

**Sediment Connectivity in the Upper Thina Catchment,
Eastern Cape, South Africa**

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Abstract

Sediment dynamics are influenced by transformed landscape connectivity in catchments worldwide. The upper Thina catchment, an important high rainfall resource in the northern Eastern Cape, South Africa, is an example of where ongoing subsistence farming on communal land has led to overgrazing and trampling that has initiated large erosive features (e.g. gullies) and river incision. The formation of gullies led to increased hillslope-channel connectivity and the resultant river incision decreased the channel-valley fill connectivity. These two changes in connectivity led to increased sediment export from the catchment that has various down-stream ecological and socio-economic impacts. This study investigates how the change in hillslope-channel and channel-valley fill connectivity has altered the sediment dynamics in the Vuvu catchment, a headwater tributary of the Thina River.

A combination of methods were used to assess the changes in hillslope-channel and channel-valley fill connectivity. High resolution aerial images were used to map source features, such as fields, gullies, sheet erosion, landslides, roads and livestock tracks. Topographic and geological characteristics of the source features were extracted using a Geographic Information System. Furthermore, hillslope-channel pathways, such as the natural drainage network, continuous gullies, discontinuous gullies, roads and livestock tracks were mapped and analysed in terms of topographic and geological characteristics. Historic aerial images were assessed to calculate the date the larger gullies began forming. Recent aerial photos and cross sectional surveys of the valley fill were combined to map the various sediment sinks.

Particle size and organic content were analysed for flood bench cores and terrace samples. The chronology of the flood benches was determined using unsupported Pb-210 and Cs-137 dating, and determined for the terraces using Optically Stimulated Luminescence dating. Quantitative and qualitative sediment tracing approaches, using mineral magnetic properties, were used to trace the origin of suspended sediment (collected during flood events), sediment stored in the flood benches and sediment stored in the terraces. Hydrological monitoring was used to assess the potential to store sediment on flood benches along the valley fill through flood bench inundation frequency. Hydrological and hydraulic modelling extended the measured inundation frequencies to a 73 year period and other cross sections along the valley fill. Furthermore, a future scenario of an increased vegetation cover and reduced hillslope-channel connectivity was assessed in terms of channel-valley fill inundation frequency.

In the Vuvu catchment, 9% of the land is severely eroded and half of the eroded areas are directly connected to the drainage network. The average soil loss rate for the catchment, based on the mapping of erosional features, was $13 \text{ t ha}^{-1} \text{ y}^{-1}$. Fields, gullies and areas of sheet erosion were calculated to be the main sediment sources that could be mapped, producing $\sim 4 \text{ t ha}^{-1} \text{ y}^{-1}$ of sediment per source type for the catchment. The areas characterised by mudstones had the highest density of erosion features, with soil loss estimates as high as $9 \text{ t ha}^{-1} \text{ y}^{-1}$ for fields, $5 \text{ t ha}^{-1} \text{ y}^{-1}$ for gullies and $5 \text{ t ha}^{-1} \text{ y}^{-1}$ for sheet erosion,

confirming the sensitivity of this geology to land use pressures. Furthermore, erosional features were more prevalent on moderate ($5\text{--}20^\circ$) and north-facing slopes. Historic mapping showed that the gullies are still active sediment sources and that the larger gullies were initiated in the 1920s; smaller gullies however were initiated in the 1950s. Gullies increased the natural drainage network by 21% in a down-slope direction, whereas livestock tracks and roads have increased the natural drainage network by 178% in an across slope direction. Continuous gullies were the most effective hillslope-channel linkages as they were directly linked to the channel, while discontinuous gullies, given their location, were buffered from the channel. Across slope drainage features intercepted and concentrated hillslope runoff, routing water directly to the drainage network, reducing chances of water infiltration as well as sediment deposition.

Sediment storage along the valley fill was limited due to the valley confinement, narrow flood benches and terraces that are active sediment sinks. Lower flood benches stored sediment for up to 60 years, higher flood benches for 92+ years and terraces for up to 4500 years. It was estimated that the flood benches only stored 2% of the sediment that was eroded over the past 100 years, confirming the limited sediment storage potential along the valley fill. The average sediment accumulation rate was higher for lower flood benches ($3.2\text{ g cm}^{-2}\text{ yr}^{-1}$) when compared to the average accumulation rate of the higher flood benches ($0.9\text{ g cm}^{-2}\text{ yr}^{-1}$), showing that the lower benches were more active in receiving and storing sediment coupled with a shorter storage period. The flood benches acted as down-river sediment buffers, whereas both the flood benches and terraces acted as a buffer for hillslope-channel linkages in the immediate vicinity of the valley fill, unless directly linked to the channel by an incised tributary channel.

Sediment tracing using mineral magnetic characteristics could discriminate between igneous and sedimentary sources. Quantitative source apportionment confirmed the results of a qualitative approach, leading to a novel semi-qualitative sediment tracing approach. Suspended sediment concentrations were high (up to $\sim 11\text{ g l}^{-1}$) and the source depended on rainfall extent and catchment shape. Local sedimentary sources dominated most of the flood water sediment, unless high rainfall in the upper catchment contributed large quantities of igneous sediment. This was reflected in the lower flood benches that predominantly stored locally sourced sediment, whereas the higher flood benches were characterised by a greater contribution from the igneous sources of the upper catchment. This suggests that the sediment stored in higher and lower sinks reflect the stage (low stage from local events and high stage when entire catchment contributes) and the source of the water and sediment. Sediment concentration decreased throughout the wet season, indicating that sediment availability decreases as vegetation cover increases or that sediment sources are depleted. The majority of the sediment stored in the valley fill was derived from local sedimentary sources, reflecting the erodible nature and good hillslope-channel connectivity of the lower catchment and that the contribution from local sources has been relatively stable over the past 4500 years. This also confirms that degradation is a catchment-wide phenomenon

that has maintained the contribution ratio from the different geologies (sedimentary sources being dominant).

Flow discharge and inundation levels were monitored using in-situ level loggers and rating curves at two sites. Estimates based on hydraulic modelling were made at an additional 13 sites. Inundation frequencies for a 73 year period were estimated using a hydrological model for the current increased hillslope-channel connectivity and a future rehabilitated condition of reduced hillslope-channel connectivity.

Intensive summer rainfall in the Vuvu catchment produced steep hydrographs of short duration (3 hours). Lower flood benches were inundated 1.5-4.5 times a year and the higher flood benches 1 in 2 years during the monitoring period (2012-2014). Where the channel was incised, inundation frequencies were lower due to the higher elevation of the flood bench in relation to the channel bed, reducing the potential to store sediment on these benches. The inundation frequency of flood benches decreased during drier periods, reducing the potential to deposit sediment on the flood benches. During wet periods the chances of sediment deposition on flood benches was increased due to the increased inundation frequency, but was dependent on sediment availability that increased during dry periods as soil is exposed to erosive rainfall due to reduced vegetation cover. Sediment storage on the valley fill could thus be greater during dry periods compared to wet periods. The potential of sediment storage on flood benches decreased for the post-rehabilitation scenario as both sediment concentration and peak flood volumes would be decreased, reducing hydrological and sediment connectivity.

Sediment connectivity has been increased in the Vuvu catchment due to anthropogenic influences such as cultivation and subsequent abandonment, overgrazing and frequent burning of rangelands. Large areas of erosion are effectively connected to the drainage network, transferring sediment directly to the channel. Once in the channel, very little suspended sediment is deposited on the valley fill, due to the small size of the depositional features and the short, high energy character of the floods. Sediment transfer rates in the Vuvu catchment are thus high from source areas to the bottom of the catchment due to increased hillslope-channel connectivity and limited sediment buffers along the valley fill. This research shows how important sediment connectivity is in understanding where sediment is sourced, how and when it is transported and where sediment can be stored. Further research should address the effective use of buffers to reduce hillslope-channel connectivity and increase storage of sediment on the slopes and valley fill.

Keywords: connectivity; hillslope-channel connectivity; channel-valley fill connectivity; erosion; gully, livestock track; Transkei; Thina River; Vuvu River; overgrazing; burning; incision; sediment deposition; flood benches; sediment tracing; Pb-210; Cs-137.

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Acronyms

ARM	-	Anhyseretic Remanent Magnetism
DEM	-	Digital Elevation Model
DFA	-	Discriminant Function Analysis
ECRP	-	Eastern Cape Restoration Programme
GIS	-	Geographical Information System
HIRM	-	Hard Isothermal Remanent Magnetism
IC	-	Index of Connectivity
IRM	-	Isothermal Remanent Magnetism
Masl	-	Meters above sea level
NDVI	-	Normalized Difference Vegetation Index
NSW	-	New South Wales
OSL	-	Optically Stimulated Luminescence
PCA	-	Principal Component Analysis
PES	-	Payment for Ecosystem Services
Q	-	Discharge
RGB	-	Red Green Blue
RMMF	-	Revised Morgan, Morgan and Finney
SAR	-	Sediment Accumulation Rate
SIRM	-	Soft Isothermal Remanent Magnetism
SWAT	-	Soil and Water Assessment Tool
UNEP	-	United Nations Environment Programme
USGS	-	United States Geological Survey
USLE	-	Universal Soil Loss Equation
WASA-SED	-	Model of Water Availability in Semi-Arid Environment with a Sediment Dynamics Component
WRC	-	Water Research Commission
WSP	-	Watershed Services Project

- XARM - Susceptibility of Anhysteretic Remanent Magnetism
- Xfd - Frequency Dependent Magnetic Susceptibility
- Xlf - Low Frequency Magnetic Susceptibility

Chapter 1: General introduction

1.1. Introduction

Soil erosion, sediment transport and deposition are natural earth surface processes occurring in any catchment (Yang et al., 2003). Accelerated soil erosion and subsequent deposition are a major threat to biophysical and socio-economic environments worldwide (Brown, 1981; Oldeman, 1994; Pimentel et al., 1995; Valero-Garcés et al., 1999; Soulsby et al., 2001; Yang et al., 2003; Greig et al., 2005; Pimentel, 2006). Soil erosion is often associated with increased surface water runoff (Battany and Grismer, 2000) and the development of erosional features, such as gullies, that enhance the transfer of water and sediment (Harvey, 1992, 1997). This increase in landscape connectivity or pathways of water and sediment transfer plays an important role in sediment dynamics as it affects the rate of sediment delivery down the system.

Landscape connectivity or coupling as an ecological concept and framework with longitudinal, lateral, vertical and temporal dimensions was introduced by Ward (1989). Building on earlier work on coupling and sediment budgets (e.g. Trimble, 1975; Caine and Swanson 1989), this framework has since been embraced by geomorphologists (e.g. Brierley and Murn, 1997; Brierley and Fryirs, 1998; Cammeraat, 2002; Hooke, 2003) as it integrates structure and function over various temporal and spatial scales (Brierley et al., 2006). Connectivity can be defined as the movement of matter through a system of interconnected units (Hooke, 2003). Where connectivity is hindered by barriers and buffers (e.g. alluvial fans, floodplains, resistant base level, etc.), matter is stored until connectivity is re-established (Fryirs et al., 2007a). Sediment movement is often symbolized as a 'jerky conveyor belt' (Ferguson, 1981), moving sediment down a slope or through a catchment in a series of overland flow events and not as one continuous event. Only a small proportion of sediment that is eroded in an event will make it to the bottom of the slope or catchment (Walling, 1983). This is due to sediment deposition down the slope and is a function of the degree of connectivity within and between the various landscapes.

Increased connectivity (e.g. development of gully networks, straightened river channels, etc.) leads to more efficient transport of materials, such as water, sediment and nutrients, slowly decreasing the natural capital of the landscape (Brierley and Fryirs, 1998). Downstream areas are affected by the increased delivery of these materials, which can have severe biophysical and socioeconomic effects such as flooding, eutrophication and sedimentation (Brierley and Fryirs, 1998; Sandercock and Hooke, 2011). Decreased connectivity, through the construction of dams, levees and elevated roads trap sediment, nutrients and water, which can have socio-economic and biophysical benefits (e.g. water provision, economic growth) as well as drawbacks (e.g. decreased fish migration) (Nakamura and Tockner, 2004; Sandercock and Hooke, 2011). Understanding the structure and functioning of landscape connectivity in a catchment is crucial to the effective management thereof (Lexartza-Artza and Wainwright, 2009).

In South Africa soils are highly erodible, especially in the central, eastern and north-eastern parts of the country, particularly areas with steep slopes (Hoffman and Todd, 2000). Land degradation is common in areas with communal land tenure which is characterised by a large number of people living in a rural environment with a frequent incidence of poverty and people who depend on the land for livelihoods (Hoffman and Todd, 2000). In the communally managed former homelands of South Africa, such as the former Transkei and Ciskei in the north-eastern areas of the Eastern Cape province (which will be referred to as Transkei and Ciskei throughout the document for practical reasons), the development of agricultural fields and subsequent abandonment, continued grazing pressure and frequent burning has resulted in severe degradation of vegetation (Blignaut et al., 2010; Schulze and Horan, 2007) and has resulted in the formation of extensive erosional features (Beckedahl and Dardis, 1988; Dardis et al., 1988; Kakembo and Rowntree, 2003). The rivers draining the Transkei, e.g. Umzimvubu, are characterised by high sediment loads due to the land degradation and associated soil erosion (Madikizela and Dye, 2003). Historical descriptions of a deep estuary at Port St. Johns that has since been silted up (Pollock, 1969) and 'gin clear' trout streams in the upper catchment, even after rainfall events (Hey, 1957), paint a different picture of past (100+ years ago) sediment dynamics and suggests that sediment erosion and transfer has been increased.

There is no exception to this degradation and extensive erosional features in the headwaters of the Thina River, a tributary of the larger Umzimvubu River system (Blignaut et al., 2010). The upper Thina catchment is characterised by steep slopes, erodible mudstones in the lower catchment, basalt in the upper catchment and high intensity rainfall (Plate 1.1). This catchment was subject to subsistence farming on communal land for 100+ years (Bundy, 1987) where overgrazing and early winter burning resulted in increased soil erosion (Blignaut et al., 2010; Watershed Services Project, 2010) and led to the formation of extensive gully networks (Rowntree et al., 2012). Evidence of the high suspended sediment transfer rates is evident in the Mount Fletcher bulk water supply reservoir, located on the upper Thina River, that lost 70% of its capacity in 4 years due to siltation (Rowntree et al., 2012).

The high-rainfall upper catchment of the Thina River is a strategic water source for downstream users and thus a key starting point for rehabilitation efforts. Working for Water, with support from United Nations Environment Programme and Joe Gqabi District Municipality, started a pilot rehabilitation project in the headwaters of the Thina River in 2011 to evaluate the implementation of Payment for Ecosystem Services (PES) through the Watershed Services Project (Watershed Services Project, 2010). The efforts are focused on the area around Lundi village which falls in the Vuvu River catchment. The aim for this Watershed Services Project is to train and employ community members in gully and sheet erosion rehabilitation work, rehabilitate eroded areas (Plate 1.1), eradicate invasive species, develop land management plans and monitor winter flow (baseflow) and water quality (sediment concentration). The upper Thina River restoration project served as a local opportunity to study sediment connectivity and called for research to answer questions such as: where does eroded sediment originate from; how is sediment transported and



Plate 1.1: From left to right, first row: steep slopes of the catchment with houses on gentle slopes; steep slopes and narrow valley fill; second row: livestock track showing signs of erosion (photo Dylan Weyer); soil conservation blankets used to rehabilitate gullies; third row: large gully network; turbid storm water runoff after heavy rain; bottom: Vuvu River in high flow after a heavy downpour.

how are the existing sediment sinks functioning? The catchment of the Vuvu River was therefore selected as the study site for this thesis.

Connectivity concepts have been applied to a few areas in South Africa, such as the Karoo (Foster et al., 2012; Rowntree and Foster, 2012; Grenfell et al., 2014) and KwaZulu-Natal (Le Roux et al., 2013; Miller et al., 2013; Grenfell et al., 2014) as well as to a geomorphological river health assessment tool (Geomorphological Driver Assessment Index) (Rowntree, 2013a), however it has not been applied to degraded high elevation, high rainfall areas. As the connectivity framework is well suited to study geomorphological processes and changes over time (Brierley et al., 2006; Kondolf, 2006), it is used in this thesis as the framework to investigate fluvial redistribution processes affecting fine grained sediment in the upper Thina River, South Africa.

1.2. The research problem

From observations in the catchment it is apparent that lateral connectivity has changed in two ways; a) extensive gullies and livestock tracks on the hillslopes increase the hillslope–channel connectivity and b) incision of the river channel has decreased channel–valley fill connectivity. How these changes in connectivity affect sediment transfer is unknown, but it is postulated that storm water runoff and suspended sediment transport has been increased.

Currently the majority of the grassland in the study area is in a degraded state (Schulze and Horan, 2007) as a result of overgrazing, trampling and frequent early winter burning. This is clearly visible near cattle dipping tanks where all the cattle are congregated once a fortnight to be dipped. The decrease in vegetation cover exposes the soil to erosive rain, decreases slope roughness and reduces sediment binding through roots (Rowntree et al., 2004; Sandercock and Hooke, 2011). The other consequence of reduced vegetation cover is that livestock are forced to travel large distances on a daily basis in order to find the necessary fodder. This leads to high traffic volumes and trampling of the land, causing linear tracks that act as water drainage features. These linear features enhance water runoff, concentrate flow and ultimately cut down into soils and aid in the formation of gully systems that are sediment sources and highly efficient conduits of water. These livestock tracks and gullies increase the hillslope-channel connectivity, increasing the volume of runoff.

From Equation 1.1 we can see that an increased volume of water can be translated into increased energy (E_{pp}).

$$E_{pp} = \gamma Vy \tag{1.1}$$

In Equation 1.1 E_{pp} is the pressure potential energy (joule), γ is weight density (kg m^{-3}), V is the volume (m^3) and y is vertical height (m) of the water column. Sediment movement is directly dependent on the

energy available in the system, thus increased energy will result in greater entrainment of sediment and decreased opportunity for deposition (Robert, 2003), ultimately leading to greater hillslope-channel sediment connectivity.

The increased hillslope-channel connectivity produces a greater volume of runoff entering the channel. This increased volume of water entering the channel is also translated into increased energy in the channel (Equation 1.1), thus crossing the shear stress or incipient movement threshold of bed material more frequently, leading to down cutting into deposited sediments (Robert, 2003). This incision enlarges the channel to such an extent that overbank flooding happens less frequently, reducing the channel-valley fill connectivity (Kondolf, 2006; Nakamura and Tockner, 2004). This changes the role of the valley fill from a sediment store to a sediment source.

Limited research has been done on sediment dynamics along the Transkei Drakensberg Escarpment. The river systems are said to be in a phase of incision (Dardis et al., 1988) and erosional features such as soil pipes, gullies, sheet and rill erosion are common throughout the Transkei (Beckedahl et al., 1988; Beckedahl and Dardis, 1988). These erosional features are formed as a result of poor soil structure and high sodium content combined with external pressures such as poor farming practices and artificial drainage networks (e.g. roads and livestock tracks). Modelling exercises by Schulze and Horan (2007) indicated that stormflow is increased and sediment erosion could be doubled as a result of land degradation in the upper Thina catchment. Rowntree et al. (2012) did some preliminary sediment tracing in the upper Thina catchment. Sampling was limited to a number of opportunistic source samples (poor access), a few river samples and a core from the Mount Fletcher reservoir that has a low sediment trap efficiency. Due to the extensive gully networks on the sedimentary geology (mudstones and sandstones), sediment was expected to be dominated by these sedimentary sources, but the results did not reflect the expected trend. A qualitative interpretation of the results showed an alternation between igneous and sedimentary source dominance throughout sequences identified as single flood events in the core. Findings were inconclusive and begged for a more in depth study of sediment dynamics.

Further afield in the less mountainous Mgwalana catchment in the Ciskei, Eastern Cape, Manjoro (2012) traced sediment stored in a floodplain deposit in a small degraded catchment. Results showed that surface soils from grasslands were the main source of sediment before 1965; whereas after 1965 sub-soils became dominant due to increases in gully erosion on abandoned fields. These studies show the extent of sediment transport related studies in the former homelands of the Eastern Cape (Transkei and Ciskei). All studies point to the increase in soil erosion and sedimentation due to land degradation, but lack detail of sediment transfer processes and connectivity structure of the steep headwater catchments along the Escarpment over a longer time period (up to thousands of years).

This landscape degradation is likely to increase hillslope-channel connectivity through gully erosion and the formation of livestock tracks and decrease chances of sediment deposition along the valley fill due to reduced channel-valley fill connectivity as a result of channel incision. The increased transfer to the

channel and increased transport along the channel leads to accelerated sediment export that poses a threat to downstream systems and the loss of natural capital in these upper catchments. In order to address the problem of high sediment yields in the upper Thina catchment it is necessary to understand the current and past structure of and processes involved in sediment delivery. This study raises the questions: has sediment connectivity changed, where and how has sediment connectivity changed and why has it changed? These questions will be addressed through this research and will contribute to the understanding of sediment connectivity in steep upland catchments in the past, present and future.

1.3. The aim of the study

The overall aim of this research is to investigate how landscape connectivity and fine grained sediment dynamics have changed in high altitude, high rainfall headwater catchments. The hypothesis for this study is that increased hillslope-channel connectivity will increase geomorphic energy, thus incise channels and reduce channel-valley fill connectivity.

1.4. Objectives

The following specific objectives were set for the research:

1. Classify, map and characterise potential sediment sources on hillslopes.
2. Map and characterise present-day hillslope-channel coupling.
3. Classify, map and characterise the valley fill sink-zones.
4. Provide a chronology of sedimentation and valley fill incision.
5. Identify the sources of suspended sediment and sediment stored in various valley fill features.
6. Assess the frequency of overbank connectivity along the Vuvu valley fill.

A framework indicating how the objectives above fit into the study of hillslope-channel and channel-valley fill connectivity is presented in Figure 1.1. For the assessment of hillslope-channel connectivity the source areas and pathways were mapped and analysed in terms of spatial and geological properties. For the channel-valley fill connectivity assessment, the various sediment stores were mapped and characterised in term of their sediment composition. The chronology and source of the various features were determined. Furthermore, sediment concentrations and overbank connectivity were monitored and a combination of hydrological and hydraulic modelling was used to assess over bank connectivity for the past 73 years and a potential future condition where vegetation cover is increased and hillslope-channel connectivity is less efficient.

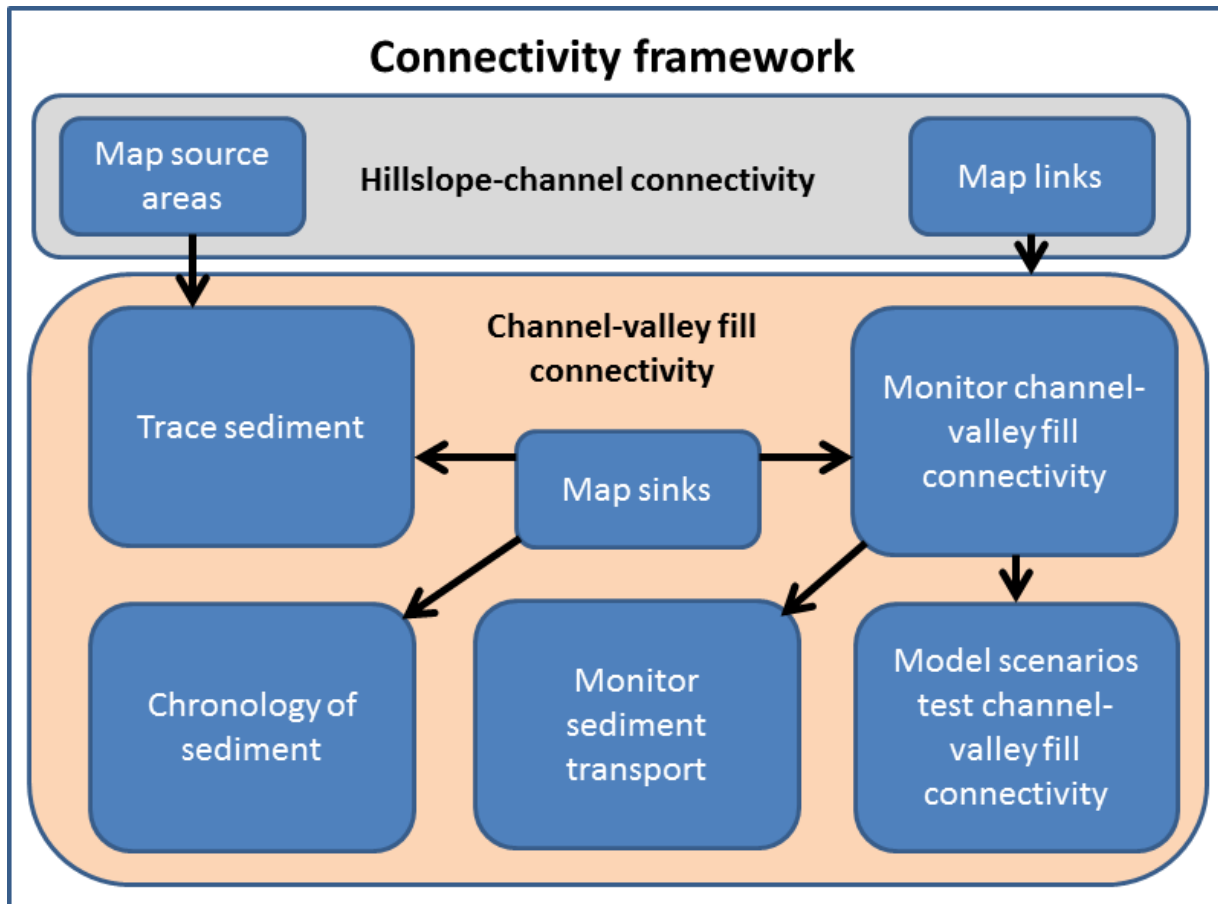


Figure 1.1: A connectivity framework showing how the objectives apply to hillslope-channel and channel-valley fill connectivity in the upper Thina River.

1.5. Outline and structure of the thesis

This thesis is composed of eight chapters. Chapter 1 provides a general introduction to the study, the research problem and the aim and objectives. In Chapter 2, literature on the connectivity concept and related drivers, such as slope gradient, vegetation cover, gully development, valley confinement and incision is reviewed. The study area is discussed in terms of topography, hydrology, geology, vegetation, soils and land use in Chapter 3. Chapter 4 deals with the mapping of sediment sources and pathways and Chapter 5 focuses on the mapping, character and chronology of the valley fill. Chapter 6 looks at the source of sediment transported in suspension and stored in the valley fill. In Chapter 7 the overbank connectivity is assessed and Chapter 8 synthesises the literature review and the findings of Chapters 4–7. Chapter 8 highlights the contribution of this study to sediment connectivity concepts and sediment dynamics along the Transkei Drakensberg Escarpment and gives recommendations for further research.

Chapter 2: Connectivity and related drivers

2.1. Introduction

Landscape development is influenced by the sediment transfer processes active in the landscape, themselves a reflection of the degree of hydrological and sedimentological connectivity (Harvey, 2002). On a global scale the majority (83%) of continental erosion is currently related to high elevation watersheds (Roy and Lamarre, 2011) due to high energy conditions associated with steep slopes, high drainage densities and erosive rainfall (Montgomery and Dietrich, 1988; Roy and Lamarre, 2011). Sediment export is accelerated in many catchments as a result of human influence (Womack and Schumm, 1977; Brierley and Murn, 1997; Walling, 1999). The entrained sediment ends up in river systems, leading to the export of valuable soil, highly turbid water, infilling of reservoirs and degradation of river health (Francke et al., 2008; Sandercock and Hooke, 2011; Boardman, 2013). With uncertain futures, research effort should be invested in understanding current and past landscape processes in order to conserve land and water for the future (Warner, 2006).

Over the past 150 years, various concepts and models were developed to study fluvial landscapes in order to answer basic questions such as ‘why does the river look the way it does?’ and ‘what would it look like in the future?’. Grant et al. (2013) gave a history of geomorphic concepts and proposed that concepts and models are mainly based on either physics, encompassing concepts such as hydraulic geometry (Leopold and Maddock, 1953) and frequency/magnitude (Wolman and Miller, 1960), or on form, such as ‘the fluvial system’ (Schumm, 1977) and channel classification (Rosgen, 1996). Both are useful to study river systems. Both departure points, whether physics or landform based, have their own limitations as discussed in detail by Dollar (2000) and Grant et al. (2013). Furthermore, geomorphologists and hydrologists have addressed some of the land and water problems with a reductionist approach by studying processes at the hillslope and river reach scale, ignoring influences of the surrounding catchment (Michaelides and Wainwright, 2002).

Significant earlier work by geomorphologists, e.g. Chorley (1969), Chorley and Kennedy (1971) and Schumm (1977), recognised influences of upstream processes. These well-established geomorphic principles were used by ecologists, such as the River Continuum Concept developed by Vannote et al. (1980), who noted that the characteristics of a river channel are a result of its position along the river continuum and related upstream processes. This concept assumes a gradual change in drivers of hydrological and sedimentological flux, but this is not always the case as abrupt changes in system drivers are frequently observed (Dollar, 2000; Grant et al., 2013).

Authors such as Leopold and Maddock (1953) and Schumm (1977) did not address connectivity *per se*, but did so implicitly. Frequently the core question revolved around sediment delivery and sediment budgets, which came with various shortcomings related to sediment transport processes over various temporal and spatial scales (Parsons et al., 2006; de Vente et al., 2007; Hinderer, 2012; de Vente et al.,

2013). These erosional, transport and depositional processes change with changes in scale and is influenced by factors such as landcover, geology, climate, etc. which are variable over time (de Vente et al., 2007).

Since the 1990s the concept of connectivity has developed as a unifying approach that allows geomorphologists to explore the internal structure and function of sediment dynamics at a range of spatial and temporal scales (Fryirs, 2013). Structure refers to the spatial layout of the various physical linkages, whereas function refers to the processes responsible for sediment movement. This connectivity concept has enabled an integrated catchment approach that embraces physics and form based concepts. In a recent review of the connectivity concept, Fryirs (2013) identified ‘thresholds of stability at various scales’ as the main shortcoming for understanding sediment connectivity and applying the concept to predict the accumulated future changes related to environmental change.

The concept has been developed internationally, but has only recently featured in geomorphological studies in South Africa, such as the Geomorphological Driver Assessment Index (Rowntree, 2013a), a review of connectivity in South Africa (Rowntree, 2012), in KwaZulu-Natal (Miller et al., 2013; Le Roux et al., 2013; Grenfell et al., 2014) and in the Karoo (Foster et al., 2012; Rowntree and Foster, 2012; Rowntree, 2013b; Grenfell et al., 2014), where research has begun to address the need to understand sediment transfer processes holistically. There is thus a need to further our understanding of sediment connectivity in other, often contrasting, geographical localities within South Africa. This study aims to apply and develop the connectivity concept in a new geographical locale where connectivity has been altered through changes in land use. As connectivity will be used as a conceptual framework for this study, the key literature related to the connectivity framework and the main factors influencing connectivity, such as magnitude, frequency, slope, vegetation cover, gully development, accommodation space and incision, will be reviewed.

2.2. *Connectivity*

2.2.1. Concept and definition

In most sediment studies, source, transfer and depositional zones and cut and fill processes are identified (Schumm, 1977; Prosser et al., 1994; Harvey, 2001; Fryirs, 2002). The distance between the source (cut process) and depositional zone (fill process) depends on the effectiveness of the transfer zone that represents the connectivity link. On a hillslope, mobilised sediment might be deposited within a relatively short distance (centimetres to metres) if the transfer linkage or connection is ineffective, such as a densely vegetated hillslope. However, where the link is more effective, such as a gully system, sediment might be transported to the bottom of the slope or to the river channel where it may be deposited as lateral bars or on the floodplains or transferred further downstream, depending on the effectiveness of the transfer process (Harvey, 2001). The general trend is that only a small proportion of eroded sediment will make

it to the bottom of the catchment as sediment is deposited along hillslopes and valley bottoms (Walling, 1983). In other disciplines, such as ecology, it is also recognised that not only the distance between the various resources, but also the degree of connectivity between units influences landscape structure and functioning (Taylor et al., 1993). Understanding connectivity proves valuable in a wide range of disciplines as it addresses both pathways through the landscape and processes involved in the transport of matter between landscape units.

In geomorphology and hydrology three types of connectivity can be identified, namely; landscape or geomorphological connectivity (physical linking of land units), hydrological connectivity (water moving from one place to the next) and sedimentological connectivity (sediment movement down a catchment) (Bracken and Croke, 2007). Michaelides and Chappell (2009) reviewed the various definitions of connectivity in terms of hydrology, geomorphology and ecology and found that the term often refers to macro connections between landscape units (structural connectivity) and their effects on water, sediment and biological processes. Hydrological connectivity relates to the ease of water movement through a landscape and how water movement is affected by the landscape and how it affects the landscape (Lexartza-Artza and Wainwright, 2009). Hooke (2003, p79) describes sedimentological or geomorphological connectivity as “the physical linkage of sediment through the system, which is the transfer of sediment from one zone or location to another and the potential for a specific particle to move through the system”. These definitions mainly refer to the process of water and sediment movement through a system, and to a lesser extent to the structural component of connectivity. As water is the driving force in sediment transport, it is thus vital to understand how the routing of water or structural connectivity affects the hydrology before one investigates the effects of connectivity on sediment transport or geomorphology (Michaelides and Wainwright, 2002; Bracken and Croke, 2007).

Connectivity in geomorphology is a relatively new concept that has recently been used extensively as an integrated approach to study complex systems (Lexartza-Artza and Wainwright, 2009; Michaelides and Chappell, 2009). Brierley and Murn (1997), Harvey (1997), Brierley and Fryirs (1998), Cammeraat (2002) and Hooke (2003) contributed to much of the ground work for connectivity as a geomorphological framework. This holistic approach improves the sediment delivery ratio method, a method where spatial and temporal lumping had a “blackbox” effect (Walling, 1983). The sediment delivery ratio is a catchment scale indicator of connectivity, determined by comparing the sediment yield against the gross soil erosion of a catchment, but ignores smaller-scale cut and fill processes, travel distances and holistic functioning of a system (Hooke, 2003; Parsons et al., 2006).

Another advantage of using the connectivity concept is that it can incorporate changes over time. By assessing the present and past processes as well as the structure in relation to environmental change, it can enable the prediction of likely outcomes (Harvey, 2002). Harvey (1997) stresses that predictions of environmental change should be studied holistically, including possible temporal and spatial changes that might occur to landscape connectivity. Thus, in order to gain understanding of sediment dynamics in a

specific catchment, one should study the location of sediment sources, the mechanics, conditions, pathways and distances of transport and the prerequisites for deposition and storage over time (Brierley et al., 2006; Sandercock and Hooke, 2011). This implies looking at the different landscape structures and how they function. This can become confusing as there are so many different landscape features in a catchment, such as rills, gullies, channels, etc. The connectivity framework brings structure to this through focusing on four dimensions at several scales.

The main challenge of the connectivity concept is that each catchment is unique and will require its own connectivity assessment (Brierley et al., 2006). Assessments are done at various scales, from a 'within landscape' scale up to a catchment scale, to get a holistic understanding of the system. Such a hierarchical range requires a large amount of research resources. Our understanding of thresholds of stability of various landscape units is still limited, especially across spatial scales, making the prediction of likely changes in connectivity challenging (Fryirs, 2013). According to Grant (2013, pers. comm.) the concept of connectivity is useful as a broad concept, but it needs a more rigorous definition and analytical framework for it to become a useful concept for geomorphic work.

2.2.2. Dimensions, scale and time

Ward (1989) presented a theoretical, four dimensional model of connectivity in lotic ecosystems. The four dimensions are: longitudinal (upstream-downstream or river network), lateral (river-floodplain), vertical (river-groundwater) and temporal (time since a catastrophe) connectivity. This model focuses on the river system only, neglecting how the wider catchment's connectivity influences connectivity of lotic ecosystems. Harvey (1997, 2001) and Brierley et al. (2006) expanded the model by adding hillslope-channel connectivity and giving it a nested arrangement as summarized in Table 2.1 and illustrated in Figure 2.1. Brierley et al. (2006) grouped different dimensions of connectivity at different spatial scales: within landscape, between landscape, sub-catchment and catchment. The latter two are combinations of within and between landscape compartments. At each spatial scale (e.g. between landscape compartment) the various dimensions at play (e.g. lateral, longitudinal) and direction (e.g. from channel to floodplain) are given. Each of these linkages can be assessed in terms of its strength through the methods listed in Table 2.1. Controls that are likely to influence the degree of connectivity are also listed, such as slope, shape of the valley bottom, inundation frequency, etc. This detailed framework by Brierley et al. (2006) gave a good overview of likely pathways and processes that sediment might encounter at various scales within a catchment.

Each of the dimensions of connectivity is influenced by spatial scale (Table 2.1). At the within-landscape scale the focus is on vertical and lateral connectivity, whereas longitudinal connectivity is introduced at the between-landscape scale and becomes dominant as catchment scale is approached (Brierley et al., 2006).

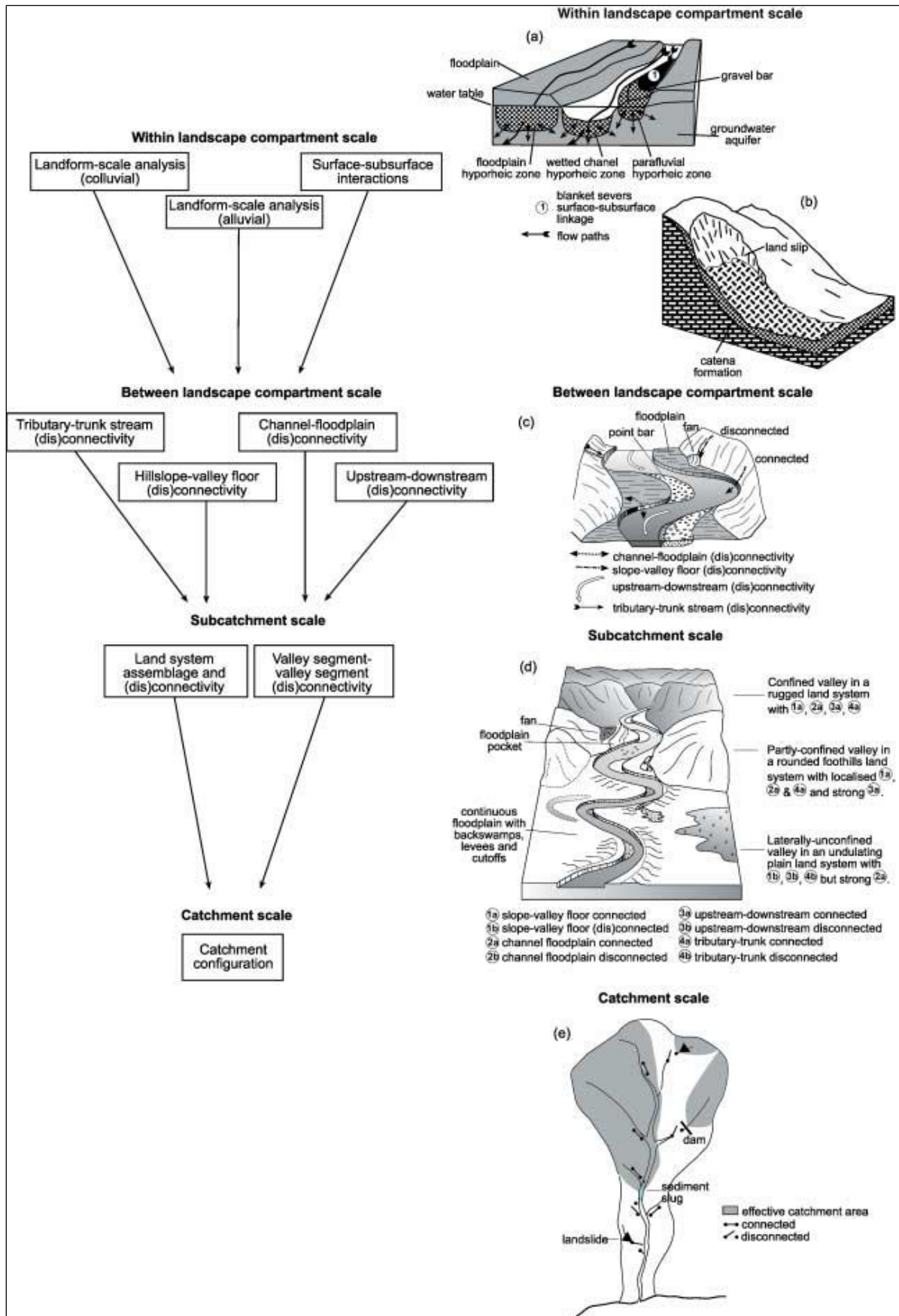


Figure 2.1: An illustration of the various types of connectivity at various scales (reproduced from Brierley et al., (2006) with permission).

Table 2.1: A summary of the different linkages and their processes, assessment measures and controls at various scales (reproduced from Brierley et al., (2006) with permission).

Type of linkage/scale	Processes	Measures used to assess strength of linkage	Controls
<i>Within landscape compartment</i>			
Landform-scale analyses (colluvial) (lateral linkage)	Development and reworking of hillslope processes along a catena.	Characterise sediment delivery within hillslope compartments through appraisal of the mechanisms, rates and downslope transfer of sediment along the catena. Assess any impediments to downslope sediment transfer in zero and first order systems.	Slope angle and morphology. Underlying geology and rates of sediment generation and reworking.
Landform-scale analyses (alluvial) (lateral linkage)	Formation and reworking of floodplains. Sediment transport and deposition in channels.	Characterise sediment storage and reworking on the valley floor. Appraisal of the mechanisms and rates of floodplain formation and reworking, and sediment transport capacity of channels.	Valley confinement and slope. Sediment supply and the magnitude-frequency of flows.
Surface–subsurface (vertical linkage)	Surface–subsurface exchange of water, sediment and nutrients. Infiltration and filtering. Maintenance of baseflow.	Characterise sediment and water exchange between surface waters and ground water compartments. Determine the presence, distribution and role of blankets that impede exchange between surface and subsurface compartments and their potential to be reworked.	Bed material texture. Sediment transport regime of the channel. Recurrence of channel flushing flows. Groundwater mechanisms.
<i>Between landscape compartment</i>			
Upstream–downstream (longitudinal linkage)	The transfer of flow through a system. The efficiency of supply, transfer and storage of sediments of variable calibre.	Appraise the pattern and role of barriers and boosters (i.e. longitudinal connectivity and continuity within the system). How readily can these barriers be reworked (i.e. the threshold conditions and recurrence interval under which they are likely to be breached)? Estimation of the ratio of transport capacity for a given range of events relative to sediment availability (and the character/accessibility of stores) involves examination of the degree of channel bed aggradation or degradation, the distribution of bedrock steps along the longitudinal profile and the degree of channel and valley confinement.	Base level. Sediment transport regime of the system (i.e. sediment supply or sediment transport limited).
Tributary–trunk stream (longitudinal linkage)	The transfer of flow through a system. The supply, transfer and storage of sediments of variable calibre.	Appraise the patterns of tributary (dis)connectivity by examining how often and over what length of river course tributaries are joined or disconnected from the trunk stream. Are buffers absent/present? Examine the impact that tributary contributions have on the trunk stream at the confluence (e.g. aggradation or degradation).	Shape of the catchment (i.e. its elongation ratio). Drainage pattern and density.
Slope–valley floor (lateral linkage)	Slope denudation and erosion via mass movement, creep, wash, etc. Colluvial footslope	Appraise how readily sediments transferred downslope are made available to channels.	Confinement of the valley floor. Channel position on the valley floor. The

Type of linkage/scale	Processes	Measures used to assess strength of linkage	Controls
	deposition and reworking. Deposition and reworking of materials on the valley floor. Channel adjustment on the valley floor.	Are buffers absent/present? What is the position of the channel on the valley floor and the nature of the hillslope–channel interface? Interpret the frequency with which impediments to sediment conveyance off hillslopes may be breached.	magnitude of flow events along the valley floor will dictate whether materials will be reworked along the channel network.
Channel–floodplain (lateral linkage)	Channel adjustment on the valley floor. Floodplain formation and reworking.	Appraise the character and volume of materials stored on valley floors, and the contemporary flux (i.e. floodplain accretion or reworking). Determine whether the reach operates as a sediment source, transfer or accumulation zone. What is the channel size and shape? What is the degree of channel aggradation or degradation relative to floodplain height? What is the floodplain inundation frequency? Is there any evidence of channel migration, avulsion, expansion, contraction?	Bed and bank material texture. Sediment transport regime of the channel relative to the floodplain. The magnitude and inundation frequency of overbank events that drive mechanisms of channel adjustment, and floodplain formation and reworking.
<i>Subcatchment scale</i>			
Valley segment–valley segment	The pattern and sequence of sediment source, transfer and accumulation zones along the valley floor.	Examine the pattern of upstream–downstream connectivity through the subcatchment as a whole. What is the sequence of valley settings (i.e. confined, partly confined or laterally unconfined valleys)? Are these sediment source, transfer or accumulation zones? Appraise the pattern and role of barriers and boosters (i.e. longitudinal connectivity and continuity within the system). Interpret the capacity for downstream propagation of sediment release from primary sediment stores, and their likely off-site impacts. Assess whether this is a transport-limited or a supply-limited system.	Valley confinement, valley slope, valley morphology which are controlled by underlying geology and landscape evolution.
Land system assemblage	Areas of relatively uniform topography measured in terms of relief, landform morphology, valley confinement and geology. Summarize slope–valley floor configuration.	Appraise tributary–trunk, slope–valley floor and channel–floodplain in the subcatchment as a whole. The role of buffers to sediment conveyance is examined.	Hillslope morphology, valley floor confinement, valley slope, valley morphology which are controlled by underlying geology and landscape evolution.
<i>Catchment scale</i>			
Catchment configuration	How valley segments and land systems fit together and are connected across a catchment to explain across	Measure the effective catchment area. Appraise how subcatchments fit together at the catchment scale through integration of subcatchment-scale relationships. Frame this in	Subcatchment variability in patterns of valley segments and land systems which are

Type of linkage/scale	Processes	Measures used to assess strength of linkage	Controls
	catchment variability in patterns of (dis)connectivity and flux	terms of analysis of how catchment shape, elongation ratio, etc. impact upon sediment conveyance, storage, etc. Determine the position of the most downstream blockage that impedes sediment output from the system. Predict the sensitivity of the landscape to change, where change will occur and be propagated from, and likely geomorphic responses.	controlled by underlying geology and landscape evolution.

Schumm and Lichty (1965) recognised the geomorphological relationship between timescales (stable, graded or cyclic time scale) and spatial scales. Dollar and Rowntree (2003) combined Schumm and Lichty's work with spatial scales and time scales taken from the geomorphological hierarchy (Rowntree and Wadson, 1999) to show how time and space scales are related. When considering short or stable timescales (less than a year), geomorphological changes are likely to happen at a localised scale, such as the hydraulic biotype. Likely changes are water level and sediment concentration. At the graded or geomorphological time scale (10-100s of years) changes are likely to happen at the reach scale, such as the channel form. When considering cyclic or geologic time scales (1 000s-millions of years), changes are likely to happen at the catchment or regional scale, such as catchment and drainage network morphometry.

The geomorphological relationships between time and space also apply to connectivity or cut and fill cycles. Local connectivity is mainly influenced by shorter term climatic and environmental factors, such as the duration and intensity of rainfall, irregularity of the soil surface and spatial arrangement of the vegetation and land units (Cammeraat, 2002; Harvey, 2002). As one considers larger systems (catchment scale), more permanent factors, e.g. tectonics and erosional history, are the main influences affecting connectivity (Prosser et al., 1994; Fryirs, 2002; Harvey, 2002). Harvey (2002) states that the frequency of the threshold exceeding event, the recovery period and the propagation time are important factors when looking at temporal changes in connectivity. Within landscape scale connectivity responds to short term events, such as a single flood (short propagation time) with a ten year recurrence interval, and has a relatively short propagation and recovery time. In larger systems, connectivity changes over longer periods (up to ca. 150 Ka) where propagation time becomes the important controlling factor.

Interactions between differing spatial and temporal scales are also possible (Bracken et al., 2015). A single localised sediment connectivity event can have an impact on sediment connectivity at a much larger spatial and temporal scale, e.g. a large landslide that dams a valley for thousands of years, effectively disconnecting the upper valley from the lower valley (Bracken et al., 2015). Feedback systems can also change how sediment is conveyed or stored in a landscape, e.g. rapid detachment and transport of

sediment can lead to the formation of an alluvial fan that will progressively trap and store sediment as it increases in size over time (Bracken et al., 2015).

The connectivity described above typically refers to ‘down system connectivity’ which is the movement of sediment from sources to sinks, but connectivity can also be propagated in the upstream direction (Harvey, 2002). Harvey notes that base level changes slowly migrate upstream throughout the channel network and onto the hillslopes where it cuts into colluvium and forms gullies. He maintains that down system propagation can be relatively quick (ten years), whereas up system propagation is a lot slower (up to ca. 150 Ka) and less affected by climatic changes (Harvey, 2002).

In well-connected systems a change in climate and resultant geomorphic response will be synchronous throughout the catchment, assuming that the effective thresholds are crossed (Harvey, 2002). In such a case increased erosion will lead to increased sediment supply to the downstream reaches; similarly, changes in stream base level will be propagated upstream, eroding into colluvium on lower hillslopes. The effects are transmitted effectively throughout the system (Harvey, 2001). On the other hand, in poorly-connected systems landscape processes could be contrasting and asynchronous, although different parts of the system could respond synchronously to climatic change (Prosser et al., 1994; Harvey, 2002). Where propagation times are slow, as with up-system connectivity, it is likely that different parts of the system are under different geomorphic regimes (Harvey, 2002; Rommens et al., 2006). Connectivity can follow a complex nonlinear pattern where a specific area switches between source and sink (cut and fill cycles) through time depending on environmental, climatic and tectonic influences (Prosser et al., 1994; Fryirs, 2002; Rommens et al., 2006).

2.2.3. (Dis)connectivity

Sediment connectivity is a dynamic process as the sediment transport linkages vary in efficiency. The degree to which the linkages are operational determines the connectivity between the various sediment stores and how a disturbance is transmitted through the landscape (Brierley et al., 2006). These linkages can also be decoupled or disconnected (Brunsdon, 1993; Fryirs et al., 2007a), a process known as (dis)connectivity. (Dis)connectivity or sediment deposition (fill process) plays a major role in the sediment budget as it explains why the sediment yield is often much lower than total erosion (e.g. low sediment delivery ratio) in the catchment (Heckmann and Schwanghart, 2013) and justifies the contention that a substantial amount of eroded material never reaches the stream channel (Walling, 1983). It further plays an important role in filtering of water and retaining water in the catchment (Fryirs et al., 2007b). While connectivity and (dis)connectivity or cut and fill processes are natural landscape phenomena, human activity can have significant impacts that alter the movement of sediment through a landscape (Prosser et al., 1994; Fryirs, 2002; Warner, 2006). Examples of natural linkages are: water movement on and in colluvium and alluvium, flood flows, overbank floods, soil creep and landslides.

Anthropogenically induced linkages include: increased runoff due to land use, fan channels, ditches, canals, irrigation return flow, channel straightening and canal off takes (Warner, 2006).

(Dis)connectivity is a result of breaks in sediment transport energy that prevent or limit movement of water and sediment (Warner, 2006). Where this happens sediment will be deposited, creating feedback systems that themselves dissipate energy and restrict further sediment movement, causing sediment not to move in a uniform and continuous manner (Harvey, 2001; Bracken et al., 2015). These depositional features or stores can go through cut and fill cycles, thus switch to being connected over time, becoming potential sediment sources. Depositional and structural features that limit sediment transfer can be grouped as buffers, barriers and blankets (Fryirs et al., 2007a). Buffers impact on hillslope-channel and longitudinal connectivity, barriers affect longitudinal connectivity and blankets hinder vertical and lateral connectivity.

Buffers are features that obstruct sediment transfer to the river channel, thus disrupting longitudinal and lateral connectivity and are found outside the channel (Brierley et al., 2006; Fryirs et al., 2007a). The various forms and characteristics are given in detail by Fryirs et al. (2007a) and are presented in Table 2.2. Floodouts, debris cones, colluvial aprons, floodplains, alluvial fans, terraces and trapped tributary fills are examples of natural buffers (Harvey, 2001; Warner, 2006; Fryirs et al., 2007b). Conservation contour banks and artificial embankments are examples of anthropogenic buffers (Warner, 2006). These features vary in size from 10–1000 m² and can consist of fine sediment, sands, gravels or a mixture of the aforementioned. Feature shape varies from conical to elongate and can store sediments for hundreds to thousands of years (Rommens et al., 2006; Fryirs et al., 2007a), but also act as potential sediment sources.

Table 2.2: The various forms and characteristics of buffers (reproduced from Fryirs et al., (2007a) with permission).

Form of buffer	Spatial scale (m ²)	Sedimentary character	Shape	Sediment cascade effect	Postulated effective timescale of disconnectivity	Breaching capacity
Intact valley fills/ floodouts	10 ² –10 ³	Fines	Elongate to lobed	Zone of dissipative flow. Sediment sink.	Thousands of years	Extreme event Infrequently reworked/ breached
Floodplain pockets	10 ¹ –10 ³	Mixed	Elongate and stepped	Localised scale geomorphic feature. Sediment sink.	Thousands of years	Overbank flow stage Infrequently reworked/ breached
Continuous floodplains	10 ¹ –10 ³	Mixed	Elongate and stepped	Landscape scale geomorphic feature. Sediment sink.	Thousands of years	Extreme event Infrequently reworked/ breached
Alluvial fans	10 ² –10 ³	Mixed	Conical	Landscape scale geomorphic feature. Sediment sink.	Hundreds to thousands of years	Extreme event Infrequently reworked/ breached
Piedmont zones	10 ³	Mixed	Planar	Landscape scale geomorphic feature. Sediment sink.	Thousands of years	Extreme event Infrequently reworked/ breached
Terraces	10 ² –10 ³	Sands and gravels	Elongate and stepped	Landscape scale geomorphic feature. Sediment sink.	Thousands of years	Extreme event Infrequently reworked/ breached
Trapped tributary fills	10 ² –10 ³	Fines	Irregular	Sediment sink.	Hundreds to thousands of years	Overbank flow stage Infrequently reworked/ breached

All buffers go through periods of aggradation, stabilization and degradation, e.g. alluvial fans may aggrade, but can also incise if the sediment budget goes negative. Harvey (2002) found that alluvial fans seem to aggrade during times of surplus sediment input, such as periods toward the end of glaciated periods (rapid weathering of materials) or periods characterised by increased availability of material (landslides during wetter periods).

Where buffers are relatively small, such as discontinuous floodplains and narrow terraces, they are likely to be breached, especially where there are high energy contributions of runoff from the slopes or tributaries (Fryirs et al., 2007b).

Fryirs et al. (2007a) describe barriers as features in the channel that delay sediment movement down the channel, such as bedrock steps, sediment slugs, woody debris and dams (Table 2.3). These features affect longitudinal linkages, mainly due to their changes in base level or bed profile. Barriers reduce the upstream slope, lead to the dissipation of energy and associated sedimentation and limit local sediment transfer. Barriers can be 1-1000 m² in size and consist of sand, gravel, bedrock or a mixture of various materials. The majority have irregular shapes and can store sediment for tens to thousands of years or permanently, depending on the elimination of the barrier (Kasai et al., 2005; Fryirs et al., 2007a). Examples of different barriers are given below (Table 2.3).

Table 2.3: Various forms and characteristics of barriers (reproduced from Fryirs et al., (2007a) with permission).

Form of barrier	Spatial scale (m ²)	Sedimentary character	Shape	Sediment cascade effect	Postulated effective timescale of disconnectivity	Breaching capacity
Bedrock steps	10 ⁰ –10 ¹	Bedrock	Irregular but stepped	Aid backfilling of valleys	Permanent over thousands of years	Extreme event Infrequently reworked/breached
Valley constriction	10 ² –10 ³	Bedrock	Irregular	Landscape scale geomorphic feature. Bottle neck to sediment transfer. Aid backfilling of valleys.	Permanent over thousands of years	Extreme event Infrequently reworked/breached
Sediment slugs	10 ² –10 ³	Sand or gravel	Elongate and lobed	Acts as a plug on sediment transfer along channels.	Tens to hundreds of years	Channel flows with the ability to entrain materials of varying sizes on the channel bed Frequently reworked/breached
Channel capacity (high width/depth ratio channels)	10 ¹ –10 ²	Mixed	Symmetrical, asymmetrical, irregular or compound in cross-section	Controls flow concentration and therefore unit stream power. Sediment store.	Tens to hundreds of years	Channel flows with the ability to entrain materials of varying sizes on the channel bed Frequently reworked/breached
Woody debris	10 ¹ –10 ²	n/a	Irregular	Acts as a blockage behind which sediment accumulates	Tens to thousands of years	Channel flows with the ability to entrain materials of varying sizes on the channel bed Frequently reworked/breached
Dams	10 ²	n/a	n/a	Anthropogenic, landscape scale geomorphic feature. Blocks sediment conveyance and enhances retention.	Permanent unless removed	Extreme event Infrequently reworked/breached No flow energy so no transport capacity

Where tributaries contribute large quantities of coarse material onto the valley fill and into the channel, the width of the trunk stream is reduced, effectively acting as a barrier for up to 100 years (Kasai et al., 2005). Resistant dolerite dykes downstream of softer sandstone valleys often form barriers as lateral

planation of the sandstone results in increased sediment accommodation space and reduced upstream valley slope and cause sediment deposition upstream of the resistant barrier (Tooth et al., 2002; Grenfell and Ellery, 2009). Increased coarse sediment supply to a river reach will lead to localised bed aggradation or sediment slug and the formation of in-channel bars and a high width-depth ratio channel (Brierley and Murn, 1997; Fryirs et al., 2007a). This build-up of sediment will reduce the unit stream power, forming a barrier that could reduce connectivity for tens to thousands of years (Fryirs et al., 2007a).

Hooke (2003) suggested five classification types of down-channel connectivity for coarse sediment; a) unconnected reaches where sediment produced from local sinks are stored locally with incompetent reaches isolating the local source or sink areas, b) partially connected reach where transfer only occurs during extreme events, c) connected reaches where sediment is flushed through during ‘normal’ flood events, d) potentially connected reaches where the reach is competent to transport the sediment, but no sediment is supplied and e) disconnected reaches where obstructions such as dams prevent sediment transport. Hooke used detailed geomorphological maps to assess the type of connectivity in the various river reaches that were studied. This gave a good indication of the dynamics within the reach and what could be expected in terms of barriers to coarse sediment transport.

Barriers can be naturally occurring features, e.g. geological nick points, riffles and woody debris, or anthropogenic, e.g. weirs, dams, barrages and dredging ponds.

Blankets are features that disrupt vertical linkages and smother other landforms, temporarily protecting the underlying landforms from being reworked or transported (Fryirs et al., 2007a). Fryirs et al. (2007a) give a detailed explanation of types of blankets and their characteristics (Table 2.4). Blankets can be found in the channel (bed armour) or on the floodplain (sand sheet), varying in size from 0.1-1000 m². It varies from thin to thick (1-100 cm) planar sheets covering large areas, but elongated features also exist. Material varies from fine grained materials to cobbles that can be stored for a year up to hundreds of years.

Table 2.4: Various forms and characteristics of blankets (reproduced from Fryirs et al., (2007a) with permission).

Form of blanket	Spatial scale (m ²)	Sedimentary character	Shape	Sediment cascade effect	Postulated effective timescale of disconnectivity	Breaching capacity
Floodplain sediment sheets	10 ¹ –10 ³	Mixed	Planar sheet — can be relatively thick	Changes soil profile and hydrological properties. Sediment sink.	Tens to hundreds of years	Overbank flow events with ability to rework floodplain surfaces Recurrently reworked/breached
Fine-grained material in interstices of gravels	10 ⁻¹ –10 ¹	Fines	Planar draped — relatively thin	Changes hydrological properties at a 1 mm to 1 m scale. Sediment store.	Individual event to years	Channel flows up to bankfull with the ability to flush fines Recurrently reworked/breached
Channel bed armouring	10 ¹ –10 ²	Mixed, but tends to be gravels or cobbles	Various — elongate, diagonal, arcuate, etc.	Changes hydrological properties at a 1 mm to 1 m scale. Sediment store	Years to tens of years	Channel flows with ability to entrain and rework surface armour and release subsurface materials. Recurrently reworked/breached

Blankets can also disrupt lateral channel-valley fill connectivity. This occurs where blankets formed on floodplains increase the height of the banks, effectively enlarging the channel capacity. This causes reduced lateral connectivity as flood flows are contained within the channel.

2.2.4. Boosters

Connectivity also can be enhanced through boosters, that are features such as sunken lanes, gorges or incised channels, where sediment is rapidly transported to the river or down river (Brierley and Murn, 1997; Brierley et al., 2006; Boardman, 2013). Boosters tend to concentrate the flow which increases the stream power (Equation 1.1) as discussed later on in this document. Human intervention may cause boosters to develop, such as levees, wetland drains and sunken lanes (Miller et al., 2013; Boardman, 2013). Boardman (2013) also found that boosters (incised roads) promote further lateral erosion, such as soil piping and mass movements.

Gergel et al. (2002) looked at the influence of levees on floodplain inundation. Levees restricted the movement of water onto the floodplain, reducing the area of floodplain inundated, effectively acting as an incised channel where channel-floodplain connectivity is reduced. However, levees also create problems when they are breached and then repaired. When that happens, the material on the floodplains no longer is available to the system. Levees also reduce sediment supply because they prevent calving (bank collapse and erosion). The construction of levees led to a slight increase in stage height in the channel for a given discharge. Flow velocity increased as a result of increased stage height, increasing the amount of work that the stream can do. In this case the floodplain section of the river, which used to function as a buffer, was turned into a booster through the disconnection of the floodplain, increasing the longitudinal connectivity and reducing the lateral connectivity of the system.

2.2.5. Magnitude and frequency

Figure 2.2 shows that larger magnitude events occur less frequently than small to moderate events. Wolman and Miller (1960) studied this relationship of frequency-magnitude on the amount of sediment moved in large basins (5000+ km²). Their work is summarised in Figure 2.2 and shows that the rate of sediment movement increases with increased magnitude, but that the frequency of these larger events decrease with magnitude. The result is that a greater total volume of sediment is moved by moderate events (e.g. two year frequency) that occur relatively frequently, than by significantly larger events (e.g. 100 year frequency) that also move large volumes of sediment but do so infrequently. This pattern changed for smaller basins (<10 km²) where the amount of sediment transported by the large, less frequent events increased (e.g. 90% of the work is done by events with a >5 year recurrence interval) (Wolman and Miller, 1960).

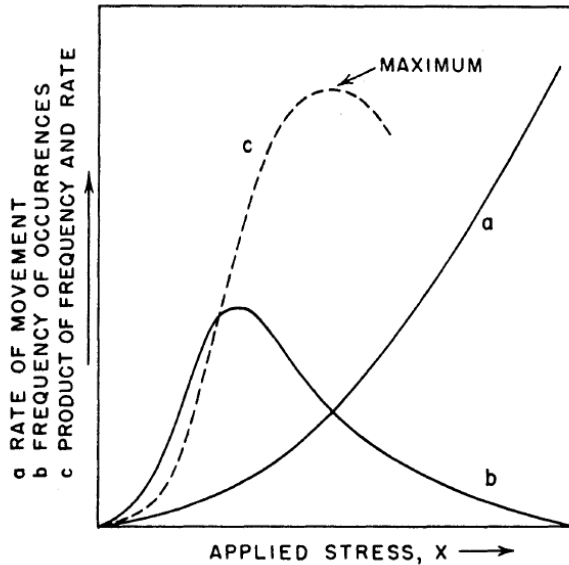


Figure 2.2: The rate of transport, event frequency and applied force (Wolman and Miller (1960); reproduced with permission).

Runoff generation is influenced by rainfall, surface and topographic characteristics, such as structural connectivity (Michaelides and Wainwright, 2002; Beel et al., 2011). Rainfall intensity proves to be the main factor responsible for runoff generation (Michaelides and Wainwright, 2002). High intensity rainfall (large rainfall amount over a short period of time) produces large volumes of runoff, but the fate of the runoff depends on the surface conditions and routing of the water.

Western et al. (2001) modelled the effects that event magnitude has on runoff generation in plots with well and poorly connected soil moisture patterns. It was found that well-connected patterns produced a higher total discharge for low, medium and high magnitude events. An unexpected result of the model was that poorly connected patterns produced the greatest peak discharge, which can be explained by the efficiency of the well-connected landscapes to drain water. Drainage of water will start sooner in the well-connected catchment, resulting in a less peaky hydrograph, while in poorly connected catchments there is a build-up until all areas are saturated effectively, only releasing the bulk of the water when barriers are breached, producing a high peak discharge. This highlights the complex nature of connectivity.

It has been shown that the difference in discharge between well and poorly connected systems decreases as rainfall magnitude increases, as buffers are breached when they become saturated, effectively linking all possible contributing areas (Western et al., 2001; Michaelides and Wainwright, 2002). Buffers such as floodplains would prevent connectivity during smaller events, but once the floodplain is saturated, it will not only act as a link, but also contribute to the discharge (Michaelides and Wainwright, 2002). This shows that increases in structural connectivity will exacerbate the propagation of overland flow due to lower flow resistance and infiltration, producing greater discharge throughout a range of events and increase functional connectivity (Western et al., 2001; Michaelides and Wainwright, 2002).

Another effect of increased discharge due to enhanced hillslope-channel connectivity is that of increased frequency of an event with a given magnitude (Cammeraat, 2002). As stated above, the general trend is that small magnitude events happen frequently, whereas large magnitude events occur infrequently (Fryirs et al., 2007a). When discharge is increased as a result of increased hillslope-channel connectivity, it results in a higher event magnitude at the same frequency or, conversely a higher frequency of the same event magnitude. As channel-valley fill connectivity is only established during larger magnitude events when banks are overtopped, the frequency of channel-valley fill connectivity is also increased as a result of increased discharge, assuming the channel dimensions remain stable. Major events almost always alter channel geomorphology, at least to some extent, unless the river banks have been armoured or there is really extensive vegetation.

As sediment is mainly transported by moving water, we need to understand how the energy of moving water impacts sediment movement. Hjulström (1935) showed that more energy is required to mobilise a particle (to overcome the shearstress threshold) than is required to transport it and that a particle will be deposited once energy is reduced below a given transport threshold. Bull (1979) used the threshold of critical power in order to determine whether sediment is deposited or entrained. The threshold of incipient movement is given in Equation 2.1 as:

$$\frac{\text{Stream power}}{\text{Critical power}} = 1 \quad (2.1)$$

where stream power is a function of discharge and gradient, and critical power is the stream power needed to move the median particle size (Bull, 1979). Although this equation was mainly intended for stream channels, it also applies to sediment on slopes. From Equation 2.1 one can see that a scenario where a factor >1 is experienced would result in the mobilisation of sediment and would happen when the power exerted by the flowing water exceeds the specific critical power of the sediment. The magnitude or discharge of an overland or channel flow event determines whether the threshold that prevents sediment entrainment will be crossed. As larger particles require larger stream power for incipient movement, sediment entrainment or connectivity for a range of sediment sizes is only established during larger events (Harvey, 2001; Kasai et al., 2005; Fryirs et al., 2007a; Beel et al., 2011). Beel et al. (2011) found that events with rainfall intensities exceeding 10 mm h^{-1} (frequent event) breached the critical power for fine sediment transport in steep headwater catchments in New Zealand. Harvey (2001) found that coarse sediment was mobilized during a large event with a return frequency of 100 years. These larger flow events are evident in sink zones where lenses of coarser material are found, indicative that stream power was sufficient to transport coarse sediment onto flood plains during overbank flow events (Miller et al., 2013).

Landscape features such as hillslopes and fans also can have intrinsic thresholds of instability related to slope angle that could alter landscape connectivity. Schumm (1979) explored the slope thresholds of stability for gullies, fan heads and river channels. Gullies formed where slopes were steeper than a specific threshold slope, showing that the development of the gully was not related to an extrinsic driver, such as land use, but related to the steepening of the slope until the threshold of slope stability was crossed. For fan head incision the pattern was similar, incision was initiated once a threshold slope gradient was crossed. For channel pattern the threshold was also related to valley floor slope, where the channel was straight to sinuous for low valley floor angles. When the valley floor slope steepened, a threshold was crossed and the channel developed a meandering pattern. Further valley steepening led to a braided channel pattern, showing that yet another slope dependant threshold was crossed. Landscape connectivity can change as a result of natural cycles, such as ongoing fan deposition and steepening, that cross a stability threshold and switch from a phase of aggradation to a phase of degradation, changing a buffer to a sediment source and conduit.

Event magnitude determines the extent of hydrological connectivity. Runoff was generated at the plot scale for events from 6-10 mm per day, whereas slope scale runoff was produced for events ranging from 20-25 mm per day (Cammeraat, 2002; Warner, 2006). Catchment-wide connectivity can be established for events bigger than 30 mm per day (Cammeraat, 2002). Fryirs et al. (2007a) used a series of switches in a catchment to illustrate the effect of event magnitude on connectivity (Figure 2.3). The switches in the catchment represent barriers, buffers and blankets. The switches are off (i.e. not connected) until larger magnitude events produce unit stream power sufficient to overcome the critical threshold and enable sediment transfer or connectivity. Kuo and Brierley (2014) demonstrated the effects of event magnitude where annual low magnitude events allowed localised lateral and longitudinal connectivity, but that catchment scale longitudinal connectivity was only established during larger magnitude events of decadal frequency. This shows that structural connectivity is independent of time, unless there is landscape change, but that functional connectivity fluctuates over time and is dependent on the hydrological state and sediment availability. Figure 2.3 also shows that the contribution of the catchment increases as more switches are on during larger magnitude events. This can be seen in the work of Francke et al. (2008) where sediment originated from the upper catchment during large events, but had a local source during smaller events. During extremely large events (i.e. all switches on), the inherent structural landscape connectivity plays a small role in the overall sediment connectivity as catchment-wide hydrological connectivity is accomplished by the sheer magnitude of the event.

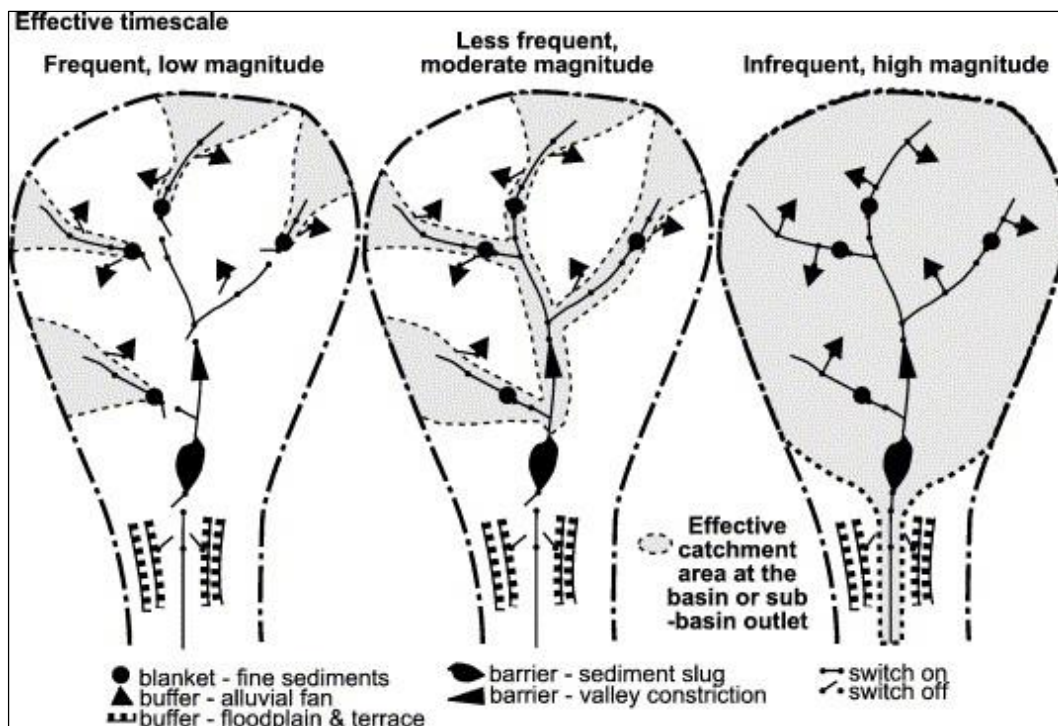


Figure 2.3: Event magnitude and the size of the effective catchment (reproduced from Fryirs et al., (2007a), with permission).

In a review of landscape change and sensitivity, Brunnsden and Thornes (1979) concluded that a specific landscape is the result of infrequent events of large magnitude. These large disturbances lead to a change in processes and connectivity that can relax over time to the previous state of functioning. Some systems are inherently more sensitive to large events, making them more vulnerable. Sensitivity in this sense refers to the state of a system or feature, where for example, a system is close to a critical geomorphological threshold, such as a steep river bank nearing the angle where the sediment is no longer stable and could collapse if the angle or weight of the sediment is increased. Where a system is close to a critical geomorphological threshold, it will respond dramatically to a large flood event by changing its functioning and structural connectivity, compared to a similar system further away from a geomorphological threshold responding minimally to the same event magnitude, maintaining its previous functioning and structural connectivity (Schumm, 1973). The rate and processes involved in the relaxation period are complex and are affected by the environment, intrinsic characteristics and the degree of connectivity (diffuse or direct) (Brunnsden and Thornes, 1979). Where systems are well connected, effects of large magnitude events will be efficiently propagated, making the system more sensitive to large events. Thus, the effect of a large event and subsequent repercussions will depend on the geomorphological structure and sensitivity of the system (Schumm, 1973).

2.3. Drivers of hillslope-channel and channel valley fill connectivity

In the Thina catchment, longitudinal connectivity has not been significantly impacted by structural interventions such as weirs or dams. The focus of this thesis is on hillslope to channel and channel to valley fill connectivity which are hypothesised to be the main pathways that have been affected over time. Hillslope-channel and channel-valley fill connectivity are influenced by a mixture of landscape factors, such as slope, land cover, gullies on slopes and the incision of channels. Trends and effects related to connectivity as a result of changes in these drivers will be discussed below.

2.3.1. Factors affecting hillslope-channel connectivity

Harvey (1997) stresses that the key to understanding upland geomorphic processes is to look at hillslope-channel connectivity, especially over varying spatial and temporal scales. Michaelides and Wainright (2002, p. 1442) defined hillslope-channel connectivity as ‘the effectiveness, direction and speed with which localized changes are transmitted away from the source (hillslope) and propagated throughout the hillslope system to the channel and ultimately to the catchment outlet’. In this section a review of the literature related to selected factors that will influence hillslope-channel connectivity, including slope, land cover, gully systems and valley confinement, is presented.

2.3.1.1. Slope gradient

The different geomorphological zones of a river, such as upper, middle and lower catchment, each has its own ability to transfer and store sediment. The upper catchment primarily transports sediment, whereas the lower catchment mainly stores sediment (Miller et al., 2013). Headwater catchments are generally described as having steep valleys with steep slopes, a high drainage density, often having a V shape, limited accommodation space, with small valley fills developing where valley slope decreases. According to Equation 2.1 steeper slopes will have more stream power that will result in more overland flow as a result of relatively short residence time of water on the surface (Kirkby et al., 2002). This means that steeper slopes have a higher potential to move sediment if all other conditions were equal. This can be seen in the work of Fryirs et al. (2007a) in headwater catchments where available sediment is efficiently transported from the steep slopes to the valley bottom. The high stream power generated on these steep upland areas result in sediment being moved and reworked relatively frequently (Kasai et al., 2005; Fryirs et al., 2007a). Slope gradient has an indirect effect through drainage density, soil thickness and vegetation cover. These are covered in more detail in the sections below.

Drainage density is another characteristic that influences connectivity. Steeper slopes have less surface storage mainly due to the effective drainage of depressions that would act as sinks on gentler slopes (Kirkby et al., 2002). The trend is that steeper catchments have denser drainage networks, and results in

the slopes being well connected to the channel, as runoff and sediment on a slope is relatively close to a drainage feature due to the high density of drainage structures (Kirkby et al., 2002). Slope can thus be used as indicator of potential connectivity, as was shown by Fryirs et al. (2007b) who used DEM based slope classifications to show potential barriers and buffers (less confined, low gradient ($<2^\circ$) areas with large accommodation space). The modelled output agreed with field observations. Slope, among a number of other characteristics, is included in numerous erosion models such as USLE, MUSLE, RUSLE, etc. This confirms the need to include and account for the effect of slope on both connectivity as well as sediment transport.

Another feedback related to slope is vegetation cover, as slope plays an important role in the thickness of sediment on a slope, which has a direct influence on the retention of water and thus influences favourable habitat for plant growth. Sediment is moved downslope to areas where it accumulates or is transported to the main channel. Where the soils accumulate, water can be stored and made available to plants. Kirkby et al. (2002) found that the thickness of soils and available water led to a scenario where steeper slopes had less vegetation cover, making the steeper slopes inherently more vulnerable to erosive rain and overland flow.

2.3.1.2. Effects of land and vegetation cover on connectivity

The geomorphic balance is often disturbed by increased discharge as a result of urbanization (increased impervious surface), reduced vegetation cover, higher drainage density or climate change (Brookes, 1994; Gore, 1994; Brierley and Murn, 1997; Cammeraat, 2002; Vanacker et al., 2005; Rommens et al., 2006; Grenfell and Ellery, 2009). The disturbance of the geomorphic balance through altered land and vegetation cover affects the connectivity of the landscape, changing the functioning of the various landscape compartments in terms of sediment production, transfer or storage (Houben, 2008). The main impacts of decreased vegetation cover is increased slope erosion, gully formation (increased connectivity) and increased sediment delivery to the river channel (Brookes, 1994; Damm and Hagedorn, 2010). Rommens et al. (2006) describe a tenfold increase in sedimentation over the last 1000 years in central Belgium, which coincide with afforestation, agriculture and extreme rainfall events. The sediment connectivity has thus drastically been changed as a result of land cover change. Where channels in upland valleys are deepened as a result of vegetation clearance, sediment is rapidly exported and can take thousands of years to refill as slope contributions can be very low (Brierley and Murn, 1997). This confirms that stability thresholds in the landscape can easily be crossed by human impact and can take a long time (if ever) to revert to the pre-disturbance levels of functionality.

In this section of hillslope-channel connectivity, the review will focus on the effect of vegetation cover on sediment connectivity. Dense vegetation or ground cover plays an important role in limiting sediment movement as it protects soils from erosive rainfall, increases the runoff threshold, reduces overland flow through increased surface roughness and acts as a sediment buffer (Gore, 1994; Cammeraat, 2002; Kirkby

et al., 2002; Fryirs et al., 2007a). Kirkby et al. (2002) found that the runoff coefficient decreased for longer slopes, mainly due to the reduction in overland flow along well vegetated slopes. The effect of dense vegetation cover on downslope sediment movement was also noted by Fryirs and Brierley (1999) where eroding upper slopes did not result in colluvial wedges on the valley fill. They relate this to the capacity of vegetation to trap and store sediment on the lower well vegetated slopes. The compartmentation of agricultural fields on hillslopes can reduce the connectivity and transfer of sediment through the interruption of the sediment cascade (sediment delivery) by terraces, strip lynchets and vegetation strips (Houben, 2008).

Where vegetation cover is reduced through cultivation, fire, drought or overgrazing, soils are exposed to erosive rainfall, which leads to local erosion (Brookes, 1994; Harvey, 2001; Cammeraat, 2002; Peticrew et al., 2006; Fryirs and Brierley, 2013). Raindrop impact can either detach soil particles or compact the soil to form a crust that will generate increased runoff (Kirkby et al., 2002). Compacted soil will increase runoff but may, in fact, decrease sediment concentrations and sediment connectivity because it limits sediment supply. Wild fire decreases vegetation cover and increases soil hydrophobicity, which leads to increased runoff volumes and the soil's vulnerability to erosion (Cerdeira and Lasanta, 2005). Where soil particles are detached and transported, a hollow will be formed that can develop into a hydrological conduit once hollows are linked. In both cases overland flow will be increased, leading to higher transport capacities and downslope connectivity (Kirkby et al., 2002; Beel et al., 2011). Several studies relate high sediment yields visible in depositional sequences to periods of decreased vegetation cover (Kasai et al., 2005; Foster et al., 2007). Steep headwater catchments will usually be well connected due to their steep nature, but changes in vegetation cover can have marked effects on sediment connectivity and yield (Brookes, 1994; Kasai et al., 2005).

