africa (Botha, 1996; Botha and Partridge, 1988), in the Chifeng region of Mongolia (Avni et al. 2010), in New South Wales, Australia (Fryirs & Brierly, 1998), as well as in Swaziland and the Eastern Cape, South Africa (Dardis, 1989). Moreover, discontinuous gully erosion as a form of natural landscape evolution has been recognised in a range of environments (Patton & Schumm, 1981; Fryirs & Brierly, 1998; Grenfell et al., 2012; Bekaddour et al., 2016). The study of landforms and their evolution falls under the domain of geomorphology (Patton & Schumm, 1981; Fryirs & Brierly, 1998; Grenfell et al., 2012) such that several studies conducted on cut-and-fill systems have included geomorphic enquiries (Patton & Schumm, 1981; Fryirs & Brierly, 1998; Grenfell et al., 2012).

### *2.10 Cut-and-fill cycles*

#### *2.10.1 Research on cut-and-fill cycles*

Cut-and-fill systems which have been investigated in various regions of the globe are observed primarily along discontinuous river channels which share similar trends in stream behaviour (Brierly & Fryirs, 1999). Common to all cut-and-fill systems is a phase of incision in the form of gully erosion followed by a filling phase, which involves sediment accumulation (Nanson & Croke, 1992; Bull, 1997). Cut-and-fill systems which have been investigated along certain discontinuous river channels of Australia are characterised by periods of long term deposition and sediment accumulation which are interrupted by short lived phases of rapid incision (Nanson & Croke, 1992; Brierly & Fryirs, 1999). In contrast, the cut-and-fill systems of North America seem to undergo gully erosion on a more frequent basis (Nanson & Croke, 1992; Patton & Schumm, 1981). Discontinuous ephemeral streams that exhibit cut-and-fill patterns are primarily restricted to arid or semi-arid environments and have been researched in North America (Patton & Schumm 1975), Australia (Brierly & Fryirs 1998), central and southern Africa (Burrough et al., 2015; Grenfell et al., 2012), South America (Bekkadour et al., 2016) as well as the Mediterranean region (Thornes, 1976). Research on cut-and-fill cycles has referred to these systems using a range of terms (Brierly & Fryirs, 1999). Some of the terms employed include: dambos (e.g. Boast, 1990), ephemeral arroyos (e.g. Bull, 1997), gullies and floodout systems (e.g. Grenfell et al., 2012), dongas (Botha et al., 1994) and finally cut-and-fill sequences (Bekaddour et al., 2016).

#### *2.10.2 Controls on the initiation of the cutting phase*

Investigations into cut-and-fill systems have often focused on identifying factors that affect the transition between the filling phase and the cutting or incisional phase (Brierly & Fryirs, 1999; Grenfell et al., 2012). Research has found that there appears to be either a threshold related to slope, or to runoff, or to a combination of both these variables, which triggers erosion (Schumm, 1973; Brierly & Fryirs, 1999). This is referred to as an intrinsic geomorphic threshold if the controls involved are within-system factors (Schumm, 1979). A geomorphic threshold according to Schumm (1973: 301) "is one that is inherent in the manner of landform change; it is a threshold that is developed within the geomorphic system by changes in the system itself through time." The concept of a geomorphic threshold was later revised to include a sudden adjustment in a landscape which transpired due to progressive change in extrinsic controls, such as tectonic activity that might lead to a lowering of base level (McCarthy & Hancox, 2000). Given this, a geomorphic threshold can be classified either as intrinsic or extrinsic (Schumm, 1979).

In North Western Colorado a number of arroyo cut-and-fill systems are thought to be controlled primarily by intrinsic geomorphic slope thresholds (Patton & Schumm, 1975; 1981). Discontinuous gully erosion in these systems propagates through knickpoint migration or bank collapse and is interspersed with stable or aggrading stream reaches; where the sediment which has been eroded is temporarily stored in the widened channel, downstream of the site of active incision (Patton & Schumm, 1981).

Considerable debate has taken place over whether the thresholds of cut-and-fill environments are controlled solely by geomorphic within system changes (Patton & Schumm 1975, Schumm, 1979; Patton & Schumm, 1981) or if additional extrinsic controls such as climate and vegetation may play a role (Balling & Wells, 1990; Prosser & Slade, 1994; Fryirs & Brierly, 1998; Bekkadour et al., 2016). Moreover, in some cases intrinsic and extrinsic controls have been observed to act in combination to trigger gully erosion, making it difficult to identify the primary causative factor responsible (Grenfell et al., 2012). Nevertheless, although these three controls (vegetation, climate and intrinsic thresholds) are viewed as the main factors which are likely to affect cut-and-fill landscapes, there are other controls that have been found to act as additional drivers of change in these systems, which can lead to a transition from deposition to erosion (Brierly & Fryirs, 1999).

In the case of Wolumla Creek in New South Wales, Australia, it was the position of the trunk stream within the trunk valley undergoing incision that acted as the primary determinant that affected incision in its respective tributaries (Brierly & Fryirs, 1999). If the incision in the trunk stream occurred at the confluence with one of its tributaries then this acted to change local base level, which triggered incision in the tributary (Brierly & Fryirs, 1999).

Conversely, if incision in the trunk stream was buffered from the tributary by a bank of non-incised valley fill, then no change in base level would take place for the tributary in question and incision would not be triggered (Brierly & Fryirs, 1999).

An additional factor that has been considered in Wolumla Creek is the effect of anthropogenic impacts on the cut-and-fill system. Impacts include the draining of swamps for cultivation purposes and widespread removal of vegetation following European settlement in the area; which has led to increased connectivity, with the resultant gully erosion becoming more continuous as opposed to discontinuous in nature (Brierly & Fryris, 1999; Brierly et al., 1999; Fryirs & Brierly, 2009).

*2.10.3 Dis-connectivity in cut-and-fill environments*

When considering cut-and-fill cycles on a larger landscape scale, the degree of connectivity or dis-connectivity of these systems is an important factor (Brierly et al., 2006; Fryirs & Brierly, 2009; Grenfell et al., 2012). Connectivity is defined as "the transfer of energy and matter between two landscape compartments or within a system as a whole" (Fryirs et al., 2007:49). Patton and Schumm, (1981) argue that cut-and-fill dynamics are part of the natural functioning of semi-arid environments with high sediment yield, and serve as a process which moves sediment through the system in a disconnected manner. Moreover, Grenfell et al. (2012) have stressed the role of floodouts and palaeo-floodouts in cut-and-fill systems, which serve to create dis-connectivity between discontinuous gullies and achieve temporary sediment storage. Dis-connectivity has also been described in cut-and-fill systems in Australia whereby alluvial and colluvial sediment have been shown to often be temporarily disconnected from the sediment transport system (Fryirs & Brierly, 1998; Brierly & Fryirs, 1999). Thus, in a number of cut-and-fill systems dis-connectivity appears to be a prominent trend.

The relevance of dis-connectivity present in cut-and-fill systems is related to the heterogeneity observed in these systems. Dis-connectivity serves to cause spatial and temporal variability in the transport and storage of sediment within a system (Brierly et al., 2006). Consequently, erosion does not become the dominant and widespread feature of these landscapes (Grenfell et al., 2012), but is rather interrupted and interspersed with sediment storage and aggradational reaches allowing greater ecosystem diversity along the river course (Brierly et al., 1999; Fryirs & Brierly, 2009).

### *2.10.4 Cut-and-fill cycles as a mechanism of landscape evolution*

In addition to the contribution to heterogeneity that dis-connectivity allows in these landscapes, the role that cut-and-fill cycles have in the evolution of landscapes of southern Africa has been under-examined (Grenfell et al., 2012). The inherent local scale episodes of gully erosion lead to lowering of the landscape (Grenfell et al., 2012), as erosion is a slope lowering process (Ellery et al., 2009). Furthermore, due to the dis-connectivity of these systems, during each successive cutting phase a different section of the valley or river channel is eroded so that over long term (geological) timescales valley widening occurs (Job, 2014). Over repeated cycles of cutting and filling a more gentle topography may emerge which is more conducive to a low energy depositional environment associated with wetlands (Job, 2014).

Hence, the presence of wetlands in highly erosive landscapes associated with gully erosion appears to be more common in semi-arid regions of the world. This is illustrated by the names assigned to describe discontinuous streams which exhibit cut-and-fill features in Australia: with names such as dells (Young, 1986), upland swamps (Fryirs & Brierly, 1998) and swampy meadows (Prosser, 1991), and is also exemplified by the presence of hydrophytes in some of the dambos of central Africa (Boast, 1990). However, rarely has the origin of wetlands in southern Africa been linked to cut-and-fill cycles (Grenfell et al., 2009a; Job, 2014; Silbernagl, 2014). This appears to be a knowledge gap which still needs to be considered.

## *2.11 Conclusion*

The potential controls on wetland formation have been outlined for both peat and clastic dominated wetlands systems as hypotheses in this review. The geomorphic dynamics and controls that lead to peatland systems are more varied and have been underexplored in comparison to the clastic dominated systems of southern Africa. By examining the potential hypotheses which may contribute to the geomorphic dynamics and origin of the Kromrivier system insights may be gained on yet another peatland in southern Africa. Given the relevance of the climate, geology, biological factors, tectonics and changes in sea level highlighted in this review, some of these factors are explored in the study area reach of the Kompanjiesdrif basin, which will be used as a case study.

# CHAPTER 3 – STUDY AREA

## *3.1 Location, topography, geology*

The Kromrivier (33°S, 24°E; Figure 1) is located in the Eastern Cape of South Africa with its upper reaches flowing past the small town of Kareedouw and ending in an estuary at St Francis Bay on the Indian Ocean coastline (Dennis & Wentzel, 2007; Rebelo, 2012; Jansen & van Veen, 2014; Rebelo, 2015). The Kromrivier is approximately 100 km in length and falls within a narrow catchment which is quite precipitous due to two mountain ranges, namely, the Tsitsikamma (max. elevation 1251 m) and Suuranys (max. elevation 1073 m) ranges (Dennis & Wentzel, 2007; Rebelo, 2012; Rebelo, 2015). These mountains flank the catchment with the former on the southern and the latter on the northern side of the Kromrivier valley, with both the mountain ranges running in an east to west direction (Rebelo, 2012; Rebelo, 2015). Most of the drainage network exhibits a trellis pattern with the exception of a limited number of tributaries that display a dendritic form (Haigh et al., 2002). The catchment spans a total area of 1125 $km^2$, receiving inflow from a number of major and minor tributaries from both ranges (Dennis & Wentzel, 2007; Rebelo, 2012). Inflow from the tributaries entering the Kromrivier from the Suuranys range is predominantly seasonal (Rebelo, 2012).

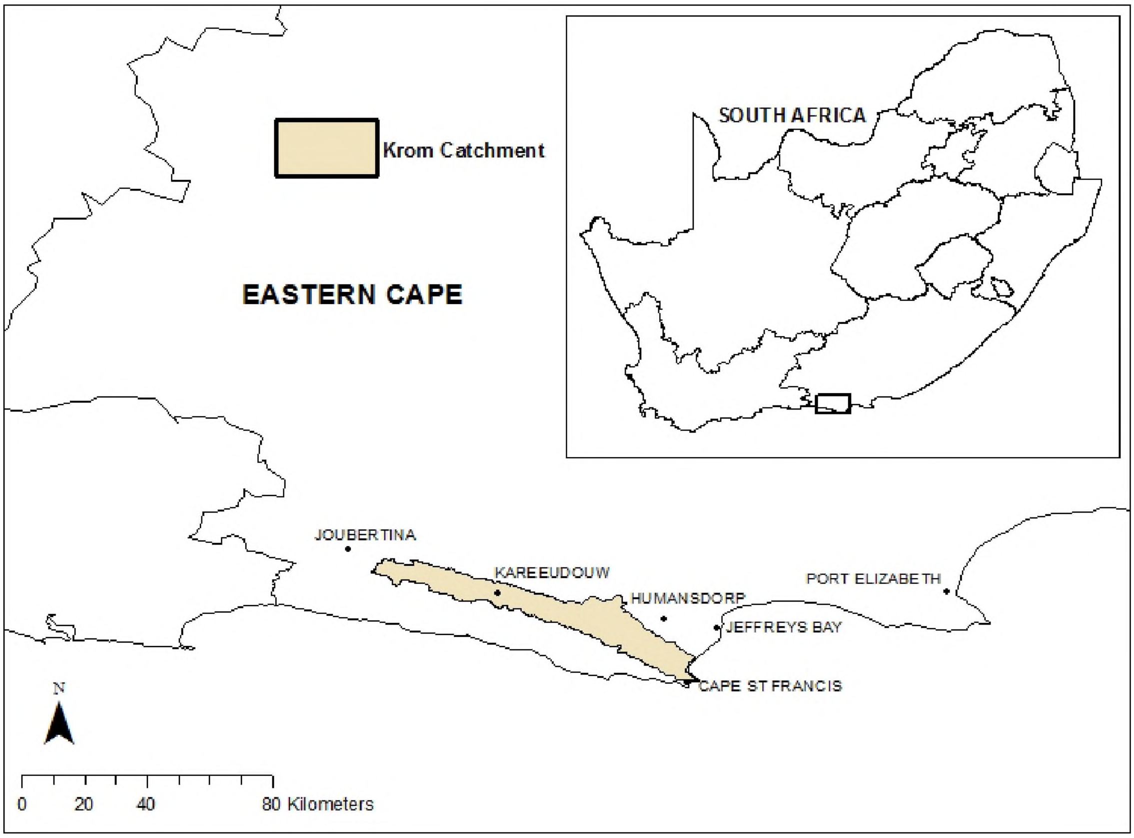


**Figure 1:** *The location of the Kromrivier catchment in the Eastern Cape, South Africa*

The geology of the Kromrivier catchment comprises predominantly rocks of the Bokkeveld and Table Mountain Groups', which are both members of the Cape Supergroup (Figure 2; Almond et al., 2008). The Cape Supergroup comprises sedimentary rocks such as siltstones, shale and quarzitic sandstone; while on the valley floor some sequences comprise Quaternary alluvium (Almond et al., 2008). Rocks of the Cape Supergroup have been folded and metamorphosed such that the quartzitic sandstone is highly resistant to weathering and erosion, forming the high-lying peaks of the mountain ranges. In contrast the shales and siltstones are more easily weathered and eroded, occurring mainly on the valley floors (Mucina & Rutherford, 2006). The Ceres Subgroup of the Bokkeveld Group comprises three thick layers of shale and three markedly thinner layers of sandstone, with the Gydo Formation being the thickest shale layer (500 m) within this Subgroup (Toerien, 1979; Toerien & Hill, 1989). The Baviaanskloof Formation in the Nardouw Subgroup of the Table Mountain Group consists of a mixture of sandstone and subordinate carbonaceous and micaceous shale (Toerien, 1979; Toerien & Hill, 1989). The Cedarberg Formation is a thin layer of shale which occurs in the sequence older than the more resistant Peninsula Formation (Mucina & Rutherford, 2006). The remainder of the formations comprise subordinate shale layers (Toerien, 1979; Toerien & Hill, 1989).

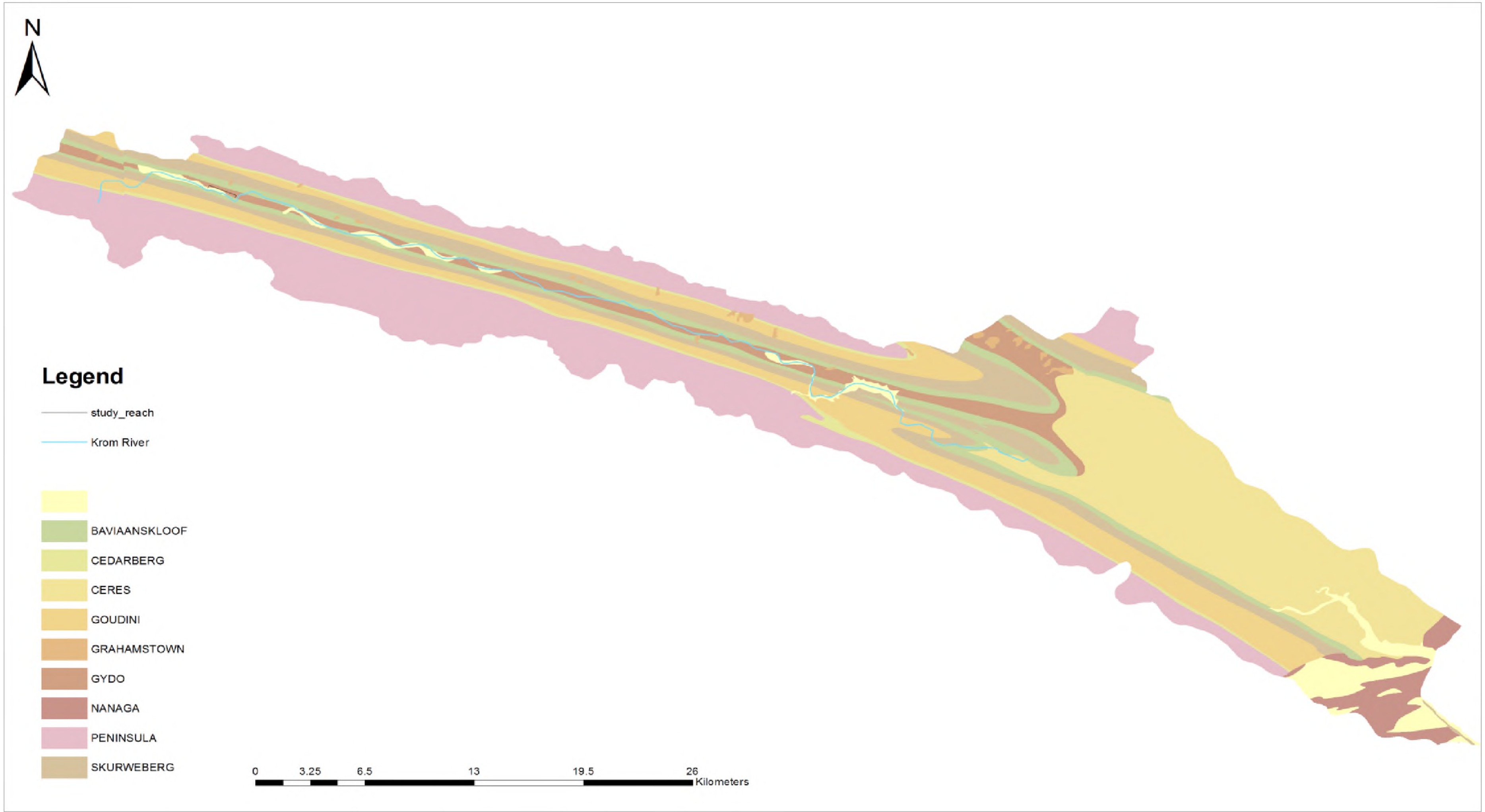


**Figure 2:** *The geology of the Kromrivier catchment*

The study site, namely the Kompanjiesdrif wetland basin, is located 20 km east of the town of Joubertina at an altitude of approximately 360 m.a.s.l. (Rebelo, 2012). It falls within quaternary catchment K90A (Haigh et al., 2002; Rebelo, 2012). The upstream catchment area above the basin is 55 km$^2$. Water flows from two palmiet peat basins on the Krugersland farm upstream of the Kompanjiesdrif basin into a channel that forms briefly at the westernmost end of the wetland where the palmiet stands are not as dense. This small channel located near a bridge separating the Krugersland basins from the Kompanjiesdrif basin becomes diffuse flow as it encounters more robust and compact stands of palmiet within a span of 10's of metres. Thereafter the majority of the wetland basin is dominated by diffuse flow through the extensive stands of palmiet with very small channels forming along the perimeter of the northern and southern sides of the study area. The basin itself is located on a broad relatively near-planar valley floor with steep valley sides. Four northern tributaries enter the wetland from the Suuranys mountain range, with their alluvial fans encroaching on the wetland basin (Figure 3). The four northern alluvial fans are active, each with a running stream, which seep into the wetland basin. The fans are all covered by pasturage grown for cattle and this is the predominant vegetation, with more riverine vegetation occurring around the steams themselves. A dirt road and the railway line laterally cut across all four alluvial fans. The westernmost alluvial fan has the R62 cutting diagonally across it, in addition to the dirt road and railway line.

Further downstream at the edge of the study area, the flow of water converges once the main stands of palmiet thins out, into a deep pool which may be a by-product of the damming effect of the gabion structure present. Water flows over the large gabion structure into another pool and a subsequent gabion structure where the water becomes a medium sized channel with relatively faster flowing water. Directly downstream of the study site two southern tributaries from the Tsitsikamma mountain range, together with additional minor northern tributaries enter the Kromrivier. Headward erosion downstream of the basin near the confluence with these tributaries has propagated upstream and threatens the integrity of the wetland basin and has eroded some of the easternmost alluvial fan (Haigh et al., 2002). This headward propagation has been halted by the gabion weir structure installed at the toe of the basin by Working for Wetlands (Figure 3; Rebelo, 2012). The bed of this eroding channel is characterised by small sand bars alternating with small boulder beds. The channel is situated in a large gully that has precipitous banks and curves quite markedly before continuing its course downstream.


