Wetland geomorphology and floodplain dynamics on the hydrologically variable Mfolozi River, KwaZulu-Natal, South Africa

by

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Abstract

Wetlands in southern Africa can be considered a rarity, forming despite a regional negative water balance and a continental background of wide scale incision. These particular characteristics lead to southern African wetlands generally forming on drainage lines, where incision has been momentarily paused and water is locally abundant. The exact evolutionary history of valley bottom and floodplain wetlands is varied. However, their development follows four main themes; 1) those that evolve due to resistant lithologies outcropping on a drainage line and acting as local base levels, 2) those that occur on the coast, with current sea level preventing drainage line incision, 3) those that arise from a particular relationship with a trunk or tributary channel that blocks a drainage line with sediment, and finally, 4) those that occur in a region of dramatic loss of confinement, resulting in the formation of a wetland alluvial fan. Despite varied histories, all wetlands share a common thread, developing along a continuum from small and steep unchanneled valley bottom wetlands to large and flat floodplain wetlands. Incision in valley-bottom wetlands is controlled by a geomorphic slope threshold, whereby for a given wetland size, a particular slope may be considered stable. Wetlands exceeding the particular slope for their size are most likely already incised, or are vulnerable to incision in the near future.

This thesis examines the general evolution of drainage line wetlands, followed by a detailed study of a large coastal floodplain, the Mfolozi River Floodplain, located on KwaZulu-Natal's northern coastal plain. The Mfolozi Floodplain is one of South Africa's largest at 19 000ha and is located just south of the world heritage site of Lake St. Lucia, with the St. Lucia and Mfolozi River mouths occasionally joining at the coast. Although once a mosaic of *Cyperus papyrus* and *Phragmites australis* permanent and seasonal wetland, approximately 60% of the floodplain has been reclaimed since the 1920's for large-scale sugar cane cultivation. A smaller percentage is used for subsistence farming, while the remaining lower portion falls in the Greater St. Lucia Wetlands Park (which was renamed iSimangeliso Park in November 2007).

The formation of the large coastal valley in which the Mfolozi Floodplain now sits was created during a period of incision during the last glacial maximum 18 000 BP when sea level was 120m below the current level. The lowered sea level resulted in regional river rejuvenation and valley down cutting. The Mfolozi River valley became deeply

incised resulting in the formation of incised meanders upstream of the Lebombo Mountains. Below the mountains, less resistant lithologies of the Maputaland and Zululand Groups allowed the development of a wide coastal valley. Following the last glacial maximum, sea level rose, reaching its present level approximately 6000 BP. As sea level rose, coastal valleys were drowned and began to infill with sediments.

Above the floodplain, the Mfolozi River follows a meandering course in an incised confined valley. Upon passing through the Lebombo Mountains, the valley widens considerably from 915 m to over 6 km in just 1.15 km. This rapid change from confinement to a broad floodplain setting results in a reduction of carrying capacity of the Mfolozi River, creating a node of large-scale deposition at the floodplain head in the form of an alluvial fan. Deposition in this region causes a local oversteepening of the valley's longitudinal profile, with a gradient of 0.1%. Contrastingly, the mid- floodplain is almost flat, with a decrease in elevation of just 1 m over almost 6 km (0.02%). The lower floodplain, where gradient is completely controlled by sea level, has a steeper gradient of 0.05%. The reason for the rather drastic slope break in the mid floodplain is currently unknown, although it may be related to faulting in the underlying Tertiary aged Zululand Group, which is currently concealed by Quaternary deposits. In addition, tributary drainage lines that once flowed into the Mfolozi River have been blocked by long-term sediment accumulation on the floodplain. As a result, these drainage lines have become drowned and provide local conditions for the formation and accumulation of peat.

Besides geological setting, hydrology is commonly recognized as being the other most important factor in valley evolution. Flow in the Mfolozi River has been characterized as highly variable relative to the rest of the globe. The Black Mfolozi has the lowest Coefficient of Variation (CV) at 61%, followed by the White Mfolozi at 69% and the Mfolozi River at 79%. In addition, catchment precipitation was shown to be variable, especially when compared to global values. As a result of variable rainfall and discharge, the Mfolozi River shows hysteresis in sediment concentration on an annual scale, and there is an indication that hysteresis may also occur on a longer time scale during wet and dry rainfall cycles. This however, needs to be confirmed with a longer-term data set. Variable discharge and sediment transport leads to different floodplain processes and dynamics than would be expected for a river of regular flow. Since flow is generally very low in the Mfolozi River, and is characterised by a series of extremely large outlier flood events, the persistence of flood features is likely to be large. In addition, it is likely that extreme flood events are the primary drivers of floodplain evolution and dynamics in such variable settings.

The Mfolozi Floodplain wetland study throws light on floodplain process rates, and the forces behind floodplain dynamics in such hydrologically variable settings.

Preface

The research described in this thesis was carried out at the School of Environmental Science, University of KwaZulu-Natal, Howard College Campus, Durban, from January 2004 to December 2007, under the supervision of Professor WN Ellery.

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

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Contents

Introduction

1. Background to research

The recognition of wetlands as important providers of a range of ecosystem services has provided grounds for wetland conservation (Brinson 1993, Kotze *et al*. 2005, Millennium Ecosystem Assessment 2005). However, approximately 50% of wetlands have already been degraded or lost globally (Zedler and Kerher 2005), predominantly through agricultural activity that recognizes large floodplains and other wetlands as productive land (e.g. van Lienden 1959, Galatowitsch and van der Valk 1994). In southern Africa, the loss of wetlands (Kotze and Breen 1994) is likely to be comparable to that cited by Zedler and Kerher (2005), with wetland drainage efforts backed by government in the early and middle twentieth century (e.g. van Lienden 1959, Department of Irrigation 1948).

The impetus now, is on regaining lost ecosystem service provision, by rehabilitating wetlands that have been damaged. In South Africa, most wetland degradation has been the result of erosion and one of the challenges has been to understand why gully erosion is so pervasive in the region (Ellery *et al*. 2008). Repairing wetlands requires a deep understanding of how and why wetlands exist, how they function, and how and why they degrade. However, not all wetlands have been created equally. Brinson (1993) and Kotze *et al*. (2005) classify wetlands in terms of hydrogeomorphology, citing variation in water inputs and outputs, sediment fill and wetland functions as factors creating diversity in wetland types. However, in addition to these differences, factors controlling the origin and evolution and ultimately persistence of all wetlands appear to be highly varied. If we accept the proposal that successful wetland rehabilitation is likely to be linked to system scale understanding of wetland formation and degradation, it becomes apparent that understanding the origin and dynamics of wetlands is imperative in enhancing current conservation efforts.

Globally, much research has been focused on wetlands. Temperate wetlands in the northern hemisphere have tended to be the main research impetus with considerable attention focused on peatlands. However, peat is generally rare in southern African wetlands due to the highly seasonal climate that promotes desiccation of plant matter during dry periods. In addition, the geomorphic setting of southern Africa is considerably different from much of the northern hemisphere, where wetlands are often the result of a positive water balance and topography created by recent glaciation (e.g. Mitsch and Gosselink 1993, Galatowitsch and van der Valk 1994). In contrast, southern African wetlands can be considered to be unusual in that they are situated upon a continent that is currently experiencing large-scale incision (e.g. McCarthy and Rubidge 2005) and where evapotranspiration generally exceeds precipitation (Schulze 1997).

This situation would seem to suggest that southern Africa should have few wetlands. However, wetlands do exist in southern Africa where local geomorphic conditions create nodes of deposition that promote shallow flooding or the presence of near surface water tables. In general, these settings are created along the drainage network, where run-off is concentrated, and as such, wetlands in southern Africa may be best investigated using the methods of fluvial geomorphology.

Considering wetlands as geomorphic entities connected to the fluvial network provides a new viewpoint from which to consider the origin and dynamics of both valley bottom and floodplain wetland systems. Often wetlands have been considered as independent ecosystems, rather than geomorphic sub-systems interacting with ongoing changes on the drainage basin. It is when one considers the suite of drainage network processes associated with wetlands within a systems context, that the impacts of climate, hydrology and inherent geomorphic vulnerability can be properly assessed.

The research described in this thesis was aimed at developing a broad understanding of the origin of valley bottom and floodplain wetlands in southern Africa and the biophysical factors that influence their persistence. Following a macro-scale analysis of drainage line wetland origin, the study focussed on an example of a coastal floodplain wetland, the lower Mfolozi River Floodplain in order to gain an understanding of the interaction between sea level and floodplain morphology and dynamics.

2. Background to study area

The Mfolozi River Floodplain is an extremely large floodplain wetland system located approximately 200km north of Durban on the Kwazulu-Natal coastal plain. The Mfolozi River is a combination of two major tributaries, the White and Black Mfolozi Rivers, which reach a confluence approximately 50km upstream of the floodplain study reach. The resulting floodplain of the combined rivers measures about 19 000 hectares in extent, much of which was reclaimed during the early twentieth century for the cultivation of sugar cane.

Physiographically, the floodplain is bounded inland by the incised meanders of the Mfolozi River which have been cut into rhyolite rock of the Lebombo Group, and to the east by the Indian Ocean. It occupies an incised depression in the coastal plain of north-eastern South Africa that extends northwards as far as Kenya in east Africa. The region experiences a subtropical climate, with large seasonal variations in precipitation. Thus, despite subtropical temperatures, inundation of the floodplain is seasonal. During the dry season, high evapotranspiration rates often lead to desiccation.

The Mfolozi River Floodplain wetland is geomorphically complex. Currently, there are four emerging models of wetland formation (see Chapter 3), including the interaction of fluvial processes and bedrock lithologies of varying resistance to erosion (Tooth *et al*. 2004), the impact of sea level, interactions of trunk and tributary streams and the development of alluvial fans due to loss of flow confinement. Of these four models, three are represented in the Mfolozi Floodplain, providing a unique opportunity for studying interactions between the models, as well as embellishing understanding of the models themselves.

In addition to its interesting geomorphic setting, the Mfolozi River Floodplain's coastal location links the system to estuarine processes. While this research has not focused on the estuarine dynamics of the Mfolozi at all (this has already been done by other authors including Lindsay *et al*. (1996 a, b)), the interaction of estuarine and floodplain processes allows a theoretical linkage of the two systems that has not been previously clarified in the context of river-dominated estuaries. Geomorphology of estuaries was well integrated by Dalrymple *et al.* (1992). This was followed by additions to Dalrymple's evolutionary model by Cooper (1993, 1994, 2001) through a study of estuaries on the KwaZulu-Natal coast. Rivers of KwaZulu-Natal characteristically fit Cooper's river-dominated estuarine model, largely due to the high wave energy of the KwaZulu-Natal coast. However, the punctuated equilibrium of these estuaries may also be related to climatic variability of the interior, which also has an impact on floodplain wetlands. As such, there is room for the development of a model that ties together the punctuated equilibrium of river-dominated estuaries and aggradation variability on KwaZulu-Natal floodplains. As such, this study also focuses on how climatic variability in the catchment may impact upon downstream geomorphology. The Mfolozi River Floodplain thus provides a useful research location for structuring these ideas.

3. Aim and objectives

Aim:

To integrate models of wetland origin in order to locate the Mfolozi River Floodplain within a broadly unifying geomorphological systems framework and to investigate the geomorphic origin and evolution of the lower Mfolozi floodplain system, associated lakes and wetlands, and to assess the dominant processes and dynamics of the system.

Objectives:

1. Develop a broadly unifying model of the origin of drainage line wetlands in South Africa.
2. Assess the origin of the Mfolozi floodplain in context of wetland formation in southern Africa.
3. Describe present floodplain geomorphology, sedimentology and hydrology.
4. Relate present geomorphology, surface characteristics and sedimentology to general models of clastic sedimentation processes.
5. Analyse and describe the geological history of deposition on the lower Mfolozi floodplain.
6. Develop a conceptual geomorphic model describing the origin and evolution of the Mfolozi floodplain wetland system and associated wetlands.

4. Structure of thesis

The thesis is structured as a set of stand-alone papers intended for publication. Given this format, there is some degree of overlap between chapters, particularly with regards to some references and concepts. Individual chapters have not been formatted in the journal style to which submission is intended in order to maintain the internal consistency of the thesis. The overall design of the thesis takes the reader from macro-scale controls on drainage line wetland formation (Chapter 2) to catchment scale processes and their impact on the geomorphology of the Mfolozi River Floodplain (Chapter 3). This is followed by an analysis of local scale geomorphology and sedimentology of the Mfolozi River Floodplain (Chapter 4). On the micro-scale, the evolution and formation of individual floodplain lakes on the Mfolozi River Floodplain is discussed in Chapter 5. The material is synthesized in an overall discussion in Chapter 6.

Since individual chapters are to be published as papers, co-authors and journals where material will be submitted are provided in Table 1. The manuscripts have not yet been submitted for publication, which will happen once the examination process has been completed.

Table 1: Thesis chapters, publication title, planned journal for submission and co-authors.

Chapter	Title	Planned Journal for submission	Co-authors
Chapter 2	Wetlands on a slippery slope: a geomorphic threshold perspective.	Wetlands	WN Ellery and MC Grenfell
Chapter 3	Variable rivers: A new geomorphology? Case of the Mfolozi River, South Africa.	Journal of Hydrology	WN Ellery
Chapter 4	Geomorphology and sedimentology of the lower Mfolozi River Floodplain, KwaZulu-Natal, South Africa	Earth Surface Processes and Landforms	WN Ellery
Chapter 5	Tributary drowning by trunk channel aggradation: the evolution of Lake Futululu on the Mfolozi River Floodplain, KwaZulu-Natal, South Africa	Geomorphology	WN Ellery

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Chapter 1. Systems, geomorphic thresholds and dynamic equilibrium as essential elements of understanding wetland formation and dynamics in South Africa.

1. Introduction and context

Brierley and Fryirs (2005, p17) wrote that, "*The scale at which observations of the natural world are made constrains what is seen*", suggesting that landscape studies should include varied spatial and temporal scales to fully appreciate landscape evolution and processes. Similarly, understanding the formation and evolution of wetlands requires a framework that is capable of focusing on minute details of within wetland processes, while also being able to upscale to broader dynamics of catchment hydrology and climate. Thus, it is important to examine localized features within the wetland, such as:

- the character of an alluvial ridge,
- particle size characteristics and mineralogy of sediment present in the influent stream,
- the morphometry and dynamics of floodplain margin lakes,

as well as features that relate to the catchment, such as:

- drainage basin extent and pattern of rainfall,
- nature of flow with respect to variations in stream flow, and
- variation in sediment inputs.

Explicit examination of processes and interactions of different spatial and temporal scales makes systems theory an ideal conceptual framework for analysis.

The integrated approach of systems theory is well suited to geomorphology, and as such, it is an approach that has been adopted by many earth scientists. For instance, the fields of geomorphology and atmospheric science are fused in Tooth *et al*.'s (2007) study of the impact of climate change on the dynamics of the Klip River Floodplain. In Ellery *et al*.'s (2003) system view of the Mkuze River Floodplain, broad, long-term processes that drive the evolution of the Mkuze River system are considered. In Fryirs *et al*. (2007) paper on geomorphic buffers, barriers and blankets, catchment scale sediment supply is considered in conjunction with relatively minor geomorphic features (e.g. alluvial fans, channel armouring) that in combination control catchment sediment delivery. This is a classic case of using the scalar properties of systems theory to further geomorphic understanding.

Thus, the main strength of systems theory is the ability to break the object of study into manageable components without considering the components in isolation, but rather as sub-systems of a larger whole. Acknowledging the geomorphic system allows one to recognize that systems interact with other systems, and that understanding these interactions is as important as understanding the inner workings of a particular system. The multitude of linkages between sub-systems of different spatial scales has frequently proved a stumbling block for environmental managers, particularly when a small sub-system is the management focus. Since landscape complexity increases with system size (Schumm 1991), failure to comprehend how a sub-system may be connected to, and interact with, a larger system, may lead to management that may even cause environmental harm.

Wetlands of southern Africa may be considered together at a broad, continental scale because of their shared geomorphic history. The geological and geomorphic evolution of South Africa was largely dormant between 100 and 20 Ma, with the most recent mountain building event occurring approximately 330 million years ago, resulting in the Cape Fold Belt, the most recent glaciation occurring 245 million years ago, and the most recent volcanic activity occurring at the onset of the break-up of Gondwana, 180 million years ago (McCarthy and Rubidge 2005). Approximately 20 Ma, southern Africa experienced an uplift event along an axis that is roughly parallel to the coastline of the eastern seaboard, just eastwards of the current day Drakensberg (Partridge and Maud 1987, Partridge 1998). The location of the Neogene axis resulted in the eastern regions of southern Africa rising approximately 200m, while the western region rose less, approximately 150m. Five million years ago, uplift occurred on the same axis at a much larger scale, lifting the eastern regions a further 900m, while the western region rose a mere 100m in comparison, creating a large westward sloping plateau.

The lack of alternative geomorphic events in recent geological history places the Neogene uplift events, in combination with recent sea level changes describe by Ramsay (1995), as the most influential events on the geomorphic evolution of wetlands in southern Africa. The uplifted elevation of southern Africa has resulted in the sub-continent entering a period of long-term incision, with the subsequent development of an extremely well integrated drainage network. Along coastal areas, changing sea level has exerted a local control on wetland formation and dynamics. Nevertheless, as most of South Africa experiences a negative water budget (Schulze 1997), the development of wetlands in the region is limited to locations on the drainage line where incision is paused by local base levels, and surface water is present for most or all of the year.

Despite sharing a common history in terms of the broad evolution of the sub-continent, there is wide diversity in wetland geomorphology and origin that is related to local landscape and climate variation. The linkages between local scale processes of a particular wetland, the hydrological, geological and climatic factors concerning the wetland's catchment, and lastly, the geomorphic evolution of the sub-continent, should all be considered in interpreting a wetland's origin and dynamics. The study of the geomorphology of wetlands, therefore, makes full use of the scalar and connective properties of systems theory.

As such, this literature review considers systems theory in light of applications to fluvial geomorphology, and in particular, wetland geomorphology. The literature review begins with a broad introduction to systems and systems theory, followed by a discussion on system types most applicable to fluvial geomorphology (i.e. process-response and nonlinear systems). Concepts of geomorphic equilibrium and geomorphic thresholds are then considered, with respect to the application of systems theory to floodplain and wetland processes and dynamics.

2. General systems theory and earth science

2.1. Defining a system

"*In instances where there are from one to two variables in a study you have a science, where there are from four to seven variables you have an art, and where there are more than seven variables you have a system*'' Van Dyne (1980, p889). This comment regarding systems theory was also elaborated upon by Huggett (1985), who found systems theory to be increasingly under attack as an unnecessarily complicated concept. While it is over 20 years since the defense of systems theory, which now appears to have been widely accepted, the fundamental application of systems theory and the reality it proclaims to represent are still under debate.

However, it is important to place the use of systems theory in context. The theory did not arise to replace existing theory, but rather to allow the more harmonious use of existing understanding to create integrated knowledge. In contrast to some scientific studies that are sometimes somewhat reductionist in their approach, systems theory recognizes that systems may be simple or complex. As Sack (1992) recognized, systems theory allows one to focus on the sometimes complicated relationship between form and process at multiple scales. Most natural or earth systems comprise many variables and components, and the simplification of such systems without careful consideration, risks misunderstanding the system as a whole. Systems theory therefore promotes synthesis of available observations, and as a result, true integration (White *et al*. 1984).

The system is essentially a number of objects with attributes, that interact with one another, or as Schumm (1977, p4) simply wrote "*a meaningful arrangement of things*". Chorley and Kennedy (1971, p1) supply perhaps the most comprehensive definition, whereby a system is "*a structured set of objects and/or attributes. These objects and attributes consist of components or variables… that exhibit discernible relationships with one another and operate together as a complex whole, according to some observed pattern.*" More recently, Phillips (1992a, p195) explains that a system is "*a set of interconnected parts which function together as a complex whole. In a geomorphic system the parts are landforms, surface processes, and factors which control or influence forms and processes. The interconnections involve flows, cycles, transformations, and storage of energy and matter."* It becomes apparent from these definitions, that almost anything can meet the criteria of a system. While this may be criticized, it may also be defended as one of the strengths of system theory, although it has led to some degree of misunderstanding. Essentially, the globe is a system that may be increasingly divided into smaller and smaller subsystems, such as from the entire fluvial system, to a drainage basin, to a tributary of that drainage basin, and eventually in our case, to a specific wetland. What determines a system in a particular study is a question of scale. Since all systems are subsystems of a larger system, it is imperative to recognize that the manner in which a subsystem behaves may be related to other subsystems of the system, and in turn, that a specific subsystem may influence the behaviour of other subsystems.

The scale at which a system is examined is particularly important in a management context, as defining the system one hopes to manage has implications for how it is managed. Brierley and Fryirs (2005) suggest the use of a nested hierarchical framework in studying fluvial systems, whereby it is recognized that processes have varying importance at different geomorphic scales, while the entire system is interconnected. For example, catchment vegetation cover may alter rates of sediment delivery to a river, while at a local scale, it may control channel shape (Brierley and Fryirs 2005). Furthermore, acknowledging the scale at which an observation is made, and the limitations that the scale of observation may impose, is imperative. Often, broad scale investigations may ignore small-scale processes, leading to generalizations of system functioning. For instance, studying catchment erosion without considering the ability of the stream to accumulate sediment in features such as point bars or floodplain wetlands may result in overestimation of fluvial sediment delivery. Contrastingly, studies conducted at the smaller end of the geomorphic scale may be prone to up scaling, whereby results are applied to larger areas than is realistic for the available data.

Thus, Van Dyne's (1980) statement is misguided in that system theory was born as an alternative to scientific reductionism, out of recognition that many systems are more complex than can be treated by analysis of just a few variables (e.g. Sack 1992). Furthermore, this does not necessarily mean that all systems must be complex, but rather that they can be. The application of systems theory requires analytical adjustment to an appropriate scale (i.e. rather than a drainage basin, one can analyse a floodplain) such that a study may be simplified to deal with appropriate variables. However, the theory puts all systems in context, as components of a larger system and with smaller subsystems. Thus system theory is essentially a method whereby a natural system can be conceptualized as a unified whole, that can be examined on the basis of natural laws and theory from classical fields of research. Phillips (1992a) suggests that while systems theory is not strictly applied to the study of geomorphic systems, it is currently widely accepted and used with respect to understanding dynamic behaviour and interactions between individual landscape elements.

2.2. Simple and complex systems

Systems may be considered on 3 levels: morphological; cascading; and process-response, which need not be mutually exclusive (Huggett 1985). In fact, useful conceptualization of a specific system may require two or all three levels in some cases. A morphological system is a set of morphological variables which are thought to interrelate in a meaningful way in terms of system origin or function, a cascading system is a system whereby energy and/or matter moves on pathways or is stored within the system (Huggett 1985). Lastly, with respect to the process-response level, a system may be interpreted as one in which energy flows through a morphological system in such a way that system processes may alter the system form and in turn, the altered system form alters the process. Bedload transport, for example may be interpreted as both a cascading or process-response system. In a cascading system, bedload transport is the movement of matter and energy along a riverbed in such a way that both may alternate between storage and transport within the system. In the process-response model, bedload transport may increase or decrease, depending on flow variables, resulting in either deposition or erosion. This may in turn alter channel geometry and hydraulics and thus the original flow variables, causing a change in the total amount of sediment transported as bedload.

Three different types of systems may be considered under this framework: simple, complex but disorganized and complex organized systems (Huggett 1985). A simple system is one in which its future and past may be predicted using mechanical laws. The application of mechanical laws suggests that processes within such systems are reversible. Such systems rarely, if ever, occur as earth systems. A complex disorganized system is one in which the many components of the system interact in a nonlinear, seemingly random manner. Since there are many components and variables, the system can only be modeled in an average way. A complex, ordered system is one in which multiple objects interact in such a way that the system is ordered, even though it is complex (Huggett 1985). Systems dominated by process-response frequently fit into this category, since negative feedback mechanisms tend to reinforce processes, thereby creating order.

Generally, system type is determined by either a variation in scale (i.e. continuum of molecular to landscape scale) or whether a system is isolated, closed or open. Isolated systems do not interact with their external environment at all, whereas closed systems exchange energy but not matter, with the external environment (Huggett 1985). Open systems may store, exchange and transport both matter and energy. Isolated systems may sometimes be simple or complex disorganized systems, but they are rarely complex ordered systems, whereas open systems are generally complex ordered systems. The reason for this variation is often explained in terms of thermodynamic entropy.

Closed systems must, by definition, reach a state of equilibrium since external influences are unavailable to disturb equilibrium conditions. This end point, termed thermodynamic equilibrium, is one of maximum entropy and no net entropy change (i.e. $dS/dt = 0$) (Huggett 1985, Montgomery 1989). This steady state does not imply that no entropy is produced, but rather that its exchange is balanced (Montgomery 1989). An example of thermodynamic equilibrium occurs once water has been heated on a stovetop. As the water begins to heat up, entropy is low since heat is differentially distributed and leads to a predictable set of convection currents in the water. However, as convection currents distribute heat, the water begins to reach a higher entropy state. Once the water is uniformly heated, entropy, or molecular disorder, is at its maximum level, and thermodynamic equilibrium is reached since there is no apparent method of predicting where a specific water molecule is likely to move. Thus, thermodynamic equilibrium cannot be maintained while energy or matter is being input or transported through a system. Furthermore, energy inputs and transport are largely responsible for the development of order (low entropy). Thus, in a sense, nonequilibrium conditions are a source of order. This type of system is termed dissipative or nonlinear, since entropy and energy are exported from the system (Huggett 1988). Huggett (1988) further suggested that dissipative geomorphic systems evolve so as to maximize total energy dissipation, or to minimize work, while also seeking to equalize energy expenditure throughout the system. This concept is frequently applied by many fluvial geomorphologists, whereby work and energy minimization are used to explain fluvial behaviour. Far-from-equilibrium systems, an extreme type of dissipative system, are characterized by high levels of entropy or energy flow that permit process-response in improbable states (Montgomery 1989). The evolution of these systems is not towards a potential end point but is rather undirected, and is therefore deterministic. Such systems may therefore 'wander'. However, this concept generally applies to biological systems rather than geomorphic systems.

In the earth sciences, thermodynamic equilibrium is extremely rare. Instead, open systems are the norm (Chorley and Kennedy 1971). These systems cannot ever reach thermodynamic equilibrium because they are subject to varied energy and matter inputs, which maintain order and keep entropy low. However, thermodynamic equilibrium is a type of attractor state for all systems, even if they are incapable of reaching that point. Conceptually, this is a useful idea in that the way in which energy and matter is likely to be moved within a system (i.e. in a manner as to attempt to move towards thermodynamic equilibrium) may enable predictions and insight into system behaviour. Thus, the importance of the idea of thermodynamic equilibrium is not in its practical application, but rather in its conceptual usefulness in understanding a system.

A current trend in systems theory is the application of nonlinear dynamical systems theory (NDS), which combines the complexity of multiple process-response systems and apparent random behaviour in the analysis of dissipative systems (Phillips 1992b).

3. Process-response and nonlinear systems

Earth systems are frequently described in terms of the process-response concept. Process-response refers to the situation where a process alters system form and, in turn the altered system form causes a change to system processes (Huggett 1985). In this way, a system's evolution may in a sense be controlled by internal mechanisms. The most commonly recognized process-response mechanism is via a negative feedback loop, which usually is associated with evolution towards equilibrium. Positive feedback loops, conversely, are associated with nonequilibrium conditions, where the system does not tend towards equilibrium, but rather moves further away from an equilibrium condition (Renwick 1992). The concept of process-response is not new, but the manner in which the concept has been applied has become increasingly complicated, yet fruitful.

In nonlinear dynamical systems theory, the complexity of geomorphic systems is recognized. Phillips (1992b) suggests that system complexity arises from the cumulative impact of multiple process-response systems, operating individually and interacting, and giving rise to apparently random behaviours and patterns. The theory only applies to open systems that are therefore nonlinear and dissipative. Although such systems are associated with self-organization and order, they may also exhibit chaotic behaviour (Phillips 1992b). Werner (1999) concurs, suggesting that simple landform patterns (i.e. highly organized systems) frequently arise from complex system behaviour. Interestingly, nonlinear relationships have been widely accepted and used in geomorphic literature (e.g. bedload sediment transport linked to velocity), but the nonlinear component has seldom been formalized.

Sources of nonlinearity in geomorphic systems have been well described by Phillips (2003), and are summarized in Table 1. Non-linearity may occur as a result of factors that control how a system may adjust to change, or by controlling how system processes occur and interact. For instance, if there are multiple attractor states for a system following disturbance, such as different styles of channel morphology after a large flood on a floodplain, the exact trajectory of geomorphic change may be difficult to predict. Within-system processes may also cause non-linearity. Thresholds may cause sudden, unpredictable changes in system processes, while the storage of sediment in a geomorphic system may create lagged responses downstream. Some processes are self-limiting, either through differential process rates (e.g. deposition can only occur at a slower or equal rate as erosion upstream) or through virtue of the setting of a geomorphic system (e.g. a river cannot erode below sea level). Non-linearity may therefore occur through the development of unpredictable responses, or by increasing complexity in process rates that prevent matter and energy outputs from always balancing inputs.

Table 1: Summary of sources of nonlinearity in open systems, summarized from Phillips (2003).

Source of nonlinearity	Description of the development of nonlinearity
Thresholds	Results in sudden process changes
Storage effects	Results in lagged responses
Saturation and depletion	Effect of differential process rates
Self-reinforcing positive feedbacks	Features reinforcing themselves result in the initial condition having a large impact on the outcome
Self-limitation	Some processes are limitted internally by the system (e.g. incision limitted by base level)
Competitive relationships	Small changes can result in a change in process dominance
Multiple modes of adjustment	Various possibilities in process-response following disturbance
Self-organization	May offset or intensify the effects of external forcing
Hysteresis	Occurs when there are multiple values of a dependent variable for a single value of an independent variable (i.e. for a given discharge, sediment concentration may be x or y)

System complexity, as already mentioned, is likely to be the result of multiple process-response systems that are operating on nonlinear scales, creating the appearance of chaos and non-random behaviour. Nonetheless, these systems may also appear to be highly ordered morphologically (Werner 1999). The extent to which irregular behaviour is attributable to chaos rather than heterogeneity is usually unknown (Phillips 1992b). One of the main ideas that separates NDS from classical system process-response theory is the unique perception of equilibrium and trajectory paths, where systems are generally perceived as moving towards some equilibrium state that is a stationary endpoint. In NDS theory, systems move on a trajectory from a repellor (unstable equilibrium states), towards an attractor (the stationary end point) (Montgomery 1989). As Phillips (1992b, p222) explains, "*In an unstable equilibrium state, the response of the system to a perturbation will not return the system to, or nearly to, its pre-disturbance state. Rather, the system will remain in disequilibrium or move to a new, different equilibrium state*."

Disturbances are conceptualized as bifurcations that may be endogenic, exogenic or system inherent (e.g. threshold). They may represent a transition from one equilibrium state to another, or from regular or periodic behaviour to chaos (Phillips 1992b). Generally, successive bifurcations occur if one system parameter is increased. Each solution, or process-response, is associated with a distinct time-space structure or trajectory (Huggett 1988). The implication of multiple solutions or responses gives support to the idea that multiple equilibria states (attractors), rather than one, must be possible. Between bifurcations, deterministic (universal) laws control the behavior of the system. However, at the threshold at which a bifurcation occurs, random fluctuations may control which trajectory is followed. Thus, non-linear dynamical systems are both deterministic and probabilistic. An example of this in geomorphology is the occurrence of an avulsion. When gradients perpendicular to an alluvial ridge become substantially greater than the long gradient of a river, an avulsion is probable. However, the exact location of an avulsion in time and space is somewhat random and following such an avulsion (a bifurcation), the system returns to a state where universal, deterministic laws apply.

Nonlinear dynamical system theory suggests that system response may sometimes be random, although the nature of randomness may vary. Systems that are dominated by random chaos are deemed deterministic (Phillips 1992b), and in this case, randomness is considered to be inherent. In reality, many systems may appear to be determinist in nature since we lack the understanding required to predict their behaviour. These systems, that appear to be random due to lack of understanding and that could be predictable if we had the required knowledge, are deemed stochastic. Thus, in stochastic systems, randomness is apparent rather than real. While the majority of systems are likely to be stochastic, it is also recognized that the existence of thresholds and bifurcations may result in the development of chaotic systems.

Huggett (1988) summarizes the implications of these variations from classical process-response ideas. Nonlinear dynamical systems may be both probabilistic and deterministic; a bifurcation may have varied results since there are a number of possible system responses. As such, change in a landform may not be uniform due to variation in susceptibility to change and bifurcations may destroy systems or their components in a catastrophic manner. Furthermore, since multiple equilibria states are possible, and because landforms may have differing response times, some components may adjust immediately following a disturbance, while other subsystems may require thousands of years to respond, such as in the case of recently glaciated regions. This further suggests that the concept of history in geomorphic systems may be important. Essentially, a system's history can only become unimportant if the system reaches equilibrium, or alternatively, is reset (e.g. by glaciation). Since this is rarely the case, the history (termed 'inheritance' by Ahnert (1994)) of a geomorphic system may influence or constrain the evolution of a system, forcing the system to adopt one set of configurations over another (Montgomery 1989).

While almost all geomorphic systems are nonlinear, not all are complex (Phillips 2003). It is this understanding that has led to the development of Complex Nonlinear Dynamics (CND), which is applied to nonlinear systems that are complex, but where complexity is not always a result of nonlinearity. Thus, complexity arises from a combination of multiple variables and multiple nonlinear process-response mechanisms, as well as the availability of information at all scales (i.e. information at multiple levels within the system, Malanson 1999). Furthermore, systems may function in variable modes of stability, instability and chaos (Phillips 2006).

The need for CND theory arose as it was recognized that in some systems small or short-lived perturbations might result in huge environmental change because of the effects of chaos and dynamical instability. Interpreting historical or geological evidence from such events can be problematic unless the effects of complexity are acknowledged. Essentially when a system is unstable, a small change might result in the 'tipping' of the system towards a different end point that would be difficult, if not impossible, to predict (Phillips 2003). For example, Knox (1993) has shown that small climatic changes may result in disproportionately large impacts on both the magnitude and frequency of extreme floods. This change may be the result of internal interactions, rather than external forcings.

Acknowledgement of nonlinear complexity may have advantages in system modeling and understanding. While it may first appear that a random or chaotic system is impossible to model or predict, this is not the case. While a nonlinear complex system cannot be known to the extent of either linear or nonlinear simple systems, use of the appropriate theory will enhance understanding. Furthermore, techniques have advanced such that chaos is detectable in systems, allowing analysis of nonlinear system behaviour (Phillips 2006). While chaotic systems cannot be modelled in the long term, they frequently may be modelled in the extreme short term (Phillips 2003). Additionally, chaos, unlike complete randomness, generally occurs within discrete boundaries. There are therefore not innumerable possible trajectories of change following a perturbation, but there tend to be just a few. When considering equilibrium in such systems, the distinction between equilibrium and non-equilibrium systems may be assessed by analyzing system stability, proportionality of response and commonality of response (Phillips 2006). These variables also give an indication of the relative importance of chaos and/or randomness in system function.

Werner (1999) suggests an alternate method of beginning to understand complex systems. A system may be simplified by drawing out variables that conform to two characteristics of non-linear, open systems. Firstly, parts of the system that undergo the most rapid change tend to be centralized. For example, the linear belts of sediment transport within a river on a large floodplain, or a rip current on a shoreline, are concentrated areas of rapid system change. Secondly, numerous system variables may be intrinsically dependent on each other by a process called slaving. Generally, fast variables may become slaved to slower variables and as such lose their independent nature. Werner (1999) uses the example of a sand dune, whereby the movement of sand grains are slaved to the overall movement of the sand dune, even though the cumulative movement of sand grains is responsible for the movement of the dune as a whole. The identification of such variables, Werner (1999) argues, will allow the building of a hierarchical model.

4. Concepts of geomorphic equilibrium

4.1. The existence and appropriateness of equilibrium

The concept of equilibrium is a general model that has been extensively used by geomorphologists to describe and explain landform evolution that appears to be balanced in space and/or time (Bracken and Wainwright 2006). Equilibrium, when strictly applied, occurs when inputs to a system equal the output, or alternately a change in input must cause a measurable change in output, immediately, or after a finite time (Phillips 1992a). This includes all types of exchanges, including matter energy and entropy. Intuitively one can see that this is unlikely to be the case when applied to most systems, and thus it has been suggested by Phillips (1992a) that in reality, equilibrium is only likely to apply to system components rather than systems themselves. Furthermore, equilibrium is likely to exist only at restricted spatial and temporal scales due to its inherent instability (Renwick 1992). The obvious question therefore arises as to what use such a concept, if it never actually occurs, can be. Furthermore, what are the practical applications of the equilibrium concept to complex systems where all inputs and outputs cannot be known? Phillips (1992a, p199) states, "*The traditional equilibrium approach implicitly offers the hypothesis that there is a single dynamically stable equilibrium state toward which a geomorphic system will evolve, given sufficient time*". In this context, it is not that equilibrium is a real state, but rather a hypothetical endpoint that enables conceptualization of system processes and behaviour. Since geomorphic systems tend to be complex, the usefulness of the equilibrium approach is in attempting to loosely predict the future tendencies or path of system evolution by envisaging an endpoint. While this may seem notional rather than scientific, the equilibrium endpoint is couched in scientific theory since the manner in which a system would move towards equilibrium is generally based on concepts outside systems theory. Renwick (1992) explicitly noted that the equilibrium concept was most beneficially applied in the context of systems theory. He suggested that the occurrence of multiple equilibria at varied system scales explained the coincidence of both variability and stability in the landscape.

Bracken and Wainwright (2006) further supported the use of the equilibrium concept as a metaphor. While some authors have suggested that equilibrium is a real endpoint to be reached (e.g. Willett and Brandon 2002), it was intended as a conceptual endpoint to aid the description, explanation and understanding of system process and response. A combination of this problem, the literal application of equilibrium concepts, and a rather confused semantics war on equilibrium definition has led to the equilibrium concept being increasingly rejected (e.g. Stott 1998). Nevertheless, the number of supporters of equilibrium theory far outweighs the critics, provided the use of equilibrium theory and terminology is appropriate (e.g. Ahnert 1994, Bracken and Wainwright 2006). Contrastingly, Phillips and Gomez (1994) argue that the semantics war over equilibrium terminology is overexaggerated, and that multiple terms for different equilibrium terms is not harmful, provided authors are particular about their implied meaning, as suggested by Thorn and Welford (1994). Currently, no one author's terminology has been more widely accepted than another's. As a result, there is no option but to heed the request of Phillips and Gomez (1994) and Thorn and Welford (1994) by ensuring that the use of a particular equilibrium term is contextualised and explained.

Nevertheless, numerous terms for different types of equilibrium have been suggested and comprehensively summarized by Bracken and Wainwright (2006) in terms of their application to form and/or process. It is still useful, however, to briefly look at the main equilibrium terms in use and their definitions. The definition of *equilibrium* is widely accepted as implying a constant relation between input and output of matter, energy or form in a time interval (e.g. Renwick 1992). This concept varies in its application, from literal, to conceptual, wherein the definition implies a state to which a system moves. The term *disequilibrium* is applied to systems that tend towards equilibrium but have not had sufficient time to reach such a condition, while *nonequilibrium* systems do not tend towards equilibrium at all, even with relatively long periods of environmental stability (Renwick 1992). These systems, undergoing frequent changes in process and form, may be the result of threshold behaviour, positive feedback mechanisms or chaos (Renwick 1992).

All terms other than these tend to be conflictingly defined. *Dynamic equilibrium* is generally defined as a progressive change of equilibrium over time. For instance, as a valley erodes following uplift, the altitude of the valley system gets progressively lower, despite equilibrium conditions of sediment erosion, transport and deposition. Ahnert (1994) objects to the term because the existence of a trend suggests a lack of equilibrium. He suggests then, that equilibrium may only be dynamic if it is maintained by a self-regulating interplay of forces and that dynamic equilibrium therefore is much the same as steady state, as described in other literature. However, *steady state* is described as a state in which the process-response system fluctuates with short-term oscillations around a mean state *without* any progressive change over time (Schumm 1979, Montgomery 1989). An example of steady state equilibrium in nature is a flow record over a long period of time, where the average shows no trend of change. Contrastingly, a dynamic equilibrium is characterized by a progressive trend. While a steady state is not a stationary state in that some fluctuation is likely, there is no long-term trend of change. In turn, the term *quasi-equilibrium* is similar to steady state equilibrium, as suggested by numerous authors (e.g. Bracken and Wainwright 2006). Langbein and Leopold (1964) describe quasi-equilibrium as being the most probable state between two opposing tendencies.

Dynamic metastable equilibrium and *metastable equilibrium* are applied in a similar way, except that in the former, metastable equilibrium occurs in conjunction with a progressive trend of change (i.e. dynamic equilibrium). In Schumm (1979), dynamic metastable equilibrium occurs when a system is in dynamic equilibrium, but there are sudden step-wise shifts in equilibrium. This occurs when external or intrinsic factors carry the system over a threshold, and into a new equilibrium regime. Cut-and-fill landscapes described by Brierley and Fryirs (1999) fit into this concept of equilibrium. Thus, a metastable equilibrium is characterised by step-wise equilibrium adjustments, rather than only gradual change. Montgomery (1989) concurs with Schumm (1979), describing metastability as when equilibrium shifts by jumps and is maintained by small fluctuations around an average without long term progressive change.

One of the least documented equilibrium terms, perhaps since the term is more widely applied in the biological sciences, is *punctuated equilibrium*. Gould (1984) argued for the acknowledgement of punctuated equilibrium since its frequent appearance in palaeontology suggested applications for geomorphology. Gould (1984) describes punctuated equilibrium as relatively rapid flips between fairly stable equilibria. Kennedy (1992) further explained punctuated equilibrium as change that is sometimes episodic and leads to rapid adjustment followed by stasis. Essentially, punctuated equilibrium is when change occurs in a series of sudden steps, rather than through long-term adjustment at a fairly gradual and constant rate. This is in contrast to the paradigm of gradualism, where processes occur at a constant rate, which is unconsciously held by many geomorphologists. Punctuated equilibrium suggests that change may be catastrophic and abrupt, and this type of change may in fact be common, rather than aberrant. Furthermore, proponents of the theory emphasise that a long period of stasis is as important as a brief period of change (Hallam 1998). However, it has emerged that theories of gradualism and punctuated equilibrium can co-exist, leading several authors to describe the phenomena as a form of pluralism in evolution (e.g. Gould 1984, Gould and Eldredge 1986, Hallam 1998, Benton and Pearson 2001).

Punctuated equilibrium was used by Cooper (1993, 1994, 2001) to describe the evolution of river-dominated estuaries along the east coast of South Africa. He suggested that accommodation space in these estuaries generally remained full. However, episodic floods scoured the estuaries, allowing sudden infilling thereafter. In intervening years, estuary systems merely acted as conveyor belts of sediment. Initially, punctuated equilibrium may appear to be similar to metastable equilibrium. However, punctuated equilibrium suggests periods of nonequilibrium (i.e. sudden catastrophic change), where after equilibrium is suddenly reestablished, corresponding to a period of stasis.

Stasis can be interpreted in two ways, a period of geomorphic equilibrium, or one of very little morphologic or process change. In the case of river-dominated estuaries on the KwaZulu-Natal coast, the lack of accommodation space resulted in a throughput of sediment, with these estuaries being described as a ‘conveyor belt’ of sediment (Cooper 1993). This conforms to the most basic of equilibrium premises, that inputs of matter and energy must balance outputs. Thus, Cooper’s periods of stasis conform to equilibrium. In this case, punctuated equilibrium is considered in terms of accommodation space, which is used as an indicator of geomorphic change.

While there are some applications of punctuated equilibrium in the literature, the majority of geomorphology and hydrology literature do not consider the idea in a practical context. It is possible that this is because many objects of geomorphic investigation support ideas of gradualism, such as regular flow rivers that constantly transport sediment downstream at relatively uniform rates, or processes of soil creep on a hillside.

Despite the clear uses of equilibrium theory in creating understanding, the reality of equilibrium and its doubtful application to complex systems requires further analysis. In general, it is useful to break systems into subsystems, in which equilibrium may in fact be possible, depending on scale. This idea is further advanced in NDS theory, where the incorporation of catastrophe theory suggests that systems may have numerous equilibrium states based on their underlying workings (Bracken and Wainwright 2006). However, the existence of multiple equilibria refutes concepts of minimization since there can be more than one attractor state, rather than one single, fixed equilibrium state. In simple systems, equilibrium is likely to be the norm, since there are few variables. When there are three or more variables, the system may show chaotic behaviour. In this case, the system may show attractor points or cycles that correspond to equilibrium behaviour.

Overall, it appears that the application of equilibrium concepts within a systems theory framework is fruitful, provided analysis proceeds with caution. By design, geomorphic systems will always strive towards a certain state of equilibrium (Montgomery 1989) because of the manner in which base levels provide a type of endpoint for geomorphic evolution.

4.2. Geomorphic thresholds

Most simply, a threshold is the point at which a system's behaviour suddenly or abruptly changes (Phillips 2003). Geomorphic thresholds occur at a wide diversity of landscape scales, and are thus intuitively considered under the umbrella of systems theory. Two types of thresholds are recognized by Phillips (2003); those that are a ratio of force and resistance (e.g. the power law governing bedload sediment transport) and those with linked processes (e.g. incision that leads to sediment production that affects deposition downstream). Thresholds are innate or intrinsic to a system in that they are self-generated and as such, they were once defined as '*the condition at which there is a significant landform change without a change of external controls*' (Schumm 1979, p485). It is now recognized that while thresholds are inherent in some systems, humans may accelerate the pace at which a system reaches a threshold (Schumm 1979).

Erosion has traditionally been considered to have been caused by only external factors, leading to the following comment of Schumm (1979, p486), that "*the assumption that all major landform changes or changes in the rates and mechanics of geomorphic processes can be explained by climatic or tectonic changes has prevented the geologist from considering that landform instability may be inherent*." The erosion cycle could not always be used to explain sudden erosional episodes, and in some cases possible external factors causing erosion could not be established. This eventually led to the conclusion that continued landform evolution could eventually reach a point of incipient instability, and instability could therefore be inherent rather than driven by external factors. However, only when the landform has evolved to this critical point, will adjustment or system failure occur.

The importance of the threshold concept is that external parameters are not always required to cause landscape change. Nevertheless, this is not to say that adjustment or failure must always be inherent, but rather that a system must be vulnerable to change in order for it to occur. In some cases, large storms may cause a sudden change in a geomorphic system, and thus appear to have generated the change. However, an abrupt change is only likely if a system was approaching or was at an instability threshold. In this way, intrinsic and extrinsic factors often coalesce (Schumm 1979). Within a landscape, landforms often react differently to environmental events, possibly due to varied proximity to a particular threshold. Thus, in some large storm events, some landforms may completely alter, while others may remain unchanged. Alternatively, vulnerable systems close to or at threshold may undergo accelerated change or threshold breaching because of human-induced catchment change. In the case of gully erosion, overgrazing in a catchment may lead to decreased vegetation cover and thus increased run-off. Increased run-off is similar to suddenly increasing discharge. Since the slope is stable only for certain discharges, an increase in run-off may push the system over a threshold, resulting in erosion (e.g. Zucca *et al*. 2006, Morgan *et al*. 2003). Run-off may be increased by urban development that creates impermeable surfaces. In addition, if deposition is accelerated through increased sediment supply, a system's slope may increase more rapidly to the point of threshold.

In essence, the concept of thresholds leads away from the traditional view of gradualism inherent in geomorphology. Since thresholds result in abrupt or sudden change, erosion and deposition on a basin level cannot always be progressive, but rather must be episodic, cyclic or step-wise. As such, thresholds are often associated with systems that are perceived to be in metastable or punctuated equilibrium.

The effect of geomorphic thresholds on drainage lines is rather complex but can be used to understand valley bottom and floodplain wetland evolution. This is because the threshold concept applies wherever incision or deposition occurs, and wetlands are located in zones of deposition and therefore naturally occurring gradient change (Ellery *et al*. 2008). From a hypothetical point of river rejuvenation, as would occur following continental uplift, a period of incision ensues. However as incision progresses, the amount of sediment delivered to a system become too great to be further transported downstream, leading to localized deposition and aggradation. However, continual aggradation leads to slope steepening laterally or longitudinally downstream of the node of deposition, and once again, erosion might occur. Rates of aggradation or incision may be accompanied by concomitant changes in river channel pattern within a particular system. Thus in this case, the response to uplift is complex and deposition and erosion co-exist in a dynamic feedback relationship within a drainage system. Those systems that are not experiencing excessive deposition likely to maintain geomorphic stasis, while those that erode are frequently at or near their geomorphic threshold, and erosion may have been triggered by natural or human causes.

Schumm's (1979) model of geomorphic thresholds is also useful in conceptualizing landscape evolution. For example, the theory suggests that uniform removal of sediment from a landscape is not possible during an erosional period. Even during periods of regional erosion and incision, some deposition must occur in discrete areas since erosion leads to flooding of a drainage line with amounts of sediment that a stream is unlikely to have the capacity to transport. Nodes of deposition in this setting are likely to move down the drainage line as if on a conveyor belt. In these instances, deposition is likely to be transient and may lead to valley bottom development for short periods only.

The geomorphic threshold concept need not only apply to valley or hillslope gradients. Threshold slopes also apply in floodplains to the cross-valley gradient. For instance, sediment deposition in the channel and on its levees causes super-elevation of the channel above the surrounding floodplain. Eventually, the lateral gradient will critically exceed the down-valley slope. At some point, usually during a large flood event, a threshold will be exceeded where erosion perpendicular to the channel will result in preferential water flow across the levees rather than down channel, resulting in an avulsion. System theory and concepts of equilibrium may therefore be used to further enhance the explanation of avulsion events.

5. Systems theory and wetland geomorphology

5.1 Geomorphic thresholds and wetland evolution

South African floodplain and valley-bottom wetlands are located primarily along drainage lines, and have been formed by the development of local base levels that induce nodal deposition (Ellery *et al*. 2008). The consequence of such deposition is multi-facetted. Not only does deposition allow the formation of a particular geohydrological setting that is conducive to the development of wetland systems, but it also causes adjustments to the longitudinal profile of the drainage line on which the wetland happens to occur. As sediment accumulates at a local base level, slope is reduced upstream of the deposition node, while conversely, slope is steepened at the wetland system's toe. It is the effect of deposition, and its link to wetland formation, that indicates the usefulness of the geomorphic threshold approach in understanding the geomorphic vulnerability and persistence of wetlands.

Since wetlands are located in depositional settings, they are inherently prone to instability. As Schumm (1979) expands, deposition within a particular drainage line eventually causes oversteepening of the longitudinal slope, ultimately resulting in incision. As incision progresses, even larger amounts of sediment are supplied to the fluvial system and at some point, sediment supply may exceed the ability of the stream to transport sediment, and deposition occurs. Current day wetlands are located within a zone of deposition that may be poised in geomorphic time to enter a period of incision. The slope at which a wetland becomes critically steep and vulnerable to adjustment and erosion is a geomorphic threshold of the system. All natural systems have an inherent tendency to establish a dynamic equilibrium due to negative feedbacks (Ahnert 1994). In this context, wetlands can be seen as components of a larger process-response system in a state of dynamic equilibrium, whereby wetlands are not the endpoint of landscape development, but rather a recurring step.

The position of wetlands in the geomorphic landscape suggests that it is the mode of their formation that leads to incision and eventually system failure (as in Schumm 1979). Interestingly, this suggests that wetlands are inherently vulnerable to change, although the magnitude of vulnerability varies according to the rapidity at which a specific system is likely to reach its geomorphic threshold. Furthermore, where wetlands incise, erosion may be natural and therefore inevitable, by virtue of an intrinsic threshold, or it may be that an external factor has exposed a predisposed vulnerability, causing incision (Schumm 1979). Thus, changes to climate or human interference cannot always be blamed for sudden wetland erosion. However, when a system reaches approaches a threshold value, external stimuli may help the system to reach or cross that threshold, resulting in gradual or sudden change, such as that initiated by erosion. Importantly, the system must be close to the threshold for the change to occur at all.

The potential applications of geomorphic threshold concepts to wetland management and rehabilitation are enormous. The proximity of a wetland to a geomorphic threshold is an indication of its inherent vulnerability to change. In addition, the threshold concept provides insight into external parameters, such as changes in climate, changes in catchment land use or activities in the wetland, which may accelerate movement towards the threshold. Information about proximity to a geomorphic slope threshold may be useful in monitoring, planning and implementing both wetland management and rehabilitation.

5.2. Floodplain processes and dynamics

Nonlinear systems are systems in which the movement of energy and/or matter into and out of the system is not necessarily equal, and where relationships between these processes is not linear, suggesting that the system as a whole is never at equilibrium. Dissipative systems occur where energy is dissipated within the system, such that order is maintained. Natural floodplain environments fall neatly within this category on a number of levels. As locations of sediment accumulation and storage, they are dissipative and nonlinear. Although the sediment load in a floodplain river fluctuates, the output of sediment from these systems tends to be more constant than inputs. During floods, the input of sediment may greatly exceed outputs such that floodplains are depositional. However, during low flows, a floodplain may temporarily reach equilibrium as sediment inputs and outputs are roughly the same, unless the stream is partially aggrading or eroding. Thus on a long-term basis, floodplains are depositional features rather than equilibrium features. In addition, floodplain sedimentation follows an ordered accumulation pattern that results in characteristic channel, levee and backswamp deposits, as has been documented by numerous authors (e.g. Pizzuto 1987, Marriott 1992, Asselman and Middelkoop 1995, Makaske *et al*. 2002, Törnqvist 1994, Magilligan 1992). The dissipation of high-energy floodwaters results in the development of physical order on the floodplain. Furthermore, the unequal distribution of entropy in nonequilibrium systems is exhibited by the concentration of entropy in river belts across the floodplain. So, it appears that floodplain environments can be characterized and explained in terms of systems theory terminology, but what does that actually mean, and how can the interpretation move forward?

The usefulness of systems theory arises when one begins to apply equilibrium and threshold concepts to a system. By design, geomorphic systems are directional in nature, in that irrespective of landform change and processes, all landscapes tend towards reducing relief. Nevertheless, the manner in which different systems do so, and how energy is managed and distributed within a system, plays a large role in altering system morphology. Even though, as nonlinear systems, most systems are assumed to be in disequilibrium, there is recognition that unless controlled by positive feedback mechanisms, there is a central tendency for systems to move towards equilibrium. In addition the traditional concept of equilibrium as some metaphorical endpoint has changed slightly, and movement towards equilibrium is increasingly envisaged as a series of ongoing mutual adjustments controlled by feedback mechanisms, which produce different results at different locations (Phillips 2002). Thus, of interest, is the rate and frequency of change to a system, which may be conceptualized through equilibrium theory. Phillips (2002) stresses that multiple types of equilibria are likely to coexist in a landscape at any one time, and that such equilibria are likely to be transient and unstable. Nevertheless, how does infilling of the Mfolozi floodplain occur? Gradually and equally over time, or sporadically in large, generous bursts? Is equilibrium likely to tend towards the system being dynamic, or metastable or punctuated? If the evolution is punctuated, how does the system alter to accommodate such change?

The concept of bifurcations is useful in terms of floodplain evolution, particularly in the context of avulsions. A bifurcation occurs when the system reaches a critical threshold and must reorganize. Generally, successive bifurcations occur if one system parameter is increased (Huggett 1988). Since the system is complex, there are multiple ways in which the system could reorganize and multiple trajectories upon which evolution could continue. In nonlinear dynamical system theory, the bifurcation is considered to be both deterministic and probabilistic. Between bifurcations, deterministic (universal) laws control the behavior of the system. However, at the threshold when a bifurcation occurs, fluctuations, which are chance like, may control the trajectory of change. As an example, avulsions are probable once the alluvial ridge reaches a certain height of super-elevation above the floodplain, or cross-valley gradients reach a certain threshold. Much research has centered on determining avulsion probabilities and resulting alluvial architecture based on physical variables (e.g. Törnqvist and Bridge 2002, Morozova and Smith 2000, Slingerland and Smith 1998, Mackey and Bridge 1995, Heller and Paola 1996, Smith *et al*. 1989, Bridge and Leeder 1979). However, the exact location of an avulsion in both time and space is essentially chance-like in character and may be completely unpredictable. This chance-like occurrence determines the trajectory of continued floodplain evolution, with sometimes unknown or unpredicted implications. Following such a bifurcation, different parts of the system may take different lengths of time to adjust. For instance, the new gradient of the stream eventually alters stream capacity and competence, as well as channel pattern and morphology, but all of these act over different time scales. As a result, there may be old (inherited) and new elements within the same landscape. This character of dissipative systems is termed susceptibility to change by Huggett (1988).

The extent to which a bifurcation is system-altering depends on the extent to which the original system is forced to change or adopt a new equilibrium. Benson (1984) arranges these disturbances on a continuum from crisis to cataclysm. A crisis occurs when an event causes sudden alteration of a system's principal structures, but through the absorption of this stress into its subsystems, the system survives. A catastrophe is an event when the sub-systems fail, but the system still survives, while a cataclysm is the complete destruction of both systems and subsystems. Benson (1984) suggests that system stress is primarily absorbed through the use of redundant pathways that were once integral to the system, such as the adoption of former channel courses to dissipate floodwater during a large flood event.

Alluvial fans within floodplain settings are more complex in terms of equilibrium history and tendency. The rapid transformation of such systems has led some authors to label them nonequilibrium features, in that they do not appear to display a tendency to move towards equilibrium (e.g. Renwick 1992). This is largely because of the effect of thresholds on the evolution of these systems, which tends to hasten system processes. It can however be argued that alluvial fans behave in a similar manner to floodplains (i.e. reach a gradient threshold and avulse), but on a much more rapid time scale due to higher sedimentation rates. This does not seem an adequate reason to suggest that there is no equilibrium tendency in alluvial fans.

Deciphering floodplain processes and evolution involves the sifting of multiple layers using various components of geomorphic system theory. It appears likely that evolution of such coastal systems will comprise both elements of gradualism and sudden disturbance overlain over time. For instance, stepwise adjustments of sea level may be seen to be in line with jumps in the context of metastable equilibria. Contrastingly, tectonic adjustment could be both gradual and catastrophic, as was the case following the occurrence of an earthquake in St. Lucia in 1932 (Krige and Venter 1933). In addition to the infrequent disturbances to the system caused by tectonic activity, infrequent, large rainfall events such as caused by occasional tropical cyclones that veer unusually far south with recurrence intervals of 300 years such as Cyclone Domoina may superimpose behaviour that could be typical of systems with punctuated equilibrium. This particular aspect is rather promising since the estuaries on which Cooper's (1993, 1994, 2001) theory was based, are primarily those of KwaZulu-Natal.

6. Conclusion

Much of current systems' theory application is in the development of mathematical models that strive to represent the real world to a certain degree of accuracy. The development of poor models is generally blamed on data inadequacy, in terms of both the number of variables that need to be considered, and the length of records available. However, a second major problem to model development lies in the acknowledgement of nested sub-systems, which together create a system of formidable complexity. In terms of wetlands, understanding their origin, formation and dynamics is intimately linked to acknowledging larger scale processes (i.e. evolution of the sub-continent), in addition to micro-scale processes that occur within the wetland, as well as processes at the intermediate scale, such as those related to the drainage basin.

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Chapter 2. Wetlands on a slippery slope: a geomorphic threshold perspective

Abstract

Wetland formation in southern Africa may be remarkable in that most of the region is characterised by a climatic negative water balance, where atmospheric water demand exceeds precipitation. Furthermore, uplift of the subcontinent 20Ma and 5Ma has resulted in the region undergoing a long-term period of incision. Despite these factors, wetlands occur throughout southern Africa, characteristically clustering along drainage lines where water is concentrated. Wetland formation in the region is associated with local base levels on drainage lines, where long-term incision can be momentarily paused and sediment deposition can occur. There are 4 main landscape settings that cause local base level creation and are thus associated with wetlands: 1) varying bedrock resistance on the longitudinal profile, 2) alluvial fan formation on a trunk river, 3) approaching sea level, and 4) aggradation on either the trunk or tributary channel that causes partial drowning of the other channel. Characteristic of all these settings is some degree of deposition, the long-term consequence of which is upstream flattening, while gradients steepen towards the wetland toe. Eventually, ongoing deposition may increase the downstream slope in excess of what is stable under prevailing run-off and discharge conditions, resulting in the crossing of a geomorphic threshold. In channelled and unchanneled valley-bottom wetlands, crossing of the geomorphic threshold is associated with incision. Data suggests that wetlands exceeding 1500ha in size are likely to be floodplains, and that for non-floodplain wetlands, the relationship between wetland size and slope determines whether or not a wetland will erode. For a given size, wetlands with a steep longitudinal slope are likely to erode. In landscape terms, crossing of the geomorphic threshold may cause wetland erosion in one part of the catchment, while wetlands may form downstream. However, the time scales involved in wetland formation may be too long in a human context, necessitating intervention in the form of rehabilitation. In this case, the threshold concept may be used to prioritise wetlands that require intervention in the near future.

1. Introduction

Southern Africa is currently a region of remarkable geological stability. The most recent mountain building event culminated in the Cape Fold Belt some 330 Ma (McCarthy and Rubidge 2005), followed by the creation of a massive basaltic outpouring 180 Ma during the break-up of Gondwana. However, since these events, tectonic movement has been limited to mid-ocean ridges that completely enclose the sub-continent. As such, topographic relief over southern Africa is largely related to a series of uplift events 20 and 5Ma that resulted in uplift on the eastern seaboard by 250 and 900m respectively (Partridge and Maud 1987). Uplift on the western seaboard was comparatively less, approximately 150m and 100m at 20 and 5Ma respectively, creating a large westward sloping plateau (Figure 1).

