ALLUVIAL SEDIMENTOLOGY AND GEOMORPHOLOGY OF A CUT AND FILL SEQUENCE ON THE DASSIEBOSKLOOF RIVER, WUPPERTAL, SOUTHWESTERN CAPE

ARABEL C. SHAW

Department of Environmental and Geographical Science

UCT

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ABSTRACT

The unique and well preserved fluvial sequences within the valleys of the Cederberg Mountains of the Western Cape, South Africa, provide the opportunity for palaeoenvironmental reconstruction which will add to the sparse evidence from across southern Africa. The location of the study site at the edge of two physio-climatic zones provides insight into the controversy surrounding the characteristics of Late Quaternary environments in this semi-arid winter rainfall region. It is already accepted that the southwestern Cape behaved out of phase with the rest of southern Africa during Quaternary climatic fluctuations and resulted in significantly different environmental conditions. The discovery of an extensive sedimentary sequence south of Wuppertal in the northern Cederberg on the Dassieboskloof River provides colluvium package overlain by a laminated and interbedded fluvial sedimentary package approximately eight metres thick, demonstrating a change in conditions from deposition to incision (cut-and-fill). Physical, geochemical and mineralogical analysis is used to characterise the sediments as well as mapping and ground truthing of the contemporary fluvial dynamics across a portion of the Moordenaarsgat River and eastward draining Tra-Tra River; a detailed investigation of the depositional environment will expand the palaeoenvironmental record for the region, and the two adjacent climatic zones (Cederberg and arid Karoo). This research will contribute to a more detailed understanding of geomorphology in the northern Cederberg and the driving mechanisms, which are extremely significant when assessing and managing potential future aridification and its impact on this fragile landscape.

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Chapter 1
INTRODUCTION

1.1. INTRODUCTION

The Quaternary, and especially the Late Quaternary, is of great significance worldwide, as major global fluctuations in the climate occurred within relatively short time periods. It is imperative to understand that, in order to facilitate management of vulnerable environments in a sustainable fashion for the present and future, one requires a more comprehensive knowledge of these fluctuations in climate and associated environments' bio-physical reactions that brought about the environmental changes during the Quaternary. Landscape, people and climate are three variables that are inextricably linked, and an understanding of the course of recent environmental changes requires an analysis, not only of the elements themselves, but also of the way in which each influences the other (Bell & Walker, 1996).

Palaeoreconstructions are an invaluable tool for understanding current global climate change (Lioubimtseva, 2004). Quaternary climatic changes in southern Africa have had a substantial influence over a wide spectrum of variables, including landform development. Dryland landscapes and sediments can be interpreted and used to establish and refine local sequences of climate change within a global context of Quaternary climate change (Rendell, 1997). Pronounced changes in arid climates in the recent geological past are well documented for many arid regions of the world. The climate of present-day arid and semi-arid regions is known to have changed at various temporal and spatial scales. It has been documented that Pleistocene climate variations had a marked effect on many presently arid zones; however, palaeoreconstructions provide a key to understanding the present trends in arid and semi-arid climates (Lioubimtseva, 2004). Landscapes have been shaped by past climates and have essentially persisted through burial or subsequent subaerial processes that fail to modify them, leaving them as effectively fossil remnants. Arguably, it is these semi-arid landscapes that bear the imprint of Quaternary changes more strongly than others (Meadows, 2001). The responsiveness of geomorphological features is therefore fundamentally evidence in itself of environmental change. The sedimentary record offers a further, more detailed impression of events during the Quaternary and an insight into palaeoenvironmental conditions. Sediments may be related to contemporary observations and depositional environments preserved in the stratigraphic record; it is possible to assess lithological changes as the sediment sequences reflect long-term accumulations over extended periods of time, therefore exacting a broader understanding of both spatial and temporal aspects of environmental change and providing palaeoenvironmental reconstruction opportunities. These accumulating sediments are also potentially significant archives of palaeoenvironmental information relating to changes in precipitation and runoff patterns during the Late Quaternary. The palaeomoisture variability is believed to be directly linked to regional climatic change and hence understanding the geomorphic system response to past changes may assist in the future management of marginal and sensitive landscapes as the position of South Africa, latitudinally, is such that much of the country is vulnerable to small scale climatic changes (Preston-Whyte & Tyson, 1993).

Evidence for Quaternary environmental change relies on various methods to aid environmental reconstructions for the period in question, including geomorphological methods where the scale-dependent principle of climate change affecting change through processes over a longer time period is applied and observed (Meadows, 2001). Quaternary palaeoenvironmental reconstructions are important in illustrating change to understand how climatic changes, coupled with ongoing geological and geomorphological variations, affected the Earth's systems in the more recent past and give an indication of how it might respond in the future to shifts or abrupt changes in climate forcing particularly in light of the growing concerns surrounding global warming and the "drying" trend in southern Africa (Meadows & Baxter, 1999). Palaeoclimatic interpretations of landforms, with emphasis on geomorphic processes as well as stratigraphic studies of deposits, aid both the regional and local establishment of the response of arid landforms to millennia and decadal time-scale climatic change.

Of the African continent, approximately 60 percent can be classified within the realm of aridity (Thomas, 1997). The contemporary environments of all drylands in Africa reflect both a long history of geologic and geomorphic evolution, in particular, the effects of Quaternary climate changes, where the origins of modern aridity in parts are the result of the development of the present-day global climatic patterns during the late Tertiary (Lancaster, 1996). The term 'landscape evolution' essentially refers to the description of surficial terrain modification through geomorphic adjustment in response to fundamentally climate-driven processes. It is perhaps better in the context of the study to rather refer to short-term modification and response of landscape, in particular of changes from aggradation/equilibrium to degradation and return to equilibrium. A record of change can be read in both erosional landforms and sequences of sedimentary deposits, within both the climatic context and that of differing potentials of sediment generation and supply and disruption in the sequence (Rendell, 1997). The more arid regions of southern Africa have all, to some extent, been impacted, although with spatial variability, by Quaternary climatic changes, with particular emphasis being placed on changes and associated resultant effects of aeolian (wind related) and fluvial (river) processes and landforms. The interaction of wind and water over time is considered as one of the key influences upon contemporary geomorphology and longer-term landscape development in more arid regions of the world (Nash, 2000). The trend within these arid systems is that geomorphic response varies, such that aeolian systems generally accentuate the effects of wet and dry phases in comparison to fluvial systems, whilst the amplitude of the climate change in question also varies, although desert margins appear more dramatically affected (Lancaster, 1996). African drylands have played an important role in the development of paradigms in research, especially with regard to processes and landforms; with the Kalahari Desert being one of the most well-known and researched areas of aridity (where drylands refers to regions characterised as semi-arid, arid and hyper-arid, although within arid regions controlling regimes such as temperature, seasonality and moisture availability can vary considerably making classification and delimitations of arid environments difficult (Thomas, 1997)).

The southwestern Cape is an environmentally heterogeneous and floristically biodiverse region of South Africa. The area experiences mainly winter rainfall and is an important southern hemisphere example of a Mediterranean-type climate. Proxy palaeoenvironmental data are crucial to understanding the climatic changes that have occurred within this region during the Late Quaternary period (Barrable *et al.*, 2002). Palaeoecological evidence suggests marked changes in the climate of the winter rainfall region over the last 25 000 years, although, as the environmental evidence for this period is relatively scarce, proxy data is often included to aid in interpretation and further elucidation. Reconstruction of environmental change over the last 20 000 years has involved analysis at a range of spatial and temporal scales within a chronological framework of variable precision. For the most part, the southern African region has experienced change consistent with the hemispheric and global experience (Tyson *et al.*, 2001). The contemporary uniqueness of the southwestern Cape environment indicates the possibility that it may have responded distinctively to Late Quaternary perturbations (Meadows & Baxter, 1999). Although a wide variety of palaeoenvironmental evidence has been examined, a comprehensive and widely accepted view of the development of its environments is lacking due primarily to a paucity in long, continuous records and a sufficiency of new investigated sites. The Cape Floristic Region's prolific plant diversity is concentrated in the southwestern Cape mountains and upland sites, although rare, provide a potential wealth of accumulation into the Late Quaternary, chiefly the Holocene. From this, the Cederberg, a mountain range on the fringe of the winter-rainfall region and potentially more sensitive to fluctuations in conditions was chosen for investigation. The Cederberg contains a rich supply of Late Quaternary information and previous along with ongoing studies reveal a propensity for evidence within the region to indicate more subtle environmental changes (Meadows & Baxter, 1999).

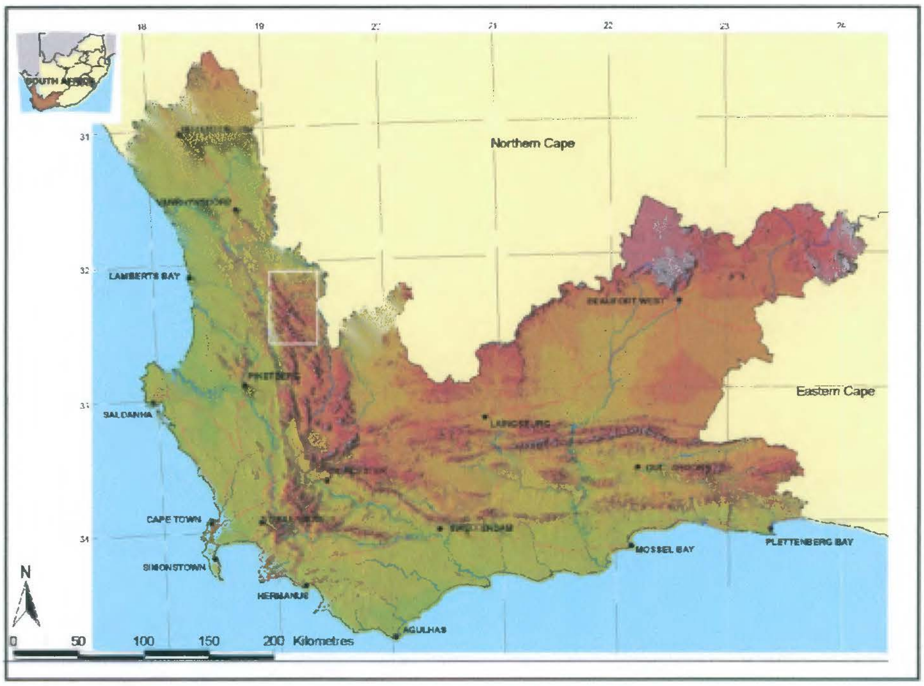


Figure 1.1: The location of the Cederberg mountain range (white block) at the northwestern extremity of the Cape Fold Belt in the Western Cape, with relation to South Africa (DEAT, 2000)

1.2. PRELIMINARY SITE DESCRIPTION

The Cederberg mountain range of the Western Cape (between 32° 00'S and 32° 45'S and 18° 50'E and 19° 25'E) forms the northwestern extremity to the arc of the Cape Fold Belt Mountains that run parallel to the coast of the southwestern Cape, as seen in Figure 1.1. Located approximately 250km north of Cape Town, the Cederberg falls close to the northern limits of the winter rainfall region and fynbos biome and is recognised as a more xeric, especially to the north, environment. Geomorphologically the range is complex with high local relief of a number of lofty peaks separated by linear valleys (Meadows & Sugden, 1991). Late Quaternary palaeoenvironmental reconstructions from organic valley-bottom sediments and accumulations of *Hyrax* midden material have already been examined and are currently underway in the Cederberg region, although there has been controversy surrounding the characteristics of the Late Quaternary environments of this semi-arid region due to varying geographical localities and palaeoecological methods employed (Meadows & Holmes, 2000). Geomorphic and sedimentary features are themselves good indicators of environmental change (Lowe & Walker, 1998), thus arid and semi-arid zone depositional settings are recognised as preserving evidence indicating the influence of abrupt global and local climate events and emphasising the punctuated nature of some arid zone environmental components (Stokes, 1997). It has already been established that a synthesis of all palaeoclimatic evidence from sites in the southwestern Cape suggests a coincidence of warmer temperatures with drier conditions and cooler with wetter conditions during the Late Quaternary, in contrast to cooler temperatures and increased aridity experienced by the rest of the country during this period (Meadows & Baxter, 1999). This creates greater interest for a study along a marginal and already semi-arid area, especially as it borders the 'anomalous' winter rainfall region (Mulock-Houwer, 2001).

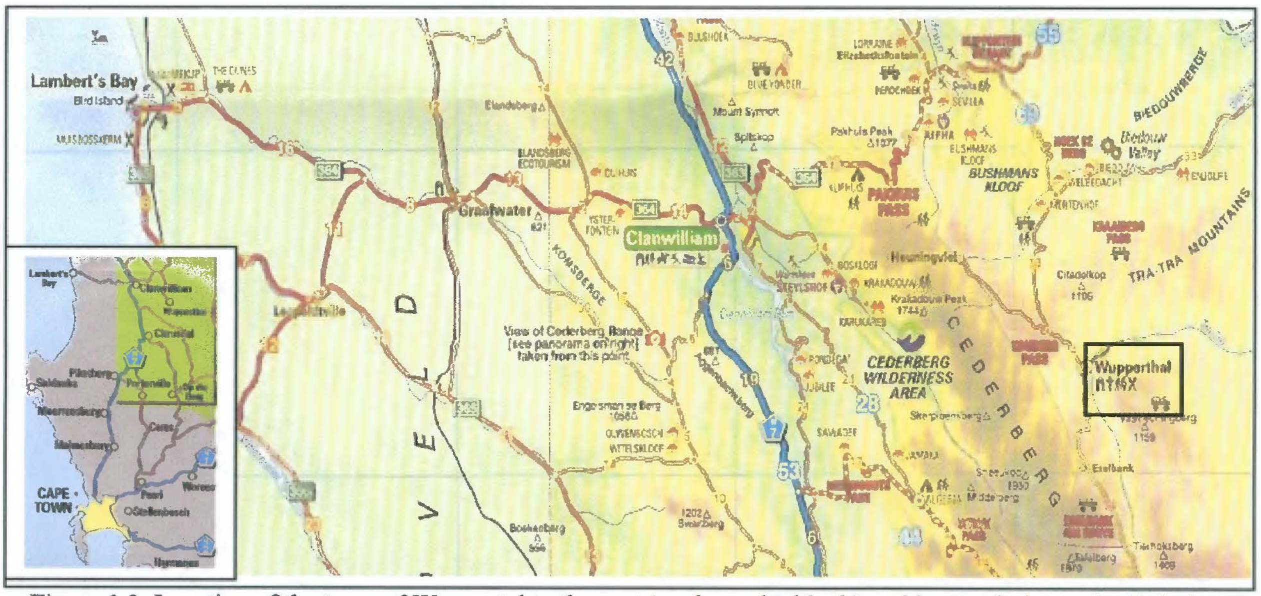


Figure 1.2: Location of the town of Wuppertal to the east (as shown by block) and its proximity to the Cederberg Mountains (1:500 000)

The study location of this project lies in the northeastern Cederberg in the Wuppertal Conservancy (Figure 1.2). Underlying geology shifts from Table Mountain Group sandstones of the Cederberg in the west, which support a variety of fynbos communities, to the more nutrient-rich shales of the Bokkeveld and associated karroid communities to the east. Climate and rainfall also varies across this region, with increasing aridity with progression eastwards towards the Tankwa and Ceres Karoo, the most arid regions in South Africa. Figure 1.3 shows the study area comprises of two rivers, Moordenaarsgat and Dassieboskloof, and their confluence to create a third, the Tra-Tra River, within bedrock constrained valleys emerging on the eastern border of the Cederberg. The valleys exhibit a contained depositional system with a now distinctly visible actively eroded palaeolandform exposed just to the west of the confluence, as the Dassieboskloof River exits a bedrock canyon.

Figure 1.3: Southerly view towards the study site, Vaalheuningberg and Moordenaarsgat River. The primary exposure is evident as a grey package in the centre of the photograph.

1.3. RATIONALE BEHIND STUDY

Environmental changes in southern Africa have occurred over a wide range of timespans and spatial scales (Meadows, 1988). The concept of change through time is one of the most fundamental factors in geomorphology, which not only involves description but also the exposition and explanation of contemporary landforms' histories, and analysis of landform development against the background of environmental change and variability. Small climatic variability within the realm of regular cyclic events, such as phases of humidity and aridity, have been included into longer period variations of the order of hundreds of years within climates and these have in turn been superimposed onto relatively major fluctuations with intervals of tens of thousands of years, such as glacials and interglacials. Along with these changes, there are also associated or sudden events of greater magnitudes that occur interspersed throughout the record or over intervals of millions of years, or rather on a geological time scale. Landform study in relation to Quaternary environmental change, specifically the Holocene, is a valuable tool towards interpreting evidence of such environmental change and consideration of their implications, both as shown in the past and possibly for the future.

Fluvial systems, in particular the processes and landforms, although very dynamic, are useful aids in reconstruction of former environments, and hence, climate, through thorough analysis and comparison with other forms of evidence, particularly for the Late Quaternary where dating can be effectively employed. Almost all Quaternary sediments contain important indications about their mode of transport and often about the climate regimes and environmental and physical conditions under which the sediments accumulated. This important information can be extracted from the stratigraphic record using various different chemical and physical methods, obtaining environmental data that can in turn be compared with analogous sedimentological processes observed in contemporary environments (Lowe & Walker, 1997).

The Mediterranean climate (winter rainfall, summer aridity) of the region is believed to be probably only of Quaternary or at most Tertiary origin, therefore greatly varied climate fluctuations have led to significant environmental changes during the last 2.5 million years, in line with the generally accepted phenomenon of Quaternary climatic fluctuations across southern Africa, as shown by the high levels of specific endemism (75% today) brought about by prolific speciation within the Cape Floristic Kingdom (Meadows & Sugden, 1993). As noted, the southwestern Cape represents a significant southern hemisphere example of a Mediterranean climate, particularly being so floristically and geomorphologically diverse, and hence, tantalisingly holds the potential to contain a wealth of information for an environment equally diverse.

The region has been shown not to respond uniformly to Late Quaternary changes, as expected due to the southwestern Cape's marked regional differences in rainfall, to a lesser degree seasonality, etc. A coincidence of warmer temperatures with drier conditions over the Late Quaternary and vice versa for the winter-rainfall region contrasts with much of the rest of southern Africa with respect to moisture availability; inferring this fact to the future suggests interesting implications for future global warming predictions and effects within the region and its response (Meadows & Baxter, 1999). Climatic changes revealed by palaeoenvironmental data thus far shows a framework of interactions between various components of the environment as well as the climate system, therefore studies of the Late Quaternary are fundamental to past environmental changes (Mulock-Houwer, 2001). Reconstructions have been hampered by the scarcity, and in some cases meagreness, of evidence and may be over simplistic as a result considering the winter-rainfall region consists of such regional differentiation.

The site is particularly applicable as high altitude valley fills have the advantage of being closer to the original sediment source and are consequently more likely to reflect the nature of the processes which were responsible for their accumulation. Theses include fluvial, colluvial, periglacial and pedogenic processes. The farther the sediments are transported from their source the more likely they are to undergo mixing with sediments derived from other sources and contaminated by leachates that complicate palaeoenvironmental interpretation (Holmes, 1998).

The discovery of an exposed section of horizons consisting of grey and orange packages as well as talus debris and distinctively divided into alluvium and underlying colluvium members, in the northeastern Cederberg, provides an excellent opportunity to use geomorphological and sedimentological evidence for Quaternary palaeogeomorphological environmental reconstructions within the realm of the winter-rainfall region and its margins. This is particularly valuable, as it is already suggested, based on mostly palaeoecological data, that this region acted out of phase with the rest of southern Africa with regards to climatic trends during the Quaternary. Furthermore, various syntheses of palaeoenvironmental conditions in the southwestern Cape have been made using primarily ecological evidence, namely palynology, fossils (micro and macro fossils), archaeology, dendrochronolgy and charcoal analysis; these include Deacon *et al.* (1983), Deacon and Lancaster (1988), Partridge *et al.* (1990), Meadows and Sugden (1991), Meadows and Baxter (1999) and Scott and Vogel (2000), however, very little besides an honours project done in the northern Cederberg (Cornell, 2001) deals with sedimentary and fluvial palaeoenvironmental evidence. With regards to future inferences, a decrease in moisture availability accompanying an increase in temperatures in a region already regarded as semi-arid to arid could be dire in light of global warming predictions, therefore understanding Quaternary and more recent Holocene changes is invaluable to such a vulnerable area. Comprehension of a small localised system, such as this, is also important in the wider context of understanding more complex systems on larger scales, even towards geohydrology, oil reservoir and mineral exploration. One must also consider the importance relative to the local inhabitants of the area in terms of using a study, such as this, to make inferences for the future taking into account recent past and planned changes in land use, change in agricultural and grazing intensities, water abstraction, dam construction, increased alien vegetation, changes in natural vegetation patterns and fire frequency.

1.4. AIMS AND OBJECTIVES

This study is based on the extraction of data from geomorphological evidence and stratigraphic records (lithological evidence), and considers both the spatial extent of the fluvial dynamics as well as the temporal scale of the sites and region as a whole. It also further examines the climate/process relationship in terms of individual geomorphic feature formation and development. In attempting to understand the development and chronology of arid and semi-arid landscapes, it is necessary to consider the evidence of both constructional and depositional and, where possible, erosional or degradational features (Stokes, 1997). During the Late Quaternary the influence of atmospheric forcing has been recognised as a primary control on environmental change over the subcontinent (Tyson, 2000) and thus has been manifested by a variety of geomorphic features within the landscape of southern Africa. It is also necessary to acknowledge more complex multi-causal explanations in which local interactions between environmental context, climate fluctuations and human impact are envisaged as appropriate explanations for landscape change.

This project aims to outline one of these features, the exposed section of sedimentary packages in the western headwater vicinity of the Tra-Tra River, a tributary of the Tankwa and eventually Doring and Olifants Rivers, and distinguish from its inherent evidence as a proxy an interpretation of conditions and change during its existence. The northern Cederberg is one of the more arid regions of the fynbos biome and this study is an attempt to establish the chronology and palaeoenvironmental indications/implications for interpretation of the sequence of exposed an alluvial and colluvial deposit along the Dassieboskloof River, southwest of Wuppertal. Hence, the primary concern of the investigation is the depositional environment along with subsequent erosion and its correlation with palaeoclimatic fluctuations in the southwestern Cape, or possible changes in local physical parameters. In an attempt to understand those geomorphic processes which have influenced the landscape of the study area during the Late Quaternary, analysis is carried out through applying sedimentological, dating and micro-geomorphological techniques.

This study was approached as a synopsis of current fluvial environments in comparison to palaeoenvironments. A secondary aim, in terms of geomorphic modification, is to elucidate or partially ascertain the forcing or controlling mechanisms at work within the system and responsible for changes from cut to fill, or vice-versa, within the valley sediments, and to possibly recognise any cyclical patterns with regard to such changes in state within the geomorphic environment. Contemporary fluvial conditions of the converging rivers in the area were also identified using a series of physical and chemical methods, surveying, mapping and ground-truthing.

In view of the more general aims of this study, the following are specific objectives:

- To review research and syntheses of past environmental changes of the Late Quaternary in the winter-rainfall region of South Africa, specifically since the last glacial maximum, and relate these findings to the significance of geomorphic response and evidence within systems.
- To examine the morphology and lithostratigraphy and present a facies assemblage as well as determine a chronology of an exposed section of a valley fill in the northeastern Cederberg at Wuppertal.
- To assess contemporary geomorphic processes and fluvial conditions along a limited section of the long profile of the Dassieboskloof, Moordenaarsgat and Tra-Tra Rivers.
- To attempt a palaeoenvironmental interpretation of the geomorphic evidence from the study area and infer, if possible, climatic fluctuations during the study's extent.
- To place the findings of the study within the broader context of palaeoenvironmental research for the winter-rainfall region of the southwestern Cape.
- To extrapolate key findings of the research to the potential aridification within the region in a future context.

The main objective of the study is to provide a palaeoenvironmental reconstruction based on sedimentary proxy evidence for the northeastern Cederberg margin with the Tankwa Karoo and to use this "palaeoperspective" to infer its significance into a future context of aridification in the region bearing in mind land-use changes affecting not only groundwater but also leading to decreased vegetation and increased runoff. This project endeavours to contextualise and understand the underlying dynamic shift in the environmental conditions within the physical setting of the Cederberg by studying a prime site within a localised fluvial system and then placing it within the macro-geomorphology of a greater regional aspect along two major rivers. A multi-pronged approach was adopted towards this project to ensure a higher degree of accuracy and more coherent interpretation of results. This research study aimed to simultaneously provide important environmental information for the southern African region and training and application of techniques not previously employed within the region as well as evaluate the significance or success of such geomorphological evidence in palaeoenvironmental interpretation within this region of the southwestern Cape.

1.5. LIMITATIONS

The Quaternary contains a rich basis for data, especially the Holocene, however, the geographical spread of analysed sites used in Quaternary palaeoenvironmental reconstruction across southern Africa is spatially inconsistent and evidence is distributed very unevenly. Southern Africa itself is extremely diverse climatically and geomorphologically and establishing a coherent view across such a wide range for Quaternary environmental change is problematic. As the region is characterised by extreme environmental heterogeneity with highly variable annual amounts, temporal and spatial distributions of precipitation coupled with high levels of biodiversity, further limitations to palaeoenvironmental elucidation are created. Conflict arises between different types of evidence and across vast areas, such as biological versus physical evidence. There is an issue with the difficulty arising from dating controls and chronology beyond radiocarbon dating and the newly utilised Optically Stimulated Luminescence (OSL) dating method, as geomorphically active or relict features (requiring a more thorough understanding of environmental conditions conducive to their formation) are problematic to date, which is further complicated by the scarcity of long continuous and viable preservation of records in arid environments that are prone to erosive processes and non-preservation of indicators (Thomas, 1997). Within semi-arid and arid regions, such as the site location, environmental conditions preclude the preservation of datable organic material leading to the omission of ^{14}C dating techniques in the study.

Geomorphological, sedimentological and palaeoecological methods that are used in Quaternary palaeoenvironmental reconstructions are spatially and temporally affected and are more accurate based on multi-proxy evidence rather than only a single proxy. Assumptions and inferences are made based on limited or shorter records due to the lack of significant or comprehensive records for specific regions resulting in an extrapolation of data. However, the response of these dryland geomorphic systems to climate change occurs over a variety of timescales and short-term changes are spatially variable and largely a result of differences in lithology and vegetation (Lancaster, 1998). Pollen analysis, although useful, has proved restrictive in elucidation of Late Quaternary environments due to the paucity of organic-rich deposits and the highly seasonal nature of the Mediterranean-type climate which entails that sediments do not appear to accumulate for any consistently long period of time during the Quaternary (Meadows & Holmes, 2000). Within the study region, pollen analysis is generally not reliable or informative enough for the period and region as localised arid conditions result in poor pollen preservation, which otherwise occurs in wetlands or vlei sediments, particularly at high altitudes and in valley bottom accumulations (Meadows & Holmes, 2000). Palaeoenvironmental proxies used in arid environments usually include lacustrine sediments, marine sediments, fluvial and aeolian processes, sedimentological, geochemical and isotopic analysis; all providing useful information towards many aspects of climate variability from a decadal to millennial scale accurately demonstrating natural fluctuations. This is important as they document changes lacking anthropogenic forcing. Although palaeoclimate records across southern Africa are sparse in time and space, Holocene climates and environments varied on scales that carried significant implications for distributions of fauna and flora, as well as human habitation. Natural variations of light stable isotopes, particularly ^{13}C, can play a vital role as a tool in spatially and temporally affected palaeoenvironmental conditions and inferences as proxy evidence, especially when coupled with other forms of evidence to provide a more comprehensive and accurate multi-proxy reconstruction.

A vital limitation to bear in mind when analysing sedimentary evidence is that it is seldom properly evaluated and representative in isolation and best applied in conjunction with other proxy evidence or integrated into further geomorphological evidence to provide a conception of landscape or climatic change (Lowe & Walker, 1997). A further major constraint to consider is that, in reconstructing palaeoenvironments of arid and semi-arid zones from relict fluvial landforms, there are a range of non-climatic factors to take into account that in combination with known aspects, such as geology, climate, relief, soil, vegetation cover and anthropogenic influences, affect the rate of fluvial activity as well as geomorphological response. There are also numerous apparent contradictions, particularly in an area on the boundary between arid and semi-arid conditions, where factors lead to varied fluvial dynamics, such as increased sediment loads resulting from increased rainfall in arid regions as opposed to increased sediment yield as a result of reduced vegetation cover during decreased rainfall in semi-arid zones. There is potential to elucidate fluvial palaeohydrology within the setting with morphological and sedimentological evidence being strongly coupled with channel-fills and terrace sequences providing the information of past flow conditions, however, as channel flow responds to a range of controls including climatic and tectonic influences, it can prove difficult to determine palaeoclimatic signals in fluvial sediments and forms (Thomas, 1997).

Regionally, data are too few and the proxy evidence is insufficiently reliable to produce a coherent picture of regional precipitation, which is fundamental for the study considering fluvial dynamics are largely related to water deposition and erosion. Coupled with this, spatial differentiation existed across the region with respect to its response to Late Quaternary climate and precipitation changes i.e. some areas were drier whereas others were wetter. Along with the uneven distribution of studied sites in the southwestern Cape, the sites are restricted, such that, for example, palynological evidence occurs in mountain wetland sites or at coasts whereas mammal fossils in caves are in regions of the fynbos biome that receive all year rainfall, therefore compiling a coherent and representative depiction of regional conditions is challenging (Meadows & Baxter, 1999). More specifically, palaeoecological data for more xeric sites are also very limited (Meadows & Holmes, 2000).

An investigation into geomorphic palaeoenvironments is only meaningful if some sort of temporal chronology can be established, although palaeoenvironmental reconstructions in semi-arid environments are hampered by a lack of ancient, preserved sediments containing material suitable for isotopic dating (Holmes, 1998). There are issues with palaeoenvironmental records across South Africa not being adequately dated to detect short-term climatic changes (Scott *et al.*, 1995).

Quaternary palaeoenvironmental studies in South Africa derive data from a number of sources, such as stable isotopes, pollen, charcoal, sediments, palaeosols and micromammalian remains. However, there exists the potential for interpretations of the data to vary showing the importance to bear in mind that inference from such data is subject to differing interpretive limitations itself and that there are gaps and inconsistencies present within the proxy evidence (Meadows & Baxter, 1999).

1.6. STRUCTURAL OUTLINE

The broad structure of this study incorporates chapters divided according to theoretical and contextual background, research methodology and interpretation, discussion and conclusions.

Chapters one, two and three introduce the broad perspective of the project and provide a theoretical, physical and literature based backdrop. Specific aims and objectives are outlined along with the reasoning behind and general limitations or constraints for this project. Chapter two provides a more detailed account of the study area and significant factors within the region as a whole. Chapter three introduces the basis of Late Quaternary, and specifically Holocene, palaeoenvironments within the southwestern Cape as well as geomorphic response and processes applicable within the setting of the three major rivers of the Wuppertal locality and overall landscape evolution.

Chapter four outlines the research methodology, namely observations and analysis, employed during this study. Field, aerial and mapping, laboratory and statistical techniques are described and their use for this study motivated, disadvantages noted and the employed procedures are explained. Chapter five presents these results according to the examined properties.

Chapter six attempts to interpret and synthesise the results and elucidate from the findings possible scenarios for the fluvial environment of the Holocene, ranging from the depositional to erosional and brought up to present day's contemporary environment, taking into consideration the theoretical and physical processes combined with past environmental change outlined in chapter three. The final chapter comprises possible future implications and the significance of the study and draws conclusions based on the findings and discussion of the project. It also suggests the possibility of a growing potential in further geomorphological and sedimentological studies for the region.

Chapter 2
PHYSICAL SETTING OF THE STUDY AREA

2.1. INTRODUCTION

This chapter introduces the regional setting of the study area selected and briefly outlines the major environmental components active within these surroundings. In so doing, the chapter summarises the relevant physical background to the study including the location of study site, geology, geomorphology, soils, climate, vegetation, land use and relevant local history. It is crucial to the understanding of the study to bear in mind the interaction of many aspects and factors relating both to the contemporary natural environment, palaeoenvironmental change and conditions as well as anthropogenic influences and changes acting now within the vicinity that play a significant role in the physical processes and dynamism of the area. A more detailed description of the precise study sites and localised drainage of the rivers follows in Chapter 5. All elevations are quoted as metres above sea level (a.s.l.).

2.2. PHYSICAL SETTING

The study area is geographically located at the northeastern edge of the Cederberg mountain range, which lies in the northwestern corner of the Western Cape Province of South Africa (between 32°00` and 32°45` S and 18°50` and 19°25` E). The highest of the northern Cederberg peaks are Sneeuberg and Tafelberg, rising to 2027m and 1969m respectively, while the highest peak of the Cederberg exists in the south, namely Sneeukop at 2070m. The Cederberg Wilderness Area, part of Cape Nature Reserves, includes almost the entirety of the Cederberg Mountain range and lies to the west of the study location. The study area encompasses the upper headwater catchment of the Tra-Tra River, a tributary of the Doring River, which is in turn a tributary of the Olifants River; and includes with it the lower reaches of the Dassieboskloof and Moordenaarsgat rivers. The topography is rugged rocky mountains with deep flat floored valleys. The site falls to the south and east of Wuppertal, a small town nestled in the rugged wilderness, 72km southeast of Clanwilliam and approximately 250km north of Cape Town. The Wuppertal Conservancy, approximately 350km^2 of protected land, runs along the eastern border of the Cederberg reserve and includes Wuppertal. The Cederberg Conservancy lies immediately to the south and the Tankwa Karoo and National

Park in the Northern Cape further to the east of the Wuppertal Conservancy (Slingsby, 2003). Figure 2.1 shows the study area extends to roughly midway along the Tra-Tra Mountains extent and encompasses a section of the Moordenaarsgat River.

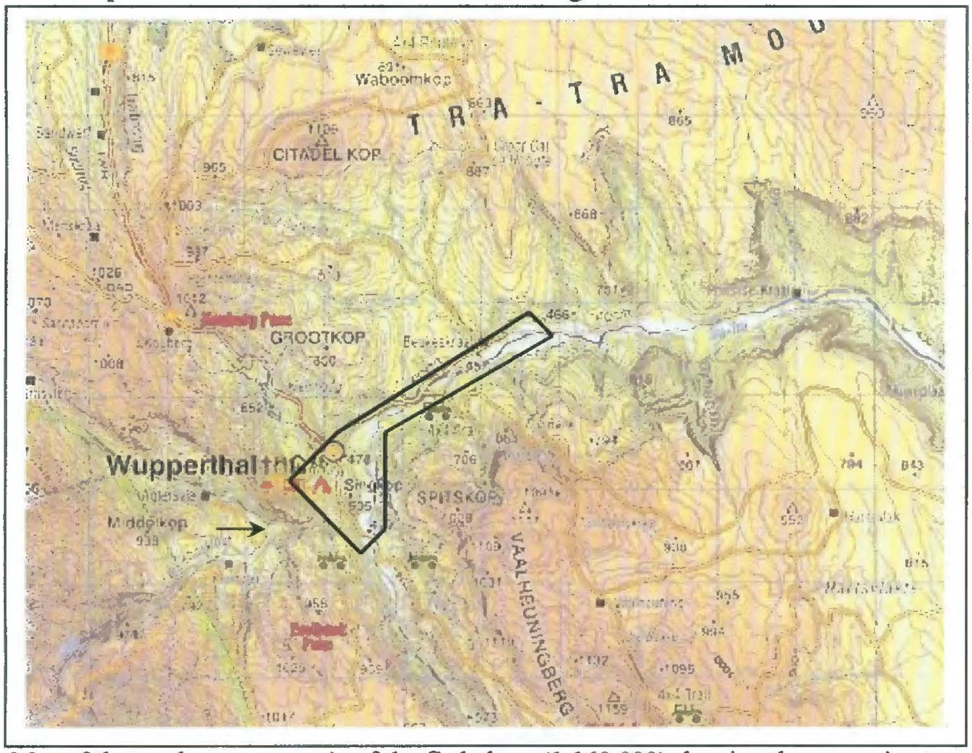


Figure 2.1: Map of the northeastern margin of the Cederberg (1:160 000) showing the approximate extent of the study area within the clearly defined Tra-Tra and Moordenaarsgat river valleys and the Dassieboskloof River (marked by arrow)

The primary site under investigation lies along the perennial Dassieboskloof River as it exits a bedrock ravine and enters the Tra-Tra River valley, approximately 1km west of its confluence with the Tra-Tra and Moordenaarsgat River. A large cutting of exposed horizons, approximately 10 meters in height from the exposed surface to the base at the edge of the river's overbank flood channel, consists of two distinct sediment packages, the upper being of loosely and partially consolidated grey silty sand sediment (alluvium) and the lower as orange harder more weathered and semi-consolidated basal clayey material (colluvium), seen in Figure 2.2. The above map displays how the study area is then extended from this site eastwards along the river valley to include the contemporary fluvial dynamics of the Tra-Tra River, a braided perennial river constrained by the Tra-Tra Mountains to the north and the Vaalheuningberge to the south, and southwards along the Moordenaarsgat River valley as far as the opening out of the ravine, roughly 8km south. The river systems and primary site straddle the transition of these two regions, crossing contrasting geology, vegetation and climatic conditions. As in nature there are no actual divides or lines designating a demarcation of regions, the gradient in climatic conditions and the transition of vegetation and varying substrates is noticeable within the study area.

Figure 2.2: The upper grey and lower orange packages of the primary site located just to the west of Wuppertal along the Dassieboskloof River

The three principal perennial rivers focused upon in this study form a "T" shape at their confluence and the central point of a series of valleys set within a region of high relief (Figure 2.3). To the south of the study area, the Moordenaarsgat River originates in the Grootberge, roughly 20km further south of the confluence with the Tra-Tra River, and flows northwards along a narrow valley out of a ravine cut between the Vaalheuningberge (1159m) and the eastern peaks of the Cederberg in the Eselbank area, with a maximum elevation of 1930m at Sneeukop. The Dassieboskloof River flows in a north-easterly direction, exiting from a bedrock canyon on the eastern edge of the Cederberg Reserve. Its tributaries drain the Skerpioensberg (1617m) to the west, another Sneeukop to the south, and the Koupoort (1628m) and Chisel (1716m) Peaks to the north. The Tra-Tra River, the resultant of these two rivers combined along with other smaller tributaries, flows eastwards along a wider flat bottomed valley between the Vaalheuningberge (south) and the more extensive Tra-Tra Mountains (north), after which the river was named, that reach an elevation of 1106m, as seen in Figure 2.1. These mountains gradually give way in an easterly direction to the incised river valleys and open plains of the Tankwa basin and Doring and Tankwa Rivers and the region known as the Tankwa Karoo. The Tra-Tra River itself converges with the northward flowing Doring River directly to the east of the valley's extremity some 30km (as the crow flies) from Wuppertal.

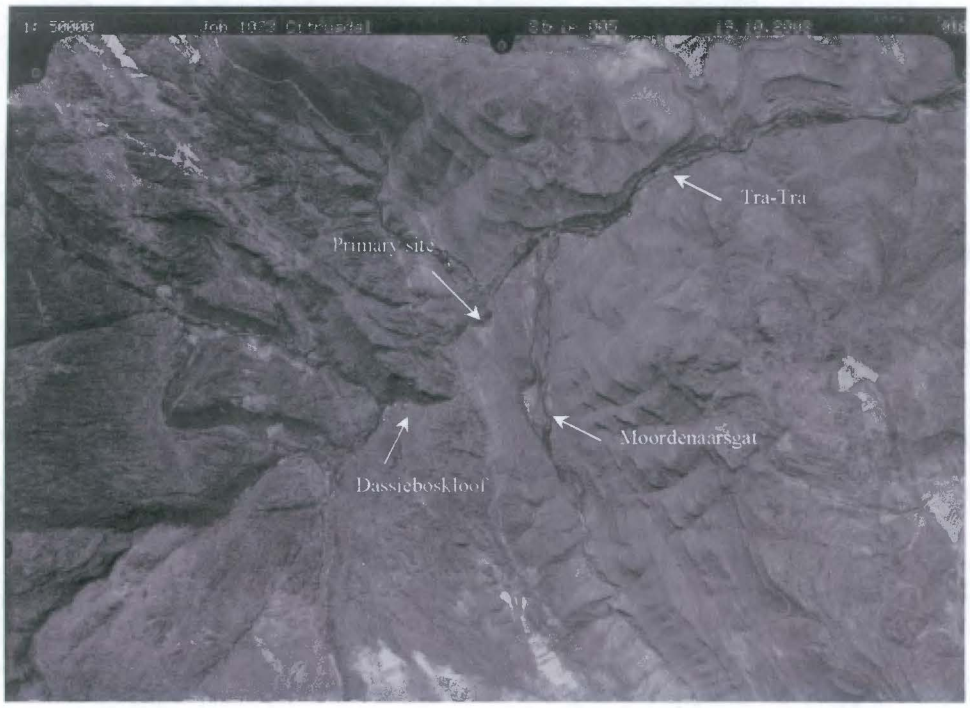


Figure 2.3: Aerial photograph showing the confluence of the Dassieboskloof and Moordenaarsgat Rivers to form the Tra-Tra River and the location of the primary site

Although the Tra-Tra and Vaalheuningberg mountains to the east of Wuppertal technically do not fall within the Cederberg Reserve and form a distinctly different environment, notable through their geology, structure, vegetation and climate, they are generally grouped together with the Cederberg Range due to their proximity, remoteness and the fact that they form the western boundary to the extensive Doring and Tankwa River plains. These open plains of the Tankwa Karoo form a large flat basin-like structure, which effectively comprises a stony desert, drained by several rivers crossing from the Roggeveld Mountains, an extension of the western limb of the Escarpment, at its eastern border, and the Klein Roggeveld Mountains to the south.

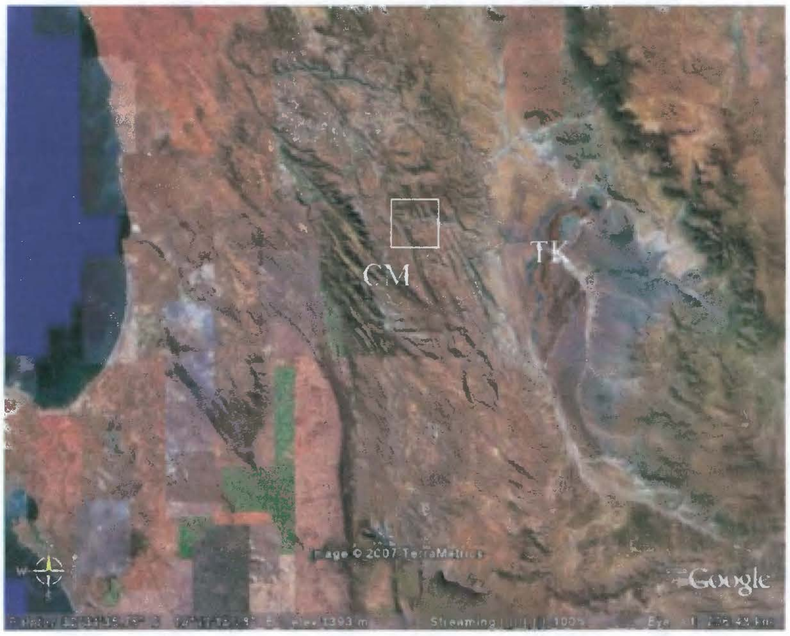


Figure 2.4: Google Earth image showing Cederberg Mountains (CM) parallel to west coast bounded by the Tankwa Karoo (TK) basin to the eastern interior (approximate location of study site shown)

2.3. GEOLOGY, GEOMORPHOLOGY AND SOILS

The underlying geology of the region is a key factor, along with other aspects, in the development, formation and appearance of numerous physical properties in the region, such as the geomorphology, soil and vegetation, although not necessarily the only determinant or control. The Western Cape has a varied and interesting geological nature incorporating igneous, metamorphic and sedimentary rocks ranging from the Late Precambrian to recent and contemporary processes of formation (Reid *et al.*, 2001). The geology of the Western Cape originates from the deformation caused by tectonic forces during the split of the southern continents comprising Gondwanaland as well as erosive, weathering and other associated factors teamed with the rock cycle (Compton, 2004). The fragmentation of Gondwanaland provided the impetus for the formation of marginal or pericratonic basins around the African coastline and the flexuring of the continental crust that resulted in the formation of the Great Escarpment (Spoenemann & Hagedorn, 2000).

The Cederberg forms the northwestern extremity of the Cape Fold Mountain Belt that runs in an arc (or L shape) at the southwestern corner of Africa and the Western Cape, mantled by the coastal plain on the outer edge. This mountain chain is comprised of erosion-resistant, quartzitic sandstone mountains alternating with plains and valleys underlain by softer shales.

The geology of the area is dominated by sedimentary rocks of the Table Mountain Group of the Cape Supergroup, which form the highest (almost north/south trending) mountain ranges. The Cape Supergroup, comprising of the Table Mountain Group and Bokkeveld Group, is widespread across the entire Western Cape and consists mainly of shale, siltstone, sandstone, grit, greywacke and conglomerate. The rocks of the Karoo Supergroup outcrop largely to the east in the Doring River catchment. This is overlain by more recent semi- to unconsolidated sediments of alluvial and aeolian origin as well as calcrete and ferricrete deposits of Tertiary/Quaternary age (DWAF, 2005). Precambrian rocks, namely Malmesbury Group and the Cape Granite Suite, provide the basement upon which these sedimentary rocks are overlain (SACS, 1980). Refer to Appendix 2 for a stratigraphic column for the Cape Supergroup.

To the north of the Biedouw Valley, a north-south fault brings the coarsening upwards shale-to-sandstone sequences of the Bokkeveld strata down into contact with the coarse grained cross-bedded sandstones of the Nardouw Subgroup. The Doring River valley itself is cut into Bokkeveld strata to the east. The Bokkeveld Group is comprised of horizontal alternating bands of resistant sandstone and more weathered shale (Reid *et al.*, 2001). These slightly eastwards dipping beds are easily apparent en route to Wuppertal along the eastern fringe of the Cederberg. Within this geological setting, topography and the Doring River tributaries are largely controlled by the Bokkeveld's horizontal strata (such as the flat-topped hills present across the plains of the Doring River valley). The Bokkeveld Group is Devonian in age and are largely, especially the lower formations, examples of coarsening upwards sequences of prograding deltaic sandstones over deltaic shales (Reid *et al.*, 2001).

The Bokkeveld Series generally lies conformably over the Table Mountain Series as sedimentary contacts, the lower contact taken to be where Table Mountain sandstone gives way to thin alternations of carbon-bearing shale, mudstone and siltstone, which is sharp at Wuppertal (SACS, 1980). The Table Mountain Group is visible in the west of the study area as quartzitic sandstone with thin shale and conglomerate lenses, such as the bedrock canyon from which the Dassieboskloof River flows, an example of the Nardouw subgroup; the Nardouw subgroup of quartzitic sandstone and minor shales extends southwards from Nieuwoudtville. Above this formation, a thin outcropping of shale, arenaceous shale, tillite, grit and conglomerate (Cederberg formation) runs horizontally below the quartzitic sandstone with minor shale and conglomerate lenses that comprises the majority of the Cederberg Range plateau and peaks. The Cederberg formation shale, siltstone and subordinate sandstone are laterally persistent and form a conspicuous grass-covered band between quartzites in the mountain range. Along the valley floor, shale and siltstone with sporadic thin limestone bands and nodules or thin sandstone banding (occasionally fossil bearing) gives way to greywacke and sandstone (Lower Bokkeveld Stage or Ceres subgroup where there is alternation of three sandstone and three shale formations characterised by their lateral continuity). The summits of the Tra-Tra and Vaalheuningberge are represented by the lower formations of the Witteberg Group, also seen as quartzitic sandstone with minor arenaceous (micaceous) shale bands, which extends along the northern flank of the Cape Fold Belt. Downslope into the Tra-Tra and eastern flank of the Moordenaarsgat valleys, two bands of sandstone with argillaceous sandstone alternate with three shale with siltstone units as part of the Upper Bokkeveld Stage (Biedouw subgroup) (SACS, 1994).

The Bokkeveld Group, as alternating mudrock and sandstone units, is wedged out northwards, extends along the Cape Fold Belt to form extensive exposures in the southwestern Cape to just south of Nieuwoudtville (SACS, 1994). The upper slopes are predominantly siltstone bands alternating with sandstone, whereas the lower slopes are predominantly shales with siltstone, sometimes fossiliferous. It is interesting to note that within the Bokkeveld Group, the Ceres subgroup has a formation named Tra-Tra (shale), and the Biedouw subgroup has a formation named Wuppertal (sandstone). The Wuppertal sandstone formation is up to 70m thick consisting of micaceous sandstone and siltstone and outcrops along the Tra-Tra Valley, while the Tra-Tra shale formation is mudstone, siltstone and subordinate sandstone 60-80m thick, in the central area, extending from northeast of Clanwilliam southwards and eastwards along Cape Fold Belt (SACS, 1994). The lower Bokkeveld succession contains marine fossils (SACS, 1980). All units are overlain and overlying conformably except for the Pakhuis formation that unconformably overlies the Bokkeveld.

There is evidence within the area surrounding Wuppertal of the effects of Carboniferous (Dwyka) glaciation from the Pakhuis ice sheet that covered most of the western southwestern Cape. This includes tillite, glacial ridges, facetted/striated pebbles and striations, evidence that is well-developed within the region and north of Clanwilliam (Reid *et al.*, 2001 and Shaw, 1997). As this area is part of the Cape Fold Belt, it has also been subjected to processes of deformation and large folds within the lower formations are evident, along the Tra-Tra valley especially where the anticline and syncline axes are obvious. There only appears to be one major fault that runs perpendicular to the Tra-Tra valley through the Tra-Tra Mountains roughly 5km to the east of Wuppertal. Approximately 1km further east, another possible smaller covered fault also crosses the valley and extends into the Vaalheuningberge.

Figure 2.5: A fold in the sandstone of the lower Bokkeveld Group and differential erosion of sandstone and shales on the slope above

In this study it is necessary to consider the primary morphological aspects of the region and its physical environment. Morphology within a landscape is to a degree the expression of processes acting upon it and in turn leads to the partial control of these processes. It is considered a controlling factor and strongly influences numerous actions, such as surface runoff and, hence, sediment transport, an important variable to take into account is this investigation. The geomorphology of the Cederberg Range is considered complex with marked local relief, from a number of high ridges and peaks separated by broad linear valleys (Meadows & Sugden, 1991). Some of these valleys contain relatively deep organic sediments, however, none were found within the study area. The geomorphology of the study site is discussed in Chapter 5 and 6.

The Dassieboskloof, Moordenaarsgat and Tra-Tra valleys appear to have been subjected to both lithological and structural control. The Tra-Tra and eastern slopes of the Moordenaarsgat valleys have been exposed to differential erosion, visible as banding down the slopes, removing the softer shales and depositing them within the valley floor and fluvial systems and leaving exposed quartzitic sandstone. Weathering of both the sandstone and shale along the valley walls, especially within the Bokkeveld group where it is more obvious, has resulted in jointing. Small-scale rock falls and mass wasting has occurred along the valleys resulting in talus slopes of large boulders and debris. The Moordenaarsgat River leaves a box canyon to the south and fans out within the broader valley to form an alluvial fan, becoming constricted at the distal end of the valley by a small Bokkeveld *koppie*, known as Singkop (606m), before its confluence with the Tra-Tra. Likewise, the Dassieboskloof River exits a rocky bedrock canyon before it crosses under a small road causeway prior to its confluence with the Tra-Tra River.

Surficial deposits, assumed to be of a Quaternary age, within the valleys and at the primary site are relatively widespread, although largely shallow in nature. They comprise predominantly of alluvial sands and gravels along the water courses. Colluvial and alluvial fills are also present, especially along areas of valley constriction and at the primary site along the Dassieboskloof River. Many of the colluvial slope deposits are partially cemented or calcretised at their base. Colluvium moving down slope has created colluvial fans, which have been truncated at the base of the talus slope by river channels. The colluvium and bedrock has in places led to channel constriction on both sides of the valley.

In this mountainous region, the lack of well-developed soil horizons and more arid physical conditions, which are less favourable to soil development, means that the soils that do occur are difficult to classify due to the exposure of the rocky terrain and variability. Since these deposits are observed to maintain their depositional stratification throughout their exposure, soil processes are presumed to be relatively weak or inactive. These alluvial surficial deposits are largely azonal with poorly formed apedal structureless or massive beds with a shallow layer of sparsely vegetated sheetwash material comprising some root networks and organics on the exposed surface. Typically these are referred to as lithosols as they occur on slopes and display permanently immature or truncated profiles due to erosion or deposition and are present widely through the Western Cape (Holmes, 1998). However, in this study, they are referred to throughout as alluvium. Colluvial exposures sourced from the shales are better developed as soil profiles showing the presence of soil formation processes. The colluvium exhibits columnar, prismatic and blocky structures, particularly well developed in places with increasing thickness of the profile. Although the beds show ped structures within, the stratigraphy is well maintained and it proves difficult to subdivide the profile into recognisable soil horizons beyond a generalised description.

According to the FAO soil classes (Greiff & Ritter, 2002), the study region is broadly categorised as L Luvisols. These are typically soils with an argillic B horizon of medium to high base status and is commonly split into high-activity clays L Luvisols and low-activity clays E Lixisols (where E refers to the eluvial horizon that is leached and bleached) (Briggs *et al.*, 1998).

Significant geomorphic features and dynamic equilibria are apparent within the study area. This is demonstrated at the main site, where a period of deposition shifted into a period of erosion and incising of the river to result in a noteworthy cutting or terrace along the river course. Along the valley, down the long profile of the rivers, incised micro-channels undercut the macro-channels that meander across the floodplain within the valley while in places there are networks of anastomosing channels. Other fluvial features observed include raised flood channels consisting of cobble lags, pebble and cobble bars, and slackwater deposits.

2.4. CLIMATE

The study area falls within the northern limits of the southwestern Cape region characterised by the so-called Mediterranean type climate. Mediterranean climates account for almost 12% of the Earth's land mass but with many and large variations within this overall framework (Le Houreau, 2004). This climate is described as having hot and dry summers with cool and wet winters. This is a unique climate type, especially within the context of the rest of the southern African region that is characterised by summer rainfall and dry cold winters. An important determinant of the weather and climate of southern Africa, particularly southern South Africa, is the cold Benguela current that runs up the west coast with its associated upwelling regime and the cold front (mid-latitude cyclone) and its associated westerly waves, depressions and cut-off lows, which produce cold snaps and are responsible for most of the winter rainfall, (80% in the southwest region) (Preston-Whyte & Tyson, 1993). These frontal progressions result in short intense periods of precipitation with peaks between June and August The climate in the Cederberg and its immediate surrounds is also classified as Mediterranean (although only within reach of larger frontal systems) with definite winter rainfall between June and August with occasional snowfalls.

With regard to rainfall, South Africa is roughly divided according to the 1000mm/yr isohyet that runs generally across the country from the northeast to the southwest. The southwestern Cape falls to the west of this isohyet and therefore, receives on average, less than 1000mm precipitation per annum (in places, such as the more arid west coast and interior, considerably less) (Preston-Whyte & Tyson, 1993), although there are marked exceptions largely related to relief and elevation. Rainfall is very variable within the Cederberg itself, depending on altitude and aspect, with an annual average between 500 and 1000mm (Meadows & Sugden, 1991). The region is characterised by a semi-permanent high pressure zone with a relatively dry mid- to upper-troposphere with the majority of moisture being sourced from the surface (Barrable *et al.*, 2002). However, the area immediately surrounding Wuppertal, on the northeastern extremity of the mountain range at its margin with the Tankwa Karoo, one of the most notably arid regions in South Africa, receives between 200 and 300mm of winter rainfall and little over 5mm during the hot and dry summer months, as averaged since 1900 based on rainfall data from the South African Weather Services (Figure 2.6) (February & Stock, 1999). Although technically within reach of the frontal systems, this lack of significant rainfall can be attributed to its position on the lee of the Cederberg range in the rain shadow of the mountains in terms of the cold fronts approaching from the west having the barrier of the imposing Cederberg; accounting for the reduced amount of precipitation. Penetration of the winter rains into the interior is restricted by the Cape Fold Belt, while summer aridity is further enhanced by the strong upwelling of the Benguela current off the west coast (Preston-Whyte & Tyson, 1993). Rainfall data from the South African Weather Services for the Clanwilliam/Cederberg region (1869 to 1999) shows April to September to be the wettest with June and July receiving the greatest amount of rain. As already noted, summer months receive comparatively little rainfall. A strong decreasing rainfall gradient occurs eastwards of the Cederberg. The eastern extremity of the study area is therefore far more arid. The western edge of the Tankwa Karoo (which is immediately to the east of the study locality) is one of the most arid regions of the entire Karoo, receiving only an average rainfall of about 50-70mm per annum (http://www.sanparks.co.za/ parks/tankwa/). Isohyets of mean annual rainfall for the Tankwa Karoo National Park fall into the 0-100mm range, with 25% of the mean annual precipitation falling in summer (Venter *et al.*, 1986). This implies that to the east of the study area, the rainfall regime changes to include both summer and winter rainfall. According to Thomas (1997), the majority of the study area falls within the category of semi-arid environments, i.e. it receives between 200-500mm mean precipitation per annum, whereas, directly to the east, the environment could be classed as arid, with 25-200mm, although, in drylands, annual precipitation frequently varies substantially from year to year.

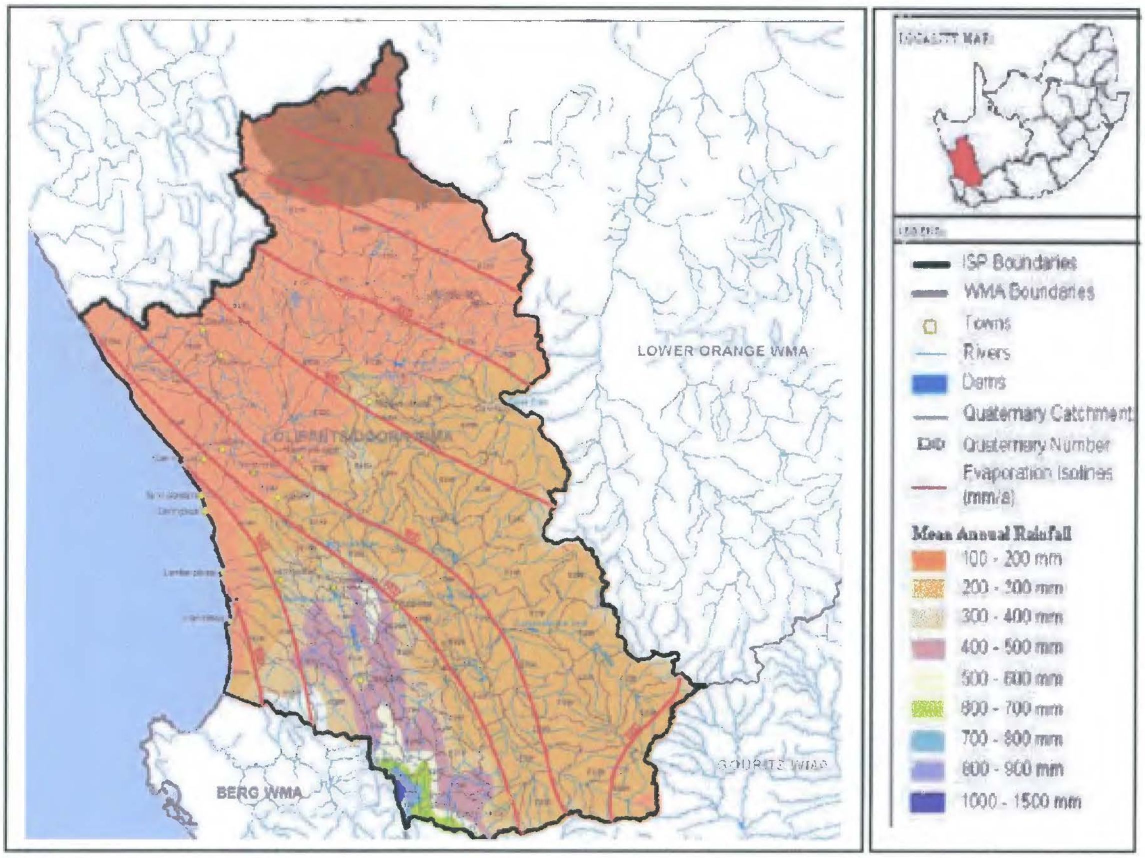


Figure 2.6: Mean annual rainfall for the Olifants/Doring Catchment (DWAF, 2005)

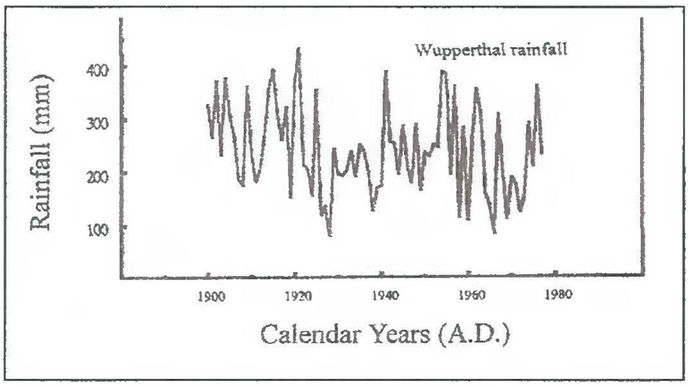


Figure 2.7: Rainfall data for Wuppertal from 1900 to 1980 (South African Weather Services)

Figure 2.6 shows Wuppertal to fall in the 300-400mm/yr category while the rainfall measurements from 1900 to 1980 (Figure 2.7) substantiate that 400mm appears as a maximum although it fluctuates greatly and can receive as little as 100mm/yr.

Other climate variables, such as temperature, soil moisture, humidity, average wind speeds and directions, etc. were not available for Wuppertal itself; instead these data were obtained for the nearest weather stations, namely Calvinia and Clanwilliam, from the South African

Weather Services, and used as a proxy for inferring contemporary climatic conditions as although Clanwilliam is in the Cederberg, it falls on the westward flank of the mountain range and still experiences some moderating effects of the ocean (approximately 45km to the west) as well as being at a lower altitude (152m a.s.l.). Calvinia, in contrast, is far inland at a high altitude (975m a.s.l.) just over the escarpment and experiences conditions likened to the Karoo. As the extremity of the study area does technically fall into the most western bounds of the Tankwa Karoo, climate information for this region was also obtained from the South African National Parks for the Tankwa Karoo National Park.