The spatial layout of vegetation also affects connectivity. Where once continuous vegetation cover becomes patchy, buffers are breached and connectivity pathways are established (Cammeraat, 2002; Kakembo et al., 2009). This allows water and sediment to move freely between vegetated patches.

When agricultural land is abandoned, a fire consumed most of the vegetation or a landslide has stripped a hillslope, soil can remain bare for several years before vegetation recolonizes it (Harvey, 2001). This can be seen where sediment yields peak after land abandonment or wild fire and yields eventually drop down when vegetation cover increases (Kakembo and Rowntree, 2003; Peticrew et al., 2006; López-Vicente et al., 2013). In the Woluma Creek, New South Wales, slope vegetation only recovered decades after cultivation, resulting in significant soil loss (Fryirs and Brierley, 1999). This shows that the disturbance of vegetation (e.g. cultivation or fire) increases both sediment entrainment and downslope sediment connectivity and can take years to decades to recover.

Herbst et al. (2012) used grazing exclosures (fenced pens) and complete removal of grazing on third order streams in the interior western United States to assess its effectiveness on river system recovery. The effects of small exclosures (established for 10+ years) on stream habitat were less pronounced than in the

catchments where grazing was completely excluded (for four years). The main geomorphological differences in grazed catchments were: steeper, actively eroding banks; higher solute concentrations in the water; streambeds with increased fine sediment; wider channels. Exclosure plots had better riparian vegetation cover and composition than grazed areas, but geomorphologically the channel was not recovering as a result of continued disturbance through grazing upstream of the exclosure plots. These data suggest that minor catchment-wide reductions in hillslope-channel connectivity (through vegetation recovery) are more effective in channel rehabilitation than major localized reductions in connectivity.

In areas where flooding or channel-floodplain connectivity is reduced by reservoirs or levees, valley fill sediment is disturbed less often and frequently colonised by invasive plants that form thick growths and stabilize sediment (Gergel et al., 2002; Nakamura and Tockner, 2004). The reduced frequency of flooding and related disturbance results in the vegetation composition changing from a disturbance related riparian vegetation configuration to a more permanent terrestrial vegetation structure (Gergel et al., 2002).

2.3.1.3. Hillslope gullies as sediment production zones and conduits

Hillslope gullies, often referred to as arroyos, wadis or semi-arid erosional channels elsewhere in the world, are landscape features that not only produce large quantities of sediment, but also increase hillslope-channel connectivity. Cut and fill processes are natural cycles in most fluvial landscapes, as can be seen in many floodplain sequences (Grenfell and Ellery, 2009). However, a certain threshold must be crossed to initiate the cut phase (Schumm, 1979), such as increased runoff as a result of reduced vegetation cover (Brierley and Murn, 1997; Grenfell and Ellery, 2009). Brierley and Murn (1997) suggest that the pre-human disturbance cut and fill phases were more localised or discontinuous. With human disturbance (e.g. agriculture) cut extent of small channels on valley bottoms was enlarged, resulting in more continuous cut features, such as incised channels and gully networks (Fryirs and Brierley, 1999). The function of gullies as sediment production zones and conduits will be discussed below, including gully formation and stabilization.

Gullies are cut features that remobilize stored sediments and are seen as point sources of sediment (Kakembo and Rowntree, 2003; de Vente et al., 2006). Gullied basins in Australia produce several times as much sediment compared to ungullied basins (Wasson, 1994). Wasson (1994) found that high sediment yields were related to small, steep gullied upland basins, with mean annual sediment yield being a linear function of gully density. This would suggest that gullies can thus be described as sediment production zones, but may simply be connecting upstream sediment sources to downstream sinks.

The spatial layout of runoff units and their connection to the channel would influence the geomorphic and hydrologic response at the catchment scale (Cammeraat, 2002). Water and sediment are optimally transported where drainage density is maximised (Fryirs and Brierley, 2013). Croke et al. (2005) traced runoff on dispersive and gullied (concentrated) pathways in New South Wales, Australia. Runoff travelled

two to three times further downslope in gullied pathways than dispersive pathways. On dispersive pathways fine sediment (<63 μm) stays in suspension until the runoff infiltrates (Croke et al., 2005). This implies that gullies effectively transfer all fine sediment directly into the drainage network as infiltration is limited along gully features. Where gullies are not linked to the drainage network, i.e. discontinuous features, water will be dispersed and sediment could potentially be deposited. Croke et al. (2005) also found that gully development effectively linked other drainage features, such as roads and trails, to the stream network and increased the drainage density by up to 10%.

Gullies act as boosters that concentrate runoff, increasing the stream power and energy of the water available to move sediment (López-Vicente et al., 2013). As seen in Equation 2.1, this implies that sediment will more likely be entrained where flow is concentrated, enabling larger volumes and larger sediment clasts to be moved compared to dispersive pathways. Where sediment is transported to a gully feature, the high stream power will prevent the sediment from being deposited, effectively increasing the movement of sediment downslope (Croke et al., 2005).

2.3.1.4. Hillslope gully development and stabilization

In former communal areas in the Eastern Cape, South Africa, severe gully and rill erosion developed on erodible soils. Such soils often are formed on sedimentary parent materials, that were ploughed and cultivated for several years and were subsequently abandoned (Kakembo and Rowntree, 2003; Vetter, 2007). Bare fields were exposed to erosive rainfall and led to the formation of gullies (Kakembo and Rowntree, 2003). Croke et al. (2005) found that gullies mainly formed on steeper slopes with a large contributing area in New South Wales, Australia; thus gullies are likely to form on sensitive materials where there is sufficient transport energy.

Erosive features form when the surface stability threshold is exceeded due to a surface disturbance or increased runoff energy (Equation 2.1). Surface disturbances, such as the formation of livestock tracks or ploughing, expose soils to erosive forces, making it a prime area for the initiation of a gully, especially if upslope runoff also is increased. Increased runoff energy can be a result of two factors, the first being an increase in slope and the second an increase in discharge (Harvey, 2002). Where base level is lowered through the removal of a barrier or the formation of a streamside bank collapse, a local lowering in elevation will make more energy available to cut into sediment (Harvey, 2001). This will form a headcut (with a steep local gradient) that slowly works its way upstream or upslope, forming a gully that is linked to the channel.

The second factor responsible for gully formation (increased runoff) can be experienced as a result of climatic or land cover changes. Wetter periods (e.g. interglacials and exceptional storm events) will lead to increased runoff volumes (Wolman and Miller, 1960; Harvey, 2002; Kakembo and Rowntree, 2003). However, under stable climatic conditions, land cover change also can be experienced and results in

increased runoff volume (Harvey, 1992; Gore, 1994; Cammeraat, 2002; Miller et al., 2013). Gullies formed due to increased runoff are not always directly connected to the main drainage network (Harvey, 2002).

Harvey (1992) stated that gullies are the result of localised, non-equilibrium processes and are characterised by progressive changes in process-form relationships that have spatial and temporal limits. He depicted the development of a gully in northwest England as a headcut moving upslope while cutting down into the slope over time, effectively lowering the slope of the lower section of the gully (Figure 2.4). The extent of the gully will increase until a barrier prevents it from spreading or when the gully slope reaches a stability threshold (Fryirs and Brierley, 1999). During the later stages of gully development, larger material is deposited along the lower channel, allowing vegetation to be established in the channel and on gently sloping areas. This means that the gullies become less efficient in their sediment transfer as they increase in size and vegetation colonises the gully channel (Harvey, 1997). Harvey (1992) found that large material will be flushed out during larger storms (2-5 year frequency), restoring the efficiency of sediment transport until it fills again. Harvey also found that gully development was strongest where the feature was directly linked to the stream channel.

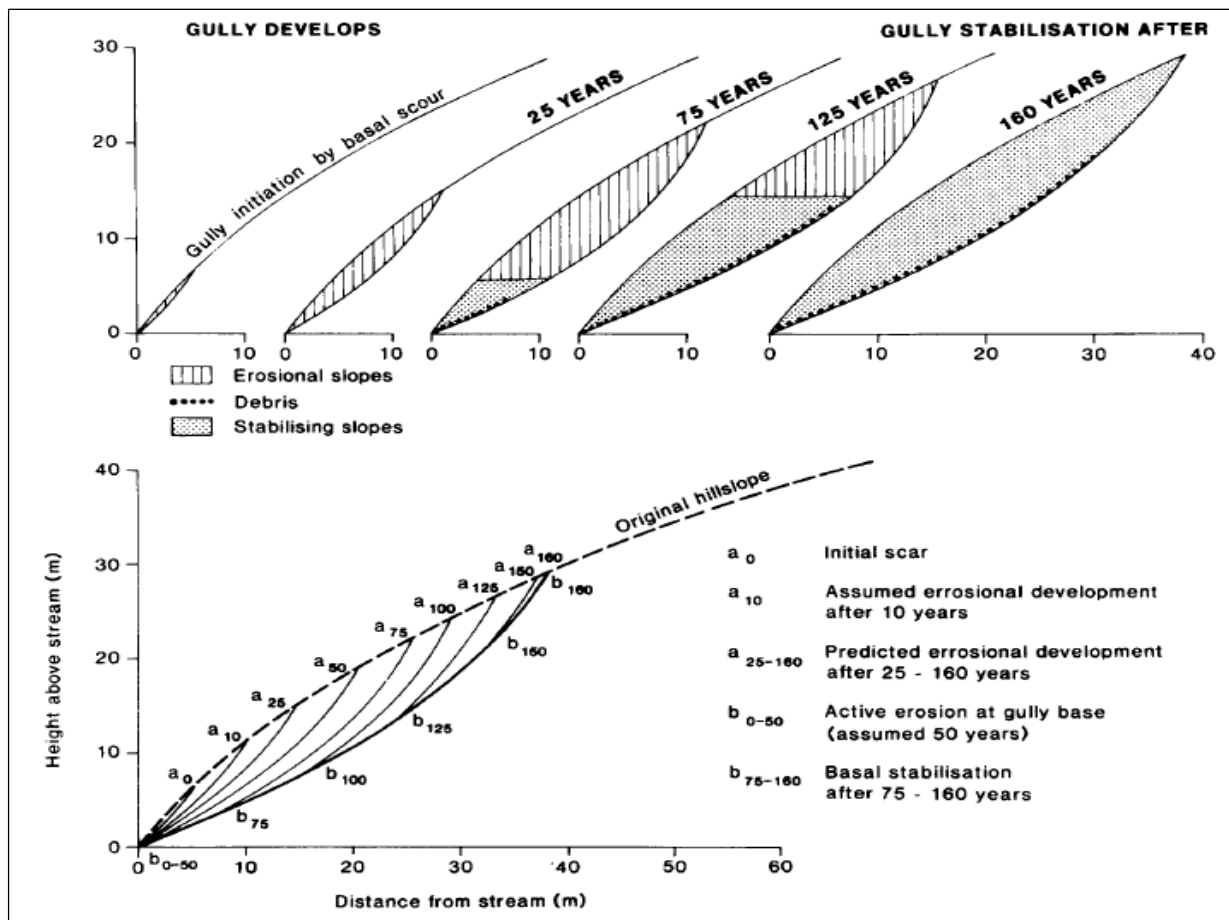


Figure 2.4: Gully development and stabilization as a result of basal scour ((Harvey, 1992); reproduced with permission).

From Figure 2.4 it is evident that the size of the gully is related to its age. Small gullies are thus relatively young, if not spatially or structurally constrained, with large stabilized gullies being 100-175 years old (Harvey, 1992; Kasai et al., 2005). Once gullies are fully developed and revegetated, they become stable and hillslope-channel connectivity is reduced (Harvey, 1992; Kasai et al., 2005; Vanacker et al., 2005).

Where gullies develop, large quantities of sediment are delivered to downstream fans, channels and floodplains (Harvey, 1992; Fryirs and Brierley, 1999; Kakembo and Rowntree, 2003; Pizzuto, 2011). Channels become wider and more unstable as a result of increased sediment load, but will adjust to a narrower, more stable phase once the sediment loads decrease (Harvey, 1997; Kasai et al., 2005). The increased sediment contribution from gullies can form barriers, buffers and blankets further downstream.

Where slopes are badly eroded, soils are thin and less favourable for vegetation establishment. If gullies are present on such slopes, however, they can offer the only suitable conditions that would accommodate the establishment of vegetation. Vanacker et al. (2005) studied sediment dynamics in the Ecuadorian Andes where a human population explosion had a marked effect on forest that was converted to agricultural land. This period was marked by high sediment yields and the formation of erosional features. Two decades later farming was abandoned, as the majority of the population moved to the cities, reducing the farming pressure, after which vegetation established in the gully networks, decreasing connectivity with a subsequent reduction in erosion and sediment supply. A response in the main channel shape was noted following the reduction in farming pressure from a wide braided channel to a narrow entrenched meandering channel. Bed sediment calibre and peak discharge also decreased over time. Analysis of aerial images indicated that land cover had not changed significantly after the abandonment of the agricultural fields. Rainfall patterns for nearby stations suggested that no major changes were observed. The position of revegetation (e.g. in gully bottoms) however, was observed to be the key to limiting the hydrological and sedimentological connectivity. Vegetation establishment in the conduits was effective in disconnecting the landscape units, even where the slopes had not been revegetated. This resulted in the vegetation retaining overland flow and associated sediment through increased roughness, water infiltration and sediment retention. This shows the dual role of gullies, firstly as sediment source and connector and, secondly, as a sediment sink, switching between mechanisms.

2.3.1.5. Accommodation space and valley confinement

Accommodation space and valley confinement determines if and where sediment is likely to be stored (Fryirs, 2002). A reinterpretation of Davis' (1899) model proves useful to understand how accommodation space is influenced by valley confinement. A river is classified in terms of its stage of development, such as youth (erosional), mature (transfer) or old (depositional) phase, in terms of the erosional and depositional cycle. The youthful reach or degradational zone is characterised by a steep river slope, high energy and a deep confined valley that is mostly degrading. Valley down-cutting rates are greater than rates of valley widening, leading to a narrow V-shaped and elongated valley with limited

sediment accommodation space (Fryirs, 2002). The mature reach or transfer zone has a moderate river slope, reduced flow energy and a less confined valley where degradation and aggradation are mostly balanced. Rates of valley down-cutting and widening are balanced for the mature phase, with a wider valley bottom that has moderate accommodation space to store sediment (Fryirs, 2002). For the old age or depositional reach, rates of valley retreat are greater than valley down-cutting, leading to a wide valley floor with ample accommodation space, resulting in a river with a gentle slope, meandering channel and large floodplain (Fryirs, 2002; Fryirs and Brierley, 2013).

Where the river is in a degradational phase, most of its load is transported downstream, but as it reaches the transport and depositional phase, transport capacity is reduced and accommodation space is increased, resulting in sediment deposition forming floodplains. Sediment accommodation space is thus increased from headwater to mouth, except where base level controls, such as escarpments and waterfalls, rejuvenate the erosional cycle (Montgomery and Buffington, 1997; Fryirs, 2002; Tooth et al., 2002; Fryirs and Brierley, 2013). Timescales of sediment storage also increase as accommodation space is increased, with sediment reworked frequently (annually to decadal) in areas with limited accommodation space, compared to areas with ample accommodation space where sediment is reworked over thousands of years (Prosser et al., 1994; Fryirs, 2002; Fryirs et al., 2007a).

Hillslope-channel connectivity also is influenced by valley confinement as the flood plain or valley fill acts as a direct buffer between the hillslope and the river channel (Brierley and Fryirs, 1999; Fryirs and Brierley, 1999; Fryirs et al., 2007a, 2007b; Michaelides et al., 2010). Hillslope-channel connectivity is highest in the source zone where confined valleys and small floodplain pockets rarely disconnect the hillslopes from the channel. As valley confinement decreases, floodplains become more continuous and more prominent hillslope-channel buffers. The buffer capacity of the valley fill or floodplain is directly linked to the width of the buffer (Brierley and Fryirs, 1999), thus a narrow floodplain will have a minor hillslope-channel buffer capacity when compared to a wide floodplain that effectively disconnects the slopes from the channel.

2.3.2. Factors affecting channel-valley fill connectivity

2.3.2.1. Incision and floodplain dynamics

The flood pulse is the main river-floodplain driving force that enables channel-valley fill connectivity and maintains the system's sediment equilibrium (Junk et al., 1989) over decades to centuries. During a flood new sediment will be transported to a floodplain, and some of the sediment in storage will be remobilised and transported downstream (Hooke, 2003). If the system is in equilibrium, the amount of sediment imported and exported would be similar. For the system to remain in this equilibrium state it is thus important that banks are frequently overtopped in order to establish channel-valley fill connectivity and to allow for sediment deposition. During a flood event in a meandering system with cohesive materials

(large clay content), the outer bank is eroded, while the inner bank is rebuilt through the formation of lateral or scroll bars (Wolman and Miller, 1960; Page et al., 2003). This continues over time, with the channel slowly migrating across the floodplain. In this equilibrium state the eroded and newly formed banks are at the same elevation (Wolman and Miller, 1960) with little changes to longitudinal connectivity. In a system with coarser material, lateral migration is accomplished through channel avulsion. Poole et al. (2002) found that paleo channels are reactivated in this process, scouring sediment from paleo channels and depositing new sediments in the abandoned channel. Where the system with coarser material is in equilibrium, the active and paleo channels are at the same elevation (Poole et al., 2002).

Where the equilibrium of sediment input or output is disturbed, the valley bottom is transformed with implications for channel-valley fill and longitudinal connectivity. Valley bottoms store sediments until sediments are reworked and is often referred to as cut and fill cycles (Prosser et al., 1994; Fryirs and Brierley, 2001; Fryirs, 2002; Kasai et al., 2005; Brierley et al., 2006). The fill cycles can dominate for thousands of years, slowly building sediment stores up over time, until a cut phase is initiated, often lasting decades to centuries and is seen as a pulsed sequence where sediment is built up and subsequently released (Prosser et al., 1994; Fryirs and Brierley, 2001; Fryirs, 2002). Sediment is stored along the valley fill during frequent contributions from the channel to the valley fill, reducing longitudinal connectivity or the sediment delivery ratio (Fryirs and Brierley, 2001).

Where sediment builds up in the valley bottoms it forms large water aquifers, which promote plant growth. These productive sinks, often swamps with deep soils and anabranching channels, were targeted by early European settlers, for grazing and agriculture (Brierley and Murn, 1997; Brierley and Fryirs, 1998; Walter and Merritts, 2008). In many cases vegetation was removed and furrows created to drain the swamps, which increased the longitudinal connectivity of the system, resulting in the rapid export of sediments, cutting down into unconsolidated sediments. In catchments such as those in New South Wales, Australia, where grassy vegetation was removed for agriculture, the main river system was incised up to 10 m deep and 100 m wide in some places, over a few decades, reducing channel-valley fill connectivity (Brierley and Murn, 1997; Brierley and Fryirs, 1998). In the transfer and depositional zone downstream of the Cobargo and Upper Wolumla Creek study sites (NSW), exported sediment choked the system by forming a thick sediment blanket (burying fence posts) on the floodplain, reducing vertical connectivity. Where valley bottoms receive increased sediment loads from the slopes, the channels often aggrade following high sediment concentrations, becoming wider and shallower (Kasai et al., 2005). This shows that the valley bottoms are sensitive areas where the stability threshold can easily be crossed by a change in land cover or land use leading to a pulsed sediment flux where the valley bottoms switch from fill to cut cycles or vice versa (Fryirs and Brierley, 2001; Brierley et al., 2006).

Where the equilibrium is affected to the extent that export dominates import and thus relative stream power is greater than the sediment load, a net reduction or cut phase will result and sediment will be lost

from the channel bed and margins (Simon and Rinaldi, 2006). This lowering or incision of the channel into the floodplain will scour the bed to a lower elevation. Where the channel migrates or switches to another location it will cut laterally relative to the newly incised bed elevation and the new floodplain will be rebuilt, abandoning the former floodplain to become a terrace (Womack and Schumm, 1977; Bookhagen et al., 2006). This new floodplain will be developed according to the new flood regime, re-establishing frequent channel-valley fill connectivity. As the channel migrates laterally over time, terraces will be formed on both sides of the river. Where further incision happens, another set of terraces will develop. Where lateral erosion cuts into an old terrace, the terrace will be removed, often resulting in unpaired terraces (Womack and Schumm, 1977). In most river systems, terraces are attributes which contain important information on changes in channel-valley fill connectivity that happened in the past (Womack and Schumm, 1977). Where channels are incised, banks will be higher and steeper, often leading to bank erosion and bank collapse, increasing the availability of sediment for downstream transport (Brierley and Murn, 1997; Simon and Rinaldi, 2006).

Incision is caused by changes of extrinsic controls such as base level, sediment input and discharge (Womack and Schumm, 1977; Maddy, 1997; White and Greer, 2006). In areas near the continental shelf, sea level will drop during colder periods (ice ages), promoting incision into deposited sediments (Maddy, 1997). Further inland, base level change can happen as a result of tectonic movement, breaching of a bedrock barrier or more local changes, such as meander cut-offs (Maddy, 1997; Harvey, 2002). This will steepen the channel gradient and lead to incision that migrates upstream, resembling a headcut (Harvey, 2002).

Climatic changes lead to dry and wet periods, resulting in cut and fill cycles (Bookhagen et al., 2006). These cut and fill cycles are caused by differing amounts of runoff (energy) and sediment production. In the north-western Himalayas, Bookhagen et al. (2006) looked at terrace formation in the Sutlej Valley. Valley infill occurred at the beginning of the Holocene when moisture availability increased and sediments were readily available. This increased the sediment flux and caused deposition where river gradient declined. Several cut and fill terraces formed after the main infill, and are linked to drier periods during the Holocene. The incision or terrace formation was attributed to reduced sediment flux during drier periods (transport limited). This will be followed by a wetter period where sediment flux is increased and a new fill level is created. Similar cut and fill cycles have been recorded in South Africa and were related to the major climatic changes during the Holocene (Temme et al., 2008; Lyons et al., 2013; Tooth et al., 2013). Cooler periods were linked to sediment being made available, with redistribution and deposition occurring during the onset of wetter periods. Where available sediment is depleted, e.g. as a result of increased vegetation cover during warmer wetter conditions, incision could be initiated and could continue throughout the wetter periods (Temme et al., 2008).

A reduction in sediment input or an increase in discharge can trigger incision. Anthropogenic activities, such as sediment mining or water storage in reservoirs can reduce the sediment input, effectively changing

the sediment equilibrium of the valley fill to such an extent that incision is initiated (Nakamura and Tockner, 2004). In Italy, sand mining has resulted in incisions of up to 8.5 m deep in cases (Surian et al., 2009). In other cases, the amount of runoff is increased through farming practices or land cover changes. In Douglas Creek, Colorado, USA, Womack and Schumm (1977) found that the introduction of cattle resulted in valley fill incision. White and Greer (2006), in a study in California, linked incision to an increase in runoff as a result of urbanisation and associated increases in impervious surfaces. Reduced sediment input and/or increased discharge can lead to valley fill incision.

Incision or a cut phase can thus be a result of geologic, geomorphologic, climatic, hydrologic, livestock and anthropogenic influences or a combination of these influences (Schumm, 1999). All these factors function in different ways, but can all be related to increased or excess transport energy relative to the availability of sediment (Simon and Rinaldi, 2006).

Where channels are incised, the channel capacity is increased relative to the discharge, resulting in reduced inundation of the floodplain (Gergel et al., 2002; Kondolf, 2006; Pizzuto, 2011). Hydraulic geometry shows that increased channel capacity will reduce the chances of overbank flooding (Leopold and Maddock, 1953). This will lead to an enlarged channel, containing larger magnitude peak flows, and reducing the frequency of channel-valley fill connectivity and chances to deposit sediment on the valley fill (Simon and Rinaldi, 2006). Where tributaries are not buffered by a floodplain, incision will work its way up the tributary until a barrier prevents upstream propagation of the incision (Fryirs and Brierley, 1999). Incised systems will thus increase sediment transport through enhanced drainage features, becoming more efficient conduits and therefore providing less sediment storage on flood plains.

2.4. Modelling and measuring connectivity

Connectivity can provide a qualitative framework, based on maps for example, indicating connectivity and (dis)connectivity, but connectivity is more difficult to quantify (Heckmann and Schwanghart, 2013) and is much needed for future geomorphic enquiry (Wohl, 2014). Quantification is not a simple task as connectivity is influenced by landscape properties and processes that vary both spatially and temporally, such as event magnitude, ground cover, topography and antecedent soil moisture properties (Harvey, 2001; Michaelides and Wainwright, 2002; Medeiros et al., 2010). Measurements can be data intensive, especially if high resolution measurements are used. Several efforts have been made to develop methods to quantify connectivity.

Borselli et al. (2008) developed a GIS based model, called the Index of Connectivity (IC), which calculates the probability of connectivity for a cell based on the upslope and downslope characteristics and contributions. Calculations were based on slope, land use, potential sediment availability (linked to land use), distance from source to sink, river and road network. Cavalli et al. (2013) applied the IC model to debris flows in alpine catchments in order to predict if sediment from hillslopes will reach a drainage

channel. The IC model proved useful in highlighting contributing areas and can be used to predict scenario-based connectivity. Various other attempts were made to quantify connectivity in order to assess how water and sediment movement will be affected by likely scenarios. This has been done using sediment and hydrological models such as the IC (Borselli et al., 2008; López-Vicente et al., 2013), a two dimensional hydrological model (Michaelides and Wainwright, 2002), WASA-SED (Model of Water Availability in Semi-Arid Environment with a Sediment Dynamics Component) (Medeiros et al., 2010), SWAT (Soil and Water Assessment Tool) (Le Roux et al., 2013), RMMF (Revised Morgan, Morgan and Finney) (López-Vicente et al., 2013) and graph theory embedded in GIS (Heckmann and Schwanghart, 2013).

These models often make use of DEMs and satellite imagery to categorise the landscape and are thus limited by the available resolution. Connectivity models are complex, and due to processing limitations, scale has to be adjusted according to the relevant extent of the catchment. The models produce results that are in agreement with field observations proving that their quantification of connectivity was realistic, or that the model somehow produced the right output although connectivity has been quantified incorrectly. There is a need to further the development of geostatistical methodologies that will enable connectivity to become a powerful, universally quantifiable characteristic that is comparable between vastly contrasting landscapes over different spatial and temporal scales (Michaelides and Chappell, 2009).

2.5. Conclusion

The connectivity concept and framework proves to be a valuable holistic approach to assess sediment dynamics as it addresses both structural and functional changes over time. The concept can be applied over various spatial scales, ranging from landscape unit to catchment scale (Brierley et al., 2006). Brierley et al.'s (2006) nested hierarchical framework provides an assessment of landscape connectivity that includes the appraisal of landscape features that could act as buffers, barriers, blankets or boosters at various scales (Fryirs et al., 2007a).

Connectivity focuses on the ease at which material can move from one place to the next. Sediment transport is largely dependent on water movement and thus, hillslope-channel connectivity can be influenced by drivers such as slope gradient, land cover, drainage pathways and valley confinement, whereas channel-valley fill connectivity can be influenced by channel incision. Steep slopes, sparse vegetation cover, high drainage density and narrow valley fills increase hillslope-channel connectivity. Incised channels lead to increased channel capacity, limiting flooding of the valley fill and reduce the associated potential for sediment deposition. Furthermore, all the drivers are affected by the magnitude and frequency of rainfall events, where large infrequent events breach most buffers, barriers and blankets. This influence of event magnitude makes it challenging to assess when linkages become active sediment conductors.

Following the review of the literature, regardless of the shortcomings, the concept of connectivity may prove to be a useful framework for a sediment study in the Thina River as limited work has been done in South Africa using this concept. By using this connectivity framework, a catchment-wide understanding of sediment transfer processes can be developed and integrated into a channel reach scale understanding of sediment transfer and deposition. This Thina River study will focus on hillslope-channel connectivity and channel-valley fill connectivity. Hillslope-channel connectivity has been studied in several settings by researchers worldwide and through various models, but channel-valley fill connectivity has largely been neglected in the current literature. This study aims to develop the current understanding of the present-day and historical sediment dynamics in these high altitude, high rainfall watersheds at a catchment and channel reach scale, and to improve the channel-valley fill connectivity concept.

Chapter 3: Study area

3.1. Introduction

The Vuvu catchment, covering an area of 65 km² (sub-catchment of quaternary catchment T34A) is located in the north-eastern Eastern Cape or former Transkei (called Transkei henceforth), South Africa, and drains the Drakensberg Escarpment towards the eastern-seaboard (Figure 3.1). The uppermost divide of the catchment forms the border between South Africa and Lesotho. The catchment is located in the mountainous foothills of the southern Drakensberg, also known as the Great Escarpment geomorphic province (Partridge et al., 2010). The Drakensberg runs northeast to southwest with long narrow spurs extending from the escarpment, creating many ridges and plateaus separated by steep valleys (De Decker, 1981; Mucina et al., 2006). The Vuvu catchment is elongated and drains from west (2920 masl) to east (1450 masl).

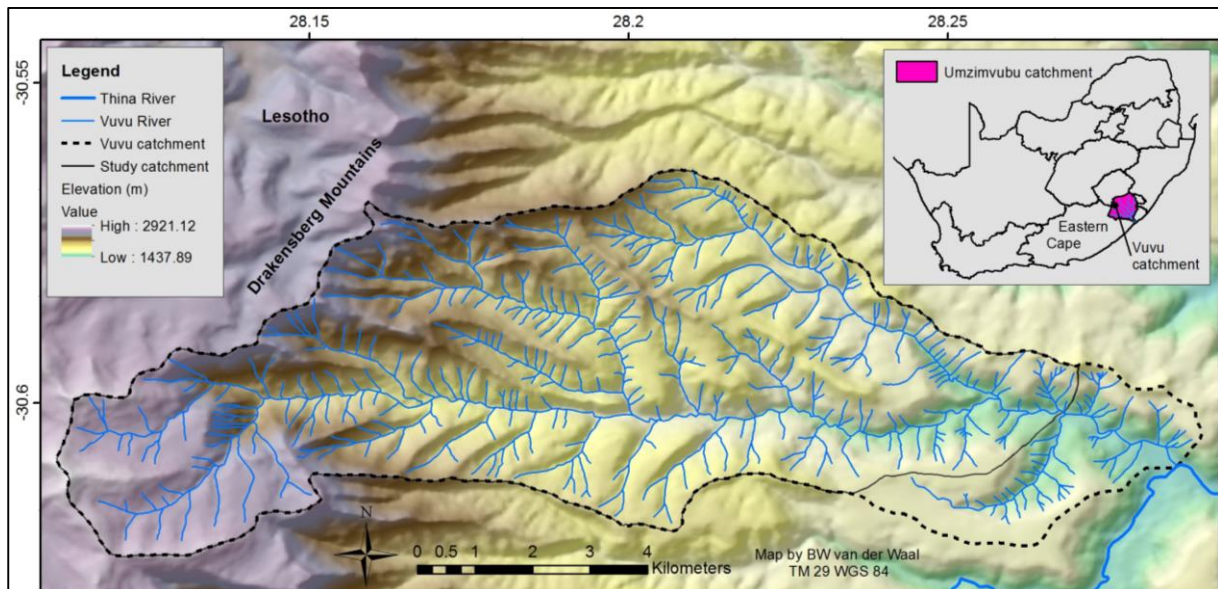


Figure 3.1: The topography and drainage network of the entire Vuvu catchment and the portion that was used as study catchment. The location of the catchment in relation to South Africa and the larger Umzimvubu catchment is indicated on the inset map.

The Vuvu River drains into the Thina River which is one of the main tributaries of the Umzimvubu River. The Umzimvubu River drains into the Indian Ocean. The Vuvu system has a high density dendritic drainage network (Figure 3.1). The long profile of rivers of the Great Escarpment geomorphic province are steep and irregular or stepped where harder layers are crossed (Partridge et al., 2010). Due to various uplift events coupled with river rejuvenation since the breakup of Gondwanaland (Mesozoic), head-ward erosion has dominated over lateral erosion, resulting in steep deep valleys with limited sediment storage or accommodation space (Partridge and Maud, 1987). A narrow valley fill exists along the lower section of the main valley. The rivers of the Transkei are incised and in a phase of degradation, cutting down into previously deposited valley sediments or bedrock (Dardis et al., 1988). This valley aggradation and

subsequent incision could be related to longer term climatic changes. Evidence from the Drakensberg Escarpment in KwaZulu-Natal suggests that localised valley aggradation occurred since the Last Glacial Maximum (~10 000-5 000 BP) (Temme et al., 2008), possibly due to surplus sediment made available during the glacial period and limited vegetation cover on hillslopes. Increases in vegetation cover since 5 000BP possibly reduced the sediment supply from hillslopes and led to valley degradation and incision. This was possibly followed by smaller localised cut and fill cycles throughout the region.

The upper Thina River is classified as a threatened river ecosystem due to land degradation within the catchment (Driver et al., 2005).

3.2. Hydrology

Rainfall varies from 707–928 mm per year and increases with elevation (Mucina et al., 2006), up to 2100 masl; whereafter, rainfall and rainfall intensity decreases (Nel et al., 2010). Precipitation peaks in summer, in the form of high intensity thunderstorms (Nel, 2008), with snow during winter (Mucina et al., 2006). Evaporation peaks in summer (December to February) and can be as high as 1628 mm per year (Mucina et al., 2006). Although evaporation exceeds rainfall, the area has a mean annual runoff of 184–209 mm per year (WR2005; Middleton and Bailey, 2011). Flow is perennial and peaks in summer. The catchment is undammed and thus has a fairly natural flow regime, with the impact of overgrazing, burning and consequent erosion being responsible for any changes to the hydrology.

Schulze and Horan (2007) modelled the hydrological effects of the current grazing and fire and related degradation in the Thina catchment using the ACRU model (Schulze, 1995). Their model showed that the current state of degradation increased the overall annual accumulated flow. Modelled baseflow decreased by 4-14% and stormflow increased by 12-44%. The average current sediment yield is estimated to be 98% higher than what would be expected as a result of natural erosion. These figures indicate that the current hydrology has a more flashy nature, increasing peak overland runoff volumes and are thus more erosive due to the greater energy available. This is highlighted by the doubling of the sediment yield under the current degraded conditions (Schulze and Horan, 2007).

3.3. Geology

The Vuvu catchment is underlain by the Karoo sequence that consists of the Drakensberg Formation (basalt) on the surface of the Clarens Formation (sandstone) with the Elliot Formation (mudstone) beneath and alluvium in the valley bottoms (Figure 3.2) (De Decker, 1981). Structurally the layers have remained stable since their deposition and the general dip of the layers varies from 2–7° to the west. A detailed explanation of each of the formations as given by De Decker (1981) follows below:

- The Drakensberg formation forms the escarpment and consists of a basalt cap that is up to 1200 m thick. The basalt was formed through a series of basalt flows of different thicknesses and varies from hard resistant crystalline rock (thick flows) to softer vesicular forms (thinner flows) that provide its stratified appearance. The basal portion can have intercalated layers of Clarens sandstones. Many fissures existed in the landscape as can be seen by the many dolerite dykes outcropping in the landscape. Most are northwest-southeast aligned and stand out as prominent fishback ridges.
- Well sorted, fine grained, feldspathic sandstones of pale orange colour and aolean origin dominate the Clarens Formation. Minor mudstone layers can be found between the sandstone layers. The formation is a maximum of 275 m thick and forms impressive cliffs, often with caves at the base.
- The Elliot formation is dominated by red and purple mudstones, with subordinate layers of medium grained feldspathic sandstone that are not laterally persistent. Upward fining sequences (sand to mudstone) are ever-present; this formation most likely was deposited by meandering streams. The red colours of the mudstones are indicative of oxidized iron that are associated with the existence of drier conditions during their formation. This layer can be up to 370 m thick.
- Quaternary deposits are present in the form of colluvial and fluvial deposits in valley bottoms. The soils are dark with sandstone and basalt rubble on the fringes.

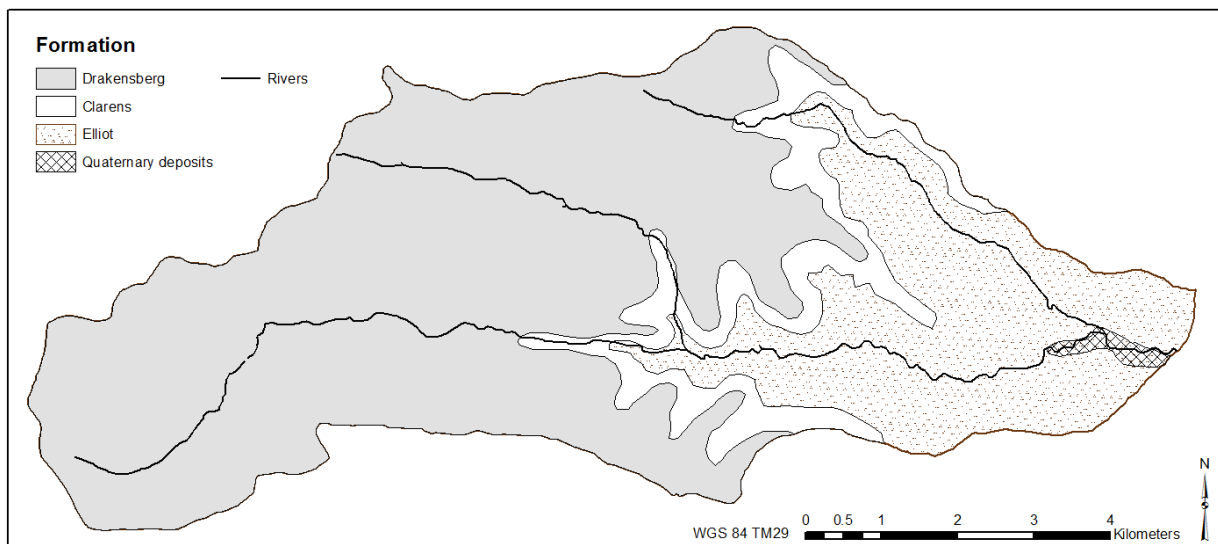


Figure 3.2: The geology of the Vuvu catchment (De Decker, 1981).

The stepped nature of the landscape is related to the presence of harder layers (e.g., sandstones, that are more resistant to erosion). The softer layers, such as mudstone, are removed by erosion, exposing the harder layers to form the steps in the landscape.

3.4. Vegetation

Grassland dominates the vegetation and varies between Lesotho Highland Basalt Grassland (1900–2900 masl), Southern Drakensberg Highland Grassland (1720–1900 masl), East Griqualand Grassland (920–1720 masl) and small pockets of Mistbelt Forest in ravines (Figure 3.3) (Mucina et al., 2006).

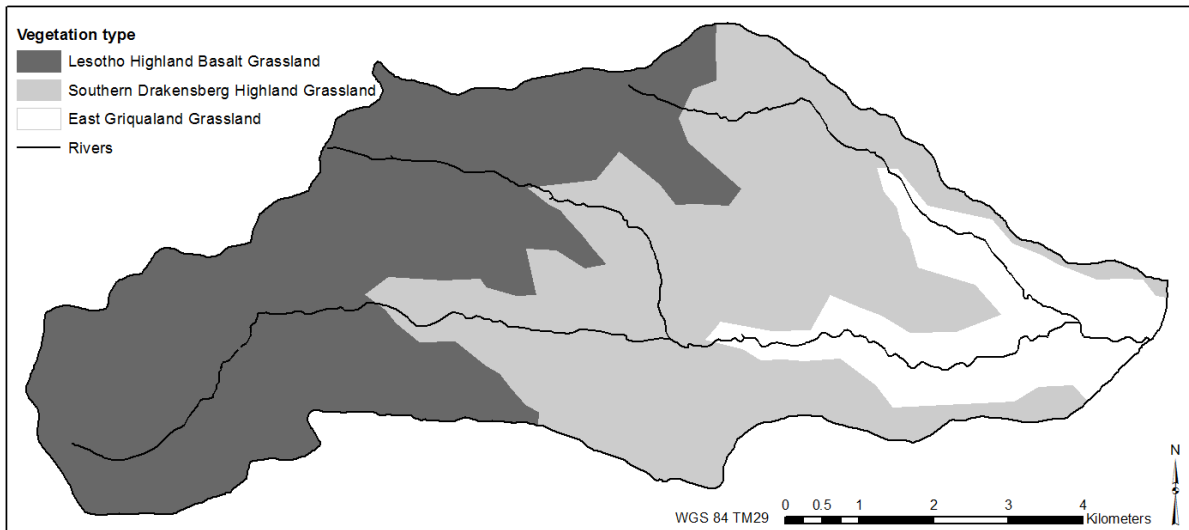


Figure 3.3: Vegetation types of the study area as per Mucina et al.'s (2006) classification.

These vegetation types and associated environmental characteristics are given by Mucina et al. (2006) as:

- The Lesotho Highland Basalt Grassland can be described as short grassveld with patches of shrubland (*Leucosidea sericea*), normally related to disturbance. This vegetation type is underlain by basalt and soils are basalt derived with almost equal parts of coarse sand, fine sand, silt, clay and organic material. The organic material is a result of slow decaying grass roots, retaining a lot of precipitation and releasing it slowly in the form of seeps. Mean monthly minimum and maximum temperatures range between -10.5 and 31.4°C, with frost occurring throughout winter.
- Southern Drakensberg Highland Grassland is found on steep slopes, where dense tussock grassland dominates with a tall shrubby component (*Leucosidea sericea*) on wet slopes. Soils are deep and fine grained. It is characterised by a cool temperate climate (mean temperature of 13°C) with 30–90 days of frost per year that increase with increasing altitude.
- East Griqualand Grassland is found on hilly country, with grassland being dominant and *Leucosidea sericea* growing on wet slopes. Associated soils are derived from mudstones and sandstones, about 500–800 mm deep, well drained with a high clay content. Mean annual temperature is 14.7°C, with frost occurring roughly 30 days a year.

3.5. Soils and erosion potential

Fey et al. (2010) gave a detailed overview of the soil groups that are found in the study area. Soil groups are dominated by lithic soils which form on convex crests and steep slopes where soils are removed at a similar pace to that which occurs during bedrock weathering, thus resulting in shallow, rocky soils. Melanic soils are relatively common on basalt and have a well-developed blocky structure, making them effective at retaining water and less susceptible to erosion. Humic, oxidic and duplex soils are common on the gentle slopes of lower lying areas. Humic and oxidic soils have a good structure and are well drained, making the materials relatively resistant to erosion; however duplex soils have a high sodium content and poor structure, resulting in highly dispersive and erodible soils. Soils of lesser dominance are plinthic, gleyic and cumulic soils, normally found in wetland or water accumulation areas. Plinthic soils are poorly drained which results in iron accumulation. Gleyic soils are poorly structured, wet most of the year, and thus anaerobic and depleted of colloidal material. Cumulic soils are unconsolidated as they have recently been deposited. Soil depth varies from 20 cm on steep slopes up to 6 m on gentle slopes (maximum depth of gullies).

Accelerated soil erosion, through features such as gullies, soil pipes, sheet and rill erosion, is common throughout the Transkei (Beckedahl et al., 1988; Beckedahl and Dardis, 1988). This erosion is a combination of intrinsic soil (particle composition, sodium concentration, etc.) and climatic properties and also is induced by external factors, such as poor farming practices and artificial drainage (Beckedahl et al., 1988; Dardis and Beckedahl, 1988).

An erosion risk map (Figure 3.4), that combines rainfall erosivity, topography, soil erodibility and landcover shows that the lower parts of the Vuvu catchment are highly susceptible to erosion (Le Roux et al., 2008; Msadala et al., 2010). This high erosion risk is caused by the duplex soils that form on the mudstones (Le Roux and Sumner, 2012). The erosion risk map suggests that north-facing slopes are more vulnerable to erosion. Huber (2013) studied gully development on mudstones in the upper Thina catchment. Gullies were mainly found on lower angled slopes (<20%), with the largest features found on the more gentle slopes with deep soils. Most gullies were continuous and active, only showing some stabilization where floodplains or bedrock acts as a base level control. The majority of the gullies were estimated to have been initiated around 1900 (Huber, 2013).

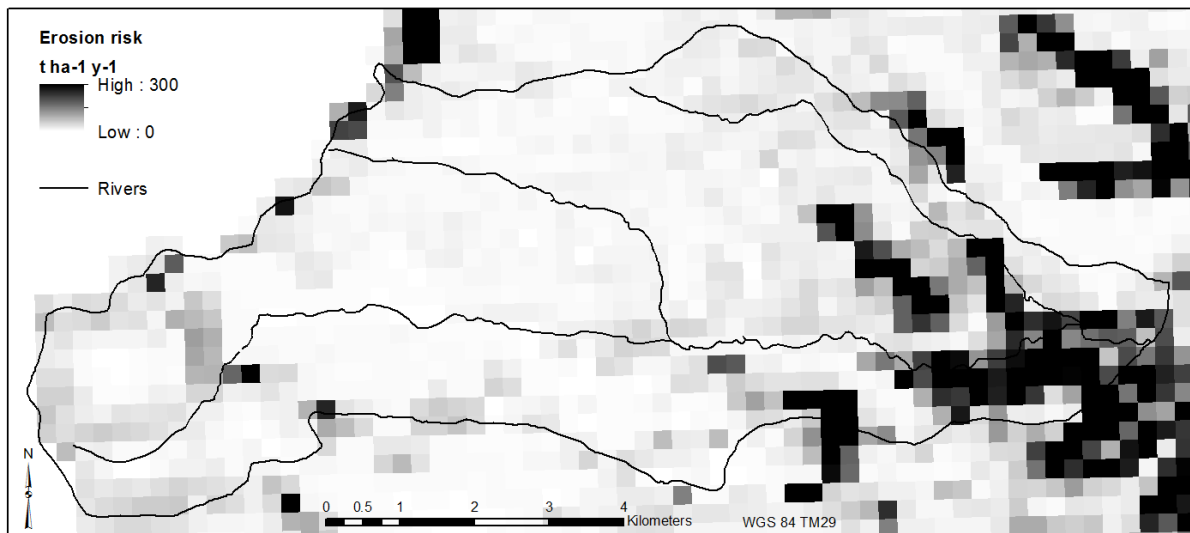


Figure 3.4: An erosion risk index (Le Roux et al., 2008) for the Vuvu area showing very high erosion potential for the lower catchment.

3.6. Land use

The Vuvu catchment is entirely located in a communal area, previously known as the Transkei homeland. Land is mainly used for livestock grazing with small scale cultivation near homesteads. Land along the Transkei Drakensberg Escarpment is managed as a communal asset, allowing livestock to graze freely (Guillarmod, 1969). Communal lands have been under pressure for the last 150 years since settlement (O'Connor, 2005). As grazing pressure increased as a result of population growth, animals are forced to higher elevations for grazing during summer months. This often led to overgrazing as rotational grazing was seldom practised (Guillarmod, 1969). The reduction in arable pasture forces animals to graze in bogs for prolonged periods of time. The duration of grazing as well as trampling make these bogs susceptible to erosion (Guillarmod, 1969). Fires often are used to bring on a flush of green growth in winter or spring, but once the new growth has been depleted, animals return to the bogs for grazing, leaving both the slopes and bogs devoid of vegetation cover and exposed to early spring erosive rain (Guillarmod, 1969).

Historical aerial photos for the Vuvu catchment dating back to 1952 show that most fields and houses on the Basalt Formation (upper catchment) were abandoned before 1952. Since 1952 crop production has been focused on low angled areas on the Elliot Formation (lower catchment), with south-facing fields showing signs of abandonment and rill erosion. Very few houses were present in 1952 and were mainly located on steeper slopes away from the prime fields. Livestock tracks were already well developed with extensive areas of sheet and gully erosion visible throughout the catchment. In the 1960s a road was developed connecting the upper Thina catchment with Mount Fletcher, via the Vuvu catchment, and soil conservation contours were improved. The increase in the number of houses that were constructed on large fields suggests a population influx in the 1960s. Smaller fields were developed around the homesteads. In the 1970s south-facing slopes on the Elliot Formation were actively cultivated. The number of huts visible on aerial images was estimated to have trebled between the 1960s and 1970s,

suggesting a great increase in population, and the new huts were positioned in clusters to form villages. The main road was upgraded by the 1990s and the number of huts was estimated to have increased by a third since the 1970s and most fields further away from villages were abandoned. In the 2000s veld fires are evident on the images, with agriculture limited to easily accessible fields. From 2000 onward there was no visible increase in the number of huts. The population is currently relatively stable for the larger Elundi Municipality with a population growth figure of 0.05% per annum (<http://www.localgovernment.co.za/locals/view/28/Elundini-Local-Municipality>; accessed November 2014). Erosional features increased in size throughout the sequence of aerial images.

3.7. Conclusion

The Vuvu catchment is characterised by a high relief landscape with a well-developed drainage network. Sediment storage along the valley floor is limited due to the confined nature of the valley fill. The upper catchment is underlain by basalt that forms soils that have a good structure and are relatively resistant to erosion. The lower catchment is underlain by mudstones that form erodible duplex soils. The high intensity rainfall, coupled with overgrazing and frequent burning, makes these soils prone to erosion, as is evident from well-developed erosional features throughout the catchment. The erosion and veld degradation has increased water and sediment transfer from the slopes to the river channel. The steep nature of the landscape coupled with erodible soils and high intensity rainfall would make gully and sheet erosion rehabilitation efforts challenging, especially under ongoing land use pressures.

Chapter 4: Sediment sources and pathways in the upper Thina catchment

4.1. Introduction

A river's physical character and functioning is a product of local physiognomies and water and sediment input from the catchment (Rosgen, 1996). Sediment input from the catchment is dependent on the quantity available and the necessary pathways or links to convey the sediment to the river (Owens, 2005). This routing of sediment is dependent on connectivity. Connectivity can be used both to describe the physical structure of the landscape and the movement of water, sediment and other materials (or organisms) through the landscape. Connectivity is therefore highly applicable to studies of sediment dynamics at the catchment scale. A well-connected system enables the flow of energy and materials and, as a result, mutual adjustment between system components (Harvey, 1997; Brierley et al., 2006; Farraj and Harvey, 2010).

Connectivity is counterbalanced by storage sites, such as buffers, barriers and blankets, that allow material to be retained in the system (Fryirs et al., 2007a). Different types of connectivity exist: hillslope-channel connectivity focuses on how sediment from the hillslope moves to the river channel; longitudinal connectivity focuses on sediment moving down the river network, lateral connectivity with a focus on sediment in the channel moving laterally on to the valley fill, vertical connectivity focuses on linkages between the channel and sediment substratum and temporal connectivity which considers changes in linkages over time (Ward, 1989; Harvey, 1997; Brierley et al., 2006). More details on connectivity and its related drivers are discussed at length in the literature review (Chapter 2).

From field observations it was noted that the Vuvu landscape was characterised by large erosive features and high sediment concentrations during rain events. The connectivity concept was adopted as a framework to investigate how landscape connectivity has changed. For this thesis there are two focus components of landscape connectivity, the first being hillslope-channel connectivity, and the second being channel-valley fill connectivity of the Vuvu catchment (Figure 1.1). In this chapter the hillslope-channel connectivity is addressed. The focus objectives are as follows:

1. Classify, map and characterise potential sediment sources on hillslopes.
2. Map and characterise present-day hillslope-valley fill connectivity.

The methods for assessing hillslope-channel connectivity, using mapping, are discussed below. Sediment source tracing, which forms part of hillslope-channel connectivity, is discussed in Chapter 6.

4.2. Methods

This section describes the methods that were used to meet the objectives set for the hillslope-channel connectivity component of the thesis. The methods were based largely on a desktop study using a Geographical Information System (GIS), with field visits to ground truth a selection of the features. High resolution images were used to identify recent sediment sources and the pathways (gullies) that connect the source areas to the channel (Owens, 2005). Characteristics of the source areas and pathways were derived from aerial photo interpretation and terrain analysis (based on 20 m contours) (Kheir et al., 2007). Historical aerial images were used to assess whether the gullies are recent features (Harvey, 1992; Huber, 2013). Details of the methods used are given below.

4.2.1. Objective 1: Classify, map and characterise potential sediment sources on hillslopes.

4.2.1.1. Classification and mapping

Colour aerial images from 2009 (0.5 m resolution; acquired from Chief Directorate: National Geo-spatial Information, South Africa) were used in ArcInfo 10.0 to digitize active hillslope sediment sources at a scale of 1:2000 for the entire Vuvu catchment. Several field visits were conducted (since 2010) before the mapping started, which aided the process of photo interpretation. Geo-tagged land based photos taken during field trips also helped with aerial image interpretation. The aerial images were colour adjusted to optimize the contrast between the various erosional features. The best results were achieved by using a RGB (Red Green Blue) composite that had a stretch value of 2.5 standard deviations. The brightness was adjusted to negative 10%. Automated extraction of bare areas was attempted as an objective method to identify or verify erosional features. GeoEye satellite images (2 m resolution) were used to calculate Normalized Difference Vegetation Index (NDVI) values in the hope that this index could be used to extract bare soil and areas with active surface erosion (Kakembo et al., 2006). The frequent occurrence of large boulders and bare rock made the method impractical as NDVI values were similar for rock surfaces and bare soil (Le Roux and Sumner, 2012). Thus, the data had to be captured manually.

The main sediment sources identified in the study area were fields (agricultural), gullies, sheet erosion, landslides, roads (gravel) and livestock tracks (Plate 4.1-4.3). Rectangular or trapezoidal areas with linear edges and vegetation varying on the inside from that on the outside were captured as agricultural fields (Plate 4.1). Fields that have not been cultivated recently (with plough lines visible) and lacked fencing were categorised as abandoned. Based on field observations the fields proved to be lower than or sunken into the surrounding unploughed landscape, suggesting that significant volumes of soil has been eroded from these features (Plate 4.1). Gullies were considered large (>2 m or 3 to 4 pixels wide), incised, linear erosional features with steep sidewalls that concentrate flow mainly in a down-slope direction (Plate 4.1 and 4.2).

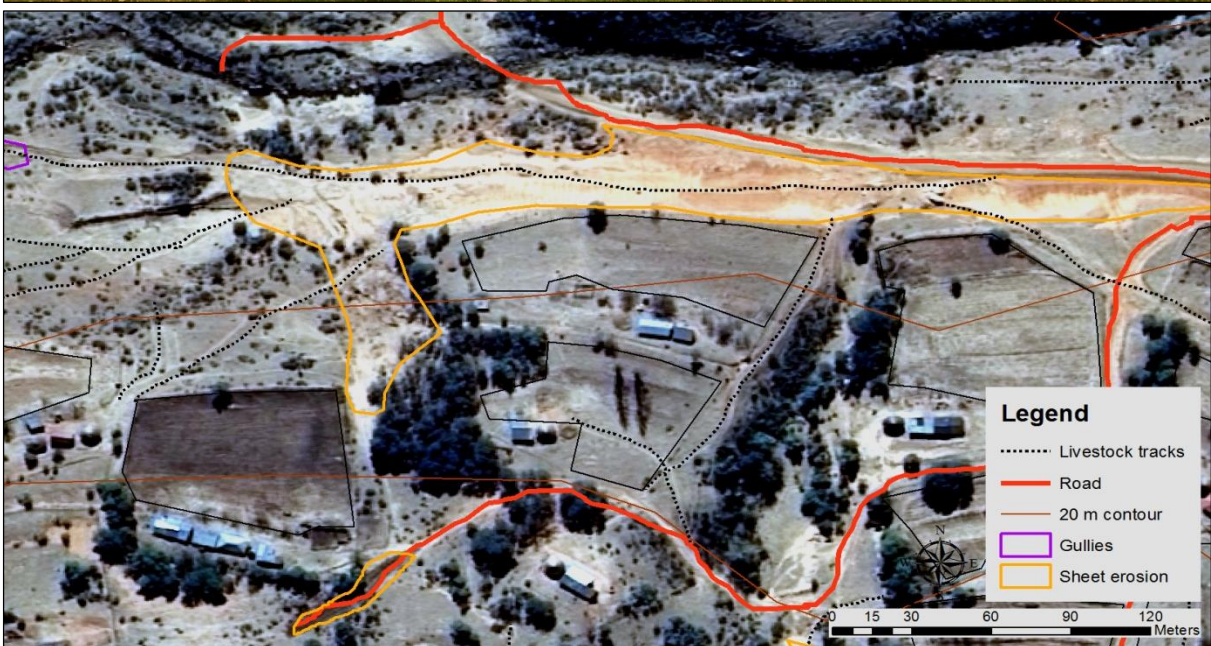
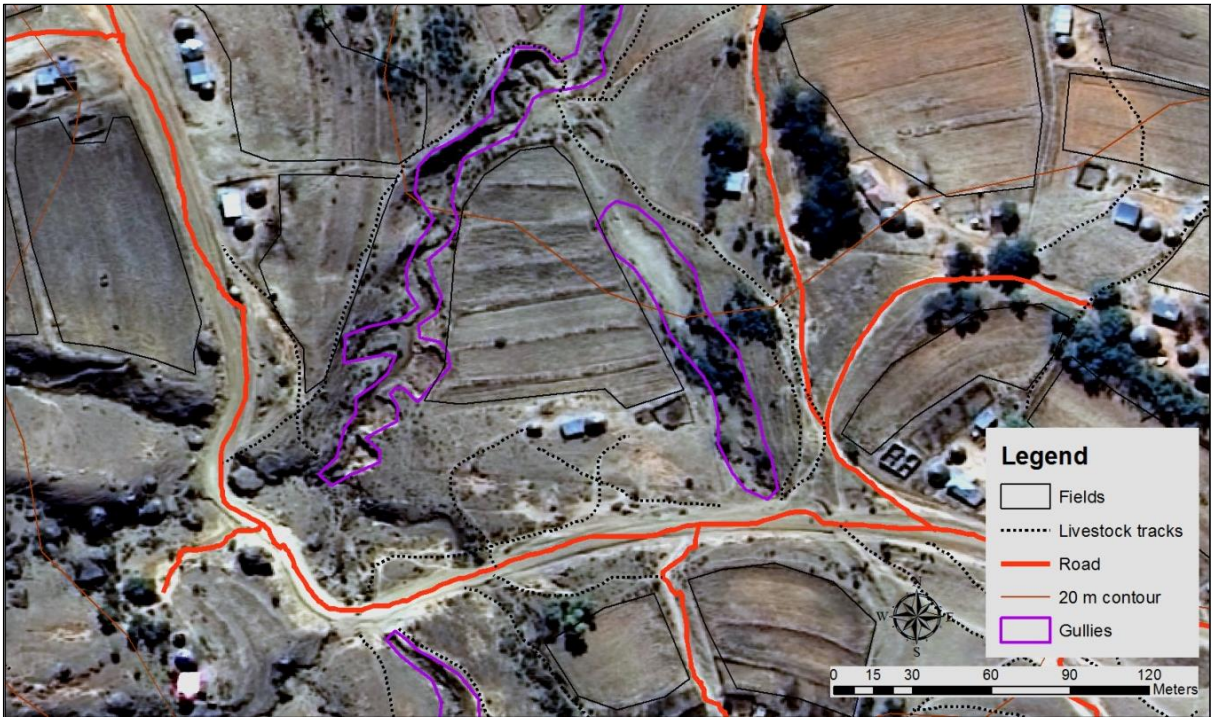


Plate 4.1: Top to bottom: map of fields, livestock tracks, roads and gullies; abandoned field, note the difference in elevation and vegetation of the field and areas surrounding the field; map of livestock tracks, roads and sheet erosion.

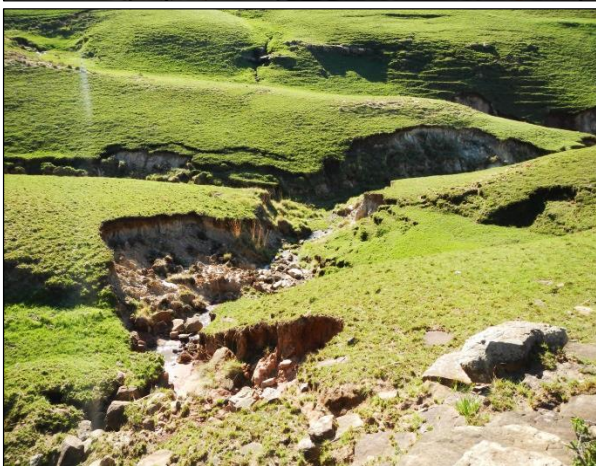
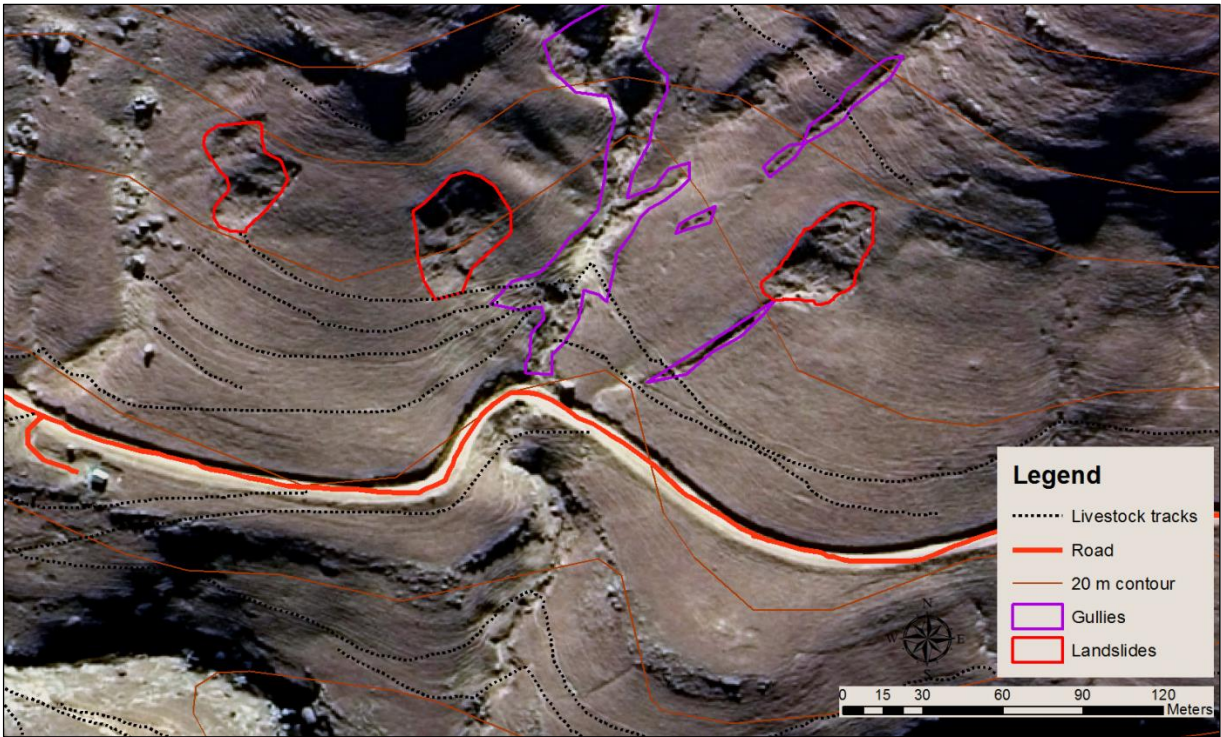


Plate 4.2: From left to right: first row: sheet erosion near cattle dip tank and Vuvu River; second row: map of livestock tracks, roads, gullies and landslides; third row: gully erosion and landslides.

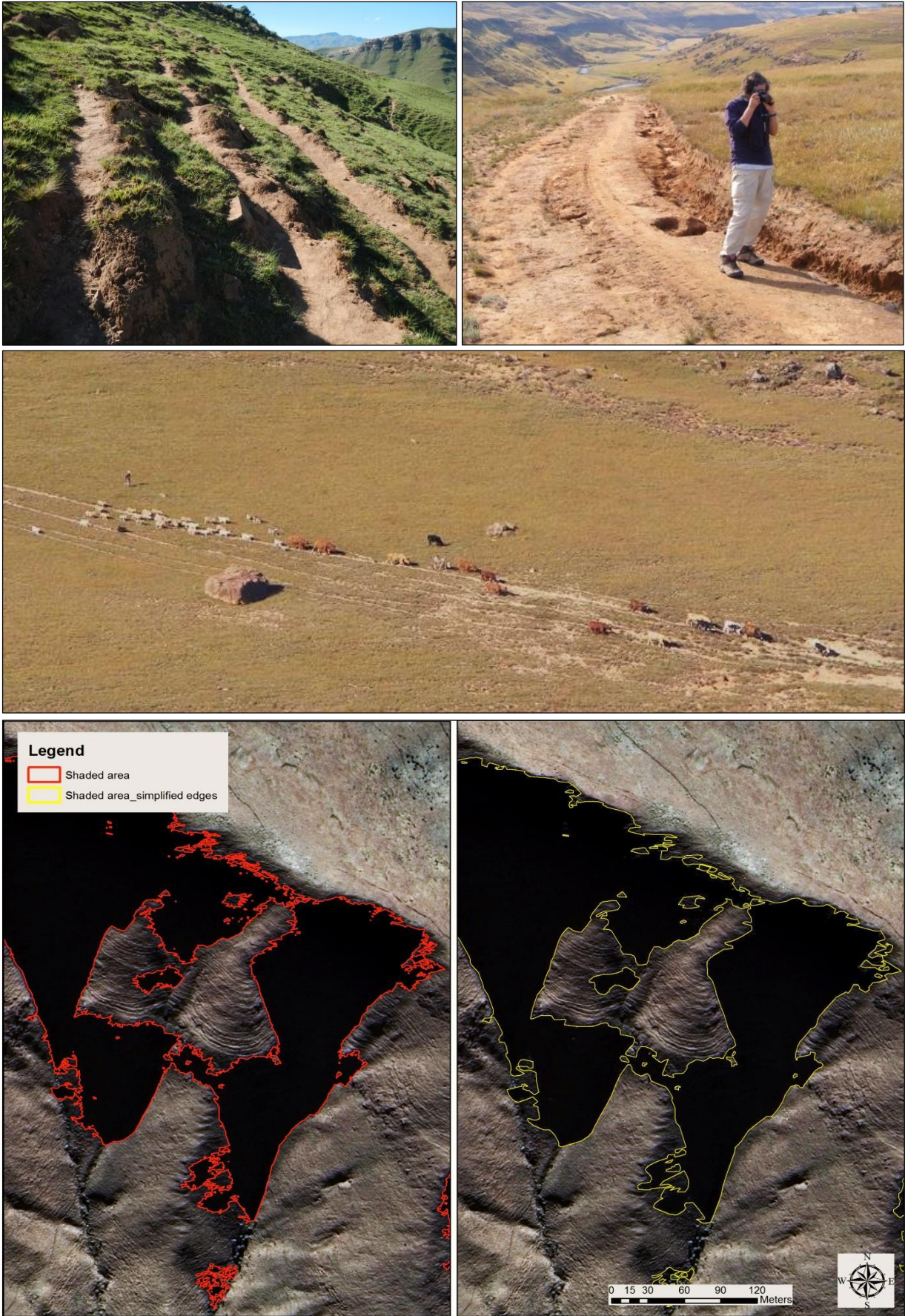


Plate 4.3: Top: livestock tracks and a road constructed in 2011 showing signs of erosion (photo date April 2012); middle: livestock using the livestock tracks; bottom: 2009 colour aerial image showing shaded areas (left) and simplified edges of shaded areas (right) on south-facing slopes.

Areas of sheet erosion were identified as shallow, ill-defined patches with bare soil that often also incorporated areas of badland or rill erosion as defined by Foster et al. (2007) (Plate 4.1 and 4.2). Gully and sheet erosion features were mapped on fields as these features were significantly deeper than the relatively thin and even lowering associated with fields (thus seen as additional erosion on fields). Recent relatively active landslides (not covered by vegetation) were digitized, which included features with a clear rounded upper rim and lower rock-waste line, where the material was deposited (Plate 4.2). Dirt roads were easily identifiable as they were larger features (>2 m wide) of a constant width, followed the contour and linked homes to the main dirt road that traverses the lower catchment (Plate 4.1-4.3). Livestock tracks were observed in the field to be narrow (<2 m wide) shallow linear features, mainly following the contours (crossing slopes) or ridgelines (Plate 4.1-4.3). Only sections that were clearly visible were digitized. Fields, gullies, sheet erosion and landslides were captured as polygons (Plate 4.1 and 4.2). Roads and livestock tracks were digitized as line features (Plate 4.1 and 4.2). In order to calculate area for roads and livestock tracks, lengths were multiplied by the average width of 4.04 m (standard deviation of 1.82 m; n = 24) and 0.79 m (standard deviation of 0.45 m; n = 21) respectively. These widths were thought to be reasonable averages based on field and air photo observations for the Vuvu catchment.

Due to the steep slopes and position of the sun (northeast) at the time of the 2009 aerial photography, a portion of the steep southwest-facing slopes was shaded and made the identification of erosional features impossible (Plate 4.3). This led to an aspect bias in the analysis. In order to identify and remove shaded areas from the analysis, the colour images were converted to greyscale images (255 values). Shaded areas were extracted from the aerial image using a greyscale raster extraction for cell values below 60 (very dark areas). The output was converted to a 'shaded area' polygon feature that was used to erase any intersecting parts of digitized features (in other shapefiles) and to correct for area calculations. This 'shaded area' shapefile included areas where black soils were exposed on north-facing slopes, but these were removed manually from the 'shaded area' shapefile so that the areas with black soils would be included in the subsequent feature analysis. Shaded areas less than 250 m² were also deleted and polygons with jagged edges were simplified to preserve the essential shape of the features using the bend option (reference baseline of 5 m) of the Simplify tool (Plate 4.3).

4.2.1.2. Topographic and geological characteristics of source areas

Statistics that described the topographic character of the catchment, each of the geological provinces (regions) and the various source types (in terms of a catchment as a whole and the individual geological provinces) were calculated using ArcInfo 10.0 GIS. Statistics included: area of features, percentage of total area (catchment or geological province), percentage that was shaded area, slope (Le Roux and Sumner, 2012), aspect (Le Roux and Sumner, 2012), connectedness to the river network, percentage of fields that were recently cultivated and percentage of features located on fields.

The catchment boundary was established based on 20 m contours (national dataset) and the underlying geology was digitized from a 1:250 000 geological map (De Decker, 1981). A 20 m DEM was created using the Interpolate tool and the 20 m contour data. A slope and aspect raster was generated from the 20 m DEM using the Surface toolset. In order to minimize the number of slope classes, only three were identified: gentle ($<5^\circ$), moderate ($5\text{--}20^\circ$) and steep ($>20^\circ$) to reflect the stepped nature of the landscape. The steep ($>20^\circ$) category represents the very steep sections, often including vertical cliffs; the moderate ($5\text{--}20^\circ$) category represents lower talus slopes and more gentle ($<5^\circ$) areas above the steep sections; the gentle ($<5^\circ$) category represents the near flat areas found between the ‘moderate’ ($5\text{--}20^\circ$) sections on hill crests or along the valley fill adjacent to the rivers. The frequency distribution for the aspect raster had two distinct peaks, one at 32° and the other at 175° . These two groupings were split to give north ($284\text{--}109^\circ$) and south ($110\text{--}283^\circ$) facing slopes. This led to north-facing slopes having a slightly wider (187°) portion than south-facing slopes (173°), but this division reflected what was observed in terms of differences in vegetation in the field, and evidence from aerial images.

Slope and aspect data were extracted for source polygons using the Extract by Mask tool. Other attributes, such as geology, connectedness and location relative to a field, were assigned to each polygon using the Intersect tool. Connectedness was assigned based on whether a feature was intersected by a continuous gully or a drainage line. The same was done for features that were located on fields. Databases were exported to Microsoft Access and the relevant data copied to Microsoft Excel for further interpretation.

Volumetric estimates of sediment loss were made for each of the source types. A single depth reading was taken for each feature visited in the field at a point that was representative of the average depth of the feature during sediment source sampling. Based on these field measurements, the average depth for an erosion type (e.g., gullies) was calculated. GIS based area data were used in conjunction with the field based depth measurements (gullies, sheet erosion, landslides, roads and livestock tracks) to calculate the approximate volume of soil lost from the various features. The volume contributed by the various features was converted to mass using a bulk density of 1.55 g cm^{-3} (Flemming and Hay, 1984; Brady and Weil, 2008). The likely time of existence of the features was estimated based on the extent of the various features on the earliest aerial images (1956) as described in section 2.1.3. It was assumed from aerial photo evidence that the majority of fields, gullies, sheet erosion and livestock tracks were all initiated at more or less the same time (± 50 years ago), whereas roads (± 30 years) and landslides (± 10 years) were more recent or short-lived features. Landslides were only active for a relatively short period as they were stabilised by vegetation several years after formation. These data allowed for a sediment loss ($\text{t ha}^{-1}\text{ y}^{-1}$) to be calculated for each of the features.

4.2.1.3. Historical change in gully area and current stability

In order to assess the evolution of gullies in the Vuvu catchment, they were mapped from aerial photos for 1956 (scale 1:30 000), 1975 (scale 1:50 000) and 2009 (georeferenced ortho photos, 0.5 m resolution). Photos with 1956 and 1975 dates were used as they were of reasonable definition (despite the smaller scale of the 1975 photographs) and allowed for a relatively equally spaced temporal succession of photos. The 1956 and 1975 photos had to be georeferenced as described below. Due to the small scale and variation in image quality across the different years, only the larger gullies (2009 length >50 m) were suitable to be accurately digitized (Martínez-Casasnovas, 2003). Nine gullies were identified as being suitable for analysis. The aerial photos for the three dates were used to map the change in gully area over time. By plotting the data as a time series it was possible to assess to what extent gullies were recent features. A logarithmic trendline was fitted to the data to predict the potential initiation date of the features (Harvey, 1992; Huber, 2013).

Historical aerial photos were georeferenced using the adjust transformation in ArcInfo. Large boulders and rocky outcrops were used as georeferencing control points as they were commonly found throughout most of the landscape, easily identifiable and proved to be stable over the 53 year period. Care was taken to use as many control points as possible, especially around the targeted gully features. Between 50-100 control points were used per image, minimising the root mean square error to below 0.01.

Visual field assessments of a stratified random selection of gullies sampled for sediment tracing (Chapter 6) were made to determine if the gullies were active (showing signs of expansion) or relatively stable (well vegetated with minimal signs of expansion). The number of gullies that were assessed was limited by the steep and often challenging terrain.

4.2.2. Objective 2: Map present-day hillslope-valley fill coupling.

Linear drainage pathways that were identified in the Vuvu catchment included: natural drainage lines, continuous gullies, discontinuous gullies, roads and livestock tracks. Natural drainage lines ranged from small, steep, grassy topographic lows to the well-developed main trunk channel. Unchannelled sections of the natural drainage network with dense vegetation cover were captured as part of the drainage network, but were specified as not connected as it was assumed that the well vegetated sections would act as sediment buffers. Channelled parts of the natural drainage network were specified as major drainage lines as these had the potential to act as sediment pathways. Gullies (>2 m or 3 to 4 pixels wide linear features) were digitized as line features and were classified as continuous or discontinuous, based on whether the feature was directly linked to the major drainage network or not (Brierley et al., 2006; Le Roux and Sumner, 2012). Gullies that had more than 20 m of vegetated slope below the outflow of the feature were classified as discontinuous. From field evidence it seemed that a vegetated strip of more than 20 m was an effective sediment buffer as there was evidence of sediment build-up on these vegetated

areas below the gully outflow. Where livestock tracks were close together or parallel (<5 m apart), only the most prominent track was digitized to prevent duplication, as water would functionally be routed in the same direction. Livestock tracks that occurred within a 5 m radius of a gully were removed (Clip tool) to prevent further functional duplication.

In order to extract slope and aspect data from the raster datasets for the linear features, the raster layers (slope and aspect) had to be converted to polygons containing integer values. The geology, slope and aspect data could then be assigned to the individual sections of line features by using the Intersect tool. The connectivity of a pathway feature to the drainage network was assigned by intersecting a feature with the continuous gully dataset (5 m tolerance). Lengths of the various line features were calculated for the Vuvu catchment and each of the geological provinces. Density was calculated by dividing the length of the features by the area for which the line features were extracted. The increase in drainage density as a result of anthropogenically induced change was expressed as a density (m ha^{-1}) and percentage of the natural drainage network.

4.3. Results

The following sections present the GIS and field based statistics for the catchment, potential sediment sources and pathways. Due to the large area and high definition of the mapped source areas and connectors, the data could not be displayed on a single map in a meaningful way. Small scale maps showing the location of the various potential sediment sources and pathways are given in the text, but larger versions and shapefile data have been included in Appendix 1 in digital format.