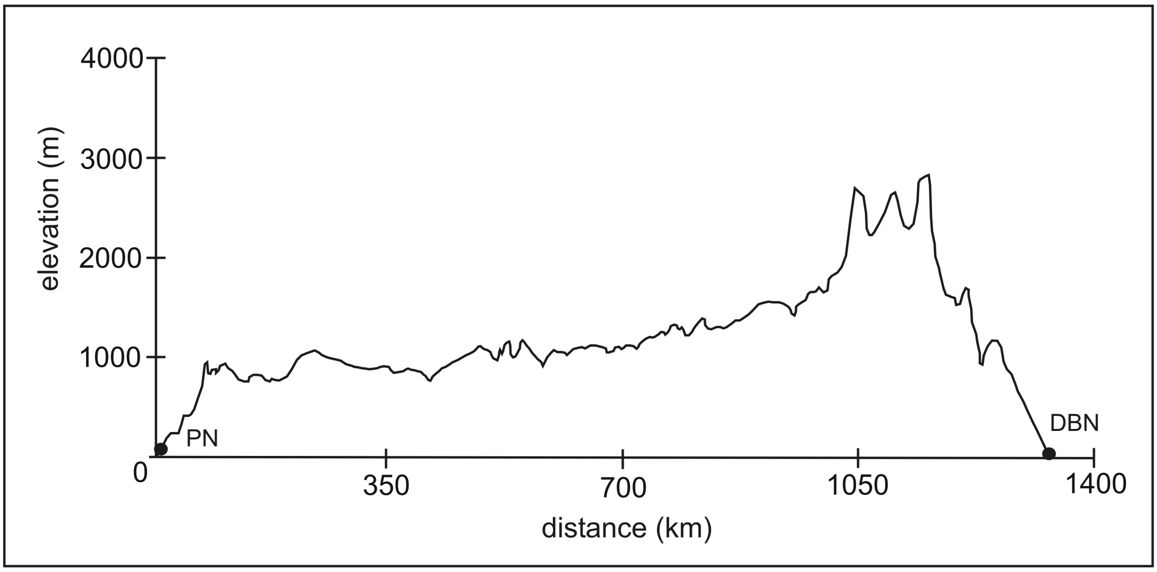

Figure 1: Topographic cross-section of southern Africa, from Port Nolloth (PN) on the west coast, to Durban on the east coast (DBN). Drawn from a 90m resolution Digital Elevation Model provided by NASA (2000).

Glaciation in southern Africa last occurred when the region was located over the South Pole between 286 and 245 Ma (Viljoen and Reimold 1999). However, no topographic features remain of this ancient glacial event as tillite and overlying sediments have subsequently been lithified and eroded. In contrast, many other regions of the globe have recently been shaped by glacial planing of the land surface, resulting in numerous shallow lakes and a flattened topography that is conducive to wetland formation (e.g. Brinson 1993, Galatowitsch and van der Valk 1994).

The lack of tectonic activity and recent glaciation on the sub-continent suggests that the most important aspect in current landscape evolution is that of Miocene uplift, which has resulted in the continent entering a period of long-term incision, with the subsequent development of an extremely well integrated drainage network.

Climatically, the development of wetlands in southern Africa may be considered rather unlikely. Mitsch and Gosselink (2000) place a high emphasis on the existence of a positive water balance when describing the global distribution of wetlands. Southern Africa has a low mean annual rainfall of 486mm/a compared to the global mean for continental areas of 900mm/a (Schulze 1997). In addition, potential evapotranspiration is also high, with the result that much of the region experiences a negative water balance (Figure 2). Thus, local rainfall is generally insufficient to sustain wetlands, which therefore must rely on groundwater and/or surface inputs to some extent. In contrast, many wetlands in northern temperate settings occur as a consequence of a water balance in which precipitation far exceeds potential evapotranspiration. In many cases, wetlands of these regions are caused by a combination of the topographic impact of recent glaciation and a positive water balance (e.g. Galatowitsch and van der Valk 1994).

Considering southern Africa's climate, which generally results in a negative water balance, and its position in a long-term cycle of incision as a result of Miocene uplift, it appears that the region should not be characterized by many wetlands at all. The macro-scale analysis of climate and geomorphology of the region suggests two things about wetland formation on the sub-continent. Firstly, for a wetland to receive sufficient water, it should be located on a drainage network such that it may receive surface and/or groundwater inputs and the water balance may therefore be locally positive for all or part of the year (Tooth and McCarthy 2007). Secondly, it must be located in a local region where incision of the drainage network has been momentarily paused.

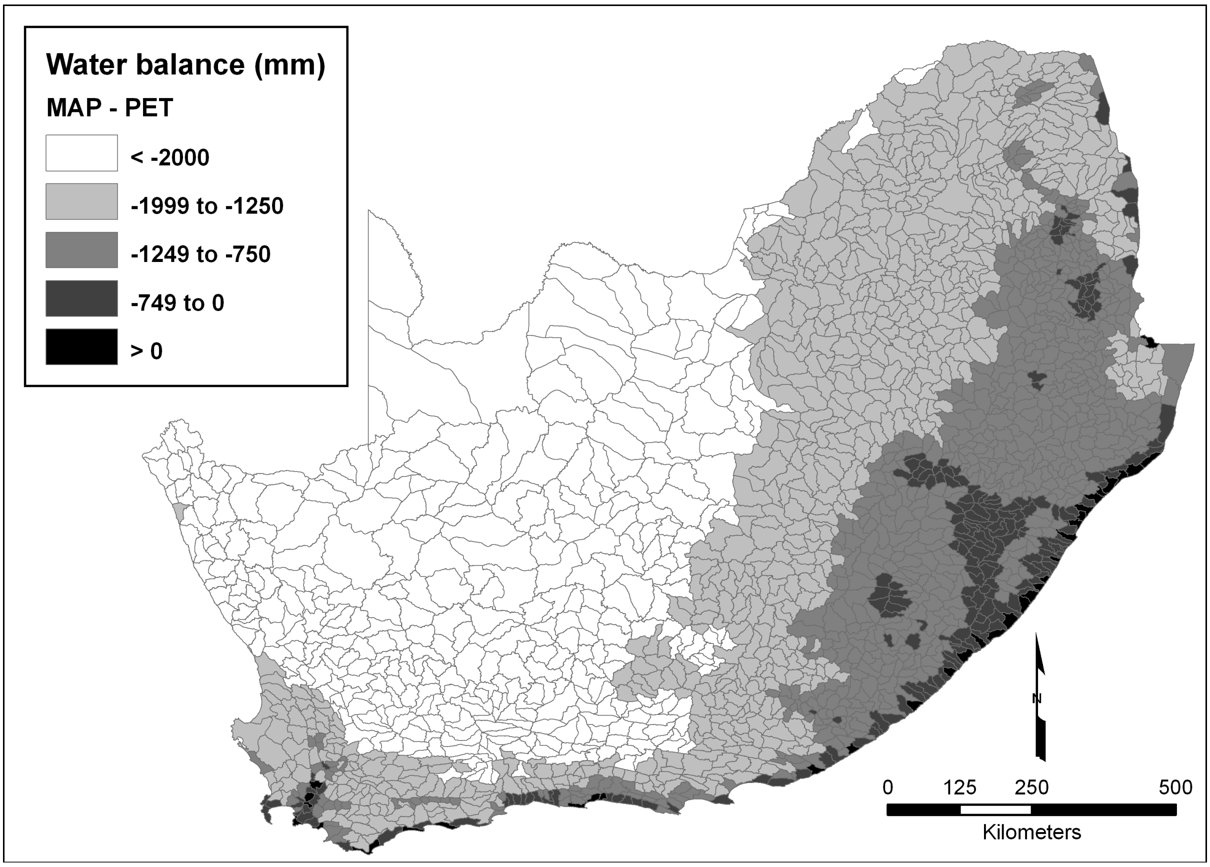

Figure 2: Annual water balance for South African quaternary (forth-order) catchments (mean annual precipitation – mean annual potential evaporation), calculated from Schulze (1997).

The aim of this paper is to further our understanding of the geomorphic origin and evolution of palustrine wetlands that are integrated within the southern African fluvial network. Since the emphasis of this paper is on wetlands that are reliant on surface and/or groundwater inputs, the analysis is focused on Brinson's (1993) valley-bottom and floodplain hydrogeomorphic wetland types, which have been adapted for South Africa by Kotze *et al*. (2005) and Ewart-Smith *et al*. (2006). Under these classification systems, both valley-bottom and floodplain wetlands receive the majority of their water from the movement of water down the drainage network, rather than through local rainfall. In the case of floodplains, water inundates the valley through the overtopping of the main channel. The active migration of the main channel in such a setting, combined with occasional overtopping, results in the formation of characteristic features such as levees, oxbow lakes, scroll bars and alluvial fill. In contrast, valley-bottom wetlands may be channelled or unchanneled, suggesting variation in water inputs from surface channels, surface run-off and groundwater. Where there is a channel in a valley-bottom wetland, features indicative of lateral migration of the channel are absent.

Nevertheless, both floodplain and valley-bottom wetlands are considered zones of deposition characterized by net accumulation of sediment (Kotze *et al*. 2005). It is this occurrence of deposition on a drainage line that is particular to the formation of valley-bottom and floodplain wetlands in southern Africa. Not only does deposition allow the formation of a particular geohydrological setting that is conducive to the development of wetland systems, but it also causes adjustments to the longitudinal profile of the drainage line on which the wetland happens to occur. As sediment accumulates at a local base level, slope is reduced on the upstream end of the deposition node, while conversely, slope is steepened at the wetland system's toe. It is the effect of deposition, and its link to wetland formation, that indicates the usefulness of the geomorphic threshold approach in understanding the geomorphic origin and persistence of wetlands.

2. Methods

34 herbaceous valley-bottom and floodplain wetlands were selected across southern Africa where sufficient field data existed (Figure 3). Of these, 21 were valley-bottom and 13 were floodplain wetlands. The majority fell within KwaZulu-Natal (n=18), a bias somewhat consistent with the countrywide distribution of wetlands. Thereafter, four wetlands were selected in both the Western and Eastern Cape, three wetlands in Mpumalanga and the remainder were located in Gauteng (n=1), Limpopo (n=2) and Botswana (n=2; Figure 3). Of the non-floodplain wetlands, an approximately equal number were incised as not incised.

Data collected for each wetland included valley longitudinal gradient, wetland and catchment size, wetland width and length, whether or not the wetland had undergone erosion through incision, mean annual precipitation and potential evapotranspiration, median annual simulated runoff into the wetland and mean catchment gradient.

Wetland longitudinal gradient was determined in one of several ways: conventional automatic level surveying for small wetlands (cm-level accuracy in the z field), differential GPS using either a local (dm-level accuracy in the z field) or remote (submeter level accuracy in the z field) base station for medium and large wetlands. For some large wetlands, valley longitudinal gradient was calculated from orthophotography (meter-level accuracy in the z field) and published data. Wetland size was determined from orthophotography and existing 1:50 000 topographic maps with some field verification in combination with onscreen digitising using Arcview GIS, or from published sources (Breen *et al*. 1994, WRI 1998, Tooth *et al*. 2004, Le Roux *et al*. 2006). Catchment area was digitised and calculated from 1: 50 000 topographic maps, or was obtained from published sources (Breen *et al*. 1994, WRI 1998, Tooth *et al*. 2004, Le Roux *et al*. 2006). Incision was deemed to have occurred where gullying was clearly evident and was verified by field visits. This definition does not apply to local incision on floodplains where erosion is a component of normal floodplain processes.

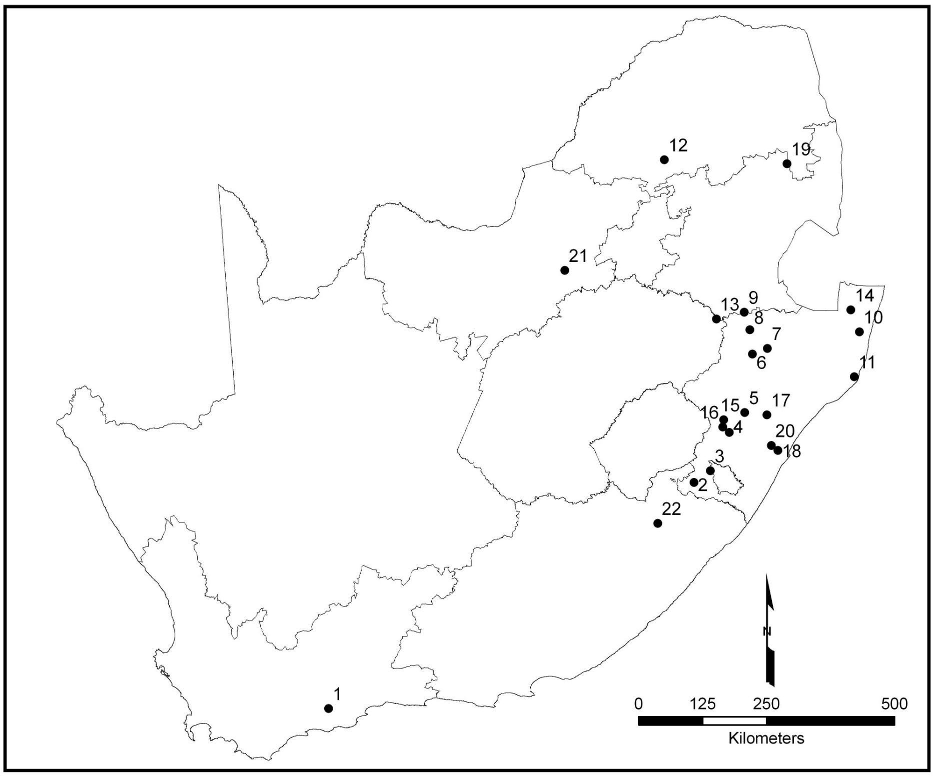

Figure 3: Map of wetland locations (1 = Grootvlei, Klein River, Kruis River and main arm of Goukou wetland, 2 = Mzimvubu, 3 = Ntsikeni and Killarney, 4 = Northern and Southern Mgeni Vlei, 5 = Kruisfontein, 6 = Gladstone Vlei, 7 = Blood River and Lynspruit, 8 = Boshoffsvlei, 9 = Wakkerstroom, 10 = Mkuze, 11 = Mfolozi, 12 = Nylsvlei, 13 = Klip River, 14 = Pongola Floodplain, 15 = Hlatikhulu, 16 = Stillerust, 17 = Mvoti Vlei, 18 = Chelmsford, 19 = Upper and Lower Craigieburn, 20 = Fredville, 21 = Schoonspruit, 22 = Weatherley).

Median annual simulated run-off, mean annual precipitation and potential evaporation were obtained from Schulze (1997). Mean annual catchment run-off or discharge, was calculated from median annual run-off data supplied by Schulze (1997) and calculated catchment area. Mean catchment steepness was determined using the 90m resolution DEM of southern Africa supplied by NASA's Shuttle Radar Topography Mission in 2000.

The geological character of bedrock underlying the wetland was examined, as was the presence of faults.

Preliminary data analyses suggested that the relationship between the wetland longitudinal slope and the independent variables was logarithmic. Prior to analysis, data were transformed to a log scale so that a linear regression could be obtained from the transformed data. A step-wise multiple regression, using a forward selection model in SPSS Version 15.0, was used to establish the relationship between the independent variables and wetland gradient of valley-bottom wetlands and then floodplain wetlands. The forward selection model enters variables sequentially into the model where the first variable entered is the one with the largest positive or negative correlation with the dependent variable, provided the probability of the F value is less than 0.05.

3. Results

Results are presented in three components. The first section assesses what wetland gradients look like in the southern African context and compares wetland gradient with wetland length. The second component analyses factors that affect longitudinal slope on wetlands using a forward step-wise multiple regression. The final section applies the results of the linear regression using the geomorphic threshold approach.

3.1. Longitudinal gradient of some southern African wetlands

The longitudinal profiles of some southern African wetlands are illustrated in Figures 4a-c in decreasing order of length. Figure 4a contains slopes of valley-bottom wetlands, while figure 4b contains slopes of small floodplain wetlands at the same scale. Figure 4c illustrates slopes of floodplain wetlands, with the Schoonspruit valley-bottom wetland included as a scale reference.

From Figure 4a it appears that there is a negative correlation between wetland length and wetland slope. As wetland length increases (sequentially down the figure), slope tends to increase. The steepest wetlands are thus usually the shortest, such as Craigieburn, Fredville, the northern arm of Mgeni Vlei and the Northington arm of Hlatikhulu Vlei. In contrast, longer wetlands tend to be more gently sloped, such as the Schoonspruit wetland, and the main arms of the Goukou and Ntsikeni wetlands. However, it does appear that there is some variability within the slope / length relationship, in that increasing length does not always result in decreased slope and vice versa.

Figures 4a-c also illustrate the changing slope down a drainage line in which a wetland occurs. All wetlands show a distinct change in slope as one enters the head and exits the toe, although this is not always visible in the figures due to the scale. Some wetlands are characterised by steepening while others are regions of slope lowering.

Wetlands have been selected for detailed description in the next section based on factors that relate to their origin. In the case of Stillerust Vlei (Section 3.1.1), the floodplain owes its origin to the rate of erosion of soft Tarkastad Formation shales being arrested by the occurrence of an erosion-resistant dolerite dyke downstream. The Mfolozi Floodplain owes its origin to valley drowning and infilling due to sea level rise during the last 8 000 years (Section 3.1.2). The Futululu wetland, a minor tributary of the Mfolozi Floodplain owes its origin to aggradation on the Mfolozi Floodplain (a trunk dominated trunk-tributary relationship) (Section 3.1.3). The Blood River Vlei is strongly controlled by the input of sediment from eroding sub-catchments that supply sediment to tributary streams (tributary dominated trunk-tributary relationship) (Section 3.1.4).

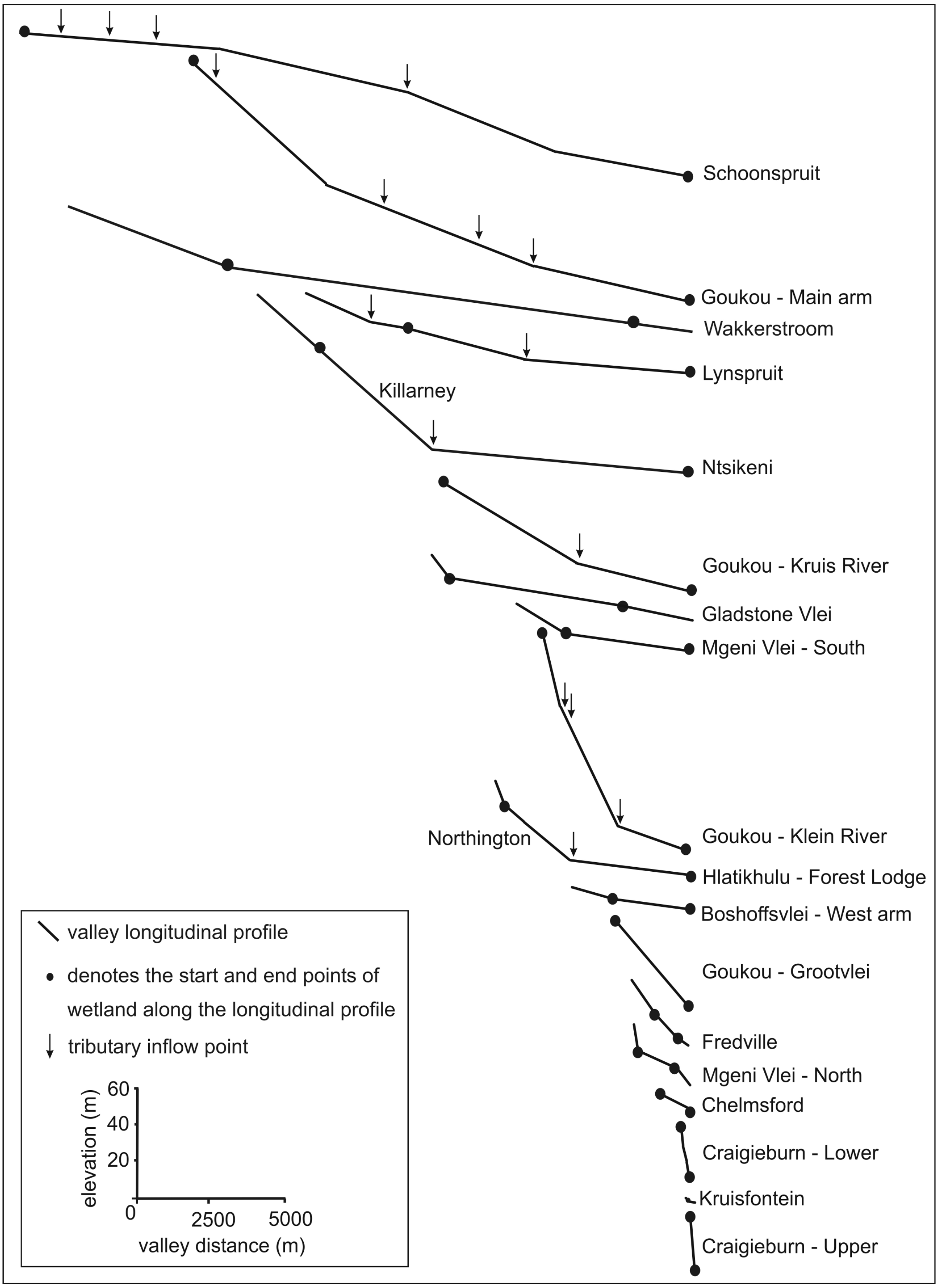

Figure 4a: Slopes of valley-bottom wetlands used in the analysis and their location on the drainage line longitudinal profile.

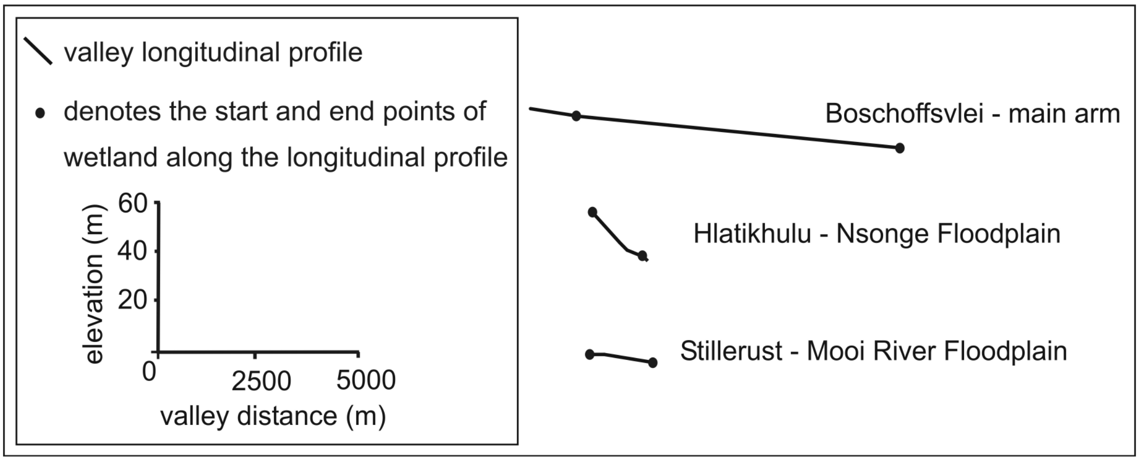

Figure 4b: Slopes of small floodplain wetlands used in the analysis and their location on the drainage line longitudinal profile. Identical scale is used in figure 4a.

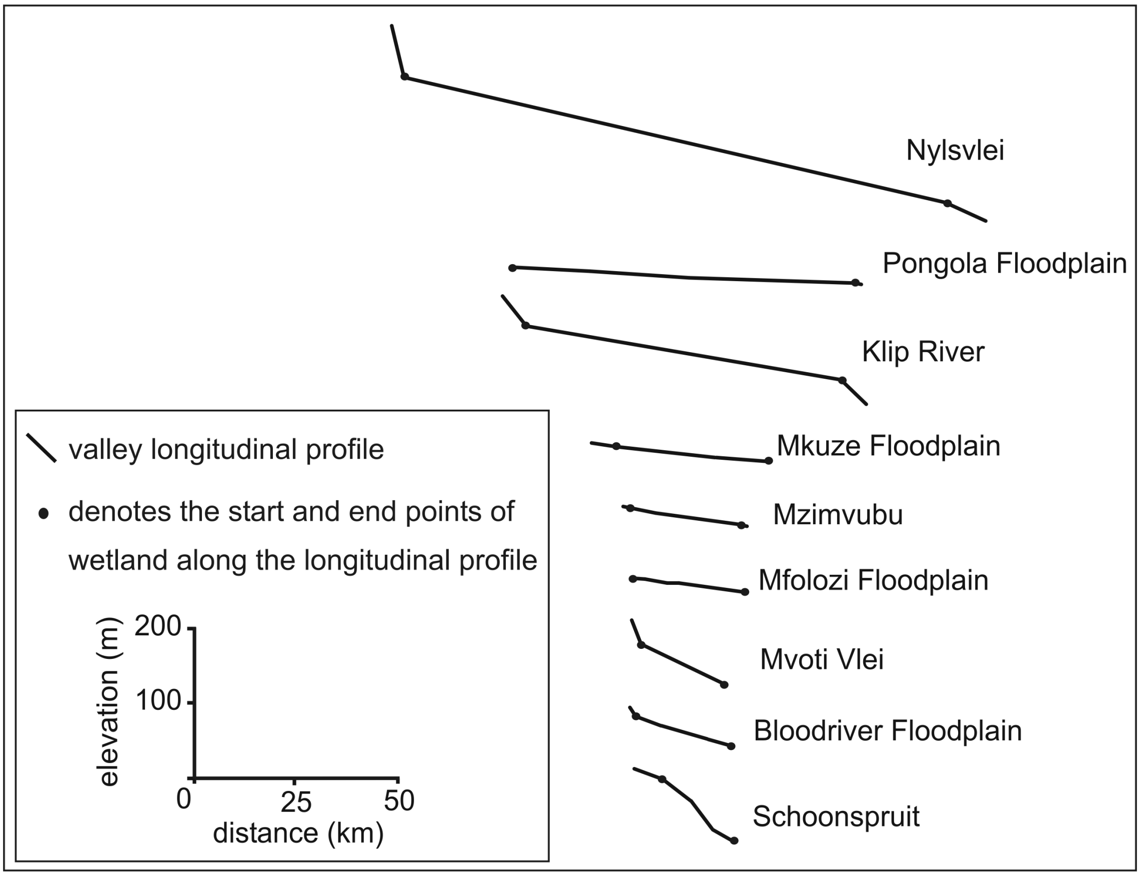

Figure 4c: Slopes of floodplain wetlands, with the Schoonspruit valley-bottom wetland as a scale reference, used in the analysis and their location on the drainage line longitudinal profile.

3.1.1. Stillerust Vlei

Stillerust Vlei is 189ha in extent, being located in the foothills of the KwaZulu-Natal Drakensberg mountains and comprising a floodplain of the Mooi River and an abutting valley-bottom wetland. Two small streams and numerous seeps feed the valley-bottom component of the system. Stillerust Vlei occupies approximately 1.6% of its 11 776ha catchment. Wetland sedimentary fill is mainly clastic, comprising relatively uniform accumulations of clay above basal gravels and weathered Tarkastad Formation sandstone, with silty features of local vertical accretion (levees and an alluvial ridge), and while organic-rich clay occurs in parts of the system that are relatively isolated from clastic sediment input, true peat is limited in depth and extent.

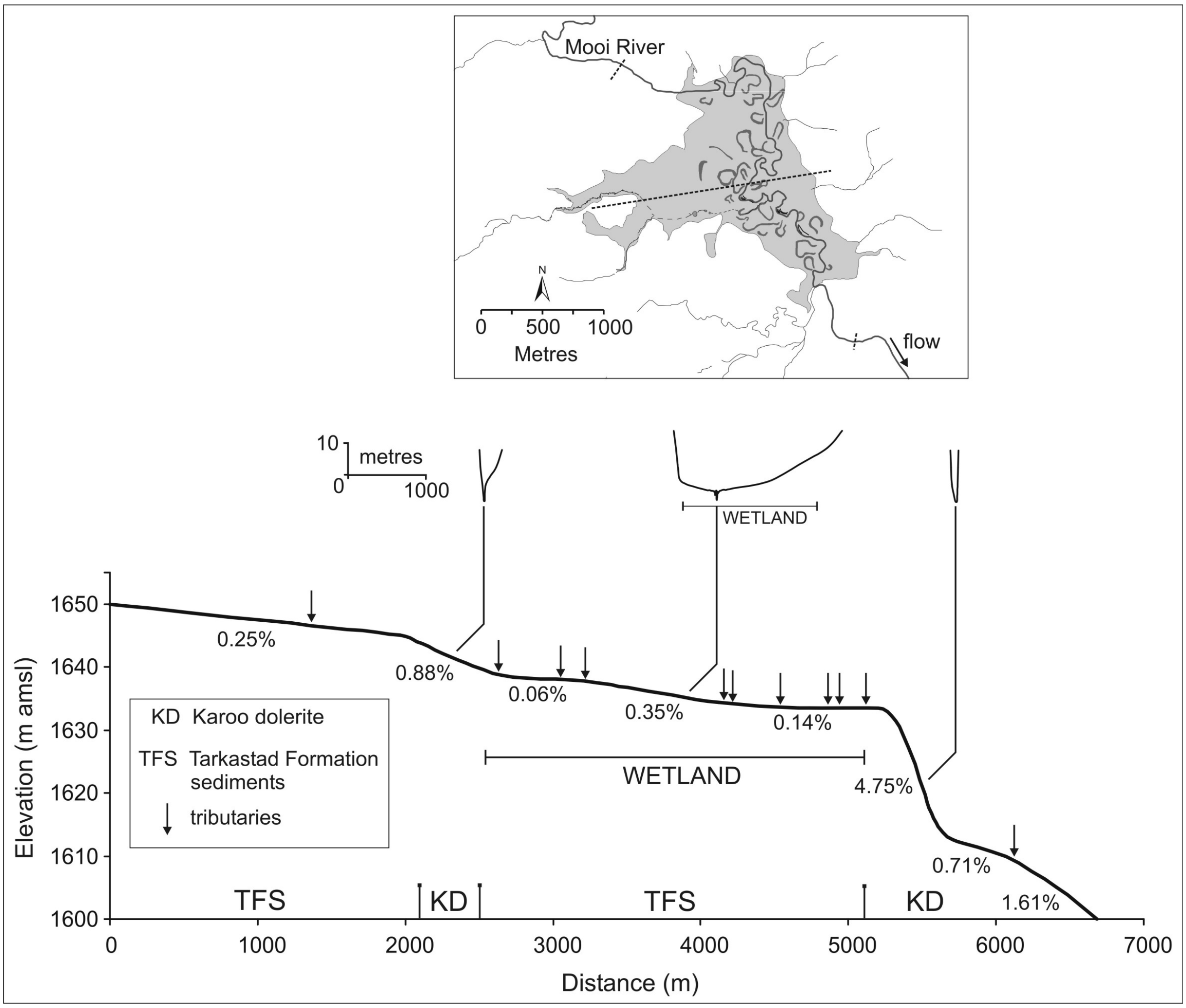

Figure 5: Longitudinal and cross-valley profiles of Stillerust Vlei. Location of tributaries, wetland and changes in bedrock are indicated on the longitudinal profile. The plan view and location of cross-valley profiles are indicated on the inset.

The system supports a diversity of short herbaceous vegetation over a range of saturation zones. The temporary to seasonally saturated outer regions of the wetland are dominated by grass species such as *Aristida junciformis* and *Setaria sphacelata*, while *Cyperus* and *Carex* sedges are common in seasonally to permanently saturated backswamp and valley floor environments. Scattered stands of *Typha capensis*, *Cyperus fastigiatus*, and *Phragmites australis* vie for dominance of flooded depressions and oxbow lakes.

The head and toe of Stillerust Vlei are marked by large dolerite sills, within which the Mooi River valley follows a steep (0.71 – 4.75%), straight and confined course (Figure 5). In contrast, the Mooi River valley hosting Stillerust Vlei (underlain by Tarkastad Formation sediments) is broad and has a low longitudinal slope (0.06 – 0.35%). The Mooi River meanders extensively through this valley, forming numerous oxbow lakes through meander cutoff. A step change in slope from 0.14% to 4.75% at the toe of Stillerust Vlei clearly marks the transition from relatively extensive lateral erosion (bedrock planing) of the weakly cemented Tarkastad Formation sediments underlying the wetland, to relatively slow vertical erosion along joints and fractures within the highly resistant dolerite sill.

3.1.2. The Mfolozi Floodplain

The Mfolozi Floodplain, one of South Africa's largest floodplains at 19 000ha, is situated on the coastal plain of northern KwaZulu-Natal. The region experiences hot summers and cool winters with no frost. Precipitation on the coastal plain at St. Lucia averages 992mm/a, with annual potential evaporation being relatively high at 1805mm. The Mfolozi River has a large catchment, with an area of approximately 1.107 million ha, extending into the hinterland of KwaZulu-Natal.

Two rivers flow through the Mfolozi Floodplain, the larger Mfolozi River in the north, and the minor Msunduze River towards the south, joining to form a combined mouth north of the Maphelane barrier dune. Historically, the Mfolozi and Msunduze Rivers shared a mouth with Lake St. Lucia, which is located just north of the Mfolozi Floodplain (see inset of Figure 6). Currently, the lower third of the floodplain falls under the iSimangeliso Wetland Park and is characterized by a mosaic of herbaceous wetland species such as *Cyperus papyrus*, *Phragmites australis* and *Phragmites* *mauritianus*, as well as small lots of subsistence agriculture. The remainder of the floodplain is under intensive sugar cane cultivation.

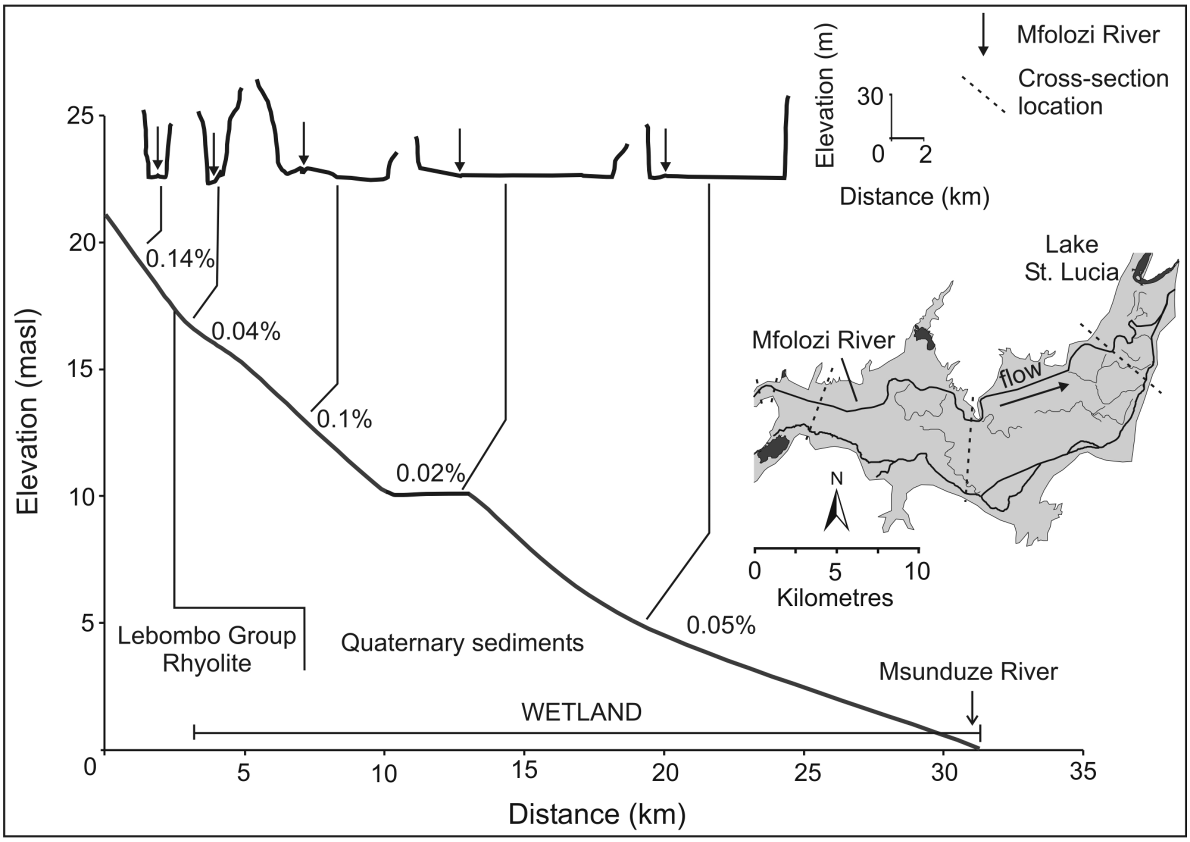

Figure 6: Longitudinal and cross-valley profiles of the Mfolozi Floodplain. Location of tributaries, wetland and changes in bedrock are indicated on the longitudinal profile. The plan view and location of cross-valley profiles are indicated on the inset.

The Mfolozi Floodplain basin contains sediment infill up to 50m deep in the lower reaches. The system is dominated by clastic sedimentation, with peat accumulation restricted to some permanently saturated floodplain margins such as Lake Futululu (Section 3.1.3). There is downstream fining of sediment with sand bodies at the floodplain head and along the channel, with fine silt and clay comprising the majority of floodplain fill.

Above the floodplain, the Mfolozi River follows a meandering course in an incised confined valley (Figure 6, first 2 cross-sections). Upon passing through the Lebombo Mountains of Lebombo Group rhyolite, the valley widens considerably from 915m to over 6km in just 1.15km. As the river enters the floodplain, the slope decreases to

0.04%. Deposition below this region causes a local steepening of the valley's longitudinal profile, with a gradient of 0.1%. However, the mid-floodplain is almost flat, with a decrease in elevation of just 1m over almost 6km (0.02%). The lower floodplain has a steeper gradient of 0.05% with a graded profile to the sea.

3.1.3. The Futululu Wetland

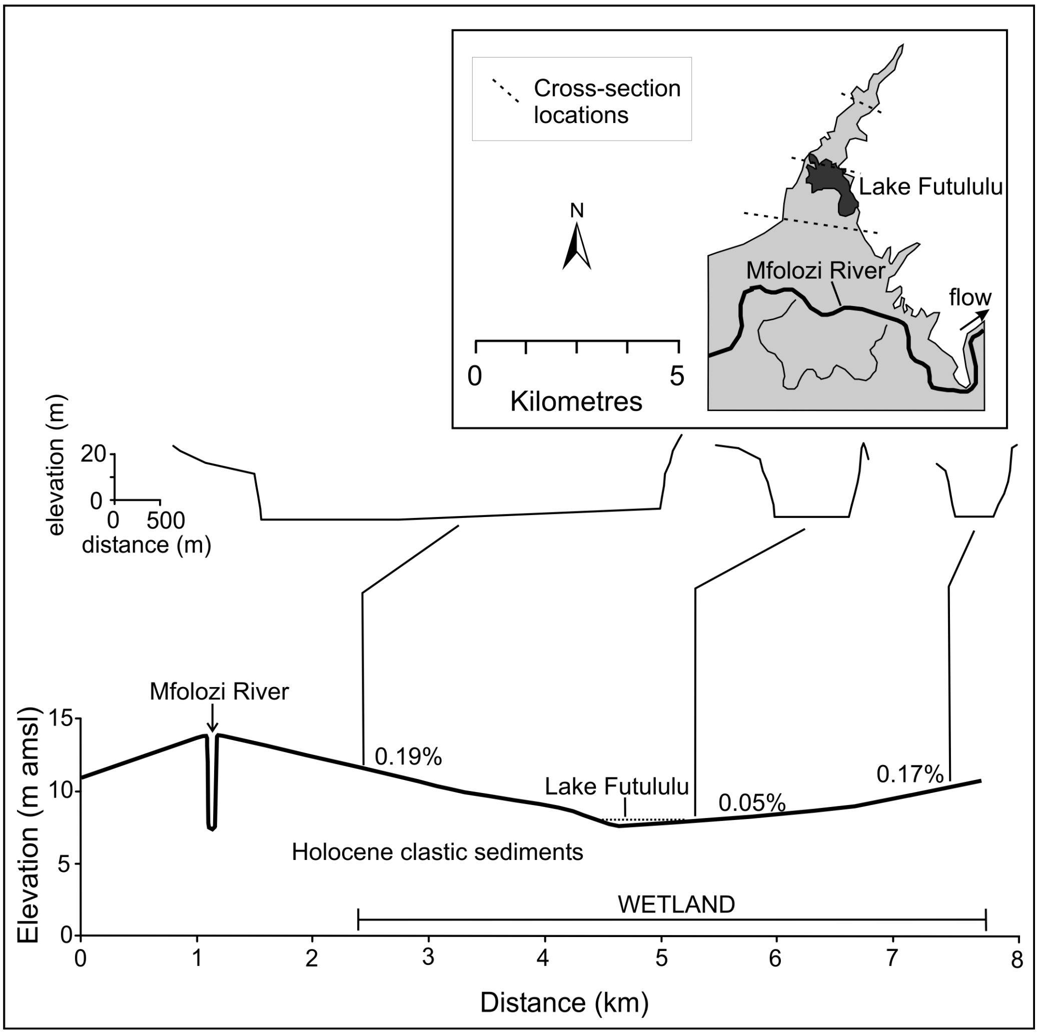

Figure 7: Longitudinal and cross-valley profiles of the Lake Futululu drainage line. Location of tributaries, wetland and changes in bedrock are indicated on the longitudinal profile. The plan view and location of cross-valley profiles are indicated on the inset.

The Futululu wetland is a southward 'flowing' drainage line on the northern margin of the Mfolozi Floodplain, and as such shares the same climatic setting (inset of Figure 7). Hydrogeomorphic characterization of the wetland depends on the scale at which it is considered. It falls within the floodplain boundary of the Mfolozi. However, it comprises a region of open water constituting a lake, while also comprising an upstream region of diffuse flow that characterizes this portion as an unchanneled valley-bottom wetland. The wetland is approximately 225ha in extent.

The Futululu drainage line is dominated by *Cyperus papyrus*, with occasional *Ficus trichopoda* trees throughout the wetland. The southernmost extent of the system has a much higher density of *Ficus trichopoda* trees. The sediment infill is variable, with peat of up to 9.5m thickness occurring towards the region of open water. Towards the north and south of the central peat filled basin, organic accumulation grades into clastic sediments (Chapter 6). The base of the basin infill sequence is marked by sand.

The longitudinal profile of the drainage line towards its intersection with Mfolozi River indicates a large alluvial ridge on which the Mfolozi River flows (Figure 7). The slope between the Mfolozi River and the lowest point on the Futululu drainage line, Lake Futululu, is 0.19%. Thereafter, the drainage line is characterized by two main slopes of 0.17% at the head and 0.05% towards Lake Futululu. Thus, the drainage line experiences a decreasing slope towards the lake. Contrastingly, the width of the drainage line increases downstream.

3.1.4. Blood River Vlei

The Blood River Vlei is located in northern KwaZulu-Natal's interior, just below the Skurweberg plateau. The system is characterised by relatively steep catchment slopes, a low mean annual precipitation (859mm) and a comparatively high potential evapotranspiration rate (1798mm). The Blood River wetland proper, excluding tributary arms, measures 4760ha in extent and accounts for 6.7% of its catchment.

The wetland comprises a central trunk stream (the Blood River) and four main tributary arms that flow from the north and north-east (Lynspruit), the south-west (Spartelspruit and Bloubankspruit) and the east and southeast (Brakfontein; Figure 8). The Lynspruit is the largest of the tributary streams, followed by the Spartelspruit and the Bloubankspruit respectively. All the tributary drainage lines have steeper gradients than the trunk stream, and many of their catchments are eroding.

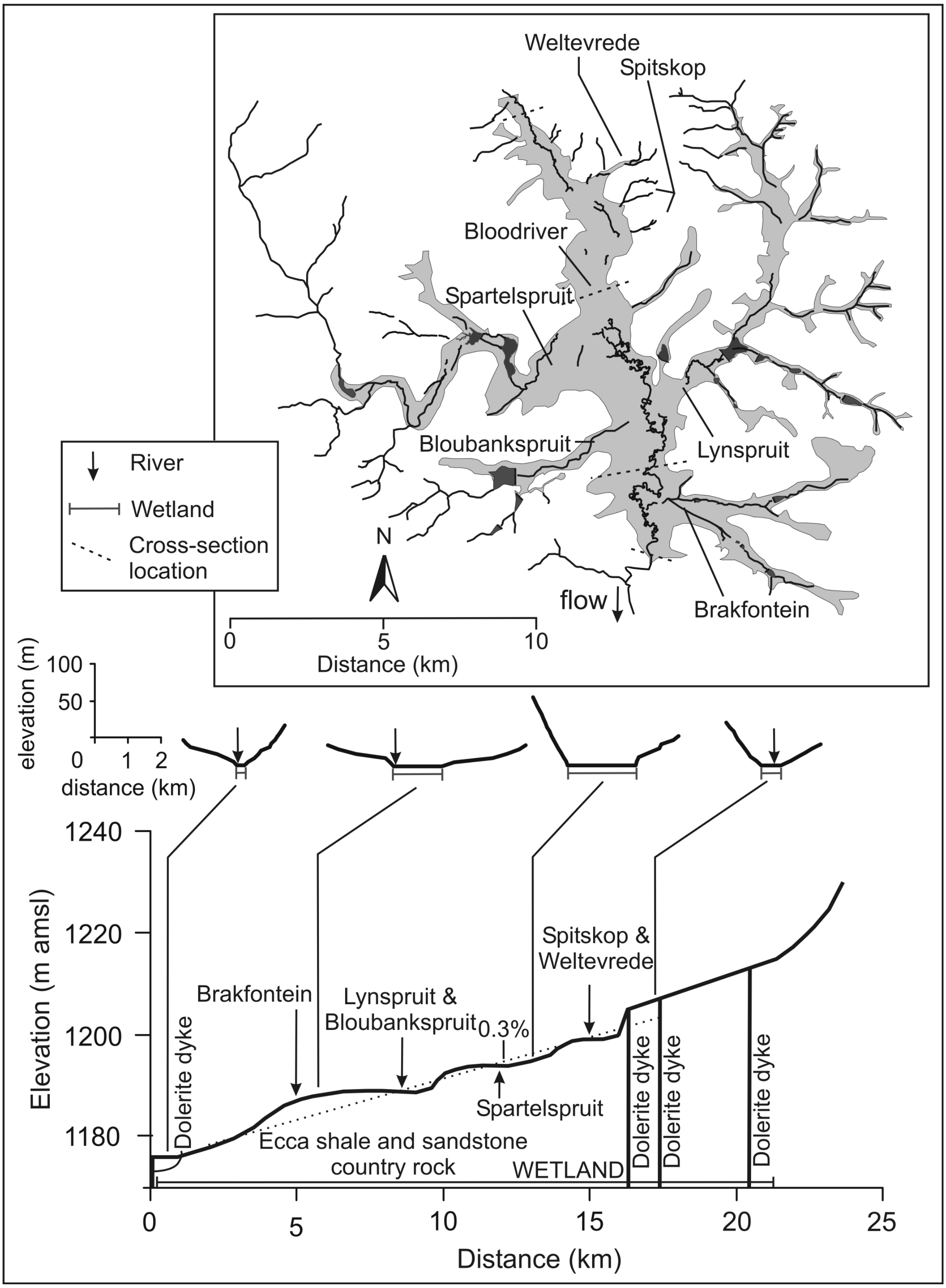

Figure 8: Longitudinal and cross-valley profiles of Blood River Vlei. Location of tributaries, wetland and changes in bedrock are indicated on the longitudinal profile. The plan view and location of cross-valley profiles are indicated on the inset.

The Blood River Vlei consists of three hydrogeomorphic units as defined by Kotze *et al.* (2005). The upper region is a channelled valley-bottom wetland characterised by a channel 2-4m deep, well-developed levees that support terrestrial indigenous tree species, and overbank marshy depressions. The wetland is dominated by grass species indicative of seasonal to temporary flooding.

Below this region the wetland is an unchanneled valley-bottom wetland that has minor channels. Two actively eroding drainage lines, the Spitskop and Weltevrede gullies, drain into the valley-bottom. This Hydrogeomorphic unit is dominated by *Phragmites australis* reed swamp. A deep channel forms in the Blood River Vlei opposite the first of the Spitskop gullies, and again looses confinement below the lat of the gullies. Flow remains diffuse for 3.1km, after which flow becomes confined again in the lower portion of the wetland. The lower 8km of the wetland is a floodplain with a meandering stream, characterised by a highly sinuous stream and numerous oxbow lakes. The floodplain becomes confined to a single channel upon intercepting a dolerite dyke at the wetland toe.

The Blood River flows out of a steep catchment into a valley dominated by Ecca shale and sandstone, crossing three dolerite dykes at the wetland head. This upper region of channelled valley-bottom wetland has a slope of 0.31%. Slope steepens to 0.97% downstream of the dolerite intrusions, the valley widens and flow becomes unconfined. As the Weltevrede and Spitskop gullies drain into the wetland, slope decreases and then steepens, which is co-incident with the deposition of sediment from tributary stream with high sediment yields. In addition, a channel reforms in this region and erodes through the mound of sediment introduced into the Blood River Vlei by the set of eroding tributaries visible in the longitudinal profile. Historically, however, this portion of the wetland was unchanneled. Two additional sediment plugs introduced by tributary streams are visible in the longitudinal profile, deposited respectively by the Spartelspruit and Lynspruit / Bloubankspruit / Brakfontein tributaries. Each plug is characterised by localised slope lowering in an upstream direction, followed by steepening in a downstream direction.

3.2. Valley-bottom wetlands and floodplains

Wetland longitudinal gradient and wetland area reveals that floodplains are generally larger than valley-bottom wetlands, and that floodplains have lower slopes (Figure 12).

Those floodplains that are particularly small include the Hlatikhulu Vlei (Nsonge River) with an area of 403ha and the Stillerust Vlei (Mooi River) with an area of 186ha. It is of interest that with the exception of the Hlatikhulu Vlei (Nsonge River), the simulated mean catchment run-off for floodplains is greater than 20 x 10^6 $m^3.a^{-1}$ (Table 1), whereas in the case of valley-bottom wetlands, with the exception of Wakkerstroom Vlei and Boshoffsvlei (west arm), it is less than this. It appears that large systems are floodplains, small systems are valley-bottom wetlands, and that intermediate sized systems are either floodplains or valley-bottom wetlands.

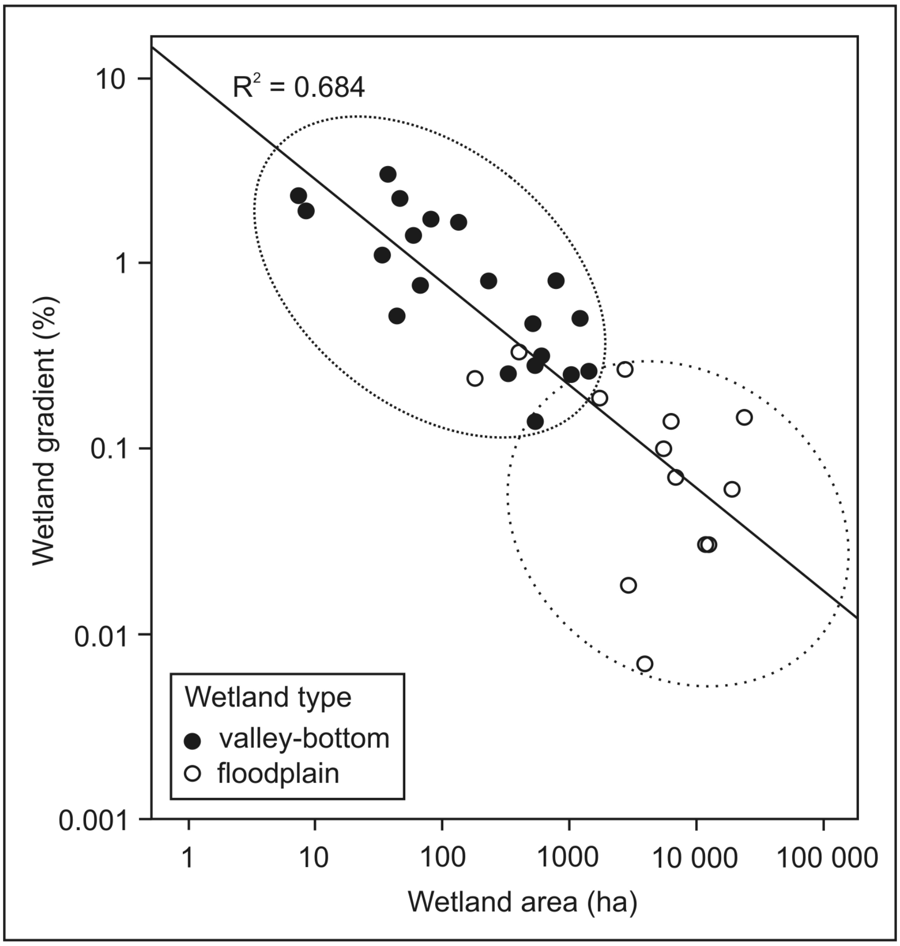

Figure 12: A comparison of wetland gradient and area of valley-bottom and floodplain wetland types.

It therefore seems that valley-bottom and floodplain wetlands fall on a continuum, from steep and small to large and flat, as indicated by the total least squares quadratic regression line, which has a higher co-efficient (R^2 value = 0.684) than either of the individual wetland types. Nevertheless, the two geomorphic types cluster together and can be visually separated based on wetland slope and size (Figure 12).

3.3. Factors affecting wetland longitudinal gradient

Wetland gradient is highly variable, as illustrated in Figures 4a-c. Considering the great variety of wetlands in the landscape and the number of modes of wetland formation, this is hardly surprising. Nevertheless, despite variety, there is a generalised systematic decrease in wetland gradient with wetland length that can be appreciated in Figure 4 a-c. The apparent systematic variation in slope is the subject of this analysis, which sought to highlight which factors were important in determining wetland gradient.

Features of climate and run-off for each catchment are variable (Table 1), with rainfall varying by more than a factor of two, potential evaporation from approximately 1650 to 2400m per annum, run-off depth by a factor of four, catchment run-off by four orders of magnitude and the ratio of mean annual precipitation to mean annual potential evapotranspiration by a factor of three. Thus, the climatic and run-off characteristics of wetlands are extremely variable.

Physical characteristics of wetlands examined in this study are also remarkably variable, with longitudinal slope varying by 2 orders of magnitude, catchment area by 8 orders of magnitude, wetland area by 6 orders of magnitude, wetland length by 3 orders of magnitude and the quaternary catchment gradient in which the wetland is situated by an order of magnitude.

Table 1: Summary of climatic characteristics of wetlands included in the analysis. NA indicates data was unavailable at the time of the analysis.

		Wetland name	Mean annual precipitation (mm)	Potential Evaporation (mm) Mean annual A-pan equivalent	Median Annual Simulated Runoff (mm)	Simulated catchment runoff (million m^3/a)	MAP : PET
Wetland type	Valley-bottom	Killarney	946.7	1531.3	208.5	3.27	1.62
		Ntsikeni - Main arm	946.7	1531.3	208.5	10.00	1.62
		Goukou - Main arm	455.8	1884.7	55.6	11.65	4.13
		Goukou - Grootvlei	455.8	1884.7	55.6	0.22	4.13
		Goukou - Klein River	455.8	1884.7	55.6	0.44	4.13
		Goukou - Kruis River	455.8	1884.7	55.6	4.39	4.13
		Lynspruit	859.3	1798.4	124.6	10.13	2.09
		Craigieburn - Upper	972.3	1933.0	171.6	0.07	1.99
		Craigieburn - Lower	972.3	1933.0	171.6	0.10	1.99
		Kruisfontein	756.2	1676.7	76.0	7.02	2.22
		Fredville	807.9	1654.5	89.7	0.30	2.05
		Hlatikulu - Forest lodge arm	954.3	1677.0	215.4	3.10	1.76
		Hlatikulu - Northington Valley Arm	954.3	1677.0	215.4	1.22	1.76
		Wakkerstroom	914.1	1715.2	167.0	34.70	1.88
		Schoonspruit	560.1	2414.5	24.3	7.90	4.31
		Gladstone	774.3	1811.6	94.3	2.55	2.34
		Mgeni Vlei -south	1008.1	1626.7	210.0	2.30	1.61
		Mgeni Vlei - north	1008.1	1626.7	210.0	0.91	1.61
		Boschoffsvlei - west arm	942.3	1764.2	209.2	30.92	1.87
		Chelmsford	922.8	1644.6	180.1	0.61	1.78
		Weatherley	812.1	1692.6	105.7	0.17	2.08
	Floodplain	Blood River Vlei	859.3	1798.4	124.6	86.09	2.09
		Mkuze	901.1	1696.6	104.9	550.13	1.88
		Nylsvlei	629.3	2250.5	46.6	NA	3.58
		Klip River	668.8	2172.2	21.5	24.51	3.25
		Okavango - Fan	NA	NA	NA	NA	NA
		Okavango - Panhandle	NA	NA	NA	NA	NA
		Mfolozi - average	922.0	1805.8	143.2	1584.94	1.96
		Stillerust	1024.6	1645.4	251.2	29.57	1.61
		Hlatikulu	879.3	1722.3	150.0	6.31	1.96
		Mvoti Vlei	920.5	1686.5	131.7	36.45	1.83
		Mzimvubu	713.5	1627.5	75.1	149.92	2.28
		Pongola Floodplain	619.9	2034.4	56.9	402.91	3.28
		Boschoffsvlei	942.3	1764.2	209.2	112.36	1.87

Table 2: Summary of geomorphic characteristics of wetlands included in the analysis. NA indicates data was unavailable at the time of the analysis.

Wetland type		Wetland name	Wetland Gradient (%)	Catchment area (ha)	Wetland area (ha)	Mean wetland width (m)	Wetland length (km)	wetland area / catchment area (%)	Mean quaternary catchment gradient (%)
	Valley-bottom	Killarney	1.70	1566.2	82.5	78.6	1.5	5.3	17.4
		Ntsikeni - Main arm	0.14	4795.2	552.0	682.2	3.1	11.5	17.4
		Goukou - Main arm	0.81	20952.2	800.0	465.1	15.38	3.8	9.6
		Goukou - Grootvlei	1.90	402.4	8.7	32.3	2.82	2.2	9.6
		Goukou - Klein River	3.00	798.5	38.4	74.3	3.93	4.8	9.6
		Goukou - Kruis River	0.80	7889.2	234.6	259.2	7.8	3.0	9.6
		Lynspruit	0.26	8129.8	1380.1	4877.2	13.0	17.0	7.9
		Craigieburn - Upper	2.30	41.6	7.4	437.9	0.5	17.8	10.2
		Craigieburn - Lower	1.10	58.8	34.1	1000.0	1.2	58.0	10.2
		Kruisfontein	0.52	9234.2	45.5	1459.2	0.3	0.5	10.6
		Fredville	2.22	337.8	47.5	208.2	3.52	14.1	20.9
		Hlatikulu - Forest lodge arm	0.25	1441.0	336.0	817.9	4.1	23.3	10.4
		Hlatikulu- Northington Valley arm	1.67	566.0	135.0	526.6	2.6	23.9	10.4
		Wakkerstroom	0.25	20775.6	1051.2	655.0	16.0	5.1	11.2
		Schoonspruit	0.50	32500.0	1200.0	518.4	23.1	3.7	2.1
		Gladstone	0.28	2700.8	531.4	726.0	7.3	19.7	6.3
		Mgeni Vlei -south	0.47	1095.8	522.8	970.4	5.4	47.7	14.1
		Mgeni Vlei - north	0.75	433.1	66.3	552.8	1.2	15.3	14.1
		Boschoffsvlei - west arm	0.31	14778.3	611.3	1415.0	4.3	4.1	12.1
		Chelmsford	1.40	338.8	59.7	557.6	1.1	17.6	13.3
		Weatherley	2.28	160.0	NA	NA	NA	NA	12.4
	Floodplain	Blood River Vlei	0.14	69092.2	4752.7	2009.6	23.6	6.9	7.9
		Mkuze	0.10	524435.3	5624.7	1413.4	39.8	1.1	10.1
		Nylsvlei	0.15	NA	24000.0	1333.3	180.0	NA	7.1
		Klip River	0.01	114000.0	4000.0	175.4	228.0	3.5	2.9
		Okavango - Fan	0.03	7.2×10^9	12000.0	86600.0	163.3	0.0	NA
		Okavango - Panhandle	0.02	7.2×10^9	3000.0	13000.0	86.6	0.0	NA
		Mfolozi - average	0.06	1106800.0	19000.0	7439.3	25.5	1.7	10.1

		Stillerust	0.24	11770.0	186.0	842.0	2.2	1.6	16.3
		Hlatikulu	0.33	4205.0	403.0	1845.1	2.2	9.6	15.3
		Mvoti Vlei	0.27	27680.1	2800.0	1152.0	24.3	10.1	11.3
		Mzimvubu	0.07	199624.2	6993.5	3851.0	18.2	3.5	7.9
		Pongola Floodplain	0.03	708100.0	13000.0	1604.9	81.0	1.8	10.1
		Boschoffsvlei	0.19	53709.1	1800.0	1766.4	10.2	3.4	12.1

The only variable accepted into the stepwise multiple regression of valley-bottom wetlands, in which wetland longitudinal slope was the independent variable, was wetland area (R=0.823, R^2=0.677, p≤0). None of the other variables met the entry requirements of the model describing wetland slope, largely because they were not independent of wetland area, such that the addition of those variables did not improve the overall performance of the model, or explain any additional variance. The relationship between wetland area and longitudinal slope for valley-bottom wetlands is shown in Figure 13.

The same analysis was run for floodplain wetlands, where wetland longitudinal slope was the independent variable. In the case of floodplain wetlands, two variables were entered into the analysis that explain wetland slope, median annual simulated run-off followed by wetland area as a proportion of catchment area (R=0.961, R^2=0.924, p≤0). An analysis of variance showed that median annual simulated run-off was the more important of the variables, accounting for 85% of the variance, while wetland area as a proportion of catchment area accounted for only an additional 7% of the variance. However, the small number of wetlands used in this analysis (n=13) results in this analysis being less useful than the previous.

3.4. Wetland incision by gullying

Given that floodplain systems are able to redistribute sediment internally in response to local variation in gradient (see Chapter 3), floodplains have been omitted from the comparison of wetlands that are incised with those that are not incised.

The R^2 value for the regression line of the incised and non-incised valley-bottom wetlands was 0.75 and 0.668 respectively (Figure 13). Both R^2 values were significant at the 99.9% confidence level. The analysis indicated that incised wetlands generally had a lower longitudinal slope for a given wetland size than non-incised wetlands.