Temperature data are based on monthly averages between 1981 and 2005 for Clanwilliam acquired from the South African Weather Services. Summer months (November to March) experience the highest temperatures, especially further inland. The study area falls inland to the lee of a significant and high relief mountain range. It therefore does not experiences the moderating effects of the ocean and more extreme temperature ranges prevail where large differences between day and night temperatures can occur. Prevailing seasonal winds aid in moderating the temperatures to a degree. Winter temperatures, although cool on average, rarely ever drop below freezing. Weather data for the Clanwilliam area shows that summer and early autumn months experience average maximum temperatures between 27.9°C and 33.6°C with minimum temperatures ranging from 13.7°C to 17.5°C. In contrast, winter (May to September) months are on average much cooler with maximum temperatures between 24.1°C and 19.5°C and minimums from 6.4°C to 10.1°C. February is the warmest month while July is the coldest. The effect of its geographical position inland is noticeable through the large range in temperatures, as much as 16°C in certain months. Within the Tankwa Karoo, the mean July minimum temperature is 5,7°C and the mean January maximum temperature is 35,9°C, with the highest average maximum temperatures occurring from November to March (http://www.sabirding.co.za/birdspot/010337.asp). Although the study area does not occur at the high altitudes synonymous with the Cederberg plateau, it is at 478m a.s.l., and its 'closed' location within relatively contained valleys can be inferred to imply lower maximum and minimum temperatures, nevertheless maintaining daily and seasonal extremes. The study area also lies in close proximity to the semi-arid dry western Karoo where the latent heat flux is small and the sensible heat flux is the dominating term in the energy balance, giving rise to annual cycles within the energy budget and overall resulting in higher daytime temperatures (Preston-Whyte & Tyson, 1993).

The southwestern Cape is dominated by the subtropical South Atlantic High Pressure system and low pressure cyclones (cold fronts) that develop in the circumpolar westerlies between 40-60°S (Preston-Whyte & Tyson, 1993). These two systems result in a strongly bimodal or bidirectional and seasonal wind prevalence of SE and NW along the coasts and within the region. SE winds are associated with the high pressure cell during summer and strong NW winds and gales during winter are due to the cyclonic frontal systems. As, however, the study site occurs further inland, although still affected by these systems, it has the additional influence of the inland dominating pressure system. The Kalahari High Pressure system dominates the southern African interior and shifts southwards during the summer months. The anticyclonic airflow accounts for further variability within the climate of the area. According to wind data extracted from the South African Weather Services (www.weathersa.co.za), the predominant wind directions, based on the percentage frequency of occurrence of wind speeds in the specified categories, for the nearest weather station at Clanwilliam, is shown to be mainly SSE with the highest frequency followed by W maintained throughout the period October to March, with S and N also occurring at higher frequencies early and later in the season respectively. NNW also occurs frequently throughout the year, however, more so in the beginning and middle of the year. S winds are well sustained during the winter months too, along with N winds, although less than S, SSE and NNW. The winter months, however, are characterised mostly by calm days. On average, the strongest winds occur in summer, although there are large fluctuations, and sporadically stronger winds occur in the winter months from other directions (e.g. NE, ENE and WSW, W). As the study area occurs within a mountainous region within valleys, it is also important to bear in mind the occurrence of katabatic (downslope at night) and anabatic (upslope during the day) winds throughout the year (Preston-Whyte & Tyson, 1993). The highest average wind speeds occur from October to March within the Tankwa Karoo (http://www.sabirding.co.za/birdspot/010337.asp).

Coastal lows and *Berg* winds also determine the characteristic features of coastal and adjacent inland climates (Preston-Whyte & Tyson, 1993). As coastal lows are initiated off the west coast, associated shifts in temperatures and winds propagate inland but seldom further than the Cape mountain chain, although occasionally as far as the Escarpment. Berg winds, which are common in late winter and early spring, result in the anomaly of high winter temperatures and strong winds from a NE direction. These are low frequency high magnitude occurrences as a result of adiabatic warming of air as it descends from the interior plateau down the Escarpment (Preston-Whyte & Tyson, 1993). As the study locality is firstly, within proximity of the coast and secondly, of the Escarpment, and therefore occupies a transitional zone to the lee of the Cape Fold Mountains, these two phenomena are an additional influence on the climate of the region, particularly the Berg wind, as the wide open plains of the Tankwa Karoo adjacent to the Escarpment provide little obstruction to the winds as they blow towards the Cederberg and the coast beyond.

2.5. HYDROLOGY

As the three main rivers under investigation are tributaries and belong to the greater catchment area of the Doring (and ultimately the Olifants) River, it is important to consider the hydrology, not only of the individual rivers, but also of the greater system of which they are a part, especially as it is difficult to separate them out. The more specific hydrology and drainage of the study area will be considered and discussed in further detail in Chapters 5 and 6.

The Moordenaarsgat River sources between the Grootberge and Langberg (consisting of the recognisable peaks of Tafelberg and Wolfberg) before entering a well carved ravine. Its main tributaries are the Langkloof and Eselbank rivers to the west. The Dassieboskloof River, as already noted, drains from a wide area within the eastern Cederberg range and has numerous tributaries, one of the primary tributaries being the Sand River (to the southwest) along with an unnamed river to the north, as well as Boontjieskloof and Skerpioenspoort further north. These two rivers provide the primary drainage system for a substantial area of the northeastern and central Cederberg Range.

The Tra-Tra River's primary water source is the Moodenaarsgat River with the secondary addition of the Dassieboskloof River. Smaller tributaries, such as Olyfshoutskloof, Groot Gatkloof and Popuilerkloof, add to the Tra-Tra, however, these are non-perennial streams and would not likely yield a great volume of water. The main contributor is the Matjiesfontein River, a more significant perennial river that flows northeast out of a converging incised valley to join with the Tra-Tra River roughly five kilometres ahead of its confluence with the Doring River. The Tra-Tra is a significant tributary of the Doring River, the largest on the northeastern flank of the Cederberg with a sizeable catchment area, draining a greater area than the Biedouw River to the north or Matjies River to the south, with a higher tributary and stream density across the plateau of the northern central Cederberg. Its main constraint is the lack of significant rainfall in this northern and eastern region, as opposed to the southern and western Cederberg.

In the greater catchment, the Doring, along with the Sout River, are the main tributaries of the Olifants River (Figure 2.7). The perennial Doring River drains the Koue Bokkeveld and Doring areas and flows strongly during the winter months. The Doring River catchment, of which the Tra-Tra and its tributaries are a constituent, is a fan-shaped catchment, with the river rising in the south and flowing in a northerly direction, joined by the Groot, then Tankwa from the east and Tra-Tra from the west (DWAF, 2005). There are no major water impoundments in the vicinity of the study area, only small and privately owned dams exist far to the east and substantially downstream on the Doring River. For the entire Doring catchment (or sub-area up until its northern limits), the total yield available is calculated at 14 million m^3/a with the total current requirements estimated at 15 million m^3/a (DWAF, 2005). The sub-area therefore has a small deficit of only 1 million m^3/a and can be regarded as essentially in balance (DWAF, 2005). Surface runoff from the drier parts of the Doring River catchment (excluding the Groot River catchment) is typically of high salinity, but with the salinity concentrations varying seasonally and with the flow in the river (CNdV Africa, 2005).

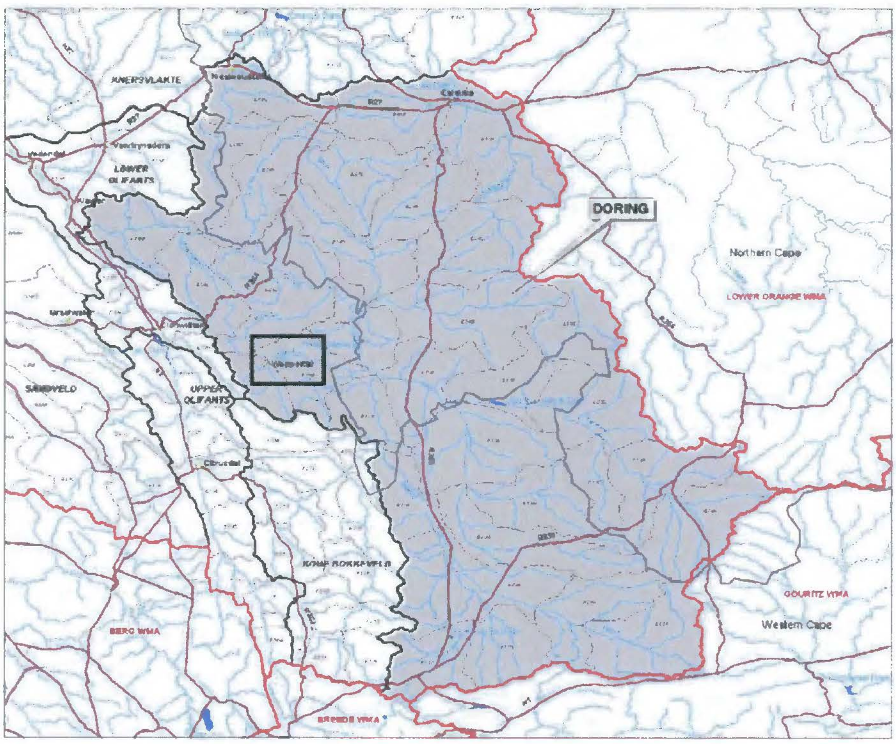


Figure 2.8: Doring River catchment (DWAF, 2005). Wuppertal area falls to the western margin.

The geohydrology of a major portion of the catchment, according to DWAF (2005), has been shown to be underlain by a shallow regolith (intergranular/-weathered-and-fractured) aquifer. The fractured-rock aquifer systems in the area include the Table Mountain Group (TMG) Aquifers in the Cederberg, and parts of the Witteberg Group in the Tankwa Karoo. The study area is on the boundary between Cape Fold Mountain rocks and those of the Karoo. The greater area to the east is underlain by formations of the Witteberg Group, Dwyka Formation, Ecca Group, and lower Beaufort Group (to the east). The Dwyka and the Ecca Formations form fractured rock aquifers. Rainfall patterns in the region are such that infrequent flood events recharge the aquifers.

The recharge of the TMG aquifers is highest in the high mountains along the southern and western catchment boundary divide of the Cederberg and Koue Bokkeveld ranges, which generally favour the exposed Peninsula Aquifer as the most sustainable source. However, in contrast to this, the estimated recharge in the northern region of the Cederberg and over a wide area of the Tankwa Karoo in the rain shadow east of the Cederberg range is less than 10 mm per annum. The yields obtained to date and the recharge distribution together indicate that the TMG fractured-rock aquifers are a main groundwater exploration target within this region for future use (DWAF, 2005). Within the Olifants/Doring greater catchment, the area of highest median yield (>5 ℓ/s) is shown on current DWAF maps to occur in parts of the TMG in the Agter Witzenberg and Koue Bokkeveld areas in the extreme south, near Vanrhynsdorp in carbonate aquifers of the Nama Group, in primary aquifers in drainage channels leading to the Wadrif primary aquifer near Lamberts Bay, and along the Tra-Tra River northeast of Wuppertal. The northern parts of the TMG and the Bokkeveld-Witteberg aquifers south of Wuppertal are associated with median yields between 0.5 and 2 ℓ/s (DWAF, 2005).

As the geohydrology and presence of aquifers in the region illustrates, the number of boreholes reflect good groundwater sources in spite of rather arid to semi-arid climatic conditions that persist. According to the Department of Water Affairs and Forestry (DWAF) (2005), the abstraction pattern shows a relatively high level of summer-season groundwater dependence from whatever aquifer sources are locally available, although there is a large amount of uncertainty regarding the groundwater usage in the area; much of the groundwater is of a very poor quality and of a relatively low yield. The quality of groundwater varies greatly over the area, at several locations the water is too mineralized for any direct use and desalination is required. In general groundwater has become highly mineralized (brackish to saline) in the drier regions (CNdV Africa, 2005). However, for small-scale supply some parts of the region are reliant on groundwater and it is believed that further exploitation potential exists. Baseflow into the rivers is of ecological importance to species that over-summer in pools along the riverbed; and invasive alien plant infestations are not yet significant in terms of water consumption within the area, however, the few and first occurrences should be controlled in accordance with good catchment management as the riparian areas are prone to rapid invasion (Low *et al.*, 2004). Prevention of such infestations would provide the benefits of maintaining the baseflows of the rivers, a vital component to the fluvial and ecological environments of the area.

2.6. VEGETATION

The study area is located within the southwestern Cape region and the Cederberg Range falls within the periphery of the Fynbos Biome or Cape Floristic Region (Figure 2.9), one of six floristic 'kingdoms' in the world and eight recognised biomes in southern Africa. Biomes are characterised as a large land community unit, distinguished primarily on the basis of dominant life form and secondarily by major climatic features (not an unnatural or anthropogenic system) (Cowling *et al.*, 1997). The southwestern Cape's winter rainfall climate, particular combination of geology, topography, soils and environmental history, have resulted in a distinctive landscape colonised by an unique vegetation formation (Meadows & Baxter, 1999). The fynbos biome is described as open to closed grassy, dwarf shrubby, shrub/woodland normally not exceeding 3m in height and dominated by life forms such as evergreen, sclerophyllous and small-leafed phanerophytes (trees), chamaephytes (shrubs) and hemicryptophytes (grasses) (Rutherford & Westfall, 1986). It is also characterised by three elements, namely restioid, erocoid and proteoid components, *Proteaceae* being the dominant species. Ericoid fynbos consists of members of the *Ericaceae* family, while restioid fynbos consists of mainly tall, grasslike restio species. The fynbos biome is renowned for its species-richness and biodiversity with over 8500 plant species over 86 000km^2 with a very high level of endemicity, in fact over 80% of plant species are confined to the Cape Floristic Region (Cowling *et al.*, 1997). Fynbos is also highly associated with nutrient-deficient soils that are weakly developed on rock, sands and lithosols and podsols and a specific fire regime resulting in successional stages (Rutherford & Westfall, 1986).

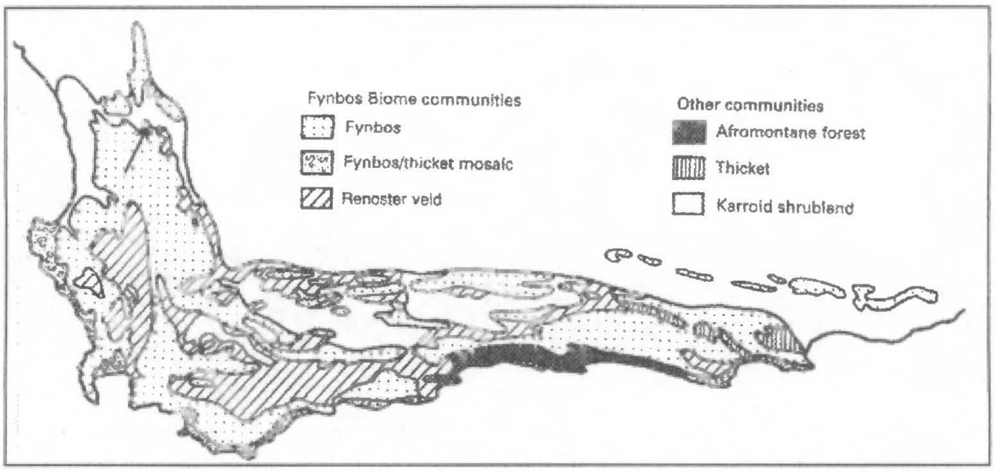


Figure 2.9: The boundaries of the Cape Floristic Region (Cowling *et al.*, 1997). Study area falls in northern extremity of Fynbos Biome (shown by arrow)

The northern Cederberg region is considered one of the more arid parts of the fynbos biome (Meadows & Holmes, 2000). The mountains are covered almost exclusively by pristine fynbos and the range almost typifies the environmental conditions associated with the Cape Floristic Region (Meadows & Sugden, 1993). The study area incorporates all three elements of the fynbos community although local environmental circumstances affect the relative components. The Cederberg fynbos vegetation has been shown by pollen analysis taken from vleis at Sneeuberg and Driehoek to shift in its community composition over time, from a more Ericaceous fynbos in the 1900s to a more Restioid fynbos recently (Meadows & Sugden, 1991). Some rare fynbos endemics are found in the Cederberg, such as the Snow Protea, rocket pincushion and the Clanwilliam Cedar. The Clanwilliam Cedar, *Widdringtonia cedarbergensis*, is a threatened endemic tree species in the Cederberg Mountains. Although once considered to have been more widespread across the mountain range, it now has a diffuse distribution and is altitudinally and geomorphically restricted, found between 1000 and 1400m a.s.l. and restricted to cliffs, rocky outcrops and very rocky slopes (Sugden & Meadows, 1990). Within the study area, only a portion is represented by fynbos vegetation, namely the western border (eastern slopes of the Cederberg) and valley floors. The type of vegetation in particular areas is associated with the underlying geological parent material, a major determinant. The oligotrophic quartzites (freely drained, sandy, acidic soils) of the Table Mountain Group (in the west of the study area) support a variety of fynbos communities in comparison with the nutrient-rich Bokkeveld shales, which outcrop in the east, that support more karroid communities (Meadows & Sugden, 1991).

The Dassieboskloof, Moordenaarsgat and Tra-Tra River banks are well vegetated with riparian vegetation in the form of restios and some larger fynbos shrubs. The Dassieboskloof River is particularly densely vegetated at the banks of the perennial primary channel, especially as one progresses further upstream (westwards) up the canyon. The Moordenaarsgat River banks are also relatively well vegetated, especially across some channel bars and the banks of the perennial main and braided channels. Its banks are very densely vegetated in the ravine to the north. In comparison, although the Tra-Tra riparian vegetation has been all but removed or greatly disturbed east of the confluence by the local population and replaced with grass and agriculture. Downstream (to the east), the riparian vegetation contrasts sharply with the sparse karroid and succulent vegetation a few metres from the channels. The amount and size of the riparian vegetation does also begin to decrease as one travels downstream further east into the Tankwa Karoo; the fynbos is slightly more xeric in appearance and replaced by more grass species.

Invasive alien plants cover an area of approximately 122 km^2 spread across the Olifants and Doring River catchment (DWAF, 2005). Much of the infestation is in the riparian zones as they are prone to rapid growth due to availability of water. Acacias, Pines, *Syringa*, *Eucalyptus* and *Prosopis* are among the top ten genera of invading alien plants, which account for about 80% of the total water use by invasive alien plants within the catchment (Cullis *et al.*, 2007). The alien plant invasion is not yet significant within the study area, although there are alien species evident, especially around Wuppertal and along the watercourse downstream of the settlement.

The succulent Karoo biome is found mostly west of the western Escarpment from Namibia to the Western Cape and inland of the fynbos biome to the Little Karoo, covering approximately 111 000km^2, (fourth largest biome in southern Africa). Topographically, much of the biome is flat or undulating with some more rugged regions occurring in the western Escarpment although it is not altitudinally restricted, having a large range from sea level to 1700m a.s.l. (DWAF, 2005). The largest drainage systems present within the biome are the Tankwa/Doring/Olifants and Gouritz and their relevant tributaries. The eastern section of the study area, along the Tra-Tra River, falls directly within the succulent Karoo biome, although the Dassieboskloof and Moordenaarsgat rivers' vegetation belongs to the fynbos biome. Soils are commonly weakly developed on rock and include sands and alluvium. The succulent Karoo is mainly limited to the winter rainfall areas with greatest summer aridity and mean annual rainfall of 20 to 290mm, as in the Tankwa-Doring Karoo where rainfall is 50-100mm. The biome has very slow natural rates of vegetation development or change and owing to the low fuel load and high frequency of non-flammable succulent plants, there is virtually no fire regime, although when fires do occur, plants are sensitive and the effects are long-lasting (Cowling *et al.*, 1997).

The western Tankwa Karoo and extending to the Cederberg is characterised by flat Karooveld with small riverine bushes, including the common *skaapbos* (*Tripteris sinuata*) (van der Merwe, 2007). Karroid vegetation dominates the Olifants/Doring River catchment, occupying some 75% of its area (DWAF, 2005). The flora is characteristically low, typically less than 1m in height, and includes scrub, bushes, dwarf trees and a few grasses. According to Cowling (1997), the vegetation of the biome is dominated by chamaephytes (often succulent, such as species of *Drosanthemum* (*vygie*)), therophytes and geophytes. Flowering displays by therophytes, such as members of the *Asteraceae* family including *Euryops annuus* and *Gazania lichtensteinii*, are usually a sign of degraded or overgrazed lands (Smuts, 2007). The biome is rich in succulent species mainly belonging to the Mesembryanthemaceae and Crassulaceae families. Succulent plants use the Tankwa Karoo as a migrating corridor. The duration and temperature of the growing season clearly separates succulent Karoo from other biomes (Cowling, 1997). Acocks (1988) described the Tankwa Karoo as terribly tramped out, and eroded down to the bedrock with the better conserved sections described as short succulent Karoo with many being of the stemless type; with non-succulents also found, and *Stipagrostis obtusa*, even becoming abundant after good rains; annuals and geophytes are numerous but rarely seen (Acocks, 1988). Milton *et al.* (1997) described the vegetation structure as very sparse shrubland and dwarf shrubland, with succulents on shallow soils, grass and ephemerals on sandy alluvium.

Paradoxically, perhaps, much of the arid and semi-arid Western Cape constitutes two biodiversity hotspots, namely the Cape Floristic Region and the Succulent Karoo, which have been recognised as such by Conservation International (Low *et al.*, 2004). There is regionally a remarkable level of endemism and diversity, especially given the aridity, with a considerable abundance of rare or endangered species in the biome and high succulent plant species diversity unparalleled anywhere else in the world (Low *et al.*, 2004). The study area incorporates both vegetation types highlighted as biodiversity hotspots.

2.7. LOCAL HISTORY AND LAND USE

The town of Wuppertal is the main settlement on the northeastern flank of the Cederberg Mountains. It is accessible only by a gravel road from the north or a 4X4 track to the south. Besides the town itself and the very minor development within it, the immediate vicinity remains relatively unchanged. Wuppertal was established as the result of a Moravian mission station set up in 1865 as German missionaries settled among the *Khoikhoi* families that inhabited the valley and encouraged farming. The population of the town grew rapidly after the abolishment of slavery as many freed slaves from nearby farms arrived (www.wupperthal.co.za). Wuppertal is a fully-functioning and productive town producing *veldskoen*, dried and rolled tobacco, dried fruit, dried beans, rooibos tea and rooibos products. The region now is also frequented by increasing amounts of tourists, however, due to the rugged terrain, Wuppertal and the Cederberg Mountains remain almost entirely unspoilt by such tourism. Such development, although on a small-scale, has the potential to have impacted the surrounding environment, even if only on a more local-scale.

A growing popularity within South Africa and worldwide has led to an increased demand for rooibos, *Aspalathus linearis*, thus resulting in an increase in its cultivation across the region. Wuppertal and the Cederberg area has become a local hub for such rooibos farming, with the area surrounding Clanwilliam being the only place where rooibos is cultivated as an agricultural crop (www.wupperthal.co.za). Seventy-five farmers from Wuppertal and the surrounding communities currently participate in a program moving from wild harvesting to sustainable cultivation of a rooibos product suitable for export. Many of farmers used to gather rooibos in the wild but there has been a recent shift towards growing it organically. Rooibos is only farmed in some valleys, but mostly across the mountaintops of the surrounding Cederberg ranges and the plateaus of the northern and northeastern region. The Biedouw Valley to the north is an example of rooibos farming using centre-pivot irrigation. This suggests that there has been some disturbance to the indigenous vegetation. Within the study area, the valleys of the Tra-Tra and Moordenaarsgat rivers remain relatively uncultivated, except for a small amount of subsistence farming that occurs in the immediate vicinity of Wuppertal and Eselbank (another small settlement nearby), where fruit and vegetables and a small amount of corn are grown, as well as down the Tra-Tra valley where small family settlements live, and beyond the ravine of the Moordenaarsgat River where there is more extensive cultivation of land. The Dassieboskloof River canyon is unsuitable for any cultivation, although there is some far upstream along the tributaries across the plateau of the Cederberg. Soils derived from the parent material (belonging to the Cape Supergroup) are not usually suitable for large-scale cultivation (CNdV Africa, 2005).

To the east of the Cederberg, nomadic pastoralism first brought sheep into the succulent Karoo about 2 000 years ago, and cattle some 1 500 years later (Boonzaier *et al.*, 2000. The European pastoralists who moved northwards from the Cape Peninsula in the 18th century moved with their flocks to suitable grazing. In the 19th century the succulent Karoo became the first biome used for settled European pastoralism (Milton *et al.* 1997). The extremely arid summers, however, make much of the succulent Karoo unsuitable for settled pastoralism, even now when boreholes provide perennial water and forage can be imported from other areas. Prior to the development of the Tankwa Karoo National Park in 1986, the area had been almost abandoned due to excessive erosion on the part of overgrazing, mostly by sheep, however, since farming and tourism in the region has been limited and mitigation measures have been implemented, the land has been turned over to a "veld recovery phase until the original vegetation re-establishes itself" (SANP, 2006). The only notable non-indigenous fauna evident along these valleys are small herds of goats and donkeys that graze, particularly in the Tra-Tra River valley.

With regard to the Clanwilliam Cedars, it is a broad preconception that the Cederberg once had a more substantial distribution of cedars than today and that they were more abundant across extensive areas of the plateau, however, pre- and post-colonial human exploitation and occupation of the Cederberg resulted in the destruction of such forests and the demise of the cedars to this day (Sugden & Meadows, 1990).

2.8. CONCLUSION

The physical setting of the study area illustrates the wide variety of elements that need to be considered when investigating the contemporary and palaeo-fluvial dynamics, geomorphological and environmental conditions within the region. These become of special importance when exploring the possible future implications of change. It is essential to bear in mind that changes over time are due to a number of factors and that no constituent can be taken in isolation.

Chapter 3

LATE QUATERNARY PALAEOENVIRONMENTS RELATED TO GEOMORPHIC RESPONSE AND CLIMATE CHANGE

3.1. INTRODUCTION

This chapter aims to provide a detailed review and consolidation of the literature associated with Late Quaternary palaeoenvironments and inferred climate change, specifically relating to the Holocene and of the Cederberg region in particular. A brief outline of the literature related to the geomorphic response of landscapes and the thresholds of systems to such changes or shifts in climatic conditions and environmental factors follows. Ages are uncalibrated and quoted in years before present (BP) or AD where otherwise mentioned.

3.2. LATE QUATERNARY PALAEOENVIRONMENTS OF THE SOUTHWESTERN CAPE

3.2.1. Overview of the Late Quaternary in the southwestern Cape

The Quaternary represents the most recent sub-period of the Neogene, a period of the Cenozoic era, approximately the last 2.6 million years, and is divided into the Pleistocene and Holocene epochs, where the Holocene covers roughly the last 10 000 years of Earth history (Summerfield, 1993). The Quaternary is a period of environmental fluctuations and characterised by repeated climatic changes of considerable amplitude that aptly demonstrate the 'fundamental' dynamism of the Earth's system and normality of its inherent change (Meadows, 2001). The Quaternary is also of importance as it is the period in which humans have become a dominant controlling factor or environmental agent within the subcontinent as well as hominid evolution and their development being closely influenced by accompanying environmental changes (Bell & Walker, 1996). These fluctuations are associated with cycles of increased and decreased rainfall and temperatures. Many causes for climate variability and environmental change over the Quaternary have been suggested focusing on inter-related and interacting forcing mechanisms with relative importance assigned on a spatial and temporal scale. Briefly, these include the astronomical theory based on orbital perturbations or Milankovitch cycles; changes in terrestrial geography, i.e. the configuration of the major continental landmasses, tectonic activities such as uplift, plate tectonics; variable solar output, atmospheric composition, volcanic emissions; and thermohaline circulation dynamics (Lowe & Walker, 1997).

The Quaternary period is defined by the cyclical nature of glaciation and deglaciation that progressed to the cold glacial of the Last Glacial Maximum (LGM) at approximately 21 000 to 18 000 years before present (BP), returning to the warmer conditions that still prevail today (Lowe & Walker, 1997). These cycles continue to the present and will carry on into the future. Glaciation is known to materialise gradually whereas deglaciation appears to be a more rapid process, as demonstrated by the period from 125 000 to 16 000BP, where in the southern hemisphere, according to Tyson (1999), higher latitudes were characterised by a series of rapid pronounced warmings followed by slow variable declines to progressively lower minima. Globally, the LGM was associated with an increase in aridity or a decrease in moisture availability and lowered sea-levels, down to -120m in certain regions. Generally southern Africa reacted to the cyclical climatic fluctuations by becoming warmer and moister during interglacials and cooler and drier during glacials (Meadows, 2003). The LGM in southern Africa has been identified as the peak of Late Quaternary aridity in spite of the paucity of high-resolution marine records in adjacent oceans, and the generally accepted large-scale forcing of southern African climates focused on in recent, sparsely distributed, discontinuous terrestrial records and few coastal sequences (Stokes *et al.*, 1997). Within these warming and cooling phases, shorter events occurred, such as the Allerod Interstadial, Younger Dryas and the Holocene hypsithermal/altithermal. Palaeoenvironmental signatures became better defined both regionally and temporally after the onset of the LGM (Partridge, *et al.*, 2004).

There is considerable complexity and variation evident in the natural environment of the contemporary southwestern Cape region and it is necessary to take this into account for any reliable accurate reconstructions of the Late Quaternary palaeoenvironments. Prevailing environmental conditions vary distinctly according to their geographical locality, therefore identified palaeoenvironmental changes should be spatially variable too. Within the region, environmental and biogeographical gradients are steep and there is the inevitable likelihood that environmental changes were not uniform throughout (Mulock-Houwer, 2001). These factors make regional summaries, particularly in light of Late Quaternary climate and environmental change, problematic (Meadows & Baxter, 1999). Modern climatic data also does not support the definitions of the "winter-rainfall zone" as presented in previous syntheses and palaeoreconstructions and this factor accentuates some of the difficulties in resolving the palaeoenvironmental record of the region along with the lack of preserved long sedimentary records (due to the seasonally arid climate) resulting in a spatially limited and temporally biased Quaternary record (Carr *et al.*, 2006). However, recent research has enhanced the understanding of the spatial and temporal complexity of climate changes affecting the region in the last glacial cycle, including a complex record of punctuated aridity, although highlighting issues as well, such as data integration and forcing mechanism controls, showing they are imperfectly understood overall (Thomas & Shaw, 2002). Southern Africa's climate is influenced by shifts in the Intertropical Convergence Zone, the westerlies, and the development and position of continental and oceanic anticyclones. Over the last glacial–interglacial cycle substantial changes in the amount and seasonality of precipitation across the subcontinent have occurred and been linked to the relative dominance of these systems especially in relation to the extent of the region's glacial climates being affected by expansions of Antarctic sea-ice, equatorward migrations of the westerlies, more frequent and/or intense winter storms and an expanded winter-rainfall zone (Chase & Meadows, 2007).

The LGM was cool and dry over most of non-equatorial southern Africa as the semi-permanent subtropical anticyclone dominating the atmospheric circulation was displaced equatorwards (Tyson *et al.*, 2001). The southwestern Cape, however, appears to have been out of phase with the rest of southern Africa and reflects varying responses to these fluctuations. Meadows and Baxter (1999) demonstrate how evidence supports the hypothesis that greater moisture availability occurred during the LGM, instead of cooler and drier conditions, and that an increase in temperature corresponds with a decrease in moisture availability. The fact that the west coast winter rainfall region underwent changes different from those elsewhere in South Africa has been largely attributed to equatorward shifts of the southern ocean frontal systems and the Antarctic sea-ice margin that increased meridional pressure gradients, strengthening the atmospheric circulation and trade-wind intensity and onshore precipitation over parts of the Western Cape during glacials and the LGM (Partridge *et al.*, 2004 citing Parkington *et al.*, 2000 and Cowling *et al.*, 1999). Proxy data from the western region in the vicinity of Elands Bay implies reduced evaporation and greater moisture availability, probably as a result of increased frequency and efficacy of frontal systems carrying moisture (Cowling *et al.*, 1999). Sediments from Elands Bay and Diepkloof Caves show major soils consisting of optimal vegetative cover and stable slopes relating to periods of non-glacial climate whilst glacial-age environments were generally geomorphologically active, drier and largely characterised by open vegetation, in contrast to otherwise suggested moister conditions (Mulock-Houwer, 2001 citing Butzer, 1982). Palaesols in the west and southwestern Cape (dated between 25 000 and 15 000BP) are interpreted as indicating conditions as moist as during the mid-Holocene, however, as stated by Parkington *et al.* (2000), it is not possible to claim with certainty the environmental conditions of the LGM in the Western Cape, but rather surmise that it was wetter, colder, cloudier and grassier than present.

The southwestern Cape forms the core region of the Cape Floristic Region, the smallest of the six large-scale phytogeographical units, with one of the richest plant communities in terms of high plant species, biodiversity and levels of endemism. Although limited in sites of studied evidence, the Late Quaternary vegetation history shows relative vegetation stability over periods of marked climatic and environmental dynamism, with only subtle shifts in vegetation community patterns in response to variations in rainfall and fire regime (Meadows & Sugden, 1993). Elenga *et al.* (2000) also shows, using pollen analysis for biome reconstruction over the last 18 000years BP, that although there were major changes in biome distribution in comparison to present day, the fynbos (xerophytic scrub) of the southwestern Cape persisted. This is also clearly demonstrated by Tyson *et al.* (2001) who shows that reconstructions have been made of biome distribution across southern Africa during the LGM and after, with the areal extent of the fynbos biome and Mediterranean shrubland of the western and eastern Cape being maintained, if not slightly increased, reflecting the combined influence of temperature depression and desiccation.

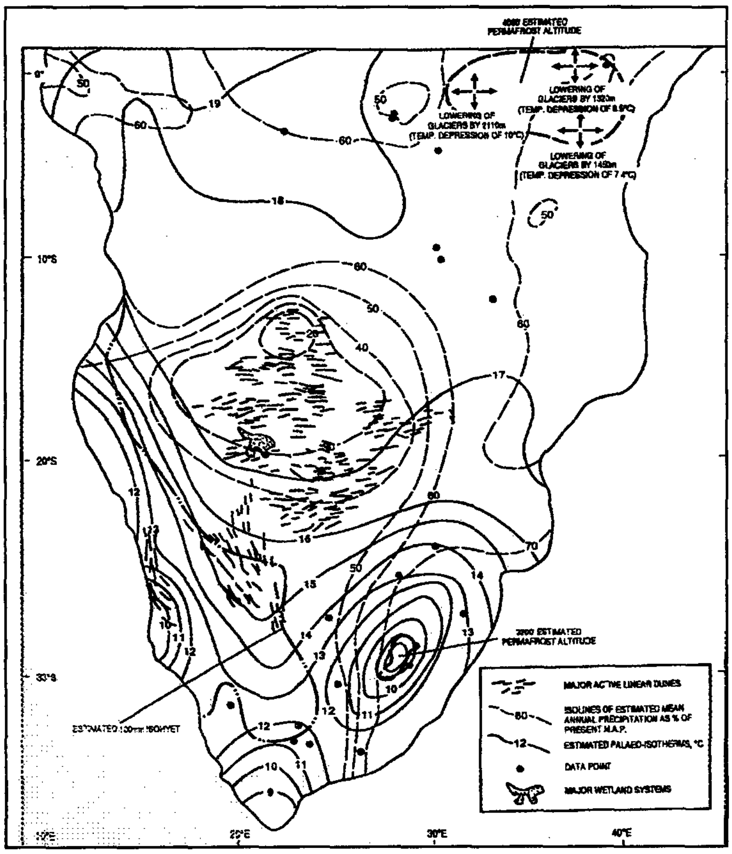


Figure 3.1: Palaeoclimatic reconstruction for southern Africa at LGM (21-18kyr) (Partridge *et al.*, 1999)

The winter rainfall zone of South Africa, and subsequently the southwestern Cape, are situated at a sensitive location between temperate and subtropical oceanic and atmospheric circulation systems (Carr *et al.*, 2006). There is a range of evidence for change over the subcontinent from the last interglacial to present, such as the general circulation changes evident through alterations over time of the alignment of Kalahari dunes of differing morphology and dune accumulation along the western margin of southern Africa (Thomas *et al.*, 2000; Chase & Thomas, 2006). The LGM and Holocene altithermal represent the two most recent extremes in global temperatures (Partridge *et al.*, 1999). During the LGM and subsequent deglaciation, events in Antarctica were "pre-eminent" in forcing changes in climate over southern Africa through variations and responses in the thermohaline circulation with temperatures gradually rising to Holocene values, although briefly interrupted by the cooling of the Younger Dryas (Partridge *et al.*, 2004). Chase and Thomas (2007) suggest that environmental changes in the region are strongly linked to variations in wind strength and atmospheric circulation systems during the Late Quaternary. It has been recognised that the patterns of changes that occurred over southern Africa fit coherently into a model of expanding and contracting circumpolar and tropically-induced atmospheric circulation adjustments, which have taken place both systematically over long time scales and occasionally abruptly to produce a high degree of variability and frequent abrupt change superimposed on longer-term oscillations (Tyson, 1999). With the improvement of the resolution of the record for southern Africa it has been shown that the climate variability has increased with abrupt, extreme changes being increasingly common. Tyson (1986) and Cockcroft *et al.* (1987) proposed an atmospheric circulation model that suggests that at the time of the LGM the circulation over most of southern Africa was dominated by the westerlies and transient weather disturbances with an annual rainfall persisting that would have been much less than present-day. The model and trend of climate transition indicates that warming and an increase in rainfall would have advanced south from the more tropical northern parts of southern Africa and would only have reached the southern coast of South Africa last (Tyson, 1999).

The conditions of the deglaciation after the LGM had stronger tradewinds that occurred during the Antarctic Cold Reversal and the Younger Dryas period (with the probable increase of upwelling off the west coast) along with a southward shift of the Atlantic anticyclone that could have resulted in both stronger tradewinds and the reduced impact of the Westerlies on the climate of southwestern Africa (Dupont *et al.*, 2004). According to Preston-Whyte and Tyson (1993) a steeper pole–equator gradient combined with a weaker Walker circulation (associated with a weaker Southern Oscillation) would have caused a northward migration of the Westerlies and an eastward shift of the ascending (rain-bringing) limb of the Walker cell into the Indian Ocean. This model reinforces for the LGM, on the one hand, a wetter winter rainfall area in the southwest of South Africa and, on the other hand, drier conditions in the rest of southern Africa that has a summer rainfall climate. Partridge *et al.* (2004) provide evidence also indicating an extension of the winter rainfall region, notably that there was a northward progression of fynbos vegetation (including Restionaceae and Asteraceae) in glacial climates, which, when considered alongside more humid conditions as presented by sedimentological records, implies a northward extension of the winter rainfall belt along the western lowlands and increased rainfall in parts of western South Africa and Namibia.

Palaeoclimatic reconstructions using regional climate modelling have shown significant differences between the southern and western coastal areas of the Western Cape for the LGM although both experienced winter rainfall, being that the western coastal region experienced cooler and moister conditions in comparison to the drier and colder conditions of the south coast region (Barrable *et al.*, 2002). During the first part of the deglaciation (until 13,000 yr ago), different measured proxies show parallel developments on land and in the ocean with the northern position of the frontal system and strengthening of the South Atlantic anticyclone bringing increased winter rains along the western margin of southern Africa during the LGM, while a strengthened Atlantic anticyclonic system would also have blocked off the Indian Ocean airflow (Dupont *et al.*, 2004). It is generally accepted that sea surface temperatures have had a substantial effect on past climates, although their influence from the southeast Atlantic is now restricted to the rather narrow strip of coastal desert along the west coast. Evidence taken from the grain size distribution of the terrigenous sediment fraction (Stuut *et al.*, 2002) from marine sediments of the Walvis Ridge off the coast of Namibia can be interpreted as the result of mixing of coarse and fine aeolian dust and hemipelagic mud; the ratio of the two aeolian components and an influx of aeolian pollen trapped in marine sediments is attributed to the stronger intensity of the southeast trade winds during glaciations compared to interglacials, with changes interpreted as variations in southwestern African aridity. According to this proxy, southwestern African climate was relatively arid during the interglacial stages and relatively humid during the glacial stages of the past 300 000 years (Stuut *et al.*, 2002). Moreover, increased relative abundance of pollen from Cape reeds (restios) indicates a northward extension of the Cape flora, probably along the western escarpment indicating a considerable northward shift of the winter rainfall area (Shi *et al.*, 2000). Marine pollen records indicate that the winter rainfall area reached far northward during glacial times (Dupont citing Shi *et al.*, 2004), as far north as the Brandberg area where now isolated insect populations indicate connections to the Cape region in earlier times. The extent to which the winter rain climate reached inland is difficult to estimate from marine records, therefore using land-based records the consensus is that the winter rainfall climate did not reach far inland but rather extended as far north as 20°S as a narrow strip along the western escarpment, leaving most of the hinterland under the influence of the summer rain climate being more arid during glacial times than during interglacial ones. This position and extension of the South African winter rainfall region is coupled with the interaction between land and ocean associated with the intensity of upwelling, the strength and direction of the tradewinds, and Atlantic and Indian sea surface temperatures influencing terrestrial climates. The decoupling of the ocean and land relationship over time could be related to the southward shift of the South Atlantic anticyclone resulting in the Westerlies being a weaker influence with the winter rainfall area decreasing and the climate in the westernmost parts became drier (Dupont *et al.*, 2004). The improved resolution of the climate variability record in southern

Africa has shown patterns of change that have occurred to support the previously mentioned model of expanding and contracting circumpolar and tropically-induced atmospheric circulation. Palaeoclimate modelling has aided in showing that both temperature and precipitation have fluctuated over time within the southwestern Cape region. More mesic conditions along with lower temperatures, roughly 1°C cooler, and greater moisture availability in the western region of the Cape is attributed to either the increased passing of frontal systems, increased cloud cover and stronger southeasterly winds and intensified oceanic upwelling along the adjacent coast around the LGM (Barrable *et al.*, 2002). Increased upwelling can be loosely coupled with aridification of the interior coastal belt, which concurs with a drier western region during these bouts. Evidence from aeolian activity on the Agulhas Plain towards the margin of the southwestern Cape winter-rainfall region indicates strong westerly winds and enhanced rainfall seasonality during both the Pleistocene and Holocene (Carr *et al.*, 2006).

All Quaternary environmental syntheses, offering an integrated and multidisciplinary perspective on change in southern Africa over time, strongly suggest developed differences between the southwestern Cape and the rest of southern Africa, particularly with respect to the development of palaeoenvironments. Climatically, depending on the amplitude of change associated with the circumpolar vortex, records of the north and south of the subcontinent are not always in phase (Tyson, 2000). In general, on an annual time scale, the shifts in the climates of the winter and summer rainfall regions of South Africa are 180° out of phase (Tyson *et al.* citing Lindesay, 2001). In contrast to the popular proposal of drier conditions across the fynbos biome during the last glacial and subsequent stages, faunal evidence from Elands Bay Cave has shown wetter conditions prior to the Holocene, while the Cederberg pollen evidence has shown limited environmental change and fluctuating vegetation belts. Wood charcoal and pollen from Elands Bay Cave showing more diverse vegetation and a wider range of plant communities points towards increased moisture availability during the last cold stage as well as there being marked contrasts between the Holocene and prior period. Pollen analysis and faunal species lists indicate increased amounts of grazers, such as springbok, with evidence pointing towards terminal Pleistocene animals that were much larger than their Holocene counterparts. However, the most notable animal present was considered the hedgehog, as the suite of animals inferred grassier conditions, the hedgehog and molerats' occurrence pointed to wetter conditions, possibly implying increased rainfall of more than 400mm, particularly at the boundary between the Pleistocene and Holocene

(9500BP) with a decrease in size through the Holocene possibly indicating a decrease in rainfall due to a close relationship between size and adult size of the molerats (Parkington *et al.*, 2000). Fossil charcoal evidence from Elands Bay Cave on the coast in the western fynbos biome winter-rainfall region provided a record for palaeoclimatic reconstruction from the late Holocene to roughly 40 000BP (Parkington *et al.*, 2000). Fossil assemblages indicated a transition from Afromontane forest to riverine communities prior to 18 000BP followed by mesic thicket and fynbos vegetation in the terminal Pleistocene and mostly xeric communities during the Holocene. This also suggests that in the western part of the fynbos biome, soil moisture conditions were higher in the LGM than during the Holocene (Parkington *et al.*, 2000).

During deglaciation periods in southern Africa, climates responded most obviously to changes originating in the Antarctic and fluctuating heat budgets of the Indian Ocean. The area of the Atlantic's influence extends northwards up the west coast by virtue of the Benguela upwelling system, although changes within this system are moderated by fluctuations within the global thermohaline circulation (Partridge *et al.*, 2004). High-resolution oxygen and carbon stable isotope data suggest that postglacial warming was first initiated around 17 000BP and was interrupted by cooling probably associated with the Antarctic Cold Reversal, followed by strong warming after 13 500BP (Holmgren *et al.*, 2003). Cartwright and Parkington (1997) indicate a raised sea-level that possibly removed the strandveld along the west coast in the vicinity of Elands Bay, whilst Klein (1991) reports an increase in available moisture from 13 600 to 9600BP, possibly associated with higher sea-levels combined with increased upwelling, however, this is in opposition to Scott and Lee-Thorp (2004) who state rather that upwelling is closely linked to aridity along the west coast.

Younger Dryas

Widespread evidence from across southern Africa demonstrates implies that rapid warming took place after 16 000BP with a sudden cooling at 11 000BP, associated with the Younger Dryas, which punctuated the overall shift of conditions post-LGM (as shown by pollen records and mollusc stable isotope data from the Atlantic coast) (Tyson, 1999). Thereafter temperatures steadily increased to the Holocene altithermal (Tyson *et al.*, 2001). The transition from Pleistocene glacial conditions to the warmer Holocene was not smooth and it has become globally recognised that the Holocene consisted of a series of oscillations (Abell

& Plug, 2000). The transition into the Holocene was a complex event with a warming trend occurring towards the end of the last glaciation that was briefly interrupted by a cold event just prior to 11 000BP (Younger Dryas), which ended abruptly at 10 000 years ago. Isotopic enrichments of mollusc shells have shown a decrease in SST around 11 000 to 10 000BP, coinciding with the northern hemisphere Younger Dryas event that occurred *circa* 11 000 to 10 000BP (Scott *et al.*, 1995). However, across South Africa, during the Younger Dryas period, the only notable and consistent evidence was of a transitioning phase from cool late Pleistocene conditions to warm Holocene environments.

3.2.2. Holocene climate change

Much of the evidence used in interpretations and reconstructions of the global Holocene palaeoconditions and environments is palynological, although proxy records from micromammals, charcoal, palaeosols, glacial and periglacial evidence, lake levels, environmental isotopes, diatoms, geomorphological evidence from changing river regimes, dune distributions and sedimentary granulometry have also played their part (Partridge *et al.*, 1999). There are inevitable problems associated with studies of the Holocene, just as they apply to all Late Quaternary palaeoreconstructions, although there is a somewhat greater abundance of sites and evidence that include preserved records from the last ~10 000 years. However, it should be noted though that evidence of moisture regimes since the last interglacial over southern Africa remains scarce (Stokes *et al.*, 1997). There is also the additional consideration that, while palaeoconditions may be inferred, regional variations within the temperature and available moisture records of the Holocene are considered generally of smaller amplitudes than those of glacial times and are therefore more subtle or minor in nature, proving more difficult to positively attribute to specific conditions or elucidate further (Scott & Lee-Thorp, 2003). Considering proxy evidence, climatic conditions adjusted following the LGM at a faster rate and with fewer fluctuations inland in contrast to the coast, resulting in greater disparities occurring in climatic conditions between the coast and inland, especially in the Western Cape (Avery, 1990).

Palaeodata shows changing moisture balances over the Late Quaternary with studies confirming shifts between relatively wet and dry conditions on a 5-20kyr timescale during the period (including the Holocene) (Stokes *et al.*, 1997). The belt of westerlies delivering moisture to the winter rainfall region was affected by latitudinal movements of the circum-

Antarctic frontal systems and climate varied overall in the southwest region of Africa in phase with the Antarctic during last glacial and Holocene climate evolution (Tyson, 1999). Holocene temperatures were reached subsequent to warming that continued after the Younger Dryas, attained shortly after ~10 000BP (Partridge *et al.*, 2004). In general, over the southwestern Cape region, the first half of the Holocene is associated with more xeric conditions than the last 5000 years, which were generally moister (Meadows & Baxter, 1999). In the west of the fynbos biome there is evidence from pollen assemblage composition at Elands Bay Cave indicating a notable change towards the more xeric nature of vegetation today; transitioning from proteoid fynbos elements dominating (12 400 and 13 600BP) to more mesic thicket elements at the Pleistocene-Holocene boundary between 8000 and 10 000BP to common asteraceous shrubland elements at 3000 to 4300BP (Parkington *et al.*, 2000). This shift in vegetation points towards a decline in moisture availability or rainfall through the Holocene.

Post glacial warming culminated in the Holocene altithermal that reached its maximum around 6000-5000BP, although it varies according to the location; this fact corresponds well to one of the most well-known logs of evidence for environmental change over the last four glacial-interglacial successions, or 400 000 years, the Vostok ice core from eastern Antarctica. In the last 6000 years, the most pronounced and sustained event was the Little Ice Age, which was characterised as five centuries of cooling from AD1300 to 1800 (Tyson *et al.*, 2001). Climatically, stable isotope analysis from a range of sites has shown the southwestern Cape to remain relatively arid throughout the Holocene due to the easterlies circulation system dominating, a reduced westerlies system and circumpolar vortex (Scott & Lee-Thorp, 2004). This appears in records as a drying trend after approximately 5000BP with high levels of *Asteraceae* pollen, increasing development of C_4 grasslands peaking around ~2000BP and a series of roughly 1°C temperature fluctuations associated with drier, cooler and stormier conditions (Scott & Lee-Thorp, 2004). This contrasts with evidence observed by Meadows & Baxter (1999), which suggests a drier first half of the Holocene, as noted elsewhere. Issues and discrepancies arise from generalisations made across regions from evidence containing poor dating controls or resolution. Records that have provided the basis for the development of a preliminary palaeoenvironmental map have given invaluable information that lends insight into the probable changes in atmospheric circulation as temperatures rose across the subcontinent by 6-7°C from the LGM to the Altithermal, which are in agreement with present-day models that have been developed to explain short-term variations in the southern

African climate (Partridge *et al.*, 1999). Models show that extended dry spells occurred with an expansion of the circumpolar vortex and an increased occurrence of westerly disturbances.

Data from widely separated localities across the world show variability and oscillations along with a tendency for quasiperiodicity within the evidence. Many studies report quasiperiods, variability or oscillations around 0.9–1.5ka years (Bond *et al.*, 1997; Lamy *et al.*, 2001). Climate oscillations in the range of 1.57-0.5ka have been thought to result from changes in the thermohaline circulation of the North Atlantic and its connectivity with the global thermohaline conveyor, from vegetation albedo positive feedbacks, stochastic resonance and variations of solar output. With retreat of tropical circulations equatorward with an expanding vortex, the southwestern winter rainfall region of the subcontinent became wetter and cooler (Holmgren *et al.*, 2003).

Early Holocene and Altithermal

Several palaeoenvironmental records across South Africa show that between 15 000 and 7000BP temperatures increased by as much as 5°C in places with vegetation in the dry Karoo and winter-rainfall grasslands, dwarf shrublands and fynbos region transitioning during this period to include more succulents and woody elements (Scott *et al.*, 1995). Some of these records show short-term oscillations implying that the transition from cooler to warm conditions was not consistently gradual. The period from 11 000 to 7500BP was characterised as relatively arid, however, it was considered more mesic in the southwestern winter rainfall region (Scott & Lee-Thorp, 2004).

Micromammalian remains serve to elicit climatic information for particular regions of South Africa, particularly rainfall, although assemblages can be skewed by factors associated with scale, specific predators, natural disasters (e.g. fires and their type, season, intensity and frequency), drought, regional, local and short-term population fluctuations of the prey and predator species, and the depositional rate of the accumulation (Matthews, 1999). Interpretation of micromammalian data, specifically in the fynbos and Karoo biomes, has shown that there has been variation in the amount and possibly the seasonality of rainfall during the Holocene (Avery, 1990). It is suggested that, essentially, the climate of the fynbos biome during the Holocene was as it is today with some relatively minor variations, notably two episodes of fluctuating temperatures at about 10 500 to 9000BP and 1600BP preceded by periods of rapid increase in temperature. During these episodes vegetation indicates relatively arid conditions with lower rainfall in the southwest between 9000 and 4000BP (although alternatively this could be interpreted as less effective precipitation due to increased temperatures closer to the boundary between the fynbos and succulent Karoo there could have been a reduction in summer rainfall) (Avery, 1990). Data presented by Van Zinderen Bakker (1976) based on faunal fossil evidence for the coastal region around Elands Bay suggest a shift from open grassland-type vegetation to more closed shrubby vegetation around 12 000 to 9000BP implying a link to warming and decreased precipitation, whilst geomorphological evidence, in the form of the last major aeolian activity, indicates a drop in precipitation also around this time (Mulock-Houwer, 2001 citing Butzer, 1982). A rise in SSTs due to less intense atmospheric circulation in the southern Benguela Current region possibly resulting in a decrease in upwelling occurred in the early Holocene (around 10 000-8000BP) also may have led to climatic warming within the region reflected as warmer temperatures by pollen data from Elands Bay Cave (Meadows & Baxter, 1999).

The most pronounced change in the pattern of variability since the LGM occurred at around 6000BP. Following the Younger Dryas cooling event, temperatures rose with records showing that the Holocene altithermal manifested itself in southern Africa between 7000 and 4500BP with temperatures rising to higher than those at present and at a maximum since the last interglacial at 125 000BP (Tyson, 1999). There is, however, conflict between the range and peak of the altithermal, for example, both Partridge *et al.* (1999) and Talma and Vogel (1992) suggest that evidence from across southern Africa indicates that both the summer and winter rainfall regions south of the equator experienced maximum Holocene temperatures at various times between 8000 and 6000BP. According to Chase and Thomas (2006), the altithermal encompassed 8000 to 4000BP. Pollen and diatom analysis in the Wilderness Lakes region of the southern Cape by Martin (1959; 1967) also shows a correlation with a defined division during the Holocene when conditions transitioned between 6000 and 5000BP while palynological evidence from South Africa and Namibia concluded that the temperature peak for the region can be placed between 7000 and 6500BP. According to PASH (Palaeoclimates of the Southern Hemisphere), during the Holocene altithermal "optimum" conditions evidently occurred earlier in the north than the south (Partridge *et al.*, 1999).

Although the relative temperature increase was subtle and of similar amplitudes over wide areas, it appears that the changes in precipitation seem to have been more complex overall

(Partridge *et al.*, 1999). There is considerable regional variation in the rainfall models for the Holocene altithermal, as shown by Tyson *et al.* (2001), Meadows and Baxter (1999), Scott and Lee-Thorp (2004), Avery (1993), Partridge *et al.* (1990) and Meadows and Sugden (1993). Collated records and evidence indicates substantial regional variation, in particular, with the semi-arid western interior of the southern part of South Africa receiving an increased rainfall in contrast to the area to the east (Tyson, 1999). According to Partridge *et al.* (1999) and Tyson *et al.* (2001), reconstructions of southern African vegetation distributions and biomes during the Holocene altithermal appear to be much the same as today. Tyson (1999) proposes that the circulation prevailing at the time of the altithermal was likely to have been dominated by tropically-induced disturbances in the easterlies. The dominating easterlies circulation system and reduced westerlies system concurrent with a reduced circumpolar vortex, as recorded in coastal Antarctica, probably accounts for the fact that the southwestern Cape was relatively arid during the period 6500 to 5100BP, which was characterised as warm and wet elsewhere (Scott & Lee-Thorp, 2004). This increased aridity is substantiated by widespread remobilisation of dune fields along the western margin of South Africa indicating triggering despite reduced wind strength but higher temperatures and reduced influence of moisture bearing westerlies resulting in increased aridity (Chase & Thomas, 2006).

Evidence from elsewhere in southern Africa shows additional periods of aridity and possible forcing factors during the mid-Holocene. Stokes *et al.* (1997) have shown variability in the SST and possible association to rainfall gradient across South Africa leading to stationary easterly disturbances to the east over the African landmass and the eastward encroachment of the South Atlantic anticyclone indicating three post-LGM periods of aridity. There could also be the consideration that such periods may correspond to windier, not neceassarily drier, periods (Meadows *pers. Comm.*). Sedimentary evidence from Verlorenvlei shows more estuarine conditions around 5500 to 5000BP and increased sedimentation from then onwards with indications that the catchment was more arid than present with reduced vegetation cover (Meadows & Asmal, 1996).

Late Holocene

Relatively little is known of the high-resolution spatial gradients of regional climatic variability in southern Africa from the altithermal until 2000BP (Tyson, 1999). Most records after 5000BP reflect an overall but variable drying trend (Scott & Lee-Thorp, 2004).

Generally the late Holocene experienced warm, evaporative conditions with fewer C_4 grasses; cooling is evident up until 2500BP, followed by warming between 1500 and 2500BP and briefly at AD1200 whilst maximum Holocene cooling occurred around AD1700. The strongest variability superimposed on more general trends has a quasi-periodicity of between 2500 and 4000; also present are weaker 1000 and 100-year oscillations, the latter probably being solar-induced. Given similarities to the Antarctic records, the proximate driving force producing millennial- and centennial-scale changes in the record is proposed to be atmospheric circulation changes associated with change in the southern hemisphere circumpolar westerly wind vortex (Holmgren *et al.*, 2003).

Two major cool events occurred between the third and fifth millennia BP in the southern part of South Africa after the altithermal (Tyson, 1999). The first took place from approximately 4700 to 4200BP with the second so-called neoglacial period (which has been extensively reported elsewhere in the world) between 3200 and 2500BP. The intervening period between the two cool periods was mild. Changes in the continuous $\delta^{13}C$ record over the last 6000 years shows subtle shifts in the proportions of C_4 grasses and C_3 shrubs; correlations of the maximum grass cover and $\delta^{18}O$ values indicating warmer and moister periods, such as that believed to have occurred between roughly 2000 and 2400BP (Talma & Vogel, 1992). A conspicuous arid episode is indicated by high $\delta^{18}O$ ostrich eggshell values from Elands Bay Cave along with more xeric thicket species and *Asteraceae* shrubs (and a distinct lack of mesic thicket) between 4700 and 3000BP (Lee-Thorp & Talma, 2000; Cowling *et al.*, 1999). High grass and *Asteraceae* pollen counts support the inference from the isotope record of a cool, grassy environment at 3000–2000BP. In contrast, more mesic conditions returned from 2700 to 1200BP as warm, wet conditions and a bushy environment occurred from 1200 to 600BP (Lee-Thorp & Talma, 2000; Holmgren *et al.*, 2003). Overall the late Holocene is interpreted as wetter at Elands Bay reflected between 4000 and 2000BP (Scott & Lee-Thorp, 2004), while moister climatic conditions after 5000BP favoured increased run-off and sedimentation until roughly 3800BP at Verlorenvlei; there is also the presence of increased regional moisture availability signal around 2500BP to the north (Baxter, 1996). Tyson *et al.* (1999, 2000 and 2001) have carried out investigations related to the Medieval warming and Little Ice Age in South Africa, especially with respect to speleothem evidence from the Cango Caves and Makapansgat Valley. Although neither site is located in the immediate proximity of the southwestern Cape, they prove to be invaluable proxy evidence that correlates with other evidence from the region specifically.

Medieval Warming and Little Ice Age

During the last millennia in South Africa, the isotope record shows two events, namely the medieval warming event from 900 to 1300AD, characterised by highly variable conditions and maximum warming occurring around 1250AD; and the Little Ice Age that prevailed for 500 years after 1300AD with maximum cooling at around 1700AD with a previous less intense pronounced "cool spell" 100 years prior (Tyson *et al.*, 2001). Holmgren *et al.* (2003) agrees that the most recent cool, dry event was that associated with the Little Ice Age although the duration differs, at about 1500 to 1800AD. Proof of these events across southern Africa in the recent record are demonstrated and revealed by tree-ring analyses from widely separated sites, such as the Cederberg and Karkloof (Dunwiddie & LaMarche, 1980), foraminifera in diatomaceous sediment off the west coast, fluvial sediment analyses in Namibia, oxygen and carbon isotope records of speleothems, molluscs and coral, high-resolution palynological analyses, fluctuating sea-levels along the western and southern coasts (indicated at Elands Bay and Knysna), changing lake levels and borehole temperature profiles (Holmgren *et al.*, 1999; Tyson *et al.*, 2001). The medieval warming period was characterised as wet in South Africa with above-average warming and variability along with drier intervals of reduced moisture availability while the Little Ice Age was considered overall as dry punctuated by extended wet spells (Holmgren *et al.*, 1999). Conditions transitioned from warm and wet to cooler and drier that appeared to have abruptly ended; this was also the case during the earlier period of the medieval warming. There were oscillations in annual mean daily maximum temperatures of 2-3°C in a few decades between the tenth and thirteenth centuries with the temperature peak, at roughly 1250AD, being the warmest event in the last six millennia (Tyson *et al.*, 2001). The Little Ice Age was also terminated by an event of similar rapidity. The medieval warming and Little Ice Age have been linked to variations in solar irradiance and observed in cosmogenic isotope records with the medieval warming corresponding to the Medieval Maximum of solar activity (Tyson, 2000).

Data have suggested that generally over much of southern Africa the period from 2000BP to 900AD consisted of a cooling period between 100 and 250AD, a period of warming from 250AD to 600AD and was followed by cooling until 900AD. Subsequently, the medieval warming period, corresponding with the Medieval Warm Epoch across many parts of the world, precedes the cooling of the Little Ice Age, with a warm episode from 1500 to 1675AD.

The Little Ice Age and medieval warming phase are distinctive features in the regional climate of the last millennium (Tyson, 2000). Using oxygen and carbon isotopes along with colour density data, a proxy climate record has been developed that indicates the climate of the interior of South Africa to have been 1°C degree cooler during the Little Ice Age and could have been over 3°C higher than the climate of present during the extremes of the medieval warming, although these periods were highly variable, it was particularly so during the medieval warming phase. A series of ~1°C temperature fluctuations across the country, associated with drier, stormier and cooler conditions, culminated around 1750AD (Scott & Lee-Thorp, 2004). Extreme events in the record correspond and show notable teleconnections with similar events in other parts of the world, in both northern and southern hemispheres (Tyson, 2000). Within the evidence, warming prevailed for longer at lower latitude sites while cooler episodes lasted longer at more southerly sites, thus the Little Ice Age in the south of the country began ~100 years before and ended ~100 years later at sites further north (Tyson, 1999). Evidence from across southern Africa has associated the Little Ice Age cooling with aridification and an accompanying increase in the frequency of major flood events. In particular, palaeoflood hydrology of the lower Orange River valley points to greatly increased flooding in the river catchment that corresponds roughly with the Little Ice Age (Tyson citing Zawada *et al.*, 1999).