**Figure 3:** *The Kompanjiesdrif basin, its associated tributaries, the upstream neighbouring peat basin and the downstream eroding river reach*

As can be seen from the geology of the basin, its steep valley sides are primarily resistant quartzitic sandstone lithologies; while the valley is situated on shale of the Gydo Formation which is easily removed by scour (Figure 4). According to Reddering & Esterhuysen (1983), the majority of the Kromrivier valley lies on a Bokkeveld shale syncline flanked by resistant quartzite from the Table Mountain Group (Bickerton & Pierce, 1988).

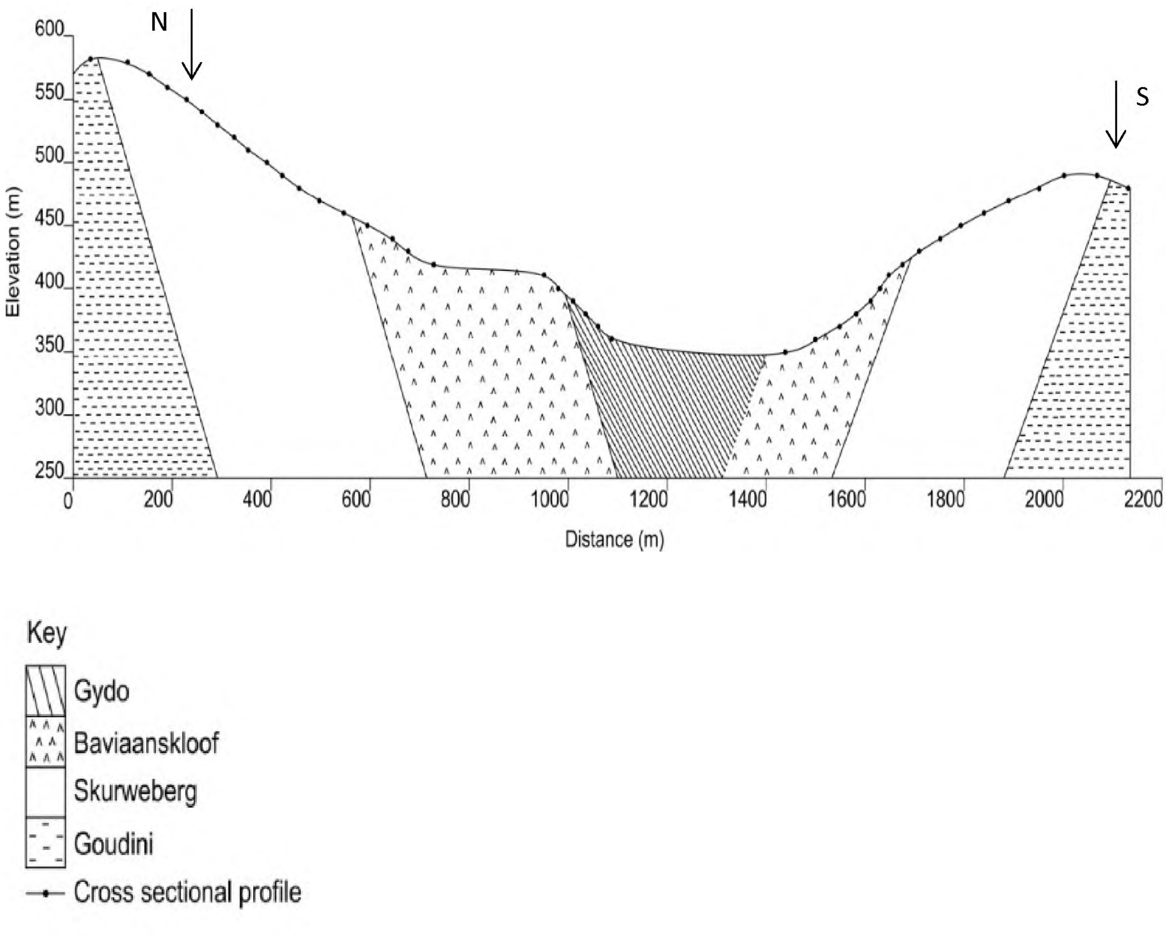


**Figure 4:** *Geological cross section of the basin compiled from the geological map (3324 Port Elizabeth)*

The wetland basin appears to lie upon the Post Africa II erosion surface which formed less than 5 million years ago, the Post Africa I surface (formed between 20 and 5 Ma) can be seen as a small level surface above the wetland basin which represents a break in slope in the northern Suuranys Mountain range. While the African Erosion surface which formed prior to 60 Ma is the level surface which forms part of the crest of the Suuranys Mountains.

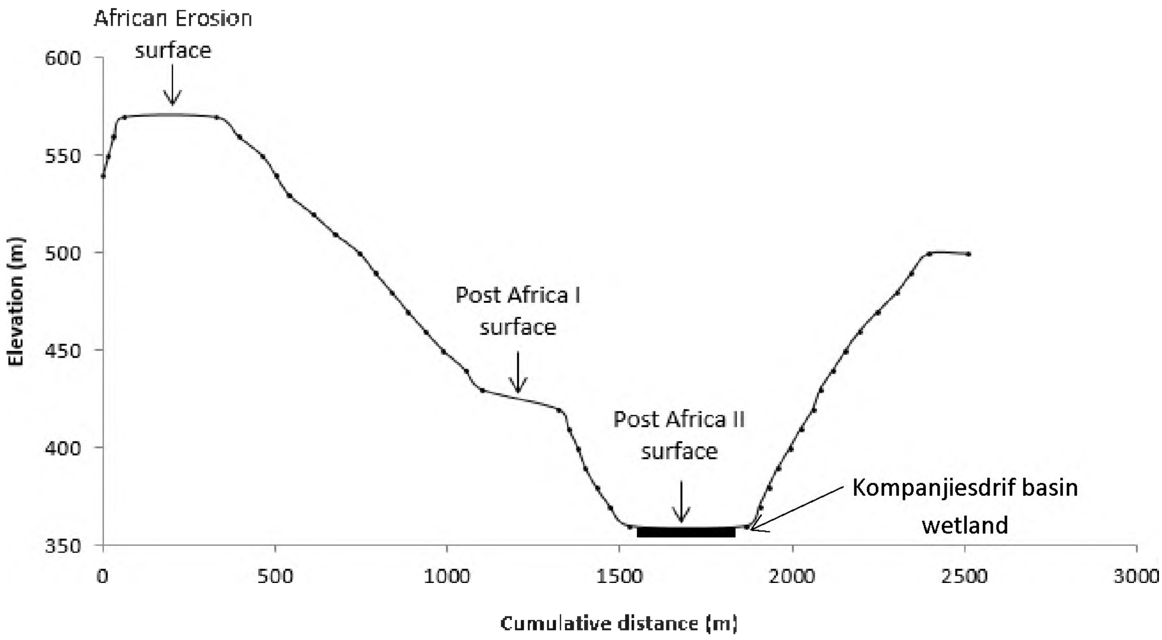


**Figure 5:** *Cross section of the basin with the position of the wetland in relation to the three erosion surfaces*

*3.2 Vegetation*

The Kromrivier forms part of the eastern Cape Floristic region which has a high diversity of biomes present (Berliner & Desmet, 2007; Cowling & Potts, 2015). According to Mucina and Rutherford (2006) the Kromrivier falls within the Fynbos biome with fragments of the Albany Thicket biome dispersed along its length.

Within the headwaters of the Kromrivier catchment vegetation falls under the Kouga Grassy Sandstone Fynbos on the southern facing slopes of the northern side of the valley although normally this vegetation colonises northerly aspects; while the Tsitsikamma Sandstone Fynbos occurs on the northern and southern facing slopes of the Tsitsikamma mountain range which borders the southern side of the valley (Mucina & Rutherford, 2006). Southern Afrotemperate Forest is distributed in small patches on the southern facing slopes of the Tsitsikamma mountain range (Mucina & Rutherford, 2006). Cultivation and alien invasive plants such as *A. mearnsii* are also known to threaten floristic diversity in the area (Mucina & Rutherford, 2006). Along the valley floor and lower extent of the slopes Langkloof Shale Renosterveld (Least threatened status) covers the landscape interspersed with azonal Fynbos Riparian Vegetation (Nsor & Gambiza, 2013; Mucina & Rutherford, 2006; Berliner & Desmet, 2007).

According to Barclay (2016) five communities of plants occur in the azonal Fynbos Riparian Vegetation of the Kompanjiesdrif and adjoining Krugersland wetland basins. The most ubiquitous is dominated by palmiet and occurs primarily in the centre of these wetlands. Other species which are abundant in the palmiet community include shrubs such as *Searsia rehmannia* and *Conyza scabrida* as well as the herb *Helichrysum odoratissimum*. Two other plant communities dominated by *Miscanthus capensis* and *Juncus Kraussii* respectively were distributed along the periphery of the wetland where palmiet thinned out. Prevalent in the *Miscanthus capensis* plant community were *Cliffortia strobillifera* and *Helichrysum odoratissimum*; while in the *Juncus Kraussii* assemblage, which acted as a transitional community, the fern *Pteris dentata* was found to be common. The fourth plant assemblage was characterised primarily by *Metalasia densa*, and occurred on the extreme margin of the wetland. The final assemblage observed was classified as mainly grassy fynbos vegetation where *Pennisetum clandestinum* was widespread and was distributed on the slopes outside the wetland.

### *3.3 Climate*

The wetland basin is positioned within an area that receives year-round rainfall, with the majority of rainfall occurring in a bimodal pattern in autumn and spring (Mucina & Rutherford, 2006; Haigh et al., 2002; Bickerton & Pierce, 1988). The Kromrivier K90A catchment experiences most of its rainfall in May (mean of 53.1 mm) and the least rainfall in January (mean of 32.9 mm; Haigh et al., 2002). The mean annual precipitation (MAP) for the K90A Catchment is approximately 716 mm while the mean annual runoff is about $142 \times 10^6$ $m^3$ (Haigh et al., 2002). Rainfall varies spatially between the higher altitude mountains that receive approximately 1140 mm per annum and the valley slopes that receive about 900 mm per annum (Mucina & Rutherford, 2006). A more marked difference is observed when these values are compared to the mean annual rainfall the valley floor receives which ranges between 770 mm and 280 mm (Mucina & Rutherford, 2006).

The Mean Annual Potential Evaporation (MAPE) is about 1980 mm for the valley floor (Mucina & Rutherford, 2006). The average daily maximum and minimum temperatures for the valley floor are 27.9 °C and 4.6 °C in January-February and July respectively, with frost prevailing between 2 – 10 days per year (Mucina & Rutherford, 2006).

The wettest year on record was 1981 where a total of 1081.8 mm of rainfall was experienced in the catchment area (Haigh et al., 2002); while the driest year on record was 1949 in which 285.8 mm was recorded (Haigh et al., 2002). The single largest flooding event which was noted at Churchill Dam (located downstream of the study area) was on the 22 November in 1997 (Haigh et al., 2002).

As a consequence of the precipitous topography of the mountains in the area, the Kromrivier possesses high and rapid runoff rates (Gull, 2012). It is thus, also a flood prone system as evidenced by flow records from the gauging station at Churchill dam (K9R01). The analysis of these flow records suggest the occurrence of several flood events in the past due to the high flow records observed (Table 1).

**Table 1:** *Flood events observed at gauging station K9R01 prior to 1985 (Modified from Bickerton and Pierce, 1988: 12)*

| **Month** | **Year** |
|---|---|
| March | 1928 |
| September | 1932 |
| May | 1935 |
| May | 1944 |
| October | 1953 |
| October | 1956 |
| June | 1968 |
| August/September | 1971 |
| August | 1979 |
| July | 1983 |

*3.4 Historical land use and anthropogenic influence*

With regards to pre-European history in the area, artefacts of archaeological significance were found in the inland dunes located near St Francis Bay and Oyster Bay (Bickerton & Pierce, 1988). The artefacts encompassed stone tools linked to prehistoric times between 200 000 and 1 000 000 years ago (Bickerton & Pierce, 1988). More recently, from the number of shell middens scattered throughout the region, occupation by two different cultural groups was noted, namely the San and Khoi people (Bickerton & Pierce, 1988).

European farming in the Langkloof valley commenced around 1775, according to the earliest record of an application for grazing rights in the area (Rebelo, 2013; Jansen & van Veen, 2014). During the 1800s land use was mainly limited to livestock farming of mainly sheep and cattle (Gull, 2012). There was limited use of artificial pastures at the time, as livestock relied primarily on the available veld for grazing (Gull, 2012). Farmers practiced frequent burning of veld to maintain this resource (Gull, 2012). In 1869 a road was constructed in the Langkloof which eventually became the R62 that forms part of a national road today (Gull, 2012). Moreover, one of the longest narrow gauge railway lines (283 km) was established in the early 1900s (Burton & Coombe, 2006; Jansen & van Veen, 2014).

These two factors led to an intensification of agricultural land use (Jansen & van Veen, 2014). From 1900 onwards major transformation of the palmiet wetlands occurred when land owners decided to use the more fertile soils of these floodplains for orchard cultivation (Rebelo, 2013). Consequent clearing of the wetlands took place which led to a decreased capability for flood attenuation in the valley (Rebelo, 2013). Many farmers were affected by a large flooding event in 1931 in which a large portion of the existing orchards were uprooted and destroyed (Rebelo, 2013). Consequently, farmers shifted to pasture cultivation for the dairy and meat industry in 1935, while orchards contributed to a smaller proportion of their original land cover (Gull, 2012; Rebelo, 2013). Around 1930 the alien invasive species *Acacia mearnsii* (Black Wattle) started colonising certain sections of the river and became a pervasive problem by about 1940 (Rebelo, 2012; Rebelo, 2013). Overgrazing and an excessively frequent fire regime as well as draining of wetlands, are the main impacts which resulted from farming practices (Haigh, 2002). Between 1954 and 2007 there was a growth in the agricultural industry of the area, resulting in the loss of large portions of palmiet wetlands along the river (Rebelo, 2012). Currently an estimated 15% of the original palmiet wetland extent still persists in the basin (Rebelo, 2013).

# CHAPTER 4 – METHODS

## *4.1 Desktop analyses*

All the desktop analyses were conducted to enable a comparison between the Kompanjiesdrif basin reach of the river and other reaches of the Kromrivier as well as to provide a catchment scale/regional context for the study area (Objective 1). With this in mind a 1:250 000 Geological map was obtained from the surveyor general of South Africa and this was cropped to the location of the study area.

A longitudinal profile of the entire Kromrivier and its major tributaries was compiled using 1:10 000 orthophotographs obtained from the Surveyor General of South Africa, with 10 m contour intervals. The width of the Kromrivier was measured every 250 m along its length using the same orthophotographs.

Using hydrology tools in ArcGIS version 10.3 and Raster DEMs of the Kromrivier catchment, the tributary catchment areas were determined, as catchment area can act as a proxy for discharge and thus the relative increase in “discharge” along the study area reach could be determined and analysed(Pareta & Pareta, 2011).

## *4.2 Field work*

### *4.2.1 Topographic surveying and location of sediment cores*

To understand and record any changes in the morphology and stratigraphy of the Kompanjiesdrif basin (Objective 2), the cross-sectional morphology of a total of eight transects was measured using Differential GPS Technology . The locations of transects are indicated in Figure 6. Transects were numbered according to the order in which they were carried out in the field. In addition to topography, this surveying allowed the locations and elevations of sediment sampling cores to be established. Examination of sediment cores was used to delineate the extent of the wetland in conjunction with observations of changes in vegetation.

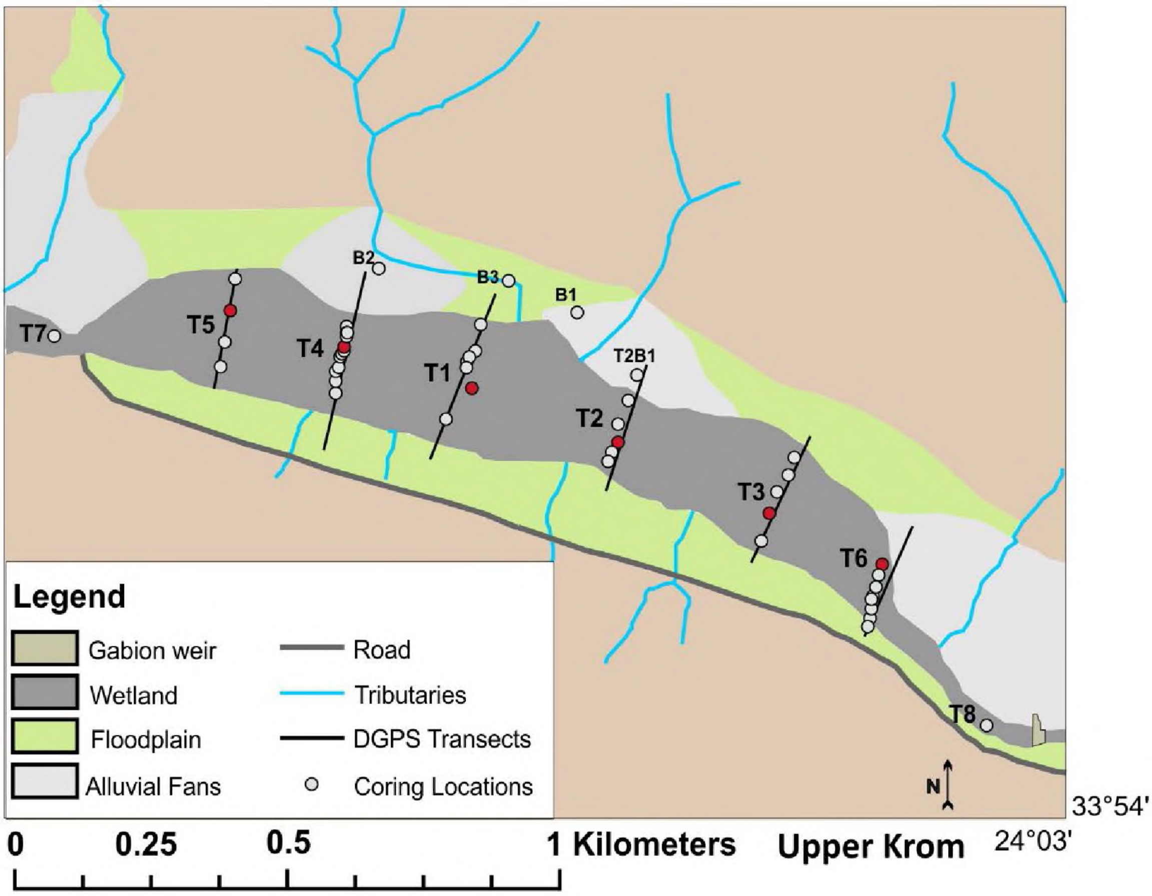


**Figure 6:** *Map of the study area showing the location of topographic surveying and the associated location of sediment cores. The location of three cores to bedrock are shown (B1, B2 and B3), the points highlighted in red were used to compile the longitudinal profile of the basin.*

*4.2.2 Sediment sampling*

Coring was carried out along all the Transects in order to describe the sedimentary fill present in the wetland basin (Objective 2; Figure 6). A small gouge corer (Grenfell et al., 2008) was used for the majority of the cores, with a Dutch Auger used for some cores to bedrock and a Russian peat corer (Watters & Stanley, 2007; Grenfell et al., 2010) for certain cores to bedrock as well as for carbon dating of organic samples where control was needed on sample depth. Three to eight cores per transect were described according to their stratigraphy along six of these transects. The variation in colour and texture of sediment cores was recorded in the field. When there were any marked changes in the stratigraphy of a core; samples were collected for analysis. In addition to these logged cores; depth to refusal (to sand or bedrock) was recorded using a small gouge corer every 10 – 25 m along each transect. Any deviations or anomalies observed in the sedimentary fill were recorded and a sub-surface cross section of such anomalies was plotted. Carbon dating of samples obtained from these sites was undertaken (Objective 3).

### *4.3 Laboratory analyses*

#### *4.3.1 Soil preparation*

Samples collected in the field and brought back to the laboratory were dried in a drying oven at 70°C for 72 hours. Thereafter, the samples were gently disaggregated using a pestle and mortar.

#### *4.3.2 Organic matter content: loss on ignition*

Loss on ignition was chosen as it is simple and cost effective. In addition the loss on ignition method is viewed as providing a sufficiently accurate approximation of the total organic carbon content of a sample (Wang et al., 2011). In this study the weight of empty, cleaned and dried porcelain crucibles was recorded to 2 decimal places. Approximately 8-10 g of each sediment sample was placed in a porcelain crucible, the crucible re-weighed and the weight recorded to 2 decimal places. The crucibles of soil were then placed in a muffle furnace which gradually reached a temperature of 450°C (Leonard et al., 2002; Donkin, 1991). Samples were removed after 4 hours and placed in a glass desiccator for an hour to avoid moisture re-absorption while they cooled (Leonard et al., 2002). Once the crucibles and ash had cooled, samples were then re-weighed and the weight recorded to two decimal places. The loss in mass following ignition was used as a representation of the organic matter content in each sample, expressed as a percentage of the original dry mass of the sample.