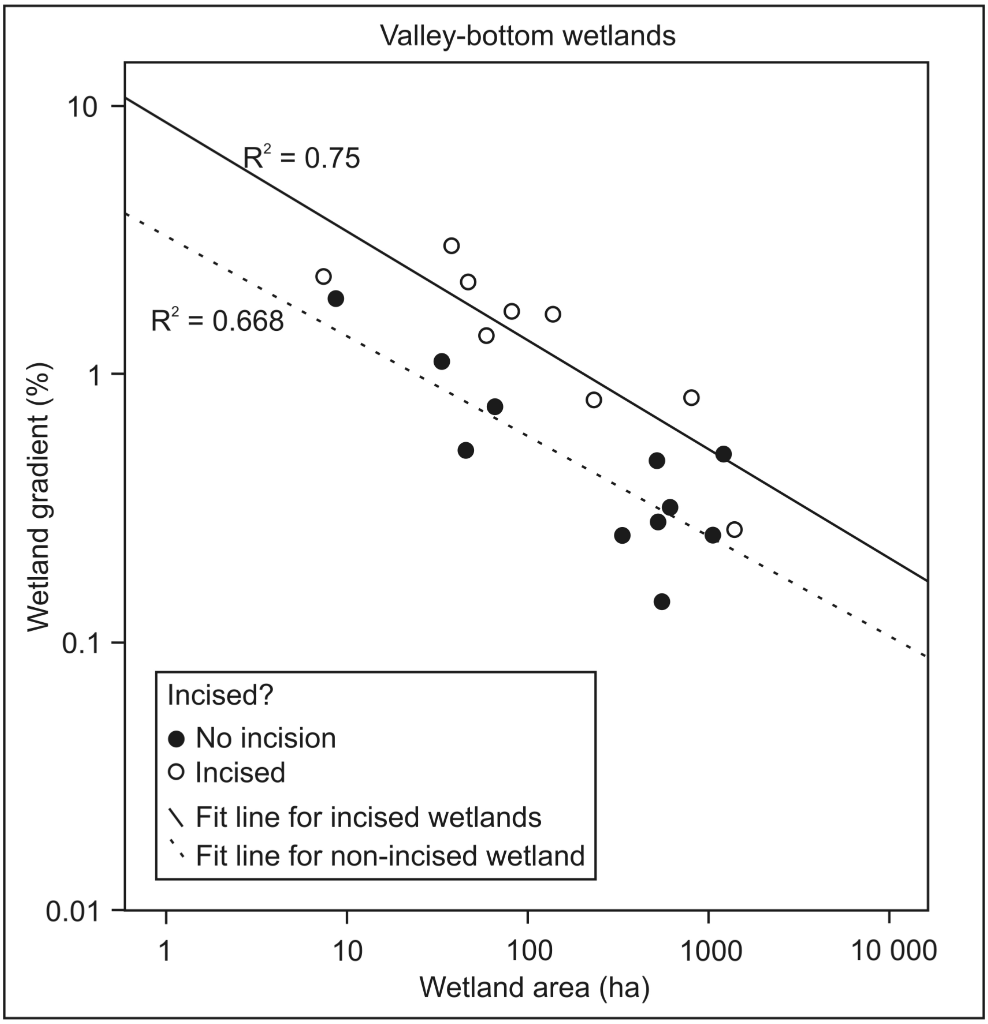

Figure 13: Valley-bottom wetland gradient as a function of wetland area (ha) on a log-log scale. The best fit least squares line for incised and non-incised wetlands is indicated.

4. Discussion

4.1. Control on wetland evolution

The water balance for any continental aquatic ecosystem is summarised by the equation $\Delta V = P + S_i + G_i - E - S_o - G_o$ (Mitsch and Gosselink 2000) where ΔV is changing water volume, P is precipitation, S_i is surface water inputs, G_i is groundwater input, E is evaporation, S_o is surface water outflow and G_o is groundwater outflow. In the southern African context, few wetlands owe their origin to direct precipitation on the wetland as the primary source of water supply, and for this reason there are few true ombrotrophic bogs in the region. In terms of water inputs, some wetlands have groundwater inputs as their primary source of water, such as many of the wetlands on the coastal plain of northern KwaZulu-Natal (McCarthy and Hancox 2000) and wetlands in headwater settings (Ellery *et al.* 2008). Wetlands in dolomitic terrain also owe their origin primarily to groundwater inputs. However, most wetlands in South Africa are linked to the fluvial network and have surface water inputs as their primary water supply – either as diffuse flow or via streams (Chapter 4, Grenfell *et al*. in press, Tooth *et al*. 2004, Costelloe *et al*. 2003, Makaske *et al*. 2002, Morozova and Smith 2000, Ellery *et al*. 2003).

The argument has been made that it is as important to consider outputs or outflows of water, as it is to consider inputs or inflows. Therefore, an important question to ask relates to those factors that slow down and limit outflows (Ellery *et al*. 2008). In this context, local base level, through its control on erosion and interruption of the prevailing processes of sediment transport along drainage lines, may temporarily limit erosion and enhance sediment deposition, thereby influencing the pattern of surface water outflows (Ellery *et al*. 2008, Tooth *et al*. 2004, Grenfell *et al*. in press, Chapter 4). Through its control on erosion and deposition, local base level creates a particular hydrogeomorphic setting that is conducive to the development of wetland systems.

The effect of local base level on deposition is multifaceted. As sediment accumulates at a local base level, slope is reduced on the upstream end of the deposition node, while conversely, slope is steepened at the wetland system's toe. Thus, wetlands usually mark regions of distinct slope change on a drainage line with respect to degree of stream confinement, erosion of bedrock, deposition of sediment and slope of the longitudinal profile.

Consideration of the settings of the 4 wetlands presented in this study, and others used in the analysis, suggests that four main mechanisms of wetland formation are represented in the southern African context of wetlands integrated with the drainage network.

4.1.1. Varying bedrock resistance

The best known model of wetland formation in southern Africa is that of Tooth *et al*. (2004). The model is most appropriate in the region because of the geomorphic evolution of the subcontinent, which has resulted in the gradual superimposition of rivers through long term incision on resistant Karoo dolerite dykes and sills. The Stillerust Vlei is a good example of a wetland that has formed through the superimposition of drainage lines on resistant lithologies (Grenfell *et al*. in press).

Long-term incision on the Little Mooi River's drainage line intercepts a dolerite sill that is more resistant to erosion than the surrounding Karoo sedimentary rock (Figure 5). The interception of the dolerite sill along the drainage line results in the resistant layer acting as a local base level, momentarily pausing incision. While incision on less resistant Tarkastad formation sandstones and siltstones continues downstream of the dolerite, the greater resistance of Karoo dolerite impedes vertical incision upstream. The retardation of vertical incision results in excess river energy, which is expended upstream of the dolerite through the lateral planing of bedrock, increasing valley width above the dolerite sill (Figure 14).

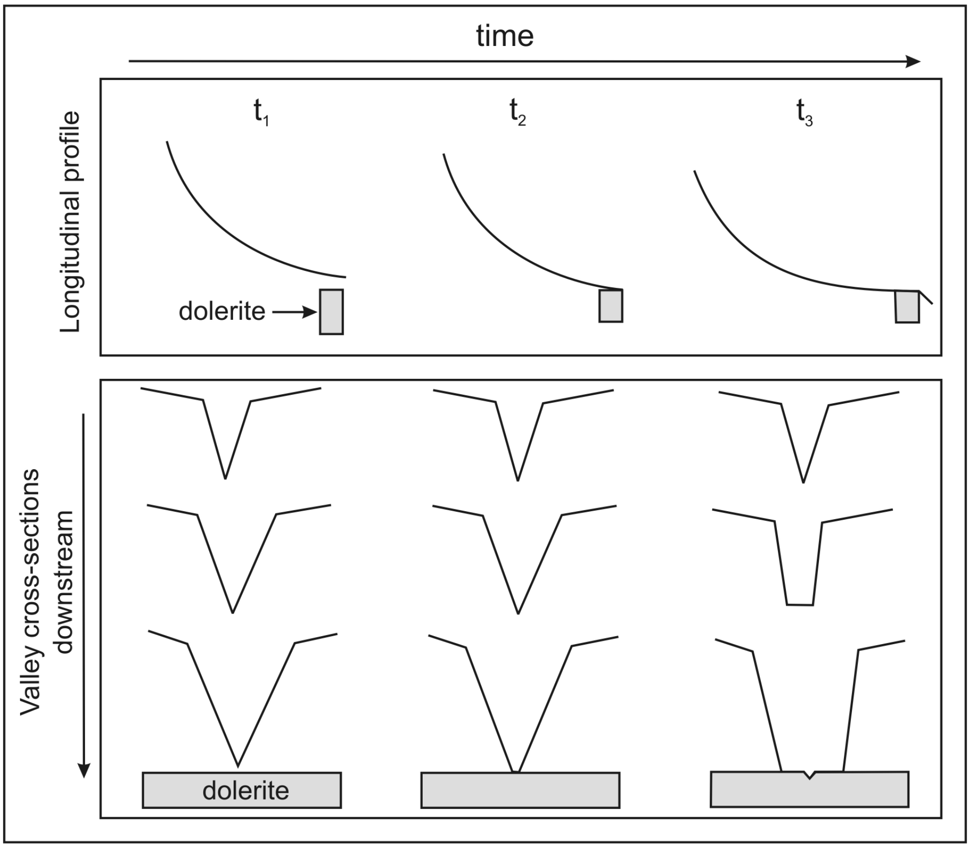

Figure 14: Conceptual model of wetland origin and evolution due to varying bedrock resistance.

Over time ($t_1 – t_3$), the effect of superimposition increases, further enhancing upstream lateral erosion, extending the uniform longitudinal valley gradient further and further upstream and across the valley floor. It is thus the development and maintenance of a local base level that results in the continued expansion of the valley at a slope that is remarkably uniform.

Mixed bedrock-alluvial channels usually characterise wetlands formed in this manner as bedrock is gradually planed and a thin sequence of sediments accumulates temporarily. Gradients steepen along the drainage line as flow passes through the dolerite sill at the toe of the wetland.

While the slope of the bedrock valley above the local base level is low (0.35% in the case of Stillerust Vlei), the slope of the overlying sediment may increase slightly downstream. This occurs because, as streams lose confinement and therefore capacity as they enter the flattened valley above the local base level, sediment is deposited. In time, this results in localized steepening of the longitudinal profile in a downstream direction. Processes of deposition therefore account for the slightly increased slope towards the toe of such wetlands. In some cases, slope steepening may result in portions of the wetland becoming vulnerable to localised incision as slope approaches the geomorphic threshold.

The Tooth *et al*. (2004) model may also be applied to lithologies other than dolerite, provided that there are lithologies of greater resistance than the surrounding country rock. Since most drainage lines in southern Africa are experiencing long term incision as a result of geologically recent uplift, exposure of any variation in resistance along a stream course may result in the formation of a local base level.

In all cases of such drainage line evolution, continued incision eventually erodes away the local base level, ending the momentary still stand in erosion and reinstating downcutting up the drainage line. This eventually removes the wetland from the landscape. Thus, the persistence of wetlands, formed via processes described by the Tooth *et al*. (2004) model, is inextricably linked to the longevity of the resistant outcrop that halts upstream incision.

4.1.2. Sea Level

The Mfolozi Floodplain wetland, located on South Africa's east coast, is a good example of a wetland controlled by variation in sea level (Figure 7). Valley profiles of the wetland suggest the existence of 4 slopes, all of which are dissimilar to that of the upstream confined valley section (gradient = 0.14%).

Upstream of the floodplain wetland, the Mfolozi River is confined to a narrow meandering valley, deepened by incision during the last glacial maximum. As the Mfolozi River flows into the wide alluvial plain loss of confinement results in the formation of a broad alluvial fan, that is prograding onto the sediment-filled valley floor from the floodplain head (see Chapter 5). The lower floodplain is characterized by a valley surface gradient of 0.05%, which is directly controlled by sea level, whereas upstream gradients are related to factors such as the loss of confinement and neotectonics.

The formation of wetlands linked to the fluvial network and located on the coastal plain may be controlled by sea level (Figure 15). During the last glacial maximum approximately 18 000 BP, sea levels dropped to 120m below current levels. Rejuvenation of drainage lines around southern Africa resulted in large-scale vertical incision, the effects of which were most pronounced towards the coastline. Many coastal valleys experienced massive erosion during this period, which ceased as sea levels again began to rise at the end of the last glacial maximum. Sea level reached current levels on the southern African coast approximately 6500 years BP, and again 3880 years BP (Ramsay 1995), leading in the first instance to infilling of the drowned portion of the Mfolozi Valley, and subsequently to sedimentation along the valley (upstream of the drowned portion of the valley) that produces a surprisingly uniform longitudinal slope towards the sea (Figure 15). In the case of the Mfolozi Floodplain, it seems that this latter process is incomplete.

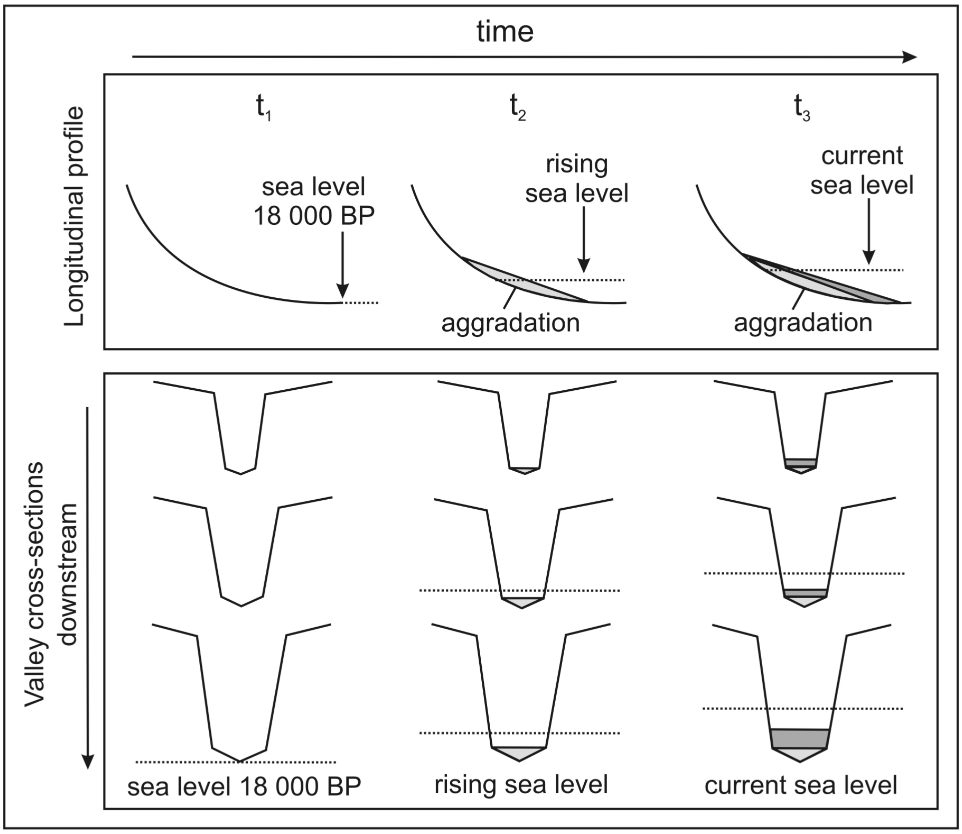

Figure 15: Conceptual model of wetland origin and evolution due to sea level change.

The erosion of such coastal wetlands is controlled by variation in sea level. Should sea level again drop, erosion of the Mfolozi River Floodplain would be unavoidable as the toe of the wetland would be oversteepened in relation to discharge and sediment supply. Unlike the wetlands described by Tooth *et al*. (2004) which are regions of *lateral erosion and valley widening associated with overall degradation*, fluvial wetlands along South Africa's coastline that are controlled by sea level are regions of *sediment infilling in a previously incised valley*. As such, they are aggradational systems.

4.1.3. Trunk-tributary relationships

4.1.3.1. Trunk-dominated wetland formation

Lake Futululu on the Mfolozi Floodplain provides a classic example of tributary impoundment by aggradation on the trunk channel (Figure 7, see Chapter 5). During the last glacial maximum, the coastal region experienced a period of widespread incision, at which time, the Futululu drainage line was a tributary of the larger Mfolozi River, and both streams were actively incising into a bedrock valley. During the erosional phase, the base level of the Futululu Stream would have been determined by aggressive erosion along the Mfolozi River, which has a larger catchment and is a much larger stream (see Brierley and Fryirs 1999). As sea level began to rise, coastal rivers lost capacity to transport sediment as valley gradients were reduced and/or drowned, causing deposition at sea level in previously incising valleys. Over time, deposition propagated upstream, enhanced by the sea level highstand between 4500 and 6000yr BP, when sea level reached +1.5 and +3.5m amsl respectively. The larger catchment of the Mfolozi River, combined with its large sediment supply, resulted in rapid aggradation of the Mfolozi Valley. As sediment accumulated, the local base level of the Futululu Valley was gradually raised, causing a lowering of valley gradient and stream capacity to the point that it no longer transported sediment (Figure 16).

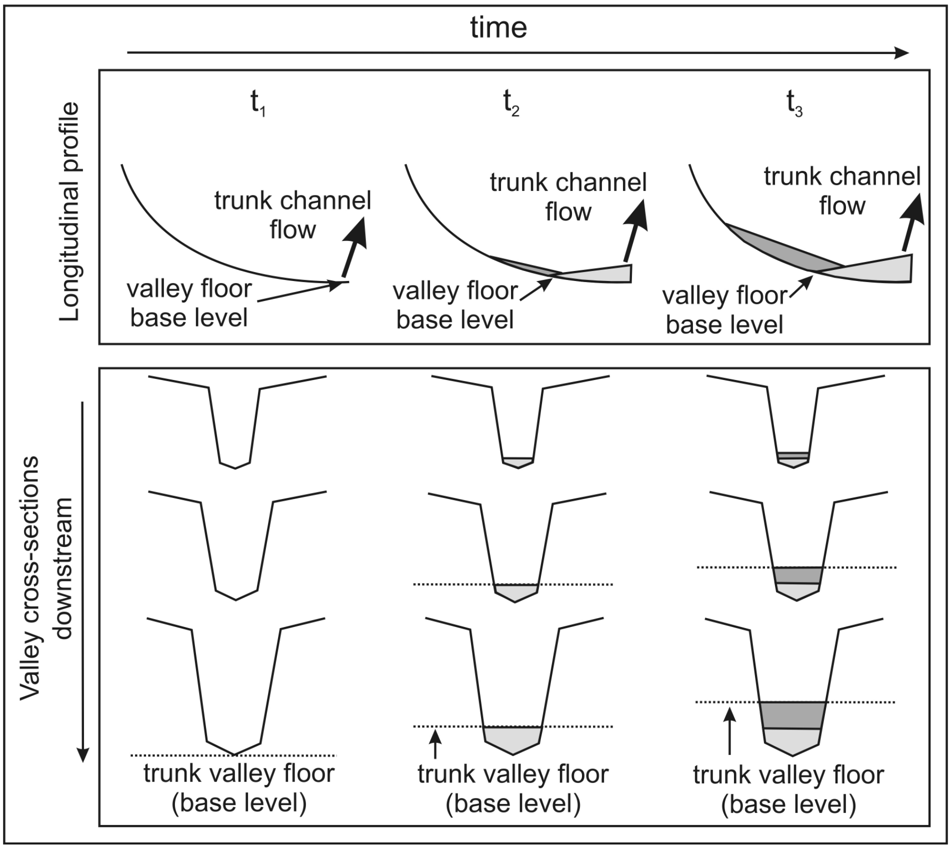

Figure 16: Conceptual model of the origin and evolution of a tributary stream wetland formed by sedimentation in the trunk channel.

Continuing sedimentation along the Mfolozi River Floodplain resulted in ponding of the Futululu Valley, which, in the absence of sufficient clastic sediment supply, became an environment conducive to the formation of peat. As such, the Futululu Valley experienced geomorphic evolution from an incising valley during the last glacial maximum, to deposition of an upwardly fining sand sequence, to peat accumulation 3980yr BP. This later sedimentation took place as a direct consequence of a raised base level caused by sedimentation along the Mfolozi River (Chapter 5).

Wetlands formed through aggradation along trunk channels are not rare, but they have not been described frequently. Grenfell *et al*. (in press) described the formation of a valley-bottom wetland as a result of aggradation on the alluvial ridge of the Little Mooi River. The cut-and-fill valleys of the Wolumla Catchment in New South Wales, Australia, described extensively by Brooks and Brierley (1997), Fryirs and Brierley (1998) and Brierley and Fryirs (1999), may be examples of alluvial ridge impounded valleys. Brierley and Fryirs (1999) described wetlands in the Wolumla Catchment, as a series of valley-bottom wetlands connected to a single large trunk channel. It was found that erosion in the smaller drainage line wetlands was related to the location of the trunk channel in the valley. When drainage lines were directly connected to the trunk channel, incision in the trunk channel propagated headward into the valley-bottom wetlands, eroding them. In cases where they were not directly connected, no erosion occurred. This suggests that some of these wetlands were maintained by the base level of the trunk river's valley sediment fill, while others were maintained by the elevation of the trunk channel bed. In addition, in describing connectivity along drainage lines, Fryirs *et al*. (2007) described tributary disconnection from trunk channels that resulted in the formation of trapped tributary fills. Wetlands formed through aggradation on the trunk channel, causing ponding on a lesser tributary, have also been recorded sparingly in the rock record by Michaelsen *et al*. (2000) and Roberts (2007).

The characteristics of wetlands formed in tributary valleys by trunk aggradation may be somewhat varied. Grenfell *et al*. (in press) and Fryirs *et al*. (2007) indicate how aggradation on the trunk channel has led to the development of valley-bottom wetlands in valleys occupied by tributary streams. In contrast, aggradation on large floodplain systems in Maputaland, such as the Mkuze and Mfolozi Rivers, frequently results in the formation of depression lakes and pans in drowned tributary valleys (McCarthy and Hancox 2000, Chapter 5). It appears that the rate of aggradation on the trunk stream, and thus the creation of accommodation space in relation to sediment supply may be one of the primary controls on the nature of tributary valley-bottom wetlands.

4.1.3.2. Tributary-dominated wetland formation

Trunk-tributary relationships are usually described in the context of the tributary's influence on the trunk channel's behaviour. Steep, sediment-laden tributaries may deposit large quantities of sediment into the trunk channel, causing changes to channel pattern, bedload characteristics and river style (Schumm 2005). Rutherford (2001) described how tributaries of the Glenelg River in Australia supplied large sediment loads to the Glenelg River, partially blocking flow and causing the formation of backwater lakes upstream of each tributary junction. In addition, Rice (1998) considered how tributaries may significantly impact upon bedload texture and as a result, may also impact upon channel form. In particular, large sedimentary inputs were found to redefine the trunk channel's grain-size distribution, hampering downstream maturation. Indeed, even in large rivers, sediment inputs from tributaries have been shown to impact highly on the geomorphology of the trunk channel. In some reaches, the Mississippi River has been shifted to the opposite side of the floodplain floor by tributary sediment inputs (Schumm 2005). This has also been described for the Okavango River in the vicinity of its confluence with the Quito River in Namibia/Angola (Diederichs and Ellery 2001).

The Blood River Vlei provides an example of how tributary sediment inputs may impact upon wetland formation and hydrogeomorphology (Figure 8). At the head of the wetland, the Blood River Vlei is a channelled valley-bottom. However, the character of the wetland is drastically altered as the Weltevrede and Spitskop drainage lines flow into the Blood River valley. The sudden change in gradient, from a steep tributary to an almost flat valley floor, results in a drastic loss of stream capacity. Sediment transported down the eroding tributary drainage lines is deposited as a prominent feature on the valley floor at the base of the tributaries. This aggradation is visible on the longitudinal profile of the Blood River valley as a mound that causes a reduction in wetland longitudinal slope upstream and by slope steepening downstream. The effect of slope lowering on fluvial processes is obvious. The Blood River loses confinement, and flow becomes diffuse as the wetland becomes completely unchanneled. This occurs because sediment loads provided by the tributaries are too great for transport relative to available discharge and slope. Unable to erode and transport the sediment, the Blood River is overwhelmed and ceases to flow. In a sense, the sediment deposited by the tributaries is damming flow on the Blood River.

Recently, erosion in the Spitskop catchment has increased substantially and sediment loads have thus vastly increased. Sediment accumulation in the wetland has caused local aggradation with the effect that incision of a channel has been initiated (Figure 17). However, this is only localised to the region of aggradation below which flow is once again diffuse.

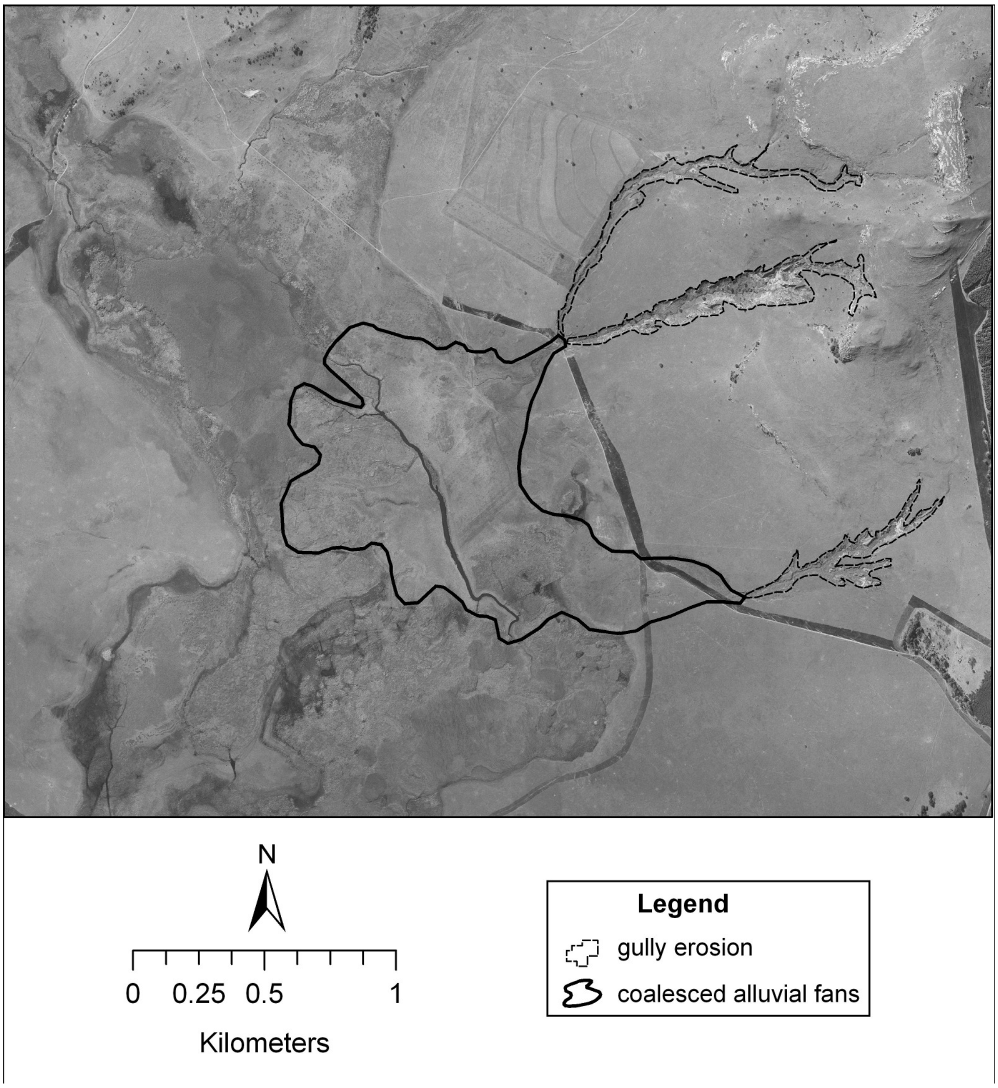

Figure 17: Sediment deposited by tributaries of Blood River Vlei creates a prominent feature in which a channel has formed.

Deposition also occurs at the base of the Lynspruit drainage line. In the case of the Spitskop, it appears that the fluvial features of the Blood River have been covered with sediment. However, in the region of the Lynspruit, a floodplain has developed. This is probably related to the relationship between discharge and sediment supply. In this region of the wetland, increased water inputs and sediment supply allows continued redistribution of the available sediment in a meandering stream.

The Blood River Vlei shows how tributaries from steep catchments may impact upon trunk channels, which, in response to adjustments in slope, affect channel form and process. While the Blood River is ultimately controlled by the location of a dolerite dyke downstream (Begg 1989), the dynamics of the central wetland regions indicated that drowning of a trunk channel by lesser tributary channels might also lead to wetland formation and to variation in wetland characteristics. The mechanisms of such formation are similar to those for the impoundment of a tributary by trunk aggradation. As such, the change in longitudinal profile of the trunk river would be the same as what would occur on a tributary impounded by trunk aggradation, which is conceptualised for trunk-dominated settings in Figure 16. In both the cases of impoundment of a tributary stream and a trunk stream, processes are aggradational and lead to a variety of wetland characteristics, from a shallow lake, to a peat filled unchanneled valley-bottom wetland, to a floodplain. The characteristics depend on the rate of aggradation produced by the primary sediment supplying stream, relative to the flow along the impounded wetland.

In the specific context of a steep catchment adjacent to a broad flat valley, small tributaries may carry large sediment loads onto the valley floor. The combination of a decrease in slope and loss of confinement as the tributary enters the valley floor results in a reduction of sediment carrying capacity. Sediment thus deposited as an alluvial fan, which may over time, prograde further into the trunk stream drainage line, eventually blocking flow if the trunk channel is unable to transport the additional sediment. In some cases, more than one alluvial fan may coalesce, forming an alluvial fan complex that may be even more conducive to drainage line incompetence. Thus, alluvial fans are of huge significance to the development of many wetlands in southern Africa, because of their capacity to flood other, frequently larger, drainage lines with sediment. This results in an upstream reduction in drainage line slope, while aggradation increases slopes downstream of the node of deposition.

The alluvial fans considered in this context refer to those arising from small drainage lines (first or second order streams). Generally, these fans form on pre-existing valley surfaces, such as those carved by larger, trunk channels. Alluvial fan formation is often contemporaneous with wetland and fluvial development, with a similar impact on valley longitudinal profiles and slopes as resistant lithologies. However, alluvial fans differ in that they may actively raise the base level over time, and therefore reduce upstream slopes through aggradation.

4.1.4. Alluvial fan on trunk channel

Unlike their arid counterparts, alluvial fans in temperate to humid environments are frequently associated with wetland formation, with the archetypal example being the Okavango Delta of Botswana (McCarthy *et al*. 1991, McCarthy *et al*. 1992, McCarthy *et al*. 1993, Stanistreet and McCarthy 1993). Alluvial fans occur where streams suddenly lose capacity to transport sediment, usually because of a loss in confinement. As floodwaters spread laterally from a confined channel, the effect of friction is increased and velocity is decreased, resulting in a reduction in both capacity and competence. The resultant geomorphological features are large nodes of sediment with multiple distributaries when in flood, usually of steeper slope than that of the upstream confined channel. As regions of aggradation, alluvial fans tend to reduce upstream gradients, despite their occurrence in steepened zones (Figure 18).

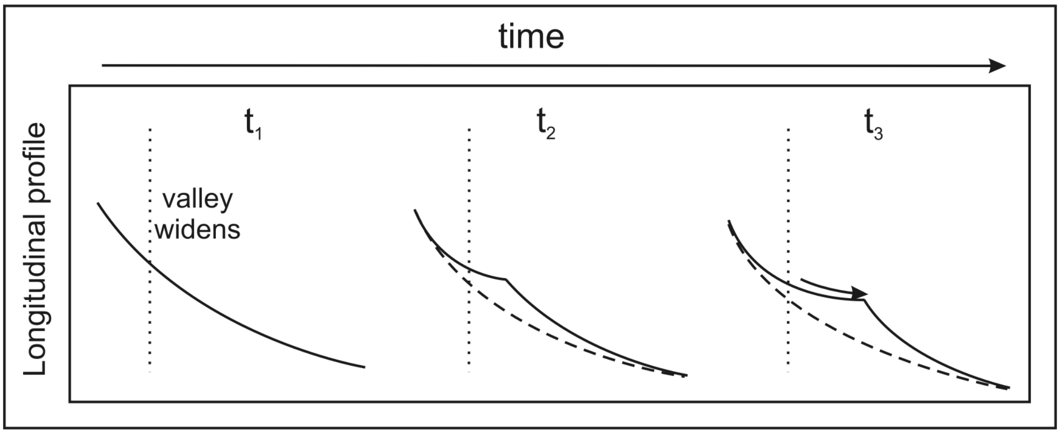

Figure 18: Conceptual model of wetland origin and evolution during alluvial fan formation. The vertical dashed line indicates the point where confinement is lost.

However, alluvial fans may differ substantially in form and process. The Okavango Delta is a large, subaerial fan that is shallowly sloping, highly vegetated and dominated by a series of meandering streams of low sinuosity that diverge at the fan apex (Stanistreet and McCarthy 1993). More typically, alluvial fans occur in dryland settings where they may vary according to the extent to which flow of the main channel is confined on the fan (Hunt and Mabey 1966, Schumm *et al*. 1987). Nevertheless, all alluvial fans are characterised by a loss of flow confinement and a slope steepening as a consequence of sediment deposition.

A more detailed illustration of wetland formation in a trunk alluvial fan setting is revealed by the Mfolozi Floodplain (Figure 6). Upstream of the floodplain, the Mfolozi River flows in a confined valley with a slope of 0.14%. As the River flows out of the more resistant Lebombo Group rhyolites and onto sedimentary rocks, the floodplain drastically widens. Sudden loss of confinement in this region causes large-scale aggradation, particularly during flood events as the Mfolozi River distributes flow and sediment load across the fan. Alluvial fans typically have strikingly uniform slopes in all directions around the fan apex (Gumbricht *et al*. 2001). However, alluvial fan evolution on the Mfolozi River results in two slopes, a slope reduction at the head where active aggradation occurred in the past but no longer occurs, and slope steepening downstream on the actively aggrading face of the alluvial fan. As the river enters the floodplain and flows on the prograding region of the fan, slope is reduced to 0.04%. However, along the actively prograding surface of the fan, slope increases to 0.1%.

4.2. The origin of gradient in valley-bottom wetlands

The origin and evolution of all wetlands is tied to a combination of two main factors, climate and geomorphology, which in turn impact upon the later stage gradient of wetland systems. In terms of climatic control, mean annual rainfall or run-off is the most frequently described variable (e.g. Zucca *et al*. 2006, Morgan *et al*. 2003, Vandaele *et al*. 1996), and is essentially related in Schumm's (1979) threshold concept using the proxy of catchment size as an indicator of catchment discharge. The current study found that in southern Africa, wetland size was a stronger indicator of valley-bottom wetland gradient than catchment area. Considering the wide variety of ways in which wetlands may form along drainage lines, the strength of the relationship between slope and wetland size appears remarkable.

It is suggested here that examination of wetland longitudinal slope offers a useful framework for interrogating wetland evolution, particularly in examining human impacts on these systems. Wetland size, the main predictor of wetland slope, is a function of both wetland length and width. As a wetland forms in the landscape, for example through erosion along a stream that is superimposed on a resistant local geological feature, the stream will rapidly erode its bed upstream of the resistant feature to a gradient that matches sediment supply and discharge such that further deepening of the valley is not possible. Any excess energy may be used to laterally plane the valley, leading to wetland formation. The length and width of the valley thus produced, reflects wetland size, and is dependent upon stream discharge. Larger rivers will erode longer and broader valleys even as the downstream local base level adjusts slowly through erosion or deposition. However, not all wetland systems may be able to reach this finite length of upstream adjustment. This is because there may be an upstream control that inhibits adjustment of the longitudinal profile, such as at Stillerust Vlei, where the upper and lower boundaries of the floodplain are marked by dolerite sills (Grenfell *et al*. in press). Nevertheless, in all of Tooth *et al*.'s (2004) wetland systems, where length is inhibited by either an upstream control, or by virtue of the long profile reaching equilibrium, excess energy is used to laterally plane the existing valley above the local base level. In contrast to a situation of having resistant lithologies controlling valley morphology, where a stream loses competence due to the development of a local base level, sediment is deposited as a function of the loss of transport capacity in a stream, increasing the areal extent of wetland development.

Data presented in this study suggests that wetland longitudinal slope is determined by wetland size, irrespective of the mode of origin of the wetland (aggradational or degradational setting). It seems ultimately that wetland size is related in some way to a feature of flow, but it is not possible based on current data to say what this is since discharge, mean annual run-off, simulated discharge and catchment size were all disregarded as indicators of wetland gradient. Nevertheless, it is likely that geomorphic inheritance is a possible basis for the lack of correlation between these factors and wetland gradient. Current valleys are partially indicative of previous stream power regimes, forming over long time periods during which climate has changed. Indicators of current climates and resulting stream discharges would not be good predictors of wetland gradient if the wetland had formed in a different climate to that present today. Current mean annual run-off estimates and discharges do not necessarily correspond to previous climate regimes under which a wetland basin may have formed. The use of catchment size as a predictor of wetland gradient is also complicated by other factors, since catchment gradient, rock and/or soil types also influence discharge. In addition, the strong rainfall gradient across southern Africa means that basin size is not a linear measure of stream discharge. Overall, despite the wide disparity in modes of wetland formation, the longitudinal gradient attained in each system was best correlated with wetland area, which is likely to be a long-term measure of stream discharge.

Figures 4a-c suggest that the larger the wetland, the lower the wetland gradient, a relationship supported by Figure 12. Thus, larger wetlands of higher stream power tend to aggrade or laterally erode flatter valleys than do smaller wetlands of lower stream power. In this context, large, steep wetlands and small, gentle wetlands are a geomorphic improbability. This is consistent with the traditional Davisian fluvial model (Davis 1902). In this model, the upper reaches of a fluvial longitudinal profile are steep, whereas towards sea level, gradients become gentler. In terms of discharge and valley gradient, wetlands with smaller catchments are generally located in first or second order catchments, and thus can be compared with the upper reaches of an ideal longitudinal profile, while more gently sloped systems, with larger catchments, are located towards the lower section of the stream profile.

4.3. Geomorphic thresholds and wetland gradient

Considering the complexity in arriving at a particular wetland gradient, there is a striking relationship between incised and non-incised valley-bottom wetland gradients relative to wetland area (Figure 13). This suggests that despite the extent of variation in environmental control in the development of wetland gradients, inherent geomorphic control exercises the greatest influence on the occurrence of wetland incision.

Wetlands with a high slope for their size are vulnerable to erosion. Non-incised wetlands between the two best-fit lines in Figure 13 are likely to be vulnerable to erosion, since this region represents a zone of critical slope threshold. The existence of a zone of vulnerability, where there is a possibility of a wetland being either incised or stable, suggests that thresholds are gradational rather than sudden, as suggested by Begin and Schumm (1984). As slope increases for a given size, the probability of a wetland being incised increases.

A feature to emerge from this study is that floodplain wetlands tend to be large (> 1800 ha), while valley-bottom wetlands tend to be small (< 190 ha). However, between 190 and 1800 ha, there is a zone of overlap where both valley-bottom and floodplain wetlands may be represented. In this region, the Hlatikhulu (Nsonge River) and Stillerust Vlei (Mooi River) Floodplains may be considered unusually small. In contrast, the Lynspruit (1380ha in size, 0.26% slope), Schoonspruit (1200ha, 0.5%) and Wakkerstroom (1051 ha, 0.25%) valley-bottom wetlands are unusually large for valley-bottom wetlands.

Floodplain wetlands are usually generated where stream discharges and capacity to transport sediment are comparatively high. In these wetlands, while sediment availability is important for the functioning of the wetland, it results in localised steepening that may lead to meander cut-offs or channel avulsion. However, in floodplain settings, incision seldom seems to result in degradation of an entire floodplain system and the system may continually adjust internally to variations in discharge and sediment supply. For instance, research on the Klip River Floodplain has shown that climatic changes have had no impact on floodplain processes such as the rate of meander migration and avulsion (Tooth *et a*l. 2007). As such, in floodplain systems, excessive steepening along the longitudinal gradient can be either removed or absorbed by adjustments to the channel pattern or localized erosion, as is the case in the Mfolozi River (Chapter 3). Therefore, deposition and erosion in floodplain wetlands are components of the system, and the geomorphic threshold approach is generally less useful for floodplain than valley-bottom wetlands because they are inherently stable for a specific set of environmental variables.

As wetlands are formed during periods of deposition, and deposition leads to oversteepening in a downstream direction, wetlands that appear in the landscape today may be poised in geomorphic time to enter a period of incision. The slope at which a wetland becomes critically steep and vulnerable to adjustment and erosion depends on the slope of the wetland relative to its size. Since natural systems have an inherent tendency to establish a dynamic equilibrium due to negative feedbacks (Ahnert 1994), wetlands can be seen as components of a larger process-response system in a state of dynamic equilibrium, whereby wetlands are not the endpoint of landscape development, but rather a temporal step in a recurring cycle of erosion and deposition.

4.4. Implications for wetland management

South Africa currently invests approximately R 70 million per year in a poverty alleviation project called ‘Working for Wetlands”. The project aims to recover wetland health and ecosystem services by returning affected wetlands to as close to their natural state as possible, while simultaneously employing people in economically undeveloped areas. Understanding of the threshold approach, and its bearing on wetland vulnerability to incision, has implications for rehabilitation planning and prioritisation.

While many wetlands are controlled by geomorphic thresholds, and may therefore be inherently prone to erosion, incision is frequently accelerated by human activities. Since gradients tend to steepen with time, changes in a wetland's catchment that either increases the rate of steepening through aggradation (e.g. bare land increases sediment supply), or increases run-off (e.g. through increased hardened surfaces) may hasten a wetland's evolution towards its geomorphic threshold. In these cases, it is a combination of natural evolution and human intervention that causes wetland incision. Contrastingly, incision may also be completely natural, occurring by virtue of the crossing of an inherent geomorphic threshold. It could be argued that rehabilitation of wetlands that are eroding due to crossing a geomorphic threshold is futile and unreasonable since it involves interrupting a natural process of wetland erosion and creation. However, current losses of wetland systems worldwide exceed 50% (Zedler and Kercher 2005), a figure that is likely echoed in southern Africa (Kotze and Breen 1994). At such rates of loss, wetlands may be rehabilitated irrespective of what the underlying causes of erosion may be, because the loss in ecosystem services is so severe. Furthermore, the length of time involved in the creation of replacement wetlands is generally considered too long in terms of a human lifetime, necessitating intervention to restore ecosystem services that contribute positively to human well-being. Nevertheless, understanding the processes of wetland incision provides information on which wetlands are most susceptible to incision, allowing cost-effective prioritisation and the development of cautionary catchment management strategies, as advocated by Patton and Schumm (1975).

Plotting a wetland's areal extent and gradient in Figure 13 can provide an indication of wetland vulnerability to incision. Vulnerability to incision has both cost and intervention success implications. Incised wetlands of lower vulnerability (wetlands in the zone of vulnerability located more closely to the non-incised best fit line) are likely to be less costly to rehabilitate. In addition, such rehabilitation is more likely to be successful in the long term. However, wetlands of high vulnerability that are well above the geomorphic threshold for their size will be relatively costly to repair, and there is a greater possibility of failure.

Cost, however, as Schumm (1994) suggests, is also related to the timing of intervention. Generally, stabilization of gullies will be most effective and economical during early phases of degradation. Thus, by plotting gradients and wetland areas, one can investigate which currently non-incised wetlands are most vulnerable to future incision. These wetlands could be monitored more closely, and their catchments managed in a precautionary manner, resulting in a more proactive approach to wetland rehabilitation and management, since one can intervene before an erosional problem has become impractical. In this way, the geomorphic threshold concept may be used to prioritise which wetlands should be most closely monitored, thus reducing the cost of rehabilitation should it be required. Wetlands furthest from a geomorphic threshold, consequently deliver the highest returns on investment through rehabilitation. This is particularly useful if budgets are limited, and specific wetlands have to be prioritised.

5. Conclusion

The occurrence of wetlands in our landscape may be seen as a rarity in a geomorphic context, considering the extent of incision on the sub-continent. As depositional or non-erosional features, valley-bottom wetlands are poised in time to reach a critical level of oversteepening that will ultimately result in erosion. However, the rate at which the slope threshold is reached may be impacted upon by numerous factors that influence catchment water and sediment discharges. By increasing the rate of sediment supply, one may enhance natural rates of system aggradation and thus bring the system more rapidly to a slope threshold.

This may be conceptualised in Figure 19, where position A represents the starting size and slope of a valley-bottom wetland, and B represents the position of the same wetland following an increase in aggradation rate and therefore increased slope. The removal of natural catchment vegetation could therefore cause wetland incision by causing catchment erosion and enhancing sediment supply, as occurred on Wolumla Creek, Australia (Brierley and Fryirs 1999).

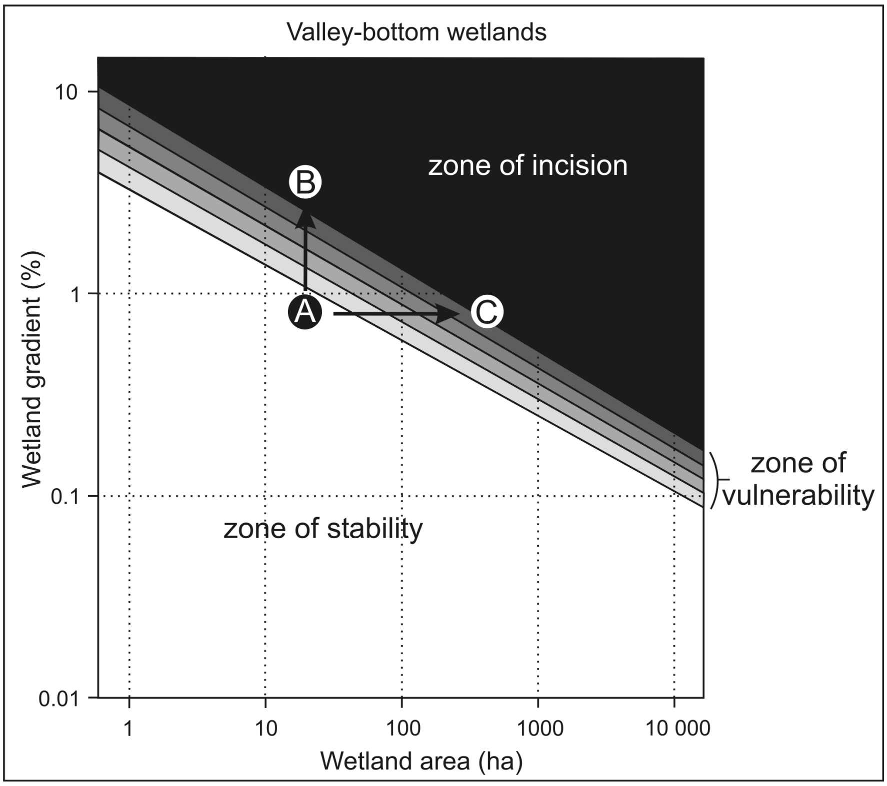

Figure 19: Zones of stability, vulnerability and incision for southern African valley-bottom wetlands, A, B and C are referred to in the text.

However, the most likely manner of causing the slope threshold to be exceeded is by increasing the size of catchment discharge. Since wetland area is an indirect measure of catchment discharge, increasing discharge into the wetland is analogous of increasing wetland area and thus moving a wetland from position A on Figure 19, to position C. Changes in catchment land use, such as from agriculture to urban, may cause wetland incision by creating run-off discharges that are too great for the wetland's slope. Of greater concern is the impact of climate change on wetlands. The climate since the last Ice Age has become warmer and wetter, which is equivalent to moving from position A to C in Figure 19. Future climate change, if it increases run-off, will have a similar effect.

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Chapter 3. Variable rivers: a new geomorphology? Case of the Mfolozi River, South Africa.

Abstract

Stream flow in southern Africa and Australia is more variable than in other continental areas, even when compared to regions in the same biome. Precipitation in the catchment of the Mfolozi River in northern KwaZulu-Natal, South Africa, is more variable than other regions of the globe. Similarly, the co-efficient of variation for inter-annual stream flow of the Mfolozi River measured 79%. Stream variability in the Mfolozi may be linked to multiple factors including a large catchment size, a seasonal climate of a dry winter and wet summer, evergreen vegetation in the catchment, variable precipitation and the occurrence of regionally pervasive climatic oscillations. This research aimed to address how stream flow variability impacted upon sediment transport and thus, geomorphology. A flow frequency analysis showed that stream flow is skewed towards low values, with a number of extremely large flood events occurring as outliers on the distribution. The impact of variability on sediment transport occurred at the intra- and inter-annual scale. Analysis of mean monthly sediment concentration and discharge showed a hysteresis effect, such that sediment concentration peaked prior to discharge in the early wet season. During the late wet season, peak discharges often had unexpectedly low sediment concentrations. Evidence suggested the existence of long-term hysteresis that may be related to decadal-scale climatic oscillations that alter sediment availability and stream capacity, resulting in discharge peaking in 2000 and sediment concentration in 2005. However, more data is required to confirm this relationship. Variability in stream flow appears to share a causal relationship with sediment transport variability, as both are linked to variation in precipitation and the resultant impacts on vegetation growth and evapotranspiration rates. Stream flow variability and sediment transport variability must have implications for stream and floodplain geomorphology, and the hydrology of variable rivers should be considered when interpreting their geomorphology.

1. Introduction

1.1. Stream flow variability

The founding premise of fluvial geomorphology is that rivers do geomorphic work, resulting in a landscape characteristic of a particular river's capacity to transport sediment. Similarly, in riverine ecology, flow is deemed the fundamental driving force of ecological form and process (Puckridge *et al*. 1998). As such, ecologists have branched successfully into hydrology in order to understand how flow variation can impact upon ecosystems, and how variation can be measured and described (e.g. Jenkins *et al*. 2005; McMahon and Finlayson 2003; Peel *et al*. 2001; Puckridge *et al*. 1998). Central to ecologists interest in flow hydrology, is that ecosystems are perceived to evolve contemporaneously with flow regimes, and as such, maintenance of such flow regimes equates to maintenance of the prevailing ecological balance.

As such, the recognition of flow variability as an important component of hydrology arose from the need to understand ecosystem dynamics. Dettinger and Diaz (2000) showed that variation in annual precipitation was generally low worldwide, with southern Africa and Australia experiencing slightly more variable precipitation than the rest of the globe. However, variability in run-off has repeatedly been shown to be far more variable in southern Africa and Australia than other continental areas (e.g. Puckridge *et al.* 1998; Dettinger and Diaz 2000), even when compared to areas of a similar climate (Peel *et al*. 2001). Coefficients of variation (CV) for southern Africa and Australia vary from 75 to 110%, as compared to global norms of between 20 and 45%. Poff *et al.* (2006), using principle component analysis, found that South African streams were characterised by flashy flows on an inter- and intra-annual scale, noting that streams in South Africa and Australia could be described as globally the most 'extreme'. As a result, Jenkins *et al.* (2005) refer to these areas as being characterised by 'boom and bust' hydrology and ecology. Variation does appear to follow some degree of cyclicity, such as the link between lower run-offs in the most variable regions during La Niña years and subsequent water years (McMahon and Finlayson 2003; Puckridge *et al*. 2003; Dettinger and Diaz 2000).

Co-efficients of variation for precipitation have been found to only modestly correlate with CV's of run-off (Dettinger and Diaz 2000). Accordingly, Peel *et al*. (2001) found that only a small proportion of run-off variability could be accounted for by variability in precipitation. The increase in variability from precipitation to run-off indicates that variability is enhanced during the process of converting precipitation to run-off. The major contributor to flow variability has been attributed to the effects of evapotranspiration (e.g. Poff *et al*. 2006; Peel *et al*. 2001). Peel *et al*. (2001) showed that evapotranspiration was greatly increased in areas that were dominated by evergreen, as opposed to deciduous, trees. In addition, he found that climates with wet summers and dry winters were more conducive to increased run-off variability. Similarly, Dettinger and Diaz (2000) reported that summer precipitation generally contributed less to stream flow than did winter precipitation. A combination of these factors is likely to be the cause of heightened stream flow variability in Australia and southern Africa. A correlation between catchment size and stream flow variability has also been noted by some authors (Peel *et al*. 2001; Dettinger and Diaz 2000; Puckridge *et al*. 1998). Overall, streams with large catchments, that receive most of their rainfall in summer, and that have greater proportions of evergreen trees as opposed to deciduous trees, can be expected to experience the greatest amount of variability in stream flow.

There has been extensive research on how such variability impacts upon ecosystems, but not on how variability impacts upon the underlying building blocks of ecosystems, the geomorphology. The impact of variable flows on geomorphology has only received adequate attention in dryland environments (Tooth 2000). However, many variable rivers do not fit into the category of dryland rivers as defined by geomorphologists, even though they may experience similar hydrology in terms of flow variability. Rivers on the eastern seaboard of southern Africa, exhibitting similar variability to those of Australia, do not experience transmission losses and are not located in areas of low precipitation (e.g. Costelloe *et al*. 2003). Furthermore, flow is generally perennial even though the majority of flow is received during a few large flood events. Despite modest precipitation on the eastern seaboard (±1000mm/a average), the majority of the region experiences a negative annual water budget due to the high evapotranspiration losses (Schulze 1997). Tooth (2000) suggests that dryland rivers are characterised by few large floods, with intervening low flows for the majority of the water year. This is similar to other authors definition of variable rivers of southern Africa, described as flashy (Poff *et al*. 2006). In Tooth's (2000) dryland rivers, large magnitude floods are the major landscape driver, as low flows lack capacity to do extensive geomorphic work in intervening periods.

There are substantial differences between the geomorphology of Tooth's (2000) dryland rivers, and those that are merely termed variable by hydrologists, despite a shared variability in stream flow. While it is accepted that geomorphic processes of dryland rivers are different from those of more regular rivers, the geomorphic processes and sediment transport characteristics of variable rivers are frequently assumed to be the same as those of regular rivers.

This chapter examines the stream flow of the Mfolozi River, a variable river on the eastern seaboard of southern Africa, which cannot be characterised as a dryland river due to lack of transmission losses and geomorphology. The aim of this chapter is to address how stream flow variability impacts upon sediment transport and suggest how this may potentially affect floodplain dynamics and processes.

1.2. Mfolozi River, KwaZulu-Natal

The catchment of the Mfolozi River drains 11 070km^2 of northern KwaZulu-Natal on the eastern seaboard of southern Africa. The KwaZulu-Natal region was rejuvenated 20Ma and 5Ma when southern Africa experienced two major uplift events that lifted the eastern seaboard by 250m and 900m respectively (Partridge and Maud 1987). As a result, the region is currently in a long-term state of incision, with rivers considered to be relatively steep and fast flowing.

The Mfolozi River comprises two major tributaries, the Black Mfolozi, which arises in the north approximately 1500m amsl, and the more southern White Mfolozi, which arises at an altitude of 1620m. The two rivers converge approximately 50km west of the Mfolozi River mouth. According to the most recent land cover inventory, the majority of the catchment is under natural vegetation cover, largely because much of the catchment falls with the Mfolozi-Hhluhluwe Nature Reserve. Of the remaining natural vegetation, 60% is grassland, with lesser areas of thicket and bush (21%) and natural forest and woodland (15%). Besides natural vegetation, just less than a quarter of the Mfolozi catchment falls under agriculture, the majority of which is small-scale subsistence and commercial forestry. 13% of the catchment has been classified as degraded through overgrazing or excessive resource use. Less than 1% of the catchment is urban, with the major industrial centers of Mtubatuba, a timber and sugar cane processing town, located in the lower catchment and Vryheid, a coal mining and quarrying district, located in the upper catchment.

Precipitation is largely restricted to the summer months when approximately 80% of the rainfall occurs, peaking between November and April (Tyson 1986). Mean annual precipitation in the catchment varies from 1288mm at the coastal town of St. Lucia, to 667mm/a in the Umfolozi Game Reserve in the mid-upper catchment, to 914mm/a at Nongoma in the upper catchment. Mean annual potential evapotranspiration is generally more than double that of precipitation, with atmospheric demands averaging 1800mm/a (Schulze 1997).

Heavy rainfall is generally associated with easterly low-pressure cells that can remain in the region for up to 10 days (Tyson and Preston-Whyte 2000). The occasional occurrence of tropical cyclones may lead to extremely high rainfall in the catchment, and therefore high recurrence interval flood events on the Mfolozi River. The most recent such event was that of Cyclone Domoina in 1987, which resulted in a peak discharge of approximately 16 000$m^3.s^{-1}$, which constitutes approximately 3 times the 100 year return period flood. During the flood, current velocities of 2.6 $m.s^{-1}$ were measured (Travers 2006). The main mechanism for winter rainfall is the passage of cold fronts and coastal low-pressure systems. The Mfolozi catchment is thus at the interface of weather producing systems derived from the south, in the form of mid-latitude cyclones, and weather systems originating towards the north in the tropical easterlies, such as easterly waves and lows, and occasionally tropical cyclones.

2. Methods

2.1. Catchment precipitation and stream flow of the Mfolozi River

To assess seasonal aspects of rainfall in the catchment, precipitation records from five weather stations in the catchment were obtained from the South African Weather Bureau. Stations were located in the upper Black and White Mfolozi River catchments (Hlobane and Goedgeloof, n=89 and n=64 years respectively), in the mid- Black and White Mfolozi River catchments (Mbhuzana and Mahlabatini, n=24 and n=89 years respectively) and below the confluence on the coastal plain (Uloa Agricultural Office, n=75 years; see Figure 1 for the location of the weather stations).

Three stream flow series were obtained from the South African Department of Water Affairs and Forestry for the Mfolozi River. Two of the data sets were of the major tributaries of the Mfolozi River, the Black and the White Mfolozi Rivers. The third flow series was of the Mfolozi River below the confluence of its two major tributaries on the lower coastal floodplain (See Figure 1 for the location of the flow gauges). The tributary flow records were much longer than that available for the Mfolozi gauge located on the lower floodplain, with only 10 years being available for analysis as compared to 40 years at the upper gauges. Characteristics of the data sets are summarized in Table 1.

Table 1: Summary of monthly discharge data availability.

River	Black Mfolozi River		White Mfolozi River		Mfolozi River	
Gauge name	W2H006		W2H005		W2H032	
Start of record	1965		1963		1993	
	%	*n (months)*	%	*n (months)*	%	*n (months)*
Accurate records	91.5	422	95.3	470	81.1	103
No record	8.5	39	4.7	23	18.9	24
Interpolated record	3.7	17	0.6	3	12.6	16
Records used in analysis	95.2	439	95.9	473	93.7	119

The median annual discharge was calculated using all available records from each of the data sets. In addition, a mean annual discharge for each year was calculated in order to allow an inter-annual comparison. A percentage deviation from the long-term median was then calculated for each year in each data set in order to establish long-term variation. The amount of correlation between the gauging stations in terms of percentage deviation was also determined. A frequency analysis of percentage deviation values allowed an investigation into the comparative number and severity of wet and dry years. Since the period of record for the Mfolozi River was insufficient, the analysis was run using data from the Black and White Mfolozi Rivers percentage deviation from the median.

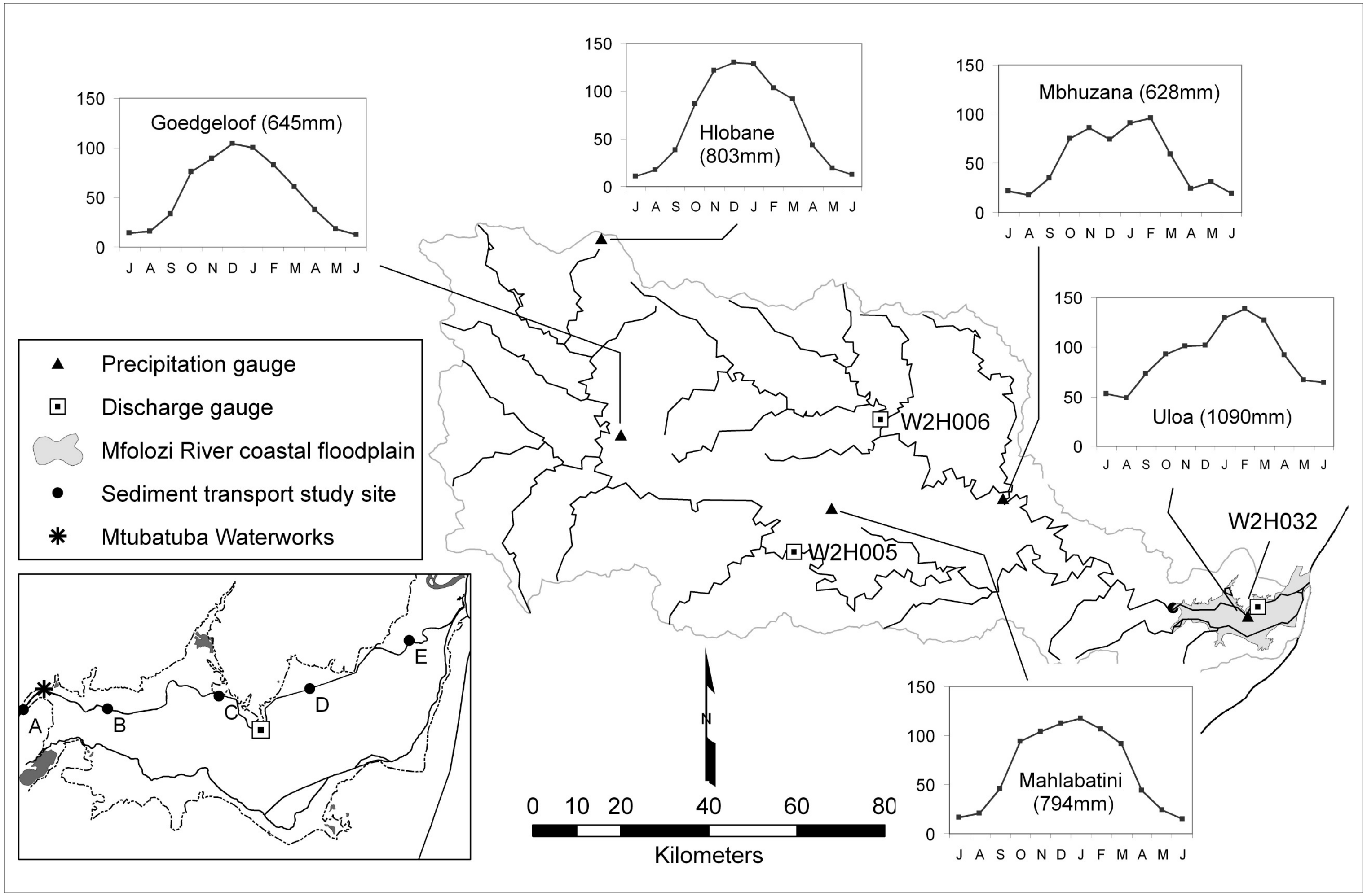

Figure 1: The catchment of the Mfolozi River, showing the location of weather stations, their average monthly precipitation (mm), and average annual precipitation (number in brackets). The location of discharge gauges is also shown. The coastal floodplain study area is displayed on the inset, with the location of the Mtubatuba Waterworks and sediment sampling sites indicated.