Rapid changes in sea-level during the Holocene are apparent with mid-Holocene coasts in particular being radically different to those of the late- and early-Holocene, both in form and process (Long, 2001). Proxy archaeological and palaeoenvironmental evidence of sea-level change from the southwestern Cape and phases of dune building on the southern coast identify minor peaks in sea-level during the last 6000 years with relative falls coinciding with neoglaciation period (Ramsay & Cooper, 2002). There is a wide range of evidence for significant fluctuations during the Holocene along the coast of the Western Cape, including sites such as Langebaan Lagoon, Rietvlei and Verlorenvlei (Baxter & Meadows, 1999; Meadows & Baxter, 2001). It is interesting to note that SSTs declined and fluctuated during the late Holocene corresponding well to the Little Ice Age, between 600 and 400BP (Mulock-Houwer, 2001 citing Jerardino, 1995; Marker, 1997). It is also valuable to note that a sea-level drop by as little as 2m, such as that occurred, is sufficient to initiate a slight drying of the interior as direct moisture from the sea does not reach as far inland as it previously did. In terms of regional aridity, the punctuated sea-level rise and fall record can be coupled with fluctuating fluvial hydrology (Parkington, 1999).

Time	Years BP	Western Sub-region	Central Sub-region	Southern Sub-region
LGM	25–17ka	Wetter Cooler	Dry Cool	Dry Cold
After LGM	16–11ka	Drier Warm	Drier Warm	More moist Warmer
Early Holocene	10–8ka	Dry Warm	Dry Warm	Dry Warm
Mid-Holocene	7–4ka	Dry to sub-humid Warm	Sub-humid to moist Warm	Moist to dry Warm to cool
Late Holocene	3ka to present	Wetter Warm	Wetter Warm	Moist to dry Warm

Table 3.1: Brief synopsis of precipitation and temperature changes for the past 25kyrs as seen from proxy evidence (Mulock-Houwer, 2001)

3.2.3. Holocene palaeoclimatic evidence in the Cederberg

A synthesis for the western region of the southwestern Cape as a whole indicates that firstly the LGM was not as cold as expected, but rather cooler and wetter, with conditions shifting to a drier and warmer early Holocene (Meadows & Sugden, 1991, Scott & Vogel, 2000). The mid-Holocene, however, begins as an extension of the early Holocene but then transitions to a sub-humid and wetter climate with similar temperatures of today (although a higher moisture availability) although levels of moisture increase into the late Holocene when conditions begin to shift towards the present-day with a decrease in moisture. Although precipitation fluctuated temporally and spatially throughout the southwestern Cape, the Holocene altithermal in the winter rainfall region is characterised by drier conditions whilst it is generally accepted that temperatures exceeded the contemporary mean by roughly 2°C (Partridge *et al.*, 1999). One should also bear in mind that the extreme conditions of LGM and altithermal did not occur everywhere at the same time.

The Cederberg has a rich supply of Late Quaternary palaeoenvironmental information and is suggestive of subtle yet important shifts in environment within the region, locally and as a whole. The Cape Floristic region's plant diversity is concentrated and remains relatively undisturbed in the southwestern Cape mountains and upland regions, therefore providing a comparatively pristine stock of Quaternary environmental change (Meadows & Baxter, 1999). Palaeoenvironmental data for the Cederberg specifically are based largely on palynological analysis from vlei sediments and *Hyrax* midden material, common throughout the Cederberg, showing particular vegetation species change, such as the decline in the Clanwilliam Cedar (*Widdringtonia cedarbergensis*) (Meadows & Sugden, 1991). Pollen preserved in organic sediments is a key instrument in palaeoenvironmental elucidation as its relative components and abundance represents efficient indicators to use as a proxy for determining past climatic and environmental conditions. Although accumulated organic sediments within these upland regions are often shallow and do not extend far back into the Holocene, palaeoenvironmental implications from such palynological analyses means that results can now be placed within the greater context of the region as a whole and compared with other forms of evidence (Meadows & Baxter, 1999). Palynological evidence taken from the Cederberg is dependable as in many respects the Cederberg typifies the environmental conditions of the Cape Floristic region as it receives mainly winter rainfall (200-1000mm/yr) and quartzitic sandstones create the characteristic sandy, freely-drained highly acidic and nutrient-poor soils in which species-rich fynbos communities thrive along with other Cape flora components. The poorly-drained valleys of the mountains along with isolated shallow pockets of organic sediment accumulation have given rise to a few excellent fossil pollen-bearing deposits of up to 6m in depth with continuous accumulation, for the purpose of palaeoenvironmental consideration, such as those at the Driehoek and Sneeuberg vleis (Meadows & Sugden, 1993). Pollen records indicate only marginal taxa fluctuations over the period of sedimentation during the Pleistocene and Holocene and minor environmental changes, perhaps in rainfall or fire frequency. Subtle shifts in composition of vegetation communities could also have occurred in response to slight increases and decreases in moisture availability and temperature. These are interpreted such that the apparent abundance of the endemic cedar earlier in the sequences could be in response to a lower fire frequency as a result of greater annual rainfall; as the environmental fluctuations are confirmed to have been relatively subtle, with moisture availability varying only ± 20% over the entire sedimentation period. This is in contrast to elsewhere in southern Africa, which during the same time period, experienced marked climatic fluctuations in both temperature and precipitation. Meadows and Sugden (1993) propose that low amplitude fluctuations, for example in mean annual precipitation, occurred over the last 15 000 years and can be coupled with patchy fire frequencies.

Measured isotope fluctuations in *Procavia capensis* (hyrax) dung samples from different vegetation zones and various ages over the last 20 000 years indicate variations in the amounts of C_4, CAM or C_3 plants consumed by these herbivores and potentially also vegetation changes that may have occurred over time. $^{13}C/^{12}C$ values for a series of hyrax middens of late Pleistocene/Holocene age, show that hyraxes favour mainly C_3 plants in their diets but do incorporate CAM or C_4 plants under certain circumstances (Scott & Vogel, 2000). Midden samples have been examined from the Cederberg Mountains, especially from the more northern region in the Pakhuis area, for both their pollen content and isotopic signatures with the isotopic data indicating diets composed almost exclusively of C_3 plants during the last 20 000 years with slightly enriched shifts (Scott, 1994; Scott & Vogel, 2000). It appears that the southwestern Cape experienced drier conditions during warmer phases in contrast to the more common increase in moisture availability as seen in the summer rainfall regions of southern Africa (Meadows & Baxter, 1999). Around 22 000BP the northern Cederberg experienced relatively cool but drier conditions as suggested by depressed vegetation zones in contrast to greater moisture implications from Elands Bay Cave (Barrable *et al.*, 2002). Scott (1994) provides a reliable record extending back to the LGM in the xeric end of the mountain fynbos spectrum generating a view of changing environmental conditions in the Late Quaternary. Late Pleistocene sequences indicate shifting and lowering of vegetation belts, therefore concurring with the cooler and moister conditions hypothesis, which correlates to the modern analogy of present day environmental gradients that cooler temperatures allude to higher rainfall. Wetland elements, such as Cyperaceae and Restionaceae, peak between 16 000 and 15 000BP, also possibly suggesting cool and moist conditions in the Cederberg, with pollen levels of tree taxa also becoming significantly more abundant in response to the shift of the climate subsequent to the LGM (Meadows & Baxter, 1999). Evidence from the relatively consistent pollen sequence taken from vleis at altitude in the Cederberg shows no indication of a Younger Dryas effect, whereas the record from the hyrax midden taken at lower altitude in the northern Cederberg does imply a marked change from before 13 000 to 8 000BP with increased succulents, tree or woody elements and larger shrub pollen, possibly suggesting a response to mild warming and slight increase in rainfall post-LGM (Scott *et al.*, 1995). The initiation of organic-rich sediment in vleis, such as Driehoek, is also an indication reflected sedimentologically of moister conditions in the interior around 14 600BP.

Palynological evidence from the northern Cederberg at the northern edge of the fynbos biome has provided a vegetation history spanning 14 500 years and suggests a record of relative overall stability for the highest and central parts of the mountain range (Meadows & Sugden, 1991). The Pleistocene-Holocene boundary has been proven as a period of marked environmental dynamism across southern Africa. Meadows and Sugden (1991) infer the LGM to have been cooler and wetter than present, interpreted through the general decline in the Clanwilliam Cedar throughout the sequences thus providing conditions that favoured the tree species. However, Scott and Vogel (2000), through the use of ^{13}C in fossil hyrax dung, argue that there has been very little variability over the last 20 000 years in the Cederberg as shown by long-term persisting negative values in the range of C_3 type vegetation typical of the region, hence the mountains continually received winter rainfall for the region or cool winter conditions persisted otherwise associated with the Pleistocene when cooler growing season conditions were possible. Their evidence also shows no marked contrast between the Pleistocene and Holocene. Only relatively small variations occur within the record, with slightly greater levels of such variability in the Holocene. Small increases in summer-rain ratios may have been responsible for some variations, although they remained relatively small in comparison to winter-rain contributions. The Driehoek and Sneeuberg vlei pollen evidence shows that vegetation patterns within the central Cederberg have only "shuffled" during the late Pleistocene and Holocene with no marked community changes or encroachment of karroid elements (Meadows & Sugden, 1991). Within the Driehoek core there are indications of shifts between periods of greater moisture availability followed by a suggestion of more xeric conditions when restios were replaced by more ericaceous and proteoid elements, although overall more mesic mountain fynbos appears to have dominated since 14 600BP (Meadows & Sugden, 1991).

Around 9640BP, the interior of the Cederberg is shown to be warmer and moister by peat initiation; pollen analysis found Proteoid fynbos with a restioid understorey (on Sneeuberg) and more ericaceous, arboreal and proteoid elements (at Driehoek) around 10 090BP (Meadows & Sugden, 1993). A hyrax midden pollen sequence in Pakhuis Pass, in the north, indicates slightly more mesic conditions with prominent *Dodonaea* and other woody elements during the early Holocene (Scott, 1994). Meadows and Baxter (1999) show pollen belonging to asteraceous shrubs and *Aizoaceae* (mainly succulents and frost-sensitive) to peak between 8000 and 6000BP indicating this to be the warmest period of the Holocene with the lowest moisture availability; higher temperatures during the early Holocene were therefore associated with more xeric climates within the Cederberg Mountains, in contrast to the prior mentioned proposal. Modelling of palaeoclimates also supports evidence for warmer and drier conditions in the western parts of the southwestern Cape during the Holocene altithermal (around 6000BP) (Mulock-Houwer, 2001). Overall, palaeoecological reconstruction for the central Cederberg, however, suggests marginally greater moisture availability during the early

Holocene, and in one site, also the mid-Holocene, although conversely studied vegetation history for the Pakhuis region of the northern Cederberg finds no evidence of greater moisture availability during these times (Meadows & Holmes, 1999). There are discrepancies between literatures on the duration of the Holocene altithermal in the region; for instance, Talma and Vogel (1992) propose it lasted from 7000 to 5000BP while Partridge *et al.* (1999) indicate it was between 8000 and 6000BP, there is an issue and controversy surrounding variations and availability in moisture and precipitation at the time.

Climatic interpretations for the Holocene altithermal derived from evidence of vegetation changes based on the remains of micromammals indicate that the winter rainfall region of the southwestern Cape experienced an extended rainfall season (Avery, 1993). A site in the northern fynbos biome suggests, for the Holocene altithermal, that there was a prominent restioid element on hillsides and a smaller scrub component whilst other fynbos sites showed predominant grass elements with a slightly reduced restioid element. From this, inferences of climate were deduced as showing relatively higher rainfall and moderately warm wet winters with a shift to more summer rainfall and reduced seasonality in the winter-rainfall region by approximately 6000BP (Avery, 1993). This agrees with the suggestion by Scott and Vogel (2000) that rainfall shifted its seasonality somewhat within the region. Again, micromammal evidence taken from the Cederberg Mountains above Clanwilliam, suggests that the grassy/restioid element was more extensive around 5000BP than present, possibly lasting until 2000BP. This higher proportion of grass would indicate, along with samples from the Olifants River valley, that the second half of the Holocene had higher rainfall than present-day (Avery, 1990). A drop in sea-level, as noted previously, has a drying effect on the interior. Such drying is noted from the vegetation composition at Driehoek, in the Cederberg, around 3230BP, as Meadows and Sugden (1991) found the assemblages to consist of dry ericaceous fynbos that signifies dry environmental conditions at the time. Late Holocene hyrax sequences suggest an increase in CAM or C_4 plants suggesting dry western and southern regions experienced more summer rains. A comparable pattern of isotope change is observed as a peak in $\delta^{13}C$ recorded from the Cango Caves stalagmite to the Namib and Cederberg hyrax middens just around 2100BP, suggesting some sort of vast regional change, mechanism or shift that plant cover was responding to although this does not necessarily imply similar seasonal rainfall shifts over the whole of the wide area (Scott & Vogel, 2000). In the Cederberg area this could be attributed to a dry episode (Scott & Vogel, 2000). Whilst more mesic conditions returned to the Elands Bay region between 2700 and 1200BP (Lee-

Thorp & Talma, 2000) hyrax midden pollen shows more Cyperaceae, "renosterbos" (*Stoebe/Elytropappus* type) and declining *Dodonaea* thicket around 2000BP in the Cederberg i.e. moving away from mesic conditions (Scott, 1994). Dietary $\delta^{13}C$ values in the hyrax dung declines sharply to their lowest around 1000 to 900BP probably reflecting cooler and maybe wetter conditions in the Cape region; another possibility is that there was a relative decline in summer rain in the north and concurrent wetter winter rain conditions in the fynbos region (Scott & Vogel, 2000). Pollen records show woody elements to have been more abundant in the Holocene, showing reductions only in the most recent levels, in contrast to the proposed decline of the Clanwilliam Cedar due to the impact of pre-colonial occupation within the Cederberg (Meadows & Baxter citing Sugden & Meadows, 1999). From 800BP there is a renewed rise in $\delta^{13}C$ values with another enrichment peak occurring at 420BP indicating an increase in CAM and C_4 plant availability and more of an influence from summer rains to the north (Scott & Vogel, 2000).

Sugden and Meadows (1990) and Meadows and Sugden (1991) have shown that the endemic Clanwilliam cedar did not dominate the Cederberg mountain range or ever be prolific enough to form closed-canopy forests. It did, however, appear to have been more widespread prior to 4000BP and evidence from before 9000BP indicates that the trees formed relatively open stands with the possibility that population numbers were three to four times greater than today. There is, however, a marked decline in the abundance of the cedars after 2000BP as the upper core levels indicate accelerated cedar decline, a probable result of anthropogenic influences (human disturbance) coupled with natural environmental changes. There is a likelihood that cooler temperatures along with the possibility of greater winter- and summer-precipitation and a more equable fire regime may have promoted and led to the relative "abundance" of cedars during the LGM and subsequent interglacial up until the late Holocene (Sugden & Meadows, 1990). This is followed by the impact of the San and Khoi, especially stock such as sheep. A more recent reduction in the relative abundance of fire-prone taxa appears to be related to human occupation during the last 2000 years within the Cederberg while increased fire frequencies promoted bulb growth (Meadows & Sugden, 1993). Human-induced disturbances in the composition, distribution and species richness of fynbos communities have been attributed to changes in the burning regimes of the Khoi-San (Meadows & Sugden, 1991). The introduction of stock though by Khoi herders from 1800BP onwards would have had a significant effect. The Khoi herders introduced animals, which were largely grazers, into a region normally associated with browsing animals. They were also responsible for pushing the San into more marginal habitats resulting in increased fire frequency through patch-burning. The European farmers arrived from 400BP with a further increase in fire frequencies by farmers (Meadows & Sugden, 1991), although the eastern fringe of the mountain range was only settled in after 1800.

Lake Bruno in the northern Cederberg (Pakhuis Basin) exhibits sedimentary suggesting a once inundated lake or marsh environment that was believed to have occurred for two periods during the first half of the Holocene, from 10 000 to 8750BP and 6250 to 5100BP, whereas today it is largely in-filled by unconsolidated alluvium sediment (Meadows & Holmes, 1999). A section illustrates 10m of alluvial and colluvial sediments exhibiting a complex sequence of intercalated medium to fine sandy and silty sediments, with some of the more surficial facies showing cross-bedding characteristic of aeolian processes. Organic sediments (occasionally highly organic), richer in clay and black in colour, with woody material and pollen, exist whilst laminations present within the section show signs of frequent local environmental fluctuations. The black and dark grey clays are consistent with formation under water or possibly under permanently moist marsh conditions. The organic units have been dated from 5100 and 10250BP (Meadows & Holmes, 1999). This type of sediment and lacustrine (or marshy) facies is unusual in the more arid northern Cederberg, and considering its extent across the localised Pakhuis Basin, strongly indicates two distinct phases in the early and mid-Holocene that resulted in greater runoff into the basin remaining as standing water, either as a result of neotectonics or alternatively higher precipitation within the catchment. Sedimentary evidence from the Bruno section provides a clear pattern of changing depositional environments and conditions over time, such that organic rich strata towards the base suggest periods of greater moisture availability during the early Holocene, whereas alternating periods of sediment mobility and aeolian activity in the upper section indicates a drier late Holocene (Cornell, 2001). From this site there is also evidence of slightly drier and significantly wetter periods within a generally moister early Holocene. Another section, Pakhuis, further upstream, similarly reflects two phases with a wetter lower half and drier upper half within the stratigraphy. Two phases of significantly moister conditions in the lower half correspond well with the Bruno section (Cornell, 2001). This evidence conflicts with palynological and micromammal evidence from elsewhere in the Cederberg. More detailed palynological analysis of fossil pollen preserved in the organic facies of the Lake Bruno site is ongoing and would shed further light on the palaeoenvironmental conditions of the Pakhuis Basin and more generally of the northern Cederberg. Preliminary pollen results indicate a slightly wetter than present Holocene as suggested by the more extensive presence of *Olea* tree species (Smith, B. *pers comm.*).

The nature of the southwestern Cape flora, being fynbos that is unique with high species richness and endemism, itself is an indicator of environmental conditions and past climatic changes. It is hypothesised that the species richness of this biome is the result of low amplitude environmental fluctuations that triggered allopatric speciation events, which were not severe enough to increase extinction rates simultaneously, as more marked climate changes would have yielded a less species-rich Mediterranean region, such as those elsewhere in the world (Meadows, 2005).

3.3. GEOMORPHIC RESPONSE TO CLIMATE CHANGE

Landscapes are the product of a dynamic and ongoing interaction between the Earth's surface and the atmosphere. Geomorphology is concerned with describing earth surface features as well as gaining insights into the complex association between geology, endogenic forces and exogenic processes, added to the role of vegetation and human impact. Geomorphology is a science that describes not only the shape of the Earth's surface but also the fundamental dynamism of landscapes, which is itself, a study of environmental change. Landscape processes are sensitive to changes in climate and other environmental parameters and the Quaternary represents an especially dynamic phase of Earth history including the substantial influence of humans in the more recent past, which plays a dominant role in determining the type and rate of operation of geomorphic processes. In particular, under arid conditions, people can impact significantly on meso-scale landforms (Meadows 2001). The inability to differentiate between the natural and human-induced signals of the palaeoenvironmental record, as well as in some contemporary environments, is a major impediment to interpretation. This holds particularly true of studies focusing around the middle to late Holocene, as seen through difficulties in interpreting the causes of rapid sediment erosion or deposition (Thomas, 2004).

During the Late Quaternary the influence of atmospheric forcing has been recognised as a primary control on environmental change over the subcontinent (Tyson, 2000) and thus has been manifested by a variety of geomorphic features within the landscape of southern Africa. Arid and semi-arid landscapes appear to preserve more evidence of former environmental conditions, although high contemporary erosion rates and the paucity of long terrestrial sedimentary sequences hinder their complete elucidation (Meadows, 2001). With the exception of the arid and semi-arid regions of southern Africa, the degree of impact of Quaternary events at the landscape scale is relatively unknown, as larger scale landforms and their morphologies are more the result of long-term geological events, such as lithological, structural and geomorphological processes over millions of years (Meadows, 2001). However, at a smaller spatial scale the importance of environmental change during the Quaternary becomes more obvious.

Many geomorphologists have drawn attention to the rapid alternations of the Quaternary as an important source of variation in the landform components within regions (Derbyshire *et al.*, 1981). Some landscapes are more obviously a product of Quaternary, and particularly Late Quaternary, fluctuations and events, such as depositional coastal landforms and the aeolian features (used especially in semi-arid climate regions due to the lack of preservation of organic proxies (Chase & Thomas, 2006)). With the exception of xeric and coastal southern African environments, evidence for environmental change has to be sought in relatively small-scale features below the landform scale. Nevertheless, a wide array of forms of evidence are available including depositional and erosional forms of periglacial, alluvial, colluvial, fluvial, lacustrine, aeolian, marine, speleological and pedological processes; all are utilised to facilitate reconstruction of the palaeoenvironments of the Quaternary (Meadows, 1988). Meadows (1988) also reviews the types of evidence, associated implications and the limitations of palaeogeomorpic features as indicators of environmental change within the field of Quaternary studies.

The approach to geomorphological process within this context requires one to bear in mind that the process, material and resulting form are interrelated (Derbyshire *et al.*, 1981). The level of resolution for the climate-landform sequence is not considered very high, although the imprint of palaeoclimates on landforms within a region can often be very strong. For example, the gross landform units, such as valleys and interfluves, may be the product of geomorphic processes that are no longer dominant (or even present) in a region. Knowledge of landform genesis is founded mainly on the study of relict rather than currently active landforms; thus the climato- or morphogenetic geomorphology (the whole complex of landscapes made under past and present environmental conditions) approach remains an important framework for the study of landscapes (Derbyshire *et al.*, 1981). The basic assumption of climatic geomorphology is that climate, varying over time and space, controls the nature and incidence of geomorphic processes, which in turn combine to produce a range of distinctive landform units. Variations in the rate and occurrence of physical, chemical and biotic reactions at the Earth's surface impact greatly on the distribution of present bioclimatic and geomorphic processes, although there is a distinct problem associated with scale, as more climatically diagnostic landforms tend to be smaller features whilst more extensive features are related more closely to tectonic and epeirogenic factors (Derbyshire *et al.*, 1981). Different climates, through their effects on processes, produce unique assemblages of landforms, thus, following the theory that climate changed in the past and will continue to in the future, one can infer that climatically controlled landforms are superimposed on each other through time, however, as climate behaves complexly and landforms are difficult to quantify, the exact relationship between the two features is usually not clear (Moon & Dardis, 1988). Attempts have been made to apply more clear-cut parameters or climatic boundaries (mean annual temperature and precipitation) to some geomorphic process realms, such as mass movements, surface runoff and wind action, however, these remain as broad superficial generalisations and disregard the functional complexity of relationships at the Earth's surface hence exceptions occur (Ritter *et al.*, 1995).

Extensive literature has been published dealing with environmental response to climate change, especially in light of recent increased attention and awareness being paid to current and future predictions of significant changes in the global climate and extreme events ensuing. Such a vast body of literature surrounding this topic, especially considering renewed interest, can only be touched on briefly in this research, and for a more detailed synthesis and overview of geomorphic response to climatic change the reader is referred to such texts as Bull (1991), Derbyshire *et al.* (1980), Schumm (1979), Millington and Pye (1994) and Thomas (1997). A wide variety of literature related to geomorphic studies, in particular of inorganic stratigraphic sequences, used to aid in the elucidation of southern African palaeoenvironments through inferences from depositional sequences, have come into being especially since the 1990s, including specific studies, such as soils (Bond *et al.*, 1994), colluvial deposits (Boardman *et al.*, 2005, Shaw *et al.*, 2001), high altitude depositional environments (Marker, 1995), sand dunes (Stokes *et al.*, 1997) and fluvial deposits (Meadows & Holmes, 1999; Zawada *et al.*, 1996; Thomas, 2004).

The geomorphological impacts of Quaternary environmental change are less clear-cut than one would expect, especially in the largely unglaciated parts of the subtropics and tropics (compared to the rest of the world). For largely semi-arid southern Africa, where conditions were not conducive to glaciation during the Quaternary, unlike South America, the impact of fluctuations in precipitation has had a greater influence on the landscapes than that of temperature, although temperature change was still significant (Meadows, 1988). The impacts of these large amplitude fluctuations in precipitation on the physical geography and the differential response between regions of the subcontinent are inadequately resolved, especially with respect to the semi-arid and Mediterranean-type climate regions, where long Quaternary sedimentary sequences are scarce. The importance of these sites is underestimated as these landscapes bear the imprint of Quaternary changes more strongly than others (Meadows, 2001).

Geomorphic evidence for environmental change in southern Africa is considered on two temporal-spatial scales. Firstly, the long-term macro-scale development of the subcontinent, which is essentially reflected by the general form and relief of the current landscape, attributed largely to the breakup of Gondwanaland and subsequently controlled by phases of tectonic uplift leading to landscape rejuvenation and increased erosional activity (Moon & Dardis, 1988). There has been a shift in geomorphic research towards an improved understanding of the second scale, namely meso- to microscale with an increased focus on regional to local processes responsible for shaping the landscape during the Quaternary; although there are a number of problems associated with such interpretation, one of the more notable ones being the largely uncertain climate-landform relationship.

In terms of lithology and geomorphic process, a variety of sediments can occur in arid and semi-arid depositional environments. Such material may be derived from *in situ* weathering and be emplaced subsequently by one or more geomorphic agents; otherwise it may derive from an extraneous source and be transported into the study area by fluvial or aeolian action (Holmes, 1998). In both *in situ* and derived material the sediments may have undergone subsequent diagenesis or been exposed and altered by pedogenic processes. Sedimentological methods are important as indicators themselves and through entrained fossils and the fact that they can reveal long and continuous sequences. Geomorphic evidence can include glacial, periglacial, fluvial, marine, lacustrine, pedogenic sediments, duricrusts, colluvial, aeolian, speleothems, palaeosols and tephra. Late Quaternary sediments can yield valuable information regarding former climatic and environmental conditions derived from the nature of the sediments themselves and from the landforms they comprise, as the physical, chemical and biological properties of the sediments can be used to infer the environment of deposition, while stratigraphic relationships and contrasts show evidence of depositional changes over time; temporal and spatial changes in sediment accumulation can be heavily influenced by climate and hence, provide a proxy record of climatic change. This form of evidence though, does have the potential to be strongly influenced and disturbed by human activity, therefore an analysis of a sedimentary sequence can also aid as a proxy record of anthropogenically induced landscape change (Bell & Walker, 1996).

With the advent of improved chronometry of Quaternary events, through such methods as luminescence dating, inorganic sediments can be more readily dated and resolve a chronology from within arid landforms and have emerged as a viable source of evidence in palaeoenvironmental elucidation (Stokes, 1997). However, the absence of long, continuous sedimentary sequences of appropriate age remains a fundamental problem in the elucidation of Quaternary palaeoenvironments in southern Africa.

The response of dryland geomorphic systems to climate change can occur over a variety of timescales and short term changes are spatially variable and largely a result of differences in lithology and vegetation (Lancaster, 1998). For example, periods of dune construction or deposition are associated with arid phases. A result of extreme climatic change leads to an alternation of different types of erosion, such that one form of geomorphic process is replaced by another, a polygenetic system (Bull, 1991). Although erosion has been accelerated and may have occurred periodically throughout the geological record in response to changing climate and base level, in southern Africa, most of the currently observed soil-erosion phenomena, sheetwash, gully formation, etc., may be considered as essentially Quaternary manifestations.

Fluvial Response

Arid and semi-arid landscapes appear to preserve an increased amount of evidence from former environmental conditions although contemporary high rates of erosion and the scarcity of long terrestrial sediment sequences proves problematic with regard to discovery and explanation. The interaction of wind and water over time is considered as one of the key influences upon contemporary geomorphology and longer-term landscape development in more arid regions of the world (Nash, 2000). The hydrology of arid and semi-arid regions, despite the difficulties in determination of rainfall, streamflow and evaporation rates, have been targeted in palaeoenvironmental/climatic studies due to their high responsivity to even small forcing factors. Rivers are not simple systems of erosion and deposition governed by climate, but rather a complex balance of numerous environmental factors. These constituent controlling aspects include elements such as catchment size, rate of sediment production, vegetation cover, initial sediment depth, runoff, underlying geology, rate of weathering, initial slope, height above sea level, discharge, fire frequency and timing in the catchment area, land management practices, precipitation inputs, evapotranspiration, infiltration capacity of soils, temperature and tectonic activity (Moon & Dardis, 1988). Although there are so many factors, coupled with the inherent difficulties of dating, fluvial sediments have achieved some success in facilitating palaeoenvironmental reconstructions, especially when considered in tandem with other forms of evidence, such as palaeoecological evidence from pollen. Changes in the amount and type of precipitation alter the magnitudes and rates of weathering, erosion, transportation and deposition, thereby changing the shape and nature of slopes and rivers comprising a fluvial system within a drainage basin (Bull, 1991). A drainage basin is regarded as an open system landscape involving the interaction of mass and energy in a physical environment within a defined space through time. Many mechanisms are at work in a fluvial system, such as "cut and fill" cycles. These cycles refer to the erosion and subsequent sedimentation of new fluvial channels cut through existing fluvial sediments, and the sedimentary structures so formed (Briggs, *et al.*, 1998). Most alluvial dynamics are controlled by the competency or capacity of the river involved; less competency results in deposition and the river aggrading, this usually occurs in drier conditions with flash flood type events dominating any high energy transport, whereas in wetter conditions there is increased flow due to increased runoff leading to increased competency and erosion and incised channel formation. Therefore sediment accumulation is related to moisture availability although it is important to determine what time scale is being viewed. Erosion and deposition provides effective fluvial geomorphic palaeoenvironmental evidence for interpretation. Erosional, or incising, and depositional processes often occur simultaneously within different parts of a river's catchment. Although climate has no designated rule with reference to these processes, semi-arid ecosystems behave differently in conditions of variable precipitation inputs, as contrasted between more humid and more arid climatic phases of the Late Quaternary, this is mostly associated with relative amounts of groundcover and runoff. Downcutting decreases

FROM	TO HUMID INCISION	ARID AGGRADATION	ARID INCISION	HUMID AGGRADATION
Humid Incision	Frequent, high-magnitude events lead to active channel incision. Channels have low sinuosity and high hydraulic radii. Humid, zonal soils develop on the interfluves.	Valley aggrades with sediment eroded from nearby slopes. Channel sinuosity increase and hydraulic radii decrease. Floodplain deflation may be widespread. Basal unconformities often may be steeply dipping and irregular.	Unlikely to occur when humid and arid climates alternate rapidly and deposition will typically intervene. Perhaps no definitive features preserved. Soil development on interfluves reverses from humid to arid trends.	Valleys aggrade and humid-geomorphic floodplains develop. Channels lengthen, becoming more sinuous. Basal unconformities may be steeply dipping and irregular.
Arid Aggradation	Valleys incise, perhaps destroying earlier alluvial fills. Channels become less sinuous and hydraulic radii increase. Response may be complex until humid, zonal soils develop on the interfluves.	Infrequent, high-magnitude events carry slope debris to fill river valleys. Sediments have "arid" characteristics and variable facies development. Channels have moderate sinuosity and, probably, low hydraulic radii. Shallow, arid zonal soils develop on interfluves.	Valley fill incision, perhaps of limited magnitude in response to decreasing sediment supply. Channel sinuosity decreases while hydraulic radii may increase. Limited soil development on interfluves continues.	Channels and sediments acquire humid-geomorphic attributes. Channels probably become more sinuous and hydraulic radii probably increase (as sediments become finer?). Soil development on the interfluves shifts from arid to humid trends. Intervening unconformities are smooth and perhaps gradual.
Arid Incision	Incision continues without creation of definitive new features. Humid, zonal soils develop on the interfluves	Unlikely to occur without an intervening humid phase when soils develop, or a change in the sediment-delivery system causes by availability of a new, less resistant rock-type. Channel sinuosity probably increase, hydraulic radii may decrease Soil development trends and unchanged.	Infrequent, high-magnitude events incise relict alluvia. Adjacent hillslopes are nearly sediment free. Channels have low sinuosity and variable hydraulic radii. Arid, zonal soils may develop on low-anglel interfluves and relict fills	Valleys aggrade and humid-geomorphic floodplains develop. Channels lengthen, becoming more sinuous. Soil development on the interfluves shifts from arid to humid trends. Basal unconformities may be steeply dipping and irregular.
Humid Aggradation	Valleys incise, perhaps destroying earlier alluvial fills. Channel sinuosity decreases. Soil development trends change little.	Remodeling of floodplains by attributes related to arid aggradation. Channel sinuosity and hydraulic radii may decrease perhaps after initial incision. Soil development on interfluves shifts from humid to arid types. Intervening unconformities may be gradual and subtle.	Humid-geomorphic floodplains are incised. Channel sinuosity decreases, and hydraulic radii may increase. Soil erosion is active on the interfluves. Arid soils may develop on slopes with low rates of erosion.	Frequent, moderate-magnitude events aggrade floodplains; channel morphologies are related to sediment type. Vertisols develop on floodplains, while zonal soils develop on interfluves.

Table 3.2: Alluvial fill environments and transitions (Helgren, 1979)

Quaternary depositional environments do not reflect a simplistic response to climate variations as other triggering and forcing mechanisms are at work. It has been argued that it is too simplistic to simply consider climate as a sole, or even more important, factor in forcing changes to geomorphic systems, specifically from sediment accumulation to degradation (and vice-versa) in headwater valleys. Deposition and erosion does not only occur under one particular set of climatic conditions with more detailed studies of Quaternary valley fill stratigraphy showing that it does not simply reflect a response to changing climate (Holmes, 1998 citing Prosser *et al.*, 1995). Studies of Quaternary colluvial and alluvial depositional environments in Australia proposed and demonstrated that the sequences could be explained by periods of stability followed by brief phases of instability, during which slopes and headwaters experienced accelerated erosion whilst valleys experienced synchronous aggradation; buried soils indicated periods of stability thus reflecting alternating cyclical stability and instability, known as K-cycles (Holmes, 1998 citing Prosser *et al.*, 1995).

The traditional approach to valleys is to view them in terms of erosional and depositional sequences resulting form channels acting within the valley network and are usually considered a result of major climatic and environmental changes, however, as recent studies in southern Africa have shown, the climatic shifts were not as large as previously implied, therefore other factors besides fluvial activity under more humid conditions within semi-arid regions could be responsible for their development (Nash *et al.*, 1994). There is a comparative lack of data for river channels and sequences within drylands as opposed to temperate regions, hence, their development is less well understood (Reid & Frostick, 1997). Research on dryland valleys, especially those that are ephemeral, fossil or contain misfit channels, generally infers formation due to erosion during periods of excess available moisture with sufficient surface water to maintain more permanent rivers, therefore, by inference, such rivers are considered indicative of large-scale changes in climate and are accorded palaeoclimatic significance (Nash *et al.*, 1994). However, as previously mentioned evidence from other landforms and proxies has pointed towards the Late Quaternary climate not fluctuating as dramatically in southern Africa but rather oscillating around a semi-arid mean.

Intra-valley forms are indicative of the role of fluvial activity within valleys, including features such as terraces, evidence of former channels and lag deposits. Sedimentary evidence of former flow can be more difficult to come across and can require remote sensing. Lag gravels and point bar deposits, containing clasts, can be found on the inside bends of river channels and can suggest either more significant perennial flow or high-magnitude flooding whilst relict meandering or abandoned channels, sometimes incised, can be located along the flanks of indistinct wider valleys, such as those studied in the Kalahari valley network (Nash *et al.*, 1994). It is, however, incorrect to oversimplify a more complex series of processes by assuming that the development of specific landforms can only be associated with either former wetter or drier periods as well as the necessity to contemplate the overall timescale and interplay of such processes. River courses and their associated relict channel bars and overbank deposits have received some consideration within southern Africa, however, their dependability as palaeoenvironmental indicators remains problematic as fluvial depositional environments are considered a function of their sediment supply, sediment properties, channel morphology and fluvial dynamics leading to complex interactions that complicates accurate interpretation (Meadows, 1988). There is also a difficulty in recognising contemporary from palaeodeposits within fluvial environments. Studies of modern arid and semi-arid rivers show that interpretation of fluvial evidence is not 'clear cut' (Thomas, 1997). As flow regimes of streams in drier areas are rarely constant and fluctuate regularly, channel forms and processes are not in equilibrium; transport of coarse sediment is very irregular, while it is also likely that the major determinant of channel morphology is the size of infrequent, large floods either causing channels to be flushed out and incised, floodplains to be stripped or channels to be clogged by large quantities of sediment from neighbouring slopes.

Overall, there is no definitive relationship between form, process and rainfall readily identifiable within more arid environment streams, making it very difficult to interpret recent form and perhaps even individualistic of different drainage basins. The scenario is further complicated by the role of base-level changes. Thomas (1997) controversially states that studies have also proved little in terms of any linkage between channel and slope systems, with overall, rivers being relatively insensitive to all but major changes in climate, while the sensitivity to climate fluctuations may be a reflection of the range of flows and floods experienced in an individual climatic regime.

Slopes, under semi-arid to arid conditions, when vegetation cover is sparse, have sediment transport that may be more active on the slopes than channels, favouring the accumulation of thick, lower-slope colluvium aprons that leads to interpretations of more widespread semi-arid conditions in the region. Alluvial fans may also act as buffers between arid zone slopes and fluvial systems with their sediments providing a valuable record of environmental changes in source areas. The proposition that alluvial fans aggrade more favourably under conditions of reduced vegetation cover has been made, however, palaeoenvironmental interpretation of fan and slope deposits may be as contentious as that of channel features and sediments, although systematic analysis of sediment changes and inferred flow variations may enhance interpretations (Thomas, 1997). Longer-term climatic change has been linked to changes in aggradational and dissectional behaviour in alluvial fans. In the Mediterranean Quaternary, more arid cold conditions with little precipitation and high seasonal hillslope erosion resulted from intense seasonal rainfall, leading to periods of major alluvial fan aggradation, followed during the Holocene by fan dissection (Harvey, 1997). Aggradation was associated with moist phases whilst dissection was with drier phases in Australia. According to Harvey and Wells (1994) (cited in Harvey, 1997), more arid conditions prevalent in the Holocene have led to fanhead trenching and distal progradation as dominant trends. Fluvial terraces are also of importance when considering environmental indicators. The most notable variable to bear in mind with regard to terraces is the adjustments in base-level that occurs as a function of tectonic activity, such as uplift or downthrusting, or eustatic changes in sea-level. Past evidence, as quoted in Moon and Dardis (1988), shows that the documentation and analysis of Vaal River terrace phases of cutting and construction correspond with periods of Quaternary aridity and humidity. Palaeohydrological studies of rivers can show temporal changes in river regimes along with changes in river erosion and deposition, sometimes allowing for inferences about palaeodischarge and palaeovelocity in former river channels; however, late Holocene alluviation is often closely related to human activity (rather than climate change). Palaeodata provides important information about the past frequency, intensity and subregional patterns of change in the world's deserts that cannot always be captured by the climatic models. Palaeoflood records provide information that is useful for better interpretation and provide unique event-scale information that is constructive in calibrating and testing geophysical models of past and anticipated future climate conditions (Knox, 2000). In the context of climatic change, there is growing interest in the study of past floods. Researchers are attempting to reconstitute the chronology of palaeofloods, with respect to the frequency and magnitude of ancient floods using several techniques to obtain the long-term chronology of flood events in relation to the specific conditions (e.g., climate, geomorphology) of a region especially in light of interest generated by global climate change and its effect on river-system dynamics (Saint-Laurent, 2004).

cover, precipitation levels and temperature regimes, thereby holding a palaeoenvironmental record relating to climatic and land-use change in the stratified sequence of hillslope sediments. Alternatively accumulation of colluvium is associated with arid phases (occasionally colder) with reduced vegetation cover and subsequent landscape instability (Marker, 1995).

Weathering and erosive processes produce particles of various sizes, which are transported and deposited under varying forces. The characteristics can be determined and as weathering, erosion and deposition are related to climatic variables, such as humidity, rainfall and temperature, interpretation of past environments can be made. The influence of erosion, as discussed by Gale and Hoare (1991), shows that the mechanisms have various capacities to entrain particles. These particles will be deposited when the erosion action loses its competency to maintain the level required to transport the particles. Various particles will have specific boundary conditions and as a result will be deposited differentially. However, it is important to realise that while one can be certain correlation of grain size with environment (Folk, 1974), grain size is not only determined by externalities, such as parent material.

There is no doubt that soil characteristics affect geomorphic processes in a number of significant and noteworthy ways and should not be underestimated, most conspicuously where specific or unique characteristics dictate, or at the very least, influence surficial processes. The identification of such an attribute as soil properties provides critical information with regards to specific environmental controls (Ritter *et al.*, 1995). Soil characteristics vary or are altered over time and through climatic changes, and such variance in soil formation and weathering plays a large role in indicating the sequence of events through the Quaternary, the last 2.5 million years, such as glaciations, relative ages of deposits, and as evidence of climate change or variability (Plummer & McGeary, 1998). Soil forming processes also change with time and factors are not constant due to this variability, therefore soil profiles preserve and can exhibit more than one forming or pedogenic factor and are termed complex or polygenetic soils (Ritter *et al.*, 1995). The varying thickness and maturity of different soils are reliable indicators of inherent developing conditions. The character of soil profiles varies with parent material, climate, biota, topography and the length of time involved in its formation. The dominant pedogenic (soil forming) regimes are podzolization, laterization and calcification to produce the major soil groups (Holmes, 2003). $\delta^{13}C$ values in soil profile organic matter or palaeosols reflect total plant cover at the time of burial and soil formation and organic residues and can easily demonstrate shifts in vegetation type. At advanced stages of decomposition, soil organic matter is enriched in ^{13}C by approximately 1-2‰ relative to the overlying flora (Koch, 1998). Climate is one factor in sediment and soil formation that is extremely variable and complex, therefore this fact coupled with South Africa's complex climatic pattern and high degree of environmental heterogeneity makes the interpretation of sediment within South Africa slightly more difficult and only locally relevant rather than of more regional use, although more general and meso-scale inferences can be drawn from the results.

There is an important overall relationship between climate, vegetation cover and sediment yield. However, there is a particularly strong connection between vegetation and geomorphological process, which in turn, is strongly correlated to climate. This relationship has often been noted as requiring further investigation (Millington & Pye, 1994). Vegetation is a useful indicator of climate change as vegetation is highly sensitive in response to climate shifts and in conjunction with resultant geomorphological processes, is a fundamental component to consider in recognising development and alteration of landscapes, especially within a semi-arid context. The geomorphological role of vegetation is not limited to microtopographical features, such as plant mound formation due to aeolian accumulation or rainsplash erosion and selective erosion of intershrub areas by overland flow, and interactions with runoff and sedimentation processes. It also plays a role in floodplain and channel morphology hence affecting the development of landforms in its immediate vicinity but also indirectly influencing landscape development by regulating the amount of water and sediment available, offering insights into the geomorphic consequences of vegetation removal or change within a region (Bullard, 1997). The effect of vegetation on the land surface has a number of implications for landscape development at a variety of scales.

The role of groundwater as a geomorphological agent has long been recognised. Subsurface water is an important factor in the processes of weathering, soil development and hillslope stability and as a component of river discharge, although, its influence in sculpting landscapes outside of carbonate rock localities has been underestimated (Nash, 1997). It is an important factor to consider in semi-arid to arid environments and to bear in mind over longer timescales. Within fluvial systems, it is possible that surface flows have been more prevalent during former periods of wetter climate with the balance between groundwater and surface erosion processes fluctuating with time (Nash, 1997). Seepage erosion may be an extremely important landforming process under semi-arid conditions, although inherent thresholds seem key to processes with different modes of operation under different lithological and climatic settings. Processes associated with groundwater, such as weathering and seepage erosion, depend on seasonality and varying climatic regimes as they comprise complex feedback mechanisms where wetter climates can be capable of either increased or decreased erosion depending on different factors (Ritter *et al.*, 1995).

Thresholds and feedback mechanisms

Feedback mechanisms are integral and fundamentally important, especially in terms of the temporal scale. Various elements or factors are interlinked, such as climate, geology, soils, vegetation and fauna, and the impact of one of these factors can have repercussions throughout the system including a feedback effect on the original factor. The system can be negative, where the effect of the original change is reduced leading to stable equilibrium conditions being resumed (this form is often self-limiting and ensures an equilibrium is maintained with small changes within certain defined limits), or it could be positive, where the consequences of the change are reinforced (usually as limits are exceeded). In systems, even relatively minor changes can have a disproportionately large knock-on effect, particularly as thresholds, even critical thresholds that set new equilibriums, are crossed (Bell & Walker, 1996). Through feedback mechanisms, the changes in landscape cause still more changes in processes and can require up to 10 000 years for the system to make a complete adjustment to a climatic perturbation; conversely climate change occurs with such frequency that fluvial systems are continually adjusting (Bull, 1991). A stream in equilibrium is one in which, over many years, slope, velocity, depth, width, roughness, pattern and channel morphology carefully and mutually adjust to provide the power and efficiency necessary to transport the load supplied from the drainage basin without aggradation or degradation of the channels and attaining a stable longitudinal profile (Bull, 1991).

A geomorphic threshold in an open system separates different reversible processes that tend to change part of the system. There is a distinction between intrinsic and extrinsic thresholds. Intrinsic thresholds occur as landforms evolve to a condition of incipient instability following which change or failure occurs; the response of a system to an external influence is referred to as an extrinsic threshold. A threshold exists within a system but will not be exceeded and change will not occur without the influence of an external variable, such as anthropogenic disturbances or climate change. The distribution of erosion, such as gullies, within semi-arid valleys is largely controlled by geomorphic thresholds, such that sediment accumulates on valley floors until a threshold gradient is reached, above which erosion or entrenchment occurs. Gullies are generally associated with accelerated erosion and landscape instability (Morgan, 1995). In tectonically inactive semi-arid regions, dynamic equilibrium reached after aggradation that has raised a streambed is followed by a degradation event that re-establishes the same longitudinal profile that was present before the aggradation event; Holocene channel downcutting rates appear to be directly related to the size of the flood discharge and the total annual stream power with degradation accelerating to a maximum and then decelerating as a new base level is approached (Bull, 1991).

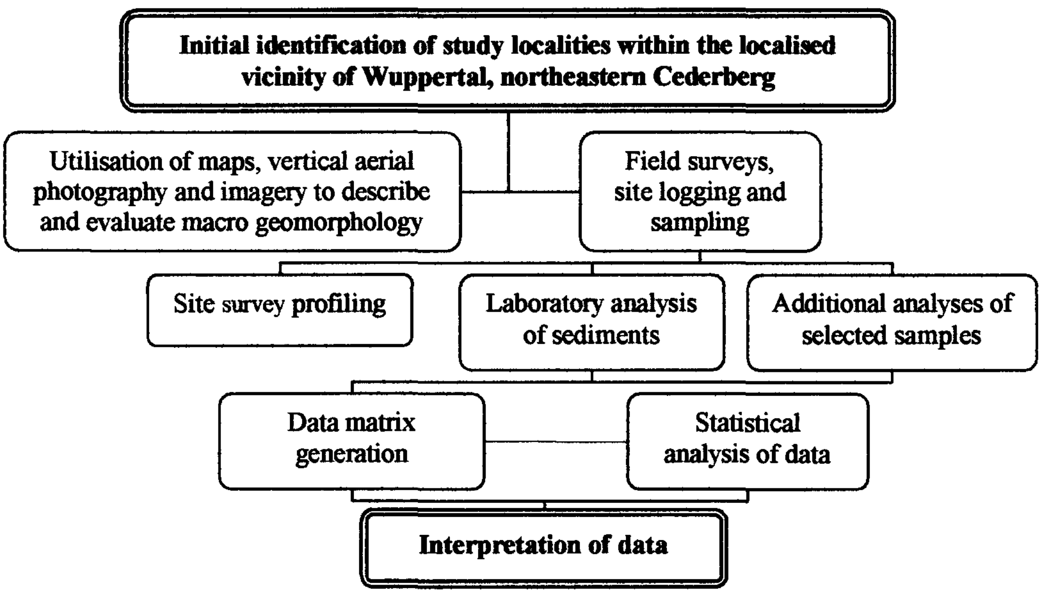
Geomorphological evidence offers a viable starting point towards the elucidation of Quaternary environments through inferences from modern analogues that can lead to a range of generalised conclusions or more specific deductions based on climatic parameters. It is essential, though, to incorporate the stratigraphic record with the landform evidence in order to establish a more coherent, detailed and particular or explicit picture towards palaeoenvironmental reconstruction. In the context of applied geomorphology, it is becoming more urgent that ecological and geomorphological responses to environmental change are understood at a variety of temporal and spatial scales (Millington & Pye, 1994). Knowledge of the timescale over which environmental changes occur is critical to an understanding of nature, cause and relationship.

Furthermore, studies of geomorphic reactions to environmental and climatic changes are integral in present-day society as they aid in disaster risk assessment and mitigation and are especially pertinent in populated or urbanised areas as well as in terms of buffeting the effects to rural agricultural or productive land, which have essentially made the land surface more sensitive and vulnerable to more dramatic geomorphic change. In a global warming scenario the effects of changing climate on geomorphological and ecological processes and functioning must be understood, predicted and managed. There is a wide range of geomorphological responses to different types of environmental changes or disturbances, however, comprehensive analysis and collaboration is required from multidisciplinary research involving not only geomorphologists but also ecologists, biogeographers, hydrologists, climatologists and remote sensors (Millington & Pye, 1994). Environmental change can be examined across a range of spatial and temporal scales, from global (macro-scale) to local (micro-scale). Natural environmental changes manifest themselves in a variety of ways and over a wide range of timescales. Within this general framework, however, it is possible to distinguish those processes that are long-term and gradual from those events, which are sudden and frequently catastrophic (Bell & Walker, 1996). Processes include climate change and soil formation whereas events range from major storms to earthquakes. It is, however, noteworthy that long-term processes may comprise of multiple of superimposed events.

Chapter 4
METHODOLOGY

4.1. INTRODUCTION

The methodology followed in this project involves a range of interdisciplinary techniques, including surveying, physical observations, statistical analysis, characterising and dating of sediments as well as spatial analysis, which were carried out over the course of the study. Although a number of limitations are apparent, an attempt to reduce human error and maintain a level of objectivity was made at all times to prevent any possible misinterpretation and to minimise error and inaccuracy within the measured and calculated results. This chapter outlines the methodology employed within this research project. The majority of the methods employed follow the protocols of the United States Geological Survey (USGS) and British Geological Survey (BGS). All reported units are SI relative to the universal standard if applicable. In addition, discussions regarding lithostratigraphy and chronostratigraphy adhere to the standards of the International Union of Geological Sciences, the International Commission on Stratigraphy (ICS) (http://micropress.org/stratigraphy/) and the South African Commission on Stratigraphy (SACS). References to Erathem Era, System Period, Series Epoch and Stage Age are based on the 2000 edition of the International Stratigraphic Chart (Appendix 1). Broadly, the procedural approach to this study can be considered as follows:



from the Tra-Tra Valley (TT) and one from the Dassieboskloof River (DBK). Prior to sampling, the valleys were first assessed using aerial photography and ground-truthing before suitable sites were chosen in applicable locations that best represented accessible contemporary channel deposits. These samples were collected by grab sampling the loose unconsolidated (sometimes water-saturated) sediment *in situ* (from the surface). Although there could be a large discrepancy with regard to where the sediments were collected, taking into account differential wind deflation, settling, rill erosion, small scale slumping, etc. generally the advantage of this procedure is that the exact method was replicated at each point, thereby creating a high level of comparability for the sediments as well as allowing for the sampler to relate the specific sample position to the general surface and depositional or erosional conditions of the immediate vicinity, therefore also permitting the consideration of specific conditions or abnormalities when analysing and interpreting the resulting data. Two MG samples were taken from active fluvial channels, while two other sites were selected along the river to sample a transect of different geomorphic features across the river (labelled A-D). These structures included contemporary inactive, side and active channels, channel bars and slackwater deposits. At each site, detailed field notes regarding any significant features and micro features were taken as well as photographs. TT and DBK samples were also taken as contemporary sediment from active fluvial channels.

The selection of appropriate study site locations most 'representative' of a depositional sequence and the choice of which section of vertical profile was sampled is largely subjective (Holmes, 1998). Samples taken in profile originated from both the Tra-Tra Valley and the main fluvial terrace exposure, WT. Prior to any actual sampling, the exposed horizons were first cleaned off to remove any slumped material or debris and a cutting made back into the face to uncover 'fresh' surface. The second sample site from TT was at a secondary lateral channel sandbank profile that was exposed and the surface cleaned to present water level, i.e. 1.37m. Samples were taken at a resolution of 20cm intervals. The main WT exposure was divided into three sampling profile sites. WT3 was located above the primary site (WT4) and represented an exposed profile uncovered by local sand mining activity. The exposure was cleaned and further excavated down to reveal a face of approximately 2.5m. This profile revealed alternating horizontal bedding of sand with dark grey/paler grey wavy laminations. Samples were taken at either a 20cm interval or at a change of texture or colour. Laminations were also sub-sampled. WT4 presented the main exposure and the primary slope of the fluvial terrace. Again the upper metre of face was cleaned and further excavation was undertaken down to the contact of the alluvium, or upper grey package, with the colluvium, or lower orange package, exposing a profile of roughly 3.5m (later surveyed). Planar bedding was again prominent with the upper 2m exhibiting more numerous laminations. Samples were taken at 20cm intervals or where lithological changes occurred; laminations were sub-sampled. WT5 was an already exposed 3m face of colluvium (orange package) with no excavation necessary, although the face was cleaned to expose new material for sampling. Samples were taken where changes in colour and appearance occurred within the profile. The faces of profiles were cleaned using either a spade or geological hammer to expose a fresh face and a trowel was used to extract samples. All samples were bagged and labelled on site, limiting contamination, for later laboratory analysis and ranged in weight from less than 100g (laminations) to 500g, depending on texture, and stored in airtight zip seal bags. A hard pedogenic substrate or chemical cementation was present towards the base of the terrace at WT and included in the samples of WT5.

Detailed field notes, logging and descriptions of the beds, in terms of their sedimentological and pedological characteristics were made along with diagrams and scaled photographs. GPS co-ordinates were taken at the location of each site. Part of the fieldwork included the creation of an extensive photographic catalogue of the three main valleys and all sites and any notable sedimentary and geomorphic features present within the vicinity for visual reference. Sedimentary structures, the larger scale three-dimensional features of sediment bodies, were also studied to ascertain characteristics of stratification i.e. the banding and layering of sediments.

4.2.2. Surveying

Primary data capture to create survey profiles for the total height, relative position and contact depth of the main WT exposure, or terrace, was conducted in the form of ground surveying, which uses measurements to determine the locations of objects (Longley, 2001). To ascertain the relief and relative three-dimensional (3D) positions of points in order to create a profile, a tripod-mounted electro-optical (Leica) Total Station Theodolite was utilised. This measured angles and distances relative to a static location or benchmark to an accuracy of 1mm (Longley, 2001). Although this is a very time-consuming activity, it is the optimal means to obtain highly accurate point locational data from which to create a series of profiles. This aspect of fieldwork involved two people, one to control the total station theodolite and another to hold the reflective prism at the point which was to be measured. Six profile positions, labelled A to F and given GPS co-ordinates, were chosen along the horizontal axis of the exposure on the southern bank of the Dassieboskloof River, beginning from the eastern corner to the bedrock outcropping at the channel constriction located at the western extremity of the WT exposure. Another reading was also taken on the north banked levee, labelled G. Profile positions were decided based on relative changes in alluvium/colluvium depth and position as well as those sites least disturbed by any human means. The theodolite position, at a high point on a channel bar to the south of the main stream, was regarded as the benchmark and considered to fall at an elevation of (relative) 0m. Readings of elevation and distance were taken relative to this fixed position. An extendable pole with the surveying reflective prism attached was maintained at a constant height of 2m that was held upright (with the aid of a spirit level) at points along the profile. These points were chosen according to the 'break of slope' method rather than the standard interval method (Nordstrom, 1986). A reading was taken along the profile at each significant or sudden change in gradient. Overall, the profile was, to a degree, generalised, i.e. not every incidental change was recorded but rather more noticeable or substantial ones. Points were measured from the top to the base of each transect and a reading was also taken at the contact between the two packages, and later alluvium and bedrock. This was so as to develop a more characteristic and graphic visual representation of the exposure and its constituent packages of alluvium and colluvium.

A database consisting of the series of readings at each transect alongside height and distance measured by the theodolite and relative to its position was created. Any reading taken below the relative height of the theodolite itself was considered as a negative value. The values obtained were entered into an Excel spreadsheet and using the difference between individual readings and adding them cumulatively, profile values in $(x;y)$ format were calculated, starting from the origin. They were not standardised to a specifically chosen point for elevation (e.g. highest or lowest reading at each site) as this varied according to the position and depended on the amount of fluvial sand and debris accumulated at the base of the terrace talus slope. The series of points for each profile was then plotted as XY scatter plots with points joined by smoothed lines to show the profile as well as corresponding values joined horizontally to represent base, top and contact position within the exposure to adequately represent reality. Profile G (the opposite bank or levee) is plotted as a negative value with reversed slope (i.e. behind the benchmark) to show it as the north bank. A problem arose as these points were plotted on a two-dimensional Cartesian plane, while in reality, the three-dimensional position relative to the theodolite shows that the exposure (A-F) is banked in the shape of an arc around the benchmark rather than a straight line measured from the static point or origin (0;0) as was assumed. The position of the total station theodolite was assumed to be a fixed point representing the "origin", meaning an elevation and distance of zero; however, it was actually a central point with some measured points below it. For the sake of creating the plots, the distortion issue was ignored and lower elevations given a negative value. This issue was taken into account for interpretation and discussion of the profiles although it does not impact on the credibility of the measurements, only making the profiles more difficult to read.

Due to the nature of the WT4 profile site, being on a very steep (although not vertical) slope with pebble and cobble lags, the need arose to excavate a further 1.5m down to the contact, although it was awkward to access in comparison to the reasonably accessible profiles WT3, TT2 and WT5. Therefore, an alternative method was devised to sample depth in this case. All other profiles could be approached from ground level up with a ladder and were easily cleaned and cut to a sharp vertical face for relatively simple sampling and dependable measurement with a tape measure held flush against and secured to the exposure. Initially a tape measure was secured to the face of the WT4 exposure to use as a rough estimate for sampling resolution and bed placement, however, once the total number of samples had been collected, the total station theodolite was set up below the site at a fixed position from which to survey the sampling points. These values were used to calculate the real depth of each sample and replaced the previously measured sample names in most of the data sets. However, due to the fact that luminescence samples were taken on a previous field visit and their depth position measured according to a tape measure, it was necessary to retain the initial measurements of the WT4 profile in order to place the OSL dates correctly relative to the bedding in the exposure.

4.2.3. Mapping and Remote Sensing

Geological, topographic, thematic and tourist maps, orthophotos, aerial photos and satellite images at various scales were used in order to develop an understanding of the relationship between the geomorphic context of the primary Wuppertal exposure, the four main water courses within the region, namely the Dassieboskloof, Kleinhoog, Moordenaarsgat and Tra-Tra Rivers, and possibly the processes responsible for the initial deposition and subsequent erosion and to visualise the contemporary fluvial environments on a small to large spatial scale. 1:50 000 topographic maps of the Wuppertal (3219AC Wuppertal), Pakhuis (3219AA Pakhuis) and Biedouw Valley (3219AD Grootberg) areas as well as aerial photographs of the north eastern Cederberg, South Africa (ranging from 1960 - 2003), were obtained from the Chief Directorate of Surveys and Mapping, Mowbray, South Africa. A geological 1:250 000 map for the west coast/Clanwilliam district, South Africa, (3218 Clanwilliam) was obtained from the Council for Geoscience, Bellville. Supplementary thematic maps of the watersheds of southern Africa (ALCOM, FAO and WWF, 1997) provided valuable information and DWAF maps of the Olifants/Doorn Water Management Area (WMA) for more regional drainage basins, land use, groundwater, topography, rainfall and evaporation, hydrogeology and dam sites. The tourist maps and accompanying information of the Cederberg were also useful.

Remote sensing imagery was obtained from a number of sources, the principal contributor being Google Earth (http://earth.google.com), MODIS (http://rapidfire.sci.gsfc.nasa.gov/), composite LANDSAT 5 images (https://zulu.ssc.nasa.gov/, http://geo.arc.nasa.gov/sge/landsat/daccess.html) and NASA Visible Earth (http://visibleearth.nasa.gov). Additional LANDSAT 7 images were obtained from the USGS (http://landsat7.usgs.gov/index.php). LANDSAT image S-34-30 was most useful in particular. Aerial photographs (panchromatic black and white vertical aerial stereophotographs at varying scales) were extremely valuable aids and utilised often for geomorphic and basin morphometry interpretation, having a superior resolution to remote-sensing imagery, as well as a base for diagrams, orientation and sketch. Topographic maps were used for cross sectional profile creation, gradient calculations and diagrams as well as for possible palaeodrainage inferences.

4.3. PHYSICAL AND CHEMICAL SEDIMENTOLOGY

4.3.1. Description

Laboratory analysis of Quaternary sediments is an integral part of palaeogeomorphological and environmental investigations as, both the physical and chemical properties of sediments supply data on the nature of former depositional environments that are often useful indicators of climatic and environmental changes (Lowe & Walker, 1997). It is standard practice in any microgeomorphological investigation and aids in elucidation of deposition and subsequent modification of sedimentary features (Holmes, 1998). This study incorporates routine laboratory analysis of sediment samples collected in the field, including physical analysis to determine textural properties, as well as limited chemical analysis for the further classification and description. In total, 66 sediment samples were collected for analysis and all subjected to the same preparation and analytical procedures, except where constraints led to selective analysis (i.e. XRD and EC). Results of physical and chemical analyses were used to compile a comprehensive data set.