4.3.1. Catchment characteristics

The largest part (65%) of the Vuvu study area (58 km^2) is underlain by the Drakensberg Formation, followed by the Elliot (25%) and Clarens (10%) Formations (Figure 4.1 and Table 4.1). The stepped nature of the landscape can be deduced from Figure 4.1, with the main steps being formed by the escarpment in the upper reaches, a second by the Clarens Formation and a third by a sandstone layer within the mudstones of the Elliot Formation. Moderate (43%) to steep (48%) slopes dominate the catchment, with a larger proportion of gentle and moderate slopes on the Elliot Formation and steeper slopes on the Clarens and Drakensberg Formations (Figure 4.1a and Table 4.1). This shift of steeper slopes for the higher lying areas reflects the expected trend along the river profile where the headwaters are on steeper slopes compared to the more gentle landscape of the lower reaches of the river. The areas with steeper slopes often had shaded south-facing slopes on the 2009 aerial images. The majority (61%) of the slopes are north-facing, with very similar north-facing proportions for the various geological

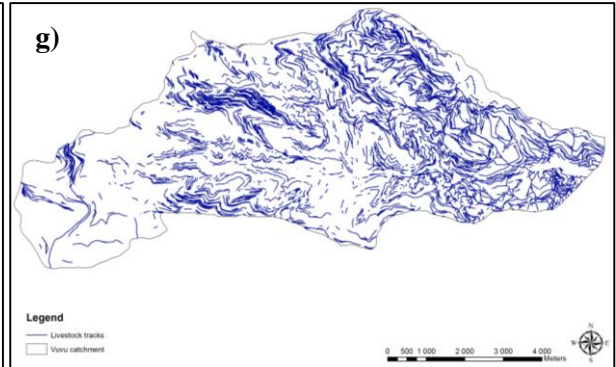
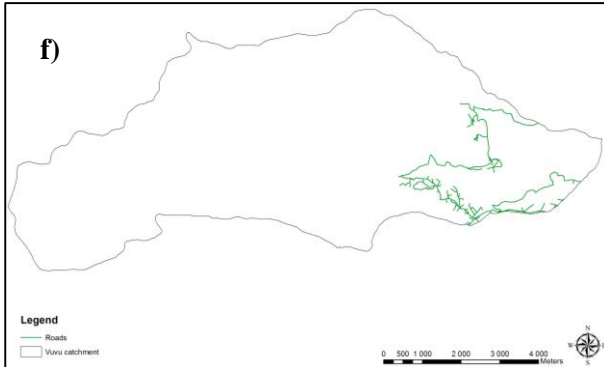
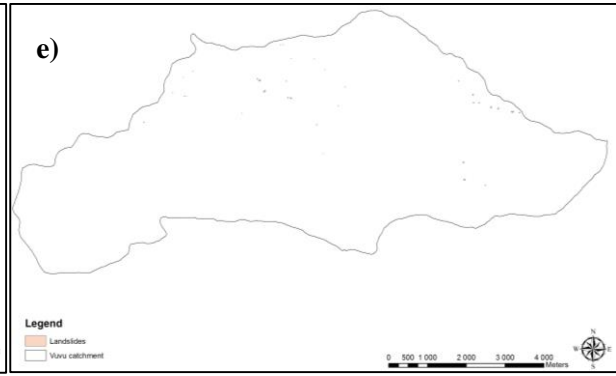
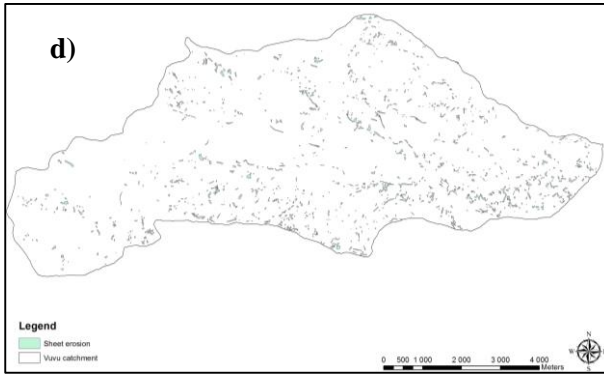
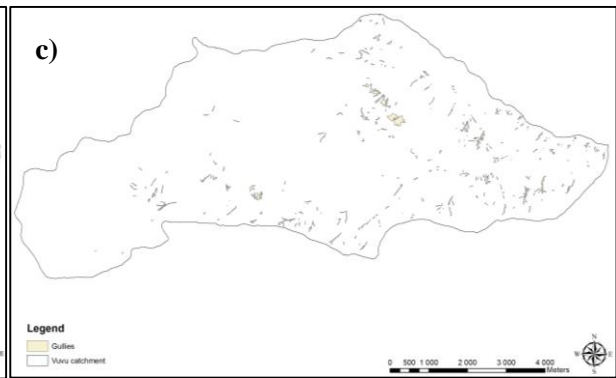
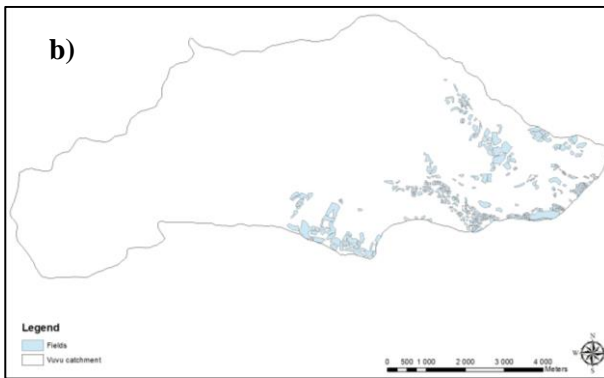
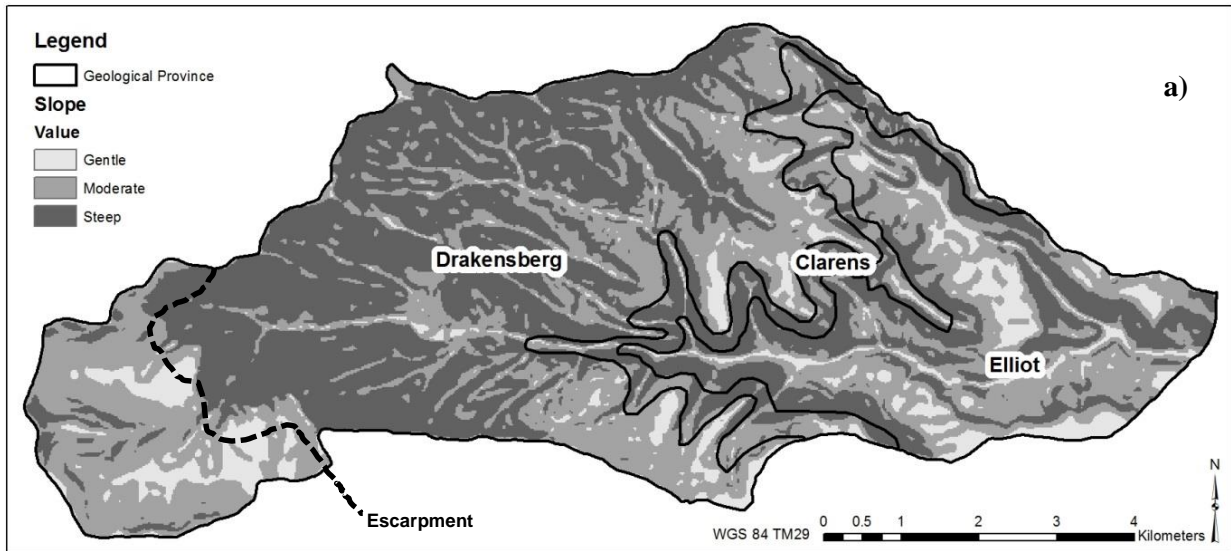


Figure 4.1: a) A slope map indicating gentle ($<5^\circ$), moderate ($5\text{--}20^\circ$) and steep ($>20^\circ$) slopes for the Vuvu catchment. The geological provinces are indicated. Potential sediment source maps of the Vuvu catchment are displayed indicating: b) fields, c) gullies, d) sheet erosion features, e) landslides, f) roads and g) livestock tracks. Maps b to g are included electronically in Appendix 1.

provinces (Table 4.1). The section above the escarpment has a very different topography, with mainly gentle to moderate slopes.

Table 4.1: Area, slope and aspect for the Vuvu catchment.

	Area	Slope			Shaded	Aspect
	ha	% gentle	% moderate	% steep	%	% north-facing
Total	5329	9	43	48	8	62
Elliot	1355	15	46	39	5	62
Clarens	529	4	41	55	10	63
Drakensberg	3444	8	42	50	9	62

4.3.2. Topographic and geological characteristics of potential source areas

4.3.2.1. Current topographic and geological characteristics of potential source areas

The topographic characteristics for fields, gully erosion, sheet erosion, landslides, roads and livestock tracks are given in this section (Figure 4.1b-g). The results are based on area calculations and exclude the bias of landscape distribution (e.g. aspect).

Previously and currently used (2009 images) cultivated fields covered less than 5% of the total catchment area and were mainly identified on the Elliot (69%) and Drakensberg (30%) Formations (Figure 4.1b). The Clarens Formation has a negligible (<0.5%) area that was classified as cultivated (Table 4.2). Based on the area calculations for each geological province, 13% of the Elliot Formation and 2% of the Drakensberg Formation were cultivated. Almost half of the area of fields was connected to the drainage network, with the proportion of fields connected being the highest (71%) on the Drakensberg formation. Fields were mainly identified on moderate (61%) and gentle (37%) slopes, with a small proportion (<2%) identified on steep slopes. A larger proportion of fields were identified on moderate slopes on the Drakensberg and Clarens Formations. The fields were mainly (80%) north-facing, with a smaller proportion (74%) facing north on the Elliot Formation. The majority (78%) of the fields had been abandoned to such an extent that only fields near homesteads were still being cultivated, all on the Elliot Formation.

Table 4.2: Summary statistics for the characteristics of fields and the proportion that was actively cultivated in 2009.

	Area	% of province	Connected	Slope			Aspect	Active
				% gentle	% moderate	% steep	% north-facing	% of area
	ha		% of area					
Total	258	4.8	46	37	61	2	80	23
Elliot	177	13.1	36	42	56	2	74	33
Clarens	2	0.4	0	25	68	7	90	0
Drakensberg	79	2.3	71	26	74	<1	93	0

Gullies covered less than 1% of the total catchment area, with the majority (62%) identified on the Drakensberg Formation (Table 4.3; Figure 4.1c), due largely to its greater area. A smaller proportion (35%) was identified on the Elliot Formation and the remaining 3% located on the Clarens Formation. The Elliot Formation had the highest density of gullies (1.3% of the area), with lower gully densities on the Drakensberg (0.9%) and Clarens (0.3%) Formations. More than 70% of the gully area was connected, with an increase in connectedness for the formations at higher altitude. Gullies (catchment-wide) were infrequently (<12%) identified on previously cultivated fields, especially on the Drakensberg and Clarens Formations. On the Elliot Formation, however, gullies were more frequently (24%) formed on previously cultivated fields. Gullies were mainly (67%) identified on moderate slopes, with an increasing proportion identified on steeper slopes at higher elevations. North-facing slopes had the highest percentage of gullies (>60%), especially on the Clarens Formation (>94%).

Table 4.3: Statistics regarding the location of gullies in terms of geology, connectivity and location with respect to fields.

	Area ha	% of province	Connected % of area	On fields % of area	Slope			Aspect
					% gentle	% moderate	% steep	% north- facing
Catchment	50	0.9	70	12	10	67	23	65
Elliot	17	1.3	55	24	23	57	20	61
Clarens	2	0.3	63	0	6	77	18	94
Drakensberg	31	0.9	80	5	2	73	26	65

Sheet erosion was identified on ca. 2% of the total catchment area (Table 4.4; Figure 4.1d). The largest percentage was identified on the Drakensberg Formation (57%), followed by Elliot (37%) and Clarens (5%) Formations. The density of sheet erosion, however, was highest on the Elliot Formation (2.9%), followed by the Drakensberg (1.8%) and Clarens (1%) Formations. Only 16% of the total area of features was connected, with the Drakensberg Formation (18%) having the highest percentage of connected features. A small proportion (8%) of the sheet erosion was identified on previously cultivated fields. The Elliot formation had the highest proportion (11%) of sheet erosion in relation to previously cultivated fields, compared to lower proportions (6-7%) on the Clarens and Drakensberg Formations. Sheet erosion was mainly (56%) identified on moderate and steep slopes (36%). A similar pattern to gully erosion was observed for sheet erosion, where steeper slopes at higher altitudes were more prone to erosion. North-facing slopes were more susceptible (>72%) to sheet erosion, especially on the Elliot and Clarens Formations (>81%).

Table 4.4: Summary statistics for sheet erosion's topographic characteristics.

	Area ha	% of province	Connected % of features	Fields % of features	Slope			Aspect
					% gentle	% moderate	% steep	% north- facing
Catchment	106	2.0	16	8	8	56	36	77
Elliot	40	2.9	15	11	6	63	31	82
Clarens	6	1.0	12	6	5	47	48	81
Drakensberg	61	1.8	18	7	9	53	39	72

Landslides occupied a small proportion (0.03%) of the catchment, with features most commonly (54%) identified on the Elliot Formation (Table 4.5; Figure 4.1e). None of the features were connected, with features mainly (78%) identified on steep slopes, especially on the Drakensberg and Clarens Formations (94–100%). None of the features were identified on gentle slopes, with a larger proportion (37%) identified on moderate slopes on the Elliot Formation. All landslides were located on south-facing slopes that are generally wetter, a necessary condition for landsliding.

Table 4.5: Summary statistics for landslides in relation to connectivity, fields, slope and aspect.

	Area ha	% of province	Connected % of area	On fields	Slope			Aspect
				% of area	% gentle	% moderate	% steep	% north- facing
Catchment	1.3	0.03	0	0	0	22	78	0
Elliot	0.7	0.05	0	0	0	37	63	0
Clarens	0.0	0.00	0	0	0	0	100	0
Drakensberg	0.6	0.02	0	0	0	6	94	0

Roads covered <1% of the catchment area, as the 29km length had an average width of ca. 4.04 m (Table 4.6; Figure 4.1f). Livestock tracks (ca. 0.79 m wide) had a total length of 637 km, which equated to <1% of the catchment area (Table 4.6; Figure 4.1g). It was estimated that more than half of these roads and livestock tracks discharged directly into the drainage network.

In Table 4.6 it can be seen that the Elliot Formation consistently has the highest density of all sediment source features. Erosion features are most commonly found on moderate slopes that are north-facing. More than 9% of the catchment has been exposed to intense erosion, but only 45% of the eroded area is connected to the drainage network.

When estimates for soil loss are considered (Table 4.7a), it can be seen that gully erosion has been the most prominent sediment source (ca. 3.9 t ha⁻¹ y⁻¹) followed by sheet erosion and fields (ca. 3.6 and 3.5 t ha⁻¹ y⁻¹). Landslides, roads and livestock tracks contributed much less (ca. 0.5–0.7 t ha⁻¹ y⁻¹). Although these sediment loss figures are approximate estimates and the time of activity is uncertain, it can be seen

that the gully, field and sheet erosion features are the main anthropogenically induced sediment sources for the Vuvu catchment.

Table 4.6: A summary table for all the mapped source features at a catchment scale.

Feature	% cover	Formation with highest density	Dominant slope	Dominant aspect	% of area connected
Fields	5	Elliot	5 - 20	North	50
Gully erosion	1	Elliot	5 - 20	North	70
Sheet erosion	2	Elliot	5 - 20	North	16
Landslides	0.03	Elliot	>20	South	0
Roads	0.22	Elliot	5 - 20	North	>50
Livestock tracks	0.94	Elliot	>5	North	>50

Table 4.7: a) Volumetric estimates and likely average soil loss for catchment-wide source features. Sample size for depth measurements is given in brackets. b) Volumetric and soil loss estimates for each of the geological provinces. Volume was calculated by multiplying area by depth. Soil loss was calculated by dividing volume by years active.

a)	Feature	Area (ha)	Depth (m±SD)	Volume (m ³)	Years active	Soil loss (t ha ⁻¹ y ⁻¹)
	Fields (n = 30)	258	0.24 ± 0.08	607 936	50	3.5±1.2
	Gully erosion (n = 46)	50	1.37 ± 0.90	678 679	50	3.9±2.6
	Sheet erosion (n = 36)	106	0.58 ± 0.19	618 037	50	3.6±1.2
	Landslides (n = 11)	1	1.72 ± 0.38	22 968	10	0.7±0.2
	Roads (n = 24)	12	0.47 ± 0.42	55 554	30	0.5±0.7
	Livestock tracks (n = 21)	50	0.20 ± 0.10	98 542	50	0.6±0.6
	Total	477		2 081 716		12.9±6.5

b)	Feature	Volume (m ³)			Soil loss (t ha ⁻¹ y ⁻¹)		
		Elliot	Clarens	Drakensberg	Elliot	Clarens	Drakensberg
	Fields	417 235	5 282	185 419	9.5±3.2	0.3±0.1	1.7±0.6
	Gully erosion	235 360	22 417	420 902	5.4±3.5	1.3±0.9	3.8±2.5
	Sheet erosion	232 228	32 165	353 644	5.3±1.7	1.9±0.6	3.2±1
	Landslides	12 370	83	10 515	1.4±0.3	0.0	0.5±0.1
	Dirt roads	55 554	0	0	2.1±2.8	0.0	0.0
	Livestock tracks	35 116	9 076	56 439	0.8±0.9	0.5±0.5	0.5±0.5
	Total	987 863	69 023	1 026 919	24.6±12.4	4.1±2.1	9.6±4.7

When soil loss is divided into the various geological provinces, it can be seen that the Elliot and Drakensberg Formations have contributed similar volumes of sediment (ca. 1 000 000 m³ respectively) despite the area of the Drakensberg Formation being ca. 2.6 times bigger than that of the Elliot Formation. These differences are reflected in the total soil loss estimates of ca. 25 t ha⁻¹ y⁻¹ for the Elliot Formation and ca. 10 t ha⁻¹ y⁻¹ for the Drakensberg Formation. The Clarens Formation was a less significant sediment source with a soil loss estimate of ca. 4 t ha⁻¹ y⁻¹. All erosional features on the Elliot Formation consistently had the highest soil loss values. Fields had the highest soil loss value of ca. 10 t ha⁻¹ y⁻¹, followed by gully and sheet erosion (ca. 5 t ha⁻¹ y⁻¹ respectively) on the Elliot Formation. Roads stood out as a more

significant sediment source on the Elliot Formation contributing ca. $2 \text{ t ha}^{-1} \text{ y}^{-1}$, as no roads were found on the other geological formations. Landslides and livestock tracks were less significant sediment sources, with soil loss rates of ca. $1 \text{ t ha}^{-1} \text{ y}^{-1}$ respectively. These soil loss estimates highlight the erodible nature of the Elliot Formation.

4.3.2.2. Historical change in gully area and current level of stability

The location of the nine gullies used in the historical aerial photo analysis is shown in Figure 4.2. A time sequence of aerial images is shown in Figure 4.3, which indicates the development of one of the gully features (B5) in the Vuvu catchment. Results for the time sequence analysis show that gully features in the upper Thina (Phiri-e-Ntso and Vuvu) catchment are still expanding in size (Figure 4.4). The majority of the smaller gullies started post 1950s, whereas the larger features were initiated from the 1920s to the 1950s. Similar results were found for gullies in the adjacent Phiri-e-Ntso catchment (Huber 2013).

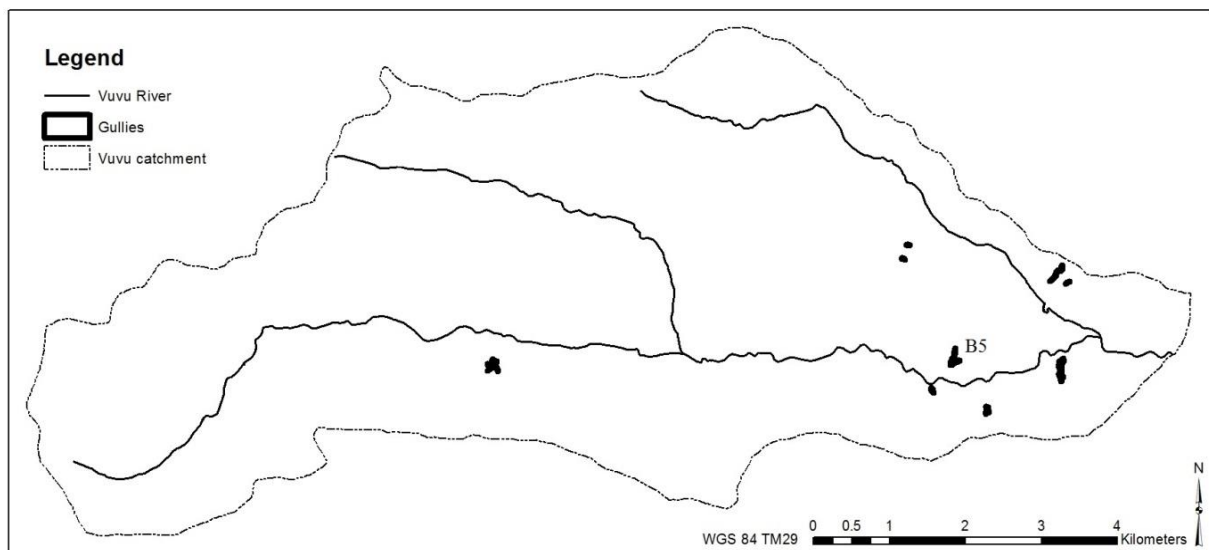


Figure 4.2: The location of the nine gullies that were used for the time sequence analysis in the Vuvu Catchment.

Of the erosion features (gully erosion, sheet erosion and landslides) that were visually assessed in the field ($n=78$), 63% were still actively eroding (Table 4.8). The remaining 38% showed signs of floor stabilization with bottoms being colonised by vegetation, although sidewalls were still actively eroding. When considering the geological provinces, it is evident that a large proportion (61–83%) of gullies are still eroding, whereas the majority (79–86%) of areas with sheet erosion have stabilized (Table 4.8). The trend for sheet erosion in the Clarens Formation and landslides in the Elliot Formation are not clear due to the small sample size, which is a function of their relatively small proportion in relation to the other features. Gullies that were connected tended to be more unstable than areas of sheet erosion that were connected to the drainage network (Table 4.8).

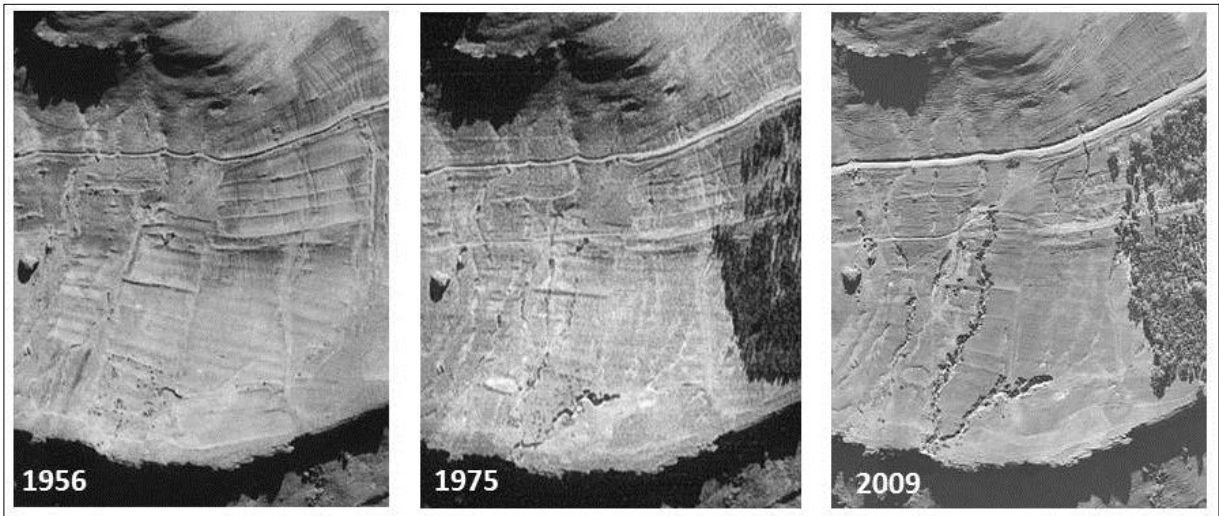


Figure 4.3: Historical aerial photos showing the evolution of gully B5 since 1956 (lower middle of photo). The gravel road near the top of the image can be used for scale; it is ca. 7 m wide in the 2009 image.

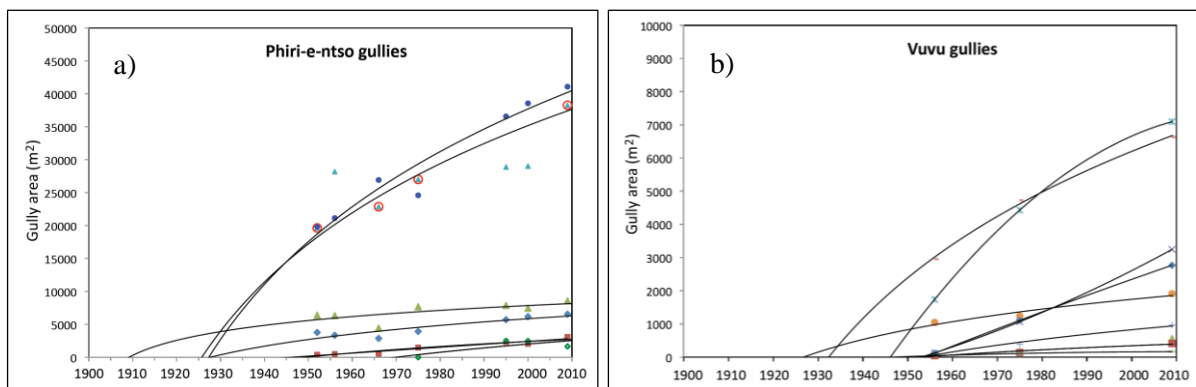


Figure 4.4: Changes in gully area over time. a) Data from the Phiri-e-Ntso catchment (Huber, 2013) fitted with logarithmic trendlines. b) Predictions of gully initiation in the Vuvu catchment are given by fitting a logarithmic curve to gully area data for 1956, 1975 and 2009. Note the difference in gully area scale between a and b.

Table 4.8: Summary of erosion features visited during source sampling. Percentages indicate the proportions of stable or unstable features and the percentage connected to the drainage network. Sample size is given in brackets.

Geology	Feature	% Stable	% Unstable	% Connected & stable	% Connected & unstable
Drakensberg	Gully (n=16)	38	63	6	38
	Sheet (n=21)	86	14	24	0
Clarens	Gully (n=6)	17	83	0	50
	Sheet (n=1)	0	100	0	0
Elliot	Gully (n=18)	39	61	6	28
	Sheet (n=14)	79	21	7	0
	Landslide (n=2)	50	50	0	0

4.3.3. Topographic and geologic characteristics of drainage features

4.3.3.1. Effective densities of drainage features

Gullies and livestock tracks added significantly to the drainage density (76 m ha^{-1}) (Figure 4.1f, 4.1g and 4.5 and Table 4.9). Downslope drainage density was increased by more than 20% overall above natural. Continuous gullies contribute ca. 10% to the drainage density and were effective conduits as these features were pathways flowing down-slope and were directly linked to the drainage network (Figure 4.5a to 4.5c).

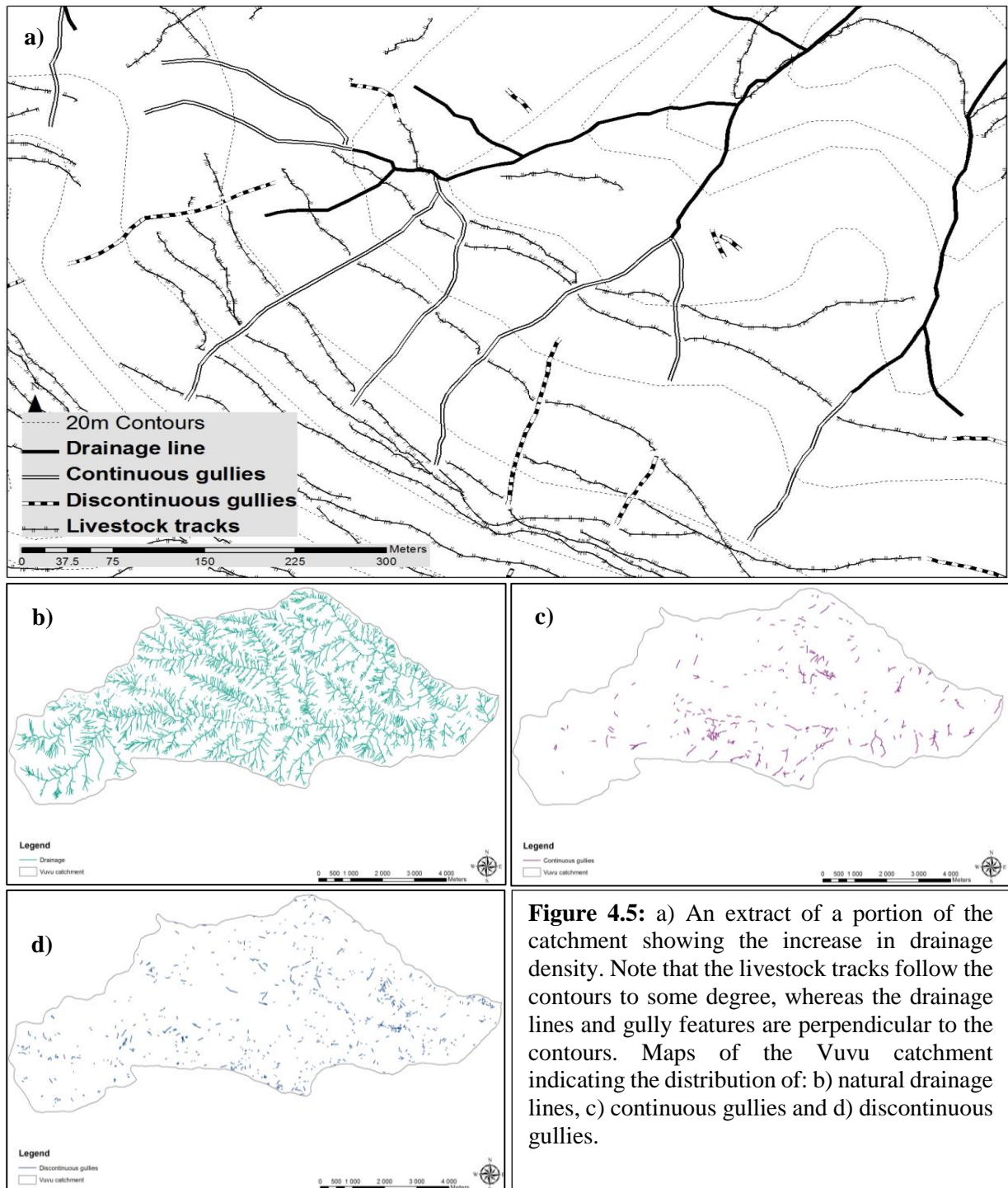


Figure 4.5: a) An extract of a portion of the catchment showing the increase in drainage density. Note that the livestock tracks follow the contours to some degree, whereas the drainage lines and gully features are perpendicular to the contours. Maps of the Vuvu catchment indicating the distribution of: b) natural drainage lines, c) continuous gullies and d) discontinuous gullies.

The 10% contributed by discontinuous gullies have less sediment and water transport potential as these features were not directly connected to the drainage network (Figure 4.5a, 4.5b and 4.5c). The Elliot Formation had the highest drainage density of 84 m ha⁻¹ compared to 73-74 m ha⁻¹ for the other formations. Continuous and discontinuous gullies both followed this trend where densities were highest for the Elliot Formation, in some cases double that of the other formations.

Roads and livestock tracks, both across-slope features, added ca. 158% to the drainage density. Roads were the lesser contributor (<1%) compared to livestock tracks (157%). Roads and livestock tracks tend to follow the contour to some degree (gently sloping) or are aligned along ridges, but frequently do cross down-slope orientated drainage features, allowing them to accumulate and route hillslope runoff towards the drainage network during overland flow events (Plate 4.3 and Figure 4.5). Roads were only found on the Elliot Formation and increased the across-slope drainage by 25%. Livestock track densities were ca 50% higher for the Elliot Formation (195%) than the Clarens (147%) or Drakensberg (142%) Formation. The total increase in across-slope drainage was more than 220% for the Elliot Formation.

Table 4.9: Densities of drainage features for the entire catchment and the various geological provinces.

	Unit	Catchment	Elliot	Clarens	Drakensberg
Area	ha	5329	1355	529	3444
% of total			25	10	65
Natural drainage	Length (km)	404	114	39	251
	Density (m ha ⁻¹)	76	84	74	73
Continuous gullies	Length (km)	42	15	3	24
	Density (m ha ⁻¹)	8	11	6	7
	% of drainage	11	13	9	10
Discontinuous gullies	Length (km)	43	15	3	25
	Density (m ha ⁻¹)	8	11	5	7
	% of drainage	11	14	7	10
Roads	Length (km)	29	29	0	0
	Density (m ha ⁻¹)	1	22	0	0
	% of drainage	1	26	0	0
Livestock tracks	Length (km)	637	222	57	357
	Density (m ha ⁻¹)	120	164	109	104
	% of drainage	158	195	147	142

4.3.3.2. Slope preferences of pathways

Natural drainage lines were prevalent on all slope classes, with a slight increase in density on the steeper slopes (Figure 4.6).

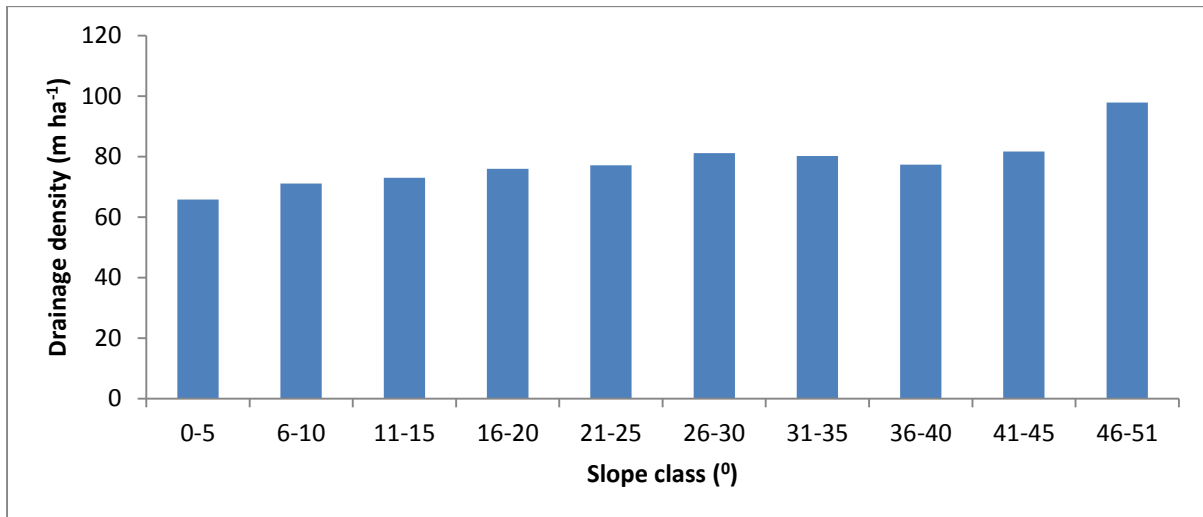


Figure 4.6: Frequency distribution of natural drainage lines within the catchment.

Continuous and discontinuous gullies were similar in their frequency distribution and formed primarily on lower slope angles (6–25°), although catchment-wide slope distribution incorporated steeper slopes (6–35°) (Figure 4.7). This indicated that gullies form where slope gradients relax. It could be a function of hydraulic working in terms of the effective catchment area needed to generate discharge sufficient to overcome the necessary shear stress to cut into slope material. This effective discharge was enhanced on the steeper slopes above the gully features by across-slope drainage features that concentrated flow on the steeper slopes and enhanced the connectivity of the drainage network on the steep slopes. The other potential reason is that the gullies only form where there is sufficient material to erode. The lower talus slopes had thick deposits that could potentially be eroded. Gullies propagated upslope to a point where soils were thin and gully formation was limited by the thickness of the soil layer.

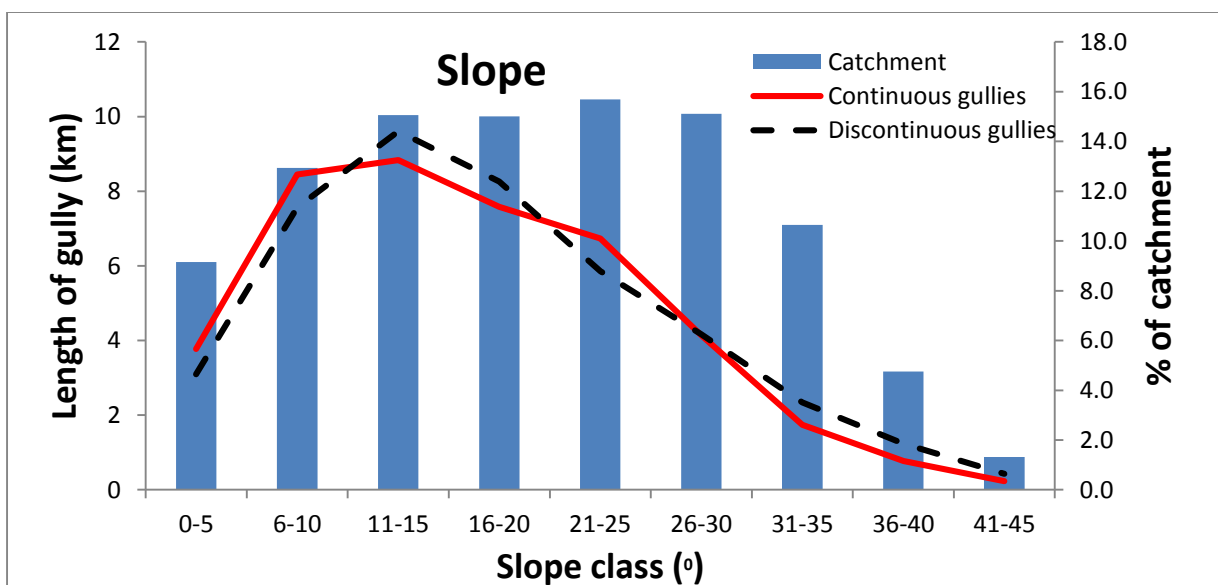


Figure 4.7: Frequency distribution of slope angles within the catchment and of the total length of continuous and discontinuous gullies.

The frequency distributions of continuous and discontinuous gullies for the different geological provinces followed a pattern similar to that observed for the entire catchment, except for continuous gullies on the Elliot Formation (Figure 4.8 and 4.9). Continuous gullies on low gradient slopes ($<5^{\circ}$) on the Elliot Formation were more than double the density of the other geological provinces. This could be a function of the presence of roads on these gentle slopes that contribute additional flow (Figure 4.10) to the system and initiated gully formation. Roads only were present on the Elliot Formation, with two thirds of the roads situated on slopes less than 10° . This hypothesis was not true in all cases for the Elliot Formation as some of the connected gullies were completely isolated from the roads. The other explanation could be linked to the highly erodible nature of the material of the Elliot Formation, which lends itself to the formation of larger gullies that reach all the way to the drainage network.

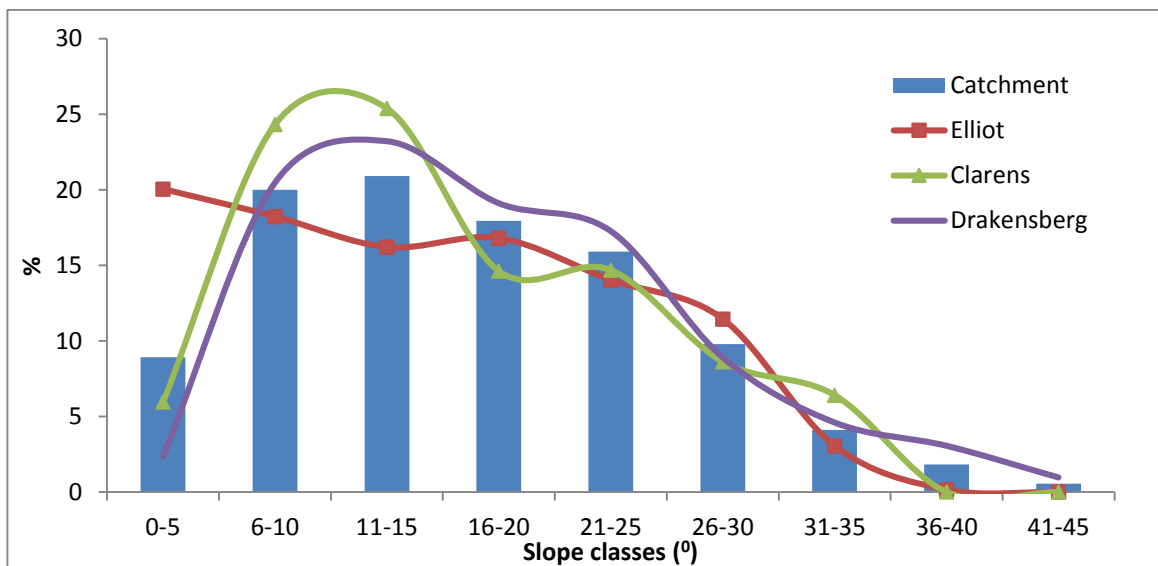


Figure 4.8: Frequency distribution for continuous gullies for the entire catchment and the geological provinces.

Livestock tracks were mainly identified on slopes ranging from $6-35^{\circ}$ for the entire catchment, but less frequently identified on steep slopes on the Elliot Formation and more frequently identified on steeper slopes on the Drakensberg Formation (Figure 4.11). This could be a function of the natural slope distribution throughout the catchment as seen in Table 4.1. The presence of these across-slope drainage features on the steeper slopes has led to the concentration of overland flow before it reaches the lower gradient slopes. This reduces the chances of infiltration and enhances drainage efficiency and connectivity.

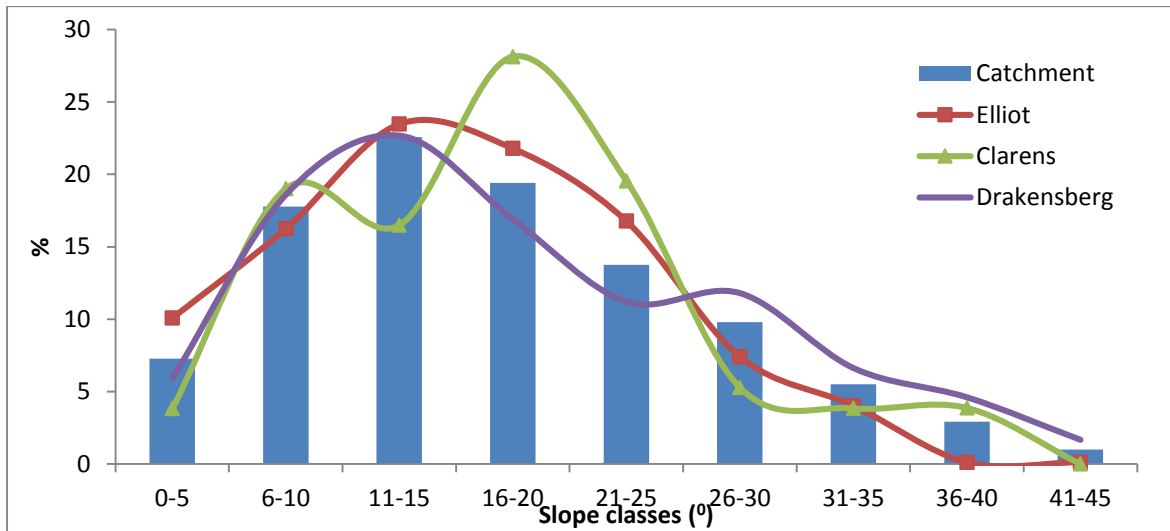


Figure 4.9: Frequency distribution for discontinuous gullies for the entire catchment and the geological provinces.

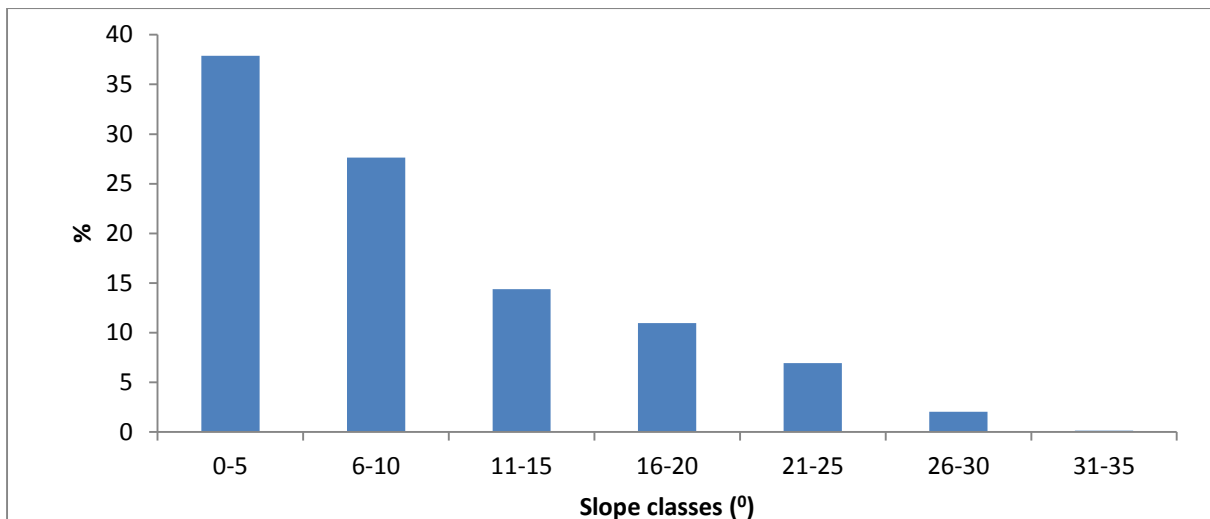


Figure 4.10: Slope frequency distribution for roads on the Elliot Formation.

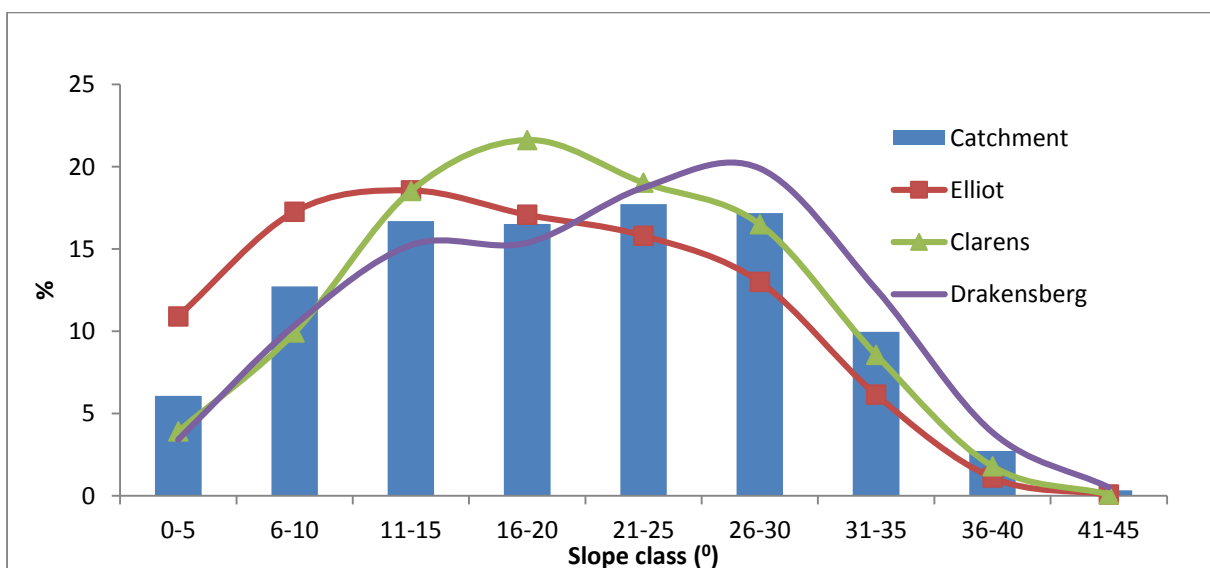


Figure 4.11: The frequency distribution for the slope classes that livestock tracks were traversing.

4.3.3.3. Aspect preferences of pathways

A large proportion (>60%) of the catchment and the individual geological provinces that were included in the analysis were north-facing (Figure 4.12). This north-facing dominance was in part due to the removal of shaded areas, which were on steep (23–45°) south-facing slopes. This removal of shaded areas accounted for a 4-6% shift that favoured north-facing slopes. Another factor that influenced the larger proportion of the slopes being classified as north-facing was the 187° that represent north-facing slopes against the 173° that represent south-facing slopes. Regardless of the apparent bias towards north-facing slopes and the removal of the steeper shaded south-facing slopes, it was perceived to be an accurate representation of what was observed in the field. To remove this bias, the total length of the drainage features was divided by the area of the aspect class, resulting in a density measurement that could be compared across various classes.

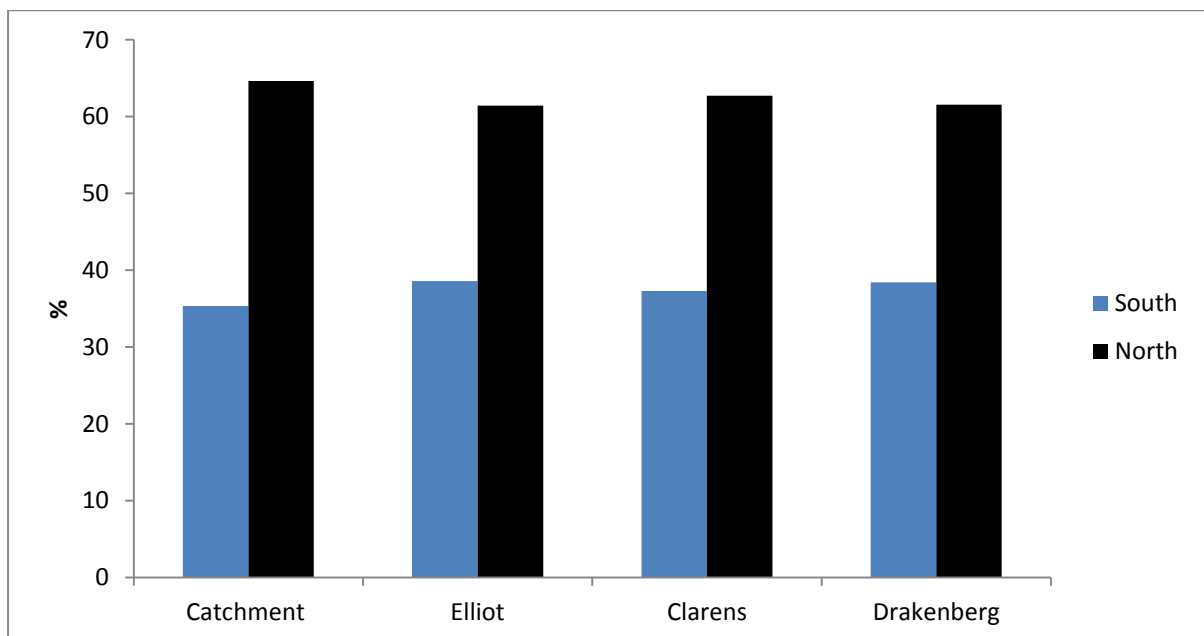


Figure 4.12: Percentage of the catchment or geological province that faces south and north.

When considering the natural drainage density for the catchment, it can be seen that the wetter south-facing slopes have the highest drainage density (ca. 85 m ha⁻¹) (Figure 4.13). This trend is reflected for the various geological provinces with the Elliot Formation having the highest south-facing and overall natural drainage density (ca. 100 m ha⁻¹). This is unexpected as drainage density normally decreases in areas of a catchment where the slopes are less steep, as would be anticipated for the lower catchment. This higher drainage density for the lower catchment could be a function of the erodible nature of the Elliot Formation.

The highest continuous gully density was found on north-facing slopes of the Elliot Formation (13 m ha⁻¹) and the lowest density (4 m ha⁻¹) on south-facing slopes of the Clarens Formation (Figure 4.14). Continuous gullies had a higher density (25%) catchment-wide on north-facing compared to south-facing

slopes. Continuous gullies on north-facing slopes were respectively 57 and 98% denser than on south-facing slopes on the Elliot and Clarens Formations. The densities on the Drakensberg Formation were not influenced by aspect.

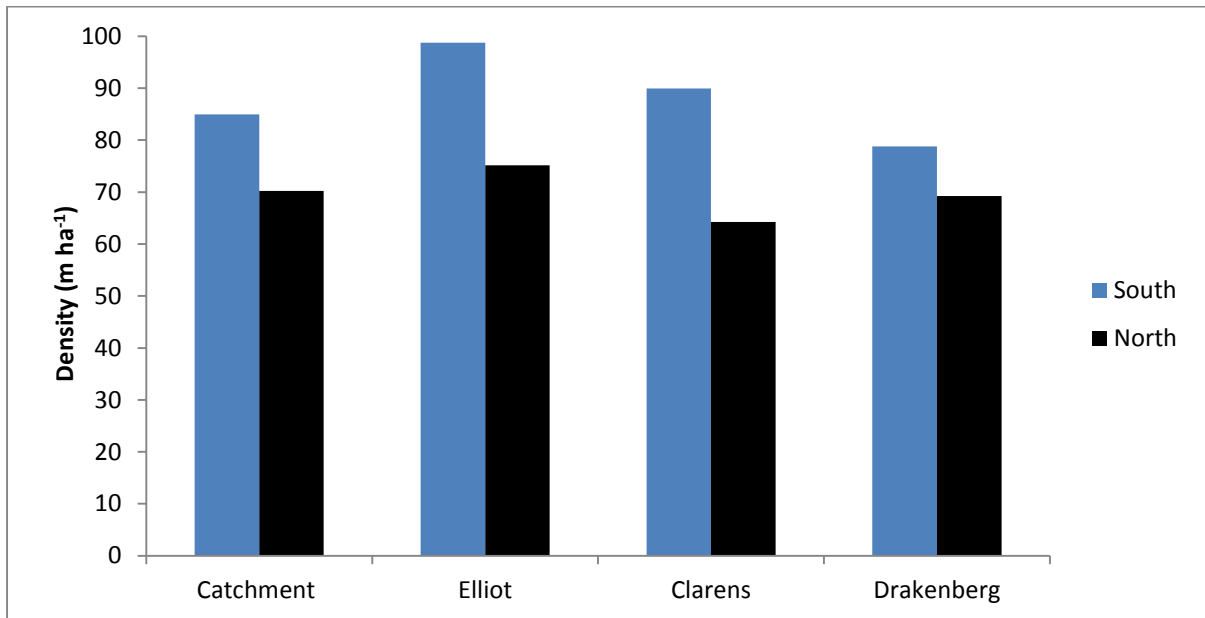


Figure 4.13: Aspect of natural drainage lines for the catchment and geological provinces.

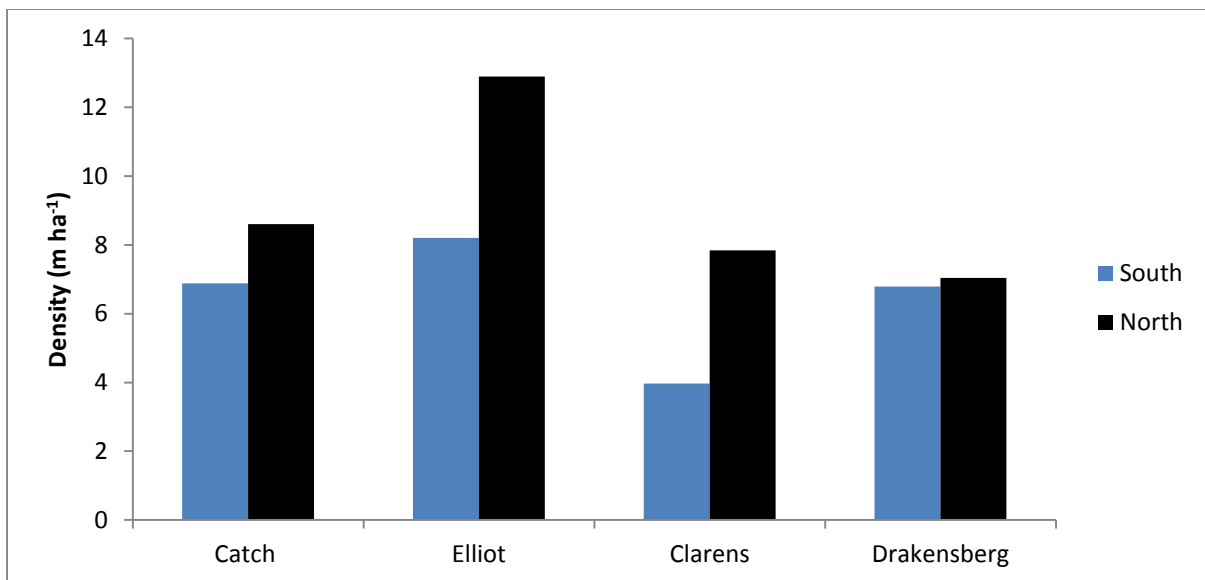


Figure 4.14: Densities of continuous gullies on south and north-facing slopes.

The Elliot Formation had the highest overall density of discontinuous gullies on south-facing slopes whereas the Clarens Formation had the lowest. Densities for discontinuous gullies were 24% higher on north-facing compared to south-facing slopes when looking at the entire catchment (Figure 4.15). The Elliot Formation had higher densities (12%) on south-facing slopes, whereas densities were higher (272% and 42% respectively) on north-facing slopes on the Clarens and Drakensberg Formations. The reversal

of the trend for the Elliot Formation could be due to the high density of continuous gullies on north-facing slopes. Gully formation was strongly influenced by aspect on the Clarens Formation and to a lesser degree on the Elliot Formation.

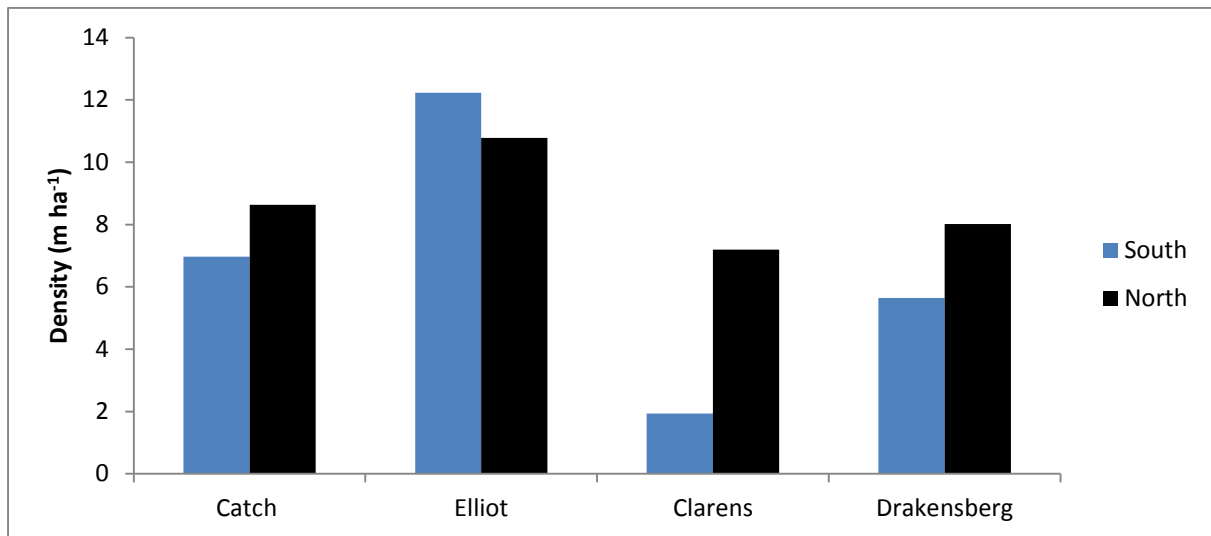


Figure 4.15: Densities of discontinuous gullies on south and north-facing slopes.

Roads were five-fold more dense on north-facing slopes and were a function of the location of houses (anthropogenic activity) on sunny north-facing slopes within the catchment.

The Elliot formation had the highest livestock track density (Figure 4.16). Livestock track densities were slightly higher (15%) for the catchment on north-facing slopes. Densities were not influenced by aspect on the Elliot and Clarens Formation, but were higher (25%) on the north-facing slopes of the Drakensberg Formation.

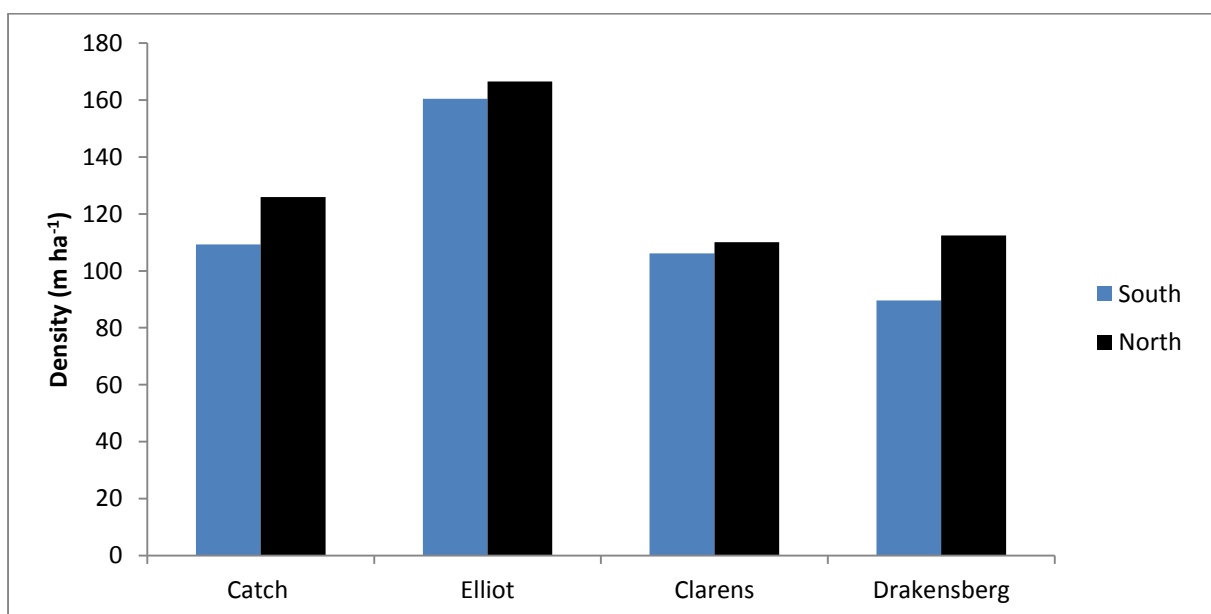


Figure 4.16: Densities of livestock tracks on south and north-facing slopes.

From the above statistics on drainage density and topographic properties, it was clear that the Elliot Formation had the highest drainage density for natural and anthropogenically induced drainage features. Natural drainage features had a higher density on south-facing slopes, whereas north-facing slopes were better connected by anthropogenically induced drainage features, especially on the Clarens and Elliot Formations.

4.4. Discussion

4.4.1. Introduction

This section discusses the mapping results presented in Section 4.3: Results. General topographic characteristics for the catchment and identified sources are discussed, including historical changes and the current state of erosion activity. Furthermore, linkages and their influence on water and sediment transport are discussed.

4.4.2. Catchment characteristics

The Vuvu catchment is a typical representation of the steep headwater areas of the Drakensberg Escarpment with steep slopes dominating the upper catchment and tier edges of the lower parts of the catchment (De Decker, 1981; Mucina et al., 2006). The step-like features are related to harder more resistant layers as can be seen in the Elliot Formation, where softer layers are removed by erosion, leaving the harder layers to form steep near vertical cliffs (Moon and Selby, 1983). Landscape relief is somewhat gentler at the lower altitudes.

A large proportion (<60%) of the catchment is north-facing. This pattern is reflected throughout all the geological provinces within the catchment. The reason for this uneven distribution could be related to erosion processes being more active on north-facing slopes that are exposed to greater temperature and moisture fluctuations over time, resulting in lower slope angles and longer slope lengths over the long term (geological time) (Meiklejohn, 1994; Boelhouwers, 2003).

4.4.3. Topographic and geological characteristics of potential source areas

Agricultural fields were mainly developed near and around villages on the relatively flat land that was available. The probable reasons for the location of the fields are mainly practical; farming near homesteads would enable easier protection of fields and would reduce travel time. The use of gentle slopes that had deeper soils and were less prone to erosion also would be more favourable for agriculture. The higher lying areas were cultivated prior to the 1950s, but fields and homes on the Basalt Formation

were abandoned by the time of the 1952 photos. Slopes on the Clarens Formation were most likely too steep, limited in size and didn't make for good ploughing and thus, were not favoured for agriculture. It is likely that agriculture dwindled in the post 1960s due to climatic and socioeconomic reasons (Kakembo and Rowntree, 2003). Betterment plans were enforced in the 1960s by government officials, where rotational grazing and rezoning of land unsuitable for cultivation were enforced. These enforcements were accompanied by a population influx to the Transkei due to forced removals from the Republic of South Africa (Vetter, 2007). The betterment plans were abandoned in the 1970s, after which farming was concentrated around villages on the Elliot Formation.

A relatively small percentage of the gully features were formed on agricultural fields on the Drakensberg and Clarens Formations (gullies <6%), but more frequently on the Elliot Formation (gullies <24%). The larger proportion of gullies (in relation to fields) on the Elliot Formation could be a function of the higher density of fields on this formation or the overlap of fields and erosion features on similar slopes. Although the Drakensberg Formation had the largest area that was gullied, the Elliot Formation had the highest density of gullies, highlighting the erodible nature of its soils (Vetter, 2007; Fey et al., 2010). Most gullies (70%) were connected to downstream pathways, with a general increase in size correlated with how well a feature is connected, similar to what Harvey (1992) found for gully systems in England. This well connected nature of the gullies explains their rapid development in the Vuvu catchment. Gully development is thus exacerbated when sediment is freely exported from the gully feature. The opposite is also true, where reduced export of sediment from the gully base leads to periods of gully stabilisation (Harvey, 1992).

In general, gullies were found on the moderate slopes at the base of steeper slopes. Vetter (2007) found gully features mainly on gentle to moderate slopes, with a reduction in area on steeper slopes. It is argued that this position, at the base of steeper slopes, is most favourable for gully development as there is sufficient runoff energy and depth of material for gully initiation and development (Schumm, 1979; Brierley and Murn, 1997; Hoffman and Todd, 2000; Kakembo, 2000; Kheir et al., 2007; Grenfell and Ellery, 2009). Road networks along the more gentle slopes also can increase the effective catchment area, increasing the volume of runoff and potential energy to initiate a gully (Huchzermeyer, 2014). The features expand laterally and upslope onto the steeper slopes where a limiting maximum size is reached as soils become thin.

The shift to a larger proportion of the gullies on the Drakensberg Formation being found on steeper slopes (in contrast to the Elliot Formation) could be a reflection of the distribution of slopes in the natural landscape. It could also be a function of soil type. In the first instance the landscape steepness is a function of rock hardness, where basalt is more resistant than mudstones and would thus form steeper slopes (Moon and Selby, 1983). The larger proportion of steeper slopes on the Drakensberg Formation makes the occurrence of gullies on steeper slopes more likely. The second reason, which is also related to the parent material, could be that soils on the Elliot Formation are more susceptible to erosion, even on gentle

slopes under low energy conditions, whereas on the Drakensberg Formation soils are only susceptible to erosion on steeper, high energy slopes (Vetter, 2007).

In very few cases the gullies were initiated by the upward migration of a base level change, rather than as a result of increased runoff from the steeper slopes. In a few localised instances base level change may have caused gully formation in the Elliot Formation where they developed on gentle slopes adjacent to the river.

Sheet erosion was more prevalent (double the area) than gully erosion and was identified throughout the catchment, especially on moderate slopes that are north-facing. Areas of sheet erosion were poorly connected and tend to occur on slightly steeper slopes compared to gully erosion (Vetter, 2007). As fields were mainly on gentler slopes, they had little overlap with areas of sheet erosion. The Elliot Formation was particularly prone to sheet erosion. Although large areas of sheet erosion were identified, the generally shallow nature of sheet erosion and the poor connectivity makes sheet erosion a less dominant source than gully erosion.

Landslides were the least extensive in area, with characteristics very different from the other source features, and were only found on steep south-facing slopes that were likely to be wetter and more prone to saturation. The Elliot Formation was once again the most susceptible to the occurrence of landslides, highlighting the poor structure of the soils found there (Fey et al., 2010). The likelihood of landslides contributing significant volumes of sediment on a regular basis is unlikely as they are poorly connected to the drainage network.

Roads and livestock tracks were relatively small in area (1.3% of the catchment), but their relatively good connectivity could mean that they could contribute sediment on a regular basis. Their connectivity was difficult to calculate due to the coarse nature of the DEM in relation to the size of the features.

When considering the soil loss data for the catchment (ca. $13 \text{ t ha}^{-1} \text{ y}^{-1}$) it is evident that gully, field and sheet erosion features were the main sediment sources (ca. $4 \text{ t ha}^{-1} \text{ y}^{-1}$ respectively). Gully erosion approximations in the Transkei by Beckedahl and De Villiers (2000) were similar and ranged between $5.6\text{-}14.2 \text{ t ha}^{-1} \text{ y}^{-1}$. The sum of the contributions of the other erosion features was less than the sediment loss of any of the above-mentioned features. When the soil loss values for the individual geological provinces are considered, it is clear that the Clarens Formation has the lowest sediment loss per area (ca. $4 \text{ t ha}^{-1} \text{ y}^{-1}$), followed by the Drakensberg Formation (ca. $10 \text{ t ha}^{-1} \text{ y}^{-1}$) and the Elliot Formation (ca. $25 \text{ t ha}^{-1} \text{ y}^{-1}$) the highest. Fields contribute the greatest proportion of sediment (ca. $10 \text{ t ha}^{-1} \text{ y}^{-1}$) on the Elliot Formation, which is almost double that of gully and sheet erosion respectively (ca. $5 \text{ t ha}^{-1} \text{ y}^{-1}$) also on the Elliot Formation. The contribution from roads (ca. $2 \text{ t ha}^{-1} \text{ y}^{-1}$) on the Elliot Formation was surprisingly high when individual geological provinces are considered, making it a more significant sediment source. These soil loss data per geological province emphasise the erodible nature of the Elliot Formation.

It is likely that calculated sediment loss from the erosion features would be even further reduced as a portion of the sediment would be deposited due to the poor connectivity of some of the features (sheet and landslide erosion), thus retaining some of the sediment on the slopes within the catchment. The total soil loss for all erosion features was ca. $13 \text{ t ha}^{-1} \text{ y}^{-1}$ which is higher than the global average of ca. $10 \text{ t ha}^{-1} \text{ y}^{-1}$ (Yang et al., 2003) and lower than the water erosion modelled for this catchment ($30 \text{ t ha}^{-1} \text{ y}^{-1}$) by Le Roux et al. (2008), both using an approach based on the USLE. The values of Yang et al. (2003) and Le Roux et al. (2008) account for sheet erosion only and does not include gully erosion, thus are potentially underestimating the total soil erosion for areas where gully erosion is prevalent. Diffuse sheet erosion was not accounted for during the current mapping study, thus was the mapping based soil erosion rates compared to the potentially underestimated USLE based modelled data to estimate what proportion diffuse surface erosion could contribute. Both approaches are likely to underestimate the total soil erosion to some degree, but as no other data are currently available and both approaches estimate soil erosion and not sediment yield, it was used for comparative purposes. If the difference of ca. $17 \text{ t ha}^{-1} \text{ y}^{-1}$ between Le Roux et al.'s and the current study's values is assumed to be surface soil erosion, it equates to 1.2 mm of surface soil lost annually for the remainder of the catchment (thus excluding the area of the erosion features already accounted for). If the values considered are accurate, despite the ~50% variability in field depth measurements, it can be concluded that catchment-wide surface soil erosion is the main sediment source for the Vuvu catchment and it highlights the importance of hillslope-channel connectivity to transfer eroded sediment to the river system.

4.4.4. Gully history and current stability

Large-gully formation seems to have started in the early 1900s in the Vuvu and agrees with similar results found in the neighbouring Phiri-e-Ntso catchment (Huber, 2013). The smaller gullies were mostly initiated post 1950. This would suggest that the largest features were located on the most sensitive areas in the landscape, mainly gentle slopes on deep soils at the base of a steep slope and were the first areas to be eroded. The smaller gullies were located on steeper slopes in shallower soils and were only starting to erode when landscape pressure reached its peak. These initiation periods of larger and smaller gullies could be related to periods of increased erosive and hydrological power, such as after a major drought (reduced vegetation cover) or intensification of livestock pressure (overgrazing and trampling), possibly combined with drought.

The above would agree with the literature that stated that rivers draining the escarpment ran 'gin' clear even after large floods in the early 1900s, but became turbid by the 1950s (Hey, 1957). Another recorded change was the measures that were enforced to prevent East Coast fever in cattle spreading throughout the Transkei region (Bundy, 1987). In 1906 cattle dip tanks were introduced to limit the potential pest carrying ticks and by 1911 it became law to dip cattle every week. Restrictions on cattle movement hampered livestock trade which, in turn, led to an increased density of cattle and altered patterns of

transhumance pasture usage. Local farmers complained about the general exhaustion of the cattle walking to and from the dip tanks on such a frequent basis, as it decreased milk production and the ploughing power of their animals (Bundy, 1987). It is surmised that this increase in numbers, and the weekly trip from the pastures to the dip tanks led to general veld degradation and the establishment of livestock tracks throughout the catchment. This would have led to increased runoff and hydrological connectivity that could contribute to gully formation. According to local elders talked to during field work (born in the 1940s) livestock did extremely well during the 1950s, due to well-timed and sufficient rainfall, putting an added wave of pressure on the already stressed landscape and could be related to the initiation of smaller gullies around 1950. During field sampling, observations revealed that a large proportion (>60%) of the erosive features was still actively eroding. Gullies were mainly unstable (>61%) compared to areas of sheet erosion that predominantly showed signs of stabilization (>78%). It also showed that areas of sheet erosion would stabilize regardless of whether they were connected to the drainage network, whereas, gullies that were connected to the drainage network were likely to be unstable. This would suggest that areas of sheet erosion would contribute sediment for much shorter periods, as they stabilized relatively quickly compared to the gullies that seem to remain active for longer periods of up to 100 years.

4.4.5. Topographic and geological characteristics of drainage features

Down-slope (21% increase) and across-slope drainage (178% increase) features increased the drainage density significantly. This increase is much higher than the 10% calculated by Croke et al. (2005) for an extensive forest road network in New South Wales and highlights the extremity of the connectivity increase in the Vuvu catchment. Continuous gullies were the most important pathway as they are directly linked to the channel network, making the timing of their flow contribution synchronous with the rest of the natural drainage network (Croke et al., 2005). Discontinuous gullies were assumed to be hydrologically and sedimentologically disconnected from the drainage network during low intensity flow events due to dispersive flow through vegetation buffers (Croke et al., 2005), but can contribute significantly once they become connected during high magnitude events (Fryirs et al., 2007a, 2007b). Both continuous and discontinuous gullies are mainly formed in the upper parts of the natural drainage network that would be vegetated under un-impacted conditions. The removal of vegetation and entrenchment of the natural drainage features alters hydrological and sedimentological behaviour, allowing water and sediment to be efficiently transported down the catchment (Brierley and Murn, 1997; Cammeraat, 2002; Vanacker et al., 2005; Rommens et al., 2006; Grenfell and Ellery, 2009). The horizontal features were not regarded as efficient a conduit as the downslope gullies, but these features do concentrate overland and upper subsurface flow. Huchzermeyer (2014) found that livestock tracks and roads added significantly to the hillslope-channel connectivity on the Elliot Formation in the Vuvu catchment, especially on gentler slopes. The horizontal features crossed downslope drainage features on a regular basis, potentially discharging concentrated flow into the drainage network (Hoffman and Todd, 2000; Croke et al., 2005; Huchzermeyer, 2014). Drainage efficiency was thus intensified by

anthropogenic influence for all slopes, increasing hillslope-channel connectivity, even on gentle slopes that could potentially act as buffers.

The well-connected nature of the Elliot Formation was not expected as the lower parts of a catchment usually have the lowest drainage density. The main contributing factor is likely to be the erodible nature of the Elliot Formation, leading to a well-developed drainage network. This trend was exacerbated by the human induced features such as roads and livestock tracks that were a result of the population living in the lower catchment underlain by the Elliot Formation. This led to larger increases of drainage densities on the Elliot Formation, making it the best connected formation in the catchment. Another reason for this result could be a function of the steep south-facing shaded areas, predominantly from the upper catchment, that were removed for the analysis.

Natural drainage density was high throughout the catchment as would be expected for a catchment with high relief draining the escarpment. The increase in drainage density for steeper slopes was thus also expected (Kirkby et al., 2002; Fryirs et al., 2007b; Miller et al., 2013). Gullies and livestock tracks were mainly found on moderate to steep slopes ($6-30^{\circ}$), but man made roads were most common on gentle slopes ($<5^{\circ}$). This is also shown in Kirkby et al.'s (2002) work which states that steeper slopes are inherently more vulnerable to erosive rain as thinner soils support less vegetation. Thinner soils and less vegetation would translate into more runoff and higher erosive energy, favouring soil erosion on unprotected or disturbed surfaces. This horizontal and down-slope increase in drainage on a range of slopes will have a marked influence on the hydrology and sediment transfer of the system (Cammeraat, 2002; Croke et al., 2005; Fryirs et al., 2007b) as buffers are breached through increased flow energy and the formation of pathways. The Elliot formation was once again the most modified formation, with distributions ranging across the spectrum of slope classes, compared to the other formations where moderate to steep slopes were more prominent. This meant that the gentle slopes that should act as sediment stores or buffers (Fryirs et al., 2007b) were connected and allowed sediment to be transferred to the channel network of the Elliot Formation. Connectivity is thus increased across all slope classes for the Elliot Formation, whereas connectivity is mainly increased on moderate to steep slopes for the Clarens and Drakensberg Formations.

The longer slope lengths and associated higher denudation rates of the north-facing slopes are reflected in the gully density data, where north-facing slopes have higher continuous and discontinuous gully densities. Only the discontinuous gullies on the Elliot Formation do not conform to this trend. The likely reason for the slightly higher proportion of discontinuous gullies occurring on south-facing slopes could be due to the large number of continuous gullies that are north-facing, meaning that a large portion of the discontinuous gullies on north-facing slopes have become continuous. This also points out that south-facing slopes are more resilient or less impacted and would have less chances of developing continuous erosion features. South-facing slopes tend to be wetter and would support denser and often woody

vegetation that binds the soil, reduces runoff and protects the soil against erosive rainfall (Holland and Steyn, 1975; Granger and Schulze, 1977).

When looking at livestock tracks it can be seen that north-facing slopes are either more susceptible to the formation of these linear features, or that north-facing slopes are preferred by livestock. The trend is exacerbated for the higher altitudes (Drakensberg Formation), which could show that livestock prefer north-facing slopes as these slopes support sweeter grasses that are more palatable than sour grasses growing on wetter, more acidic south-facing slopes (Ellery et al., 1995; Mucina et al., 2006). This effect would be exaggerated at higher elevations where soils are naturally more acidic due to higher rainfall. Another possible reason for the higher proportion of livestock tracks on north-facing slopes could be related to the removal of shaded steep south-facing slopes from the analysis.

4.5. Conclusions

A GIS proved to be a useful tool to address connectivity through identifying and assessing potential source areas and pathways using high resolution aerial imagery. Historical aerial photos could only be used to determine the history of the larger erosion features (length >50 m) due to lower resolution. Field visits and land based photos proved invaluable to assess the current state of the erosion features.

The Vuvu catchment is characterised by steep confined valleys with steep slopes in the upper catchment. Along the lower sections of the catchment (Clarens and Elliot Formations), the landscape becomes more step-like, with gentle slopes being separated by moderate to steep slopes. Along the Elliot Formation, the valley is less confined, making space for a narrow alluvial fill. South-facing slopes are steeper than north-facing slopes, resulting in shorter slope lengths and a larger proportion (>60%) of the catchment being north-facing. Natural drainage density increases on steeper slopes.

Erosion features were widespread throughout the catchment, with agricultural fields, gully and sheet erosion being the main sediment sources that could be identified. When modelled data by Le Roux et al. (2008) are considered, general surface erosion is estimated to be the main sediment source for the catchment. Fields covered the largest portion of the landscape and contributed similar volumes of sediment to gully and sheet erosion respectively, on an annual basis. Most of the fields were found on moderate slopes ($5-20^{\circ}$) that were fairly well connected to the drainage network, making them potential sources if soil was mobilised. Sheet erosion was mainly found on moderate slopes and had the largest total surface area, but the poor connectivity of these features made them less significant sediment sources. Gullies also were mainly found on moderate slopes and remained active for up to 100+ years, whereas areas of sheet erosion seemed to stabilize over shorter periods. It is likely that gullies are the dominant sediment source due to their high level of connectivity to the drainage network, size, depth and lifespan. Landslides, roads and livestock tracks did not feature as significant sediment sources due to their relatively small volume. Roads and livestock tracks, however, play an important role in connecting slopes

to the drainage network as they concentrate and route stormflow to the drainage network. Erosion feature densities were highest for north-facing slopes, except for landslides.

The Elliot Formation was highlighted as the geological province that was most susceptible to sheet and gully erosion and that has suffered the highest landscape pressure in terms of cultivation, grazing, trampling and hydrological changes (e.g. roads, livestock tracks), making it the main source of sediment per unit area of the catchment. The erodible nature of the Elliot Formation is demonstrated by the Elliot and Drakensberg Formations that contributed similar volumes of sediment (~1 000 000 m³ each) over the past 50 years with the area of the Drakensberg Formation being more than double that of the Elliot Formation.

Gully features developed mainly on moderate slopes at the base of steeper slopes. The likely reason for the location of these features could be due to an increased hydrological contribution from upslope areas (reduced grass cover, hardened soils, increased pathways and increased catchment area due to road construction) and that moderate slopes have thick soil deposits that were easily eroded. Livestock tracks contributed significantly to the horizontal drainage network, concentrating flow and linking slopes to the rest of the drainage network. Evidence from historical photos suggests that erosion intensified during 1910-1970 when gullies were initiated.