In order to test relationships between rainfall and discharge in different areas of the catchment, correlations between rainfall (percentage deviation from the median of total annual precipitation) and discharge (percentage deviation from the median of annual mean precipitation) were also calculated. The co-efficient of variation (CV) for each of the discharge and rainfall gauges records were computed. In addition, the water budget for the quaternary catchment in which each rainfall gauge was located, was calculated from Schulze's (1997) values of mean annual precipitation and potential evaporation.

2.2. Sediment transport

Sediment flux was assessed at five straight reaches on the lower Mfolozi Floodplain between the 6th and 11th of March 2006 (A to E on the inset Figure 1). Sites were selected such that they were equally spaced and represented areas of different floodplain slope. The same sampling method was used at each site A to E. Data collection at each site took approximately 8 hours.

2.2.1. Bedload sediment

Bedload was measured using a Helley-Smith bedload sampler constructed to the specifications of Emmett (1980) with a weight of approximately 30kg and a sampling bag with a 0.25mm mesh size. The weight of the Helley-Smith sampler ensured there was no frictional drag as it was lowered into the water. At sites A to E, 3 traverses were completed 10m apart to overcome the effect of channel bedforms (Emmett 1980; Carey 1985). Each traverse was divided into 5 subsections and bedload sampling was conducted at each one. Initially, 4 samples per subsection were collected, but this was reduced to 3. This was largely to ensure that the entire data set was collected within the day, such that error due to changing discharge could be reduced (e.g. Kleinhans and Ten Brinke 2001). Unfortunately, this is likely to increase error associated with flow and transport variability (e.g. Pitlick 1988, Gaweesh and van Rijn 1994, Kleinhans and Ten Brinke 2001), but was considered to be less important than error incurred through changing discharge. Overall, between 45 and 50 samples were collected at each sample site, exceeding Gomez and Troutman's (1997) recommendation of 40 samples to reduce random and systematic error. Sampling time per sample was between three and four minutes, and was measured to the nearest second.

Bedload samples were dried and weighed. The dry weight of the combined samples at each subsection was used to calculate sediment flux and sediment discharge respectively. Dried samples were then sieved in order to establish particle-size distribution.

2.2.2. Suspended sediment

Suspended load was sampled on the first of the three traverses at each sample site. Samples were taken at each of the 5 subsections at variable depths using an Eijelkamp Watertrap sampler, with a cylindrical volume of 1.22L. The aim was to sample at 0.5m depth intervals, but it was sometimes necessary to sample more frequently when the channel was shallower than 1m. The samples were transported in 2L plastic bottles for laboratory analysis.

Particle-size of 10 of the suspended sediment samples was measured using a Malvern Mastersizer.

Turbidity was measured for each of the 38 samples using a calibrated turbidity meter. The sediment concentration of each sample was calculated by evaporating the sample, and then weighing the remaining sediment. It was assumed that the addition of dissolved solids to the suspended sediment through evaporation was negligible, particularly since sample mass was accurate to only 2 decimal places. The suspended sediment concentration was used to calculate overall suspended sediment discharge for each sample site in kilograms per second. However, this data was of limited use in terms of determining long term suspended sediment trends as the data set had too few data points for the development of a sediment rating curve (e.g. Horowitz 2002; Ferguson 1987). Furthermore, a sediment rating curve was considered inappropriate for the analysis. Sediment rating curves, representing the average relationship between discharge and sediment concentration are not sensitive to seasonal variations in sediment transport, antecedent conditions and differences in sediment availability (Aselman 2000), which were the effects under investigation. Furthermore, it seemed likely that strongly seasonal rainfall in the catchment would further increase scatter around the regression line, causing inaccuracy (Ferguson 1986; Aselman 2000).

To overcome the problem of developing a long-term understanding of sediment transport using a sediment rating curve, an indirect measure of sediment concentration, using the relationship between sediment concentration and turbidity was used (e.g. Walling 1977). A six-year record of turbidity, with records maintained every one to two hours every day, was obtained from the Mtubatuba Waterworks located at the head of the floodplain. Sediment concentration for the turbidity record was calculated using the relationship between the known sediment concentration and turbidity of the 38 samples collected in this study. Lenzi *et al*. (2003) and Riley (1998) noted that using turbidity as a proxy for suspended sediment load could be erroneous if there was large variation in sediment mineralogy and particle size, or if the water contained high amounts of organic matter. However, Walling (1977) suggested using turbidity as a proxy when sediment particles were clay and silt sized and where catchments consisted of relatively homogeneous rock types. However, since the primary aim of this study was to compare sediment transport between years and months and not calculate absolute sediment yields, and the R^2 value for turbidity and sediment concentration was greater than that of Riley (1998), it was assumed that the error would remain constant, and therefore would not materially affect the results. Gippel (1995) suggested that this is a reasonable assumption, since sudden temporal changes from purely organic to purely mineral loads are rare in nature. Furthermore, Gippel (1995) also argued that the close correlation between sediment concentration and turbidity suggests that particle size variation in streams is either not usually great, or that particle size variations do correspond with changes in sediment concentration. As such, turbidity records were used to calculate mean annual and monthly sediment transport rates.

2.2.3. Depth profile, flow velocity and channel gradient

Along each transect, depth was measured using a weight attached to a measuring tape, while velocity was measured at frequent (every 0.5m or less) depth intervals using a SEBA-current meter with a 125mm diameter propeller. Both depth and velocity were measured at 4-6m intervals across the channel such that an accurate cross-section and velocity profile could be drawn. Mean current velocity was calculated using an area-weighted average for each channel cross-section, which was then used to calculate discharge.

Channel gradient was calculated from the sample site's position on a longitudinal profile surveyed in April 2005 using a differential GPS with an on-site base station and roving GPS receiver. Results were accurate to within approximately 0.1m in x, y and z fields following processing and correction.

3. Results

3.1. Seasonal variation in stream flow

Precipitation in the catchment varies substantially in terms of timing and amount (Figure 1). The Hlobane station, located in the north on the boundary of the Black and White Mfolozi catchments, experiences peak rainfall in the month of December. Rainfall at the Goedgeloof station, in the White Mfolozi River catchment, similarly reaches a peak in December. Mahlabatini station contrastingly experiences peak rainfall during January. Precipitation in the Mbhuzana and Uloa regions occurs as two distinct peaks; the first (lesser) peak occurs in November, while the second (larger) peak occurs in February.

Seasonal discharge patterns of the Black and White Mfolozi Rivers are relatively similar (Figure 2). Both rivers experience two peaks in discharge, one smaller peak in December, followed by a second larger peak in February. The White Mfolozi River has a larger discharge than the Black Mfolozi River, and has a larger catchment (see Figure 1). The Mfolozi River, as measured at the coastal gauge, shows only one distinct peak that occurs in January. There are substantial ungauged tributary inputs below the two tributary gauges (see Figure 1). As such, there is generally a substantial difference between the combined flows of the Black and White Mfolozi Rivers, and flow of the Mfolozi River. However, in 16% of the months on record, the combined discharges of the Black and White Mfolozi Rivers were found to exceed that measured on the Mfolozi River at the W2H032 gauge, suggesting that within these months discharge decreased downstream.

The relationships between rainfall stations and flow gauges were found to be positively correlated, generally with significance between p=0.01 and p=0.05. The positive correlation between rainfall variation at Mahlabatini and Goedgeloof rainfall stations, and the Mbhuzana and Hlobane stations was significant at the 99% confidence level. Similarly, the correlation between discharge measured at the Black Mfolozi gauge and rainfall at Mahlabatini was also highly significant ($p < 0.01$). Discharge on the Mfolozi River and rainfall at the upper Goedgeloof gauge was significant at the 95% confidence level. Contrastingly, the positive correlation between discharge on the White Mfolozi River and rainfall at Mbhuzana was not significant ($p > 0.05$). Discharge measured at the Mfolozi gauge was also correlated positively with precipitation at Hlobane, Mahlabatini, Mbhuzana, and Uloa, although this was not significant ($p > 0.05$).

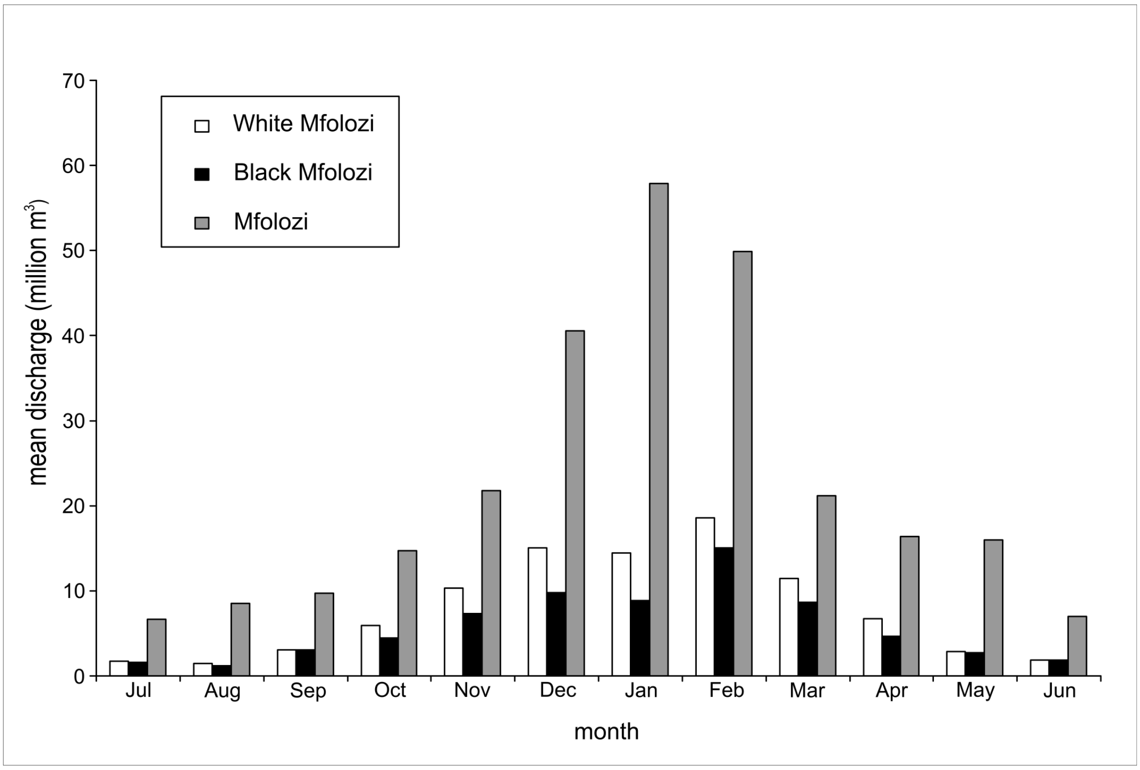

Figure 2: Seasonal flow variations on the Black, White, and Mfolozi Rivers.

3.2. Inter-annual variation in stream flow

The short duration of the Mfolozi River flow record made it unsuitable for long term flow analysis. However, the strong correlation between the Black and Mfolozi Rivers, and the White and Mfolozi Rivers suggests that the long-term pattern of the combined flows is likely to be similar to that of its tributaries. As such, flow measured at the Black and White Mfolozi gauges is used as a proxy for variation of the combined flow. Variation from the median was plotted as a time series for the Black and White Mfolozi Rivers (Figure 3). The moving average trend lines for each station display a similar pattern to each other. The most notable error in the data is the absence of a very high discharge in 1984 for the White Mfolozi River. However, in that year, a large deviation from the median is obvious on the Black Mfolozi River. The flood that it records damaged the gauge of the White Mfolozi and this result is therefore incorrect. Similarly, damage to both the gauges during a large flood in 1987 has resulted in the deviation from the median in 1987 being underestimated.

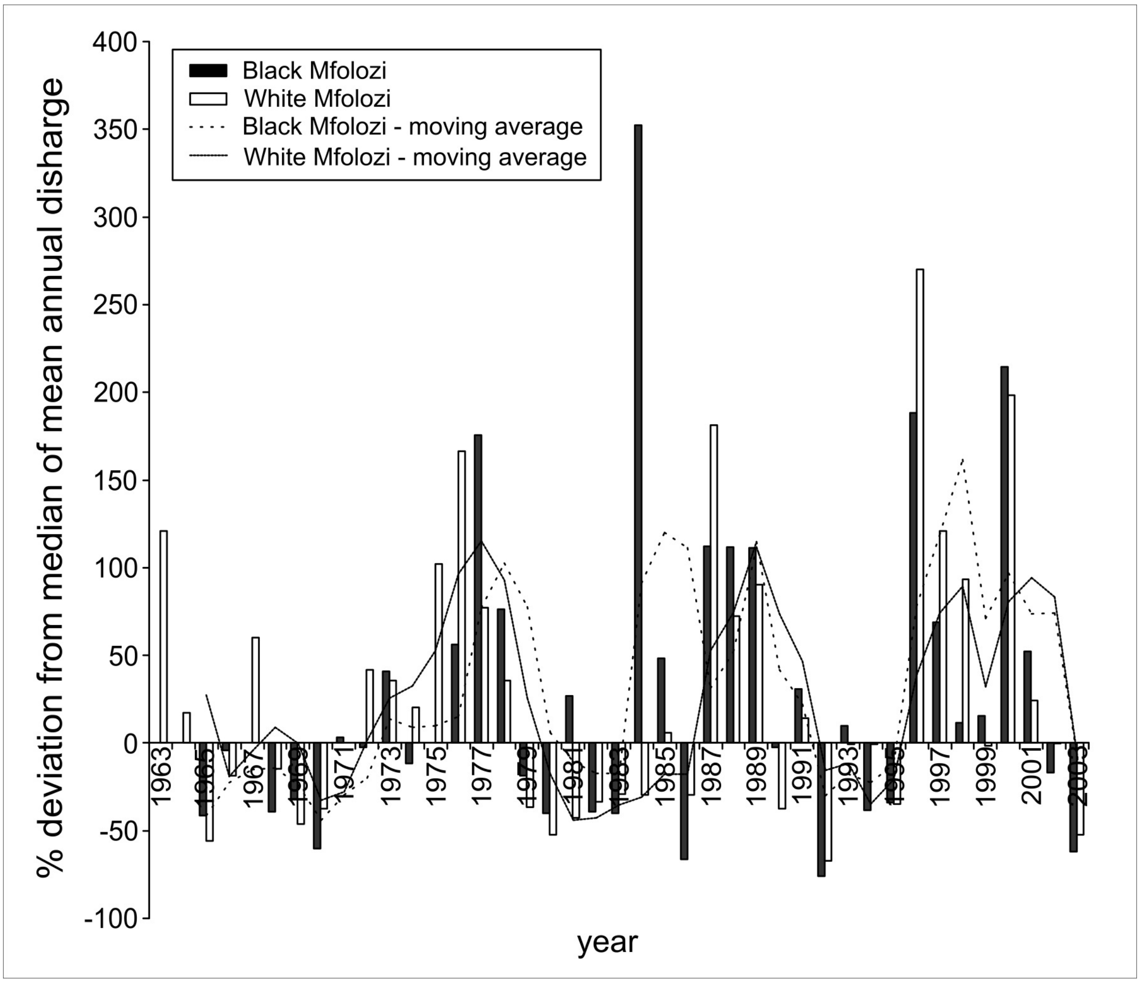

Figure 3: Percentage deviation from median of annual discharge for the Black and White Mfolozi Rivers for the period of record. Moving average trend lines (3 years) are also depicted. The 1984 and 1987 floods are under-represented due to gauges being damaged.

Despite these difficulties with inaccuracies in the data during and following large floods, there are several succeeding periods of dry and wet years clearly represented in the data. Flow is below the median between 1965 and 1971, above the median from 1972 to 1978, and then below the median again between 1979 and 1983. 1984 was marked by a large flood, and 1985 was also characterised by flows greater than the median flow, whereas in 1986 flow was less than the median flow. Two more generalised periods of greater than median flow occurred from 1987 to 1991 and between 1996 and 2001. The intervening period, 1992 to 1995 was a period of relatively low flow, as was 2002 and 2003.

A frequency histogram of percentage deviation from the median of mean annual flow on the Black and White Mfolozi Rivers is shown in Figure 4. The frequency distribution differs quite strongly from a normal distribution, with a definite negative skewness. Despite the strong skew towards values lower than median, it is the set of positive deviations that have the greatest number of outliers (i.e. at + 250% and above). There are no similar outliers with a negative deviation.

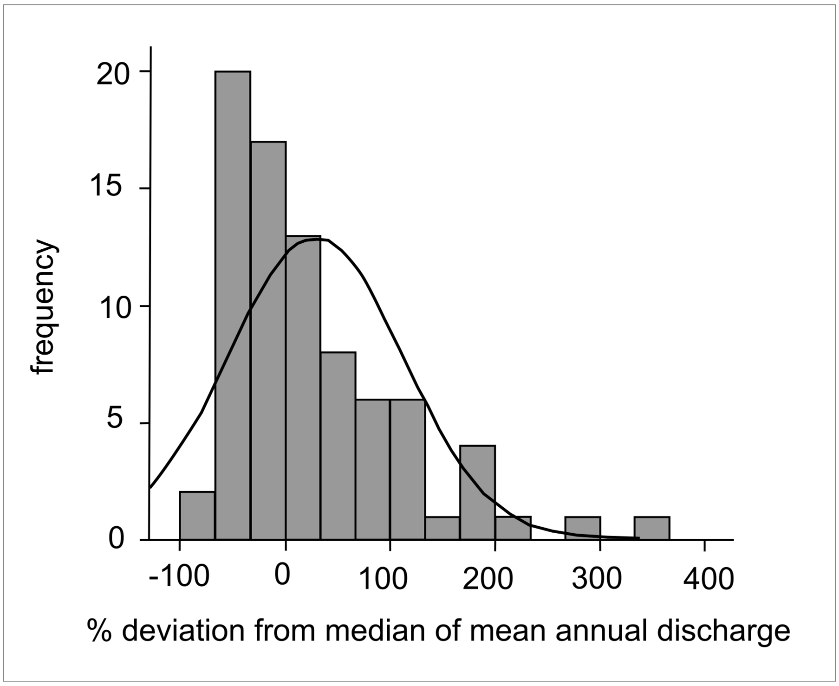

Figure 4: Frequency histogram of percentage deviation from the median of mean annual discharge on the Black and White Mfolozi Rivers over the period of available record.

The co-efficients of variation for all of the discharge gauge records were high. The Black Mfolozi had the lowest CV at 61%, followed by the White Mfolozi at 69% and the Mfolozi River at 79%.

The Mbhuzana precipitation gauge showed the greatest amount of inter-annual variation, with a CV of 36.6%. This was followed by Mahlabatini (30.8%), Uloa (29.9%), Hlobane (26.6%), and Goedgeloof (22.6%). All the precipitation gauges are located in areas of a negative water budget, whereby atmospheric demand exceeds precipitation. The water budget, calculated for the quaternary catchment of each gauge from Schulze (1997), showed that Mbhuzana had the greatest atmospheric water demand in relation to precipitation, with an annual water deficit of –1145mm. This was followed by Goegeloof (-1132mm/a), Mahlabatini (-1077mm/a), Hlobane (-1019mm/a), and Uloa (-621mm/a).

3.3. Channel morphology and hydrology during low flows

Data collection during a week of March 2006 was found to correspond with a falling limb of a flood wave, as indicated by stage height measurements taken from the Uloa Bridge gauge (Figure 5). Decreasing stage height appears to correspond with decreasing discharge measurements taken at various parts of the floodplain over the time period of the study.

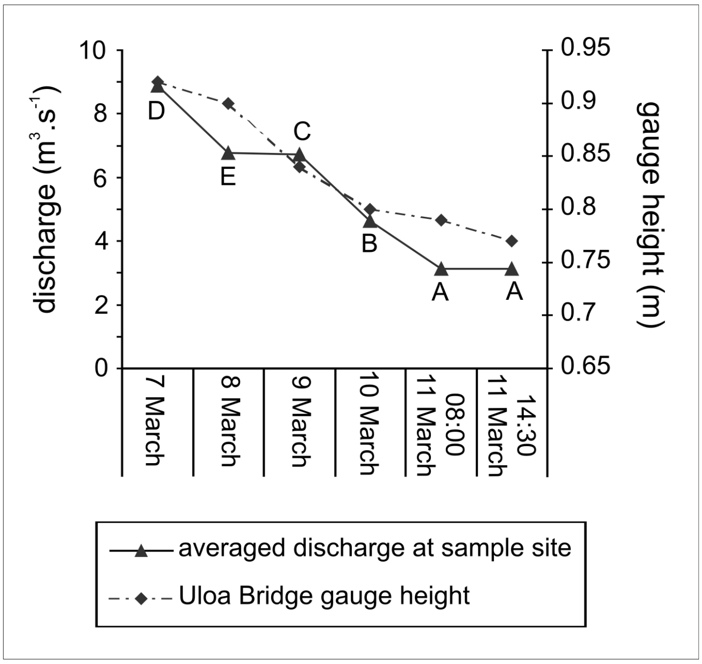

Figure 5: Discharge measured at each sample site (A – E) and gauge height as measured at the Uloa Bridge. Locations of each site are indicated on inset of Figure 1.

Channel width varied from 29m at sample site B to 41m at site D (Table 2; Figure 6). Transects B and C were found to have the fastest average flows, with a velocity of greater than $0.5 m.s^{-1}$ at each (Figure 6).

River gradients for each sample site were calculated from their location on an accurate floodplain longitudinal profile. These gradients therefore represent a regional stream gradient, rather than a localized stream gradient. Sample site A had the highest stream gradient of 0.06%. This was followed by sample site C at 0.05% and B at 0.03%. Sample sites D and E were located in an area of relatively uniform stream gradient, with a slope of 0.02%.

Table 2: Summary of sedimentary and hydrological data.

Sampling date	Sample site	Transect	Cross-sectional area (m^2)	Weighted average velocity ($m.s^{-1}$)	Discharge ($m^3.s^{-1}$)	Channel width (m)	Average depth (m)	Average discharge for sample site ($m^3.s^{-1}$)	Average velocity for sample site ($m.s^{-1}$)	Average turbidity (NTU)	Total suspended sediment discharge ($kg.s^{-1}$)	Average suspended sediment concentration ($kg.m^{-3}$)	Total bedload discharge ($kg.s^{-1}$)	Bedload discharge per width unit ($kg.m^{-1}.s^{-1}$)	Suspended load ($kg.s^{-1}$): bedload ($kg.s^{-1}$)	Channel gradient (%)
11-Mar-06	A	A	11.20	0.40	4.44	38	0.26	3.13	0.31	297.33	1.95	0.44	0.45	0.0154	4.38 : 1	0.06
		Ax	10.03	0.35	3.47											
		Ay	7.97	0.19	1.49											
10-Mar-06	B	B	9.04	0.53	4.79	29	0.32	4.64	0.50	422.71	2.75	0.57	1.02	0.0428	2.70 : 1	0.03
		Bx	9.31	0.51	4.74											
		By	9.65	0.46	4.39											
09-Mar-06	C	C	13.17	0.57	7.62	32	0.39	6.72	0.54	520.29	4.36	0.57	1.08	0.0423	4.05 : 1	0.05
		Cx	11.60	0.53	6.17											
		Cy	12.21	0.52	6.37											
07-Mar-06	D	D	22.08	0.41	9.00	41	0.51	8.89	0.42	880.40	8.47	0.94	0.82	0.0218	10.32 : 1	0.02
		Dx	22.53	0.42	9.48											
		Dy	18.42	0.44	8.19											
08-Mar-06	E	E	17.14	0.37	6.25	40	0.40	6.78	0.43	673.00	4.57	0.73	0.84	0.0254	5.41 : 1	0.02
		Ex	14.70	0.46	6.74											
		Ey	15.89	0.46	7.34											

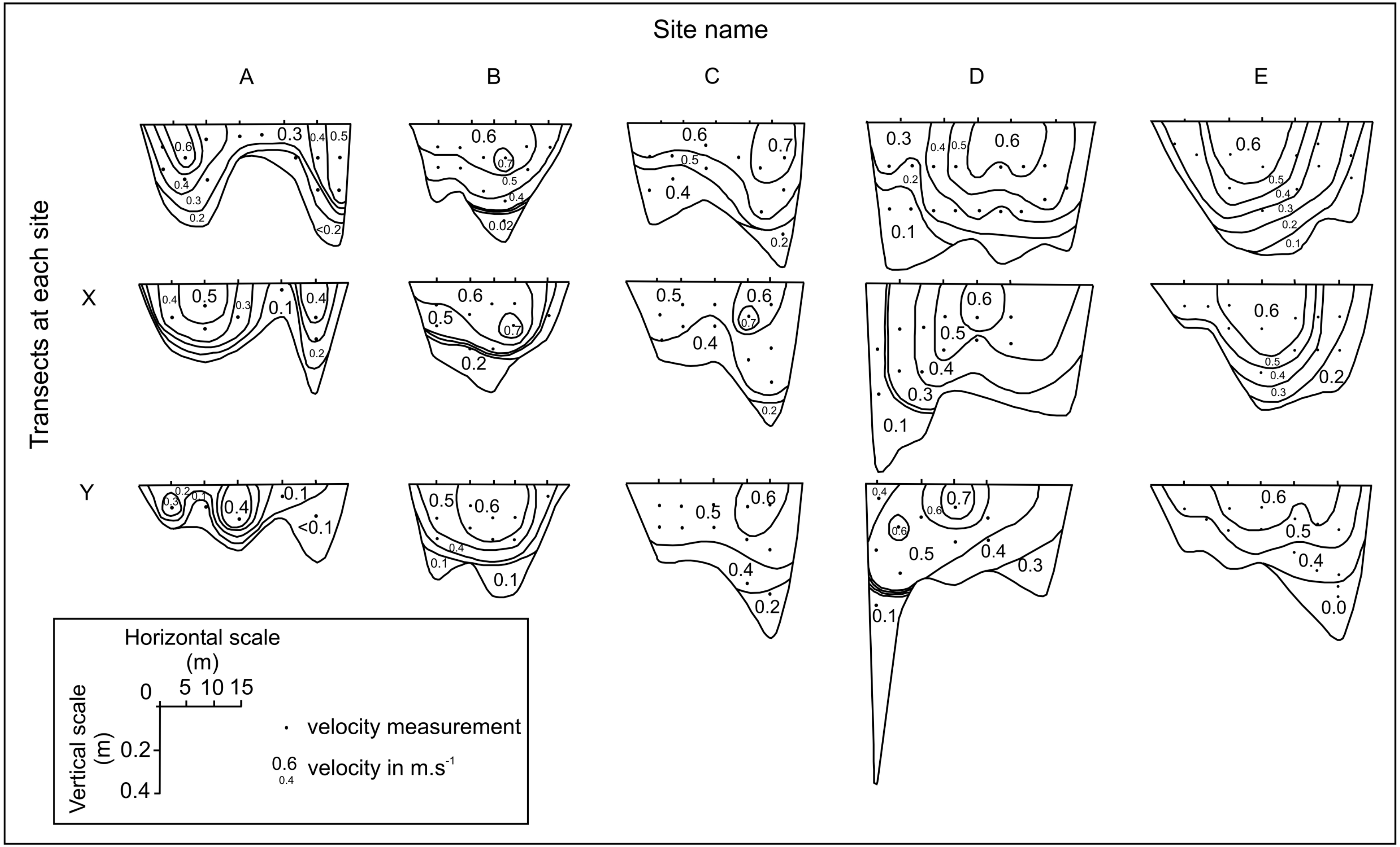

Figure 6: Cross-section and velocity profiles of transects completed at sample sites A to E.

3.4. Bedload sediment transport

Bedload discharge per unit width was found to correlate strongly with mean velocity (R^2 = 0.9086), displaying a linear relationship when plotted logarithmically. Sediment transport was most closely related to velocity using a power function.

The particle size distribution of bedload sediment during the period of study was relatively constant, with all transects having an average D_{50} particle size of medium grained sand, with the average D_{50} bedload particle size being 0.35mm in diameter. All particle size distributions were unimodal.

Comparisons of bedload discharge between sample sites was complicated by the falling flood wave over the study period (Figure 5). Nevertheless, bedload sediment discharge was greatest at sample sites B (1.02kg.s^{-1}) and C (1.08kg.s-^{1}), where velocities were greatest. Site A had the lowest bedload sediment discharge, with 0.45kg.s^{-1}, while sites D an E had bedload discharges of 0.82 and 0.84kg.s^{-1} respectively.

3.5. Suspended sediment transport

The D_{50} particle size of suspended sediment averaged 0.0052mm (very fine silt). Suspended sediment concentration was most closely related to discharge on a linear scale, with a regression co-efficient (R^2) value of 0.6473. However, the small number of values on the regression line makes the development of a sediment rating curve invalid, as discussed previously.

Average suspended sediment concentration (kg.m^{-3}) of the 38 samples was most closely related to turbidity (R^2 = 0.9567) when a power equation was applied. The power equation, where T equals turbidity (ntu), is:

$$[\text{suspended sediment}] = 0.0093\ T^{\ 0.674}$$

Thus, the turbidity record from the Mtubatuba waterworks was used to construct a record of suspended sediment concentration. The suspended sediment discharge was then calculated by multiplying the suspended sediment concentration by the discharge. When the relationship between suspended sediment concentration and turbidity was applied to the years of record, it was found that sediment concentration was not significantly correlated to discharge (R^2 = 0.2087). Figure 7, indicating variation in sediment concentration and discharge between 2000 and early 2006, further highlights the wide disparity between sediment concentration and discharge. Sediment concentrations were generally highest in 2004, followed by 2003 and 2005. High sediment concentrations were not always coincident with high discharges, as the highest discharges were recorded during 2001.

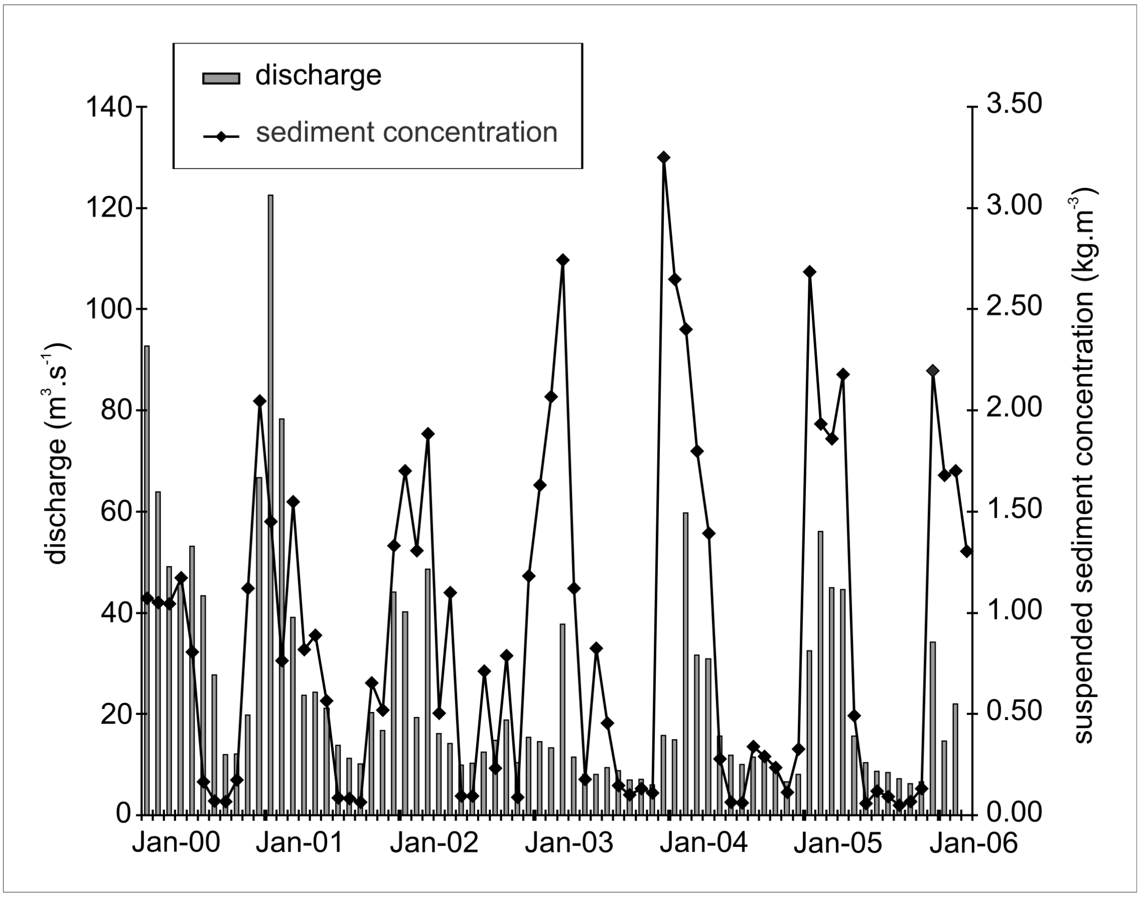

Figure 7: Monthly variation in sediment concentration and discharge for the period 2000 to February 2006.

4. Discussion

4.1. Stream flow hydrology of the Mfolozi River catchment

The Mfolozi River catchment comprises three distinct regions characterised by differences in timing of maximum precipitation. In general, rainfall occurs later in the summer season as one moves towards the coast. In the interior, rainfall peaks in December. Towards the central catchment, peak rainfall occurs in January as represented by the Mahlabatini gauge. At the coast, and towards the mid-catchment in the north, peak rainfall occurs in February. The latter regions are also characterised by a less marked precipitation peak in November. There does not appear to be an obvious regional trend in terms of amount of precipitation experienced in the catchment, except that rainfall is highest near the coast.

The January discharge peak of the Mfolozi River is probably related to a combination of high rainfall in the mid-catchment in January and groundwater flow from December rains feeding into the drainage network. The reasons for the two peaks in discharge in the Black and White Mfolozi River is not well understood, although it may be related to base flow from December and January rains in combination with February inputs.

The lack of correlation between the Mfolozi gauge and catchment rainfall records is likely an indication of the heterogeneity in rainfall over the catchment.

Discharge assessed on a monthly scale showed that the Mfolozi River was on occasion, susceptible to transmission losses downstream. This mostly co-incided with summer periods. Of the transmission losses, 60% occurred during December and January. In 53% of the cases, transmission losses exceeded 30%, while in 26% of the cases, losses exceeded 50%. However, due to the short record of the Mfolozi River, combined with abstraction of water for irrigation on the Mfolozi River floodplain, it is difficult to characterise this as part of the river's hydrology. Nevertheless, the loss may reflect a combination of withdrawal for the cultivation of sugar cane on the floodplain and the effects of evapotranspiration.

4.2. Stream flow variability

Precipitation in the Mfolozi River catchment was found to be more variable than many other parts of the globe (e.g. Dettinger and Diaz 2000), with co-efficients of variation for rainfall in the catchment varying from 36.6% at Mbhuzana to 22.6% at Goedgeloof. These translated to even higher CV's for stream flow, which seem to be related to catchment size, with the smallest catchment of the Black Mfolozi River experiencing the least variation in discharge (61%), while the larger Mfolozi River catchment, experienced the greatest variation (79%), a relationship that has also been described by other authors (e.g. Puckridge *et al*. 1998). In global terms, the CV values can be considered extremely high and indicate highly variable stream flow (e.g. Dettinger and Diaz).

Peel *et al*. (2001) found that in addition to increasing basin size, catchments with summer rainfall and dry winters were most susceptible to experiencing variable stream flow. Peel *et al*. (2001) argue that continued evapotranspiration during the dry months results in continual depletion of water supplies. The Mfolozi River catchment is characterised by a range of factors that exacerbate stream flow variability, including high rainfall variability, a large proportion of evergreen vegetation in the catchment, a large catchment, as well as summer rainfall combined with a dry winter.

Dettinger and Diaz (2000) have suggested that variable rivers are characterised by relatively large inter-annual variation, with typically small base flows that persist throughout the year. A few, large but brief floods usually determine much of the total annual flow in a given year. This statement holds true for the Mfolozi River, with a high frequency of persistent low flows well below the median. Of particular interest is the occurrence of high positive deviations from the median, which represent sporadically occurring large flood events.

A visual analysis of percentage deviation from the median suggests a discharge cycle in the region of 6 - 7 years, with considerable variation around this figure. This is consistent with what other authors have found regarding variation associated with ENSO (e.g. Pasquini and Depetris 2007; Zhang *et al*. 2007). However, more sophisticated analysis, such as that done by Amarasekera *et al*. (1997) for tropical rivers, is required to confirm whether ENSO could be responsible for some of the discharge variability. The impact of cyclical variability could be critical in a similar fashion suggested by McMahon and Finlayson (2003), who have shown that quasi-cycles usually impact on stream flow in variable rivers, resulting in the characteristic persistent sequence of below median flow followed by above median flow.

4.3. Character of bedload sediment transport on the lower floodplain

The lower Mfolozi River is dominated by suspended sediment loads, with ratios of suspended load to bedload varying from 2.7 and 10 to 1.

Nevertheless, variation in bedload transport could provide a means for the adjustment of stream channel morphology. Since bedload transport is related to velocity, one might expect the variables of slope, hydraulic radius and roughness, as provided in Manning's equation, to have an indirect impact on the transport of bedload sediment. In order to determine the influence of each variable in maintaining a particular stream channel shape for a specific discharge; width, average depth and average velocity were plotted against discharge for each of the transects (Figure 8).

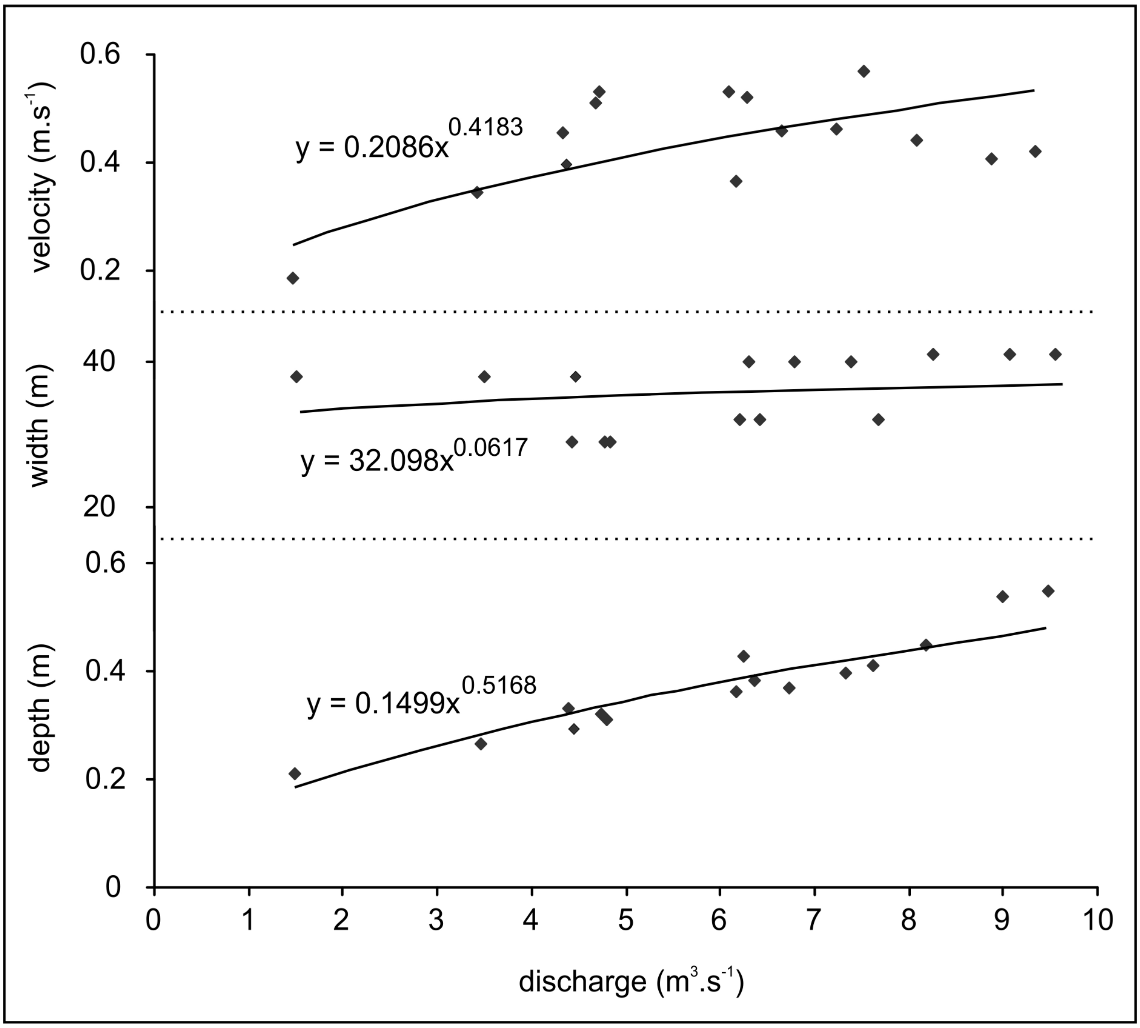

Figure 8: A comparison of width, average depth and average velocity with discharge at each of the 15 transects.

Leopold and Maddock (1953) describe the relationship between these variables and discharge as follows: $W=dQ^a$, $U=eQ^b$ and $D=fQ^c$, where W = channel width, U = stream flow velocity, D = depth, Q = discharge, and a, b, c, d, e and f are empirical coefficients.

The relative contribution of each variable to discharge is described by the values of a, b and c. The variable that is most capable of altering to accommodate discharge along the study reach was depth (c = 0.5168), followed by velocity (b = 0.4183) and then width (a = 0.0617). This suggests that variation in depth is an important mechanism for accommodating discharge. Velocity is the second most important factor in that varies with variation in discharge.

In contrast to Ellery *et al*. (2003) who found that variation in discharge was accounted for primarily by variation in channel width in the Okavango Delta, stream width is the least important variable in terms of accounting for variation in discharge. In Ellery *et al.'s* (2003) study, constricting vegetation on the channel margin could alter channel width, causing concomitant changes to stream velocity and channel roughness. In this study however, width is not easily altered, partly because of the low flow regime of the Mfolozi River, which decreases the potential time available for erosion or deposition along the stream banks. As a result, channel depth is the variable that may be altered most easily to accommodate available discharge.

Over a long period of time, ongoing local erosion and deposition through differential bedload transport will have an impact on the floodplain channel slope. The current regime of erosion and deposition along the study reach will, over the long term, lead to the rivers longitudinal profile approaching a uniform grade (Figure 9). Therefore, at sites B and C, bedload transport increases, indicating erosion of the channel bed. However, at sites A, D and E, aggradation is occurring. However, the impact on slope at each site is varied. At site A, ongoing deposition will cause channel steepening, while at sites D and E, channel slope will decrease as deposition occurs. Similarly, erosion and enhanced bedload transport at site B will cause channel slope steepening, while the channel slope at C will lessen. Erosion at sites B and C and deposition at sites D and E, provides an example of the longitudinal effects of differential rates of erosion and deposition, whereby the slope of the bed at sites B, C, D and E become more uniform. However, the upstream region becomes steeper.

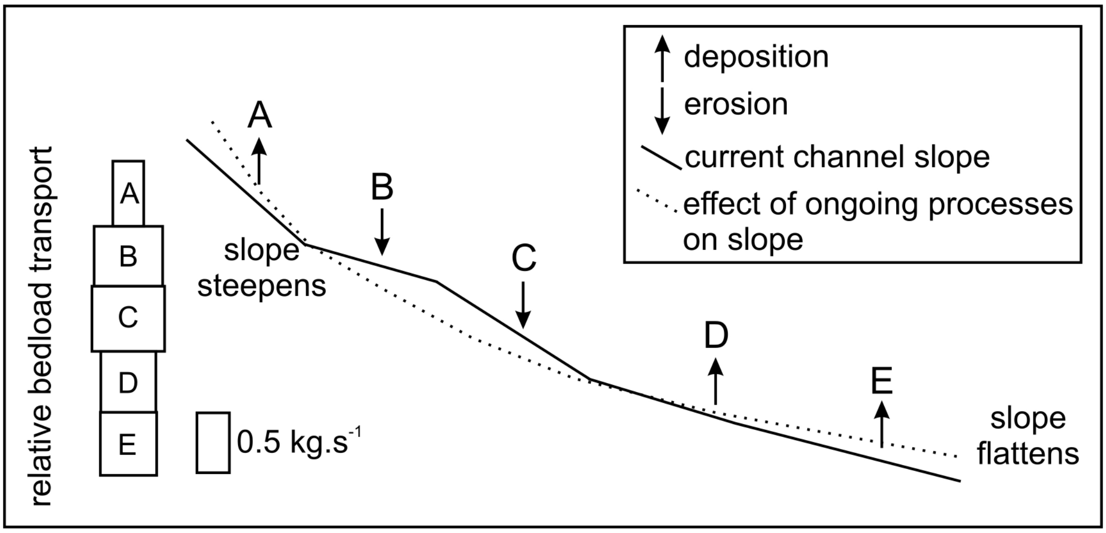

Figure 9: Relative bedload sediment transport along the study reach. The resultant long-term effect of local deposition and erosion on channel slope is shown schematically.

4.4. Sediment transport variability

The turbidity – sediment concentration relationship revealed a high degree of variability in sediment transport that was not related to discharge. Using discharge as a predictor of sediment transport would be completely invalid on the Mfolozi River due to high variability in sediment transport at the inter-annual and seasonal scales.

Sediment transport on the Mfolozi River is characterised by an annual hysteresis loop (Figure 10), where concentration usually peaked during the early wet season prior to peak discharges, with sediment concentrations generally highest in November and December. Thereafter, sediment concentrations generally decreased during the peak discharge months of January and February. Sediment concentrations between January and September were relatively low for the discharges, with the lowest concentrations experienced during the lowest discharge month of August. Thus, hysteresis results in sediment concentration being greater than one would expect during the months of October to December. Contrastingly, the months from January to September usually have lower than average sediment concentrations for their respective monthly discharges.

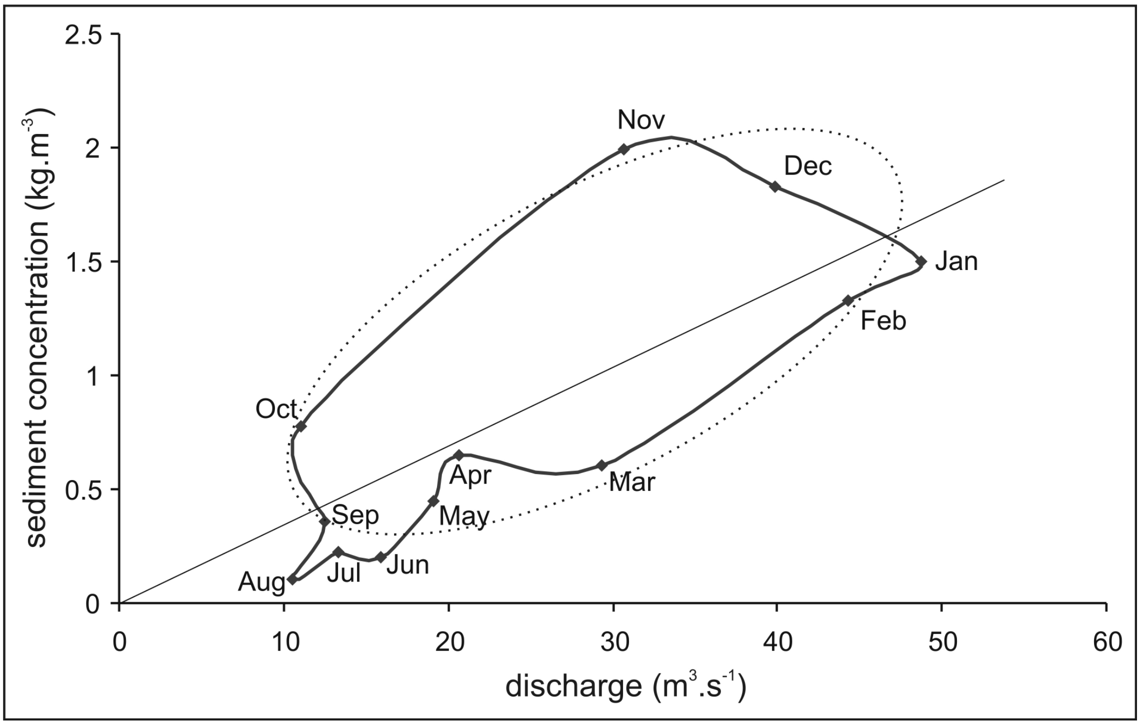

Figure 10: Mean monthly discharge and suspended sediment concentration of the Mfolozi River plotted against the linear trend line (2000 –2006). The dotted line represents the probable long term average.

Variations in sediment transport can generally be interpreted in two ways. Either, the low discharge of the dry months leads to sediment accumulation in the catchment, or alternatively, decreased vegetation cover during the dry months leads to increased sediment availability during the early part of the wet season. In the first scenario, it is the lack of transport capacity of over land run-off and the Mfolozi River in the dry months that leads to increases in sediment transport at the beginning of the wet season as capacity increases. In the second scenario, it is not that the river gains capacity to transport sediment, but rather that the actual amount of sediment available for transport increases. There is a connection between the two scenarios in that sediment may become available during the dry months as vegetation dies back, but cannot be transported to river channels without over land flow from precipitation. Since some months show much higher sediment concentrations than other months with the same discharge, sediment transport capacity of the Mfolozi River cannot be a limiting factor in sediment transport. However, sediment transport capacity during the dry season into stream channels may be lacking. As a result, hysteresis on the Mfolozi River is likely to be caused by a combination of the two scenarios. During the dry season, sediment becomes available as vegetation cover decreases. However, movement of available sediment into river channels can only occur at the onset of the rainy season through overland flow. This results in the characteristic high sediment concentrations of the early wet season, which subsequently drops as the sediment available from the dry season is completely transported. As such, accumulation of available sediment during the dry months can be considered in terms of a 'reservoir' (Picouet *et al*. 2001). Sediment concentrations drop as the 'reservoir' is depleted.

In 2000, 1600 million m^3 of water transported 897 581tonnes of sediment down the course of the Mfolozi River, which approximates 561tonnes of sediment transported for every million m^3 of water. In 2001, total discharge and sediment transport both dropped substantially. However, suspended sediment concentrations were higher and the sediment discharge to water discharge ratio increased from 561.2 in 2001, to 652.8 in 2002. Sediment transport was lowest in 2002, while discharge only reached the lowest value of all years in the following year. In 2003 and thereafter, annual sediment transport gradually increased, despite generally low but increasing discharges on the Mfolozi River. The sediment / discharge ratio in 2004 and 2005 were 1068.4 and 1423.1 respectively.

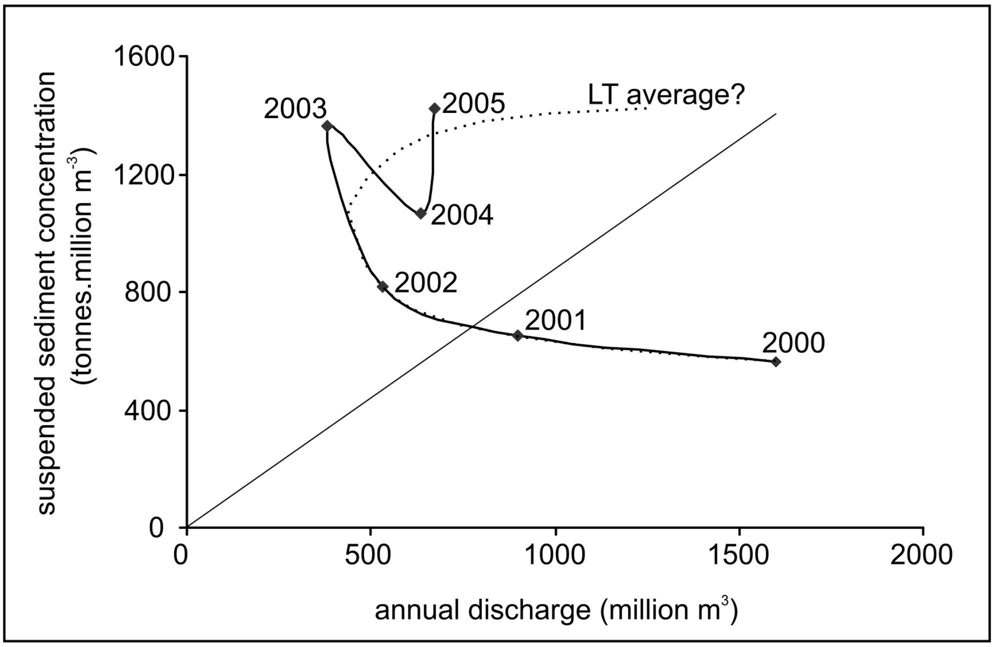

Figure 11: Annual discharge compared to mean annual suspended sediment concentration from 2000 to 2005, plotted against the linear trend line. The dotted line represents the probable long-term average.

Thus, in addition to the existence of an annual hysteresis loop, there is substantial variation in sediment transport on an inter-annual scale. Indeed, it appears that there may be hysteresis effects in the long term (Figure 11). Unfortunately, additional data would be required to confirm this relationship. Nevertheless, Figure 11 shows that during 2000 and 2001, annual discharges were greatest, but sediment concentrations were comparatively low. Contrastingly, between 2002 and 2005, sediment concentrations were high relative to annual discharge.

Since it has already been shown that variations in sediment transport on the seasonal scale are related to sediment availability, it follows that inter-annual variation is also related to temporal changes in sediment availability, except on a longer time scale. Tucker and Slingerland (1997) suggest that increases in run-off and decreases in vegetation have a similar impact on drainage basins, with both resulting in sudden increases in sediment supply through expansion of the channel network. As such, many authors assume a relationship between precipitation and vegetative land cover (e.g. Coulthard and Macklin 2001). In the Mfolozi River catchment, persistent increases and decreases in precipitation would result in changes in vegetation cover. Following a dry period, one would expect sediment concentrations to suddenly increase as sediment accumulated on slopes becomes available for transport by flowing rivers.

Long-term variability in sediment transport may therefore be related to similar factors as discharge variability. Firstly, variability in precipitation directly impacts on the amount of sediment that is likely to become available for transport in any year. And secondly, variability in catchment run-off (not stream flow) impacts upon how much sediment is likely to reach a stream channel for transport. Run-off variability is impacted upon by several factors such as antecedent conditions, vegetation cover, and high evapotranspiration demands that give rise to variability in southern Africa and Australia as is described by Peel *et al*. (2001).

Since sediment transport variability and stream flow variability have similar roots in high evapotranspiration demand, one should expect that rivers with variable discharges be characterised by variability in sediment transport that exceeds variations in discharge.

In this case, variable hydrology and variable sediment transport should go hand in hand.

4.5. Sediment yield

In the past, inter- and intra-annual sediment transport variability has not been acknowledged, with the result that suspended sediment loads have been overestimated. Previous estimates of suspended sediment loads were based on catchment size and run-off estimations, rather than direct measurement. Considering our current understanding of the relationship between sediment availability, run-off, discharge and resulting hysteresis, the use of catchment factors and run-off to estimate suspended load is awkward. Lindsay *et al.* (1996) estimated a suspended sediment transport of $1.24 \times 10^9 kg.a^{-1}$, which was based on measurements of suspended sediment on one day in January. Rooseboom (1975) estimated suspended sediment transport at 2.36×10^9 kg.a$^-$1. The current estimation, based on the relationship of turbidity and sediment concentration over a 6-year period is 6.8×10^8 kg.a^{-1}. This translates to an average suspended sediment yield of 61t.km^{-2}a^{-1}. Thus, Lindsay *et al.* (1996) and Rooseboom (1975) exceeded the current estimate by 560 million and 1680 million kg.a^{-1} respectively.

In global and southern African terms, sediment discharge from the Mfolozi River is extremely small, contributing an average of 0.68 million tonnes annually. In comparison, the Orange and Zambezi Rivers contribute 17 and 20 million tonnes each year respectively (Milliman and Meade 1983). However, a comparison of global sediment yields versus discharge (from Milliman and Meade 1983) and data for the Mfolozi River, suggests that the sediment yield is relatively high considering discharge. Overall though, it appears that the misconception that sediment transport is not variable over longer time periods, and that the Mfolozi River transports sediment in the same manner as regular rivers, has led to overestimations of sediment transport.

4.6. Variable rivers: A new geomorphology?

The Mfolozi River, a variable river in terms of hydrology, may also be considered to be variable in terms of sediment transport. As emphasised earlier, variability in sediment transport is unrelated to changes in discharge. Since the causes of sediment variability and stream flow variability are linked, it seems likely that most variable rivers are characterised by sediment transport variability. Sediment transport can be seen as an indication of a rivers capacity to do geomorphic work, depending on the definition one adopts. It follows then, that variable rivers may not only be characterised by variability in stream flow and sediment transport, but also by variability in terms of geomorphic change.

While the concept may seem outlandish, the approach has already been adopted by geomorphologists to understand variable dryland rivers. It is accepted that variable stream flow has recognisable impacts on the geomorphology and ecology of such river systems. For instance, variable flows in dryland rivers sometimes results in river channel sizes that are not comparable to mean flows, but are rather designed for flood flows. This is largely because dryland rivers are sensitive to the effects of large floods in that they are limited in their capacity to recover (Tooth 2000).

If we accept variable hydrology as a driver of geomorphology in dryland environments, can the same not be said for variable rivers that are not in dryland environments? It is not argued here that dryland rivers should be considered as variable rivers, but rather that all variable rivers may exhibit different geomorphic processes, features and rates of processes than may be expected for more regular rivers with frequent and predictable flood pulses.

Flow on the Mfolozi River is usually low and impounded by high levees on the coastal floodplain. While the actual proportion of the time that the Mfolozi River may have exceeded bank height is unknown due to management of the floodplain for sugar cane cultivation, it is known that for the majority of every year, flow is a mere trickle down the Mfolozi channel. Stream flow is characterised by ‘normal’ low flow years that are frequently below the long-term median. However, in addition to long periods of low flow, the frequency distribution of discharge also indicates a tail of high flows. It is these large flood outliers that are likely to mark periods of geomorphic change, for the same reason as in dryland rivers. Years between large flood events lack the capacity to ‘heal’ or ‘undo’ features created by flood flows.

The extreme flood event of 1984 on the Mfolozi River may be used to explore what is meant by flood event features. The 1984 floods followed the unusually far southward movement of tropical cyclone Domoina, resulting in discharges greater than three times the 100-year flood recurrence interval of the Mfolozi River. These flood discharges resulted in the river avulsing towards the south of the floodplain and depositing a lobe of sediment 3km wide and 10km long, and averaging 5m deep (Chapter 4). The flood event has permanently altered the morphology of the upper floodplain region, and without human intervention, would have resulted in the development of a new Mfolozi River course. Furthermore, since 1984, the effect of floodplain deposition on surface topography and floodplain dynamics has been minimal. It appears then, that on variable rivers, where sediment transport may also be variable, major geomorphic change occurs in spurts corresponding with 'outlier' flood events. Furthermore, geomorphic change in intervening years, as a result of lower recurrence interval floods, is generally negligible.

5. Conclusion

The Mfolozi River may be described as a variable river hydrologically. In addition to this variability, the river is also characterised by sediment transport variability on the intra- and inter-annual scale. More data is required to clarify how sediment transport changes in the long term and to confirm whether there is indeed long term hysteresis on the Mfolozi River. However, as it stands, it can be readily seen that sediment transport variability is not related to changes in discharge. This variability raises questions about studies that use the sediment rating curve approach to determine sediment transport regimes on hydrologically variable rivers.

However, the main aim of this chapter was to call for recognition of variable rivers as geomorphically distinct from more regular rivers, on the basis of variability in geomorphic work. More research on the geomorphology of variable rivers, with particular emphasis on process rates, is required to ascertain the validity of this claim.

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Chapter 4. Geomorphology and sedimentology of the lower Mfolozi River Floodplain, KwaZulu-Natal, South Africa

Abstract

The Mfolozi River Floodplain and estuary are located in the subtropical region of northern KwaZulu-Natal, South Africa. Longitudinal valley profiles of the Mfolozi River Floodplain indicate the existence of four main geomorphic zones, differing in morphology and dominant processes. Confinement of the Mfolozi River above the floodplain has led to the development of an alluvial fan at the floodplain head, characterised by relatively high sedimentation rates and avulsion frequencies with a gradient of 0.10%. The lower floodplain is controlled by sea level, with an average gradient of 0.05%. Between the two lies an extremely flat region with an average gradient of 0.02%, which may be controlled by faulting of the underlying bedrock. The upper floodplain is dominated by fine to medium grained sand. There is a general downstream fining in median particle size. This trend is somewhat interrupted by the occurrence of abandoned alluvial ridges and inter ridge depressions that are regions of fine sediment accumulation.

Avulsion occurrences on the Mfolozi Floodplain are linked to the two main zones of aggradation, the alluvial fan and lower floodplain regions. On the alluvial fan, normal flow conditions result in scour due to local steepening. However, during flood conditions the channel becomes overwhelmed, resulting in large scale deposition and frequently avulsion. Contrastingly, on the lower floodplain, reaching of the avulsion threshold is not necessarily linked to large flood events, but rather to long-term aggradation on the channel that decreases the existing channels gradient while increasing its elevation above the surrounding floodplain. Thus, it appears that the impact of flow variability on geomorphic processes may be varied across the floodplain. Infrequent large floods are primary drivers of geomorphic change in the upper floodplain region, but have little impact on long-term geomorphic processes of the lower floodplain.