#### *4.3.3 Particle size analysis*

A much smaller sub–sample of 2 – 4 g of the dried soil samples was used for particle size analysis. About 5 ml of 30% hydrogen peroxide was added to each sub-sample (Gray et al. 2010). The samples were then placed on a hot plate until organic matter had been removed (Gray et al. 2010). Approximately 5 ml of 3 % sodium hexametaphosphate solution was added to each sample beaker in order to disperse sediment. Samples were then additionally dispersed by 2 minutes of ultrasonic dispersal. Thereafter, a Malvern Mastersizer 3000 laser granulometer was used to measure particle size (Grenfell et al., 2010). The median (dsv50) particle diameter of each sample was recorded.

### *4.3.4 Carbon dating*

Carbon dates were obtained for selected samples by Beta Analytic Inc, Miami, Florida, USA, using Accelerator Mass Spectrometry Carbon 14 dating. Organic sediment was used as opposed to plant material found in these samples, as plant roots potentially represented more modern contamination. The 'acid washes' pre- treatment of samples was applied. The mathematic approach used to calibrate the 14-C dates was after that of Talma & Vogel (1993). Calibration of the 14C-ages was determined using the SHcal13 data set (Hogg et al., 2013).

# CHAPTER 5 – RESULTS

## *5.1 Longitudinal profile*

The longitudinal profile is initially logarithmic, with the upper reach possessing a steep initial slope of 14.7 % (0.0 -2.0 km) which transitions into a slope of 3.2 % (2.0 - 6.8 km, Figure 7a). Subsequently, a steep step that has a slope of 42.8 % and an approximate height of 80 m interrupts the logarithmic trend of the profile. This step ends at about 8 km along the profile and the profile continues with an average slope of 2.2 % which terminates at 10.8 km. Thereafter, the profile becomes more planar with a slope of 0.8 %, which further decreases to a slope of 0.4 % along the remainder of the profile section plotted as far downstream as Churchill dam.

There are just over 30 major tributaries which feed into the Kromrivier. Two of the larger southern tributaries' confluences with the Kromrivier are immediately downstream of the abrupt step noted on the longitudinal profile at 9.6 km and 10.8 km (Figure 7a).

With regards to the underlying geology most of the formations present comprise quartzitic sandstone, with subordinate layers of shale. The Gydo formation is the only exception as it comprises a 500 m thick layer of shale. From 5.4 km to 10.0 km there is a transition from the more resistant lithology of the Goudini, Skurweberg and Baviaanskloof formations to the less resistant Gydo formation.

The location of the study area in the upper reaches of the Kromrivier catchment is highlighted in Figure 7a. Although it appears as if the study reach occurs on the more resistant quartizitic sandstone of the Baviaanskloof formation, from Figure 4, it is clear that the basin lies on the less resistant shale of the Gydo formation and is flanked on either side by the more resistant Baviaanskloof formation. More evident from the profile is the occurrence of two major northern tributaries which enter the Kromrivier on the eastern and western margins of the study area and two southern tributaries that enter the Kromrivier wetland on the eastern end of the study area. From the width vs distance graph, the broad nature of the Kompanjiesdrif basin river reach in relation to the width of other reaches is also highlighted (Figure 7b).

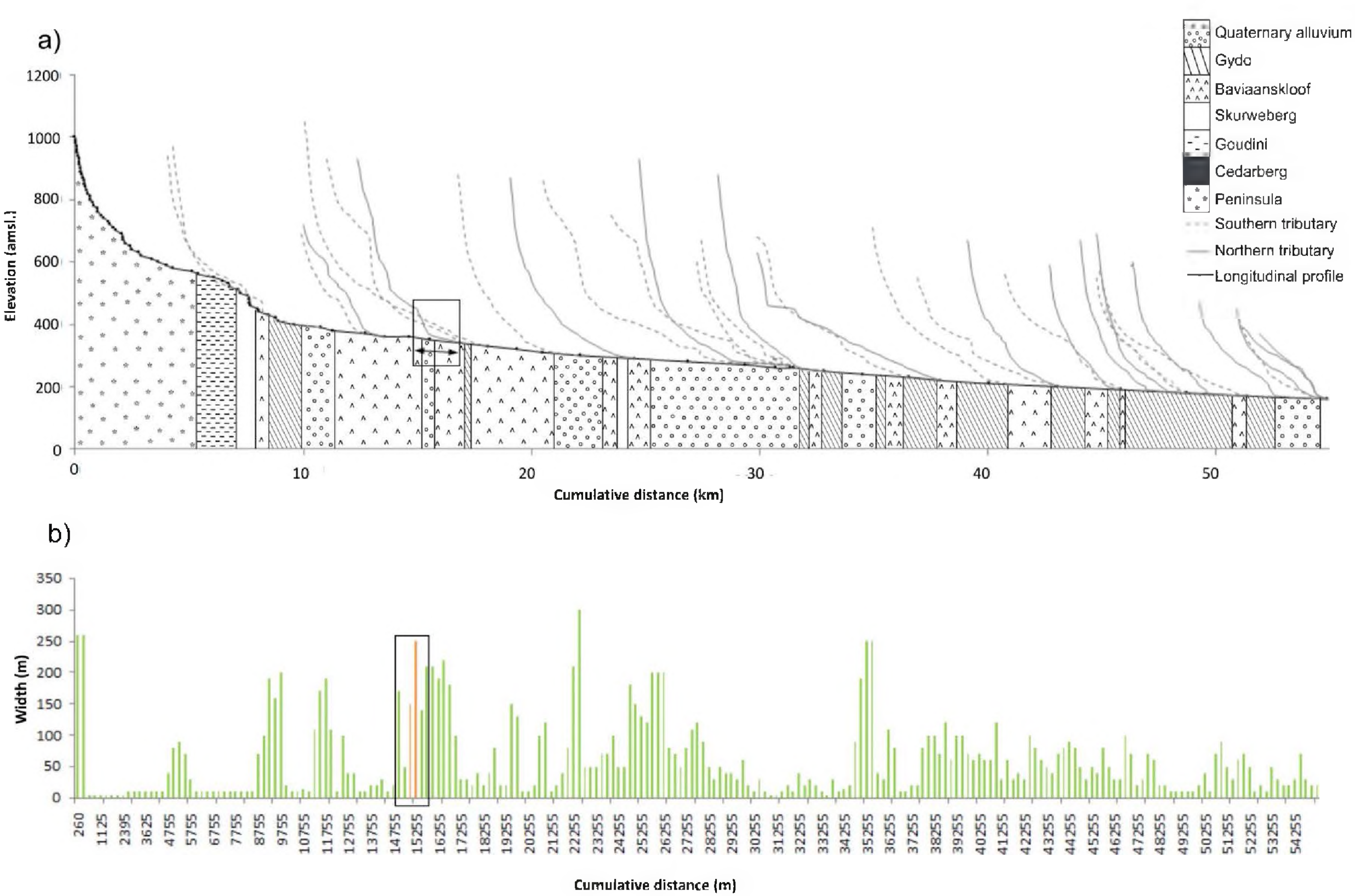


**Figure 7: a)** *Longitudinal profile of the Kromrivier and* **b)** *wetland or channel width in relation to distance on the Kromrivier from its head to Churchill Dam*

The cumulative catchment areas of the northern and southern tributaries increase at approximately the same rate over the cumulative distance of this reach of the Kromrivier i.e. 0.0 – 76.3 km (Figure 8). There are fewer northern (45) than southern (71) tributaries draining into the Kromrivier over the section examined. Nevertheless, although there are fewer northern tributaries; the total cumulative catchment area of the northern tributaries is greater (352.1 km²) than that of the southern tributaries (339.3 km²), demonstrating that southern tributaries have smaller average catchment sizes than northern tributaries.

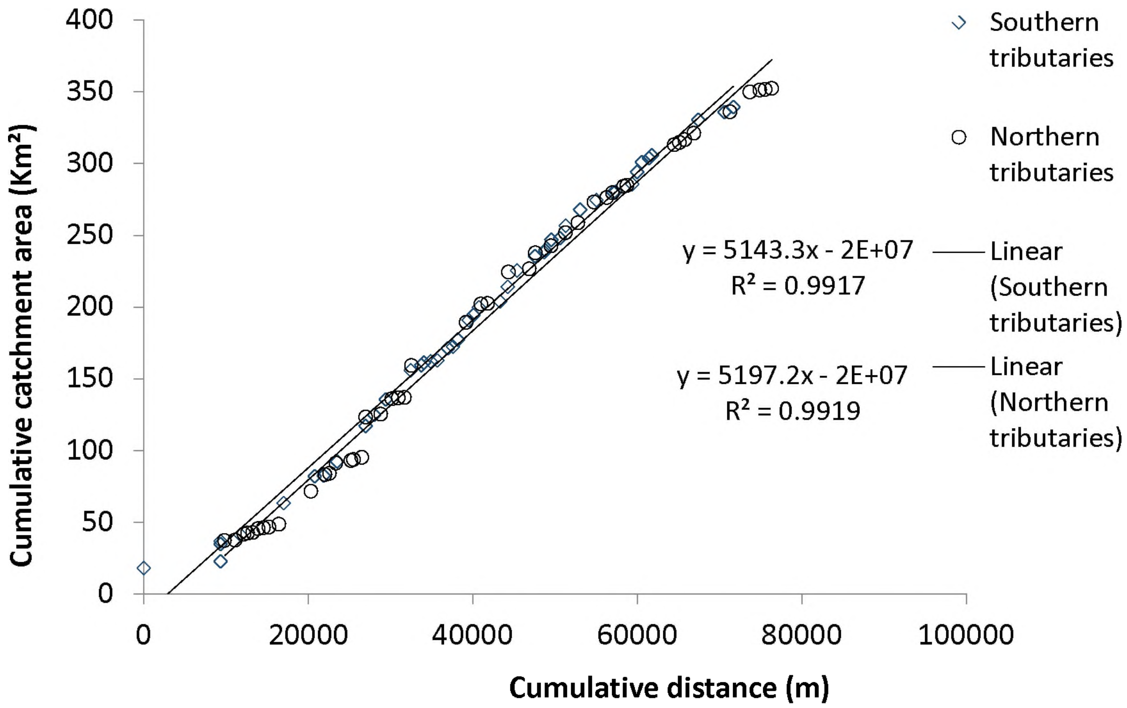


**Figure 8:** *Cumulative catchment area of the Kromrivier's tributaries*

### *5.2 Kompanjiesdrif basin cross-sectional morphology*

The morphology of the wetland basin was examined in order to investigate the influence of alluvial fans on the topography and sediment dynamics of the study area. The width of the basin was variable along its length with the surveyed cross sections at Transects 1, 3 and 5 (Figure 9) representing broader sections of the wetland basin between alluvial fans. In the broad zones, widths ranged between 200 m and 264 m. The basin is bounded on the northern side by a grassy floodplain region which progresses upslope into steep mountainous terrain of the Suuranys Mountains. On the southern side the transition is more abrupt and the gradient increases rapidly from the valley bottom to the Tsitsikamma Mountain Range. The valley floor, which is covered primarily by the wetland, has local relief of approximately 1 m over a distance of more than 100 m for Transects 1, 3 and 5 which may be caused by the manner in which palmiet grows across the water's surface in a uniform mat (Figure 9). The water table at Transects 1 and 5 slopes predominantly towards the northern margin of the wetland; as the highest point of the land surface is located relatively close to the southern margin. Depth to refusal, which was often represented by a sand layer that collapsed and could not be penetrated by coring, was commonly found at a relatively consistent depth beneath sedimentary fill with higher organic content. The sand layer in Transects 1, 3 and 5 is present beneath the land surface at depths between 1.20 m and 3.30 m; 1.65 m and 2.30 m; 1.50 m and 3.00 m respectively.

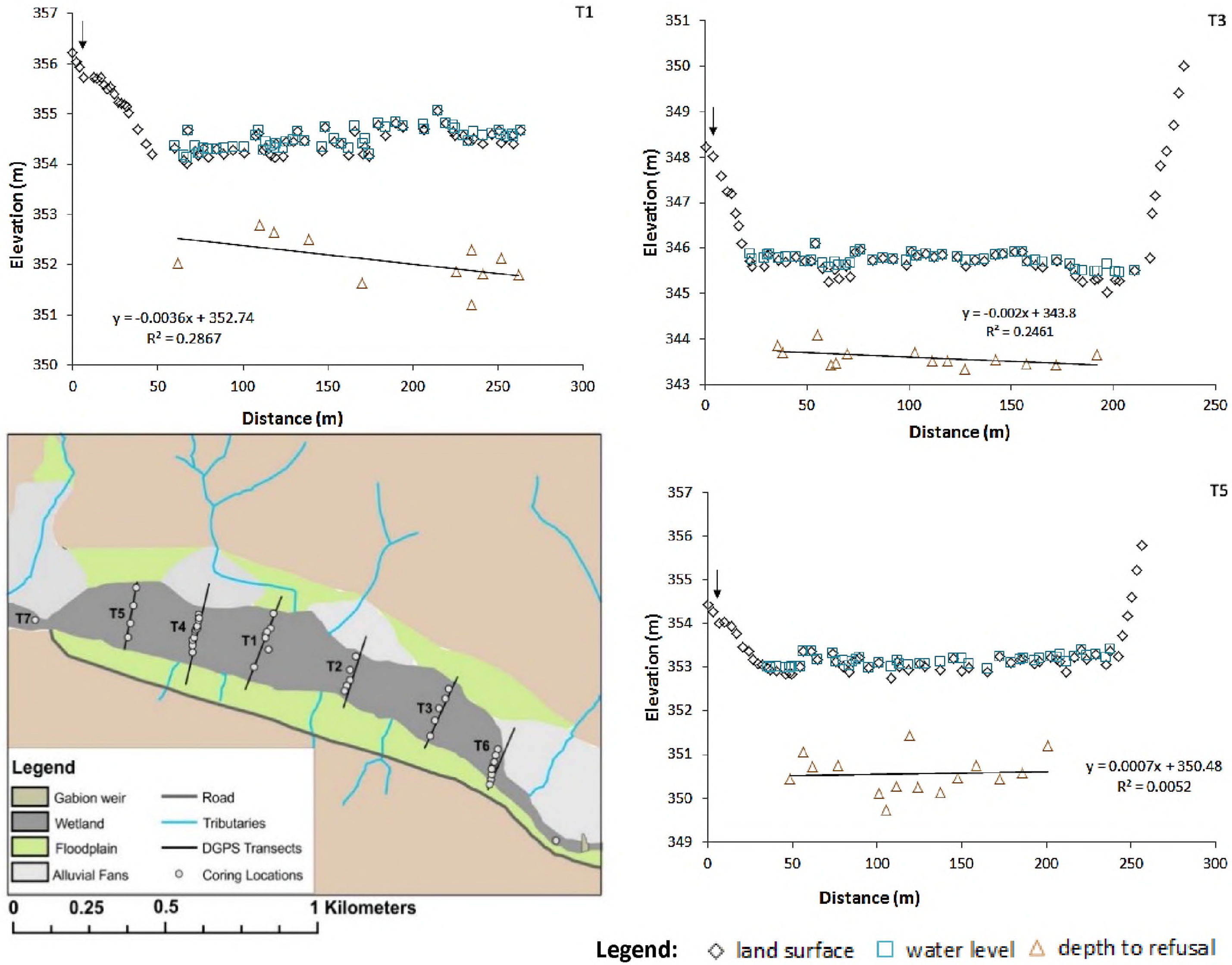


**Figure 9:** *Cross-sectional profiles of Transects 1, 3 and 5 showing the depth of a sand layer across the entire wetland*

Compared to Transects 1, 3 and 5; Transects 2, 4, and 6 are narrower since they occur adjacent to tributary alluvial fans. Transect widths range between 115 m and 212 m (Figure 10). Rather than the wetland having steep sides as for Transects 1, 3 and 5, the land surface to the north of the wetland slopes gently down to the wetland surface, which is near planar again due to the ecological processes governed by palmiet (Figure 10). The water table at Transects 2, 4 and 6 slopes predominantly towards the southern margin of the wetland, as the highest point of the land surface is located relatively close to or at the northern margin of the wetland. The sand layer below the wetland surface in Transects 1, 3, and 5 is also present in Transects 2, 4, and 6 at depths ranging between 2.10 – 2.70 m; 1.00 – 2.50 m and 1.30 – 2.40 m respectively.

The depth to refusal on Transects 2, 4, and 6 were characterised by localised features where the sand layer was much deeper than was expected. Furthermore, at points where the depth to sand was greatest, valley-fill sediment was absent and palmiet was found growing across a column of water that extended to depths between ~5.00 m (Transect 4) and 8.20 m (Transect 2). The base of these features contained organic-rich sludge. An identical feature was found in Transect 7, which unfortunately could not be surveyed using DGPS. Nevertheless, as with the other features of this kind, this was also located adjacent to a very large alluvial fan and had a depth of 8.00 m. These features appear to be narrow and deep, and seem to be longitudinally discontinuous and occur opposite impinging north-bank tributary alluvial fans (Figure 10).

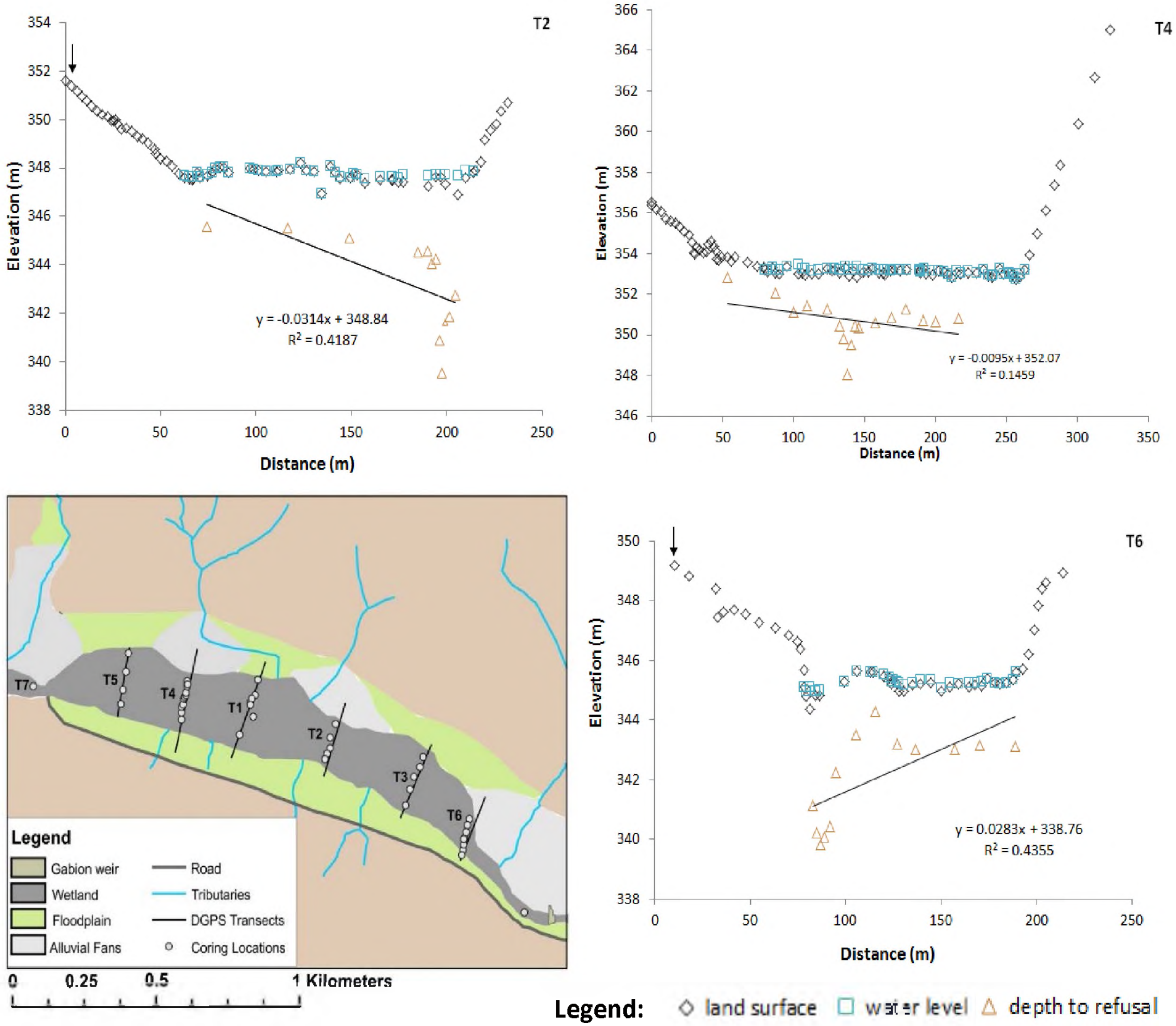


**Figure 10:** *Cross-sectional profiles of Transects 2, 4 and 6 showing the depth of a sand layer across the entire wetland*

### *5.3 Characteristics of valley fill*

Valley fill sediments were sampled from about 30 logged cores along the six surveyed transects, with 3 – 8 cores along each transect. The majority of the cores were fairly similar stratigraphically, with organic layers of up to 16.0 % organic matter content that contained fine sediment of silt and clay. These organic-rich layers were interrupted by less organic, coarser layers of sediment (with a median (dsv50) particle size up to 360.0 μm). The sand layer mentioned previously was present in most cores.