Future management should address the influence of livestock on the landscape. Grazing pressure should be reduced to allow eroded areas to stabilize and re-vegetate. Once connectivity is reduced, grazing can be continued in a sustainable manner. Moving livestock over large distances on a regular basis should be minimized to prevent trampling and the formation and perpetuation of livestock tracks. Rehabilitation should target the lower parts of larger active gullies on gentle slopes in order to create a vegetation buffer that would retain sediment. This might not be accomplished easily due to the high hydrological energy along the lower parts of the continuous gullies, but once vegetation is established it should promote the upslope stabilization of the gully features.

To summarise the findings of this terrain and aerial photo analysis, it is evident that the Vuvu catchment is a steep headwater catchment with a high drainage density. It is naturally well connected, but due to anthropogenic influences such as ploughing of fields and subsequent abandonment leading to gullying, erosion and landscape connectivity have been increased over the past 100 years. Fields, gully and sheet erosion features were identified, through mapping, as the main sediment sources. North-facing slopes ranging from 5–20° were the most susceptible to erosion. Pathways linking slopes to the channels have increased both in the across-slope (roads and livestock tracks) and down-slope (gully) directions. This increase in connectivity results in sediment that has been mobilized off the slopes having a higher potential of being transported to the main river, even during lower intensity rainfall events. The potential to store sediment in the catchment has thus been reduced due to anthropogenic activity. From historical photo analysis it was predicted that the major gullies were initiated between 1910 and 1970. It is thus likely that other erosion features would have developed during the same time. Gullies and livestock tracks

were the main contributors to increased hillslope-channel connectivity. The Elliot Formation was the most impacted in terms of erosion and connectivity and that is likely to be a function of the erodible nature of the soils and ongoing anthropogenic activity.

Chapter 5: Valley fill character and history

5.1. Introduction

A river is the product of its catchment, reflecting the integrated effects of climate, geology, land use and basin physiography (Knighton, 1984; Fryirs and Brierley, 2013). Sediment that is deposited in the valley fill would reflect the processes that are active in the catchment. Clues such as coarser layers in overbank deposits indicate higher energy events (Tooth et al., 2013), whereas changes in river channel shape and spatial layout indicate changes in flow and sediment dynamics (Schumm, 1977; Knighton, 1984; Erskine, 1986). Thus, the valley fill features store a historical record of past events and catchment dynamics (Fryirs and Brierley, 2013). Sediment stores, such as fluvial terraces, are indicative of varying erosional stages and can thus be used to unravel some of the history of the valley fill development (Pavlopoulos et al., 2009). Terraces are mostly step like features with a plane indicating a previous fill level that has since been eroded (Erlanger et al., 2012). Various valley fill levels can exist, with the older features found at the highest elevations relative to the stream (Pavlopoulos et al., 2009; Erlanger et al., 2012). Lower flood benches keep an archive of the more recent dynamics of sediment deposition (Erskine and Livingstone, 1999). The various features store sediment that can be used as clues to assess how sediment dynamics and landscape connectivity have changed over time.

Landscape connectivity plays a major role in how and when sediment is redistributed within the valley fill (Brierley et al., 2006; Fryirs et al., 2007a). Sink features are shaped by cut and fill processes that vary spatially and temporally (Knighton, 1984). This connection of the valley fill with the rest of the catchment determines the shape, size and character of the various valley fill features (Fryirs and Brierley, 2013).

The geologic and landscape setting of the Vuvu catchment leads to narrow deep valleys with limited accommodation space. Due to the steep topography and limited accommodation space it is expected that sediment delivery ratios are high, thus storing relatively little sediment in the catchment. Harder lithological layers result in lower hillslope and valley floor slope angles that have the potential to act as localised sediment stores. From evidence elsewhere in South Africa (Okhombe valley, KwaZulu-Natal) it can be deduced that sediment that was made available during the Last Glacial Maximum was redistributed and stored on these localised low angled surfaces during the relatively stable warmer and wetter Holocene (Temme et al., 2008). Sediment residence times in the larger sediment stores could thus be up to thousands of years during stable conditions. It is hypothesized that the recent anthropogenic related changes in connectivity (Chapter 4) has initiated valley fill incision.

The aim of this chapter is to characterise the Vuvu valley fill in terms of the spatial layout and physical nature of the sediment sink features and the timescales over which the various features have developed in order to understand if these features act as sediment sinks or sediment sources. The following objectives were the focus of this chapter:

- Objective 3: Classify, map and characterise valley fill sink zones
- Objective 4: Provide a chronology of sedimentation and valley fill incision

The results of these objectives would give a good understanding of the structure of the features that are present, their likely function in storing fine grained sediment, the likelihood of stored sediment being remobilised and the potential of these features to store sediment in the future. It will also allow for changes in the sediment regime to be determined. This will contribute to the connectivity framework in terms of the function and structure of the valley fill as a sediment sink or source.

5.2. Methods

The detailed methods are outlined for each of the objectives below.

5.2.1. Objective 3: Classify, map and characterise valley fill sink zones

Large scale (1:1 000) geomorphological maps of the Vuvu valley fill features were constructed based on high resolution colour aerial photographs (0.5 m resolution; 2009) and topographic surveys (Pavlopoulos et al., 2009). An Epoch 35 Differential Global Positioning System (centimetre accuracy) was used to survey valley cross sections and along valley features (terraces, flood channels and main channel), noting clear breaks in slope, landscape features and changes in sediment composition (Gordon et al., 2004). Fifteen transects were surveyed along a 2 km section of the lower Vuvu valley fill (Figure 5.1). By overlaying the surveyed transects and features on a high-resolution aerial image in ArcInfo 10, a detailed map was drawn of the valley bottom. The features were classified based on survey results, field observations and remote sensing (Pavlopoulos et al., 2009). Evidence such as slope, sediment composition and size of features was used, together with changes in vegetation, to extrapolate field based evidence onto the map.

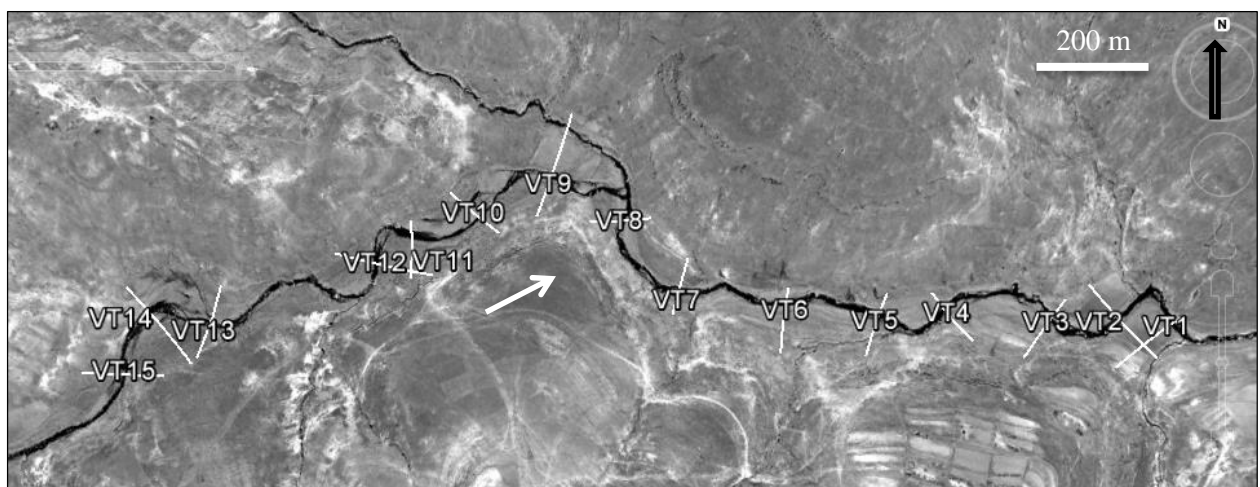


Figure 5.1: The location of the Vuvu valley fill transects (Google Earth image). The white arrow indicates flow direction.

The following sink features were identified: alluvial fans, terraces and flood benches and instream cobble bars. Cobble bars mainly store gravel and larger particles that do not form part of the suspended sediment load and were omitted from this study as suspended sediment is a major problem downstream. Alluvial fans were defined as sediment deposits with a conical shape that radiates downslope from the point where a stream loses confinement (Bull, 1972). Fans are seen as buffers that store sediment and reduce hillslope-channel connectivity (Fryirs et al., 2007a). Terraces were defined as flat elevated alluvial levels or old flood benches that were no longer thought to be inundated (this is tested in Chapter 7) during large floods (Pazzaglia, 2013). Terraces lined the outer parts of the valley fill and were located high above the current water level, thus indicative of former flood levels that were abandoned as a result of river incision. Terraces often act as buffers as the level surface of the feature reduces hillslope-channel connectivity. Flood benches were defined as relatively flat sediment deposits adjacent to the river channel that is inundated during contemporary floods (Rosgen, 2008). The flood benches have a similar function to flood plains in larger lower-gradient systems, but are not as continuous nor extensive in size. Flood benches are seen as buffers that store sediment and reduce hillslope-channel and longitudinal connectivity (Fryirs et al., 2007a). Both terraces and flood benches consisted of horizontally layered fluvially deposited sediment. Flood channels were identified on some of the flood benches and as they act as flood water pathways, they were included on the maps. The total area of flood benches was calculated in order to determine the volume of sediment that has recently been stored in the valley fill.

As flood benches were the most recent main sediment sink in these valley fills, four vertical cores were taken of flood benches using an Eijkelkamp percussion corer (nose diameter 4.5 cm), targeting two frequently inundated lower flood benches and two higher lying flood levels that were likely to be inundated less frequently. Only limited undisturbed sites were available, as large areas of the valley fill has been ploughed for agriculture. The location of cores was spread along the lower reach of the valley fill (VT2, VT4 and VT7; Figure 5.1) and care was taken to core undisturbed parts of the flood bench where horizontal sediment layering was clearly visible throughout the sequence. All cores were taken from the upper sandy sequence until a cobble layer was encountered (often visible in the bank). The cores ranged between 60 and 120 cm and were cut into 2 cm slices (30 to 60 samples per core).

In the laboratory, slices were dried for 48 hours at 40°C to preserve the magnetic character of the sediment. Every second slice was analysed, to reduce sample numbers due to time and budget constraints, for particle size distribution using a Malvern Mastersizer 3000 (<2 mm fraction). Organic content was calculated using loss on ignition at 450°C over 12 hours (Rowntree et al., 2012). Grade scales for particle size were used as given by Gordon et al. (2004). Further sampling was done to characterise and date terraces and sediment from abandoned channels (sampling described in section 5.2.2.). A sorting index (using the D10, D50 and D90) as given by Andrews (1983) was used to evaluate the degree of sorting of each sample. The poorer the sorting the more varied the sources.

5.2.2. Objective 4: Provide a chronology of sedimentation and valley fill incision

The chronology of sedimentation and incision was based on rainfall records, historical aerial image analysis and sediment dating. As rainfall is the main driver for sediment connectivity (flood and sediment transport events), long term rainfall data (dating back to 1937) for Matatiele (station T3E001; 65 km north-west of the Vuvu rain gauge), the nearest rainfall station along the escarpment, were analysed to determine if rainfall records agree with sediment chronologies.

Nel (2008) and Nel et al. (2010) describe rainfall along the Drakensberg escarpment as highly erosive and is produced by orographic uplift and thunderstorm activity occurring mostly in the late afternoon in summer. As Matatiele is 65km away, it is likely that rainfall might vary across this distance. In order to test what the difference might be for daily and monthly data, recently measured daily and monthly rainfall data were tested for significant differences between the Vuvu (station details presented in section 7.2.1.) and Matatiele datasets using a regression analysis. A longer term trend was calculated by fitting a 24 month running mean to the dataset. This indicated wet and dry cycles that could be used to interpret the sediment chronologies.

Recent valley fill and channel dynamics were further assessed using historical aerial images. Aerial images for 1956, 1966 and 1975 were georeferenced to georectified aerial images for 2009. The active channel along the Vuvu valley fill was digitized for each set of images. Sections where the river has abandoned a channel or has been straightened were identified. It was assumed that, when a channel is abandoned as a result of river straightening, the abandoned channel bed, consisting of cobble, would remain stable and fine sediment would slowly accumulate above it, thus preserving the abandoned channel bed elevation. Transects were surveyed where the channel was straightened, thus limiting the chances of cobble bar formation along the outside of these reaches (cobble expected to be deposited on the inside of bends), as a build-up of cobble post river straightening would give false readings of the abandoned channel elevation. A gouge corer was used to systematically core down to a cobble layer where the abandoned channel had been identified from the aerial images, recording distance along the transect and depth below the surface. The coring data were combined with the surveyed profile to determine the buried profile of the abandoned channel in relation to the current channel. As only two suitable sites could be identified along the Vuvu River, the neighbouring Phiri-e-Ntso River (with similar hillslope-channel connectivity increases and catchment area of 78 km²) also was assessed. An additional four sites were identified and surveyed along the Phiri-e-Ntso River. The difference in elevation between the active and abandoned channel was calculated based on the average of the five lowest points along each channel profile (active and abandoned). The difference in abandoned and active bed elevation was calculated for all six sites to determine the extent and timing of incision.

Various dating techniques were used to date the valley fill features. Pb-210 in combination with Cs-137 were used to date the more recent flood benches (maximum dating range <88 years due to loss of Rn-

222) (He and Walling, 1996; Saxena et al., 2002; Saint-Laurent et al., 2010; Du and Walling, 2012) and historical aerial images were used to verify flood bench dating and river dynamics. Optically Stimulated Luminescence (OSL) dating (dating range of 300–100 000 years) was used to date paleo-channels and terraces (Olley et al., 1998). Pb-210 is a radionuclide that has continuous natural fallout with a half-life of 22.3 years and is detectable for up to 100 years in stable lake environments (Appleby et al., 1986; Du and Walling, 2012). Sediment is dated based on the decay of Pb-210, assuming that the rate of supply from the atmosphere is constant. Cs-137, on the other hand, is an anthropogenic radionuclide that was first present in the atmosphere during the time of atmospheric thermonuclear testing (late '50s to mid '60s) and only appeared in the Southern Hemisphere around 1958 (Foster et al., 2007). Cs-137 can thus be used as a marker horizon since all sediment containing Cs-137 has been deposited post 1958 (Foster et al., 2005). OSL dating is based on the bleaching of quartz grains by sunlight and subsequent burial of the grains allows the time of last exposure to be dated (Duller, 2004).

Pb-210 dating is based on the activity of the unsupported Pb-210 in a sample, that is influenced by the radioactive decay of Pb-210 and the dilution of Pb-210 by sediment contributions. There are two sources of Pb-210 in sediment samples, both related to the U-238 decay series. The first is produced by *in situ* decay of Ra-226 (supported Pb-210), the second by Rn-222 (daughter isotope of Ra-226) that escapes into the atmosphere and forms Pb-210 that is a wet fallout assumed to be uniform through time (unsupported Pb-210). Pb-210 dating was first used for dating sediments from lake environments, but recently the method was successfully applied to floodplain environments using the Constant Rate of Supply (CRS) model (Saint-Laurent et al., 2010; Du and Walling, 2012; Manjoro, 2012). As floodplain sediment is not constantly inundated by water and the sediment calibre is likely to be greater than that of lake sediments, a proportion of the Rn-222 is expected to escape from the sediment and thus underestimate the supported Pb-210. An emanation coefficient of 0.3 was used as suggested by Du and Walling (2012) to adjust or correct the supported Pb-210 values in order to compensate for the loss of Rn-222 (mother isotope). Pb-214 activity was measured as surrogate for Ra-226 activity and was subtracted from the total Pb-210 activity to calculate the unsupported Pb-210 activity in the sample (Foster et al., 2007). This result was used in the composite CRS model. The CRS model compares the cumulative unsupported Pb-210 down to the specific slice in a core against the total Pb-210 in the core to calculate the age of that specific slice. This model proves to be reliable in areas with varying sedimentation rates, especially if Cs-137 is used to adjust the CRS age (Appleby, 2001; Du and Walling, 2012).

Pb-210 dating has a few shortcomings that might limit its use in this context. Unsupported Pb-210 delivery is dependent on rainfall, thus drier areas with a higher coefficient of variation will have a varying unsupported Pb-210 input (Appleby, 2008). Although the Thina catchment falls within a temperate climatic zone and the coefficient of variation is lower than for drier parts of South Africa, rainfall can follow natural wet and dry cycles and be spatially heterogeneously distributed; this could lead to the uneven delivery and distribution of Pb-210. Another shortcoming noted by Appleby (2008) is that in areas

with rapid sediment accumulation Pb-210 concentrations are diluted and could limit the dating horizon to three to four half-lives (e.g. 66–88 years) instead of six half-lives if the surface (peak) activity is lower than 100 mBq g⁻¹. This is the case in the Thina catchment where peak counts are in the region of 60 mBq g⁻¹, thus dating was limited to a maximum of 88 years with a high levels of uncertainty.

Cs-137 was used as an absolute marker in floodplain cores. The first detections of Cs-137 from the bottom of the stratigraphy marked 1958 as it was the start of the Cs-137 peak fallout in the southern hemisphere (Foster et al., 2007). The Cs-137 fallout peak was assumed to occur during 1965, but this peak was not clearly discernible in sediment cores in the southern hemisphere (Foster et al., 2007). As clear Cs-137 peaks were hardly ever detected in the Vuvu cores, the first detection of Cs-137 was used as the 1958 marker in the composite Cs-137 and Pb-214 CRS model. Down core migration of Cs-137 might have been possible, introducing dating error, but the potential downward migration could not be quantified and was disregarded as it could introduce further uncertainty. Historical aerial images were used to verify the Cs-137 marker horizons. Photos were assessed from the most recent date to the oldest date. If the core location was in the channel in an aerial image, the date of the bottom of the core was adjusted to the date of the image, assuming that sediment deposition at the core location started soon after the date of the photo. The resultant chronology of floodplain sedimentation was used to: calculate the Sediment Accumulation Rate (SAR) (He and Walling, 1996); confirm recent river channel-flood bench connectivity; calculate the timing of changes in source material (Chapter 6).

Gamma vials were filled with <250 µm sediment to the well depth (4 cm) of the hyper pure germanium crystal of the Ortec gamma spectrometer. Samples were sealed with paraffin wax and left for 21 days in order for the unsupported Pb-210 activities to equilibrate with Rn-222 (daughter isotope of Ra-226) before being counted (Foster et al., 2007). A background spectrum was generated by counting an empty vial for 10 000+ seconds. The background was stripped from counted spectra and the resultant spectra were analysed using GammaVision[®] software. Emissions were counted for 180 000+ seconds to allow for the detection of low radionuclide activity at the 95% level of confidence. Detectible limits for Cs-137, Pb-214 and total Pb-210 were 0.8, 2.2 and 2.5 mBq g⁻¹ respectively.

Paleo-channels in exposed river banks and high lying terraces along the valley fill were up to 7 m above the current river level and thus assumed to be much older than the lower flood benches. These features were relicts of former valley fill levels; thus, if dated successfully, they would indicate when these features were abandoned or became sediment sinks (Rodnight et al., 2006). OSL dating was used on these features and proved useful as OSL dating spans a large temporal range and sediment from the Vuvu catchment contains sufficient quartz grains for analysis. As OSL dating was expensive, sample numbers were limited to five samples (two palaeo-channels, two terraces and one terrace/high flood bench).

The bottom of sand-filled palaeo-channels or the lower section of a sand deposit on a high lying cobble terrace was targeted for OSL sampling (Figure 5.2). Care was taken to sample layers that were clearly sorted and striated, thus being true fluvial deposits. The OSL sampling was done at night (using red light)

to prevent white light from bleaching the quartz grains. The vertical face, 15 cm above the cobble layer, was cleaned in order to remove any sediment contaminated by white light (removed a 5 cm deep layer of sediment from the vertical surface). A 30 cm horizontal core of sediment was sampled using an Eijkelkamp percussion corer and the sample was packaged in tough black light-tight packaging. A bulk sample was collected from around the cored hole for gamma analysis, magnetic characterisation, particle sizing and loss on ignition analysis. All samples were screened for the presence of Cs-137 or unsupported Pb-210 before they were sent to the Geo-luminescence Laboratory at WITS University for OSL dating. Samples containing Cs-137 or unsupported Pb-210 were younger than ca. 88 years and were thus excluded from OSL dating. One abandoned-channel (1 metre above the current channel) tested positive for unsupported Pb-210 and was omitted from the OSL analysis.

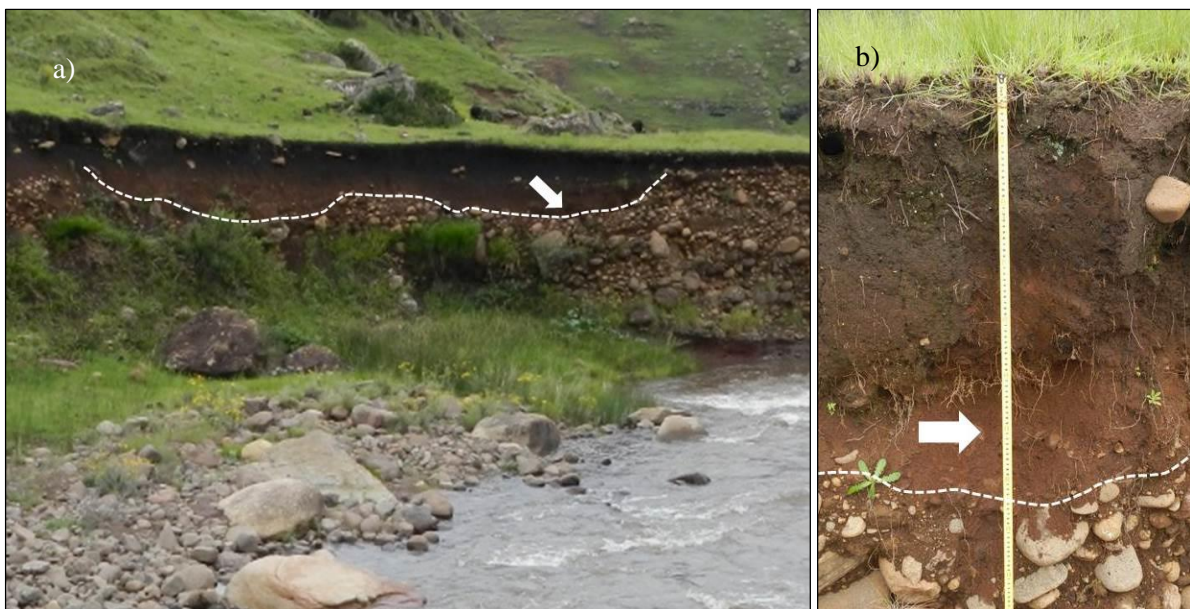


Figure 5.2: Photos showing the interphase between cobble and sand (dashed line) and the position of OSL samples (arrows) in a: a) palaeo-channel and b) terrace.

At the Geo-luminescence Laboratory 2 cm of sediment was removed from either end of each core for water content and dosimetry analysis for Thorium, Uranium and Potassium (gamma analysis done by iThemba Labs, South Africa) (Evans, 2014). OSL measurements were done on ca. 30 grains per sample, ranging from 180–212 μm , following the single aliquot regenerative protocol as given by Murray and Wintle (2003). Samples were treated with additional infra-red light before OSL measurement to reduce any contaminating signal from feldspar grains (Evans, 2014). The minimum-age model was applied with a sigma_b of 12% (Cunningham and Wallinga, 2010). Further details can be found in the OSL dating report by Evans (2014) in Appendix 2.

5.3. Results

5.3.1. Valley fill characteristics

5.3.1.1. Description of valley fill sinks

Sediment along the Vuvu valley fill was mainly stored in flood benches, terraces, alluvial fans and cobble bars (mostly containing gravel and boulders) (Figure 5.3; Table 5.1). As the valley fill was narrow (± 100 m wide including the channel), features were relatively small but well defined. The average slope of the channel was 0.016 and varied between 0.007 and 0.028 and was defined by Rowntree and Wadeson's (1999) classification as a mountain river (0.01–0.1) or foothill river type (0.005–0.01).

Table 5.1. Summary table of valley fill landforms.

Landform	Shape and dimensions	Position in landscape	Dominant sediment composition	Present day buffer function
Cobble bar	Elongate, 2-10 m long	In channel or bank attached	Cobble and gravel	Longitudinal connectivity
Flood bench	Relatively flat sediment deposits adjacent to the river channel, 0.5-3.5 m above thalweg channel, up to 100 m long	Located between valley margin or terrace and channel	Cobble base covered with fine grained material up to 1.2 m thick	Longitudinal and hillslope-channel connectivity
Terrace	Flat elevated alluvial levels, up to 7 m above thalweg channel and 200 m long	Located between valley margin and flood bench	Cobble base covered with fine grained material	Hillslope-channel connectivity
Alluvial fan	Conical shape, up to 40 m long	Forms where tributary loses confinement	Fine grained	Hillslope-channel connectivity

Instream deposits consisted mainly of cobble bars that were formed on inside bends and below deep pools (Table 5.1). The bars consisted of cobbles and boulders ranging from 0.1–0.5 m in diameter. Only minimal fine sediment was stored in these features and thus, these features were omitted as a significant fine grained sediment sink.

Flood benches were mainly narrow features lining the channel (Table 5.1). Two levels were identified, a lower level that ranged between 0.5–2.5 m and a higher level ranging between 1.5–3.5 m above the thalweg channel. Lower flood benches were mostly visible on one side of the channel only, suggesting that they build up as the channel slowly migrates across the valley fill, reworking the higher flood bench. The flood benches were composed of a cobble base with horizontal sand and silt layers (up to 1.2 m thick) that builds up over time as sediment is deposited during flood events. The flood benches were covered in short grazed grasses and were lined with sparse sedges closer to the channel's edge.

Terraces were well preserved and clearly visible along the upper reaches (VT9–VT15) of the valley fill (Figure 5.3). Along the lower reaches (VT1–VT8) terraces often were sloped, indicating that they were either alluvial fans (located where tributaries contributed sediment), a combination of alluvial fans formed

on or draped over terraces or fields that were created by ploughing steeper slopes. The description of the terraces was thus based on the upper reaches of the valley fill. Two levels were identified, a lower level ranging from 3–4 m and a second, higher level at 5–7 m above the thalweg channel. Along the upper reaches of the valley fill, remnants of a higher terrace were visible on both sides of the river. This higher terrace consisted mainly of a cobble base (up to 6 m high) with a sand deposit (up to 1 m thick) on top of the cobble. Mudstone bedrock was seen in sections under the cobble layer (up to 1.5 m above the present channel), suggesting that the river channel has been incised to a new level that is lower than the level that was present during the deposition of the cobble layer. Flood benches had a very similar composition to the terraces, but had a thinner cobble component. Both levels of terraces were covered with short grazed grasses throughout the year.



Figure 5.3: Downstream view of VT10. Note the lower flood bench on the left hand bank and the higher flood bench and higher terrace on the right hand bank.

Alluvial fans were small features that measured 10 to 40 m in length from apex to toe (Table 5.1). Smaller fans lacked channels, whereas larger fans had a channel that connected them to the main channel. The channel would often be incised, a process that remobilised some of the stored sediment. The fans mainly consisted of fine material (sand and smaller) and were found where steep tributaries cross the valley fill.

The cobble dominance of the channel, and formation of cobble bars indicated an active channel bed. Cobble bars form the base of the lower flood benches as cobble is deposited as the channel migrates across the valley fill. These cobble benches are further built up by finer sediment during flood events. Due to the elevation difference it would be expected that the lower benches are inundated more frequently than the higher benches that would only be overtopped during larger flood events.

5.3.1.2. Map of the valley fill sinks

It can be seen that the active channel occupies a significant area of the valley fill of the Vuvu valley (Figure 5.4). The river switches from the left to right hand bank on a regular basis in the upper and lower

reaches and on a less regular basis along the middle lower reach. Flood channels across many of the higher flood benches act as pathways for flood water during high magnitude events as has been observed after heavy downpours in March 2013 (video footage; Appendix 3). Freshly deposited sand was evident on the lower flood benches and along the flood channels during field surveys, indicating that these benches and channels were still active in storing and transporting sediment. No visual evidence of recent flood deposits has been found on any of the terraces. This would confirm that these higher levels have been abandoned and might only be inundated during extremely high magnitude events. The river was cutting into these higher levels on some of the river bends, reworking sediment stored in the terraces. Alluvial fans formed where steep tributaries met the valley fill. Sediment was deposited where the water's carrying capacity is reduced by the change in slope as it spills onto the relatively flat valley fill. The alluvial fans, formed by small tributaries, were disconnected from the channel in most cases by the valley fill. Where larger tributaries formed alluvial fans, a small incised channel linked the tributary to the river channel, reworking stored sediment and making these larger fans less efficient sediment sinks.

In terms of sediment storage, flood benches and terraces play a different role from that of alluvial fans. Terraces and flood benches mainly store sediment that has been transported down-river. The terraces have been disconnected from the main river through incision and are thus no longer receiving inputs through overbank flooding. Currently, the terraces are likely to receive sediment inputs from the adjacent slopes. The flood benches are still active in receiving and storing sediment that is transported by the river during high flow events. The alluvial fans, on the other hand, only store sediment that was transported by the tributaries. Where channels were present along the alluvial fans, the potential for sediment storage was reduced as the channels act as sediment pathways.

All these valley fill features are potential sediment sources as the river migrates across the valley fill and reworks the stored sediment. In Figure 5.4 it can be seen that the outside bends of the river were located along flood benches and terraces. Sediment stored in these features has the potential to be reworked on a regular basis. The alluvial fans, on the other hand, were located along the outer edge of the valley fill where the channel infrequently reaches, thus reducing the chances of sediment stored in fans to be reworked.

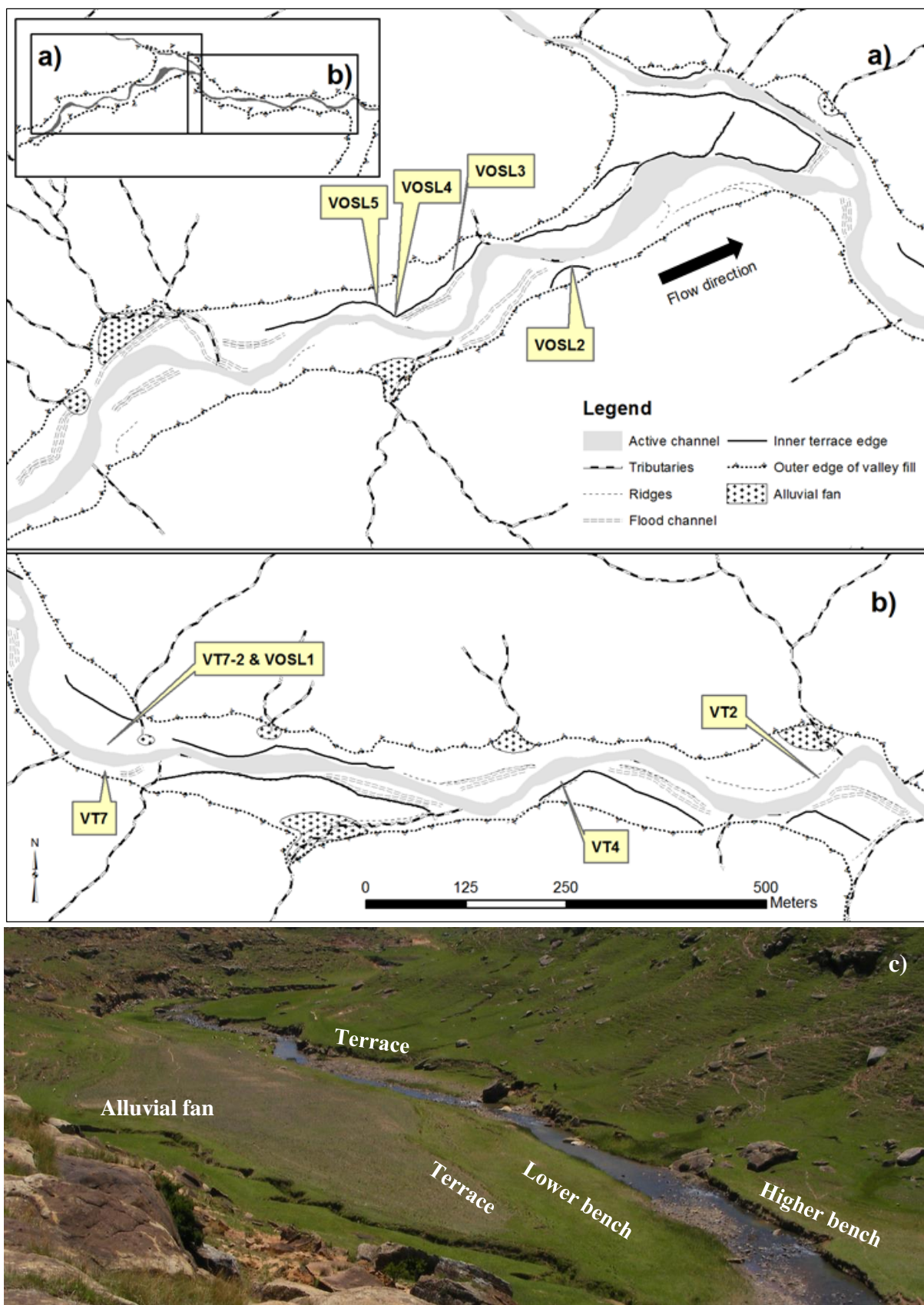


Figure 5.4: Plan view of the a) upper and b) lower Vuvu valley fill. Labels indicate the location of the four cores (VT) and five OSL samples (VOSL). Tributaries cutting across the alluvial fans indicate incised alluvial fans. c) Oblique photo showing an alluvial fan, terrace, high and low bench.

5.3.1.3. Valley fill sink composition

5.3.1.3.1. Flood benches

The cores were positioned to target two higher and two lower flood benches (Figure 5.4 and 5.5). VT2 and VT7-2 were located on higher flood benches and field evidence (e.g., debris and fine sand deposits) confirmed that these levels were still active. The lower levels, VT4 and VT7 were located on lower flood benches that were inundated more frequently. The lower flood benches had thick (up to 4 cm) fresh deposits of sand after the 2012/2013 wet season. This confirms that both levels are still active and that the lower level was more active and more frequently inundated during flood flows.

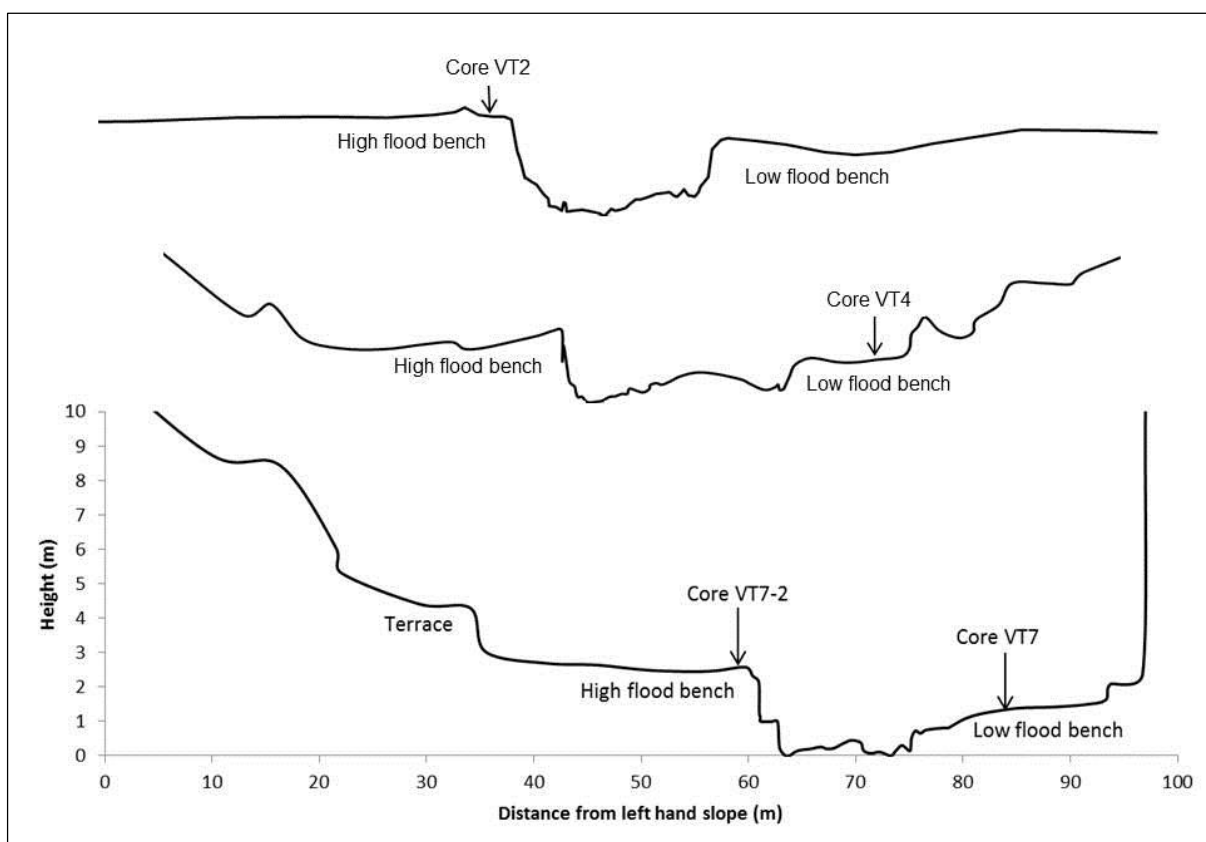


Figure 5.5: Transect VT2, VT 4 and VT 7 indicating the location of core VT2, VT4, VT7 and VT7-2. The horizontal and vertical scales are identical for all transects.

Median particle size (D50) for the core samples ranged between 58 and 343 μm (coarse silt to medium sand) (Figure 5.6). Sediment size varied throughout all the cores, indicating that sediment has been deposited during different events, each having a unique sediment composition, or that sediment size varies throughout a depositional event. Coarser layers that can be seen in Figure 5.6 indicate larger magnitude flood events that deposited larger particles. The expected fining towards the top was not clearly visible in most cores, except for core VT2. This trend would suggest that VT2 has lately only been inundated by shallow slow flowing flood waters that mainly transported finer particles. A similar trend was observed for VT7-2, although the coarser nature of the lower part of the core was missing (Figure 5.6). As both of

these cores were on the higher flood benches, it can be assumed that floodwaters still contributed sediment to these levels, but that these levels were covered by shallow floodwaters, mainly transporting fine sediment (fine sand and smaller).

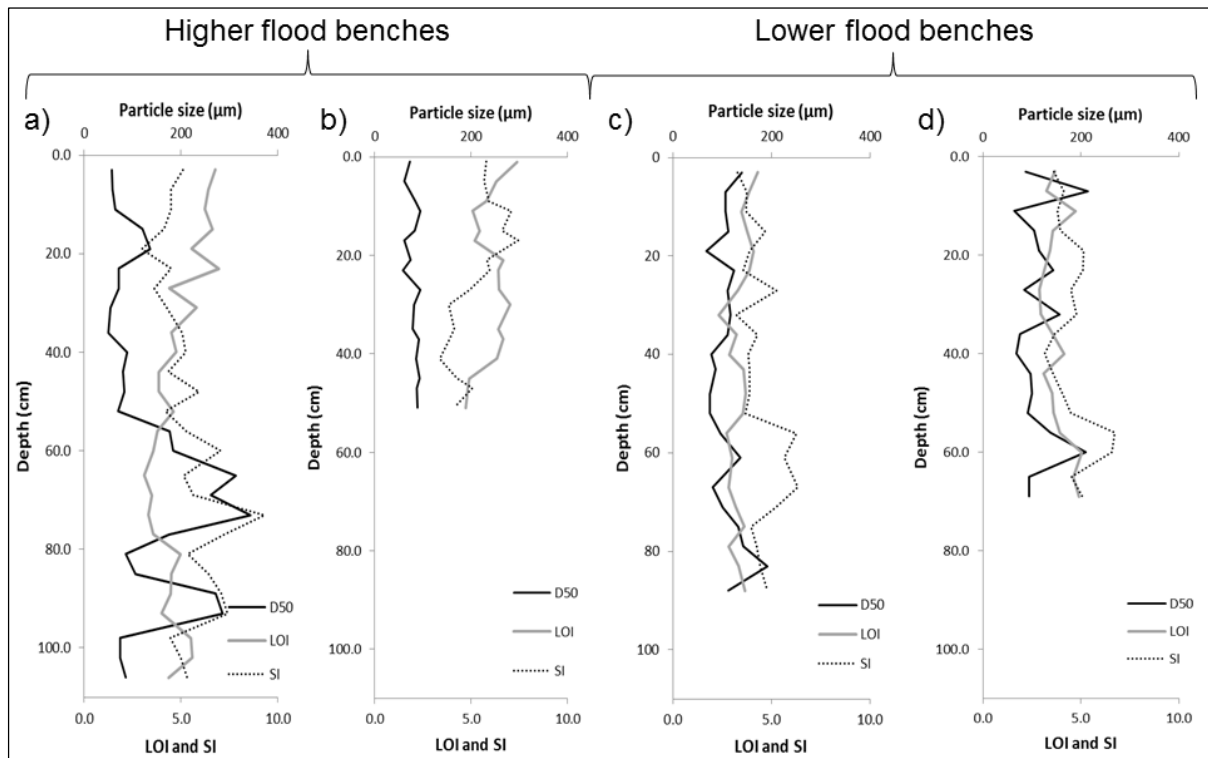


Figure 5.6: Particle size data (D50), loss on ignition (450°C) and Sorting Index (SI) data for core: a) VT2, b) VT7-2, c) VT4 and d) VT7.

Sorting Index values vary between 2.9 and 13.9, showing poor sorting ($SI > 2$) for all core samples (Andrews, 1983) (Figure 5.6). The degree of sorting varies throughout each of the cores, mostly showing better sorting for sections with larger median particle sizes. A weak positive correlation ($R^2 = 0.19$) existed between particle size and Sorting Index.

Down-core, it can be seen that organic content increased throughout the top sections of the higher lying flood benches VT2 and VT7-2 (Figure 5.6). This could indicate more stable conditions that favoured soil formation and subsequent plant growth and thus increased the organic concentration. The higher organic concentrations lower down these cores could be the result of similar stable conditions, or increased moisture availability during frequent overbank wetting, experienced during earlier stages when these flood benches formed. VT4 and VT7, which were located on the lower flood benches, had a lower organic content (3.4–3.8%) than the cores taken from higher flood benches and did not show an increase in organic content towards the top. This could indicate that these were relatively new features that were rapidly accumulating sediment and had not yet experienced stable conditions ideal for soil formation. The weak negative correlation ($R^2 = 0.13$) between organic content and particle size suggested that a higher organic content was hardly associated with a lower median particle size (Figure 5.6 and Table 5.2).

5.3.1.3.2. Terraces

The OSL samples had a relatively high organic content (4.5–7.3%), which were likely as these sediments were located at the base of sediment infilling old channels and cobble bars (Table 5.2). Wetter conditions during infilling would have permitted temporary wetland type conditions that would increase organic content as a result of vegetation growth.

Table 5.2: Location, loss on ignition, median particle size (μm) and Sorting Index data for the five OSL samples and average values for the cores.

Core/sample	Location	LOI%	D50	SI
VOSL1	Higher flood bench/terrace	4.5	344	14.0
VOSL2	Higher terrace	7.3	87	4.9
VOSL3	Higher terrace	5.8	156	4.3
VOSL4	Channel in lower terrace	5.1	455	5.2
VOSL5	Channel in higher terrace	6.9	83	5.8
VT2	Higher flood bench	4.8	133	5.4
VT7-2	Higher flood bench	6.0	81	4.2
VT4	Lower flood bench	3.4	109	4.4
VT7	Lower flood bench	3.8	114	4.9

The median particle size for the OSL samples ranged between 83–455 μm (very fine to medium sand) (Table 5.2). The terrace samples had larger particle sizes than those of the flood benches (based on average values), but this difference was to be expected as the OSL samples were from the bottom chronologies and were similar to bottom sections of the cores (Figure 5.6). Terrace samples were all poorly sorted, similar to that for the flood benches (Table 5.2).

In general the valley fill sinks had an overall low organic (LOI) content ranging from 2.3–7.3% (Figure 5.6 and Table 5.2). This indicates that these sinks were well drained and seldom experienced prolonged periods of inundation that would promote wetland formation.

5.3.2. Sediment sink chronology

5.3.2.1. Long term rainfall patterns: wet and dry periods

Long term rainfall records were analysed in order to determine the timing of dry and wet periods and to determine if these events relate to the chronology of sinks. It was expected that sediment dynamics would vary between wet and dry periods and that these cycles would be visible in the sediment records. Rainfall events proved to be highly localised with large differences in daily rainfall amounts over the 65 km distance between Matatiele and the Vuvu catchment. Daily rainfall data for Matatiele and the Vuvu catchment correlated poorly ($R^2 = 0.278$) over the period from December 2011 to January 2014. This highly localised rainfall can be shown by the event of the 17th of December 2012 when the Vuvu station

received 89 mm compared to the 0.7 mm measured for Matatiele. When total rainfall was considered, it can be seen that the Vuvu station received 50% more than the Matatiele station. When monthly data trends were considered for December 2011 to January 2014, the Vuvu and Matatiele data were reasonably well correlated ($R^2 = 0.796$). This indicated that the monthly data for the Matatiele site could be used as proxy for longer term trends in rainfall volumes. From the longer-term monthly rainfall data for Matatiele, the following wet periods were identified: 1940–1944; 1954–1964; 1985–1988; 1996–2002; 2004–2007; 2009–2013 (Figure 5.7). Long dry periods were identified from 1967–1973; 1978–1985 and 1989–1996.

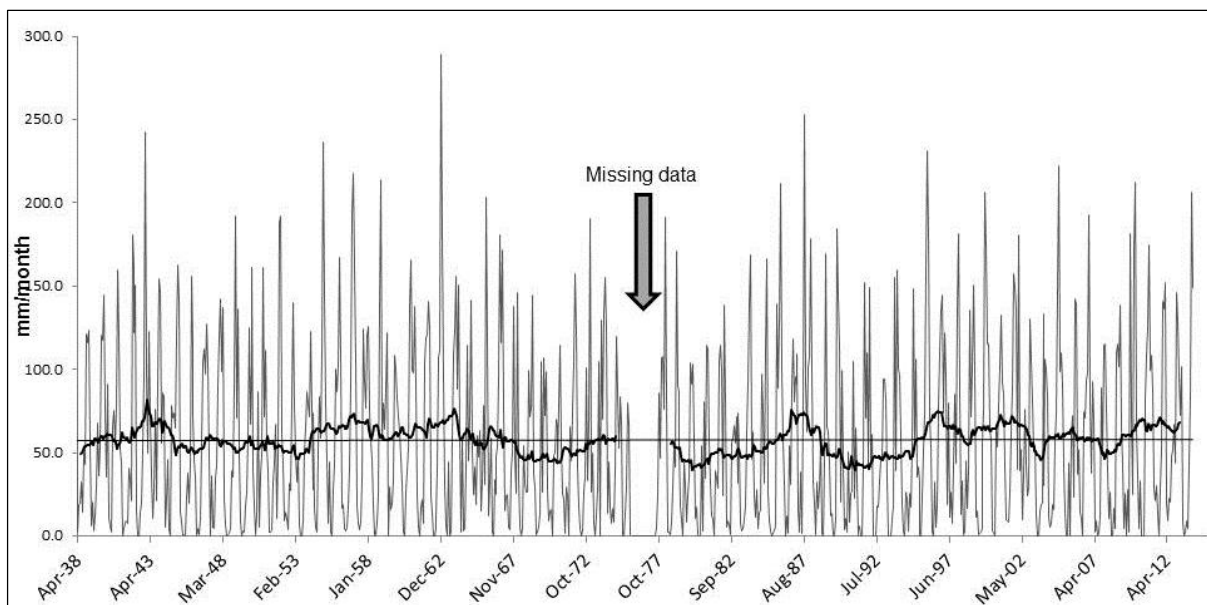


Figure 5.7: Monthly rainfall data for Matatiele from January 1938 to January 2014 (provided by Department Water Affairs). The black line is a 24 month running average, whereas the horizontal grey line is the average. Note the missing data from April 1975 to July 1977.

5.3.2.2. Evidence for river incision

From the historical images, river migration could successfully be tracked, and allowed field measurements to be made of river incision for the observed period (Figure 5.8 and Appendix 3). In Figure 5.8b it can be seen that the current channel of VT7 for 2012 is ca. 50 cm deeper than that for 1956. As this could be a local phenomenon, another suitable site (VT14) was surveyed 900 m higher up along the Vuvu River. The bed elevation difference for VT14 was 1.4 m (Table 5.3). This large difference in elevation could be related to the localised straightening of the channel and the removal of a highly sinuous section of channel or higher levels of incision experienced along the upper parts of the valley fill. Both the Vuvu and Phiri-e-Ntso Rivers experienced a lowering of the current bed elevation, compared to that of \pm 1956 (Table 5.3). Values ranged from 0.5–1.4 m (average of 0.95 m) for the Vuvu and 0.28–0.89 m (average of 0.54 m) for the Phiri-e-Ntso River. As resistant dolerite dykes cross both rivers at various places and should act as local longitudinal base level control points, it would suggest that incision took place along both rivers and, although to varied degrees, throughout both the valley fill reaches. Although

this observation was based on limited evidence, the parallel incision along both rivers would suggest that synchronous incision took place post 1956.

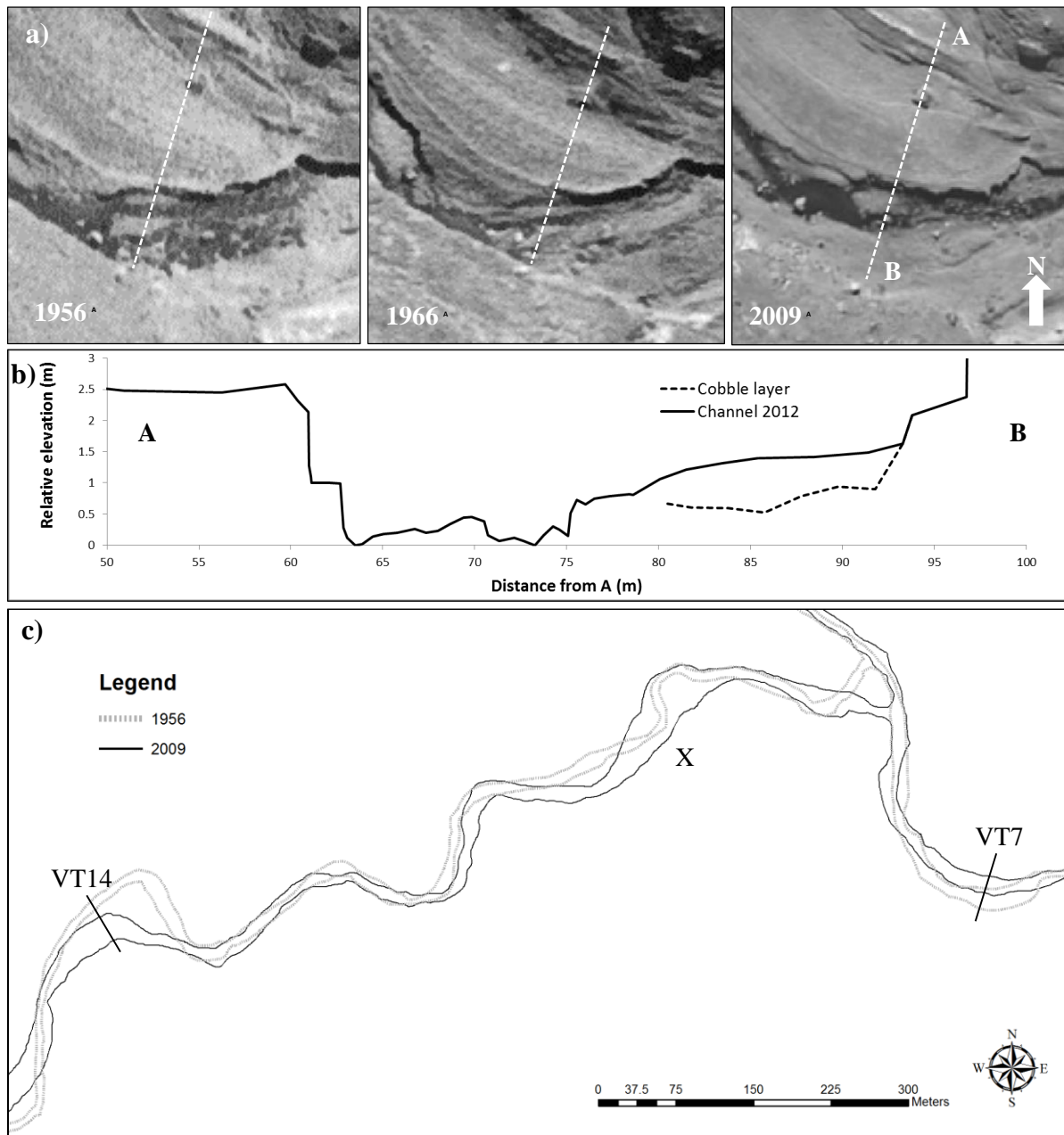


Figure 5.8: a) Aerial images for VT7 showing the movement of the channel in a northerly direction from 1956 to 2009 and the accumulation of sand on the southern lower flood bench (1966 to 2009), b) the cross section AB shows the channel profile for 2012 and the depth of the cobble layer in the southern flood bench, and c) shows how the channel has straightened from 1956 to 2009 (cross section VT7 and VT14 indicated). Note the channel widening at VT14 and X.

Both rivers experienced river straightening after 1956 (Table 5.4, Figure 5.8c and Appendix 3). The Vuvu's sinuosity decreased from 1.21 to 1.13 and the Phiri-e-Ntso from 1.48 to 1.32 over the period 1956 to 2009 (Table 5.4). The channel also has widened along sections (VT14 and X) as can be seen in Figure

5.8c. The decreased sinuosity is linked to incision and river steepening that will ultimately increase the transport efficiency of the channel.

Table 5.3: Summary table of bed elevation differences between 2012 and 1956 for the Vuvu and neighbouring Phiri-e-Ntso River

River	Cross section	Difference in bed level (m)	Average (m)
Vuvu	VT7	0.50	0.95
	VT14	1.40	
Phiri-e-Ntso	P1	0.65	0.54
	P2	0.28	
	P4	0.89	
	P7	0.33	

Table 5.4: River sinuosity changes between 1956 and 2009 for the Vuvu and Phiri-e-Ntso Rivers

River	Valley fill length (m)	Date	Length (m)	Sinuosity
Vuvu	1977	1956	2391	1.21
		2009	2241	1.13
Phiri-e-Ntso	905	1956	1340	1.48
		2009	1194	1.32

5.3.2.3. Cs-137 and unsupported Pb-210 dating of cores

Cs-137 values for the cores ranged between 0 and 4.3 mBq g⁻¹, with multiple peaks or detectable activities along each core, except VT7-2 which had a single peak (Figure 5.9). All cores except VT4 showed a clear zero Cs-137 section at the bottom of each core. The first occurrence of Cs-137 can thus be used as the 1958 marker with confidence, except VT4 and VT7, and was confirmed by the location of the core sites away from the channel since 1956, using historical aerial images. Core VT7 was in the position of the channel in the 1956 images, whereas the flood bench started forming in the position of core VT7 before the 1966 images, suggesting rapid sediment accumulation before the 1958 Cs-137 marker or low Cs-137 activity of the lower section of the core due to the coarse nature of the sediment, possibly sourced from a sub-surface soil or reworked channel sediment. The date of the base of core VT7 was adjusted to fit the aerial image date of 1956. This changed the date for the base of VT7 from 1920 (based on Cs-137 marker) to 1956. On the 1966 aerial photos the location of core VT4 was situated in the active channel, but in the 1977 images core VT4 was situated outside the channel. This would agree with the Cs-137 found near the bottom of the core, but the first occurrence of Cs-137 could not be used as a 1958 marker and the bottom of the core VT4 was adjusted to 1966.

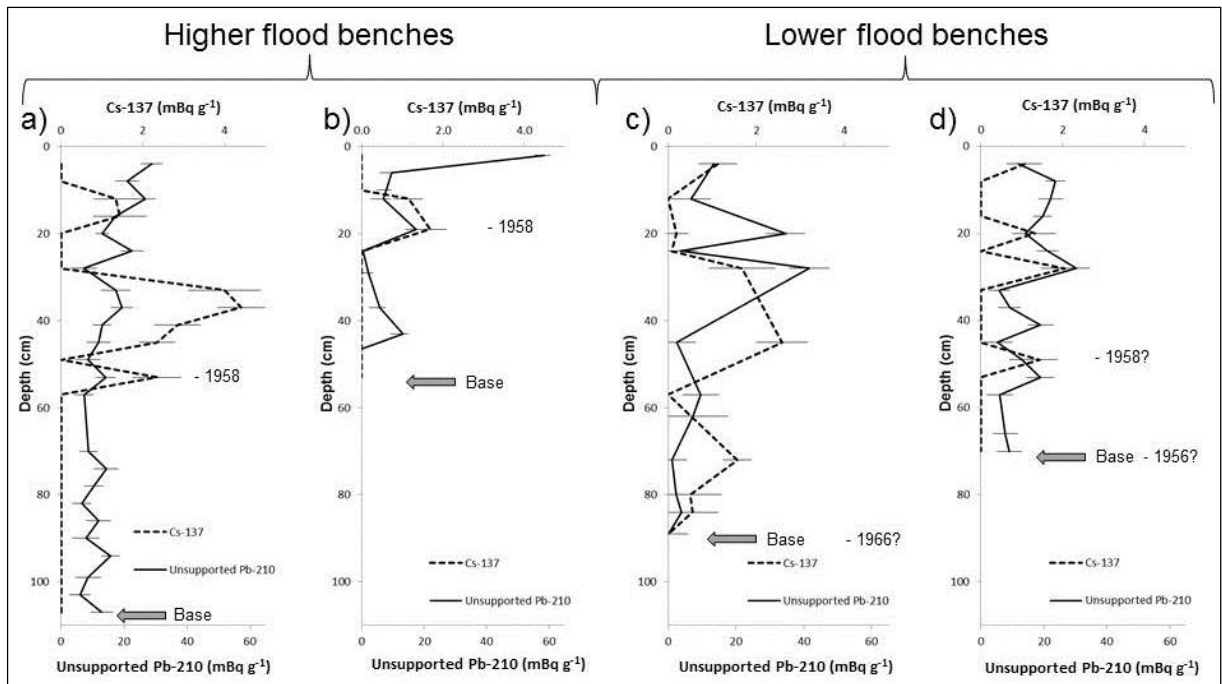


Figure 5.9: Cs-137 and unsupported Pb-210 data for cores: a) VT2, b) VT7-2, c) VT4 and d) VT7. Measurement error given by horizontal lines.

Unsupported Pb-210 values increased towards the surface for the higher lying cores, VT2 and VT7-2. This showed that sediment has been accumulating steadily at these higher sites, whereas the lower VT4 and VT7 sites showed large variations in values with minimal unsupported Pb-210 accumulation towards the top, suggesting rapid sediment accumulation. The dates produced for the higher lying cores estimated the initiation of accumulation as pre-1920 (Figure 5.10). The dates that were produced for all cores were in agreement with the historical images that were assessed.

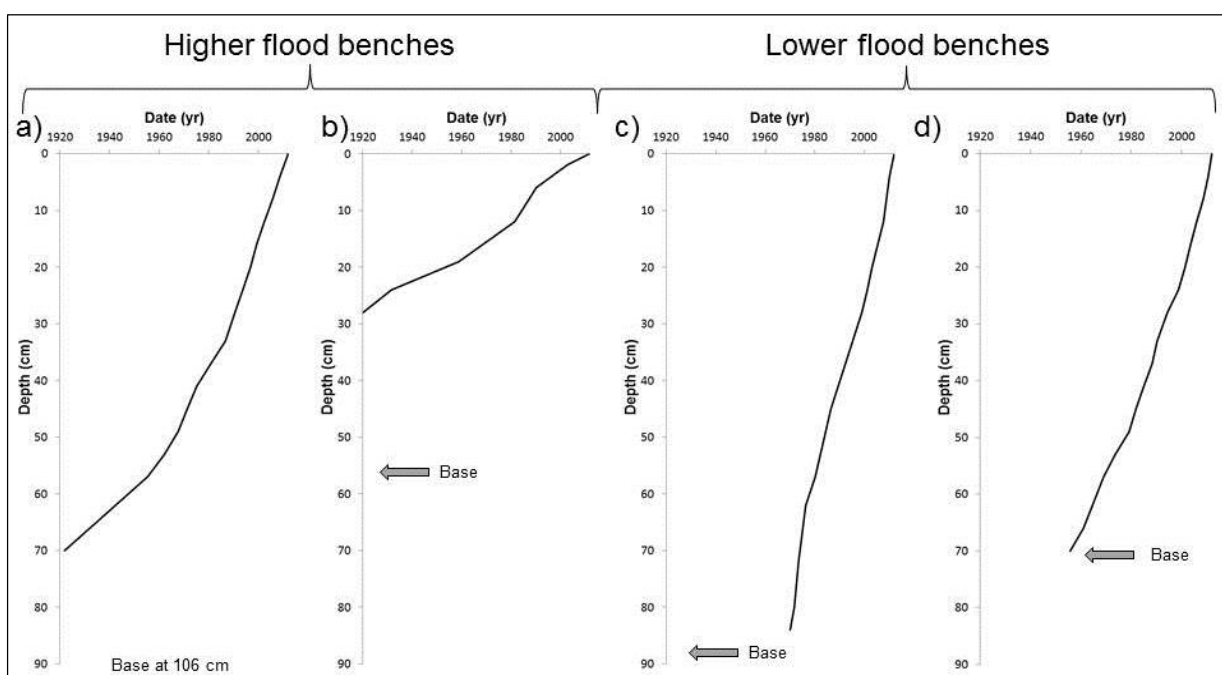


Figure 5.10: CRS dates for the cores: a) VT2, b) VT7-2, c) VT4 and d) VT7.

Sediment accumulation rates varied between 0.2 and 3.3 g cm⁻² yr⁻¹ for the higher flood benches and 0.6 and 11 g cm⁻² yr⁻¹ for the lower flood benches (Figure 5.11). The lower rates were observed for the higher lying cores and the higher rates for the lower lying cores. This would suggest that the higher flood benches: 1) were connected to the main river less frequently compared to the lower levels; 2) have accumulated during times where sediment deposition was less; 3) higher levels were inundated infrequently compared to lower levels, which would be related to the depositional environment (elevation relative to the channel or velocity of water), or 4) were connected when lower sediment concentrations in the water during flooding.

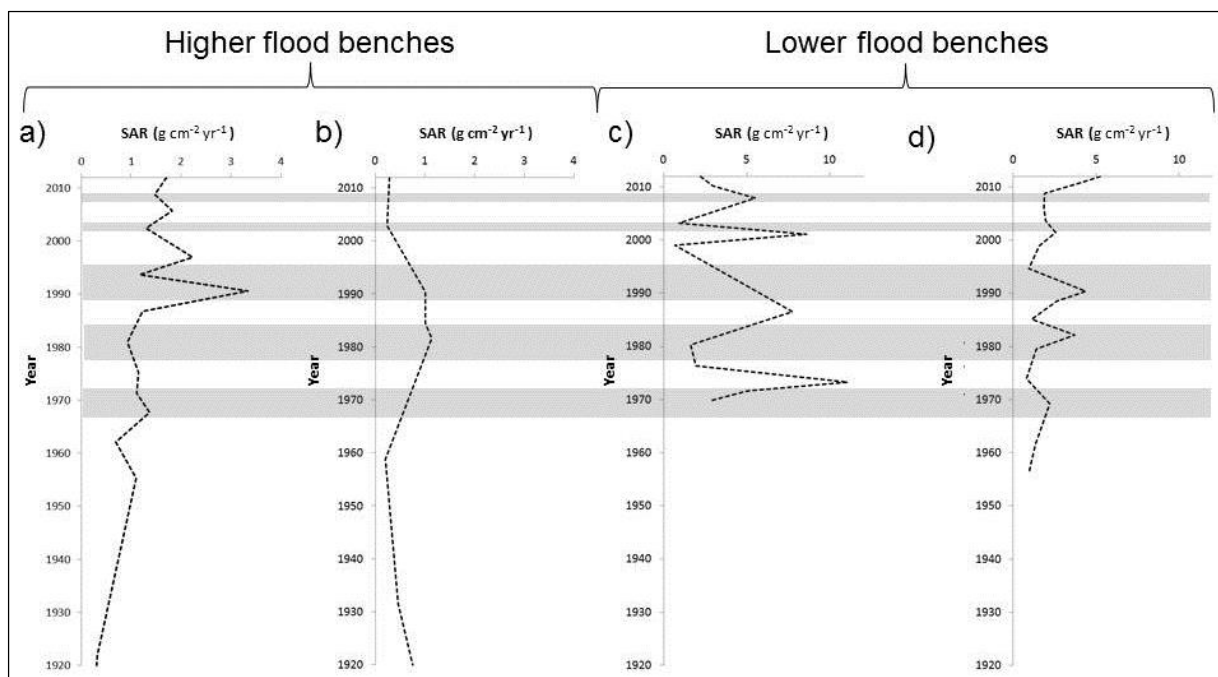


Figure 5.11: Sediment accumulation rates for: a) VT2, b) VT7-2, c) VT4 and d) VT7. Rates were based on the Pb-210 dating using the composite CRS model. Dry periods are shaded in grey.

The peaks in sediment accumulation rate do not correspond well between the cores which could be due to two main reasons: 1) that sediment accumulated at different times on different flood levels; or 2) it was difficult to produce detailed sediment accumulation rates using the composite CRS model. Linking the sediment accumulation rates back to the rainfall trends proved challenging as the rainfall data do not provide reasonable event data for the Vuvu catchment. The data do not indicate clear trends between sediment accumulation rates and rainfall, but if the peak sediment accumulation rates for core VT2, VT7-2 and VT7 are considered, it would suggest that sediment accumulation is highest during the dry periods.

Flood benches covered an area of 72 000 m² for the mapped valley fill. A depth of 0.8 m (based on the average core depth) was used as the average depth of fine sediment in the flood benches and equated to 58 000 m³ of sediment that has been stored. The dating of flood benches would suggest that these were

active sinks over the last 60+ years (higher benches are older, but could have acted as a sediment source, thus the average age was taken as 60 years). Mapped catchment-wide erosion peaked over the last 50 years, mobilising 2 000 000 m³ of sediment (Table 4.7). When these volumetric values were compared, it was calculated that the flood benches stored 2.3% of the sediment that was mobilised. These values exclude surface erosion and other sinks that might store sediment along the way to the valley fill.

5.3.2.4. Terraces and paleo-channels: longer term dating

Terraces and paleo-channels varied in age from 4 570 to 1 400 years before present (BP) (Table 5.5 and Figure 5.12). The dates follow chronologically except VOSL3 that is >1 590 years younger than the samples at a similar height above the active channel (>4 m). VOSL3 was excluded from further analysis as it is assumed to be in error, possibly due to light contamination or more recent materials that were introduced through bioturbation. The highest lying channel dated 4 570±490 BP (VOSL5) and the corresponding terrace dated 2 990±720 BP (VOSL2), both at a height of >4 m above the current active channel. The lower lying channel dated 2 100±1 550 BP (VOSL4) and the corresponding terrace dated 2 270±490 BP (VOSL1) and both were at >1.5 m above the active channel.

The channel sediment indicates the date when the channel was abandoned due to lateral migration or incision. Sediment could have been reworked prior to final deposition (as it was sampled), thus actual dates for abandonment could be older than determined by the dating. The base of the fine sediment on terraces indicates when that terrace was first active as a fine sediment sink. The following stages can thus be inferred from the dates (Table 5.5):

- The high level channel was abandoned some time before c. 4 570 (±490) BP.
- High level floodplain (>4 m) sediments were accumulating around c. 2 990 (±720) BP. Another 0.94 m of sediment was deposited after this date. The channel probably migrated laterally post 4 570 (±490) BP as the same level was actively storing sediment
- Incision of 2+ m occurred prior to 2 270 (±490) BP. This incision must have been relatively rapid as the high level terrace was still active post 2 990 (±720) BP.
- The lower level channel was abandoned and incised by 1.86 m post 2 100 (±1550) BP, but this date has a wide margin of error.

Table 5.5: OSL dates (BP) for terraces and paleo-channels (Appendix 2). Sample depth below terrace surface and sample height above the active channel is given.

Sample	Type	Depth below surface (m)	Height above active channel bed (m)	Age in years (std. dev.)
VOSL1	Terrace/flood bench	0.67	1.52	2 270 (490)
VOSL2	Terrace	0.94	4.60	2 990 (720)
VOSL3	Terrace	1.55	4.77	1 400 (750)
VOSL4	Paleo-channel	1.34	1.86	2 100 (1 550)
VOSL5	Paleo-channel	1.40	4.03	4 570 (490)

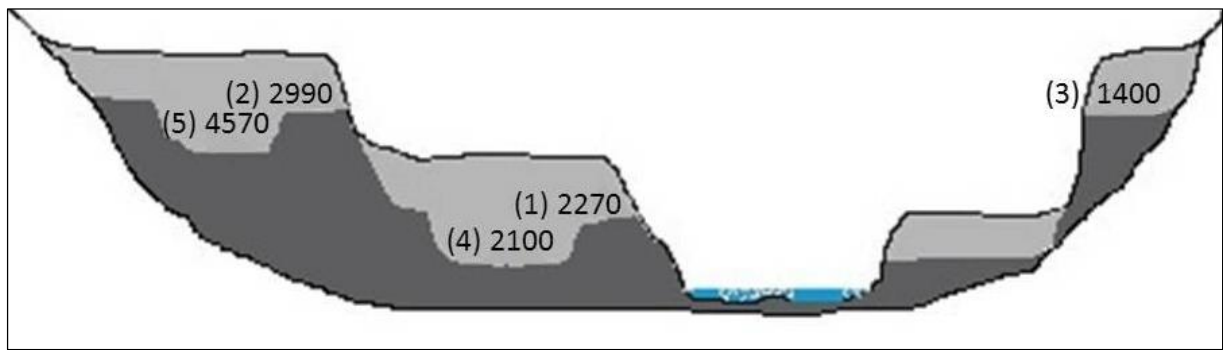


Figure 5.12: Schematic cross section of valley fill indicating the relative location of OSL samples (Sample numbers given in brackets). Dates are indicated for the various features and correspond to Table 5.5. Light grey indicates sandy deposits and darker grey cobble deposits.

5.4. Discussion

5.4.1. Function and spatial layout of valley fill features

The Vuvu valley fill had three main types of sediment sinks (flood benches, terraces and alluvial fans) that store fine sediment. These features influence how sediment moves to the channel and along the channel in terms of tributary-trunk, hillslope-channel and channel-valley fill connectivity (Brierley et al., 2006). By mapping the spatial extent of the various valley fill features, their likely functioning and roles in storing and contributing sediment could be deduced.

The Vuvu did not have an extensive valley fill, as was expected for a steep headwater catchment (Fryirs and Brierley, 2013). According to Rowntree and Wadeson's (1999) classification system the channel along the Vuvu valley fill is classified as a 'mountain stream' (based on the average channel slope). This did not fit what was observed along the Vuvu valley fill. A mountain stream was described as a bedrock and boulder dominated stream with cascades and bedrock steps, lacking a floodplain (Rowntree and Wadeson, 1999). When the range of channel slopes along the Vuvu fill was considered, sections fell into the 'river in the foothills' category which agreed with what was observed along the Vuvu valley fill. Their description of the foothill river was a mixed bedrock cobble bed with a pool riffle sequence and includes a narrow floodplain storing sand, gravel and cobble. This confirms the relatively small extent of the valley fill as there was limited accommodation space due to the confined valley. As the river switched from side to side along most of the valley, it created an opportunity for stored sediment to be reworked along the outside bends and transported sediment to be deposited on the inside of bends and on flood benches. Along the middle lower reach the straightened section would increase sediment transport and reduce the potential for sediment deposition, acting as a booster (Brierley and Murn, 1997), especially as it was confined by terraces on both banks. Channel straightening is often attributed to increased sediment load as it adjusts to a plan form that is more efficient in transporting bedload (Erskine, 1986).

Flood benches have shown evidence of active sediment deposition over the two year fieldwork period. Deposits were identified on both lower and higher flood benches, with thicker sandy deposits on lower flood benches compared to the thin layer deposited on higher flood benches. This contrast was a result of more frequent connectivity of the lower bench with the main channel, allowing a more frequent sediment input. These flood benches were thus active sediment sinks and acted as buffers that store sediment moving down the main channel for a period of time (Fryirs et al., 2007a). In places the stored sediment was remobilised where tributaries or the active channel cut into the flood benches. The likelihood of sediment stored in flood benches being mobilised was high, as the majority of the river was lined by flood benches.

Terraces lined the outer edges of the valley fill, with higher terraces along the outer edges and the lower terraces often extending closer to the main channel. The terraces were indicative of a previous active level, which has been incised and was no longer connected to the main channel. Their loss of connectivity also precluded them from functioning as sediment sources for the main river channel. At present the terraces can only receive wash from the adjacent slopes and act as a buffer between the slope and the river channel. The function of the terraces as a buffer has not changed, although the source of sediment has shifted from river dominated to adjacent hillslope dominated. Terraces were thus mainly sediment stores, except where sediments were remobilised by incision by tributaries or lateral bank erosion by the main river as was evident along the outsides of bends.

The alluvial fans were storing sediment that originates from sources along the adjacent hillslopes and were transported and deposited by the tributaries. The fans formed where tributaries crossed terraces and flood benches, possibly creating buffers across the valley fill that disconnected the tributary from the main river (Fryirs et al., 2007a). Where larger tributaries crossed the valley fill fans developed, but fans often appeared incised. Fan incision could be a result of a) over steepening of the fan surface due to rapid deposition or b) increased stormflow from the degraded catchment crossing a stability threshold that erodes the fan. The fan incision also could be a combination of the two factors as was seen with gully erosion in the catchment that was triggered by crossing a stability threshold related to a surface disturbance (e.g. trampling or ploughing) and increased energy from stormflow.

This incision of fans resulted in connectivity between the tributary and main river and reduced the potential of sediment to be deposited on the incised fan. Where the incision occurred, sediment that was stored in the fan and underlying terrace or flood bench was eroded and transported to the main river. In these cases the sink has turned into a source and has become an efficient conductor linking local slopes and the tributary to the main river. The smaller fans still acted as buffers, whereas the larger fans have lost that buffering function. As the fans were located along the outer edges of the valley fill, away from the active channel, their potential to contribute stored sediment to the main channel is limited, except where tributary channels cut through the fans (Fryirs et al., 2007b).

5.4.2. Composition of valley fill features

A mixture of fine and coarse sediment was stored along the Vuvu valley fill. The bed and lower banks were mainly composed of cobble and gravel, topped by fine grained materials. The Vuvu River can thus be classified as a mixed load river (Schumm, 1977). The coarser load would be transported as bedload along the channel by processes such as rolling, sliding and saltation (Schumm, 1977). As these larger particles (cobble) are close to the transport threshold, they are easily trapped and attract similar clasts to form a storage bedform such as a cobble bar (Fryirs and Brierley, 2013). These coarse particle storage units build up to a point where limited bedload sediment can be added due to hydraulic threshold constrains (Fryirs and Brierley, 2013). Finer sediment is then deposited on these coarser sediment stores by over-bench flooding, forming the cobble base flood benches visible along the Vuvu valley fill.

All flood bench cores showed variation in sediment grain size distribution throughout the cores. This indicated that sediment availability and flow energy varied during deposition throughout the sediment sequence of each of the cores (Tooth et al., 2013). The upwards fining trend in the higher level VT2 would suggest that channel-flood bench connectivity was waning as the flood bench built up vertically, reducing the chance of coarser particles being transported onto the flood bench. The generally fine nature of VT7-2 would suggest that some change in river elevation (incision) took place soon after the deposition of the cobble base as the coarser bottom section, visible in VT2, was not present. This abrupt change in elevation allowed only finer sediments to be transported and deposited due to hydraulic threshold constrains (Fryirs and Brierley, 2013). The cores of the lower flood benches consisted of a sediment grain size distribution similar to the bottom of VT2. The lower section of sediment in core VT2 was thus deposited when VT2 had a similar inundation frequency or elevation relative to the channel as for VT4 and VT7.

Furthermore, the poor evidence of upwards fining in cores of the lower flood benches would be related to the poor sorting of the sediment throughout the cores. Poor sorting throughout the cores was indicative of short duration high energy flows, entraining and dumping sediment over a short period of time (Fryirs and Brierley, 2013). As layers of larger particles in sediment chronologies were associated with higher discharge/inundation (Fryirs and Brierley, 2013), and lenses of coarse sediment were often better sorted, it can be inferred that larger events have a better sorting ability, which would be related to longer inundation times and sufficient entrainment energy to remove the smaller particles. This would be likely for the lower flood benches, or initial low elevation when features were formed, but was not always the case, pointing to sediment deposition being variable over time.

The flood benches had an overall low organic content which was related to the well-drained nature of the sediments and short inundation times that limited wetland formation. The higher organic content towards the top indicated lower sedimentation rates that were conducive to soil formation, whereas sections with a lower organic content were related to higher accumulation rates that limited soil formation. The observed relationship, that fine sediment generally has a higher organic content, would point to stable conditions (shallow inundation and limited sediment deposition) that favour soil formation or that coarser

sediment has either lost some of the organic material through winnowing, or that the coarse sediments were derived from sources with low organic content.

Terraces had a similar physical structure to the flood benches. A cobble base was covered by fine sediment. The median particle size, Sorting Index and organic content was similar to the lower sections of the floodplain cores, suggesting short high energy flows and frequent wetting and drying. Although limited samples were collected from the terraces, it would seem as if they were formed under similar conditions to the flood benches.

5.4.3. Rainfall trends and sink chronology

5.4.3.1. Rainfall trends

Rainfall data for Matatiele and Vuvu proved to be quite different in terms of daily rainfall and annual totals. The Vuvu rainfall station measured 50% greater annual volumes and was situated along the lower foothills of the Drakensberg Escarpment, whereas the Matatiele rain gauge was situated 20 km away from the Drakensberg Escarpment. The higher volume of rainfall produced for the Vuvu station can be ascribed to the orographically produced thunderstorms that are typical along the foothills of the escarpment (Nel, 2008). The Matatiele station would miss the majority of the rain produced by these orographically produced events, due to its distance from the escarpment, which explains the lower annual rainfall. The mismatch in daily timing would largely be related to the localised nature of the orographically produced rainfall. Daily rainfall was only synchronised when extensive rain events were produced.

Monthly rainfall data showed similar trends for the Matatiele and the Vuvu catchment and proved to be a reliable indicator of long term wet and dry cycles. Cycles varied in timing and intensity with clear wet and dry cycles that could be compared with patterns in the sediment chronology. The wet and dry cycles detected for Matatiele corresponded to those given by Nicholson (2000) for southern Africa (record overlap 1938-1994), indicating that the long-term rainfall trend was regional and was likely to be the same for the Vuvu catchment.

5.4.3.2. Channel incision

Historical aerial photo analysis combined with cross sectional surveys proved successful to assess the incision that has taken place in the Vuvu and Phiri-e-Ntso Rivers. As incision occurred over the wider area of both river systems it would suggest that input–output relationships have been altered for the area and that it is unlikely that the incision was related to a localised extreme event (Brunsden and Thornes, 1979). The change in input could be related to changes in climate, vegetation and rainfall-infiltration-runoff relationships that influence the flow and sediment regime or changes in base level (Knighton, 1984; Schumm, 2005). Increases in the flow regime were likely as hillslope-channel connectivity has been

increased (Chapter 4). Booth (1990) measured similar levels of incision (0.2 m) in the Soos Creek, western Washington, USA, and related it to increases in discharge as a result of increased runoff efficiency.