1. Introduction

The origin and evolution of floodplain wetland systems are varied, both in a global and a southern African context. The evolution of many floodplains in southern Africa is attributed to the Tooth *et al*. (2004) model, whereby an erosional surface becomes superimposed on a resistant lithology of localized extent that retards incision upstream. Stream energy is instead used by laterally planing the underlying bedrock and thus widening the valley. In these wetlands, rivers generally comprise mixed bedrock-alluvium, and sedimentary infill is minimal. The most important aspect of this model is that even though incision may be momentarily paused, in the long term, net erosion of the drainage line is ongoing. This is largely because of a peculiarity in the physiography of the southern African subcontinent, whereby two periods of uplift were experienced approximately 20 and 5 Ma. Uplift in the eastern region of southern Africa was 250 and 900m for the two periods of uplift and was not as great on the western side of the subcontinent. These uplift events created an anomalously high continental region that is tilted towards the west (Partridge and Maud 1987). The result has been the widespread rejuvenation of river networks, and the concomitant incision of the subcontinent. Thus, almost all rivers and wetlands in the region are predestined to erosion. Superimposed on these uplift events are variations in sea level caused by global climate changes associated with variation in the amount of water in the solid phase trapped in polar ice caps.

Variations in sea level have greater impacts on freshwater wetlands in close proximity of our present coastline than those located further inland. In general, coastal wetlands degrade during periods of low sea level but they undergo renewed formation and growth as sea level rises. The Mfolozi Floodplain owes its origin to a rise in sea level, from 120m below present sea level since the last ice age (Figure 1 from Ramsay 1995). Following the Last Glacial Maximum, sea level rose relatively steadily at 8 $mm.yr^{-1}$, until 8000BP. Sea level continued rising, reaching a highstand of +3.5m approximately 4480BP, consistent with the early Holocene warming documented in Antarctic ice cores (Masson *et. al*. 2000). 3880BP sea level regressed to its present level, remaining stable for approximately 500 years (Ramsay 1995). A neoglacial event occurred between 3200 and 2500BP, resulting in global sea level lowering, and cooler and dryer conditions in the Maputaland region (Talma and Vogel 1992, Ramsay 2005). This cooling event caused extensive peat fires in Lake Futululu, a drowned tributary in the northern portion of the Floodplain (Chapter 5). Following a second short-lived highstand, sea level reached its present elevation 900BP, remaining stable since then.

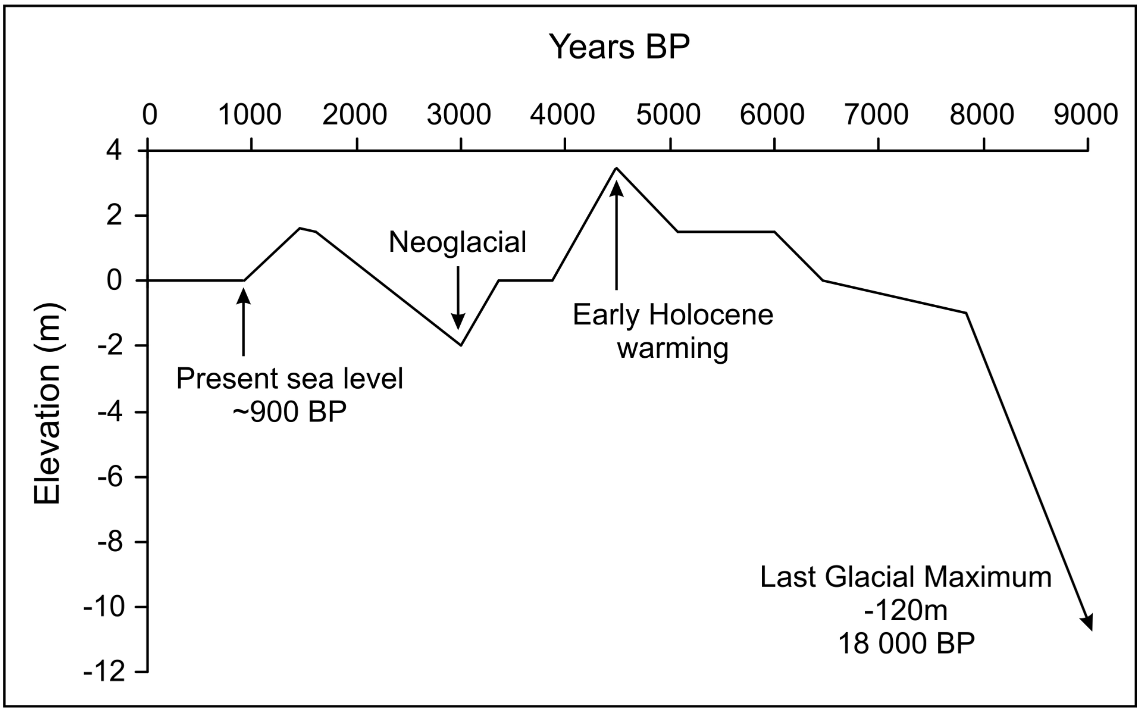

Figure 1: Holocene sea-level variation along the southern African coast (from Ramsay 1995).

This chapter explores the evolution of a subtropical floodplain located on the coastal plain of the eastern seaboard of the northern KwaZulu-Natal coast, South Africa (Figure 1). The catchment is characterized by a steep hinterland, with the headwaters in the Drakensberg Mountains at an altitude of 3000m. In addition, the KwaZulu-Natal region is South Africa's wettest, although the northern Maputaland region in which the study area is situated is slightly dryer. Rainfall in the Mfolozi Floodplain region averages 1090mm/a, declining to just 645mm/a towards the upper watershed. Potential evapotranspiration on the floodplain is 1805mm/a (Schulze 1997).

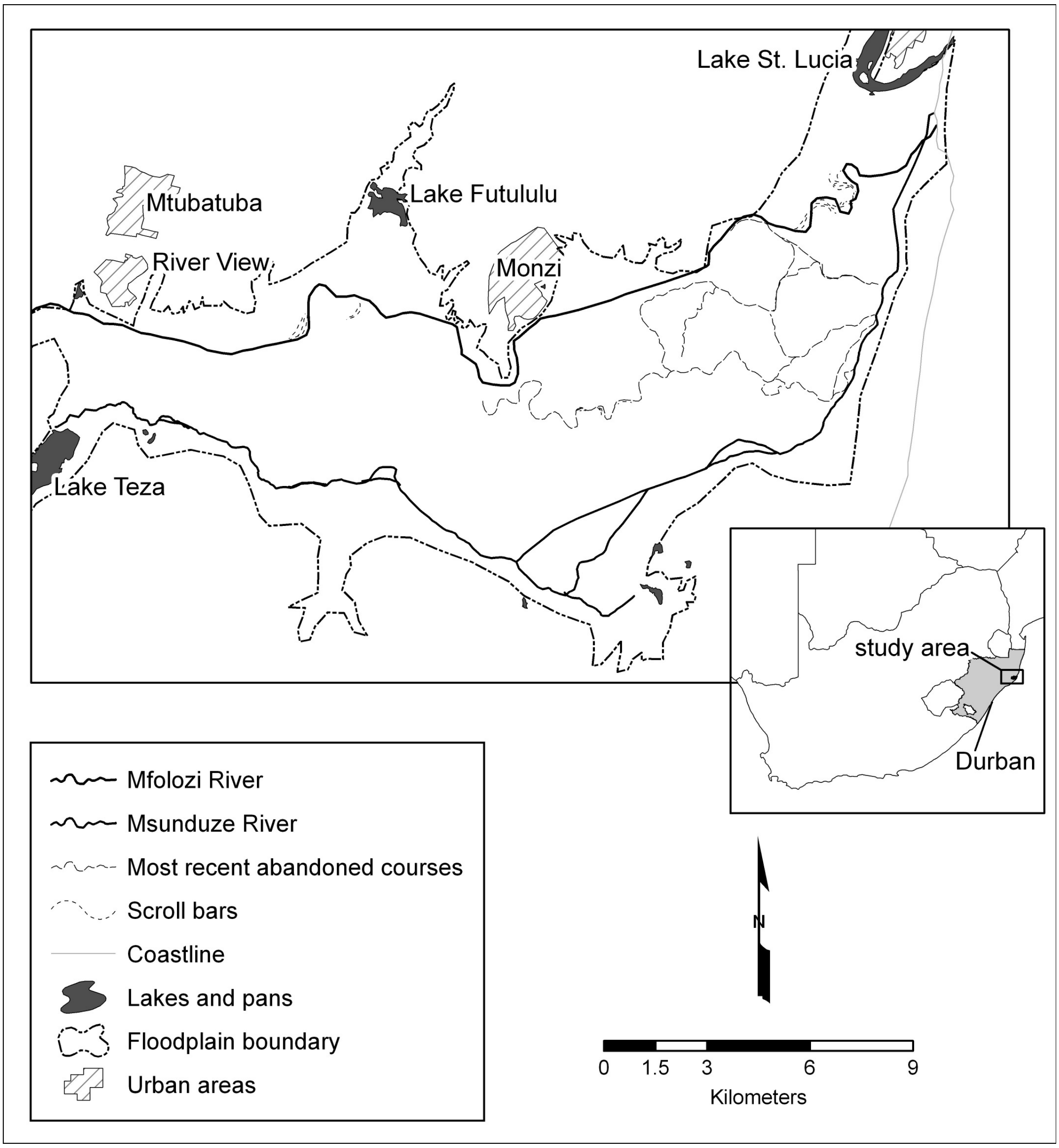

Figure 1: General morphology of the Mfolozi Floodplain, its location is shown on the inset.

The floodplain is one of South Africa's largest, approximately 30km long from its head at the foothills of the Lebombo Mountains to its entry into the sea just south of Lake St. Lucia. The floodplain is bordered on its southern and northern sides by relatively steep outcrops of Zululand and Maputaland formation rocks, while the Maphelane dune cordon, which is over 100m in height, blocks its eastern boundary. Three large lakes occur on the floodplain periphery; Lake Teza towards the southwest, Lake Futululu in the north and Lake St. Lucia in the northeast. The floodplain comprises two rivers; the

Mfolozi River which drains a vast catchment extending to the Drakensberg mountains and flows eastward along the northern part of the floodplain, while the Msunduze River, which drains a localized catchment, flows along the southern edge before turning northwards at the Maphelane dune cordon. The Msunduze River joins the Mfolozi River near its mouth. Occasionally, long shore drift forces the Mfolozi – Msunduze estuary to combine with the Lake St. Lucia estuary to the north, such that they share a common mouth.

Two thirds of the floodplain is currently transformed from a natural state to commercial sugar cane cultivation, while the remainder falls within the iSimangaliso Wetland Park boundary. However, despite being in the Park, the lower third is extensively cultivated by small-scale subsistence farmers. Historically, the floodplain was a mosaic of permanent and seasonal herbaceous wetland, with *Cyperus papyrus* and *Phragmites australis* as the dominant plant species. Where vegetation has not been cleared, *Ficus trichopoda* trees occur on the banks of abandoned and active channel courses.

In general, fluvial geomorphologists have tended to overlook floodplain geomorphology and sedimentology (Dollar 2000) and there are thus substantial gaps in our understanding of these systems, especially with respect to subtropical floodplains. To date, research conducted on the subject has tended to focus on major anastomosing or meandering river systems situated in North America, such as the Saskatchewan River (Cazanacli and Smith 1998, Morozova and Smith 2000; 2003), Columbia River (e.g. Makaske *et al*. 2002) and Mississippi River (e.g. Gomez *et al.* 1997, Hudson and Kesel 2000, Kesel 2003). In addition, Törnqvist and Bridge (2002) and Törnqvist (1994) have compared sedimentation on the Rhine-Meuse and Mississippi alluvial floodplains. Rowntree and Dollar (1996) considered channel instability in the Bell River with respect to bed-material size and riparian vegetation. These studies, excepting that on the Bell River, have tended to focus on the evolution of anastomosing systems and single channelled rivers, as well as the development of crevasse splay complexes. Furthermore, most are characterized by substantial peat formation in the floodplain. Contrastingly, the Mfolozi River Floodplain is dominated by clastic sediments and backswamp areas are very seldom suitable for peat formation.

In addition, the Mfolozi River has highly variable flow. Furthermore, the Mfolozi River is not a dryland river, such as those described by authors such as Tooth (2000) and

Costelloe *et al.* (2003). This research therefore focuses on investigating how variable rivers in non-dryland settings may be different from more regular rivers in terms of their geomorphology and sedimentology, while also investigating system origin and evolution.

2. Methods

2.1. Floodplain morphology

The most recent set of 1:10 000 orthophotos with 5 metre contour intervals (2.5 metre accuracy) in conjunction with aerial photographs from the years 1937, 1960, 1970, 1988 and 1996, were used to map geomorphic features. The orthophotos were used to georeference and mosaic all the aerial photographs for each year in Arcview 9. This allowed the development of an historical time sequence in terms of land use, river course and behaviour, and other geomorphic elements.

A longitudinal valley profile and a series of north-south cross-sectional profiles were constructed from the orthophotographs. To obtain greater accuracy than could be obtained from orthophotographs, real time differential GPS surveying, using a base station and a GPS receiver, was used to survey river water and levee elevation over the length of the Mfolozi River. This allowed the construction of a river longitudinal profile accurate to 0.01m in the z-field. Core locations were also surveyed using this technology. The floodplain surface was mapped using a Trimble GPS with differential correction capabilities using a remote base station in Durban, with accuracy in the z-field of approximately 1m or less. Data collected using the Trimble GPS and the base station survey were integrated into a digital elevation model using the *topo to raster* facility in Arcview. The digital elevation model was constructed to indicate general surface elevation trends as well as it was possible to do, although insufficient data was collected in some inaccessible areas of the floodplain for it to be considered very accurate.

A series of cores and the surface topography of an abandoned channel were surveyed using an automatic level.

2.2. Floodplain sedimentology

2.2.1. Sediment sample collection

Fourteen cores were augered using standard auger equipment along a series of three systematic north south transects across the floodplain (B – upper floodplain, A –central floodplain and C – lower floodplain; Figure 2) until sediment could no longer be retrieved due to limitations imposed by equipment. Core lengths varied from 3.5 to 6.9m. Three additional holes were augured in order to obtain some sense of variation in sediment longitudinally on the floodplain (Core names UC, MC and LC). At the site of Core UC, 6 additional holes were augered in a transect across an old alluvial ridge.

At each auger site, the core was logged. Sediment samples were taken every half metre, or whenever there was an observable change in sediment characteristics.

Seventy-four surface samples were collected across the upper two thirds of the floodplain and a GPS point taken at each sample site. These were augmented with surface samples taken at each auger site.

2.2.1. Sample treatment

All sediment samples were treated in the same manner. Samples were oven-dried at 105^{o}C and then incinerated at 450^{o}C for 4 hours to measure organic content (Heiri *et al*. 2001). The remainder of each sample was used for particle size analysis. Coarse organic fragments were removed by sieving through a 2mm sieve and the sample was then ground using a pestle and mortar. If the sediment had a high organic content (>35%), organics were first digested using hydrogen peroxide and the sample was dried and crushed, taking care not to damage grains. The Malvern Mastersizer 2000, which employs laser diffraction methods, was used to determine particle size.

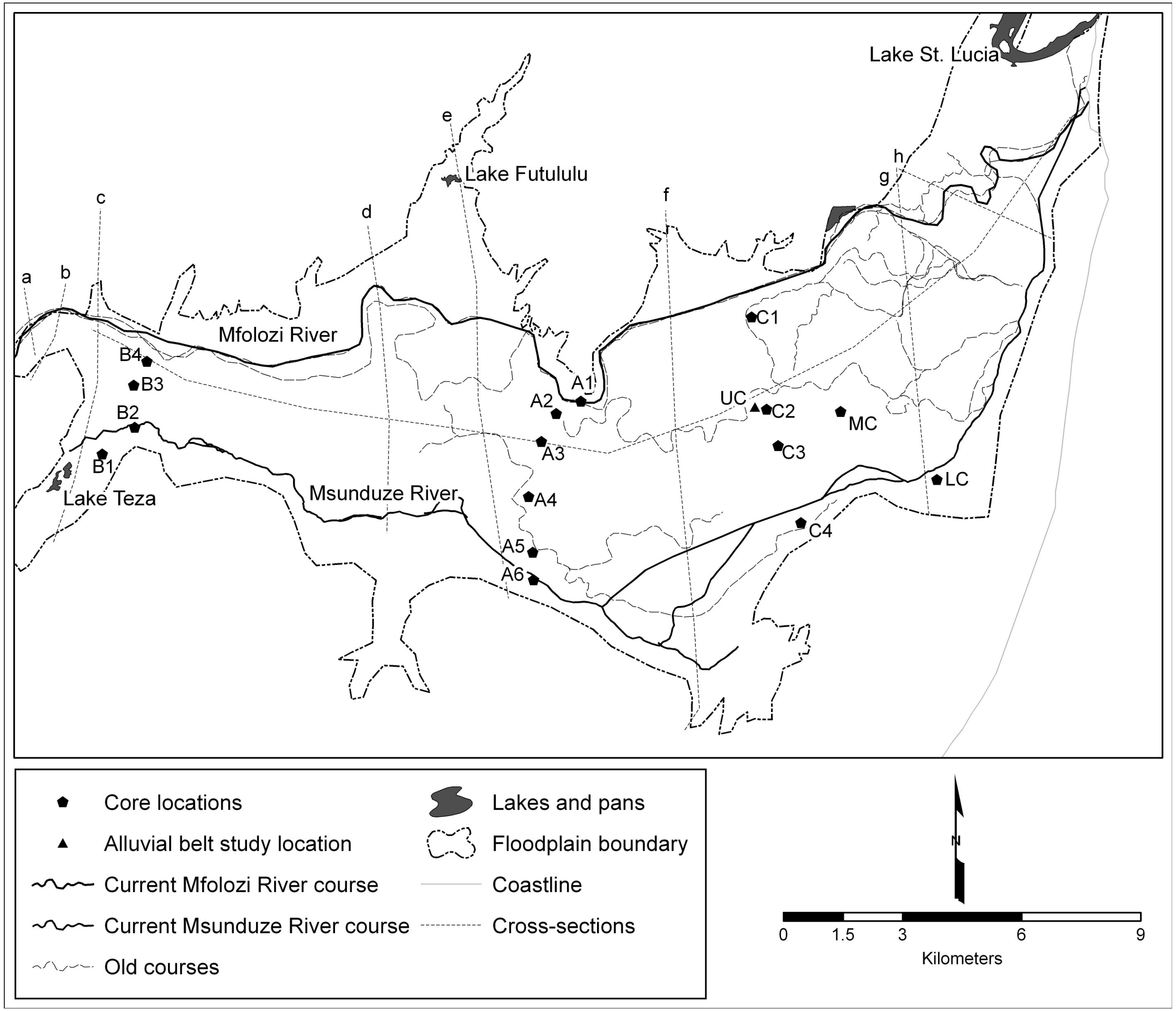

Figure 2: Location of sediment cores, orthophotograph cross-sections (a-h), orthophotograph longitudinal valley profile and alluvial belt study (UC).

3. Results

3.1. Floodplain morphology

3.1.1. Changes in floodplain characteristics over time

Aerial photograph coverage in 1937 was not complete as some areas of the lower floodplain were not photographed and clouds were present in others. Nevertheless, the available 1937 photographs provide useful information on the likely original appearance of the floodplain, even though channel straightening by agriculturalists had already commenced (Figure 3).

In the upper floodplain, the Mfolozi River course was relatively straight, and as far as can be understood from UCOSP archival material, it had not been straightened at this stage. In addition to the relatively straight course of the Mfolozi River in the upper floodplain region, this part of the floodplain was also marked by numerous southeastward trending lineations oriented away from the Mfolozi River. There were no visible abandoned channel courses in the upper part of the floodplain. South of Lake Futululu the Mfolozi River became sinuous with a series of large bends that ended immediately to the east of the peninsula of land extending into the floodplain from the north. This series of bends is known as the 'Uloa Loop'. In the western part of the Uloa Loop were indications of a meandering fluvial style in the form of a series of scroll bars on the banks of the convex channel bends, but evidence of this fluvial style in the upper floodplain region was limited to this area only.

Downstream of this, the channel straightened with some degree of sinuosity developing before the Mfolozi River joined the Msunduze River. Downstream of the confluence of the Mfolozi and Msunduze Rivers, the channel had the appearance of a large meandering stream.

Some abandoned channel courses were visible towards the lower floodplain. Although the coverage of the lower floodplain was incomplete, it seems likely that much of the area had indigenous wetland vegetation.

Sugar cane cultivation was generally restricted to the upper floodplain and to an abandoned alluvial ridge orientated towards the southeast. At this stage, the network of artificial drains and furrows designed to reduce waterlogging of soils was limited in extent. Lake Futululu, also on the northern floodplain boundary, was the largest floodplain lake, partially occupying a drainage line entering the floodplain from the north, and partially occurring along the floodplain boundary itself.

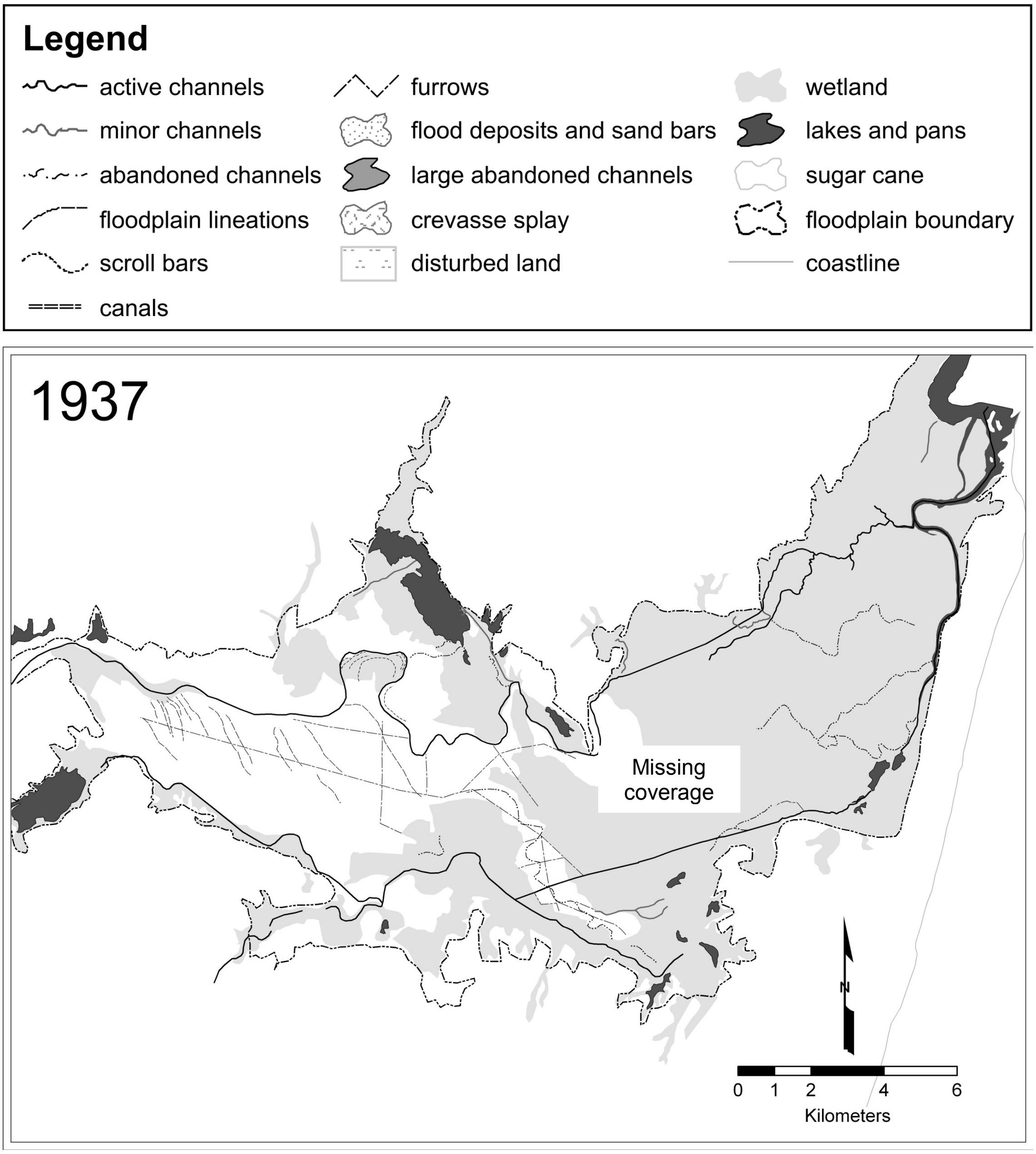

Figure 3: Geomorphic features and land use of the Mfolozi River Floodplain in 1937.

Lake Teza, on the southern side of the floodplain occupied the Msunduze drainage line entering the floodplain from the southwest. The lake was orientated with its long axis in the same direction as the drainage line. Further open water appeared in the region where the Msunduze River turned north adjacent to the Maphelane coastal dune.

The 1960 photograph indicates that more than 50% of the floodplain had been converted to sugar cane cultivation by this time (Figure 4). The Uloa Loop had been straightened to shorten the Mfolozi River course, presumably to increase the efficiency of water flow down the floodplain. The extent of Lake Futululu was substantially reduced from what it was in 1937 due to the disappearance of much of the southern part of the lake. The furrow network in the region of cultivation had also been extended in order to lower the water table and limit flooding of sugar cane during floods.

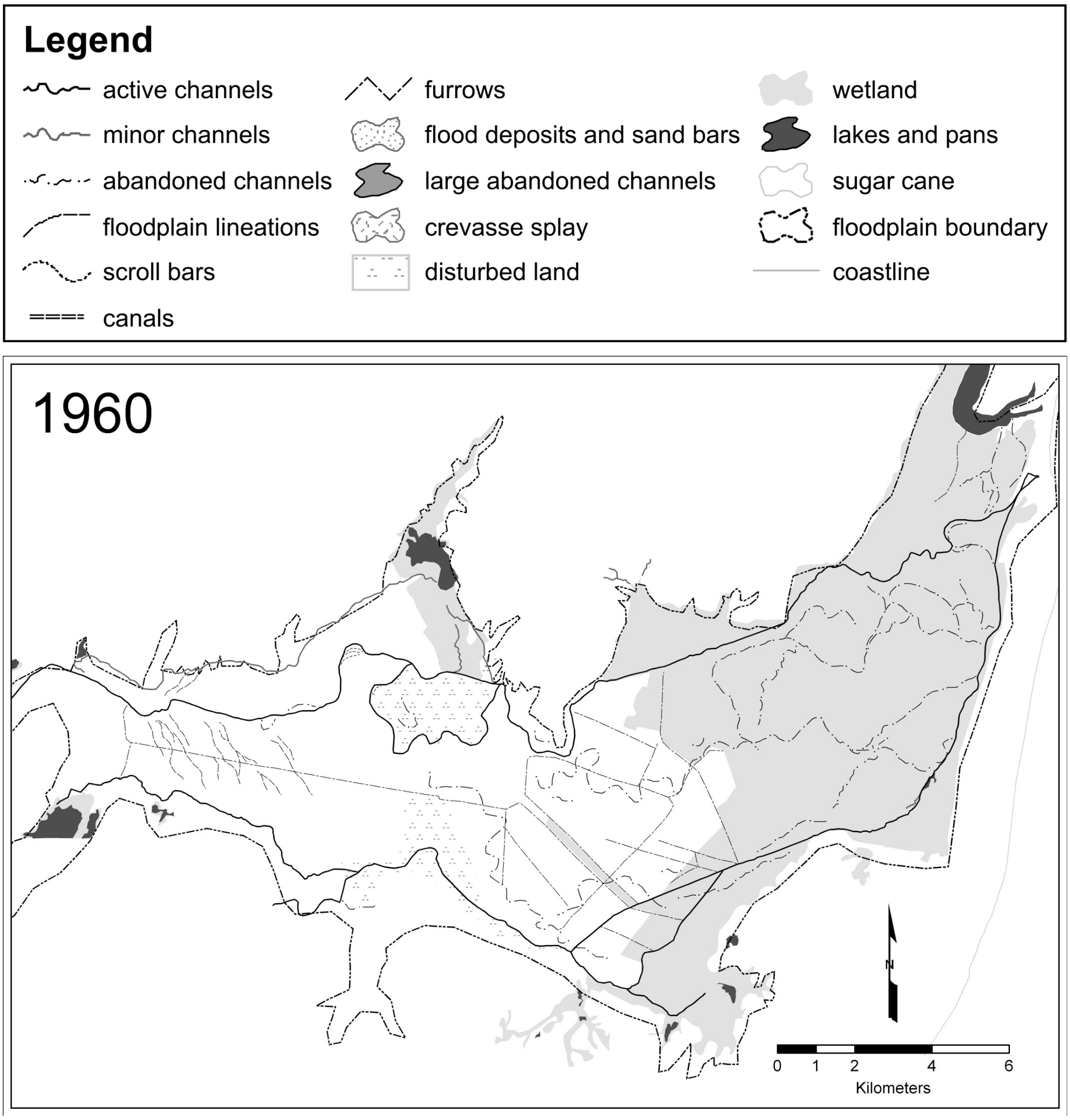

Figure 4: Geomorphology and land use of the Mfolozi River Floodplain in 1960.

As in 1937, the 1960 photography showed a straight Mfolozi River course on the upper floodplain, with many southeast trending lineations on the floodplain. Above the Uloa Loop, scroll bars were preserved. The full coverage of the floodplain in the 1960 photographs indicated numerous abandoned channels on the lower floodplain that were not visible in the incomplete imagery of 1937.

By 1970, sugar cane cultivation had been associated with increasing land reclamation towards the east (Figure 5), where a substantial network of drains was developed, presumably due to the fact that this area was very wet. In addition, a drain was formalized in the former bed of Lake Futululu in order to make cultivation possible. The southern boundary of Lake Futululu had moved marginally north. The upper floodplain was again characterized by numerous southeast trending floodplain lineations, as well as scroll bar development just west of the Uloa Loop. There was some evidence of attempting to straighten channels in the lower floodplain towards the Mfolozi River mouth.

Aerial photography for 1988 only covered the lower two thirds of the floodplain (Figure 6). Between 1970 and 1988, KwaZulu-Natal experienced the two largest floods on record up to this time. The first occurred in January 1984 as the consequence of a tropical cyclone, Cyclone Domoina, which moved southwards in the Mozambique Channel, reaching as far south as Durban. Discharge on the Mfolozi River peaked at 16 000$m^3.s^{-1}$, which equates to approximately 3 times the 100 year return period flood. Flood current velocities of 2.6 $m.s^{-1}$ were measured (Travers 2006). In September 1987, flooding once again occurred, this time as a result of a cut-off low pressure system in the interior of KwaZulu-Natal. The limited coverage of the 1988 aerial photography is thus very unfortunate. Nevertheless, the lower portion of a large depositional feature is visible on the western edge of the area covered by aerial photography. Van Heerden (1984) reported the avulsion of the Mfolozi River from its northern course into the southern Msunduze River near the floodplain head during the Domoina flood event. However, following the flood, agriculturalists returned the river to its original northern course. As such, there were no major river adjustments visible in the 1988 imagery.

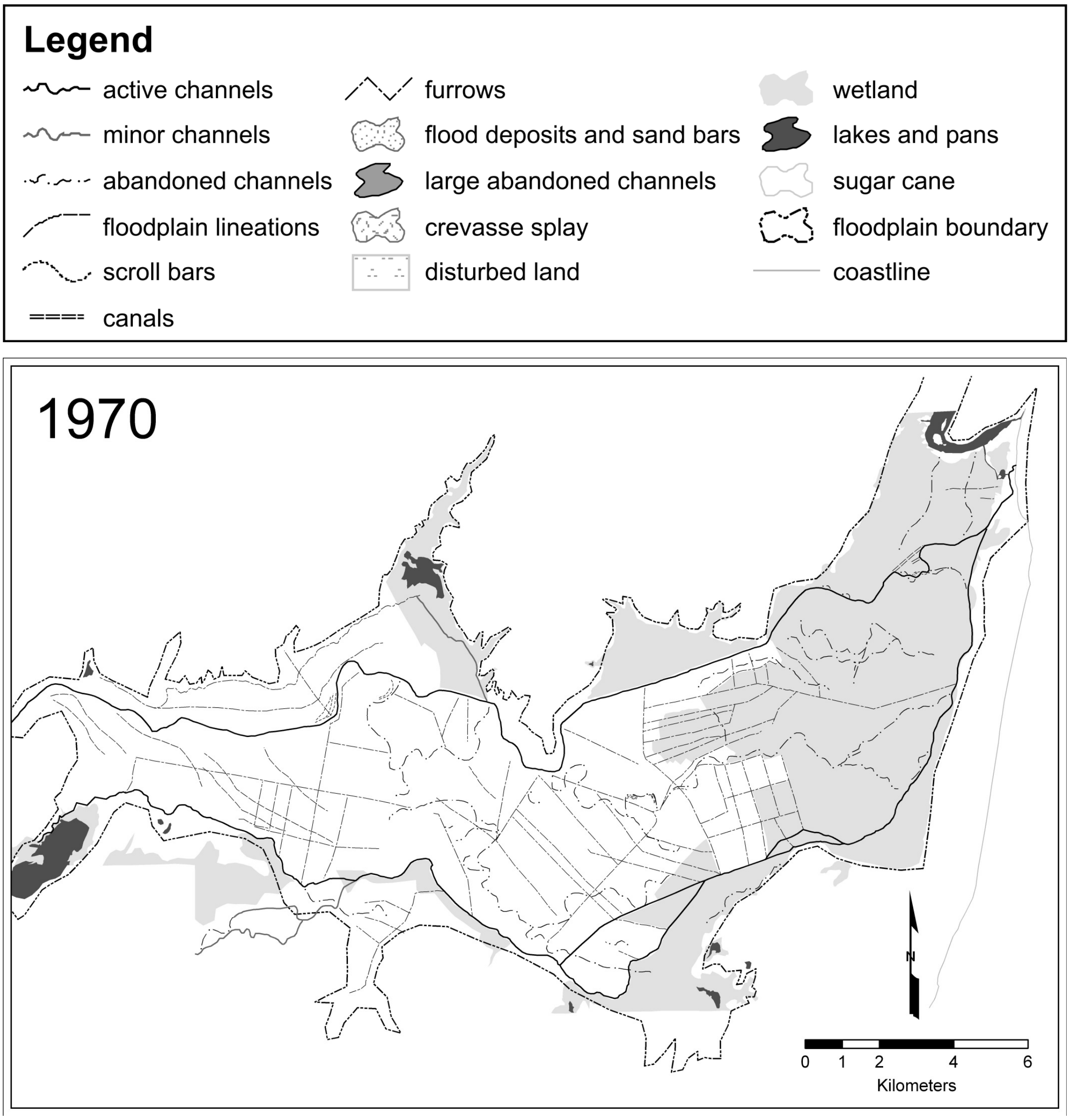

Figure 5: Geomorphology and land use of the Mfolozi River Floodplain in 1970.

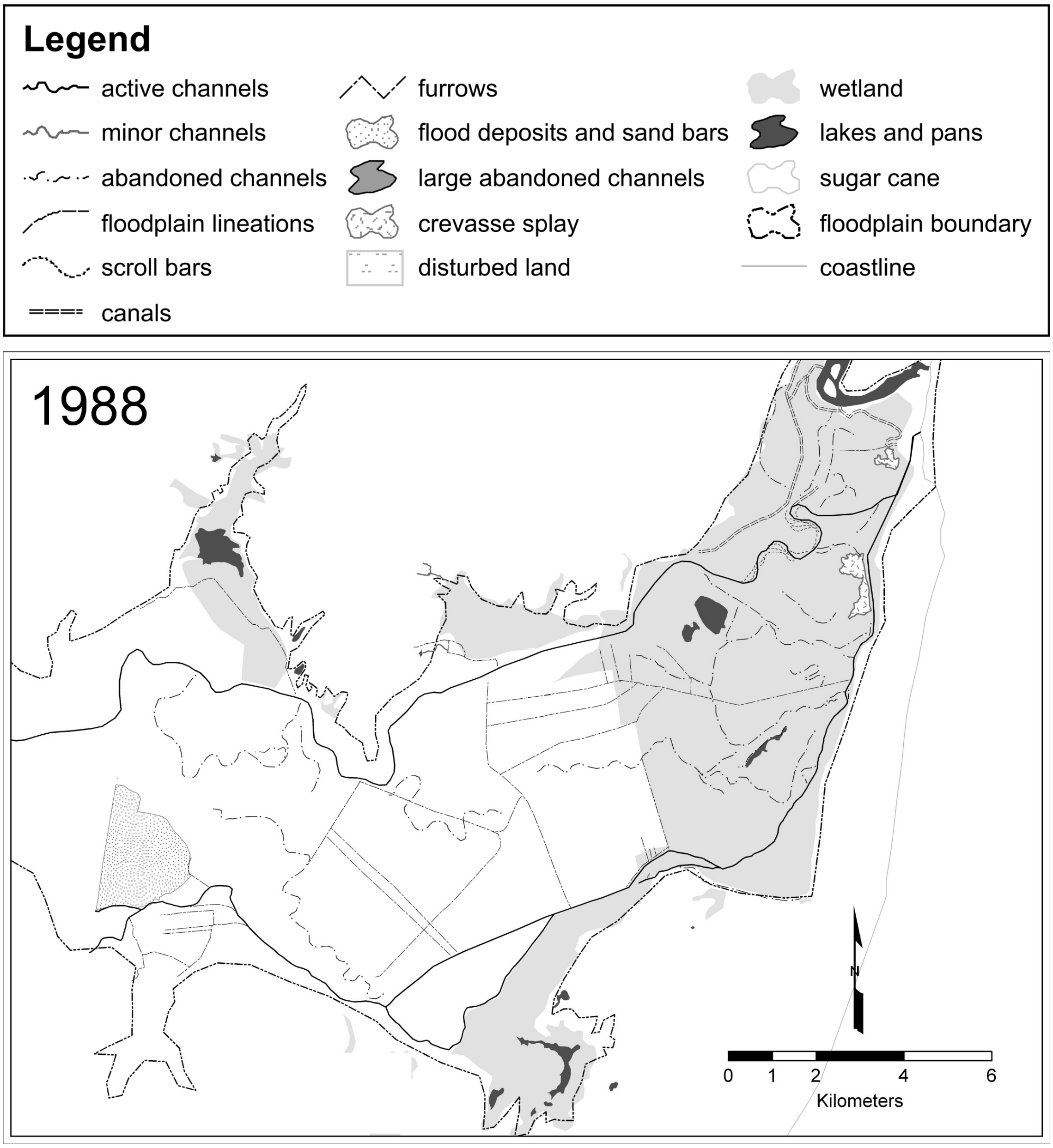

Figure 6: Geomorphology and land use of the Mfolozi River Floodplain in 1988.

In addition to the flood deposit, two crevasse splay complexes were on the Msunduze River upstream of its confluence with the Mfolozi River. A third crevasse splay had formed on the conjoined rivers, presumably also during flood conditions of either 1984 or 1987. Evidence of meandering was visible on the lower floodplain in the form of scroll bars. Several open water features were also visible on the floodplain, three of which occurred between abandoned river courses of the Mfolozi River. Between the imagery of 1970 and 1988, KwaZulu-Natal Wildlife officials built the Link Canal between Lake St. Lucia and the Mfolozi River estuary in an attempt to introduce fresh water to the lake on an ongoing basis.

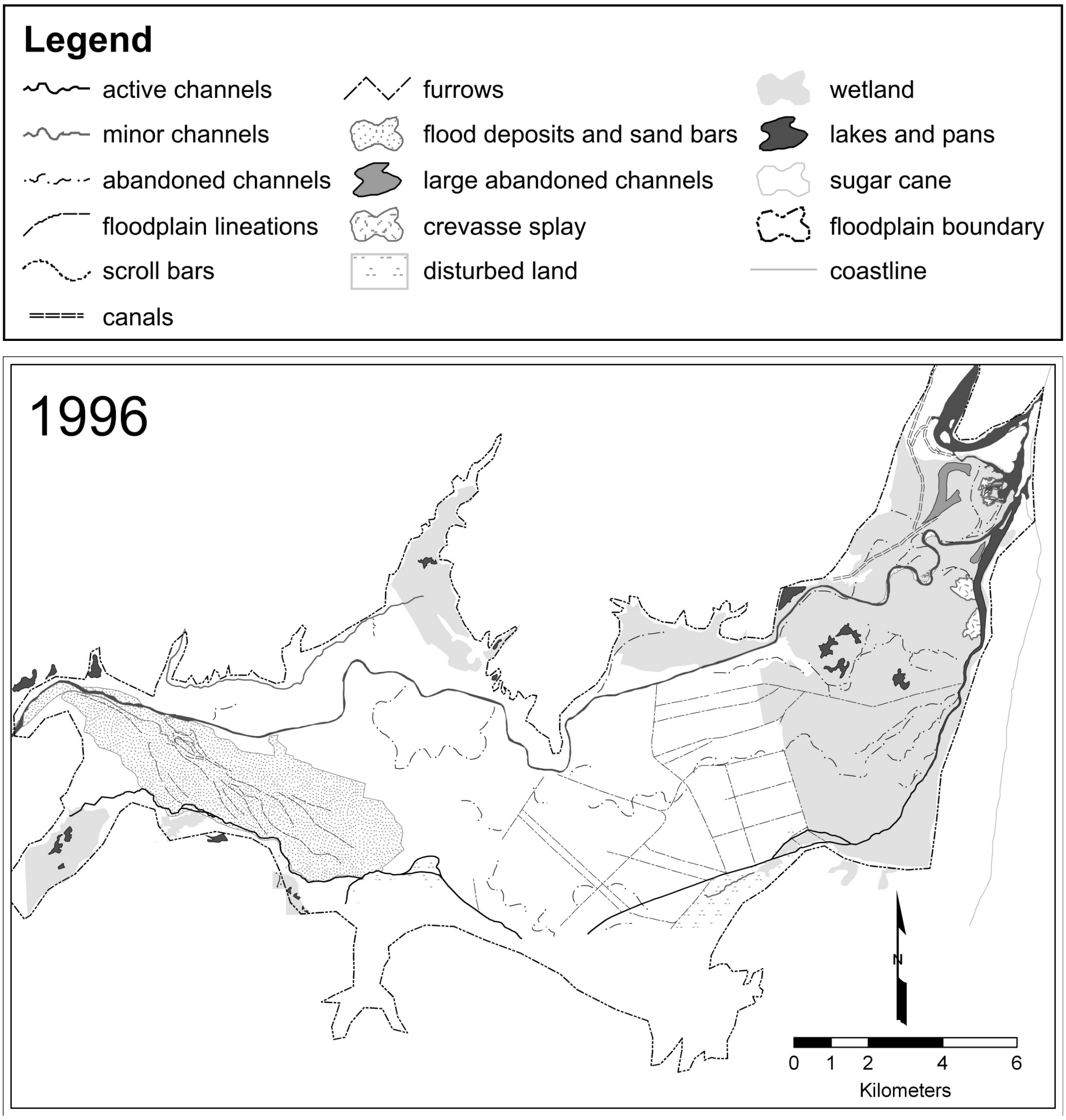

Figure 7: Geomorphology and land use of the Mfolozi River Floodplain in 1996.

Imagery from 1996 is the first generation of aerial photographs that fully depicts the extent of flood deposits from Cyclone Domoina in 1984 (Figure 7). A large lobe of sediment approximately 10km long and averaging 3km wide was deposited in a southeastward trending swathe across the floodplain. The deposit is riddled with similar trending lineations indicating distributaries. Sugar cane cultivation on this lobe of sediment had been abandoned, while cultivation had further encroached on the lower floodplain. The crevasse splay complexes visible in 1988 were still preserved, and the Link Canal was closed. Five lakes were visible on the lower floodplain.

By 2007, some of the sugar cane fields in the lower floodplain had been abandoned due to excessive wetness (Figure 8).

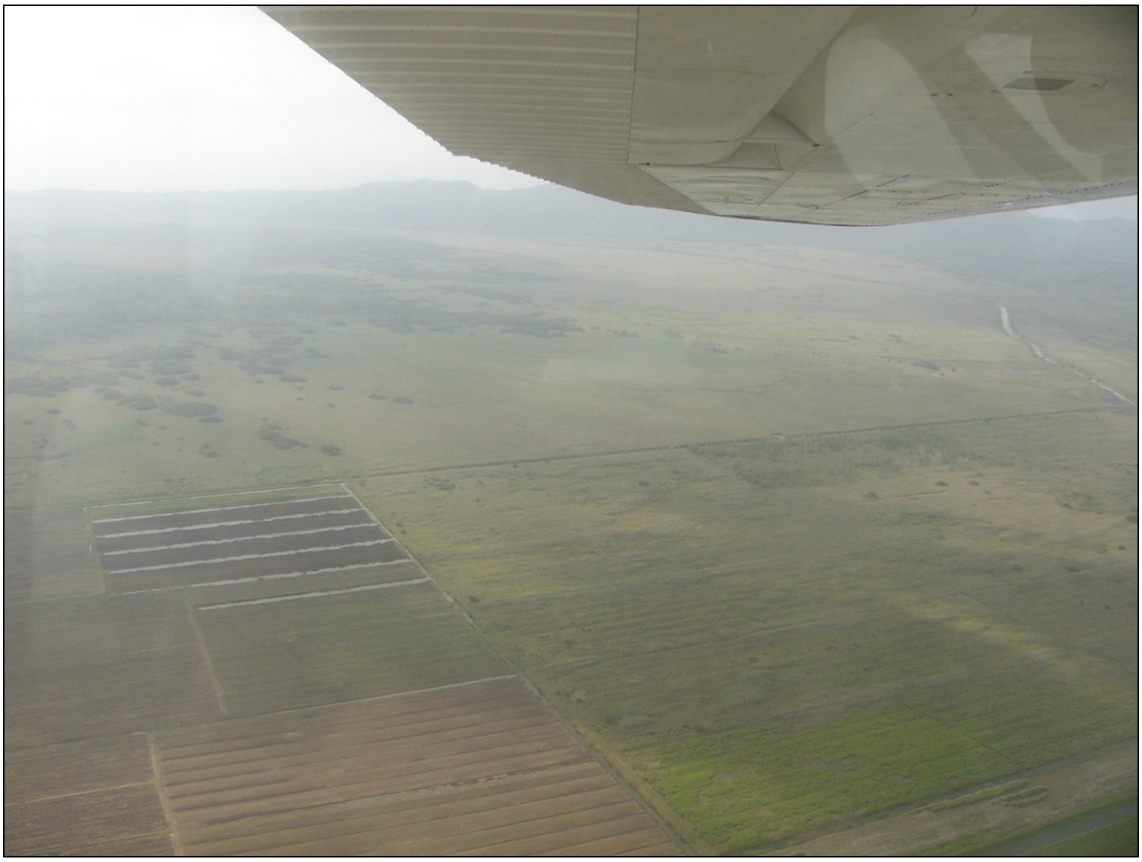

Figure 8: Abandonment of sugar cane farms on the lower portion of the Mfolozi Floodplain in August 2007.

3.1.2. Valley morphology

Floodplain cross-sections constructed using orthophotos and survey data are presented in Figure 9. In the uppermost region of the floodplain, the Mfolozi River is confined to a deep (> 60m), narrow valley of approximately 1.0km in width which narrows further (to 0.6 km) at the floodplain head (Figures 9a, b). Downstream of this confined reach the floodplain widens rapidly to a width of approximately 6.5km over a downstream distance of just 1.15km (Figure 9c). At this point, the floodplain is bounded by steep features of greater than 60m depth in the north and 17m in the south.

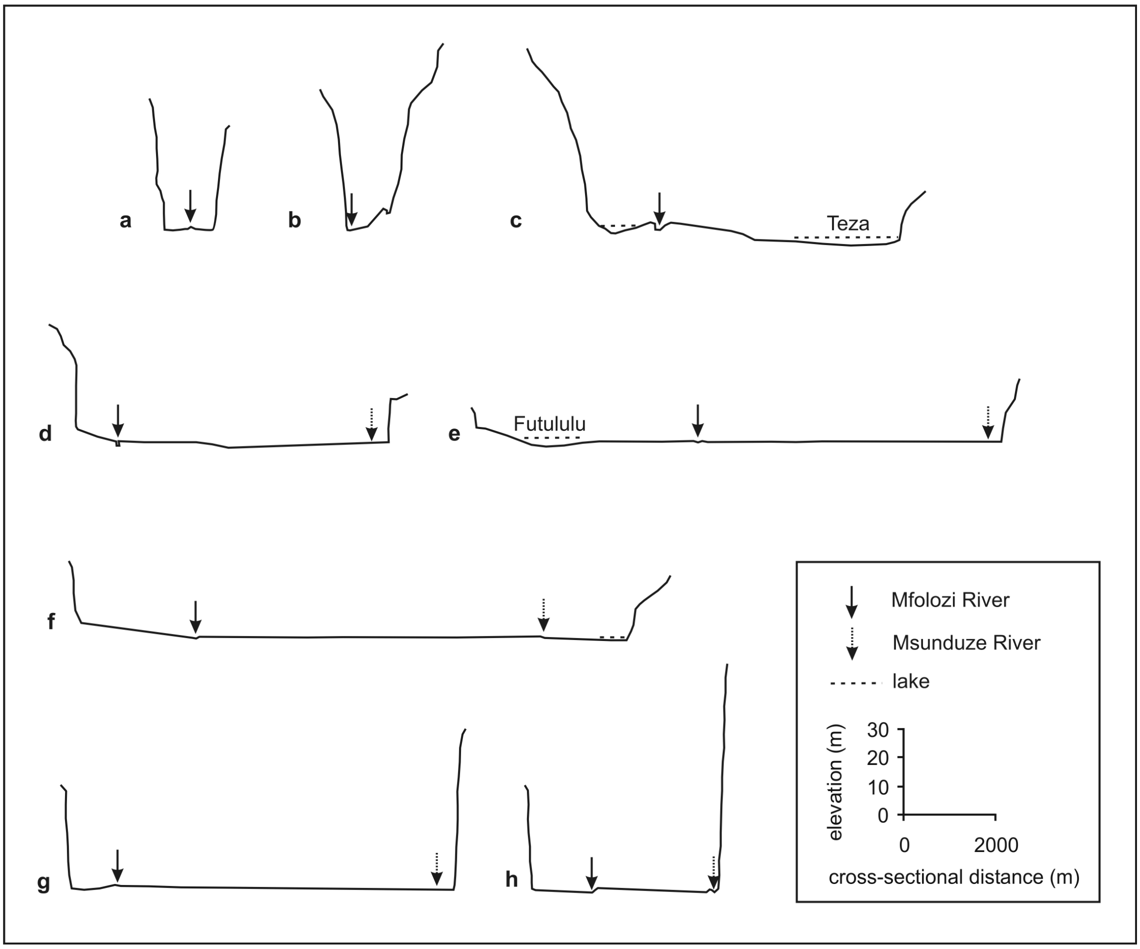

Figure 9: Valley cross-sections drawn from orthophotos, location of cross-sections indicated in Figure 2.

Also visible in the upper floodplain is the presence of an alluvial ridge containing the Mfolozi River, which is situated an average of 2-3m above the rest of the floodplain over a width of approximately 2.2km (Figure 9c). The depression to the south of the ridge contains Lake Teza, while a minor lake is situated in the depression north of the ridge. These two water bodies occur at different elevations, Lake Teza is flooded at an elevation slightly less than 14m amsl, while the northern lake, which is much smaller in extent, is situated at an elevation of 17m amsl. The alluvial ridge of the Mfolozi River extends eastwards on the floodplain, where it is set up against the high-lying ground adjacent to the floodplain (Figure 9d). At this point the floodplain reaches a width of 5.8

km, and the presence of the alluvial ridge can be clearly seen by the change in elevation approximately 2.3km south of the Mfolozi River. The Msunduze River occupies the lowest region of the floodplain.

The floodplain continues to widen downstream such that in the region of Lake Futululu it is 10.5km wide (Figure 9e). Lake Futululu occupies a depression less than 2m lower than the alluvial ridge. Apart from this, the floodplain is remarkably flat, with a fairly consistent elevation of approximately 10m. Here the Mfolozi River is centrally located on the floodplain while the Msunduze River hugs the southern floodplain boundary. The floodplain reaches its maximum width of 11.7km to the east of the Uloa Peninsula (Figure 9f), where it is characterised by a gentle north-south gradient. The southern edge of the floodplain contains a small lake, north of which is the Msunduze River. At this point, the floodplain is still confined within 2 steep sided escarpments with approximate elevations of 32m and 23m in the north and south respectively.

The lower floodplain, represented by Figures 9g and h, becomes progressively narrower and flatter such that the floodplain at the lowermost cross-section is approximately 4km wide. At this point it is also remarkably flat in cross-section. The Mfolozi and Msunduze Rivers are blocked from the sea on the eastern end by a large dune cordon that reaches 90m in height. This cordon is a depositional feature created by a northward moving system of longshore drift, forcing the river system northwards to a break in the dune cordon, which allows the Mfolozi to drain into the sea just south of Lake St. Lucia.

The river longitudinal profile as measured using differential GPS was completed (April 2005) when discharges were slightly above average due to good rainfall in the catchment in the weeks prior to conducting the survey. The water surface profile is generally concave, although there is substantial deviation from an idealised logarithmic longitudinal profile (Figure 10). At the head of the floodplain the gradient on the water surface is remarkably uniform at approximately 0.06%. Downstream of this region of uniform slope, the stream steepens (0.09%) and then flattens (0.03%) relative to the slope in the upper part of the floodplain. The stream then exhibits properties of a graded profile with a steady decrease in gradient downstream as far as the mouth. The profile is truncated, with an elevation of 0.6m measured at the estuary mouth.

The longitudinal profile of the bank is somewhat irregular with a linear slope in an overall sense (Figure 10). There is a region of greater than expected elevation in the region between approximately 32 and 18km from the mouth. The height of the levees above the water surface varied somewhat at the head of the floodplain, although all were generally between 4 and 6m high (Figure 10). From approximately 25km upstream from the mouth, there is a general decrease in levee height.

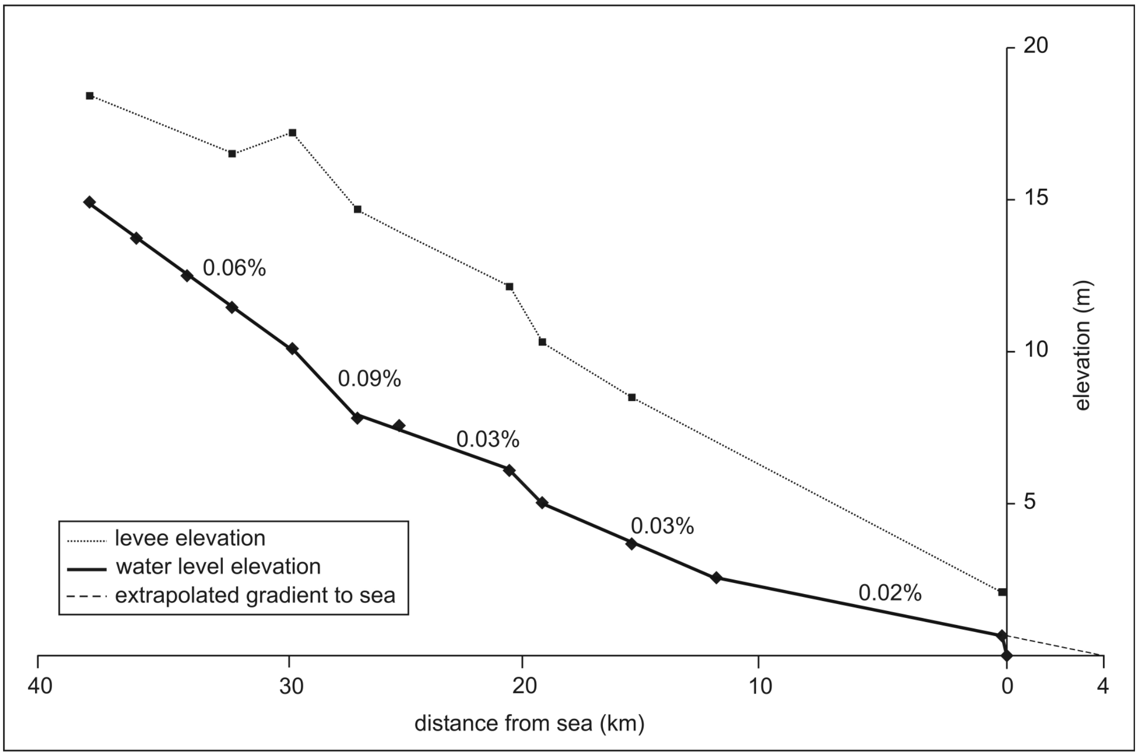

Figure 10: River longitudinal profile of April 2005.

3.2. Floodplain surface characteristics

Figure 11 depicts general trends in floodplain surface elevation, from which it is clear that the floodplain surface dips in an overall sense towards the ocean. At the floodplain head, the northern region of the floodplain along the current river course is at a greater elevation than the south. Thus, in the mid and upper floodplain, the local gradients on the floodplain surface are downwards towards the south and southeast. In addition, the alluvial ridge of the Mfolozi River is elevated above the surrounding floodplain in this region, with a height advantage of between 5 and 10m. Lakes Futululu and Teza occupy depressions adjacent to the alluvial ridge of the present day Mfolozi River course.

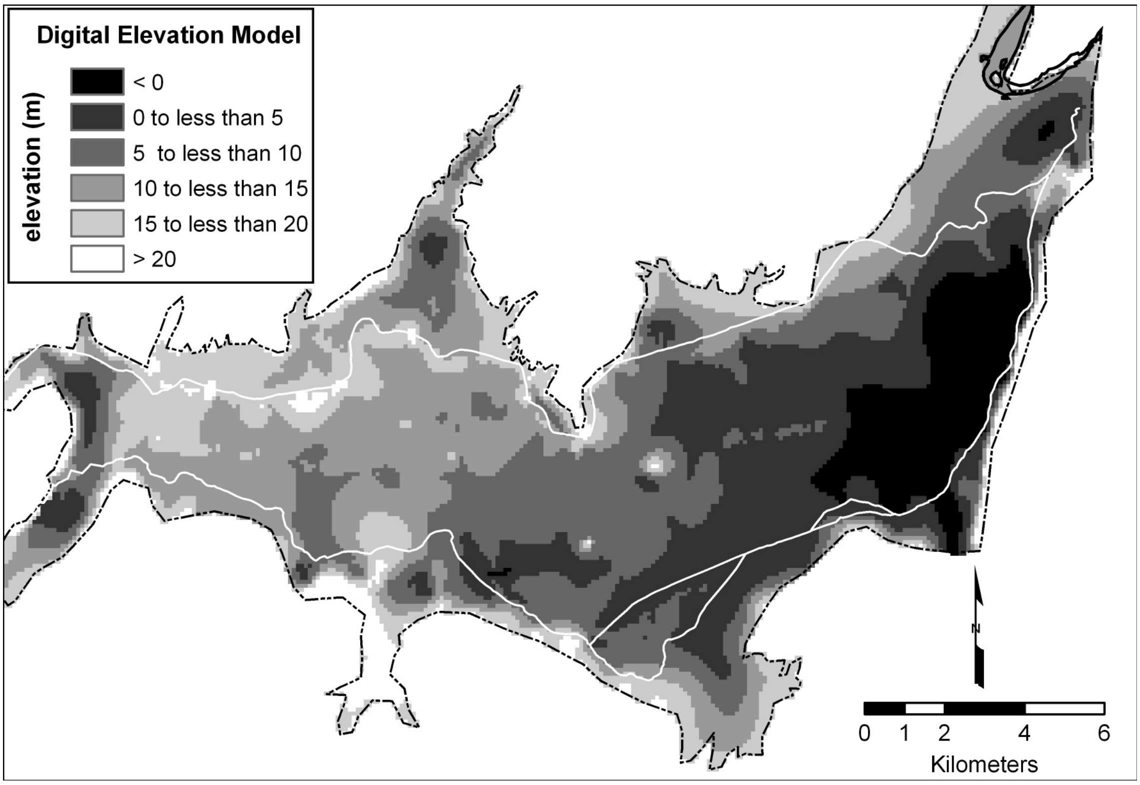

Figure 11: Digital elevation model created using differentially corrected Trimble GPS data and the *topo to raster* facility in Arcview 9.

Towards the central floodplain, two prominent rock pinnacles are visible as elevated features in the surrounding alluvium. Once again, the central floodplain is also characterized by a downward slope towards the southeast. However, there is substantial variation within this trend, with localized depressions and areas of high ground. There appears to be a weakly developed linear feature from the southernmost point of the Uloa Peninsula towards the south. Some portions of the lower floodplain appear to be below sea level, particularly along the lower Msunduze River. Lake Teza and Futululu are located in depressions, although the depression of Lake Teza extends further north than would be expected.

The median particle size was interpolated for the entire floodplain from surface samples (Figure 12). Silt-sized sediments dominate the floodplain surface, with the majority of sample medians being characterized as medium silt. There is substantial variation in particle size towards the head of the floodplain and less towards the lower floodplain in the east.