The profiles of six of the cores in Transect 1 are outlined in order to show major trends and any variation in the stratigraphy across the wetland basin. The first transect was located between two fairly small alluvial fans and consisted of 8 cores orientated roughly north-south across the wetland basin, with T1C1 in the north and T1C9 in the south (Figure 11). The depth of cores to the point of refusal increased systematically from north to south across the wetland.

Core T1C1 was located approximately 5 m from the northern margin of the wetland (Figure 11). The base of this core was characterised by very fine sand at a depth of 1.20 – 1.63 m. This lower unit was overlain by clay and compact organic sediment. The sequence as a whole fined upwards with increasing organic content in the core from <1.0 % near the base to 5.6 % at a depth of 0.72 m. The upward fining sequence ended at 0.04 m where a thin layer of sandy clay covered the underlying finer sediment.

A thin layer of sandy clay at a depth of 1.80 – 1.85 m comprised the base of Core T1C2. The remainder of the core was composed of clay and organic material. Most of the core exhibited an upward fining sequence. This sequence ended at 0.10 m where a thin layer of brick-red clay covered the underlying finer sediment. Organic content increased upwards in the sequence from a low of 1.0 % at the base. The highest organic content was at 1.45 m, while organic content generally remained above 6.4 % in the shallower parts of the core.

At a depth of 1.50 – 1.90 m Core T1C3 contained sandy clay. This was overlain by compact organic material, interrupted by two minor sandy clay layers. Overall the core exhibited an upward fining sequence with organic content increasing from 0.3 % near the base to 4.7 % at a depth of 1.47 m. The upward fining trend ended at 0.10 m where a layer of brick-red clay covered the underlying finer sediment.

Sediment in Core T1C6 was 0.80 m below water level and comprised an overall upward fining sequence of clay and silt mixed with organic material, although there were two minor coarse layers at the base. Organic content was low at the base of the core at generally less than 3.0 %. The highest organic content was at a depth of 2.20 m, (12.0 %) and remained above 3.0 % at shallower depths in the core.

The base of Core T1C8 contained sand at a depth of 2.30 m – 2.42 m, which was overlain by a thin layer of clay. The remainder of the core was characterised by an organic slurry (very watery fine sediment), organic sludge (thicker than slurry in consistency but viscous rather than solid) and more compact organic material mixed with silt and clay. The entire profile constituted an upward fining sequence with organic content rising from a low of 1.4 % near the base, to generally above 7.0 % in the upper parts of the core.

Two major trends emerged from the sedimentary fill in the upper parts of the cores in Transect 1. There was no apparent systematic variation in median particle size (dsv50) across the valley as the calibre of sediment comprised predominantly finer particles. Yet there was generally a consistent increase in organic content from north to south observed across the basin.

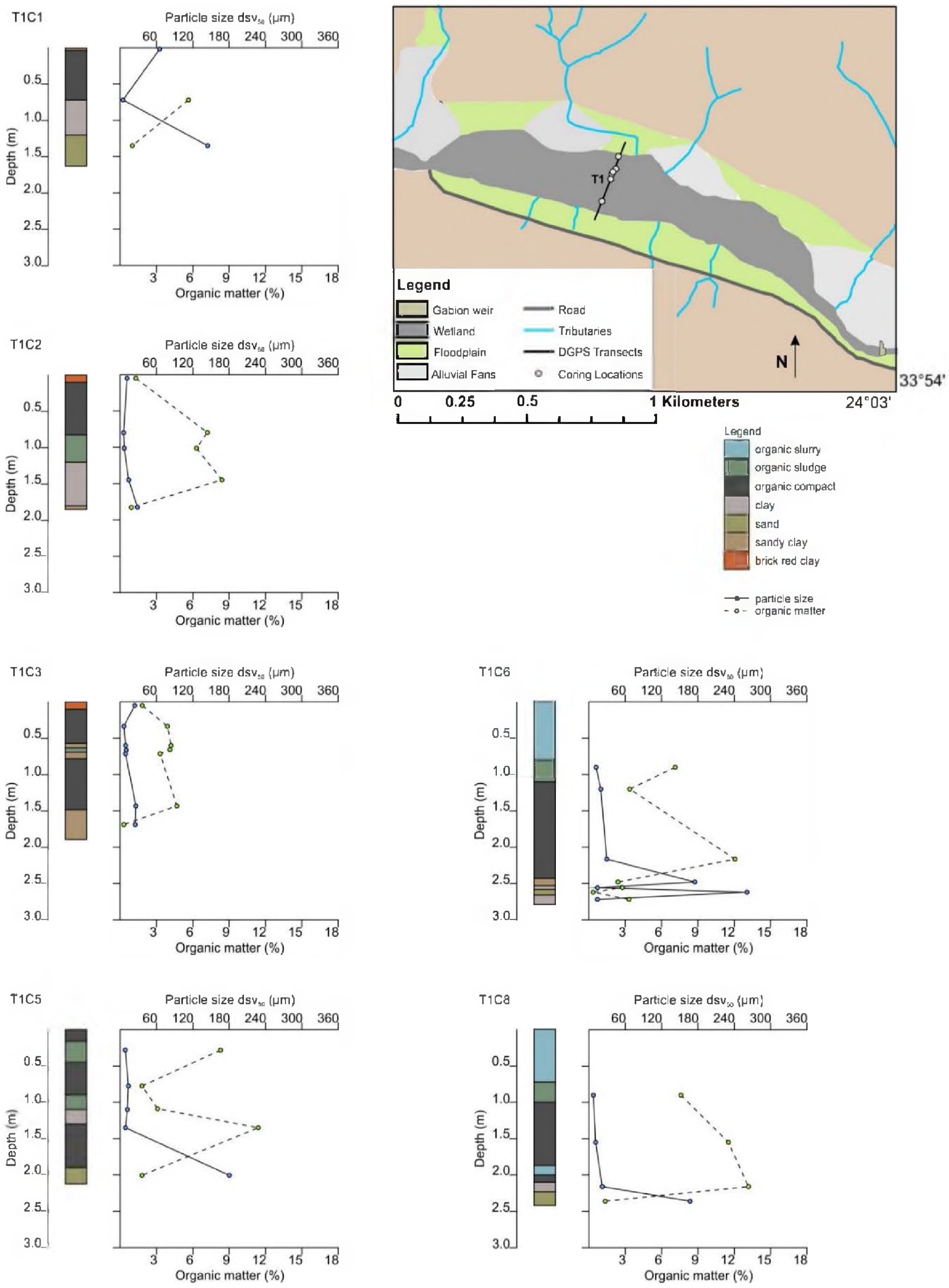


**Figure 11:** *Location and stratigraphy of cores in Transect 1 showing variation in particle size and organic matter content with depth*

The remainder of the cores that were augured in other transects were similar to the cores found in Transect 1. Slight differences were observed but the main trends persisted throughout the majority of the wetland basin. Differences included the absence of the brick-red clay layer occasionally observed at the surface of the wetland. A further disparity, was the occurrence of thin layers of sand or sandy clay (T6C7 see Figure 12, T6C8, T6C12, T6C14), or multiple layers of sand mixed with silt similar to those observed in T1C3, for example T6C10. The remainder of the core profiles not presented in the results are shown in Appendix 1.

Transect 6 was augured very close to the largest alluvial fan present in the wetland basin downstream of the other transects, with Core T6C7 in the middle of the transect presented as Figure 12. The sediment in Core T6C7 comprised predominantly unconsolidated dark silty-clay layers with between 4.0 % and 9.0 % organic material (Figure 12). These darker layers were interrupted by 3 lighter layers at depths of 1.06 m - 1.27 m, 1.37 m - 2.06 m and 2.80 m - 3.30 m, which contained a mixture of very fine sand, silt, clay and organic material with organic contents between 0.6 % and 1.3 %. These layers were larger than any layers containing sand previously encountered in other transects, which suggested that a different trend or process took place along this transect. Moreover, there were multiple layers as opposed to the more prevalent single thin layer of fine to very fine sand.

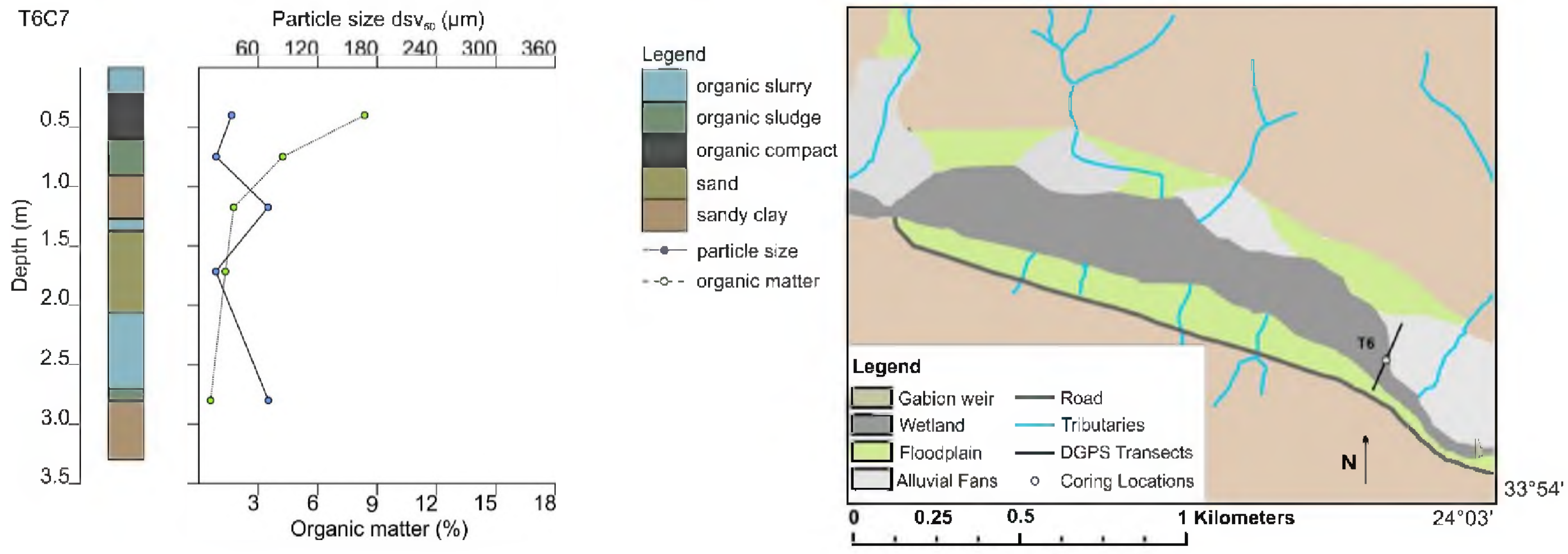


**Figure 12:** *Location and stratigraphy of a single Core (T6C7) in Transect 6 showing variation in particle size and organic matter content with depth*

The profiles of six cores one from each transect were compared in order to show major trends in the stratigraphy longitudinally along the wetland basin. The cores were located close to the northern margin of the wetland orientated roughly west to east across a lengthwise section of the wetland basin, with T5C1 in the west and T6C7 in the east (Figure 13). There was no apparent systematic trend in median particle size (dsv50), organic matter content or depth to refusal observed in a downstream direction along the wetland basin. However, there were coarse sediment layers present in all the cores.

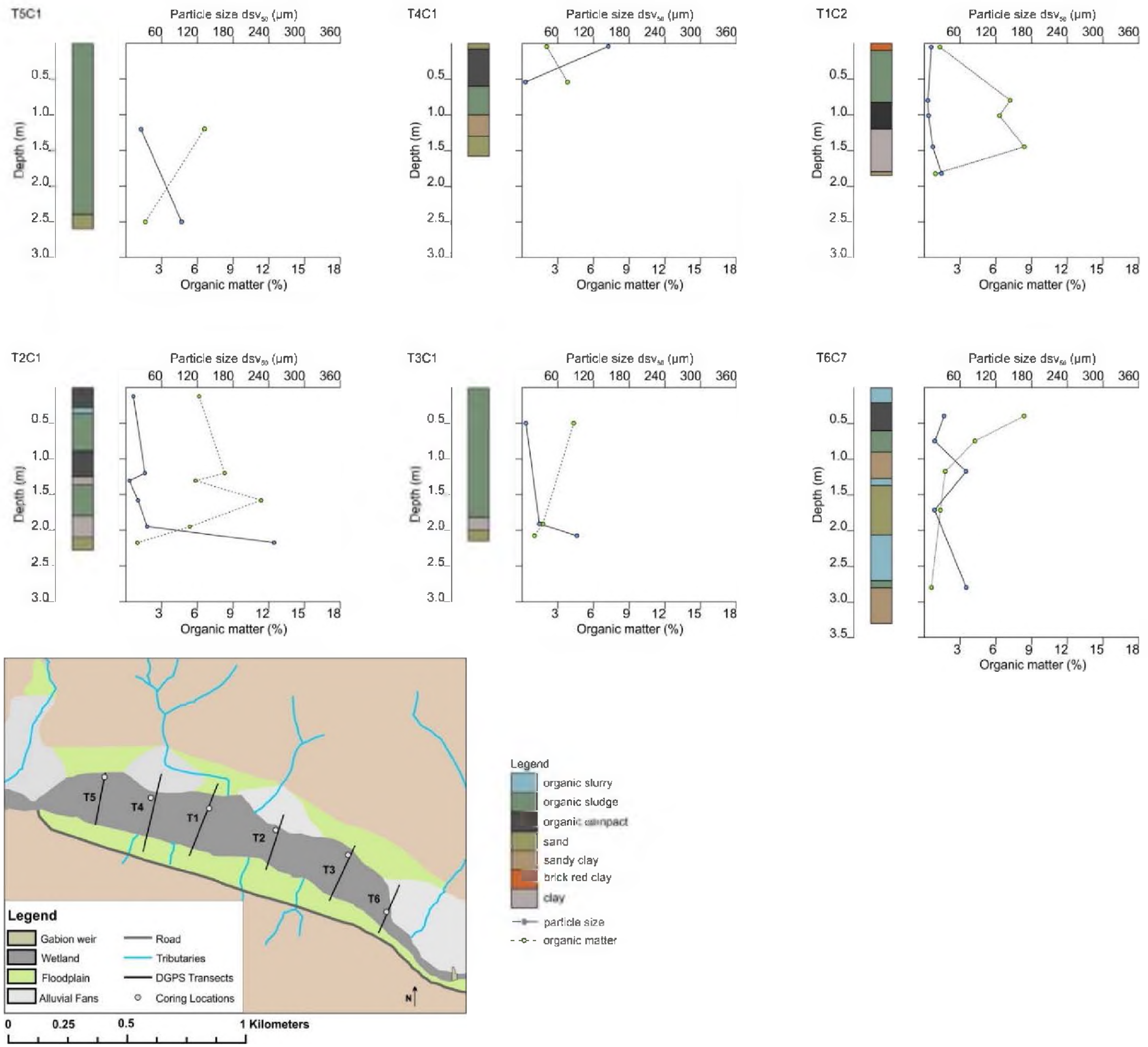


**Figure 13:** *Location and stratigraphy of cores across a longitudinal section of the wetland showing variation in particle size and organic matter content with depth*

Four cores were sampled from the deep narrow discontinuous features mentioned earlier and were strikingly different from the cores described thus far. Each of these was found in a different transect, although all were found in transects positioned opposite an alluvial fan, namely Transects 2, 4, 6, and 7.

Located about 25 m from the southern edge of the wetland (Figure 14); Core T2RPC was 7.50 m deep. This core was covered in a dense stand of palmiet, which, once penetrated, gave way to water to a depth of 7.00 m (Figure 14). From 7.00 m to 7.20 m the core consisted of a layer of very fine sand mixed with organic matter with an organic content of 8.4 %. Below this a sandy clay layer was found to a depth of 7.50 m, with organic material that ranged between 8.7 % and 9.5 % organic matter content.

Transect 4 spanned a section about 210 m wide and Core T4RPC was located about 140 m from the northern edge of the wetland on this transect (Figure 14). Floating palmiet was growing over a water column which extended to a depth of 4.50 m, below which three layers of sediment occurred between 4.50 m and 5.00 m. The first and last layers were composed of silt/clay mixed with organic material (26.0 % organic content) in the first layer and 8.8 % - 12.0 % organic material in the third layer. The central layer was a thin layer of sand (dsv50 = 348.0 µm) with 3.0 % organic content.

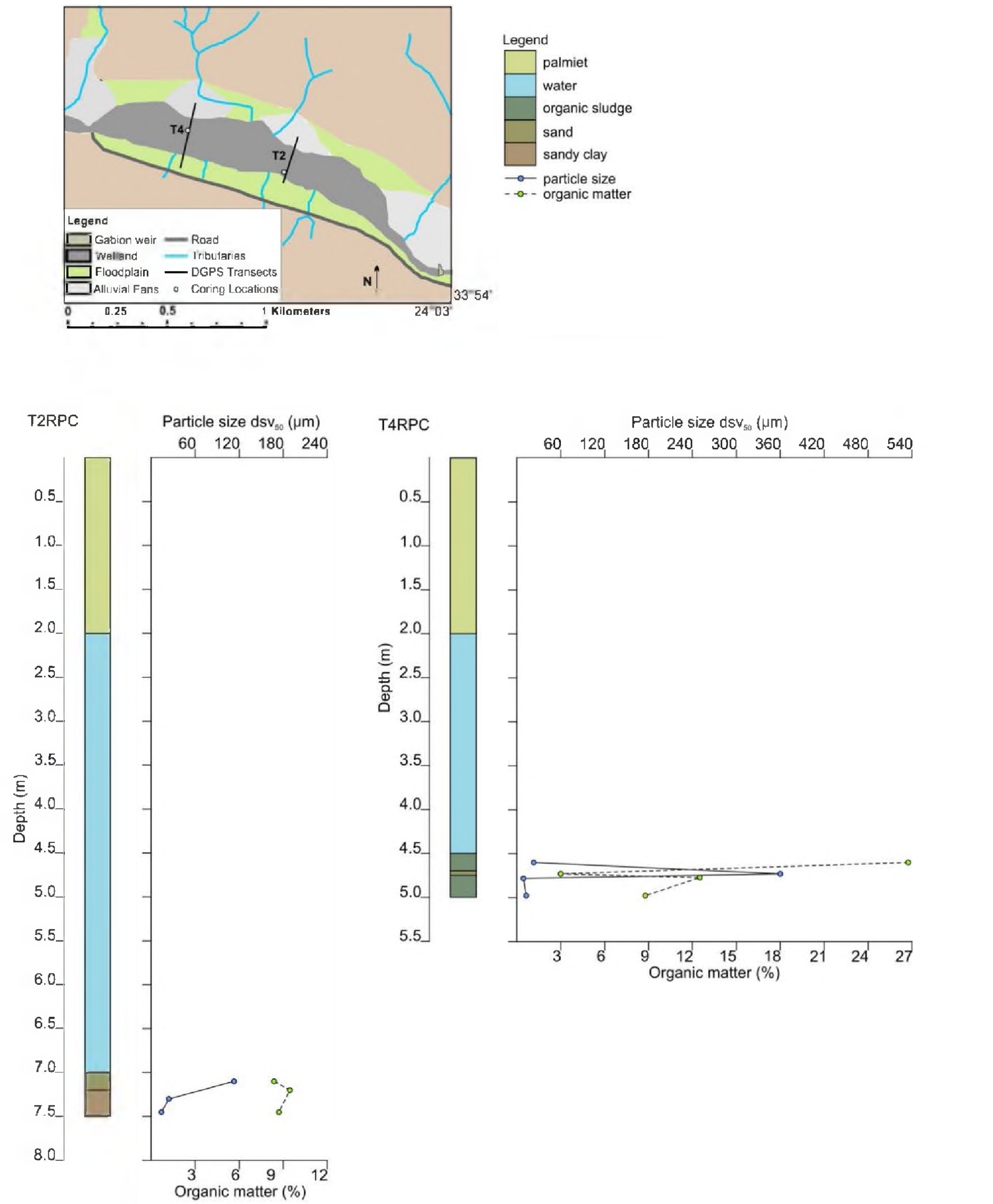


**Figure 14:** *Location of cores in Transects 2 and 4 that had palmiet growing across deep trench-like features, with associated core depth, particle size and organic matter content*

Near the northern edge of the wetland on transect 6, Core T6RPC comprised a thick mat of palmiet that floated above a water column which extended to a depth of 4.50 m (Figure 15). From 4.50 m – 4.90 m the core was characterised by fine sand with an organic content of 0.9

to 1.3 %. Below this a silt/clay layer was found to a depth of 5.00 m, mixed with organic matter with an organic content of 3.0 %.