A decrease in sinuosity is often associated with increases in sediment load (Knighton, 1984; Fryirs and Brierley, 2013), river widening and river steepening (Rosgen, 1996). The straightening of the Vuvu and Phiri-e-Ntso channels could be related to a combination of the mentioned factors, with increased sediment loads being the most likely variable responsible for the river straightening due to increased hillslope-channel connectivity increasing sediment delivery to the channel. The widening of the channel would suggest a shift towards a more braided character that is either associated with an increased coarse sediment load (Knighton, 1984) or previously buried coarse material that is exposed by channel incision with the channel currently flowing around the exposed bars of coarse material.

5.4.3.3. Flood bench sediment dynamics and connectivity

5.4.3.3.1. A critical appraisal of the suitability of the flood benches for dating using the composite CRS model

The composite CRS model produced dates and sediment accumulation rates that showed variability as would be expected for flood benches with varying degrees of connectivity over time. The emanation coefficient of 0.3 as used by Du and Walling (2012) proved effective, and made the use of the composite CRS method possible in these flood benches. A combination of Cs-137 and historical aerial photos were used to calibrate the age of the model. Conditions for using the method were not ideal as Cs-137 activity was relatively low, as would be expected for the Southern Hemisphere (Foster et al., 2007). The use of photos proved invaluable for cores VT4 and VT7. If the model only was based on the first occurrence of Cs-137 from the base upwards, the dating at the bottom of cores VT4 and VT7 would have been inaccurate by more than 8 and 36 years respectively. The calibration date for that slice was changed from 1958 (based on first occurrence of Cs-137) to 1966 +, as the slice below the Cs-137 marker slice was given the date 1966 based on historical aerial images. Similarly for VT7 where the 1958 marker was two thirds down the core, the date for the base was calculated as 1920. Aerial images proved that the base of core VT7 was closer to ~1956. This shows the importance of using historical aerial images when using Pb-210 and Cs-137 to date flood benches.

Unsupported Pb-210 concentrations for the higher flood benches increased toward the top of the sequence as was expected for undisturbed sediment chronologies that accrued steadily during the accumulation phase. The higher fluctuation in unsupported Pb-210 values for the lower flood benches show the rapid accumulation of sediment as was expected due to more frequent channel flood bench connectivity. Extrapolating the dating beyond 1958 proved unreliable as was seen for VT7 on the lower flood bench (bottom of the chronology was initially predicted as 1920, although aerial images confirmed that the core

site was located in the active channel in 1956). This shows that historical aerial images proved invaluable to adjust and verify composite CRS dates and that dates extrapolated beyond 1958 should be used with caution.

5.4.3.3.2. Flood bench sediment dynamics and connectivity

The sediment sequences on the higher flood benches were 92 + years old compared to the lower flood benches that were 46–60 years old. The accumulation rates for the higher lying flood bench VT7-2 showed low accumulation rates that reflected infrequent channel-flood bench connectivity. The increase in SAR for the higher level cores since 1958 was not to be expected as connectivity usually becomes less frequent as a flood bench is built up vertically. This increase in SAR towards the top could be the result of increased connectivity between the channel and the higher flood bench (increased stormflow over the last ~50 years), higher sediment concentrations during over bench connectivity over the last ~50 years, or the model being sensitive to coarser sediment layers, skewing the results.

Sedimentation trends also varied for the lower flood benches, with an overall increase in SAR as the lower flood benches were built up, indicating an increase in over bench connectivity and sediment deposition. This was not to be expected for the connectivity reasons mentioned earlier.

SAR rates post 1958 differed between the lower and higher flood benches. The average SAR rate for higher flood benches was $0.9 \text{ g cm}^{-2} \text{ yr}^{-1}$ compared to $3.2 \text{ g cm}^{-2} \text{ yr}^{-1}$ for the lower benches. This reflects the greater chances for sediment to be deposited on lower flood benches compared to higher flood benches (Erskine and Livingstone, 1999). If the lower benches are ± 50 years old it suggests that incision took place approximately 50 years ago and was confirmed by the aerial photo and cross sectional surveys (Section 5.3.2.2.). This means that the lower section of the high benches would have been equal to the present lower benches and would have been inundated more frequently. Yet the SAR is higher post 1950 according to Figure 5.9. This indicates significantly higher sediment loads post 1950 and is in agreement with the formation of erosional features (e.g. sheet erosion, gullies, etc.) and increased pathways (gullies, livestock tracks, etc.) that is linked to the greater amount of anthropogenic activity over the past 50 years.

The sediment accumulation rates amongst the individual cores did not match up temporally and made it difficult to compare to the identified wet and dry periods. This mismatch could be due to dating inaccuracies, the variance in connectivity and conditions that were optimal for sediment deposition, or it could be that a section of the flood bench, above the section used as a marker horizon, has been scoured during localised cut and fill cycles of the flood benches which would lead to the ‘stretching’ of the sediment chronology that was subsequently deposited on the flood bench. The observation that sediment accumulation rates appeared highest during dry periods would agree with observations where sediment erosion is increased as vegetation cover is reduced during dry periods, exposing soils to erosive rainfall (López-Bermúdez et al., 1998; van der Waal et al., 2012). Increased quantities of sediment could be available during dry periods, with large quantities of sediment possibly being deposited on flood benches during large infrequent flood events.

When the overall percentage of sediment mobilised in the catchment was considered and compared to the volume that was stored in the active flood benches, it was clear that a small proportion (~2%) of the sediment transported by the river was stored in the valley fill. This showed that the majority of the sediment was exported from the catchment as was expected for upland catchments that have high energy flow conditions and limited sediment storage space (Rowntree and Wadson, 1999; Fryirs and Brierley, 2013).

5.4.3.4. Terrace dynamics

The dates produced for terraces and palaeo-channels were up to two orders of magnitude older than those for flood benches. This highlights the long period over which the Vuvu valley fill has been active in storing sediments. Although there were large error margins on the OSL dates produced, in cases up to 74% of the date, it indicated that the terraces were active between 2 100 ($\pm 1 550$) and 4 570 (± 490) years ago. The dates suggested an overall phase of valley fill incision since ~3 000 BP, probably interspersed with periods of limited valley filling.

Schumm (2005) related incision to increases in discharge or decreases in sediment load or a lowering of base level or a combination of the three. Climate and sediment dynamics were considered by looking at longer term records of these two variables for South Africa. Base level changes could play a role in the Vuvu River, but no evidence or literature could be found that documented sudden changes in base level at high altitude.

Various methods have been applied to reconstruct historical climate data for southern Africa. The 200 000 year rainfall record, based on rates of salt formation in the Pretoria saltpan, suggests that rainfall increased over the last 10 000 years (Partridge et al., 1997), but this record is not detailed enough for in-depth analysis of more recent trends, but would suggest wet conditions that would favour sediment transport. Ramsey and Cooper (2002) combined sea level evidence for South Africa that can be used as an indication of past climate. The record shows that sea levels were higher for periods ca. 5 080–4 240, ~3 820, ~3 740 and 1 610–1 450 BP, with highest levels experienced during the two last mentioned periods (+1.5 to +3.5 m). Levels were similar to present sea levels for the periods in between the high sea level stages according to the available records (Ramsay, 1995; Ramsay and Cooper, 2002). Sea levels were below current levels before ca. 6 460 BP (Ramsay, 1995). Higher sea levels were indicative of a warmer climate when sea water expands and ice melts. The warmer wetter periods would be associated with increased vegetation cover and reduced erosion, leading to valley fill degradation. These warm periods from ca. 5 080 BP would agree with the incision of the Vuvu valley fill over the last ~3 000 years.

Work done on the Okhombe valley, KwaZulu-Natal found that sediment that was produced during the Last Glacial Maximum was fluviually redistributed during warmer and wetter early Holocene (10 000–5 000 BP) (Temme et al., 2008). Due to the surplus of sediment, deposition occurred where river slope

relaxed. This can be linked to the formation of the highest terrace level along the Vuvu River. During the later Holocene sediment supply was reduced, possibly due to increased vegetation cover on hillslopes, and marked the onset of mass valley evacuation (Temme et al., 2008). This onset of the incision phase around the late Holocene (5 000–1 800 BP) would agree with evidence from the Modder River, Free State (Tooth et al., 2013). This erosional phase would confirm that the palaeo-channels and higher and lower terraces of the Vuvu valley fill were formed, and abandoned in the case of the higher terrace and channels, during the time 4 570 (± 490)–2 100 ($\pm 1 550$) BP.

Lyons et al. (2013) found evidence along the Blood River, KwaZulu-Natal for a second erosional phase that was ascribed to abrupt climatic changes associated with the warmer period of the Medieval Climate Anomaly (~1 060-760 BP) and cooler phase of the Little Ice Age (~510-210 BP). This erosional phase along the Blood River could be linked to the second incisional phase along the Vuvu River marking the abandonment of the lower terrace and incision to the current active valley fill level.

The physical properties of the upper sections above the cobble base of terraces were similar to those of the lower chronology of current day flood benches, pointing out that the sediment available 4 570 (± 490) years ago was of a similar size and as poorly sorted. Although the climate varied over this time period, with periods of accelerated sediment export from the valley bottom during wet periods, overbank events and sediment available were similar in nature to those responsible for recent flood bench formation.

5.4.3.5. Tributary-trunk connectivity over time

From the evidence presented above it would seem that tributary-trunk connectivity has been increased over the past 5 000 years. The higher valley fill level (to the level of the terraces) ~5 000 years ago potentially allowed for larger alluvial fans to have formed due to the wider and generally flatter valley fill, possibly with a river with a shallow braided flow pattern. These large fans could potentially have disconnected the sediment supply of the hillslopes from the trunk stream. As the climate and sediment supply stabilised, incision was potentially triggered, cutting into the valley fill. This steepening of the fan toe could potentially trigger local fan reworking, contributing stored sediment to the trunk stream. The length of the fan was likely shortened with reductions in hillslope-channel buffering capacity. The exact time of fan incision is unknown, but it is likely that headcuts formed in the fans due to the incision of the main channel and increased runoff from the hillslopes due to anthropogenic activities. Small isolated alluvial fans were not incised and still act as a buffer between tributaries and the main channel, but the incision of the larger fans increased the tributary-trunk connectivity.

5.5. Conclusions

Sediment was stored in the Vuvu River valley fill mainly by two processes: 1) the main process being channel-flood bench sediment transfer (connectivity) and 2) hillslope-valley fill sediment transfer (connectivity). The first process was responsible for the deposition of the majority of sediment in storage

and relates to the terraces and flood benches. From the composition of these storage features, it was clear that the Vuvu River has been a mixed load river transporting cobble and gravel as bedload and sand and finer sediment as suspended load (deposited on flood benches). The valley fill was typical of a river along the foothills with a narrow valley fill consisting mainly of cobble and gravel on the bed and base of the valley sink features. Sand and silt carried in suspension was deposited on the cobble bars, building up over time to form flood benches. The valley fill went through a phase of incision (since ~3 000 BP) that disconnected a former active fill level from the channel, forming terraces with a similar topography to the current flood benches.

The composition of the terraces was similar to that of the current flood benches and thus it can be concluded that processes responsible for the formation of the terraces were similar to what we see today in the active channel.

The valley fill caused tributary-channel disconnectivity and acted as a buffer that stored sediment transported by tributaries to be deposited as alluvial fans. Where tributaries were large and had enough energy to cut a channel through the valley fill, they became reconnected with the channel. This channel formation along the fans reworked sediment stored in both the fan and valley fill and increased the potential for sediment to be transported to the main channel. The valley fill also acted as a buffer that prevented slope-channel connectivity along the margin of the valley fill.

River migration and incision reworked sediments stored in flood benches, terraces and fans on the outsides of river bends. This allowed stored sediment to be transported downstream. Sink features distant from the active channel are less likely to be re-worked in the near future, compared to features adjacent to the active channel. Terraces and alluvial fans on terraces are thus seen as longer term sediment sinks, compared to flood benches that are shorter term sediment sinks. Dates produced for the terraces and flood benches suggest that flood benches store sediment for up to 100+ years, whereas terraces can store sediment for up to ~4 500 years.

Channel incision and straightening of the Vuvu River is likely to be related to increases in hillslope-channel connectivity that increase peak discharge, sediment supply from hillslopes and sediment entrainment capacity in the channel. The effect of the channel incision and straightening would be increased sediment transfer through the system, changing the function of the valley fill from a sediment buffer to a booster.

The composite CRS method combined with an emanation coefficient of 0.3 proved to be effective for dating the flood benches, although marked with high error margins. Historical aerial images proved invaluable for interpreting Cs-137 marker horizons that were used to calibrate the composite CRS model. When the composite CRS model is used to date flood benches, dates older than the 1958 Cs-137 marker should be used with caution and preferably verified by field or photographic evidence.

Although the Vuvu valley fill is currently incised with reduced sediment storage potential, it is still active as a sediment sink, storing sediment transported down river in flood benches and sediment transported from localised tributaries and slopes on terraces and higher flood benches, often as alluvial fans. A large proportion of lateral sediment input was stored by the valley fill (although small if considered at a catchment level), compared to the small portion (~2%) of sediment stored by the flood benches of sediment transported down river. The valley fill can thus be described as an important sediment buffer for localised hillslope-channel connectivity and as a lesser important buffer for catchment-wide longitudinal connectivity.

Chapter 6: Suspended and valley fill sediment sources

6.1. Introduction

Sediment connectivity was used as the framework to study sediment dynamics in the Vuvu catchment. Sediment connectivity focuses on the flow of material from one component of the landscape to another (Brierley and Murn, 1997; Croke et al., 2005; Bracken and Croke, 2007; Fryirs et al., 2007a, 2007b). This transport between components is dependent on pathways to make transport possible or more efficient. Assuming sediment that is transported from source areas to sinks would retain its physical character, it could be possible to trace sediment stored in sinks (Collins et al., 1997a; Hancock and Pietsch, 2008; Rowntree et al., 2012). Sediment accumulation on flood benches and flood plains takes place by vertical accrual over time, thus sediment tracing can be used to construct a record of sediment origin to assess changes in spatial and temporal connectivity between sources and sinks (Collins et al., 1997a).

Sediment fingerprinting is a well-established method which has proven its value in revealing the dynamics of catchment erosion processes and downstream delivery of sediment (Walling et al., 1979; Walling and Woodward, 1995; Collins et al., 1997a, 1997b; Foster et al., 1998; Wallbrink et al., 1998; Carter et al., 2003; Collins and Walling, 2007; Foster et al., 2007; Collins et al., 2010; Manjoro, 2012; Fryirs and Gore, 2013; Walling, 2013; Wilkinson et al., 2013). The method has been applied to a range of catchment types ranging from natural to agricultural to urban at various spatial scales (from plot to basin scale). Temporal scales vary from within event to hundreds of years. The method aims to identify and estimate the percentage contributions from various sources documented in the catchment, thus the researcher has to collect source and sediment samples, analyse the samples for tracers and statistically select a combination of tracers able to differentiate between sources, before unmixing the contributions. In a review of sediment tracing Walling (2005) identified various tracers that have been effectively used, namely: geochemical properties, mineralogic properties, mineral magnetic properties, sediment colour, plant pollen, fallout radionuclides and isotopes. The choice of tracers used is mostly influenced by the question at hand, e.g. Cs-137 for discriminating surface and subsurface sources, resources available and the number and complexity of sources.

The Vuvu catchment has had two main land use types over the past 100 + years, the upper catchment has been used mainly for grazing and the lower catchment used mainly for housing, cultivation and grazing. From Chapter 4 it is clear that the lower catchment has been affected most by land use change and increased hillslope-channel connectivity and thus could be a significant source of sediment. As the mineral magnetic signature of the parent material of the upper catchment (basalt) would be expected to differ considerably from that of the lower catchment (sand and mudstone) (Oldfield et al., 1979; Lees, 1999), mineral magnetic signatures were used in combination with Cs-137 for sediment tracing (Oldfield et al., 1979; Lees, 1994; Foster et al., 1998; Lees, 1999; Foster et al., 2007; Manjoro, 2012). Magnetic characterisation is a relatively quick, cost effective and non-destructive method (Oldfield et al., 1979;

Walling et al., 1979) that should produce good discrimination, given the contrast expected between the igneous and sedimentary sources.

In this chapter the focus is on identifying the sources of suspended sediment and sediment stored in various valley fill features to determine if there has been a change in source over time that might be related to changes in connectivity. The majority of the sediment is transported during high magnitude events (Gonzalez-Hidalgo et al., 2013) that should leave traces of the dominant source of sediment on higher lying flood related features, such as flood benches. As these valley fill features have been dated (Chapter 5), potential changes in sediment source and the timing of the change can be related to changes in landscape connectivity. Modern day source dynamics in the Vuvu catchment can be derived from suspended sediment samples, whereas historical source information can be derived from valley fill features, such as fluvial terraces, that are indicative of changing erosional stages (Pavlopoulos et al., 2009). Terraces and flood benches keep an archive of more recent dynamics of sediment deposition or channel-flood bench connectivity. The objective of this chapter was to determine the source of sediment stored in sinks (Objective 5). By tracing the source of the various features and linking it to the chronology (Chapter 5) we can determine how and when sediment dynamics, and connectivity, have changed.

6.2. Methods

6.2.1. Source and sediment sample collection and processing

Sediment sources were classified into three groups based on parent material, e.g. basalt (Drakensberg Formation), sandstone (Clarens Formation) and mudstone (Elliot Formation). Each group consisted of samples collected from gully, sheet erosion features, landslides and hardened roads. Both surface and subsurface samples were collected for these features, except for roads (only surface samples). A total of 100 sites were targeted for source sampling as this would allow a manageable number of samples per geological area (limited resources and challenging terrain). Stratified random sampling was used as it would preserve the ratio of geological and slope proportions that were calculated in Chapter 4 and allowed for random sampling within each of the groups (Kheir et al., 2007). The largest proportion of samples was thus located on moderate slopes and the Drakensberg Formation, respectively as presented in Table 6.1. The low numbers for the Clarens Formation could have implications for the accuracy of source tracer description.

All mapped gully, sheet erosion and landslide features were converted to points based on the centroid of each feature. These points were merged to create a single point shapefile. The attribute data for the point file were used to sort the samples according to their groupings as in Table 6.1. The cases within a group were given numbers and the target number of features was selected using the random number generator in Microsoft Excel. The selected points were uploaded on Garmin GPSmap devices and were used to

navigate to the sites. After a visual inspection of a feature, a sample site was chosen that would be representative of the feature in terms of depth and soil colour. Five of the higher sites on the Drakensberg Formation could not be reached due to potentially dangerous and challenging terrain.

Table 6.1: A matrix showing the sampling groups (based on parent material and slope) and number of samples per group that was used for the random site selection.

	Formation		
	Drakensberg	Clarens	Elliot
Steep	17	4	14
Moderate	25	5	20
Gentle	7	2	6

A stainless steel (non-magnetic) corer (2.5 cm inside diameter) was used to collect source samples, as the same depth (10 cm) could be cored for all samples and gave a sufficient volume of soil for analysis. A vertical surface core was extracted away from the edge of the erosion feature where there was dense grass cover. It was assumed that the grass cover was indicative of a relatively undisturbed surface soil. This sample would represent the surface material that was eroded from the erosion feature and the general surface material of that area. A horizontal subsurface sample was taken half way to the bottom of a feature. For large features (deeper than 1.5 m), several subsurface samples were extracted at one metre intervals. A total of 200 samples were taken, 93 surface and 107 subsurface.

Samples were dried for 48 hours at 40°C (Yu and Oldfield, 1989). Particle size analysis and organic content were determined using a Malvern Mastersizer 3000 and muffle furnace (450°C) (Chapter 5). It is important to target the sediment particle size that is dominant in sinks as differences in particle size between source and sink will alter tracer signatures, introducing error in the fingerprinting (Collins et al., 1997a; Walling, 2013). Reconnaissance field data were collected for 10 sink features. The <250 µm fraction was the dominant (>85%) grain size contained in the valley fill and the downstream Mount Fletcher bulk water supply reservoir and was thus identified as the fraction used to match sediment to sources (Collins et al., 1997b).

Mineral magnetic signatures were used as tracers for sediment sources as basalt is associated with magnetite (ferimagnetic, thus high magnetic affinity) and sedimentary rocks with haematite (canted antiferromagnetic, thus low magnetic affinity) (Dearing, 1999; Oldfield, 1999). Mass specific magnetic susceptibility and remanence properties were determined for 5-10 g of <250 µm sediment packed into 10 ml pots. The Bartington® susceptibility meter with an MS2 B dual core sensor was used to measure high and low frequency magnetic susceptibility (Maher, 1986; Dearing, 1999). The susceptibility readings were corrected for the organic content using the LOI data and expressed as minerogenic low (Xlf) and high (Xhf) frequency magnetic susceptibility (Foster et al., 1998). Associated measurement errors for magnetic susceptibility were <1%. Frequency dependent susceptibility (Xfd) was calculated based on the Xlf and Xhf (Foster et al., 2007). Remanence properties, such as anhysteretic susceptibility (XARM),

saturation remanence (SIRM), hard isothermal remanence (HIRM), were measured using the Molspin® rotating magnetometer, Molspin® a.f. demagnetiser and a Molspin® pulse magnetiser (maximum pulse strength of one tesla) as described by Maher (1986), Walden (1999) and Foster et al. (2007). Xfd, XARM and ARM are sensitive to small magnetic grains, whereas SIRM, IRM and HIRM are sensitive to large magnetic grains (Table 6.2). Xlf is sensitive to a range of particle sizes. Associated error for remanence signatures were <16% possibly due to the non-linear additivity as a result of grain size interactions.

Table 6.2: The properties of mineral magnetic tracers and associated units (Anderson and Rippey, 1988; Walden, 1999; Foster et al., 2007; Yang et al., 2010; Wang et al., 2012).

Magnetic measurement	Unit	Minerals measured	Grain size
Xlf	$10^{-6} \text{ m}^3 \text{ kg}^{-1}$	Diamagnetic, paramagnetic, canted anti-ferromagnetic, ferrimagnetic	Small - large
Xfd	$10^{-9} \text{ m}^3 \text{ kg}^{-1}$	Ultrafine super paramagnetic grains	Small (<0.02 μm)
XARM and ARM	$10^{-9} \text{ m}^3 \text{ kg}^{-1}$	Stable single domain ferrimagnetic minerals	Small (0.02-0.4 μm)
SIRM	$10^{-3} \text{ Am}^2 \text{ kg}^{-1}$	Minerals carrying remanence	Large (<100 μm)
IRM	$10^{-3} \text{ Am}^2 \text{ kg}^{-1}$	ferrimagnetic minerals	Large (<100 μm)
HIRM	$10^{-3} \text{ Am}^2 \text{ kg}^{-1}$	antiferromagnetic minerals	Large (<100 μm)

Cs-137 was used to indicate surface soil contributions in sediment cores and to assess the extent of surface soil loss since 1958. A selection of surface samples (n = 28) was analysed for the presence of Cs-137 as Cs-137 adheres strongly to finer particles in surface soils. Cs-137 fallout is assumed to be spatially homogeneous at a relatively small scale (Walling, 2004, 2005). Absence of Cs-137 in the surface soil would therefore indicate that the top soil horizon (upper 10 cm) has been totally eroded since 1958. As the <250 μm fraction was used for Cs-137 detection, it could be possible that the coarser particles within the <250 μm sample might dilute the Cs-137 activity. Samples with very low activities could thus fall below the detection limit and be classified as not containing Cs-137. This could lead to an over estimation of the number of samples not containing Cs-137.

6.2.2. Selecting and verifying the best discriminators for quantitative sediment tracing

In this section quantitative and qualitative sediment tracing approaches were tested on the most complete core (VT2). The quantitative approach was based mainly on the work of Lees (1994) and Collins et al. (1997b) that has two stages: 1) to test if tracers can distinguish between the source groups and 2) to compare the composite fingerprint between source samples and sediment samples using a multivariate mixing model. The outcome of the quantitative approach was compared to that of the qualitative approach, using a single tracer, in the hope of simplifying sediment source tracing for other sediment cores and suspended sediment samples.

The tracer signatures of the source samples were interrogated statistically to test if meaningful groups could be identified based on the measured tracers (magnetic susceptibility, magnetic remanence and median particle size). Particle size was not used as tracer, but included in the statistical analysis to assess if there are large size differences between the sources. Main differences were expected to exist between the different geological provinces (Collins et al., 1997b). The null hypothesis, that there was no significant difference for tracers between the various geological provinces, was tested using the two-tailed Kruskal-Wallis H-test. Another set of null hypotheses, that there was no difference between the slope classes, also was tested using the Kruskal-Wallis H-test.

As magnetic signatures are not only influenced by parent material, but also by biogeochemical processes (Oldfield, 1991), further nonparametric testing was done using the Mann-Whitney U-test, to test the null hypotheses that there were no differences for the tracers between: surface and subsurface samples; gully and sheet erosion; north and south-facing slopes, both within the catchment and within each respective geological grouping. Null hypotheses were rejected if $p < 0.05$. Groupings with less than five cases were not included in the hypothesis testing, but were used in the sediment tracing section. Range tests were performed on each tracer by plotting source and sediment on a scatterplot in Excel to verify that tracer signatures in the sediment falls within the range of signatures found in the potential sediment sources. If sediment signatures fell outside the source signature range, it would indicate that a potential source was missed during sampling or that tracer signatures changed during transport or after deposition and would invalidate the tracing.

Due to the selective transport between source and sink, median particle size and organic content were removed from the list of tracers used in source apportionment. Furthermore, Xfd% was removed as it violates a key assumption of the sediment fingerprinting methodology, that the tracers are linearly additive (Lees, 1999).

6.2.3. Quantitative and qualitative sediment tracing approaches

For the quantitative approach, tracers that could distinguish between source groups were used in a forward stepwise Discriminant Function Analysis (DFA) to identify the composite fingerprint of tracers able to best differentiate between the source groups (Collins et al., 1997b; Foster et al., 2007). The composite fingerprint of tracers that produced the highest percentage of sources correctly classified was used in a multivariate mixing model to produce a probability density function that would determine the proportional contribution of sediment from each source group along with associated uncertainty (Equation 6.1). The probability distribution was based on 3000 iterations per sample using the Monte Carlo algorithm (Collins and Walling, 2007). The measured value for sediment was changed for each Monte Carlo iteration to a random value within the range of the measurement error.

$$\sum_{i=1}^n \left\{ \left(C_i - \left(\sum_{s=1}^m P_s S_{si} \right) \right) / C_i \right\}^2 \quad (6.1)$$

where C_i = concentration of fingerprint property in sediment sample; P_s = the optimised percentage contribution from source category; S_{si} = median concentration of fingerprint property in source category; n = number of fingerprint properties comprising the optimum composite fingerprint; m = number of sediment source categories (Collins et al., 2010). No weightings or correction for particle size was applied.

Goodness of fit (GOF) was calculated to define the difference between actual and modelled values using Equation 6.2.

$$GOF = \left\{ 1 - \left(C_i - \left(\sum_{s=1}^m P_s S_{si} \right) \right) / C_i \right\} * 100 \quad (6.2)$$

For the qualitative approach the tracers that could distinguish between source groups were used in a Principal Component Analysis (PCA) in Statistica 12 as it would optimize the variance for this multi-dimensional dataset (Smith and Blake, 2014). The variable factor loadings were assessed for any meaningful groupings. The resultant groupings and strongest corresponding variable was displayed down each core and interpreted in a qualitative manner based on the variable range for each of the source groupings.

The results of the quantitative tracing results for core VT2 were compared to that of the qualitative tracing (using regression analysis), and if both approaches gave similar results, the simplified qualitative approach would be adopted for further sediment tracing.

6.2.4. Collecting and tracing suspended sediment

Present day sediment dynamics were assessed for a wet season (December 2013–January 2014) by sampling flood peaks and monitoring flood stage. Water samples were collected by a community member, Motlatsi Mabaleka, living near the river at the bottom of the Vuvu valley fill near VT1 (Chapter 5). Mr. Mabaleka collected 121 water samples in total during the rising and falling limbs where possible, at 10 minute intervals for 11 floods (6 to 13 bottles per event). A >350 ml bottle was filled along a turbulent section of river, and the time and date were recorded on the bottle. Bottles and contents were weighed and numbered for further laboratory analysis.

Particle size analysis was done on a diluted extraction from the sample as the sediment concentration of flood samples was too high and obscured too large a portion of the laser used by the Malvern 3000. Sediment was re-suspended using a magnetic stirrer and a pipette was used to extract 10 ml of water

which was diluted to 500 ml using distilled water. The magnetic grains in the sample stuck to the magnetic stirrer and could not all be recovered after stirring, thus changing the mineral magnetic signature of the sample. As this would compromise the tracing of sediments, every other sample was excluded from particle size analysis and solely used to measure suspended sediment concentration (57 samples) and mineral magnetic signature (64 samples).

Sediment concentration was determined by vacuum filtration by passing the entire sample through a weighed 90 mm diameter Macherey-Nagel glass fibre disk (average pore size of 0.6 μm). The disk plus sediment was dried at 40°C, weighed and packaged in magnetic pots for magnetic analysis. The sediment was not dried to 105°C to retain the magnetic signature and this can lead to marginal overestimations in sediment weight. Sediment concentration was determined by first calculating the volume of the water in the bottle (weight of water, calculated by subtracting the empty bottle weight from full bottle weight, was converted to volume assuming 1 kg = 1 l), the weight of the extracted sediment, and then dividing the sediment weight by the volume of water to give a concentration in g l^{-1} . Magnetic susceptibility and magnetic remanence were measured for the packaged suspended sediment. Data were extrapolated linearly to create a dataset with 20 minute intervals for each flow event.

Stage was measured at 20 minute intervals at three stations along the valley fill from December 2012 to May 2014 (section 7.2.1). A Solinst Levellogger was secured onto a large (1 m long axis) boulder at VT1, VT7 and VT15. The level data were corrected for barometric changes using data recorded onsite by a Solinst Barrologger. All loggers were set to record at the same time-steps to prevent unnecessary extrapolation of the data during barometric correction. Unfortunately the logger for VT7 was ripped off the boulder and the entire boulder was removed at VT1 during the monitoring period from December 2013 to January 2014. Stage data from the logger at VT15 was used as a proxy for VT1 and was adjusted by +20 min to match the time interval for the suspended sediment data measured at VT1. Data for previous flood events indicated a 20 min delay in the flood peak over the 2 km of river between VT15 and VT1. Level data were converted to discharge based on a rating curve that was developed for VT15 (Chapter 7). As early discharge measurements increased from VT15 to VT1 and catchment area increased from 47 km^2 for VT15 to 58 km^2 for VT1, due to the contribution from a significant tributary, the discharge for VT15 was up-scaled by 23% to match the proportional area increase for VT1.

Time to concentration for the flow events was calculated based on the start of the downpour (when rainfall was >2 mm per 5 min) to help understand where sediment might be sourced from. If rainfall is uniform throughout the catchment, the first flood water should be generated from nearby slopes first with an increase of contribution from distant sources over time.

This dataset on suspended sediment was used to trace sediment qualitatively and combined with discharge and sediment concentration, was used to interpret source changes during flood events.

6.3. Results

6.3.1. Selecting the best tracers

The null hypothesis, that there was no significant difference between the geological provinces, was rejected for all tracers measured as at least one parent material grouping was significantly different from the other groups using the Kruskal Wallis H-test (Table 6.3). The largest difference was present between the Xlf values of the source groups ($H = 142$). Further expansion of these results, using multiple comparisons, showed that the largest differences were present between the Drakensberg and Clarens/Elliot Formations (Table 6.3). Significant differences were measured between samples with parent material derived from the Drakensberg and Clarens Formation for all tracers except median particle size. For the Drakensberg and Elliot Formation, significant differences were detected for all tracers. No significant differences were detected for the tracers between samples derived from the Clarens and Elliot Formations. This would suggest that only two groups can be distinguished, i.e. igneous (Drakensberg Formation or basalt) and sedimentary (Clarens and Elliot Formations or sandstone and mudstone).

Table 6.3: Kruskal Wallis H-test and multiple comparisons for magnetic, organic and median particle size for samples from the various parent materials. Mean values and standard deviation are given for the variables. Significant p values in bold and italicized.

Variable	H			Drakensberg- Clarens	Drakensberg -Elliot	Clarens -Elliot	Drakensberg		Clarens		Elliot	
	(n=193)	df	p				Mean	SD	Mean	SD	Mean	SD
Xlf	141.9	2	<i>0.000</i>	<i>0.000</i>	<i>0.000</i>	0.233	5.4	2.1	0.8	0.4	0.4	0.3
SIRM	140.4	2	<i>0.000</i>	<i>0.000</i>	<i>0.000</i>	0.482	121.4	64.2	7.0	3.4	4.7	2.8
IRM	136.1	2	<i>0.000</i>	<i>0.000</i>	<i>0.000</i>	0.484	70.7	40.4	3.9	2.7	2.2	1.9
XARM	134.5	2	<i>0.000</i>	<i>0.000</i>	<i>0.000</i>	0.619	48.0	25.6	7.4	3.9	5.3	7.7
ARM	134.5	2	<i>0.000</i>	<i>0.000</i>	<i>0.000</i>	0.619	1.5	0.8	0.2	0.1	0.2	0.2
LOI	125.0	2	<i>0.000</i>	<i>0.000</i>	<i>0.000</i>	0.300	16.2	5.4	7.7	3.1	6.0	3.8
HIRM	129.1	2	<i>0.000</i>	<i>0.000</i>	<i>0.000</i>	1.000	15.5	9.9	1.3	0.4	1.1	0.5
Xfd%	100.6	2	<i>0.000</i>	<i>0.000</i>	<i>0.000</i>	1.000	4.5	2.6	10.3	1.2	9.5	2.5
Xfd	96.9	2	<i>0.000</i>	<i>0.006</i>	<i>0.000</i>	0.123	237.6	172.6	82.0	45.7	46.0	35.6
D50	18.4	2	<i>0.000</i>	0.775	<i>0.000</i>	0.807	93.5	53.4	66.5	20.9	60.0	22.2

When differences between slope classes were tested using the Kruskal Wallis H-test, the null hypothesis, that there was no difference between slope classes, was only rejected for HIRM ($H = 6.51$; $n = 180$; $p = 0.039$), with the significant difference existing between steep and gentle slopes. This would suggest that slope class did not have significant influences on the tracers.

Differences between surface and subsurface soils were detected in terms of organic content, Xfd and median particle size, thus rejecting the corresponding null hypotheses that there were no differences between surface and subsurface soils (Table 6.4). Organic content was higher for surface soils (12.4% versus 9.1%) and this trend also was observed within the Drakensberg and Elliot Formations, with the Drakensberg having the highest overall organic content. Xfd% could not distinguish between surface and subsurface soils as found by Dearing (1999), but although not statistically significant, values tended to be higher for surface than subsurface samples. Xlf, ARM, XARM, SIRM and IRM could only distinguish

between surface and subsurface within the Elliot Formation, with surface samples having higher values than subsurface values. Xfd could distinguish between surface and subsurface for all samples and samples from the Elliot Formation, with a general trend that surface samples had higher values than subsurface samples. Median particle size distinguished differences between surface and subsurface for all samples combined, and the Drakensberg Formation. Samples were finer for surface samples as would be expected as surface samples have been exposed to soil formation processes for a longer period. In conclusion, organic content, Xfd and median particle size were the tracers or variables that could most often distinguish between surface and subsurface soils.

Table 6.4: Mann-Whitney U-test results for surface and subsurface groupings for source samples according to geology. Mean values and standard deviation are given for the variables. Significant values in bold and italicized

	Variable	U	Z	p-value	Surface		Subsurface	
					Mean	Std.Dev.	Mean	Std.Dev.
All samples (surface n = 93, subsurface n = 107)	<i>LOI</i>	3581.0	3.4	0.001	12.4	7.5	9.1	5.5
	Xfd%	4443.0	1.3	0.193	7.7	3.2	6.9	3.9
	Xlf	4607.0	0.9	0.367	2.8	2.7	2.8	3.0
	<i>Xfd</i>	4093.0	2.2	0.031	154.3	161.9	122.2	139.8
	<i>ARM</i>	4317.0	1.6	0.107	0.8	0.8	0.8	1.0
	<i>XARM</i>	4317.0	1.6	0.107	25.9	24.3	23.6	29.9
	SIRM	4561.0	1.0	0.311	54.2	62.2	59.4	78.4
	IRM	4391.0	-1.4	0.152	31.3	36.6	34.9	47.9
	HIRM	4725.0	0.6	0.540	7.0	8.2	7.8	10.7
	<i>D50</i>	3217.0	-4.3	0.000	63.1	32.1	87.0	46.2
Drakensberg (surface n = 41, subsurface n = 46)	<i>LOI</i>	444.0	4.2	0.000	18.6	5.6	14.1	4.3
	Xfd%	813.0	1.1	0.271	4.8	2.4	4.3	2.8
	Xlf	933.0	0.1	0.936	5.4	1.9	5.4	2.3
	Xfd	778.0	1.4	0.162	264.3	184.4	213.8	159.7
	<i>ARM</i>	820.0	1.0	0.298	1.5	0.5	1.5	1.0
	<i>XARM</i>	820.0	1.0	0.298	48.6	16.4	47.5	31.8
	SIRM	869.0	-0.6	0.532	114.0	47.6	127.9	75.9
	IRM	903.0	0.3	0.737	66.1	29.0	74.8	48.3
	HIRM	876.0	-0.6	0.572	14.3	7.7	16.6	11.5
	<i>D50</i>	474.5	-4.0	0.000	70.9	43.1	113.6	54.0
Clarens (surface n = 7, subsurface n = 7)	LOI	11.0	1.7	0.097	8.4	2.0	7.0	3.9
	Xfd%	23.0	0.1	0.898	10.6	0.6	10.0	1.5
	Xlf	17.0	0.9	0.371	0.9	0.5	0.6	0.3
	Xfd	14.0	1.3	0.201	98.9	50.8	65.2	35.8
	<i>ARM</i>	10.0	1.8	0.074	0.3	0.1	0.2	0.1
	<i>XARM</i>	10.0	1.8	0.074	9.2	3.8	5.6	3.4
	<i>SIRM</i>	8.0	2.0	0.041	8.6	3.3	5.4	2.7
	IRM	8.0	-2.0	0.041	5.3	2.7	2.5	1.9
	HIRM	19.0	0.6	0.523	1.4	0.4	1.2	0.5
	D50	16.0	-1.0	0.307	61.0	11.0	71.9	27.5
Elliot (surface n = 43, subsurface n = 49)	<i>LOI</i>	609.0	3.5	0.001	7.2	5.0	4.9	1.5
	Xfd%	989.0	0.5	0.616	10.0	1.4	9.1	3.2
	<i>Xlf</i>	742.0	2.4	0.015	0.5	0.3	0.4	0.3
	<i>Xfd</i>	777.0	2.2	0.031	53.1	34.5	39.7	35.8
	<i>ARM</i>	746.0	2.4	0.016	0.2	0.3	0.1	0.1
	<i>XARM</i>	746.0	2.4	0.016	6.9	10.5	3.9	3.6
	<i>SIRM</i>	715.0	2.6	0.008	5.5	2.9	4.0	2.4
	<i>IRM</i>	613.0	-3.4	0.000	2.8	1.9	1.6	1.7
	HIRM	971.0	0.6	0.521	1.2	0.5	1.1	0.5
	D50	805.0	-1.9	0.052	55.2	17.7	64.2	24.8

HIRM was the only variable that could distinguish statistically between gullies and sheet erosion based on all samples. Values of HIRM were higher for samples from sheet erosion (Table 6.5); the null hypothesis was rejected for this variable. The Clarens Formation had only one sample representing sheet erosion and was thus not included in the test. For the Drakensberg Formation, organic content, ARM, XARM, SIRM and HIRM could differentiate between gully and sheet erosion samples and thus the associated null hypotheses were rejected. Samples from sheet erosion tended to have higher organic content, SIRM and HIRM values, whereas ARM and XARM values tended to be lower for sheet erosion. None of the null hypotheses were rejected for the gully and sheet erosion on the Elliot formation.

Table 6.5: Mann-Whitney U-test results for gully and sheet erosion. Mean values and standard deviation are given for the variables. Significant values in bold and italicized

	Variable	U	Z	p-value	Gully		Sheet erosion	
					Mean	Std.Dev.	Mean	Std.Dev.
All samples (gully n = 106, sheet n = 73)	LOI	3266.0	-1.8	0.077	10.1	5.9	12.3	7.8
	Xfd%	3343.0	1.5	0.123	7.5	3.4	6.6	3.7
	Xlf	3416.0	-1.3	0.184	2.6	2.7	3.3	3.1
	Xfd	3805.0	-0.2	0.852	148.4	167.7	137.8	137.6
	ARM	3753.0	-0.3	0.734	0.9	1.0	0.8	0.7
	XARM	3753.0	-0.3	0.734	27.1	31.6	25.6	22.6
	SIRM	3205.0	-1.9	0.051	52.6	70.8	75.0	75.4
	IRM	3302.0	1.7	0.096	31.6	43.6	41.7	44.7
	HIRM	3069.0	-2.3	0.019	6.7	9.3	10.2	10.5
	D50	3850.0	0.1	0.957	76.8	42.8	79.7	44.3
Drakensberg (gully n = 48, sheet n = 39)	LOI	602.0	-2.8	0.004	14.7	4.2	18.1	6.2
	Xfd%	725.0	1.8	0.072	5.1	2.8	3.9	2.2
	Xlf	770.0	-1.4	0.158	5.1	2.1	5.7	2.1
	Xfd	840.0	0.8	0.415	255.0	191.2	216.2	146.1
	ARM	677.0	2.2	0.027	1.7	1.0	1.3	0.5
	XARM	677.0	2.2	0.027	52.8	30.8	42.2	15.7
	SIRM	596.0	-2.9	0.004	109.3	72.0	136.2	50.0
	IRM	709.0	1.9	0.053	66.3	44.7	76.1	34.2
	HIRM	599.0	-2.9	0.004	13.3	10.5	18.1	8.5
	D50	905.0	-0.3	0.795	92.8	55.8	94.3	51.1
Elliot (gully n = 43, sheet n = 33)	LOI	604.0	-1.1	0.271	5.9	4.3	5.7	1.8
	Xfd%	626.0	-0.9	0.384	9.2	2.6	9.7	2.5
	Xlf	609.0	-1.0	0.295	0.4	0.3	0.5	0.3
	Xfd	611.0	-1.0	0.304	40.7	35.8	48.2	34.2
	ARM	596.0	-1.2	0.236	0.1	0.1	0.2	0.4
	XARM	596.0	-1.2	0.236	4.0	3.7	6.7	11.9
	SIRM	611.0	-1.0	0.304	4.4	3.0	4.7	2.4
	IRM	625.0	0.9	0.379	2.0	2.0	2.2	1.8
	HIRM	570.0	-1.5	0.145	1.0	0.5	1.1	0.5
	D50	664.5	0.5	0.641	61.5	19.3	62.6	27.6

Null hypotheses were rejected for organic content, Xfd and median particle size for north and south-facing features when all source samples were considered (Table 6.6). Organic content was significantly higher on the south-facing slopes for all samples and the Elliot Formation. Particle size was significantly smaller on the south-facing slopes. Differences for Xfd were detected between north and south-facing slopes for all source samples, whereas differences in HIRM could be detected between north and south-

facing slopes within the Drakensberg Formation, with the respective null hypotheses rejected. Xfd and HIRM values were higher on south-facing slopes.

Table 6.6: Mann-Whitney U-test results for north and south-facing slopes. Mean values for the groupings are given. Significant values in bold and italicized

	Variable	U	Z	p-value	North		South	
					Mean	Std.Dev.	Mean	Std.Dev.
All samples (north n = 116, south n = 67)	<i>LOI</i>	3186.0	-2.0	0.043	10.2	6.4	12.2	7.3
	Xfd%	3783.0	-0.3	0.767	7.3	3.6	7.4	3.5
	Xlf	3279.0	-1.8	0.079	2.6	2.7	3.3	3.2
	<i>Xfd</i>	3197.0	-2.0	0.046	121.1	126.1	172.8	187.8
	ARM	3277	-1.8	0.078	0.7	0.9	0.9	0.9
	XARM	3277	-1.8	0.078	23.5	28.6	28.7	26.7
	SIRM	3409	-1.4	0.167	54.0	69.0	67.3	78.5
	IRM	3559	1.0	0.344	32.6	43.4	37.0	45.2
	HIRM	3381.0	-1.5	0.144	6.6	8.6	9.4	11.2
	<i>D50</i>	2605.5	3.7	0.000	82.1	43.5	65.2	37.7
Drakensberg (north n = 50, south n = 32)	LOI	657.0	-1.4	0.176	15.8	4.5	17.9	6.2
	Xfd%	675.0	-1.2	0.237	4.3	2.5	5.0	2.5
	Xlf	664.0	-1.3	0.198	5.2	2.1	5.8	2.2
	Xfd	610.0	-1.8	0.072	208.5	137.2	291.7	210.4
	ARM	671.0	-1.2	0.222	1.5	0.9	1.6	0.6
	XARM	671.0	-1.2	0.222	47.3	29.3	50.3	20.2
	SIRM	720.0	-0.8	0.450	118.1	61.5	131.0	69.2
	IRM	797.0	0.0	0.981	71.9	40.3	71.4	42.5
	<i>HIRM</i>	561.0	-2.3	0.023	13.7	9.0	18.3	10.4
	<i>D50</i>	493.0	2.9	0.004	103.6	52.6	75.3	51.1
Elliot (north n = 53, south n = 30)	<i>LOI</i>	533.0	-2.5	0.013	5.6	3.9	6.3	2.0
	Xfd%	647.0	-1.4	0.162	9.3	2.6	9.9	2.3
	Xlf	616.0	-1.7	0.091	0.4	0.3	0.5	0.3
	Xfd	611.0	-1.7	0.082	39.4	32.7	54.6	38.7
	ARM	600.0	-1.8	0.065	0.1	0.1	0.2	0.4
	XARM	600.0	-1.8	0.065	4.1	3.6	7.4	12.4
	SIRM	631.0	-1.5	0.121	4.4	3.0	5.0	2.5
	IRM	669.0	1.2	0.234	2.0	2.0	2.3	1.6
	HIRM	666.0	-1.2	0.223	1.0	0.5	1.2	0.4
	<i>D50</i>	464.5	3.1	0.002	65.2	25.6	52.9	12.5

Variable values for sink samples were within the range of the source samples, indicating that all possible sources were represented by the sampling and that tracer properties had not been significantly transformed during sediment transport and storage (Table 6.7).

Table 6.7: Summary of range test results showing that variable values for sinks fall within the source range.

Variable	Source				Sink	
	Igneous		Sedimentary		Sediment	
	Mean	Std. dev.	Mean	Std. dev.	Mean	Std. dev.
Xlf	5.4	2.1	0.5	0.3	1.8	0.9
Xfd	237.6	172.6	50.7	38.8	57.8	49.9
ARM	1.5	0.8	0.2	0.2	0.5	0.7
XARM	48.0	25.6	5.6	7.4	14.8	20.6
SIRM	121.4	64.2	5.0	2.9	50.0	32.9
IRM	70.7	40.4	2.4	2.1	29.9	19.8
HIRM	15.5	9.9	1.1	0.5	8.8	6.0

The statistical tests showed that most of the tracers can be used to distinguish between samples derived from one of two groups: the Drakensberg and the Clarens/Elliot Formations. Although differences can be detected within geological provinces for surface/subsurface, feature type and aspect, the trends would be helpful only if tracing was done within that particular geological province. The detected differences were less apparent once sediments from other geologies were introduced as within group (i.e. surface/subsurface of particular geology) variance was less than between group (different geologies) variance, making mixtures from various geologies impossible to distinguish in terms of surface/subsurface, feature type and aspect. Sediment tracing was thus limited to igneous and sedimentary sources. All tracers passed the range test, confirming conservative behaviour of the tracers and suggest that the main sources within the catchment were sampled.

6.3.2. Quantitative sediment tracing of core VT2

Results from the Discriminant Function Analysis showed that the best igneous and sedimentary source discrimination was achieved using a composite fingerprint consisting of Xlf, ARM and HIRM. A total of 96% of source samples was correctly classified into their respective groups.

Using these tracers (Xlf, ARM and HIRM) in a Monte Carlo model with 3000 iterations per sediment sample resulted in an apportionment result for VT2 that was dominated by sedimentary sources throughout most of the core (Figure 6.1). Sedimentary sources contributed on average 64% of the sediment, with the remainder of the 36% contributed by igneous sources. Average mean uncertainty for the contributions was 10% with error margins ranging between 3 and 27%. The average goodness of fit for the apportionment was 87%, ranging between 76 and 96%.

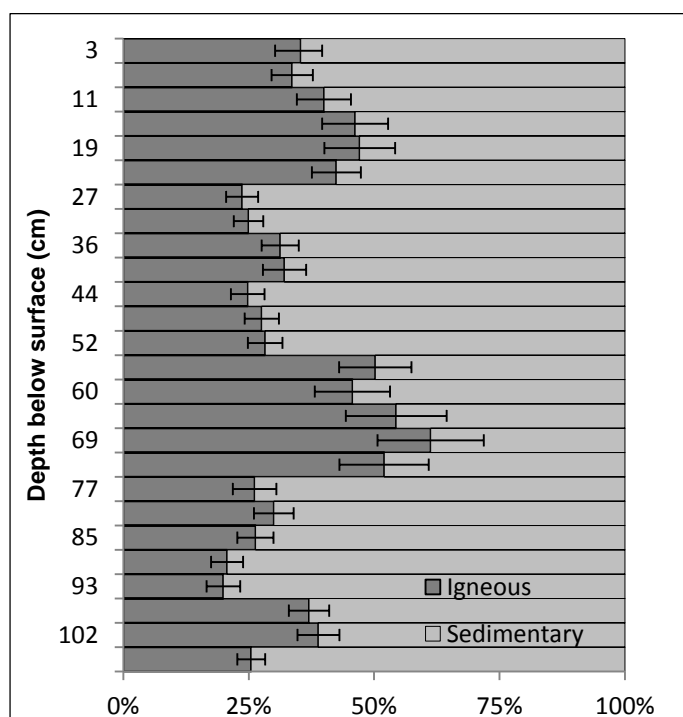


Figure 6.1: Source apportionment for core VT2.

6.3.3. Qualitative sediment tracing of core VT2

Sediment and source samples were plotted using PCA to optimize variable variance (Figure 6.2 and 6.3). Factor 1 (eigenvalue of 5.1) explained 71.7% of the variance with Xlf being the most significant variable loading (Figure 6.2; $R^2 > 0.92$). Factor 2 (Eigenvalue of 1.2) explained 16.6% of the variance with Xfd having the highest correlation ($R^2 = 0.54$). No meaningful grouping was achieved for source groups by Factor 2 (Figure 6.3). Xlf was thus chosen as the single variable to trace sediment from igneous and sedimentary sources. Interestingly, the tracers plotted as three groupings Xfd, ARM and XARM at the top, Xlf in the middle and IRM, SIRM and HIRM at the bottom in Figure 6.2. Each grouping is sensitive to different grain sizes: small (Xfd, ARM and XARM), a combination of sizes (Xlf) and large grains (SIRM, IRM and HIRM) (Table 6.2).

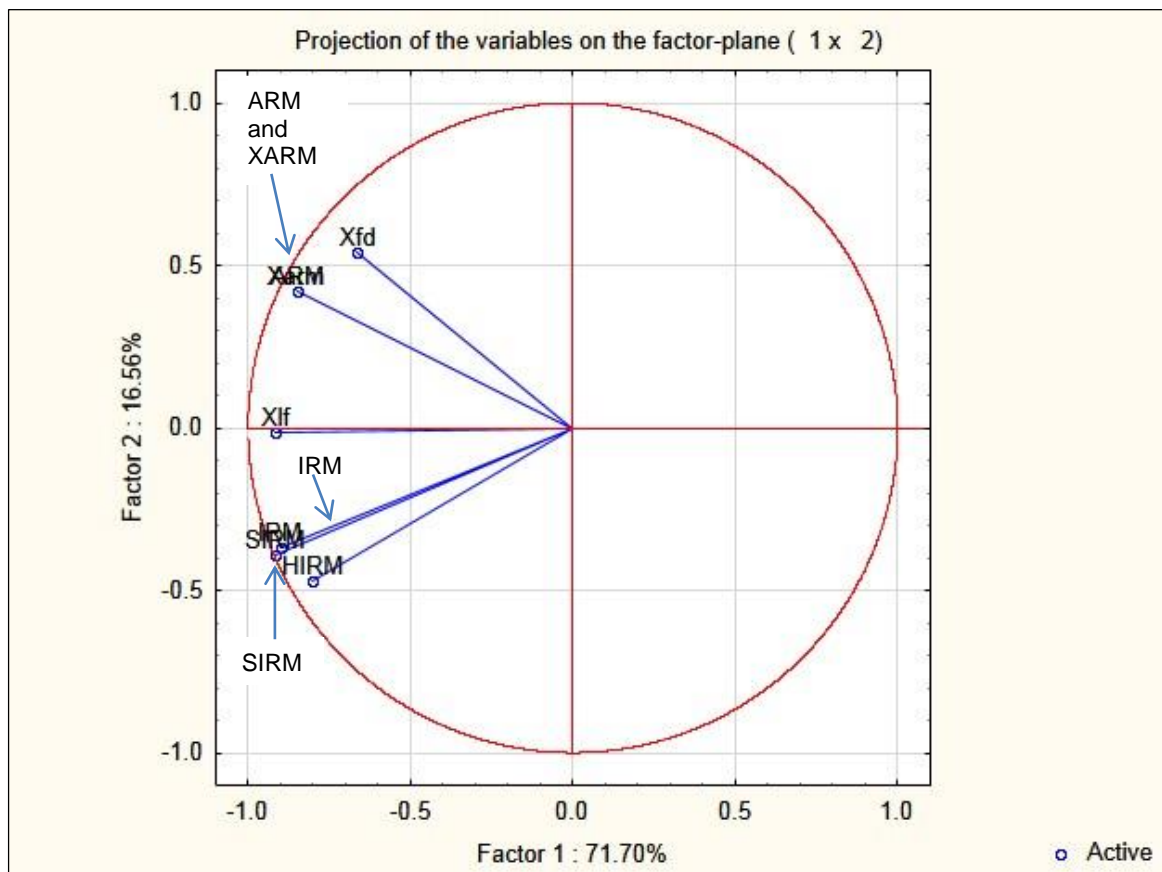


Figure 6.2: Factor loadings for source and sink samples using a Principal Component Analysis

The flood water, flood bench and terrace samples clustered between the sedimentary and igneous source samples, when considering Factor 1, indicating that these samples were a mixture of these two main source groups (Figure 6.3).

Median Xlf values of sedimentary ($0.4 \times 10^{-6} \text{ m}^3 \text{ kg}^{-1}$) and igneous sources ($5.1 \times 10^{-6} \text{ m}^3 \text{ kg}^{-1}$) (Figure 6.4) showed large differences between the two source groups with minimal overlap between values. From the Xlf values of core VT2 it can be seen that the sediment was a mixture from both source groups. The Xlf values were mostly nearer to the sedimentary source range, indicating that the majority of the sediment was of sedimentary origin, with layers that have larger igneous inputs.

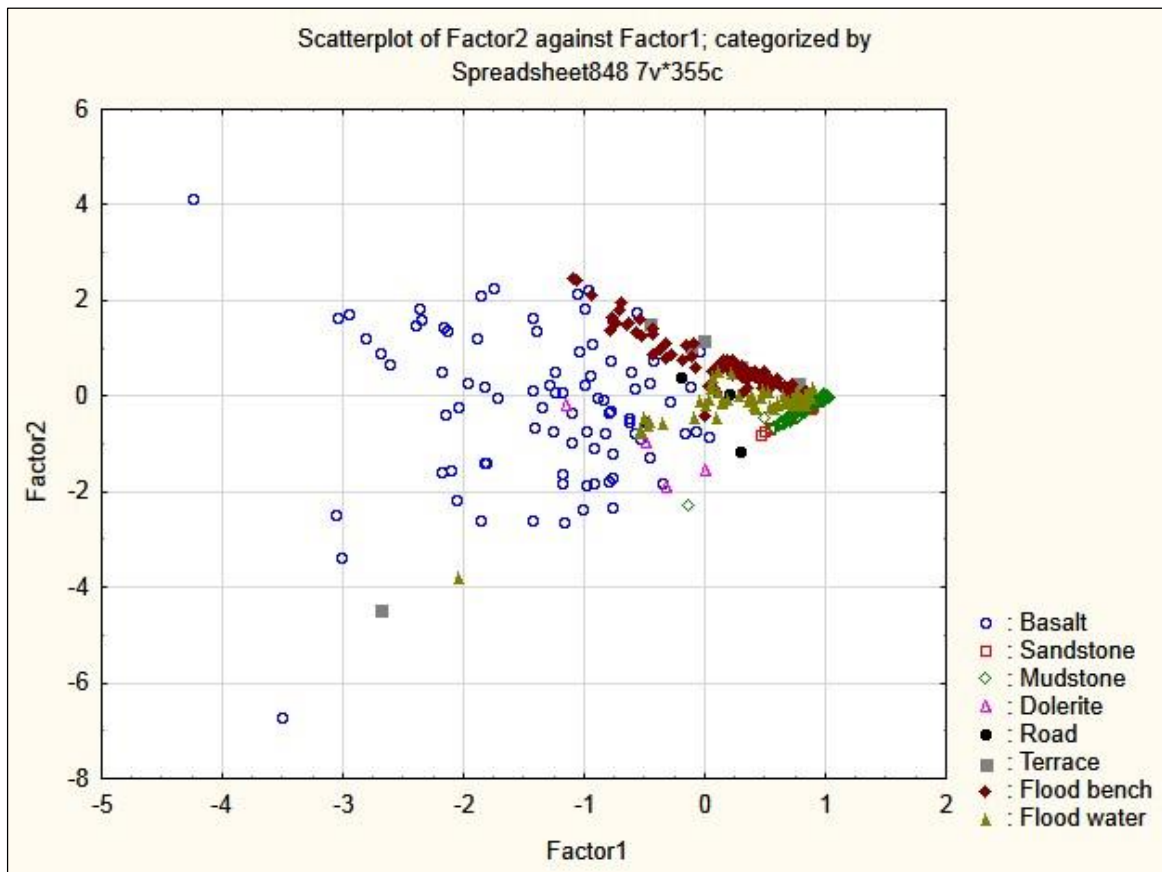


Figure 6.3: Principal Component Analysis factor scores for source, sink and suspended sediment samples.

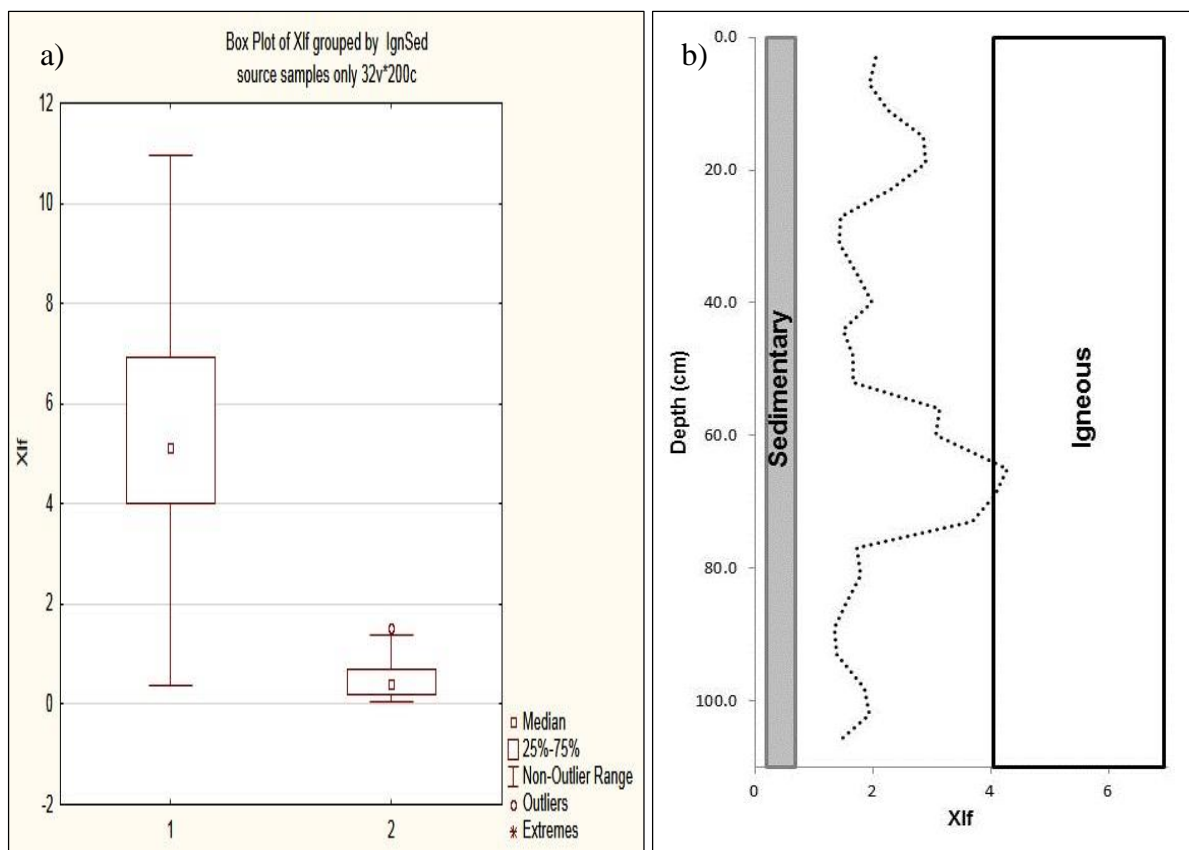


Figure 6.4: a) A boxplot showing the range of Xlf values for igneous (1) and sedimentary (2) source samples and b) Xlf values for core VT2. Xlf quartiles (25–75%) for sedimentary and igneous source samples are indicated for VT2 using boxes.

6.3.4. Comparing quantitative and qualitative sediment tracing

From Figure 6.5 it can be seen that the sediment makeup follows a very similar trend for both the quantitative and qualitative sediment tracing of core VT2. The quantitative approach produces estimated proportions with the amount of error related to the apportionment, whereas the qualitative approach gives an indication of an increased or decreased contribution from a specific source.

The agreement between the two approaches is confirmed by the strong positive relationship ($R^2 = 0.95$) between percentage contribution from igneous source and Xlf (Figure 6.6). The equation:

$$y = 32.077\ln(x) + 12.402 \quad (6.2)$$

where x is Xlf ($10^{-6} \text{ m}^3 \text{ kg}^{-1}$) and y is % contribution from an igneous source.

This equation can thus be used to calculate the contribution from an igneous source based on the Xlf value of the sediment. Equation 6.2 would allow for sediment apportionment for other stored sediment and future monitoring based on Xlf values alone. A sediment Xlf value of $3 \times 10^{-6} \text{ m}^3 \text{ kg}^{-1}$ would define equal

contributions of igneous and sedimentary sources (Figure 6.6). This benchmark, in combination with Equation 6.2, was used to interpret further sediment tracing as the focus of the tracing was to see if there was a shift in sediment source that could be related to changes in connectivity.

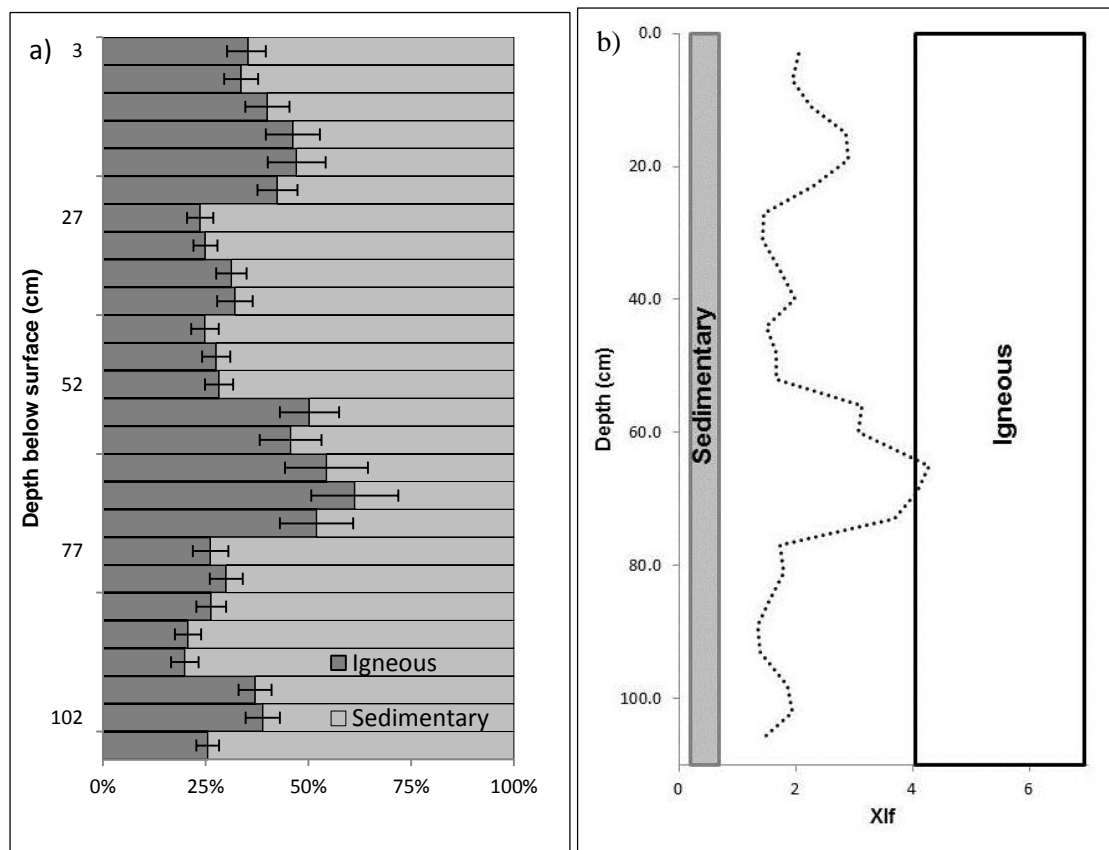


Figure 6.5: Comparison of the outcome of a) quantitative and b) qualitative sediment tracing.

The negative relationship between goodness of fit and median particle size indicated that apportionment accuracy for layers with larger particles declines with an increase in particle size ($R^2 = 0.64$) (Figure 6.6). This trend could be related to the selective transport and deposition of sediment, removing a fraction of the sediment that would be expected to be present when compared to the source material. Sediment deposited during larger floods was thus less representative of the source sediment, introducing increased error. Results of sediment tracing for coarser layers should thus be interpreted with caution and in the light of the particle size range for the various sources. Particle size was however determined not to be the primary control for the magnetic signature ($R^2 < 0.08$; Figure 6.6c and d).

From this section it was determined that sediment source tracing can be done in the Vuvu catchment in a semi-qualitative way using Xlf values alone as a tracer. This simplifies any further tracing without compromising the results.

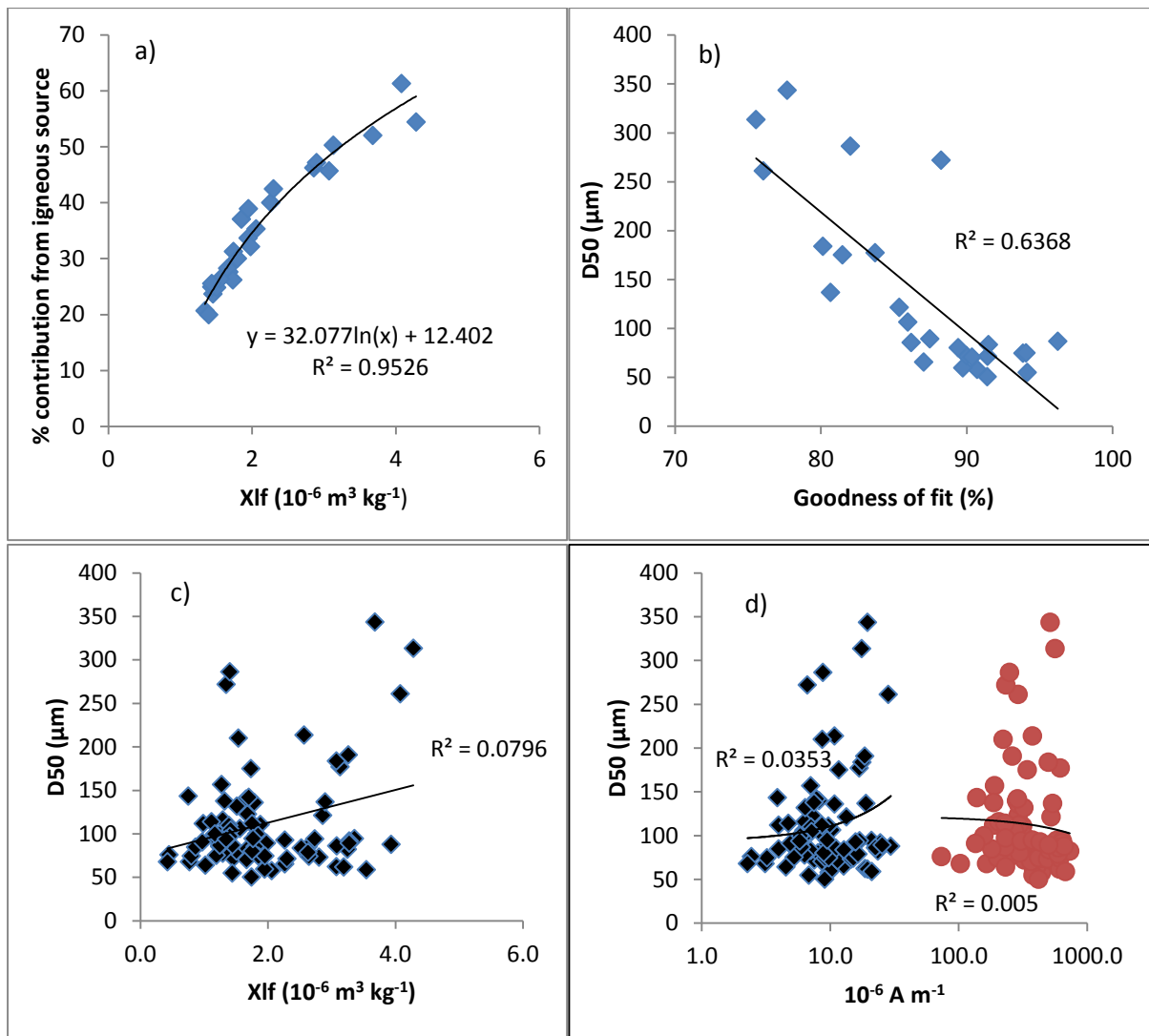


Figure 6.6: a) Relationship between Xlf and the contribution of igneous material and b) the relationship between median particle size and goodness of fit and c) the relationship between median particle size and Xlf and d) the relationship between median particle size and HIRM (diamonds) and ARM (circles).

6.3.5. Present day sources for suspended sediment

Only two of the 11 flood events that were sampled captured both the rising and falling limb of the flood event (26 December 2013 and 29 January 2014). Flood peaks were short lived (60–100 minutes) with sharp rising and falling limbs (Figure 6.7). This was to be expected for a small steep catchment, and confirmed its hydrologically well-connected nature. Measured suspended sediment concentration was very high (up to 11 g l^{-1}) for the Vuvu River and preceded or followed the flood peak (Figure 6.7). The peak in sediment concentration also was short-lived, although concentrations proved to stay high directly after the main flood pulse ($1\text{--}6 \text{ g l}^{-1}$). Suspended sediment consisted of clay sized particles with an average D50 of $2 \mu\text{m}$ (min D50 = $1.76 \mu\text{m}$ and max D50 = $2.36 \mu\text{m}$). No relationship between median particle size and discharge was observed.

The event on the 26th of December 2013 was the largest discharge recorded for the wet season. Although only 22 mm of rainfall was recorded (max intensity of 8 mm in 5 min), discharge peaked at 32 m³ s⁻¹ and inundated 93% of the lower flood benches and 20% of the higher flood benches during peak flow (Chapter 7 and Appendix 5). There was a clear change in the source dominance of the suspended sediment with time. During the first 20 minutes of the flood, the suspended sediment concentration was dominated by local sedimentary sources (Xlf < 1 or < 12% igneous material), whereafter igneous sediment was introduced and led to a sharp rise in sediment concentration from 6 to 11 g l⁻¹ during the 20–60 min period (Figure 6.7). The discharge peaked at 40 min (Xlf = 1.5 or 26% igneous material) and the suspended sediment concentration at 60 min. Thus the peak sediment concentration followed the peak flow and displayed lagging hysteresis, although normally expected before the flood peak (Walling and Webb, 1982). The rising limb of the flood (20–60 minutes) marked the transition phase from sedimentary dominated (Xlf = 1 or 12% igneous material) sources to an almost equal mixture of igneous and sedimentary (Xlf = 2.5 or 42% igneous material) sources. From 60–120 min the basalt signature became stronger (Xlf = 3 or 50% igneous material) while the discharge and suspended sediment concentration decreased steadily. The number of benches that were inundated during the increased basalt contribution decreased from 60% to 6% (60–120 min). If it is assumed that the rainfall intensity and timing was uniform throughout the catchment, it took the waters from the upper catchment ca. 40 min longer to arrive at the monitoring point than the waters delivered from the lower catchment. The overall time to concentration for this event was ca. 80 min.

The event that occurred on the 29th of January 2014 was smaller in magnitude (14 m³ s⁻¹), although 38 mm of rainfall was recorded (max intensity 3.8 mm in 5 min), and generated a lower peak sediment concentration (up to 5.3 g l⁻¹) (Figure 6.7). This flood peak overtopped 60% of the lower flood benches and none of the higher flood benches (Chapter 7 and Appendix 5). The flood peak consisted mostly of suspended sediment from locally derived sedimentary sources (Xlf < 1 or < 12% igneous material). The slight drop in Xlf (from 0.7 to 0.5) during the peak discharge of the event suggested a dominance of sediment with very low Xlf values (typical of the Elliot Formation) (Table 6.3). Towards the end of the flood the Xlf values increased to levels similar to that before the flood, suggesting that locally produced sediment from the Elliot Formation dominated during the flood as opposed to the baseflow containing some igneous material. This flood event represented a localised rain event that only produced floodwaters and sediment from within the lower catchment. The time to concentration for this localised event was ca. 40 min.

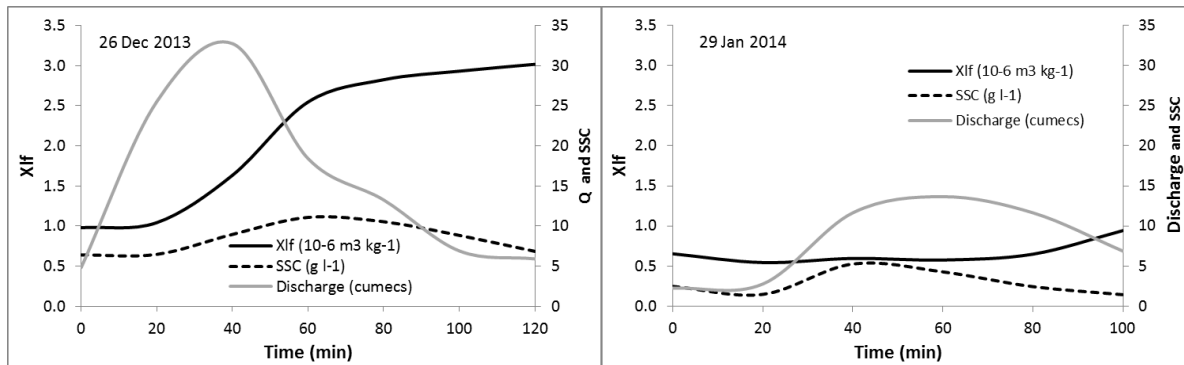


Figure 6.7: Xlf, discharge and suspended sediment concentration the lower Vuvu River for 26 December 2013 and 29 January 2014.

The sediment concentration and discharge plot for all events (Figure 6.8a) (64 samples) showed a moderate positive correlation ($R^2 = 0.3475$) for December 2013 and a weak positive correlation ($R^2 = 0.105$) for January 2014, indicating that sediment was generally transported during higher flows. The greater correlation for December showed that higher flow events produced higher sediment concentrations on a more consistent basis, whereas during the later stages (January) this relationship became less consistent, often producing lower sediment concentrations for comparable flow events at the beginning of the season. This trend was clear when the average suspended sediment concentration and discharge values per event were considered. The average suspended sediment concentration for similar flow events decreased throughout the season. This would indicate that sediment available for transport was reduced during the wet season, thus a supply limited system.

There was no correlation between suspended sediment concentration and % contribution from igneous sources ($R^2 = 0.014$).

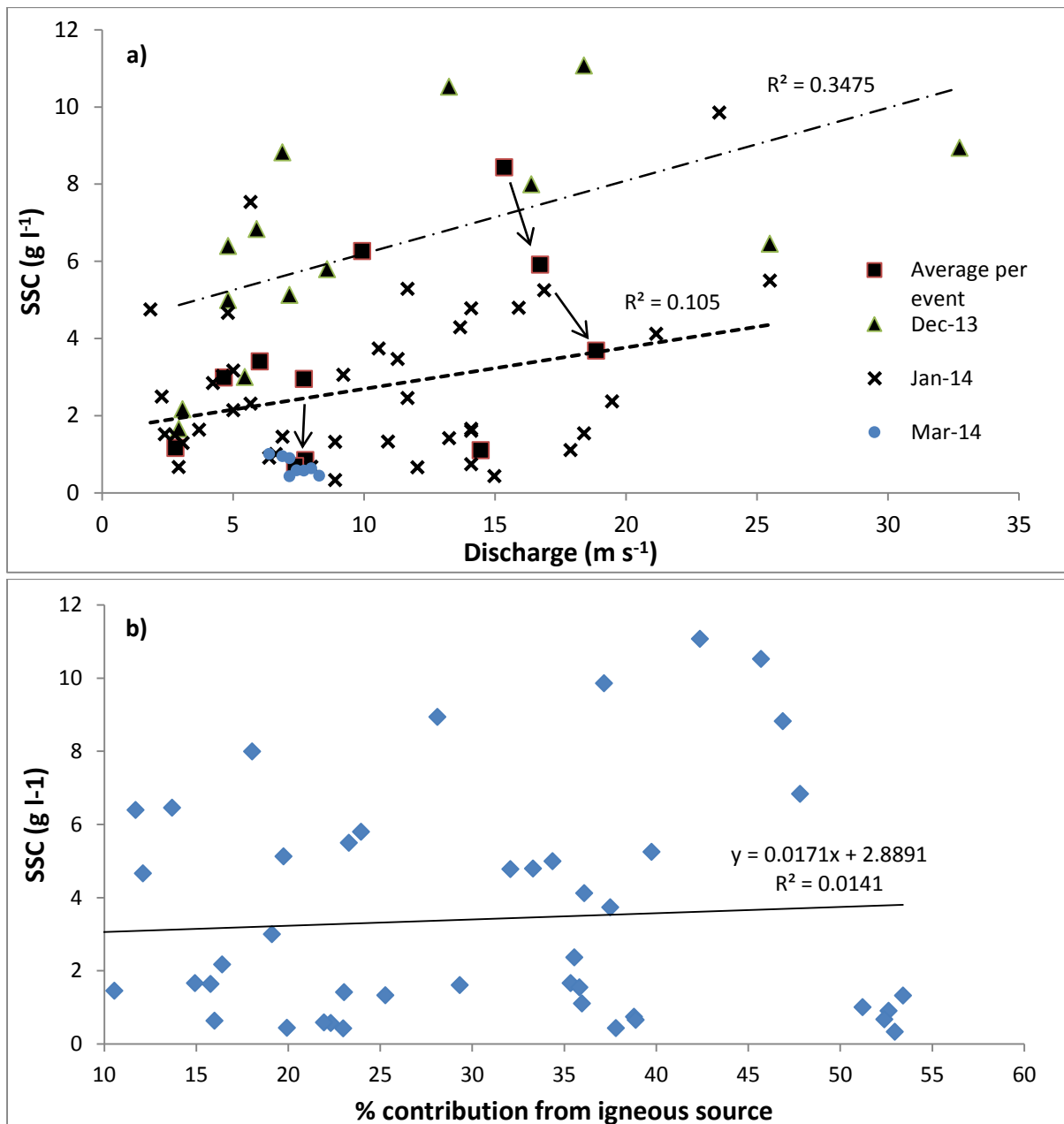


Figure 6.8: a) Bi-plot showing discharge against suspended sediment concentration data for 11 high flow events (December 2013 to March 2014). Arrows indicate the decrease in average suspended sediment concentration for similar magnitude events throughout the wet season. Linear trendlines are given for December 2013 and January 2014. b) Suspended sediment concentration against the % contribution from igneous sources.