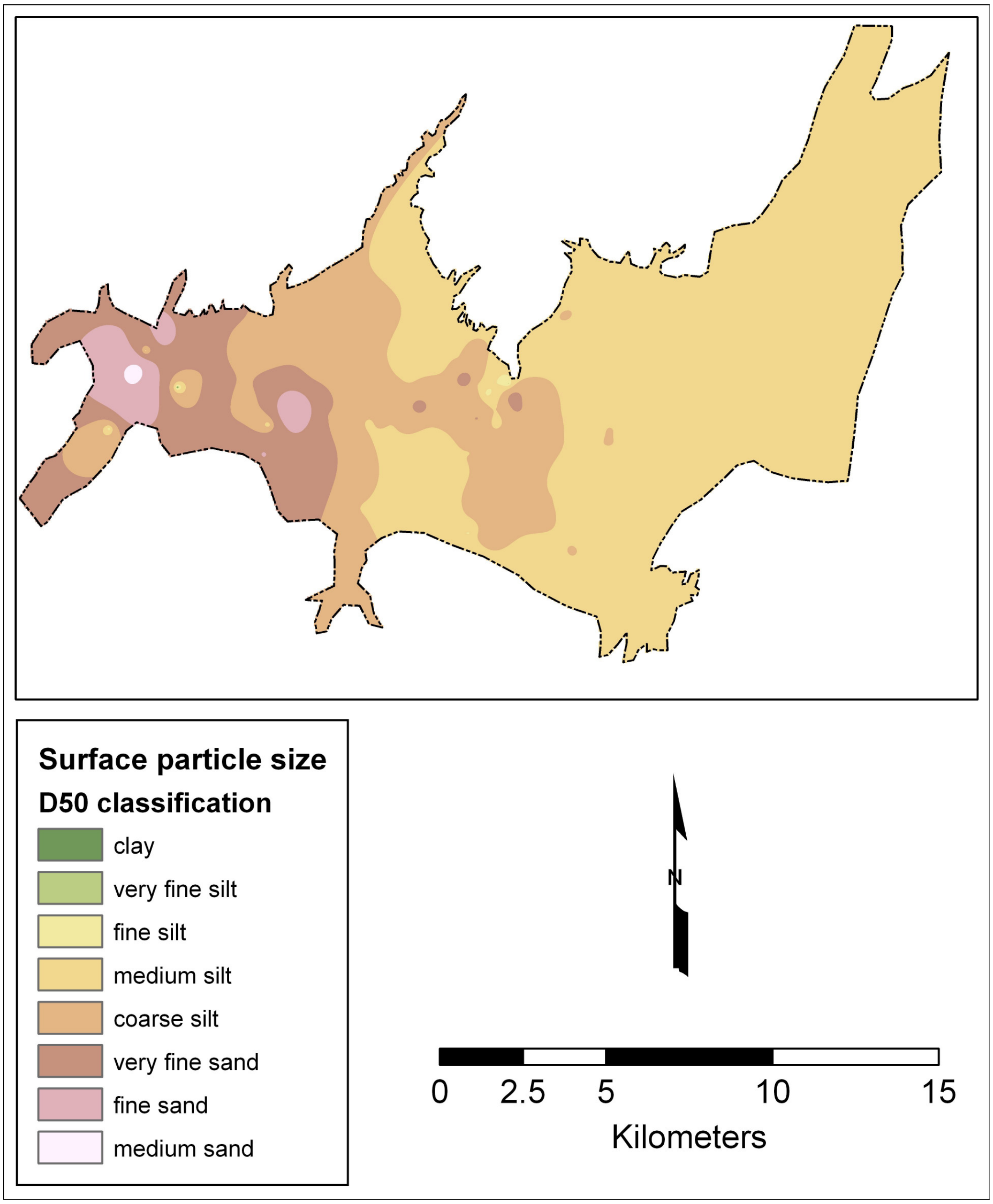

Figure 12: Interpolated median particle size on the floodplain surface, classified according to the Wentworth-Udden scale.

The upper floodplain is the coarsest in terms of particle size, with median particle size varying from coarse silt to medium sand. Surface grain size is not associated with the current day alluvial ridge at the floodplain head, but coarser sediments (very fine sand) are roughly orientated as a linear feature from the point where the Mfolozi River enters the floodplain towards the south-east. An additional linear feature of coarser sediments, this time predominantly of coarse silt size, is located with a southwards orientation in the central floodplain.

3.3. Floodplain sedimentology

The uppermost transect of cores (Transect B) consisted of four holes orientated roughly north-south, with B4 in the north and B1 in the south. The corresponding water table of the transect slopes away from the highest point on the floodplain surface at B3 towards the Mfolozi River in the north, and Lake Teza in the extreme south (Figure 13).

The northernmost core, B4, was augured on the southern levee of the Mfolozi River. Coarser sediments of coarse and medium grained sand mark the base of core B4. This lower unit is overlain by a varied sequence that is generally upward coarsening, and then upward fining from 16m amsl. However, within these general trends, the sequence is interrupted by a number of coarser (fine to very fine sand) and finer (coarse to medium silt) layers that are typically thin. Ash contents generally varied in a similar fashion to particle size, with the coarser layers having the highest ash contents. Nevertheless, there was some variation in the percentage of organic material, with a peak being reached towards the surface at 15-20 cm depth, with an organic content of 7.65%. Similarly, at 2.2m depth, an organic content of 7.4% was recorded.

Core B3, which was located towards the south of B4, showed very little variation in terms of particle size with depth. The entire core was relatively coarse, comprising a sequence of medium and very fine sand. The core consisted of a lower upward fining sequence, followed by an upward coarsening sequence, and a second upward fining unit. The upper part of the core comprised an upward fining sequence of medium sand. The percentage organic content was relatively consistent throughout, with a maximum ash content of 2.3% at 1m depth.

Core B2 may be divided into two sequences, a lower upward fining unit from 4m to 1.9m depth. Between 1.9 and 2.8m depth the sediment is much finer than the rest of the core, comprising fine and medium silt. The upper part of the core comprises an upward coarsening sequence that grades from very fine sand to fine sand to medium sand. Ash contents were generally high, and mostly uniform. As in the other cores, ash content was associated with particle size, with the lowest ash contents being associated with the finest particle sizes.

The final core of this transect (B1) was augured on the boundary of Lake Teza in the south. B1 is dramatically finer throughout than any of the other cores on the transect. The predominant particle size class was fine silt. Ash contents were also lower and more variable than the cores to the north. A medium silt layer marked the base of the core. Thereafter, the sediment coarsened upwards until a depth of 4.2m, where a sudden change from medium silt to fine silt occurred. Within the middle fine silt layer, a sequence of upward coarsening and then upward fining sediment was found. This fine silt layer was separated from an overlying fine silt layer by a thin layer of very fine silt. The uppermost fine silt layer also tended to coarsen upwards. An unusually coarse layer of medium silt was found at the top of B1. Organic content was highest in the top 10cm of the core, with an organic content of greater than 10%.

Transect A was augured towards the central floodplain, with core A1 just south of the Uloa Peninsula and A6 next to the Msunduze River in the south (Figure 14). A comparison of particle size between Transects B and A shows A to comprise finer particle sizes than the upper floodplain.

The first core, A1, was augured north of the Mfolozi River, in a region that was once intermittently flooded by Lake Futululu before the Uloa Loop was circumvented by agriculturalists. The core comprises three main sequences, the first of which is a lower upward coarsening sequence that coarsens from very fine silt to medium silt. This is followed by two upward fining sequences; both of which fine upward from fine silt, culminating in clay and very fine silt respectively. *Cyperus papyrus* plant fossils occur between 2.5 and 3.5m below the surface, corresponding to an increase in organic content to over 8%.

Core A2 was taken on the current Mfolozi River levee. The core comprised two sequences, a lower upward fining sequence followed by an upper upward coarsening sequence. As such, median particle size in the core varies from very fine silt at the base, to fine silt, to very fine silt and then fine silt towards the top. Ash content varied little, and was once again related to median particle size.

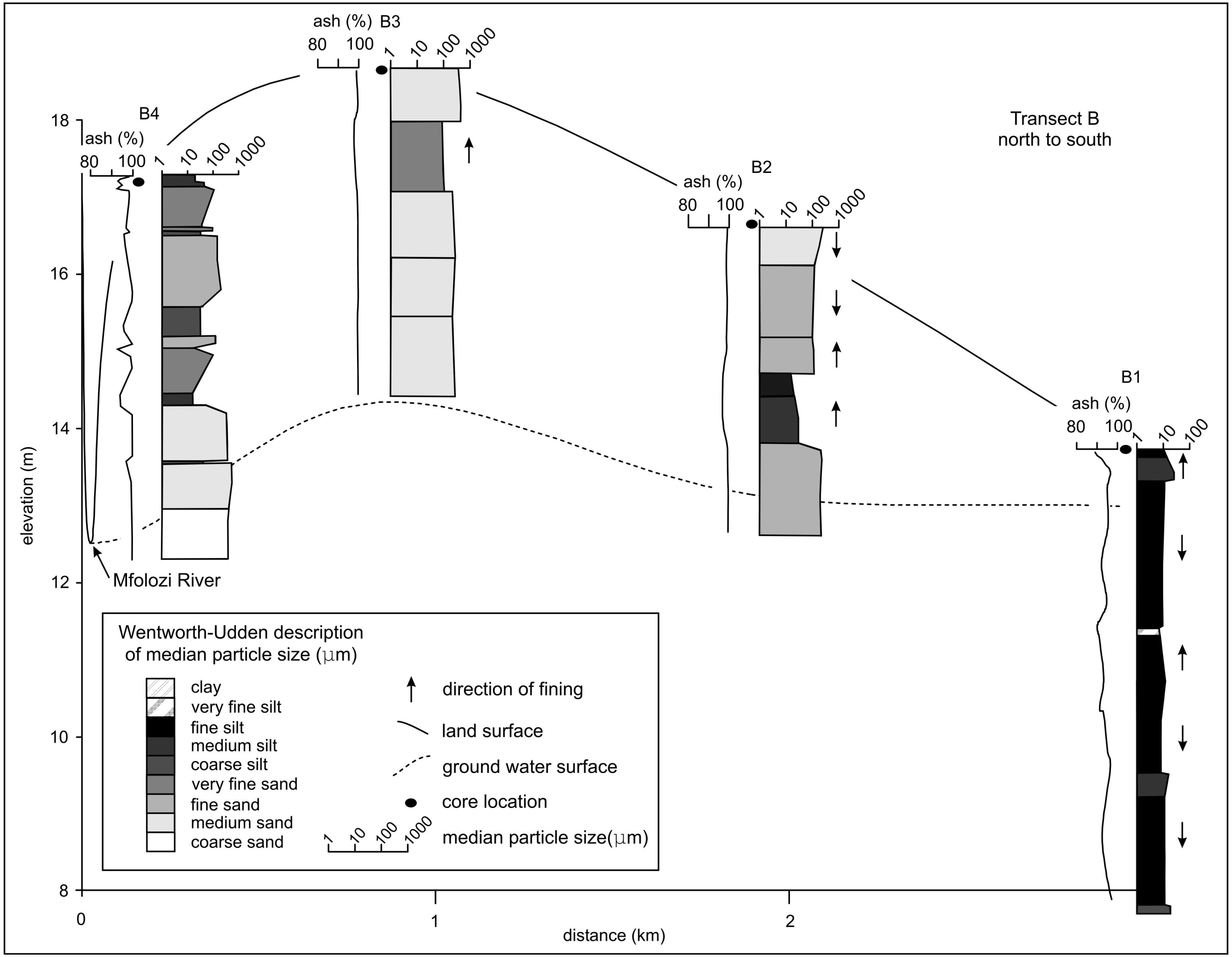

Figure 13: Sedimentology overlain on the topography of Transect B.

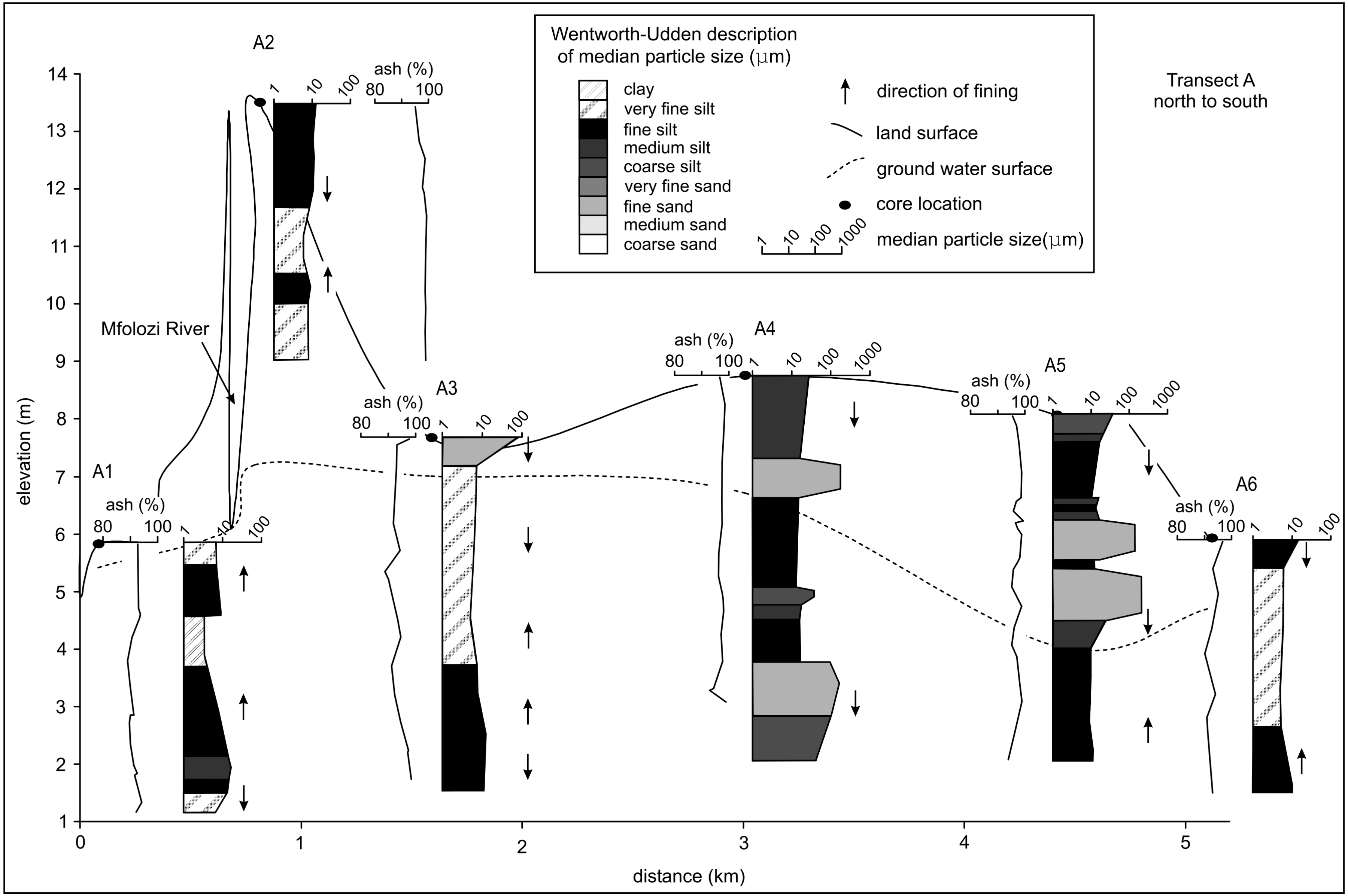

Figure 14: Sedimentology and topography of transect A in the central floodplain.

Core A3 was sampled just south of the current day Mfolozi River course in a depression between cores A2 and A4. The core comprised three units of different median particle size. The lowermost unit of fine silt was approximately 1.5m thick and comprised an upward coarsening unit overlain by an upward fining sequence. Contrastingly, the overlying very fine silt unit comprised an upward fining unit at the base, overlain by an upward coarsening sequence. At a depth of 30cm, particle size changed suddenly from very fine silt to a surficial deposit of very fine sand.

Core A4 was sampled close to an abandoned alluvial ridge visible in aerial photographs. Consequently, it is elevated above surrounding auger holes on the same transect. Both cores A4 and A5 were slightly coarser than others from the same transect, with no samples characterized as finer than fine silt in respect of median particle size. The base of A4 was an upward coarsening sequence of coarse silt overlain by fine sand. This was overlain by a general upward coarsening unit with a median particle size of fine silt, with intervening layers of coarser sediment.

A similar pattern was observed in Core A5. At the base of the core, there was a small layer of upward fining sediment. Thereafter, there was a trend of upward coarsening, with distinct intervening layers of coarser sediment. The coarsest of the layers (fine sand) occurred at an elevation of between 5.9 and 4.5m amsl.

Organic contents of both A4 and A5 were generally below 5%. However, organic content peaked in Core A4 at 7% at a depth 5.5m below the ground surface, while in A5, an organic content of 6.1% was measured at a depth 2m below the ground surface.

The southernmost core, A6, was very similar in terms of sedimentary sequences to Core A3. As in the case of Core A3, A6 comprised three units. The lower unit of fine silt fined upwards into the second unit of very fine silt, which was overlain by a thin upwardly coarsening sequence of fine silt.

The piezometric surface of transect A at the time of study sloped away from the Mfolozi River to the north and south. Further floodplain groundwater recharge occurred from the Msunduze River that is located south of Core A6.

The lowermost transect, Transect C, was located on the easternmost accessible part of the floodplain, and consisted of four core holes, with C1 just south of the Mfolozi River and C4 just north of the Msunduze River. The piezometric surface in this transect generally followed the topography as it sloped down from the highest elevation in the north (Figure 15).

The northernmost core, C1, consisted of two major sequences: a lower upward fining sequence that grades from fine silt to very fine silt, and an upper upward coarsening sequence, which grades from very fine silt to medium silt at the top. Within the sequence there are intermittent layers of coarse silt. Organic contents were generally low throughout the core, with a maximum value of 9% at 3.5m depth due to numerous plant fossils and roots.

Core C2 was sampled on an elevated terrace on the floodplain. The base of Core C2 was characterized by relatively coarse sediments of fine and very fine sand. Three metres below the surface, particle size changed suddenly from very fine sand to an upward coarsening medium silt layer that culminated in coarse silt. Above the upward coarsening sequence, an upward fining sequence of fine silt gave way upwards to a layer of medium silt. Organic contents were relatively uniform throughout the core, except for a maximum value recorded 50cm below the surface where organic content reached 8.9%.

As in Core C2, the base of Core C3 was unusually coarse, consisting of coarse silt and very fine sand. Above this layer, particle size suddenly declined in an upward coarsening sequence of fine silt. Above the fine silt, an upward fining sequence of medium silt gave way abruptly to a layer of coarse silt. Organic contents were unusual in core C3, with a sudden increase 2.5m below the surface where organic contents were 55%. This was related to a deposit rich in non-decomposed wood.

The southernmost core, C4, comprised two upward fining sequences. The lower portion fined upwards from medium silt at 4.5m to fine silt at 0.8m. The top 80cm were slightly coarser deposits of coarse and medium silt. Between depths of 0.8 and 1.5m, ash contents were relatively low with organic contents reaching almost 10%.

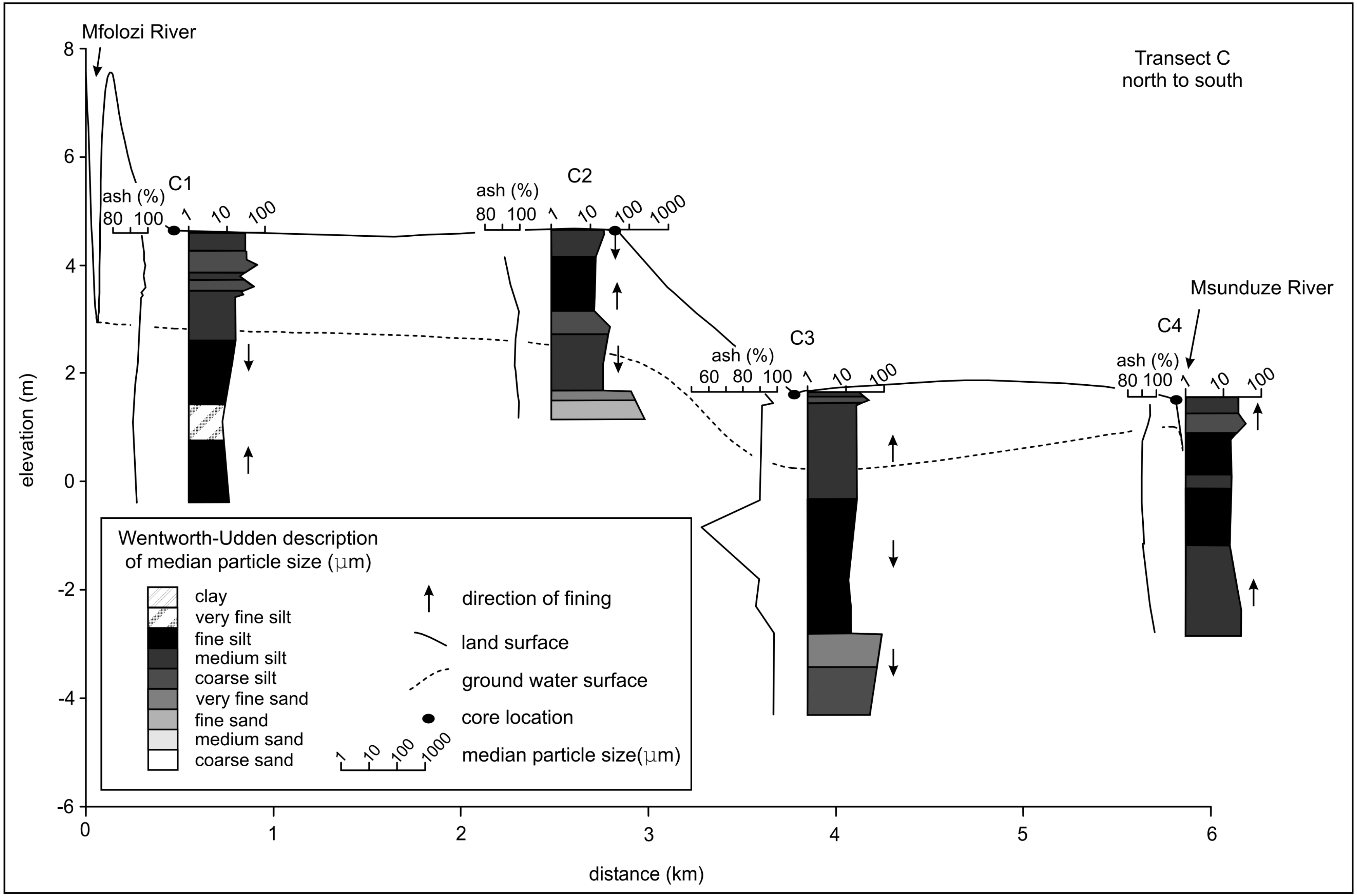

Figure 15: Topography and sedimentology of transect C in the lower floodplain.

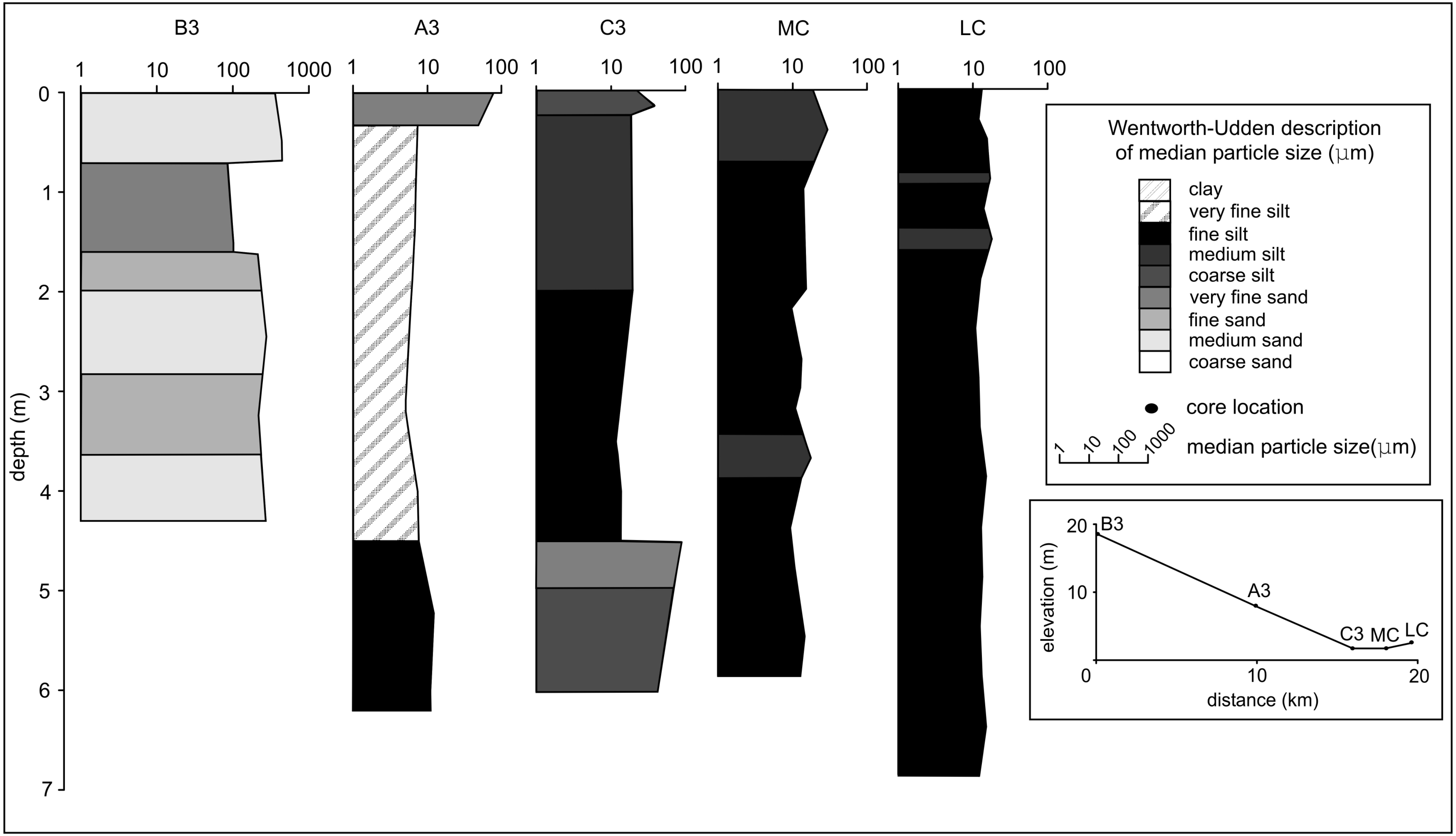

Figure 16: Longitudinal variation in sedimentology. The longitudinal profile of cores that are illustrated is in the right inset.

In order to establish differences in particle size and sediment sequences along the longitudinal axis of the floodplain, the central core from each of the transects were illustrated together with two additional cores, MC and LC (Figure 16). It is immediately obvious that the floodplain head (as represented by B3) is much coarser than the remainder of the floodplain. There appears to be a general fining trend towards the east, but there is substantial variation around this trend. Core A3 is, on average, the finest of all the cores. However, comparison of the uppermost portions of each core suggests that sediments became progressively finer downstream.

In terms of overall particle size variation with depth, there is no consistent trend in cores B3, A3 or C3. However, in cores MC and LC, there is a slight upward coarsening trend but this is not significant.

3.4. Morphology of the alluvial belt

Figure 17 indicates the stratigraphy and topography of an abandoned alluvial ridge on the Mfolozi Floodplain. The alluvial ridge is elevated above the surrounding floodplain by at least 2m, sloping more steeply from the abandoned channel towards the north (4.3%) than towards the south (0.6%). The elevation of the floodplain is also greater in the north (3.1m amsl) than the south (2.2m amsl). The morphology of the pre-existing channel is not clearly discernable because of extensive ploughing and cultivation on the alluvial ridge that has removed details of the channel.

Cores T1 and T2 were very similar, consisting primarily of medium silt sediments, with intervening layers of fine sand and coarse silt. T3 was much coarser than any of the other cores and was dominated by medium to fine sand that consistently collapsed at a depth of 2.2m. The top of the core was marked by a medium silt layer some 40cm thick. A similar deposit of similar thickness occurred on top of all of the cores. In cores T4, T5 and T6, coarse sediments were encountered at greater depth. In T4, silt was encountered until a depth of 1.7m, whereupon silt changed suddenly to very fine sand. The same unconformity was encountered at 1.2m and 1.56m depths in cores T5 and T6 respectively. No similar sandy layers were encountered in cores T1, T2 or T7. Core T7 was markedly finer than the other cores, and was dominated by medium silt sediments with two intervening layers of coarse silt. The base of the core was fine silt, a sediment type that was encountered nowhere else on the alluvial ridge.

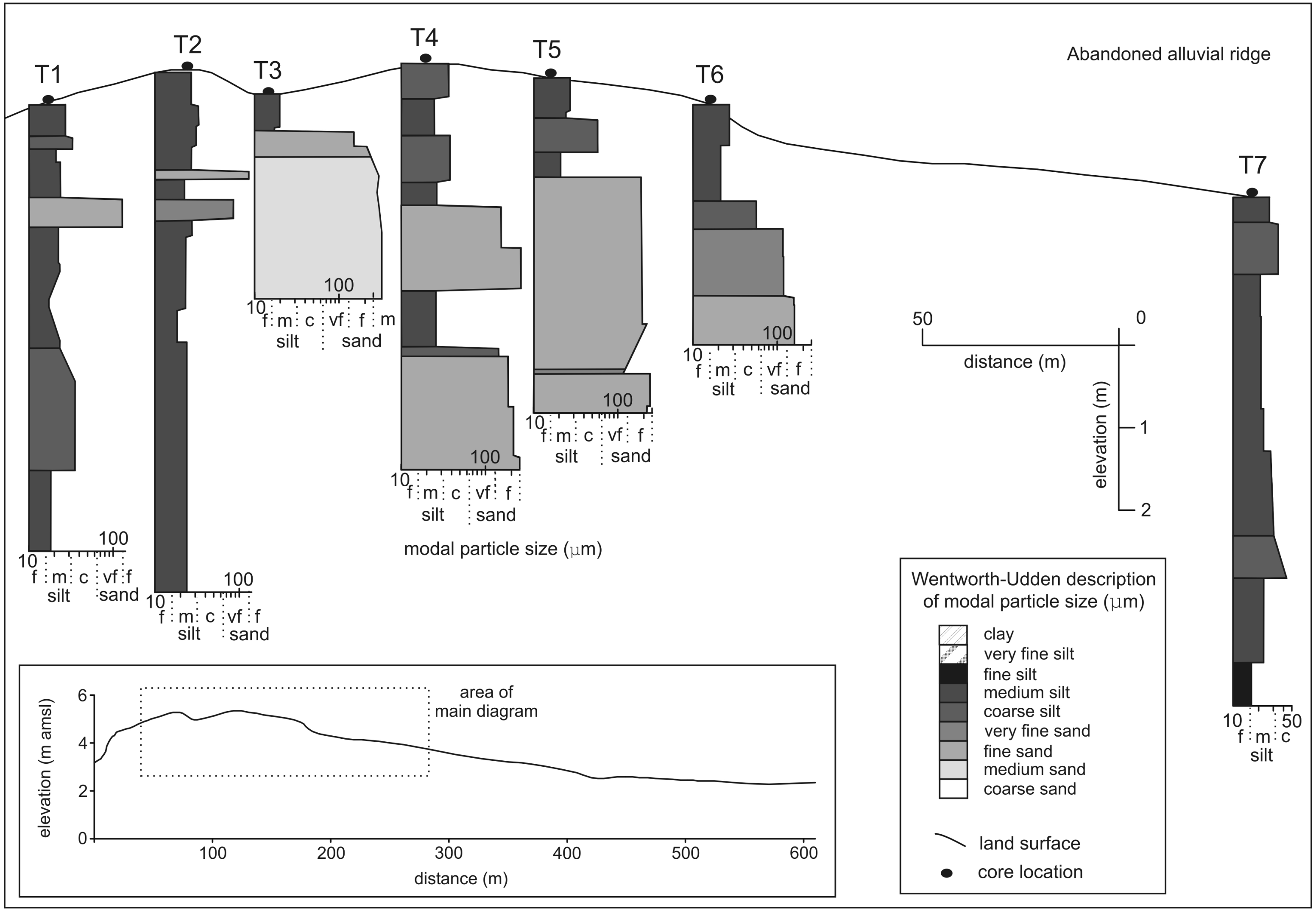

Figure 17: Stratigraphy and morphology of an abandoned alluvial ridge (location indicated on Figure 2).

4. Discussion

4.1. Floodplain origin and evolution

Upstream of the Mfolozi Floodplain, the Mfolozi River occupies an incised valley characterised by entrenched meanders over 30m high. This square shaped valley is cut into rhyolites of the Lebombo Group that erupted during a failed rifting event 179Ma (Watkeys *et al*. 1993). This early rifting event, occurring on the supercontinent of Gondwana, was the first episode to lead to the formation of a sedimentary basin in northern KwaZulu-Natal in which the Mfolozi Floodplain is now situated. The crust was further extended during rifting events that eventually separated the continents of Africa, Antartica, South America and Australia, leading to localised subsidence. The bedrock of the valley in which the floodplain is now situated, was deposited in the newly formed basin, initially as alluvial and fluvial deposits, but later marine facies as the ocean transgressed. Deposition of the Zululand Group ceased as the ocean regressed, initiating an erosional period. 50Ma later, when a transgression again initiated continental deposition, the Maputaland Group, comprising numerous fossiliferous beds and palaeodune deposits, was deposited in the basin.

More recently, sea level again dropped, reaching 120m below MSL approximately 18 000BP, initiating a new erosional phase (Ramsay 1995). Coastal rivers that had been graded to the previous sea level were locally rejuvenated and the sudden increase in gradient resulted in widespread incision on the coastal plain. In regions where lithologies were resistant, deep square valleys were incised. In the case of the Mfolozi River, this resulted in the development of entrenched meanders upstream of the floodplain. However, as the Mfolozi River flowed over the Lebombo Group rhyolites, it encountered less resistant rock of the Maputaland and Zululand Groups. These were more conducive to erosion, and a wide, deep valley was carved.

Following the Last Glacial Maximum 18 000 year BP, sea levels began to rise. Beach rock evidence suggests that sea level along the Maputaland coasts was 3.5m above current levels until 4880yr BP, reaching current levels 3780yr BP (Ramsay 1995). Palynological evidence from Lake Teza suggests that higher than current sea levels drowned portions of the Mfolozi River Valley (Scott and Steenkamp 1996). In addition to drowning, the gradient of coastal rivers decreased, causing a shift from being erosional, to depositional. For a period after the Last Glacial Maximum, the rate of sea level rise exceeded valley aggradation, resulting in valley drowning. As sea level dropped and thereafter remained static, sediment accumulation eventually prevented intrusion of the sea into these formerly eroded valleys. However, some portions of the floodplain, particularly those where rates of overbank sedimentation are low, are still located below sea level (Figure 11).

The evolution of the Mfolozi River Floodplain is different from that of other wetlands studied in southern Africa that primarily owe their origin to portions of their drainage lines overlying resistant rocks (Tooth *et al*. 2004, Grenfell *et al*. in press). Floodplains of this type are typically characterised by mixed bedrock-alluvial rivers that experience long-term erosion of the valley floor, and as such, the depth of sediment accumulation is usually less than 8m. In contrast, portions of the Mfolozi River Floodplain are known to be underlain by at least 50m of sediment (van Heerden and Swart 1985). Thus, while the mechanism of formation of the Mfolozi River Floodplain is determined by a base level that limits incision due to a rise in sea level, instead of incising in the long-term, it is aggrading.

4.2. Floodplain geomorphology and sedimentology

The Mfolozi River floodplain may be subdivided based on variation in valley slope, which reflect different controls and floodplain processes. Four geomorphic zones can be identified. The sedimentology and geomorphology of each zone is described in the following section, with the upper two zones discussed in tandem due to their interconnected nature.

4.2.1. Upper Floodplain and alluvial fan

Two distinct slopes mark the upper floodplain region. The upper slope river gradient is 0.06% and extends from upstream of the floodplain to the upper reaches of the floodplain. Below this region, the river slope steepens to 0.9% over a 4km long stretch. This gradient represents the prograding face of an alluvial fan that has formed due to sudden loss of confinement at the floodplain head. Above the floodplain region, the Mfolozi River flows in a confined valley composed of Lebombo Group rhyolites. In the vicinity of the floodplain, the Mfolozi River Valley is composed of less resistant lithologies, and widens dramatically from 0.6km at the floodplain head to 6.5km wide over a downstream distance of just 1.15 km. As the Mfolozi River flows from the confined valley onto the floodplain, the river widens and stream capacity, particularly during flood events when the stream loses confinement as inundation of the floodplain takes place, declines. Resultant deposition in the region, particularly during flood flows, has built a lobe of sediment at the floodplain head that currently slopes towards the southeast.

The alluvial fan is dominated by sand size sediments that vary from very fine to coarse. As a result of the permeability of the sediments, the fan is a groundwater recharge area since rainfall easily infiltrates the sandy sediment and drains freely to recharge the groundwater table. As such, the piezometric surface slopes away from the central region of the lobe of sandy sediment. The dominant manner in which relief is built on the alluvial fan must be accredited to overbank flood flows rather than channel aggradation. While the maximum depth of the Domoina deposits is not represented in Transect B (Domoina deposit depths vary from 0.7m at B3 to 1.8m at B2), they are still substantial for a single flood event lasting a few days. Overbank flood flows must be responsible for most relief building of the alluvial fan. The channel pattern of the Mfolozi River in this region may also be used as an indication of the importance of flood flows in this region. Usually, floodplain regions of steepening and aggradation are characterised by high sinuosity (Schumm 2005). Contrarily, the Mfolozi River is relatively straight in this region. During flood events, the river course would straighten to accommodate flood flows. However, the Mfolozi River does not adjust between flood events because the most frequent flows lack the stream power required to do geomorphic work. As such, in this region, the Mfolozi River is not in equilibrium with its most frequent flows and sediment loads.

Rapid aggradation on the alluvial fan has implications for tributaries flowing onto the floodplain from its boundaries. On the southern boundary, the Msunduze River flows into the floodplain, and is essentially impounded by the alluvial fan, resulting in the formation of Lake Teza. The Msunduze River flows out of the lake as a yazoo stream. While the lake may appear to be too close to the western boundary of the floodplain to receive sediment from flood flows of the Mfolozi, Domoina deposits are represented in the upper unit of Core B1 as a 40cm deep layer of medium grained silt. Domoina deposits in Lake Teza have also been described by Scott and Steenkamp (1996). Other layers of medium silt at 4.2m and 5.9m depths must be indicative of previous large flood events on the Mfolozi Floodplain.

The depression at the head of the floodplain north of Lake Teza is anomalous and reflects the nature of sedimentation in the vicinity of an alluvial ridge. An alluvial ridge is a consequence of deposition during overbank flooding due to the reduction in current velocity as water flows out of the channel and over the levees, leading to localised aggradation close to the channel edges. Given that this deposition takes place more-or-less uniformly along the banks of the channel and along the alluvial ridge, gradients are steepened perpendicular to the channel. However, as floodwaters move away from the alluvial ridge and levees, flow will respond to regional gradients, which are oriented towards the depression along the southern margin of the floodplain. Given this combination of processes, it should be expected that depositional features close to the Mfolozi River would be oriented towards the south, but that further away from the river, depositional features would be oriented towards the southeast.

In the particular case of the depositional feature created by Cyclone Domoina its location and nature may also be due to manipulation of the stream course by the agricultural sector in order to manage floods. A surprising feature of the Domoina deposits is that they occurred some distance down from the head of the floodplain and deposition seems to have created a depression upstream of the main depositional feature.

4.2.2. Central Floodplain

Below the alluvial fan, the stream gradient flattens to 0.03%, then follows a step-wise drop of just over 1m in 1.3km. Thereafter, gradient decreases again to 0.03%. A valley longitudinal profile showed a similar region of flattening, suggesting gradient in this region is controlled structurally across the entire floodplain. In conjunction with the extremely low slope, the floodplain is anomalously wide (approximately 10 km) just upstream of the central region. The combination of the width and change in gradient are frequently an indication of faulting in an alluvial setting (Schumm *et al*. 2000). Anomalies in river sinuosity may also be interpreted as an indicator of faulting, but the location of the alluvial fan upstream of the region complicates the use of sinuosity as an indicator.

The location of the probable fault is visible on satellite images of the St. Lucia area as a linear feature extending across the central floodplain and bordering False Bay of Lake St. Lucia. A second lineament comprises the western boundary of the main water body of Lake St. Lucia and is roughly parallel to the coastline. The area is currently seismically active, with the most recent report indicating the existence of WNW-ESE striking fault lines (Umvoto Africa 2004). However, in Krige and Venter's (1933) study on the St. Lucia earthquake of 1932, isoseismal lines drawn using onshore observations suggested a fault line approximately parallel to the coast. If this is so, lineations observed in the Lake St. Lucia area would be consistent with this strike. However, it is difficult to infer the direction of fault movement. Furthermore, coastal features follow a similar strike to the potential fault. Nevertheless, the combination of data, from adjustments in gradient, floodplain width and visible lineations in the area of interest, strongly suggest the existence of such a fault.

The sedimentology of the central and lower floodplain is dependent on various factors besides structural control, the most important of which is distance from the channel. Many studies have highlighted the importance of distance from channel as a determinant of sediment particle size (e.g. Pizzuto 1987, Asselman and Middelkoop 1995, Makaske *et al*. 2002). However, in this case it appears to be distance from abandoned channels and the concomitant mosaic of inter ridge depressions that plays an important role in resultant sediment particle size. As a result, the central floodplain is marked by large variability in particle size, as indicated by Transect A (Figure 14). Cores A4 and A5 were located on or near abandoned channel courses and were characterized by slightly coarser sediments than the other cores, with fine silt and fine sand layers well represented. Contrastingly, despite the proximity of Core A3 to the channel, the majority of the core is extremely fine, with a thick sequence of very fine silt, capped by recent coarse deposits, presumably from recent flood deposits. The very fine sediments of Core A3 are representative of those of small depressions created by basins that occur between alluvial ridges. Asselman and Middelkoop (1995) also found that ponding in closed depressions was an important factor in determining the nature of sediment accumulation.

4.2.3. Lower Floodplain

The longitudinal profile reveals a concave slope, measuring between 0.03% to 0.02%, between the central floodplain and the coast. The slope of the lower region resembles a graded profile that is controlled by sea level. The lower floodplain has been a region of enhanced aggradation due to its proximity to sea level since the last transgression, whereupon sea level reached a still stand approximately 3780BP (Ramsay 1995). As such, Van Heerden and Swart (1984) estimate the sediment depth of the lower Mfolozi Floodplain to exceed 50m.

In addition to controlling the occurrence of depositional or erosive periods through transgression and regression, sea level and tidal patterns also have an influence on channel pattern close to the river mouth. This is largely because channel sinuosity in this region is impacted upon by the extent of the tidal prism (Dalrymple *et al*. 1992), resulting in frequent changes in sinuosity in this region. Furthermore, the wave dominated nature of the KwaZulu-Natal coast (Cooper 2001) results in offshore sedimentation being retarded. As a result, the lower portion of the Mfolozi River profile is truncated (Figure 12). This suggests that if wave energies were lower, the Mfolozi River would build a delta out to sea some 4km long.

The sedimentology of the lower floodplain region is represented by cores MC and LC, as well as to some extent, by the lowermost Transect, C. Unexpectedly, the median particle size of all the samples was never finer than very fine silt in the lower floodplain region, either at the surface or at depth, and the typical median particle sizes were fine and medium silt. This seems somewhat surprising considering that overbank areas generally consist of clay to silt sized sediments, and in many cases, peat (e.g. Makaske *et al*. 2002, Törnqvist 1994, Magilligan 1992). However, Magilligan's (1992) study of the Galena River Basin in Wisconsin showed that local geology and sediment provenance could play a major role in determining the sedimentology of floodplain alluvium. Particle size determinations of suspended sediment loads showed that the average median particle size was 5.2µm, while the minimum median size recorded was 4.0µm. These both classify as very fine silt according to the Wentworth-Udden scale (Wentworth 1922). It seems likely that the lack of clay in overbank deposits reflects a lack of supply.

4.3. Floodplain processes and dynamics

4.3.1. Fluvial style

The sedimentology of the abandoned alluvial ridge clearly indicates that in the central and lower Mfolozi River floodplain region, the Mfolozi River was a meandering river. The river historically (and currently) occupied a position of elevation on the floodplain, flowing more than 2m above the surrounding floodplain. The alluvial ridge was also noticeably coarser than the surrounding floodplain (e.g. Core T7), and showed coarsening with depth on the inner bend (right bank) that is consistent with the deposition of point bars. Cores T1 and T2 were finer, and had thin intervening layers of fine and very fine sand. These two cores are located on the outside bend of the meandering course. Finer sediments are indicative of normal overbank flooding and resultant deposition, while the coarser layers may be interpreted as larger flood events that were capable of carrying coarser sediments over the levee and onto the floodplain. Thus, the sedimentology of this reach is also marked by larger than usual flood events, in conjunction with aggradation in the channel and levees of the alluvial ridge.

4.3.2. Floodplain dynamics

Figure 18 illustrates the avulsion sequence of the Mfolozi River as interpreted geomorphically from aerial photography. The sequence acknowledges two types of change to the channel’s course; natural avulsions and human induced channel straightening events. Figure 18a is the earliest known course of the Mfolozi River that is still preserved on the floodplain surface. This course flowed relatively straight down the alluvial fan, completing a heart shaped loop towards its base, before flowing towards the southeast and along the current day Msunduze River course. This alluvial ridge is visible in floodplain surface particle size distributions (Figure 14) as a coarse southeast trending silt ridge on the floodplain surface that is parallel to the western edge of the Uloa Peninsula. The parallel orientation of the ridge and the Uloa Peninsula may also have been responsible for the original surface water pattern of Lake Futululu, which was also southeast trending (Figure 4).

At some stage, aggradation on the southeast trending ridge resulted in channel longitudinal gradients that were exceedingly low as compared to the gradient away from the alluvial ridge in a fashion described by Heller and Paola (1996). This resulted in a channel avulsion towards the east at what is now called the Uloa Loop (Figure 18b). This actively meandering feature resulted in the river flowing due east from the location of avulsion. During this period, the Futululu drainage line may have undergone accelerated erosion due to the increased proximity of the alluvial ridge. Eastwards of the Uloa Peninsula, the newly formed sinuous Mfolozi River flowed in the centre of the floodplain, north of its previous position. The following avulsion (Figure 18c) resulted in a portion of the river being directed even further north and decreasing its sinuosity. This early sequence indicates a series of avulsions progressing northwards, suggesting that the floodplain surface dipped towards the northeast at this stage.

Between 1910 and 1937, sugar cane farmers began cultivating on the floodplain and began to straighten portions of the channel to improve channel efficiency. It appears that the first major alteration was the movement of the channel towards the extreme north such that it flowed around the Uloa peninsula and then hugged the floodplain boundary. It then occupied the existing channel approximately 4km downstream of the Uloa Peninsula (Figure 18d). This was followed by the complete removal of the Uloa Loop from the rivers course between 1937 and 1960 (Figure 18e).

Continual aggradation on the northern region of the floodplain resulted in slopes towards the south becoming steeper than the channel longitudinal slope, bringing the system close to an avulsion threshold. The impetus for sudden change as the threshold was crossed occurred in 1984 when floods associated with cyclone Domoina completely altered the course of the Mfolozi River (Figure 18f). The progressive northward migration of the Mfolozi River in previous years had resulted in the floodplain surface gently dipping towards the southeast (Figure 11). The sudden huge discharges and sediment loads overwhelmed the existing channel, and existing gradients favoured an avulsion at the floodplain head. The Mfolozi River switched from hugging the northern floodplain boundary to flowing to the south and occupying the existing Msunduze River course.

The resulting location of an avulsion is a combination of super-elevation of the channel above the floodplain (Heller and Paola 1996), the relative channel versus cross-valley gradient, the occurrence of previous avulsions (Mackey and Bridge 1995), and also stochastic events. Much research and probability modelling has been focussed on determining where and when avulsions are most likely to take place. In order to simulate a rising base level, such as would occur in coastal settings if sea level rises (and as would have occurred along the South African coast during the Holocene transgression), Mackey and Bridge (1995) increased floodplain and channel aggradation rates down-valley. This was associated with a down-valley decrease and a cross-valley increase in slope gradient, creating the situation whereby avulsion probabilities increase downstream in this simulation, because the gradient between channel and floodplain is greatest there. However, while initial avulsions seem to have occurred downstream initially, consequent avulsions tended to occur progressively upstream from the initial avulsion.

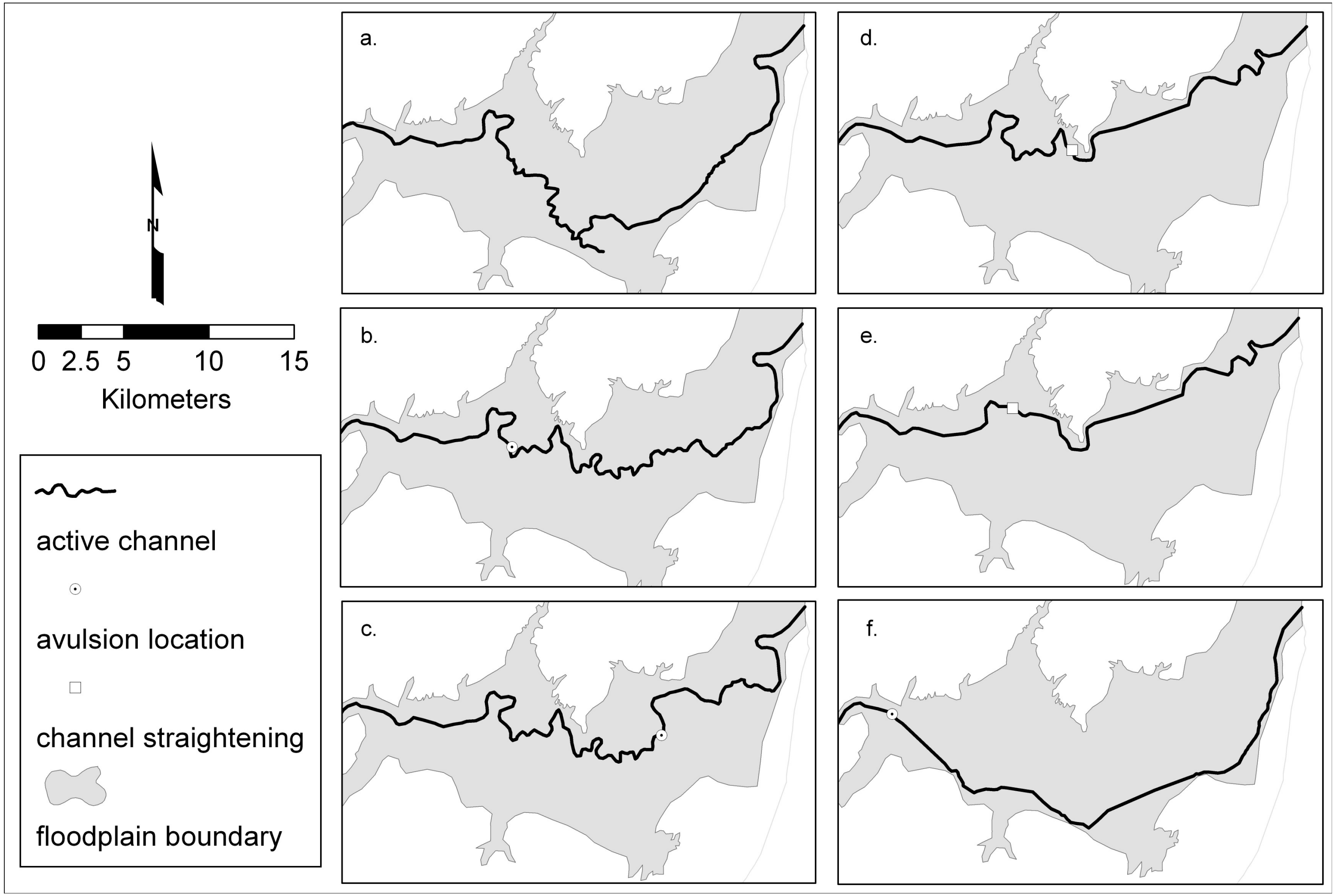

Figure 18: Interpreted recent avulsion history of the Mfolozi River: a-c - pre-human intervention, d - 1937, e - post 1937 and f -following Domoina. The westernmost quarter of the floodplain’s history is the most uncertain due to the lack of preservation of channel features.

Contrastingly, a down-valley decrease in aggradation rate simulates the behaviour of an alluvial fan, as occurs at the head of the Mfolozi Floodplain. In this simulation, it was found that avulsions are likely to occur in the upper reaches of the floodplain, around the apex of the alluvial fan. Bryant *et al*. (1995) showed that increased aggradation increases super-elevation of the channel above the surrounding floodplain, resulting in more frequent avulsions, a result that was consistent with Törnqvist's (1994) study in the Rhine-Meuse delta.

Despite the limited availability of information regarding the Mfolozi River's avulsion history, these models provide insight into the dynamics of the Mfolozi River floodplain. The floodplain's current geomorphic state suggests two main nodes of ongoing sedimentation. The first is at sea level, where the floodplain is under ongoing adjustment. As the river approaches the sea, gradients decrease, the river loses both capacity and competence, and aggradation of the channel occurs. The result of aggradation at the coast is further lowering of the river's longitudinal gradient, which in turn causes feedback that further encourages sedimentation. Unless the estuary and lower floodplain area are scoured, as described by Cooper (1993, 1994, 2001), aggradation on the lower floodplain channel may reach a critical elevation causing local avulsion. Thus, the lower floodplain region must be one of frequent avulsion, as the channel jostles the effects of transgression and loss of sediment transport capacity. The evidence for frequent avulsions in this region is the large number of preserved abandoned channels in the lower floodplain region, which is also indicative of low rates of overbank aggradation.

The second node of enhanced aggradation, and thus avulsion frequency, is on the alluvial fan at the floodplain head. As the Mfolozi River exits the Lebombo Group rhyolites upstream of the floodplain, sudden loss of confinement results in a loss of sediment transport capacity, particularly during flood events, resulting in sediment deposition. In the Mfolozi, the majority of alluvial fan building is not through gradual ongoing (daily) channel aggradation, but rather through sudden rapid aggradation associated with infrequent flood events. During cyclone Domoina, discharges exceeded the channel's flow capacity. As a result, the river overtopped its banks, and began depositing sediment on the alluvial fan as it flowed southwards towards the Msunduze River course. Deposition occurred as a series of lobes that were deposited and subsequently eroded, disrupting the characteristic flood sequence of upward coarsening followed by upward fining as the flood wanes. The result is a varied sequence of upward coarsening and/or fining within the Domoina flood deposits, as shown in Figure 13. At Core B3, only a portion of the Domoina deposit is represented, while at Core B2, only the upward coarsening unit is visible. At Core B1, a thin unit of upward fining is visible, indicating the waning flow. Using this information, it seems that during the flood, the sediment lobe was eroded and redepositted at least twice between Cores B1 and B3. The current topography of the upper alluvial fan is also indicative of the process of deposition followed by scour, where the sediment lobe is transported progressively further from the avulsion node. This suggests that episodic scour is an important part of alluvial fan evolution, particularly during large flood events.

Avulsion on the alluvial fan is likely to be more common than can be assessed from the location of abandoned courses, particularly since flood deposits continuously hide abandoned channel courses in this region. However, the overall southerly slope of the floodplain surface suggests that there has been a substantial period of time since the last successful avulsion of the Mfolozi River to the south.

In general, avulsions on the Mfolozi Floodplain are caused by a combination of a lowering of the existing channel longitudinal slope as aggradation occurs at sea level, and an increase in slope perpendicular to the channel due to alluvial ridge aggradation. The combination of these processes result in the Mfolozi River constantly evolving close to an avulsion threshold, the location of which may be determined by chance and the relative effects of downstream aggradation versus alluvial ridge and alluvial fan aggradation.

The process of alluvial ridge development results in the formation of a wide variety of depositional environments. In particular, the coalescence of alluvial ridges at different points on the floodplain creates a mosaic of inter-ridge basins that are conducive to the very slow accumulation of fine sediments and organics, and sometimes results in the formation of small, shallow lakes. Examples of inter-ridge depressions may be seen in Figures 6 and 7, where lakes occur in between abandoned channel courses. The effect of this process on sedimentology is well represented in Figure 14, where Core A3 occupies a depression between the current alluvial ridge and an abandoned channel course at Core A4. The majority of Core A3 is finer than would be expected considering its position on the mid floodplain (Figure 16). Thus, the development of inter-ridge depressions interrupts broad scale sedimentological trends and produces features dependent on local topography and geomorphic setting.

Furthermore, the location of existing alluvial ridges may have implications for future channel development, as they limit the movement of new channels across the floodplain surface. Mackey and Bridge's (1995) modelling suggested that alluvial ridges frequently resulted in channels clustering preferentially on one side of the floodplain. This appears to have been the case in the Mfolozi River, particularly when avulsions occurred in the lower floodplain. Here, channel development is impeded by pre-existing alluvial ridges that resulted in new channels forming mainly towards the north of the course that occupied the center of the floodplain.

Floodplain aggradation may also have been altered by human intervention in the last century. A closer analysis of core LC indicates sudden sediment coarsening in the upper 1.5m of the core. In addition, the core appears to have been taken on a levee of the Msunduze River, resulting in this lower core being elevated by 0.7m above core MC, some 1.6km upstream. The change in sediment characteristic is sudden and corresponds to more than the elevation difference between Cores MC and LC. It seems most likely that the sudden increase in rate of aggradation on the lower Msunduze River is related to sudden increased sediment availability. The majority of the furrows that drain the Mfolozi Floodplain empty into the Msunduze River, because of the southeast slope of the basin. Furthermore, the Msunduze River just eastwards of Lake Teza is eroding, presumably as a result of Mfolozi flood flows being forced into the channel by a spillway constructed by the Umfolozi Co-operative of Sugar Planters (UCOSP). The combined erosion of the Msunduze River and the supply of sediment from furrows have substantially increased the sediment load of the Msunduze River. However, the gradient of the Msunduze River is substantially reduced as it begins to travel north alongside the barrier dune. Survey data supplied by UCOSP indicates that upstream of the point where the Msunduze River turns northwards, the river gradient is 3 times (0.006 %) that of the river gradient below the dogleg (0.002 %). Thus, at this point, the Msunduze River experiences a sudden loss of capacity, which sometimes results in ponding (as is visible in Figure 3) and aggradation. Aggradation in this region is likely to propagate upstream, decreasing the efficiency of furrows on the floodplain. As such, human induced erosion has caused aggradation on the lower floodplain that will, in time, have a negative feedback that decreases erosion upstream.

5. Conclusions

The Mfolozi River Floodplain expands our understanding of wetland formation within southern Africa. The Floodplain is unlike its northern hemisphere compatriots due to a combination of climatic factors and continental uplift followed by variation in sea level. The absence of peat in regions of overbank flooding can be mostly attributed to the highly seasonal nature of flood flows, and indeed, the irregularity of overbank flooding on the floodplain. This results in long periods of desiccation that cause aerobic decomposition of plant matter. Furthermore, there is evidence to suggest that the Mfolozi River floodplain has adapted geomorphically to its highly variable flow regime, and that the morphology of the river is not in equilibrium with normal flow conditions. It appears that irregularly large flood events may be responsible for the majority of geomorphic change in the upper and middle floodplain, causing avulsions and massive deposition.

The evolution of the floodplain also differs considerably from the dominant model of floodplain evolution in southern Africa of Tooth *et al.* (2004). The Mfolozi Floodplain is a region of sediment infilling and aggradation, rather than lateral erosion and long-term incision.

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Chapter 5. Tributary drowning by trunk channel aggradation: the evolution of Lake Futululu on the Mfolozi River Floodplain, KwaZulu-Natal, South Africa

Abstract

Trunk-tributary relationships have more commonly been described in terms of the impact of the tributary on the geomorphology of the trunk channel. Sediment laden tributaries, flowing through steeper catchments than their trunk channel counterparts, deposit their sediment load at the stream confluence. The geomorphic impact is variable, and is dependent on the trunk's ability to assimilate and/or transport the tributary's sediment. In some cases, tributary sediment loads may result in changes to the trunk channels shape or pattern, while in more extreme cases they may result in the formation of palustrine wetlands or lakes. This research describes trunk-tributary relationships at the other end of the continuum; where aggradation on trunk channels may overwhelm tributary valleys, once again causing either palustrine wetland formation, or lakes. Lake Futululu is an impounded drainage line of the Mfolozi River. During the last glacial maximum, rivers of KwaZulu-Natal were rejuvenated, resulting in wide-scale vertical incision. During this period, the Mfolozi Valley was carved in soft sedimentary rocks of the Zululand and Maputaland Groups. The resulting valley was drowned during the Holocene highstand approximately 4000 BP, but sea levels regressed during a neoglacial period 3200 BP, rising to current levels 900 BP. This series of sea level changes has defined the evolution of the coastal Mfolozi Floodplain and its associated drainage lines. As sea levels rose, incision in the valley ceased, and aggradation commenced. The high relative sediment load of the Mfolozi River resulted in aggradation on the Mfolozi Floodplain outstripping aggradation in the Futululu drainage line, causing the formation of a basin environment conducive to peat formation by 3980 BP. Rates of peat accumulation since the development of the basin have been constant, indicating disconnectivity between the rate of basin filling, and changing climate and sea level. This suggests that the Mfolozi River replaced sea level as the base level of the Futululu valley. Nevertheless, despite the disconnection in climate and rate of peat accumulation, the most recent neoglacial, when southern African climates were dryer and cooler, is recorded in the Futululu peat record as a 0.3m thick charcoal layer, indicating wetland desiccation during this period.

1. Introduction

Trunk-tributary relationships have more commonly been described in the context of the tributary's influence on trunk channel character and behaviour. Steep, sediment-laden tributaries may deposit large quantities of sediment into the trunk channel, causing changes to channel pattern, bedload characteristics and river style (Schumm 2005). Rutherford (2001) described how tributaries of the Glenelg River in Australia supplied large sediment loads to the trunk channel, partially blocking trunk flow and causing the formation of backwater lakes upstream of each tributary junction. In addition, Rice (1998) considered how tributaries might significantly impact upon bedload texture and as a result, impact upon channel form. In particular, large sedimentary inputs were found to redefine the trunk channel's grain-size distribution, hampering downstream maturation. Indeed, sediment inputs from tributaries have also been held accountable for the shifting of the Mississippi River to the opposite side of the floodplain floor (Schumm 2005).

In contrast, the relationship between a large aggrading trunk channel and smaller tributaries that have gentle gradients and carry only small amounts of sediment is the subject of this chapter. When aggradation on the trunk channel exceeds sediment delivery from a tributary, the trunk river exerts an influence on the tributary's geomorphic behaviour, resulting in the formation of a drowned valley. The resultant lake should not be confused with an overbank floodplain lake, as it is fed by water from it's own drainage line, rather than through overtopping of the trunk river. These types of systems have been sparingly acknowledged in the rock record. Michaelsen *et al*. (2000) described levee-dammed lakes abutting vertically stacked channel sandstones in the Permian Rangal Coal Measures in Australia. The lake deposits were characterized by well-sorted carbonaceous siltstone with rhythmical varve-like laminations and rare plant fossils. In terms of geometry, the deposits were sheet like and extended up to 17km longitudinally to a maximum depth of 23m. The rarity of plant fossils suggested that water was sufficiently deep to preclude standing vegetation, and that aerobic conditions occasionally prevailed (Michaelsen *et al*. 2000). Lenticular bodies of sandstone amidst carbonaceous siltstone were interpreted as representing occasional crevassing of the trunk channel into the lake. Similar finely-laminated calcareous siltstone deposits were described by Roberts (2007) in the Cretaceous-aged Kaiparowits Formation of Utah, USA. Also sheet-like in geometry, these deposits varied from just a few 100m to kilometers long. Unlike those of the Rangal Coal Measures, plant fossils were abundant, suggesting the occurrence of emergent vegetation.