One of the deepest cores, Core T7RPC was located opposite the westernmost alluvial fan, upstream of all the surveyed transects (Figure 15). A mass of palmiet covered a water column which extended to a depth of 7.50 m. A silt/clay layer was found from a depth of 7.50 m to a depth of 8.00 m, with organic material that ranged between 21.0 % and 34.0 % organic content.

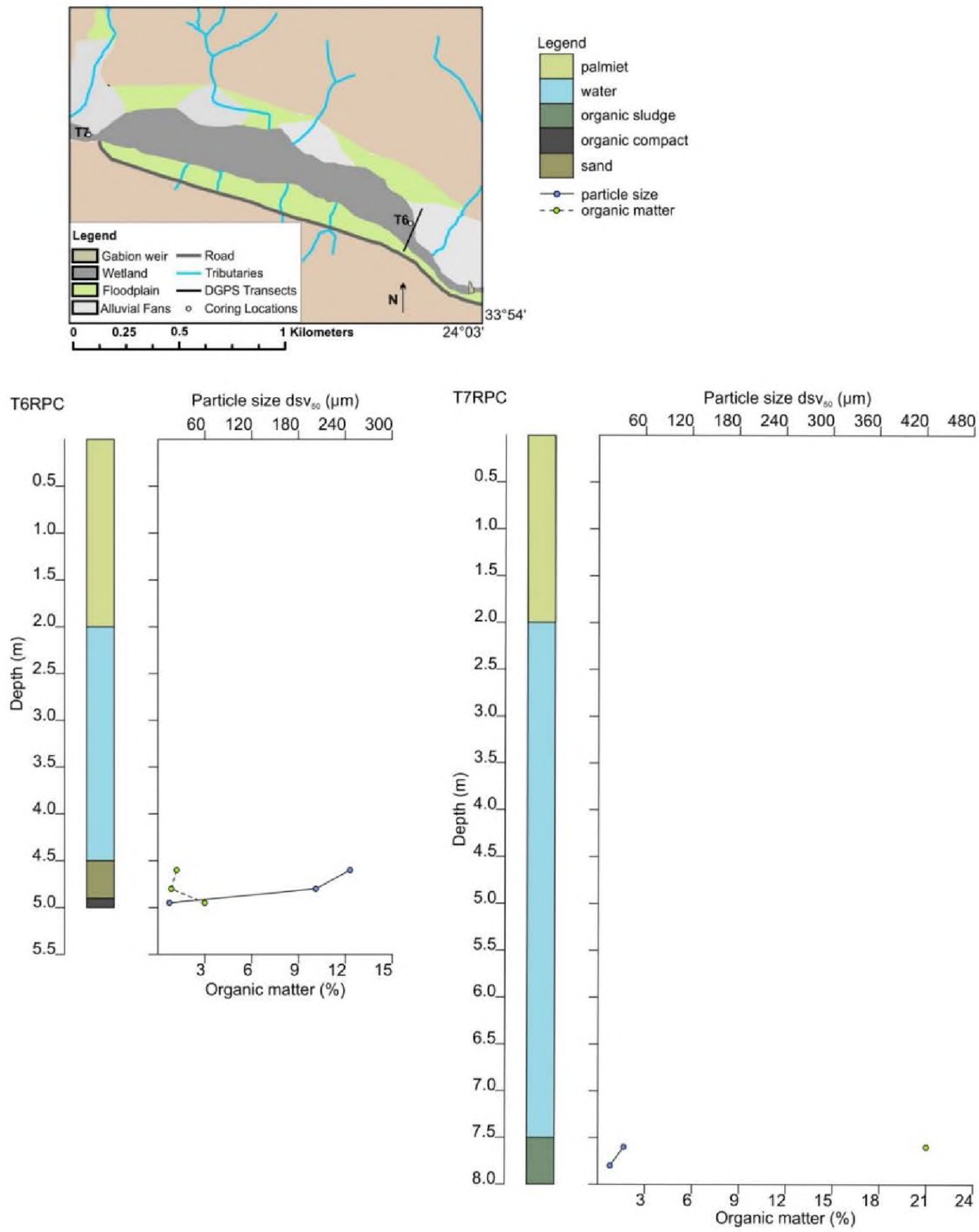


**Figure 15:** *Location of cores in Transects 6 and 7 that had palmiet growing across deep trench-like features, with associated core depth, particle size and organic matter content*

### *5.4 Cores to bedrock*

These cores were either located on the northern floodplain where alluvial fans fed into the wetland or within the wetland itself beneath a dense mat of palmiet.

From the stratigraphy of Core B2 one upward coarsening sequence (0.00 – 0.90 m) and three upward fining sequences (0.90 – 1.38 m; 1.38 – 1.93 m and 5.90 – 7 .40 m) were observed. Throughout the core, sediment ranged from clay to silt or sand or a combination of these different sediment clast sizes. This core was located on an alluvial fan on the northern floodplain (Figure 16)

Core B3 was located between two alluvial fans on the northern side of the floodplain (Figure 16). There were two upward fining sequences present in this core at depths of 2.10 – 2.55 m and 3.00 – 4.00 m. The core in question did not reach a depth comparable to that of B2, which may be explained by the presence of a single layer of coarse (cobble to boulder) material that limited penetration of the corer before bedrock could be reached.

As in Core B2, the type of sediment present was highly variable with clay, silt and sand present in varying degrees. Sand mixed with clay was common as was sediment comprising separate layers of sand and clay. For example, the upward fining sequence from 2.10 – 2.55 m comprised only sand particles, whereas the upward fining sequence from 3.00 – 4.00 m contained clay mixed with coarse sand at its base, which became progressively finer and was overlain by mottled clay mixed with minor amounts of sand.

The shallowest of all the bedrock cores, Core B1, was augured on an alluvial fan on the northern floodplain (Figure 16). Given the shallow nature of Core B1, it may have reached bedrock at 4.50 m but it is more likely that the corer hit a layer of boulders as the coring site was on an alluvial fan. The sediment at the base of the core exhibited an upward fining sequence of sand at a depth of 2.85 – 4.50 m. Two additional upward fining sequences characterised the remainder of the core at depths of 1.00 – 2.85 m and 0.00 – 1.00 m respectively. The second sequence initially comprised sand and thereafter transitioned into sediment which was more clay-like in texture. The final sequence near the upper end of the core contained pebbles at the base and graded into silt mixed with sand, which was followed by dark organic topsoil.

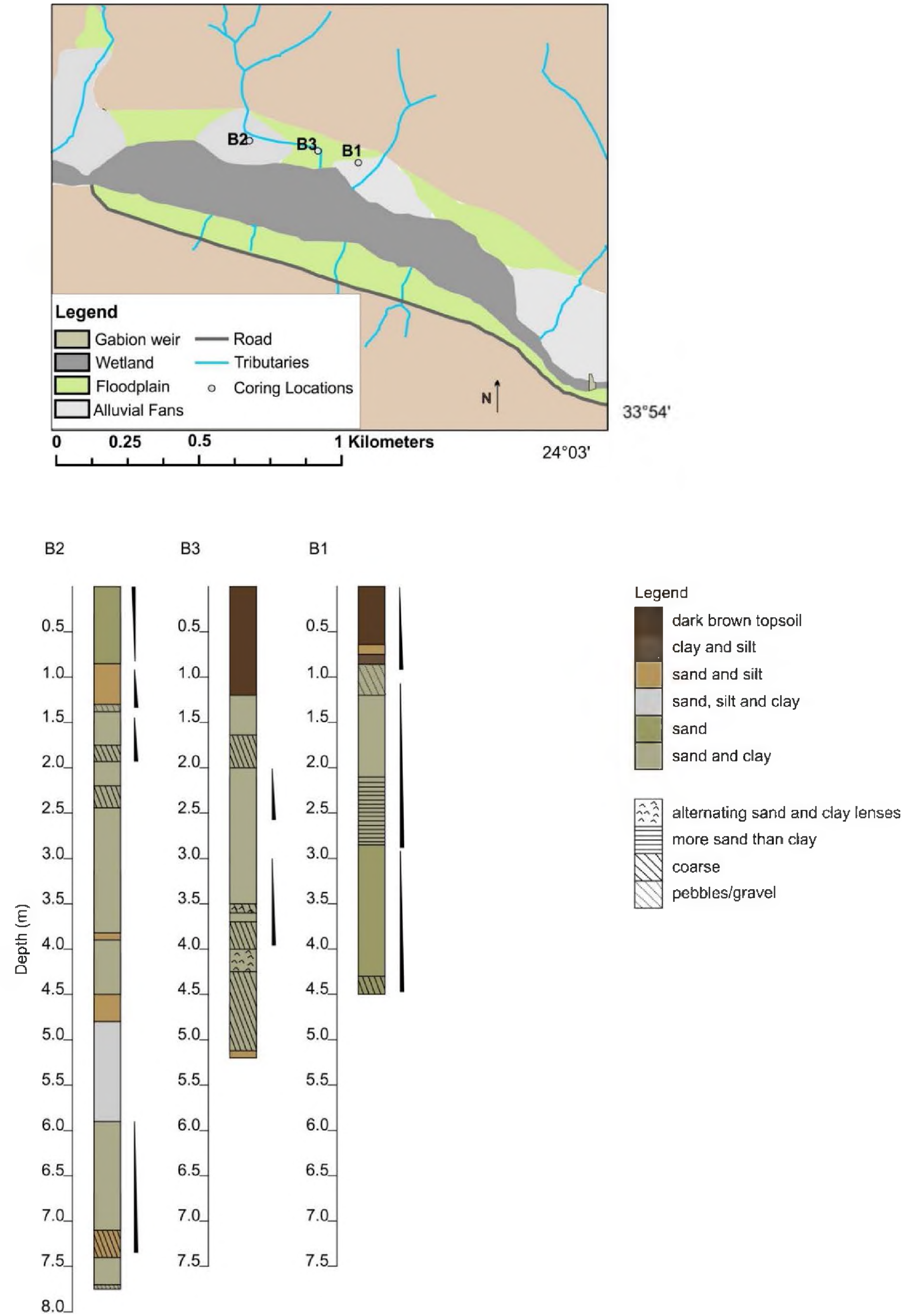


**Figure 16:** *Location of bedrock cores on the northern floodplain with associated core depth, particle size and organic matter content*

Situated extremely close to the edge of the wetland on the northern floodplain was Core T2B1 (Figure 17). In this core, which was augured to bedrock, layers of grey clay to silt sediment alternated with sandy layers. Three of the sand layers were sampled at depths of 4.00 – 4.30 m, 5.00 – 5.30 m, and 5.70 – 6.00 m with median particle sizes of 170.0 µm, 134.0 µm and 192.0 µm respectively.

Two of the trench-like features found in Transects 7 and 2 respectively reached bedrock and have hence been included as bedrock cores. Located within the wetland itself, Core T7B1 was cored to bedrock beneath the surface of the thick mat of overlying palmiet (Figure 17). Once through the thick layer of vegetation (about 0.00 – 2.00 m), the auger entered the water column which extended from 2.00 to 7.89 m. Sediment was only encountered 0.11 m above bedrock (depth of 8.00 m). The sediment sample collected consisted of unconsolidated highly organic material with a sediment particle size of 19.2 µm (silt/clay) and an organic matter content of 19.90 %.

The depth to bedrock for Core T2B2 was 8.20 m and this core was also located within the wetland (Figure 17). The sediment found just above bedrock had a median particle size of 31.0 µm (silt/clay), which appeared highly organic in nature, but there was insufficient dry sample available to measure organic matter content as all the material not used for particle analysis was sent for carbon dating.

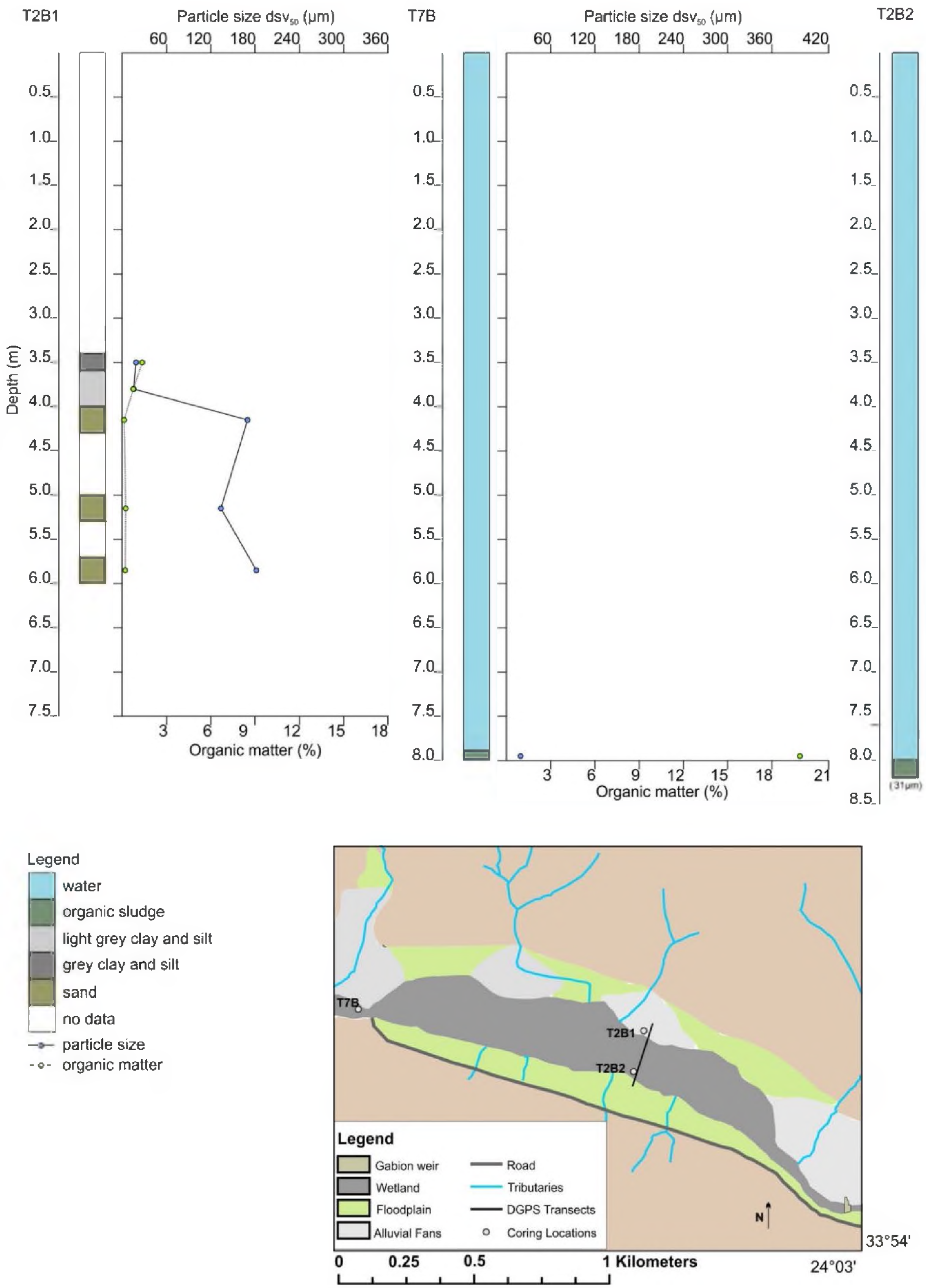


**Figure 17:** *Location of Bedrock Cores in Transect 2 and 7 with associated Core depth, particle size and organic matter content*

*5.5 Longitudinal profile of the Kompanjiesdrif basin*

The profile of the basin shows localised changes in slope associated with two of the larger alluvial fans (a & d) such that slope is lower upstream of the locus of deposition associated with the laterally impinging alluvial fans, and increases downstream of these alluvial fans. The fans that are related to changes in longitudinal slope are situated on the western and eastern ends of the Kompanjiesdrif basin (Figure 18). In the central region of the basin there is a large localised increase in slope that may have been formed from the cumulative deposition on the two smaller alluvial fans (b & c) encroaching on the wetland. Also noticeable on the profile are two troughs present between the larger alluvial fans (a & d) and the two smaller central alluvial fans (b & c). The points recorded as the base of cores, which is usually the depth to refusal, are closer to the surface near the two larger alluvial fans (a & d).

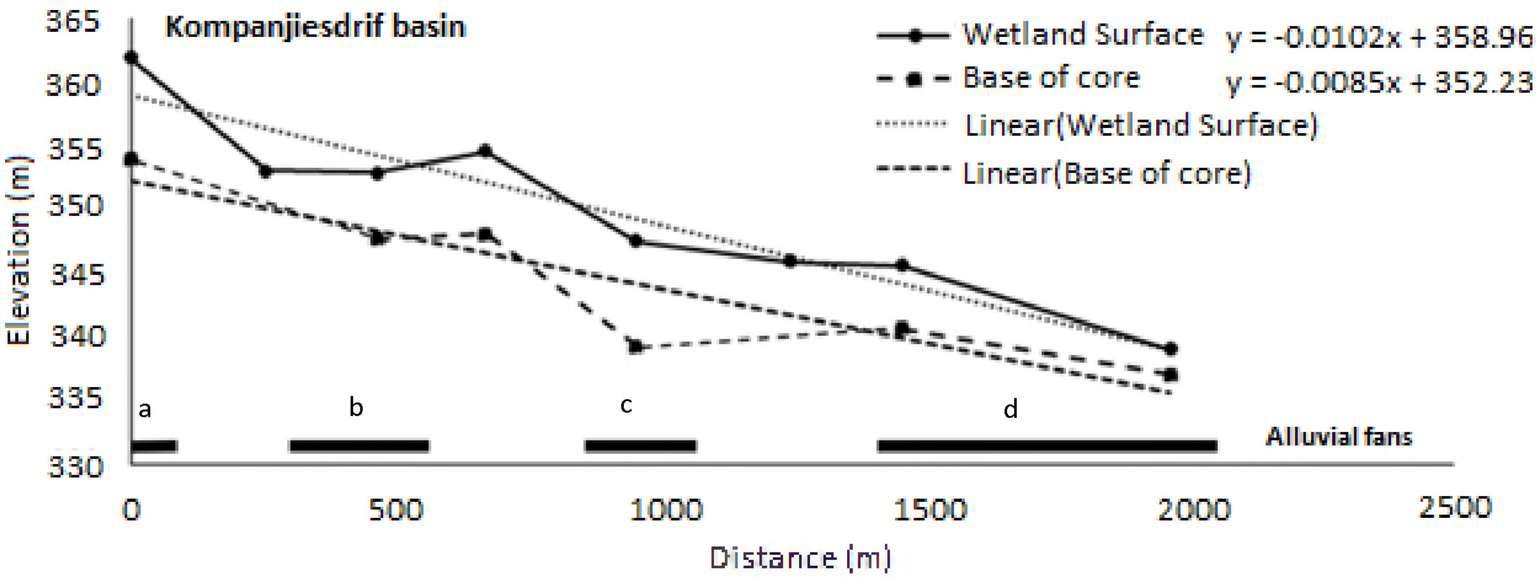


**Figure 18:** *Longitudinal profile of the Kompanjiesdrif basin*

*5.6 Carbon Dates*

Samples of sediment from the bottom of a bedrock core (Core T2B2), which had palmiet growing across a deep water column (Cores T2RPC, T4RPC, T6RPC and T7RPC), and one taken below the sand layer (Core T2C1; depth of 3.3 m) were dated (Figure 19). The trench-like feature in Transect 6 (T6RPC) was the most recently formed feature with a conventional radiocarbon age of 460 years BP (+/-30 y), while the feature in Transect 7 was dated at 1290 years BP (+/- 30 y). The core in Transect 2 where palmiet had grown over a deep column of water (T2RPC) was dated at 4940 years BP (+/-30 y), while the oldest trench-like feature was found in Transect 4 and dated at 7040 years BP (+/- 30 y; Figure 19).

The intervals between the potentially discrete erosional events that formed these features are also variable with the earliest interval spanning approximately 2100 years, the second interval spanned approximately 3650 years, while the most recent interval was 830 years. Furthermore, it is possible that there are some undiscovered deep narrow trench-like features below the wetland surface that occurred during the intervals mentioned above, which were not dated.