6.3.6. Flood bench sediment origin

When looking at down-core contributions from igneous and sedimentary sources in flood benches (Figure 6.9), it could be seen that the contribution from both sources varied over time. There was no clear trend that would have suggested a definite shift in sediment source over the past 50-100 years for both the cores on higher and lower flood benches (Figure 6.9).

Cores on the higher flood benches had igneous contributions ranging from 22-59% (Xlf of $1.3\text{--}4.3 \times 10^{-6} \text{ m}^3 \text{ kg}^{-1}$), thus a mixture of sources (mainly sedimentary) with greater igneous contribution over periods as can be seen for VT2 from 55–77 cm. The average contribution of igneous material for VT2 was $36 \pm 11\%$ (Xlf of $2.2 \pm 0.8 \times 10^{-6} \text{ m}^3 \text{ kg}^{-1}$) and for VT7-2 was $47 \pm 5\%$ (Xlf of $3 \pm 0.4 \times 10^{-6} \text{ m}^3 \text{ kg}^{-1}$), showing that VT7-2 stored more sediment from igneous sources than VT2. VT2 also showed more variability than VT7-2, suggesting that VT2 received sediment during high flows where igneous material content varied, whereas VT7-2 mainly received sediment during high flows with almost equal contributions from igneous and sedimentary sources.

Sediment on the lower flood benches originated from sources with a greater contribution from sedimentary parent materials (range of igneous contribution 3–50%). VT4 (average igneous contribution of $24 \pm 10\%$; Xlf of $1.5 \pm 0.5 \times 10^{-6} \text{ m}^3 \text{ kg}^{-1}$) contained more of a mixture of source sediments than VT7 (average igneous contribution of $19 \pm 9\%$; Xlf of $1.2 \pm 0.5 \times 10^{-6} \text{ m}^3 \text{ kg}^{-1}$) which consisted mostly of sediments from sedimentary sources. The lower flood benches stored a smaller proportion of igneous material than the higher flood benches.

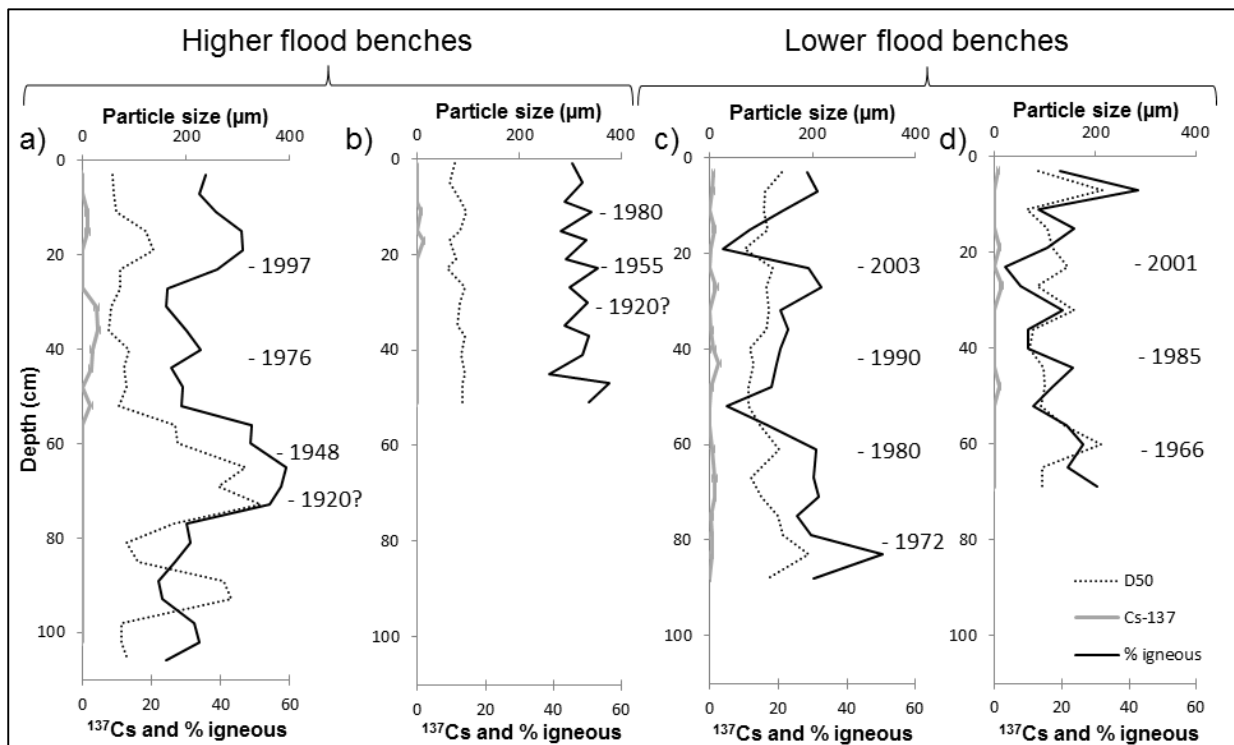


Figure 6.9: Down core median particle size, Cs-137 (mBq g^{-1} ; note horizontal error bars) and % igneous sediment for core: a) VT2, b) VT7-2, c) VT4 and d) VT7.

When considering down core values for median particle size, Cs-137 and % contribution from igneous sources, it can be seen that particle size and % igneous follow a similar trend. This was to be expected as the igneous source samples generally had a larger particle size ($D50 = 93 \mu\text{m}$) and Xlf values than sedimentary source samples ($D50 = 60\text{--}66 \mu\text{m}$) (Table 6.3) and there was a poor positive correlation between D50 and Xlf ($R^2 = 0.31$). The median particle size falls mostly within the range of the source

samples, but lenses of coarser sediment, with a higher contribution of igneous sources, were indicative of larger floods when selective transport took place, in contrast to shorter duration ‘dumping’ that would occur during smaller floods. For the coarser lenses, it was suggested that the flood benches were inundated to a greater depth and probably for a longer time as would be expected for larger magnitude events. In terms of connectivity, the data indicated that the upper catchment contributed a larger proportion of sediment during the larger floods, whereas smaller floods transported sediments that were more locally sourced. It is likely that the upper catchment also generates small floods due to isolated lower energy storms, but attenuation of these smaller events due to longer travel distances results in lower magnitude flood peaks in the lower catchment compared to local stormflows generated in the lower catchment. These small events generated in the upper catchment do not overtop the lower benches in the lower catchment and no igneous sediment is deposited.

Although a clear chronology could not be established, it seemed that an increased contribution from the upper catchment was stored on flood benches around 1997 for VT2 and VT4, and 1920?–1950 for Core VT2. The 1997 increases in igneous material coincided with the beginning of the wet period following a long dry period seen in Figure 5.7 of Chapter 5. This indicated that large quantities of igneous sediment were made available during drier periods but were stored locally and only transported to the lower catchment during subsequent wet periods when a portion was deposited on flood benches.

Organic content, which was found to be a good discriminator between surface and subsurface soil, could not be used as a tracer as selective transport, in-situ soil formation and contributions from the more organic rich igneous sources (Table 6.4) would confound these observations. As sediment trapped in the lower benches was traced mostly to local sedimentary sources and Xfd could discriminate between surface (53.1 ± 34.5) and subsurface (39.7 ± 35.8) soils within the Elliot Formation, Xfd values were used to determine top soil contributions (Table 6.4) despite the overlap of the error ranges. Average Xfd values for VT4 were 20.1 ± 6.5 and for VT7 were 27.2 ± 4.1 , indicating that the deposited sediment was potentially sourced from subsoil. The low Xfd values could be a function of selective transport as finer sediment associated with topsoil would stay longer in suspension and thus be less likely to be trapped on the lower benches.

Cs-137 was used as an alternative tracer for surface soil contributions, as it can be used as a marker of intact topsoil (Bajracharya et al., 1998). Cs-137 data for surface source samples (upper 10 cm) suggested that half of the catchment’s surface soils had lost the top layer containing Cs-137 (Table 6.8). When considering individual geological provinces, the data suggested that the samples from the Clarens Formation has not lost any topsoil, the samples from the Elliot Formation has lost 63% and the samples from the Drakensberg Formation has lost 40% of topsoil since 1958. This confirmed earlier results on erosion mapping that the Elliot Formation is the most active geological province in terms of being a sediment source (Chapter 4). It seems as if the Clarens Formation was very stable, but limited sample numbers ($n = 2$) confined further extrapolations. The Drakensberg formation was relatively active and

has potentially lost a significant volume of topsoil, especially if the size of this formation is considered (Drakensberg 1378 ha eroded and Elliot 854 ha eroded).

Table 6.8: The absence of C-137 in surface samples

Formation	No. of samples	No. of samples with no Cs-137 present	% of sampling sites with topsoil eroded
All	28	14	50
Drakensberg	10	4	40
Clarens	2	0	0
Elliot	16	10	63

In the flood bench cores Cs-137 was detected in layers with igneous contributions ranging from 12 to 51% ($X_{if} 1-3.2 \times 10^{-6} \text{ m}^3 \text{ kg}^{-1}$) and smaller particle sizes ($<150 \mu\text{m}$) (Figure 6.9), suggesting that the majority of these sediment layers was derived from sedimentary surface soils. However, layers containing Cs-137 showed increases in the contribution from igneous sources compared to layers not containing Cs-137. When the surface contributions were linked to the wet and dry periods detected in the long term rainfall data, it would appear that relatively intact surface soils were contributing during or straight after these drier periods, as surface cover was reduced. During wetter periods, it appears that surface soils contributed to a lesser extent, and that the increased contributions from sedimentary sources during wetter periods tended not to contain Cs-137, indicating that they originated from subsoils or that the Cs-137 rich top soil had already been lost. This was supported by the low X_{fd} values in the lower flood benches and low frequency (37.5%) of Cs-137 in surface soils of the Elliot Formation (Table 6.8).

6.3.7. Terrace sediment origin

Sediment dynamics over the past 1 400–4 500 years seemed to follow similar trends to recent dynamics (past ~100 years), where sediment was dominated by sedimentary sources (52–73% sedimentary contribution), with greater igneous inputs around 3 000 years ago (48% igneous contribution) (Table 6.9). These trends should be treated with caution as it was clear from the sediment in flood water and flood benches that source dominance can vary in time from within a flow event to decadal wet and dry cycles. Coarser samples did not show the higher X_{if} values associated with igneous sources as was found in the flood benches (Figure 6.9 and Table 6.9). Organic content increased with age and particle size was significantly larger 1 400–2 270 years ago. This might not be a true reflection of the dominant sediment size for that time, as the particle size was most likely a function of the varied depositional environments (e.g. velocity) for the samples. X_{fd} values for terrace samples with a smaller contribution from igneous sources ($<40\%$) would suggest that the sediment was sourced from subsoils of the Elliot Formation (Table 6.4) as X_{fd} values ranged from 21.9 to 29.4.

Table 6.9: Median particle size, organic content, Xlf, % igneous contribution and Xfd for the terrace samples.

Sample	Age (a)	D50 (μm)	LOI (%)	Xlf	% igneous contribution	Xfd
VOSL 3	1400 \pm 750	156	5.8	2.4	41	46.3
VOSL 4	2100 \pm 1550	455	5.1	2.0	35	25.6
VOSL 1	2270 \pm 490	344	4.5	2.1	36	21.9
VOSL 2	2990 \pm 720	87	7.3	2.9	48	51.2
VOSL 5	4570 \pm 490	83	6.9	1.6	27	29.4

6.4. Discussion

6.4.1. Selecting the best tracers

Differences between basalt (igneous) and at least one of the sedimentary (sandstone and mudstone) source geologies were significant for all magnetic properties (Oldfield et al., 1979; Foster et al., 1998, 2007), but no significant differences were detected between source samples derived from mudstone or sandstone parent material. The contrast in mineral magnetic properties between the igneous and sedimentary sources proved ideal for tracing (Oldfield et al., 1979; Foster et al., 1998, 2007). Only being able to discriminate between two sources would limit the amount of error introduced and having magnetically contrasting sources proved ideal for robust sediment tracing (Lees, 1997).

The higher organic content for the higher lying igneous source samples could be the result of the type of soil that promoted vegetation growth and related increases in organic content or that organic rich soils have been removed through erosion along the lower catchment. The latter is unlikely as source samples were often taken from subsoils, suggesting that igneous soils have an inherently higher organic content. Mucina et al. (2006) described soils formed on basalt as well-structured and retaining moisture (which leads to an increase in organic content) when compared to poorly drained fine-grained clay-rich soils of sedimentary origin that are less likely to increase in organic content. The larger particle size for igneous sources could also be related to the parent material, indicating that sandstone and mudstone break down to smaller particles, compared to basalt (Mucina et al., 2006; Fey et al., 2010). The larger proportion of fine material (silt) combined with a high sodium content and poor structure makes the mudstones dispersive and highly erodible (Fey et al., 2010). As organic content and particle size were dependent on fluvial transport and subject to selective transport and sorting, it was likely that direct comparisons between source and sediment would be compromised and thus should not be used for primary tracing, but could be considered as supportive evidence (Collins et al., 1997b).

Slope classes could not be meaningfully distinguished based on the tracers measured for the source samples and were thus excluded from the sediment tracing.

Surface and subsurface samples could be differentiated based on organic content, Xfd, and median particle size. Organic content was higher in surface soils and was associated with soil formation in surface layers. The lower organic content in subsurface soils could be a result of bacterial action that break it

down. Higher Xfd values for surface soils were associated with increases in secondary ferromagnetic grains, ranging from single domain to smaller viscose grain sizes (supraparamagnetic) as would be expected for upper layers exposed to soil formation (Yu and Oldfield, 1989; Dearing, 1999). Weathering played a significant role in median particle size as was noted in the subsurface samples closer to the parent material having a larger median particle size compared to finer particles nearer the surface.

The distinction between gully and sheet erosion features proved limited, only HIRM values being significantly higher for sheet erosion features. HIRM values are associated with antiferromagnetic (goethite and hematite) concentrations (Liu et al., 2007). The higher HIRM values associated with areas with sheet erosion indicate larger magnetic grain sizes (Oldfield, 1999) that would be associated with subsoils that are closer to its parent material state than a surface soil. This would suggest that the thinner layers of soil on slopes that were affected by sheet erosion, due to stripping of finer topsoil, have not been exposed to processes of soil formation for as long as slopes with deeper soils, prone to gully formation. Gullies were formed in areas where water accumulated, thus wetter conditions could promote soil formation and the formation of smaller magnetic grains. This contradicts results of surface and subsurface soils, where surface soils had greater HIRM values than subsoil. HIRM being usable to discriminate between gully and sheet erosion features was rejected as error margins were very large. Furthermore, the trend in the data where sheet erosion had a higher organic content than gullies contradicted what would be expected for gullies that were associated with wetter conditions and longer time scales of soil formation. This trend where samples from sheet erosion had a higher organic content than samples from gullies could not be validated due to high error margins.

Higher values for organic content and Xfd were all associated with wetter conditions on south-facing slopes that favour soil formation. Bacterial magnetite could be responsible for the increase in the concentration of fine-grained magnetite (higher Xfd values) (Oldfield, 1999). The finer particles on south-facing slopes would also be expected on wetter south-facing slopes which are thus exposed to longer cycles of soil formation and less subject to sheet wash removing finer particles (due to increased flow resistance associated with greater vegetation cover on south-facing slopes).

Differentiating between source samples in terms of sample depth, erosion feature and aspect was often possible, especially within each lithology. However its use in terms of catchment-wide tracing was limited, as variability within each geological grouping (e.g. surface and subsurface within igneous or sedimentary) was much lower than the variability between the geological groups. This would introduce noise into any tracing and probably confuse the conclusion of the tracing exercise.

The outcome of the statistical source differentiation showed that only one significant differentiation could be determined, thus sediment tracing was limited to igneous and sedimentary sources, unless one source was dominant, in which case it would be possible to make some deductions about contributions from topsoil or subsoil.

6.4.2. Quantitative sediment tracing

Tracing two contrasting source groups using a quantitative approach proved effective with relatively small uncertainties (<10%). Although magnetic tracers are generally inter-correlated (Lees, 1999), source groups could be differentiated based on Xlf, ARM and HIRM. Xlf and HIRM are often used as tracers as they are not well correlated to each other and are indicative of the two main magnetic components: ferrimagnetic and canted-antiferromagnetic components (Lees, 1999).

The results of the apportionment for the oldest and most complete core (VT2) on a higher flood bench indicated that sedimentary sources were dominant (64%), with only one section where sediment contribution was equal between igneous and sedimentary sources. No linear shift in the relative dominance of igneous versus sedimentary source was observed through time.

6.4.3. Qualitative sediment tracing

Magnetic susceptibility (Xlf) stood out as the main discriminator between igneous and sedimentary sources. Mzobe (2014) found similar results and ascribed it to the dominance of iron containing minerals (magnetite) in basalt compared to the very low iron concentration (maghaemite and goethite) in sedimentary sources. All the sediment samples plotted fell within the range of igneous and sedimentary sources for Factor 1, showing that sink samples were mostly a mixture of igneous and sedimentary sources. The second axis, Factor 2, correlated best with Xfd but did not give any significant grouping, thus Xfd was discarded as a tracer.

Sediment in VT2 was mainly related to sedimentary sources. An increase in igneous source was apparent for one section in the core, with no clear shift in source material for the core. These results were similar to the results for the quantitative sediment tracing.

6.4.4. Comparison of quantitative and qualitative approaches

The quantitative and qualitative approaches presented closely matched trends in the data ($R^2 = 0.95$); both showing that sedimentary sources were dominating core VT2 with an increase in igneous material in the same section of the core. The strong correlation between the two approaches indicates that the more complex quantitative approach confirms the results of the simpler qualitative results. Comparing the two methods allowed for an equation to be developed that could estimate the percentage igneous contribution based on a single Xlf measurement. The equation was not linear, indicating likely grain size interactions (Lees, 1997; Oldfield, 1999). Furthermore the relation produced a single Xlf benchmark, $3 \times 10^{-6} \text{ m}^3 \text{ kg}^{-1}$, that indicates equal contributions from both sources. Having a single reference point further aids the process of rapid source ascriptions.

Particle size plays a role in the accuracy of sediment tracing. Lenses with larger particles produced larger error ranges as selective transport could have removed some of the finer particles that were characteristic of the various sources. Reducing the particle size range for future tracing would increase the accuracy of sediment tracing.

The quantitative approach performed well using a few magnetic tracers and has the benefit of predicting actual source contributions and uncertainty margins (Collins and Walling, 2007). Increases in error margins can be accounted for by comparing affected layers to other tracers, i.e. particle size.

In comparison, the simpler qualitative approach performed well, but lacked the quantification of source and uncertainty. Using a single tracer (Xlf) that is easily measured in the laboratory with very low measurement errors (Lees, 1997), proved a practical option for further tracing in the Vuvu catchment, especially when combined with the apportionment equation (Equation 6.2).

6.4.5. Present day sources of suspended sediment

Floods generated during the wet season (December 2013–January 2014) along the Vuvu River were flashy with a peak duration of less than an hour. This was expected for a steep catchment with a high drainage density, receiving high intensity rainfall (Gregory and Walling, 1968).

Sediment produced and transported along the upper reaches of a river system was expected to be dominated by coarser material (Fryirs and Brierley, 2013). As no data on bedload transport volumes were available this could not be assessed, but measured suspended sediment concentration was very high (up to 11 g l^{-1}). Opportunistic measurements reported by Madikizela et al. (2001) (summers of 1996-1998) recorded suspended sediment concentrations of 3 g l^{-1} for the Thina River (the Vuvu River is a tributary of the Thina River) and 9 g l^{-1} for the neighbouring Kinikra River, both feeding into the larger Umzimvubu River system. As these were opportunistic samples taken when the chance presented itself, and not specifically over large flood events, concentrations were expected to be higher during peak events. Therefore measured suspended sediment concentrations for the Vuvu River were in the same order of magnitude as larger rivers in the same river system.

Suspended sediment particle size (average D50 $\sim 2\mu\text{m}$) was significantly smaller than that of source samples (average D50 ranged 60 and $94 \mu\text{m}$) and that stored in sinks (D50 ranged between 58 and $343 \mu\text{m}$). This is an indication of the selective transport of fine grained materials (clay) during lower flow events as those observed during the study. This would indicate that these relatively small events that were observed (inundating the lower flood benches) transport finer material than what was observed in the sediment sinks. Particle size is thus likely to increase for larger events as is evident from sediment stored in the lower and higher flood benches.

Within-flood event data revealed the variability of sediment sources and concentration in relation to discharge. The event on the 26th of December 2013 provides evidence that locally sourced sediments with a moderate suspended sediment concentration dominated the first flood waters, but water was later injected from the distal parts of the catchment containing very high sediment concentrations. This type of sediment peak flow is termed “anti-clockwise hysteresis” and is associated with larger elongated catchments where sediment is sourced from distal parts of the catchment during the latter part of the flood hydrograph (Heidel, 1956) or where sediment transport is slower than the propagation of the flood peak (Asselman, 1999; Fryirs and Brierley, 2013). As the Vuvu is not a large catchment, the likely reason for the sediment concentration to peak after the peak in flow discharge is attributed to the lack of sediment availability from local sources or the delay in sediment from the upper catchment that is a function of travel time due to the elongated shape of the Vuvu catchment (Fryirs and Brierley, 2013). Due to the elongated shape of the catchment, locally derived flood waters arrived first at the monitoring point (40–80 min after start of downpour), followed by sediment laden waters from higher up in the catchment towards the end of the flood (120 min after the start of the downpour).

Discharge peaked as the first flood water from the upper catchment was combined with the runoff from the lower catchment. The sediment was still dominated by sedimentary sources at the time of the highest discharge (26% igneous material) with sediment concentrations around 8 g l^{-1} . During the peak flow the channel was connected to 93% of the lower flood benches and 20% of the higher flood benches, making over bench sediment deposition with a mainly sedimentary composition possible. As the igneous rich water (>42% igneous material) was introduced, 60% of the lower benches were inundated, decreasing to 6% as the flood receded, reducing the potential to store igneous sediment on the flood benches. If the flood peak was reached earlier, it was likely that sediment deposition would have been dominated by sedimentary sediments; if the flood peak arrived later it would be dominated by an equal mixture of igneous and sedimentary material. This shows the sensitivity of the source make-up of deposited sediment. The concentration of the suspended sediment would play a role in the potential depth of sediment deposited, which would also vary depending on the timing of overbank connectivity as sediment concentration increased during the flood wave. The December event showed evidence that the entire catchment received high intensity rain and contributed sediment.

The event of the 29th of January 2014 was significantly smaller in magnitude and sediment concentration as a result of less intense rainfall compared to the abovementioned event in December. The suspended sediment data showed that sediment was derived from local sedimentary sources (<12% igneous) with the sediment concentration peaking just before peak discharge (Walling and Webb, 1982; Rovira and Batalla, 2006). This pattern is referred to as “clockwise hysteresis” and is typical of small catchments where sediment was relatively close to the main trunk and transported by the rising limb of the flood (Asselman, 1999; Fryirs and Brierley, 2013; Wilkinson et al., 2013). Sediment available for transport was exhausted during the clockwise hysteresis flow event (Rovira and Batalla, 2006). Due to the low discharge for this measured event, overbank connectivity was only established with 60% of the lower flood benches

and limited sediment could be deposited on the inundated lower flood benches due to low sediment concentrations.

Although the timing of the peak flow and peak sediment concentration was rarely synchronous, a general trend where sediment concentration increased with increased discharge was observed. This trend is well documented in the literature with a general decrease in sediment load occurring towards the end of the wet season as sediment supply is exhausted (Asselman, 1999; Rovira and Batalla, 2006; Grenfell and Ellery, 2009; Fryirs and Brierley, 2013). Grenfell and Ellery (2009) related higher sediment concentrations at the beginning of the wet season to the readily available sediment stored in the catchment during the dry winter season (lack of transport, thus stored) and/or the lack of vegetation cover after the winter. Suspended sediment concentration decreases throughout the wet season as sediment sources are exhausted and increases in vegetation cover reduce soil vulnerability.

These data on within event and seasonal sediment dynamics suggest that the majority of the sediment is transported during short pulses at the beginning of the wet season during catchment-wide high-intensity rainfall. This confirms the hydrological well-connected nature of the catchment that leads to high energy flow events that are likely to breach most buffers that exist in the catchment, transferring both water and available sediment effectively down system. This is depicted by Fryirs et al.'s (2007a) concept of switches that are on, transferring both stormflows and sediment throughout the catchment. Towards the end of the wet season stormflows may still have the capacity to transport sediment to the channel and onto the flood benches, but the depletion of available sediment would limit sediment transport and deposition. Sediment deposition on flood benches is thus not only dependent on flood bench inundation, but also on sediment availability. This would suggest that although sediment connectivity is dependent on hydrological connectivity, sediment connectivity tapers off during the wet season as the sediment supply is depleted.

6.4.6. Flood bench sediment origin

Higher and lower flood benches differed in their sediment source makeup. Although both higher and lower flood benches consisted of a mixture of igneous and sedimentary sources, higher flood benches stored sediment with a greater igneous contribution, whereas lower flood benches stored sediment with a greater sedimentary contribution. This difference can be attributed to the timing of sediment delivery from the different sources during the flood peak as noted in the previous section on suspended sediment (Section 6.4.5.). This shows that the interpretation of a sediment record is not only a linear cause-and-effect exercise, but that flood effectiveness and climate variability play an important role in what sediment is stored in sinks. The higher lying levels were only inundated when discharge was sufficient to connect the channel to the higher flood benches. This overbank connectivity was only possible when the upper and lower catchment contributed water simultaneously, possibly with a greater contribution of flow from the upper catchment due to its relatively large size. The input from the upper catchment would suggest

that the suspended sediment has a greater contribution of igneous sediment, which is seen in the higher flood bench cores.

Due to the lower relative height of the lower flood benches, these benches would be inundated more frequently and during all flood events. Smaller events were likely to transport localised sediments, thus making sediment storage in lower flood benches possible. The coarser nature of the sediment in lower flood benches could be a result of selective transport during higher flows, entraining the finer particles (Wolman and Leopold, 1957). The coarser layers also had a higher contribution from distant igneous sources as is expected for high flows when a larger part of the catchment is contributing water and sediment (larger particle size of igneous sources) (Fryirs et al., 2007a).

Cs-137 data confirmed the high erodibility of the Elliot Formation as most of the surface soils have been eroded, compared to the Drakensberg Formation that has lost less of its surface soil. Sediment containing Cs-137 in flood benches would be from sedimentary surface soils as most of the topsoil has been lost between 1958 and 2012 and was deposited on the valley fill. Present day sources of sediment containing Cs-137 are likely to be igneous surface soils as Cs-137 is still present in the majority of that material. Surface soils are eroded after long dry periods as surface cover is reduced and surface soils are exposed to erosive rainfall (Rooseboom and Harmse, 1979; Kakembo and Rowntree, 2003).

6.4.7. Terrace sediment origin

Fine grained sediment stored in the upper section of terraces was mainly of sedimentary origin and did not show a longer-term trend for sediment source dominance over the last 2 100 ($\pm 1 550$)–4 500 (± 490) years. This suggests that since the formation and subsequent incision of the highest level of the valley fill in the early Holocene, fine grained sediment sourced from the sedimentary rocks dominated the flood waters. The contribution from surface soils could not be determined using Cs-137 as it predates the fallout in 1958. Xfd was used to determine surface soil contributions for samples dominated by sedimentary sources and indicated that subsoil from the Elliot Formation was the main sediment source in these sedimentary samples. This highlights gully erosion being an important source well before recent human occupation (discounting early hunter-gatherers). It suggests that gullies always were present on the hillslopes, possibly going through various cut and fill phases, with a new extensive cut phase that was initiated over the past 50 years.

6.5. Conclusion

Igneous and sedimentary sources proved to have contrasting magnetic properties that could be used to distinguish between the two lithologies. Although other differences (such as aspect and depth) were discernible using organic content, magnetic properties and median particle size within each geological

province, between geological group differences dominated and obscured these within geological group differences, unless the dominance of a sedimentary sources could be confirmed. Low X_{fd} values could then be used to determine subsurface soil contribution. Using an additional tracer, Cs-137, which was independent of the magnetic properties, extended the surface-subsurface tracing of more recent samples (post 1958) to sediment from igneous sources.

Both quantitative and qualitative sediment tracing gave similar patterns of source dominance. The advantage of the quantitative approach is the calculation of source proportions and an estimate of error associated with the apportionment. A semi-qualitative approach was developed based on X_{lf} values and a regression equation. This semi-qualitative approach was useful for further monitoring as source dominance can be based on a single magnetic measurement. The qualitative sediment tracing approach proved sufficient for this study that investigated the change in connectivity and related sediment sources.

Suspended sediment was dominated by sedimentary sources during low and moderate events, but received large (up to 50%) contributions from igneous sources during high flow events. The results showed that suspended sediment sources changed throughout flow events and were dependent on rainfall distribution, intensity and catchment shape. Rainfall distribution affected the likely sources that were activated and rainfall intensity determined the amount of sediment that was potentially mobilised from a source. Catchment shape (distance from source to monitoring site) played a role in the timing of sediment delivery from different sources, with sediment from the distal parts of the catchment arriving during the falling limb of the flood wave. Timing differed by 40 min between proximal and distal sources. This has large implications for the character of the sediment stored on the flood benches as sediment could only be stored when the channel was connected to the flood bench. The sediment source that dominated the flood water during the time of overbank connectivity was likely to be preserved in the sedimentary record of flood benches. The timing of the flood wave and dominant sediment source was dependent on the season, rainfall distribution and connectivity. To have accurate data on all these variables was often impractical for various reasons (remoteness, safety, cost, etc.), making the interpretation of sediment chronologies challenging and often subjective. The smaller flow event proved to be dominated by locally sourced sediments. Furthermore, sediment concentration decreased throughout the wet season as a result of sediment exhaustion.

Sediment stored in flood benches was mainly from sedimentary sources, with higher flood benches storing finer sediment with a larger input from distal sources, whereas the lower flood benches mostly stored coarser sediment from local sources. This interaction was attributed to the dynamic nature of connectivity as sediment from distal sources was only transported, and the channel was only connected to the higher flood bench, during the larger flow events. The lower flood bench was connected to the channel during both large and smaller events, but sediment deposits were associated with the smaller flow events mobilizing locally sourced sediment. Locally sourced soils tended to be subsoils, as the majority

of surface soils were previously eroded off the sedimentary sources, whereas distal sediment sources contained top soils.

Terraces contained a similar mixture of igneous and sedimentary sources to that of the flood benches. Terrace samples that were dominated by sedimentary sources received a large proportion of sediment from subsoils, indicating that gullies were also a significant sediment source over the past 4 500 years. Sediment stored in the Vuvu valley fill did not show a shift in source over the last 4500 years, indicating that the entire catchment contributed sediment over this period, with expected variability in source dominance that was related to climatic changes and channel-flood bench connectivity. It can thus be concluded that the recent (last 50 years) increase in connectivity throughout the catchment has not changed the proportions of sediment eroded from the upper and lower catchments and deposited on the Vuvu valley fill over the last 4500 years.

Chapter 7: Overbank connectivity along the Vuvu River valley fill

7.1. Introduction

Overbank or lateral connectivity is important for the maintenance of river morphology (Shields Jr et al., 2000), biological processes (Amoros and Bornette, 2002; Thoms, 2003), habitat formation (Ward et al., 1999), nutrient cycling (Meyer and Likens, 1979; Junk et al., 1989; Thoms et al., 2005) and sediment storage (Erskine and Livingstone, 1999; Walling et al., 1999; Fryirs et al., 2007a). Sediment storage along the Vuvu valley fill is dependent on two main processes: sediment entrainment and sediment deposition. Sediment entrainment occurs mainly along the outer meander bends through lateral migration of the channel. Entrainment might also be induced where banks collapse or where tributaries cut through the valley fill. Sediment deposition occurs in the channel (mainly as coarse sediment such as cobble and gravel) and on the lower and higher flood benches as fine sediment. Sediment is potentially stored on the flood benches when they are inundated during high flow events, thus being dependent on overbank connectivity.

Flood magnitude and channel geometry will determine the degree of connectivity between the channel and the lower and higher flood benches. High magnitude events occur less frequently, whereas smaller magnitude events occur more frequently (Wolman and Miller, 1960; Pickup and Warner, 1976; Fryirs et al., 2007a). When this is translated into channel hydraulics it means that higher flood benches will be inundated less frequently than lower flood benches (Leopold and Maddock, 1953; Erskine and Livingstone, 1999; Amoros and Bornette, 2002; Bowen et al., 2003).

Catchment land use change (e.g. urbanisation, agriculture, etc.), degradation (decreased vegetation cover and hardened surface soils) and increased hillslope-channel connectivity lead to increases in peak discharge and increases in the frequency of large events (Booth, 1990; Schulze and Horan, 2007). The channel adjusts to this increased energy by channel deepening and widening (Leopold and Maddock, 1953; Neller, 1988; Booth, 1990; Leigh, 2010). In some cases a geomorphic threshold is crossed where incision and subsequent widening leads to a channel enlargement that is greater than the corresponding discharge. This would mean that channel-floodplain connectivity is reduced, unless a new floodplain or flood bench is formed in the enlarged channel (Leigh, 2010).

The 'flood pulse' concept of Junk et al. (1989) highlights the importance of overbank connectivity, but was extended by Tockner et al. (2000) to smaller more frequent within-channel flow pulses as these smaller pulses play an important role in system maintenance. Average overbank connectivity is assumed to vary on average between 1 to 2 years (Wolman and Leopold, 1957; Leigh, 2010), whereas several smaller flow pulses per year would be expected for free flowing rivers.

In this chapter the frequency of overbank connectivity is considered for 15 sections along the Vuvu valley fill. Measured water level data for two of the Vuvu sections were assessed to evaluate the observed inundation frequency of lower (within channel) and higher (river bank) flood benches. Longer-term rainfall records were used in hydrologic models to provide longer-term discharge records. These data were analysed for longer-term channel-flood bench connectivity for all 15 sections. The objective of this chapter is to report on the monitoring and modelling of the hydrological connectivity between channel and valley fill sediment sinks (higher and lower flood benches) (Objective 6).

7.2. Methods

The methods used to monitor shorter term (2 year) and model longer term (73 year) channel-flood bench connectivity are described in this section. An overview of the components involved in the short and longer term approaches to test the channel-flood bench connectivity frequency are presented in Figure 7.1.

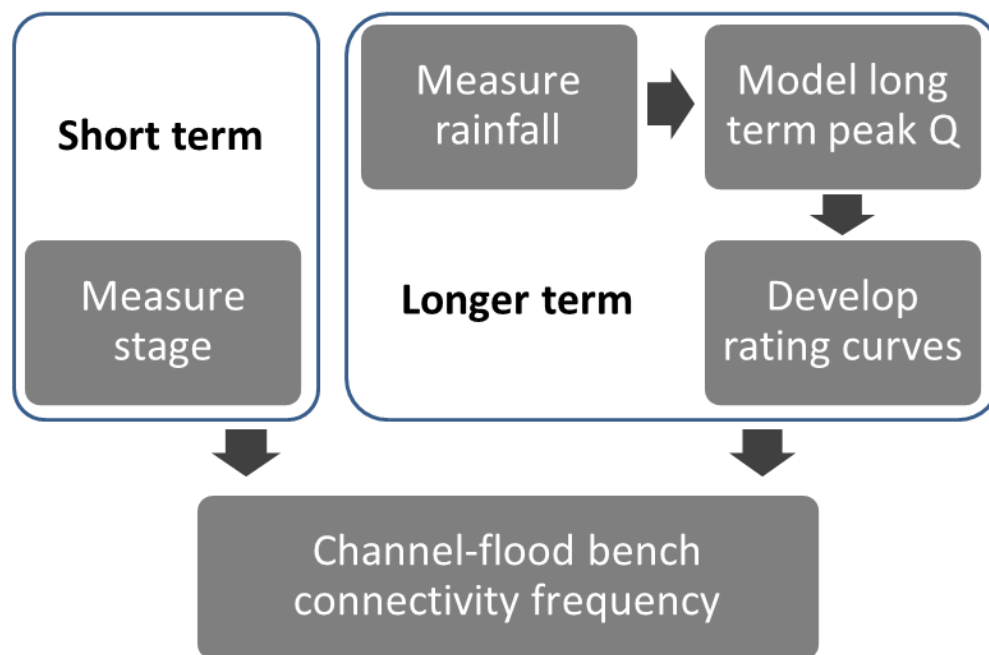


Figure 7.1: The approaches followed to measure the short term and model longer term frequency of channel-flood bench connectivity.

7.2.1. Hydro-meteorological monitoring

Rainfall was measured in the Vuvu catchment (Lundi Village; 30.60871S, 28.22882E) using a 150 mm orifice funnel tipping bucket rain gauge and HOBO® pendent event logger (Figure 7.2). The rain gauge was installed in December 2011 and data were used to model discharge for two river reaches along the

Vuvu valley fill (VT15 down to large tributary and large tributary to VT1 in Figure 7.2) as discussed in Section 7.2.2.

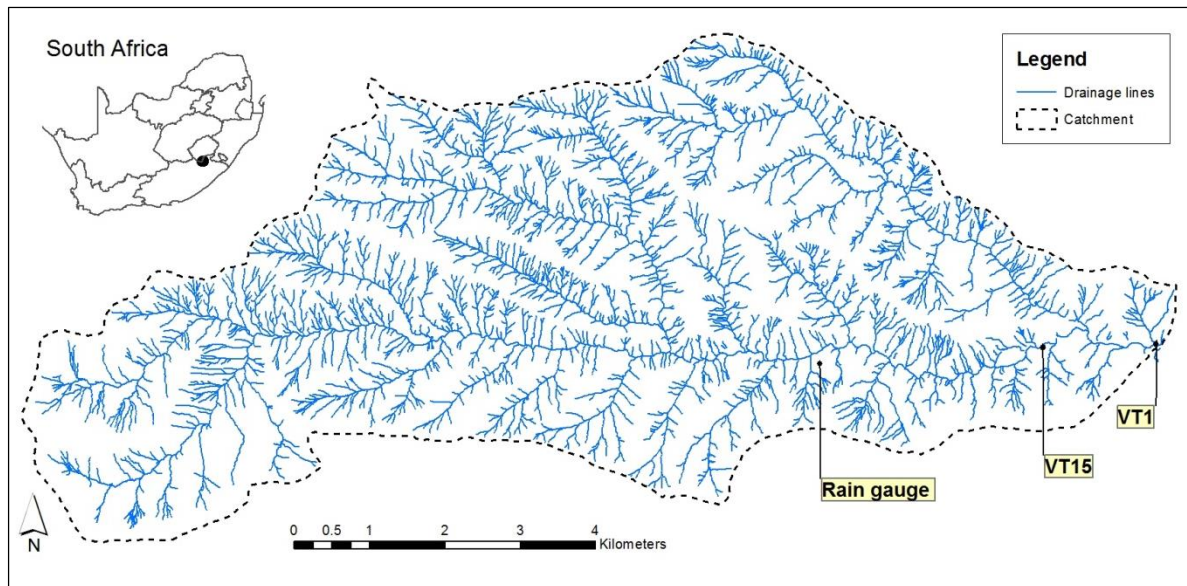


Figure 7.2: The drainage network of the Vuvu catchment and location of the rain gauge and hydraulic monitoring stations.

A hydraulic monitoring site was established at the bottom of the Vuvu valley fill at VT1 (Figure 7.2 and 7.3; 30.59900S, 28.26816E). Guidelines developed by the USGS were followed for site selection and flow measurements (Rantz, 1982). The natural bed shape was used in the hydraulic section, which was located at the top of a riffle where flow was relatively uniform throughout a range of water levels. The hydraulic section was surveyed using an Epoch Differential Global Positioning System (centimetre accuracy). Repeat discharge measurements were done for nine different stage heights, ranging from 0.31 to 1.06 m, in order to develop a rating curve for each hydraulic section (Whiting, 2003). A portable electromagnetic Flo-Mate 2000[®] was used to measure flow velocity at 0.6D (60% down the water column) in water less than 0.5 m deep and a combination of 0.2D, 0.6D and 0.8D for water deeper than 0.5 m (Gordon et al., 2004). Flow velocity was measured at more than 20 vertical points across the section for each discharge measurement.

To monitor water level, a self-logging water level sensor (Solinst Levelogger[®] with 0.5 cm accuracy) that records stage height at 20-minute intervals was installed at VT1 (installed in December 2012). The sensor was placed in a protective metal casing that was bolted onto a large boulder (Figure 7.3) on the hydraulic transect. In addition to the level sensor installed at the hydraulic section VT1, sensors were installed at VT7 and VT15 to measure stage or water level. Unfortunately the logger at VT7 was ripped off the boulder during December 2012 to March 2013 and was not replaced. The entire boulder and logger at VT1 was washed away during the monitoring period December 2013 to January 2014. A new logger was installed at VT1 in January 2014. Level data from VT15 was used to complete the dataset for VT1 (as

described in Section 6.2.4.). These stage data for December 2012 to May 2014 were used to calibrate the relevant hydraulic models and assess the frequency of observed flood bench inundation.



Figure 7.3: Hydraulic site VT1 and underwater image (bottom left) of the level sensor within a metal casing bolted to a large boulder.

The level readings were corrected for changes in air pressure by using data captured at the same time intervals by an air pressure sensor (Barrologger[®]) located near the rain gauge. These barometrically corrected stage data were converted to instantaneous discharge using rating curves as discussed under hydraulic modelling (Section 7.2.3). The measured instantaneous discharge was converted to mean daily discharge and regressed against the maximum daily discharge to assess the relationship between mean daily and peak daily flow. Measured peak daily discharge also was regressed against the daily rainfall to assess the relationship between rainfall and peak discharge.

7.2.2. Hydrological modelling

Hydrological modelling was performed to calculate peak discharge volumes for various event frequencies under present day conditions and to simulate a rehabilitated scenario of a future condition where vegetation cover is improved and hillslope-channel connectivity (drainage efficiency) is decreased.

As there were only short-term daily rainfall data available for the Vuvu catchment (2012-2014), daily rainfall data for the Matatiele station (1942–2014; 65 km northeast of the Vuvu station) were used to extend the rainfall record. Both the daily Matatiele and Vuvu rainfall data were sorted from large to small and plotted on the same frequency distribution plot. The daily Matatiele data were manually up-scaled, using a correction factor ranging from 1 to 1.98 depending on the difference between the Matatiele and Vuvu dataset, to match the daily rainfall frequency distribution pattern measured for the Vuvu catchment. This scaled long-term daily rainfall data for Matatiele were chronologically ordered and converted to

monthly rainfall data. The monthly Matatiele data were used as input to the monthly Pitman rainfall-runoff model (Hughes, 2013) with parameter values based on existing regional information (Midgley et al., 1994). Monthly discharge volumes were calculated for catchment areas of 47 and 58 km² for VT15 and VT1 respectively. Simulated monthly discharge volumes were disaggregated to mean daily flows using the scaled daily rainfall data for Matatiele and a model developed by Slaughter et al. (2015).

As instantaneous peak flows were required to test channel-flood bench connectivity it was necessary to further disaggregate the mean daily flows that were larger than $>5 \text{ m}^3 \text{ s}^{-1}$ (the lowest threshold observed to inundate the lower flood bench at VT3; Appendix 4). The first step in this process was to separate the baseflows from the total mean daily flows using a digital filtering method (Smakhtin, 2001). In this context, the baseflow is assumed to include the recession flow volume from previous days within an event lasting several days and therefore can be higher than traditional definitions of baseflow that assumed an almost constant contribution from baseflow (Figure 7.4). The mean daily baseflow calculated according to Smakhtin (2001) was deducted from the mean daily flow and the resultant volume was used to determine the peak discharge.

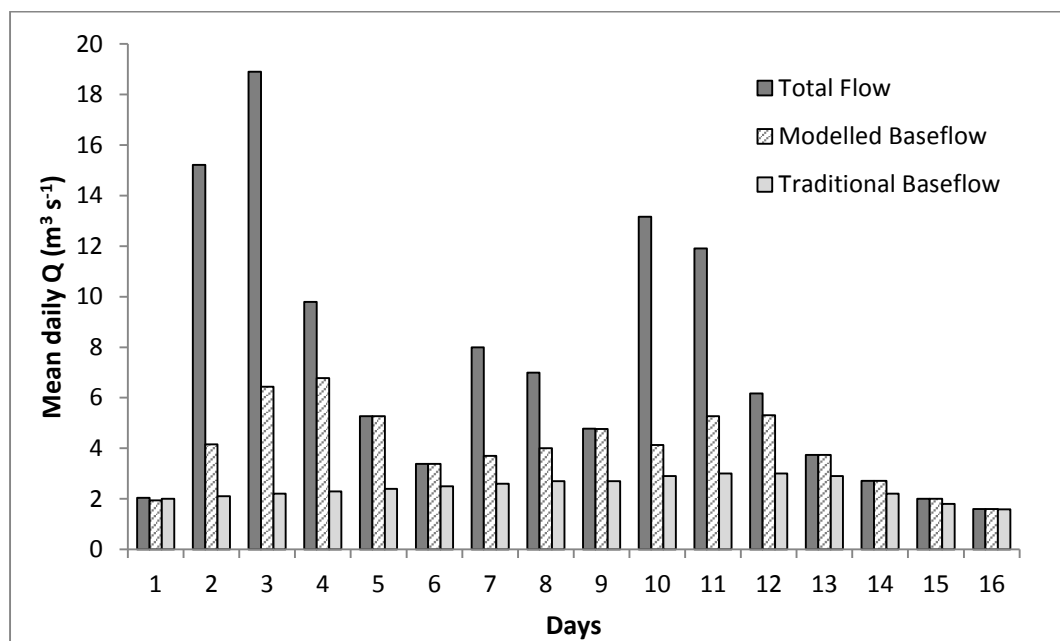


Figure 7.4: An example of total flow and modelled baseflow for the Vuvu River. Modelled baseflow using traditional definitions is included to show how it differs from the modelled baseflow.

Observed flood peaks (Figure 7.5) mostly had a triangular shape, with a steep rising and falling limb, followed by a relatively flat recession limb. Two of the flood peaks (marked X on Figure 7.5a) were drawn out instead of having a steep peak climax. These two events were a result of continued low intensity rain that produced lower ($<15 \text{ m}^3 \text{ s}^{-1}$) peak discharges. A correction factor was calculated to compensate for the higher baseflows that result from recession flows following previous wetter periods. This portion was calculated based on the ratio between mean daily baseflow and mean daily discharge (Equation 7.1). The peak discharge was calculated using Equation 7.2 that consists of two parts that are added together,

a triangle that represents the flood flow and a base that represents the mean daily baseflow. The triangle height was determined by the volume of water available (daily flow volume minus baseflow volume) and the length of the base of the flood flow triangle. The length of the base of the flood flow triangle was set at 2, 4 and 6 hours to account for uncertainties and modelling error (Figure 7.5b). The resultant triangle height was added to the height of the mean daily baseflow to produce a total peak discharge (Equation 7.2).

$$\text{Correction factor} = \frac{\text{mean daily baseflow}}{2 \times \text{mean daily } Q} \quad (7.1)$$

$$\text{Peak } Q = \frac{\text{mean daily } Q - \text{mean baseflow}}{2 \text{ base} \times \text{correction factor}} + \text{mean baseflow} \quad (7.2)$$

Base is the length of the flood peak (2, 4 or 6 hours).

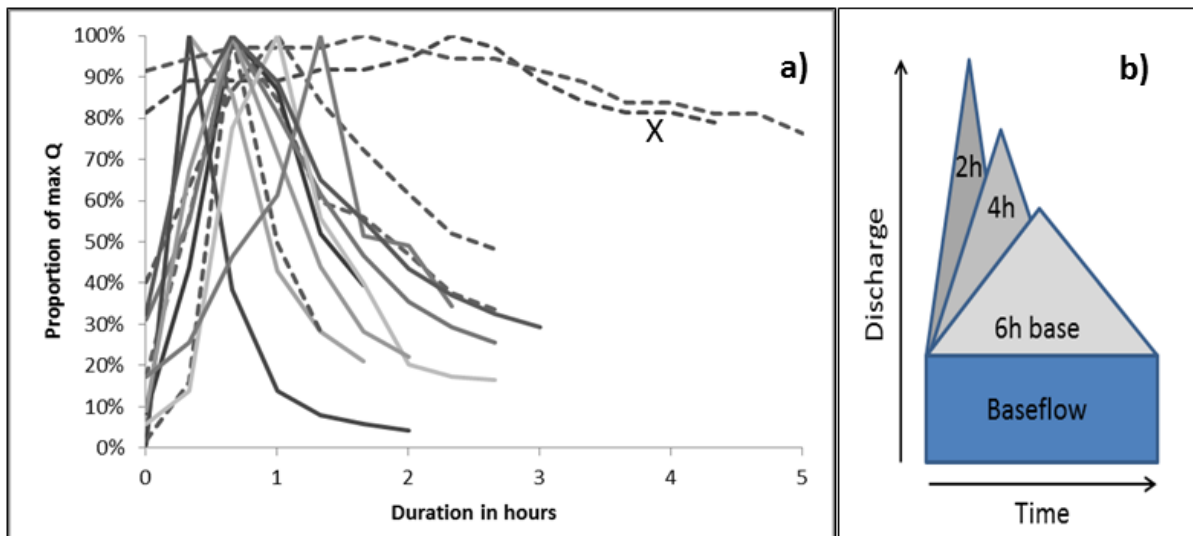


Figure 7.5: a) Observed flood peaks for the Vuvu River. Flood peaks were standardised against maximum discharge for each event. X marks the extended flood events. Events $<15 \text{ m}^3 \text{ s}^{-1}$ are indicated with dotted lines. b) A diagram showing how flood duration influences the height of the flood peak.

Observed flood peaks, including the two extended floods marked X on Figure 7.5a, with a maximum discharge of more than $7 \text{ m}^3 \text{ s}^{-1}$ (stage of 1 m), had a mean flood duration of 2.98 ± 0.95 hours. As there was uncertainty introduced by the various steps of scaling the rainfall data (generating and disaggregating monthly flow data) it was reasoned that disaggregating the mean daily flow to instantaneous flow should have a base ranging from 2–4 hours, which will account for the total uncertainty in the modelling.

A future post-rehabilitation scenario was introduced where flow delivery rates were assumed to be reduced as reduced hillslope-channel connectivity would increase the runoff time. The disaggregation of

the mean daily to instantaneous flow was adjusted so that the base of the peak discharge triangle was 2 hours wider than for current conditions, thus 4–6 hours. The largest independent peak was selected for each rainfall event, as multiple peak flows (over 2–5 days) existed during times of sustained rainfall. The resulting peak daily discharge data were ranked, giving a partial duration discharge-frequency dataset. The Weibull formula was used to calculate the average recurrence interval (Gordon et al., 2004). This disaggregation process using a flood base of 4-6 hours only accounted for the reduction in hillslope-channel connectivity and not for the increase in infiltration that is associated with the rehabilitation scenario.

The partial duration discharge-frequency dataset was used to determine the frequency of channel-flood bench connectivity for the current condition (2-4 hour flood base) and a rehabilitated condition (4-6 hour flood base).

7.2.3. Hydraulic modelling (assessing lateral connectivity)

Stage and discharge data were collected in the field and longer term (73 years) peak discharge datasets were generated as described in Section 7.2.2. An additional 13 sections were surveyed between VT1 and VT15 as described in Section 5.2.1. In order to link all these data in terms of channel-flood bench connectivity, a stage-discharge curve was developed for each section, based on the Manning’s equation (Table 7.1).

Table 7.1: Details of hydraulic sections and approach followed to develop rating curves.

Section	Stage recorded	Roughness (Manning’s n)	Slope	Method to develop rating curve
VT1	Yes	0.03-0.12	0.0099-0.0188	Measured discharge at 9 intervals up to 1.06 m or 7.3 m ³ s ⁻¹ , modelled rating curve up to 4 m based on Manning’s equation
VT2	No	0.03-0.12	0.0140-0.0160	Rating curve based on Manning’s equation
VT3	No	0.03-0.12	0.0093-0.0160	Rating curve based on Manning’s equation
VT4	No	0.03-0.12	0.0080-0.0160	Rating curve based on Manning’s equation
VT5	No	0.03-0.12	0.0070-0.0110	Rating curve based on Manning’s equation
VT6	No	0.03-0.12	0.0074-0.0118	Rating curve based on Manning’s equation
VT7	No	0.03-0.12	0.0161-0.0173	Rating curve based on Manning’s equation
VT8	No	0.03-0.12	0.0161-0.0173	Rating curve based on Manning’s equation
VT9	No	0.03-0.12	0.0164-0.0173	Rating curve based on Manning’s equation
VT10	No	0.03-0.12	0.0231-0.0173	Rating curve based on Manning’s equation
VT11	No	0.03-0.12	0.0091-0.0173	Rating curve based on Manning’s equation
VT12	No	0.03-0.12	0.0170-0.0172	Rating curve based on Manning’s equation
VT13	No	0.03-0.12	0.0205-0.0265	Rating curve based on Manning’s equation
VT14	No	0.03-0.12	0.0265-0.0284	Rating curve based on Manning’s equation
VT15	Yes	0.03-0.10	0.0167-0.0260	Rating curve based on Manning’s equation

A rating curve was completed for each of the 15 transects using the hydraulic sub-model (based on the Manning's equation) that is part of the Revised Desktop Model (RDM) (Hughes et al., 2014). The hydraulic sub-model required a surveyed channel cross section (assumed to remain unchanged for the modelling exercise), minimum and maximum channel slope reading and minimum and maximum channel roughness value (Manning's n). Slope and roughness were scaled by the hydraulic sub-model as stage was increased. Minimum slope was related to the surface slope of flood water and was calculated based on an average low flow water level over a 500 m (250 m on either side of the hydraulic section) section of surveyed river channel. Maximum slope was related to the water surface slope during low flows, thus slope was calculated over a distance stretching from the bottom of a riffle upstream to the bottom of a riffle downstream. The channel roughness values were based on the rating values for VT1 and expert knowledge (Chow, 1959) and ranged from 0.03 to 0.12 (Table 7.1).

The modelled rating curve for VT1 was then fine-tuned based on measured stage and discharge data, using the gradient shape factor and roughness shape factor in the hydraulic sub-model, until the modelled curve and measured points matched. A hydraulic model was developed for each transect based on measured transect and gradient data, whereas gradient and roughness shape factor data, 10 and 5 respectively, were copied from that used for VT1 (various combinations of gradient and roughness shape factors are given for VT1 in Appendix 4).

Higher and lower flood benches were identified on each hydraulic section and stage threshold levels were determined for inundation of lower and higher flood benches. These threshold stage levels were converted to discharge volumes based on the rating curve for the cross section. This gave a discharge threshold at which flood bench inundation was initiated. The threshold values were used to assess channel-flood bench connectivity concepts based on the shorter term observed discharge data and longer term modelled peak discharge data for the 73-year record (including flood peak values based on a 2, 4 and 6 hour flood duration base).

Inundation frequencies for VT1 were also assessed for dry, moderate and wet years (water years starting on 1 October). The years were ranked based on the total discharge volume for that year. Inundation frequencies were extracted for the 10 driest, 10 moderate (10 years centred around the median total discharge per year) and 10 wettest years. The difference in inundation frequency for the current increased hillslope-channel connectivity (flood duration ranging from 2 to 4 hours) and reduced hillslope-channel connectivity (flood duration ranging from 4 to 6 hours) was determined using Equation 7.3. Flood frequencies were limited to 10 years as the data were subsampled.

$$\% \text{ difference} = \left(\frac{A+B}{B+C} - 1 \right) * 100 \quad (7.3)$$

where A, B and C are the inundation frequencies for flood peak with a 2 hour base, a 4 hour base and a 6 hour base respectively.

7.3. Results

7.3.1. Hydro-meteorological monitoring

The measured total rainfall for Matatiele was ca. 30% less than that for the Vuvu village (over period 6 December 2011 to 31 January 2014), and rain days were not always experienced at the same time. The scaled Matatiele data fitted the measured Vuvu data well with similar exceedance percentages for corresponding rainfall volumes (Figure 7.6). The maximum peak daily rainfall for the measured Vuvu data was 89 mm and for the scaled Matatiele data was 103 mm, a 16% difference.

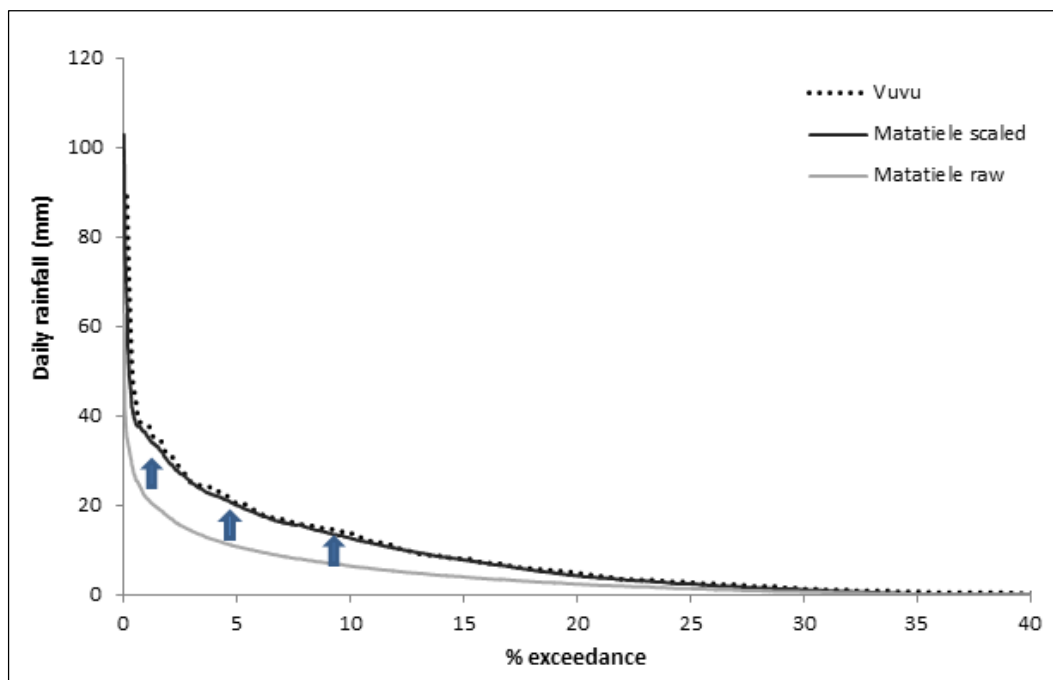


Figure 7.6: A comparison of the % exceedance of scaled daily rainfall for Matatiele and the measured rainfall for the Vuvu village.

The observed daily mean-peak discharge relationship for the Vuvu correlated moderately ($R^2 = 0.6537$) and followed an exponential curve (Figure 7.7a). For the large magnitude events, the daily maximum flow could be up to 10 times greater than the daily mean flow, showing the flashy nature of the Vuvu catchment. A few anomalies existed, such as Point A far above the trendline (Figure 7.7a). This point indicated a very flashy flow during a day with generally low river levels (>20 times average daily flow). Point B fell far below the trendline, a result of sustained gentle rainfall (indicated by an X in Figure 7.5a), leading to a high mean daily discharge relative to the peak daily discharge (<2 times mean daily

discharge). In Figure 7.7b the lower edge of the data cloud indicates the 1:1 ratio where peak and mean daily flow were equal. Points above the 1:1 ratio indicate greater daily flow events, with distance perpendicular to the 1:1 ratio indicating the flashiness of the flow event.

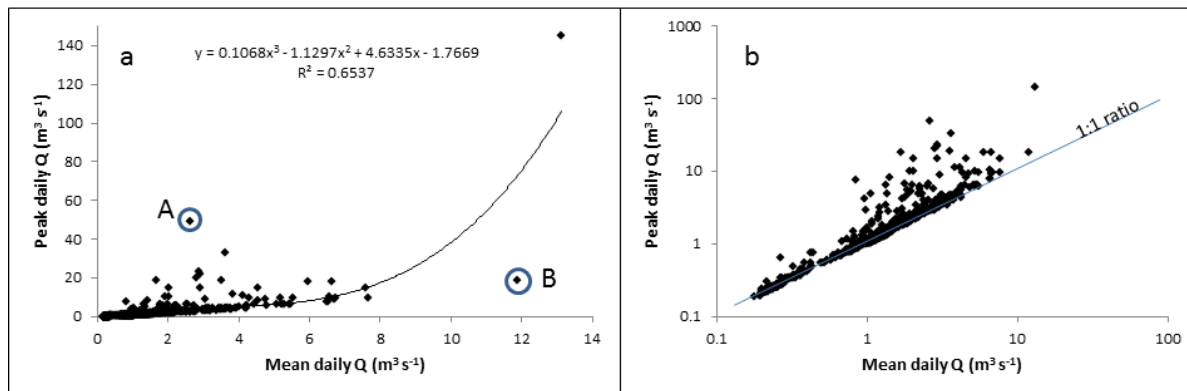


Figure 7.7: A biplot showing the relationship between mean daily discharge (Q) and peak observed Q for the lower Vuvu at VT1 on a) a normal axis and b) a log axis.

The relationship between daily rainfall and peak discharge correlated well ($R^2 = 0.8273$) (Figure 7.8). Similar outliers to those marked in Figure 7.7 can be seen and reflect similar trends of high and low intensity rainfall which influence the peak discharge. High intensity rainfall events, such as the event on the 17th of December 2012 that had a maximum intensity of 10 mm over 5 min (total daily rainfall 89 mm), were associated with flashy flow regimes ($144 \text{ m}^3 \text{ s}^{-1}$) (A in Figure 7.8). In Figure 7.8 it can be seen that daily rainfall events of 20-30 mm act as a threshold where discharge increases significantly for larger daily rainfall events as can be seen by the upward curve of the trendline around 20-30 mm.

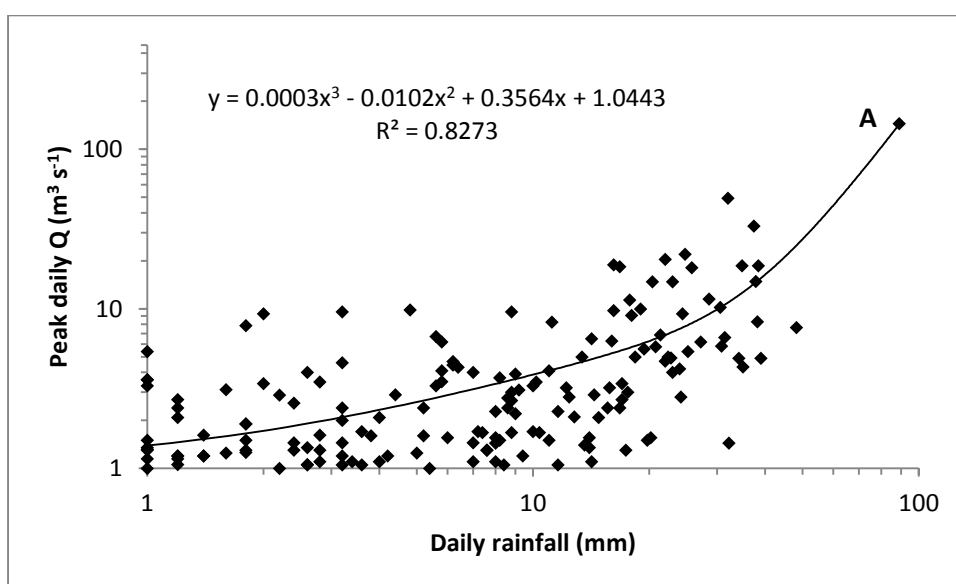


Figure 7.8: A biplot showing the relationship between the daily rainfall for the Vuvu and the observed peak daily discharge for the lower Vuvu River (VT1). A polynomial trendline was fitted.

7.3.2. Hydrological modelling

The modelled discharge data (4 hour base) fitted the observed data well for the higher discharge volumes ($>140 \text{ m}^3 \text{ s}^{-1}$), but more inaccuracy was observed for the lower discharge values with over- estimations of modelled data for lower discharges ($<100 \text{ m}^3 \text{ s}^{-1}$) (Figure 7.9). The 2-hour flood base overestimated the flood peak discharge for all discharge volumes, whereas the 6-hour base underestimated the peak flow volumes for observed volumes $>100 \text{ m}^3 \text{ s}^{-1}$. For lower discharge volumes the modelled data were higher than those estimated from observed levels, with the 6-hour flood base modelled data being the closest to the observed data.

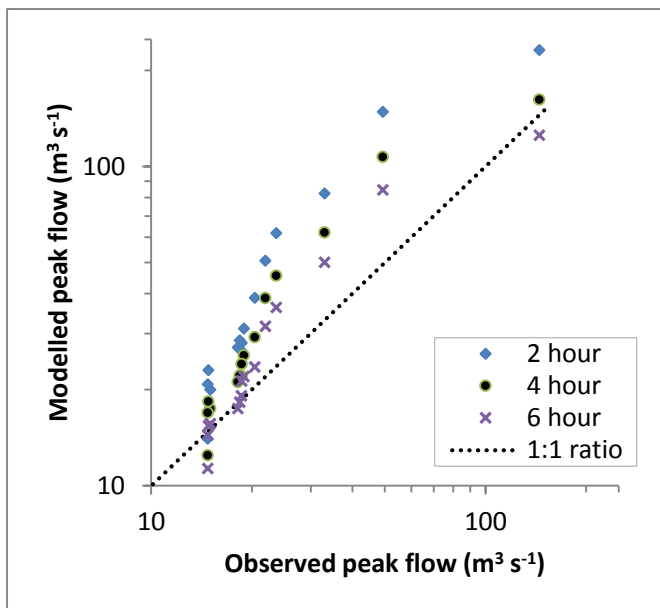


Figure 7.9: Observed versus modelled peak daily flow with a 2, 4 and 6 hour base. The dotted line indicates a 1:1 ratio.

Various flood return frequencies for transect VT1, from one in 73 years to less than a year, are given in Figure 7.10. Peak discharge decreased for events with a longer flood duration base, e.g. a flood with an estimated return period of 5 years will produce a peak discharge of ca. $160 \text{ m}^3 \text{ s}^{-1}$ (2 hour base) , $110 \text{ m}^3 \text{ s}^{-1}$ (4 hour base) and $90 \text{ m}^3 \text{ s}^{-1}$ (6 hour base) (Figure 7.10).

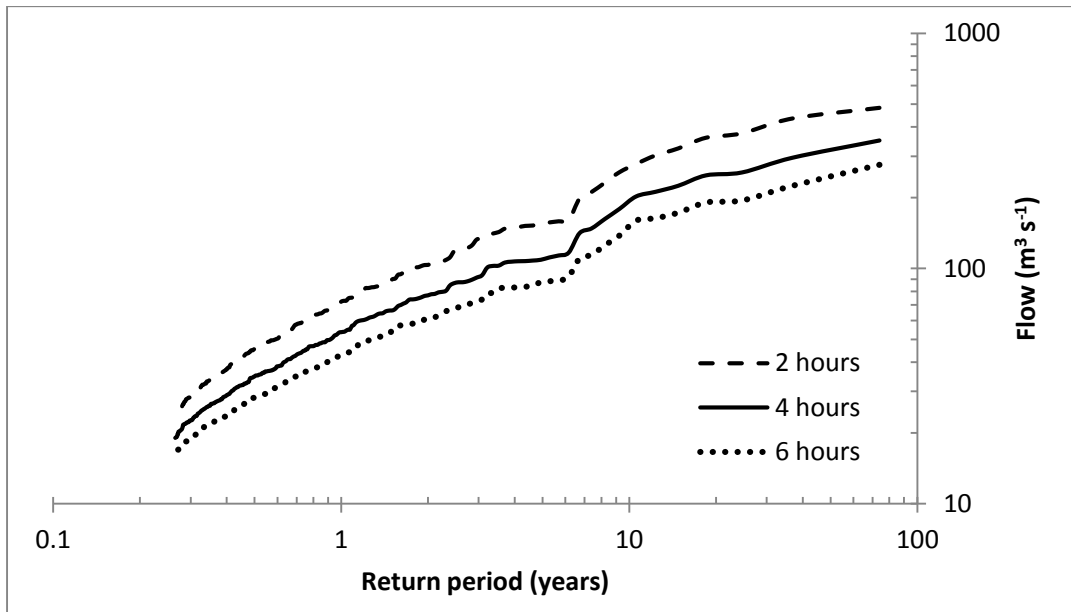


Figure 7.10: Partial duration discharge-return period graph with a flood duration of 2, 4 and 6 hours for the Vuvu River.

7.3.3. Hydraulic modelling and lateral connectivity

7.3.3.1. Hydraulic cross sections and rating curves

The channel cross-section and rating curve for VT1 can be seen in Figure 7.11. The higher flood bench is clearly visible at 3 m, whereas the lower flood bench is less obvious at 1.4 m due to vertical exaggeration and the lower bench sloping down towards the channel (Figure 5.5). The calculated rating curve matched the measured discharge and stage points.

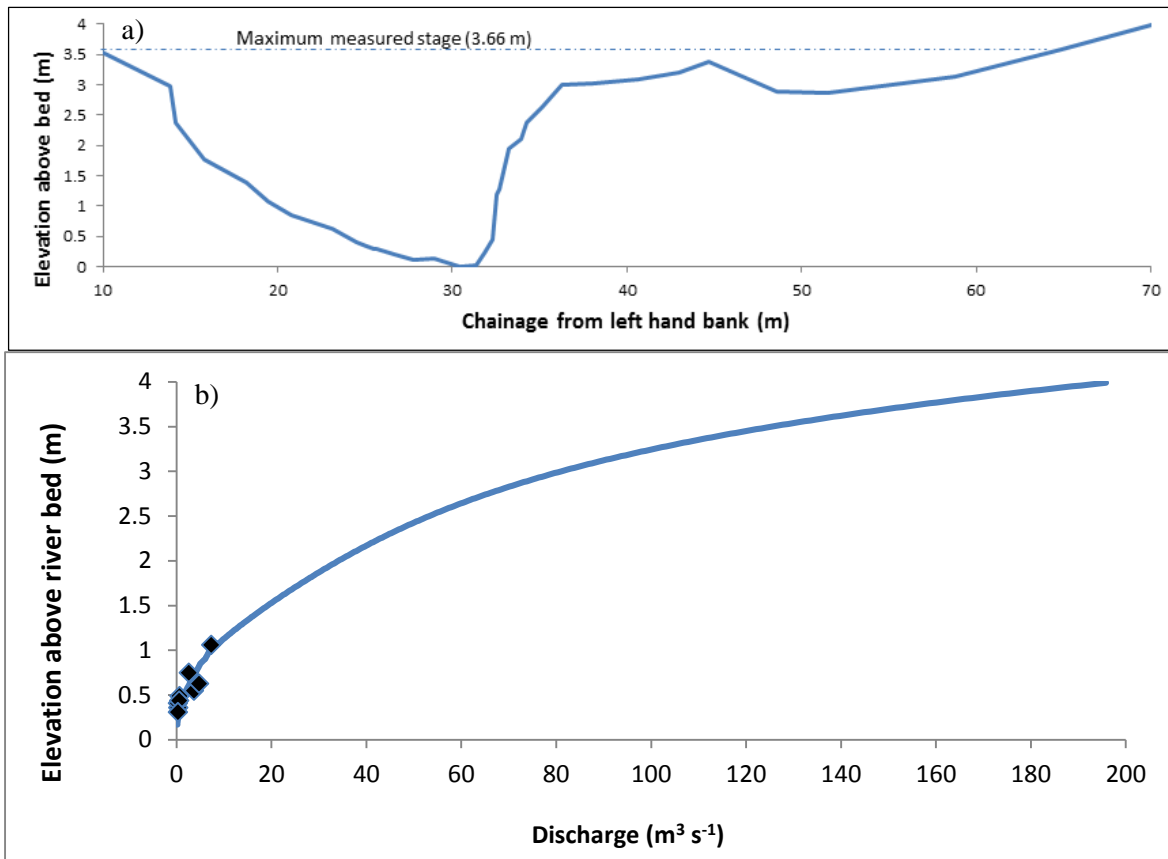


Figure 7.11: The a) channel cross section and b) rating curve for VT1. Note the measured discharge and stage points (diamonds) on the modelled rating curve (line).

7.3.3.2. Assessing the frequency of channel-flood bench connectivity

Higher and lower flood bench inundations were initiated at flood discharges that differed factors from 3 to 4 times (the range for all transects was 5–40 m³ s⁻¹ for lower benches and 15–167 m³ s⁻¹ for higher benches) (Table 7.2 and Appendix 5). Observed data recorded lower flood benches being inundated 1.5–4.5 times a year, whereas higher flood benches were inundated 1 in 2 years (monitoring period of 2 years). Modelled data for a 73-year period showed similar inundation frequencies for the lower flood benches (2.5–4.4 for base lengths 2–6 h), with decreasing frequency as the flood peak base was extended from 2 to 6 hours (Table 7.2). Inundation frequency dropped from 4.4 to 3.8 for the lower flood bench for VT1; 3.7 years to 2.5 years for VT15 and 4.4 years to 3.8 years, on average, for VT2–VT14 over the range of flood base durations (2–6 h). The higher flood bench inundation showed a similar trend with a decrease in frequency (1 in 1.2 to 1 in 3.5 years for VT1; 1 in 3.6 to 1 in 10.4 years for VT15; 1 in 1.6 to 1 in 4.3 years on average for VT2–VT14) for a longer flood base (Table 7.2).

Table 7.2: Observed and modelled (average for 73 years) inundation frequencies for the lower and higher flood benches. Average values were given for VT2–VT14 (values for all 13 transects given in Appendix 5).

Section	Feature	Inundation threshold		Average per year or recurrence interval			
		Stage (m)	Q (m ³ s ⁻¹)	Observed	2 hour	4 hour	6 hour
VT1	Lower flood bench	1.4	16	4.5	4.4	4.2	3.8
	Higher flood bench	3	82	1 in 2	1 in 1.2	1 in 2.4	1 in 3.5
VT15	Lower flood bench	1.5	24	1.5	3.7	3.2	2.5
	Higher flood bench	2.9	144	1 in 2	1 in 3.6	1 in 7.3	1 in 10.4
Ave for VT2-VT14	Lower flood bench	1.2	15	-	4.4	3.9	3.8
	Higher flood bench	2.2	76	-	1 in 1.6	1 in 3	1 in 4.3

Flood bench inundation frequency varied between dry, moderate and wet years (Table 7.3). Note that the recurrence frequency was limited to 10 years due to the extraction of 10 years with dry, moderate and wet conditions. During modelled dry years the lower bench (VT1) experienced an inundation frequency of 1.0 to 1.2 per year depending on the flood base duration and the higher flood bench had an inundation frequency of less than 1 in 10 years for a 2-6 hour flood base. During the scenario of moderate years the lower flood bench had an inundation frequency of 3.5 to 4.2 events per year, whereas the higher flood bench had an inundation frequency of 1 in 1.0 years to 1 in 1.7 years (2-6 h hour flood base). For modelled wet years the inundation frequency increased to 5.9 to 7 events per year for the lower flood bench and 1 in 0.6 to 1 in 1.6 years for the higher flood bench over the range of flood base durations. These data suggested that the observed monitoring period was moderately wet as observed inundation frequencies for VT1 were similar to those for moderate to wet years (Table 7.2 and 7.3).

Table 7.3: Modelled inundation frequencies for the lower and higher flood bench of VT1 during the 10 driest, moderate and wettest years. Percentage change in inundation frequency was given for the 2 to 4 and 4 to 6 hour flood duration.

Year		Inundation		Average per year or recurrence interval			Change %
		Stage(m)	Q(m ³ s ⁻¹)	2 hour	4 hour	6 hour	
Dry	Lower flood bench	1.4	16	1.2	1.2	1.0	9.0
	Higher flood bench	3	82	<1 in 10	<1 in 10	<1 in 10	-
Mod	Lower flood bench	1.4	16	4.2	4.1	3.5	9.2
	Higher flood bench	3	82	1 in 1.0	1 in 1.4	1 in 1.7	23.5
Wet	Lower flood bench	1.4	16	7.0	6.5	5.9	8.9
	Higher flood bench	3	82	1 in 0.6	1 in 0.9	1 in 1.6	37.0

The inundation frequency was shown to decrease for longer flood durations (Table 7.2 and 7.3). For the current catchment conditions with increased hillslope-channel connectivity (flood duration of between 2 and 4 hours), the frequency of inundation of lower flood benches was ~9% higher than for a modelled rehabilitated catchment condition with reduced slope to channel connectivity (flood duration of between 4 and 6 hours; Table 7.3). For the higher flood bench the difference was ca. 23-37 percent between increased and reduced hillslope-channel connectivity for moderate and wet years respectively. Wet years showed a larger difference between increased and reduced hillslope-channel connectivity. This difference

could not be determined for the dry years as no inundation of the higher flood bench was detected in the modelling for the 4 and 6 hour flood base.

7.4. Discussion

7.4.1. Hydro-meteorological monitoring

Differences in rainfall volume and timing between the Matatiele and the Vuvu monitoring station indicated the spatial and temporal variability and localised nature of rainfall along the Drakensberg Escarpment of the eastern seaboard. The rainfall volume was greater towards the foothills of the escarpment and was linked to the topographic influence of the escarpment as the Vuvu rain gauge was situated at a higher altitude (100 m difference) and closer to the physiographic barrier (escarpment) than the Matatiele gauge (20 km away from the escarpment). Nel and Sumner (2006) described a similar orographic influence of the Drakensberg Escarpment where altitude and distance from the escarpment played an important role in increasing rainfall below 2100 masl. Larger rainfall events were mainly localised late afternoon thunderstorms related to the orographic influence of the escarpment (Nel, 2007; Nel and Sumner, 2007; Nel, 2008). Maximum rainfall intensity (5 min rainfall intensity up to 120 mm h⁻¹) was similar to that measured for the northern Drakensberg by Nel and Sumner (2007) (5 min rainfall intensity up to 144 mm h⁻¹).

Despite scaling the 73-year Matatiele dataset to the 2-year observed Vuvu dataset for a wet period, it produced maximum daily rainfall that was only 16% higher than that observed for the Vuvu catchment. This would suggest that the scaling exercise produced relatively conservative maximum rainfall data, as only rainfall with >0.1% exceedance was upscaled as the maximum rainfall for Matatiele with <0.1% exceedance was greater than that of the Vuvu catchment and thus not upscaled (correction factor of 1).

The flashy nature of the Vuvu catchment was highlighted by the short flood duration (3±1 hours) and the high ratio (<1:10) of peak daily flow to the mean daily flow. Taguas et al. (2008) measured peak daily to mean daily ratios of <5:1 for steep headwater catchments in southeast Spain receiving intense summer rainfall. Higher ratios were associated with high intensity rainfall and catchments with sparse vegetation cover (Taguas et al., 2008). The high ratio observed in the Vuvu catchment indicates that the slopes are well connected to the channel, routing surface water quickly to the lower catchment (Patton and Baker, 1976; Costa, 1987). Rainfall intensity played a large role in the flashiness of a flow event as gentle sustained rain reduced the peak to mean daily discharge ratio and increases in rainfall intensity increased the ratio (Patton and Baker, 1976; Taguas et al., 2008). High intensity rainfall events were dominant during the monitoring period, producing flashy high energy flows.

Good hydrological hillslope-channel connectivity was further confirmed by the positive relationship between daily rainfall and peak daily flow, as the majority of the rainfall events resulted in high flows at the basin outlet within the same day.