The description of presently existing alluvial ridge dammed lakes and valleys is sparse. Knighton (1989) observed the development of tributary lakes as tin mining in the Ringarooma River catchment artificially enhanced aggradation rates of the trunk channel. Grenfell *et al*. (in press) describe the formation of a valley-bottom wetland that abuts the levee of a mixed bedrock-alluvial river in the KwaZulu-Natal Drakensberg Foothills of South Africa. Outgoing flow in this system is impeded by a combination of impermeable clay plugs in abandoned oxbows on the alluvial ridge, as well as by the elevated nature of the Mooi River's levees. The cut-and-fill valleys of the Wolumla catchment in New South Wales, Australia, described extensively by Brooks and Brierley (1997), Fryirs and Brierley (1998) and Brierley and Fryirs (1999) may be examples of alluvial ridge dammed valleys. In describing connectivity along drainage lines, Fryirs *et al*. (2007) refer to tributary disconnection from trunk channels that resulted in the formation of trapped tributary fills. The specific focus of these authors is the impact that geomorphology has on sediment cascades along drainage lines.

In this chapter, the sedimentology of alluvial ridge dammed features, and implications for the origin and evolution of wetlands, are explored. The primary focus of the chapter is the origin and evolution of a floodplain lake, Lake Futululu, on the Mfolozi River Floodplain, KwaZulu-Natal north coast. The lack of organic sediment accumulation on the Mfolozi Floodplain proper makes measurement of the rate and timing of fluvial evolution difficult and expensive, while the thickness of alluvial sediment (>50m) prevents direct sedimentological study of palaeo-fluvial dynamics. Since the evolution of alluvial ridge dammed systems, such as Lake Futululu, is closely tied to the evolution of the trunk channel that they abut, the sedimentological archive of levee dammed lakes and valley fills may be invaluable in understanding palaeo-fluvial dynamics of the trunk channel. Thus, the contemporaneous evolution of Lake Futululu and the Mfolozi River Floodplain may be used as a point of departure from which to study both systems.

In addition to providing insight with respect to system evolution, a conceptual model of alluvial ridge dammed features along a continuum from those with extensive regions of open water, to completely vegetated valley fill wetlands, is presented.

Study area

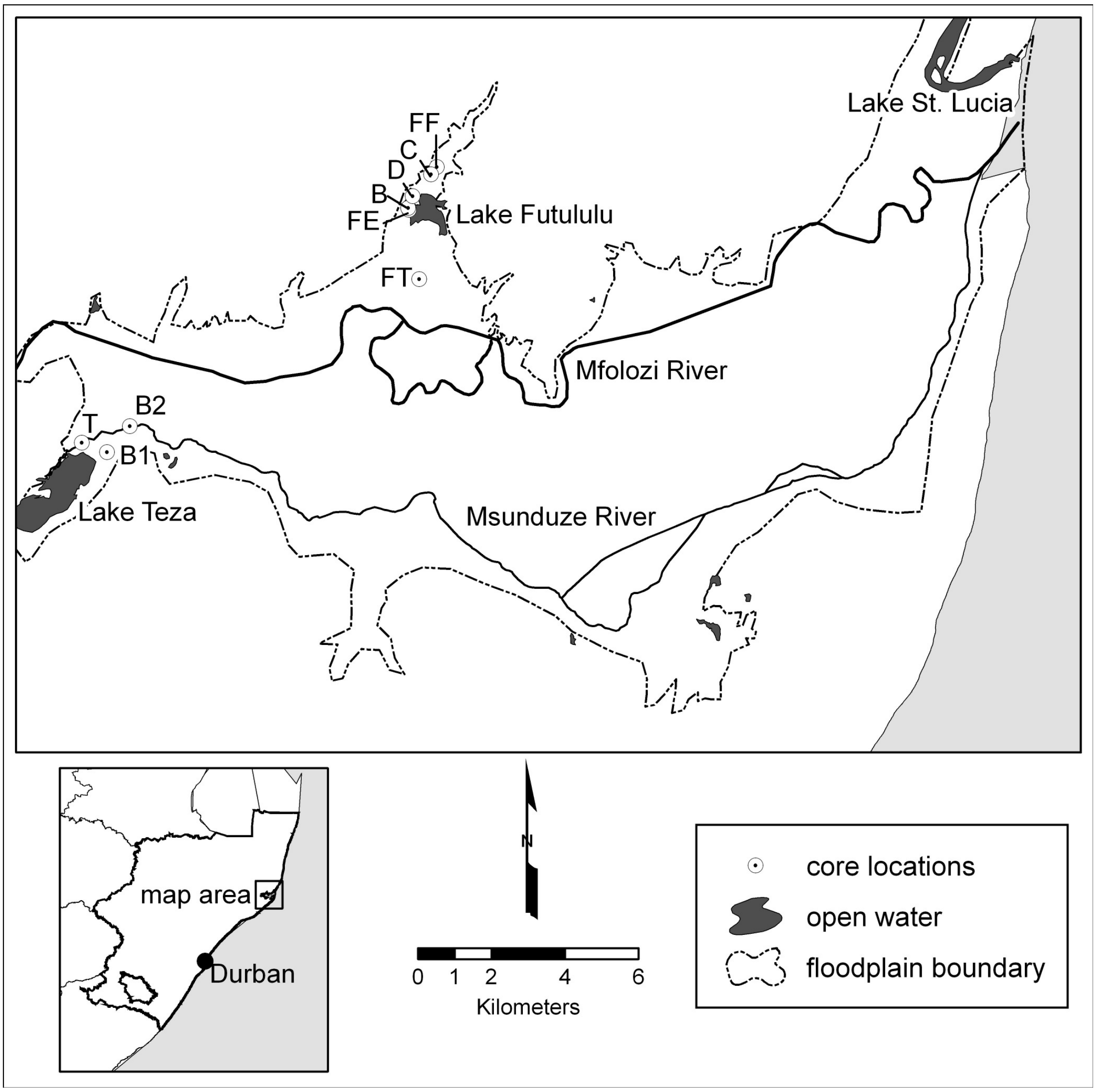

Figure 1: Location of the Mfolozi River Floodplain and the floodplain lakes; Lake Futululu, Lake Teza and Lake St. Lucia. The location of cores is also shown.

The Mfolozi River Floodplain is located on South Africa's eastern seaboard approximately 200km north of Durban on KwaZulu-Natal's coast (Figure 1). Upstream of the floodplain, the Mfolozi River occupies a confined valley underlain by rhyolite as it passes through the southernmost portion of the Lebombo Mountains, upstream of which the valley is on Karoo sedimentary rocks. As the river enters the coastal plain east of the Lebombo Mountains, the floodplain widens drastically, which is associated with the formation of an alluvial fan at the floodplain head (Chapter 4). To the south of the alluvial fan, a lesser river, the Msunduze, flows onto the floodplain, feeding into Lake Teza and then exiting the lake in the east and joining the Mfolozi River near the coast (Figure 1). Lake Futululu occupies a tributary drainage line on the northern floodplain boundary, while at the coast, the combined Mfolozi/Msunduze River flows just south of the Lake St. Lucia estuary. The Mfolozi River and Lake St. Lucia historically shared a common mouth as the Mfolozi River was forced northwards under the influence of longshore drift. The floodplain is bordered on its southern and northern sides by relatively steep outcrops of Zululand and Maputaland formation sedimentary rocks which are predominantly of marine origin, while the Maphelane dune cordon, which reaches over 100m in height, marks its eastern boundary.

Precipitation is largely restricted to the summer months when approximately 80% of the rainfall occurs, peaking between November and April (Tyson 1986). Mean annual precipitation in the catchment varies from 1288mm at the coastal town of St. Lucia, to 667mm in the mid- to upper- catchment, to 914mm in the upper catchment. Mean annual potential evapotranspiration is generally more than double that of precipitation, with atmospheric demands averaging 1800mm (Schulze 1997).

The Mfolozi Floodplain is dominated by clastic sedimentation, predominantly of silt size, that fines progressively towards the coast. The Mfolozi River is a meandering alluvial river dominated by suspended load. Aggradation on the alluvial ridge of the river is substantial, and results in it being elevated above the surrounding floodplain by ± 2.5m (Chapter 4). Peat accumulation on the floodplain is limited, largely because of the large clastic sediment input and highly seasonal nature of discharges that cause overbank flooding.

A notable exception in terms of peat accumulation is the Lake Futululu drainage line. This drainage line, flowing south from a small catchment (~1600ha in size) of *Eucalyptus* plantations and indigenous sand forest, is an unchanneled valley-bottom wetland (classified according to Kotze *et al*. 2007). *Cyperus papyrus*, *Phragmites mauritianus* and occasional *Ficus trichopoda* trees dominate the vegetation of the wetland. The drainage line culminates in a region of open water that averages 100ha in extent, although surface water area varies from year to year.

In contrast to the small catchment of Lake Futululu, Lake Teza in the south is supported by a constant water influx from the Msunduze River, which has an approximate catchment area of 30 000ha. Water accumulates in Lake Teza, evaporating and slowly draining into artificial drains that comprise much of the current day Msunduze River. The average extent of open water at Lake Teza is 240ha.

The largest of the lakes, Lake St. Lucia, is fed by five rivers, namely the Mkuze, Mzinene, Hluhluwe, Nyalazi and Mpati, that have a combined catchment area of ~665 000ha. From its estuary mouth just north of the Mfolozi River estuary, Lake St. Lucia extends some 60km north, trapped behind an ancient dune barrier complex. The average surface water extent of the Lake is approximately 24 700ha.

2. Methods

Lake Futululu was the most accessible of the floodplain lakes. In addition, the abundant peat deposits made it the preferred study area, as radiocarbon dating was possible. In contrast, Lake St. Lucia and Lake Teza have little or no peat accumulation and are populated by hippopotamus, hampering accessibility. The relationship between Lake St. Lucia and the Mfolozi River is further complicated by tidal influences. As such, the focus of this study was Lake Futululu. Nevertheless, three cores near Lake Teza allow further characterization of the relationship between the trunk Mfolozi River and smaller tributary drainage lines.

Six cores were sampled along the western margin of Lake Futululu, using a combination of a Russian peat corer and a clay auger. The cores close to the lake were retrieved to the depth of the sand bed of the previous valley surface. Two additional cores were sampled to the depth allowed by available equipment in Lake Teza (T and B1), and one north of the lake (B2). In the Futululu Cores B, C and D, the complete core was retained in a PVC pipe and sub-sampled in the laboratory. All the other cores were subsampled on site by placing contiguous 0.10m sections in plastic bags. The location of each core was recorded using a GPS with differential correction using a remote base station (sub-metre accuracy in the x, y and z fields), while cores B1 and B2 were recorded using a GPS with differential correction using a local base station (sub-centimetre accuracy in the x, y and z fields). Samples were characterized in terms of percentage organic matter content (%) and particle size distribution. Dried samples were burned in a muffle furnace for 4 hours at 450°C, following the methods of Heiri *et al*. (2001) and the organic content calculated as a percentage. Samples were classified as peat when the organic content was 30% or more. Particle size was determined using the Malvern Mastersizer 2000 after samples were prepared with hydrogen peroxide to remove organic matter. Particle size was described according to median particle size, as some samples showed bimodal distribution. The median particle size was classified according to Wentworth-Udden particle size classes.

Radiocarbon dates were determined for 10 samples, eight from Core FF and two from Core FE, by the CSIR Quaternary Research Laboratory in Pretoria.

A longitudinal profile of the Lake Futululu drainage line and a cross-section from the Mfolozi River into Lake Teza, was compiled from an assortment of data collected using the GPS instrumentation with differential correction facilities (sub-metre accuracy for a remote base station, sub-centimetre accuracy for a local base station). Corrections from spheroidal to geoidal surfaces were not made, as the variation between these surfaces was considered insignificant due to the localised nature of the data being collected. Valley cross-sections were drawn from 1:10 000 orthophotographs with 5m contour intervals (2.5m accuracy).

In order to get a sense of the topography of the pre-existing Futululu valley surface, the depth of recent sediment infill was determined by probing the sediment with steel poles until refusal. This method had been verified at core sites where the sediment could be examined at depth. Probing sediment infill was completed systematically along valley cross-sections at discrete intervals of 25m, which were measured using a measuring tape.

In addition, 5 generations of aerial photographs from 1937 to 1996 were studied in order to determine temporal variation in the extent and distribution of lake surface water. Aerial photographs were rectified based on orthophotography using control points, and digitisation was completed in ArcView GIS.

3. Results

3.1. Lake Futululu

3.1.1. Geomorphology

The Futululu Lake occupies a southward flowing drainage line that abuts the Mfolozi River. In 1937, the lake was extensive with a surface area of ~300ha (Figure 2). The lake's long axes was orientated from the north-west to the south-east, onto the floodplain periphery, while the Mfolozi River was characterised by a large loop towards the south away from the Futululu drainage line, before turning north just west of the Uloa Peninsula. Lake Futululu was characterised by two channels at this time, one flowing along the south-eastern portion of the basin, and a second called Crocodile Creek, on the south-western portion of the lake. In the 1937 aerial photograph, Crocodile Creek flowed into the lake basin as suggested by an arcuate sediment lobe in Lake Futululu. The south-eastern channel flowed into Lake Futululu, resulting in the formation of a delta. Between 1937 and 1960, sugar cane farmers on the Mfolozi Floodplain had begun to shorten the river's course by circumventing the large meander loop. In 1960, water flowed down both courses. The shorter Mfolozi River course flowed directly over previously flooded areas that comprised part of Lake Futululu in 1937. The area of open water in the lake was subsequently reduced by half to ~150ha. In addition, Crocodile Creek was extended upstream towards the west by sugar cane farmers to enhance drainage of cultivated fields.

By 1970, the Mfolozi River flowed only on the straightened course, and the old loop had been completely abandoned. The region of open water in Lake Futululu decreased to approximately 100ha. In 1988, the area of open water once again measured about 100ha and the two lake channels had been completely formalised as drainage furrows. Nevertheless, by 1996, the river course remained unchanged and surface water was substantially reduced to less than 10ha. In addition, the southern margin of open water had moved north by 400m, while the northern margin remained the same.

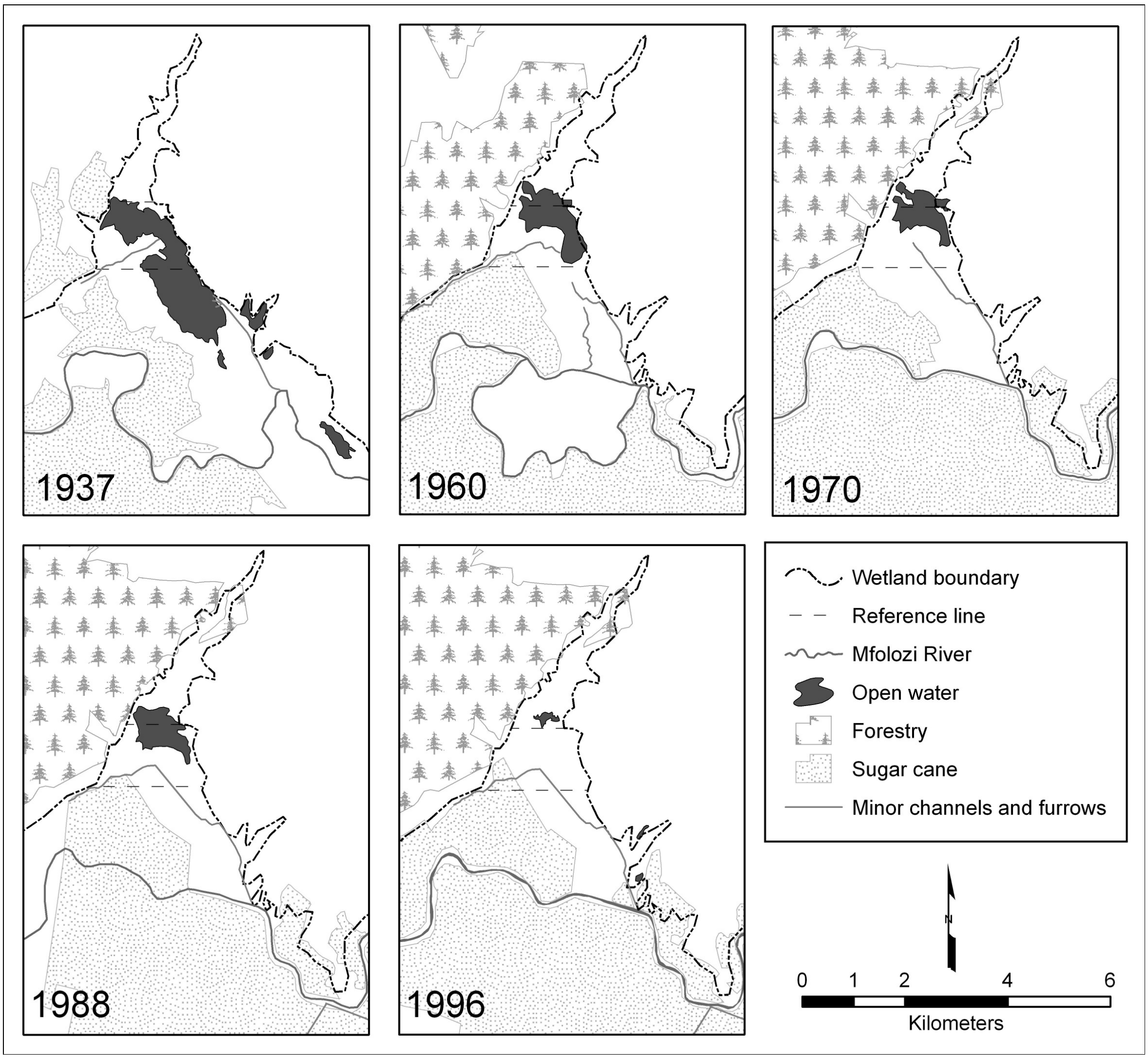

Figure 2: Aerial photograph representation of the extent of open water in Lake Futululu and the changing course of the Mfolozi River relative to the floodplain boundary.

Overall, the trend in the extent of open water over the period of record is a gradual movement of the southern lake margin towards the north. Using the reference lines in Figure 2, a comparison between years was made (Figure 3). Between 1937 and 1996, the southern shoreline moved northwards by over 3km. The largest change occurred between 1937 and 1960, when the southern margin moved north by 2.3km. Thereafter change occurred more gradually, except between 1988 and 1996 when movement north accelerated, resulting in the southern margin moving 705m north. Contrastingly, the northern margin has remained relatively static during historical times (Figure 3). However, between 1937 and 1960, the northern margin moved northwards by 370m. Thereafter, no systematic change of the northern margin northwards occurred.

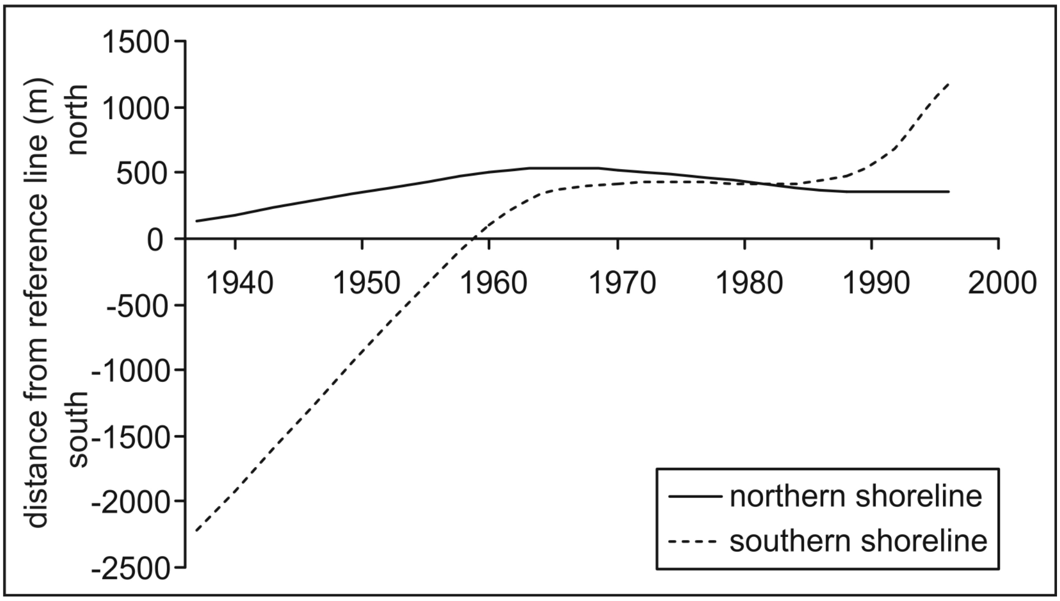

Figure 3: Movement of the northern and southern shorelines, measured as a distance from the upper and lower reference lines indicated in Figure 2, during the period of historic record.

The current Futululu valley surface is effectively level in cross-section and the wetland is situated between relatively steep valley walls (Figure 4). Towards the valley head, the valley surface is less than 1km wide (Figure 4a), and it widens progressively downstream to 5.5km toward the middle of the drainage line (Figure 4d), to more than 11km wide south of the lake (Figure 4f). In addition to an increase in width downstream, the valley slopes gradually downwards towards the southeast (Figure 4f).

The valley surface prior to peat formation was reconstructed for cross-sections a to e (Figure 4a-e). Unlike the current valley surface, the previous surface was u-shaped with a clearly defined thalweg. In the uppermost cross-section, one thalweg position is evident as a depression on the pre-existing valley floor (Figure 4a). In the following three cross-sections (Figure 4b-d) the presence of two thalweg positions is evident, while in cross-section e this is reduced to one (Figure 4e).

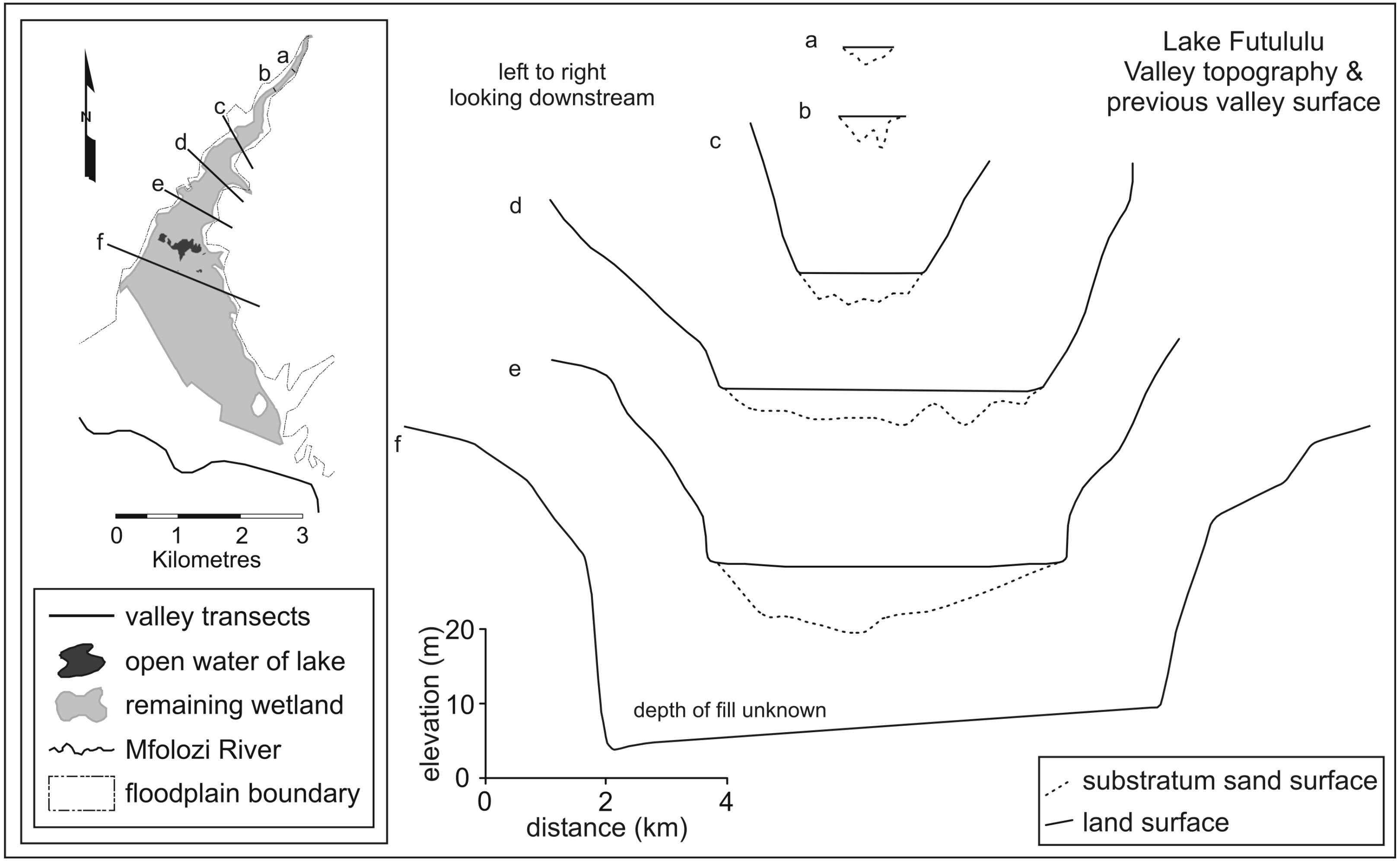

Figure 4a-f: Topography and bathymetry of the Futululu Valley looking downstream. Topography of valley sides was unavailable for cross-sections a and b.

A longitudinal profile of the Futululu valley (Figure 5, inset) illustrates the valley's position relative to the Mfolozi River. The Mfolozi River occupies a large ridge that is currently aligned perpendicular to the Futululu drainage line. The levees of the Mfolozi River are 6.4m above the valley's lowest point (Core B on inset of Figure 5). The elevation of Core B and the elevation of surface water in the Mfolozi River at the time of survey (water in the river was approximately 0.5m deep) were similar.

Open water usually occurs between Cores FE and D (planform location of cores is indicated in Figure 1). This is associated with a depression in the lake bed (at Core B). In addition to this depression, there is substantial variation in elevation along the longitudinal gradient of the valley that cannot be attributed to GPS error. However, the variation could conceivably be attributed to the proximity of the measurements to the valley side. Nevertheless the average current valley gradient upstream of the lake was determined as 0.02%, half of the gradient of the previous valley base determined by probing, which was 0.04%.

3.1.2. Sedimentology

The stratigraphy and organic contents of the 6 Futululu cores are illustrated in Figures 5 and 6 respectively.

Core FT was closest of the Futululu valley cores to the Mfolozi River, and was located close to the western lake shoreline in the 1937 photography. This 7m deep core was dominated by clastic sedimentation and organic content never rose above 10%. In addition, there was no systematic variation in organic content with depth. The core comprised alternating layers of fine to coarse silt with no peat.

Core FE was taken just south of the present day open water body of Lake Futululu, to a depth of 5.7m amsl. The core comprised numerous fine to medium silt layers that were characterised by varve-like laminations less than 5mm in thickness. A peak in organic content occurred between 7 and 8m amsl where organic content reached 58%. Below this depth, organic content was less than 10%. The change from low to high organic content typically occurred over less than 10mm in depth. Unfortunately the core was abandoned at 5.7m amsl as no more material could be retrieved.

Core B was characterised by a series of fine rhythmic laminations of clastic sediment and peat. An upwardly fining sand layer with organic contents less than 6% marked the base of the core. Above this basal sand layer, organic contents rapidly increased to 59% at 3m amsl. Thereafter, organic content decreased somewhat systematically up the length of the core, with a second subsidiary peak encountered at 6.1m amsl (Figure 6).

While the top portion of the core was too loosely consolidated to retrieve, the upper section of the core was characterised by alternating layers of organic clay, fine silt and very fine silt. At 4.4m amsl, the core coarsened to medium silt, below which another sequence of alternating clay peat and fine silt layers occurred. In general, the thickness and frequency of organic sediments increased with depth.

The previous valley surface in Core D was again marked by an upwardly fining sand and silt layer with very low organic content (<8%). Above the basal sand layer, organic content increased sharply, reaching a peak at 5m amsl (organic content 66%). Thereafter, there is a systematic decrease in organic material towards the surface with the thickness and frequency of peat layers decreasing upwards in the core. As such, thick peat layers mark the lower portion of the core, whereas towards the surface, clastic layers become more numerous. The median particle size of clastic layers varies from clay to fine silt. Once again, the clastic and peat layers were rhythmically varved in appearance.

Core C was located north of the current open water region of the lake. The core, 8.26m in length, reached the basal sand at 2.5m amsl. The upward fining sequence began with medium sand at the lowermost position, and fined upwards to fine sand and then very fine silt. Organic contents in this sequence were less than 2%. Above the basal sand unit, the core was predominantly peat, with thin interlayers of very fine silt at 6.6 and 4.8m amsl. At the top of the core, a floating peat mat was encountered within which a sample could not be retrieved. Two trends in organic contents with depth were visible in this core. Upwards from the basal sand (organic content < 2%), organic content increased to 57% at 2.8Ma amsl. Thereafter, there was a gradual decrease in organic content, to an elevation of about 6.5m amsl (organic content 11%). Above this low, organic content again rose to a peak of 81% at 7m amsl.

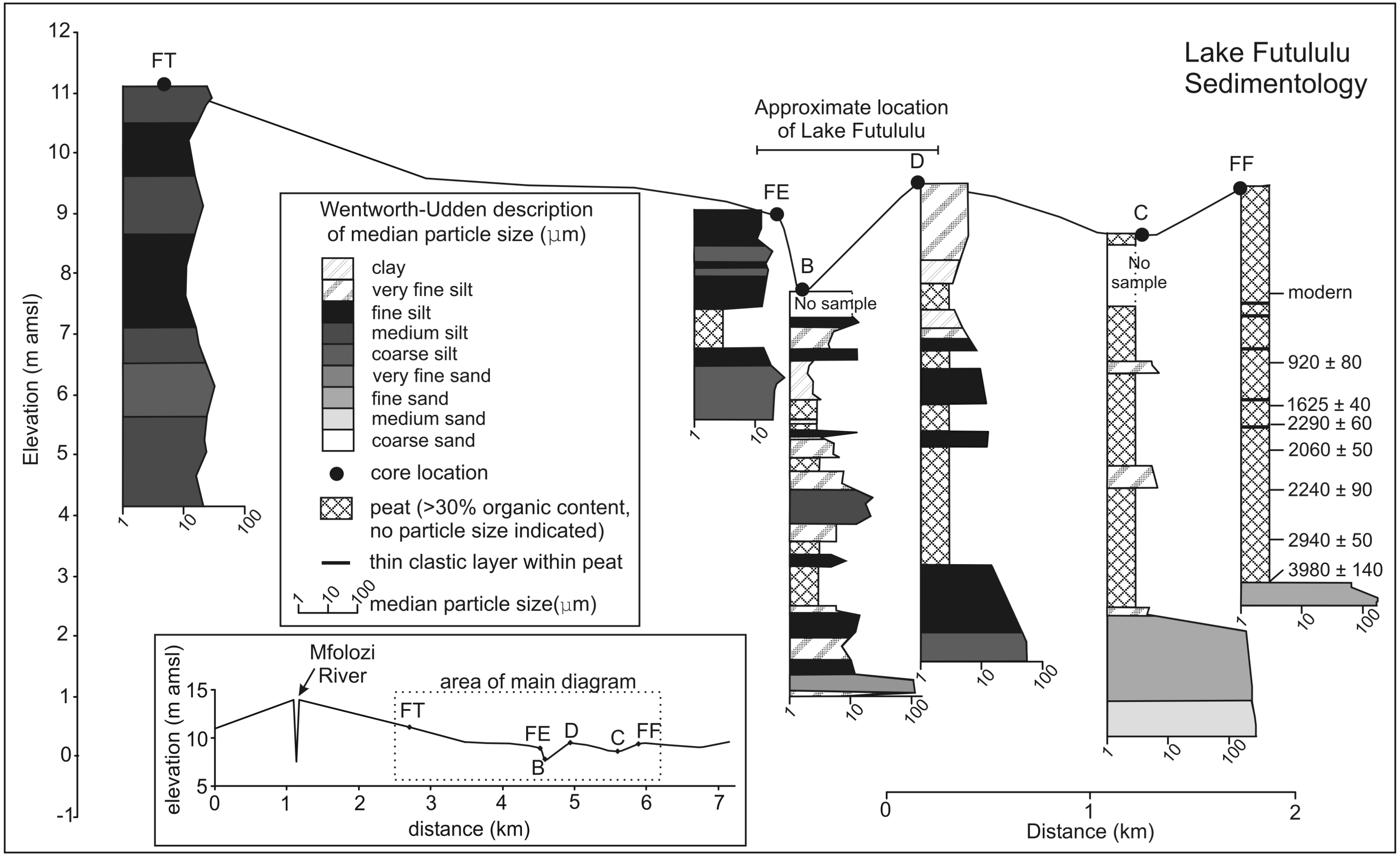

Figure 5: Sedimentology of the Lake Futululu drainage line. Thin clastic layers within peat are only indicated for Core FF. Longitudinal profile of the drainage line and the location of core holes on the longitudinal profile are indicated on the inset.

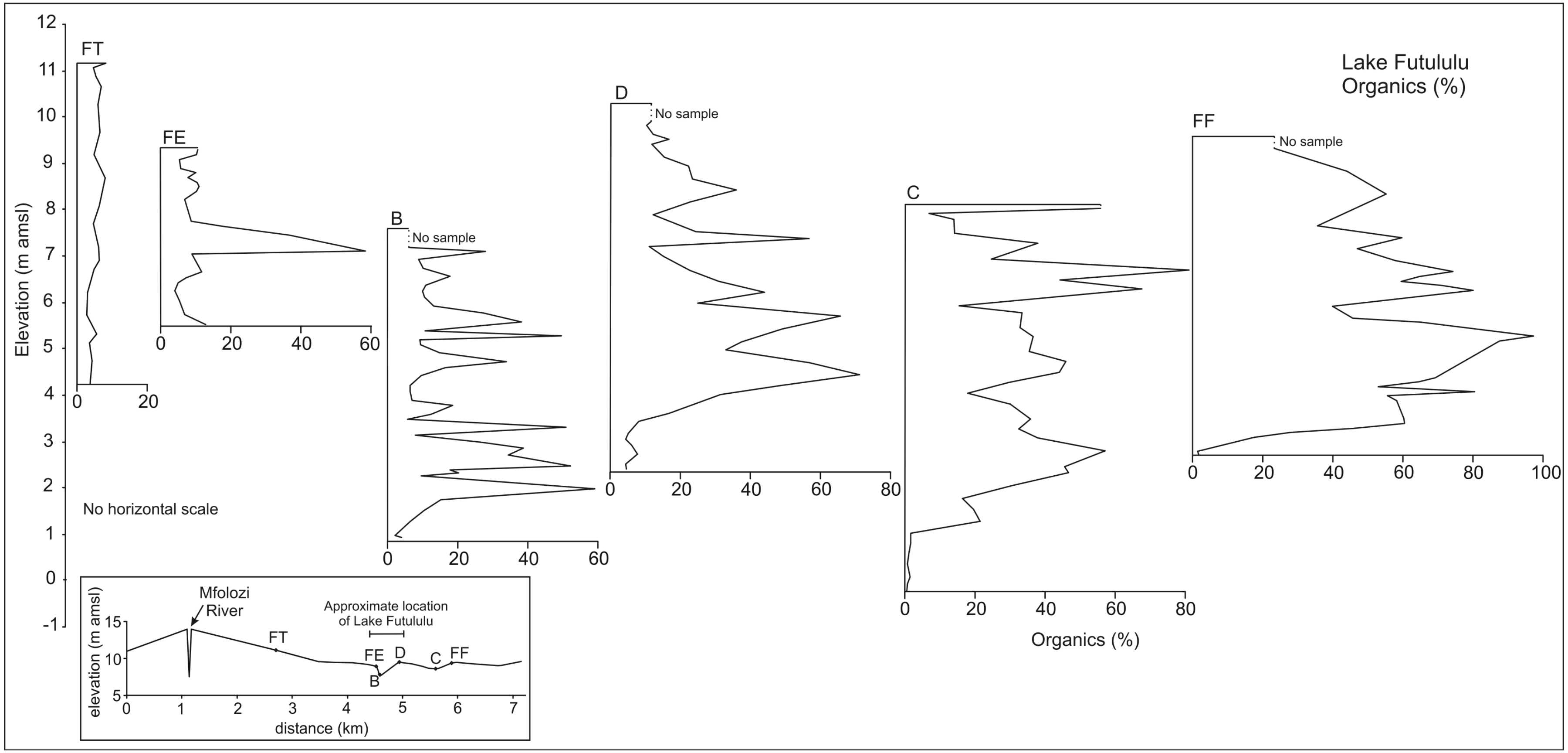

Figure 6: Organic contents with depth along the Lake Futululu drainage line. Location of the core holes indicated on the inset.

The final Core FF was 7m in length and reached the upwardly fining basal sand sequence at 3.2m amsl with an organic content of less than 2%. Up to this depth, organic contents were sufficiently high to be classified as peat. Organic content followed a fairly systematic trend with depth. From the base of the core, there was a systematic increase in organic percentage until about 5.5m amsl. At this depth, a compressed charcoal layer approximately 0.30m deep was encountered (organic content 98%). Above the charcoal layer, organic content gradually decreased towards the surface.

There were a number of similarities between the sedimentology of the cores. Most noticeably, several cores were characterised by an upwardly fining sand to silt base with very low organic content (typically < 2%). The change in organic content, from very low in the basal sand sequence to a peak in each core's organic content generally occurred between 2 and 3m up from the sandy base. The peak in organic content near the base of the core was usually followed by a systematic decrease in organic content upwards, towards the ground surface. As such, most cores were characterised by the greater occurrence of peat at depth, with increasing amounts of clastic sediments towards the surface. Similarly, the amount of peat increased up-valley. Core FT had no peat, and relatively low organic contents throughout. Cores FE, B and D exhibited inter-fingering of peat and clastic sediments, while Core C had only two thin layers of very fine silt, and Core FF was entirely peat. The particle size of clastic sediments decreased from the Mfolozi River towards the Lake. Varve-like laminations, suggesting processes of cyclic sedimentation, were characteristic of almost all the cores and were especially noticeable in clastic sediments.

In addition to the systematic trend of decreasing organics, Cores FF, C, D and B all displayed organic content minima at a similar elevation. At Core FF, this minimum occurred just above the charcoal layer, at approximately 6m amsl. Cores B, D and C had minima at elevations of 5.5, 7 and 6m amsl respectively. Considering the GPS error, it appears that the minima may be at similar depths, within a range of less than 1m.

3.1.3. Radiocarbon ages

The stratigraphic position of samples from Cores FE and FF that were dated using ^{14}C methods is illustrated in Figure 5. A brief description of each sample's characteristics, together with the ^{14}C age, is given in Table 1. All the samples were peat, although they varied in degree of humification and compaction. Generally, compaction increased with depth. Many of the samples were characterised by *Cyperus papyrus* plant fossils, particularly rhizomes. Contrastingly, sample FF 4.2-4.3 consisted of highly compacted charcoal that measured some 0.30m in thickness.

Table 1: Radiocarbon ages of samples selected from Cores FE and FF. A brief description of each sample is also given.

Core	Analysis number	Sample Name	Description	^{14}C result	Depth (m)
FE	Pta 9693	FE 1.375-1.5	Compressed peat	880 ± 150	1.40
	Pta 9696	FE 2.37-2.5	Fibric peat	155 ± 40	2.40
FF	Pta 9699	FF 1.5-1.9	Compressed fibric peat	modern (i.e. >1950)	1.70
	Pta 9687	FF 2.8-2.9	Peat with *Cyperus papyrus* rhizome material present	920 ± 80	2.85
	Pta 9698	FF 3.5-3.65	Peat with *Cyperus papyrus* rhizome material present	1605 ± 40	3.55
	Pta 9690	FF 3.8-3.9	Peat	2290 ± 60	3.85
	Pta 9688	FF 4.2-4.3	Charcoal (0.30m thick layer)	2060 ± 50	4.25
	Pta 9697	FF 5.0-5.1	Peat with *Cyperus papyrus* rhizome material present	2240 ± 90	5.05
	Pta 9692	FF 5.8-5.9	Peat with *Cyperus papyrus* rhizome material present	2940 ± 50	5.85
	Pta 9694	FF 6.7-6.8	Peat with *Cyperus papyrus* rhizome material present	3980 ± 140	6.75

In Core FF, all but one of the dates was stratigraphically consistent (sample FF 3.8-3.9). In Core FE, radiocarbon ages for the two samples processed were reversed stratigraphically. Processing difficulties, and the rather unlikely age outcome, suggests that these samples may have been contaminated and have thus been disregarded.

The relationship between depth and ^{14}C age was used to interpolate the ages of peat layers in Core FF that had relatively high clay contents.

3.2. Lake Teza

3.2.1. Geomorphology

Between the Mfolozi River in the north and Lake Teza in the south, the floodplain is characterised by a large sediment mound that slopes towards the south (Figure 7, inset). At the mound's greatest elevation, it exceeds the height of Lake Teza by just less than 6m. At the time of a floodplain survey in April 2005, with sub-centimetre accuracy in the z field, the water surface of Lake Teza was slightly elevated above the water level in the Mfolozi River. The groundwater surface also sloped away from the centre of the mound, towards the Mfolozi River in the north and Lake Teza in the south.

3.2.2. Sedimentology

The sedimentology and organic contents of the three cores associated with Lake Teza are illustrated in Figure 7. Core B2 was collected just over one kilometre north of Lake Teza. It was found to be markedly coarser than any of the other cores from Lake Futululu and Lake Teza. The depth of the core was unfortunately limited by the repeated collapse of the core 4m below the ground surface. The lower 1.1m of the core comprised fine sand. This was overlain by an upward fining sediment sequence that ranged from medium to fine silt. Contrastingly, the upper 1.9m of the core was an upwardly coarsening sequence of very fine to medium sand. Organic contents throughout the core were very low.

Core B1 was collected on the north-eastern lake periphery amongst a stand of *Phragmites australis*. In contrast to B2, B1 was extremely fine grained. The majority of the core was fine silt, with one thin (< 0.1m) layer of very fine silt at a depth of 2.3m. There were three regions of coarsening in the sequence, with layers of coarse silt at depths of 0.3, 4.2 and 5.9m. Organic contents were once again low throughout, at less than 12%. The highest organic content was at the surface (ash content 11%), but below this point, organic contents were always less than 10%.

The last core, Core T, was collected on the north-western margin of Lake Teza in reclaimed wetland that has been planted with sugar cane. The base of the core constituted a 1.5m deep upwardly fining sequence that graded from coarse to fine silt. Above the upwardly fining sequence, a sequence of upwardly coarsening sediment, from coarse silt to very fine sand to fine sand was evident. The very top of the core comprised a layer of medium silt 0.3m thick. Organic contents of Core T were very low throughout, with the highest content recorded at a depth of 3.5m (organic content 7%).

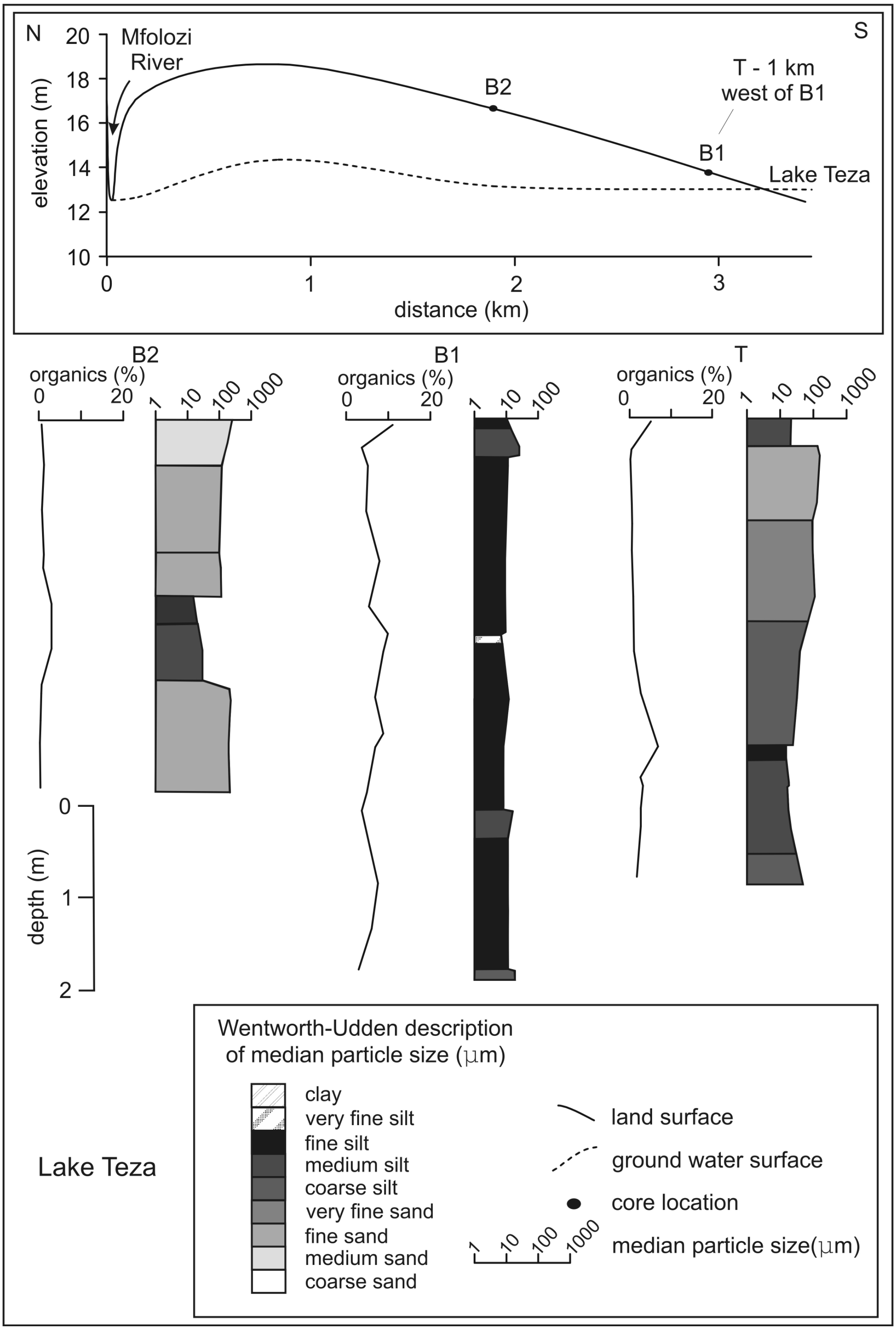

Figure 7: North-south cross-section from the Mfolozi River in the north to Lake Teza in the south. Location of cores is indicated on the cross-section, sedimentology and organic content is presented below.

In general, the set of Lake Teza cores show a marked fining towards the lake. Core B2 is much coarser than any of those described, primarily comprising sand. In addition Core B1, which was located towards the north-eastern boundary, was much finer than Core T, which was on the north-western boundary. Only Core B1 also displayed varve-like sedimentation at some depths. All the cores displayed very low organic contents. However, organic contents did increase towards the lake, and higher values were associated with a decrease in particle size.

4. Discussion

4.1. The evolution of Lake Futululu

During the last glacial maximum (18 000BP), sea levels along southern Africa's eastern seaboard were approximately 120m below the present level. The lowered base level caused large-scale vertical incision along coastal areas of KwaZulu-Natal in particular, as rivers were rejuvenated. Incision of the Mfolozi River, upstream of the current day floodplain, resulted in entrenched meanders developing in erosion resistant Lebombo Group rhyolites. Downstream, the Mfolozi River encountered softer and less resistant sedimentary rocks of the Zululand and Maputaland Groups, consisting of siltstone, fossiliferous beds and palaeodune deposits. These easily eroded rocks were vertically incised and laterally eroded, resulting in the confined meandering river course opening into a deep wide valley, in which the current day Mfolozi Floodplain is now situated.

As the Mfolozi River adjusted to the lowered sea levels through erosion, tributaries of the trunk channel were simultaneously rejuvenated. The Futululu Valley and Msunduze River Valley of Lake Teza were such valleys, and during this period they also carved out deep wide valleys into underlying soft sedimentary units of the Zululand and Maputaland Groups. The elevations of the beds of these streams would have been determined by the local base level established by the bed of the Mfolozi River.

Sea level rose steadily from the low attained during the Last Glacial Maximum, until approximately 8000BP, when sea level reached –1m amsl (Ramsay 1995, Figure 8). Thereafter, the rate of sea level rise slowed, reaching current levels approximately 6500BP, followed by the Holocene highstand that persisted for approximately 2500 years. Sea level reached a maximum height during this period of +3.5m amsl 4480BP (Ramsay 1995). Following this, sea level dropped to –2m amsl during the Little Ice

Age approximately 3000BP rising again to +1.5m amsl 1610BP before settling to its current level.

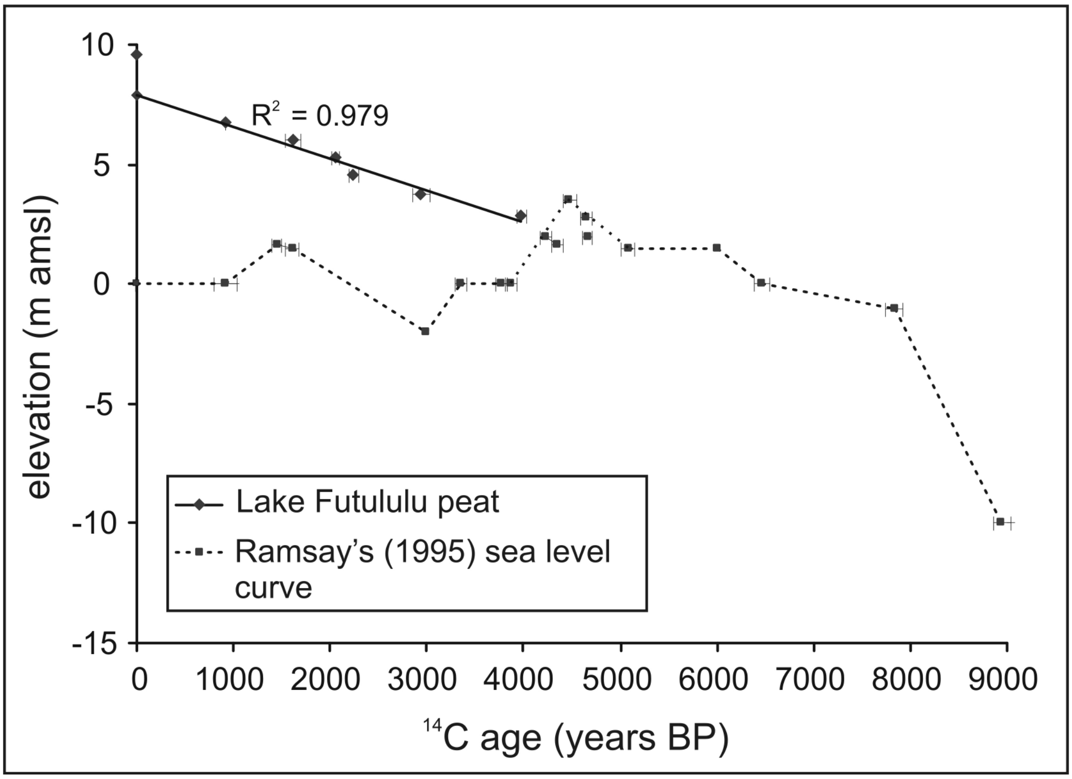

Figure 8: Peat accumulation in Lake Futululu (excluding sample FF 3.8-3.9) compared to Ramsay's (1995) sea level curve.

Rising sea level caused rivers, which had previously been rejuvenated by the lowering of sea level during the Last Glacial Maximum, to deposit their loads in the bay of the drowned valley, and this reduced slope in an upstream direction leading to aggradation upstream of the newly formed Mfolozi Bay. Wide-scale incision was thus replaced by deposition as rivers began to lose sediment transport capacity and competence as they approached sea level from the eroded interior. In a gross sense, the Mfolozi River thus filled or partially filled the lower Mfolozi River Valley with sediment. In the same way that a rising sea level caused deposition of sediment in the lower Mfolozi Valley, aggradation and floodplain development along the Mfolozi River caused deposition along the Futululu Valley by raising its base level. It is of interest that sea level rose above the bed of the Futululu drainage line without causing deposition in the area, as evidenced by subsequent peat formation 500 years after the highstand. A large portion of the lower Mfolozi Floodplain was below sea level during the highstand 4480BP, resulting in the formation of a bay. Despite the Futululu Valley (at Core FF) being situated below sea level at this time, the lack of lacustrine infilling suggests the ocean did not drown this area. Clearly aggradation of the Mfolozi River upstream of the Mfolozi Bay had raised the elevation of the floodplain above sea level, blocking off incursion of seawater as far upstream as the Futululu Valley. Interactions of sea level, sedimentation on the Mfolozi River Floodplain, and pre-existing topography would have influenced the patterns and extent of incursion of seawater up the Mfolozi Valley following the Last Glacial Maximum.

Despite not being flooded directly by seawater, tributaries such as the Futululu drainage line were affected by the rising base level. The rapid formation of the Futululu Basin suggests that drowning of the Mfolozi Floodplain, and thus the formation of a protected bay, allowed deltaic sedimentation in the central floodplain region. As deposition of sediment along the Mfolozi River blocked the lower Futululu Valley, slope up the tributary valley was reduced, leading to aggradation of the Futululu Valley. Two streams that arose on the sandy coastal plain in the upper region of the Futululu Valley at this time, each with a distinct thalweg evident in cross-sections b to d (Figure 3), probably exhibited logarithmic longitudinal profiles in equilibrium with discharge and available sediment supply. As the base level at the toe of the Futululu stream rose with rising sea level, these became depositional systems – probably with little or no clastic sedimentation on the valley floor due to limited capacity to transport sandy clastic sediment. Thus, rising sea level altered the evolution of the Futululu valley from an erosional to a depositional environment, by raising base level and lowering valley slope. 4480BP, sea level regressed. However, the Futululu Valley remained a depositional environment, indicating that the Mfolozi River had replaced sea level as the base level of the Futululu Valley by this time.

Since the Mfolozi River drains a steep, sub-tropical hinterland, sediment supply is relatively high. In contrast, the Futululu catchment is very small and gently sloped. The superior sediment supply and transport capacity of the Mfolozi River resulted in rapid vertical aggradation that resulted in the Mfolozi River filling the Mfolozi Valley in the area of the floodplain. The super-elevation of the Mfolozi River prevented smaller tributaries from joining the main stem. The already aggrading Futululu system lacked the energy to erode a channel through the Mfolozi alluvial deposits, and eventually lost all capacity to transport sediments. By 3980BP, deposition on the Mfolozi Floodplain and alluvial ridge had created a basin in the Futululu Valley. As capacity to transport sediment was lost, peat accumulation began relatively rapidly as *Cyperus papyrus* dominated the wetland valley. The formation of peat in the Futululu Basin was linked to low clastic sediment inputs of the two tributary streams, combined with the formation of a basin that allowed permanent standing water to halt aerobic decomposition. In southern Africa, these conditions are relatively rare due to strongly seasonal climatic conditions. In addition, the geomorphic evolution of the subcontinent has resulted in the region undergoing a period of long-term erosion, a situation that causes most fluvial networks on the eastern seaboard to carry large amounts of sediment, while also inhibiting the formation of closed basins.

Aggradation on the Mfolozi Floodplain and alluvial ridge continued, increasing accommodation space and causing the accumulation of peat to propagate further upstream in the Futululu valley. Organic deposits in the Futululu Valley are thus characterised by a decrease in depth upstream towards the valley head (Figure 9). In addition, the base of the peat deposits towards the head of the valley would be more youthful than those towards the Mfolozi River, where the Futululu Basin is oldest.

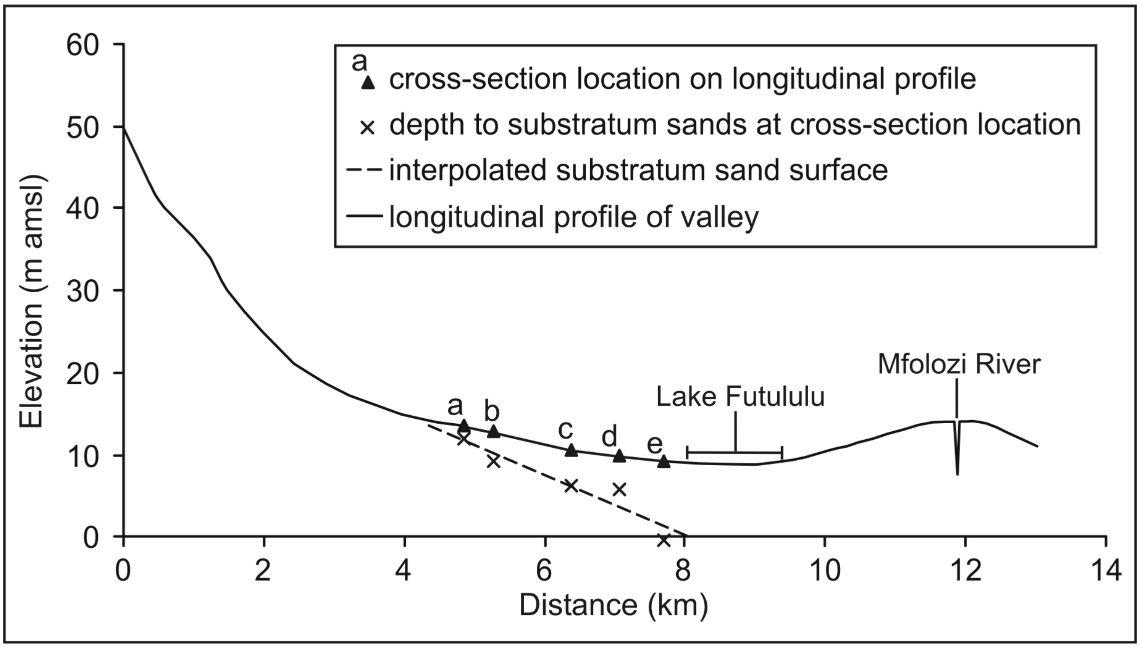

Figure 9: Shape and depth of peat accumulation in the Futululu Valley.

Peat accumulation continued at a constant rate of approximately 13mm.a^{-1}, despite variations in sea level and climate. Between 3360BP and 1610BP, sea level fell below the present level, reaching –2m amsl approximately 3000BP (Ramsay 1995). Lee-Thorp *et al.* (2001) attributed vegetation changes in the interior of southern Africa to a cooler environment between 3200 and 2500BP, a change that was also reflected in the temperature index of the Cango stalagmite (Talma and Vogel 1992). This neoglacial period, causing a cooler and drier climate in southern Africa, has been recorded across the globe (Ramsay 2005). This period is marked in the Lake Futululu Core FF by a 0.30m band of charcoal at a depth of 4.2m. These remains of a large peat fire were dated at 2060 ± 50 BP, a date reasonably consistent with the latter period of climatic cooling. Decreased rainfall and groundwater inputs resulted in water table lowering and shrinkage of the surface area of the lake. The resulting desiccation of the peat deposits caused extensive burning.

While the rate of peat accumulation remained relatively constant during the last 2000 years, it appears that the influence of the river in supplying clastic sediment to the lake became greater towards modern times. Each time a large flood occurs on the Mfolozi River, clay and very fine silt are injected into the Futululu Basin. The floods are preserved in Core FF as thin clastic bands bounded by peat, the age of which can be interpolated using the age-depth relationship established from ^{14}C dating. The first layer of clastic sediment in the predominantly organic core occurred approximately 1700BP, followed by floods again at 1470, 700, 550 and 180BP.

Considering that impoundment first resulted in peat accumulation 3980BP, the impact of floods is relatively recent, being confined to the last 2000 years. The clustering and increasing frequency of floods may be suggestive of the onset of a warmer and wetter climate as the globe came out of the neoglacial period. An alternative, and perhaps more likely scenario, is that as aggradation continued on the Mfolozi River's alluvial ridge, clastic sediment that originally impounded the drainage line began to encroach on the Futululu Basin. The increasing elevation of the Mfolozi Floodplain has resulted in sedimentation along the Mfolozi River extending further and further up tributary valleys, such as the Futululu Basin. With time, one might expect that the distance at which sediment may be deposited up valley will increase as fluvial facies begin to overlap onto lacustrine sediments. The observable result of increasing clastic inputs is the movement of the southern lake margin northwards up the valley, assumedly resulting in the slow loss of lake depth with time (Figure 3).

However, the last flood that caused clay accumulation at Core FF occurred 180BP. It is likely that agricultural authorities on the floodplain have controlled more recent floods, preventing the deposition of clastic sediment in the upper reaches of the Futululu Valley. Furthermore, the tendency for the Mfolozi River to now avulse towards the southern region of the floodplain upstream of the Lake Futululu basin suggests that under natural conditions, clastic inputs from the Mfolozi River would in all likelihood decrease, essentially lengthening the geomorphic life of the basin.

Between 1937 and 1960, the Umfolozi Sugar Planters Co-operative straightened a section of the Mfolozi River such that it flowed over the southernmost portion of Lake Futululu. This event drastically reduced the size of the lake as the alluvial ridge began aggrading onto the lake bed. In 1937, Lake Futululu was aligned along a northeast axis, and occasionally shared a link to the Mfolozi River when the lake reached its maximum level. Following straightening of the Mfolozi River course, sedimentation during large floods caused the southern boundary of the lake to move northwards while the northern margin has remained largely static. In addition, Lake Futululu no longer has a northeast axis. Since the 1960's, the lake has not reached a maximum retention level and water has not flowed from Lake Futululu into the Mfolozi River. The changing morphology of the lake suggests that in historic times, the movement of clastic sediment into the basin from the southwest has changed to a more rapid sediment input from the south.