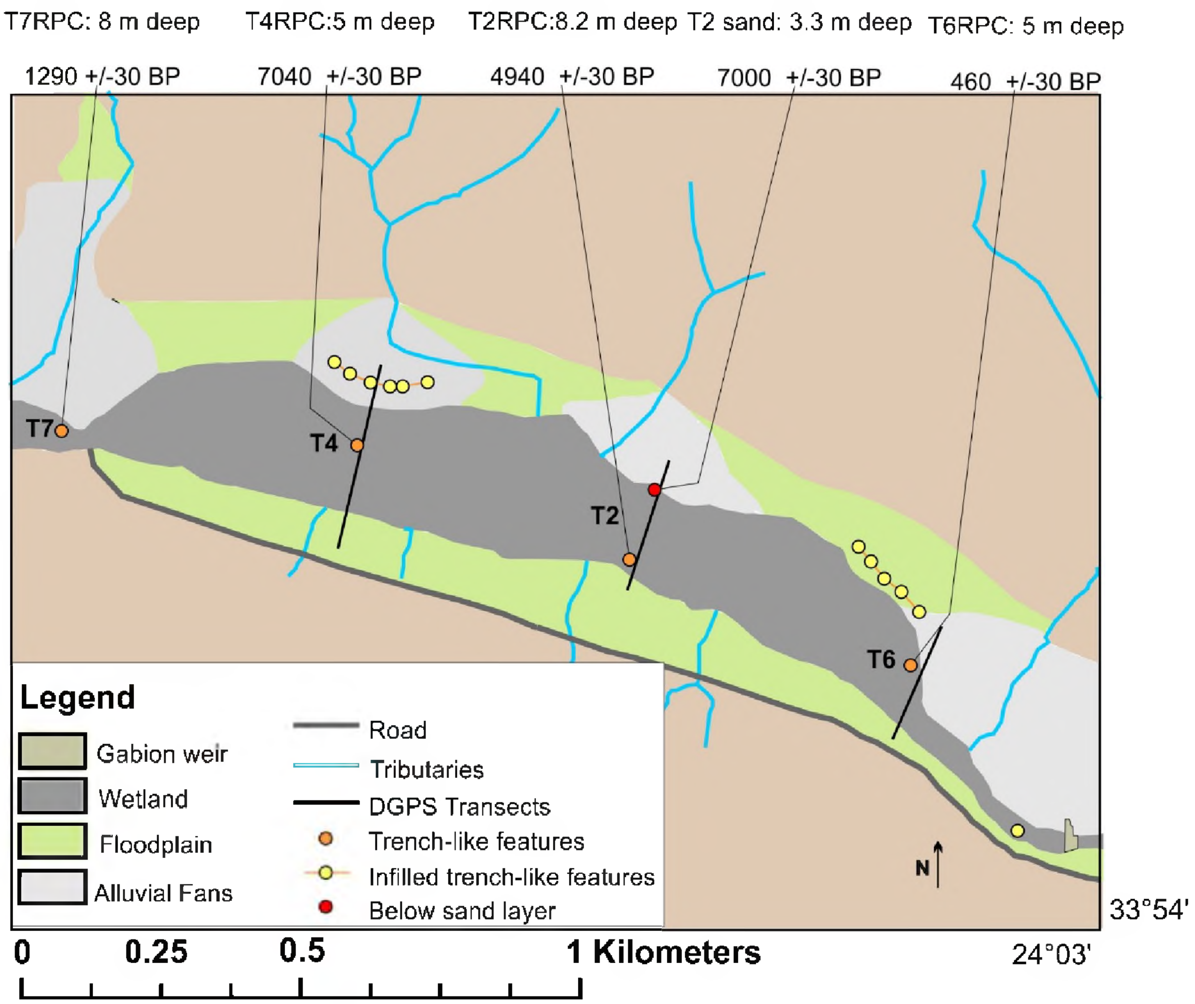


**Figure 19:** *Location of carbon dated sediment samples from the trench-like features and the sand layer, with their respective ages and depths, as well as the location of what appear to be additional previously infilled trench-like features*

# CHAPTER 6 – DISCUSSION

## *6.1 The geomorphic context of the Kromrivier*

Measurement of the longitudinal profiles of rivers and wetlands has been undertaken in the past based on the fundamental notion of a graded stream profile being a stable equilibrium condition (Goldrick & Bishop, 1995; Ellery et al., 2009). The shape of a classic graded stream profile is logarithmic in shape (Schumm, 1977; Goldrick & Bishop, 1995; Ellery et al., 2009) with the steepest region being in the headwaters and with slope gradually declining as stream discharge increases due to tributaries joining the trunk stream. Any deviation from this equilibrium condition is often attributed to one or more variables affecting the equilibrium state of a graded stream, including variation in the geology across which the stream flows, tectonics, climate and tributary influence, among others (Goldrick & Bishop, 1995; Brierly & Fryirs, 1999; MacGregor et al., 2000; Partridge & Maude, 2000; Crosby & Whipple, 2006; McCarthy et al., 2011). Several South African rivers have been found to display marked variations in longitudinal slope which can be attributed to one or more of these factors (King, 1963; Partridge & Maude, 2000; McCarthy & Rubidge, 2005). Similarly, the Kromrivier exhibits notable changes in slope which need to be examined within the South African context in order to assess whether it can be considered a graded profile i.e. at equilibrium (King, 1963; Partridge & Maude, 2000; Ellery et al., 2009).

The steep upper reach of the Kromrivier longitudinal profile (~14%) appears to be upon what remains of the African Erosion Surface which has weathered and eroded over tens of millions of years to produce a gently undulating surface with an appropriate gradient for prevailing climatic conditions (King, 1963; Partridge & Maud, 2000; Ellery et al., 2009). This led to lowering of the African Erosion Surface and the consequent lowering of the longitudinal slope of the Kromrivier (Ellery et al., 2009; Edwards et al., 2016) to form a graded stream profile. Uplift approximately 20 million years ago led to stream incision and valley widening to produce what is referred to as the Post-Africa I erosion surface (Partridge and Maud 2000, McCarthy and Rubidge 2005). A second uplift event occurred about 5 Ma, which also led to incision, in this case of the Post-Africa I erosion surface to form the Post-Africa II erosion surface (Partridge & Maud, 2000; Ellery et al., 2009). Most of the mid and lower reaches of the Kromrivier are found on the Post Africa II erosion surface.

The presence of the Kromrivier on the Post-Africa II erosion surface provides insight into the origin of this wetland as it narrows the range of possible relevant scenarios that might be invoked. In particular, the model developed concerning Dartmoor Vlei (Edwards et al. 2016), in which wetland formation has been attributed to valley-widening due to meander migration followed by sagging associated with deep weathering, may be discounted. Dartmoor Vlei is located on the African Erosion Surface where bedrock has undergone deep weathering (> 10 m), suggesting that this mode of wetland formation takes tens of millions of years. The Kompanjiesdrif basin lies on the Post Africa II erosion surface, which is relatively young. Moreover, although Dartmoor Vlei has accumulated a thin layer of peat (~ 1 m deep), the Kromrivier has accumulated peat to a considerable depth. Furthermore, the Kompanjiesdrif wetland basin is not positioned on a dolerite lithology and the geology of the area is not prone to deep weathering (King, 1963). Therefore, this model can be regarded as inapplicable to the Krom.

The Kromrivier has an abrupt step which separates the upper distinctly logarithmic longitudinal profile from a lower relatively uniformly sloping middle reach. This abrupt step in the profile is possibly a consequence of lithological controls as the underlying geology transitions from a more resistant to a less resistant lithology (Seidl et al., 1994; Tooth et al., 2004). The change in lithology may have caused a localised steepening of the longitudinal profile (MacGregor et al., 2000; Tooth et al., 2004). Furthermore, the confluence of the Kromrivier with two of its first major southern tributaries roughly coincides with this abrupt step in the profile and may have contributed to the marked reduction in slope downstream of the confluences (McCarthy et al., 2011). An alternative explanation is that this step represents an erosional knick-point between the Post Africa II erosion surface and the Post Africa I erosion surface (Partridge & Maud, 2000; Ellery et al., 2009).

### *6.2. Tributary-trunk interactions in the Kromrivier: the Kompanjiesdrif basin*

Two steeply sloping major northern tributaries feed into the lower gradient trunk valley of the Kromrivier on either side of the Kompanjiesdrif basin. Previous studies have shown that in some cases the rate of sedimentation within a tributary is greater than the rate of sedimentation within the trunk stream (Rice & Church, 2001; McCarthy et al., 2011; Pederson & Tressler, 2012). This appears to be the case in the Kromrivier, and consequently the incursion of tributary sediment into the trunk stream results in sediment deposition; to the extent that a tributary alluvial fan may form, which either encroaches on the trunk stream or blocks it completely (McCarthy et al., 2011; Joubert and Ellery, 2013). This process leads to temporary impoundment of the trunk, creation of accommodation space, long term inundation and enhanced sedimentation due to decreased stream flow velocity upstream of this impoundment. In the absence of clastic sediment input along the system, organic sedimentation is favoured and conditions would be more conducive for peat formation (McCarthy et al., 2011; Edwards et al., 2016).

The decrease in slope, upstream of the impoundment created by tributary sediment deposition, is also associated with a local steepening of the longitudinal profile elsewhere. This local steepening occurs downstream of the locus of deposition of the tributary fan impinging on the trunk stream (Hermon 2016). The local increase in gradient can be identified along the long profile of the Kompanjiesdrif basin study reach. The localised steepening of the longitudinal profile for a given discharge will increase the amount of sediment a stream is able to lift and transport (Ellery et al., 2009; McCarthy et al., 2011). Such localised steepening of the slope of the trunk stream would increase the likelihood of geomorphic threshold slope conditions being transgressed (Schumm, 1973; Schumm, 1979). Once transgressed, erosion and knickpoint formation would be initiated (Schumm, 1973; Schumm, 1979).

The discharge of the Kromrivier along the reach occupied by the Kompanjiesdrif basin, and hence its transport capacity at this point, need to be considered when examining the potential impact of tributary sediment input into the Krom. The cumulative catchment area, which is known to exhibit a strong correlation with discharge (Patton & Schumm, 1975), increases by 6.7 % along this reach. The southern and northern tributaries entering the trunk river along this reach contribute to its increased cumulative catchment area over a relatively short distance of ~ 2.5 km. This further supports the hypothesis that tributaries play an important role in the geomorphic dynamics of this basin. Moreover, the larger catchment sizes observed for the northern tributaries of the Krom signifies their greater influence on the dynamics of the Kromrivier, particularly in the Kompanjiesdrif basin where four northern tributaries enter the trunk river.

Although the model of McCarthy et al., (2011), in which tributary stream sedimentation on the trunk stream does seem to apply to the Kompanjiesdrif basin, it fails to explain the formation of the wide relatively planar valley observed at the study site. A flat near-planar valley floor (in cross-section) is often accompanied by the presence of alluvial meanders and floodplain wetlands set in less resistant lithology, where lateral planing of the valley floor is achieved by meander migration at a lower relative stream power (Tooth et al., 2002; Tooth et al., 2004). However, in the case of the Kompanjiesdrif basin the distinctive geomorphic features that would indicate a floodplain wetland are absent, including a meandering channel, cutoff lakes, and an alluvial ridge (Tooth et al., 2002; Tooth et al., 2004; Joubert & Ellery, 2013; Job, 2014). Therefore, the geomorphic mechanism by which this valley has evolved into its current relatively wideplanar form remains uncertain.Based on empirical observations made in the field as well as the data collected; the presence of tributaries and their alluvial fans is the starting point from which an explanation to this conundrum may be derived.

The first indicator which suggested a strong influence by tributaries was detected in the sedimentary fill of the wetland basin. As can be seen from a number of the cores, there are several coarse sediment deposits which interrupt finer, more organic wetland deposits in the stratigraphy. One of these coarser layers, which was predominantly composed of fine and medium sand, was observed to be present across the entire wetland; signifying that a large flood event took place that deposited sediment across the entire Kompanjiesdrif basin. This coarse sediment layer, which was found consistently across the entire wetland, will be referred to as “the sand layer”. The sand layer was often associated with a silt and clay layer directly above it, which was potentially deposited in the same flood event. In order to estimate the age of the sand layer, a sample from directly below it (T2 C1 300-330 cm) was dated at 7000 +/- 30 BP; indicating that the flooding event occurred thousands of years ago. Additional coarse sediment deposits (some with upward fining sequences) were found above and below the more prevalent “sand layer” flood event in some of the cores (T1C1, T1C3, T1C6, T6C7, T2B1, Core B1, Core B2, Core B3, T4C1 and T6C10, T6C12, T6C14 in appendix 1); suggesting additional, less extensive and in some cases, more recent flood events have taken place.

### *6.3 Valley widening and longitudinal slope reduction*

Despite making considerable effort to sample uniformly in the Kompanjiesdrif basin of the Kromrivier wetland to determine wetland morphology and stratigraphy, trench-like features were only found opposite alluvial fans. Based on their form they are likely to have been gullies that formed in the wetland in the distant past. Moreover based on the evidence obtained in this study a tentative model could be conceived about the role of alluvial fans in the Kromrivier. The fans feeding into the valley create a localised steepening of the longitudinal slope of the trunk stream; which occurs downstream of the node of deposition of these features. This is due to the gradual sediment deposition that takes place as the tributaries escape the narrow confinement of the adjacent mountain front (Lecce, 1990; McCarthy et al., 2011). As slope increases over time the trunk stream approaches a geomorphic threshold slope; above which local incision will take place (Schumm, 1973; Patton & Schumm, 1975; Patton & Schumm, 1981). Once a tributary's alluvial fan builds sufficiently across the trunk stream valley, the trunk stream may approach this threshold; such that flooding events could trigger initiation of gullying along the trunk stream (Schumm, 1973; Patton & Schumm, 1975; Patton & Schumm, 1981).

Therefore, the four trench-like features found below dense floating mats of palmiet are thought to be former gullies that have been drowned. Two of the four gully features have reached bedrock and, combined with two of the floodplain bedrock cores, suggest a uniform elevation for bedrock beneath the wetland basin. Accordingly, the gullies that form as a consequence of this process may plane bedrock and lower longitudinal slope. The morphology of these gullies seems to be consistent with gullies found in other semi-arid regions such as Piceance Creek, North Western Colorado; which have been shown to have a gradient less than the original valley floor, associated with lowering of longitudinal slope (Patton & Schumm, 1975). Moreover, as with some other deep and narrow gullies found in other environments, the sedimentary fill which has been eroded is primarily silt and clay (Patton & Schumm, 1981). It is suggested that due to the cohesive nature of the silt and clay that has been deposited, erosion leads to the deep and narrow morphology of gullies observed (Patton & Schumm, 1981).

The differing depths and locations of the gullies raised the possibility that they were formed during separate erosional phases and are gradually being filled with sediment over time (Patton & Schumm, 1981; Grenfell et al., 2012). This possibility was confirmed by dating post erosional sediment at the base of each gully with conventional radiocarbon ages and calibrated radiocarbon ages for Transect 6, Transect 7, Transect 2 and Transect 4 of 460 +/-30 BP (1455 cal AD), 1290 +/-30 BP (770 cal AD), 4940 +/-30 BP (3655 cal BC) and 7040+/-30 BP (5890 cal BC) respectively.

The presence of erosional gullies dating back over 7000 years BP suggests gully formation occurred prior to European settlement in the area which took place between 1775 and 1787 AD (Rebelo, 2012). Other studies have also found evidence of gully formation prior to European settlement (Botha et al., 1994; Dardis, 1989; Fryirs & Brierly, 1998; Avni et al., 2010). The confirmation that these gullies formed at different ages supports the concept of erosional or cutting phases, which led to the lateral planing of bedrock and valley widening; but also to gradual relative longitudinal slope reduction, given that the slope of the bed of a gully is lower than that of the original land surface (Patton & Schumm, 1975).

### *6.4 Palmiet and the deposition phase*

Although the postulated role of the alluvial fans provides an explanation for the existence of these former gullies; gully erosion as envisaged here does not completely account for how, after incision has occurred, the system reverts back to a peat wetland. Palmiet, which is regarded as an ecosystem engineer in this system, may play a role by facilitating the transition from an eroded, bare channel reach into a peat wetland (Sieben, 2012; Barclay, 2016). It is proposed that after a period of erosion, palmiet is capable of colonising the bottom of, or growing across the top of these gullies (Figure 20). This proposition is strengthened by previous observations regarding palmiet's colonising capabilities, as it has been found to stabilise levees within several different river channels (Boucher & Withers, 2004; Sieben, 2012; Job, 2014).

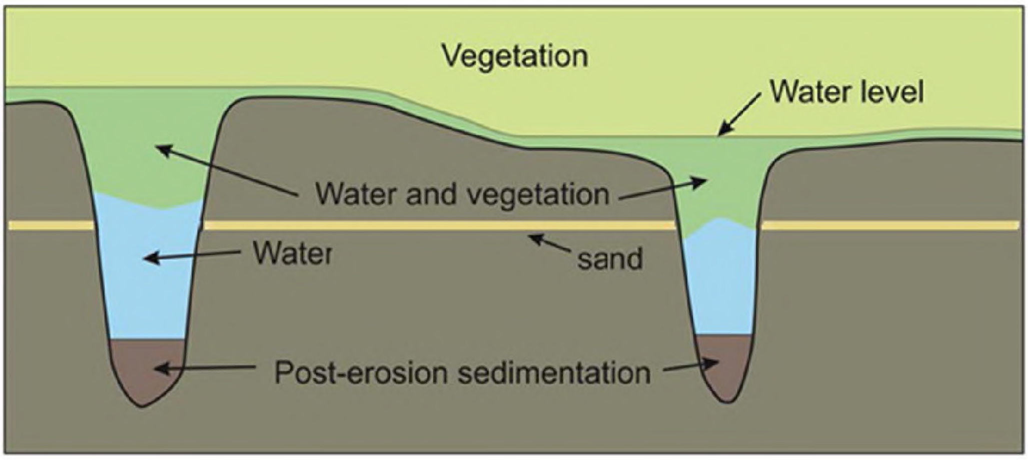


**Figure 20:** *Schematic diagram of gullies below the wetland surface*

Palmiet colonises the bank as well as any small sedimentary bars (Figure 21a) and grows in an intertwining manner, eventually blocking streams and channel/s; thus causing diffuse flow of water (Figure 21b; Sieben, 2012; Barclay, 2016). Combined with its excellent sediment trapping capabilities in the web-like leaf bases that surround the stem (Figure 21c), colonisation by palmiet leads to the gradual infilling of these gullies with organic matter and sediment that has been trapped in their root and trunk systems (Barclay, 2016). While simultaneously transforming the environment from an eroded river reach into a peat wetland system (Figure 21d; Barclay, 2016).

a) b) c) d)

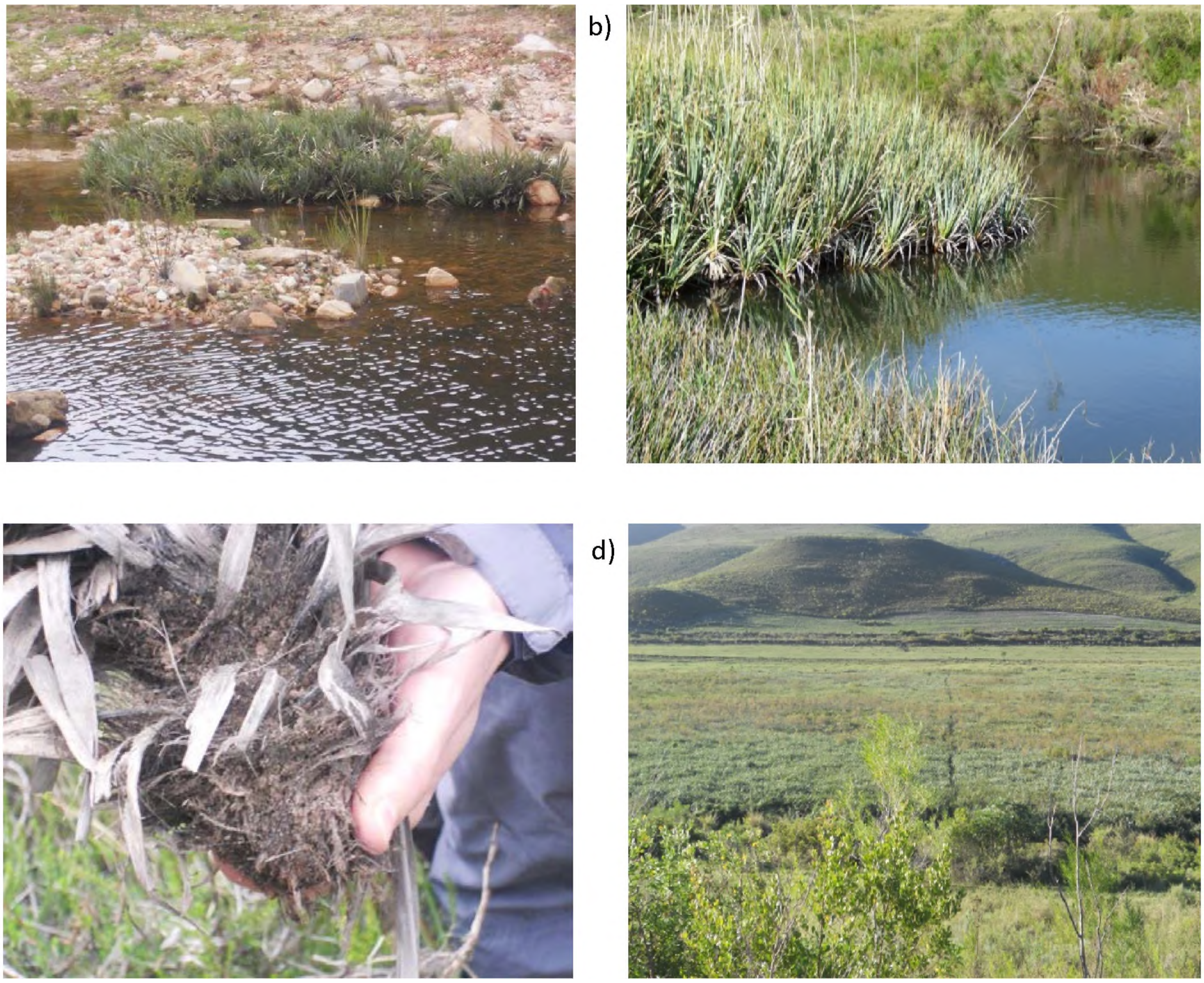

**Figure 21: a)** *An eroding reach downstream of the Kompanjiesdrif basin with palmiet starting to colonise banks and deposited sedimentary bars* **b)** *palmiet growing across a channel in an intertwining manner* **c)** *the highly effective sediment trapping stem and root system of palmiet,* **d)** *palmiet completely blocking the trunk stream in the Kompanjiesdrif basin creating diffuse flow across the valley*