7.4.2. Hydrological modelling

The disaggregation of the modelled mean monthly and mean daily data overestimated the flow data (base ranging from 2-6 hours) for discharges $<140 \text{ m}^3 \text{ s}^{-1}$, but produced data that were comparable to those observed for the Vuvu flow gauge using the 4 hour base for larger events ($>140 \text{ m}^3 \text{ s}^{-1}$). The 2-hour flood base overestimated all discharges and the 6 hour base underestimated discharges above the observed $100 \text{ m}^3 \text{ s}^{-1}$. Although modelled peak flows were mostly higher than observed peaks for the present day conditions (2–4 hour base), they were the best available estimates as data for further calibration were limited.

The modelled partial duration discharge-frequency data showed the expected trend where a shorter flood base increased peak flood flows. Nash (1957) showed a similar trend where a shorter flood hydrograph produced a higher flood peak for the same total flood volume. This is further noted in the literature where increases in flow pathways to the main channel are known to increase the efficiency of runoff routing (Wemple et al., 1996) and results in greater peak flow events (Harr et al., 1975).

From the modelled Vuvu discharge-frequency data, peak flood flows differed by 42% between the 2 and 4-hour base and by 25% between the 4 and 6-hour base. The peak flow model was thus more sensitive to shortened flood duration compared to lengthened flood duration. The larger difference between the 2-4 hour base compared to the 4-6 hour base would suggest that there were larger uncertainties associated with a shorter flood base compared to a longer flood base. Similar increases in stormflow were modelled by Schulze and Horan (2007) for the larger Thina catchment, where stormflows were increased by up to 30% due to the present day decreased infiltration resulting from land degradation. Stormflow differences of up to 50% were recorded by Jones and Grant (1996) for catchments in western Oregon where hillslope-channel connectivity was increased and vegetation cover was reduced. This shows that increased hillslope-channel connectivity will increase the effectiveness of runoff routing and result in greater peak flows and flow energy.

7.4.3. Hydraulic modelling and lateral connectivity

Hydraulic models proved useful to test the channel-flood bench connectivity frequency, although the modelling was based on natural channels and limited flow observations, conditions which can be expected to introduce uncertainty (Rantz, 1982). For VT1 the observed flow readings were limited to $7.3 \text{ m}^3 \text{ s}^{-1}$,

thus higher readings (up to $200 \text{ m}^3 \text{ s}^{-1}$) had to be extrapolated by modelling. Calculations for the other transects made use of modelling which resulted in larger uncertainty when compared to VT1.

Flood benches were mostly clearly defined horizontal bank-attached depositional features, but in some cases, especially for the lower flood benches, the depositional features were small and the depositional surface sloped towards the river channel. The sloped nature of some features made defining the inundation level challenging as the inner and outer edge of the bench were at different levels. Erskine and Livingstone (1999, p446), working in Australia, defined benches as “depositional landforms that are essentially tabular, often vegetated, elongate, discontinuous, sometimes paired, usually bank-attached sediment bodies”. The smaller sloped features do not strictly conform to Erskine and Livingstone’s (1999) definition of benches, but were included as functionally they were sediment stores that were inundated at intermediate flow levels. In a field study by Erskine and Livingstone (1999), up to 41% of the benches had sloped surfaces, showing that it is relatively common. They also found that there was no significant difference in the inundation frequency results when using the inundation threshold for the lower edge of a bench or the average bench level (Erskine and Livingstone, 1999).

Sediment sink inundation thresholds for the Vuvu varied significantly between lower and higher flood benches. Although there was some overlap between lower and higher flood benches, on average the inundation threshold differed by a factor 3 - 4.

The inundation threshold also varied within the group of lower benches and the group of higher benches. This variability shows some variation in the depositional environment, such as channel slope, velocity and accommodation space for the various transects (Erskine and Livingstone, 1999; Vietz et al., 2004). The age of the depositional feature also plays a role in the inundation threshold as benches are affected by cut and fill processes (Gupta and Fox, 1974). Benches often are removed by large events and rebuilt by smaller floods (Gupta and Fox, 1974; Erskine and Livingstone, 1999). Young features are often inundated at lower levels as sediment build-up has not increased the feature’s elevation to that of more mature benches.

Another possibility is that certain features were higher relative to the channel bed as a result of bed lowering or incision (Erskine and Livingstone, 1999; Leigh, 2010). This often is the case where a geomorphic threshold is crossed, such as increased stormflow that has greater energy to entrain bed material (Leigh, 2010). This is likely in the Vuvu as increased hillslope-channel connectivity increased stormflow and incision was noted for the Vuvu where incision was a recent (<60 years) phenomena (Chapter 5) that was likely to increase the inundation threshold of features as the channel capacity has been enlarged. Where the Vuvu channel has been incised without lateral movement, the inundation threshold for the lower flood bench will be higher than the previous unincised thalweg level. However, where the Vuvu channel has migrated laterally, a new lower flood bench was formed at a lower level that corresponded to the present day thalweg channel, allowing similar channel-flood bench connectivity.

The measured incision was deeper for the upper VT14 than for the lower VT7 of the Vuvu (0.5 m for VT7 and 1.4 m for VT14; Table 5.3); this translates into a higher inundation threshold for the upper VT14 ($24 \text{ m}^3 \text{ s}^{-1}$) compared to the lower VT7 ($10 \text{ m}^3 \text{ s}^{-1}$). A similar increase in channel capacity for VT15 was observed as the lower flood bench at VT1 (inundated 4.5 times per year) was inundated three times more frequently than that at VT15 (inundated 1.5 times per year) for the same monitoring period. This would suggest that the upper reaches of the Vuvu valley fill were incised to a larger extent than the lower reaches. For the higher flood benches, the observed flow was sufficient to inundate both VT1 and VT15 once during the 2 year monitoring period. It also is evident in the exponential shape of the rating curve that a small increase in channel capacity (e.g. bed lowering to increase the channel capacity by $15 \text{ m}^3 \text{ s}^{-1}$) will have a much greater effect on water level in the region of the lower flood benches (e.g. 1-1.5 m) compared to the higher benches (2.5-3 m) (Figure 7.11). This suggests that the inundation frequencies of lower flood benches are more sensitive to incision than the higher flood benches. At VT15 channel incision reduced the potential to store sediment on the valley fill, transforming the section of channel to a booster with increased sediment conveyance (Brierley and Murn, 1997; Brierley et al., 2006). Where the incision was less pronounced, such as VT7, the lower flood bench-channel connectivity was reduced to a smaller degree.

The longer term modelled data showed similar trends where the lower flood bench of VT1 was inundated 3.8-4.4 times a year and the lower flood bench at VT15 was inundated 2.5-3.7 times a year.

These intra year inundation frequencies of lower benches are often reflected in vegetation and geomorphological studies. Graminoids and pioneer plants often are found on lower depositional or wet zones that are associated with frequently disturbed sedimentary environments and regular inundation (Reinecke et al., 2007; Sieben and Reinecke, 2008). Erskine and Livingstone (1999) recorded the annual inundation of the lower benches in the Hunter River system in New South Wales, using the annual maximum flood series and were thus limited to one event a year. In the Sabie River in South Africa, lower bars and depositional features were inundated several times a year (Heritage et al., 2001). From the literature and observations it is clear that lower flood benches are inundated several times a year, making within-year sediment stripping and deposition on these features possible.

The modelled data for the higher flood benches showed less frequent inundation than the observed data. This was due to the moderate to wet cycle experienced during the monitoring period. Modelled inundation frequency data showed that higher flood benches were inundated every 1.2-7.3 years (2-4 hour flood base). Erskine and Livingstone (1999), Heritage et al. (2001) and Chalmers et al. (2012) measured similar inundation frequencies of >2 years for higher depositional features. As the higher bench was assumed to be related to bankfull stage, it would be expected to be inundated every 1-2 years (Wolman and Leopold, 1957; Wolman and Miller, 1960; Leigh, 2010). It can thus be deduced that the higher flood benches were effectively river terraces or well developed vertically accreted flood benches (Erskine and Livingstone,

1999) as their longer term flood frequency was less than that expected for bankfull stage. This finding could also be due to the overestimation of discharge from the hydraulic rating curve.

Australian and southern African rivers have a more complex morphological nature than temperate rivers (Erskine and Livingstone, 1999; Dollar, 2000; Heritage et al., 2001). Erskine and Livingstone (1999) ascribed this to the high flash flood magnitude index of rivers in Australia and southern Africa (McMahon et al., 1992) with various depositional features related to a range of flow levels. Erskine and Livingstone (1999) documented up to six separate bench levels for the Hunter River system in New South Wales where the highest and most extensive flood bench was inundated <1 in 24 years. This river had compound channels with a larger macro channel related to large flow events and a smaller active channel related to average flow events (Erskine and Livingstone, 1999). The most extensive bench in the Hunter River system was much higher than the smaller benches that were related to bankfull discharge and were inundated less frequently than floodplains for most rivers on a global scale. Bench 2, the second lowest bench out of four within the active channel, resembled bankfull discharge as it had an inundation frequency of 1 in 2 years (Erskine and Livingstone, 1999).

Similarly, Heritage et al. (2001) described several levels within the incised macro channel of the Sabie River, a lowland river in South Africa. They identified depositional features in the active channel that were frequently inundated (several events a year), active channel bars that were inundated every 1–1.5 years, lateral bars within the larger macro channel that were inundated every 2+ years and depositional features outside the macro channel bank that were inundated every 2–20 years. Dollar (2000) found similar results for the Mkomazi River, a steep South African river, where depositional features in the active and macro channel were related to a wide range of flow frequencies. Dollar found that low benches were the most consistent features that were related to return frequencies of 1 to 10 years, emphasising the large variability within the system. This shows the complexity of channels and depositional features that are formed by highly variable flows. This complexity also was evident to an extent in the Vuvu River where benches corresponded to a range of flows. What is not yet certain is whether the various flood bench levels that were observed are simply a result of channel incision or a result of the highly variable flow conditions experienced throughout southern Africa. It is likely that it is a combination of the two factors, but due to the limited number of observations made during this study, it will remain a debate.

When hillslope-channel connectivity was decreased by slope re-vegetation and gully rehabilitation (modelled), channel-flood bench connectivity was decreased, given the current channel geometry. This is due to the reduction in peak stormflow volumes associated with the increase in flood duration as a result of decreased hillslope-channel connectivity leading to reduced flood levels (Leopold and Maddock, 1953). Modelled data showed that lower benches could have up to one less inundation per year, whereas the higher flood bench inundation frequency can be decreased by up to 6 years. Slope rehabilitation could reduce the chances of sediment deposition on flood benches in the present day channel due to the reduced inundation frequency. Increased vegetation cover and reduced sediment delivery to the channel, due to

rehabilitation, should decrease the suspended load, further reducing the volume of sediment that could be deposited on the valley fill. The need for sediment storage would be decreased for a rehabilitated condition due to lower sediment loads.

The results also showed that inundation frequencies were influenced by dry and wet cycles. Modelled data showed that during dry years lower benches could be inundated once a year, four times for moderate years and seven times for wet years. Similar results were presented by Erskine (2013) where flooding frequency decreased during dry periods and increased during wet periods. Higher benches in the Vuvu were inundated <1 in 10 years during dry periods, 1 in 1 to 1 in 1.4 years during moderate years and once a year during wet periods, showing increased channel-flood bench connectivity during wetter years and decreased channel-flood bench connectivity during drier years.

When reductions in hillslope-channel connectivity, due to rehabilitation, were brought into account it can be seen that inundation frequencies were even further decreased. During dry periods lower flood benches would be inundated once a year, four times in moderate years and six times during wet years. For higher benches inundation frequency for dry years would be <1 in 10 years, for moderate years 1 in 1.4 to 1 in 1.7 years and 1 in 1 to 1 in 1.6 years for wet years. The change in hillslope-channel connectivity would decrease the lower flood bench inundation frequency by up to 9% and the higher flood bench inundation frequency up to 37%.

The present day channel geometry and flow regime showed that the current benches are connected to the channel at the expected frequencies, although incised sections reduce the channel-flood bench connectivity and the potential to store sediment. Modelled results showed that the decrease in hillslope-channel connectivity could decrease the channel-flood bench connectivity and chances to store sediment on these features. Reductions in hillslope-channel connectivity could further reduce the suspended sediment loads, reducing the volume of sediment available to be stored on flood benches.

As suspended load makes up a large part of the sediment load of the Vuvu River (Chapter 5 and 6), and the fact that the Vuvu is a high energy river, it would be unlikely that the incised sections of channel would be filled in by fine sediment to recent (50 years) pre-incision levels when hillslope-channel connectivity is reduced. The likelihood of transporting large material would also be decreased as flow magnitude is reduced and further incision would be unlikely. Lateral migration of the channel could rework the higher benches, where contributions from stored cobble layers could elevate the channel and new flood benches could be formed that would relate to the post-rehabilitation flow regime (reduced peak flow). Due to the incised and narrower area that is inundated due to incision, the chances of storing suspended sediment on the valley fill is reduced.

It was shown that wet and dry cycles influenced hydrological channel-flood bench connectivity, with more frequent inundation of flood benches during wet periods compared to dry periods. This would suggest that hydrological connectivity peaks during wet periods, whereas sediment connectivity peaks during dry periods when sediment accumulation rates are highest (Chapter 5). This indicates sediment

connectivity being greatest during infrequent high magnitude events during dry periods, whereas sediment connectivity decreases during wetter periods, despite the more frequent hydrological connectivity, as a result of increased vegetation cover and subsequent decrease in sediment availability. This also shows how complex the Vuvu system is in terms of hydrological and sedimentological connectivity, a characteristic that should be kept in mind when further research or monitoring is done in this catchment.

7.5. Conclusions

Flow and rainfall data were successfully monitored for the Vuvu catchment. The increased hillslope-channel connectivity acted as a booster that routed the rainfall efficiently to the channel to produce high-energy flashy or short-lived flows. The rainfall was mostly of high intensity, adding to the short-lived nature of the flows.

The modelled peak flow data provided interesting results that could be used to examine channel-flood bench connectivity for present day conditions, a future scenario with reduced hillslope-channel connectivity and wet and dry cycles. The observed and modelled data showed that the lower flood benches were inundated up to 4 times a year, whereas the higher flood benches were inundated up to once a year. The future scenario of reduced hillslope-channel connectivity showed a decrease in peak flood flows and the corresponding flood bench inundation frequency. Wet and dry cycles had a large influence on channel-flood bench connectivity, with increased inundation frequency during wet years and limited (1 event for lower flood benches) channel-flood bench connectivity during dry periods. Although hydrological channel-flood bench connectivity frequency peaks during wet periods, sediment contributions to the valley fill peak during dry periods and suggests that sediment connectivity is highest during these infrequent flood events. The effect of wet and dry cycles should be kept in mind for future research and monitoring of channel-flood bench connectivity as the results of short term observations can be confused by longer term natural variability.

It also was noted that channel incision reduced inundation frequency and potential for sediment deposition of the lower flood benches, but to a lesser extent reduced that of the higher flood benches as inundation thresholds of the higher flood benches were relatively insensitive to small changes in channel area during high flows.

Rehabilitation of hillslope-channel connectivity could reduce the potential of the Vuvu valley fill to store sediment until the channel has adjusted to the post-rehabilitation flow and sediment regime. This process could take a number of years (10–50+ years) to allow for the reworking of higher flood benches and terraces (by means of lateral channel migration) that would result in the formation of new benches that would act as sediment stores. Subsequently, sediment would be released (by the reworking process) from current stores to create space for the formation of new stores. An increased sediment contribution from

the valley sinks could be expected as the channel adjusts to the post rehabilitation flow regime and reduced sediment contribution from the slopes. However, increased sediment contributions from the valley fill could negate the reduced sediment input from the slopes due to rehabilitation.

Chapter 8: Synthesis

8.1. Introduction

A synthesis of the main findings with respect to the scientific objectives of the thesis is presented in this chapter. The most relevant results on changes to hillslope-channel and channel-valley fill connectivity from preceding chapters are integrated in this synthesis to form a broader perspective of how landscape connectivity has changed in the high rainfall mountainous headwaters of the northern Eastern Cape Province, South Africa. In each section the implication of the change in landscape connectivity on sediment dynamics is emphasized and the contribution of the work to the field of geomorphic analysis of river systems and sediment dynamics is highlighted. A conceptual model of altered landscape connectivity is presented, together with research opportunities and management recommendations.

8.2. Summary and general discussion

8.2.1. Connectivity as a concept and framework for the study of sediment dynamics

Fluvial geomorphology often is approached from either a purely physical (e.g. Leopold and Maddock, 1953; Wolman and Miller, 1960) or a purely structurally based approach (e.g. Schumm, 1977; Rosgen 1996) and focuses on a specific reach, ignoring how the reach fits into the larger catchment. The connectivity concept and framework integrates and unifies these approaches to allow geomorphologists to explore the structure and function of fluvial landscapes at a range of spatial and temporal scales (Fryirs, 2013). The connectivity framework is used to anchor the well-developed physics- and structure-based geomorphic approaches in order to study catchment-wide sediment processes. This holistic approach appeared appropriate for the often complex systems encountered in South African rivers (Dollar, 2000).

Connectivity is defined as the ease with which material is moved from one landscape unit to the next (Hooke, 2003). Brierley et al. (2006) presented a clustered hierarchical framework that can be used to identify pathways, buffers, barriers, blankets and boosters at a range of scales. This structural and functional view of the catchment and river system enables the researcher to draw conclusions about processes at a range of scales, from channel reach scale to catchment scale, with a holistic understanding of how the links between the landscape units influence the morphology of specific features over time.

In systems where landscape units are poorly connected in terms of pathways, sediment transfer is relatively slow (hundreds to thousands of years) between landscape units, with the different landscape units being nonsynchronous in their sediment delivery processes (Fryirs et al., 2007a). In well-connected systems sediment transfer is rapid (on an event basis) with changes in one part of the catchment often manifesting in other portions of the catchment (Harvey, 2001). The degree of connectivity also is

influenced by the magnitude of an event. For smaller events (annual frequency) a barrier or buffer can be effective in disconnecting a sediment pathway (depicted as a switch in the off position), but the pathway becomes activated (switch on or connected) during large magnitude events that occur less frequently (50 year return frequency) (Fryirs et al., 2007a).

For this study of sediment dynamics in the Vuvu catchment, the main changes in connectivity are addressed at two scales, 1) hillslope-channel connectivity at a catchment scale and 2) channel-valley fill connectivity at a channel reach scale. The following common themes are drawn from the literature. Firstly, at the catchment scale the focus is on linear pathways downslope and through the catchment (cf. Harvey, 1997, Brierley et al., 2006; Fryirs et al., 2007a). From the literature it is clear that steeper slopes (Kasai et al., 2005; Fryirs et al., 2007a), reduced vegetation cover (Gore, 1994; Cammeraat, 2002; Kirkby et al., 2002; Fryirs et al., 2007a), gully formation (Croke et al., 2005; López-Vicente et al., 2013) and narrow valley fills (Brierley and Fryirs, 1999; Fryirs et al., 2007a; Fryirs and Brierley, 1999; Michaelides et al., 2010) all contribute to hillslope-channel connectivity and make transport of material to the channel more likely. Secondly, at the channel reach scale the focus is on longitudinal, lateral and vertical connectivity between the channel and the adjacent valley fill (cf. Ward, 1989; Kondolf et al., 2006). Channel-valley fill connectivity is decreased by channel incision, making material transfer to buffers less probable if the channel dimensions are enlarged (Gergel et al., 2002; Kondolf, 2006; Pizzuto, 2011). In the literature no studies were found that encompass both changes in catchment scale connectivity and channel reach scale connectivity in depth. This research will add to the body of connectivity research by integrating these two scales.

Several shortcomings of the connectivity concept have been noted, such as: connectivity is unique in every catchment and assessments are thus resource intensive (Brierley et al., 2006); no universally quantifiable measurement unit has been developed for connectivity, making comparisons among different catchments challenging (Michaelides and Chappell, 2009; Grant, 2013 pers. comm.); and magnitude-frequency switches leak, thus not blocking sediment entirely, especially in systems dominated by suspended sediment (Fryirs et al., 2007a). Despite these apparent shortcomings, the connectivity concept proved to be a powerful tool to study catchment-wide sediment transfer processes and structures. Limited work has utilised the connectivity concept in South Africa, especially in high rainfall, high altitude catchments, thus the novel application of this holistic approach can be valuable to guide management of this precious high water-yield region and contribute to local and international understanding of connectivity in these catchments.

8.2.2. Hillslope-channel connectivity: present day sediment sources and pathways

Given the regional setting of the Vuvu catchment in the headwaters of a steep eastern seaboard river draining the Drakensberg Escarpment, it would be expected that drainage density and hillslope-channel connectivity of sediment is high due to the steep topography. The anthropogenic influence on this hillslope-channel connectivity was not well understood and formed the focus of this section. As the connectivity approach is not well developed for quantifying hillslope-channel connectivity, new methods were developed in this study to assess and compare hillslope-channel connectivity within and between various regions in the Vuvu catchment. Field visits, ground based photos and high resolution aerial images proved a useful combination of resources to map and classify potential sediment source areas in the Vuvu catchment. The large extent of gullies, areas of sheet erosion, landslides, roads and cattle tracks made it possible to digitize these features in the mountainous terrain using GIS. Using historical photos to calculate changes in the area of larger gullies over time proved useful in determining the formation date of these features. Drainage pathways also were assessed and allowed for the differentiation between features that are directly linked to the drainage network from those where sediment transfer is hindered by a vegetated buffer. As no universal measurement unit is available for hillslope-channel connectivity, the density per unit area that often is used for drainage density was adopted. This gave a result that could be compared to that of the existing drainage network and to that of other features that increase hillslope-channel connectivity.

Results from the mapping of source areas showed that more than 9% of the catchment has been exposed to severe erosion, and that 45% of the eroded area is directly connected to the channel, thus having a high sediment yield potential. The Elliot Formation had the highest density of eroded areas due to the erodible nature of the soils (Vetter, 2007; Fey et al., 2010) and the ongoing land use pressures in the lower catchment. Moderate slopes ($5-20^{\circ}$) that were north-facing were most affected by erosion throughout the catchment. Although the Elliot formation was shown to be the most erodible and most active sediment source, the overall greater area of the Drakensberg Formation (65% of catchment area) disproportionately increased contributions from this area.

Fields covered about 5% of the catchment and were estimated to contribute on average $3.5 \text{ t ha}^{-1} \text{ y}^{-1}$. Gullies only covered 1% of the catchment area, but produced the highest sediment loss of ca. $3.9 \text{ t ha}^{-1} \text{ y}^{-1}$, on average, due to their greater depth. Values were similar to that estimated by Beckedahl and De Villiers (2000) for gully features in other areas of the Transkei ($5.6-14.2 \text{ t ha}^{-1} \text{ y}^{-1}$). Areas of sheet erosion covered ~2% of the catchment area and produced on average ca. $3.6 \text{ t ha}^{-1} \text{ y}^{-1}$ of sediment. The landslides, gravel roads and livestock tracks covered less than 1.2% of the catchment and contributed less than ca. $1.8 \text{ t ha}^{-1} \text{ y}^{-1}$ collectively, making them less significant sediment sources. When averaged modelled erosion data by Le Roux et al. (2008) are considered, catchment-wide diffuse surface erosion is estimated to be the main sediment source ($17.1 \text{ t ha}^{-1} \text{ y}^{-1}$) for the catchment.

The Clarens Formation had the lowest sediment loss per area and the Elliot Formation consistently the highest. Fields on the Elliot Formation had the greatest soil loss ($9.5 \text{ t ha}^{-1} \text{ y}^{-1}$), which is almost double that of gully and sheet erosion on the Elliot Formation respectively (5.4 and $5.3 \text{ t ha}^{-1} \text{ y}^{-1}$). The contribution from roads on the Elliot Formation ($2.1 \text{ t ha}^{-1} \text{ y}^{-1}$) was surprisingly high when calculated on a geological province basis. These soil loss data per geological province emphasise the erodible nature of the Elliot Formation as the Elliot formation and Drakensberg Formation have lost similar volumes of soil over the last 50 years ($1\ 000\ 000 \text{ m}^3$ respectively) despite the Elliot Formation being less than half the area of the Drakensberg Formation.

Historical mapping shows that the larger gullies were initiated in the 1920s, whereas the smaller gullies were initiated in the 1950s, similar to that found by Huber (2013) in a neighbouring catchment. These features are mostly human induced as ploughing and grazing were intensified during the early 1900s, leading to increased runoff energy and soil vulnerability. From field assessments it was clear that the majority of the erosion features were still active (63%). Gullies proved to still be active (61-83%), but sheet erosion features were mostly (79-86%) stable with vegetation colonising localised areas where sediment is deposited. Gullies that were connected to the natural drainage network tended to be more active than those that are buffered. Erosive features on the Elliot Formation were the most active, mainly due to its erodible nature and continued pressure on the land. Results also showed that gullies remain sediment sources for periods up to 100+ years, where after they stabilize.

The Vuvu catchment had a well-developed drainage network as expected for a high relief headwater catchment. Limited natural buffering exists due to the steep nature of the slopes, except for the gentle slopes in valley bottoms and occasionally between the steeper hillslopes (stepped landscape). Hillslope-channel connectivity was significantly increased throughout the Vuvu catchment. Down-slope erosional features, such as gullies, increased the drainage density by 21%, whereas across-slope features (e.g. livestock tracks and roads) increased the drainage density by 178%. The down-slope features would be more efficient in routing water to the channel, but across-slope features play a significant role in intercepting hillslope runoff, concentrating flow and routing it to the drainage network, decreasing the chances for water to infiltrate and sediment deposition along the slope.

Gullies were formed in colluvium on lower slope angles at the base of steeper slopes due to increased runoff contributions from upslope as a result of reduced vegetation cover, hardened surfaces and increased pathways (Cammeraat, 2002). The gentle slopes possibly acted as buffers between the steeper slopes and the river system, storing sediment, but gully development has formed a link between the steeper slopes and the channel, thus effectively overriding the potential for the gentler slopes to act as buffer areas. Continuous gullies were the most important hillslope-channel linkages in terms of sediment delivery as water and sediment are directly discharged into the channel. Buffers below discontinuous gullies reduce the velocity of flows and deposit sediment during low magnitude events, however these buffers will be breached during high magnitude events (Croke et al., 2005; Fryirs et al., 2007a).

The Elliot Formation was the geological province most affected by increases in hillslope-channel connectivity, with increases occurring on gentle to steep slopes. The Drakensberg and Clarens Formations were affected to a lesser degree, with increases in hillslope-channel connectivity mostly occurring on moderate and steep slopes (slopes steeper than 5°). Hillslope-channel connectivity increases were greatest on north-facing slopes owing to the resilient nature of the moister south-facing slopes that support a greater vegetation cover (Holland and Steyn, 1975; Granger and Schulze, 1977) and the increased pressure on north-facing slopes due to settlement preference and more palatable grasses (Ellery et al., 1995; Mucina et al., 2006).

Landscape connectivity within the Vuvu catchment is naturally high due to limited accommodation space in the generally steep landscape, but due to anthropogenic influence and accelerated erosion this connectivity has been increased over the past 100 years. This means that sediment that is currently mobilized on the slopes is more likely to be transported to the main river, even during the lower intensity rainfall events. The Elliot Formation was the most impacted in terms of erosion and hillslope-channel connectivity; this is a likely function of the erodible nature of its mudstone derived soils and ongoing anthropogenic land use pressures.

8.2.3. Longitudinal connectivity: the valley fill character and history

Sediment that is stored along the valley fill affects the longitudinal connectivity of the Vuvu catchment. During high flows sediment can either be stored in or eroded from flood benches through lateral connectivity, altering the transfer of sediment down the system. The Vuvu valley fill is narrow (up to 100 m wide) with a wandering channel pattern (channel up to 30 m wide) that switches from side to side along the valley fill similar to the ‘foothill river’ classification by Rowntree and Wadeson (1999). The channel consists of a cobble pool-riffle sequence with bedrock outcropping in places and can be classified as a mixed bed river (Rowntree, 2013a). Cobble makes up the majority of the bed, with limited fine sediment stored along the channel bed. Suspended sediment is mostly stored in flood benches, terraces and alluvial fans along the valley fill. The flood benches are narrow (up to 50 m wide) and not continuous, often dissected by poorly defined flood channels. The flood benches and terraces have a cobble base with a sequence of horizontal sand and silt layers up to 1.2 m thick, typical of a river with a mixed load transporting both bed load and suspended load (Schumm, 1977). The cobble base of the terraces is an indication of a previous fill level that formed before 4 500 BP, with cobble contributions most likely from the upper basalt parts of the catchment as the mudstones of the lower catchment breaks down to fine grained sediment fairly rapidly. The valley fill has limited accommodation space due to the steep topography as the river is mainly cutting into the escarpment instead of eroding laterally.

Higher and lower flood benches were both active in storing sediment as fresh sand and debris deposits were identified at both levels. Particle size varied down the cores that were taken on the flood benches,

with median particle size ranging from coarse silt to medium sand, where coarser layers indicate higher energy events (Tooth et al., 2013) and overall variability of the flow regime. The general upwards fining towards the top of the higher benches shows that energy to transport larger particles is waning as the bench builds up vertically. Terraces had a similar sediment composition to that of the benches with all bench and terrace sediment being poorly sorted. This suggests that sediment and energy regimes have not changed significantly over the past 4 500 years.

Pb-210 dating using an emanation coefficient of 0.3 (Du and Walling, 2012) proved effective, especially when Cs-137 detection and channel position from historical aerial images were used to calibrate the CRS model. Results showed that the lower benches stored sediment for 40-60 years, whereas the higher benches have stored sediment for ~92+ years. The average sediment accumulation rate was ca. $0.9 \text{ g cm}^{-2} \text{ yr}^{-1}$ for the higher flood benches compared to ca. $3.2 \text{ g cm}^{-2} \text{ yr}^{-1}$ for the lower flood benches, showing that the lower benches were more active in receiving, storing and possibly contributing sediment (Erskine and Livingstone, 1999). The increase in sediment accumulation rate over the last ~50 years would suggest that sediment availability has increased, probably due to accelerated erosion and efficient sediment delivery as a result of increased hillslope-channel connectivity. The sections with greater sediment accumulation rates were associated with dry periods when vegetation cover is reduced and soils are exposed to erosive rain, resulting in greater sediment availability (López-Bermúdez et al., 1998; van der Waal et al., 2012).

Calculations indicated that the flood benches stored ~2% of the sediment that has been eroded over the past 50 years, making the valley fill an insignificant sediment store, as would be expected for a steep headwater system with limited accommodation space (Rowntree and Wadeson, 1999; Fryirs and Brierley, 2013).

Furthermore, river incision, channel straightening and channel widening over the last 50 years reduced channel-flood bench connectivity. Erskine (1986) found that river straightening would steepen the river slope and the transport efficiency of the channel, reducing chances of storing sediment on the valley fill. Incision in the Vuvu system was likely a response to increased hillslope-channel connectivity resulting in increased peak discharge (Knighton, 1984; Schumm, 2005), whereas river straightening and widening was the response to an increased coarse sediment load (Knighton, 1984; Erskine, 1986). From the information above one can deduce that channel incision, straightening and widening reduce the buffering capacity of the valley fill.

Terrace dating revealed cut and fill cycles in the Vuvu valley with incision being dominant over the last ~3 000 years. Warmer and wetter paraglacial conditions during the early Holocene (5 000–10 000 BP) could have redistributed sediment that was produced during the Last Glacial Maximum (Temme et al., 2008) and led to the build-up of sediment in the Vuvu valley fill (up to 4+ m). As vegetation cover increased during the middle to late Holocene, sediment supply was reduced and led to mass valley sediment evacuation (Temme et al., 2008). The findings of the Vuvu valley cut and fill cycles [infill and

subsequent degradation during the late Holocene (~3 000 BP)] agree with other studies from South Africa such as the Modder River, Free State (Tooth et al., 2013) and Okhombe valley, KwaZulu-Natal (Temme et al., 2008).

The Vuvu valley fill acts as a buffer for both hillslope-channel and longitudinal connectivity. Terraces and higher flood benches disconnect the slopes and smaller tributaries in the direct vicinity of the valley fill from the channel, whereas flood benches store a small proportion of sediment that is transported downstream. The Vuvu River is in a phase of degradation, reworking old deposits on an ongoing basis, mainly through lateral migration, straightening and incision of the channel.

8.2.4. Sediment dynamics over time

The discrimination between igneous (Drakensberg Formation) and sedimentary (Clarens and Elliot Formations) sources proved to be effective using mineral magnetic properties (Oldfield et al., 1979; Foster et al., 1998, 2007). A quantitative and qualitative approach was tested. The quantitative sediment apportionment approach (using mineral magnetic properties in a Discriminant Function Analysis) could differentiate between source groups, based on Xlf, ARM and HIRM values, with relatively small uncertainties (<10%). Apportionment results had relatively low error margins (3-27%). For the qualitative approach Xlf was selected as the best discriminator between igneous and sedimentary samples. A comparison between the two approaches showed the same trends, with sedimentary source sediment being dominant (~64%) in the core (VT2). An equation was developed that could calculate the proportion of igneous material in a sediment sample based on an Xlf value. Increases in particle size reduced the certainty levels of the tracing. Surface-subsurface tracing, using Xfd, proved possible where mudstones dominated the sediment sample. Using an additional tracer, Cs-137, which was independent of the magnetic properties, extended the surface-subsurface tracing to more recent samples (post 1958) of sediment from igneous sources.

This qualitative approach could be adopted in catchments with a distinct magnetic difference between the various geologies as it can produce tracing results relatively quickly, not requiring elaborate source ascription modelling. Managers and citizen scientists can also use the qualitative approach to determine the general source of sediment as it requires relatively simple equipment and is more easily understood.

Suspended sediment concentrations were very high during summer peak flows (up to 11 g l⁻¹ consisting of clay; D50 of 2 µm) and were comparable to concentrations for the larger Umzimvubu River system (Madikizela et al., 2001) of which the Vuvu River forms a part. The difference in particle size between the flood water sampled and bench material (D50 of 2 µm and 110 µm respectively) indicates that the observed floods were relatively small as they did not carry the larger particles in suspension. This was confirmed by the water level readings for the suspended sediment monitoring period that showed that the

sampled floods did not inundate the higher benches. It is thus expected that the particle size will increase during larger flood events.

Results from the suspended sediment tracing showed that sediment source changed throughout the flood hydrographs and was dependent on rainfall distribution, event intensity and catchment shape. As channel-flood bench connectivity is established during peak flows, clockwise and anticlockwise sediment hysteresis played an important role in the source signal that was stored in the flood bench. Clockwise hysteresis is associated with sediment sourced from the lower parts of the catchment whereas anticlockwise hysteresis is associated with sediment mainly sourced from the distal parts of the catchment. In the Vuvu catchment, for flood events with the same magnitude but opposing hysteresis directions, the event experiencing clockwise hysteresis could leave a stronger sedimentary signature compared to the event with anticlockwise hysteresis that could deposit more igneous sediment on flood benches. These flood hydrographs and linked sediment source tracing results highlighted how variable the Vuvu system is and that the dominance of a source can change over a short time (e.g. within a flood event). This variability has implications for sediment tracing as flows (responsible for overbank links), sediment concentrations and sources differ between and within events. Sediment tracing results should thus be interpreted with care and an understanding that a layer of sediment that is deposited on a flood bench is a mere 'snapshot' or portion of the sediment transported during an event.

A general decrease in sediment concentration was evident towards the end of the wet season as sediment supply is exhausted (Asselman, 1999; Rovira and Batalla, 2006; Grenfell and Ellery, 2009; Fryirs and Brierley, 2013). Sediment concentrations decreased during the wet season as sediment available for transport is depleted and vegetation provides cover from erosive rainfall (Grenfell and Ellery, 2009). These data on within event and seasonal sediment dynamics suggest that the majority of the sediment movement occurred during short pulses at the beginning of the wet season during catchment-wide high-intensity rainfall. The Vuvu River is thus a supply limited system due to the frequent high energy events and good landscape connectivity. This has implications for downstream water users as fine-grained sediment made available in the Vuvu catchment is likely to be exported to the downstream users.

Higher and lower flood benches stored mostly locally sourced sedimentary sediment, with the higher benches storing finer material with a greater igneous contribution (up to 59% igneous material). This is due to the difference in elevation between the lower and higher flood benches that require different sized flows for inundation (Wolman and Leopold, 1957). Sediment is deposited on the higher benches when the whole catchment is contributing water and sediment (high magnitude), whereas sediment is deposited on the lower benches during more locally derived low magnitude events. Low magnitude events in the upper catchment do not contribute significant amounts of sediment to the lower benches, pointing to both the lower erosion potential of the upper catchment and sediment storage on slopes and along the higher order drainage channels. The sediment from the upper catchment is transported to the lower catchment

during high magnitude events, similar to the switch concept by Fryirs et al. (2007a) where switches are on (letting sediment through) during large magnitude events.

Results show that the lower catchment has lost the majority of its surface soil as is reflected by the dominance of subsoil in the lower benches whereas higher benches contained more surface soil from the upper catchment. This confirms the mapping results that pointed to the mudstones of the Elliot Formation as the geological province with the greatest soil loss and highest hillslope-channel connectivity.

Terraces were also dominated by sedimentary subsoil sources. The erodibility of the mudstones, due to the high sodium content and poor structure (Fey et al., 2010), was confirmed by its dominance over the past 4 500 years, suggesting that the recent increase in hillslope-channel connectivity has not changed the proportions of sediment stored in the Vuvu valley fill. It confirms the mudstones as an important suspended sediment source, regardless of land use and degradation, that experienced greater hillslope-channel connectivity over the past 4 500 years relative to the basalt and sandstone areas. This could mean that suspended sediment yield from mudstone areas can remain high despite rehabilitation efforts.

8.2.5. Channel-valley fill connectivity

Intensive summer rainfall in the Vuvu catchment produced steep hydrographs of short duration (3 hours). This is expected for steep, well connected catchments that receive intensive rainfall (Costa, 1987; Taguas et al., 2008). Short term (2 years) discharge was measured for the Vuvu catchment which was extended by disaggregating monthly flows to daily and subsequently peak instantaneous flows. This allowed for the assessment of channel-flood bench inundation frequency for current conditions and a potential future rehabilitated condition (re-vegetation of bare areas and gullies) where hillslope-channel connectivity is reduced.

Results showed that the peak discharge will be reduced in the future as hillslope-channel connectivity is decreased by rehabilitation efforts, similar to results by Jones and Grant (1996) and Schulze and Horan (2007).

Measured and modelled hydraulic results showed that higher and lower flood benches vary significantly in inundation frequency. River level monitoring at two transects showed that the lower benches were inundated 1.5–4.5 times a year and the higher benches 1 in 2 years. The monitoring was done during a moderate to wet period, thus inundation frequencies were higher than expected. Modelled discharge data for 15 transects over a 73 year period produced inundation frequencies of 2.5–4.4 events a year for lower benches and 1 in 1.2 to 1 in 7.3 years for higher benches. The large range of frequencies is due to the differences in inundation threshold (bench level) between the various transects (n=15). Some of the features were elevated due to channel incision or near vertical accrual of sediment, thus increasing their inundation threshold and decreasing their inundation frequency.

The modelled inundation of higher flood benches was less frequent than the 1-2 year frequency expected for bankful discharge (Wolman and Leopold, 1957; Wolman and Miller, 1960; Leigh, 2010). This suggests that the higher flood benches were effectively terraces (Erskine and Livingstone, 1999) or that the hydraulic modelling overestimated discharge requirements for inundation or that the channel has adjusted to the highly variable flows that are characteristic of southern African and Australian rivers (McMahon et al., 1992). Similar results of low inundation frequencies for higher benches were recorded in the literature for South Africa by Dollar (2000) and Heritage et al. (2001), and for Australia by Erskine and Livingstone (1999) who related such low inundation frequencies to the highly variable flows that result in a more complex channel composition where an active channel is situated in a much larger macro channel. A very large flood would remove the lower benches, whereafter the benches are built up again by smaller events, resulting in localised cut and fill cycles (Gupta and Fox, 1974). The effects of a large flood are expected to vary throughout the system, leading to benches at different stages of development, as seen along the Vuvu valley fill. As the Vuvu River has experienced incision over the last 50 years, it is challenging to tease out whether the different flood bench levels are simply due to the incision, or related to the highly variable flows experienced throughout southern Africa. It could also be a combination of incision and variable flows, but identifying the main driver of channel complexity remains debatable with only the current limited observations and data sets.

Wet and dry cycles also play an important role in inundation frequency. In the Vuvu, inundation frequency is increased during wet years (lower benches inundated 7 times a year and higher benches inundated on an annual basis), and decreased in dry years (lower benches inundated once a year and higher benches inundated <1 in 10 years). It was expected that frequent inundation would result in greater sediment accumulation rates. This was contradicted by observations in the sediment history of the Vuvu catchment where dry cycles had greater sediment accumulation rates than wet cycles. This would suggest greater sediment availability during dry periods (reduced vegetation cover). Although average channel-flood bench connectivity was reduced, the greater sediment availability during these dry periods made large contributions to the flood benches during infrequent large events that do inundate the flood benches. This would suggest that hydrological connectivity peaks during wet periods, but that sediment connectivity peaks during infrequent large events over dry periods. This implies that there is a need to separate hydrological and sediment connectivity.

Modelled results for a potential rehabilitated state with reduced hillslope-channel connectivity showed reduced channel-flood bench inundation frequencies as peak flow volumes are reduced. This would reduce the potential to store sediment on the flood benches as sediment deposition is influenced by both sediment concentration and channel-flood bench connectivity. It is likely that the suspended sediment concentration will be reduced due to increased buffering in the catchment as a result of improved vegetation cover and reduced hillslope-channel connectivity. Sediment availability for deposition during the short inundation time of benches will thus be limited. Furthermore, reduced channel-flood bench connectivity will be the case until the channel reworks the flood benches, through lateral migration or a

large flood, and builds up new benches that are formed and maintained by the post rehabilitation flow regime. These new benches should be inundated more frequently as they are a product of the post rehabilitation flow regime. A release of valley fill sediment is likely as the channel adjusts to the post rehabilitation flow regime where less sediment is supplied from the hillslopes to the channel as hillslope-channel connectivity is reduced. Only once the hillslope-channel and channel-valley fill connectivity have stabilized due to rehabilitation efforts will the sediment yield be significantly reduced. This reduction in sediment yield can possibly take 10-100+ years.

As coarse sediment was not considered in this study a brief overview of how coarse sediment is likely affected by changes in landscape connectivity is given. Coarse sediment transport should be augmented by increases in hillslope-channel connectivity. Increased runoff, concentrated flow and direct pathways to the river network will contribute to more efficient transport of coarse material. Whether more coarse material is made available due to the anthropogenic influences is largely unknown, but it is assumed that the majority of the coarse material originates from the upper catchment's more resistant rocks, thus human influences could be negligible. It is more likely that most of the coarse material was made available during the last Glacial Maximum (10 000+ BP), whereafter the increase in rainfall moved the coarse material to the valley fill where a large proportion was deposited. As vegetation stabilised the slopes and the coarse sediment availability declined, the coarse sediment stored in the valley fill started moving downstream. This could be around 5 000 BP as is suggested by the dating of a few terraces in the Vuvu valley. Since then conditions were more stable with smaller fluctuations in coarse sediment availability and transport. The incision of the valley fill, coupled with increased runoff, will possibly increase the efficiency of coarse sediment transfer along the valley fill. As rehabilitation efforts should reduce both the runoff and transfer of coarse sediment to the river, coarse sediment transport and export will be reduced. Locally sourced coarse sediment from scoured benches and terraces will not be transported as readily and could help form more stable structures such as cobble bars that can initiate the formation of fine grained sediment stores. Due to the confined nature of the valley fill, both fine grained and coarse sediment storage will remain low.

8.2.6. Conceptual model of sediment connectivity

A conceptual model of changes in connectivity in the Vuvu catchment is presented in Figure 8.1. The Vuvu catchment is characterised by a high altitude, high rainfall basin. The overall catchment is in a phase of degradation or incision as headward erosion slowly cuts into the Drakensberg Escarpment. Minimal lateral erosion is taking place as can be seen by the generally steep topography. Limited storage space exists due to the steep topography and the well-developed natural drainage network contribute to the good hillslope-channel connectivity. The gentler colluvial slopes of this stepped landscape lends itself to small buffers between the steeper slopes. A narrow valley fill in the lower catchment stores both coarse and fine grained sediment. The coarse grained sediment is likely derived from the harder basalt rock in the

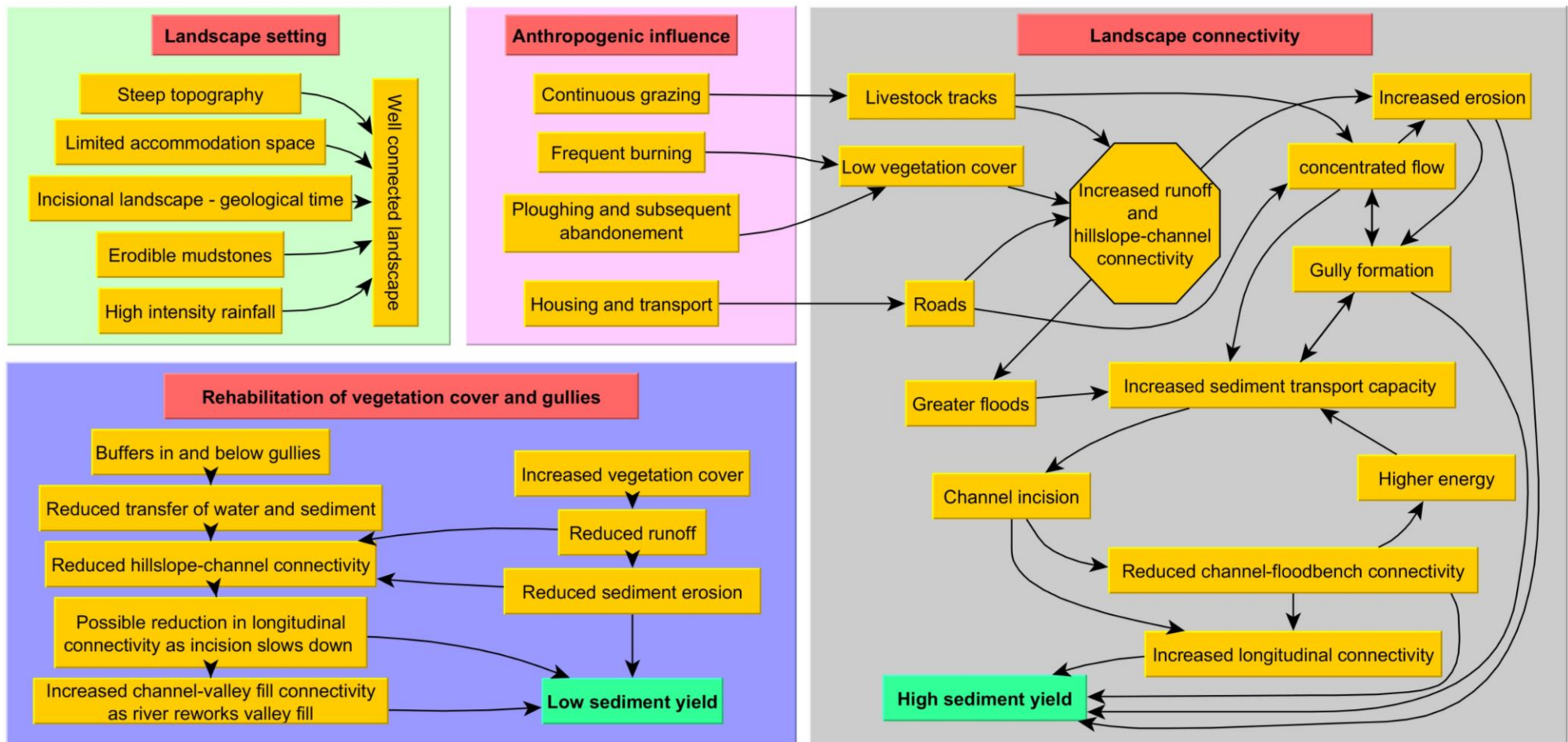


Figure 8.1: A conceptual diagram of how changes in landscape connectivity have altered sediment dynamics in the Vuvu catchment.

upper catchment whereas the fine grained sediment is likely to be a mixture from basalt, sandstone and mudstone derived soils. Catchment-wide high intensity rainfall coupled with erodible mudstones in the lower part of the catchment contribute to high suspended sediment concentrations in the Vuvu River. This high energy system is inherently prone to efficient sediment transfer, but human induced pressures have further increased the sediment transfer potential. Cut and fill cycles are evident on the hillslopes and valley fill as sediment is redistributed and stored in new locations, such as alluvial fans, if not exported from the catchment.

Livestock grazing and trampling, frequent burning, the development and the subsequent abandonment of agricultural fields and the introduction of housing areas and associated road networks have led to the disturbance and reduction in vegetation cover and successive increases in runoff and erosion in the catchment. Gullies formed where slope angles relax and thicker colluvial soil deposits are found. Livestock tracks formed where animals frequently traverse the catchment. Roads link the various homesteads to the main road that traverses the catchment and links the various scattered villages. Gullies, livestock tracks and roads increased the down-slope and across-slope routing of water, limiting water infiltration, and concentrating flows. Increased runoff volumes have resulted in increased energy available to erode soils and transport sediment. Furthermore, the generally fine nature of the eroded sediment and the high energy system made transport in suspension possible, thus limiting both sediment deposition and the effectiveness of existing buffer features on the hillslopes and the valley fill. The increase in hillslope-channel connectivity would increase both fine and coarse grained sediment transport.

The valley fill has limited storage space due to its confined nature. Narrow flood benches were shown to be stores of mostly fine grained sediment with coarse grained sediment or bedrock forming the base of the flood benches. Increases in runoff and hillslope-channel connectivity resulted in larger flood magnitudes and increased sediment concentrations. This increased flow energy enhanced the sediment entrainment and transport capacity and resulted in channel incision, straightening and widening. This reduced the channel-valley fill connectivity, but due to increased sediment concentrations, sediment deposition on flood benches has increased during the few events that do inundate the flood benches. The overall result is increased down-valley or longitudinal connectivity due to the small storage potential of the narrow flood benches. This reduction in channel-flood bench connectivity and greater flow energy regimes reduced the potential to store sediment in the valley fill, enhancing an already high sediment yield from the Vuvu catchment. This shows that the degraded nature of the catchment reduced the buffering capacity of the landscape, enhancing sediment production and sediment transfer, leading to high sediment yields.

Wet and dry cycles played a role in sediment availability, transfer and deposition. During dry cycles vegetation cover is reduced and soils are exposed to erosive rainfall. Large rainfall events are fewer during dry periods, but when intensive events happen, great quantities of sediment is eroded and transported.

This possibly results in high sediment concentrations during the infrequent events that deposit large quantities of sediment on flood benches that are inundated. Wet periods are different in that vegetation cover is greater and soils better protected, leading to low sediment concentrations during the more frequent high flow events and results in less sediment deposition despite the greater channel-valley fill hydrological connectivity. This indicates that sediment connectivity peaks during dry cycles, whereas hydrological connectivity peaks during wet cycles.

Rehabilitation of the catchment's vegetation cover and gullies will lead to reduced hillslope-channel connectivity. This will limit rain drop related soil erosion, the volume and energy available of runoff to erode and transport soils. Buffers around and in gullies will further facilitate sediment deposition and water infiltration. These reductions in hillslope-channel connectivity will reduce flood energy. Incision will slow down with reductions in net sediment export from the valley fill. The river could continue to rework stored valley sediment until a new stable phase is developed as flood benches are rebuilt to the new channel level and supports channel-valley fill connectivity that facilitates greater sediment storage. It is envisaged that rehabilitation will immediately reduce sediment contributions from hillslopes, but that valley fill degradation could continue for another 10-100+ years while a new sediment storage phase is developed on the valley fill. The effects of the catchment-wide rehabilitation is mere speculation and warrants further research.

The Vuvu catchment will remain a non-equilibrium system where sediment export dominates over sediment deposition. This is mainly due to its landscape setting and geological makeup. Localised cut and fill phases were always present in the landscape within larger more general cycles of increased cutting or filling, mainly due to climatic variations. The recent anthropogenic influence has initiated an overall accelerated cut phase on the hillslopes and valley fill, but with the appropriate rehabilitation strategies an catchment-wide fill phase can be initiated with net reductions in sediment yield.

8.2.7. Sources of measurement error

The wide range of methods used and general lack of detailed data for South Africa made the introduction of error during the research highly likely. A brief overview of possible inaccuracies and approximate estimates of error is given.

Maps and aerial photos are replications of reality and are influenced by scale, geometric transformations and projections. Possible inaccuracies due to photo and map representations and digitizing could be as high as 10%. Sediment volume calculations were based on mapping and field measurements that had standard deviations of ~50% due to the variability of depth and width measurements of the features in the field. Sediment erosion estimates could thus be up to 60% in error.

Rainfall measurement is only representative of the area measured by the rain gauge, thus upscaling a single point measurement to catchment scale will introduce error. Scaling of rainfall from the Matatiele

area will add further error. The accumulated error associated with the long term rainfall data could be as high as 50%. The hydrological modelling and disaggregation process could introduce a further 50%, making the total error associated with the modelled flow data 100%. Measuring discharge, stage and developing stage-discharge relationships could introduce additional error of 20-50%. Due to the large amount of potential error, a range of flood base duration (i.e. 2-5 and 4-6 hours) was used to account for the likely uncertainty introduced by both data and modelling.

Sediment tracing and determining the chronology introduced significant error. The measurement error of laboratory equipment was in the range of 2-10%. Using the data in models introduced further error of 10-30%. Dating and tracing related error could thus be as high as 40%, which excludes the representativeness of the actual field sampling.

The study thus experienced high levels of error associated with data and modelling. Error related to hillslope-channel connectivity is ~50%, sediment tracing and dating error is ~50 and channel-valley fill connectivity modelling error is 50-100%.

8.3. Conclusion

The following is a summary of key findings of the thesis:

- Due to the steep topography and limited accommodation space within the Vuvu catchment, it is expected that erosion will always dominate deposition, sediment yield should generally be high. Localised cut and fill cycles, climate and changes to landscape connectivity could influence the sediment yield and is expected to be highly variable over time.
- From the mapping of erosion features, gullies, sheet erosion and agricultural fields were the main sediment sources. The total sediment contribution from mapped erosion sources was ca. $13 \text{ t ha}^{-1} \text{ y}^{-1}$. A comparison of these erosion estimations with that modelled by Le Roux et al. (2008), based on the RUSLE, suggests that diffuse surface erosion could contribute a similar volume to that produced by gullies, sheet erosion, fields, cattle tracks and roads combined. The seemingly insignificant catchment-wide surface erosion stands out as the key sediment source, but is challenging to verify. Whether or not the sediment reaches the channel depends on the degree of connectivity between the hillslope and the channel.
- The Vuvu catchment has been exposed to erosion throughout the catchment, with the lower part of the catchment underlain by erodible mudstones being the most affected, partly due to greater land use pressures. Moderate slopes ($5\text{--}20^\circ$) that were north-facing were most affected by erosion throughout the catchment.
- Gullies were initiated from 1910 to 1970 and coincide with increased grazing and agricultural pressure.

- Gullies increased the down-slope drainage network by 21%, whereas the livestock tracks and roads increased the drainage density by 178%, mostly in an across slope direction. This increase in hillslope-channel pathways escalates the energy available for further erosion, perpetuating erosion and sediment transfer.
- High sediment concentrations during short stormflow events highlight the lack of buffers in the landscape due to increased hillslope-channel connectivity.
- The narrow depositional features along the confined Vuvu valley fill limit significant sediment storage (stores ca. 2% of eroded sediment), transferring the high sediment concentrations produced on the slopes.
- The Vuvu channel has incised and the channel has straightened since 1956 and coincides with the increase in hillslope-channel connectivity. This would suggest that channel incision and straightening is a result of the increased runoff.
- A semi-qualitative approach to sediment tracing using a single mineral magnetic property (Xlf) and an equation proved effective in discriminating between igneous and sedimentary sources.
- Sediment stored in the valley fill is dominated by locally sourced sedimentary sediments, with minimal source variation over the past 4 500 years. Localised sedimentary sources produced constantly high sediment yields. This was observed in flood events that were dominated by sedimentary sourced sediment. This could be a result of the generally better veld condition of the upper basalt catchment, the less erodible nature of the basalt derived soils, or that sediment is deposited en route or a combination of these factors.
- Lower level sediment sinks are inundated more frequently (up to 4 times a year) and receive more sediment ($3 \text{ g cm}^{-2} \text{ yr}^{-1}$) compared to higher level sediment sinks (inundated once a year to 1 in 7 years; sediment accumulation rate of $1 \text{ g cm}^{-2} \text{ yr}^{-1}$), but store sediment for shorter periods of time (~60 years compared to ~92+ years for higher flood benches).
- The sediment stored in valley sink features of a steep, well-drained system does not necessarily represent the dominant sediment that was transported during the various events. Channel-flood bench hydrological connectivity plays an important role in the type of sediment that is deposited as sediment sources can change throughout a flood event and overbank events are short lived. Thus, only sediment from a short phase of the flood hydrograph could be deposited on the flood bench, not reflecting the sediment that was transported throughout the entire flood event. This shows that the interpretation of a sediment record is not only a linear cause-and-effect exercise, but that flood effectiveness and climate variability play an important role in what sediment is stored in sinks.
- Hydrological channel-valley fill connectivity is reduced during dry periods and increased during wet periods.

- Channel incision has a negative effect on the potential for low elevation sediment sinks to store sediment until lateral migration reworks the stored sediment and builds new levels that correspond to the channel elevation and flow regime.
- Reductions in hillslope-channel connectivity as a result of slope rehabilitation and subsequent declines in peak discharge will limit the potential for sediment to be deposited on the valley fill features until the channel adjusts through lateral migration to the post rehabilitation flow regime.

8.4. Contribution and directions for future research

This study contributes to the understanding of landscape and sediment connectivity at a catchment and a channel reach scale along the upper reaches of the Thina River. The connectivity concept was embraced and developed to assess the sediment dynamics in the Vuvu catchment at these two scales. Using the connectivity framework contributed to the understanding of why, where and how the various sources and pathways contribute to sediment delivery. Brierley et al.'s (2006) framework of likely pathways and processes that sediment might encounter at various scales proved useful in this study of sediment connectivity as this framework gave the study a scale related or nested hierarchical structure. In this Vuvu study a similar approach was followed, but the understanding of catchment-wide connectivity was used to interpret within landscape scale or channel reach changes in connectivity along the valley fill at the bottom of the catchment. This study also added a) methods to quantify hillslope-channel connectivity and b) an understanding of how changes to catchment-wide hillslope-channel connectivity affect channel-valley fill connectivity at a channel reach scale.

An original approach to quantify increases in drainage density at the catchment scale made it possible to compare the influences of geology, slope and aspect on hillslope-channel connectivity. This method to quantify increases in drainage density revealed the influence that each drainage pathway type had on overall hillslope-channel connectivity. At the channel reach scale the characterisation of the valley fill defined the current sediment stores that are present. Determining the chronology and source of the sediment in storage revealed sediment dynamics during the past 4 500 years, with more detail over the past 100 years. The study contributed to the understanding of current sediment loads and sources, as well as sediment dynamics during flood events and during wet and dry cycles. Another contribution to the study of landscape connectivity at a reach scale was the assessment of channel-valley fill connectivity using hydraulic models to simulate the current and future post-rehabilitation potential to store sediment along the valley fill through channel-flood bench connectivity. The approach also explored the influence of wet and dry cycles on sediment storage potential along the valley fill through changes in channel-valley fill connectivity.

Future research should address the increase in hillslope-channel connectivity in terms of the most effective approaches and locations to buffer sediment. Sediment monitoring and sediment tracing at

several stations along the Vuvu River would address the question of how much sediment is stored between the upper and lower catchment. Furthermore, existing buffers, such as floodplains and wetlands, should be assessed to determine the current functioning of sediment traps and the potential to improve the sediment trapping efficiency. Hydrological and sediment connectivity should be separated in future work. Further investigations of how catchment wide rehabilitation will affect valley floor sediment dynamics would be valuable in terms of expected sediment yields after rehabilitation. Long term monitoring should be conducted to validate the conclusions that were drawn from this study.

8.5. Recommendations for further rehabilitation

The following recommendations could contribute to reduced hillslope-channel connectivity and therefore suspended sediment in the Vuvu River:

- Increase the vegetation cover, especially during dry cycles, throughout the catchment by: 1) having strict grazing rules that will allow the veld to rest and recover and 2) reducing the frequency and extent of fires.
- Improve contour agriculture with well vegetated buffers equally spaced along and directly below fields.
- Establish vegetation buffers, such as Vetiver grass, in and at the base of gullies to trap sediment. This will help with the stabilization of gullies by trapping sediment and building up a sediment sink that will store water and enhance vegetation growth and sediment trapping efficiency. Sediment trapping should propagate upslope leading to further gully floor stabilization and sediment trapping.
- Prevent new across-slope pathway formation and the deterioration of existing pathways by formalising access routes for livestock and reducing up and down catchment movement by establishing dipping stations higher up in the catchment.
- Create runoff diversions, using non-flammable materials, along livestock tracks and roads to channel water onto vegetated slopes in order to disperse runoff, allow for infiltration and sediment trapping.
- Protect seeps, wetlands, vegetated drainage lines and flood benches from grazing, trampling and fire as dense vegetation growth will aid sediment trapping and the potential release of sediment. Consider grazing these areas during winter when rainfall is less erosive.
- Restore degraded wetlands and alluvial fans in order to optimise sinks.
- Construct artificial wetlands on very low angled slopes ($<1^{\circ}$) to trap sediment.
- Consider gully restoration (constructing gabions) only for low angled ($<1^{\circ}$) sections along gullies.
- The soil along the sharp edges of gullies are mostly dry due to the increased drainage efficiency of the incised feature and does rarely support good vegetation cover. Replanting along these

edges, even if resloped and covered by a biomat or mulch before planting is unlikely to restore the necessary soil moisture regime needed for plant proliferation due to the ongoing desiccation of the soil. Only consider using this method for smaller features where the difference between the top and bottom of the feature is less than 0.5 m. The desiccation effect should be less for these smaller features, thus could the replanting be more successful.

- Construct non-flammable silt fences at regular intervals of 10-50 m along a hillslope to trap sediment and promote water infiltration.
- Construct check dams and contour walls where soils are not prone to pipe formation, but avoid this method where pipe formation is likely as it will increase the concentration of free water in the subsoils and promote pipe erosion (van Zijl et al., 2013). All duplex soils are prone to pipe erosion due to the contrasting soil texture and dispersive character (van Zijl et al., 2013).
- It is worthwhile restoring a gully at the beginning of an erosional cycle, but this is less useful to restore a feature that is nearing the end of the erosional cycle as erosion slows down (van Zijl et al., 2013). Indications of features that near the end of the erosional cycle are: they are continuous or linked to the drainage network, have reached bedrock throughout most of the feature and has reached the steep back slope with the headcut (Nordstrom, 1988). This is the case unless new pipes and gullies are formed. Existing gullies can trigger further pipe erosion in duplex soils as it drains and mobilizes accumulated free water and by so doing initiate pipe formation (van Zijl et al., 2013).
- Restore gully heads of systems that are in the initial erosional phase and protect inter gully areas as they are susceptible to sheet erosion (van Zijl et al., 2013).

The use of the connectivity framework proved beneficial to understand where sediment is eroded from, how sediment is transferred through the landscape and where and how much sediment is stored. The outputs of the research can be used to plan and guide which areas should be targeted in rehabilitation work. The Elliot Formation stood out as the geology with the greatest sediment yields, as was confirmed by the mapping of erosion features and sediment tracing; thus it is the key area to focus rehabilitation efforts. Fields, sheet erosion and gully erosion produced the highest sediment yields and should be targeted first. Gullies, livestock tracks and roads act as important conduits of water and sediment, especially where these features breach sediment sinks (e.g. gentle slopes, wetlands, etc.). These features should thus be targeted to reduce down-slope sediment delivery. By focusing rehabilitation work on these key sediment sources and pathways, suspended sediment delivery hopefully can be reduced to sustainable levels that would improve the livelihoods in the catchment and for downstream water users.

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Appendices

Appendix 1: Sediment source and pathway shapefiles for the Vuvu catchment

The spatial data for the sediment sources and pathways are included in digital shapefile format on the attached CD. All spatial data were digitized of colour 2009 ortho photos (0.5 m resolution) at a scale of 1:2 000. All data are projected to Transverse Mercator 29 for area calculations. Detailed metadata for each feature is embedded into the shapefiles. The following shapefiles are included:

Fields

Continuous gullies

Discontinuous gullies

Sheet erosion

Landslides

Roads

Livestock tracks

Vuvu catchment.

Appendix 2: OSL dating report



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**School of Geosciences
Geo-luminescence Laboratory**

OPTICALLY STIMULATED LUMINESCENCE DATING REPORT

THINA and VOSL SAMPLES

for

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Compiled by

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University of the Witwatersrand

2014

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Introduction

Luminescence dating emerged as a technique for determining the age of ancient ceramics in the early 1960's and 1970's, through the use of thermoluminescence (TL) or the light emitted by a sample when it is heated. Wintle and Huntley (1979; 1980) presented a workable TL dating method for determining the time of deposition of sediments; first for deep sea sediments, then for a range of aeolian and aquatic clastic sediments. Huntley et al. (1985) showed that light could be used to stimulate the dating signal from sedimentary minerals and this work suggested that luminescence dating methods would offer additional practical and methodological advantages to dating sediments. Subsequently the luminescence technique has undergone major developments and significant improvements in instrumentation, measurement, procedures and selection of preferred minerals for dating (see Bøtter-Jensen *et al.*, 2003; Murray and Wintle, 2000).

Luminescence dating falls within the category of radiation dosimetric dating techniques which are based on the time-dependent accumulation of radiation damage in minerals (Aitken, 1998; Murray and Wintle, 2000), which is the result of exposure to natural low-levels of ionising radiation. The intensity of the radiation damage is a measure of the total dose or the total amount of energy absorbed from the ionising radiation by the mineral over a certain period of time. In luminescence dating the intensity of the radiation damage is detected as a small amount of light called *luminescence*. This latent luminescence signal can be removed or "zeroed" by exposure to heat or light. Thus for sediment dating the zeroing event or "bleaching" is the exposure to daylight during erosion, transport and deposition of the mineral grains. Once the sediments are buried again and no longer exposed to sunlight, the luminescence signal can start to build up again (Duller, 2004). The latent luminescence signal can also be released under laboratory conditions using heat (thermoluminescence) or light (optically stimulated luminescence) and this signal can be recorded. This luminescence signal is related to the dose the mineral has received since the last zeroing event (or last exposure to sunlight); and if the rate is determined at which the dose was absorbed, it is possible to determine an age for the sediment. However, this age refers to the time that has elapsed between the moment the sediment was deposited and the moment of sampling for analysis (Fig. 1). The OSL signal acquired during the previous burial period is reset by sunlight during sediment transport prior to deposition. Quartz or feldspar minerals are commonly used for OSL dating of sediments, but, quartz is the preferred dosimeter as the physical basis of OSL production in quartz is better understood; and feldspar may be subject to phenomena such as anomalous fading (e.g. Bøtter-Jensen *et al.*, 2003).

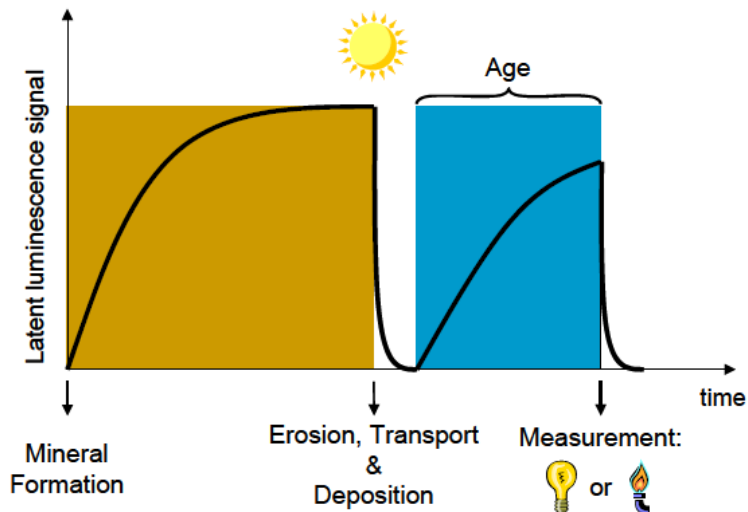
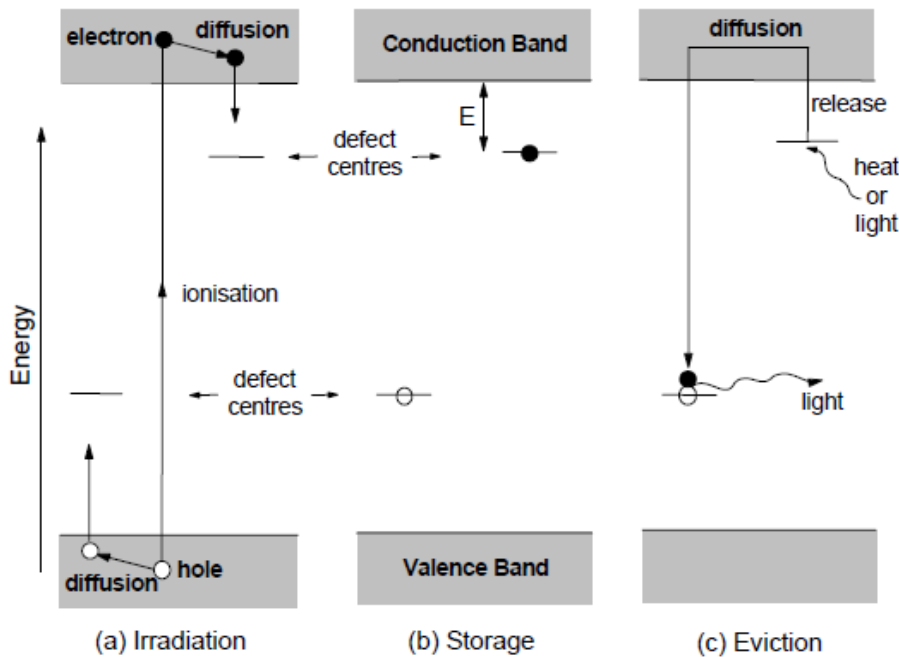


Fig. 1. A schematic representation of the event that is being dated in the luminescence dating of sediments where minerals are continuously exposed to a low-level of natural radioactivity, through which they can acquire a latent luminescence signal. During erosion, transport and deposition, the minerals are exposed to sunlight and all the previously accumulated luminescence is removed. Once shielded from the sunlight, the signal starts to build up again, until the moment of measurement in the laboratory. The age that is being determined is consequently the time that has elapsed between these two zeroing events (Duller, 2004)

The main process causing luminescence can be described in terms of the energy level diagram for non-conducting ionic crystalline materials (Fig. 2). Aitken (1998) indicated that electrons are associated with discrete ranges of energy called "bands"; where the lowest energy band is the valence band, and the highest energy band is the conduction band. The gap between the two is referred to as the "forbidden zone" which in an ideal crystal, no electron will occupy a position in this zone. However, in a natural crystal defects such as impurities or missing atoms are present and disturb the ordered crystalline structure, which give rise to the presence of energy levels in the forbidden zone. The conduction and valence bands extend throughout the crystal, but the defect states are associated with the defect itself and are referred to as localised energy levels which are significant to the luminescence phenomenon, as they carry the memory of exposure to nuclear radiation. Therefore luminescence requires the existence of lattice defects within the crystal structure (Aitken, 1998).



- hole
- - electron

Fig. 2. Energy level representation of the luminescence process (modified after Aitken, 1998).

In nature low-level radiation, through its ionising effect can raise an electron from the valence band into the conduction band (Fig. 2a). For every electron created, an electron vacancy (hole) is left behind and both the electron and the hole are free to move through the crystal and in this way the energy of nuclear radiation is taken up. This energy can be released again by recombination and most charges do recombine easily. However, the electron and the hole may be trapped at defect centres. In this instance the nuclear energy is stored temporarily in the crystal lattice and the system is said to be in a metastable situation (Fig. 2b). Energy is required to remove the electrons out of the traps and return the system to a stable situation, where more energy is required to empty deeper traps. By exposing the crystal to heat or light the trapped electrons may absorb enough energy to bridge the barrier to the conduction bands (Fig. 2c), where once excited they can be trapped again, or can recombine with holes in recombination centres (Aitken, 1998; Duller, 2004).

The recombination can result in either the emission of heat or light and the defect sites where radiative (light) recombination occurs called luminescence centres and the resulting light is called TL (if heat is used to release the electrons from the traps) or OSL (if light is used). The amount of light emitted is proportional to the amount of electrons stored in defects and therefore the amount of energy received from irradiation. Since the energy is absorbed at a certain rate, the intensity is related to the time of accumulation, i.e. the longer the material is exposed, the more signal is acquired (Aitken, 1998). The process of stimulation and eviction of the electrons is termed signal bleaching, with stimulation by heat or light resulting in a TL or OSL signal, respectively. If the grain is stimulated sufficiently to evict all the trapped electrons, it is described as being fully bleached. Whilst the TL glow curve contains contributions from traps of various depths, the main source of the OSL signal is thought to be the traps associated with the readily bleachable 325°C TL trap (Aitken, 1998). As OSL is generally very much more rapidly reset than TL, it is more suitable for dating sediments where the bleaching occurs by exposure to light.

The radiation to which a grain is exposed during burial derives from a number of different sources, and consists of alpha and beta particles, and gamma rays. Quartz grains contain very little internal radioactivity, so the majority of the radiation they receive is derived from uranium, thorium, and potassium contained within the surrounding sediment, as well as a small level of cosmic radiation. The radiation flux at a sampling location is termed the environmental dose-rate, and can be obtained either by field or by laboratory measurements. The amount of radiation that a grain has received during burial can be determined by measuring the OSL signal from the natural dose (i.e. that received during burial), and the OSL signals from a series of laboratory irradiations of known dose. The laboratory measurements are used to calibrate the OSL signal derived from the natural dose, and thus determine the laboratory dose that is equivalent to the dose received by the grains in nature; this is termed the equivalent dose (D_e), with the unit of measurement in Grays (Gy). Since the OSL signal is bleached on exposure to light during sediment transport, the D_e normally represents the amount of radiation received following deposition and burial after the last transport event. By deriving the burial dose (D_e) of a sample, and the environmental dose-rate at the sampling location, the age of a sample can be found using the equation:

$$Age(yr) = \frac{EquivalentDose(Gy)}{DoseRate(Gy.ka^{-1})}$$

where the equivalent dose (in Grays) is the laboratory radiation dose that produces the signal equivalent to the natural luminescence signal for the grains being measured (Duller, 2004). The dose rate (in Grays per thousand years) is the annual dose received by those grains in their depositional environment (Duller, 2004). This annual dose rate refers to the rate at which mineral grains in sediment absorb energy from the

surrounding flux of radiation which comes from the naturally occurring radionuclides ^{232}Th , ^{238}U and ^{235}U and their daughters, and ^{40}K and ^{87}Rb as well as cosmic radiation.

Ideally, the OSL signal is derived from a single type of mineral grain such as quartz or potassium feldspar mineral separates. The dating technique can be applied to the age range from present to approximately 300 000 years ago and precision varies from ~3-10% of age for heated materials and ~5-20% for sediments (Wintle *et al.*, 1993).

Initial studies of the application of OSL dating to sediments concentrated on depositional environments where the transport process ensured that sufficient exposure to sunlight to bleach the material had occurred, such as aeolian deposits (e.g. Huntley *et al.*, 1985; Stokes, 1992). In some other depositional settings (e.g. fluvial or colluvial), not every grain receives light exposure of a sufficient strength and/or duration to bleach fully the dose from the previous burial period (Fig. 2b). If a residual trapped charge remains in the grains, the sediment is regarded as partially or incompletely bleached. Where partial bleaching is present, careful assessment of the distribution of De values is necessary to obtain the correct age. Simply taking some measure of the average from the De values for a partially bleached sample is not appropriate because the grains with residual trapped charge will cause overestimation of the age (Wallinga, 2002; Kim, 2009).

Sample Details

Six samples were submitted for analysis although sample VOSL was withdrawn.

Sample	Latitude	Longitude	Altitude (m)	Depth below surface (m)	
THINA 1	-30	28	30.6	0.47	
THINA 2	-30	28	30.6	1.45	
THINA 3	-30	28	30.8	2.43	
VOSL 1	-30.60094	28.25921	1499	0.67	
VOSL 2	-30.60037	28.25497	1506	0.94	
VOSL 3	-30.60055	28.25360	1511	1.55	
VOSL 4	-30.60095	28.25306	1508	1.34	
VOSL 5	-30.60079	28.25274	1514	1.4	
VOSL 6	-30.60146	28.25049	1513	1.57	Pb-210 present younger than 100 years

Sample Preparation

Sediment was removed from each sampling tube under controlled, laboratory safe-light conditions. The sediment located within 2 cm of both ends of each tube was removed to isolate any sediment potentially exposed to light during sampling. This material was used to measure the water content and to determine the dosimetry of the sample. The remaining 'bulk' content of each sample was wet sieved to remove any particle sizes smaller than 63 μm . The remaining sediment was treated with 33% hydrochloric acid and 20% hydrogen peroxide to remove any carbonate and organic components. Quartz grains were isolated from any other minerals by heavy liquid separation using sodium polytungstate. After rinsing, drying and sieving, the fine sand (180-212 μm) fraction was then etched for 40 mins in 40% hydrofluoric acid, in order to remove the outer 10-15 μm layer affected by alpha radiation and degrade each samples' feldspar content. 30% hydrochloric acid was added to remove acid soluble fluorides. Each sample was dried and re-sieved in preparation for equivalent-dose determination.

Determination of Equivalent Dose (D_e)

Measurement apparatus

All luminescence measurements were performed on a Risø TL/OSL Reader Model -DA-15, an automatic measurement system that enables measurement of both thermoluminescence and optically stimulated luminescence of quartz, feldspar and poly-minerals. The reader is programmed through a sequence protocol written on Sequence Editor (Risø National Laboratory) which is directed through the MINI-SYS computer attached to the reader. The reader then carries out the measurements fully automated. The system allows up to 48 samples to be individually heated to any temperature between room temperature and 700°C, individually irradiated by radioactive beta or alpha ($^{90}\text{Sr}/^{90}\text{Y}$) source and optically stimulated using various light sources. The emitted luminescence is measured by a light detection system comprising a photomultiplier tube and suitable detection filters. The sample is placed in a light tight sample chamber which can be programmed to be evacuated (vacuum) or have a nitrogen atmosphere maintained by a nitrogen flow. The standard photomultiplier tube (PMT) in the Risø TL/OSL luminescence reader is bi-alkali EMI 9235QA PMT, which has maximum detection efficiency at approximately 400 nm. A 7.5 mm Hoya U-340 detection filter prevents the stimulation light from reaching the PMT.

Optical stimulation is achieved using an array of light emitting diodes (LEDs), which are compact, fast and enables electronic control of the illumination power density. The LEDs are arranged in 7 clusters each containing 7 LEDs (i.e. a total of 49 LEDs). The distance between the diodes and the sample is approximately 20 mm. The Risø TL/OSL luminescence reader is equipped with two stimulation sources: 1) Infrared (IR) LEDs emitting at 875 nm arranged in three clusters each containing seven individual LEDs. The maximum power from the 21 IR LEDs is approximately 135 mW/cm² at the sample position; and 2) Blue LEDs emitting at 470 nm arranged in four clusters each containing seven individual LEDs. The total power at the sample position from the 28 LEDs is ~40 mW/cm². A green long pass filter (GG-420) is incorporated in front of each blue LED cluster to minimise the amount of directly scattered blue light reaching the detector system.