A combination of consolidated, unconsolidated and partially consolidated sediment samples were collected from logged profiles and surface contemporary fluvial systems during June 2005, December 2005 and March 2006 using standard geomorphological protocols (Goudie, 1994). Details of sample location, profile thickness, number of samples collected and geomorphic setting are shown in Appendix 3. TT1, DBK and all MG samples consisted of unconsolidated sandy material. WT5 samples were consolidated and sometimes cemented belonging to the lower orange package of the main exposure. WT3 and WT4, along with TT2, were partly consolidated bedded sandy material with root, leaves and stem fragments in the upper substrate and charcoal occasionally present within certain beds. Descriptions of site profiles and samples follow in Chapter 4.

4.3.2. Grain Size Analysis

4.3.2.1. Introduction

Particle size analysis is an essential diagnostic property of sediments as any changes in average grain size or the range of grain sizes, clasts and content through a section or between horizons may reflect significant changes in the sedimentary environment responsible for their deposition. Grain size is a fundamental attribute of siliclastic sediments and important when considering descriptive properties, in particular, the techniques for measuring and expressing it in terms of a graded scale; presenting them in a graphical or statistical form; and analysing their genetic significance (Boggs, 1995). Grain size analysis is central to interpreting the energetics of clastic sedimentary environments (Syvitski, 1991). Particle size distributions in sediments are a function of the parent material from which the sediment is derived, the type of transport and processes since deposition it has been subjected to (Holmes, 1998 citing Leeder, 1991).

Particle size distribution should reflect as closely as possible the granular composition of the sediments as they occurred in the field, therefore pre-treatments were minimised and preparation methods standardised. As all samples were taken from a fluvial environment characteristic of the upper to middle reaches, moderately sorted sand was expected for the majority of the samples, excepting the colluvium (Briggs *et al.*, 1998). Particles in sediments can range from a few microns to a few metres; therefore sizes are expressed on logarithmic or geometric scales and based on the method proposed by Udden (1914) and Wentworth (1922, 1929), where each successive size frequency covers twice that of the previous (Boggs, 1995). This scale was modified by Krumbien in 1934 for graphical plotting and statistical calculations and expressed as the logarithmic phi scale, based on the following relationship: $\phi = -\log_2 D$, where ϕ is phi and D is the grain diameter in millimetres; negative values of ϕ correspond to particle sizes coarser than 1mm and as the ϕ unit increases, so the size of the particle decreases (Syvitski, 1991).

Particle size analysis was performed using both wet and dry sieving techniques and respective measurements, a computerised settling column and, in the case of clay and silt, or fine materials, the pipette method. Between 50 and 70g of sediment was sub-sampled using the mechanical splitter or riffle box. The sample to be analyzed was placed in a pre-weighed 100 ml beaker, weighed, and dried in a convection oven at 40°C so as not to destroy any clay structures. When dried, the samples were placed in a desiccator to cool and were then re-weighed. The loss in weight due to water evaporation was taken to be the equivalent of the original moisture content. Because, however, some of the samples had been stored for a few weeks while others were water-saturated at the time of sampling (active channels and sandbanks) and some were taken the day following significant rainfall, this content was considered unreliable and not representative, therefore not included in any calculations.

In arid environments, salts tend to accumulate at or near the surface; therefore there is the need to remove them to prevent any skewing effect they may occur depending on their relative amount within the samples (as it can lead to adhesion and incrementally distort the sample weight) (Rogers *pers. comm.*). This is achieved through the use of cellophane dialysis tubing, which, when filled with distilled water washed sample and submerged in water, allows the process of osmosis to draw the salty water (and hence salts) out of the tubing. All MG and TT, DBK and WT5 samples were subjected to this form of dialysis, and wet-sieving and pipette method of analysis. For this set, a sub-sample of 5 to 10g of homogenised dried grain size sample, was washed into dialysis tubing, tied closed and submerged in a bucket of fresh water for at least four hours. Upon removal, the sample was agitated (to assure the mud fraction was in suspension) and washed out of the tubing using tap water into a 63μm sieve and collecting bowl. It was wet sieved to remove the mud fraction from the sand and gravel fraction. The suspended mud fraction was collected and decanted into a 1000cc cylinder. The coarse fraction was washed into a beaker, allowed to settle and the excess water and floating organic matter decanted off. This was then placed in the oven at 70°C to dry overnight. The fine fraction then had precisely 5.0g of the deflocculant calgon (sodium hexametaphosphate – $NaPO_3$) added and thoroughly agitated to prevent flocs aggregating and skewing clay to become silt sized.

4.3.2.2. Fine Fraction Analysis

Granule- to silt-sized particles can be measured by sedimentation techniques on the basis of the settling velocity of the individual particles. The theory states that particle size relates to settlement rate, as "particles will settle in a fluid at a rate dependent upon their shape, size and density, and the density and viscosity of the fluid" (Gale & Hoare, 1991). In the case of clay and silt, the grain size can be determined based on sedimentation methods utilising Stoke's Law where the settling velocity of small particles can be measured at a particular temperature through calculation of the particles' diameter using an arrangement of the mathematical equation (Appendix 4). Pipette analysis separates out clay and silt as samples taken at a particular depth in the fluid after a specific time will only contain particles that are finer than a given hydraulic equivalent diameter. This law, however, does assume that the settling particles are spheres, therefore, as most particles are not, this usually results in the analysis producing finer skewed values than the actual particle sizes (Boggs, 1995). It assumes the particle density of quartz.

The fine fraction (< 63 μm) was analyzed using pipette analysis. To alleviate biologic or chemical changes, storage of the fine fraction in containers prior to analysis never exceeded five days. Firstly the pipette was prepared and the calibrated pipetting factor ascertained as 43.853. The cylinder containing the mud fraction was filled so the base of the meniscus rested on the 1000ml mark Pre-weighed beakers were labelled for silt plus clay, and clay. The mud and water mixture was agitated again for exactly one minute continuously to create a homogeneous suspension. The pipette was then lowered into the cylinder. As the clay/silt boundary used for pipette analysis was 4μm, the time elapsed between the first and second sampling was one hour. The 25ml aliquot first sample was emptied into the silt plus clay beaker; precisely an hour later the second 25ml sample, at a predetermined depth dependent on the temperature of the suspension, was released into the clay beaker. The ambient temperature of the suspension is vital to this form of analysis as it affects viscosity and thus settling rate. Temperatures at the time of analysis ranged from 24.5 to 22°C. Both beakers were then placed in the oven at 75 to 80°C to dry and then weighed. To establish the mass of clay and silt, the resultant weights were multiplied by the pipetting factor and the 5g of calgon subtracted (silt is calculated by subtracting the clay value from the total fines dried in the first beaker).

A disadvantage associated with this method of clay and silt analysis is that if the fine fraction of the sample is very small, results rendered by the method are void (Rogers, *pers. comm.*). In this case, there are numerous samples with almost negligible quantities of fines, therefore there was no need to further split the fine fraction.

4.3.2.3. Coarse Fraction Analysis

The remaining raw bulk samples were disaggregated on a shaking table for 15 minutes using a solution of 10% calgon (sodium hexametaphosphate) in order to deflocculate the clay fraction or eliminate inter-particle cohesion. The solution was then wet-sieved through a number 230, (63 μm; 4ϕ) sieve using distilled water to separate the coarse and fine fractions. The fine fraction was collected in a pre-weighed beaker and placed in the oven at 75°C to dry and then re-weighed to give the total fine fraction weight. The coarse fraction was rewashed in tap water, reintroduced into a pre-weighed beaker, dried in the convection oven at 75°C and then reweighed. The weight of the coarse (> 63 μm) fraction is equal to the weight of the sand plus gravel. The weight of the fines (silt and clay) can also be calculated and checked by subtracting the coarse weight from the sample weight. The coarse fraction was then put through a 2mm (-1 ϕ) sieve to separate out gravel from the sand fraction. Each fraction was weighed and respective components noted. The sand fraction was then split using a fine splitter (to avoid mechanical sorting and ensure a random representative sample) until a sub-sample of 2-3g was obtained. This sample was used for the settling column.

The settling column or tube also uses the principle of grains settling through a column of water at a specified temperature in a settling tube and measuring the time required for the grains to settle (Figure 4.1) (Boggs, 1995). This method, based on empirically related settling times to a standard size distribution curve to obtain the equivalent phi or millimetre size, does, however, also assume spherical particles, although in reality grains are not perfectly spherical and non-spherical particles settle slower. It also assumes the fall velocity distribution of particles is neither hindered by other particles nor involved in convective plumes (Syvitski, 1991). This can lead to different value yields if compared to the sieving technique. Automatic recording settling tubes measure changes with time in the weight of the sediment that collects on a suspended pan in a column of water and compares the weight versus time curves to a calibration curve, hence the frequency distribution of the sample as a function of settling duration is obtained (Syvitski, 1991 and Boggs, 1995). Again, this takes into account the temperature and dynamic fluid viscosity of the water in the column (Appendix 4). Density of the particle also plays a vital role in its rate of descent through the water. The settling column was previously calibrated to assume the average density of the sedimentary material as that of quartz, 2.65g/cc; specifically quartz as it is a very common mineral comprising the sand fraction (Rogers, *pers. comm.*). The settlement tube was prepared to automatically weigh the sand accumulating on the balance, set to three decimal places and was linked to a computer that plotted the cumulative frequency curve and performed various statistical procedures on the captured data, based on 0.1 phi intervals, with five intervals smoothed together. The previously split 2-3g sand fraction was spread evenly across the magnetic disc, with a small amount of 'teepol' (weakly diluted washing-up liquid) used to break surface tension of the grains, and held slightly above the water. It was gently lowered into the water, which sets off an automatic device to begin recording the procedure at the exact moment of contact and release of the sediment. Care was taken to carefully submerge the disc twice to ensure all grains (especially fine sand particles) were released.

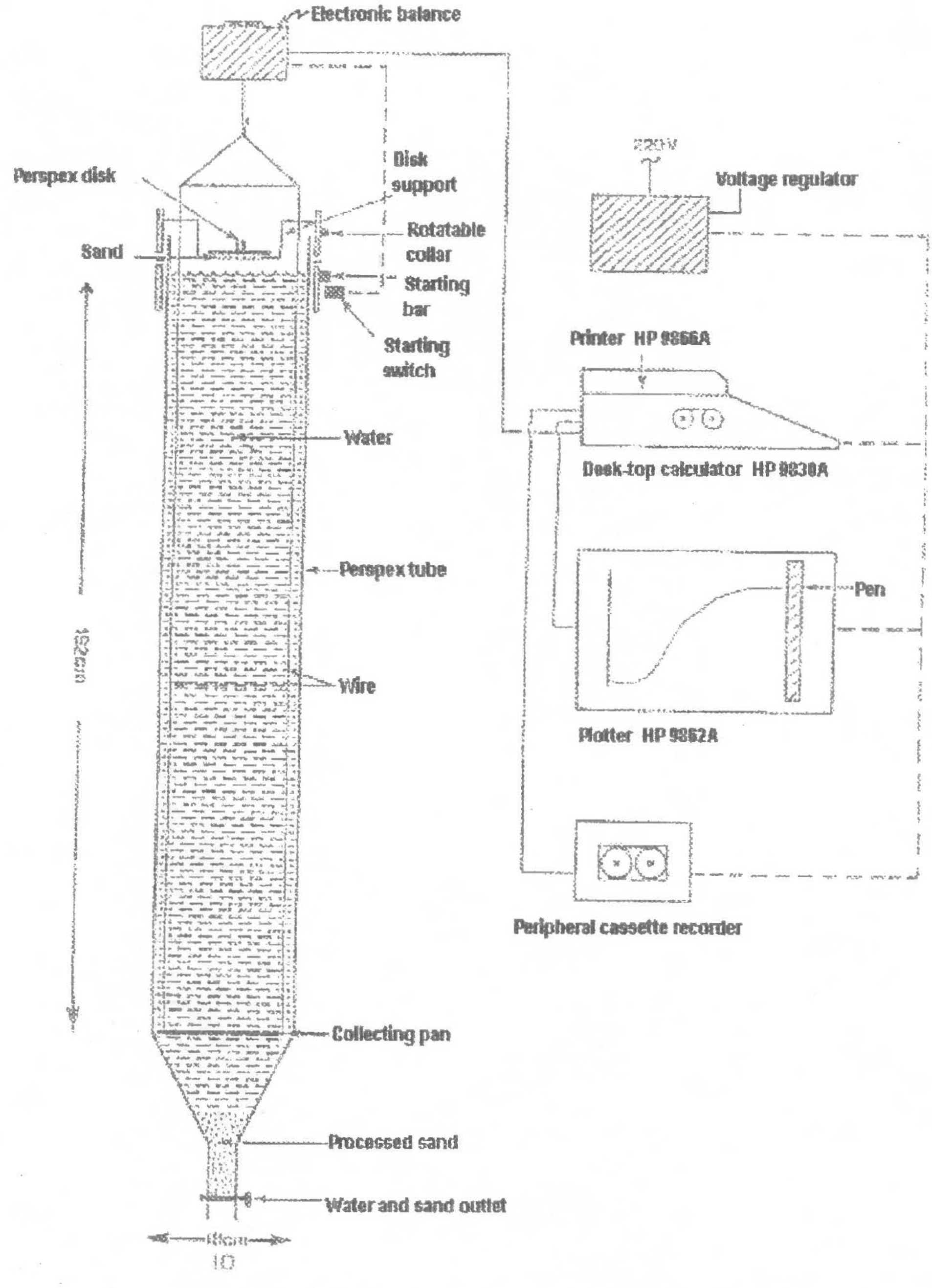


Figure 4.1: Configuration of the settling column used during analysis (author unknown)

The settling column procedure was chosen for sand fraction analysis over the dry sieving method as it is faster and, considering the number of samples, gives a large amount of information relatively quickly and is less labour intensive. It also minimises human contact and error. There are, however, disadvantages associated with this method. As already mentioned, there are assumptions of sphericity and density that can skew the results. If the disc is lowered too vigorously, a vibration could be set up down the water column to the weight pan which mimics the presence of a gravel fraction and distorts the results. Another issue is that, during the descent of the grains, they can bounce off the sides of the tubing thereby delaying settling; otherwise, if they are coarse, there is the possibility that they could bounce on the actual weight pan itself or simply fall to the side or off the pan. The sensitivity of the pan also plays a role as the start and finishing balance values tend to fluctuate as the pan gradually settles.

4.3.2.4. Final Data Compilation

The initial data were used to calculate the relative proportions of fine and coarse fractions for each sample. These fractions were broken down into clay and silt, and sand and gravel respectively. The settling column data were then processed using the GRADISTAT version 4.0 macro in Excel (Blott, 2000). Statistics using the GRADISTAT macro were calculated using the method of moments (Griffiths, 1967) and the inclusive graphical statistical method (Folk and Ward, 1957; Folk, 1974). The Folk and Ward method was used in final results.

4.3.3. Morphometric and Microscope Analysis

Particle shape, or the morphometric properties of particles, is employed to distinguish the different depositional and energy environments under which the sediment accumulated and is particularly effective for fluvial sediments. This is using a visual chart for descriptive classification of the particles based on roundness, sphericity and long axis elongation (Powers, 1982).

Particle shape, i.e. morphometric properties, was analysed by making grain mount slides and viewing under an optical microscope at both 1x and 2x magnifications. Particle shape is defined by varying aspects of the grains, namely the overall proportions (length of the long, short and intermediate axes) or sphericity, which is how closely the grain approximates to a sphere; roundness, or how angular the edges of the grain are; and surface texture (which was only briefly examined for certain samples under a light microscope at a magnification x10) (Pye, 1994). These properties are considered independent of each other; although these properties demonstrate characteristics associated with transport and deposition, their form is mainly a function of the mineral composition of the grain (Boggs, 1995).

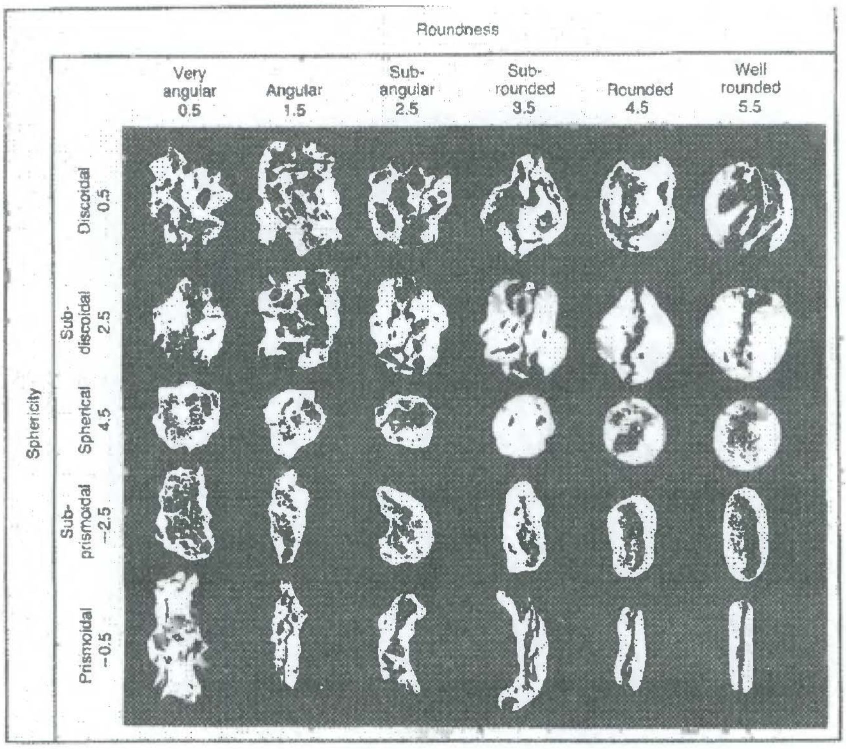


Figure 4.2: The visual comparison chart used for estimating roundness and sphericity (Powers, 1982)

Sands are silica-rich, and in the Western Cape region, are predominantly composed of quartz and feldspars (potassium and plagioclase) (Reid *et al.*, 2001). Quartz (SiO_2) is a very hard mineral at 7 Moh with no cleavage and feldspars (both orthoclase and plagioclase) are also relatively hard (6 Moh) with prefect cleavage (Whitten and Brooks, 1972). Quartz is not especially impacted by transport, except when in the pebble size class (Boggs, 1995). Changes in shape are due to impact-induced fractures through abrasions and breakages as the grains strike each other. Samples were described by the range of sphericity and roundness the various minerals fell into according to the Powers (1982) diagram. Where necessary or present, clasts were measured using a tape measure to ascertain long-axis length, while roundness was estimated based on the Zingg (1935) classification (Figure 4.3).

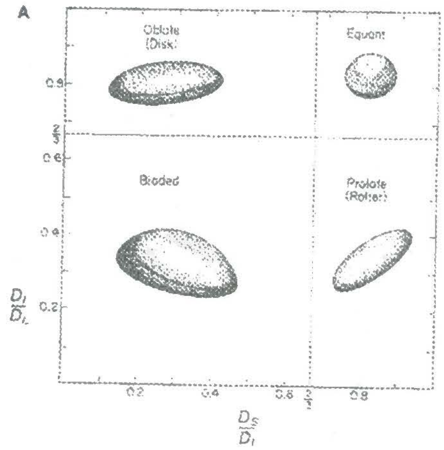


Figure 4.3: The visual comparison chart used for estimating the shape of clasts (Zingg, 1935)

The colour for each sample was described using the Munsell Colour Chart (1985) as presented in the "Revised Standard Soil Colour Charts" by A.H. Munsell. This was performed on each of the oven-dried bulk samples. Munsell colour provides a very broad indication of mineralogy, organic content or the presence of certain salts. It reflects aspects of both major and minor composition components. It can also provide useful insight into post-depositional processes as staining or weathering of sediment subsequent to deposition can occur (Briggs *et al.*, 1998). As these sediments derive from terrestrial environments, more oxidation and, hence, a more orange colour is expected for some. Munsell classification consists of separate notation for hue (colour), value (lightness) and chroma (strength), which, when combined, forms a standardised and universally recognised colour designation (Paton *et al.*, 1995). It is used in conjunction with colour names to increase the precision of the descriptions. Subsequent to the furnace burn for LOI, the sample colour was noted as to whether it had oxidised (turned to a deep reddish brown or orange) or been bleached (turned to a whitish grey tone) by the high temperatures.

Nodules, or naturally occurring small aggregations of sediment, were present in numerous samples from the profiles. As these nodules are representative and indicative of embryonic cementation, thereby possibly providing evidence of soil processes where cementing agents, such as clay, organic matter or oxides, may be present leading to compound particles or clusters of primary particles, they were evaluated. The size of the largest nodule present was measured from each oven-dried raw pre-treated sample and classed according to Munsell (1985) granular and crumb structures scale, from very fine (less than 1mm) to very coarse (more than 10 mm diameter). Size was classed as being from the specific class and smaller as the nodules decreased in size. Frequency of nodules was also noted and classed qualitatively with increasing frequency, from absent, sporadic to numerous.

Texture	Size diameter (mm)	Class no.
Very Fine	< 1	1
Fine	1 ~ 2	2
Medium	2 ~ 5	3
Coarse	5 ~ 10	4
Very Coarse	10 <	5
-	none	0

Table 4.1: Scale used for classing nodule size

4.3.4. LOI Organic Carbon

As organic matter naturally occurs and accumulates in all sediments, the study of carbon content in sediments is a means of estimating primary productivity and the abundance of plant life. In this sense, carbon content is a proxy for past productivity. Carbon in sediments occurs in two diagnostic forms: organic and inorganic (carbonate). Organic carbon occurs as the product of photosynthesis and decomposition. Inorganic carbon comes from the dissolution and precipitation of carbonate compounds as sediments. According to Holmes (1998), the presence of organic material may infer a reducing-type environment and, within specific beds, organic carbon is a good indicator of pedogenic processes influencing the profile. Organic matter has the potential to hold moisture and affect the structure by contributing to the physical condition of the sediment. Fluvial bed sediments represent an important sink and source for a variety of organic and inorganic compounds with constituent organic matter and its primary component, organic carbon (Sutherland, 1998). Destruction of the organic material in the samples by oxidation and the estimation of its original abundance are though the loss of ignition (LOI) method by comparing initial and final weights. The widespread adoption of the LOI method in soil science reflects its ease of use, as it is inexpensive, rapid and requires no specialized training.

The organic carbon content of the sediments was determined by the weight loss on ignition method (LOI) at 550°C in a muffle furnace (Heiri *et al.*, 2001). Approximately 15 g of fresh sediment from each sample was placed in pre-weighed porcelain crucibles. The samples were then dried in an air-circulation oven at 105°C for 12 hours to drive off all moisture. They were then cooled to room temperature in a desiccator, re-weighed to yield the dry-weight of sediment, and then placed in a furnace at 550°C for four hours in order to destroy all the organic matter. As the organic matter combusts, the carbon is changed to carbon dioxide

(CO_2) and carbon monoxide (CO). The samples were then allowed to cool to room temperature in a desiccator and reweighed. The percent organic matter was calculated by the difference in weight before and after heating in the oven at 550°C. Depending on the character of the sediment, various losses of volatile salts, structural and chemically bound water of clay minerals or metal oxides and ammonia will occur, which will have an effect on the final result. This results in an incremental error on the estimation of carbon content from LOI values, although the general trends of LOI_{550} and LOI_{950} show a good correlation with carbon content (both organic and inorganic) allowing use of LOI as a qualitative test for carbon content (Santisteban *et al.*, 2004).

LOI is a reasonable indication of the amount of organic matter in the sample, and the correlation with organic carbon is regarded as acceptable. Various conversion factors, such as that organic carbon represents 12/30 of the organic content ($(CH_2O)_n$), have been reported for converting loss on ignition to organic carbon (Sutherland, 1998). As the factor depends on the material being analysed, the use of conversion factors should be handled with care even within a single sediment profile. It is very difficult when using the loss on ignition method to determine organic carbon, not to avoid carbonate interference (Byers, *et al.*, 1978). The method is further complicated by the fact that carbon occurs in several forms: (a) carbon-containing compounds, little altered from their initial composition in living tissues; (b) highly altered and fairly resistant decomposition products of the original tissues, such as coal or graphite; and (c) carbonate minerals, such as $CaCO_3$, or rarely, as soluble salts containing HCO_3 (Carver, 1971).

4.3.5. LOI Inorganic Carbon

Loss on ignition (LOI) has been widely used as a method to estimate the amount of organic matter and carbonate mineral content (and indirectly of organic and inorganic carbon) in sediments. The relationships between LOI at 550 °C (LOI_{550}) and organic carbon content and between LOI at 950 °C (LOI_{950}) and inorganic carbon content are currently accepted as a standard (Santisteban *et al.*, 2004).

The allogenic, endogenic and authigenic minerals which comprise the inorganic carbon content of a sample can be measured by LOI or digest. This study used the weight loss on ignition (LOI) at 950°C method (Goudie, 1994; Heiri *et al.*, 2001). The weighed crucible with the residual ash from the organic carbon LOI analysis (weight A) was placed in a muffle furnace for 8 hours at 950°C. The crucible with the ignited sample was then cooled to room temperature in a desiccator and the weight of ash and crucible (weight B) determined. The inorganic carbonate content (CO_3) is equal to:

$$\frac{(A - B)}{\text{dry-weight of sediment (DW)}} \times \frac{60 \text{ (molecular weight for } CO_3)}{44 \text{ (molecular weight for } CO_2)}$$

As carbonate compounds become volatile at different temperatures, the carbonate content estimation is only a rough determination. Various conversion factors have been given by different authors, but as a general rule the ash-weight loss obtained between 550°C and 950°C should be multiplied by 1.36 to obtain the carbonate content. Weights were measured on the available balance (to an accuracy of 0.01g), although, a balance with a greater degree of resolution would have ensured more precise results.

There are associated difficulties with the LOI method. The loss is dependent on the sample size and samples continue to lose weight at a slow rate during exposure of up to 64 hours, therefore time lapsed affects the percentage change in weight. Results vary according to the position of the crucible within the furnace. It has been revealed that at 550 °C, samples in the centre of the furnace lose more weight than marginal samples. At 950 °C this pattern was still apparent but the differences became negligible (Heiri *et al*, 2001). These factors, such as sample size, exposure time, position of samples in the furnace and laboratory measuring, affects results, with LOI at 550 °C being more susceptible to these factors than at 950 °C. It is therefore vital to be consistent in the method used in relation to the ignition temperatures, exposure times, and the sample size and to include information on these three parameters when referring to the method (Heiri *et al*, 2001).

4.3.6. Electrical Conductivity

Electrical conductivity values provide a rapid and reasonably accurate estimate of the concentration of ionisable salts in sediment (Krauskopf & Bird, 1995). A higher electrical conductivity reading indicates that the sediment is more conductive, therefore correlating the concentration of ionisable salts to also be greater. This method is only approximate due to the relationship between salinity and its affect on the osmotic pressure of the sediment (Holland, 2000). Conductivity values through a profile allude to the variable leaching down through the horizons. Leaching is a major external factor that controls the mobility of constituent ions through a profile, the determining factor associated with the extent of chemical weathering that alters the parent mineralogy (Ritter, 1995). Some ions are easily removed (high mobility) from the weathering system under normal groundwater conditions, whereas others are not (high immobility). The presence of highly mobile ions in a mature weathering zone indicates that some or other factor is impeding their transfer out of the system. These mobile ions are distributed across common cations, with the most mobile being Ca^{+2}, Mg^{+2} and Na^{+2} and the least as Al^{+3} (Ritter, 1995). Leaching, also loosely defined as "the washing out of materials in solution or suspension from a soil horizon or profile" (Briggs *et al.*, 1998) can occur in conjunction with eluviation, or the removal of suspended solids or mineral colloids from higher to lower soil horizons by water percolation and reflects a range of soil processes. In more arid environments soils are typically unaltered and lacking in humus; coupled with a lack of significant rainfall, it would follow that less intense leaching and chemical weathering results in poorly develop soils with clays of higher cation exchange capacities (Briggs *et al.*, 1998). This means profiles are generally saline, especially so depending on the type of rock-type influencing groundwater salinity and the fact that constant evaporation at the surface draws water from the lower to the upper layers of the profile, leading to an accumulation of salts. Hence, the retardation of leaching, usually due to insufficient rainfall, causes immobility of ions in weathering systems and an abundance of mobile constituents in arid-climate soils.

In order to determine the electrical conductivity (EC), approximately 10mg of fresh sediment was placed into a glass vial. 10ml of deionized water was then added to the sediment in order to form a 1:1 sediment/water solution, which was agitated on a shaking table at 150rpm at 21°C for 2.5 hours in order to allow the sediment and deionized water to equilibrate. The conductivity of the solution was then measured using a WTW Multi 340i conductivity meter and electrode that had already been calibrated (therefore no need for standards). The electrode was immersed in the supernatant for 5-10 minutes and the reading recorded in microsiemens.

4.3.7. X-Ray Diffraction Analysis

The 1930's saw the rise of X-ray diffraction (XRD) as a method to unequivocally establish the crystalline nature of clay minerals to class the sample according to various mineral groups based on their characteristic basal spacings (White, 1997). The technique involves directing beam of X-rays (electromagnetic radiation of wavelength from 0.1 to 10nm) at the bulk and clay fraction of a sample. A natural sediment or soil sample contains significant contributions of three types: crystalline (mineral) phases, an organic amorphous phase (organic matter), and an inorganic amorphous and nanoparticulate phase (Rancourt & Dang, 2005). The clay sample can be a powder sample or a suspension dried onto a glass slide giving the orientation of the plate-like crystals. The X-rays penetrate a crystal, with a small amount being absorbed by atoms that then become excited and emit radiation in all directions. Radiation that is in phase will form a coherent reflected beam of parallel X-rays of wavelength λ, striking a crystal surface at an angle θ, the necessary condition for the reflected radiation from atomic planes to be in phase is, a mathematical statement of Bragg's law, $n\lambda = 2d\sin\theta$, where n is an integer and d is the characteristic atomic spacing (White, 1997). X-ray diffraction has application in most fields dealing with solid materials. XRD identifies crystalline compounds as opposed to X-ray Fluorescence (XRF) or other spectro-chemical methods that identifies just the elements. Areas of application are quite wide and include metals, organic and inorganic compounds. XRD of clay minerals is now an accepted method of mineral evaluation among geomorphologists.

The clay mineralogy of sediment provides constructive information with regards to the origin of the material and any chemical changes that may have occurred due to the effects of different weathering processes since deposition and is therefore indicative of conditions. It is carried out by XRD that involves the measurement of the physical values concerning the crystal structure of clays as the intensity of X-ray diffraction and the angle of diffraction is related to chemical composition allowing for identification of clay mineral species (Goudie, 1994). Although definitive explanations of clay minerals are difficult and identical species of clay minerals are formed by different processes, such as diagenesis (processes occurring at the surface) and weathering, slight differences are retained by their chemical compositions and structures due to the genetic processes.

The definitions of clay are slightly different in various sciences. In terms of XRD analyses, clay is defined as (a) aggregates of particles finer than 2 μm, (b) adhesive substances and (c) natural inorganic materials mainly composed of silicon, aluminium, iron, magnesium, alkali-metal elements, alkaline earth elements and water. Clay minerals generally include phyllosilicates composed of sheet structures, fibrous palygorskite and sepiolite, as well as some amorphous substances. However, amorphous silica, iron oxide and alumina gels are not classified as clay minerals (Briggs *et al.*, 1998). Clay mineralogy may be utilised as an indicator of the degree of weathering to which a profile has been exposed, hence palaeoenvironmental interpretations can be derived from the relative proportion of clay minerals and their behaviour. Important considerations taken into account with clay mineralogy include inheritance (parent material), neoformation (weathering and solution chemistry) and transformation (diagenetic and hydrothermal) (Holmes, 1998 citing Eberel, 1984). As the study locality borders the Karoo and occurs across the Table Mountain Sandstone/Bokkeveld Shale contact, there are complications involving the inclusion of shale-derived sediments as they themselves are the product of previous cycles of weathering and diagenesis.

To ascertain whether clay mineralogy may uncover evidence of pedogenesis or obvious differences between the horizons of the profiles, XRD samples were chosen from the main terrace exposure from both the alluvium and colluvium packages to identify clay mineral species from both bulk and clay samples. Samples were selected on the basis of relatively higher percentage fine fraction. Selected bulk samples for XRD were dry sieved at 0.25ϕ intervals from -1-4ϕ using a Ro-Tap shaker with the fraction retained in the pan (<63μm) being sent for analysis. Clay mineralogy was investigated using random powder bulk and clay size fraction XRD analysis. This was conducted at the Agricultural Research Council, Institute for Soil Climate and Water (ARC-ISCW), Pretoria, using a Siemens D500 X-ray goniometer equipped with Cu tube, variable slit and secondary graphite monochromator (Kirsten, *pers. comm.*). The latter eliminates K ß radiation and additionally reduces fluorescent radiation. A fine grained powder (± 10 μm) of each sample is pressed in an aluminium frame against a rough filter paper. Such random powder preparation is scanned from 2 to 65° 2Θ $Cu_{K\alpha}$ at a speed of 0.02° 2Θ steps size/sec^{-1} with generator settings of 35 kV and 25mA.

Powder X-ray diffraction (pXRD) has the advantage that any type of material that is already powdered or can be powdered could be used, without particle size or shape limitations other than certain manageable complexities related to homogeneity and micro-absorption. The potential of pXRD in treating complex solid mixtures and environmental samples in particular, has not been fully developed, although practical difficulties have been reviewed, such as the significant effort required, during laboratory calibration, in controlling, measuring, and monitoring all the needed instrumental parameters, such as incident beam intensity, counter efficiency, beam-path attenuation and scattering, effective beam widths as determined by various slits, degree of incident beam polarization, etc. (Rancourt & Dang, 2005).

The results were evaluated qualitatively using SIEMENS DIFFRAC^{Plus} software. The XRD evaluation program (EVA) allows matching with the PDF-2 Database and semiquantitive estimates are based on peak-height percentages. Ouhadia and Yong (2003) have experimentally reviewed the application of the four most popular methods of XRD analysis, including the method used for mineral quantification purposes. They suggested that: (a) direct quantitative XRD analysis based on peak intensity or areas under the peaks is likely to cause an over- or under-estimation of the quantity of clay minerals; (b) quantitative evaluation based on only the major reflection line is likely to induce mistakes in clay mineral identification; and (c) quantitative X-ray analysis using different reflection lines of minerals that take into account the impact of clay microstructure have been shown to be the most accurate method of mineral evaluation. In light of the potential over- or under-estimation of specific clay mineralogies, values represent proportional percentages.

In total, 25 bulk samples were analysed using XRD in the Department of Geological Sciences, University of Cape Town. This was useful in determining mineral content of the sediment. Clay was, however, not present in sufficient quantities within the samples to be consistently isolated and identified during analysis. Organic content was negligible as the proportion was low and does not diffract due to a lack of crystalline structure. Preparation could, however, have included combustion to burn off any organics and calcium carbonate, or through the use of an acid leach. This has associated problems in that sulphate has the tendency to precipitate out (Nel, *pers. comm.*). The samples were dried overnight in an oven at 70°C, so as not to thermally alter any crystalline structures, and crushed as finely as possible with a pestle and mortar. Slides were cleaned with acetone (CH_3CO), packed with the sample powder and mounted in the X-ray diffractometer (Philips PW3830) at 40kV and 25mA. The slide was positioned so that the specimen surface pressed exactly through the axis of the goniometer. The sample was run from 5° to 76°, with a step size of 0.020° and scan speed of 0.040°/s. The scan is graphically represented as counts/s versus °2 Theta showing Bragg peak heights (Philips operation manual, unpublished). Search-match parameters used peak heights to search for a list of possible known mineral and chemical substances. The parameters for this search and identifying are based on intensity thresholds, confidence threshold, number of strong lines, length of score list and minimum and maximum displacement (XRD Laboratory Handbook, unpublished). These results were not used quantitatively, rather qualitatively, to present a generalised view of mineral content across the study area.

Figure 4.4: Philips X-ray diffractometer (Department of Geological Sciences, UCT)

4.4. NUMERICAL DATING TECHNIQUES

4.4.1. Introduction

Absolute age determinations are vital to this project in order to provide a chronological context for the beds and sediments. By providing numerical age controls on stratigraphy, luminescence dating affords determination of the timing (or approximate time-scale) of erosional and depositional events and makes possible the calculation of the rate and magnitude of environmental, geomorphic and sedimentological change (Makaske *et al.*, 2002). As semi-arid to arid sedimentary environments act as an archive to the punctuated natural environments of the past and its components, moderate- to high-resolution chronologies are of great value to interpretation and significant to inferences. Although luminescence dating has primarily, in the field of geomorphology, been utilised in providing chronologies for aeolian deposits, it does also apply to the context of sandy fluvial deposits such as that at the main terrace site at Wuppertal. The study site sediments were presumed to encompass the Holocene (at least), and within the context of physical parameters dictated by the sediments, the luminescence dating method would be indispensable and prove an integral part of the project that aims to provide a palaeoenvironmental reconstruction based on sedimentary proxy evidence for the north eastern Cederberg margin with the Karoo.

4.4.2. Optically Stimulated Luminescence Dating

Luminescence dating, like radiocarbon dating, is an absolute radiometric dating method based on the measurable and negatively exponential radioactive decay of unstable isotopes. Luminescence dating is used in establishing "age since burial" is a very valuable method and particularly applicable to inorganic materials, especially in arid and semi-arid environments with ages from 1000 to 500 000 years (Walker, 2006). As the primary site falls within an arid region and is notably quartzitic as a result of predominantly quartz- and feldspar-rich Table Mountain Group sandstone parent rock, luminescence dating, especially optically-stimulated luminescence, appeared as a specifically appropriate means of radiometric dating. Dating of alluvial channel and fan sequences has provided additional insights into periods of pluvial activity (Stokes, 1997).

When sediment is deposited, the surrounding rock and actual sediments contain radioactive isotopes of uranium, thorium and potassium. The sediment is exposed to the ionising radiation from these elements' decay in the surrounding matrix as well as a small additional amount from cosmic rays. Common (luminescent) minerals, such as quartz and feldspar, absorb this radiation and act as dosiometers and record the damage caused by this radiation. This results in the displacement of the electrons from the atoms within the crystal lattice, which become trapped within lattice defects or luminescence centres (Walker, 2006). Upon exposure to heat or light, these electrons are freed from their traps, leading to luminescence being emitted in proportion to the dose of ionising radiation assimilation since the time of deposition (Aitken, 1998). In this study, light was used as the stimulus. This signal is highly sensitive to light exposure, and during transportation and exposure to sunlight, the sediment is removed of its luminescence, thereby having a signal close to zero. Once burial occurs, the process of radiation absorption begins again and over time, grows until it is exposed to light again (either naturally in the environment or artificially in the laboratory).

The total amount of accumulated radiation dose can be determined since burial, is measured in Grays (absorbed dose) (Gy) and is termed the palaeodose or referred to as the equivalent dose (D_e). The D_e is an estimate of the total dose absorbed during the irradiation period. Dividing the palaeodose by the rate of exposure or received ionizing radiation (Gy/ka or Gy/yr), the age of the deposit can be established. This is also referred to as the dose rate (D_R) and is commonly comprised of K, U, Th, Rb and cosmic ray components (Aitken, 1998). The annual radiation dose can be calculated on-site or from a sample taken within the depositional environment. If the dose rate is known the duration of the "dosing period" can be calculated:

$$\text{Age (a)} = \frac{\text{Total accumulated radiation dose (Gy) (Palaeodose)}}{\text{Annual radiation dose or dose rate (Gy/a)}}$$

Alternatively the luminescence age equation can be written:

$$\text{Age (a)} = \frac{\text{Palaeodose, } D_e \text{ (Gy)}}{\text{Dose rate, } D_R \text{ (Gya}^{-1}\text{)}}$$

The palaeodose or equivalent dose can be determined through the additive (using multiple aliquots), regenerative (single aliquot), partial bleach (assuming a mix of grains with partial bleaching) or total bleach (assuming complete zero) methods. Bleaching refers to the release of trapped charge by light (Murray & Olley, 2002). The dose rate may be determined using a number of methods, namely neutron activation, Y spectrometry, Inductively Coupled Plasma–Mass Spectrometry (ICP-MS) and atomic absorption. OSL can use either method by attempting to characterise the growth of the luminescence signal by irradiating the sample in the laboratory using equipment such as that shown in Figure 4.5. Unlike thermal luminescence, this procedure is relatively quick and complete bleaching occurs. Until the late 1990s, the great majority of OSL dates were based on infrared stimulation of feldspars, however, readily accessible measurement technology has become widely available with the development of cheap green and blue light sources meaning quartz is now the preferred mineral (Murray & Olley, 2002). Measurement protocols have also developed and are now more commonly based on single-aliquot regenerative-dose (SAR) protocol to establish D_e of quartz (Murray & Wintle, 2000). Table 1 below sets out a typical SAR measurement cycle.

Step	Treatment	Observed[d]
1	Give dose[a], D_i	-
2	Preheat[b] (160-300° C for 10 s)	-
3	Stimulate[c] for ~100 s at 125° C	L_i
4	Give test dose, D_t	-
5	Heat[b] to 160° C	-
6	Stimulate[c] for ~100 s at 125° C	T_i
7	Return to 1	

Table 4.2: Generalised single-aliquot regenerative-dose protocol (Murray & Wintle, 2000)

The protocol runs as follows: the sample is given a dose during burial (before sampling) then in the laboratory, the sample is first preheated to a temperature in the range of 160 to 300°C (for 10 seconds) and the natural OSL signal measured. The sample is then given a test dose, heated to 160°C and the test dose OSL signal is measured, completing the first (natural) measurement cycle. To begin the second cycle, a regenerative dose is first administered; the sample is then heated to the same preheat temperature as in the first cycle and the OSL signal measured again. The sample is then given the same test dose as before, heated to 160°C and the test dose OSL signal measured; this regenerative cycle is then repeated as many times as desired, with the regenerative dose being the only variable (Bailey *et al.*, 2001; Murray & Wintle, 2000).

It is suggested by Bailey (2000) that multiple OSL components, both technical and physical, can affect the form of the D_e plots. The moisture content of the sample has an affect on the amount of radiation attenuation that occurs. Cosmic radiation contribution, however small, can significantly influence the age determinants and must be taken into account. Latitude, altitude and the depth of the soil or sediment overburden also impacts on the intensity of the cosmic radiation dose to which sediments are exposed (Aitken, 1998). Other difficulties include the proximity of the sample to the surface, it cannot be less than 30cm, and handling in the field and necessary precautions to prevent any contamination of the sample. An additional issue is that quartz grains in nature are rarely "clean" and the transport medium transparent therefore if exposed to sunlight, the fast component of the OSL signal is not reduced to a negligible level within a few seconds as expected thereby affecting subsequent signals (Murray & Olley, 2002).

Luminescence dating samples were taken from WT3 profile between 0.9-1m and 1.67-1.75m and WT4 profile between 1.85-1.95m and 2.70-2.80m (sandy alluvium). No OSL samples were taken at the WT5 profile (colluvium) as the substrate was in a state of calcretisation in places towards the base and partial lithification and it proved almost impossible to hammer the tubing into, and many complications arise from colluvial based OSL dating (Boardman *et al.*, 2005). Samples were taken using cut sections of opaque aluminium irrigation piping of a diameter of approximately 10cm hammered into the freshly cleaned partially consolidated sediment face at the selected measured depth and plugged. It was then carefully extracted and sealed, labelled and packed in black bags for transportation and storage. The tube is later cut open under dark room conditions and the exposed ends discarded to prevent accidental exposure of the sample to light which would almost instantaneously zero the luminescence signal (Boardman *et al.*, 2005). OSL dating of these four samples was carried out at the Oxford Luminescence Dating Laboratory (OLD) in the Oxford University Centre for the Environment. The altitude, latitude/longitude and depth of each sample were also provided. For this study, D_e was determined using the five-point SAR analysis protocol and dose rate was established using ICP-MS. The five-point SAR protocol is commonly used in dating of various sediments, as outlined by Chase and Thomas (2006) and Murray and Wintle (2000). Dose determination for the analysed samples also used both single and multiple aliquots in an attempt to refine the accuracy of the data as the coarse nature of the sandy material and a lack of fine sediment yield below 250μm required more thorough preparation for adequate results to be established (Telfer *pers. comm.*). Age results are expressed in calendar years before present (BP).

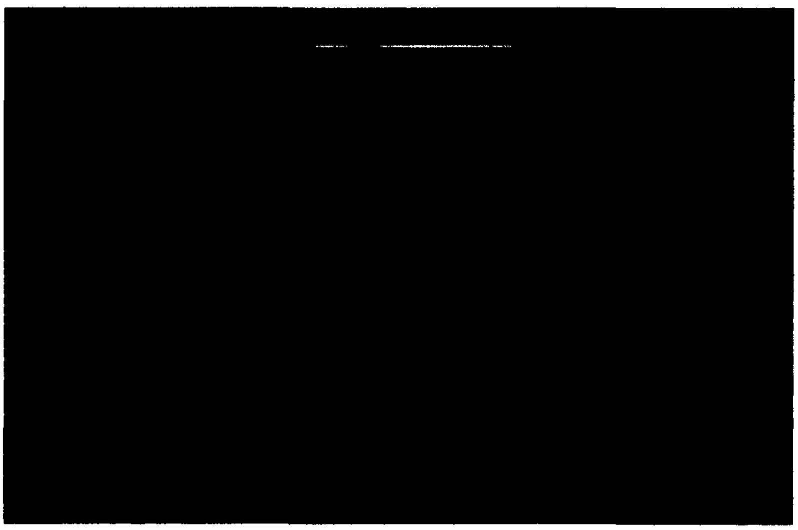


Figure 4.5: Facilities at the Oxford Luminescence Dating Laboratory where OSL was carried out on samples and handled under subdued lighting conditions using Riso TL-DA-series TL/OSL readers (OLD, 2006)

4.5. STATISTICAL ANALYSES

4.5.1. Introduction

Quantitative information that has been measured, sampled, perceived/observed, etc. is presented as sets of data requiring interpretation and analysis to be better understood and applied within an appreciated context. The rapid computation facilitated by computer processing that resulted in an array of statistical values and properties, such as grain size means, medians, standard deviations, skewness, etc., require that they be further examined by other statistical methods so that they may reveal more information and provide some order to the data. This is done through a series of analyses that aim to make generalisations and reduce complexities within the data for interpretation as this is more difficult or unlikely with raw data. This investigates the data structure in terms of identifying what is being measured, for example, and to establish if the data are symptomatic of underlying or more discreet fundamental aspects that are not immediately obvious or otherwise to prove these; and then assess the relationships between the variables. Such techniques aim to reduce noise and dimensionality and establish any patterns or processes.

Nevertheless there is a degree of subjectivity inherent in this kind of statistical analysis as generalising is never indicative of perfect classification. There are always varying interpretations dependent on the investigator as well as the skewing influence of outliers or exceptions. One can over estimate relationships between components and variables or misinterpret the correlations and loadings as being representative of a concrete relationship. Therefore, it could be problematic in attempting to source a relationship and conveniently establish one within a dataset even though it may be a result of human error, measurement or skewing due to temporal and spatial resolutions (Rollinson, 1998).

The data set created for this statistical analysis is multivariate and multidimensional containing observations over a wide range of parameters (where multivariate refers to data consisting of multiple variables from varying parameters but where each variable represents a 'dimension' of the data set that could be orthogonal or considered independent). The temporal scale in this data set is negligible and the spatial scale is given through the sample names (i.e. depth). It contains a large collection of data with assorted measures, units and scales, values range from single units to double digits with some involving decimal places and others not. The data set itself is 17 columns by 66 rows, and represents a significant number of variables from which relationships or groupings are expected to be identified where variables are similar or dissimilar. The data matrix was imported into the programme STATISTICA with null values or empty cells replaced with -9999 (as the programme and procedures do not recognise "no value").

In the generated data set for this study, it is desirable to establish if there are any significant links or relationships, besides the already recognized ones, between various elements that are not immediately obvious. Holmes (1998), Holland (2000) and Cornell (2001) have used these methods, specifically standardisation of a dataset followed by cluster analysis and principal component analysis (PCA), with success to aid in elucidating further relationships between sedimentological and geochemical results. Therefore these statistical methods lend an added value to understanding, analysis and interpretation of the quantitative data obtained during field and laboratory work.

4.5.2. Normalisation

One of the most significant factors in multivariate exploratory techniques is the data itself. The classification depends on it and if one variable is very large in magnitude or has large variance or range, it will have a disproportionate influence or dominate the grouping; variables can be measured in different scales creating discrepancies that distort the proximity or distance relationships leading to highly skewed results towards variables of high values, thus misinterpretation ensues. To avoid this it is recommended that one standardises the data set by converting all the values of the data to the same proportional or similar scales before the analysis is performed. The scale attributes of the original data (%, phi units and μS/cm) are removed and no single variable is weighted at the expense of the other variables in the data (Holmes, 1998). This is done by subtracting the mean from the variable and dividing by the standard deviation or using normalising assumptions. Choosing whether or not to standardise data can have a large impact on its interpretation. Prior to any type of analysis, standardisation is carried out on the entire raw data set pertaining to each stratum from the profiles and surface samples at each study site, using the standardise function in the 'data' drop-down menu of STATISTICA. This procedure produces a new data set with a mean of 0 and a standard deviation of 1 for each variable.

4.5.3. Cluster Analysis

Cluster analysis is an exploratory data analysis technique used to solve classification problems and to organise observed data into meaningful structures. It uses a multivariate procedure for detecting or identifying natural groupings, or clusters, of associated objects in the data. It is a mathematical technique that summarises or classifies large amounts of multidimensional data into groups where members or profiles of the objects in the same group share common properties or are very similar, yet distinct, sometimes substantially, from those objects in different clusters. Classification is on the basis of the similarity of the characteristics the objects possess and the observations or data entities are either partitioned or aggregated into clusters. Cluster analysis aims to minimise within-group variance and maximise between-group variance by generating a number of heterogeneous groups with homogeneous contents. It is based on the procedure of generalised distance or proximity between objects versus their measures of likeness or similarity (correlation). The actual technique involves a range of algorithms that can be used to create the clusters and there are namely two methods or categories of analysis: clustering where the user defines the number of groups after the procedure, referred to as hierarchical; or clustering where the number of clusters is chosen or pre-defined by the user beforehand, non-hierarchical (StatSoft Electronic Textbook, 2004). Its objectives include discovering types of groups within the set and reducing the number of cases by facilitating consideration of several types instead of numerous records. Overall, cluster analysis is a very useful tool for classifying large amounts of information or data into more manageable and meaningful groups that can be explored further and provide insight for the researcher (http://www.mathworks.com/access/helpdesk/help/toolbox/stats/multiv14.html, 2004).

There are a number of problems associated with cluster analysis. Methods are not clearly established and there are various options to decide between when undertaking this procedure. This leads to disputes over the fact that it could be considered convenient that the observer or researcher may use varying methods of computing the proximities or linkages until the structure that was originally believed to be contained by the data is 'discovered'. Another factor is that cluster analysis does not complete and produce a finished product in its entirety; rather it is up to the researcher to determine and resolve groupings that cannot be validated by the statistical programme (StatSoft Electronic Textbook, 2004). Identification of the optimum number of clusters is one of cluster analysis's major problems especially bearing in mind that as larger clusters are successively created, increasingly dissimilar clusters are classified together making the result increasingly artificial. Deciding on the number of clusters is mostly subjective. A dendogram shows sudden jumps in the level of similarity as more dissimilar groups are linked. A further difficulty is the presence of outliers within the data and their detection. Outliers can be single data observations, values or small clusters far removed from the main groups.

In this study hierarchical clustering was used. This method is based on measures of distance between the observations or objects within the data. The hierarchical system calculates as many clusters as there are data points and then displays their relative closeness, by determining the distance between every pair of data points, on a tree diagram or dendogram.

The decision to cluster according to variables (columns) as opposed to cases (rows) would alter the interpretation through linking similar columns or rows respectively, therefore showing relationships in different dimensions in space (17 versus 66 dimensions). The choice of algorithm is also of importance (and subjective) and is a contributing factor towards interpretation as depending on what kind of grouping or relationship one wants to determine, will establish which algorithm one uses. Clustering was executed according to cases i.e. sample (rows) on the raw standardised data and on the principal component factor scores (http://www2.chass.ncsu.edu/ garson/pa765/cluster.htm, 2004).

The so-called single linkage and Ward's methods were chosen as minimum variance clustering from the various procedures and algorithms to utilise and are based on a Euclidean distance measure (StatSoft Electronic Textbook, 2004). 'Single linkage' is when two clusters are linked together when any two objects in the two clusters are closer together than the respective linkage distance i.e. start with two closest points and link them, followed by linking next two closest points, etc. until all data points are linked therefore with minimum dissimilarity. This algorithm produces clusters that are chained together by single objects that happen to be close together and therefore form long chains of loose clusters. It is effective at identifying outliers.

Ward's algorithm is based on the analysis of variance to evaluate distances. It aims to minimise the within-group variance and maximise between group variance. Objects in a cluster are decided based on calculating the total sum of squared deviations from the mean of

equalling no correlation. The matrix represents a set of vectors; if these vectors are plotted, a structure of similarity, showing that the variables are possibly responding to some underlying process or relationship, becomes noticeable. The second step is the extraction of structure inherent within the similarity matrix through the use of eigenvalues and eigenvectors. Eigenvectors explain the variance given by the eigenvalue. Eigenvalues of 1 show that eigenvectors explain as much variance as an individual variable, whereas eigenvectors with eigenvalues above 1 show more information than an individual variable; eigenvalues also account for the total variance of a component (Rollinson, 1998). Only eigenvalues above 1 were retained. Each eigenvector is called a principal component. Each variable in the dataset can be assessed concerning its contribution to the overall distribution. High correlation between the first principal component and a variable indicates association with the direction of the maximum amount of variation in the data set. More than one variable may correspond highly with it as more than one may be having a strong influence on the distribution of data. A strong correlation between a variable and the second principal component indicates that the variable is responsible for the next largest variation in the data perpendicular to the first, and so on. Conversely, if a variable does not correspond to any principal component axis, or corresponds only with high-number principal component axes, this usually suggests that the variable has little or no control on the distribution of the data set. Therefore, PCA may often indicate which variables in a dataset are important and which ones may be of little consequence. Some of these low-performance variables might can be "weeded out" and removed from consideration in order to simplify the overall analyses to reduce noise (StatSoft Electronic Textbook, 2003).

The third step is the expression of the variable relationships to the eigenvectors, known as the loadings. A negative loading is the relationship in the opposite direction. A loading squared is the variance of the variable that is described by the component. A high loading shows a close alignment with the eigenvector and hence, the underlying component has a strong influence, whereas a low loading shows that the underlying component has a weak influence (http://www2.chass.ncsu.edu/ garson/pa765/factor.htm, 2006). Subsequent weak features or components to the strongly influencing first components may actually represent noise within the dataset structure. The data has now been distributed and aligned within the underlying data structure. However, as not all of these components are significant, superfluous eigenvectors are removed and the remainders may be better positioned for more accurate interpretation. Therefore, the eigenvectors can be rotated to maximise the loadings. Usually it is best to do this prior to interpretation as pre-rotated loadings may be misleading (StatSoft Electronic Textbook, 2003). Factor scores created subsequent to loadings shows a matrix of the original raw data expressed as underlying components that gave rise to the initial variance. These factor scores are particularly useful when one wants to perform further analyses involving the factors that one has identified in the factor analysis. The final step, factor scores, will be used for a cluster analysis to compare these results, which are a 'filtered' or cleaner orthogonal version of the original data with reduced variance, noise or redundancy, to the initial cluster analysis of the raw data.

4.6. CONCLUSION

Geomorphological studies and palaeoenvironmental reconstruction depend not only on field surveys and observations, but also a range of laboratory techniques and further statistical analysis of these results. This chapter has described the variety of techniques utilised, as well as the motivation and expectations in relation to this study and the procedures carried out. Some methodology was based on value judgments, however, it was hoped that the methodology implemented would aid elucidation and specific insights with regard to particular properties and features. Time and financial constraints did, however, limit various techniques to fewer samples analysed and precluded some altogether (e.g. AMS radiocarbon dating). The large data set developed by the field and laboratory methodology required proper data handling and organisation for it to be properly and meaningfully used in interpretation, and that the statistical techniques employed aid in this. The results and application of these methodologies are presented in Chapter 5 and discussed in context in Chapter 6.

Chapter 5
RESULTS

5.1. INTRODUCTION

It is the aim of this chapter to present and state objectively the results of the field study, profiles and geochemical and sedimentological analysis of the samples. Subsequent statistical analyses of these results are also described along with chronology and remote sensing. Interpretation and synthesis of the raw data and findings is reserved for the discussion in Chapter 6.

5.2. SITE DESCRIPTIONS

The choice of study sites used in this investigation was largely subjective based on assessed aerial photography, maps and physical observations in the field; however, there was an attempt to choose a representative cross-section of contemporary fluvial environments of the two convergent Moordenaarsgat and Tra-Tra valleys in conjunction with the primary site on the Dassieboskloof River. The number and location of sites was constrained by accessibility and the timeframe of the study. This section aims to provide a diagnostic description of the fluvial environments within the study area as well as pertinent factors and features related to the greater geomorphic system. The specific sites used in the palaeo-interpretation as well as those used as contemporary environment benchmarks are described. A general photograph of each site used is included for the purpose of context and information. Qualitative data on sediment characteristics accompany the photographs, while quantitative graphical representations of the profiles' sediment texture and colour (Munsell notation) follow in section 5.4 along with positioning of the OSL age determination samples. For a detailed summary of sample site position, location, etc., refer to Appendix 3. Figure 5.1 below provides a spatial framework for the three river catchments under investigation in terms of supplying a regional drainage map, a basic 1:50 000 catchment map with relevant geomorphic features and a site specific sketch map, in order to present a context for the study itself and primary site in particular. A three dimensional perspective provided by a long profile shows the Dassieboskloof and

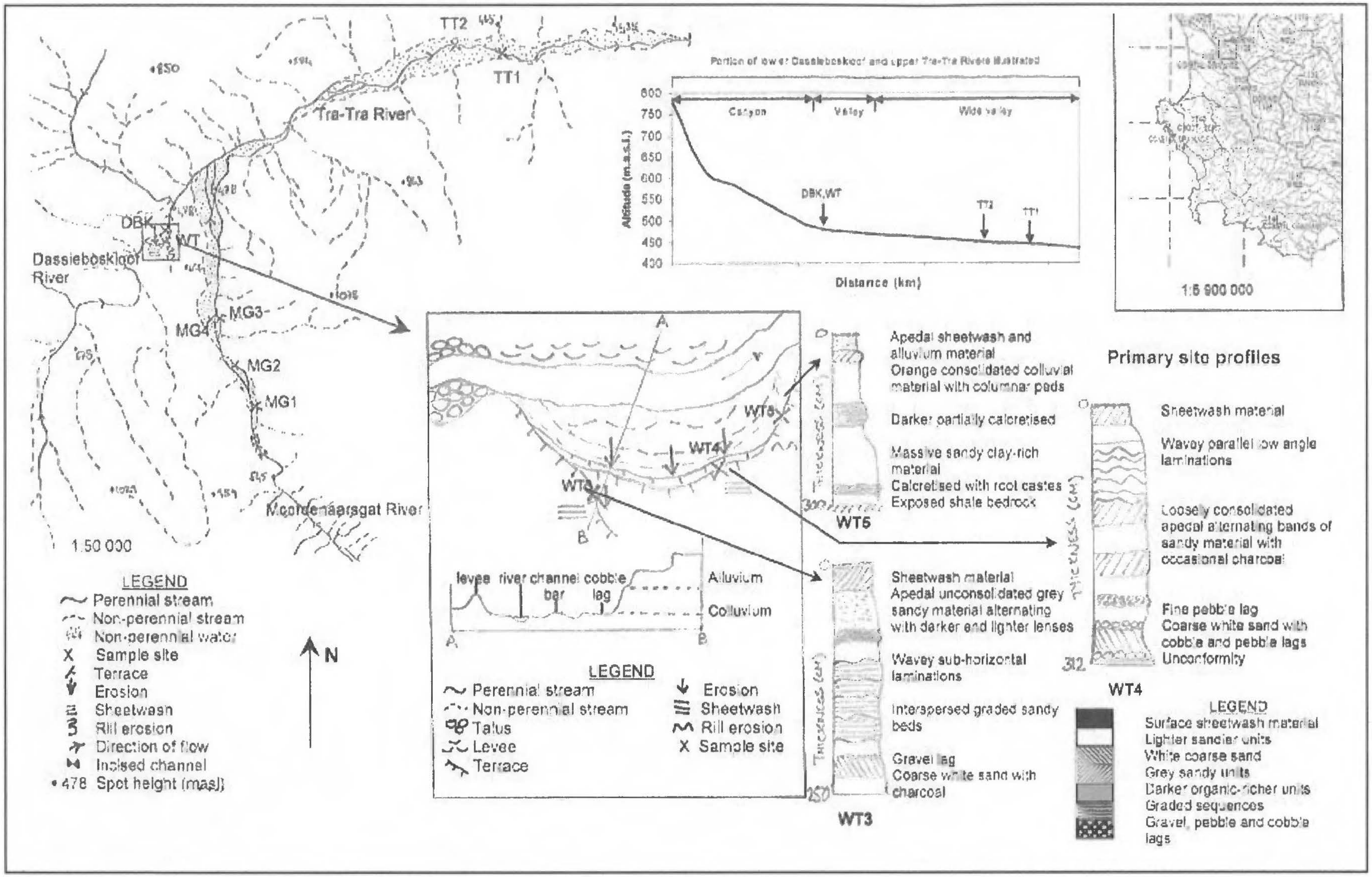


Figure 5.1: Overview of Moordenaarsgat, Tra-Tra and Dassieboskloof River catchments within study locale showing sample site distribution, pertinent geomorphic features and the long profile of the Dassieboskloof and Tra-Tra Rivers displaying its transition from upper to middle reaches and relative distribution of sites. A sketch map and cross-section shows the primary site, sample localities and basic profile stratigraphy.

Tra-Tra Rivers transitioning from an upper to middle reach of the fluvial system in relation to the study vicinity; the location of the sample sites demonstrates that the primary site (WT) occurs in the transition section between the high gradient canyon and lower gradient wide valley of the Tra-Tra River. Figure 5.1 also includes a cross-section of the primary site and shows the orientation of sample sites and their distribution relative to notable features in the area and the overall morphology of the site's morphology; basic profile stratigraphy for WT3, WT4 and WT5 illustrates the nature of the sediments and key units and sedimentary features as well imparting a further spatial context of the study sites.