The current proximity of the Mfolozi River has elevated the base level of the Futululu Valley, further creating accommodation space, and has thus increased rates of peat accumulation. At a depth of 1.7m, peat deposits were found to be of a modern age. While the upper portions of peat are uncompressed, an accumulation rate of 23mm.a^{-1} exceeds the preceding rate.

4.2. Origin and description of sedimentary facies of drowned tributary valleys

The low slope and small, well vegetated catchment of the Futululu valley results in almost no sediment being transported down the Futululu valley into the catchment, as can be verified by peat formation up valley (Figure 5). The lack of sediment inputs from the Futululu valley allows the anatomy of the Mfolozi Floodplain / Futululu valley drainage line interface to be more easily characterised.

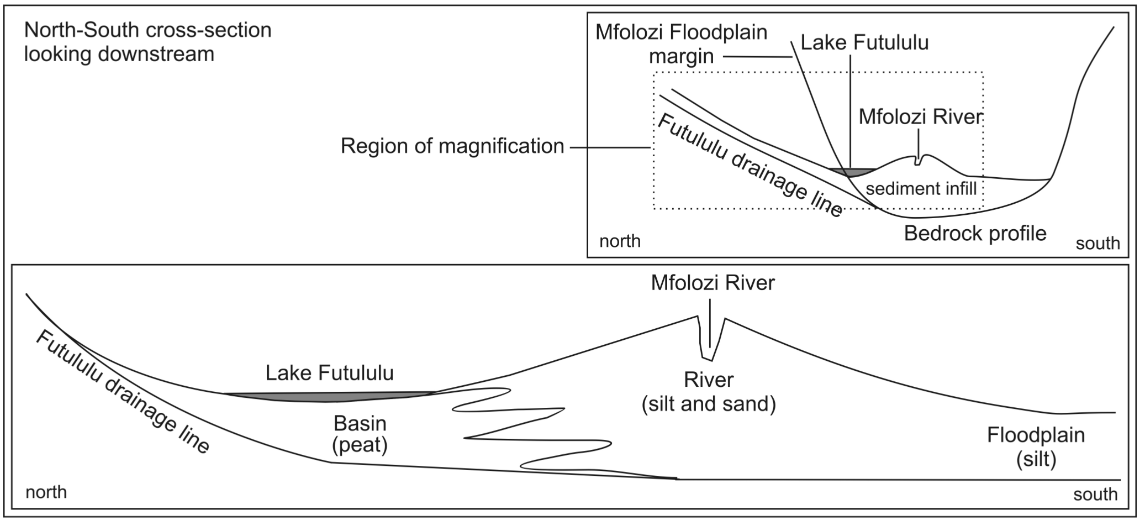

Figure 10: Generalised sedimentary facies development in the Futululu Basin.

An understanding of the geomorphic co-evolution of the two systems suggests there are two main processes influencing sedimentary facies in the drainage line. The first is that sediment fines with distance away from the Mfolozi River, an observation frequently described in the literature (e.g. Pizzuto 1987, Magilligan 1992). Secondly, ongoing aggradation on the Mfolozi River is increasing the height and width of the alluvial ridge, essentially extending the region in which the river may deposit sediment during a flood. Thus, over time, the area that may receive clastic sediment input is increasing. The combination of these processes creates the characteristic sedimentary facies of the Futululu valley / Mfolozi Floodplain interface (Figure 10).

During flood events, the Mfolozi River overtops its banks and pushes sediment-laden water into the Futululu Basin. Sediment is sequentially deposited with coarser sediments closer to the river, and finer sediments transported up the vegetated valley. The finest particles, suspended clay and very fine silt, are transported furtherest up the valley, where eventually they fall out of the water column and are deposited on the peat valley floor. Between floods, clastic sedimentation ceases and peat formation predominates wherever water is permanent and aerobic decomposition is prevented. This process, of clastic sedimentation during flood events, with sediment fining away from the Mfolozi River, results in an inter-fingering of peat deposits and clastic sediments at the floodplain-valley interface.

If aggradation on the Mfolozi River were static, one would expect the region of inter-fingering to remain in a static position over time. However, ongoing aggradation on the alluvial ridge widens it, while also increasing the height of the river above the valley. As a result, floods laden with clastic sediment reach progressively further north with time, creating the characteristic stacking pattern of peat, peat inter-fingering with clastic sediments, followed by overtopping levee silt. The region of inter-fingering therefore moves progressively north as regions closer to the river eventually become overwhelmed with clastic sediment, and accommodation space is filled. The basin is also pushed northwards up the drainage line, effectively moving the locus pf peat accumulation.

The resulting geomorphic processes allow the characterisation of three sedimentary facies in the Futululu drainage line; trunk fluvial, tributary basin and tributary valley. The trunk fluvial facies, characterised by silt sized clastic sediment deposited during flood events on the trunk, creates elevation on the alluvial ridge and allows progradation of the Mfolozi River sediments over the Futululu valley sediments. The tributary basin is characterised by organic sedimentation of *Cyperus papyrus* peat, with inter-layers of clastic sediments towards the south. The basin facies is loosing accommodation space in the south, but this is currently replaced with increased accommodation space towards the north of the valley. The tributary valley facies is limited towards the valley head where peat accumulation is limited by seasonal water supply. In this region, a small stream is reworking sands that were deposited during the Holocene highstand.

4.3. Trunk-tributary relationships: a continuum

Trunk-tributary relationships appear to operate at two different scales. The first is where tributary sediment supply is high, usually due to a steep catchment. As such, the rate of aggradation in the region where the tributary meets the trunk channel is relatively high. When the sediment supplied by the tributary channel exceeds the sediment transport capacity of the trunk channel, sediment accumulates at the stream confluence. Over time, the sediment supplied by the tributary may have a perceivable impact on the trunk channel, as it is unable to transport the sediment that the tributary supplies. In this case, the confluence is tributary dominated (Figure 11). On the opposite scale, some tributaries have relatively gentle gradients, and the amount of sediment they supply to a trunk channel is negligible as compared to the amount of sediment transported by the trunk channel. In these cases, trunk channels have the ability to drown the confluence with sediment, and prevent the tributary stream from joining the flow. In this case, the trunk channel controls the geomorphic outcome, an the relationship may be considered to be trunk dominated. These two different scales should not be considered as isolated responses to different tributary sediment regimes, but rather as two opposite ends on a continuum of trunk-tributary relationships.

The evolution of Lake Futululu represents the trunk dominated end member of the trunk-tributary relationship continuum (Figure 11). Since sediment inputs from the Futululu stream were extremely low as compared to sediment transport on the Mfolozi River, aggradation on the trunk channel overwhelmed the Futululu drainage line, effectively cutting it off from the drainage network and creating a lake (lacustrine environment).

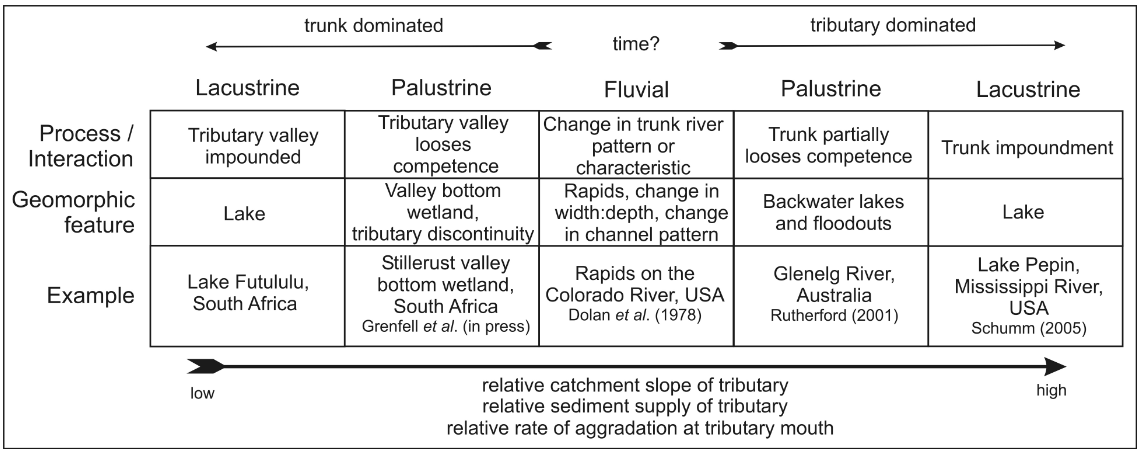

Figure 11: Conceptual continuum of trunk-tributary relationships, from trunk dominated systems on the left, to tributary dominated systems on the right.

In some cases, the trunk channel does not carry very large amounts of sediment, and the rate of alluvial aggradation is slower. However, aggradation still overwhelms inflowing tributaries and results in the tributary streams loosing competence even though a basin may not have formed behind the trunk channel's alluvial ridge. In such a system, the greater aggradational potential of the trunk channel results in the formation of a valley-bottom wetland (as defined by Kotze *et al*. 2005) or tributary channel discontinuity. An example of the development of a palustrine environment at a trunk dominated tributary confluence is the formation of the Stillerust valley-bottom wetland abutting the trunk Mooi River alluvial ridge (Grenfell *et al*. in press).

Where tributary sediment inputs are not so great as to overwhelm the trunk channel, or the trunk is not so large that it overwhelms the tributary channel, normal fluvial processes will dominate. In these systems, tributary sediment inputs may be assimilated by adjustments to the trunk channel. For example, most major rapids on the Colorado River, USA, have been attributed to sediment discharge from steep tributaries (Dolan *et al*. 1978). Furthermore, Schumm (2005) describes how sediment from tributaries may cause the channel pattern or width-depth ratio of the trunk channel to change to accommodate increased sediment supply. Nevertheless, despite increased sediment inputs from tributaries, or the high sediment load of the trunk channel, both systems are able to compensate and accommodate the change to the sediment transport regime, without palustrine or lacustrine type drowning occurring.

In contrast, trunk channels that are dominated by tributary sediment inputs may be liable to palustrine or lacustrine drowning. In the tributary dominated end member, extremely large sediment discharges, from a usually steep tributary, may completely or partially block the drainage line of a trunk channel, causing the trunk to loose competence. The trunk channel becomes impounded by sediment supplied by the tributary as it lacks the stream energy to transport it. For example, Lake Pepin, on the upper Mississippi River, was formed as glacially derived sediments transported by the Chippewa River, partially blocked flow in the Mississippi River (Schumm 2005).

Sometimes sediment discharge from tributary streams is less severe, and may serve to intermittently block trunk channel flow resulting in the formation of a palustrine, rather than a lacustrine, environment. Sand supplied by the tributaries of the Glenelg River accumulates in the trunk channel at the stream confluence, creating backwater lakes and wetlands (Rutherford 2001).

4.4. Trunk dominated tributaries: the Mfolozi Floodplain lakes

Three large lakes occur on the peripheries of the Mfolozi Floodplain, the smallest is Lake Futululu in the northern central floodplain. Lake Teza is the second largest lake, fed by the Msunduze River. Towards the coast, the Mfolozi River flows just south of Lake St. Lucia, a major lake system fed by five rivers. Each of these lakes is extremely different in character. Lake Futululu is a peat filled basin, which currently has no channelled tributary. Lake Teza, on the other hand, has very little peat, and the Msunduze River is a source of permanent fresh water supply. Lake St. Lucia is a clastic dominated basin, which also receives marine sediment when it has an open connection to the ocean (Wright and Mason 1993). While all three lakes have the Mfolozi River in common, not all three are classic versions of impounded tributary basins and indicate a large degree of variability within classes of trunk-tributary relationships.

The effect of a larger tributary on the evolution of a tributary impounded valley is well documented by the characteristic differences between Lake Teza and Lake Futululu. Sediment inputs from the Msunduze River prevent the formation of peat in Lake Teza, and recently, enhanced catchment erosion has led to more rapid infilling of the lake basin (Scott and Steenkamp 1996). Clastic sedimentation is much faster than organic sedimentation, with 18m of sediment accumulating in Lake Teza since 8330 ± 120BP (Scott and Steenkamp 1996). This translates to $46mm.a^{-1}$, as compared to organic accumulation rates of $13mm.a^{-1}$ in Lake Futululu. In addition to sediment inputs from the Msunduze River, the south-eastward tilt of the Mfolozi Floodplain and current avulsion tendencies of the Mfolozi River (Chapter 4) make the Teza Basin more susceptible to infilling from large floods on the Mfolozi River. Cyclone Domoina deposited approximately 0.5m of fine sand at Core T, a deposit that was also noted by Scott and Steenkamp (1996).

Despite the comparatively larger catchment water and sediment discharge of the Msunduze River into Lake Teza, as compared to the Lake Futululu drainage line, elevation data clearly shows that Lake Teza is also situated in a basin created by flood deposition of the Mfolozi River (Figure 7). However, unlike Lake Futululu, Lake Teza has a stream outlet as the Msunduze River exits Lake Teza towards the northeast of the lake. This suggests that it is not just aggradation on the floodplain that causes tributary drowning, but rather relative rates of aggradation. Lake Teza has not been completely blocked by Mfolozi Floodplain aggradation, as tributary inputs of sediment and water are large enough to maintain the slope required for a channel. The maintenance of an outlet channel allows the continual removal of sediment from the lake, preventing the lowering of channel gradients upstream of the river. In turn, this prevents complete loss of stream competence upstream of the lake. The maintenance of an outlet channel therefore positively reinforces itself. Similarly, positive feedback mechanisms are highly influential once a tributary becomes drowned by trunk channel aggradation. Once drowned, slope is reduced upstream, stream energy is further reduced, and the slope is further reduced by sediment deposition within the lake.

Lakes Teza and Futululu may legitimately be called tributary lakes as their formation is strictly linked to aggradation on the Mfolozi River Floodplain. The same term cannot be applied to Lake St. Lucia, as aggradation on the Mfolozi River has not led to development of the St. Lucia basin. Lake St. Lucia is situated behind a barrier dune complex, through which the Maputaland Rivers have been unable to erode a channel to the ocean since sea level rose. However, the location of the St. Lucia estuary mouth just north of the Mfolozi River mouth is not coincidental.

The Mfolozi River has a larger discharge than any of the rivers flowing into Lake St. Lucia. As a result, when the Mfolozi River estuary is open, it is flood dominated (Lindsay *et al*. 1996). In comparison the estuary of Lake St. Lucia is ebb-dominated (Wright and Mason 1993), and under most conditions, struggles to maintain an open mouth. Considering this information, it is likely that the Mfolozi River maintained a channel to the ocean following rising and then lowering sea levels, rather than Lake St. Lucia. Furthermore, aggradation on the Mfolozi River fixes the southernmost extent of Lake St. Lucia at Honeymoon Bend. Digital elevation models of the Mfolozi River Floodplain show large expanses of the lower floodplain are below sea level (Chapter 4). If Lake St. Lucia maintained the channel outlet, and was not fixed in position by aggradation on the Mfolozi River, it is likely that these regions would also be flooded by Lake St. Lucia. Thus, while Lake St. Lucia is not a tributary lake per se, the stream flow and sediment discharge of the Mfolozi River has and continues to play a large role in the Lake's geomorphic evolution.

5. Conclusion

Trunk-tributary relationships and the features they create can be considered on a continuum from those that are dominated by aggradation of the trunk channel, to those that are dominated by comparatively large sediment inputs from tributaries. Both end members of the scale are characterised by lacustrine settings, of which Lake Futululu provides a well-documented case of evolution. Where relative differences in aggradation rates are less, palustrine systems are frequently the result. In contrast, many trunk-tributary interactions may not be dominated by either system, resulting in fluvial processes of sediment transport dominating.

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Chapter 6. Discussion: Wetland formation and system processes in a context of Fluvial Geomorphology

1. Introduction and context

Wetlands of southern Africa occur throughout the sub-continent against all climatic and geomorphic odds as high evapotranspiration rates commonly result in negative water balances (Schulze 1997). Furthermore, two unusual uplift events occurred 20Ma and 5Ma, leaving the region in a long-term state of erosion (Partridge and Maud 1987). It has been argued by Ellery *et al*. (2008), that wetlands occur in non-erosional settings, such that wetlands should be rare in southern Africa. Given this combination of circumstances, wetlands in the region favour drainage lines where groundwater seepage is concentrated or where fluvial inputs of water lead to a positive water balance, and/or erosion may be momentarily halted by the formation of local base levels where the surrounding rock is more easily weathered and eroded.

Many wetlands in southern Africa form where valleys situated in areas of relatively weatherable and erodible rock are crossed by a resistant lithology such that erosion is limited in the valley upstream of the resistant lithology (Tooth *et al*. 2004, Grenfell *et al*. in press). Others form through complex trunk-tributary relationships, where the stream with the greatest sediment load tends to impound the other, such as the drowned tributary of Lake Futululu. Loss of confinement may result in the formation of alluvial fans, the most spectacular example of which is the Okavango Fan (McCarthy *et al*. 1992). Lastly, sea level provides the base level for many wetlands occurring on the coastal plain, such as the Mfolozi River Floodplain. Despite the wide variety of modes of formation and the multiple factors that shape the evolution of a wetland from the development of a local base level, all southern African drainage line wetlands share an evolutionary theme. All of them evolve to a particular gradient that is a product of their catchment discharge and sediment flux history. Furthermore, due to the climatic gradient of southern Africa, with wetter regions in the east and dryer regions towards the west, wetland size rather than catchment size, is the best predictor of wetland gradient.

This discussion considers the implications of the continuum of drainage line wetlands in terms of hydrogeomorphic wetland types, their morphology and reasons for differences in morphology. From this broad-scale view of drainage line wetland evolution, the evolution of the Mfolozi River Floodplain is evaluated in terms of equilibrium concepts and process rates. The linkages between coastal floodplains and river-dominated estuaries of KwaZulu-Natal are then reflected upon in terms of equilibrium and system dynamics. Finally, the problems that variable rivers and drainage line systems (as systems of multiple scales) pose to management are contemplated, particularly with respect to likely changes in climate in the future.

2. The continuum of drainage line wetlands and geomorphic thresholds

The relationship between wetland size and slope creates a hydrogeomorphic continuum of wetlands, whereby larger wetlands with bigger catchments are low gradient floodplains, while smaller, steeper wetlands are characteristically valley bottom wetlands. Floodplains and valley bottom wetlands differ in their response to sediment influxes. Floodplains alter their gradient through local erosional and depositional processes in response to variation in discharge and sediment supply in the floodplain stream. Floodplains are therefore fluvially controlled in a fairly strict sense. In contrast, valley bottom wetlands are unable to adjust their gradient to the amount of incoming water and/or sediment because these are diffuse and sediment is not strictly dispersed via the fluvial system. As a result, valley bottom wetlands tend to progressively steepen since significant hydrological and clastic sediment inputs enter these systems from their heads. Ultimately, steepening results in erosion as the geomorphic slope threshold is reached, either entirely by natural means, or by increased hydrological and/or reduced sediment influxes due to the effect of catchment land uses or possibly climate changes. It appears that there is thus scope to further refine the geomorphic continuum of drainage line wetlands, perhaps including hillslope seepage wetlands that have the smallest catchments and the steepest slopes, followed by unchanneled valley bottom wetlands with larger catchments, channelled valley bottom wetlands with even larger catchments and lastly, floodplains, with the largest catchments and therefore significant fluvial inputs.

The geomorphic continuum of drainage line wetlands is conceptualised in Figure 1. Data suggests that wetland slope is negatively related to wetland size such that they generally occur within a narrow range of slopes for their size. Therefore, for the wetland types being described small, gently sloped wetlands and large, steep wetlands are geomorphically impossible. The envelope of slope relative to size may be divided into three major wetland types as previously described, where the location of the zone of transition between valley-bottom and hillslope seepage wetlands has yet to be determined, and would be a useful point of departure for future research.

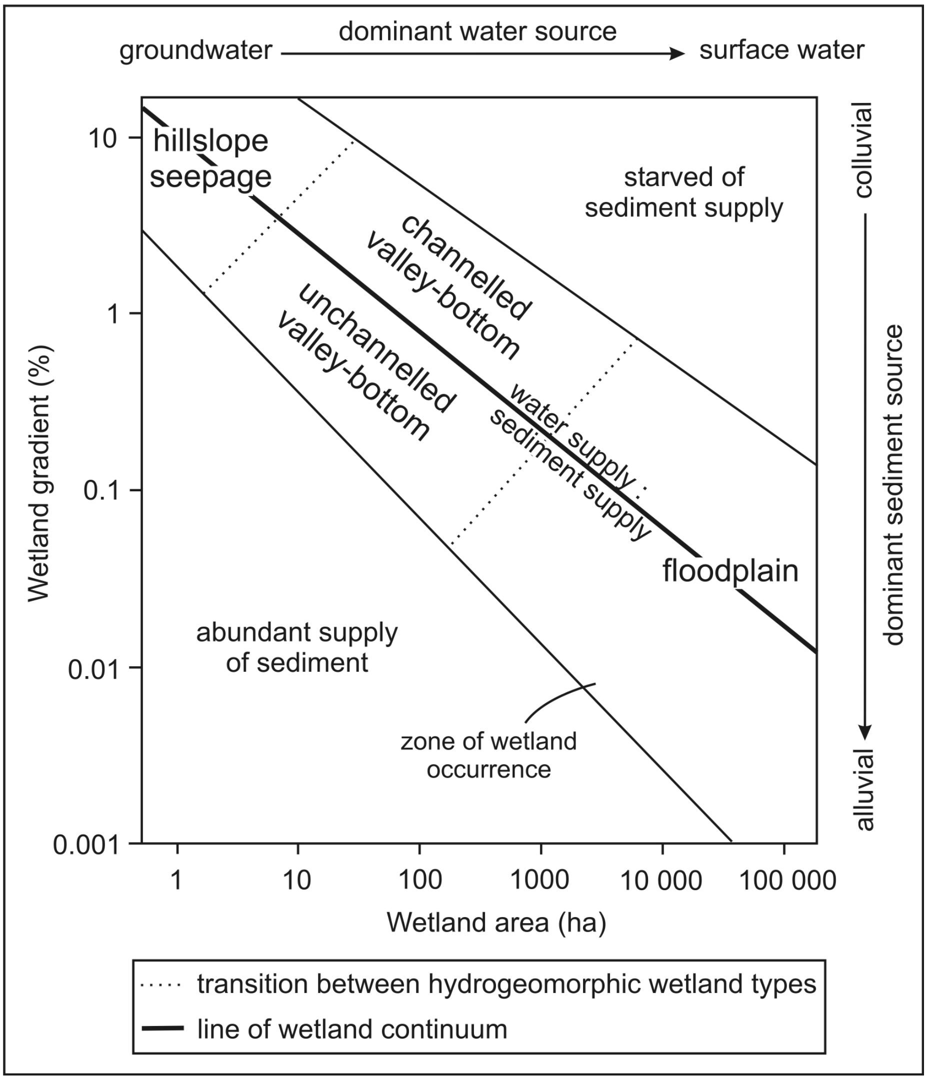

Figure 1: Conceptual diagram of the continuum of wetland hydrogeomorphic type. The central line is a trendline describing wetland gradient relative to wetland area.

The continuum of wetland size and slope also reflects the characteristics of wetland processes, particularly with respect to dominant sources of water and sediment inputs. Small wetlands, including hillslope and some valley-bottom wetlands, experience a high proportion of water input from groundwater sources or diffuse surface water inputs. Large wetlands, including some valley-bottom wetlands and all floodplain wetlands, receive a high proportion of water inputs as surface flow (stream flow) as compared to groundwater inputs. Wetlands of intermediate size experience a variable proportion of groundwater and surface water inputs. Wetlands that receive greater proportions of surface water than groundwater inputs tend to have larger total water inputs, probably reflecting the considerably greater efficiency of stream flow than either groundwater flow (very inefficient) or diffuse flow (moderately efficient). As such, the groundwater-surface water continuum is also representative of overall discharge.

Dominant sediment sources are also represented along a continuum of colluvial sediment supply from hillslopes, to alluvial sediments from valleys and rivers. Hillslope wetlands, towards the top of Figure 1, are primarily reliant on sediment moved under the influence of gravity, as discharges are seldom high enough to actively transport sediment in this setting. Valley-bottom wetlands may receive colluvial and alluvial sediment, depending on the connectivity between the wetland and the surrounding hillslopes. The importance of alluvial sediment inputs increases as wetland size increases, and as such, floodplain wetlands are settings of alluvial sedimentation. In addition to indicating sediment source, the continuum also gives an indication of the amount of sediment supply relative to water input. As colluvial sources tend to be limited in supply by processes of weathering, wetlands that receive alluvial sediment primarily, tend to receive a greater amount of sediment than those reliant on colluvial sources.

The slope of the trendline describing gradient in relation to wetland area may be conceptualised as indicating the mean ratio of water supply as compared to sediment supply to a wetland. In Figure 1, those wetlands that lie above the line receive larger water inputs relative to sediment inputs and may thus be considered to be sediment starved as compared to the mean condition. In contrast, wetlands that fall below the mean line receive an abundant supply of sediment relative to water inputs. It can be seen, therefore, that those wetlands that tend towards being sediment starved are more vulnerable to incision than those where water supply is predominantly via groundwater. In contrast, where sediment supply to a wetland is abundant, it is unlikely to erode.

Figure 1 provides a useful framework in which to conceptualise the impacts of catchment land use and management on a wetland's geomorphic stability by considering how a wetlands sediment and water source is likely to change. For instance, increasing water supply without increasing sediment supply may lead to erosion. In this case, capacity exceeds available sediment load, and erosion occurs. Similarly, decreasing sediment supply to a wetland without changing water supply has the same result. Wetland systems are finely tuned ecosystems occurring where there is a geomorphic balance between water and sediment inputs and outputs. Increasing or decreasing either variable may lead to wetland degradation through incision.

The evolutionary link between floodplains and valley bottom wetlands is perhaps unsurprising as most differences between hydrogeomorphic types can be attributed to varying water and sediment inputs that then impact upon wetland morphology and processes. Floodplain wetlands are regions where a meandering or sinuous river occasionally overtops its banks, resulting in the characteristic morphology of a floodplain through differential sediment deposition that creates levees and overbank depressions. However, floodplains may vary substantially in form. Floodplains formed upstream of resistant lithologies are characterised by mixed bedrock-alluvial rivers, where the stream banks are composed of alluvium while the base comprises bedrock. Alluvial depth is usually limited in such settings (e.g. Tooth *et al*. 2004, Grenfell *et al*. in press). In contrast, the origin of the Mfolozi River is extremely different, and rather than undergoing incision at its toe (such as would happen at the downstream end of the mixed bedrock-alluvium wetlands), the Mfolozi River valley is a region of infilling and aggradation as sea levels have risen since the last glacial maximum.

As such, the Mfolozi River Floodplain is different from many southern African wetlands in terms of its evolutionary history. However, a shared climatic setting results in many floodplains of southern Africa differing strongly from their northern temperate counterparts. The major anastomosing or meandering river systems situated in North America, such as the Saskatchewan River (Cazanacli and Smith 1998, Morozova and Smith 2000; 2003), Columbia River (Makaske *et al*. 2002) and Mississippi River (e.g. Gomez *et al*. 1997, Hudson and Kesel 2000, Kesel 2003) have been widely studied. In addition, Törnqvist and Bridge (2002) and Törnqvist (1994) compared sedimentation on the Rhine-Meuse and Mississippi alluvial floodplains. Gradziński *et al*. (2002) studied the upper Narew River in Poland, an anastomosing system characterised by peat banks and bedload dominated channels. All of these studies have tended to focus on the evolution of anastomosing systems into single channelled rivers, as well as the development of crevasse splay complexes that have led to river avulsion (e.g. Slingerland and Smith 1998, Smith *et al*. 1989, Morozova and Smith 2000, Törnqvist and Bridge 2002, Makaske *et al*. 2002). In the temperate northern hemisphere, all studies classified floodplain sediments as levee, overbank or channel based on particle size and organic content, an idea not inconsistent with southern African studies. Overbank regions are frequently characterised by peat (e.g. Makaske *et al*. 2002, Smith *et al*. 1989, Morozova and Smith 2000, Gradziński *et al*. 2002, Törnqvist and Bridge 2002). Since peat accumulation is linked to the formation of basins and requires low sediment influx, one should expect floodplain systems with high sediment influxes and limited peat formation to differ substantially in form and function. For instance, Michaelsen *et al*. (2000) found that compaction of peat by overlying aggrading clastic sediments proactively altered floodplain evolution by differential creation of accommodation space.

The strongly seasonal subtropical climate of the Mfolozi Floodplain, which is characterised by a hot, wet summer followed by a cool, dry winter, is not hospitable to peat formation as desiccation occurs between annual periods of flooding. As such, the Mfolozi River Floodplain is dominated by clastic rather than organic sediment accumulation. The Mfolozi River is variable hydrologically and variable flow is associated with inter- and intra-annually variable sediment loads that arise from changes in catchment sediment availability that are associated with climate. Thus, climate has a large impact on the Mfolozi River Floodplain, not only by preventing accumulation of organic sediment, but also by controlling the frequency and amount of water and sediment delivered to the system.

The climatic and geomorphic context of the Mfolozi River Floodplain places it at the centre of several ongoing theoretical debates. The first considers the relative rate of processes and the contrasting views of gradualism and punctuated equilibrium with respect to the processes and geomorphology of the Mfolozi River Floodplain. The relationship between the functioning of river-dominated estuaries and coastal floodplains will be considered based on Cooper's (1993, 1994, 2001) analysis.

3. Equilibrium in the context of a variable flow river: the Mfolozi River Floodplain

Equilibrium concepts provide a framework from which to conceptualise dynamics and processes of the Mfolozi River Floodplain. A frequency analysis of mean annual flows of the Mfolozi River showed a skewed distribution with a high occurrence of below median flows with a small number of years with extremely high flows. Not only was stream flow of the Mfolozi River variable between years, but flow was also shown to be flashy within years, as flows are generally low, while flood flows are large and infrequent. Thus, during the majority of the year, the stream flow of the Mfolozi River is generally low and does not result in overtopping of the levees. In these circumstances, the amount of inflowing suspended sediment equates to the amount exiting the floodplain at the estuary mouth. In addition, bedload measurements suggest a region of channel bed erosion in the central floodplain while deposition, which seems to be related to slope adjustment, occurs towards the lower reaches. Matter inputs and outputs on the Mfolozi River are generally near or at equilibrium during normal flood conditions, conforming to the idea of stasis.

Periods of stasis are interrupted by large flood events, during which the Mfolozi River overtops its banks and loses confinement. Flooding results in a loss of sediment transport capacity as the wetted perimeter is increased, leading to large-scale sediment deposition, as occurred during Cyclone Domoina. Sediment deposition during smaller events is apparently negligible compared to large, infrequent flood events.

Thus, considering the ratio of sediment inputs to sediment outputs as an indication of floodplain equilibrium, the long-term equilibrium status of the Mfolozi River Floodplain may be conceptualised (Figure 2). During normal flow conditions, which are generally low, the Mfolozi River does not overtop its banks and the amount of sediment transported into and out of the floodplain remains roughly balanced, with internal variation related to localised slope adjustment. Extremely large flood events occur episodically, either as a response to tropical cyclone atmospheric circulation or to the occurrence of cut-off low pressure systems that occasionally (annually or more than once annually) reside over KwaZulu-Natal. Long-term hysteresis, over a number of years, suggests that maximum sediment availability occurs during the early onset of a wet spell, following a dry spell. When a flood event occurs in tandem with a high rate of catchment sediment supply, the period of stasis is interrupted as the river overtops its banks and wide scale deposition occurs. During events such as this, inputs to the floodplain greatly exceed outputs, and the system is not at equilibrium in terms of matter or energy balances.

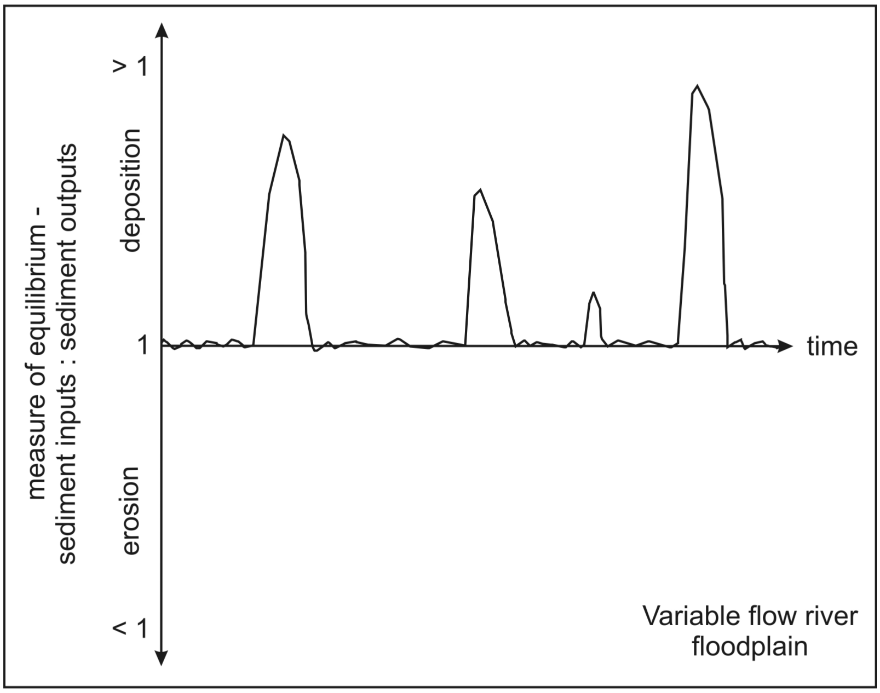

Figure 2: Schematic illustration of equilibrium with respect to sediment budget on the Mfolozi River Floodplain over time.

Patterns of sediment accumulation in the Mfolozi River Floodplain are therefore strongly linked to climatic controls. Rainfall variability, in conjunction with the large catchment of the Mfolozi River, its highly seasonal rainfall with a wet summer season, and a large proportion of evergreen vegetation in the catchment, all contribute to increased stream flow variability. In addition to influencing stream flow variability, the occurrence of wet and dry years influence the amount of catchment sediment available for stream transport.

Variable flows and variable sediment transport result in sediment accumulation in the Mfolozi River Floodplain occurring as a series of steps, rather than as a gradual, ongoing process (Figure 3). Available accommodation space reached a maximum following the last glacial maximum. Thereafter, rising sea levels resulted in a gradual decrease in available accommodation space. For the last 4000 years, sediment accumulation has been sporadic and linked to large flood events, rather than ongoing.

Sedimentation on the floodplain given a fairly constant sea level, is gradually reducing the available accommodation space.

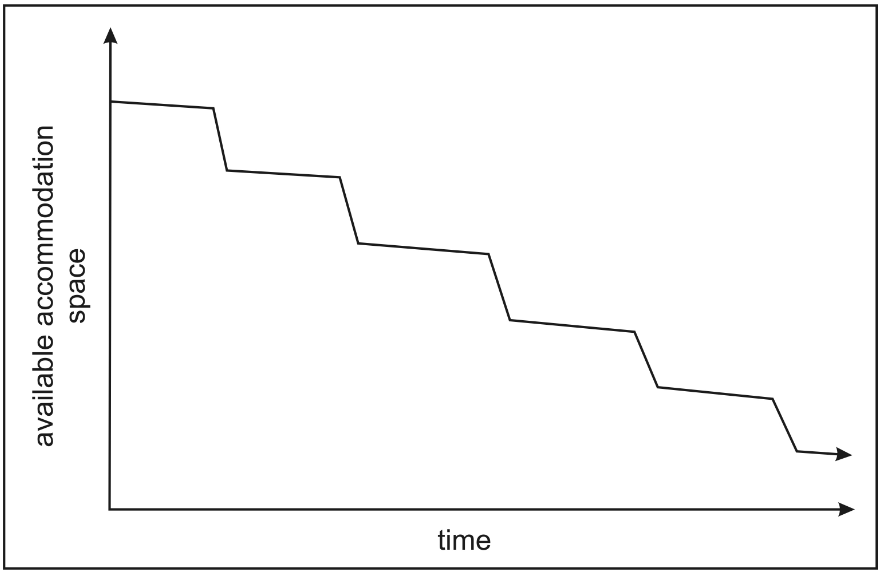

Figure 3: Schematic conceptualisation of the reduction of available accommodation space due to sedimentation on the Mfolozi River Floodplain.

While the clastic region of the floodplain aggrades sporadically, there is pluralism in floodplain process in that some regions of the floodplain are characterised by steady, gradual processes consistent with gradualist notions. Peat accumulation in the Lake Futululu drainage line has seemingly been unaffected by the sporadic flood events of the Mfolozi River or by changes in climate, accumulating instead at a steady rate of 0.17 $cm.a^{-1}$. This suggests that the alluvial fill in the Mfolozi Floodplain, which acts as the base level of the Futululu drainage line, is also aggrading at a relatively constant rate when considered over hundreds of years. This is in contrast a to large portion of the floodplain that aggrades mostly during large flood events, when 4m of sediment can be attributed to a single event. The occurrence of both gradualism and punctuated equilibrium in a single system should not come as a surprise (see Gould 1984, Jensen 2004).

In addition to questions of frequency of infill, the perceived rate of aggradation, as measured as a change in elevation over time, is also dependent on the shape of the Mfolozi River Basin. In Scenario 1 (Figure 4), the infilling over time of a v-shaped bedrock valley is considered. The same volume of sediment is assumed to have been accumulated during each time unit (t_{0-6}). At the onset of infilling, between t_0 and t_1, the elevation of the valley floor rapidly rises due to the shape of the valley floor. Over time, the valley floor rises progressively more slowly.

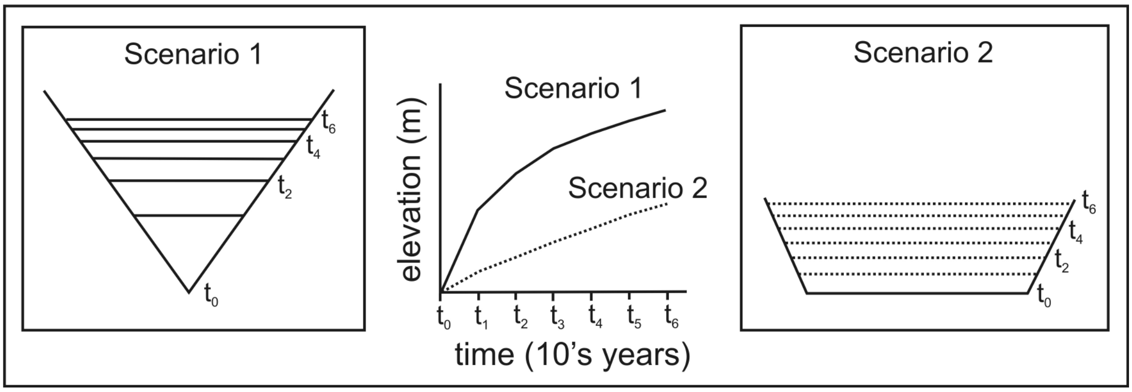

Figure 4: Perceived aggradation rates caused by variation in basin shape.

In Scenario 2 (Figure 4), the shape of the bedrock valley is assumed to be flat with steep, almost vertical, sides. With the same volume of sediment accumulating between time periods, the elevation of the valley floor rises at an almost constant (but slightly decreasing) rate. By t_6, the rate of rise in valley floor in both scenarios is approximately the same as the valleys are of equal width. Unfortunately, the shape of the underlying bedrock valley of the Mfolozi River has not been verified.

4. Evolution and dynamics of the Mfolozi River Floodplain system

The concept of underfit is usually applied to streams in which discharge has decreased and the resulting valley is oversized relative to the stream (Schumm 2005). In the case of the Mfolozi River Floodplain, the result is similar in that despite lowered discharges in the past (associated with global cooling and drying, Ramsay 2005), the stream had greater erosional power due to increased valley gradient. Thus, the Mfolozi River valley is a classic inherited feature that is not altogether in equilibrium with current conditions. Furthermore, the inherited valley has influenced the evolution of the valley, as suggested by Ahnert (1994), by leading to floodplain aggradation in order for the Mfolozi River to reach grade. This suggests, according to Montgomery (1989), that the Mfolozi River Floodplain system cannot be considered to have reached equilibrium at the landscape scale because of the occurrence of inherited components. While the inherited valley has been the precursor to valley aggradation, the floodplain appears to have reached equilibrium in that sediment throughputs are generally maintained during periods of stasis. Thus, while geomorphic history may shape future floodplain evolution, it is currently fairly unimportant. Instead, stream flow variability currently plays the major role in the dynamics and evolution of the Mfolozi River Floodplain through controlling sedimentation rates and processes of system change.

Long-term aggradation of the Mfolozi River Floodplain is reliant on the movement of the Mfolozi River alluvial ridge across the floodplain since sedimentation rates in the backswamps close to the floodplain boundary are relatively small compared to aggradation of the alluvial ridge. Successive avulsions, which are a response to local differences in floodplain slope produced by sedimentation during large flood events, rather than ongoing depositional processes related to stream meandering, are responsible for large-scale movement of the alluvial belt across the floodplain, and thus floodplain aggradation. During periods of stasis, equilibrium processes prevail, resulting in the Mfolozi River behaving predictably in accordance with natural laws. However, avulsions represent a sudden change in a floodplains' equilibrium state, causing the system to follow a new trajectory of change. Mechanisms of avulsion have been well documented by numerous authors (e.g. Bridge and Leeder 1979, Smith *et al*. 1989, Mackey and Bridge 1995, Heller and Paola 1996, Slingerland and Smith 1998). Essentially, an avulsion occurs when ongoing aggradation in the channel causes gradients upstream to decrease and lateral gradients to increase, making the channel inefficient for the stream discharge and sediment load. At some position along the channel course, the gradient perpendicular to flow (i.e. across the alluvial ridge) becomes critically greater than the downstream channel gradient. Thus, ongoing floodplain processes cause a crisis in system evolution, forcing the system to adapt to changing conditions. Avulsions represent the crossing of a geomorphic threshold, caused ultimately by inherent geomorphic processes. However, it is usually external environmental conditions that push the system over the geomorphic threshold.

Normal flows on the Mfolozi River are low and therefore cannot overtop the levee to force an avulsion, even though the channel may have become decreasingly efficient. However, during a large flood, the channel is overwhelmed and floodwaters may overtop the levee. If the floodwaters find a more efficient path and the crevasse splay continues to erode (e.g. Slingerland and Smith 1998), the entire river may switch course and abandon the less efficient channel. Thus, although avulsions occur because of ongoing floodplain processes, they are associated with large flood events that enforce change. Renwick (1992) suggested that if thresholds (in this case an avulsion) were major discontinuities at relatively high magnitudes of process operation, then the resulting landform might be static for most of the time, with its morphology a reflection of the last threshold-exceeding event. Thus, evolution of the Mfolozi River Floodplain is somewhat static during normal flow conditions, and dynamical change occurs during relatively short time periods. As a result, the overall morphology of the Mfolozi River Floodplain is usually of the last large flood, an idea that is supported by the longevity of Domoina deposits on the floodplain.

The importance of avulsions may be contextualised using non-linear dynamical systems theory, which deals with sudden changes to system processes as bifurcations. Bifurcations represent a transition from one equilibrium state to another, or from regular or periodic behaviour to chaos (Phillips 1992). They may occur as a result of external influences or inherent thresholds, or a combination of both (Huggett 1988). Prior to an avulsion, the Mfolozi River is essentially at equilibrium and universal laws control the behaviour of the system. However, when a flood occurs, stream flow is usually not at equilibrium with regards to channel shape and slope. A system in non-equilibrium to an external influence (e.g. sudden catchment rainfall) must reorganize such that the effect of the external influence is dissipated (Huggett 1988). In rivers, avulsions are thus a response to non-equilibrium that through self-organization, may result in the occurrence of a new equilibrium.

At the critical threshold, chance-like fluctuations may control where and when an avulsion occurs, influencing the direction that evolution of the floodplain follows. As such, in nonlinear dynamical systems theory, a bifurcation is considered to be both deterministic and probabilistic. While an avulsion may be likely in terms of probability theory and it may be more likely to occur in a particular area due to local floodplain slopes, the exact location of an avulsion may be difficult to predict such that at a local scale it appears stochastic. Nevertheless, geomorphic systems are always somewhat predictable in that they always tend towards an equilibrium state of low relief where entropy is minimized.

5. Coastal floodplains and estuaries in KwaZulu-Natal: shared geomorphology

The sedimentary dynamics of KwaZulu-Natal's estuaries are well documented by Cooper (1993, 1994, 2001, 2002). On the micro-tidal coast of KwaZulu-Natal (and indeed southern Africa), wave energies are too high to allow for the development of a delta according to the classification of Dalrymple *et al*. (1992), even when sediment loads are high. Instead, sediment is transported northwards under the influence of longshore drift. The exception is following large magnitude flood events that build an ebb tidal delta in the ocean. However, these are rarely persistent for longer than a few months, as shown by Cooper *et al*. (1989). The effect of a wave-dominated coast on estuarine facies development is well illustrated by the perched elevation of the Mfolozi River bed (1m amsl) at the estuary mouth.

However, Cooper's (1993) river dominated estuaries were conceived from the apparent importance of rivers in maintaining accommodation space. Cooper (2002) suggested that the cohesiveness of estuarine mud and vegetation required extremely large magnitude floods to scour the estuary, and thus recover accommodation space. Following a flood, river sediment loads were generally high enough to quickly fill the available accommodation space. Thereafter, the estuary acted as a conveyor belt, with incoming sediment transported to the ocean.

The Mfolozi River Estuary is an example of a river-dominated estuary according to Cooper's (1993) classification. Study of the Mfolozi River hydrology suggests that the reason high magnitude floods are important geomorphologically is not because of estuary resistance to erosion, but rather because stream powers large enough to erode the estuary are variable in occurrence. The frequency analysis of flows of the Mfolozi River indicated a skewed distribution of flows towards lower than median values. A small number of extremely high positive deviations (up to +300% from the median flow) indicate that extremely large floods sporadically occur on the Mfolozi River. These large, infrequent floods are primarily responsible for estuary scour and accommodation space recovery. Thus, during normal flow conditions, that are generally relatively low, sediment is transported through the estuary, maintaining equilibrium in terms of sediment mass balance (Figure 5). During low frequency flood events, sediment is eroded from the estuary, resulting in sediment outputs locally exceeding sediment inputs.

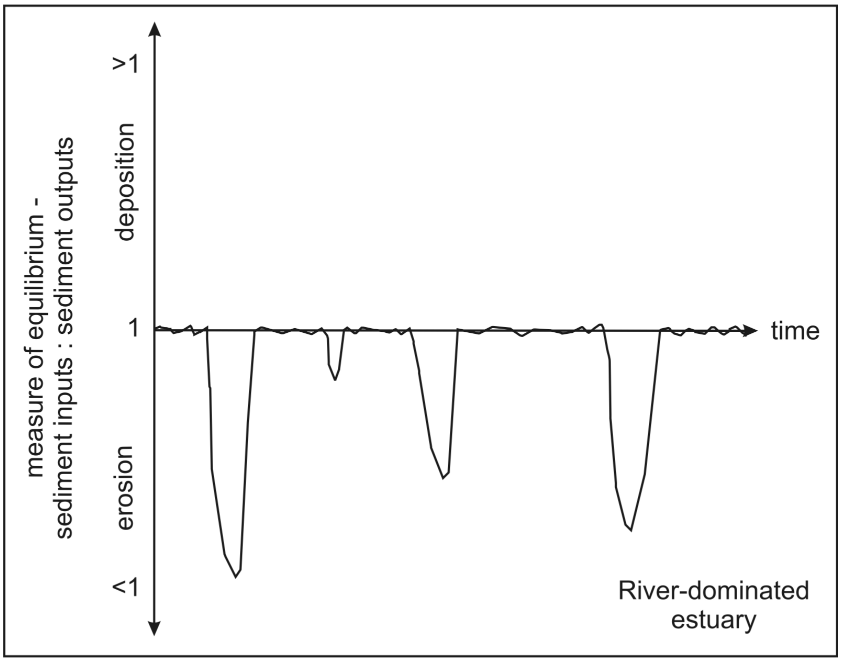

Figure 5: Conceptual illustration of punctuated equilibrium, in terms of sediment throughputs, in river-dominated estuaries of KwaZulu-Natal.

The effects of a large flood on coastal floodplains and estuaries in the region are polar (Figure 2 and 5), with floodplains experiencing deposition and loss of accommodation space, while estuaries are intensively scoured and accommodation space is momentarily recovered. The frequency at which these changes occur is related to stream flow variability. Estuarine erosion and floodplain deposition are limited during normal flow conditions, as flows are too low to scour or overtop levees, resulting in a period of stasis. However, during large flood events, the relationship between sediment inputs and outputs becomes unsettled and equilibrium is momentarily stalled.

The factors that cause flow variability on the Mfolozi River occur throughout southern Africa's eastern seaboard, where summer rainfall occurs. As such, the apparent polarity, yet shared response, of linked coastal floodplains and estuaries to large floods is likely to be widespread throughout KwaZulu-Natal.

6. The problems of scale and variability: implications for fluvial system management

6.1. Misconceptions of flow variability

While it is known that many rivers of southern Africa are variable, in that flow is flashy and dominated by periods of low flow interrupted by extremely large floods, this is not commonly acknowledged by management agencies. Instead variable flow rivers are often regarded as regular flow rivers, resulting in failure to anticipate and understand the geomorphic implications of large floods in such systems.

Management of the Mfolozi River and Lake St. Lucia estuarine system has been fraught with such misunderstanding. Calculations of sediment loads on the Mfolozi River system have been grossly over-estimated in the past as scientists assumed the relationship between sediment concentration and discharge remained constant throughout the year (e.g. Lindsay *et al*. 1996, Rooseboom 1975). However, as previously discussed (Chapter 3), sediment concentration variability is intrinsically linked to the seasonal and annual variation in run-off that causes flow variability. Thus, failure to recognise a variable flow regime will always lead to incorrect estimations of sediment transport. Considering the historical link between the Lake St. Lucia estuary and the Mfolozi River mouth, this misunderstanding has had consequences for the management of the two systems. Management agencies believed that the Mfolozi River was carrying extremely large loads of sediment into the lake, causing excessive sedimentation and loss of depth, while also raising turbidity beyond ecologically acceptable levels (Anon 1970). Scientists confirmed what they believed to be extremely high sediment loads in global terms. In fact, a plot of the sediment yields of Rooseboom (1975) and Lindsay *et al*. (1996) and catchment area of the Mfolozi River, against global data from Milliman and Meade (1983) suggested that sediment concentrations of the Mfolozi River well exceeded global norms. In reality, this was not the case and the true sediment yield of the Mfolozi River, calculated over a 6-year period, is generally consistent with its catchment size. Unfortunately conservation authorities, believing sediment yields to be absurdly high, adopted a strategy of maintaining separation of the historically linked St. Lucia and Mfolozi River estuaries that has remained in place for over 60 years.

The second major misconception regarding variable flow rivers is under-estimating the geomorphic impact of a large flood event. Brierley and Fryirs (2005) describe sand-bed alluvial rivers as sensitive, purely because they have the capacity to respond to changing discharge fairly rapidly. Their characteristically large valleys and sand bed channels result in a capability to adjust channel morphology, planform and slope, all within a single flood event. The Mfolozi River Floodplain may in this context be defined as 'sensitive'. In addition to being geomorphically sensitive, the flow of the Mfolozi River is variable, a situation that is compounded by its latitudinal position on the east coast of Africa that makes it vulnerable to occasional, yet infrequent, tropical cyclone incursions or floods related to cut-off low pressure systems.

The flood associated with tropical cyclone Domoina in 1984 resulted in rapid transformation of the Mfolozi Floodplain, causing the Mfolozi River to avulse towards the south. Loss of flow confinement resulted in deposition of bedload sediment up to 4m deep in a large, south-east trending lobe. The immensity of the flood and its effect on infrastructure is documented by van Heerden (1984) and Travers (2006). In addition, the resultant loss to commercial sugar cane farmers was in the region of R57 million (Begg 1987).

The MFA, an association representing sugar cane farmers of the Mfolozi River Floodplain at the time, asserted that damage to sugar cane production was primarily as a result of accelerated soil erosion in the Mfolozi catchment that was associated with overstocking and poor cultivation practices in the KwaZulu homeland regions (Watson 1993). Furthermore, the perceived increase in catchment erosion was considered to be responsible for an increased incidence and severity of floods (Anon 1984).

However, there is no evidence to suggest that floods have increased in magnitude, or occurrence in the Mfolozi catchment. In contrast, the hydrology of the Mfolozi River and the geomorphology of the Mfolozi River Floodplain suggest that the system has been and continues to be characterised by infrequent, but large flood events. Hydrologically, flow is variable and punctuated by large flood events. In addition to hydrological variability, the shape of the floodplain is such that flood flows would naturally have caused widespread sedimentation and flooding. Upstream of the Mfolozi River Floodplain, flood flows are confined in a narrow valley. As floodwaters enter the floodplain, the valley widens dramatically and bedrock valley walls no longer confine the channel. Consequently, flow is spread laterally during large floods over the upper floodplain region, causing widespread sedimentation as stream capacity is lost. During Cyclone Domoina, sediment deposits up to 4m deep were encountered in some regions of the floodplain. Using available flow data, floods of this magnitude are likely to re-occur approximately every 300 years. Thus, in the context of a human life, such geomorphic change may be considered unlikely. However, in the case of variable flow rivers, large, extreme flood events and the sudden geomorphic change they enforce are to be expected.

6.2. Recognizing temporal and spatial system scales

System understanding is limited by the boundaries we apply to a system of study, which in a sense, may be considered to be somewhat arbitrary. The spatial and temporal boundaries that are imposed on a system determines what can be known about that system by framing the scale of data collection in both time and space. Furthermore, the scale of research determines to what extent understanding gained from each study may be applied to other regions in time and space. The failure to recognize the limitations of scale in research can lead to erroneous generalizations about geomorphic processes.

Geomorphic analysis occurs across a wide diversity of spatial and temporal scales. Spatially, geomorphic problems may be considered from individual gullies to individual floodplains, to local stream networks, and then to entire catchments and continents. Temporally, for instance, geomorphologists consider what happens to erosion cycles over time, how deposition affects slope and how climate change has and will affect geomorphic processes.

An interesting example of consideration of the implications of scale in geomorphic problems concerns understanding whether wet or dry spells are related to erosion. Meadows and Hoffman (2003) investigated the impact climate change would likely have on land degradation in southern Africa. Under a drier climatic regime, land degradation caused by erosion was considered to be likely to increase. Erosion in this context was thought to be associated with dry periods because vegetation cover in bottomlands was low and rainfall irregular, leading to erosion. However, increased sediment supply caused by erosion in the upper catchment, is likely to result in downstream deposition as the stream sediment transport capacity is exceeded. Thus, erosion is always likely to be accompanied by deposition in another downstream setting.

If upstream settings are likely to degrade during dry periods, it is likely that downstream reaches are likely to *aggrade* during dry periods. However, as a catchment enters a wet period, vegetation in the catchment begins to recover, and erosion in the upper catchment slows and ceases. In valley-bottoms and other depositional settings, sedimentation during the dry period, from sediment supplied by upper catchment sources, results in steepening at the system toe. As discharges increase with the onset of a wet spell, the steepened toe of the depositional setting in combination with increased stream sediment transport capacity is likely to result in erosion, even though erosion has decreased in the upper catchment.

The result is that at a system scale, processes of erosion and deposition occur at different temporal and spatial scales. Thus, the application of process rules to entire catchments is generally not feasible. Furthermore, Fryirs *et al.* (2007) suggest that sediment cascades may not always be connected to the drainage line due to geomorphic buffers or barriers. In such cases, there may not be a simple relationship between erosion in one part of a catchment and consequent deposition downstream as there may be a substantial time lag in process response. Thus, broad-scale connectivity is vitally important in considering sediment cascades and the impact of changing environmental variables.

A useful indication of drainage line system complexity is provided by measurements of suspended sediment concentrations at the lower extremity of the Mfolozi River. The hysteresis effect recorded in flow and suspended sediment concentration, whereby sediment concentrations increase to a peak prior to the peak of stream discharge, indicates the importance of wet and dry seasons, and possibly years, in a variable flow river. During the dry season (or dry spell), sediment becomes available for transport as vegetation cover decreases in the catchment. However, despite sediment availability, the lack of connectivity between hillslopes and channels during the dry season (or dry spell), as well as lack of stream capacity, results in suspended sediment concentrations remaining relatively low. However, at the onset of the wet season (or wet spell), connectivity and stream sediment transport capacity is restored. This results in a sudden flushing of available sediment sources from the catchment, resulting in high sediment concentrations relative to stream discharge. If one were to consider erosion just in terms of sediment concentration measured on the lower floodplain of the Mfolozi River, it would be appear as if all erosion occurred during the wet season (or wet spell). However, hysteresis reflects the reality that erosion occurs in different temporal and spatial scales throughout the catchment, and that a measurement of sediment concentration at the toe of a fluvial system is a cumulative measurement of multiple processes. Erosion occurs in upper catchment regions during dry spells to some extent, when sediment becomes available for transport. However, the actual movement of sediment away from its source occurs during wet seasons (or wet spells). Furthermore, data from the Mfolozi catchment suggests that erosion will be greatest when an extremely wet period follows a drought, a conclusion shared by other authors (e.g. Watson 1996, Bull 1997).

It can be seen that separating processes from the system scale to which they apply may lead to simplification of system understanding that often generates poor generalizations. Acknowledging and identifying the spatial and temporal scale of study is imperative to avoid such generalizations.

6.3. Climate change: implications for depositional fluvial settings

Wetlands, and indeed all fluvial systems, evolve towards an equilibrium determined by current regimes of rainfall and temperature. As a result, geomorphic systems that have evolved under current climatic conditions are likely to undergo significant changes in form and process if global warming, as a result of anthropogenic emissions of carbon dioxide, occurs. While global warming has arguably already occurred, what climatic changes can be expected in the future has been under debate. However, there has been increasing clarity and confidence in predicted climate changes in southern Africa.

Climate models of the late 1990's predicted an overall drying of the southern African climate. Hulme (1996) suggested the entire region would become drier, while rainfall would become more variable. As a result of decreases in precipitation and an increase its variability, Arnell (1999) predicted substantial reductions in run-off. Flow in the Zambezi River, for example, would decrease by 40%, while the Limpopo and Orange Rivers would be affected to a lesser degree (decreases of 30% and 4% respectively (Arnell 1999).

Meadows and Hoffman (2003), reviewing climate models at the time, suggested that the western regions of southern Africa would become drier while the eastern regions of KwaZulu-Natal, Mpumalanga and portions of the Eastern Cape, could expect to become wetter. Hewitson and Crane (2005) concurred, finding that regions that experience convective rainfall would experience an increase in rainfall and rainfall intensity, while the western regions, influenced primarily by frontal systems, would experience a reduction in winter rainfall. In contrast, New (2002) predicted that land degradation in southern Africa would occur due to the region experiencing longer dry spells with more sporadic and intense rainfall events. Although rainfall would not increase in the sub-continent, rainfall intensity and variability would increase.

The most recent IPCC report has indicated that with current information it is no longer possible to predict whether the eastern regions of the subcontinent will get wetter, although the western regions will certainly get drier (IPCC 2007). In the eastern regions of southern Africa, it now appears likely that rainfall may decrease (as early predictions suggested) or stay the same. However, irrespective of the actual amount of precipitation in the region, it is clear that rainfall will become more variable and the magnitude and incidence of large flood events will increase (IPCC 2007). Furthermore, the duration and spatial occurrence of dry spells will increase.

All of these changes are likely to have geomorphic impacts at different scales. Increasing rainfall intensity in the eastern regions is likely to have the greatest implications, particularly for valley fill systems. Increasing rainfall intensity will increase rates of run-off, resulting in valley fill systems experiencing larger than usual discharges. The new discharge regime will likely have greater sediment transport capacity, and in some systems, the slope may be too steep to remain stable. In such cases, increased rainfall intensity will likely push the system past a geomorphic threshold prematurely, causing incision.

If dry spells persist for longer periods, the geomorphic impact will be similar to that of increasing discharge, although the cause of incision differs slightly. Longer dry spells will lead to a more than normal decrease in vegetation cover, which will in turn lead to increases in run-off and sediment supply when precipitation does fall. Thus, longer dry periods will also result in incision of valley-bottom settings. In the western regions of southern Africa, it is expected that annual precipitation will decrease. Similarly to increasing the length of dry spells, less precipitation will lead to a decrease in vegetation cover, thus increasing the risk of erosion.

7. Conclusion

The dynamics of the Mfolozi River Floodplain are intrinsically linked to flow variability of the Mfolozi River and its tributaries. Episodic accumulation of clastic sediment occurs as a result of erratic, extremely large, flood events with intervening periods of normal flow. The Mfolozi River Floodplain experiences plurality, in that gradual processes of sediment accumulation occur in some regions of the floodplain, while processes of punctuated equilibrium dominate the majority of the floodplain. Climate change is likely to increase rainfall variability, resulting in the Mfolozi River and floodplain becoming even flashier in terms of hydrology and system processes.

While it has been shown that flow variability is extremely influential in terms of the manner in which a floodplain system functions, and that it is certainly linked to sediment transport variability, the exact mechanisms of how sediment transport variability arises is uncertain. The work of Fryirs *et al*. (2007) focuses on how geomorphic processes influence the movement of sediment through a drainage network, and the application of this approach with respect to sediment transport variability in time and space would certainly be fruitful. In addition to understand landscape connectivity, it would be useful to locate sediment sources in time and space (i.e. is there a seasonality in the supply of sediment from different sources?), especially in view of sediment transport hysteresis.

On a broader scale, a more complete understanding of how the morphology of different wetlands arises would be useful for future management and rehabilitation planning. The threshold approach has not yet been applied to hillslope seepage wetlands although they are conceptually consistent with current ideas. In addition to understanding processes of sediment delivery in relation to discharge in hillslope areas, differences in sediment and water supply between channelled and unchanneled valley bottom wetlands also requires clarification. Currently, it appears that unchanneled valley bottom wetlands arise in conditions where sediment supply is abundant and connectivity between the wetland and its sediment sources is high. In contrast, channelled valley bottom wetlands are likely to occur in valleys where sediment supply relative to water supply is low. However, this requires further study.

In addition to understanding the movement of sediment through the drainage network, an understanding of how these processes have, and will, change through time due to changes in climate and catchment land use is essential. Understanding future processes requires fully understanding the past.

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