This transition from an eroded river reach to a valley-bottom wetland system may also be facilitated or accelerated by the deposition of sediment from tributary alluvial fans; which may aid in filling eroded gullies, as evidenced by the presence of what appear to be infilled gullies on the northern floodplain alluvial fans (Figure 22a). Northern tributary alluvial fans may completely or partially impound the trunk stream, which allows for the gradual elevation of a local base level of the trunk stream to occur, creating accommodation space on the trunk stream, where the accumulation of organic matter is promoted (Grenfell et al., 2010, McCarthy et al., 2011). Taking into consideration that many wetlands tend to be influenced by multiple controls including geomorphic, tectonic and biological factors, it is possible that both palmiet and the northern alluvial fans contribute to the accumulation of organic material in the Kompanjiesdrif basin (McCarthy et al., 1998; McCarthy et al., 2002; Job, 2014).

a) b)

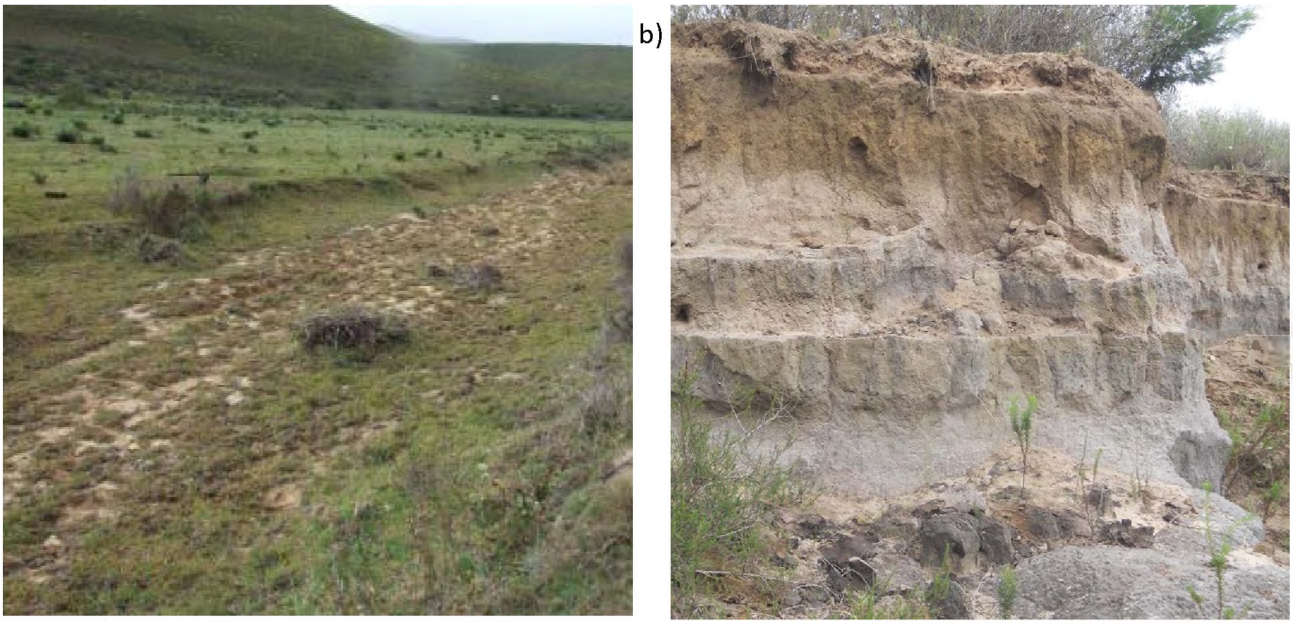

**Figure 22**: **a)** *What appears to be an infilled gully on one of the northern tributaries' alluvial fans situated on the margin of the wetland (wetland positioned to the right of the photo)* **b)** *A channel sidewall/ gully side wall downstream of the wetland basin showing cut-and-fill stratigraphy*

*6.5 Cycles of cutting and filling*

It is evident from the above hypotheses that the Kompanjiesdrif basin may undergo prolonged periods of sediment filling through deposition, which can be attested to by the thick layers of clastic sediment throughout the basin (Figure 22b). These periods of filling are interrupted by briefer periods of incision, which is supported by the presence of trench-like features throughout the Kompanjiesdrif basin. Periodic cycles of cutting and filling have been observed in other dryland river systems (Patton & Schumm, 1981; Bull, 1997; Grenfell et al., 2012) as well as, in swamps, floodplains, dambos and various wetland features in the Australian and African landscapes (Boast, 1990; Brierly & Fryirs, 1999; Grenfell et al., 2009a). However, recognition of the erosional phase in these systems as a precursor to wetland formation in the southern African landscape has been limited with the exception of Grenfell et al. (2009a), Job (2014) and Silbernagl, (2014). As an attempt at addressing this oversight a conceptual model outlining the theorised geomorphic dynamics of the Kompanjiesdrif basin has been developed (Figure 23).

The first phase of this model portrays a valley with a trunk stream located on the valley floor flanked by steep valley sides (cross section; Figure 23a) and a logarithmic longitudinal profile (longitudinal section; Figure 23a). A tributary stream at its confluence with the trunk stream has deposited sediment in the form of a tributary alluvial fan, thus confining flow along the left side of the valley (cross section; Figure 23a). The longitudinal section illustrates the reduction in slope upstream of the alluvial fan and increase in slope downstream of the alluvial fan (longitudinal section; Figure 23a). Sediment deposition on the alluvial fan leads to a localised increase in the longitudinal slope of the trunk stream, downstream of the area of deposition, while upstream of the laterally impinging tributary fan, slope along the trunk stream is reduced. The reduction in slope upstream leads to a loss in relative stream power and sufficiently low flows along with shallow inundation can potentially promote wetland formation. In contrast, downstream, the gradual localised increase in longitudinal slope for the size of the trunk stream catchment increases the likelihood that a geomorphic threshold slope will be crossed. The concentration of trunk stream flows due to the encroachment of the tributary alluvial fan increases the already unstable state of the landform.

Once this geomorphic threshold is transgressed, the trunk stream is more vulnerable to erosion. In the second phase of this model, gully erosion is initiated. The slope on the bed of the eroding gully is lower than the original pre-erosional land surface, i.e. erosion along the trunk leads to lowering of longitudinal slope (longitudinal section; Figure 23b). Furthermore, a section of the trunk stream channel has been widened by this gully erosion (cross-section; Figure 23b).

The third phase of this conceptual model describes the sediment eroded at the head of the gully, being deposited at the toe of the gully, due to loss of confinement of flow as it emerges from the gully (Figure 23b & 23c). These deposits cover much of the valley floor as indicated by the extensive sand layer encountered across the length and width of the Kompanjiesdrif basin, which provides a suitable environment for the growth of palmiet and leads to the resumption of a filling phase (Figure 23c). The morphology of the feature which forms is such that an elevated mound is formed in the vicinity of the former gully, and depressions occur along both valley margins such that the next cycle of erosion is most likely on the edge of the wetland (Figure 23d).

After a cycle of cutting and filling, tributary stream sediment accumulation on the trunk valley floor, via alluvial fan sedimentation where a tributary impinges the trunk stream, eventually leads to a localised increase in the longitudinal slope of the trunk stream and phase 1 of the cycle through to phase 3 (Figure 23 a-d) of the cycle are repeated. During each re-iteration of this periodic cut-and-fill cycle, a different portion of the trunk valley is widened and slope lowered, such that over time scales of thousands to tens-of-thousands of years, a wide valley with a planar floor is formed, with a very low longitudinal slope resulting in lower relative stream power along this reach. These conditions create an environment suited to wetland formation (Figure 23e). Thus, this model provides an alternative mechanism for valley widening and longitudinal slope reduction in the Kompanjiesdrif basin, to the mechanism outlined by Tooth et al. (2002) and (2004).

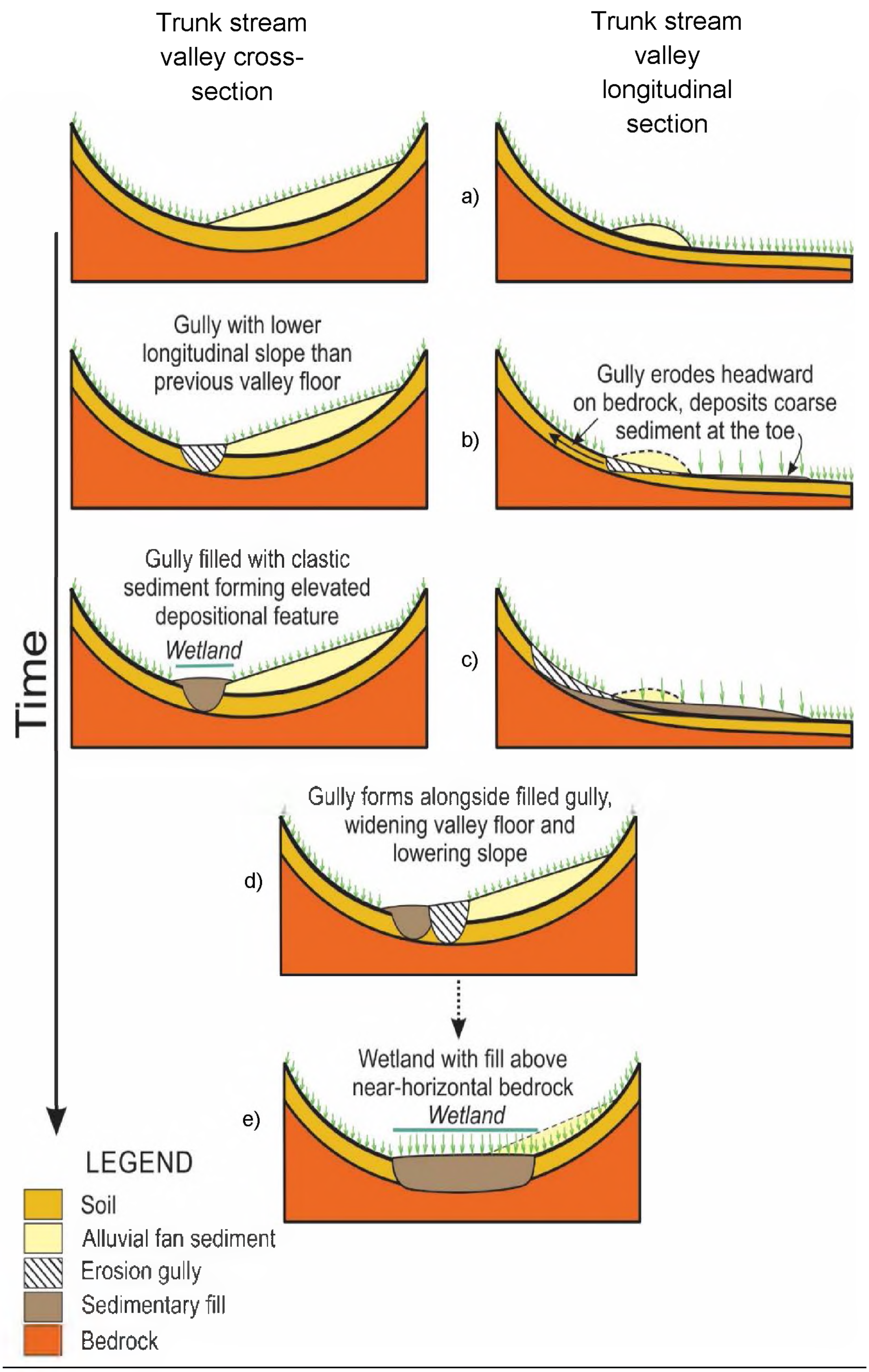


**Figure 23:** *A simplified diagram of the Kompanjiesdrif basin mechanism of wetland formation by gully erosion*

Cycles of cutting followed by filling, may take place in the Kompanjiesdrif basin over 100 to 1000 year timescales. The estimation of these time intervals is based on the gaps between the dated ages of gullies' in this study. Processes of cut-and-fill in Kanab Creek in northern Colorado appear to have cycles that take place over similar time scales (Nelson & Rittenour, 2014). Nevertheless, there are a number of marked differences between Kanab Creek and the Kromrivier, which, along with other examples, illustrates the variability between different cut-and-fill systems (Patton & Schumm, 1981; Nelson & Rittenour, 2014; Burrough et al., 2015; Bekkadour et al., 2016).

For instance, the rate at which deposition takes place varies with the environmental context and factors such as sediment yield. This point is illustrated when the Kromrivier's dynamic is compared with that of a cut-and-fill system in Piceance Creek, North Western Colorado. In Piceance Creek; a greater sediment yield may contribute to increasing the likelihood and frequency of transgressing intrinsic geomorphic threshold slopes, and, as a result, cutting phases may take place over shorter time scales relative to cutting phases in the Kromrivier (Patton & Schumm, 1975). Although intrinsic geomorphic threshold slopes may act as the primary control on some cut-and-fill systems as demonstrated by the case of Piceance Creek (Patton & Schumm, 1975), it may not be the only control on cut-and-fill systems like the Kromrivier. A number of other cut-and-fill systems, (Fryirs and Brierly, 1998; Burrough et al., 2015; Bekkadour et al., 2016) as well as, the cyclic nature of stratigraphy observed by McCarthy et al. (2011) in their trunk-by-tributary impoundment model, have been linked to changes in climatic regimes and thus there is a need to explore this possibility in the Krom.

Adding to the difficulty presented by interpreting cut-and-fill cycles in the context of climate variability is the number of discontinuities and discrepancies that have been observed in other systems (Patton & Schumm, 1981; Fryirs & Brierly, 1998; Brierly & Fryirs, 1999; Bekkadour et al., 2016). For example, as mentioned previously, during cutting phases only select sections of a valley are eroded. Consequently, it has been notoriously challenging to find agreement between different terraces or sedimentary facies that act as historical records of cut-and-fill cycles within one river system or between rivers in the same region, given the fragmentary nature of these records (Patton & Schumm, 1981; Fryirs & Brierly, 1998; Brierly & Fryirs, 1999; Bekkadour et al., 2016). Evidence of this limitation being applicable in the Krom is illustrated by the multiple flood events recorded in the stratigraphy near Transect 6, as opposed to one flooding event being recorded in Transect 5. This limitation needs to be acknowledged, when examining any potential correlations between radiocarbon ages of post-erosional sediment deposition and the local and regional paleo-climatic record.

### *6.6 Dry and humid climatic phases*

According to the work done on Featherstone Kloof wetland, near Grahamstown in the Eastern Cape of South Africa, wet and dry climatic cycles may influence whether cutting and filling is predominantly influenced by extrinsic climate controls (Silbernagl, 2014). It is hypothesised that cooler and more arid periods, which occurred during glacial maxima, were dominated by limited vegetation cover on hillslopes, which in turn increased hillslope erosion and sediment supply from tributary streams; leading to greater rates of sediment deposition on the valley floor in the form of tributary alluvial fans impinging on trunk streams (McCarthy et al., 2011; Silbernagl, 2014). Portions of the valley which are relatively narrow are thought to be the most affected by the incursion of tributary alluvial fans and are more likely to be impounded (McCarthy et al., 2011; Silbernagl, 2014). Impoundment was thought to lead to floodplain environments in the trunk stream, upstream of the impoundment (McCarthy et al., 2011). During subsequent wetter climates the vegetation cover on surrounding hillslopes is thought to increase, and the rate of erosion by tributary streams and concomitant sediment supply to the trunk stream would be diminished (McCarthy et al., 2011; Silbernagl, 2014). Reduced tributary sediment input in association with greater discharge, would likely lead to greater capacity in the trunk stream, increasing the chances of erosion (McCarthy et al., 2011; Silbernagl, 2014).

As with the Featherstone Kloof system the Kromrivier wetland and particularly the Kompanjiesdrif basin have exhibited similarities in that they both possess cut-and-fill stratigraphy and are situated in long narrow steep-sided valleys (Silbernagl, 2014). In addition to this similarity, the Krom and Featherstone both fall within the year-round rainfall zone (YRZ) which lies between the extensive north-eastern summer rainfall zone (SRZ) and the narrower south western winter rainfall zone (WRZ; Bateman et al., 2004; Chase & Meadows, 2007; Silbernagl, 2014). As has been quoted in many an article on paleoclimate, there is a paucity of information on climate dynamics within southern Africa (Carr et al., 2006; Reinwarth et al., 2013; Chase et al., 2015; Wündsch et al., 2016). This is particularly true of the year-round rainfall zone, which has just a few paleoclimate sites that have been studied, which include the Cango Caves (Talma & Vogel 1992), Boomplaas Cave adjacent to the Cango Caves (Deacon, 1979), Groenvlei (Irving & Meadows 1956; Wündsch et al., 2016), Vankersvelsvlei (Quick et al., 2015), Nelson Bay Cave (Deacon & Lancaster, 1988; Partridge & Maude, 2000) and Eilandvlei (Reinwarth et al., 2013).

Moreover, aside from limited paleo-climatic records, an additional limitation is the potential margin of error associated with ages derived from carbon dating, which can be attributed to a number of factors (Talma & Vogel, 1993). Thus, linking the carbon dated ages obtained from the Krom to climate, should be considered with a degree of caution. Nevertheless, when these ages are examined in conjunction with existing paleoclimate records in southern Africa there are tentative teleconnections that may be drawn.

The age of the post erosional sediment from the gully in Transect 6 has a conventional radiocarbon age of 460 +/-30 BP and calibrated ages of 495 cal BP and 1455 cal AD. This has been cited in the literature as a time-period, either during or very close to the beginning of the Little Ice Age in the paleo-climatic records; i.e. colder, drier conditions noted at Groenvlei (610 – 149 BP, YRZ) and Makapansgat (1500 – 1800 cal AD, SRZ; Figure 24). The sediment dated was potentially deposited during a filling phase (drier, colder conditions) after an erosional phase (warmer, more humid conditions), which indicates that the age of this sediment supports the proposed hypothesis that cutting and filling phases may be linked to periodic climatic transitions.

The sediment from the gully in Transect 7 has a conventional radiocarbon age of 1290 +/-30 BP with calibrated ages of 1180 cal BP and 770 cal AD. Climate records from Makapansgat (1200 – 600 cal BP, SRZ) and Groenvlei (1210 – 700 cal BP, YRZ) indicate that this age falls within a more humid, warmer climatic stage whereas data from Chase et al., (2015) at Katbakkies Pass in the Cape Fold Mountains suggest a more humid climate slightly earlier than this age (2700 – 1700 cal BP, WRZ; Figure 24) followed by a period of high climatic variability which fluctuated between drier and more humid conditions from 1700 cal BP onwards. This radiocarbon age potentially suggests that gully formation took place as the environment transitioned from a drier cooler climate (filling phase) prior to 2700 BP towards the more humid climate (erosional phase) between 1200 – 600 BP.

The sediment from the former gully in Transect 2 has a conventional radiocarbon age of 4940 +/-30 BP, with calibrated ages of 5605 cal BP and 3655 cal BC. This age is coincident with the end of a more humid phase at Katbakkies pass (~6900 – 5600 BP, WRZ) and occurs at the beginning of a phase of gradual cooling at Makapansgat caves (6000 – 3000 BP, SRZ; Figure 24). At Vankervelsvlei there was a limited pollen record with some pollen samples present for ~6900-5000 BP (YRZ); which suggests a warmer climate, along with an increased aridity inferred from the increased amount of charcoal present, which would lead to increased incidences of fires. All the data suggests that there may have been a transition from a more humid, warmer phase towards a cooler, drier phase and possibly that this transition period may have been characterised by warmer, yet, arid conditions as described at Vankervelsvlei (Quick et al., 2015). Thus, erosion may have taken place during the more humid, wetter climate before 5600 BP (Katbakkies pass, WRZ) and thereafter more arid conditions were observed (Makapansgat, Vankervelsvlei), which are associated with a filling phase.