5.2.1. Moordenaarsgat River Sites (MG)

Within the study area that extends to the opening of the valley from the canyon, the Moordenaarsgat River forms a valley-constrained alluvial fan (Figure 5.2). The proximal fan or fanhead exits an asymmetrical box canyon, owing to the dipping strata of the bedrock to the south from a minorly entrenched single channel characterised by coarse sediment, small channel bars and elevated flow flood banks with boulder, cobble and pebble lags. The fan spreads out within the valley with the midfan consisting of a network of branches within the trunk channel. The flat valley floor covered by unconsolidated sediments, some partially consolidated, clearly shows the braided nature of the Moordenaarsgat River, as seen in Figure 5.2. A notable feature is that the primary channel runs closer to the western flank of the valley at the proximal end and then meanders across the centre. At the point of valley constriction it is closest to the western slope and then transitions from a constricted channel back into an open channel and meanders again towards the eastern flank eventually exiting the valley to the northeast where it converges with the Dassieboskloof River. Secondary and lateral channels are also braided and contain sections of incised channels. Channel sediment is generally coarse sand with pebbles, cobbles and numerous boulders. There is some indication of imbrication but it is not well defined and most boulders are sandstone. Even dry and abandoned channels exhibit some braiding and small-scale incision. The river gradient within the valley is gentle with the upper half (above the valley constriction) at roughly 1:50 whereas below the constriction the gradient decreases to approximately 1:70 at its confluence with Dassieboskloof River. The height difference between the proximal end of the fan and valley margin at its confluence is 60m (535m and 475m) and the difference between the proximal and distal fan ends is 45m (535m and 490m).

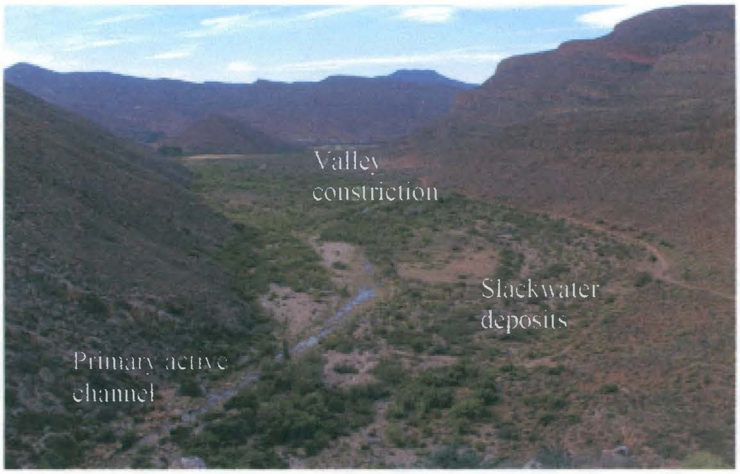


Figure 5.2: Moordenaarsgat River looking north (downstream) towards Wuppertal (white buildings in the distance) showing the nature of the braided river network as a valley constrained alluvial fan (note slackwater deposit to the eastern edge of the proximal fan end)

It appears that, in addition to fluvial transport, sediment is fed into the valley parallel to the Vaalheuningberg, and Moordenaarsgat fluvial system by mass wasting from the valley walls and slopes in the form of slope debris and colluvium. There are clearly defined talus slopes at the base of the mountain slopes feeding into the colluvium and in places these form small colluvial fans. These colluvial fans at the base of the talus slopes are truncated by the Moordenaarsgat river channels. This is especially obvious towards the midpoint of the valley where a valley constriction occurs owing to the presence of a more significant colluvial fan. The toe of the fan is scarped into a steep vegetated and bouldered slope on the edge of the river channel as the secondary braided channels are forced to converge before fanning out again to the lee of the constriction. Occasional non-perennial tributaries dissect these fans and there is evidence of rill erosion on the eastern slopes of the valley.

The road along the eastern flank of the valley crosses over slackwater deposits, which are especially obvious at the proximal end of the valley and upstream of the valley constriction. Flood and recirculation eddies and overbank flow have deposited finer materials that are subsequently being eroded by a more energetic system over time. They are indicative of flood activity along the valley. Coarser sediment is buried below the slackwater deposits and remains so as it is not an active component within the system as are the coarser elements present in active channels (Figure 5.3).

Figure 5.3: Slackwater deposit at MG1A, note buried clasts (Cl) and mud drapes (MD)

The contemporary braided channels are separated by channel bars consisting of boulders and smaller sized poorly-sorted sediment. Interfluves between the braided channels spread across the valley. Towards the distal section of the fan, the active channels indicate anabranching separated out by floodplain material. These networks of channels bifurcate or split and later regroup downstream; some of these active channels are incised into the floodplain deposits and even undercut the braided channel banks. As the valley widens, there appears to be greater prevalence of these incised channels in comparison to shallower channels separated by cobble bars in places upstream. There is, notably, a higher presence of shale upstream; where the shale and sandstone clasts remain relatively unweathered, although occasional boulders on the higher slopes of the valley exhibit exfoliation. Flood channels occur on either sides but predominantly on the eastern fringe of active perennial baseflow channels. Within these perennial streams there is significantly higher flow and even some ponding. In the north of the valley the channels are wider with more substantial channel bars and small-scale debris dams. There is even evidence of such debris dams within flood channels.

A well-defined small hill or *koppie*, Singkop (606m), an outcrop of Bokkeveld shale, occurs at the northwestern distal end of the valley. The saddle, over which the main 4X4 track passes to enter either the Moordenaarsgat River valley or to travel on to Eselbank, demarcates the division between the Moordenaarsgat and Dassieboskloof rivers and their catchments. It is on the opposite side of this saddle that the primary site or terrace occurs. Passing over this saddle along the road, there is sheetwash and rill erosion on the koppie side with a shallow gully running on the opposite side of the road. A small farm, growing wheat and vegetables, is located at the northwestern extremity of the Moordenaarsgat River valley against Singkop. There is evidence that channels of the Moordenaarsgat River meandered across both sides of the valley as a wide, shallow dry channel occurs at the gate of the farm and along the southern margin of Singkop it appears to have undercut the slope, although in contemporary conditions the channel is quite small. Meanders of active channels elsewhere have eroded into talus fans against the base of the slopes. This is seen along the eastern flank where undercutting is occurring.

There are no obvious evident relict alluvium packages evident along the valley that show any resemblance or even similarity to the primary site along the Dassieboskloof River. Based on the location of extra channel and slackwater deposits and inactive contemporary flood channels, floods disperse to the eastern flank of the river. To the lee of the valley constriction there are obvious flood channels containing considerable cobble lags with signs of imbrication, whilst elevated flow channels are more spread out with boulders and cobbles scattered across the surface with increasing distance away to the east of the channels.

- *MG1*

Surface samples were taken from the slackwater deposit (A), inactive contemporary channel (B), channel bar (C) and extra channel deposit (D) approximately 3.25km southwest of Wuppertal at the proximal end of the alluvial fan. It is noteworthy that towards the edge of the slackwater deposit, exposed by side channel incision, coarse buried materials (i.e. cobbles, etc.) were observed, although not sampled as they are not an active component of the system. Microfeatures were noted on the fringe of the slackwater deposit, namely mud drapes and wavey and flaser bedding.

- *MG2*

Surface samples were collected from the slackwater deposit (A), channel bar (B) and side channel (C) roughly 2.5 km south west of Wuppertal just ahead of the valley constriction, shown below in Figure 5.4. The active side channel contained pools and possibly seeps and springs. The gradient between MG1 and MG2 is relatively gentle at 1:75.

Figure 5.4: Location of sample site MG2A, a slackwater deposit upstream of colluvial fan valley constriction

• ***MG3 and MG4***

Surface samples were taken from contemporary active channels to the north of the approximate midpoint of the alluvial fan, some 1.8km southwest of Wuppertal (Figure 5.5). MG3 is an active lateral contemporary channel on the eastern flank of the fan while MG4 is an active primary channel from the centre of the fan. Notable features include the substantial boulder and cobble lag within the channels and to a lesser degree across the raised flood deposits. Small channel bars comprising of cobbles and coarse sand are obvious along with debris dams within the active channels. The gradient between MG2 and MG3/MG4 increases subtly to 1:70.

Figure 5.5: Sample sites MG3 (left) and MG4 (right) along active lateral and primary channels respectively

5.2.2. Tra-Tra River Sites (TT)

The Tra-Tra River displays similar characteristics to the Moordenaarsgat River except on a larger scale. It is typical of an upper to middle course of river with channel bars, braiding, etc. The valley also has a flat valley floor but is wider than that of the Moordenaarsgat River valley; it also contains sections of bedrock constrictions along the valley where deformation of the geology, manifested as folding, is clearly obvious and exposed. The braided perennial trunk channel of the Tra-Tra River flows centrally within the valley. The gradient is very low and gentle along the profile of the river, and the sandy sediment displays anastomosed channels. Downstream the larger and deeper channels meander across the valley floor and bifurcate with some rejoining forming anabranching channels that incise through floodplain material. This is clearly evident in Figure 5.6. The channel bars and interfluves spread across the valley comprise of poorly sorted sediment with pebbles and cobbles and occasional boulders. The height difference of the valley within the study area is between 475m and 435m, representing a gradient of 1:115.

Figure 5.6: Tra-Tra River viewed westward (upstream) towards Wuppertal showing elements of braided active channels (B), anabranching (An) and anastomosing (As) lateral channels through floodplain material and pooling

Secondary and lateral channels are also braided and are clearly incised in places. Channel sediment is generally coarse and medium sand with pebbles and cobbles with some indication of imbrication, although not well defined. Dry and abandoned channels also display some braiding and small-scale incising. As the valley widens, the channels meander and widen themselves in places with vegetated and raised islands of floodplain material occurring. Flood channels and flood deposits occur along the flanks. This is particularly obvious in places where elevated flow channels with partly rounded boulder and cobble lags infilled by sand (as floods recede) are evident along the edges of the contemporary active channels (Figure 5.7). There are also occasional pockets of finer sediment deposits adjacent to these lags resembling slackwater deposits. Flood and lateral channels occur on both sides of the valley but predominantly on the northern fringe of perennial baseflow channels. Within the perennial streams across the valley there is considerably higher flow with numerous sections of ponding and pooling and possibly springs or seeps. Again, there is a lack of any obvious relict alluvium packages present along the valley that resemble the primary site on the Dassieboskloof River.

Figure 5.7: Elevated flow channel along the northern bank of the Tra-Tra River showing evidence of cobble and boulder lag infilled by finer sediment during flood and subsequent recession

As the valley walls are further apart and less steep, colluvial fans are not as noticeable, although they are larger, more dispersed and diffuse with a lower gradient. The talus slopes at the base of the rock faces feed sediment and debris material into the colluvium. The road on the northern flank of the valley runs along this colluvium and, in places, talus slopes. It is raised above the floor of the valley and a steep slope falls towards the more lateral flood and side channels that have scarped the toe of these colluvial fans.

- ***TT1 and TT2***

TT1 is a contemporary active fluvial sediment sample taken from a channel deposit adjacent to an elevated flow channel displaying an extensive boulder and cobble lag approximately 5.5km north east of Wuppertal. Sample site TT2, shown in Figure 5.8 below, was taken at a secondary or lateral channel sand bank profile. The sandy sediment appeared uniformly unstratified with a cobble lag present at the base of the profile at 1.37m in depth (from the surface). The gradient is very gentle between TT2 and TT1 downstream, at 1:150.

Figure 5.8: Location of sampling site TT2 (beneath the tree) that comprised a sand bank profile

5.2.3. Dassieboskloof River (DBK)

The Dassieboskloof River leaves a relatively open asymmetrical bedrock canyon as it exits the eastern edge of the Cederberg Mountains. The contemporary trunk bedrock channel is relatively undisturbed and flows through significant boulder, cobble and pebble sandstone material. Figure 5.9 shows the canyon to be filled with coarse elements, especially boulders, along its flanks and at the base of the valley walls as talus slopes and bedrock screes. The stream sediment is coarse in nature with cobble and pebble elements present within the shallow channel. There is no braiding, although lateral flood channels that are not well defined are present along the fringes with evidence of pooling within the rocks. These channels are largely identifiable due to the vegetation pattern and the presence of imbricated cobbles and pebbles central to coarser material on their edges (noted in Figure 5.9). There is also evidence of debris dams and chatter marks (pitting on the surface of the cobbles) on the coarse material from transport. Finer sandy sediment is banked and stabilised by vegetation along the edges of the perennial stream.

Figure 5.9: Boulder and cobble elements of Dassieboskloof River upstream of the main exposure

As the river exits the mouth of the canyon, the channel width potential increases dramatically. The gradient of the active channel decreases slightly although it continues to flow eastwards within its minorly incised channel. A notable feature is that there are numerous lateral flood and dry secondary channels comprised of pebble and cobble beds and lags evident separated by bleached and vegetated sandy channel bars. These lags exhibit imbrication and some pebbles are well rounded within the dry channel bed. These channels meander within the area, even undercutting the terrace that runs along the southern flank. There is also evidence of pooling and debris dams within the active as well as abandoned flood channels.

Figure 5.10: Primary site (grey package (WT4) overlying orange package (WT5) extending across the centre) viewed east towards the western flank of Singkop (note 4X4 car for scale towards midpoint)

The primary site of the study falls along the southern fringe of this section of the Dassieboskloof River as an exposed terrace that extends for roughly 300m with a maximum height or elevation of approximately 8.5m (Figure 5.10). The exposure is clearly divided into two packages of grey, sandy partially consolidated alluvium above, and orange, more consolidated colluvium below, with a sharply defined boundary between the two. The contact, or unconformity, varies in height from just over 3m at the distal, more easterly end of the terrace, to 4m towards the centre and then falls to 2m and eventually the level of the channel at the most westerly or proximal point. The grey upper package becomes shallower and peters out towards the distal end of the exposure near the road. This upper package, for convenience, was separated into an upper and lower sampling unit. The upper surface of the terrace has been greatly disturbed by vehicles, dumping and local sand mining. The face of the terrace has experienced slumping and avalanching of the upper package and differential weathering exposing cobble and pebble lags, laminations and in places has been used as a nesting site for birds and some clasts have been washed out of the exposed face. There is a cobble lag near the surface of alluvium in the proximal corner comprised of angular shale, sandstone and conglomerate (where the conglomerates are made up of granules and small pebbles). Figure 5.11 shows a cobble and pebble lag is observed persisting along the upper surface of the contact between the alluvium and colluvium. Coarse subangular to subrounded pebble and cobble lags and lenses comprised of small fragments of Bokkeveld parent material are evident with no imbrication within the lower package, especially as a laterally persistent bed at roughly 2m from the channel bed. The flood channel has truncated the toe of the colluvium and in places undercut the exposure and exposed bedrock. A levee has been built up along the northern flank of the river to protect vegetable and cultivated fields behind it that extend towards the town of Wuppertal. The levee, at roughly 1.5m in height, consists of grey vegetated sandy material that resembles the alluvium of the opposite bank. The fields used for agriculture are also grey and sandy in nature and appear to be developed on alluvium (Figure 5.12)

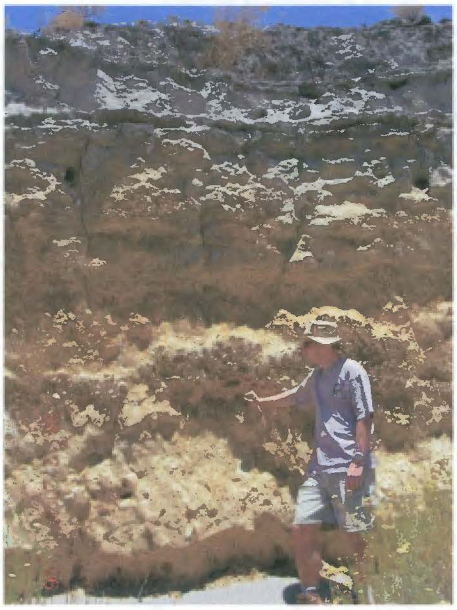


Figure 5.11: Exposed profile of lower colluvial package overlain by upper alluvium package (note the unconformity and pebble lags)

Figure 5.12: Upper package of primary exposure visible from behind the levee on the opposite bank

- ***Wuppertal 3 (WT3) – Upper unit of upper package***

The upper unit of alluvium, shown in Figure 5.13, consists of stratified sandy sediment. It is normally layered and can be divided into upper brown beds and grey beds below with wavey laminations. The laminations are fine between 10 and 20cm below the sheetwash material of the surface and become more clearly defined from 43cm downwards. The sediment is largely unconsolidated and apedal. There are occasional coarser lenses of sediment below 1.5m.

- ***Wuppertal 4 (WT4) – Lower unit of upper package***

The lower unit of alluvium bears strong similarities with the upper unit (Figure 5.123. Grey sandy sediment is normally layered with wavey parallel low angle laminations extending laterally along the exposure. The laminations persist through the upper 1.5m of the profile being slightly thinner and closer together nearer the surface in contrast to being thicker and more dispersed with depth. Below 1.5m, the loosely consolidated apedal stratified sandy sediment forms alternating beds defined by different shades of grey. Lenses of coarser sediment with gravel lags occur below 2m with occasional pebble lags around of 2.5m and below. These lags become more frequent, coarser and more clearly defined towards the contact with the lower package.

Figure 5.13: Upper grey alluvium package (WT3 on left, WT4 on right) with obvious laminations

- ***Wuppertal (WT5) – Lower package***

Extensive colluvial deposits are a feature of both the Cape Fold Mountains as well as the Great Escarpment (Boardman *et al.*, 2005). The lower package comprises of partly consolidated orangey brown sediment with columnar and occasional prismatic peds , shown below in Figure 5.14. The surface is partially consolidated apedal sheetwash and alluvium material. Shale bedrock belonging to the Bokkeveld Group is exposed at channel depth at the base of the exposure face and as occasional small outcrops along the base of the exposure. The river channel has scarped the base of the colluvium deposit. The sediment contains poorly sorted clasts with an unordered fabric. The exposure appears massive in nature, although colour banding indicates some stratification within the matrix-supported sediment. Below 40cm the sediment is calcretised, particularly from 240 and 300cm. Calcified root castes are also observed.

Figure 5.14: Lower orange package showing columnar, prismatic peds and apedal surface material

5.3. MAPPING AND REMOTE SENSING

5.3.1. Profile Surveys

In the Figure 5.15, the star represents the position of the theodolite on a raised channel bar within the dry lateral channels on the southern flank of the Dassieboskloof River, while the contemporary active channel of the river is indicated by an arrow. The levee along the northern flank of the river is maintained at approximately 1.5m in height above the stream bed and comprises of vegetated sandy sediment embankment. At this point the width of the Dassieboskloof River bed is 155m. This plot illustrates the problem of perspective in representing three-dimensional elements on a two-dimensional graph. As the exposure is curved laterally relative to the basal position of the theodolite, this is not considered and not apparent when viewed as a set of profiles below. There is also the problem that as the face itself changes angle and slopes away at the top surface, it again distorts the perspective. The plot does, however, aid in providing an overall or generalised view of the measured profiles as if viewed from the side and shows the relative position of the basal channel level as it varies according to the amount of debris or material at the base of the slopes as well as collected modern fluvial deposits, such as the sand that is evident across the dry flood beds and channel bars and extra channel deposits. This basal measurement also indicates a relative incline in the exposure towards the proximal or canyon end as well as showing the channel bar to be raised. The contact line is relatively difficult to read from this so an amended diagram (Figure 5.16) aims to help better identify its relative position.

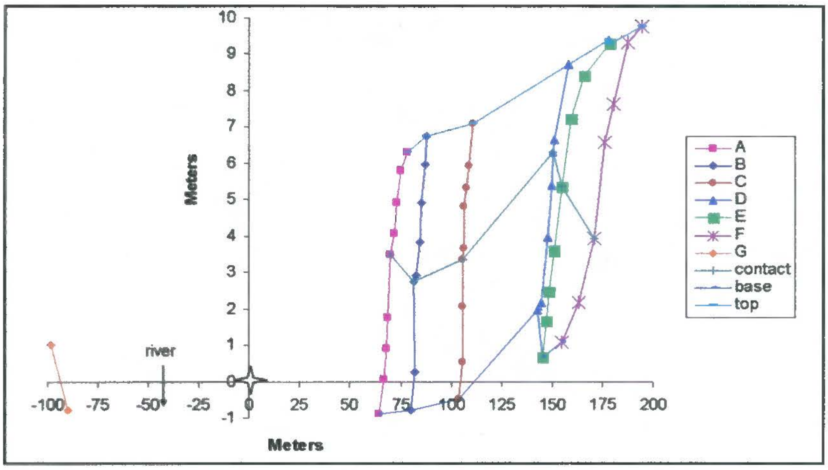


Figure 5.15: Surveyed profiles of the terrace

The contact between the upper and lower packages extends at varying elevations from 3m above the contemporary exposure base or channel bed, to just over 4m at the centre of the terrace and then falls to less than 3m towards the proximal end until it reaches channel level at the exit of the canyon. The plot in Figure 5.16, although slightly clearer, is also misleading as the contact itself reaches channel bed level as the alluvium tapers off and it becomes rather a contact between the talus slope and outcropping bedrock covered by alluvium at the proximal end at the head of the canyon. This is also the point where the channel bed has undercut the terrace.

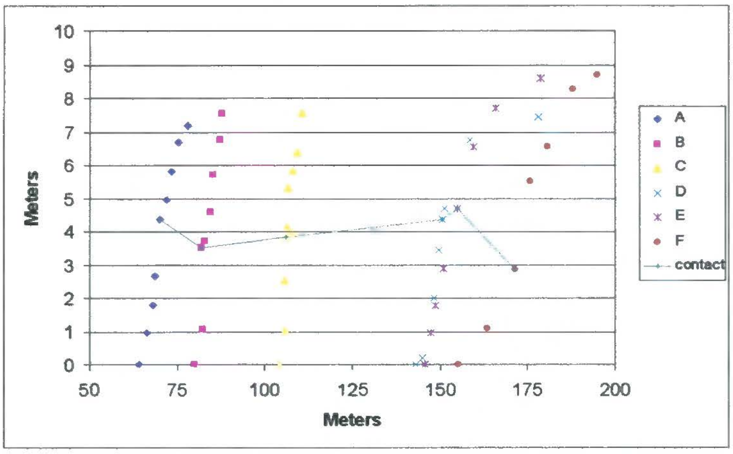


Figure 5.16: Profiles adjusted to display the contact depth

Both types of surveyed plots depict the shape of the exposure or terrace face and the relative heights of the deposit, its maximum elevation and point of contact where the unconformity lies between the alluvium and colluvium packages.

Cross-sectional profiles of the Moordenaarsgat, Tra-Tra and Dassieboskloof river valleys shows the varying nature of the valley shape along the river profile between sites as well as showing the separation of the adjacent valleys by Singkop are presented as Figure 5.17. These also identify the relative active trunk channel position. The first profile shows a V-shaped ravine at the proximal fanhead within the Moordenaarsgat River, subsequently transitioning to a narrow flat valley floor shallowly incised by braided active perennial channels. The Tra-Tra River valley is clearly seen as wider and slightly flatter although also with deeper more incised active channels.

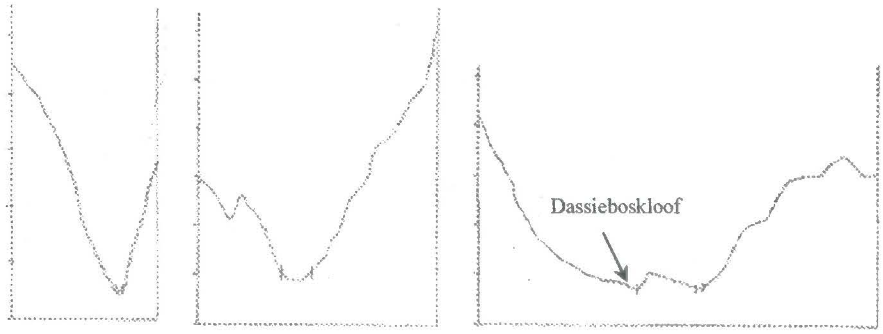


Figure 5.17: Cross sections of the Moordenaarsgat valley from the ravine (left) to before its convergence with the Dassieboskloof River

5.3.2. Drainage

The drainage basin is the fundamental landscape unit concerned with the collection and distribution of water and sediment and can be viewed as a geomorphic system or unit. Every basin and network possesses a set of descriptive, measurable geometric and topological properties (basin morphometry) that define the linear, areal and relief characteristics of the watershed basin, density and pattern; these reflect the principal catchment attributes including stage of development and catchment shape (Ritter *et al.*, 1995; Briggs *et al.*, 1998). Drainage is structurally controlled by the underlying geology and its inherent structure (jointing, faults, dipping strata, etc.) across the region. Overall, the Moordenaarsgat and Tra-Tra rivers exhibit similar drainage characteristics and properties, most likely as a result of the common presence of the Bokkeveld group geology, lower relief and relatively smaller non-perennial immediate catchment (and to a far lesser degree, similar vegetation). The Dassieboskloof River appears to have more unique drainage properties as it flows eastwards out of the higher relief Cederberg Mountain Range with distinctive geology (Table Mountain Series) and a greater catchment, as well as markedly different precipitation and vegetation patterns.

According to the Strahler (1952) scheme of stream ordering, both the Moordenaarsgat and Tra-Tra rivers are third order rivers, while the Dassieboskloof River is a second order stream. The simpler unambiguous Strahler system is firmly established and generally accepted and was therefore utilised instead of the Shreve or link magnitude scheme, which is more commonly employed in runoff models (Knighton, 1996). The stream magnitudes have the potential to provide a false representation of the relationship between rainfall and runoff. This is due to the short ephemeral nature of tributaries along the Moordenaarsgat and Tra-Tra rivers in comparison to the much more developed and extensive tributary networks that extend westward into the high plateau of the Cederberg belonging to the Dassieboskloof River, and to a lesser degree, the Eselbank River, a more significant tributary of the Moordenaarsgat River. The sinuosity of a river is a function of the ratio of the measured length of the stream channel against that of the thalweg of its valley (Chorley, 1973). An approximation of both the Moordenaarsgat and Tra-Tra rivers is that they exhibit a transitional stream pattern (low sinuosity) whereas the Dassieboskloof River has a more regular stream pattern (higher sinuosity).

The Dassieboskloof River catchment is approximately 110km^2, with its most distant point on the perimeter of the drainage basin at 12km, and the Moordenaarsgat River's is 130km^2, where its furthest point is 21km. In contrast, the Tra-Tra River has a catchment, when including the major south eastern tributary of the Matjiesfontein River but excluding the watersheds of the Dassieboskloof and Moordenaarsgat rivers, of just over 400km^2. If these two main acute tributaries of the Tra-Tra River are included as part of its greater catchment, it then exceeds 640km^2. These are considered as relatively micro-scale catchments as there are less than 5000km^2, however, the greater catchment of the Doring and Olifants rivers could be regarded as a meso-scale catchment (Bull, 1991). The Moordenaarsgat River catchment is adjacent to the Driehoek River, which is a westerly and interior Cederberg tributary of the Matjies River, an upper reaches tributary along with the Groot River of the Doring River that drains into the western fringe of the Ceres Karoo. The Dassieboskloof River watershed is adjacent to the Heuningvlei River to the north, which forms the western upper reaches of the Biedouw River, and the west draining Jan Dissels River. The Tra-Tra River is separated from the Moordenaarsgat River by the Vaalheuningberg interfluve to the west whilst the Tra-Tra Mountains form a substantial divide between it and the Biedouw drainage basin to the north. The Matjies and Doring River catchments are adjacent to the south and east respectively.

Geology influences drainage density, which is inversely proportional to permeability and is central in the creation of a drainage pattern, which in turn also strongly reflects the underlying geology (Briggs *et al.*, 1998). The Moordenaarsgat and Tra-Tra basins exhibit a trellis drainage pattern. This is expected as it is characteristic of dipping or folded sedimentary rocks (Ritter *et al.*, 1995) and the alternating harder and softer bands of the underlying Bokkeveld Series. The drainage pattern of the Dassieboskloof catchment can chiefly be described as dendritic with some trellis elements obvious on certain tributaries to the north. This is accounted for by the uniformly resistant geology of the quartzitic sandstone of the Table Mountain Series over most of its basin with a transition over a shale and conglomerate Cederberg formation to sandstone with shale where the trellising occurs. This trellis pattern occurs across the study area where small tributaries are exploiting the faults and jointing through the shale and sandstone formations. Drainage densities also reflect catchment variables, channel network efficiency and maturity. All three drainage basins are characteristic of moderate drainage densities assessed by approximation and according to Chorley (1973), although the Dassieboskloof River exhibits a higher drainage density than the Moordenaarsgat River, while both have higher values than the Tra-Tra River.

The catchment shape, defined by plan circularity or elongation and size, affects both the channel network and the hydrographic character of discharge (Briggs *et al.*, 1998). The drainage basin character of both the Moordenaarsgat and Tra-Tra rivers is elongated as a result of structural effects, whereas the Dassieboskloof basin shape that covers more homogeneous bedrock from one area to another is more "tree-like" and shows a greater degree of circularity. The degree of branching, or bifurcation ratio, for a given density of drainage lines is largely controlled by basin shape and influences the landscape morphometry; it is also an important indicator and control of the 'peakedness' of the runoff hydrograph. For the Dassieboskloof River, the bifurcation ratio is relatively high at 9-11, due to the ephemeral nature of first order streams in contrast to the mostly perennial second order streams. There is a high prevalence of non-perennial streams within the catchment of all three rivers, especially the Tra-Tra watershed where predominantly only the trunk tributaries, namely Tra-Tra and Matjiesfontein rivers, are perennial. Most short tributaries of the trunk streams within all three drainage basins are non-perennial. There is also a high frequency of obtuse side tributaries along the Tra-Tra and Moordenaarsgat, with greater prevalence of both acute and obtuse side tributaries within the Dassieboskloof basin. Figure 5.18 provides a basic drainage basin map for the localised study area of the Moordenaarsgat, Dassieboskloof and Tra-Tra rivers including spot heights and the primary site's location.

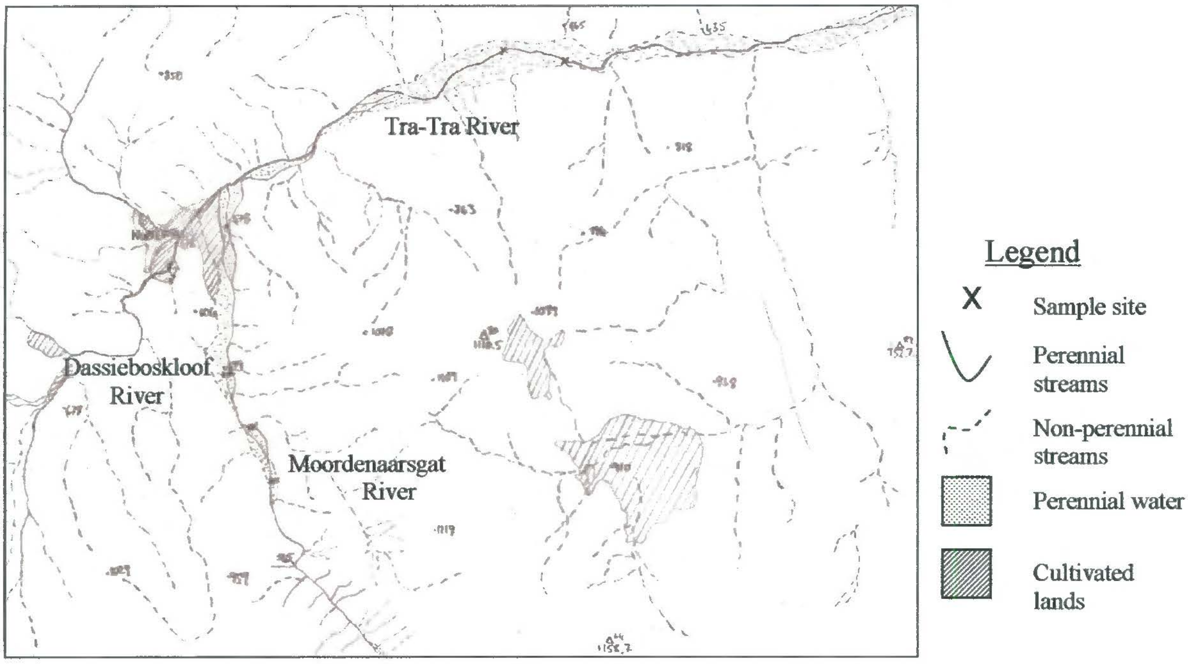


Figure 5.18: Sketch map of three rivers' drainage in the study area including spot heights (1:50 000)

Through a broad assessment of the drainage characteristics of the three linked drainage basins one can distinguish the unique features of each system along with their similarities and combined properties as a whole geomorphic unit. It is, however, clear that, while the Tra-Tra and Moordenaarsgat catchments share many attributes, Dassieboskloof is distinct and is likely, therefore, to behave and respond differently to and independently of the two relatively larger systems. For this study the focus falls on the Dassieboskloof watershed as it provides the main point of entry for water input and erosive potential at the primary site to the south west of Wuppertal. Moordenaarsgat is viewed more in terms of a contemporary analogue of fluvial conditions of a tributary of the Tra-Tra River, while Tra-Tra is the result of the confluence, input and transition of the two other rivers. Another key point to consider subsequent to briefly assessing the drainage of the immediate region surrounding the study sites is that both the Moordenaarsgat and Dassieboskloof rivers have a significant potential for input that converges on and passes through narrow constrictions directly before converging to form the Tra-Tra River. This is particularly pertinent to the WT locales on the Dassieboskloof River.

5.4. SEDIMENTOLOGY

5.4.1. Introduction

The results of the sedimentological analyses for the samples taken are presented according to the location/site from where they were sampled. Detailed data are visualised in graphical formats to aid comparative and spatial reference within the site. For a full data summary matrix of all variables, refer to Appendix 5. It should be mentioned that, according to the South African system, soils and sediments are described in terms of "textural classes" although the material is not necessarily 'soil'.

Particle size is a fundamental property of sedimentary materials and provides vital information about its origins and history; in particular, the dynamical conditions of transport and deposition of the constituent particles are usually inferred from their size and shape. It is a very useful descriptive property and analytical tool. The size distribution is also an essential property for assessing the likely behaviour of the sediment under fluid or gravitational forces (Syvitski, 1991). The Folk and Ward method was utilised in grain size analysis and the modal and mean values are given as well as sorting, kurtosis and skewness (symmetry).

Computer processing facilitates the rapid computation of mean and median grain size, standard deviation, skewness and other statistical properties. These properties of grain size distributions are fundamental descriptive parameters when used in sedimentological analysis as part of a study (Leeder, 1991). Grain size analysis is an indicator of subtle differences which are exposed within the statistics revealed by measurement i.e. change in mean/median grain size, sorting, kurtosis, modality, etc. and with the aid of additional plots of one variable against another to test their co-variance, further information to the nature of the sediments is revealed.

5.4.2. Textural Analysis

The sediments within the study area consist predominantly of a mixture of quartz and feldspars, a small amount of precipitated carbonates, mud and some weathered material as well as clasts of quartzitic sandstone and arenaceous shale. All the fluvial system samples comprised of majority sand-sized particles while the colluvium contained a smaller proportion of sand. Silt and clay was present in all samples, predominantly in the colluvial sediments, laminations and slackwater deposits; although throughout the alluvium the quantity of clay and silt collectively remained generally below 10% with slightly higher proportions in contemporary flood deposits and at the base of the upper unit of alluvium recorded. As the sand fraction of the sediments comprised in most cases the principal component (more than 90%) of the sample and would therefore provide the most informative data, textural analysis was principally applied to this fraction only (Folk & Ward, 1957). In samples where clay and silt percentages were notably high, such as the colluvium and slackwater deposits, pipette analysis was used to further analyse the relative proportions of clay and silt. There is, however, a lack of accuracy associated with the pipette measurement method in smaller proportioned samples that requires one to maintain a degree of objectivity and to decide if the results are applicable so as to limit any skewing of the results (Rogers *pers. comm.*).

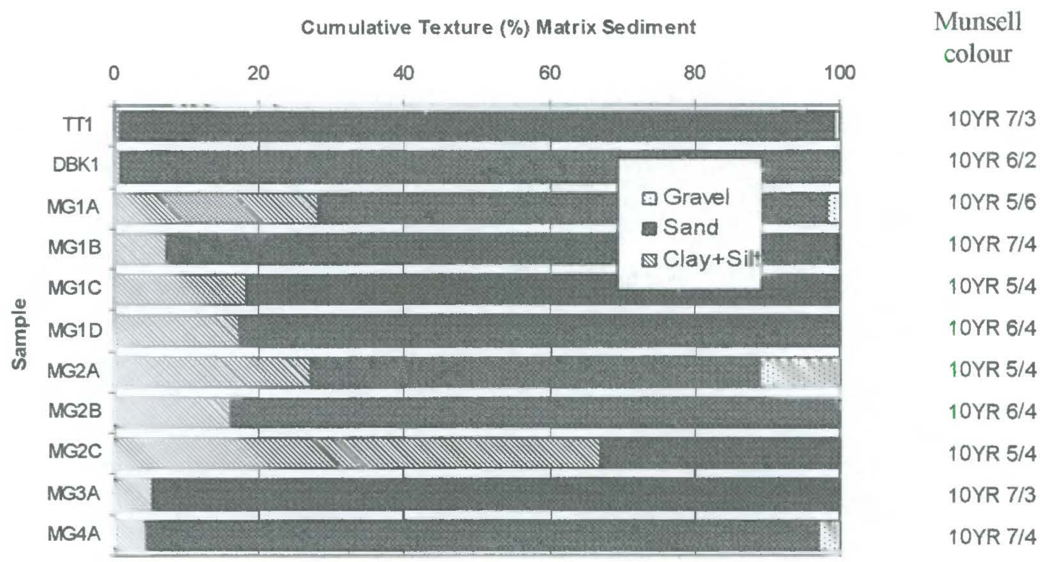


Figure 5.19: Contemporary channel sample texture and colour from Moordenaarsgat (MG), Tra-Tra (TT) and Dassieboskloof (DBK) rivers

The contemporary samples from across the three valleys exhibit a variety of textures, as shown by Figure 5.19. Notably the contemporary active channels of the Dassieboskloof and Tra-Tra rivers comprise of 98% sand with only tiny fractions of gravel, silt and clay. In comparison, the active channel sediments of Moordenaarsgat River (MG3A and MG4A), although also predominantly sand, consist of 4-5% silt and clay. The slackwater deposits (MG1A and MG2A) contain a relatively higher percentage of silt, clay and gravel, especially MG2A. Surprisingly MG2C, a side channel, comprises of 67% clay and silt together. Gravel is mostly absent from these samples with the exception of the slackwater deposits and active channels. In the above plot, it would not be logical to join points between samples as it is not representative of any depth within a profile rather that they are individual surface samples.

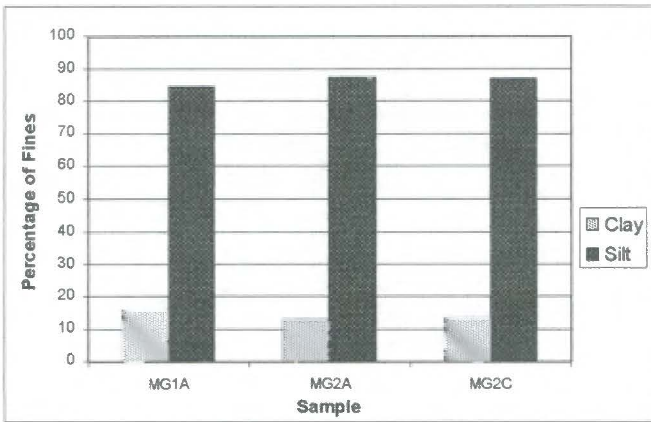


Figure 5.20: Comparative clay and silt fractions of contemporary channel deposits with >20% fines

Figure 5.20 above shows that the fines fraction of the contemporary samples taken within the Moordenaarsgat River valley, regardless of the relative proportion of the total sample, all exhibit a very similar percentage of clay and silt. Silt dominates with 84% (MG1A – slackwater deposit), 86% (MG2C – side channel) and 87% (MG2A – slackwater deposit) of the total fines.

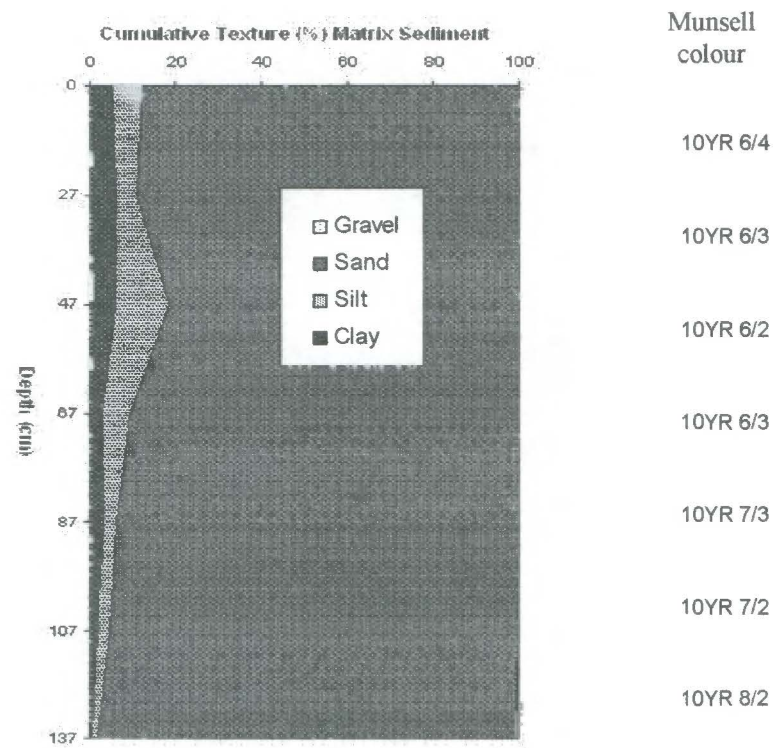


Figure 5.21: Cumulative texture and colour of TT2 sand bank profile along Tra-Tra River

The TT2 profile is sandy with almost no gravel fraction except for the basal sample that contained a very small percentage (0.083%). Overall the silt and clay (mud) fraction is relatively small throughout the profile with a slightly muddier lens occurring at 47cm and a general decrease in their component with depth. Silt is again more dominant in composition than clay, particularly between 27cm and 87cm.

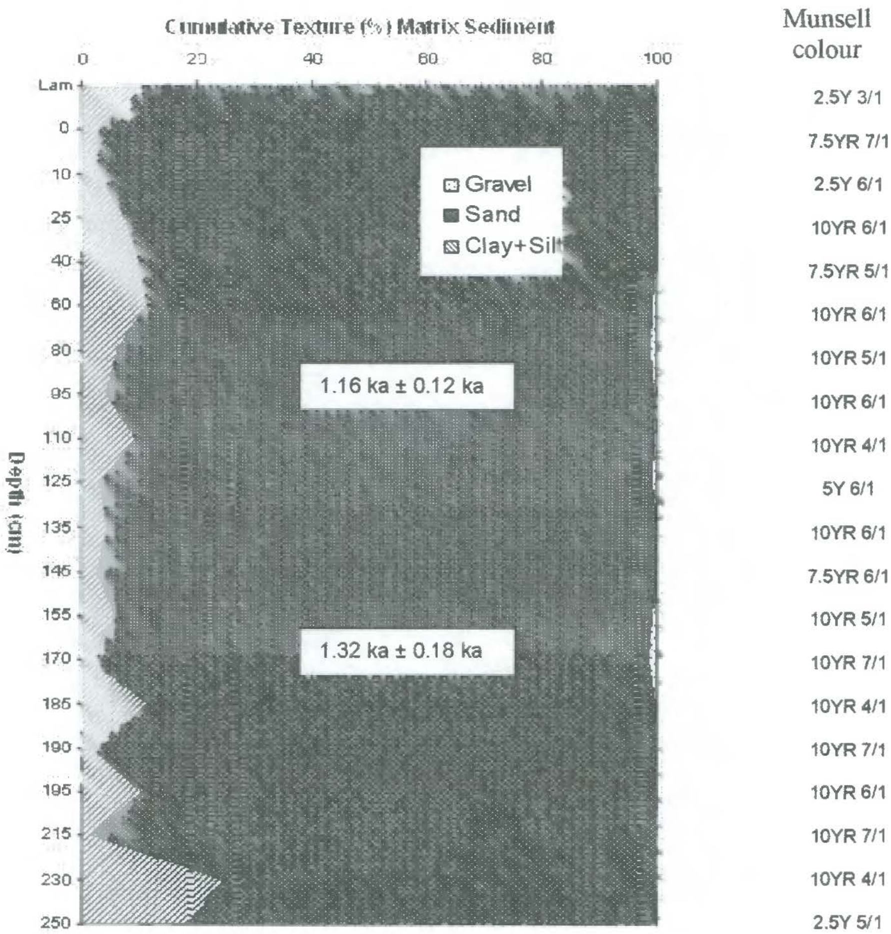


Figure 5.22: Cumulative texture and colour of WT3 including OSL ages at appropriate depths

WT3 profile (Figure 5.22 above) exhibits the sandy nature of the sediment with very small gravel components and small yet fluctuating amounts of clay and silt present within the beds. Where gravel does occur, the percentages remain below 1% of the total sample weight for all samples except at 80cm and 170cm. The fines component fluctuates but maintains a relatively small percentage, generally less than 10% of the total. Exceptions occur at 60cm, 110cm, 185cm, 195cm, 230cm and 250cm. Clay and silt appears to increase with respect to bedding structure in cycles. The lamination also has a higher combined clay and silt fraction (10.75%).

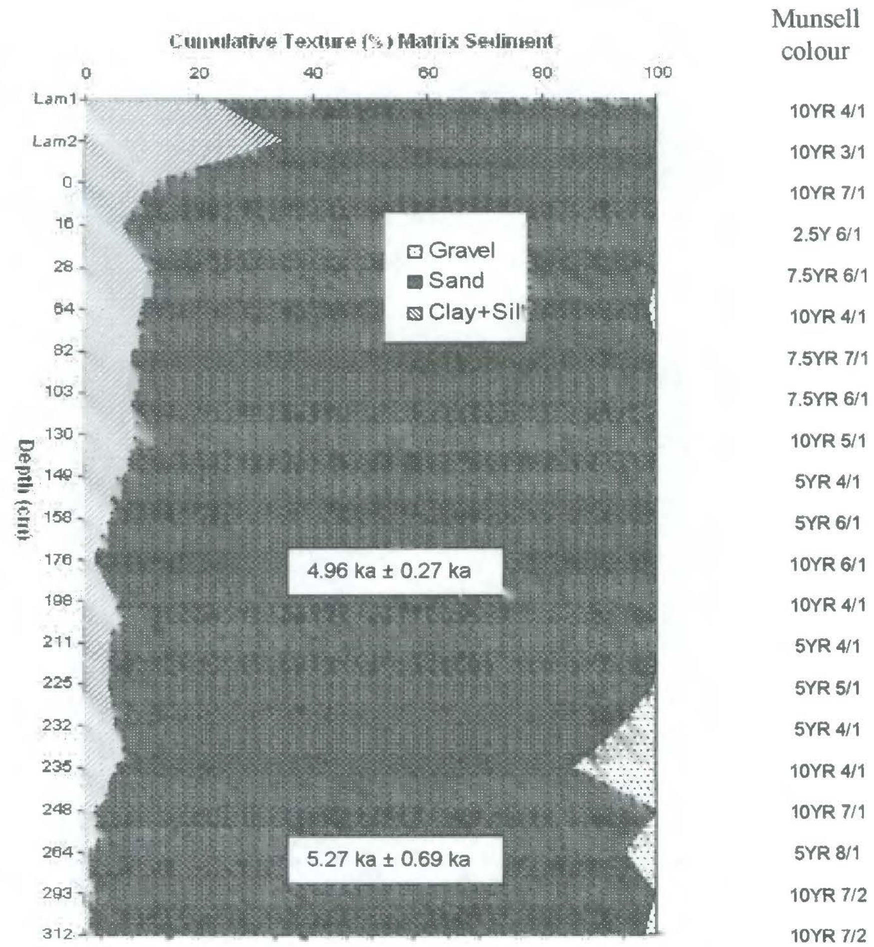


Figure 5.23: Cumulative texture and colour of WT4 including OSL ages at appropriate depths

Figure 5.23, the WT4 profile, also demonstrates the sandy nature of the sediment. The clay plus silt fraction overall has relatively higher and more constant values and decreases with depth to the base, just above the contact between the upper and lower units (alluvium and colluvium) at less than 1% of the total. The lamination samples from this profile have significantly higher fines percentages at 23 and 35%. Again the gravel component is very small at mostly less than 1% until closer to the base of the profile where there is a sharp increase at 235cm with lenses of more gravelly sand interspersed below along with the notable presence of pebble and gravel lags.

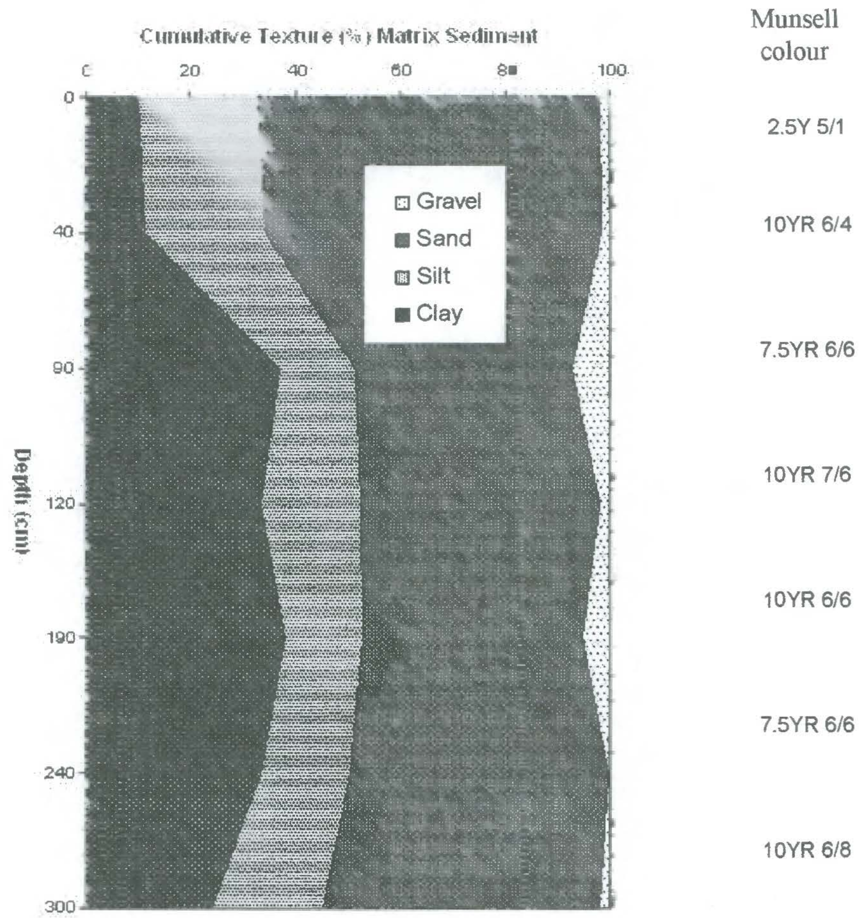


Figure 5.24: Cumulative texture and colour of WT5 lower package of exposure

WT5 exhibits a wider distribution of fractions, although it is still predominantly sandy. The clay and silt component increases considerably at 90cm where the sand fraction decreases and gravel increases. Downwards from this point, the textural nature of the profile remains very similar with a decrease in the cumulative clay and silt fraction towards the base, clear in Figure 5.24. Gravel fluctuates and ranges from 0.5 to 7%. In comparison to the more surficial samples, where the silt percentage is greater than clay, the remainder of the profile displays notably higher clay from 90cm again with a smoothed decrease towards the base.

Statistical analysis of the sand fraction grain-size distribution from the settling tube analysis yielded distinct characteristics within the deposits under investigation. The most observable trend was that all samples had unimodal distribution curves, even within the fluvial channel, and all cases were classed texturally as slightly gravelly sand. This indicates, despite the occurrence of two distinct deposits or packages (colluvium and alluvium), that the sand fraction at least, possesses broadly homogeneous grain-size characteristics. This is probably due to the provenance of the sand sized fraction, namely that the Cederberg Mountains consist of prominent Table Mountain Group sandstones as well as the sandstone formations within the Bokkeveld Group, although a lesser contributor. When disintegration and decomposition occurs to these quartz-arenite lithologies, the formed sediment retains the grain-size characteristics of the parent material and according to Boggs (1995) polycyclic quartz grains retain their form, and often grain-size characteristics, even after numerous sedimentary cycles. There was the expectation and presence of some skewing of the sample distribution, especially those samples taken within the vicinity or from lag horizons, due to a higher concentration of gravel-like fragments (although sieved, some samples still contained a large amount of coarse sand that penetrated the 2mm sieve); this fraction tends to skew the overall distribution as, while it may not constitute a dominant fraction, its mass (associated with its larger size and the nature of its material, i.e. quartz and feldspar, both dense minerals) is consequently measured as greater.

The grain-size distribution of sand contains much information about transport, deposition, or both. Within one profile of sand, different samples may exhibit distinctive characteristics as a result of several controls or effects, such as mass-density contrast between mineral in air and mineral in water, supply rates, through-flow, trapping, and winnowing, grain-grain interactions and transport directionality (Syvitski, 1991).

The sand grade of the samples ranged from fine to coarse sand (no very fine or very coarse sand) with the majority being medium sand. All colluvial and contemporary active channel samples were medium sand, except MG4A that was coarse. It is interesting to note the relative proportions of the sand-sized sediment as it is the dominant fraction and also provides an essential indicator.

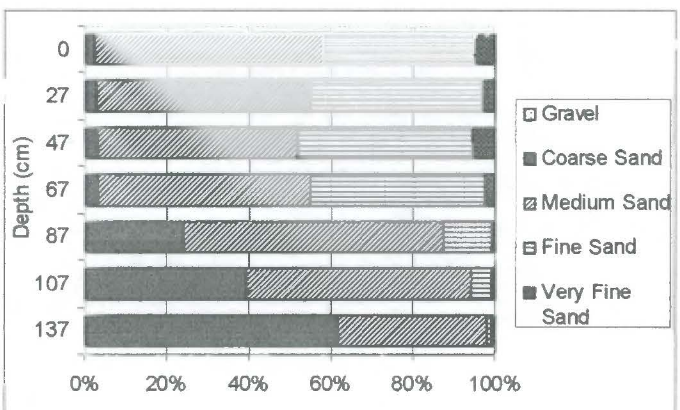


Figure 5.25: Relative sand textures of TT2 (sand bank profile along Tra-Tra River)

In the contemporary active channel sand bank of the Tra-Tra River, there is a distinct trend with increasing depth. The coarse sand percentage increases whilst the medium sand value remains relatively unchanged, until the base, however, it is the fine and very fine sand that decreases significantly.

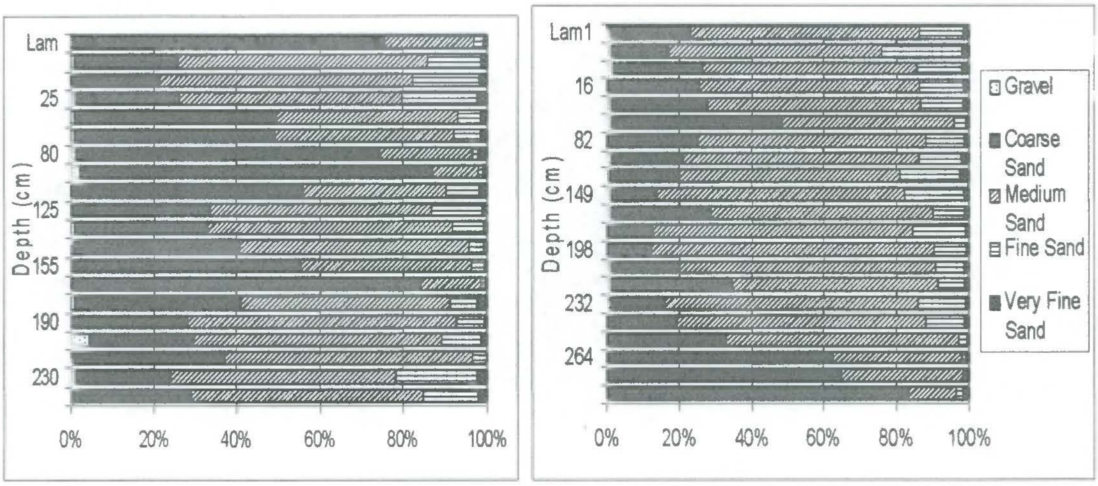


Figure 5.26: Relative sand textures of upper and lower units of the upper package of the primary site, WT3 (left) and WT4 (right) respectively

In the WT3 profile it is obvious that the sand grade is predominantly coarse and medium sand. The most notable feature is the presence of a pattern or cycle of increasing coarseness with depth followed by a rapid decrease and then subsequent increase again. There are a few lenses with more gravelly elements, close to the peak coarse proportion, while the percentage of very fine sand remains less than 2.5% and is maintained throughout the profile. In contrast, fine sand fluctuates greatly in keeping with the medium and coarse pattern, being greatest within the realm of the medium sand peak. The lamination is very coarse in nature.

A similar pattern is observed in the lower unit of the upper package (WT4). The profile again shows, although slightly less obviously, a fluctuating cycle of increasing and decreasing coarse and medium sized sand proportions. Again, a similar percentage of very fine sand (less than 2.5%) is maintained and present throughout the profile whilst fine sand varies in tune with the cycle. However, at the base of this profile coarse sand becomes dominant and fine and very fine sand is practically absent or at least very small in relative proportion. The laminations are less coarse in this profile.

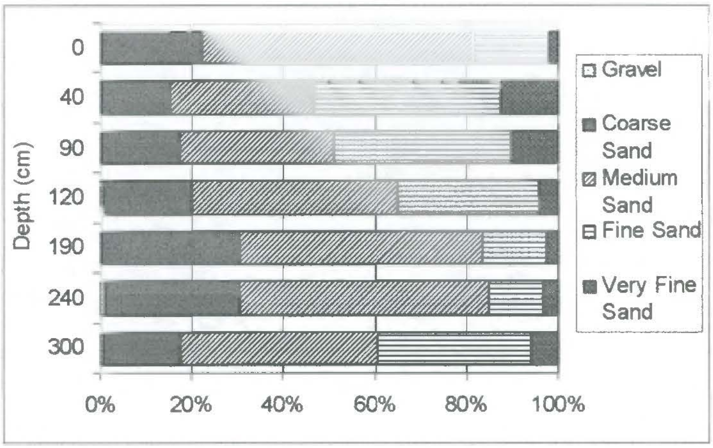


Figure 5.27: Relative sand textures of the lower package of primary site (WT5)

In contrast, WT5 comprises a wider relative range of various sand sized sediments. It clearly consists of more fine and very fine sand as well as occasional gravelly elements. It is predominantly medium sand. Interestingly, it does also appear to coarsen with increasing depth although there is a noticeably finer bed at the base of the profile.

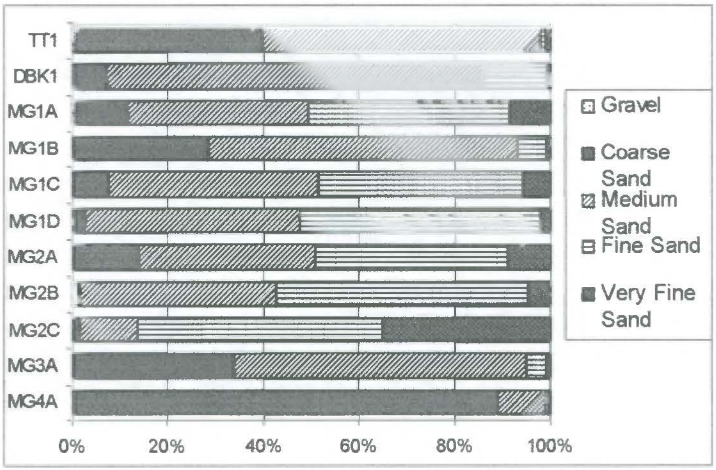


Figure 5.28: Relative sand textures of contemporary surface samples from all three rivers

The only samples classified as fine sand due to the dominant presence of fine and very fine sand grade sediment belonged to the channel bar and side channel of MG2, which comprised of high relative proportions of fine and very fine sand respectively. Very fine sand was almost absent in active contemporary channels as expected, which were mostly coarse and medium, especially MG4A, whilst very fine sand was evident in higher percentages along with fine sand in the slackwater deposits. The channel bar and extra channel deposit of MG1 also comprised a higher proportion of fine sand.

5.4.3. Results Based on Mean Grain Size, Sorting, Skewness and Kurtosis

Variables of grain size analysis, such as mean, standard deviation, kurtosis, etc. can be plotted and become more useful as they show certain aspects of sediment supply, hydrodynamics, etc. For classification and statistical parameters utilised in this study refer to Appendix 4. Textural analysis assists in the discrimination of depositional environments, showing how physical processes during transportation and at the depositional site impart a distinctive fingerprint to sediments (and particularly sand) texture (Leeder, 1991). The sediment grain-size, although containing numerous similarities, holds a distinctive signature inherited from the various modes of transport, aeolian or fluvial, and mechanism of deposition, mass movement or laminar. Mean grain size is an excellent indicator towards coarsening or fining sequences.

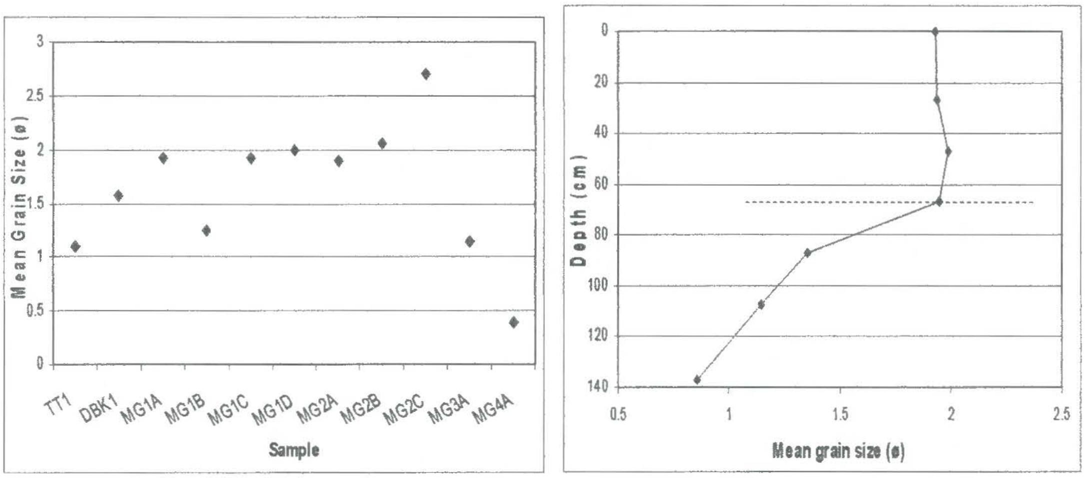


Figure 5.29: Mean grain size of surface samples (left) and active channel sand bank profile TT2 (right)

The mean grain size of the sand fraction from the surface samples (Figure 5.29) again demonstrates the wide range of sediment and texture as displayed previously. Contemporary active and lateral channels are coarser with slackwater deposits and channel bars displaying similar mean grain sizes just below 2 phi. In contrast, MG2C (a side channel) is a clear outlier with a relatively fine mean grain size. In comparison, the secondary active channel bank profile of TT2, another contemporary site, shows a similar mean around 2 phi units from the surface until roughly 70cm from where it begins to coarsen downwards.