Measurement protocol

OSL measurements were carried out the grain-size fraction 180 – 212 μm . A monolayer of grains was prepared on steel discs, over an area of 1 mm diameter; resulting in aliquot sizes of ~30 grains. The OSL signal was collected from the initial 0.20 s of the decay curve, with the subsequent 0.50 s used for the ‘background’. This channel choice maximizes the dominance of the ‘fast’ OSL component while keeping the signal-to-noise ratio as high as possible (Cunningham & Wallinga, 2010). The measurement

procedure follows the Single Aliquot Regenerative dose (SAR) protocol of Murray & Wintle (2003), with an additional IR stimulation prior to OSL measurement used to reduce any contaminating signal from feldspar grains

The steps of the SAR protocol are given in Tables 1 and 2. An appropriate preheat temperature was determined using a ‘thermal-transfer test’, in which the OSL signal is repeatedly measured after progressive heating steps. The TT-test allows the onset temperature of unwanted thermal transfer to be estimated for samples where poor bleaching makes the usual ‘preheat plateau’ test unfeasible. The D_e was evaluated using linear fit to a single regenerative dose point. The error term on the D_e assumes poisson statistics, with an additional 1% uncertainty added to each OSL measurement (Galbraith, 2002; Duller, 2007). The protocol includes a zero and repeated dose point; aliquots were accepted if recuperation was negligible and the recycling ratio between 0.9 and 1.1.

Table 1. The SAR protocol. for THINA

<i>Step</i>	<i>Treatment</i>
1	Give dose (Natural,1,2,5,10,20,0,1 Gy)
2	Preheat 200°C for 10 s
3	IRSL 40 s at 0°C
4	OSL 40 s at 125°C
5	Test dose (1 Gy)
6	Preheat 180°C for 0 s
7	IRSL 40 s at 0°C
8	OSL 40 s at 125°C
9	Illum 40s at 280°C

Table 2. The SAR protocol for VOSL

<i>Step</i>	<i>Treatment</i>
1	Give dose (Natural,11, 22, 44, 0, 11 Gy)
2	IRSL 235°C for 10 s
3	OSL 40 s at 125°C
4	Test dose (8 Gy)
5	IRSL 10 s at 225°C
6	OSL 40 s at 125°C
7	OSL 40 s at 210°C

The different measuring protocols were used based on the characteristics of the samples. The THINA samples were well bleached and sensitive samples, whereas the VOSL were less sensitive and with more scatter in the data. The VOSL samples presented with an ultrafast component which was not present in the natural samples but did appear in the test and regenerative OSL measurements. The infra-red preheat was used to eradicate this ultrafast component but that did not work entirely. The decay curves were analysed using channels 6-20 for the initial signal, and 21-57 for the background subtraction to remove the effect of the UF component as it would have decayed after five channels. The minimum-age model, with a σ_b of 12%, and bootstrapped errors was applied (Cunningham and Wallinga, 2010).

Dose rate determinations

A portion of the bulk sediment was sent to iThemba Labs, South Africa, for gamma-spectroscopy measurements, from which the isotopic abundances of the parent radionuclides has been estimated (Table 2). From these, the dose rates to quartz grains have been calculated using the latest conversion factors (Guerin et al., 2011). Allowance was made for the attenuation of the beta dose due to grain size (Mejdahl, 1979). Dose attenuation due to water follows Zimmerman (1971), using the measured water content values with a minimum of 5%.

The cosmic dose rate was determined as a function of altitude, latitude/longitude and depth, according to Prescott & Hutton's (1994) equations, plus the soft component from Madsen (2005). A constant sedimentation rate was assumed for the cosmic component.

Luminescence Ages

The luminescence age was obtained by dividing the palaeodose by the mean total dose rate and is shown in Table 3. The error on luminescence age estimates represents the combined systematic and experimental error associated with both the D_e and dose rate values. There is no datum for luminescence dates therefore the age reported is taken from date of sampling, i.e. AD 2012.

Table 3. Results

	<i>Th</i>	<i>U</i>	<i>K</i>	<i>WC (%)</i>	<i>Depth (m)</i>	<i>De (Gy)</i>	<i>Dr (Gy.ka⁻¹)</i>	<i>Age (a)</i>
THINA 1	9.2±0.3	1.95±0.05	1.34±0.05	10	0.47	2.27±0.12	2.30±0.13	996±78
THINA 2	6.8±0.3	1.37±0.05	1.03±0.04	10	1.45	1.83±0.10	1.79±0.10	1 019±81
THINA 3	9.1±0.3	1.94±0.05	1.17±0.05	10	2.43	0.25±0.08	2.21±0.13	113±37
VOSL1	3.43 ± 0.1	0.69 ± 0.06	0.89 ± 0.003	13.49	0.67	3.22 ± 0.68	1.45 ± 0.06	2270 ± 490
VOSL2	2.90 ± 0.1	0.28 ± 0.05	0.18 ± 0.002	18.27	0.94	3.51 ± 0.83	1.18 ± 0.04	2990 ± 720
VOSL3	3.40 ± 0.1	0.62 ± 0.05	0.86 ± 0.002	12.67	1.55	1.89 ± 1.01	1.35 ± 0.05	1400 ± 750
VOSL4	1.73 ± 0.08	0.16 ± 0.05	0.73 ± 0.002	10.66	1.34	2.23 ± 1.64	1.06 ± 0.042	2100 ± 1550
VOSL5	4.43 ± 0.12	0.54 ± 0.06	1.00 ± 0.003	20.00	1.4	6.54 ± 0.65	1.43 ± 0.05	4570 ± 490

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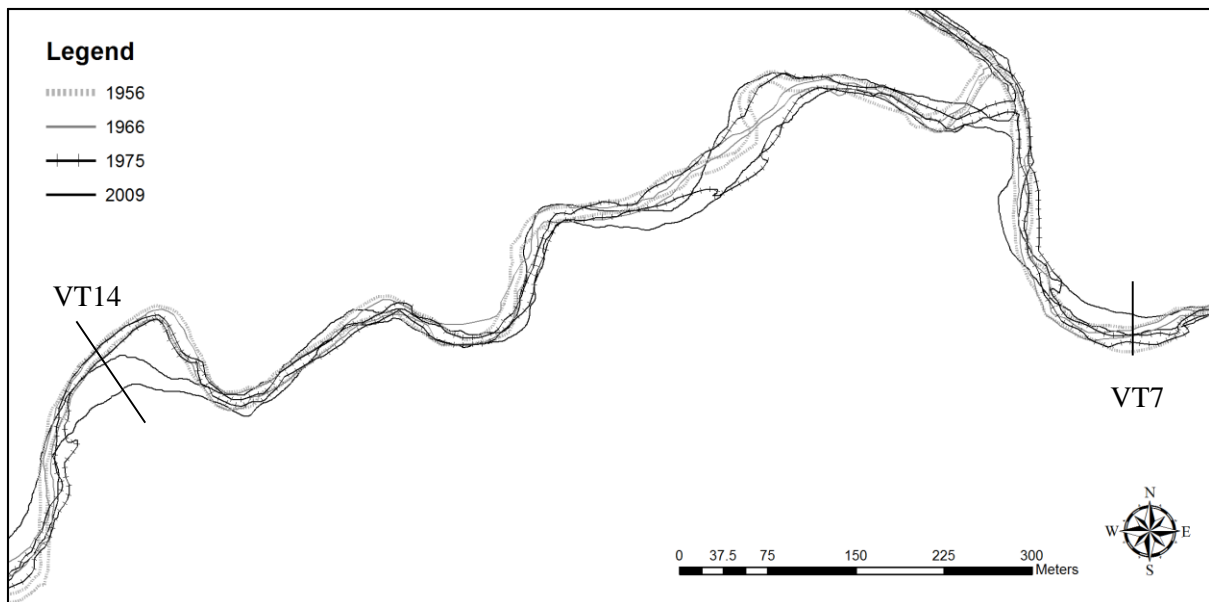
Appendix 3: Video footage of the Vuvu flood of March 2013 and maps showing channel straightening and transects where incision was measured

Vuvu flood

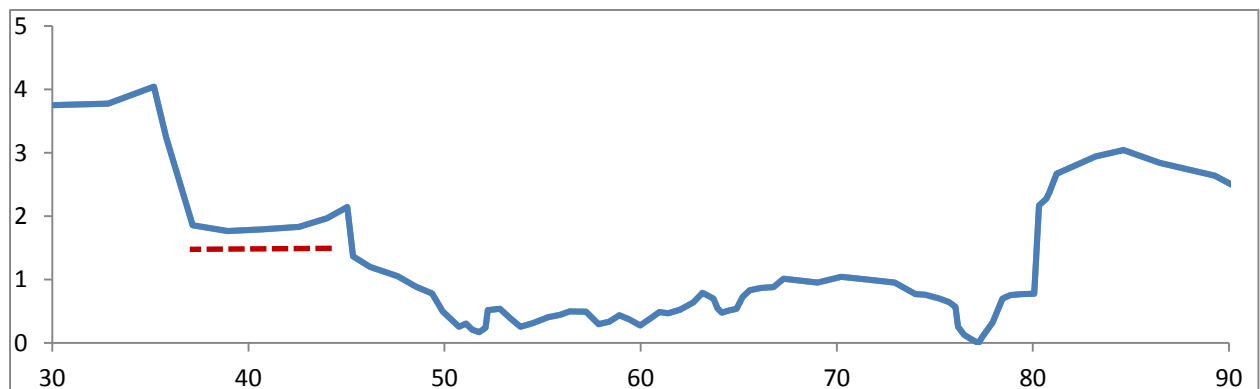
A 90 second video of a flash flood coming down the Vuvu valley fill. The Video can be found on the CD and is titled ‘Vuvu flood’.

Channel straightening and transects where incision was measured

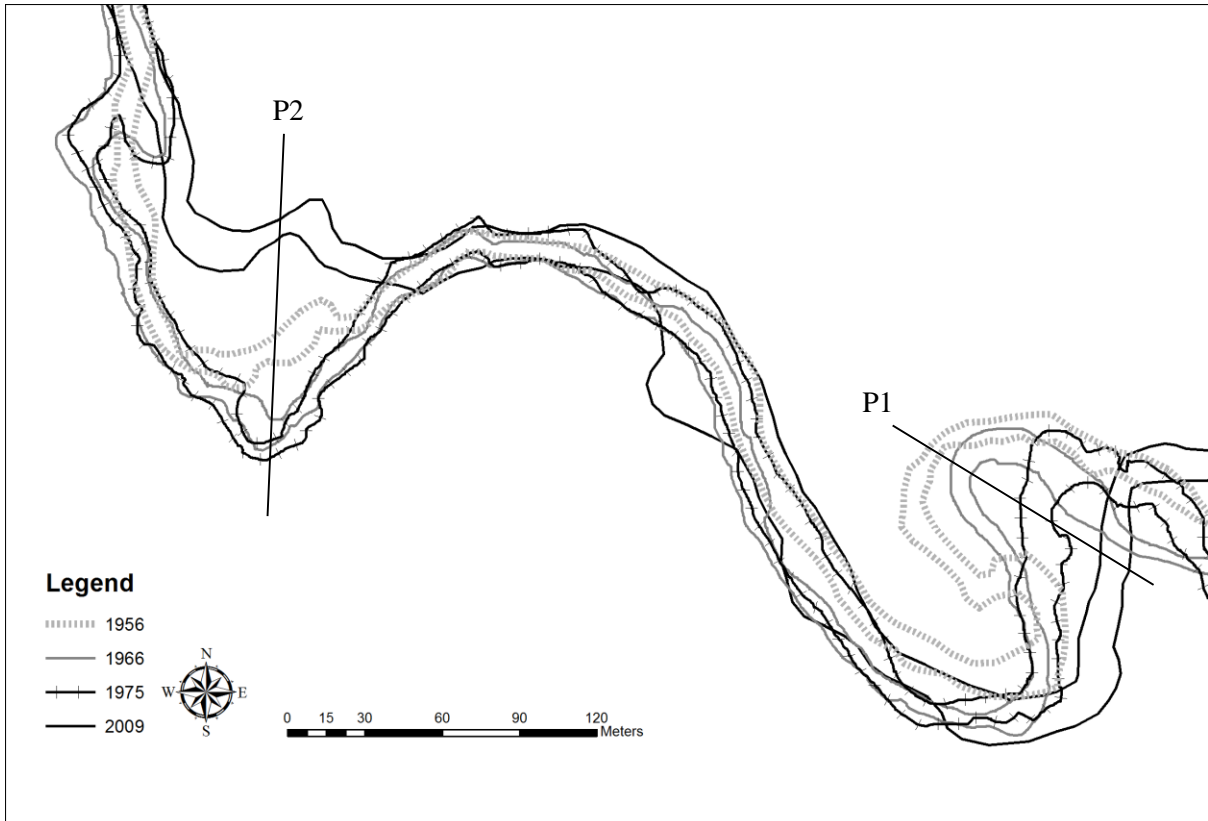
Maps showing channel straightening from 1956 to 2009 and cross sections where incision was measured in the Vuvu and Phiri-e-Ntso Rivers are shown below. Cross sections of the 1956 channel and current channel are also included. All measurements are given in metres.



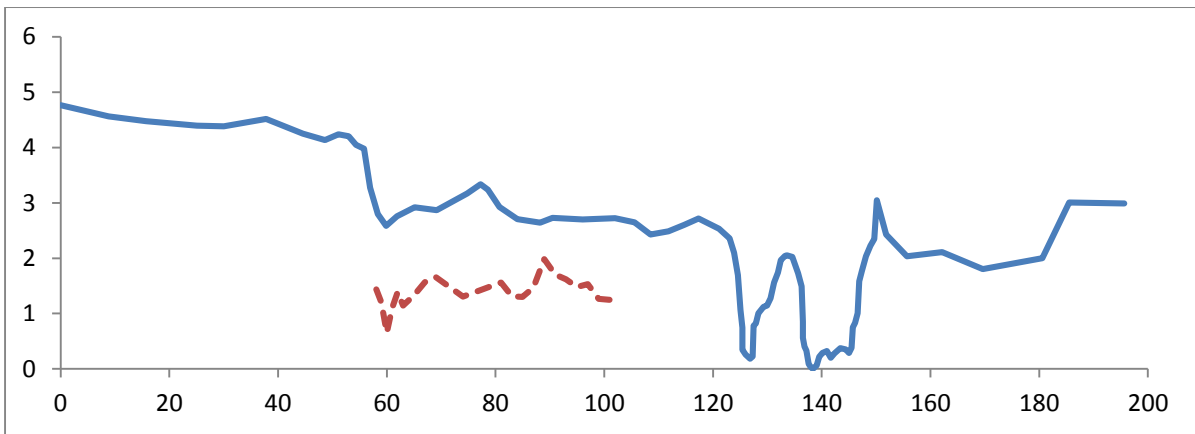
Channel straightening and the location of cross sections VT7 and VT14.



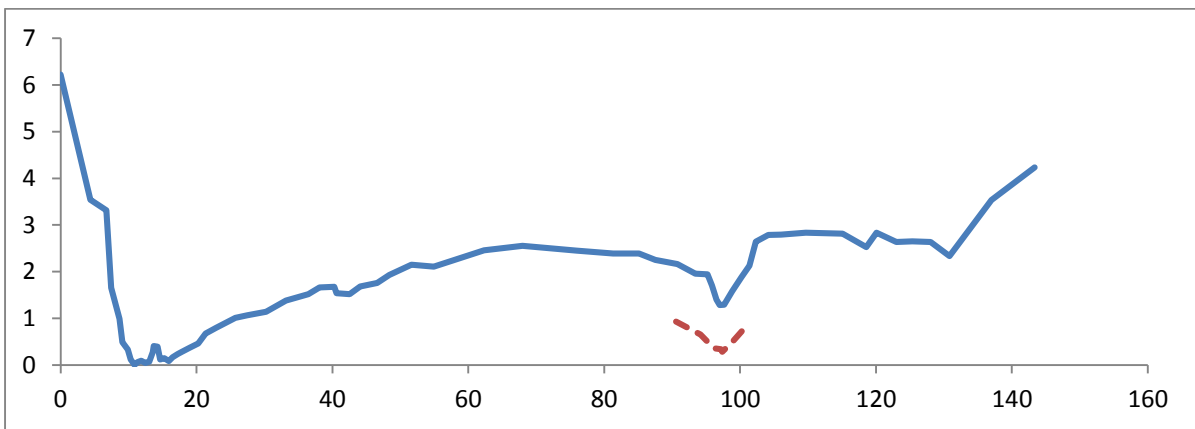
Cross section for VT14, 1956 level indicated by a dotted line.



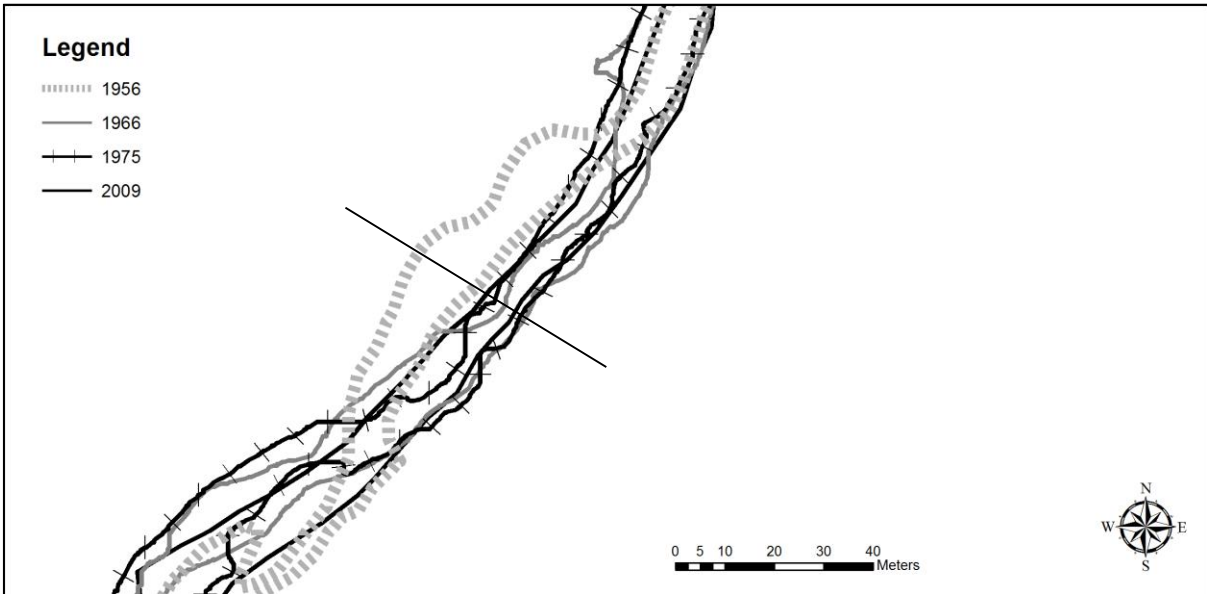
Channel straightening and channel cross sections for P1 and P2.



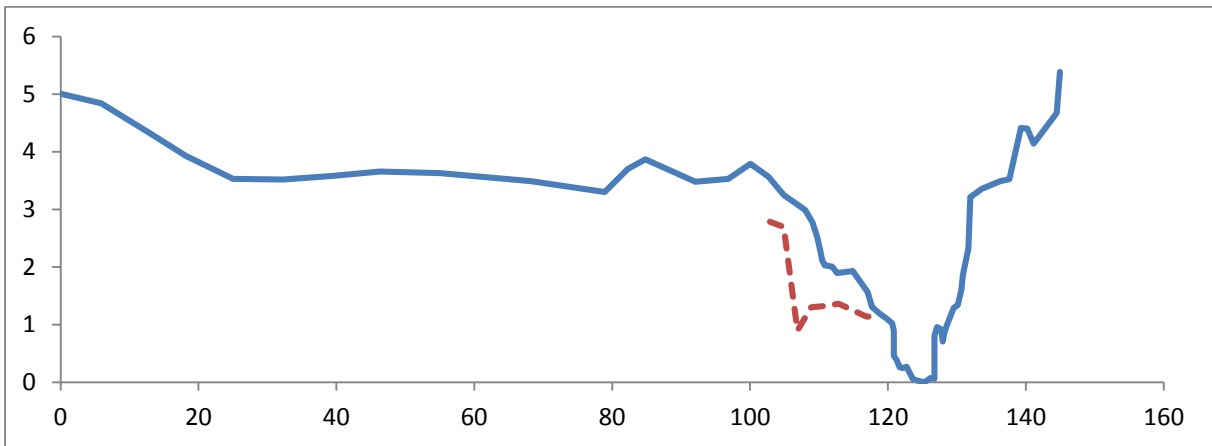
Cross section for P1, 1956 level indicated by a dotted line.



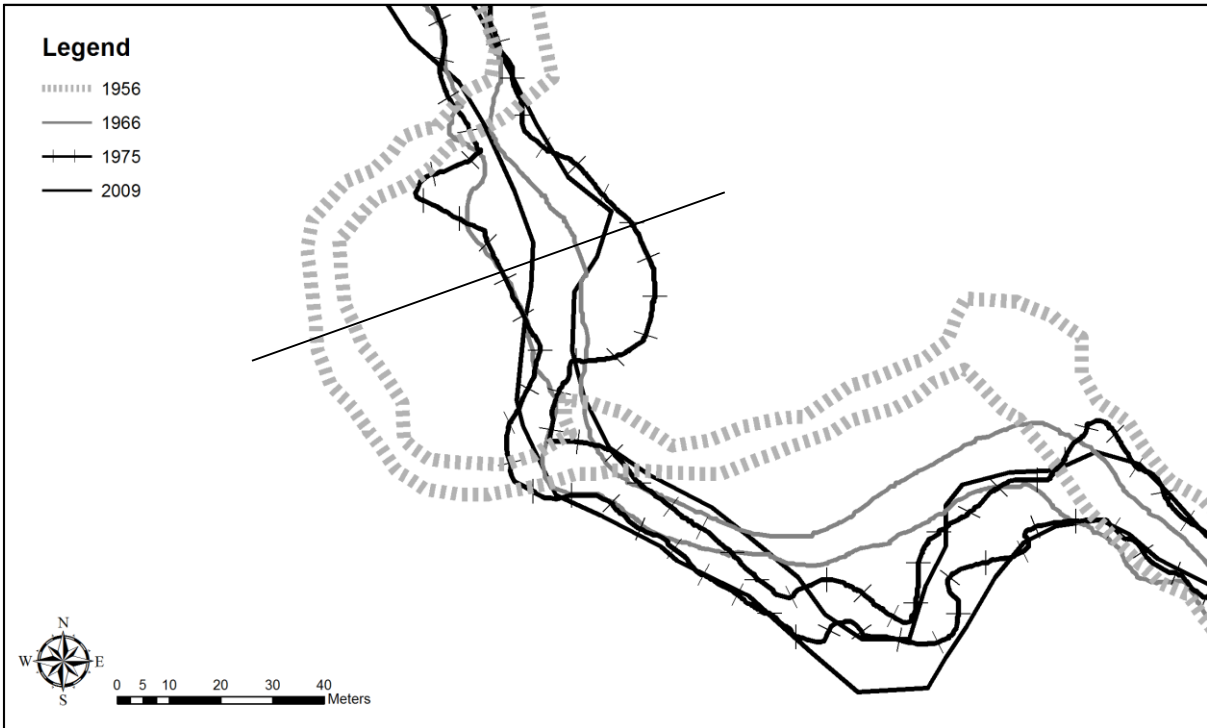
Cross section for P2, 1956 level indicated by a dotted line.



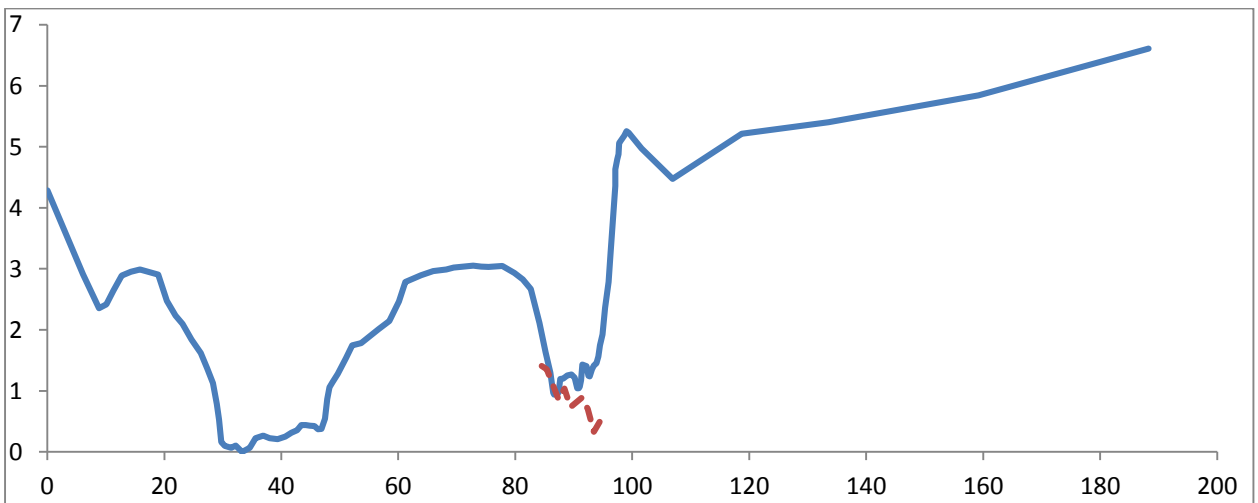
Channel straightening and channel cross sections for P4.



Cross section for P4, 1956 channel indicated by a dotted line.



Channel straightening and channel cross sections for P7.

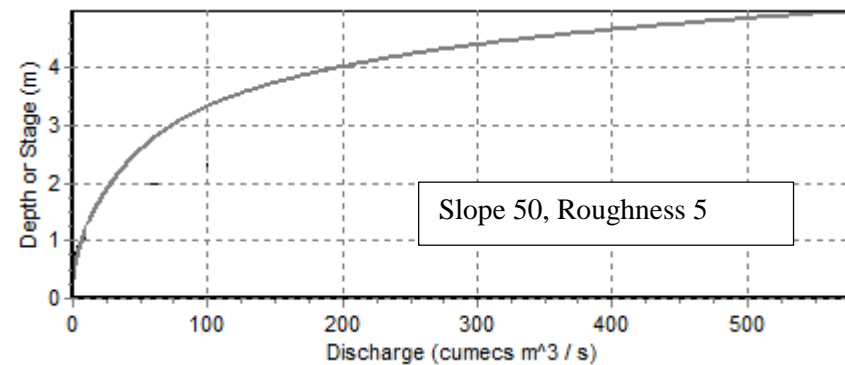
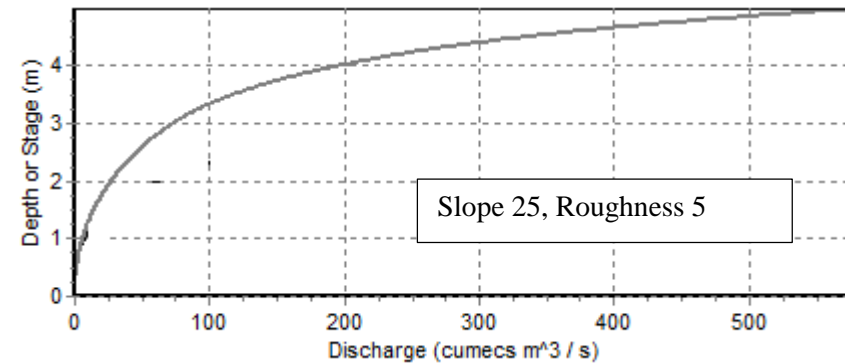
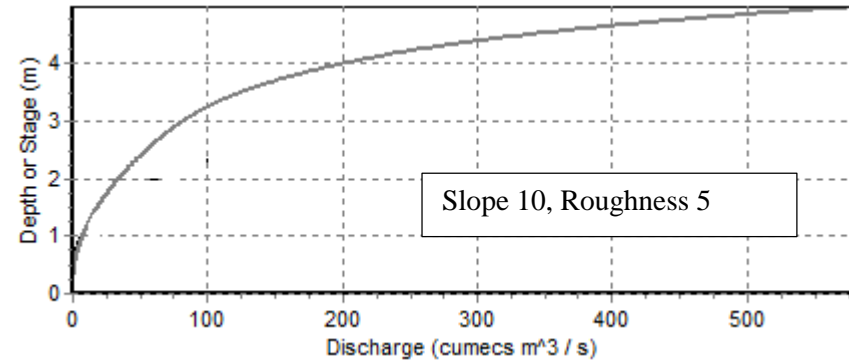
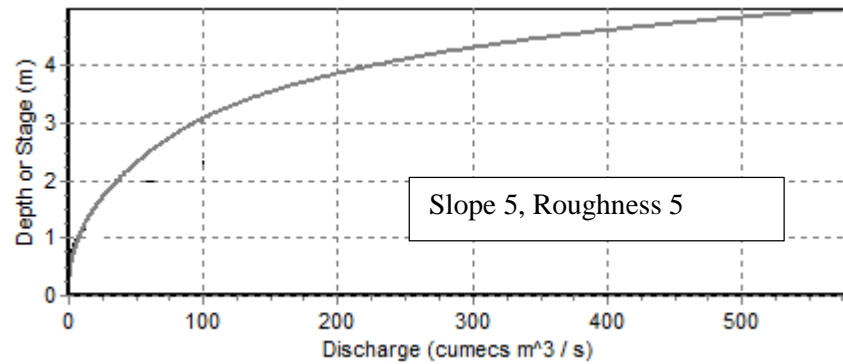


Channel cross section for P7. 1956 channel indicated by dotted line.

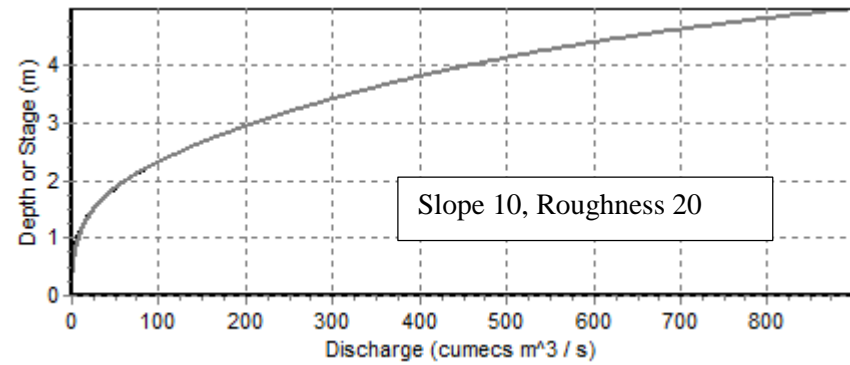
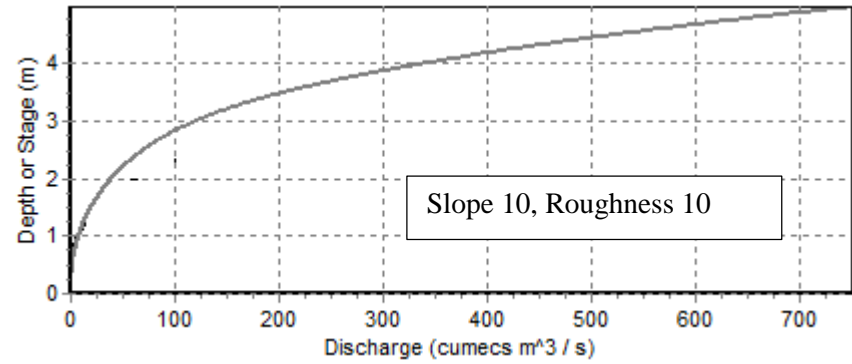
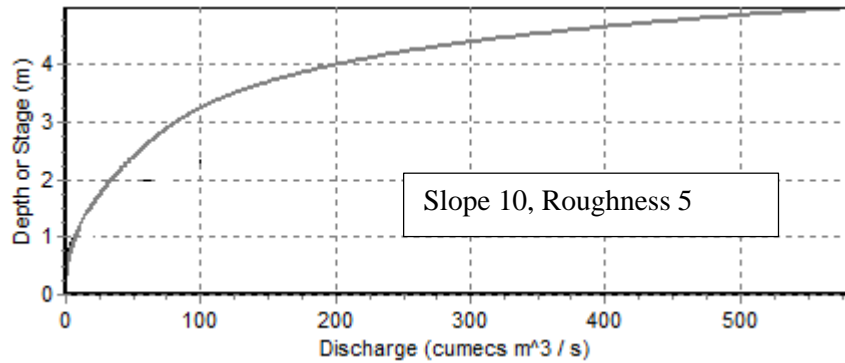
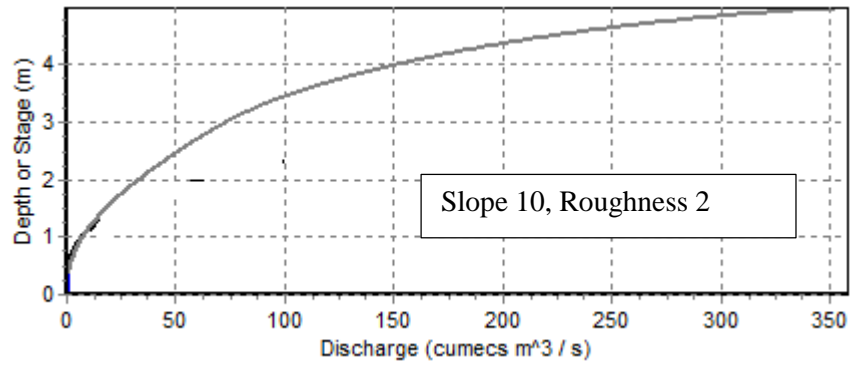
Appendix 4: Hydraulic sections and rating curves

Rating curves for VT1

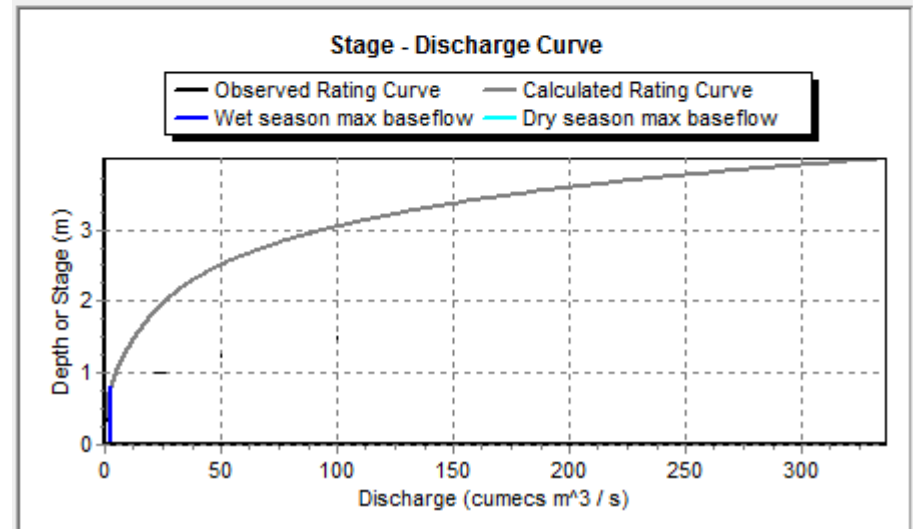
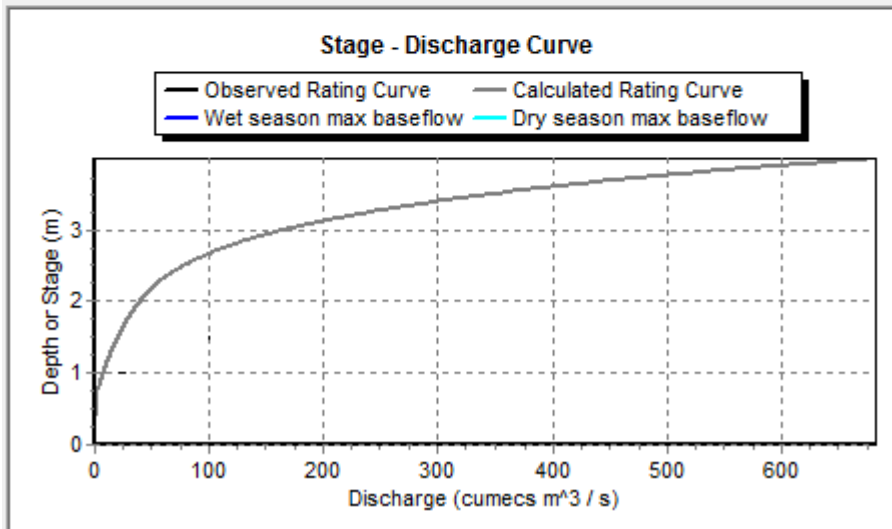
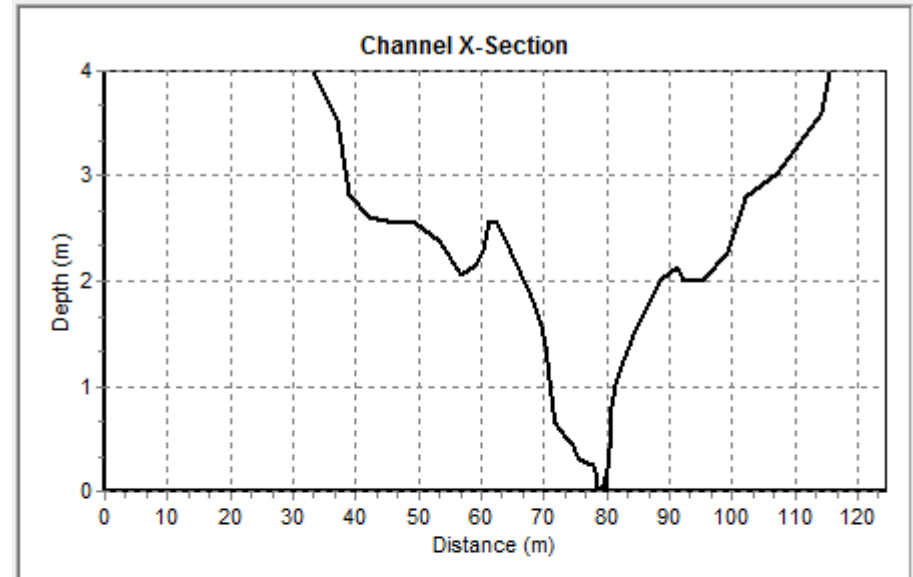
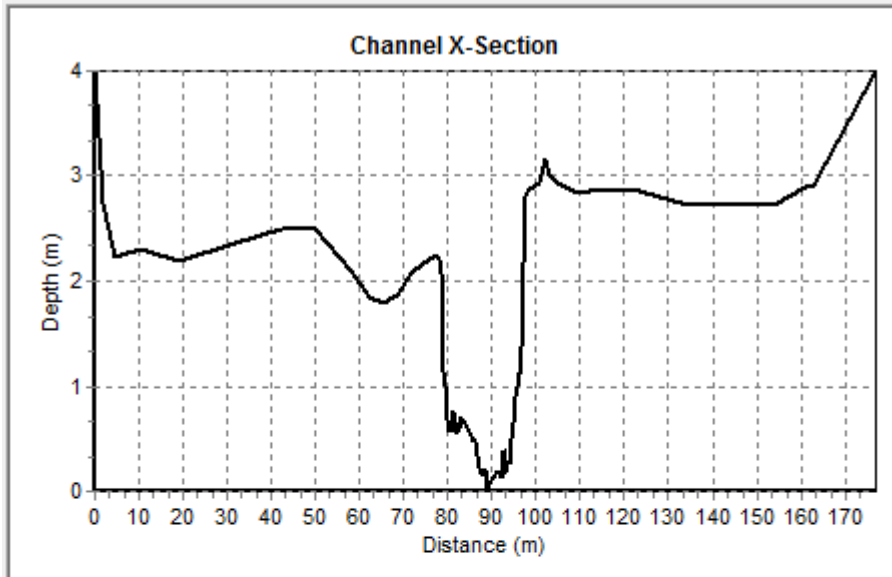
Various rating curves for VT1 showing the influence of the slope and roughness shape factor on the rating curve in SPATSIM's hydraulic module. A slope shape factor of 10 and roughness shape factor of 5 produced a reasonable fit between observed and modelled discharge given the uncertainty involved in the modelling. Variations in the shape factor is given below:



Variations in the roughness factor given below:

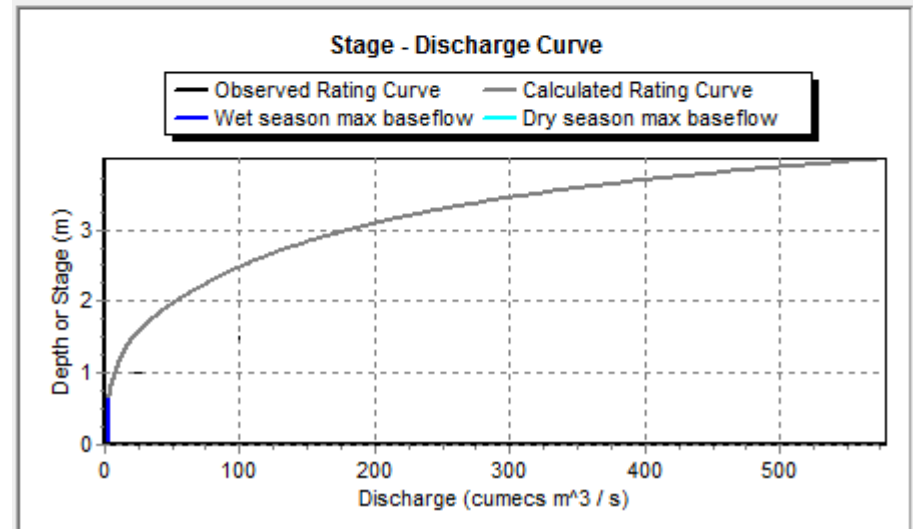
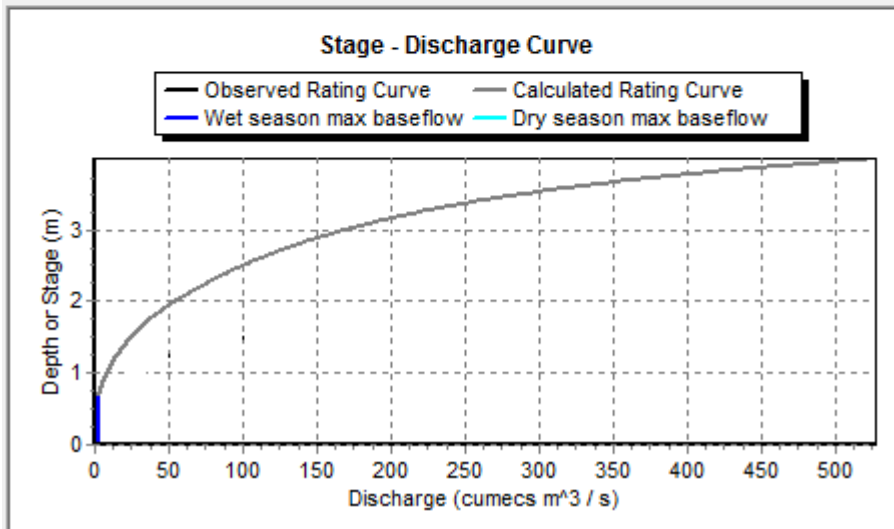
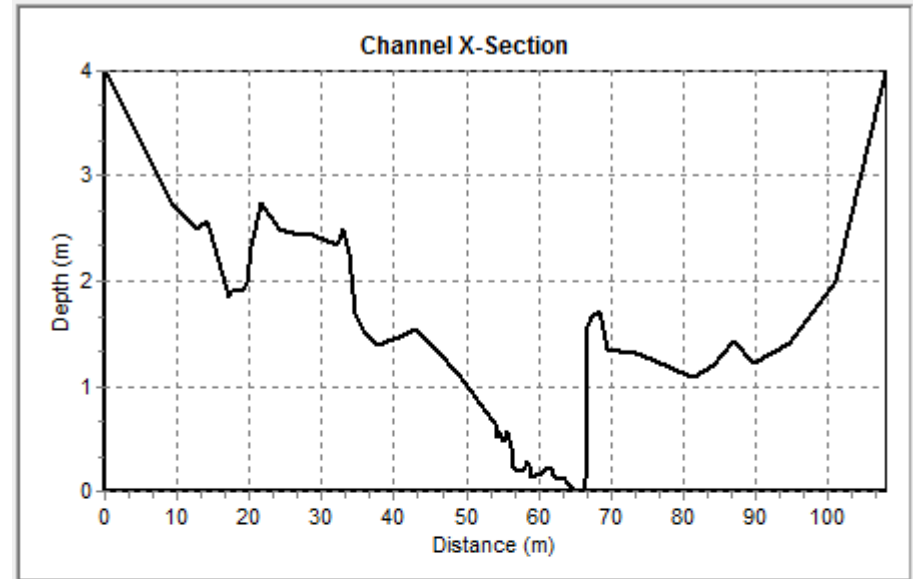
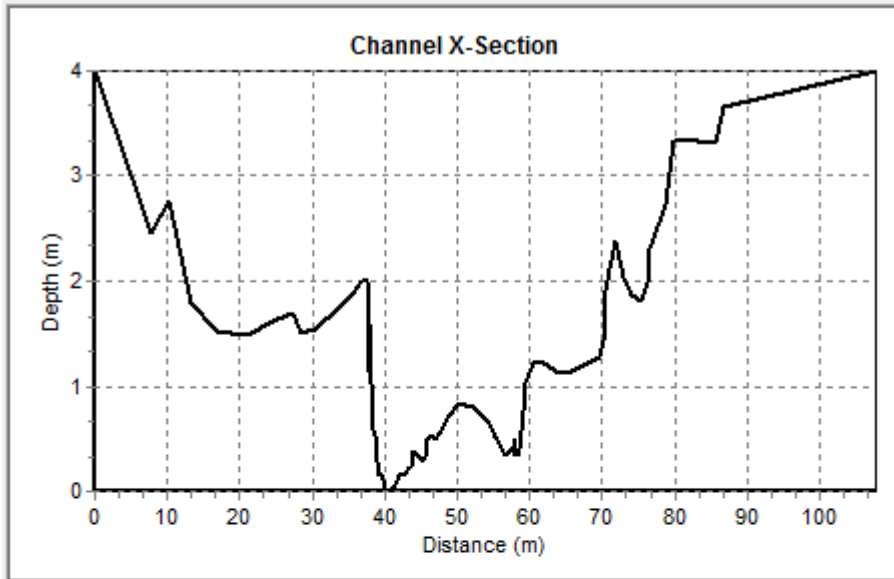


Screen prints from the SPATSIM Hydraulic module have been included below to show the hydraulic sections and rating curves for VT2-VT15. VT1 has been given in Chapter 7.



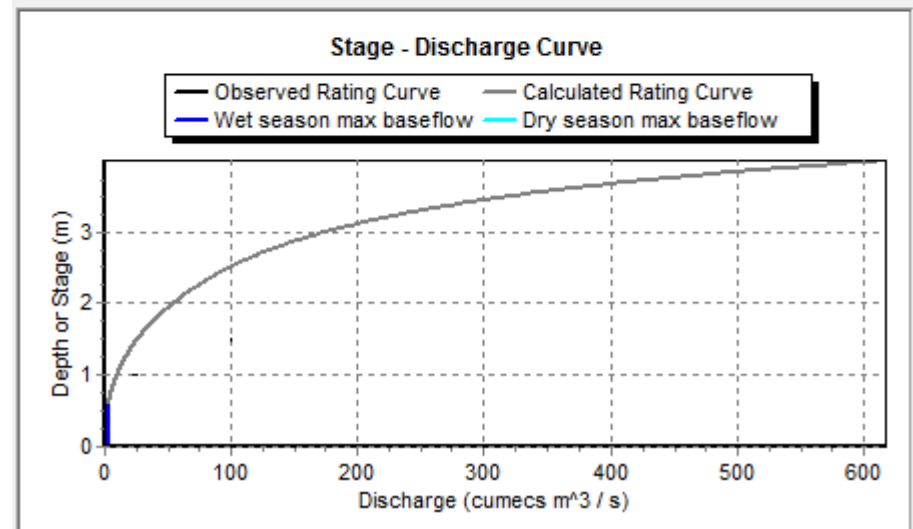
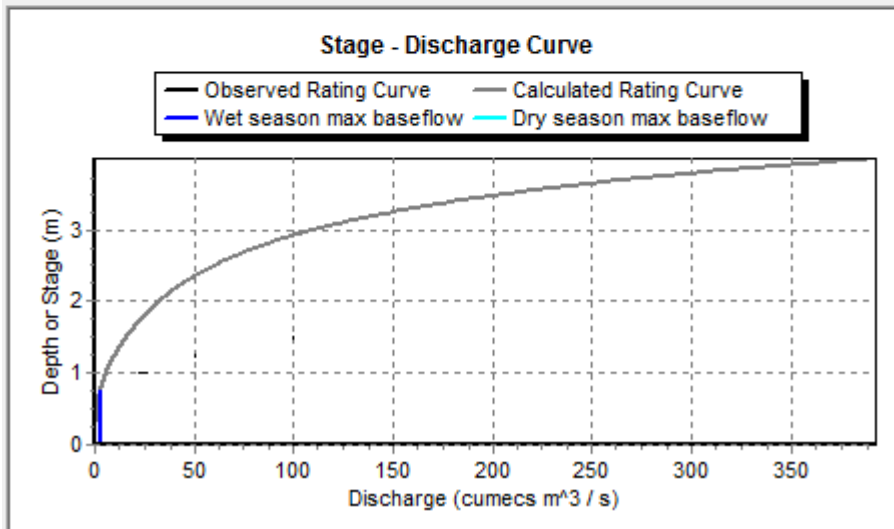
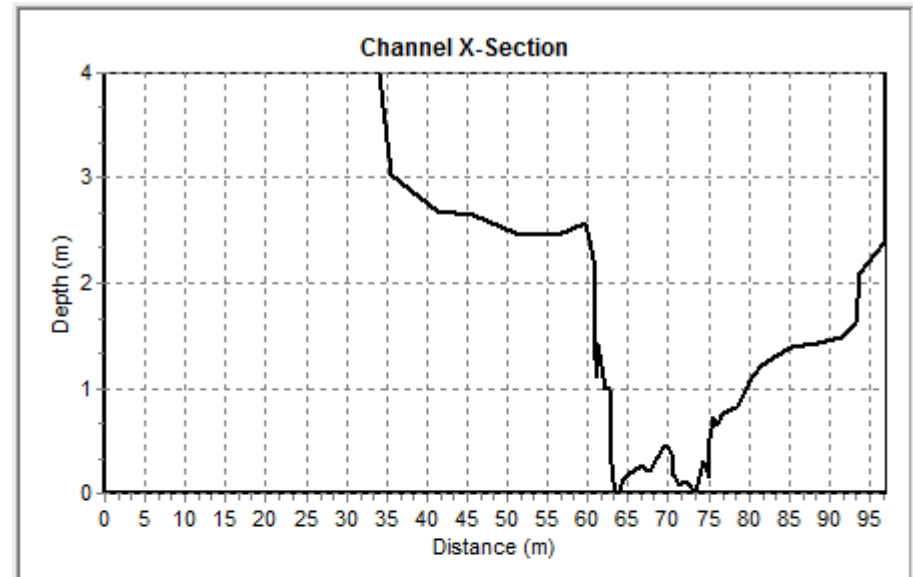
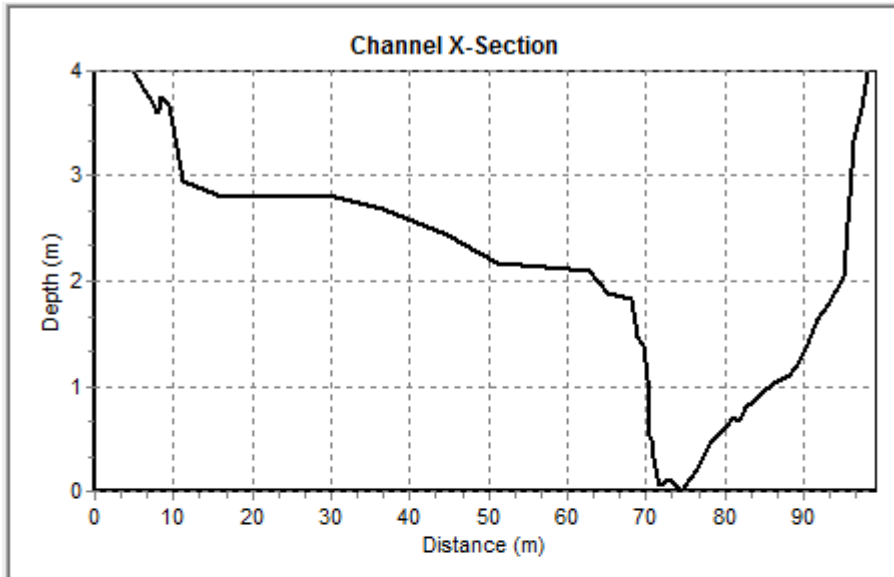
Hydraulic transect and rating curve for VT2

Hydraulic transect and rating curve for VT3



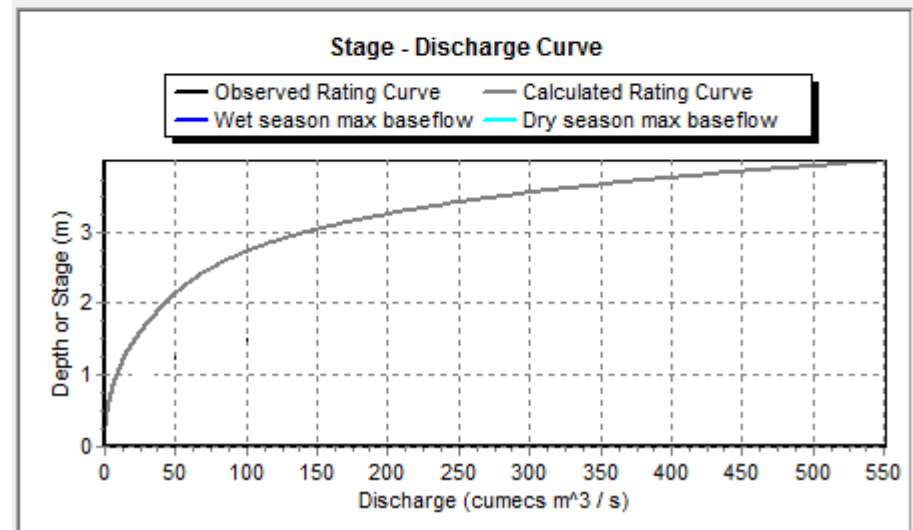
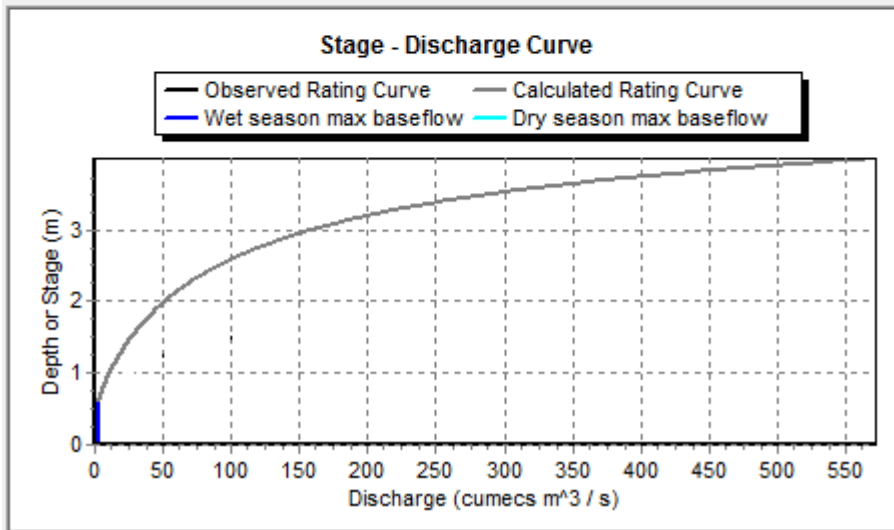
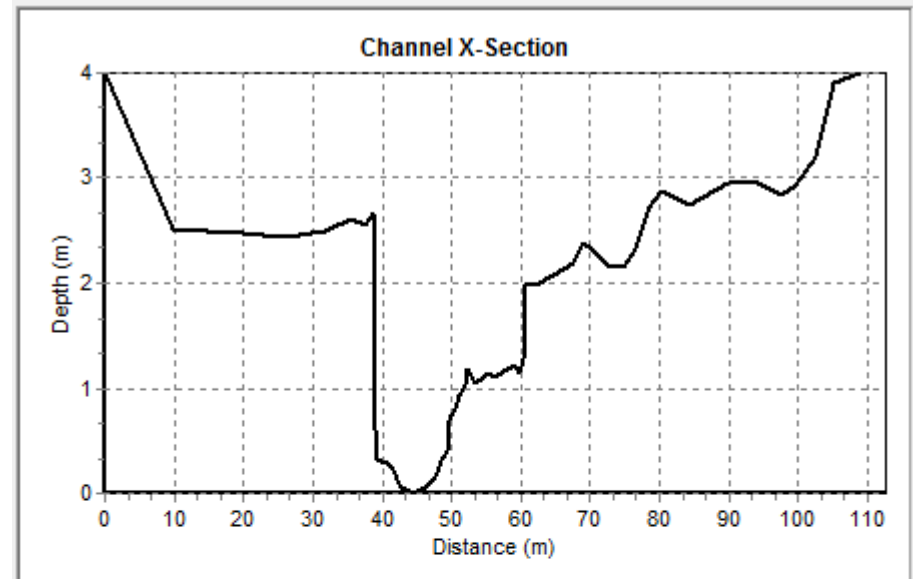
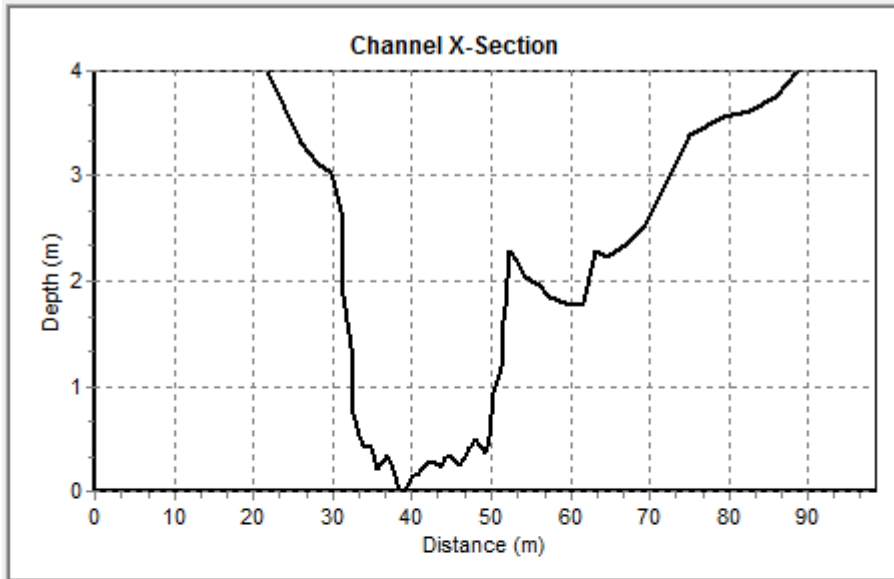
Hydraulic transect and rating curve for VT4

Hydraulic transect and rating curve for VT5



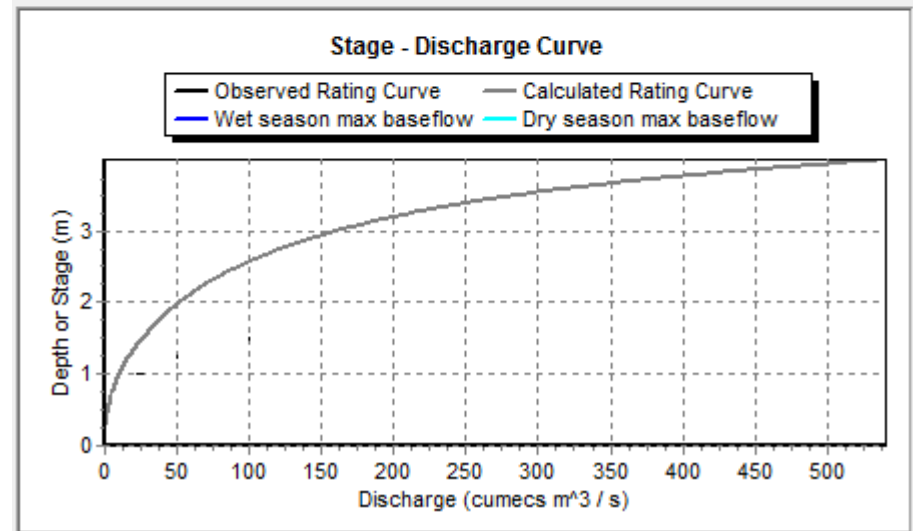
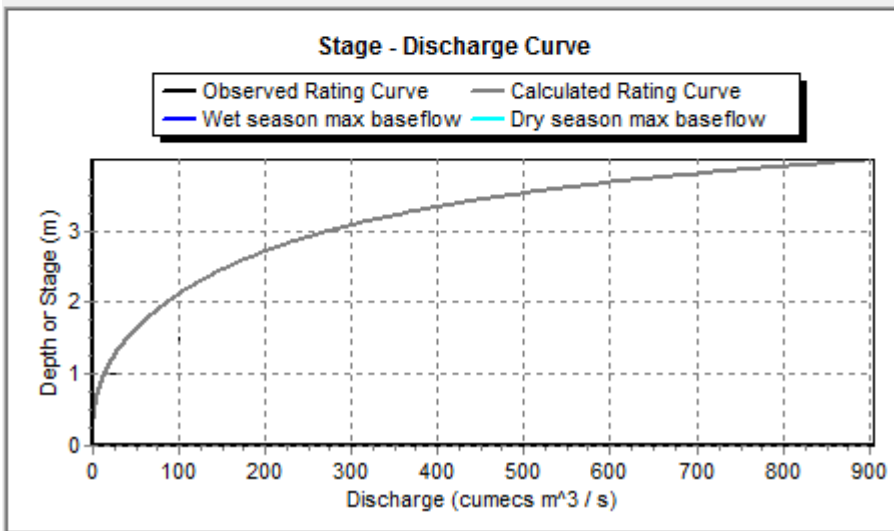
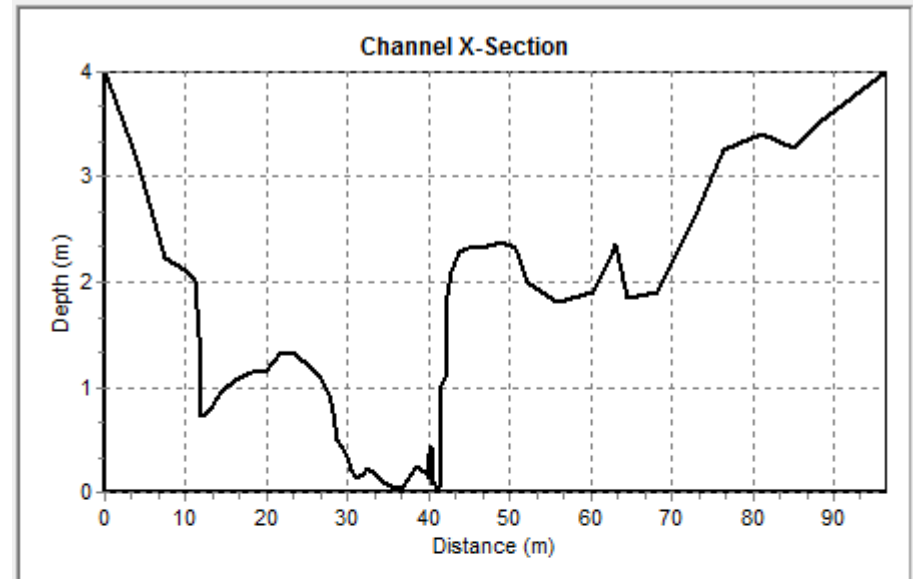
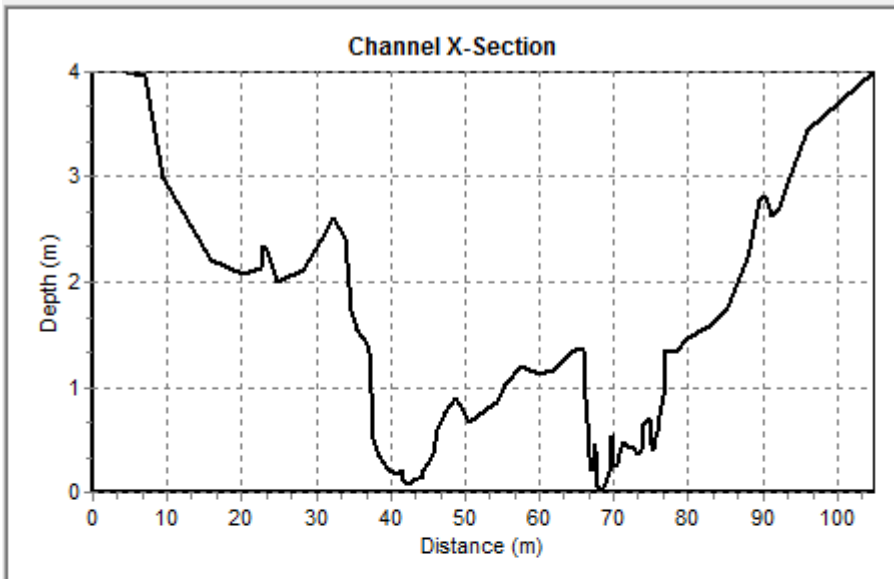
Hydraulic transect and rating curve for VT6

Hydraulic transect and rating curve for VT7



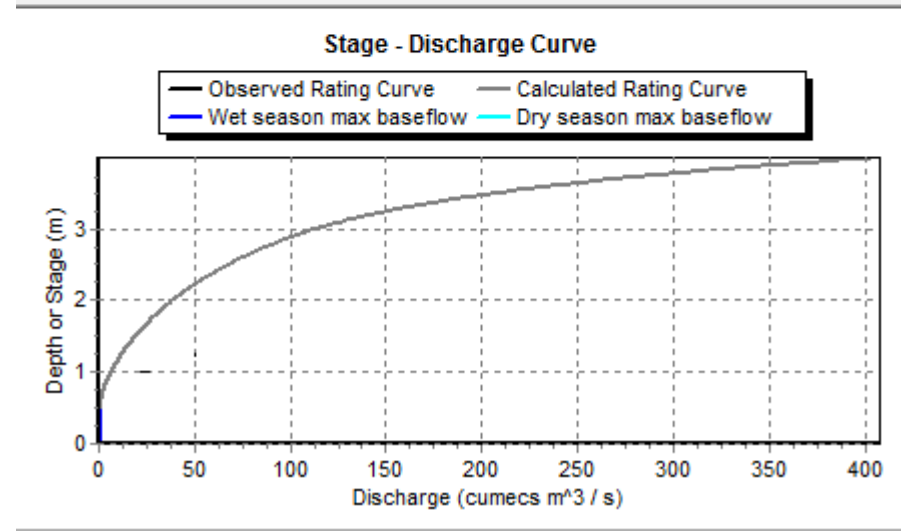
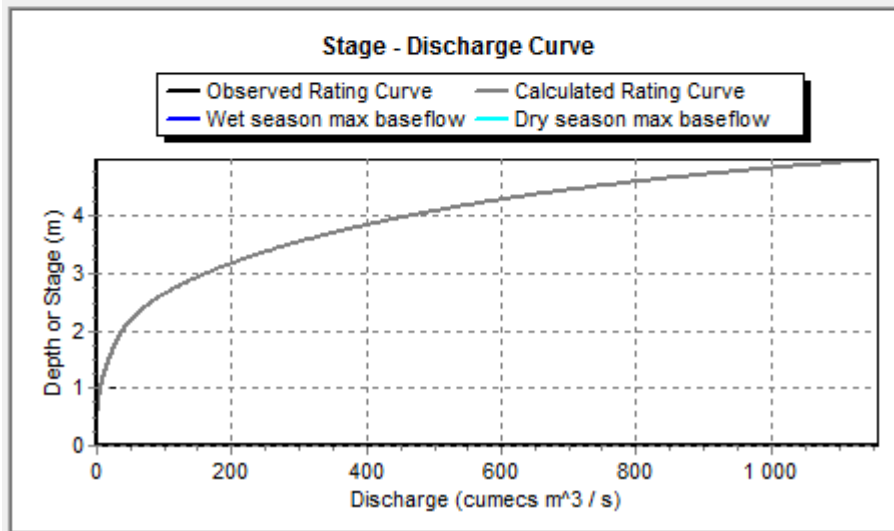
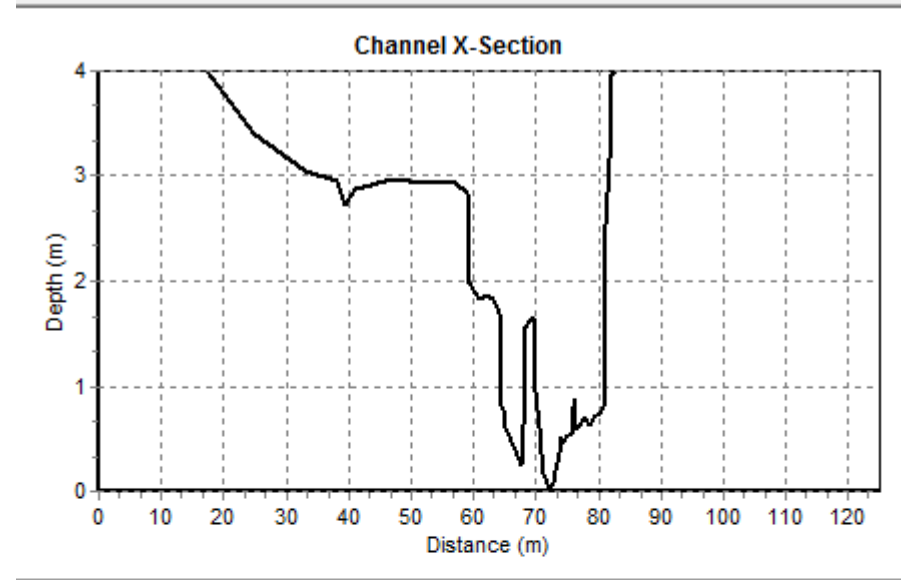
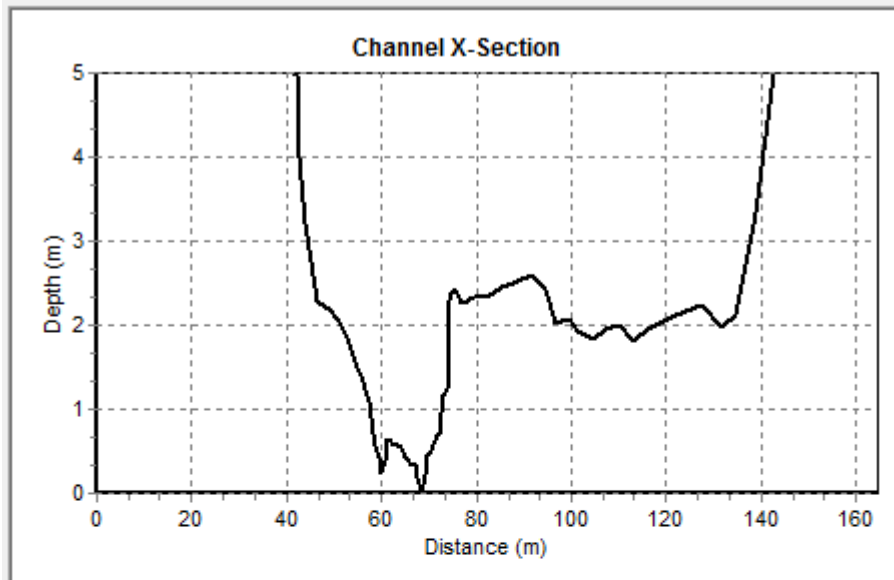
Hydraulic transect and rating curve for VT8

Hydraulic transect and rating curve for VT9



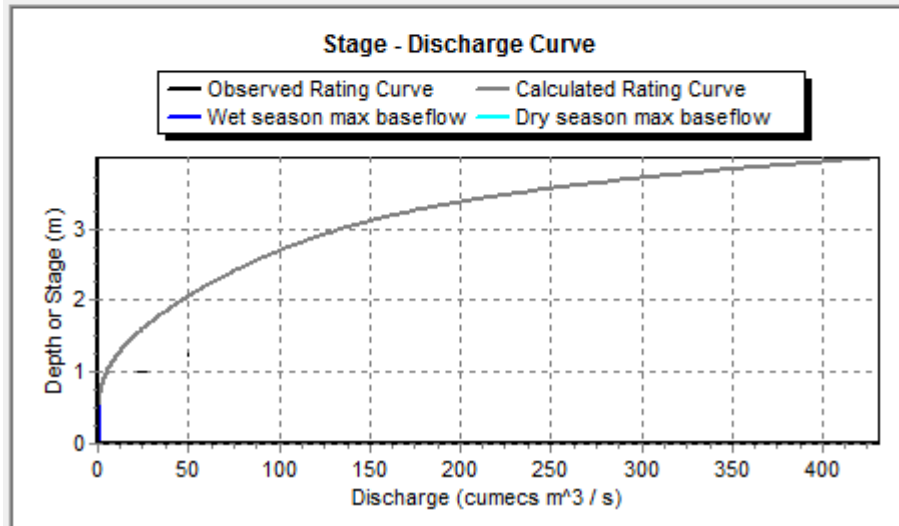
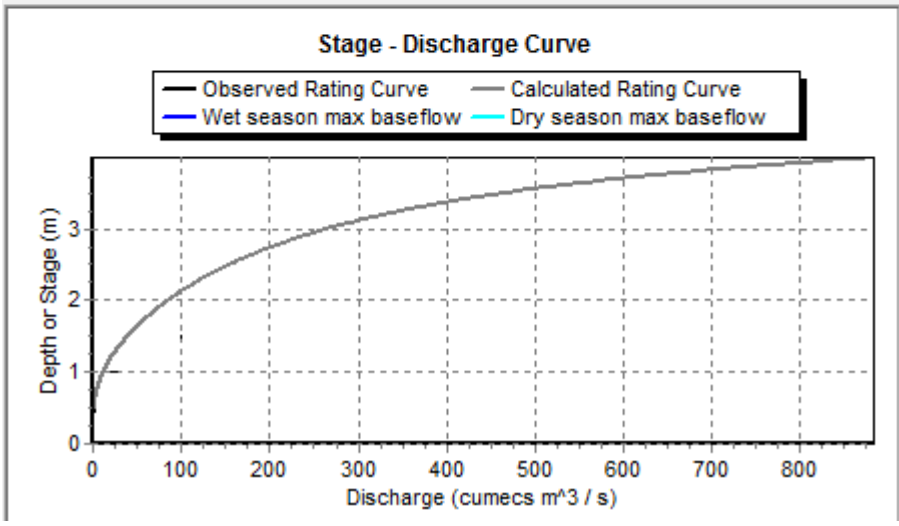
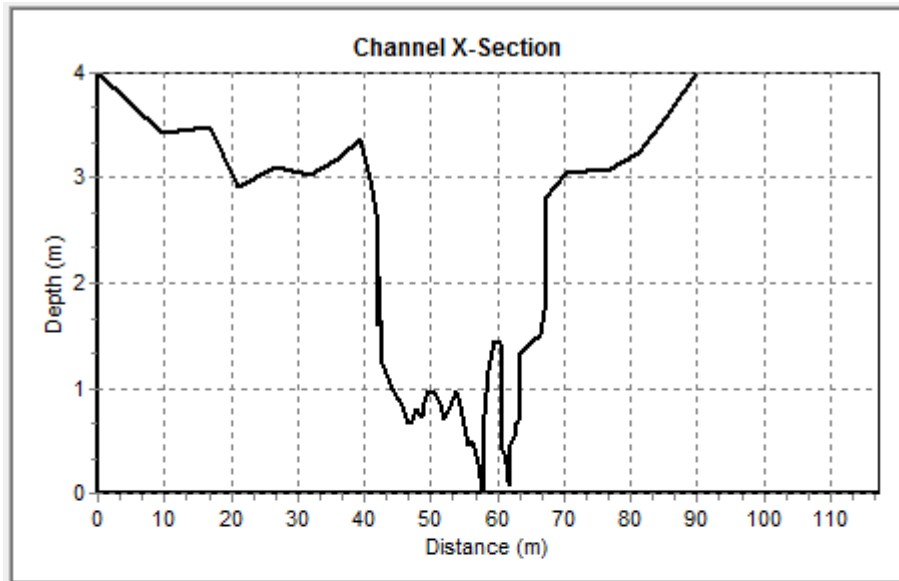
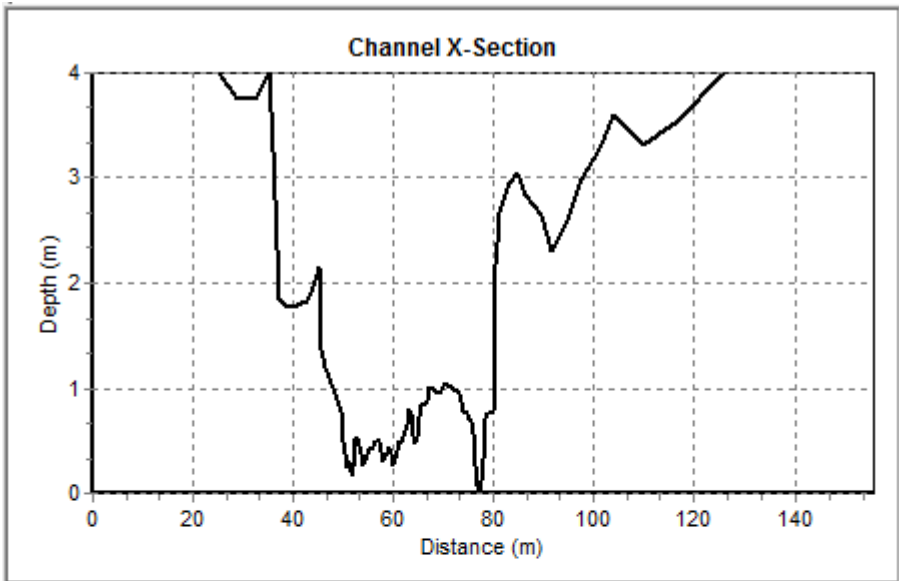
Hydraulic transect and rating curve for VT10

Hydraulic transect and rating curve for VT11



Hydraulic transect and rating curve for VT12

Hydraulic transect and rating curve for VT13



Hydraulic transect and rating curve for VT14

Hydraulic transect and rating curve for VT15

Appendix 5: Modelled inundations frequencies for VT2-VT14.

Transect					Lower flood bench			Higher flood bench		
	Stage m		Q (m ³ s ⁻¹)		Freq. per year			Recurrence interval (years)		
	Lower FB	Higher FB	Lower FB	Higher FB	2	4	6	2	4	6
2	1	2.8	8	120	4.6	4.5	4.5	2.6	6.6	8.1
3	0.9	1.6	5	15	4.6	4.6	4.58	0.22	0.23	0.25
4	1	2	8.5	53	4.5	4.5	4.5	0.6	1	1.5
5	1.1	1.5	9	22	4.5	4.5	4.5	0.25	0.3	0.35
6	1.3	1.8	12	25	4.5	4.4	4.4	0.27	0.3	0.42
7	1	2.5	10	97	4.5	4.5	4.5	1.6	3.2	6.6
8	0.9	2.3	7	72	4.6	4.6	4.6	1	1.7	2.9
9	1.1	2	12	52	4.5	4.4	4.4	0.6	1	1.4
10	1.4	2.4	40	167	2.4	1.6	1.1	6.6	9.1	14.6
11	1.2	2	19	62	4.2	3.8	3.4	0.8	1.3	2.1
12	1.2	2.4	12	82	4.5	4.4	4.4	1.2	2.4	3.5
13	1.7	2.8	31	109	3.2	2.3	1.7	2.4	5.2	6.6
14	1.2	2.1	24	115	3.7	3.2	2.5	2.4	6	8.1
Average	1.2	2.2	15	76	4.2	3.9	3.8	1.6	3	4.3