The sediment from the gully in Transect 4 has a conventional radiocarbon age of 7040+/-30 BP and calibrated ages of 7840 cal BP and 5890 cal BC. This age falls between two drier, cool periods in the paleoclimate record at Makapansgat Cave (Holmgren et al., 2003; SRZ) i.e. 9000 – 8400 BP and 6000 – 3000 BP. The hyrax midden records from Katbakkies pass (Chase et al., 2015; WRZ) cease at ~ 7000 BP; with increased humidity recorded ~6900 yr BP as the oldest trend observed from this record (Figure 24). It is possible that the phase separating the two drier, cool climatic phases at Makapansgat were interrupted by a humid and warmer climate (cutting phase) between 8400 and 6000 yr BP which coincides with the radiocarbon age of the gully. However, this is highly speculative and more robust data pointing to a humid phase would need to be acquired in order to confirm whether a correlation exists.

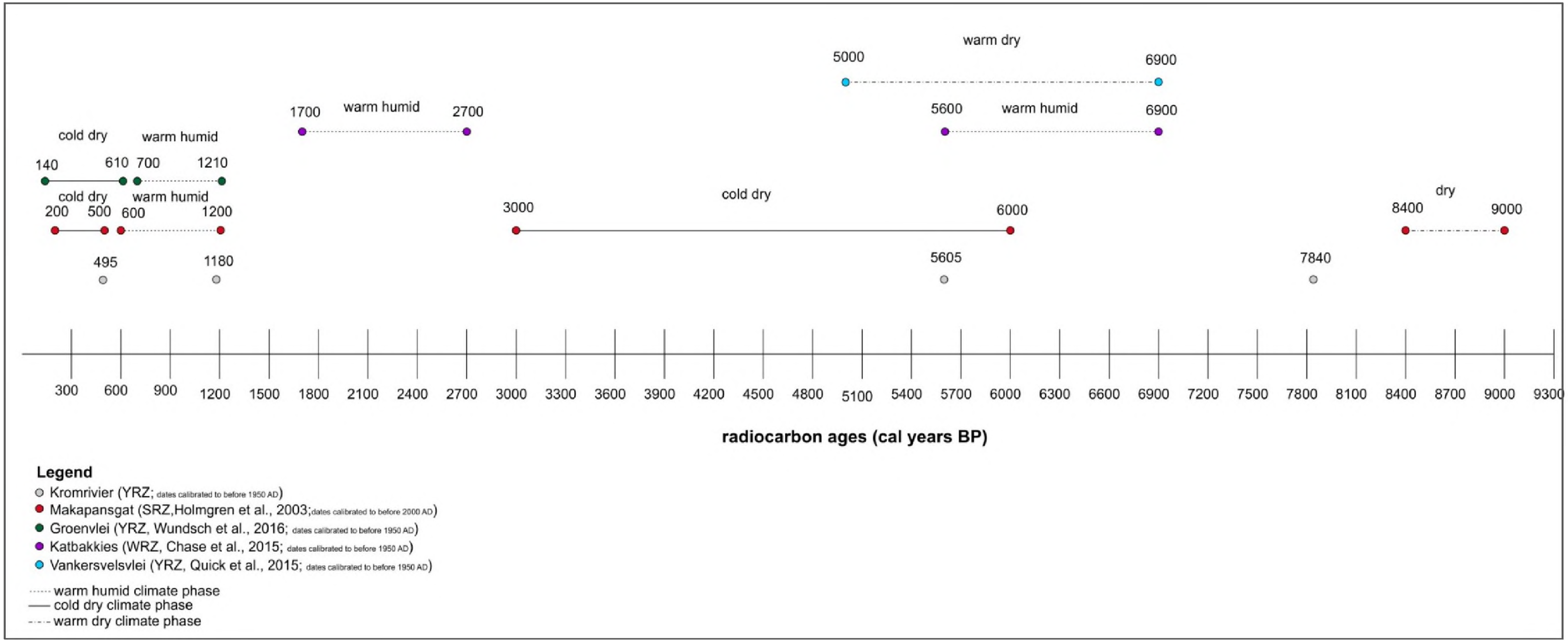


**Figure 24:** *A timeline with the postulated and dated climatic dry or humid climatic phases from four studies (Modified from Holmgren et al., 2003:2313-2323; Chase et al., 2015:141-147; Quick et al., 2015:11; Wündsch et al., 2016:298-306); as well as the four carbon dates obtained from the bottom of the drowned gullies in the Kompanjiesdrif basin of the Kromrivier*

*6.7 Implications of cut-and-fill cycles in the southern African context*

Whether cut-and-fill cycles in the Kompanjiesdrif basin are stochastic or regulated by climatic dry and humid phases requires more research and a greater agreement between paleo-climatic data and the effects which were experienced in the year-round rainfall zone. However, it does appear that these cut-and-fill processes are natural within the Kompanjiesdrif basin. More data and sedimentary sequences would need to be examined to determine if this process is applicable to the entire Kromrivier. Further consideration is also required for the possibility that this process acts as a mechanism of valley widening and relative slope reduction in other South African rivers.

Despite the confirmed recurrence of natural gully formation in the Kompanjiesdrif basin, it is important to acknowledge that as in other cut-and-fill systems, anthropogenic influences can have a profound impact on landscape-level processes (Prosser & Slade, 1994). Additionally, once these systems are disrupted it is thought that it is very difficult, if not impossible, to revert these systems back to their pre-disturbance state (Brierly & Fryirs, 1999; McCarthy et al., 2010). Although the anthropogenic factors that affect gully erosion in the Krom have not been the focus of this study, it would be remiss not to highlight some of the most notable human impacts which have potentially increased the severity of what was an originally natural process of gully erosion (Rebelo, 2012). As in the study conducted by Brierly & Fryirs (1999), the land use in the Kromrivier after European settlement has been primarily for cultivation, including fruit orchards, timber plantations and dairy farming, which has been well documented by Rebelo (2012). As a result, wetlands have been drained to create pastures, natural vegetation has been cleared and extensive transport infrastructure such as railways and roads have been built in order to facilitate transport of goods and increase accessibility to and from the Langkloof (Rebelo, 2012). Clearing of natural vegetation for other land use as well as increased runoff caused by the presence of roads has been linked to head-cut erosion along some reaches of the Krom (Rebelo, 2012). More research would be needed to investigate the degree to which human impacts have affected cut-and-fill processes in the Kromrivier (Brierly & Fryirs, 1999). Thereafter, once the extent of their impact has been assessed, management would be better equipped to determine whether the system is beyond repair through costly restoration initiatives (Brierly & Fryirs, 1999; McCarthy et al., 2010). Such assessments would provide a clearer understanding of whether the measures put in place to halt gully erosion and headward propagation of erosional knickpoints, have been successful or not (Hermon, 2016). However, given our current understanding, the long term persistence of these structures and their effectiveness in addressing the issue of erosion are questionable (Hermon, 2016).

# CHAPTER 7 – CONCLUSION

The majority of wetlands in South Africa that have been studied have developed under the climatic and tectonic influence of the Pliocene. The Kompanjiesdrif basin along the Kromrivier is no exception and appears to have evolved on the Post Africa II erosion surface. Therefore, the warmer more humid phases of climate experienced prior to 2 Ma, as well as notable interglacial periods in the more recent past, may have played a critical role in facilitating erosion of the valley which hosts this wetland basin. The more dominant process of erosion present in this valley was the process of gully erosion, as opposed to the lateral erosion produced by meander migration of a sinuous channel in the model of Tooth et al. (2002, 2004). The quartzitic sandstone lithology and their reputation as groundwater aquifers, as well as, the steep and rugged topography of the surrounding mountains is thought to play a major role in maintaining a water supply to this valley bottom wetland and is a topic which warrants further research.

This study proposes that repeated cycles of cutting has led to valley widening and relative longitudinal slope reduction over geological time scales. The filling or long term depositional phases, which follow the more rapid incision episodes in the basin, appear to be controlled by tributary alluvial fans and the remarkable biological capabilities of palmiet. Both these controls promote the accumulation of organic material by creating accommodation space behind a rising local base level, thereby promoting diffuse flow. The role of alluvial fans and their associated geomorphic threshold slopes that trigger each subsequent phase of incision could potentially be explored further by comparing dated stratigraphic fill from both a tributary and the dated samples from the Kromrivier.

The model of cutting and filling presented in this study combined information gained from a thorough appraisal of the existing topography, sedimentary fill and the elevation of bedrock of the Kompanjiesdrif basin, as well as, the dating of sediment at the base of the drowned gullies. Given that the initial aim of this study was to develop a conceptual model of processes giving rise to this basin, the model developed appears to be well supported. It can be concluded that gully erosion in the context of the Kompanjiesdrif basin formed part of the natural process of valley widening and relative longitudinal slope reduction that favours wetland formation. Having said that, it is likely that the wetland is naturally vulnerable to the negative influence of human drivers which have become manifest in more downstream reaches of the Kromrivier.

While cut-and-fill cycles have been identified as contributing in important ways to landscape evolution in other semi-arid environments, this is only the fourth time this process has been linked to the formation and dynamics of a valley bottom wetland in South Africa and the first time such an in depth and targeted investigation on a cut-and-fill wetland system has been achieved. This conceptual model consequently has implications for any valley bottom wetlands that occur in similar settings, as well as for the rehabilitation of these systems.

The relevance of cut-and-fill cycles as a process that exerts a control on the origin and dynamics of the Kompanjiesdrif basin has been demonstrated through this study. Nevertheless, one notable limitation that has been observed, is the narrow and specific scope of investigation undertaken in this study. Therefore, more research is needed to investigate whether this mechanism of valley widening and relative slope reduction is applicable to the remainder of the Kromrivier wetlands, or to other valley bottom wetlands which occur under similar tectonic and geological conditions.

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# APPENDIX 1

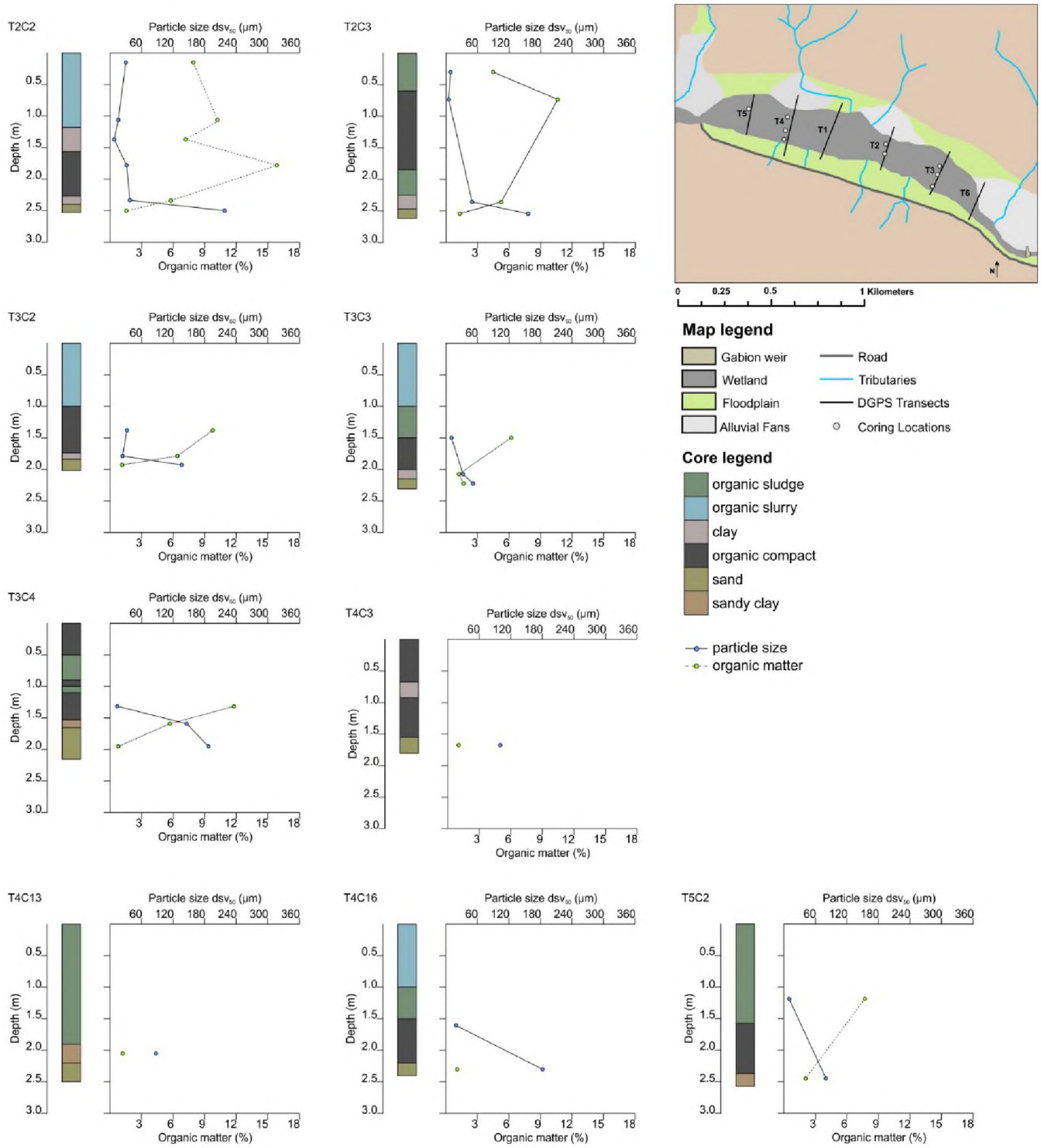


**Figure 25:** *Location and stratigraphy of cores in Transects 2, 3, 4 and 5 showing variation in particle size and organic matter content with depth*

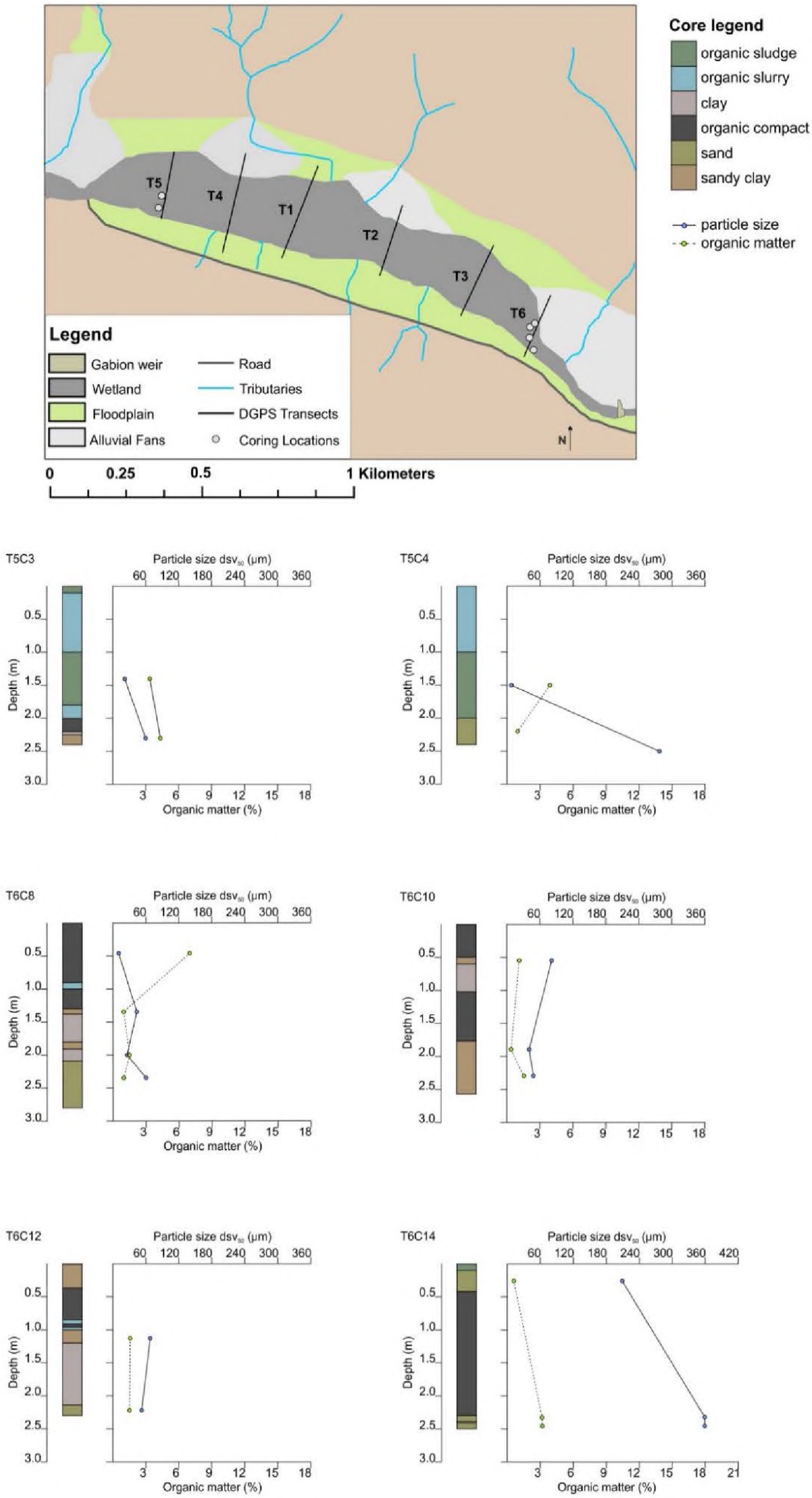


**Figure 26:** *Location and stratigraphy of cores in Transects 5 and 6 showing variation in particle size and organic matter content with depth*

**Table 2:** *Sediment samples and their associated conventional and calibrate ages*

| Beta sample no | Submitter Sample no | Conventional age (BP) | Intercept of radiocarbon age with calibration curve (Cal AD/BC) | Intercept of radiocarbon age with calibration curve (Cal BP) |
|---|---|---|---|---|
| 426639 | T7 RPC BOTTOM | 1290 +/- 30 BP | (Cal AD 770) | (Cal BP 1180) |
| 426638 | T6C4 RPC2.3 | 460 +/- 30 BP | (Cal AD 1455) | (Cal BP 495) |
| 426637 | T4 RPC47-52 | 7040 +/- 30 BP | (Cal BC 5890) | (Cal BP 7840) |
| 426636 | T4 RPC0-20 | 5980 +/- 30 BP | (Cal BC 4795) | (Cal BP 6745) |
| 426635 | T2 300-330b.s | 7000 +/- 30 BP | (Cal BC 5840) | (Cal BP 7790) |
| 426634 | T2 8.2m2 | 4700 +/- 30 BP | (Cal BC 3495) (Cal BC 3465) (Cal BC 3375) | (Cal BP 5445) (Cal BP 5415) (Cal BP 5325) |
| 426633 | T2 RPC1.3 | 4770 +/- 30 BP | (Cal BC 3620) (Cal BC 3610) (Cal BC 3520) | (Cal BP 5570) (Cal BP 5560) (Cal BP 5470) |
| 426632 | T2 RPC1.2 | 4940 +/- 30 BP | (Cal BC 3655) | (Cal BP 5605) |
| 426631 | T1C2 85-100 | 920 +/- 30 BP | (Cal AD 1180 ) | (Cal BP 770) |

**Table 3:** *Sediment samples and their associated 2 Sigma Calibrations*

| Beta sample no | Submitter Sample no | 2 Sigma Calibration (Cal AD/BC) | 2 Sigma Calibration (Cal BP) |
|---|---|---|---|
| 426639 | T7 RPC BOTTOM | (Cal AD 680 to 880) | (Cal BP 1270 to 1070) |
| 426638 | T6C4 RPC2.3 | (Cal AD 1435 to 1495) | (Cal BP 515 to 455) |
| 426637 | T4 RPC47-52 | (Cal BC 5980 to 5940)<br>(Cal BC 5925 to 5835)<br>(Cal BC 5825 to 5810) | (Cal BP 7930 to 7890)<br>(Cal BP 7875 to 7785)<br>(Cal BP 7775 to 7760) |
| 426636 | T4 RPC0-20 | (Cal BC 4895 to 4865)<br>(Cal BC 4850 to 4725) | (Cal BP 6845 to 6815)<br>(Cal BP 6800 to 6675) |
| 426635 | T2 300-330b.s | (Cal BC 5900 to 5745) | (Cal BP 7850 to 7695) |
| 426634 | T2 8.2m2 | (Cal BC 3520 to 3365) | (Cal BP 5470 to 5315) |
| 426633 | T2 RPC1.3 | (Cal BC 3635 to 3555)<br>(Cal BC 3540 to 3495)<br>(Cal BC 3455 to 3375) | (Cal BP 5585 to 5505)<br>(Cal BP 5490 to 5445)<br>(Cal BP 5405 to 5325) |
| 426632 | T2 RPC1.2 | (Cal BC 3710 to 3640) | (Cal BP 5660 to 5590) |
| 426631 | T1C2 85-100 | (Cal AD 1050 to 1080)<br>(Cal AD 1145 to 1220) | (Cal BP 900 to 870)<br>(Cal BP 805 to 730) |