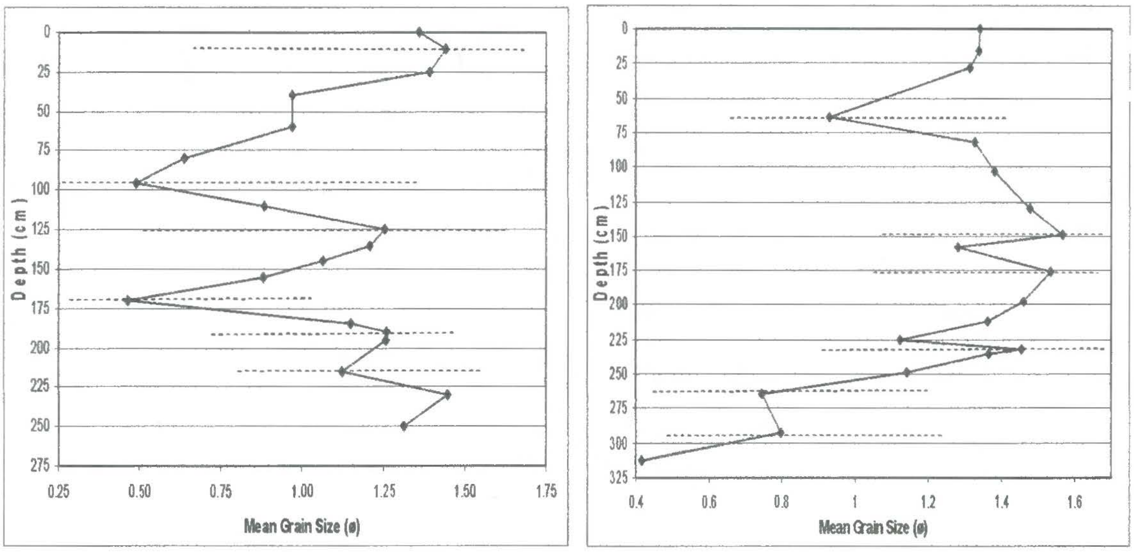


Figure 5.30: Mean grain size with depth of WT3 (left) and WT4 (right) profiles of upper package at primary site

Within the profiles of the upper package, using mean grain size, there is a clear presence of coarsening and fining upwards sequences across the beds of the units. Overall the lower unit (WT4) shows a fining upwards trend across the beds whilst the upper unit (WT3) generally displays coarser sediment towards the surface although there is a significant fining sequence from 1m upwards. Although not following directly on from WT3, the upper values for WT4 are 1.33 and 1.31 phi units whilst the basal value for WT3 is 1.31 phi, this could indicate a similarity in bed properties and proximity between the two units. Within the structure of the units there is a notable feature, namely the numerous sequences where the mean grain size fluctuates through cycles. These are seen as sharp changes in direction of the joined points between subsequent samples. If sampled at a higher resolution, the profile may more closely resemble the exact pattern as well as the depth of the boundary between such cycles or sequences. Beginning from the surface below the sheetwash material, within WT3 there are three alternating fining and coarsening sequences. Fining upwards occurs from approximately 1m with a coarsening sequence below this. Another significant fining upwards sequence follows between 170 and 125cm and the third 215-185cm. Each fining cycle is followed immediately by a shorter coarsening sequence. There is a fourth fining cycle at the base of this profile.

WT4 exhibits a similar set of alternating fining and coarsening upwards sequences. The surface sediment again represents a level of sheetwash material, however, starting from the base, or contact between the upper and lower packages, are four sets of alternating sequences followed by a fifth fining upwards sequence. In the first three sets of sequences the fining sequence is again longer, however, between 64 and 149cm there is a well developed coarsening upwards sequence. It is notable that except for the significant coarsening of sediment around the depth of 1m in WT4, the fining sequences are not only longer but cover a greater range of size, especially in WT4 and the latter half of WT3.

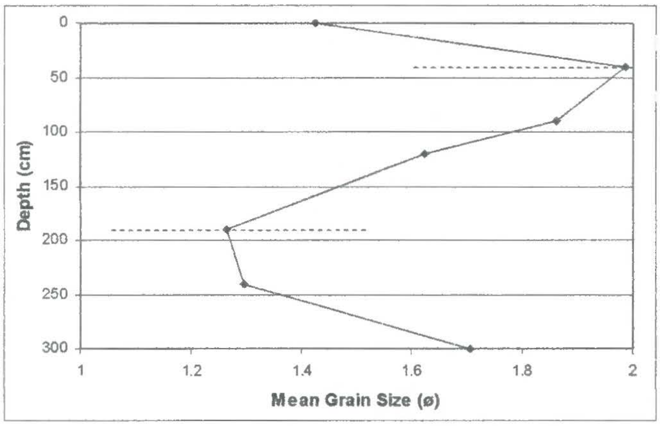


Figure 5.31: Mean grain size with depth of WT5 lower package profile

The mean grain size of WT5, as depicted in Figure 5.31, shows two distinct sequences within the profile. Aside from the surface material, which was of a different nature to the sediment below as it belonged to the package above and some sheetwash material too, there is a fining upwards sequence from 190cm. Sediment at the base of the profile is finer and coarsens towards this depth.

The standard deviation of grain size distribution, otherwise referred to as the sorting of a sediment, is a measure of the range of grain sizes and its spread of these sizes around the mean size within a grain population (Leeder, 1991). As it is one of the most commonly used grain size parameters many regard it as an indicator of the maturity of sediment and is associated with the sediment's distance from its provenance and its transport mechanism although some argue it as an inherited property of sediment (Pye, 1994). It is still widely recognised as providing useful information about sedimentary environments and the processes controlling their formation and is an important parameter in facies model approach to modern depositional environments (Reineck & Singh, 1980). The sorting values of the analysed samples do not vary greatly, ranging from moderately to well sorted, with the majority being moderately well sorted, falling between 0.45 and 1.0. Kurtosis, which displays the level of peakedness, is regarded as an indication of the environmental energy at the time of deposition, with low kurtosis representing high energy while high kurtosis represents low energy (Syvitski, 1991). Samples varied between mesokurtic and leptokurtic but the majority were the former. Skewness within the sample set is generally near symmetrical with few being coarsely or finely skewed. TT2 and WT3 are almost entirely symmetrical in distribution with the exception of 2 finely skewed beds. WT4 and WT5 are also generally symmetrical with occasionally coarse skewed beds. A positive skewness indicates a 'tail' of fine grains and coarse grains for negative skewness (Syvitski, 1991). This grain size parameter is best used in analysis in conjunction with sorting and mean grain size.

When two component grain-size variation diagrams are drawn in which one statistical parameter is plotted against another, according to Friedman (1967, 1979), distinct fields become apparent. These fields can correspond with the depositional environment from which the sample was taken.

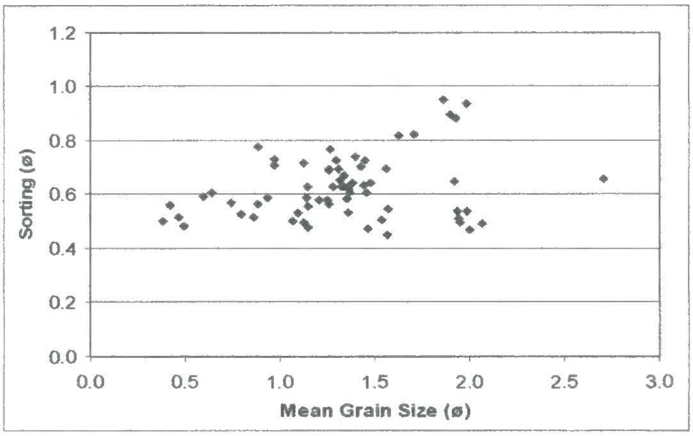


Figure 5.32: Mean grain size versus Sorting of all samples

Mean grain size was plotted against sorting to aid in ascertaining the depositional energy of the environment and transport mechanism of the samples (Figure 5.32). This is useful as sorting is not independent of sediment size or particle shape yet is a helpful parameter in investigating the sorting characteristics of sediment in relation to depositional processes. Figure 5.33 shows a distinct central cluster with two groups on each side and a single far outlier (MG2C). The clusters show that finer sediment is divided into two groups of sorting. According to Syvitski (1991), the groupings all fall within the position of "river" sediment.

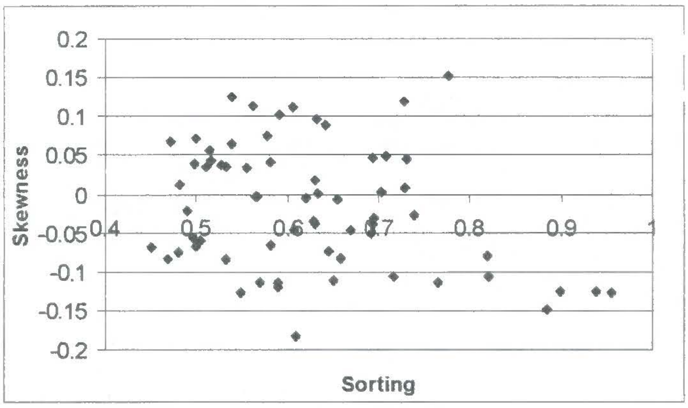


Figure 5.33: Sorting versus Skewness of all samples

Numerous studies use the bi-variate plot of graphical skewness against standard deviation as an aid to offer an additional mechanism for the classification of distinct groupings that can be related to depositional processes or mechanisms (Boggs, 1995). This plot (Figure 5.33) shows a large dispersed group with outliers to the right, which when correlated to the sampling sites are identified as a combination of WT5 site samples and contemporary MG samples. This is useful in loosely identifying two types of depositional environments. According to Leeder (1991), these samples fall close to the boundary or discriminant line separating beach and river sand, probably as a result of the coarseness of some samples, such as DBK1.

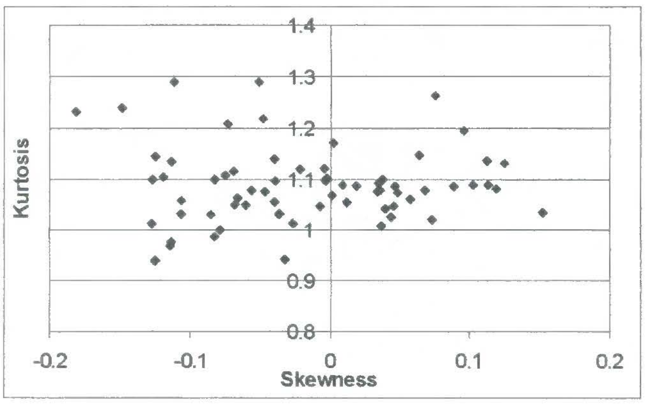


Figure 5.34: Skewness versus Kurtosis of all samples

Another beneficial bi-variate plot is skewness versus kurtosis. In this instance fluvial deposition is expected to result in a coarse tail (Syvitski, 1991). There is a wide dispersion of samples across this plot, as demonstrated in Figure 5.34, with two clusters evident to the left and right with outliers, particularly to the negative skewing (coarse-skewed). This demonstrates the potential that finer sediment is deposited under lower energy environments with the inverse also being true whilst most symmetrical sediment is deposited under more similar energy environments.

5.4.4. Structures within the Sediment

Sedimentary structures reflect environmental conditions that prevailed at or very shortly after the time of deposition (Boggs, 1995). The upper package, as already described, is characterised as loosely and partially consolidated sandy material with horizontal stratification of beds and the notable presence of parallel and low angle wavey laminations. Normal or concordant graded layering dominates the upper package of alluvium. Poorly sorted gravel, pebble and, at the base, cobble lags occur. Occasional trace fossils in the form of root castes are noticeable. Bedding within both the upper and lower packages is laterally continuous and persistent and horizontal. The lower package is largely massive with colour banding indicating minor but not well developed stratification. The massive matrix supported diamicton is thought to be the result of gravity driven sediment flows resulting in a relatively homogeneous matrix containing clasts of a wide size range limited to siliclastic and argillaceous lithology. There are no erosional surfaces or unconformities evident within the deposit to represent any hiatus thus providing evidence of a more continuous process of deposition, although a significant laterally persistent pebble and cobble lag is evident. Towards the base of the deposit there are secondary sedimentary structures in the form of numerous calcium carbonate concretions formed by mineral precipitation and chemical replacement within the semi-consolidated matrix, consisting of calcite and micrite. Mud drapes are also obvious along the exposed surface of the colluvium. There are occurrences of biogenic trace fossils in the form of root penetration structures and castes that are exposed.

According to the Wentworth size classes, lags within the primary site's profiles of the upper package consisted, besides those granules measured as gravel during size analysis, of few pebbles and cobbles, measuring less than 25cm in length. Small pebbles occurred towards the base of WT3 whilst significant deposits of coarse grained material along with both pebbles and cobbles were evident as lags at approximately 2.35m and 2.65m above the contact between the upper and lower packages. Aside from the sampled profiles of the primary sites there are also occurrences of interspersed cobbles extending across the upper package, especially towards the proximal end or west of the exposure. There is also a distinct presence of both pebble and cobble beds within units of the lower package or colluvium. Sporadic boulders occur more frequently in the colluvium but also in the lower unit of the upper package at the centre of the exposure but nearer the surface to the west. Again this is more evident to the west or proximal half of the lateral extent of the exposure.

5.5. SHAPE, COLOUR AND MICROSCOPE ANALYSIS

5.5.1. Morphometric Properties

Grain size and shape are not truly independent variables within sediment, and it is recognised that shape is probably one of the most fundamental elemental properties of any particle and its form is the sum of all morphological properties of a grain (Syvitski, 1991). Results based on the shape and form of grains using microscope analysis aid in providing additional evidence towards their deposition and transportation mechanisms. Natural sediment grains in sedimentary deposits, such as those in this study, vary greatly in shape with roundness being a function of grain composition, grain size, provenance and transport history i.e. transport process and distance of transport whilst form is more directly related to the original shape of the minerals (Pye, 1994; Boggs, 1995). Larger fragments and pebbles' shape are also a function of the shape inherited from the parent rocks, however, abrasion and breakage means that pebbles can be modified to a greater extent than sand-sized particles during transport (Boggs, 1995). Rough estimation of grain shape and form using a binocular microscope is only possible for the sand, and very limited coarse-silt, fractions of a sample as fine-silt and clay particles need to be carefully dispersed and viewed using a scanning electron microscope. There was a relatively large degree of variation between grains of all the samples, as well as high variance within the samples themselves, especially between different minerals and size fractions. The samples were considered more on the basis of overall shape as a majority or average was judged qualitatively regardless of any particular differences due to varying minerals, source, etc. whilst irregularities and unusual characteristics were noted. As such, this phase of analysis was carried out by describing qualitatively in terms of overall shape and form to readily recognisable geometric shapes based on Powers (1953, 1982), for sand-sized fractions, and Zingg (1935), for pebbles and cobbles, visual comparison charts.

Grain roundness varied slightly throughout the samples and within profiles and across river transects, although there was a definite trend in shape and form. Almost all of the grains studied fell within a certain range on the Powers chart. This scope incorporated low to medium roundness i.e. more angular, and a full extent of sphericity, however, predominantly low sphericity. Contemporary channel deposits, such as TT1, DBK1, MG1A and MG3A, etc. exhibited a more diverse collection of shapes, however, mostly subdiscoidal to spherical with occasional subprismoidal elements with an overall higher level of sphericity. Angularity ranged from few angular to sporadic rounded with the majority falling within the realm of subangular and subrounded. There was a definite higher presence of prismoidal and rounded elements within contemporary deposits than the primary site profiles of WT3 and WT4. Samples from cross sections of Moordenaarsgat River pointed towards a wider variety of sphericity and roundness occurring in modern slackwater deposits, again with more noticeable subprismoidal to prismoidal elements but with a range from discoidal to subprismoidal and angular, sporadically very angular, to subrounded. The majority of the fine material falls within the subangular subdiscoidal or spherical region whilst coarser grains are more angular. Channel bars and extra channel deposits appear to also have a wide range with more subrounded, especially in channel bars where prismoidal elements are almost lacking, whilst the latter has more subprismoidal and is majority subangular subdiscoidal with less spherical grains. Grain shape of the contemporary secondary channel profile along the Tra-Tra River shows a very similar and consistent pattern but with slightly higher sphericity and lower angularity.

In profile at the primary sites, WT3 and WT4 showed similar morphometric characteristics. The laminations present throughout the WT3 site and the upper section of WT4 showed a large variety of shapes with a high proportion of angularity, i.e. very angular to angular, although the range extended to subrounded. Sphericity encompassed discoidal to subprismoidal. WT3 included mostly subangular to angular with some sporadic subrounded elements or occasional very angular, with relatively slight sphericity varying from subprismoidal to subdiscoidal with sparse discoidal occurrences. Higher sphericity does prevail throughout. This does vary with position within the profile, as at greater depth the grains appear less angular or discoidal but rather more spherical and subrounded with few prismoidal elements, while some coarser grains towards the centre of the exposed profile (1m and lower) are more spherical and subangular in nature. Near surface samples from WT4 also show a full range of shapes but are mostly between angular to subrounded and subdiscoidal to subprismoidal with the majority being angular spherical, however, sporadic very angular, discoidal and prismoidal shapes occurred. One metre and below yielded only occasional very angular traits with most being subrounded to angular and sphericity being low; subprismoidal, subdiscoidal and discoidal, with spherical shapes only sporadically occurring in rarer minerals. This location, however, is not as angular or discoidal in nature as below, towards 3m, where, particularly the quartz grains, had high angularity and low sphericity (very angular to angular and subdiscoidal to discoidal and sub prismoidal). Overall shapes were more subangular and sporadically spherical.

WT5 exhibited a wider range of shapes than WT3 and WT4, although in general it appeared more rounded with rarely occurring angular grains; most fell between subangular to subrounded and subdiscoidal, spherical and subprismoidal. Finer grains showed higher sphericity and roundness in comparison to larger more angular and discoidal natured grains. With increasing depth grains appeared to become less discoidal with fewer angular elements.

Sandstone parent material results in silica- and feldspar-rich sand sediments (predominantly quartz, plagioclase and orthoclase) with intermittent shale fragments along with other plentiful minerals present within both sandstone and shale, namely mica (muscovite), such as the characterised micaceous sandstone and siltstone Wuppertal formation that occurs along the Tra-Tra Valley (SACS, 1994). Overall, quartz grains, which have no cleavage (Schumann, 1998), exhibit low sphericity, mostly discoidal to subdiscoidal, with a lesser degree of roundness, normally being subangular to angular or even very angular, in contrast to feldspar grains that are definably more rounded and spherical, being mostly subrounded to subangular and spherical to subdiscoidal or even subprismoidal. In most cases, although exceptions occurred in places, the finer sand fraction appeared to have a higher sphericity when compared to larger or coarser grains, whilst they exhibit mostly subangular traits. Quartz, which is a major component of all samples taken within the study area (see section 5.5) and was therefore commonly examined throughout for shape analysis, has a superior hardness and durability (7 on the Moh scale of hardness) and hence, is less significantly modified during transport when small grains as they are rounded less readily, although slight changes can occur in early stages of transport (Boggs, 1995). In contrast, feldspar, another frequently occurring mineral within samples, is weakly durable and more easily rounded. Shale is considered as relatively soft and its pebbles are more readily rounded than quartzite pebbles. As a rule of thumb, the larger the grains or pebbles, the more readily they are rounded, especially within fluvial environments. Other minerals, such as biotite and mica, due to being softer and exhibiting cleavage, as well as bedding orientated shale fragments, have more prismoidal and rounded properties. This is obvious throughout all the samples as fine mica and biotite grains appear overall as prismoidal and subangular or subrounded with the latter's higher sphericity. It is important to bear in mind the cleavage and breakage patterns of various minerals as well as to consider that the frictional behaviour exhibited by grains is partly due to microscopic irregularities on grain surfaces in addition to the obvious macroscopic edges or corners (Leeder, 1991).

According to the Zingg classification of pebbles, clasts from the primary sites WT3, WT4 and WT5, along with cobble and boulder sized fragments are mostly subangular prolate or bladed, very occasionally subrounded equant. The cobbles and boulders present along contemporary channels throughout the three valleys show a range of shapes but are more consistently subrounded to subangular prolate and bladed. Again there are occurrences of more compact or equant forms, particularly in primary channels, with even some oblate forms being present towards the base of channels.

5.5.2. Colour

Variations in colour of sediment were used as key indicators in profile and across river transects, particularly in sampling. Again, colour is closely associated with the parent material of the sediment, mineral components as well as the environmental setting and processes at work and diagenesis, such as leaching, organic input, cementation, etc. Colour changes were mostly subtle within profile and as expected, no dramatic differences in colour were noted, and rather a case of short gradual transitions to lighter or darker shades of the same hue. A more marked modification in conditions within the environment could be surmised by abrupt colour changes to a very different hue, however, as the conditions appear to have slowly fluctuated, the colours remain relatively closely related. Munsell colour notation demonstrates the mostly homogeneous colours of horizons within a profile and between the two alluvium sites, as well as the majority of contemporary environment samples. All contemporary fluvial sites (from Moordenaarsgat and Tra-Tra rivers) are categorised as belonging to the same hue, 10YR, and vary from pale brown to yellowish browns. Modern channel sediment is pale or light brown with some yellowish/dull orange elements whilst contemporary slackwater deposits appear more yellowish brown. In comparison, Dassieboskloof channel sediment is a lighter brownish grey. The shallow secondary channel profile along the Tra-Tra River clearly shows the transition with depth from more yellowish brown sediment to paler brown followed by grey (pale). This is a helpful indicator towards the profile transitions within the WT3 and WT4 sites, although sediment input along the Tra-Tra River includes a higher shale proportion, hence browner (and more yellow elements). Although not uniform, the entire of the upper package at the primary site (WT3 and 4) exhibits sediments with variations of grey. WT3 shows an overall lighter colour to that of WT4, although in both packages of sediment there appears to be a transition of colour bands in that there is a division between two distinct groupings of colour in the lower and upper sections of the profiles. The upper section of WT3 is a browner/yellowish grey in comparison to a lighter grey with interspersed darker grey horizons towards the lower units of the profile. WT4 also exhibits a similar pattern with darker and occasionally brownish grey beds towards the profile surface followed by a transition to grey, punctuated by a darker unit, and then especially lighter and even whitish grey beds at the base of the package. It is also notable that the clasts, i.e. pebbles and cobbles, present within these lower whiter beds are white and light grey in colour and appear bleached. Laminations in both the upper and lower sites are particularly obvious as they stand out as thinner dark grey bands interspersed against mostly lighter sediment (Figure 5.35). The laminations of WT3 are darker in comparison to WT4, being darker grey to black versus darker brown and grey respectively.

Figure 5.35: Laminations present in WT3 (left) and WT4 (right) (upper and lower profiles of upper package respectively)

The lower package of the primary site, WT5, described as colluvium, except for the capping sediment that comprises partially of alluvium, sheetwash and organic material, is a distinctly different colour to that of the upper package and the sediment resting immediately above it. Although there is less variation in colour, colour banding is apparent within the colluvium giving the appearance of beds or horizons within the otherwise more uniformly coloured units. The colluvium is characterized generally as orange brown. As horizons were only described up until the basal point of exposure at which the contemporary flood channel dissected or scarped the face or bedrock outcropped, colour was not noted below the surface. The darker redder beds are positioned closer towards the surface of the package with lighter horizons both above and below. The surface unit is grey in contrast to the light yellowish brown of the bed immediately below. An orange reddish yellow bed occurs below this, at approximately 1m, which is particularly apparent when viewing the profile; below this colour ranges around yellowish brown. A lighter reddish bed does occur again at approximately 2.5m; however it is not as discernable as the upper horizon. Clastic material is mostly dark brown shale and lighter sandstone fragments.

5.5.3. Microscope Observations

A light microscope was used briefly to ascertain the general nature of the surface texture of the grains. The degree of microrelief, or surface roughness or smoothness, can be used under favourable conditions along with specific surface textural features to aid in providing clues to transport and postdepositional alteration (Pye, 1994). It has been found that a number of surface patterns, fractures or impact markings are diagnostic of particular environments (Leeder, 1991).

The susceptibility of surface texture to change during sediment transport and deposition appears to be higher than that of sphericity and roundness, as well as it being more likely to record the last cycle of sediment transport or the last depositional environment, however, the usefulness of such surface features and markings in environmental analysis can be misleading as similar types are produced in different environments or retained from previous conditions (Boggs, 1995). Interpretation of palaeoenvironments using surface texture is complicated due to the fact that surface microtextures can be changed during diagenesis by addition of cementing overgrowths or by chemical etching and solution.

As the magnification was insufficient to realise a high level of detail for the surface texture throughout the range of grains across all samples, it was necessary to make generalisations concerning the overall patterns, microfeatures and relief. Two diagnostic textural features that can be identified using a binocular microscope are frosting and polish (Pye, 1994). As there is a large component of quartz grains, it was important not to confuse the vitreous or glass-like lustre of light reflected by quartz minerals with a polished grain appearance. All contemporary and alluvial samples contained grains that were a majority of frosted, while the colluvial samples appeared to contain a more noticeable combination of both frosted and polished grains. This is important to note in interpretation of grain origin, transport and postdepositional processes. Another notable feature was that of relatively regular pitting on the surface of coarser quartz and feldspar grains. These pits appeared as almost circular depressions sporadically occurring across the surface, however, on occasions in clusters. A small number of conchoidal fractures, typical of quartz, were also evident, along with some ridges.

Microscope analysis revealed the apparent existence of noticeable grain coatings, particularly at higher magnification and on those samples viewed subsequent to furnace burning. In some samples, especially those of the upper package, these coatings are interpreted as clay residue and coatings, whereas in most samples, especially the contemporary channel and colluvium, the coatings are an orange to red hue and oxidised upon burning leading to the conclusion that this was due to the presence of iron compounds. This is attributed to the nominal presence of biotite within the argillaceous (shales) and to a lesser degree, the arenaceous (sandstones and greywacke) parent material (sourced from the surrounding geology) which accounts for the sporadic appearance of iron coatings on few of the particles, where the iron components of the minerals leach out. Biotite is much less stable than muscovite due to its ferrous iron content (White, 1997). Argillaceous rocks may also contain small amounts of iron pyrites adding to the iron content of the fine fraction of samples, upon which they were more obvious, and hence, leading to more intense iron coatings, which become especially obvious subsequent to furnace burning, when the iron becomes oxidised to a dark orange or red. The occurrence of thermal oxidation was noted for each sample and it was found that it did occur in all contemporary fluvial samples i.e. those from both the channels, side channels, channel bars and slackwater deposits of the Tra-Tra and Moordenaarsgat rivers and the Dassieboskloof River. The upper samples taken in profile from the secondary channel of the Tra-Tra River were also oxidised, whereas those from 107cm and 137cm did not. The most obviously thermally oxidised set of samples was that belonging to WT5, the colluvium. All bar the surficial sediment oxidised to deep reds and oranges. Microfeatures, such as the pitting on the surface of some grains, particularly quartz and feldspar, were highlighted where such coatings accreted. In contrast to those samples that contained grain coatings associated with iron compounds, some samples subsequent to furnace burning became bleached (colour markedly reduced to whitish or very light grey, pink or yellow). This was most evident for the alluvium (WT3 and WT4) with only occasional lenses of sediment sampled turning slightly pale orange after burning.

5.5.4. Nodules

Nodules are described for the purpose of this study as small (less than 256mm), spherical, oval or similarly rounded concretions or aggregates of sediment including from very fine up to very coarse sediment, sometimes incorporating pebble and very rarely cobble sized fragments. These aggregates or nodules could be either permanent or non-permanent, depending on their nature and material of cementation, which is usually an insoluble compound that irreversibly cements or encloses soil particles, and position within the profile, i.e. formed at or near the surface or susceptible to mechanical disturbance (White, 1997). In this study, nodules are interpreted as an indication of embryonic soil formation, pedogenesis and the beginnings of partial cementation and ped formation.

Measurement of nodules was mostly a qualitative exercise in order to add further descriptive information and an additional dimension regarding the overall nature and microfeatures of the sampled sediment from the different locations. The nodules were considered as those aggregates obvious to the naked eye that occurred within a bulk sample of relatively undisturbed sediment taken *in situ*. The sample was carefully and thoroughly examined for any presence of so-called nodules. It was important to note that these aggregates were naturally occurring throughout the sediment subsequent to oven-drying at 80°C in order to assure they were not the result of any residual moisture or pore water. Measurement also needs to take into account any possible mechanical disturbance to the sample, which may have broken larger nodules into the smaller ones. Nodules are widespread throughout the samples and are only absent in very few, namely MG1B, MG4A, WT3-10, and WT4-300 to 355. In all samples except the colluvium (WT5), the nature of the nodules was found to be crumbly and they disintegrated under pressure between the fingertips. As the colluvial samples consisted originally of consolidated and, in places, partially cemented material, it was difficult to ascertain the true nature or presence of any nodules as samples were comprised of broken off chunks of sediment and had to be physically fragmented for analysis, hence it was judged to assume consolidation instead of nodules occurred except in WT5-240 where they were particularly evident. Nodules were generally less frequent and smaller in contemporary fluvial environment samples taken from the 3 valleys, with the exception of MG2A and MG2C and between 47 and 87cm in profile at TT2 where they were numerous and larger. Within WT3 profile there was a lower frequency and again overall smaller size classing, however, the base of the profile provided few that were slightly coarser. Similarly WT4 stayed within a range of size but with coarser less frequent nodules, which became less with increasing depth. Overall, as evident on the graph below, nodule size varied but was maintained between 1mm-10mm, and a more cyclical pattern is noticeable within the alluvial profiles (TT2, WT3 and WT4) at similar depths. The laminations did also contain small occasional nodules.

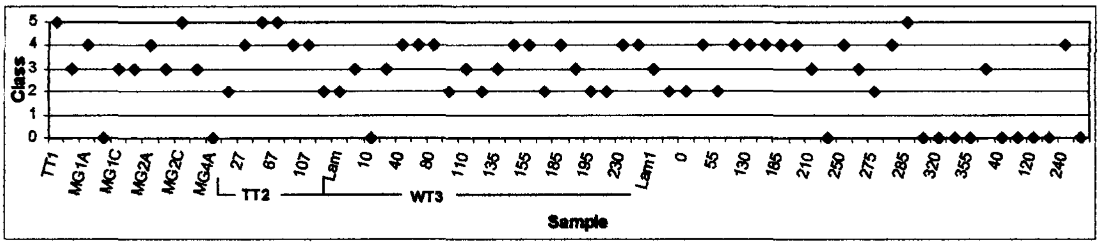


Figure 5.36: Nodule size according to class for all samples

Although not included in Figure 5.36, it is important to note that whitish cemented sediment nodules found within the consolidated material of the colluvium were tested with HCl acid at varying solution concentrations and discovered to react instantaneously without any agitation leading to the conclusion of the presence of calcium carbonate or calcite. This occurred occasionally below 40cm and most noticeably towards the base at 240cm and 300cm. Using the definition of calcrete as a near surface, terrestrial accumulation of predominantly calcium carbonate, which occurs in a variety of forms, and based on typical calcrete formation principles (Watson & Nash, 1997; Shaw & Goudie, 2004), the calcrete present in the colluvial profile ranges from pedogenic micrite towards the surface to larger sparry calcite towards the base of the exposure.

5.5.5. Charcoal and Organics

Charcoal occurred widely throughout the samples, especially within the middle regions of the WT3 and WT4 profiles. It is assumed to be the product of biomass burning within the region. Its presence to the naked eye was noted (although microscope observations yielded further presence but in very fine small quantities throughout most samples except the laminations and colluvium); however, its poor preservation and fine nature of fragments meant measuring its relative abundance was difficult so a scale of qualitative frequency was established. Interestingly, charcoal was absent from contemporary slackwater deposits along the Moordenaarsgat River valley although it did occur in both sampled channel bars and some of the contemporary active and secondary channels. It was also noted throughout the secondary channel profile and primary channel of the Tra-Tra River. Charcoal was occasionally present within WT3, notably at the surface and below 1m. It was conspicuously lacking in the browner beds towards the surface and at the base along with a darker grey lens at 185cm. Charcoal was consistently found throughout WT4 bar near the surface and at 130 and 185cm.

Fragment size varied but was generally very small and fragile. Sporadic larger chunks occurred with the largest fragments found being up to 5mm in length in WT4 profile between 2 and 3m.

Organics in the form of rootlets, root and stem fragments, leaves and seeds were also recorded. Seeds, fine fibrous roots and woody material persisted through the contemporary samples of the Moordenaarsgat and Tra-Tra River valleys. Fine root fragments only extended to a depth of 25cm in WT3 after which organic matter was absent until some larger roots were discovered at 230cm. Roots, leaves, seeds, etc. extended to a greater depth in WT4, up to 2m, and again reappeared towards the base as a few seeds and wood fragments. Organics were mostly lacking in WT5 except for the surface material and at the base of the exposure where sporadic roots encroached.

5.6. GEOCHEMISTRY

5.6.1. Organic and Inorganic Carbon Content

Fluvial bed sediments represent an important sink and source for a variety of organic and inorganic compounds (Sutherland, 1998). The organic and inorganic carbon content was established for all 66 samples within the study locality. The overall carbon contents were relatively low (all < 4% for organic carbon, with the exception of MG2C, and < 2.5% for inorganic carbon) and displayed a high level of homogeneity. It is interesting to note that there is a distinct lack of A horizons in the profiles, hence organic carbon content is not highest for the sample taken immediately below the surface (0cm).

Organic carbon content within the contemporary channel and extra-channel deposits peaks at 10.31% for MG2C, a side channel, and drops to its minimum at 0.34% and 0.35% for DBK1 and TT1 respectively. Organic carbon content for contemporary deposits is generally within line of the other samples sites, although the presence of the high value for MG2C does skew the overall impression of the graph (below). The average range (excepting MG2C) is mostly below 2%. Active river channels within the valleys appear to have the lowest values (TT1, DBK1, MG1B, MG3A and MG4A) as the organic matter is permanently being flushed out or removed by more rapidly and perennially flowing water. Channel bars and slackwater deposits appear to contain a small accumulation of organic carbon as material has the opportunity to settle and be deposited as well as the presence of a riparian vegetation and organic matter litter that becomes incorporated with the sediment.

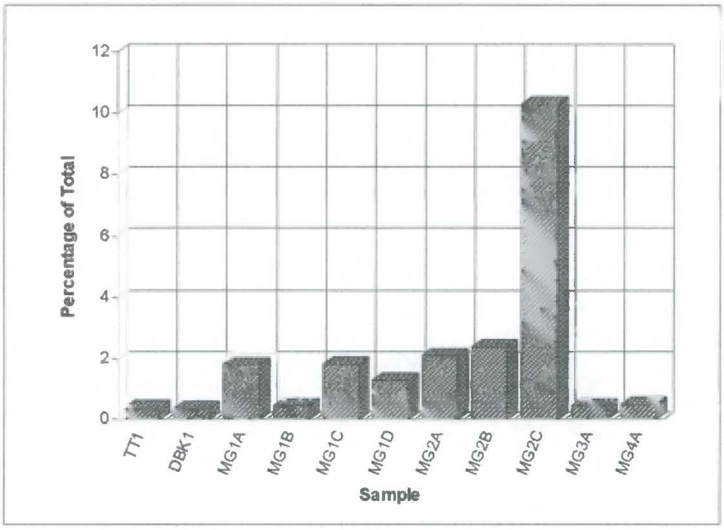


Figure 5.37: Organic carbon content of contemporary channel deposits of all three rivers

Figure 5.38 shows that in terms of inorganic carbon content, MG2C again represents the maximum value, at 0.96%. On this occasion slackwater deposits and channel bars also reflect high inorganic carbon content, especially MG2A. Again contemporary active channels have the lowest values as the minerals that represent inorganic carbon content are either winnowed out or only present in very small quantities.

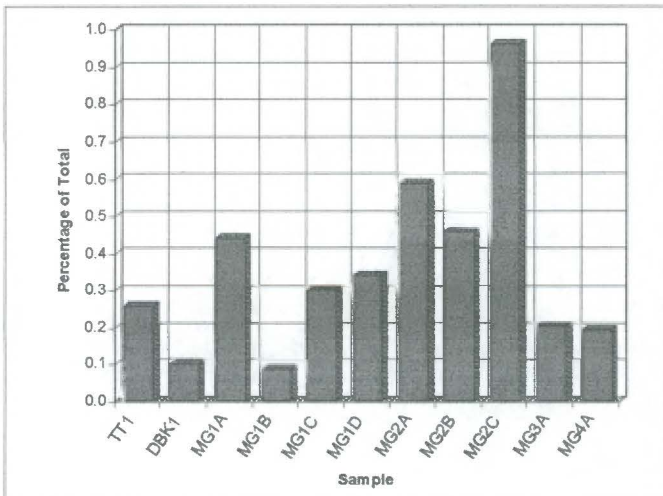


Figure 5.38: Inorganic carbon content of contemporary channel deposits

With the exception of the sample taken at depth 47cm, TT2 exhibits a decreasing trend in organic carbon content with increasing depth. TT2-47 has the highest value at 2.38%, however, as of the surface sample at 2.20%, organics gradually decreases to 0.19% at 137cm. The most marked change in content is between 47 and 67cm. As this site was an exposed channel bank profile, it would follow that organic matter accumulated at the surface with some surficial burial and depending on vegetation roots and then continued decrease with depth. Alternatively fluctuating flow levels within the adjacent channel could have gradually removed organic matter, although, conversely this would also lead to some accumulation when flow retreated or slowed.

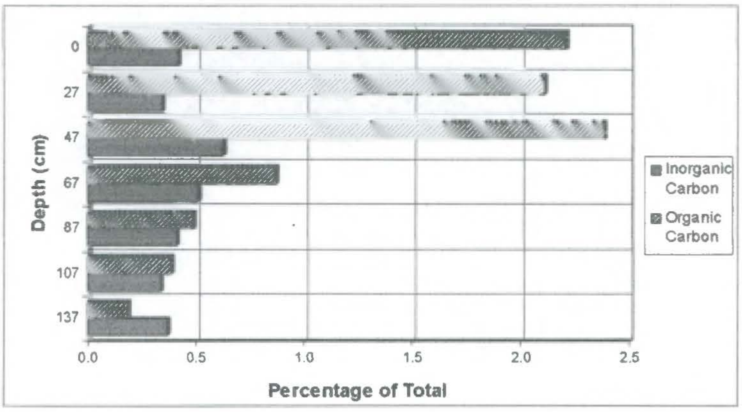


Figure 5.39: Organic carbon content of TT2 sand bank profile on the Tra-Tra River

Inorganic carbon content fluctuates more notably throughout the profile. The values are significantly smaller than organic carbon content at the surface and just below, however, at the base, inorganic carbon exceeds organic carbon. There is also a decreasing trend towards the middle of the profile with an apparent accumulation present towards the base at 137cm. The inorganic carbon content ranges from 0.62 to 0.34% with its peak coinciding with that of the organic carbon, at 47cm.

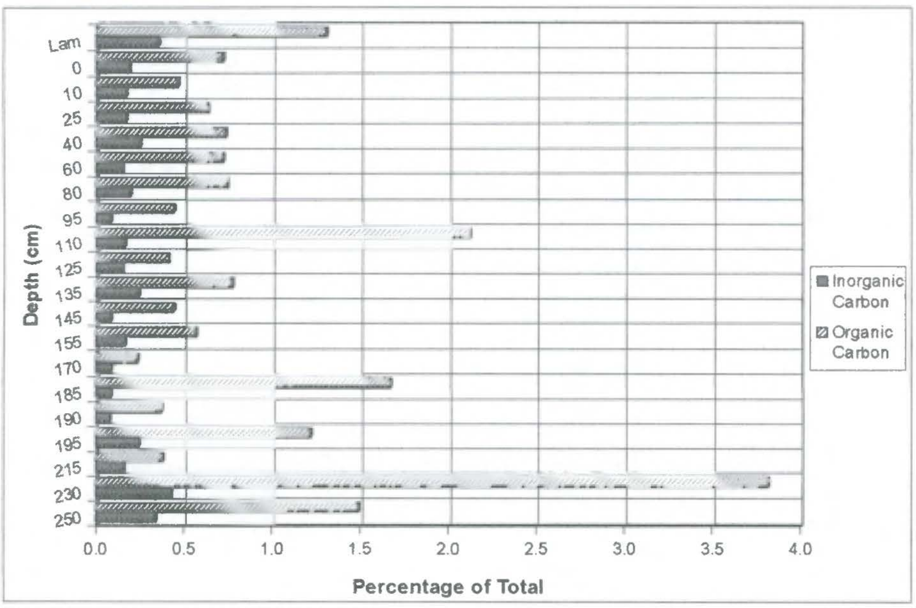


Figure 5.40: Carbon content of the upper profile (WT3) of the upper package at the primary site

Figure 5.40 clearly displays that both organic and inorganic carbon content fluctuate greatly throughout the WT3 profile. Organic content is consistently higher than inorganic content although they do appear to behave in tandem with few exceptions. This could be coupled with sediment size to be indicative cycles of deposition as one notices groupings of alternate increases followed by subsequent decreases, such as 135cm to 145cm followed by 155 and 170cm, and so on. The organic carbon is at its maximum at 230cm (3.81%) and minimum at 170cm (0.23%). Inorganic carbon content corresponds with its peak also at 230cm with 0.43% and its lowest value at 190cm (0.08%). The lamination has relatively high organic and inorganic carbon content in comparison to the majority of the WT3 sediment at 1.30 and 0.35% respectively. The trend in values shows more consistent values for both towards the surface, however, 110cm is an exception with more organic carbon. The basal percentages show an increase in organic and inorganic content. Both variables content is relatively low within the upper horizons of the sandy alluvium package.

The laminations present within the WT4 profile display the highest organic and inorganic carbon content present throughout the depth of the profile. This is in keeping with the lamination at the WT3 profile site that also demonstrated high values for both. They have significantly higher inorganic contents at 0.78 and 0.60% and organic contents at 1.91 and 1.76% for lamination 2 and 1 respectively. The rest of the profile exhibits an overall decreasing trend in inorganic carbon content with increasing depth and percentages range from higher values towards the surface to lower ones at the base (0.31-0.07%). Organic carbon content fluctuates somewhat more readily and behaves less consistently with inorganic carbon. Percentages increase towards the middle of the profile with a peak at 265cm (1.81%) and decreases towards the base where its lowest value is 0.16% at 340 and 355cm. Overall organic content percentages are very low indicating very little organic matter is present within the profile beyond surficial organics and besides some charcoal fragments at depth.

Generally the organic and inorganic carbon contents of WT4 are marginally lower although mostly in line with those of the WT3 profile (with the exception of WT3-230). An important trend evident within WT4, similarly to WT3, is the presence of so-called deposition cycles within the lower sandy alluvium package. One sees a sequence of gradually increasing organic carbon content followed by a rapid decrease and subsequent increase again. This is particularly obvious in the upper half of the profile through two sequences until a peak at 205cm. In the lower half the sequence appears to be in reverse with a decreasing trend with depth. This is apparent in a series of three cycles until the base (or two cycles if one considers between 250 and 355cm as one cycle in itself). Inorganic carbon content fluctuates around more stable percentages and displays only subtle trends within that generally follow those of the organic carbon.

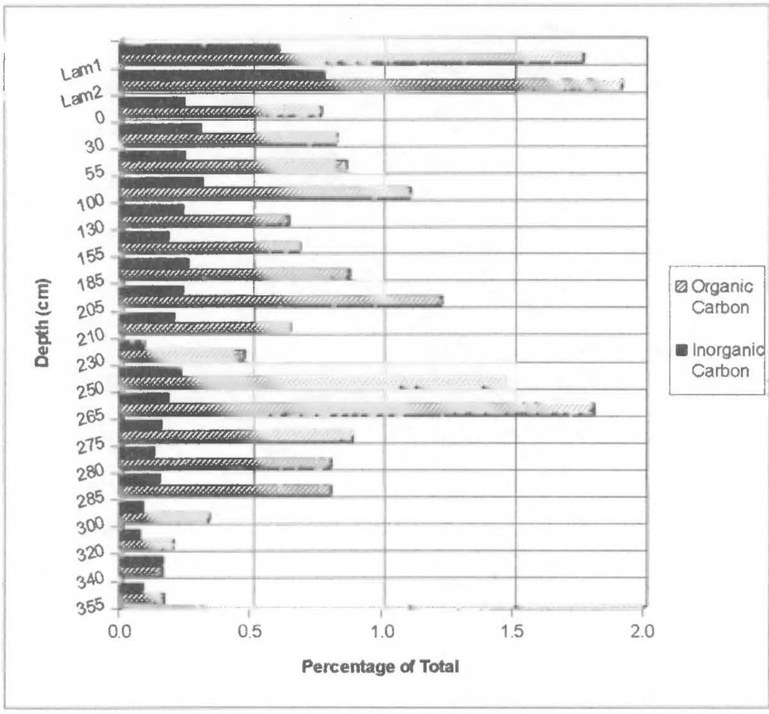


Figure 5.41: Carbon content of WT4 (lower profile of upper package of the primary exposure)

Organic and inorganic carbon contents of the colluvial package are relatively higher than those of the sandier alluvial packages above. Organic carbon percentages range from 1.24% (40cm) to 3.52% (90cm) whilst inorganic carbon values are between 0.33% (0cm) and 2.60% (190cm); the surface sample was on the contact between the two packages and was more alluvial in nature, so it is more appropriate to say the minimum value is 0.63% at 40cm for the colluvium. In general the organic and inorganic carbon content acts in tandem, as one increases, so does the other. The difference between the percentages of the total weight of each sample for the two types of carbon are also less i.e. the values are more similar with no large disparities as shown in the upper package. The trend within the profile resembles that of the alluvium in that clear sequences are apparent. Within this profile there are a series of successive coupled increases within strata, as shown by three cycles, namely 40 and 90cm, 120 and 190cm, and 240 and 300cm. The high organic carbon content at 0cm could be attributed to its surficial position and proximity to vegetation and organic matter and litter.

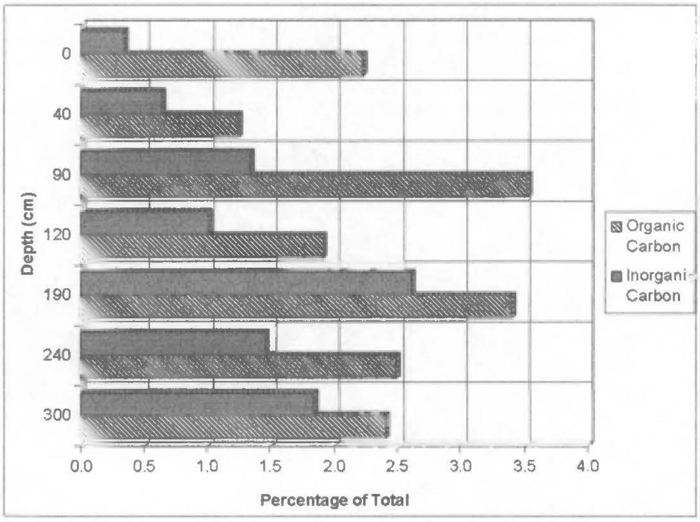


Figure 5.42: Carbon content of the lower package (WT5) profile

5.6.2. Electrical Conductivity

Electrical conductivity (EC) is the ability of a sediment or soil to conduct an electric current. It is used as an index of soil salinity (Briggs *et al.*, 1995). Selected samples from profiles were measured for their EC using saturated pastes of a 1:1 ratio. The values are given in microsiemens per centimetre ($\mu S\ cm^{-1}$). High concentrations or large EC values are considered restrictive to productivity whereas low values indicate excessive leaching. As a reference value, the EC of tapwater measured on 31/10/2006 was 161 $\mu S\ cm^{-1}$ and compared relative to sample results. Overall, the WT3 and WT4 profiles had the same conductivity for the surface horizon (120 $\mu S\ cm^{-1}$) and similar very low conductivity values, while WT4 conductivity had greater fluctuations with some extremely low conductivity values. This indicates that the sandy alluvial upper package is extensively leached. In contrast WT5 exhibits very high conductivities, with the exception of the surface sample, which indicates less leaching.

WT3 has very low conductivities ranging from 122 to 36 $\mu S\ cm^{-1}$. There is a general decreasing trend from the surface towards 170cm, with the exception at 60cm (122 $\mu S\ cm^{-1}$). At the minimum values, between 170 and 215cm, conductivity fluctuates and then drastically increases to 121 $\mu S\ cm^{-1}$ at the next point, 250cm.

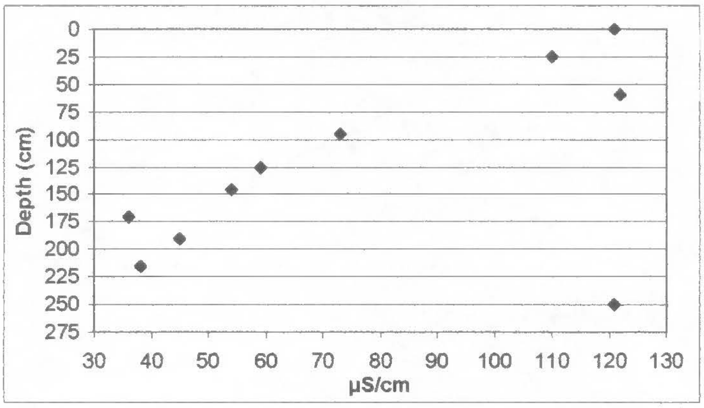


Figure 5.43: Electrical conductivity of WT3 profile (upper unit of upper package of exposure)

WT4 has a wider range of conductivity but does display extremely low values towards the base of the profile (above the contact with the colluvial package). The highest conductivity is present at 205cm (289 µS cm^{-1}), which is followed by a dramatic jump to a significantly lower conductivity value. There are slight fluctuations below 230cm with a decrease to the lowest values at the basal area of the profile (37-27 µS cm^{-1} at 300-355cm respectively). The general trend, excluding the exception at 205cm, is of decreasing conductivity with increasing depth.

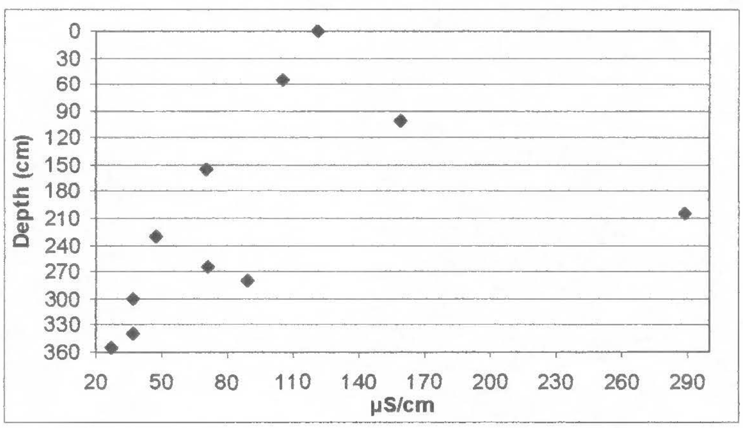


Figure 5.44: Electrical conductivity of WT4 (lower profile of upper package)

The WT5 profile, as seen in Figure 5.45, displays an exceptionally wide range of conductivity, from 243 µS cm^{-1} at 40cm to 3430 µS cm^{-1} at 240cm. There are substantial jumps in values, particularly between the basal two samples, from 3430 to 783 µS cm^{-1} at 300cm. Although there is a slight decline from the surface to 40cm (285-243 µS cm^{-1}), this is followed by a stepwise increasing trend with increasing depth that stabilises at 190-240cm followed by a rapid decrease towards the base.

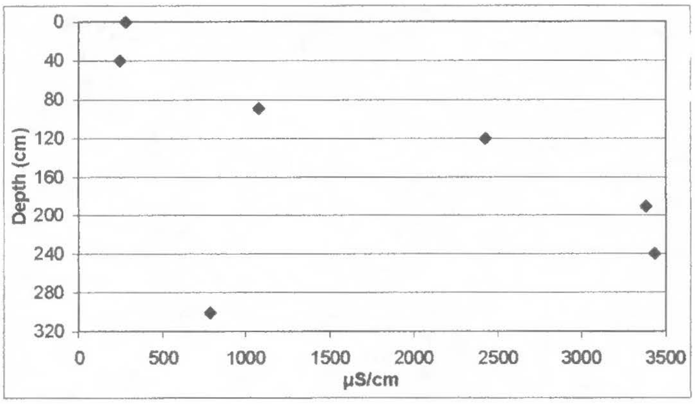


Figure 5.45: Electrical conductivity of WT5 profile (lower package of exposure)

EC values were plotted against the silt and clay fractions and the fines and sand components of WT5 to establish if there was any kind of relationship evident between the variables that would help determine the considerably higher conductivities of the profile (Figure 5.46). A primary consideration is the fact that WT5 comprises a more significant fines fraction than either WT3 or WT4. There does appear to be a subtle coincidence between increasing conductivity with an increasing proportion of fines. Higher sand percentages are synonymous with lower conductivity, as in the case of the upper two and basal samples.

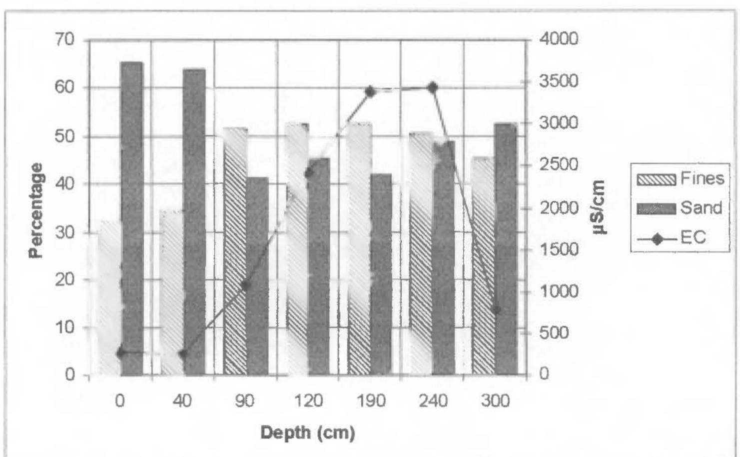


Figure 5.46: Comparative EC, fines and sand fraction of the lower package of the exposure (WT5)

The relationship is less clear when considering clay and silt, demonstrated in Figure 5.47. It appears that higher silt/lower clay percentages are associated with lower EC values (in the case of 0 and 40cm) in comparison to more substantial clay fractions being linked to higher conductivity. The decrease in clay fraction coincides with the decrease in conductivity. This presumes that increased fine fractions, specifically clay, result in the increased presence of ions and poorer leaching.

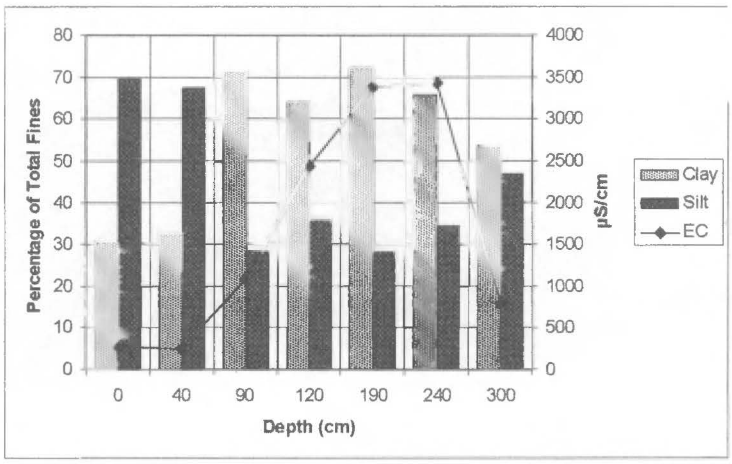


Figure 5.47: EC compared to silt and clay fractions of the lower package (WT5)

5.7. X-RAY DIFFRACTION

The relative abundance or proportional percentages of a suite of clay minerals and total minerals from ten selected samples within the study area are represented below in Figures 5.48 and 5.49. Two samples were selected from contemporary fluvial environments, one from a slackwater deposit on Moordenaarsgat River, the other from a depth of 47cm in profile of a side channel bank of the Tra-Tra River. The eight remaining samples were strategically selected from within the three individual profiles of the primary site to represent any differences in mineralogy in relation to depth and relative clay content within the alluvium and the colluvium (taken towards the surface, centre and base where possible). Clay minerals are largely layer-lattice or phyllosilicate in nature comprising the bulk of oxygen as silicate (SiO_4^{-4}) ionic groups in combination with metallic cations (White, 1997).

The most common minerals occurring in all samples at varying abundance are quartz and feldspar, predictable in this region through the parent material of the sediment being largely Table Mountain Group sandstone, which is high in these mineral quantities and are hard and less readily eroded or weathered. The siliciclastic sediments of the Cape's rocks consist of major stable and less stable minerals, some accessory minerals and some chemical cements. Quartz is a stable mineral with the greatest resistance to chemical decomposition, making up approximately 65% of average sandstone, 30% of average shale; less stable major minerals include feldspars (mostly plagioclase and orthoclase) comprising 10-15% of sandstone and 5% of shale; and clay minerals and fine micas, including kaolinite group, smectite group (montmorillonite a principal variety), chlorite group and fine micas are principally muscovite (and biotite) making up >60% of minerals in shales (Boggs, 1995). Abundant chemical cements are carbonate minerals, predominantly calcite. All samples exhibit a very high proportion of quartz, with the exception of WT3-25 with 61%, in comparison to WT3-110 and 230 with 97% and 98% respectively, and WT5-190 with 80%. All others contain 90-95% quartz mineral. The second most abundant mineral overall, feldspar, is most evident in WT3-25, consisting of 39%, by far the highest proportion. Due to the micaceous nature of some of the sandstone, siltstone and mudstone formations present within the region, namely the Wuppertal formation along with the Bokkeveld and Witteberg Group formations (SACS, 1994), mica is also an obvious mineral with consistent occurrences. Smectite (TT2-47), pyrophyllite (WT4-250) and calcite (WT5-190) are readily distinguishable in only singular samples, whereas kaolinite occurs in three.

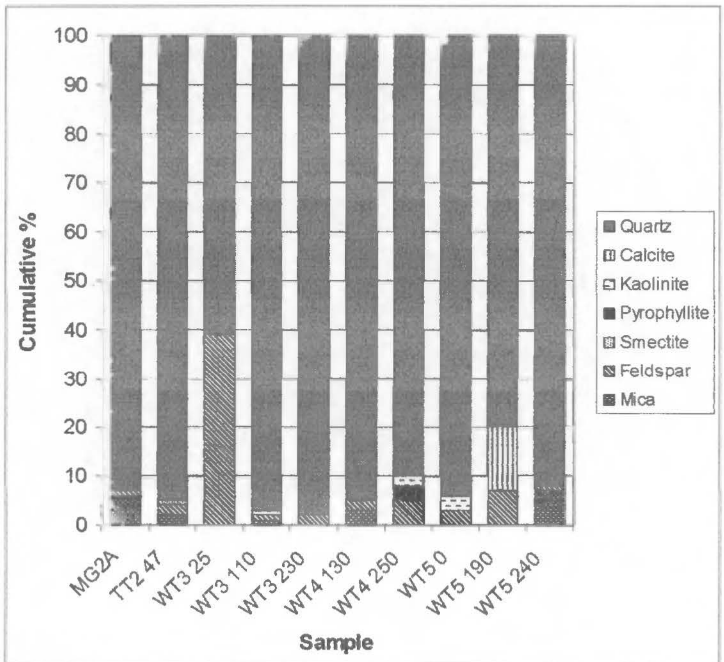


Figure 5.48: Mineralogy total analysis by XRD for selected samples

Clay mineralogy carried out on the clay fraction all ten samples exposed more significant detail regarding the dominant clay minerals present within varying localities and depths. Again, quartz is the major mineral in all samples except WT5-190, where instead mica is the highest proportion at 55% with the entire difference being represented by chlorite. The quartz values were, however, significantly less (54-69%) than previously and only ranged between 33 and 38% in the WT5 samples. The presence of mica is also considerably increased with higher abundances present, particularly in MG2A. Clay fraction XRD analysis showed a more substantial appearance of kaolinite across all samples. Values above 10% show TT2-47, WT3-25, WT4-130, WT5-0 and WT5-240 to have noteworthy components of kaolinite. There is also a strong presence of pyrophyllite across all samples, bar three, the contemporary locations and WT5-190. Smectite occurs in relatively high abundance in the samples in which it appears; WT3-110, WT3-230, WT5-240 and to a lesser degree WT4-130. Chlorite occurs only in two samples, namely WT5-0 and WT5-190, in a relatively high proportion at 28% and 45% respectively. Vermiculite, the fifth clay mineral identified, appears in three of the analysed samples, TT2-47 (15%), WT3-25 (12%) and WT4-250 (7%).

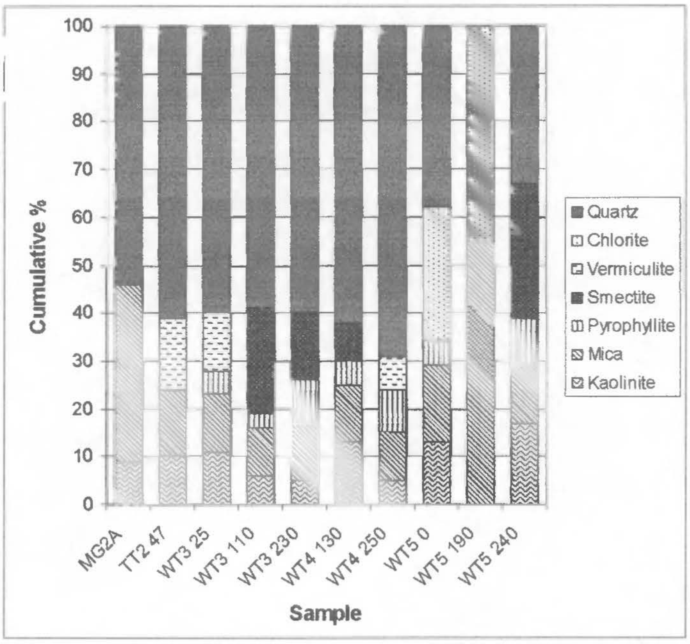


Figure 5.49: Mineralogy clay analysis by XRD for selected samples

There appears to be little spatial patterning of either the total (bulk) or clay minerals throughout the analysed samples. The most obviously notable feature is the distinct and relatively high proportional presence of chlorite in two of the three colluvial samples, being the upper and not basal samples, revealed by XRD analysis of the clay fraction. Quartz, mica and feldspar occur throughout the samples, as well as kaolinite (with the exception of WT5-190). Pyrophyllite is well distributed throughout the alluvium (WT3 and WT4) and upper and basal colluvium. A clear outlier in comparison to the other samples, is WT5-190, it contains almost only quartz and mica elements with a substantial proportion of chlorite. Interestingly, it is also the only sample to contain calcite, however, this may be a by product of micritic calcrete coatings evident within the analysed bulk sediment.

Parent materials influence the permeability, base content, hardness, grain size and mineralogy of their weathering products i.e. the sediment (Briggs *et al.*, 1998). Smectite belongs to the montmorillonite group of clay minerals. Vermiculite is related both to the montmorillonite and to the chlorite clay groups. It appears as a constituent of clay in certain soils and seems to form mainly as a result of the alteration of biotite flakes or, more rarely, chlorites, hornblende, etc. Pyrophyllite occurs in low- and medium-grade metamorphosed aluminium-rich sediments (Whitten & Brooks, 1972). The argillaceous component (shales) of the parent material is primarily clays and silts, from which the colluvium is primarily derived, in contrast to the predominantly sandy alluvium that is fundamentally mostly quartz with feldspathic and mica elements along with only a very small clay fraction consisting largely of kaolinite, smectite, vermiculite and pyrophyllite. The mica is preferentially weathered out of the sandstone and shales of the Cederberg range upstream, as well as a small amount from the surrounding valleys. Kaolinite is a 1:1 dioctahedral phyllosilicate whereas mica, smectite and chlorite are 2:1 structurally. Vermiculite is a 2:1 trioctahedral mineral (White, 1997). The relative abundance and widespread nature of kaolinite can be attributed to the fact that it is a pseudomorph after feldspar and other aluminosilicates and the result of hydrothermal alteration and weathering and occurs on account of feldspar in the surrounding geology where it is plentiful. Pyrophyllite is another aluminosilicate that exists alongside kaolinite. In aqueous solutions, pyrophyllite is only stable in supersaturated quartz solutions, otherwise the mineral bonds with water to form kaolinite; or, it can result from the break down of kaolinite under heating into pyrophyllite, alumina and water, although under natural conditions this doesn't occur but rather that kaolinte combines with quartz to form pyrophyllite, or if the clay adsorbs potassium ions (as most clays do), muscovite would form instead of pyrophyllite (Krauskopf & Bird, 1995). These two processes may be occurring within the fluvial and profile environment as there is a consistent presence of pyrophyllite, although mica is far more dominant. The presence of pyrophyllite does raise questions as it is normally associated with metamorphic rock material, of which there is none in the immediate vicinity or even elsewhere within the drainage of the Moordenaarsgat or Dassieboskloof rivers. The mineralogical properties of sediments are important in terms of providing additional palaeoenvironmental indicators through pointing towards their parent material source, the degree of weathering, sediment mixing, provenance and the extent of soil development within the localised vicinity at a small scale as well as inferences for the transport and deposition within the system itself. As these results represent only ten of the 66 samples, inferences obtained through XRD analysis of the mineralogy of the sediment for the entire locality and within the full the extent of the profiles is limited and assumptions are more reserved.

5.8. SEDIMENT AGES

Absolute ages for the deposits are integral to this study in order to give a temporal constraint and place the primary site within a context of deposition and provide a point after which the incision must have occurred i.e. to give some sense of event timing within the fluvial environment, as well as to provide an approximation of the accumulation rate within the alluvium.

Dating was carried out on four samples using the optically stimulated luminescence dating technique (OSL). This contributed two ages in the upper alluvium (WT3) and two from the lower alluvium (WT4) packages ranging from 5.27 ka years closely above the contact between the two packages and 1.16 ka years close to the uppermost surface of the alluvium package (Table 5.1).

Site	Depth (cm)	Age BP
WT3	90 - 100	1.16 ka ± 0.12 ka
	165 - 175	1.32 ka ± 0.18 ka
WT4	185 - 195	4.96 ka ± 0.27 ka
	270 - 280	5.27 ka ± 0.69 ka

Table 5.1: Measured age determinations for OSL samples

OSL sample WT3 90-100 proved to be well-behaved and distributed sample, despite the low dose and dose rate. A wide size range of sediment (90-250 μm) was used in analysis to ensure adequate material was prepared as the sediment was overall very coarse and yielded very little material below 250 μm (Telfer *pers. comm.*). Samples WT3 165-175 behaved similarly to WT3 90-100. WT4 185-195 was similarly well-distributed and well-behaved. For the final OSL sample at WT4 270-280, the distribution was untidy with multimodal data showing a strong skew and required that additional aliquots were run in order to confirm this. This suggests evidence of reduced bleaching (and poor distribution) or possibly (as is more likely given the known age of the unit above), substantial mixing with grains of an older D_e, therefore a suggested mean was not adequately representative of the data and resulted in a revised date in comparison to the preliminary proposed date, which employed the utilisation of an age model, namely the minimum age model of Galbraith *et al.* (1999), to isolate the youngest incident of bleaching that yielded a D_e of 3.30 ± 0.42 Gy (Telfer *pers. comm.*).

Although a crude method of estimation, based on consecutive dates provided by OSL and known or measured thickness of sediments or beds, it is possible to roughly determine an overall rate of sedimentation that occurred during deposition of the alluvium. The approximation of sedimentation rates is based on dividing the measured depth between age samples by the difference in age. As cycles of deposition, particularly within alluvium and slackwater, usually result in the initial removal of the surface and upper sediment prior to subsequent deposition upon it, there is a parallel unconformity or small-scale hiatus in between deposition, coupled with the fact that slackwater deposits depend on occasional, seasonal or infrequent events (depending on environmental conditions of the time) and these rates average out any events of the entire duration, the use of sedimentation rates in interpretation is limited and rather utilised as a guide towards overall view of the rate and size of deposition within the dynamic fluvial environment. The sedimentation rates that were calculated for the two alluvial sites were as follows:

- WT3 - 4.69mm/yr
- WT4 – 2.74mm/yr

These sedimentation rates are relatively high and above, substantially in the case of WT3, the average rate of accumulation, which is roughly 30cm/1000yr (Boggs, 1995). From these sedimentation rates one can extrapolate, assuming semi-constant continuation of vertical successional accumulation within the profile, the approximate timespan under which the final upper 90cm of alluvial sediment was deposited, hence, an extremely coarse, although helpful, age resolution of the near surface deposition and a narrower temporal control giving the estimated age after which incising of terrace commenced. For this purpose, the WT3 sedimentation rate was loosely extrapolated for the upper 90cm which yielded a depositional period of ± 192 years (obviously this is debatable as numerous factors coupled with lithological and stratigraphic characteristics could argue for either slower or faster rates of accretion as well as any subsequent sheetwash, slide or slumping, further cyclic erosion and deposition and even anthropogenic disturbance) and therefore an age approximation of ± 960 years BP. However, this at least provides a maximum age for the break of threshold and commencement of the switch in dynamic from accumulation (deposition or fill) to incision (erosion or cut) within the fluvial system at this point along the Dassieboskloof River. Likewise, such a rough estimate can be extrapolated downwards in the sequence from the basal date and used as the lower sedimentation rate. As above, there are numerous concerns when applying this; however it is purely suggested as a gross approximation upon which to very loosely base the beginnings of initial deposition within the system. As the basal OSL date provided occurs at 280cm from the surface, roughly 75cm above the contact between the alluvium and colluvium, using the WT4 accumulation rate, this yields a depositional period of ± 274 years, which, when added to the oldest age, gives an age approximation of ± 5 540 years BP.

5.9. STATISTICAL ANALYSIS

5.9.1. Principal Component Analysis

Principal component analysis (PCA) was performed in an attempt to elucidate some form of order within the dataset, especially when considering the fill cycle of deposition in order to group or classify strata that were identified within the alluvium. PCA was undertaken on the standardised data set comprised of 66 cases (rows representing sediment samples) and 17 variables (columns). The full statistical data matrix used in analysis is included in Appendix 6.

The first stage of PCA incorporated the generation and inspection of the eigenvalues, upon which it was determined how many loadings should be retained for further analysis. The first four eigenvalues (values above 1.0) or principal components account for a cumulative 79.03% of total variance within the dataset and were retained as they accounted for the greatest variance and lack of marked loadings on subsequent components.

	Eigenvalue	% Total variance	Cumulative Eigenvalue	Cumulative %
1	7.663742	45.08083	7.66374	45.08083
2	2.813929	16.55252	10.47767	61.63336
3	1.613370	9.49041	12.09104	71.12377
4	1.343885	7.90521	13.43493	79.02898

Table 5.2: Eigenvalues of retained principal components

The four principal components were then subjected to a non-orthogonal rotation (unrotated) in an attempt to establish any inherent variables underpinning the distribution or spread of the data. This step was followed by another rotation, varimax raw, to discern any underlying pattern within the component loadings and so identify the most important variables that account for the behaviour of the data. Loadings between -0.7 and -1.0 and 0.7 and 1.0 are highlighted in bold and represent variables with a more considerable influence on the data.

Variable	Factor 1	Factor 2	Factor 3	Factor 4
Organic Carbon %	**-0.780**	0.079	0.334	-0.244
Inorganic Carbon %	**-0.736**	0.423	-0.298	0.017
Conductivity µS/cm	-0.330	0.535	-0.641	0.056
Mean (ø)	**-0.820**	-0.551	0.053	0.069
Median (ø)	**-0.829**	-0.538	0.019	0.068

Sorting (ø)	-0.576	0.320	-0.205	-0.211
Skewness (ø)	0.369	0.182	0.494	0.078
Kurtosis (ø)	0.070	-0.520	-0.079	-0.525
Total Clay %	**-0.796**	0.187	0.036	0.258
Total Silt %	**-0.797**	0.172	0.059	0.252
Total Fines (Clay&Silt) %	**-0.896**	0.325	-0.079	-0.170
Gravel %	0.122	-0.137	-0.218	**-0.809**
Coarse Sand %	0.669	0.694	0.162	-0.127
Medium Sand %	0.051	-0.655	-0.591	0.236
Fine Sand %	**-0.814**	-0.357	0.237	0.020
Very Fine Sand %	**-0.764**	0.030	0.455	-0.178
Total Sand %	**0.888**	-0.339	0.099	0.165
Expl Var	7.664	2.814	1.613	1.344
Prp Totl	0.451	0.166	0.095	0.079

Table 5.3: Component loadings and explained variance from PCA (unrotated) of all samples

There is a marked difference in loadings between the unrotated and rotated components. The rotation using varimax raw does, however, appear to weed out less influential variables from component and modifies the loadings of component 2 and 3 to provide more meaningful data. Although the values are changed somewhat, component 4's information remains unchanged communicating particular variables' significance. Unrotated loadings show a great deal of significant variables within the first component with none in 2 and 3 and only one in 4.

Variable	Factor 1	Factor 2	Factor 3	Factor 4
Organic Carbon %	0.389	0.514	0.606	-0.057
Inorganic Carbon %	**0.858**	0.213	0.111	0.122
Conductivity µS/cm	**0.839**	-0.184	-0.258	0.069
Mean (ø)	0.164	**0.976**	0.052	0.010
Median (ø)	0.195	**0.971**	0.031	0.005
Sorting (ø)	0.672	0.155	0.181	-0.110
Skewness (ø)	-0.387	-0.335	0.341	0.202
Kurtosis (ø)	-0.256	0.249	-0.096	-0.649
Total Clay %	0.568	0.471	0.235	0.371
Total Silt %	0.548	0.484	0.251	0.367
Total Fines (Clay & Silt) %	**0.808**	0.397	0.363	-0.011
Gravel %	0.030	-0.091	0.009	**-0.853**
Coarse Sand %	-0.091	**-0.951**	0.239	0.038
Medium Sand %	-0.131	0.378	**-0.817**	-0.094
Fine Sand %	0.183	**0.853**	0.288	0.058
Very Fine Sand %	0.283	0.554	0.661	0.020
Total Sand %	**-0.821**	-0.381	-0.348	0.008
Expl Var	4.414	5.288	2.223	1.510
Prp Totl	0.260	0.311	0.131	0.089

Table 5.4: Component loadings and explained variance from PCA (varimax raw rotated) of all samples

On the other hand, component 1 of the rotated loadings shows four principal variables, four in component 2 and one in components 3 and 4. Values have also shifted from negative to positive relationships. Ordinarily, components can be regarded for interpretative purposes as reflecting various properties of the dataset. However, the conflict between the loadings derived from rotation creates difficulty in doing this. Loosely, component 1, which accounts for 45.08% of explained variance and whose values changed substantially between rotations, is taken to represent sediment chemical properties (organic carbon, conductivity, etc.) and total granularity of the sample (total fines, total sand). Principal component 2 (16.6% of variance) reflects the "sand fraction" as well as the granularity or particle size of the data set, i.e. fine and coarse sand along with the mean and median. Component 3 (accounting for 9.5% of explained variance) is interpreted as reflecting the "sand fraction" again with component 4 (7.9%) the gravel nature of the sediment within the dataset. The lack of any clear or distinct distribution of the data around specific variables besides the sand fraction lends to the relative importance of the sand within the environment, specifically the alluvium and contemporary fluvial dynamics.

5.9.2. Cluster Analysis

Cluster analysis was executed on the standardised variables for statistical analysis to identify inherent groupings as well as outliers within the data. Two methods of cluster analysis were used primarily, namely single linkage, for its propensity to recognize outliers and demonstrate more hierarchical clustering, and Wards method of clustering, which is more adept at identifying groupings. These two methods were used as the both groupings and unique cases are relevant to the study and provide useful information. Clustering of both the raw data as well as the PCA factor scores were carried out to obtain the most representative and apparent groupings and cases of outliers. It was hoped that cluster analysis could aid identifying samples at depths indicating similar properties and like characteristics or conditions of deposition within the strata, particularly within the alluvium.

Cluster analysis was firstly employed by clustering according to the variables used within the statistical analysis to show any predisposition of groupings around linked aspects of the measured properties of the sediment samples. Single linkage showed an outlying group consisting of total sand, coarse sand and skewness with total sand as the most extreme outlier. It also shows medium sand as an outlier to the larger group represented in green, shown in Figure 5.49, comprised of smaller groupings of median and mean and clay and silt, which is partly expected.

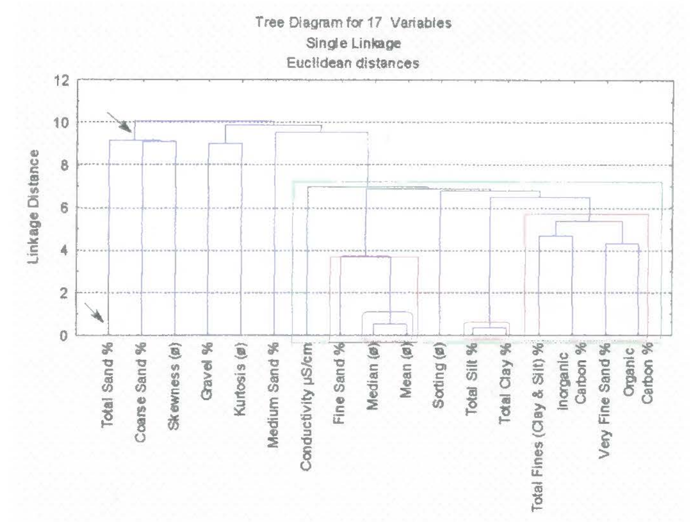


Figure 5.50: Single linkage cluster analysis by variable of entire dataset

Wards method, seen below in Figure 5.51, clearly divides the variables into two groups but more correctly as three sets of groupings at a linkage distance of 16-12. Again silt and clay; and median and mean; and total sand, coarse sand and skewness are more finely grouped with the now more closely linked variable grouping of inorganic carbon with total fines and organic carbon with very fine sand being reinforced. This indicates some sort of correlation in behaviour of these properties.

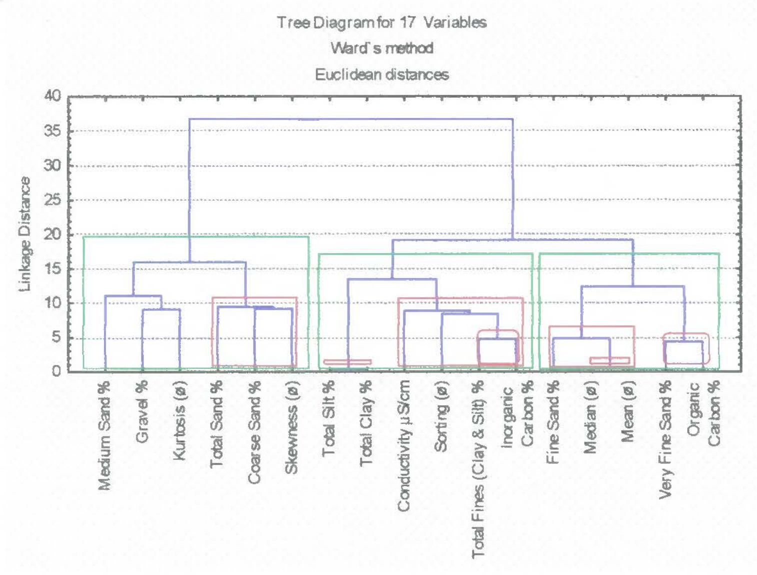


Figure 5.51: Ward's method of cluster analysis by variable of entire dataset

Figure 5.52 displays single linkage clustering of the full dataset showing a relatively close clustering of all but two of the cases at a linkage distance of 4 or less. Samples MG2C and WT3-195 are clearly distinguished as outliers with the furthest distances at 8.6 and 6.2 respectively. Smaller more obvious groupings occur throughout the analysis, however, these become clearer under Wards method of clustering. It is interesting to note the position of TT1 to the far right, an apparent outlier of the final group at a close linkage distance.

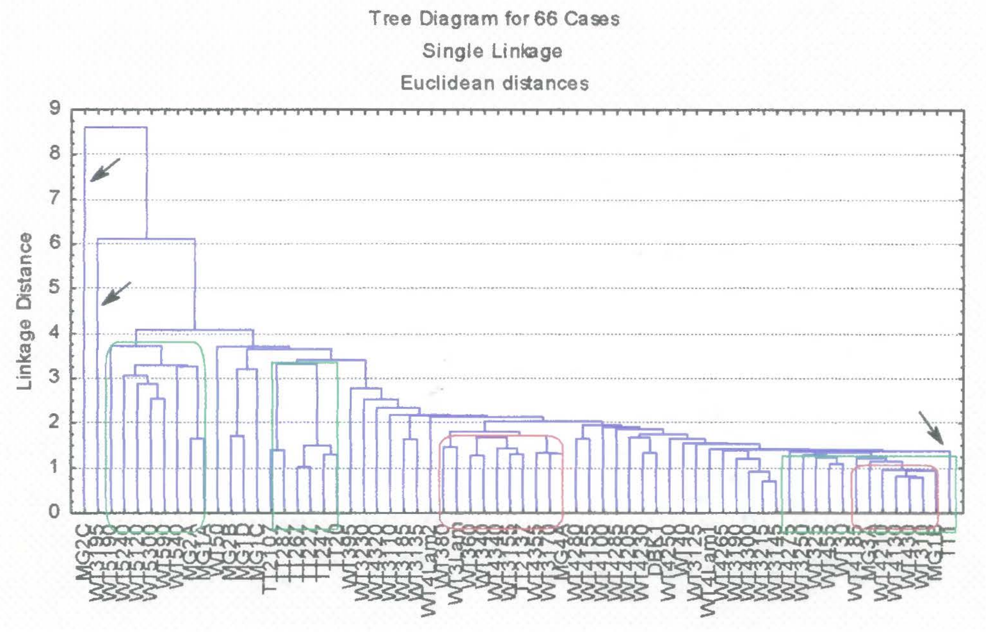


Figure 5.52: Single linkage cluster analysis for entire dataset by sample

Ward's method of clustering results in the presentation of five overall groups, as demarcated below in red, at a linkage distance varying from 36 to 16. These linkage distances, however, are far greater although the groups are more clearly defined in terms of sample proximity and relation. It is important to note the apparent tendency of grouping in places of samples according to location rather than depth or position within the profiles (shown by green in Figure 5.52), as seen by the clusters of TT2 profile samples, all the colluvial samples (WT5) and most of the contemporary MG sites. As it appears deposition occurred within the alluvium as sequences, depth associations would be misleading, therefore one notices that some clustering rather indicates a propensity for clustering of samples from within WT3 and WT4 together in places, inferring that these samples share similar properties or have inherent likeness. There are sporadic inclusions of some of the contemporary samples within these groups, possibly indicating fundamental correspondence to the sediment in profile, such as DBK1, MG4A and MG1B. It is also interesting to note that the WT4 lamination samples are grouped together, whilst WT3 lamination is coupled with WT3-80, showing distinctive yet comparable properties.

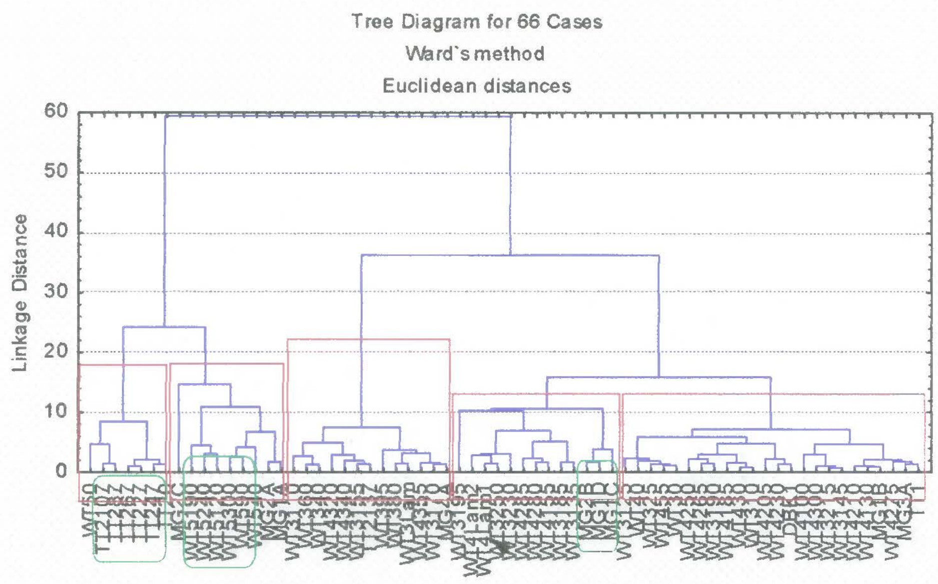


Figure 5.53: Ward's method of cluster analysis for entire dataset by sample

Cluster analysis was also carried out on the two alluvial profile site samples (as shown in Figures 5.54 and 5.55) to assess groupings that may infer deposition cycles and hence, associations at variable depths, or to establish if sediment properties would be linked according to depth more specifically. For site WT3, Wards method identified more variability in the size of groupings and their linkage distances. Again WT3-195 is noticed as an outlier from the other samples whilst WT3 lamination is consistently clustered alongside WT3-80. The primary linkage distances range from 10 to 5 with the most notable group highlighted in green. Smaller more closely related groups are circled in red and show similarities in sediment properties of the surface samples with a depth of 125cm, whilst the more central samples appear to be clustered together, with the exception of 145cm.

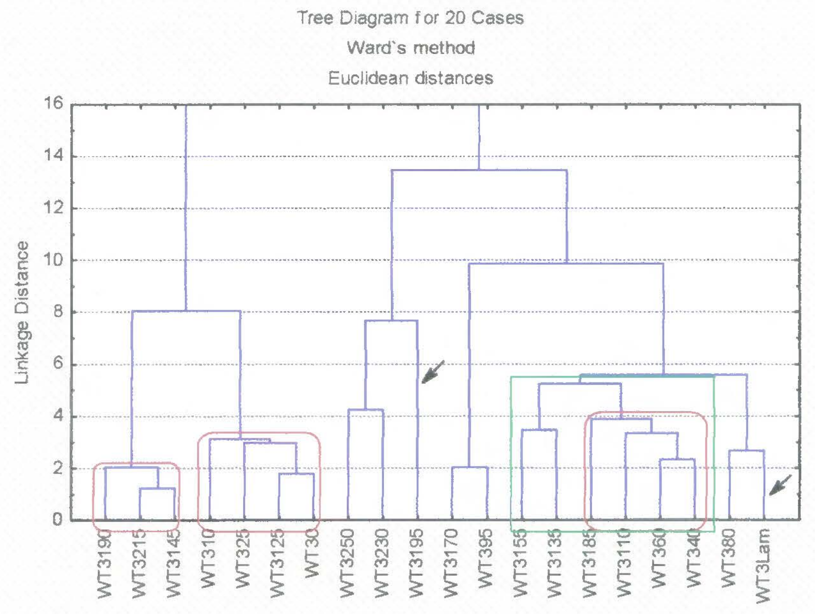


Figure 5.54: Ward's method of cluster analysis for WT3 site by sample

The WT4 site shows more defined clusters, with three groups consisting of two small clusters, visibly the lamination samples together again and the group to the far left (355, 340 and 320), and the much larger cluster in the centre (shown in red below in Figure 5.55). This significant group can be further divided at linkage distances from 7 to 4 to reveal four clusters. Again the samples at a more central depth of the profile are grouped together (205-265cm), as well as at the base (320-355cm). Notable similarities in samples from varying depths occur in the grouping at the left highlighted in green.

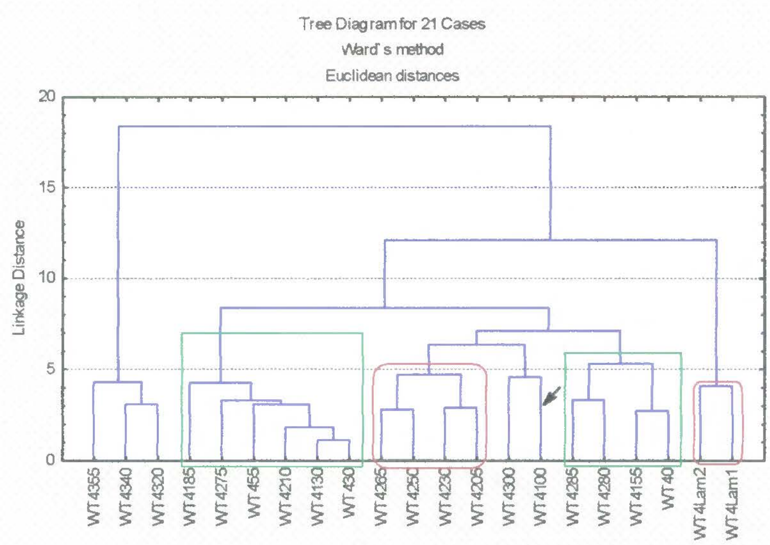


Figure 5.55: Ward's method of cluster analysis for WT4 by sample

Cluster analysis of the PCA factor scores shows much closer linkage distances, but overall, similar groupings. Single linkage still clearly identifies WT3-195 and MG2C as outliers but at a smaller linkage distance (4.3 as opposed to 8.7), although the remaining linkage distance now ranges from 1.5 and below. The clusters are still mostly by location for WT5 and the contemporary sites.

Ward's method of clustering on the PCA factor scores yields a much tighter version of grouping within the samples at smaller linkage distances too (11-3). More or less the same clusters are identified with the tendency to group according to location being maintained for WT5, TT2 and few of the MG sites. A clear difference between Wards clustering of the raw data against the PCA factor scores is the grouping of the two outliers together as well as more defined smaller groupings at all near enough the same linkage distance ranging from 0.5 to 1.0.

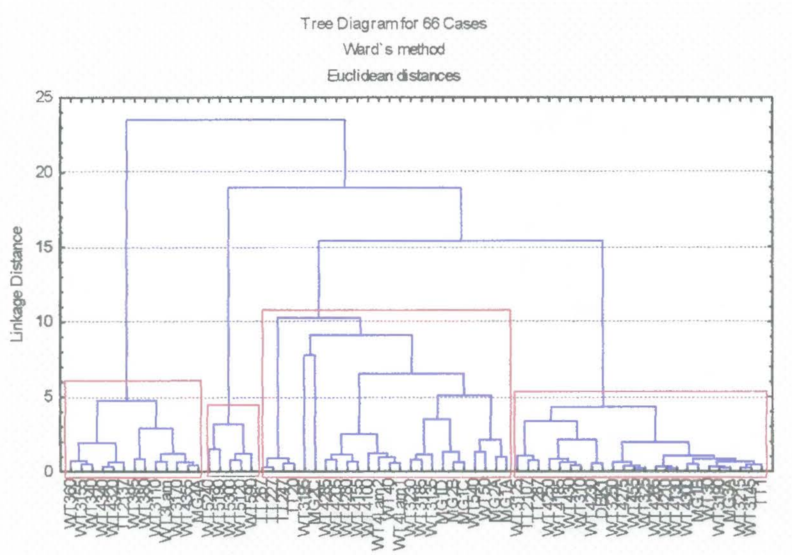


Figure 5.56: Ward's method of cluster analysis of PCA factor scores for all samples

5.10. CONCLUSION

The results presented within this chapter form only a single aspect of this investigation. When combined with field observations and the associated disciplinary theory of facies models, a more detailed stratigraphic history begins to emerge. From this, possible interpretations of depositional and erosional theories can be considered or facilitated aiding in the elucidation of further palaeoenvironmental reconstructions and related to the broader environmental context of the region in terms of the contemporary setting and future implications.

Chapter 6

DISCUSSION

6.1. INTRODUCTION

In this chapter the results of the laboratory, statistical and surveyed data are interpreted and synthesised with the possible implications of these findings then discussed in terms of their palaeoenvironmental and contemporary significance. The arguments proposed in this discussion should be considered in conjunction with theoretical background and other forms of palaeoenvironmental evidence presented in Chapter 3. Firstly it is important to outline the theoretical approach taken towards the results and leading to the discussion.

To deduce a satisfactory explanation for the palaeogeomorphological and contemporary features of the study locale, a nested approach was used working according to scale. Firstly the site itself was considered, then the greater catchment (including climate and land use), followed by a more regional view in terms of assessing tectonics and finally on a longer-term scale to reflect climate as a whole and long-term climate shifts inferred from evidence. In the context of this study, the term landscape evolution, which essentially refers to the description of surficial terrain modification through geomorphic adjustment in response to essentially climate-driven processes, is replaced to infer rather short-term modification and response of landscape, in particular of changes from aggradation/equilibrium to degradation.

From the evidence presented in Chapter 5 it is apparent that the area within the immediate vicinity of Wuppertal and the Dassieboskloof river and upland depositional environments exhibit various phases of sediment generation, transport and accumulation. In particular, valley fills reflect within their stratigraphic sequences evidence of fluctuating environmental conditions during the Holocene. Geomorphic evidence serves as a proxy indicator for processes that have influenced or impacted upon the landscape. Subsequent to deposition of this fill, the area has been further influenced by processes that have modified the morphology of landscape elements such as slopes and river channels as well as the sediment themselves. This chapter attempts to deal in more detail with the interpretation of the geomorphic evidence for environmental change preserved in the landscape at the primary site and along the Tra-Tra and Moordenaarsgat River valleys. Specifically the issue of climatic influences during the late Pleistocene and Holocene are addressed.

6.2. HYPOTHESISED DEPOSITIONAL ENVIRONMENTS

Firstly, it is necessary to identify the depositional source of the sediment under investigation. Textural, sedimentological and geochemical analysis, bivariate plots and statistical analysis as well as field observations and comparisons with contemporary environments convey the primary site at Wuppertal along the Dassieboskloof River to be divided into an upper package of fluvially derived alluvium and a lower package of colluvium. Grain size analysis (texture, mean grain size, etc.) as well as morphometric and microscope analysis ruled out the possibility of any aeolian influence.

Periods of stream aggradation and incision in response to climate change during the Quaternary have been noted across southern Africa. Accumulation may be due to conditions considered as reduced stream discharge and competence or embayments or slackwater deposits during major flood events. During this study, contemporary environments and deposits are used as an analogue for recreating and theorising past depositional and erosional environments within the region as controlling factors within the system could not have changed or varied significantly over time (the Holocene) or location to that large a degree. Sedimentary sequences and deposits within headwater valleys can be interpreted as representing a wide range of palaeoenvironmental conditions that prevailed throughout the phases of the Pleistocene and Holocene (Holmes *et al.*, 2003). The sequences reflect processes and particular sedimentary environments.

Sedimentary features are determined by the texture of grains as well as aggregates of grains arising from the size, shape and orientation of individual sediment. This texture has long been assumed to reflect the nature of transport and depositional processes, (such as current direction and velocities) thus aiding in interpretation of ancient environmental settings and boundary conditions (Boggs, 1995). Cluster analysis proved very useful in identifying the fact that various beds within units of the upper package of alluvium and sites along the Moordenaarsgat and Tra-Tra rivers shared similar sedimentological and geochemical characteristics, as well as outlining the separate nature of the colluvial deposit and some outliers of the contemporary deposits, such as the side channel at fine-grained MG2. This aids comparison of the samples to indicate similar depositional conditions. Sediments where dated using OSL, which indicates the time of deposition of the sediment.

The alluvium package of the terrace is comprised largely of homogeneous, coarse to medium textured sediment, both laterally and vertically, reflecting a consistent set of dynamics responsible for deposition throughout its accretion. Sedimentological and geochemical evidence suggests cyclical accumulation of the alluvial sediment during the late Holocene under moderate but fluctuating amounts of energy. There is the need, however, when interpreting the terrace to consider the spatial proximity of sites for sampling. The fact that WT3 is approximately two metres behind WT4 increases the likelihood of WT3 sediment being finer due to lateral accretion and belonging to an incrementally different depositional environment, while WT4 is also downstream from WT3. There are no real structural or topographical valley constrictions at any point downstream of the Dassieboskloof River terrace or along the Tra-Tra River to account for the sediment accumulation and retention within the valley. The exposure appears as the only situation of marked alluvium accumulation, with the possibility that it occurred to a lesser degree on the opposite bank but was subsequently eroded or removed for small-scale agricultural farming (vegetables) and then artificially built up again to create a levee.

A theorised depositional environment has been proposed to explain the process of accretion of alluvium along the southern bank of the Dassieboskloof River. The terrace consists of phases of upper-flow-regime turbidity current sequences with deposition by a previous event being followed by initial erosion of the surface of the deposit followed by subsequent deposition of further sediment. This terrace has been hypothesised to have been deposited by overbank flow associated with periods of high flow regimes and flooding of the fluvial system above and on top of a previously deposited small-scale colluvial fan. Over time, as the dynamic shifted from the depositional colluvial feature dominating and constricting the palaeochannel with higher flow regime, the toe of the colluvium may have become scarped and later truncated by the palaeochannel, such as the situation currently occurring at the valley constriction mid-way along the Moordenaarsgat River. With a more equable flow regime and higher frequency, possibly associated with more seasonal behaviour of the catchment discharge pattern, along with palaeochannel migration, an elevated or raised flood channel may have been initiated and deposited above the colluvium, similar to the contemporary elevated flow channels evident along the Tra-Tra valley and presently dry lateral channels of the Dassieboskloof itself. These extra channel deposits began as cobble and pebble lags infilled by coarse and medium sand as the flood receded; successive flooding and overbank flow account for the coarse nature of the sediment and occasional pebble lags depending on the magnitude and capacity of the flood event and its inherent sediment yield. In this way a natural levee may have begun to form. Levees develop on the margins of the channel as a result of rapid fall-out of the coarser components of suspended load, thicker sand sequences accrete during this process and usually more rapidly than flood basins beyond thus becoming areas of positive relief (Pye, 1994). Although the lower unit of the alluvium package does not appear as a levee with a lower lateral relief, it does indicate that it would have exhibited positive relief. Within the facies there is an overall fining upwards sequence as deposition is dominated by rapid accretion of coarse material, and in places, lags of partly rounded pebbles indicative of flood events, with a gradation from coarser to medium sand with increasing height. This normal grading corresponds to contemporary examples of this process of aggradation as seen in TT2 along the Tra-Tra River as a side channel bank profile as well as in the Moordenaarsgat extra channel deposit MG1D. Fining upwards sequences and darkening of colour upwards indicates waning or slowing flow regime and loss of energy. Rounded cobble and pebble grade gravels indicate a high-energy fluvial environment at the starting point of deposition.

As the relief of the deposit increased and the capacity of the stream changed with time, each successive flood event deposited progressively finer sandy material. At some point towards the upper unit of the terrace the dynamic shifted towards accretion through slackwater deposits rather than merely overbank flood deposits. The presence of increased organic content along with laminations and a wider range of sediment size indicate a more varied energy input regarding the process of deposition. However, these slackwater deposits are not characterised as such fine sediment as those found in the Moordenaarsgat valley. This could in part be due to the parent material and sediment transport distance as the overall terrace is coarser to more medium sand texture. However, grain size is not only determined by externalities (Folk, 1974). Fining upwards sequences display the consistent and regularity with which flood events occurred throughout the record as deposition continued both vertically and laterally. There may also have been an additional influence of the terrace area as slackwater deposits if, as the unnamed tributary converged with the Dassieboskloof at Wuppertal, larger flood events and higher discharge resulted from backflooding. The upper unit of the alluvium package was rapidly accreted, according to sedimentation rates. An additional theory is that the upper alluvium may have accreted not strictly as slackwater deposits, but as a result of an embayment to the lee of the southern bank resulting in recirculation eddies and subsequent deposition out of coarse material and sandy sediment. Periods of high runoff would also incorporate sheetwash off the saddle and immediately surrounding slopes into the deposit. Such alluvial deposits forming "benches" and "terraces" are commonly exposed along the concave banks of the meanders and on both sides of straight, stable reaches of a channel. Overbank deposits, according to Brooks (2003), forming the upper two to three metres of the bank are generally weakly defined and composed of massive beds and laminations. Underlying these are oblique low-angle accretion deposits that are similar texturally and structurally, with the transition between the two being gradational and indistinct. A decrease in sedimentation is interpreted to represent a shift from oblique/proximal overbank accretion to distal overbank deposition (or reverse) probably relating to the relative height of the aggrading surface above the river channel and an associated decrease in frequency of inundation.

Based loosely on sedimentary subfacies descriptions by Hiroki & Terasaka (2005), interbeds of parallel bedded sand and gravels containing scattered pebbles and sporadic cobble lags (especially in the lower facies) with parallel (minorly wavy) laminated coarse to medium sand beds exist in the upper to basal facies of the exposed and excavated WT3 profile; the laminations are more defined and commonly occurring in the middle to lower facies. Wavy laminated medium sand beds containing scattered pebbles occur in the upper facies of the exposed WT4 profile followed by interbeds of parallel bedded medium sand and gravels with frequent cobble lags in the middle to lower facies. A lack of erosional surfaces in the upper facies of WT3 and lower facies of WT4 indicates continuous accumulation of sand in flat beds under planar bed flow. The wavy laminated sands of the upper facies exhibit erosional surfaces indicative of repeated deposition and erosion.

The kurtosis of the samples taken from the terrace (mesokurtic to leptokurtic) indicate a moderate to lower energy environment while, in contrast, the bivariate plot consisting of very large variations suggests a high-energy stream. Slackwater deposits are characterised as lower energy environments than active channels. Coarse skewed distributions indicate higher energy environments as fines are winnowed out subsequent to flooding and consistent with high energy flushing of the system, while fine skewed distributions suggests receding flood leaves some silt sized and fine sediment. Coarse and fine skewed distributions alternate throughout the terrace in accordance with the hypothesised flooding events. Morphometric properties and microscope analysis revealed conchoidal fracturing, breakage patterns, ridges and evidence of percussion marks and pitting also indicative of high-energy environments. They may possibly also be the result of chemical or solution etching during diagensis during post-depositional processes, however, this may apply to the colluvium but the alluvium is too well leeched and exhibits little chemical weathering in contrast to the colluvium where cementation is occurring. First cycle weathering debris and sediment grains that have been transported short distances from their source areas are typically angular to sub-angular, in contrast to grains that have experienced a longer transport history, even involving more than one cycle of erosion and deposition, exhibit a much higher degree of rounding, although the rate of rounding during transport is largely governed by the hardness of the minerals concerned (Pye, 1994). The grain properties of the terrace suggest first cycle deposition. Quartz grains that have become rounded can preserve this property through several sedimentation cycles and hence is not easily lost. Stream transport is relatively ineffective in rounding small quartz grains (Boggs, 1995). Pebbles, depending on their composition and size, can become rounded by fluvial transport in vastly different distances. Pettijohn (1975) proposes that quartz pebbles require 300km to reach a roundness of about 0.6. As all the river systems in this study and within the region considered do not extend as far as 300km in total (up until the study boundary along the Tra-Tra Valley) it aids in explaining the general lack of rounded or well rounded size pebbles or larger sized fragments. Using grain-size techniques to identify depositional environments can be difficult due to the variability in depositional conditions within major environmental settings. The energy conditions and sediment supply within river systems can differ markedly from one river to another and even within different parts of the same river system. Grain-size distributions reflect processes, not environments, and sediment transport processes are not unique to a particular environment (Boggs, 1995).

Using Moordenaarsgat River contemporary fluvial deposits as an example, the valley constrained alluvial fan deposits a range of flood deposits, from a slackwater deposit to extra channel, lateral flood channel and channel bars to the right bank while the primary trunk stream channel flows in closer proximity to the left bank. The same may have applied in the case of the terrace. As the Dassieboskloof River discharged out of the bedrock canyon, at the proximal end of the exposure, and lost its energy and capacity, it may have flooded towards the right (or south) bank as it maintained its primary active channels closer to the left bank. Such a pattern is obvious at present as lateral flood channels are maintained in the vicinity of the base of the terrace to the south of the active channel.

A similar case to the study site is the headwater basin of the Klein Seekoei River to the west of the Great Escarpment, which displays evidence of multiple episodes of deposition of colluvial and fluvially derived sediments, like that of the sediment packages along the Dassieboskloof River. The earliest depositional phase is represented by deeply weathered, calcretised gravel deposits which were likely emplaced by debris flow and fluvial processes in the form of a fan (Boardman *et al.*, 2005). Similarly, the colluvium present as the lower orange package, WT5, using the law of superposition, represents the earliest depositional landform at the primary site. The colluvium is evident in the form of a scarped shallow footslope fan deposited off a combination of the northern slopes of Singkop, the eastern edge of the exterior wall of the Dassieboskloof River canyon and the saddle that separates the two. The finer-grained colluvial deposit, comprised mostly in origin from shale and sandstone, was subsequently buried by medium- to coarse-grained, partially to unconsolidated sediment, with emplacement occurring during the Holocene, and an ongoing phase of sediment transfer from hillslopes into the drainage system. A chronology for the Holocene alluvium emplacement has been established from 5100BP to 1000BP.

To piece together evidence from past environments and conditions of the area is to attempt to establish the cause that initially created such a well developed and significant deposit and then an erosional feature present along the banks of the contemporary active fluvial system. In common with many rivers in the interior of South Africa, minor channel incision and erosion followed by extensive deposition has broadly coincided with relatively arid intervals identified from other environmental proxies. Characterization of magnitude and frequency of exceptionally large floods involves utilization of a variety of techniques for estimating palaeoflood peak discharge: (1) indices of flow strength derived from studies of flood-transported gravel and boulders, (2) erosion of tributary debris fans, (3) floodplain vegetation, and (4) slackwater sediments (Kochel & Baker, 1988). Where bedrock has sufficient erosional resistance, narrow, deep confined channels may result and exceptional flow depths accommodate the high discharges of extraordinary floods. The rapid fall in stage that follows from the flood peak through narrow, deep bedrock channels causes suspended sand and silt to be deposited at high flood levels in slackwater areas, and preserved as the flood rapidly recedes. Channel margin areas may accumulate slackwater deposits of relatively coarse suspended load and wash load with appropriate geometry and erosional resistance. It is also important to examine processes of tributary backflooding and associated slackwater sedimentation (Baker & Kochel, 1988). High-level flood deposits and SWDs accumulate rapidly from suspension during large floods in areas where flow and current velocities are locally reduced, such as in back-flooded tributary mouths, channel expansions, downstream from bedrock spurs and/or slump blocks, and along bedrock walls in shallow caves (Saynor & Erskine, 1993). However, studies involving relief peels have shown that slackwater sediments are, for the most part, not massive, exhibiting mainly flat lamination but rather indicate that slackwater sedimentation is characterised by moderate rates of deposition instead of punctuated or rapid rates (Hattingh & Zawada, 1996).

Colluvium is more massive in small-scale appearance as primarily sourced from Bokkeveld Group, which is dominantly alternating argillaceous shale and arenaceous sandstone bands. The shales and siltstone, which comprise a large constituent, and its inherent clays and silts, has no well-developed parting along bedding planes, it does, however, display larger scale banding. Shales do tend to exhibit some bedding planes due to parallel orientation of clay minerals, whereas siltstones lack bedding plane fissility (Whitten & Brooks, 1972). The Bokkeveld shales in combination with Cederberg sandstones have weathered to form colluvium (soil parent material) that in turn has weathered into a matrix supported diamicton with high clay content. This is characterised as argillic brown earths or luvisols typically in sequence of A/E/Bt soil profile (Briggs *et al.*, 1998) and exhibited here as commonly reddish to orange in hue with calcareous deposition at or near the base. Unless an age was established for each individual bed at their base and top it is impossible to say that there is an unbroken period of deposition with conformable beds, however, within the short period that spans the duration of deposition to the apparent switch in dynamic (roughly 5 200BP) one would assume that, although small or short hiatuses inherent between cycles of deposition, that overall the beds of the alluvium are conformable in comparison to the contact between the colluvium and alluvium. It appears that the colluvium may have been previously eroded followed by the deposition of the alluvial sediment but as no date from the surface of the colluvium has been ascertained, one must be cautious in attributing any time period or unconformity. As the process of downslope movement of colluvium is gravity driven and continually ongoing with occasional sporadic input events due to mass movements such as slope failures, rock falls, etc. The process of deposition throughout the colluvium is regarded as more or less consistent through time, although dependent on particular environmental conditions that may encourage or inhibit such deposition.

Quartz, smectite, chlorite, mica, calcite, kaolinite and vermiculite occur within the samples. Micas and chlorites remain dominant as primary layer silicates under mild, generally physical, weathering conditions, whereas under more intense weathering the primary minerals are transformed to secondary clay minerals, such as vermiculite, by the leaching of interlayered potassium from primary micas. High soluble silica concentration in the weathering environment results in the formation of 2:1 layer minerals like smectite. Leaching and the removal of silica produces kaolinite (White, 1997). Calcite ($CaCO_3$) is indicative of more arid conditions as it occurs in the clay fraction of soils, particularly in high pH soils of arid regions, where it accumulates within the profile to form calcrete. This applies to the lower package where micrite and calcrete coatings are evident. Chlorite is a regular mixed layer mineral with a high magnesium content rendering it susceptible to dissolution. Mica, vermiculite and smectite differ mainly in their extent and location of isomorphous substitution and type of interlayer cation that predominates. Vermiculite minerals also exhibit predominance towards magnesium and calcium cations and aluminium substitution (source of calcium for precipitation out as calcrete) (White, 1997). Most of the clays present within the profiles show a relatively high Al, Mg and Ca composition, which can be removed through leaching. From the minerals identified in XRD analysis, one can infer the stage of weathering in which they were present. Calcite is indicative of early weathering stages and occurs mostly in sand, silt and clay fractions of young soils of arid regions where a lack of water inhibits chemical weathering and leaching; quartz, muscovite, vermiculite and smectite belong to the intermediate weathering stage and usually indicate more temperate regions (White, 1997).

Weak interlayering bonding and free movement of water and cations within the montmorillonite smectite minerals results in expanding-lattice clay, with more irregular layers and a smaller size than mica and kaolinite minerals; therefore it plays a key role in shrink-swell cycles of soils or sediment (White, 1997; Holmes, 2003). Vermiculite is more reactive than muscovite with a higher cation exchange capacity (CEC), whereas smectite's CEC is reduced and its ability to swell is decreased by the removal of Mg and Ca cations by leaching. Strong leaching results in Al being left as the only available cation to replace the interlayer potassium of smectite, which in combination with oxygen and hydroxyl, changes the structure to become chlorite (Paton, *et al.*, 1995). Mesodiagenesis results in clay mineral authigenesis, where one kind of clay mineral is altered to another. It is also the stage during which the precipitation of carbonate (calcite) cements occur with accompanying porosity reduction and when mineral replacement takes place with partial to complete replacement of some silicate grains and clay matrix by new minerals, such as the replacement of feldspars by calcite (happening within the colluvium). Dissolution, replacement and oxidation occur within telodiagenesis, in this case where feldspars are altered to clay minerals and oxidation of iron carbonate minerals occurs as a result of uplift or exposure (Reineck & Singh, 1980). In overall terms the process of leaching is responsible for the segregation of more mobile elements, such as K, Ca, Na and Mg, from less mobile, Fe and Al, with silicon occupying an intermediate position. However, biospheric reactions play a vital role in this process and can alter profiles accordingly substantially (Paton *et al.*, 1995). The upper packages are very leached, containing few ions. This may account for the lack of weathering and pedogenesis within the profile, although small aggregates and nodules to suggest embryonic soil formation. In comparison the colluvium does not experience much leaching and weathering processes are obvious.

In terms of stages of deposition, it would follow that the proximal western end would experience primary deposition with a lateral extension of fluvial sediment being deposited out towards the distal eastern extreme of the exposure. Along with this fact, and perhaps distorting the assumption, is that swirling overbank flow would firstly deposit furthest away from the trunk channel, hence, against the lee of the saddle towards the centre of the terrace as well as in the proximal corner close to the head of the canyon where energy was lost quickest and eddies may have formed to the lee of the flow. From here there would be a subsequent progression of sediment accumulation forward towards the bank of the secondary flood channel and gradually across extending to the distal edge if the volume and velocity of the water was sufficient along with its load. This would account for the relatively thinner package of alluvium that tapers out towards the flank of the *koppie* along the eastern periphery of the deposit. There also appeared to be a reduction in coarse elements along the eastern verge suggesting an associated decrease in capacity and load.

Floodplains are considered to be essentially depositional features with laterally and vertically deposited sediment. Lateral accretion tends to more coarse-grained and cross-stratified. Vertical accretion tends to be finer-grained with horizontal stratification and occurs mostly when the stream floods, thus developing natural levees that are coarse-grained neared the stream channel and progressively finer away from the channel. Meanders develop as lateral erosion dominates over downcutting. A terrace is characterised as a step cut into alluvium (or bedrock). The alluvium is first aggraded as it is deposited in valley bottoms. Simplistically, either base level drops or discharge increases to result in downward erosion or incision. Aggradation generally occurs due to infilling of valleys from an oversupply of debris with respect to the stream's capacity. This surplus can be from accelerated slope erosion through severe storms, fire, vegetation removal, long-term climate change, increase rates of erosion through process change, etc., or input from external sources, such as mass movements, or a decrease in discharge associated with climate change. Incision into the alluvium or previously deposited material occurs as the stream obtains an increased capacity relative to supply, possibly due to numerous reasons, such as base level change, tectonic uplift, changes in discharge (again climate change affects load-discharge relationship) and decreased slope erosion reducing load input (through increased vegetation cover or a change in sediment). Terraces can be either erosion or depositional, although generally considered an erosional feature, the terrace under investigation is a depositional terrace.

Rates of sediment deposition on cut-and-fill floodplains strongly determine the degree to which pedogenic features develop within aggrading alluvium. Three pedofacies common to semi-arid alluvial deposits correspond with increasing aggradation rates: (1) cumulic soils (2) multiple buried soils and (3) no pedogenic features (Daniels, 2003). Radiocarbon age determinations from alluvial fills indicate that floodplain aggradation greater than approximately 5mm/yr limits soil formation: this represents a threshold rate of pedogenic assimilation. Floodplain soils formed under aggradation rates lower than this exhibit a strong positive relationship between aggradation rate and total $CaCO^3$ percent, and a negative relationship between aggradation rate and organic carbon percent (Daniels, 2003). WT3 sedimentation rate is close to the boundary and accounts for the lack of any indication of soil formation. The sedimentation rates for the primary site's deposit at Wuppertal are relatively high, although in line with few other alluvial terrace rates of aggradation (Makaske *et al*, 2002 and Knox, 2005). They do, however, exceed, substantially in the case of WT3, the average rate of accumulation, which is roughly 30cm/1000yr (Boggs, 1995). As the profiles were not continuous it was not possible to calculate or even infer a sedimentation rate between the two sites, especially taking into account their relative vertical and lateral positions from the trunk channel and terrace face. Higher sedimentation rates can be seen as a tendency for increasing vertical magnitudes of sedimentary facies with increasing size of the contributing watershed, but may also result from continuing very high magnitude influxes of tributary sediments that were previously stored during previous regional climates (Knox, 2005). As Knox (2005) describes, discrete multi-centimetre thick sandy deposits can be associated with former large floods, indicating that average vertical accretion rates do not properly represent the more typical episodic nature of the floodplain evolution. The relative thickness of vertical accretion sedimentation is both a function of the length of time that deposition has been occurring and the relative elevation of the alluvial surface.

Laminations are produced by less severe, or shorter-lived, fluctuations in sedimentation conditions and result from changing depositional conditions that cause variations in grain size, content of clay and organic material, mineral composition, or microfossil content of sediments (Boggs, 1995). Boundaries between the laminae and beds are sharp, not gradational, in the terrace, especially in WT4. Changes in clay content or mineral composition, such as alternating mica-rich or darker heavy-mineral sand laminae in mica-poor lighter beds, or else as a result of higher presence of fine, dark coloured organic matter, can create obvious laminae in otherwise even-sized coarser sand beds. Such colour changes accentuate the laminations, as noted in the exposure as dark grey against grey and lighter grey beds. The wavy laminations present within the stratigraphy do not comprise of any higher proportion of clay than the immediately surrounding beds, nor did bulk XRD establish any significant heavy-minerals or different mineral composition in comparison to other alluvium samples. Although WT3 sampled lamination exhibited a coarser grain-size, it did not apply to WT4 and was not substantially dissimilar to overall textural trend of the terrace. The one key point, which does set the laminations apart, and in so doing, imparts an explanation for their formation and indicates a particular depositional environment, is their dramatically higher proportion of organic carbon. Laminations are formed in environments of deposition where alternating periods of moving and slackwater occur or the process of sediment supply is rhythmic or periodic (Leeder, 1991). Laminae can be produced by deposition from suspension and traction currents and have the greatest potential for preservation where organic activity is minimal, or in environments where deposition is so rapid. Laminations can also be produced in the plane-bed phase of high flow regime as a result of bed load segregation, or alternatively by low-relief bedforms in shallow depths such as low-amplitude sand waves in very shallow depths of a river under conditions of low or transitional flow regime (lateral pinching may oocur) (Reineck & Singh, 1980). Natural levees often exhibit ripple drift lamination, evidence of the very rapid nature of deposition (Reid & Frostick, 1994). The laminations present within the terrace suggest a combination of depositional factors. Their wavy coarse to medium sandy nature and lack of segregation, particularly in WT4, indicates periods of transitional flow in shallower depths, whereas the sub- to parallel thinner laminations of WT3 indicate more rapid deposition. It is possible that these laminations formed erosional surfaces between successive flood events or periods of deposition and therefore represent short hiatuses in the stratigraphic record before subsequent deposition is resumed, at points along the disturbed more distal points of the terrace laminations are exposed as being more resistant to weathering and erosion as well as the fact that decreasing energy would have resulted in sand laminations being overlain by finer sand cross bedding and subsequent planar laminations of silt, however, these are conspicuously absent, thus reinforcing the suggestion that they could be past erosional surfaces.

The evidence accumulated throughout analysis of the terrace suggests an overbank deposit facies for the upper alluvium package. The accumulation of layer upon layer of overbank deposits relies upon the passage of successive flood waves of sufficient magnitude to overtop the channel banks; vertical accretion is self-limiting since vertical growth of the floodplain will eventually take it beyond the reach of all but very infrequent events. However, erosion of the floodplain either by laterally migrating channels or during large flood events can strip off overbank deposits and rest the system for a further period of deposition. Where vertical accretion is the dominant floodplain process, this periodic floodplain stripping can lead to complex cut-and-fill structures within the deposit. Vertical accretion can persist over longer periods under conditions of active aggradation. Increased sediment supply as a result of climatic modifications of the hinterland can produce a wave of sediment that passes down the system and results in aggradation. Where gorges cut are wider, coarse-grained alluvium accumulates between large and infrequent flood events, as seen at the primary site. In wider, less steep valleys where the river is still relatively energetic but where migration is restricted for some reason, such as by bedrock outcrop, vertical accretion with typical fining-upwards sequences of gravel and sand will develop with episodic erosion. This is evident throughout the deposition. Flood events of a high magnitude but low frequency are responsible for stripping out accumulated deposits and resetting the depositional cycle (Reid & Frostick, 1994).

As described in Chapter 5, the Moordenaarsgat and Tra-Tra rivers exhibit both braiding and anabranching channel patterns respectively. Braided channels are characterised by the division of a single trunk channel into a network of branches and the growth and stabilisation of the intervening islands, or bars. Braiding is enhanced by erodible banks, sediment transport and abundant load along with channel shifting and temporary deposition of bars, and rapid and frequent variations in discharge that leads to an increased magnitude in the alternating patterns of erosion and deposition (Ritter *et al.*, 1995). As channel bars are deposited, flow is deflected around it and bank erosion ensues. As braiding occurs in the region, it implies that the slope and width-depth ratio are under high conditions of dominant discharge; the bars do not denote instability and the pattern is simply a response to external controls and may be maintained for a long period of time, possibly even being as close to a true equilibrium as a meandering pattern (Ritter *et al.*, 1995). Anabranching is a term used to describe a system of multiple channels characterised by vegetated or otherwise stable islands that divide flows at discharges up to around bankfull level. The interconnected network of low-gradient, relatively deep and narrow channels with variable sinuosities and relatively stable banks are noted as transporting predominantly suspended load (Ritter *et al.*, 1995). Tributaries typically supply gravelly sand during low flows, such as in the Orange River, the sediment accumulates at the mouth of the tributary near to where it enters an anabranch. Development of such anabranches includes channel aggradation, stable, cohesive banks limiting widening, and a flood-prone regime; they can also independently exhibit straight, meandering and even braided patterns. In many alluvial anabranching rivers, a highly variable flow regime is an important factor promoting anabranching, particularly where this occurs in conjunction with mechanisms to block or constrict channels (e.g., channel sedimentation, vegetative blockages, flow ponding). These factors periodically promote overbank flows and cutting of anabranches on floodplains or islands (Knighton, 1996). Such multichanneled rivers are associated with aggradation represent disequilibrium, although Knighton (1996) suggests they can exist for long periods of time. According to Tooth and McCarthy (2004), many authors in fact regard alluvial anabranching (including anastomosing) essentially as a transitional river pattern that only forms when geomorphic or hydrological conditions change, such as following base-level rise, avulsion of a meander belt to a new position on a floodplain, and consider that multiple channels will eventually be replaced by a single channel. Thus, it may be concluded that both the Tra-Tra and Moordenaarsgat Rivers are in a transitional stage and have experienced some change in conditions and undergo variable flow regimes prone to episodic flooding.

6.3. HYPOTHESISED EROSIONAL MODELS

6.3.1. Introduction

Rivers do not simply erode under one given set of conditions and deposit under another; there is a more complex interplay of a variety of factors at play (Holmes, 1998). Fluvial systems may respond variably within the same region to a combination of factors including changes in base level, tectonic activity, climate change and alluvial morphology (Bullard, 2004). The landscape of the northeastern Cederberg region is a function of numerous factors acting over an extended period of geological time and, to a lesser degree, of human impact over the past 200 years. The dominant factor is the geology and the juxtaposition of rocks of differing relative resistance to erosion. Other significant factors include the post-Gondwana tectonic history, Quaternary changes in base-level, climate fluctuations and the present climate and natural vegetation.

Within fluvial systems, adjustment of slope can be made by a major change in planform rather than by vertical filling or trenching. The initial fluvial response may not be the same as its end response, while regulation of a river may generate responses that do not remain local but alter channel morphology over greater lengths of the river system. There is a high tendency in fluvial response for a change in discharge to be counterbalanced by a simultaneous change in the character of the load (Ritter *et al.*, 1995). A significant geomorphic feature indicating dynamic change and fluctuating conditions is shown in the headwater valley of the Tra-Tra River along the Dassieboskloof River tributary within the Cederberg region, where a period of deposition shifted into a period of erosion and incising of the river to result in a significant exposure of colluvium overlain by alluvium. The severity of erosion occurring depends upon the quantity of material supplied by detachment and the capacity of eroding agents to transport it (Morgan, 1995). The extent and pattern of flow in many dryland rivers is largely determined by the relationship between discharge and transmission losses (Bullard, 2005). There are several concepts about the formation of cut-and-fill terraces and alluvial fans. Amongst others, climate change is an important factor, which may trigger the transition between aggradational and degradational behaviour of a fluvial system.

With regards to erosion, there are issues providing an age for the exposure. In this particular instance there are no varnish coatings or lava seals or cosmogenic isotope dating carried out. Although OSL ages were used to reveal the age since burial, i.e. time and roughly the duration of deposition, there was no method available to determine the age, with relative certainty, that the incision of the terrace was initiated or the duration of the entire process of incising. Inferring from the most recent age nearest the surface of terrace, and using the law of superposition in conjunction with calculated sedimentation rates, the incision seems likely to have occurred after approximately ± 960BP.

6.3.2. High Magnitude Flood Events

High energy fluvial events may have been responsible for the large scale removal of material. Such high magnitude low frequency events have been proven to have occurred both prior to and during the Holocene (Holmes, 1998). There is difficulty determining and assessing the magnitude of rainfall intensity within the basin, which is largely responsible for initiating and maintaining channels; especially as networks appear to be closely related to conditions of maximum runoff. In semi-arid regions, the proportion of precipitation receipt that flows as surface runoff and which is therefore immediately available for erosion is greater than in humid environments. Rainfall intensity is a more appropriate parameter that influences short-term availability of water for channel cutting, especially during infrequent storms. Seasonality in combination with rainfall intensity is a marked factor of importance with regard to drainage and efficiency, fluvial dynamics and specifically flood events. Certain conditions seem conducive to cutting and expansion and thus erosion across otherwise uneroded terrain, the most important of these conditions was found to be very high runoff rates from infrequent storms (Holland, 2000).

A study of overland flow erosion rates in semi-arid Mediterranean environments of Spain shows that at a small-scale, sediment yield and erosion is spatially variable with varying responses to rainfall whereas at a larger-scale and within catchments for individual events there is a greater relationship between rainfall and sediment yield. High magnitude, low frequency rainfall events impact heavily on sediment transport in semi-arid environments and response is largely based on models resulting in the understanding of such geomorphic events being somewhat limited (Bullard, 2004). When considering such events, it is also essential to incorporate additional processes operating, such as hail erosion. The hydrological response of hydrologically similar surfaces to changes in topography and storm duration emphasizes the importance of stream–slope coupling, or connectivity, in determining spatial patterns of runoff-response in semi-arid areas (Bullard, 2004). Studies by Romero-Diaz *et al.*, (1988), Gallart and Clotet-Perarnau (1988) and Mena *et al.* (2002) (as cited by Bullard (2004)) indicates that storms provide a high correlation with erosion rates although singular large events show relatively low annual erosion rates in comparison to medium sized events; suggesting that the magnitude and frequency of medium events are integral in terms of sediment yield and erosion. This is largely attributed to vegetation cover, high surface roughness and slope geometry within the environs. The long duration of effective rainfall may result in high or very high water stages in the main rivers and runoff is affected by rainfall intensity and soil type. It was also shown that effective soil sediment size changed during a rainfall event due to aggregate breakdown and soil crusting, thereby impacting sediment transport. The impacts of individual events on flow and sediment transport have been found that the nature of the relationship between suspended sediment and discharge during flood events is unpredictable, suggesting that variation in suspended sediment concentration cannot always be attributed to the flushing of sediment at the beginning of an event and the exhaustion of sediment supply at the end, but that it is also governed by spatial and temporal variations in rainfall–runoff and sediment supply (Alexandrov *et al.*, 2003 quoted by Bullard, 2005).

Flood peaks are closely related to catchment size and the number of tributaries, and flow is lightly related to hillslope gradient which can reduce the concentration time and reflect short periods of intense rain within the continuous rainfall event. Studies from the Mediterranean climate of Spain show observed changes in the low order streams to be deepening by scour of alluvium and widening by bank erosion (Gallart & Clotet-Perarnau, 1988). This appears to loosely match the scenario present at Wuppertal. Bank erosion with deposition of alluvial bars can result in channel widening by a significant amount. Rare events thus, seem to be of chief importance in the transport of coarse alluvium and in shaping valley bottoms within mountainous areas such as the Cederberg.

Drainage density is broadly correlated with mean annual precipitation. High values usually occur in semi-arid areas where the range of drainage density is also greatest. The proportion of precipitation receipt, which flows as surface runoff and is therefore immediately available for erosion, is greater in these regions than more humid environments. Drainage density decreases in more arid and more humid areas because of reduced runoff potential and the impeding effects of vegetation respectively (Knighton, 1996) whereas drainage density increases during wet spells as the permanent network expands to incorporate intermittent channels (Briggs, 1998). This is clearly evident when considering the reduced drainage density as one travels eastward within the catchments of all three trunk rivers, especially along the Tra-Tra River, towards the more arid conditions of the Tankwa Karoo. Knighton (1996) also points out how there is no coincidence that this global pattern of drainage density variation is similar to that for sediment yield; as high sediment yields reflect increased channel development and a more efficient drainage system. Bifurcation ratios of streams is an important control over the 'peakedness' of the runoff hydrograph (Chorley, 1973). The bifurcation ratio is a useful indicator but essentially represents merely an empirical relationship, which is too insensitive to variations in physical controls and therefore tends to obscure fundamental properties of the drainage network (Knighton, 1996). There is a relative importance to maritime and continental influencing climate factors in the Cederberg region as coastal lows and particularly significant frontal systems have the potential to deliver a vast amount of precipitation to the upper reaches and more westerly drainage of both the Moordenaarsgat and Dassieboskloof rivers, particularly of the Dassieboskloof river as its catchment extends further westwards into the plateau of the Cederberg mountains. It is also essential to consider the effects of snow and meltwater on changes in the volume and runoff within the basin. The Dassieboskloof River's westward extending catchment has the potential to experience not only higher levels of precipitation due its position incorporating some of the more significant central and northern peaks of the Cederberg plateau, but due to this, has a higher input or yield of sediment.

The presence of lateral flood channels at the base of the exposure containing cobble lags with some, but not well defined, imbrication suggests that there has been episodic flood flow and the poorly sorted contemporary fluvial sediment of the primary active channel indicates a contemporary environment that is subjected to fluctuating energy and discharge. This evidence is supported by the Moordenaarsgat and Tra-Tra rivers that exhibit well-defined flood lateral channels containing elevated flow channels, cobble and boulder lags, extra-channel deposits, such as slackwater deposits as well as the active channel component consisting of a significant presence of partly rounded boulders denoting occasional or episodic high magnitude flood events within the contemporary catchment. These may correlate with short-term intense events due to substantially large mid-latitude cyclones (cold fronts) and cut-off lows extend northwards and into the interior. The morphology and drainage of many fluvial systems were drastically altered during July 2006 and May/June 2007 when the Western Cape experienced abnormally large frontal systems that resulted in numerous floods and a large amount of damage across the province, most notably as the Olifants River along the western boundary of the Cederberg flooded.

Incised valley fills from rivers in the Kwazulu-Natal coastal region that exhibit successive cut and fill sequences, have shown single severe flood events to scour up to 6m (Ramsay & Cooper, 2002 citing Cooper *et al.*, 2001). If considering the extreme flood event as a potential for the incision and origin of the terrace, one must bear in mind the geomorphic consequences of continuous rainfall events in relation to geographic variables, such as altitude, topographical gradient, lithology and land use, and the spatial distribution of the rainfall. Contemporary evidence downstream and into the Tankwa basin shows that rivers have overbank flow, although relatively infrequently, as there are artificial, mud-walled impoundments, known as *saaipanne*, built by stock farmers to retain water during high flows (Smuts, 2007). There is also evidence provided by Smuts (2007) indicates more extreme infrequent rains within the region as local inhabitants of the western Tankwa Karoo vicinity had only experienced heavy rainfall as mentioned previously in June 2006 (South African Weather Service), once before in 25 years. An additional indicator of flooding in the area is the notable present of the man-made levee present on the opposite bank to the terrace, which protects vegetable fields from inundation. Its elevation, at close to two metres, suggests a maximum flood discharge in the region of two metres above the channel base level.

6.3.3. Vegetation and Fire Regime Changes

The large differences in erosion rates between vegetated and unvegetated slopes point to a major influence of the presence or absence of vegetation on surface water erosion (Knighton, 1996). If there is an overall change and removal of vegetation within the catchment, one would expect an increase in discharge as infiltration rates are lowered and direct runoff increases; there would also be an increase in coarse sediment normally stabilised on slopes by vegetation that makes its way into the channel, although it will also be moved more frequently as peak discharge increased. With these increases, an increase in channel width, wavelength and width-depth ratio would occur while sinuosity would decrease. Coarser sediment and higher vegetation covers than surrounding areas contribute to higher infiltration rates on beds but these depositional areas are also vulnerable to erosion, especially during periods of low rainfall and in areas where the protective vegetation cover is reduced (Bullard, 2004). Reduced vegetation cover increases the susceptibility of valleys to channel incision. Figure 6.1 below illustrates the theoretical relationship between precipitation, vegetation, hillslope and geomorphic work as well as demonstrating the net effects of factors during humid and arid phases, namely that there is an initial increase in sediment yield during a humid phase due to combined precipitation and vegetation factors, which decrease towards a slight increase and levelling off in the arid phase.

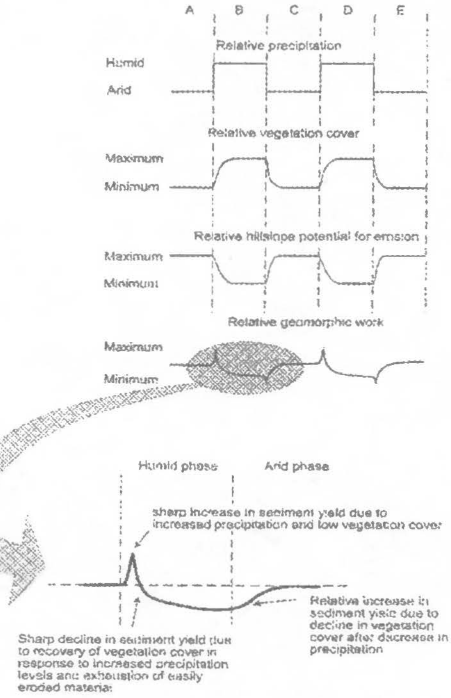


Figure 6.1: Sediment yield in terms of climate driven landscape response (after Knox, 1972 and Botha, 1992 in Holmes, 1998)

Taking into consideration the substantial impact vegetation has on sediment supply and runoff within a catchment, along with the fact that human populations began to migrate into and impact on the environment after 1000BP, there may have been changes in the fire regime on the surrounding slopes thus resulting in vegetation removal in places and hence, more sediment input into the system. Both plant cover and surface sealing is an important influence on runoff. Gabet (2003) (as cited by Bullard, 2005) found that substantial sediment erosion could take place on hillslopes following a fire, particularly where a hydrophobic layer has been deposited beneath the soil surface. As Meadows and Sugden (1991) outline, fire regimes in the Cederberg have changed over time, as indicated by evidence from the history of the Clanwilliam cedar and greater frequency of resprouting *Protea nitida* that suggests a possible increase in fire frequency; this would follow that such a factor as a shift in fire frequency and intensity during the last 1000 years may have played a role in impacting on sediment mobilisation and discharge of the Dassieboskloof both prior to, and after, human habitation of the Cederberg mountains. It is possible that fire frequency and intensity coupled with changes in vegetation and combined with a shift in climatic conditions could result in a shift in the dynamic of the fluvial system along the Dassieboskloof River.

6.3.4. Influence of a Small-Scale Alluvial Fan

Most rivers, such as Dassieboskloof River, flow in a single channel either because the river is forced to follow the thalweg of a narrow constricted valley, or because of the positive feedback that exists between erosion and channelised flow. It appears that both these factors could be at play along the Dassieboskloof River up until its confluence with the Tra-Tra River. For instance, once erosion is initiated by flowing water, the depressions so created capture more of the flow, leading to the concentration of stream power and encouraging further erosion that usually consumes the areas between the depressions, so that a single channel is eventually formed (Tooth & McCarthy, 2004). At the point of exit from the bedrock canyon, the Dassieboskloof River experiences an opening to a previously restricted channel and thus, a loss of energy. Although the contemporary active trunk channel consists of a single slightly incised and well-vegetated channel, secondary channels and flood channels, as indicated by cobble lags and vegetation patterns in the vicinity, indicate a splaying out of the system. This is similar to a micro-scale mountain front but valley constrained alluvial fan, especially as alluvial fans are characteristic landforms of arid and semi-arid mountain areas (Harvey, 1997). Although clearly not an alluvial fan *per se*, as it exits the bedrock canyon the channels (perennial and non-perennial) do exhibit a fanning out structure. Alluvial fans process dominance is by rare events and distal loss of discharge, particularly in semi-arid to arid regions. Associated with this feature are headward incising and fanhead trenching at the proximal end, often modified by contemporaneous tectonism or by climatically induced changes in sediment supply (Harvey, 1997).

6.3.5. Possible Base-Level Adjustment

A fall in base-level can be associated with an increase in channel sinuosity, and narrowing and incision of channels down the long profile of the river (Bullard, 2004). The Tra-Tra valley and Dassieboskloof River lies roughly 60km inland but substantially further if considering the distance includes the Olifants, Doring and Tra-Tra river extents. Its continental position to the lee of the Cederberg denotes that the location of the exposure at Wuppertal is outside the reach of sea-level changes resulting in base-level adjustment for the river system. If base-level adjustment was a contributing factor to the erosion of the terrace, one should notice incision working its way back up the numerous valleys from the distal point (working progressively upstream) towards the Dassieboskloof tributary. Although base-level adjustment can occur where adjacent rivers undergo deep incision, such incision would not be isolated in only the single tributary of the Tra-Tra, and would also manifest itself along the Moordenaarsgat River. There is also no evidence of knickpoints along the river profiles.

Network evolution is also time dependent as networks become better integrated through channel capture and progressive denudation of the landsurface, as well as representing an adjustment of drainage to the underlying geological structure. This network efficiency is related to the stage of development, just as channel flow is more efficient than overland flow, fully integrated networks drain catchments far more efficiently than immature networks (Briggs *et al.*, 1998). As channels extend and incise themselves, landscape mass is removed and the zone of maximum erosion tends to migrate headward as sediment is derived from progressively further upstream while downstream parts must continually adjust to this supply, sometimes in the form of pseudo-cyclic aggradation and incision (Knighton, 1996). The possibility of drainage extension and elongation within the drainage system by headward growth and tributary addition or capture is largely discounted as the time-scale does not allow for such rapid changes in catchment, usually associated with tectonic activity, and subsequent adjustment of the entire long profile of the stream.

6.3.6. Impoundment

An alternative theory resulting in the sedimentation and subsequent erosion at the mouth of the Dassieboskloof River canyon could be attributed to impoundment or damming within the valley. However, this mechanism can be discounted due to the absence of any evidence anywhere along the length of the valley indicating damming i.e. remnants of a significant feature that could have represented a dam wall, any rockfalls, debris flows or other mass movements large enough to sustain a complete blockage of the channel beds to the height of the terrace. For such deposition to occur, damming would have needed to have occurred relatively near to the upper reaches and upstream of the confluence with the Moordenaarsgat River. There is also no evidence of lacustrine environmental conditions within the vicinity of the site or a matching terrace on the opposite side of the valley or indication of a location to break through a dam wall. Even if small-scale impoundment did occur through the formation of channel blockages by tree trunks, organic litter, etc., the terrace elevation would not be as substantial. In addition, the nominal presence of a smaller terrace in the substantially wider Biedouw valley excludes this theory from consideration. A key consideration to this theory is that there would have been sedimentation further up the profile of the river instead of halting at the canyon mouth.

6.3.7. Channel Migration

There is no evidence of a defined secondary channel (aside from raised lateral flood channels) to suggest that the primary channel split, especially considering channel constrictions both above and below the exposure. However, there is the possibility that a palaeo-channel migrated across the narrow area within the constrained section as a result of energy, thus meandering to the southern bank and enhancing incision within the shifted dynamic.

6.3.8. Slumping

There is the possibility that rivulets draining off the saddle would result in rill and subsequent erosion down through both packages. An ephemeral river is marked on the 1:50 000 topographical map indicating a high likelihood of additional runoff from this southerly direction traversing the generally west-east running terrace, although there is little evidence of any obvious non-perennial channel across the alluvium package surface. Nevertheless, such flow may account for rill erosion within colluvium and the initiation, along with vegetation removal and human disturbance, of a *donga* or gully towards the distal margin of the terrace as well as influence the upper package in terms of slumping and small-scale slope failures.

Shallow fluid slides or debris flows are mass movements of less than 2 metres depth in surficial deposits and commonly spoon-shaped. The movement usually starts as a failure, promoted by high pore pressure, but there is little accumulation of material at the foot of the scar (only in small scattered piles and levees) that evidences a rapid liquid movement of mass down the hillslope. These flows can be slow or rapid (Gallart & Clotet-Perarnau, 1988). Using this principle, such small-scale slides may have occurred along the face of the terrace, particularly during heavy rainfall events. There is contemporary evidence of avalanching perhaps supporting this proposed more minor erosion. Interdigitated fingers of grey alluvium with orange colluvium laminations occur along the face of terrace, as seen in Figure 6.2. This may also reflect cyclical processes of small-scale slope failures along the face as well as indicate the influence of sheetwash and rill erosion.

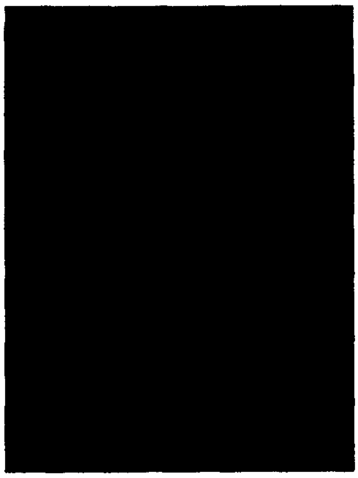


Figure 6.2: Interdigitated fingers of alluvium (grey sandy material) and colluvium (orange more consolidated clay-rich material)

6.3.9. Neotectonics

The relationship between tectonics and landforms is dialectic. Much of tectonic geomorphology is concerned with using evidence from landforms to infer rates of operation of tectonic processes; conversely some geomorphological studies invoke tectonics in order to explain landform evolution (Rendell, 1997). Epeirogeny refers to the uplift or depression of continental or sub continental landmasses as a result of widespread adjustments of level where movements are generally even in character producing little more than tilting, slight warping and minor faulting of rocks (Whitten & Brooks, 1972). Rates of uplift and fault patterns exert a major influence on both patterns of drainage and rates of incision in an area. Neotectonics has been shown to disrupt drainage over time, such as the influence of faulting in Namibia and outside Oudtshoorn (Marker, 2003). Surface uplift is often widely observed or revealed in regions of continental crust by river terraces spanning the Quaternary with the characteristic geomorphology being explained by flow forcing in weak lower crusts by surface processes such as non-steady-state erosion and cyclical loading by sea-level fluctuations (Westaway *et al.*, 2003). Due to problems associated with radiometric dating of sediment sequences in more arid regions, there are difficulties substantiating the length of records of sedimentation or erosion and justifying its relationship to its tectonic settings (neotectonics). The relative impact of neotectonics and Quaternary changes in climate on landform evolution are frequently difficult to distinguish (Berger, 1997). A tentative suggestion of neotectonics as an explanation has been largely discounted due to a lack in tectonic evidence for the area. Neotectonics within the region, which could have attributed to initial fill and subsequent cut of the terrace through tectonic uplift and an adjustment to a change in relative base-level cannot be conclusively proven as the dominant factor in the geomorphological feature.

Morphotectonic aspects of the southwestern African continental margin, and including the Karoo and Cape region, have been investigated by Spoenemann and Hagedorn (2000). According to their study, the landforms of the continental margin are of polygenetic origin, induced by tectonics and isostasy and sculptured under different climatic conditions. Considering the tectonics of the entire region within the locality of the Cederberg, the surface vergences of the Namaqualand Escarpment and of the Piketberg can be correlated with the Great Escarpment as monoclinal flexures; along the eastern margin of the Tankwa Karoo (the latter one is interpreted as a counterpart of the Doringrivier Flexure and both of them are attributed to the Tankwa Basin). Tectonics take on strong importance factor in the long-term, however, in the short-term other factors become more important in controlling fluvial dynamics and development. As exhibited by the Moordenaarsgat River, rivers tend to erode along a plane of weakness and according to gradients downslope, such as faults, conformities, etc. There are, however, no signs of faulting across or along the Dassieboskloof River catchment while conformities are mostly horizontal or slightly tilted and the river has exploited these to create bedrock canyons. When inferring neotectonic influences, resolution of timescales is critical. Although there are several normal faults crossing the Tra-Tra Valley, there is no visible evidence pointing towards any tectonic activity in the recent past (subsequent to the Tertiary). There is also a distinct lack of any obvious knickpoints along the long profile of the lower reaches of the Dassieboskloof River or within the Tra-Tra River valley itself. The last significant stage of regional tectonism dates from the Pliocene (Cornell, 2001) and as Baxter (1996) and Lancaster (1996) state, the west coast region of southern Africa is regarded as tectonically stable. In addition, if the effects of more localised fault activity had resulted in such an erosional feature as the terrace, there would not be evidence of a similarly corresponding terrace in the Biedouw valley to the north. Even considering such a hypothesis as stream capture at the headwaters of the Dassieboskloof River due to neotectonics and accounting for a suddenly greater volume of discharge down its course, must be disregarded as there is again a lack of any evidence of tectonism during the late Holocene to allow for such rapid changes in drainage and headward extension of the catchment.

6.3.10. Further Erosional Processes

Other discounted processes include gullying and groundwater seepage erosion, although gullying may take place relatively rapidly and does occur in the area, as seen towards the distal end of the terrace; especially as soil types associated with the region are readily erodable and there are many examples of deep gully and sheet erosion that has occurred in historical times due to overgrazing of the originally sparse vegetation cover. The generally low kinetic energy of rainfall in this region, however, slows the rate of erosion (Rutherford & Westfall, 1986). There is no match to the gully walls and the bank width is too expansive. The amount of surface-flow influences the slope morphology and hence the operation of seepage erosion with conflicting indications that wetter periods would lead to increased groundwater discharge and hence increased erosion, or lead to less erosion as increased outflow hinders accumulation of minerals and salts which act as heave mechanisms (Nash, 1997). Groundwater seepage erosion has been dispelled as the process is too lengthy and the timeframe under consideration is too short and there is an insufficient source in the immediate vicinity of the site. There is also no indication of piping to suggest groundwater movement (Campbell, 1997). With regard to the confluence of the Dassieboskloof River with an unnamed tributary from the north at Wuppertal and the Moordenaarsgat River to the east of Wuppertal, their influence has been eliminated as they are too spatially distinct to have had an impact.

6.3.11. Conclusion

Using the principle, according to Bell and Walker (1996) that most erosion does not take place imperceptibly slowly but is concentrated during specific observable events and effects depend on their recurrence, the terrace formation is attributed to high magnitude low frequency events (during roughly the last 1000 years). From geomorphic evidence, the fluvial terrace along the Dassieboskloof River at Wuppertal exhibits a cut-and-fill sequence and from contemporary conditions, remains in the "cut" dynamic of the cycle with the "fill" being preserved as the terrace obvious today. Aggradational terraces, such as the primary site, formed by cut and fill cycles, are indicative of periods of overall dissection (Harvey, 1997); while threshold conditions between cutting and filling modes of stream behaviour operate within very narrow margins (Campbell, 1997). Erosion may be part of positive feedback mechanism (as opposed to negative). In a negative feedback system erosion could have reduced the slope thus enhancing later deposition, whereas a positive mechanism sees erosion increasing unit stream power which enhances further erosion, however an increase in unit power is usually compensated for by rapid increase in channel width, thus accounting for likely lateral erosion of the channel bank. A climatically induced reduction in sediment supply could also result in increased incision. In a shorter-term, climatically induced sediment supply and its effect on fluvial processes has an overwhelming influence on dynamics.

The incision of the terrace is hypothesised to be the result of a combination of factors, namely an increase in high magnitude low frequency events coupled with a change in the fire regime and subtle shifts in vegetation patterns, while more local channel morphology may have led to channel migration and subsequent incising resulted in small-scale slumping or bank caving of the partially consolidated alluvium and weathering of the colluvium.

6.4. PALAEOCLIMATIC INTERPRETATION OF THE GEOMORPHIC EVIDENCE

Rapid warming episodes followed by gradual cooling, associated with sub-Milankovich cycles, implies an asymmetry in the behaviour of climate that is likely to be reflected in landscape responses (Thomas, 2004). Slope failures, floods and colluvial/alluvial sedimentation may reflect short-term changes in the record, but reorganisation of slope and fluvial systems involve significant time lags or delays, often on a millennial scale, and require changes in vegetation cover. While some records of sedimentation indicate major landscape instability during the last 20 000 years, or especially in the early Holocene, others indicate pulses of activity throughout the last glacial cycle. These differences may reflect regional patterns of climate change, but also illustrate the importance of landscape sensitivity to our understanding of the impacts of rapid environmental change (Thomas, 2004).

River channels and alluvial fans are common features in many mountainous dryland regions, and their landforms and sediments have been widely used for palaeoenvironmental reconstruction. In many cases, however, there is considerable debate as to the nature of the palaeoenvironmental signal recorded, such as whether phases of erosion and sedimentation are primarily controlled by tectonic, climatic or land-use change and, if climatically-controlled, whether these phases reflect conditions of relative aridity or relative humidity. However, the relative tectonic stability of the African continent results in the dominance of climatic effects on the behaviour fluvial systems (Lancaster, 1996). Within southern African arid environments there are abundant cut-and-fill sequences and river terraces as well as extensive evidence for the expansion of integrated drainage systems during periods of humid climates. As previously mentioned, the terrace present along the Dassieboskloof River displays both periods of sediment accumulation, in the form of colluvium overlain by alluvium, as well as a period of incision or erosion, possibly both prior and subsequent to its deposition. This pattern of underlying geomorphic change is also evident along the higher order Moordenaarsgat and Tra-Tra Rivers. The interaction of wind and water over time is considered as one of the key influences upon contemporary geomorphology and longer-term landscape development in drylands of the world with the dynamics of interaction and flow as well as the type of system and its morphology being of great significance (Nash, 2000). Tributaries, such as the Dassieboskloof River, provide links between lithology and climate and are adjusted to both. Channel characteristics vary in response to the external variables of sediment and water discharge, which are influenced naturally by climate, tectonic and lithologic factors; human influence also modifies these variables through land-use alterations (Ritter, 1995).

The northeast region of the Cederberg and transitional boundary margin between the Cederberg Mountains, to the west, and the Tankwa Karoo basin to the east, is considered presently as semi-arid, although a decreasing precipitation gradient exists towards the east, where it becomes more arid. Palaeoenvironmental interpretation aims to assess whether there has been any changes or fluctuations within this semi-arid environment towards either periods of increased or decreased moisture availability. The region under investigation is, ideally located, along a dryland boundary, to perhaps accentuate any environmental shifts that may have occurred and preserve such conditions within geomorphological features of the fluvial system. The sediment yield and fluvial erosion are greatest in climates with seasonal aridity, including climates classified as semi-arid to sub-humid (Derbyshire *et al.*, 1981). Although the Late Quaternary environmental changes appear to have been of insufficient magnitude to produce any marked shifts in vegetation patterns within the Cederberg (Meadows & Sugden, 1993), it is possible that such subtle fluctuations could have impacted more readily with a more immediate response time on geomorphological environments, depending on the proximity of the palaeoenvironment to its threshold.

River systems within the Namib exhibit sequences of fluvial deposits that indicate periods of aggradation and incision. An interpretation of the deposits (offered by Heine and Heine (2002)) suggests they are slackwater deposits reflect extreme flood events in the Kuiseb River Valley during wetter climate phases rather than a record of more arid conditions as had previously been thought (Bullard, 2004). Such flood evidence as the aforementioned, however, does essentially indicate an important dynamic, such as a shift towards or away from arid conditions during the Holocene to more humid conditions, as well as suggesting that extreme flash floods occurred more frequently during the Little Ice Age. This evidence is key to assisting in elucidating palaeoclimatic interpretation of the terrace at Wuppertal and surrounding study area.

Global correlations of time periods characterised by large floods are noted from northern and southern hemisphere rivers. A late 19th century flood period is recognised from the Mississippi (and some of the southwestern Rivers of the USA), the River Dee (Scotland) and the Orange River (South Africa). The period of the Little Ice Age can clearly be recognised from the flood record of several rivers and some indicate a double pulse (900-1200AD; 1300-1600AD), which straddles the Medieval Warm Epoch. Although data becomes less reliable and telescoped with time, the global flood record appears to show further flood periods between 3000-1800BP, 4000-5000BP, and possibly 7000-9000BP and it is likely that the Younger Dryas was also a flood period (Smith, 1992). Alluvial records of palaeofloods show that natural floods resulting from excessive rainfall, snowmelt, or from combined rainfall and snowmelt, are highly sensitive to even modest changes of climate equivalent or smaller than changes expected from potential future global warming in the 21st century. Holocene palaeoflood chronologies from the Upper Mississippi Valley and Colorado River drainage show that recurrence frequencies of large floods have been subject to abrupt changes over time. Flood chronologies, observed for other mid-latitude regions, suggest that recurrence frequencies of large floods are increased when there is an increase in the number of waves and their amplitudes in the middle and upper tropospheric circumpolar westerly circulation. However, some mid-latitude regions on the western margins of continents experience increased frequencies of flooding during strong onshore zonal westerly circulation. Flood chronologies from several regions suggest that times of rapid climate change have a tendency to be associated with more frequent occurrences of large and extreme floods.

The conclusion that flood evidence reflects climatic shifts corresponds and agrees with evidence from elsewhere in southern Africa, as seen in the Orange River. Palaeoflood hydrology from the Orange River, specifically the lower Orange River valley, as reported by Zawada (1996), also points to increased flooding within the river catchment at the time of the Little Ice Age with evidence suggesting high magnitude palaeoflood events. Amongst the several palaeoflood events identified, a maximum flood was recognised as well (Boshoff *et al.*, 1993). The Orange River palaeoflood record correlates with dry cool phases in the late Holocene record from 300–0BC and a brief cool, dry interval in the mid-fifth century (Zawada *et al.*, 1996). The palaeohydrology of Nahal Zin, a catchment in the Negev Desert, inferred from slackwater deposits and palaeostage indicators, also shows clustering of floods around 1000BP and again during the last 60 years characterized by high flow magnitudes (Greenbaum *et al.*, 2000). Therefore, it appears that deposition of alluvium, such as the study site, suggests flood conditions, relatively medium frequency and high magnitude, however, these factors have not been quantified but rather inferred from sedimentological and geochemical evidence. In contrast, geochemical and palynological results from Two Mile Lake in southwestern Australia, which shares a similar environment to the study site, indicates that the vegetation and environment were unresponsive or unaffected by early Holocene climatic changes (Itzstein-Davey, 2004). A difficulty associated with this particular study also involved sedimentation rates, as evidence could be interpreted as either rapid sedimentation during the Holocene or suggests a lack of environmental change as a result of complex environmental processes. The accumulation and preservation of slackwater sediments appears to be controlled by tributary-mainstream junction morphology and tributary basin drainage efficiency; significant morphological changes occur during large floods, but only if there has been adequate recovery time between floods for the channels to readjust to more frequent low magnitude flow conditions (Kochel, 1988). This fact is of key importance to the study site as obviously there has been sufficient time within the channel to for recovery and readjustment with intermittent successive floods leading to consistent slackwater deposition on the southern bank of the Dassieboskloof River resulting in the build-up of such a terrace.

Important recent advances have been made in the reconstruction and interpretation of ancient floods, particularly in the use of SWD and palaeostage indicators. For certain appropriate geomorphic settings, relatively accurate estimates of palaeoflood discharges and ages can be made over time scales of centuries and millennia particularly with regard to flood frequency in the drainage system; hydroclimatic change is a likely cause of such dynamics (Baker, 1987). Clusters of periods with many floods correspond well to periods with probably relatively wetter conditions, while periods with few floods indicate a drier climate, contrary to prior evidence. Fluctuations in the frequency of floods are typical of periods of transition from one climate regime to another (Greenbaum *et al.*, 2000).

River and alluvial-fan terraces may form by a variety of climatic, tectonic, and autogenic mechanisms. Models for small- to large-scale alluvial fans, as demonstrated by Pelletier (2004), show transient variations in drainage density as responsible for the largest fluctuations in sediment supply to the fan, and hence, the dominant control on the occurrence of cut-and-fill cycles on the fan. Alluvial fan terraces created by the model, firstly correlate with humid-to-arid transitions, secondly, have volumes and dips that correlate with the duration of the previous humid interval, and thirdly, are areally preserved with an inverse relationship to the duration of the subsequent dry interval. These observations provide a preliminary blueprint for correlating regional patterns in alluvial-fan deposits with Quaternary climatic changes (Pelletier, 2004).

Arid conditions favour channel aggradation. Reduced rainfall often accompanied by marked seasonality means diminished stream power and the clogging up of channels with relatively coarse sediment, derived from slopes because of reduced vegetation cover; conversely, wetter conditions favour incision and terrace development (Thomas, 1997). An increase in storm magnitude and/or frequency would result in a stripping of the regolith from the bedrock, before vegetation cover could be established. Erosion of the regolith cover on the hillslopes and a subsequent establishment of vegetation cover may result first in an increase and then in a decrease of the downslope flux of sediment. The response of the trunk stream is sediment aggradation followed by incision and terrace formation (Steffen *et al.*, 2007). An increase in mean annual precipitation above the threshold results in headwater sediment accumulation, a subsequent decrease results in a flushing of the accumulated sediment as runoff increases as a consequence of reduced vegetation cover. Cooler and drier climates lead to erosion of sediments due to reduced vegetation cover, particularly applicable to semi-arid landscapes (Meadows, 1988). It is important when considering fluvial erosion to determine the recurrence interval of big floods, i.e. time-scale is vital as the threshold events are not bankfull rivers but flooding rivers with inundated high terraces or inundated braided river bottoms. Particularly wet years can result in high levels of material being transported out of upper basins and accumulating on valley bottoms. A large amount of this material can come from landslides in the headwater regions and is transported only during flood events (Ergenzinger, 1988).

Slopes, under semi-arid to arid conditions, when vegetation cover is sparse, have sediment transport that may be more active on the slopes than channels. This favours the accumulation of thick, lower-slope colluvium aprons leading to interpretations of more widespread semi-arid conditions within the region. Gravels and diamicton are indicative of colder more arid conditions (as a result of instability from k-cycles) (Marker, 1995). The transition from colluvium to alluvium deposition is abrupt and there are no buried soils or palaesols evident within the sequence. The stone lag on top of the colluvium along the contact below the alluvium is indicative of a weathering surface. Changes in the amount of moisture available have occurred through time since deposition of the colluvium, as calcrete within the colluvium provides evidence of water table fluctuations. Calcretes are important features of drylands ranging from powdery to massive and are produced by pedogenic and non-pedogenic processes that cause the dissolution, mobilisation and precipitation of carbonates in the near-surface environment (Shaw & Goudie, 2004). Soil forming processes are present within the deposit along with *in situ* weathering. Mud drapes are laid down in the last stages of flood flow (Reid & Frostick, 1997) therefore their presence along the face of the colluvium indicates such flood events.

No other packages of alluvium overlying colluvium are evident within the nearby catchment or anywhere along the long profile considered within the study area, therefore the only possible correlation or corresponding/comparable deposit occurs in the Biedouw valley, however, it was not considered in detail during this study. The presence, however, of this deposit, albeit disturbed by farming along the Biedouw River, suggests a more regional geomorphic response to some changing dynamic within the greater system. This could be interpreted as identifying a link between the two adjacent, eastward draining basins of the Tra-Tra and Biedouw rivers that constitute a substantial area of drainage systems within this marginal northeastern region of the Cederberg Mountains on the fringe with the Karoo. Based on the type of geomorphic evidence investigated during this study, hypothetically the lack of any conspicuous valley fills across the region, in particular to the arid east of the study area as well as only a smaller exposure present to the north in Biedouw in itself is a possible indicator of already more arid conditions prevailing or developing more noticeably during the Holocene.

Reduction in the competence of streams is usually associated with a decrease in precipitation, as shown in the Namib Desert in streams from the western Escarpment crossing the desert that experienced a decrease in rainfall between 8000 and 4000BP, as well as a similar reduction in the southern Karoo (although a modest increase occurred in the coastal areas of the southern Cape) (Partridge *et al.*, 1999 quoting Vogel, 1987). Drier conditions were probably a result of the reduced influence of the westerlies as temperatures increased and the circumpolar vortex contracted. This may concur, especially considering the site's proximity to the Karoo, that the basin's streams also experienced a relative reduction in stream power during the early Holocene as a result of decreased precipitation and more arid conditions.

During the accumulation of alluvium subsequent to approximately 5000BP, a lack of significant organic material present within any particular beds suggests, even if a wetter more seasonal environment prevailed, that the fluvial system was more energetic and did not allow for deposition of finer organics amongst the coarser nature of the sediment, although charcoal fragments, a result of biomass burning of trees and reeds (*restios*), were entrained and deposited across most of the units. This reflects consistent fire regimes during the period of sediment aggradation. The presence of charcoal fragments within a number of the horizons valuable towards noting possible relationships between natural fires leading to decreased vegetation and increased runoff and hence, sedimentation. It is difficult to use charcoal as a means of evidence for inferring fire regime within the catchment, however, due to the fact that it could not only have originated from elsewhere in the catchment by smaller scale fires as opposed to significantly more widespread burning of vegetation across the upper reaches of the Dassieboskloof catchment within the Cederberg plateau, but it may also have undergone previous cycles of transport and deposition and been recycled previously through the system. Localised erosion could lead to entrainment of previously buried charcoal and its subsequent inclusion into the channel through runoff thus ultimately resulting in its deposition within the terrace units. This would present a false impression of burning in the region during that particular phase of the cycle, or by exaggerating the amount of burnt biomass resulting in a misinterpretation of the intensity, frequency and extent of the fire. The arrival of farmers and controlled fires in the area also reduces the incidence and extent of grasses and causes removal of fuel for grazing, farmland, etc. and accounts for low microscopic charcoal counts.

Increased moisture availability during the depositional period of the terrace (~5000-1000BP) may have led to denser riparian vegetation and increased litter within the channel, which may have in turn promoted overbank flow by choking the channel in places. The consideration that adjustments and reactions within the Dassieboskloof fluvial system may only be local in nature was taken into account as the response may not be a widespread shift as result of climate change, although extension of the sites or examples northwards like Lake Bruno and the Biedouw valley would eliminate this possibility.

According to Joyal (2003), the transition to the altithermal, ~5000BP, may have generated disturbances suitable to manifest as a regional response evident in the fluvial systems, such as the Dassieboskloof catchment. Intrinsic thresholds responding to weaker climatic fluctuations and/or changes in fire frequency and vegetation could be accountable for a shift in localised conditions resulting in a lack of stratigraphic continuity across the study area and throughout the northeastern region. A dynamic equilibrium may have been reached during deposition where the equilibrium shifted with long-term changes while in the shorter-term, fluctuations occurred around this equilibrium, evident as flooding the overbank area in episodic or seasonal periods. The intrinsic threshold of the deposition threshold was breached by an extrinsic force, such as a shorter-term notable climate shift, resulting in the dynamic changing towards erosion instead of deposition. This dynamic is being maintained into the present-day fluvial environment of the Dassieboskloof River. Overall the system, including the erosional force, represents a dynamic metastable equilibrium (theoretically demonstrated graphically in Figure 6.3 below) where long-term trends are separated by a threshold to a new level (Bell & Walker, 1996).

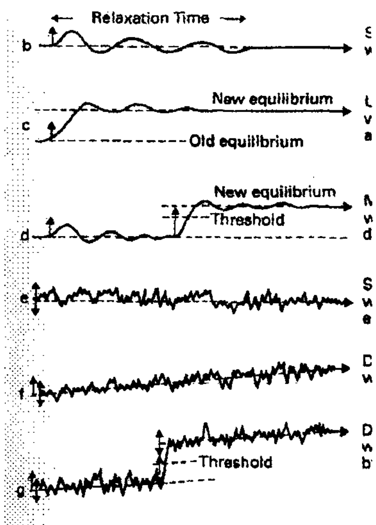


Figure 6.3: Diagrammatic representation of various forms of environmental change and equilibria (Bell & Walker, 1996)

The palaeoenvironmental interpretation includes the colluvium, although no chronology was established for the deposit. A small colluvial fan developed during a more arid phase of the early to mid-Holocene and constricted the active channel of the Dassieboskloof River during which time fluvial deposits were flushed out of the vicinity of the terrace area by larger magnitude flood events. In turn, these events may have begun to scarp the toe of the colluvial fan and lead to a truncated colluvial fan, exemplified in a contemporary setting at the valley constriction mid-way along the Moordenaarsgat valley. A shift to more consistent, perhaps, seasonal flooding and higher flow regime during the mid-Holocene, just prior to 5270 ± 0.69ka BP, along with possible palaeochannel migration, may have resulted in the initiation of raised or elevated flood channels consisting of cobble and pebble lags infilled by coarse sand overlying the colluvium. As the more energetic palaeochannel migrated away from the southern bank deposition became more indicative of regular, or more consistent in frequency and magnitude, flood events and resembled coarser grained slackwater deposits. Successive flood events resulted in the cyclical accumulation of sand. Increased terrace relief coupled with continued episodic events accounts for the further accretion of more organic-rich sediment to its maximum elevation around 960BP. Subsequent to this a single continued dominant phase of late Holocene incision is apparent in the valley fill, which manifested as the terrace along the Dassieboskloof River.

From this brief proposal of palaeoenvironmental conditions, it appears that the geomorphic evidence of the study site concurs with other forms of proxy evidence from the northern

Cederberg, such as Lake Bruno (Cornell, 2001), and Pakhuis Pass (Scott, 1994; Scott and Vogel, 2000). During the interpretation of the evidence in conjunction with literature, however, it was necessary to bear in mind the fact that evidence from the northern Cederberg did not show any definitive increase in moisture availability or significant changes in vegetation through the Holocene. Rather it is inferred that subtle fluctuations within the Holocene were too rapid to be well established and reflected in vegetation but instead showed better in other proxy evidence (Scott, 1994). Therefore, it is beneficial to rather compare the site to Lake Bruno and Pakhuis Pass sections as they are more likely to reveal climatic shifts or signals preserved in sediments as a direct response to changes in catchment. These sites are also better correlated to the Dassieboskloof site due to their proximity and location on the fringe of the Cederberg Mountains, as opposed to central or upland/plateau sites used in other palynological vlei and Hyrax midden evidence. The geomorphic evidence of the study site suggests colluvial deposition occurred during a more arid early Holocene. The advent of the altithermal, approximately 6000BP, influenced the fluvial system to shift its dynamic towards aggradation controlled by palaeofloods that transpired from the mid- to late Holocene (as dated by OSL to reflect time of deposition). This period of the Holocene is mostly characterised as experiencing increased moisture availability. Avery (1990) notably hypothesises the possibility of an increase in winter rainfall on the boundary between the fynbos and succulent biomes; this is a possible indication of increased rainfall in the catchment with increased flow and sediment. Moister conditions that generally prevailed in the late Holocene would have favoured run-off in the upper reaches of a catchment, such as at the conditions that existed at Verlorenvlei, which brought down higher quantities of finer grained material from the upper catchment, pointing towards increased river capacity due to higher rainfall conditions (Meadows & Asmal, 1996).

There is, however, conflict with the palaeoclimatic interpretation for the late Holocene. Well-sorted sandy sediment indicates greater sediment mobility and drier conditions (such as Lake Bruno in late Holocene) in contrast to the more obvious presence of lags suggesting moister conditions (Cornell, 2001). This would suggest that subsequent to the shift towards aggradation, climate was initially less arid and more moisture was available, however, a shift towards more well-sorted medium textured thinner stratified sand beds and laminations in the upper units of the alluvium may reflect a shift in the direction of less moisture availability and more arid conditions. Widespread organic-rich sediments are unusual in the contemporary environments of the semi-arid Cederberg, thus the noticeable lack of any significantly organic-rich beds present within the alluvium, especially towards the upper surface, supports the possibility of increasingly arid conditions within the catchment characterised by more frequent flood events, hence the rapid sedimentation of the upper unit of the alluvium package. Alternatively, this rapid aggradation may be the result of increased moisture availability as suggested by Scott (1994) *circa* 1000BP.

Scott and Vogel (2000) suggest cooler winter rain conditions in the Pakhuis Pass coinciding with medieval warming whereas evidence from Elands Bay, to the west, suggests more arid conditions for last 1000 years. Around this time there was another shift in the dynamic of the Dassieboskloof fluvial system as a period of incision of initiated. The probable chronology of this switch corresponds to the medieval warming period, which was associated with not only warmer temperatures but a decrease in precipitation. This would impact on the flow-regime and palaeoenvironment of the study area and may have manifested in a succession of more high magnitude low frequency extreme flood events, as opposed to seasonal flooding, coupled with increased aridification to present-day. This is supported by Scott and Lee-Thorp (2004) who indicate that conditions of the Little Ice Age in South Africa culminated around 1750AD with conditions that were drier and stormier with associated temperature fluctuations (becoming cooler). There were also shifts in the intensity and extent of frontal systems associated with the Antarctic circumpolar vortex, therefore there is the additional possibility of stronger and larger frontal systems at points in the late Holocene, which high magnitude palaeofloods could be attributed to. The hypothesis of large palaeofloods is also supported by the contemporary morphology of the Moordenaarsgat and Tra-Tra rivers, being braided and anabranching respectively. These channel patterns are indicative of transitional climatic phases as well as being indicative of highly variable flood-prone flow regimes, as reflected by the presence of elevated flood channels containing significant cobble lags, cobble and boulder elements in active channels and slackwater deposits. It appears that the proposed palaeofloods were larger than any recent historic floods as there is no indication in any literature of recent or historically mentioned pertinent high magnitude events in the region.

There are numerous limitations and controls present within this interpretation of palaeoclimatic conditions through the mid- to late Holocene in the region of the northeastern Cederberg. One of the main limitations is the lack of any high resolution chronologies of alternative proxy evidence in the region as well as the low resolution proposed by this study itself, where only four sediment ages were established. Although the palaeointerpretation is largely consistent with Scott (1994) and Scott and Vogel (2000), their investigation is again characterised by poor chronology thus demonstrating a more generalised inference of palaeoclimatic conditions. A more detailed study of the Quaternary valley fill stratigraphy with improved chronological resolution would show that it does not simply reflect a response to changing climate (Holmes, 1998 citing Prosser *et al.*, 1994).

As climate is recognised as a controlling factor on sediment supply in streams, it is tempting to assume a "simple causal relationship" (Baxter, 1996 as quoted by Holmes, 1998). However, it is imperative to factor in the impact of human activity that also impinges on sediment supply, especially in terms of rural land-use and agriculture or stock-farming practices, such as fencing, point water source installation, boreholes and damming, let alone the introduction of non-indigenous stock such as sheep, goats and donkeys specifically in the Wuppertal vicinity of the Cederberg impacts greatly on natural vegetation which in turn indirectly influences sediment supply. Even within the more general Cederberg region, the initial harvesting of wild rooibos followed by the introduction of rooibos cultivation and even centre-pivot irrigation systems in some places, such as the Biedouw valley, has an effect on vegetation patterns as well as runoff patterns within the respective catchments. This is not to mention the incremental increase of the small communities' populations, which is intensified by the influx and transient movement of tourists and visitors across the Cederberg Mountains, coupled with increasing demands for fuel and water. Under these circumstances, although the majority of the northern Cederberg has not undergone substantial development, humans have still had some kind of influence on the environment, therefore when considering inferences from the available geomorphic evidence, it is necessary to employ some caution whether climate or human activity has dominated environmental change over the more recent past as the signals are far from clear and present difficulties in discerning between the two.

6.5. CONCLUSION

Reconstructing former climatic conditions on the basis of proxy records from geological or biological sources is a far from straightforward process. When considering palaeoenvironmental interpretation of evidence it is important to bear in mind that there are often multiple catalysts or scenarios that can result in the same or similar outcome. This fact is often overlooked in preference to rather suggest with certainty and assumed clarity that one factor was responsible, however, this leads to misconceptions and over simplification of situations. Instead, although more difficult to fully explain and thoroughly incorporate into research, a potential unbiased multivariate explanation is preferential to include and take into account elements liable for change. For this reason it is perhaps more ideal to approach such a study as a complex system comprised of interrelated factors and constituent feedback mechanisms which impact, alter and ultimately result in change. Therefore, it is not possible to single out one particular factor responsible for the overall change, but rather that that factor behaves as part of a system and catalyst to other factors that imparts a signal of environmental change on evidence.

With respect to geomorphic palaeoenvironmental indicators within the southwestern Cape region, one cannot know if localised environment reacted to changes or shifts in a spatially or chronologically consistent manner, especially in light of sub-regional variations. This chapter, however, has aimed to elucidate the nature and context of firstly, the contemporary fluvial environment across the Moordenaarsgat, Dassieboskloof and Tra-Tra River valleys within the study area, secondly, the depositional and erosional environmental setting of the primary site at Wuppertal, and thirdly, to use this geomorphic evidence to explore the palaeoenvironmental conditions for the marginal northeastern region of the Cederberg.

Chapter 7
FURTHER DISCUSSION AND CONCLUSION

7.1. INTRODUCTION

This chapter considers the findings of the research and integrates the aims and objectives of the study with its significance and conclusions along with the potential future implications. It also considers the broader context of the study in terms of the Cederberg region in particular and arid environments of the southwestern Cape in general.

7.2. OVERVIEW

Chapters 1 and 2 introduced both the realm of this study's investigation, aims and objectives as well as the study site, an area in the vicinity of Wuppertal in the northeastern Cederberg, in detail with regard to pertinent physical factors. The literature pertaining to possible palaeoenvironmental proxy data for the southwestern Cape, and specifically the Cederberg region, reviewed in Chapter 3, has revealed a distinct pattern of change in the spatial and temporal characteristics of the winter rainfall region associated with various forcing factors during the Late Quaternary. Geomorphic evidence has been shown to preserve and reflect such palaeoenvironmental changes that have occurred through numerous geomorphic features, landforms and processes, thus substantiating the use and applicability of such geomorphic evidence within the context of the study area.

Chapter 4 outlined the methodology and analytical procedures adhered to for the duration of this study as well as considering limitations and constraints inherent to the employed exploratory techniques. The methodological approach was also justified in terms of the relevance to elucidating meaningful results and data from the evidence for subsequent interpretation. Chapter 5 presented objectively the results of the analyses.

Chapter 6 proposed, through a study of the temporal and spatial extent of geomorphic evidence recorded within the Late Quaternary sediments at Wuppertal, hypothesised scenarios of deposition and erosion for the primary study site and placed the exposure within a contemporary context with relation to the study area as a whole. From this, palaeoenvironmental conditions were inferred to suggest possible changes or shifts in significant forcing factors responsible for such change, whether it was climate-driven, thereby implying a natural cyclicity or progression, or largely anthropogenic, and subsequently extend this theory to incorporate the wider northeastern region of the Cederberg Mountains and more arid margin of the southwestern Cape.

7.3. SUMMARY AND SYNTHESIS OF RESULTS

The first objective of this study has been to evaluate Late Quaternary environmental change in the southwestern Cape from geomorphic evidence present in the northeastern Cederberg, drawing on a multidisciplinary approach using sedimentological, geochemical and geomorphological techniques. The investigation was placed within a context for the palaeoenvironmental conditions of the southwestern Cape based on proxy evidence, specifically for the Holocene, and related to geomorphological response (Chapter 3). Overall, the climatic changes of the winter and summer rainfall regions of South Africa are 180° out of phase (Meadows & Baxter, 1999) and further studies indicate that the winter rainfall region and its inherent climatic features have fluctuated spatially, as outlined in chapter 3. Therefore, it is beneficial to examine more closely the winter rainfall region in order to compare its past conditions to the current thereby establishing a better understanding of future climatic changes. Furthermore, collecting and combining data from a wide range of proxy evidence, as well as modelled projections, to create a more comprehensive synthesis of palaeoenvironmental change would be more useful towards past and future inferences. It is considered a more reliable method of ascertaining an overall synthesis of past environmental and climatic changes (Mulock-Houwer, 2001).

Subsequent objectives involved the utilisation of a range of methodologies and techniques to examine and assess the morphology and lithostratigraphy of the cut and fill sequence in the Dassieboskloof River valley and the contemporary fluvial dynamics of the Moordenaarsgat and Tra-Tra Rivers, as well as determine a chronology for the exposure at the primary site. Data revealed an episodically high to medium energy sandy environment typical of the middle to upper reaches of a semi-arid fluvial system, which has undergone, during the recent past, fluctuations in flow regime and sediment input due to variable factors. A coarse-scale palaeoenvironmental interpretation has been undertaken on the basis of the geomorphic evidence and palaeodeposit of the marginal northeastern Cederberg to infer hypothesised climatic fluctuations and possible anthropogenic influences along with additional secondary factors associated with the complex interplay of features within the semi-arid feedback system. This proposition of environmental change and fluctuation has been extended to consider its relevance, significance and consistency within the greater winter-rainfall region of the southwestern Cape. The geomorphic evidence investigated during this study suggests a palaeoclimatic interpretation and shift in palaeoenvironmental conditions within the northeastern Cederberg consistent with literature from the region, namely Scott (1994), Meadows and Holmes (2000), (Cornell, 2001), Scott and Vogel (2000) and Scott and Woodborne (2007). Early Holocene conditions are inferred as more arid with the altithermal inducing fluctuations within the fluvial systems and the aggradation of alluvium, which mantles the colluvium below, along the Dassieboskloof River, at approximately 5 500BP. More seasonal moister conditions are indicated during the late Holocene. A further shift in the dynamic of the fluvial systems of the region is indicated subsequent to 1000BP, coinciding with the medieval warming period as well as human migration into the region. Evidence reflects an overall change towards more arid conditions with an increased occurrence of extreme floods or high magnitude low frequency events.

There are many issues that remain unresolved in the reconstruction of past semi-arid to arid environments, including the evaluation of morphological and sedimentological evidence. It is of importance when considering reconstructions, that to better understand a landscape's response to change, one could understand the environmental conditions preceding the climatic change or shift. This concept of "geomorphological inheritance" also illustrates that the direction of an environmental change influences the geomorphological response. A shift from aridity to humidity would be represented differently in geomorphological record than one in the opposite direction. Also it is difficult to determine the relative contributions of temperature (linked with evaporation) and precipitation changes to overall aridity and its spread (Thomas, 1997). In a global warming scenario the effects of changing climate on geomorphological and ecological processes and functioning must be understood, predicted and managed.

The absence of more conspicuous valley fills over a wider spatial extent within the region; the temporal control of the study site i.e. lack of alluvium predating the Holocene; as well as the poor chronological resolution of the primary study site, restricts a higher resolution interpretation on a more regional scale. A major limitation present within both this study, as

It is considered certain that future changes in the southwestern Cape region will be a reflection of past heritage as well as present and future natural and anthropogenic forcing combined (Tyson *et al.*, 2001). Given the similarities between earlier warm periods from the Quaternary and present-day conditions, research may imply that without human intervention, a climate similar to the present one would extend well into the future (Augustin *et al.*, 2004). Evidence from across southern Africa has provided a picture that suggests that biota and humans across the subcontinent were strongly influenced by relatively mild, when compared to glacial periods, climate fluctuations that occurred during the Holocene (Scott & Lee-Thorp, 2004). If the patterns of change observed during the Late Quaternary are maintained, and even accentuated or amplified and accelerated, then it would follow that any increase in mean annual temperatures will result in a reduction of precipitation and lower moisture availability in the southwestern Cape (Meadows & Baxter, 1997). There is the need to take the knowledge gained from studies of the Late Quaternary and what they have shown regarding environmental change, even globally, and consider it seriously, especially in lieu of land management and planning. Palaeoclimatic interpretations of landforms, with emphasis on geomorphic processes as well as stratigraphic studies of deposits, aid both the regional and local establishment of the response of arid landforms to millennia and decadal time-scale climatic change. Palaeoreconstructions are an invaluable tool for understanding current global climate change (Lioubimtseva, 2004). Bearing in mind the response to Quaternary climate change, although a different forcing mechanism to today, any increase in mean temperature within the winter-rainfall region seems to equate to decreased precipitation and a reduction in moisture availability. This is a critical factor with regards to future land management and planning and suggests more acute water shortages and problems alongside growing agriculture, industry and urbanisation demands in the future (Meadows & Baxter, 1999).

7.4. RECOMMENDATIONS FOR FUTURE RESEARCH

A comparison of present climate variability with the past would provide a sound basis for understanding possible future conditions (Tyson, 1999). There is a wealth of potential study sites and palaeoenvironmental evidence within South Africa and already successful documenting and understanding of climatic and environmental change has been carried out, however, taking the constraints and limitations of palaeoenvironmental interpretations into consideration, and the applicability of the Cederberg region to the elucidation of southwestern Cape palaeoconditions, the following research avenues are suggested.

This project has the potential to draw on more widespread and regional conclusions, thus including more significant ramifications for the Cederberg/Karoo transition locality and in turn for semi-arid and arid regions as a whole in the Western Cape. In order to elucidate such general inferences, a more extensive array of sites and set of analyses could be employed. For example, there is the potential to include new sites of significance to the study that share striking similarities and resemblance to the main Wuppertal exposure (although some still anthropologically altered and disturbed), such as the Biedouw Valley cut and exploration further upstream of Wuppertal on the Dassieboskloof River. Previously considered and studied sites, such as those at Pakhuis Pass (Cornell, 2001) and Lake Bruno (Smith, 2007), could be more extensively investigated to include further detailed sedimentary analysis and larger scaled geomorphology in conjunction with GIS methods (digitising, etc.) to create a comprehensive model and mapping of the area. Additional techniques that could be employed include X-ray fluorescence, pH, exchangeable cations and cation exchange capacity, CBD extraction, radiocarbon dating, mapping vegetation and land-use changes, catchment variations, etc. Trace geochemistry and analysis is also a possibility for a better elucidation of the erosional versus depositional environments through the sediments. This is done using leaches or mass spectrometers, such as the inductively coupled plasma atomic emission spectrometry (ICP) method that establishes the variations in proportions of metallic ions. These can act as indicators for changing erosional histories of catchments (Lowe & Walker, 1997). Attention could be given to stable isotope studies involving charcoal, calcrete and organic matter. Stable isotopes of light elements are useful tracers of a variety of processes that occur at the Earth's surface and geomorphology and offer a number of applications, particularly with regard to determining climates of the recent past (Goudie, 1994). This principle is based on the fundamentals of fractionation. The charcoal present within three horizons of the exposed section could be analysed for stable carbon isotopes. Charcoal, as it is assumed to be the product of biomass burning, when broken down into elemental carbon, is particularly useful after chemical treatment in facilitating carbon isotope determination to establish the origin of the carbon by separating out the isotopic signatures of various photosynthetic pathway plants, namely C4 (grassland) and C3 (forests and shrubs) (Williams *et al*, 1998). Fynbos and the Clanwilliam cedar are both C3 but the close proximity of the arid Tankwa Karoo and CAM succulents that use both C3 and C4 pathways may play a role in identification and changes in vegetation leading to inferences for moisture availability. Calcium carbonate is chemically precipitated in various forms but usually, when in isotopic equilibrium, temperatures can be inferred from $\delta^{18}O$ values although the values of surficial carbonates largely reflect changes in $\delta^{18}O$ values for surface waters and humidity; $\delta^{18}O$ of calcretes and local meteoric precipitation are correlated (Goudie, 1994). Carbon isotopes in terrestrial carbonates also reflect, to a degree, the influence of organic matter on soil CO_2 (and so C3 versus C4 plants) and variation in aridity and soil productivity where an increase in local rainfall results in a decrease of $\delta^{13}C$ (Talma & Vogel, 1992). $^{18}O/^{16}O$ and Deuterium/Hydrogen composition of waters (soil water versus meteoric and groundwater) can also be used in indicating past temperatures (Lee-Thorp & Vogel, 2000). Close microscope inspection of sections of sediment might also include fossils (or microfossils) such as mollusc shells from fluvial environments and so incorporate a palaeoecological aspect into a project as an additional proxy for a more complete or comprehensive study.

7.5. POTENTIAL IMPLICATIONS OF ARIDIFICATION FOR THE REGION

The final objective of this study has been to extrapolate key findings of the research to the potential aridification within the region in a future context. Deserts and semi-deserts are the most extensive of the world's land biome types, occupying more than 30% of the Earth's surface; they are often predicted to be among the most responsive ecosystems to global climatic change, however, there are still major uncertainties regarding the potential effects of increasing concentrations of either CO_2 or future climate change in arid ecosystems with the results of general circulation models (GCMs), in relation to arid environments under a future 'greenhouse effect' climate, being complex and contradictory (Lioubimtseva, 2004 citing Hulme, 2001 and Intergovernmental Panel on Climate Change (IPCC), 2001). It is indispensable to attempt prediction of future conditions and their impact not only on human living conditions and resource availability, such as water and soil, but also ecosystems as well as contemplate the effect dramatic climate change associated with an increased propensity for extreme events will have.

Climatic fluctuations and human activities in the twentieth century have caused the expansion of arid surface conditions, especially a decrease in vegetation cover (and in many regions an associated increase in fire frequency and magnitude) into semi-arid environments (Thomas, 1997). For the most part, the southern African region has experienced change consistent with the hemispheric and global experience (Tyson *et al.*, 2001). However, shorter but no less significant events have occurred in the last 100 years, largely as a result of anthropogenic forcing. Anthropogenic activity in the temperate mid-latitude zones has arguably become the dominant agency in determining the direction and nature of landscape change (Bell & Walker, 1996; Meadows, 2001). The Department of Water Affairs and Forestry for South Africa (2005) suggest a scenario of climate-change over the next 50-100 years in the study area potentially receiving up to 15% less rainfall. Within the realm of fluvial dynamics and this study, it is therefore essential to consider natural versus human-induced changes in flow regime.

Human influence within the study area has been delayed in comparison to other regions of the southwestern Cape. Although aboriginal human populations have inhabited the region, their impact was initially marginal with populations transitioning from nomadic hunter-gatherers to small-scale *Khoikhoi* herders. As Meadows and Sugden (1993) propose, their effect on the environment, although still of significance, may have been overestimated for the Cederberg region as a whole. Their primary influence was to modify the fire regime by making it more frequent and patchy while subsequent stock introduction to the area resulted in numerous additional degradational or disturbance factors being introduced to the environmental system, from vegetation removal through grazing, soil erosion and trampling by stock and associated vegetation removal, to increased water abstraction and consumption. Fire is a necessary component of fynbos life cycle with hot, dry and strong windy summer conditions increasing the likelihood of fires at regular intervals (Cowling, 1997). A change in fire regime i.e. increased frequency and intensity, can be very detrimental to the overall recovery of the fynbos resulting in periods of stunted and reduced vegetation and in more severe cases, total absence. This may either lead to increased surface erosion or the encroachment of opportunistic alien vegetation.

With the advent of post-colonial farmers and Europeans in the area approximately 400 years ago, settlements were constructed and larger-scale agriculture, beyond subsistence, was introduced to the region. Within the context of the study, however, it is not possible to attribute post-colonial exploitation of the region to be solely responsible for change in surfaces, vegetation, sediment loads, water extraction etc. Changes in land-use, resulting in more intensive land utilisation, leads to additional pressure and favours erosion; land use change is currently ongoing within the northern Cederberg. Small settlements, such as

Wuppertal, do create demands on the surrounding environments for increased subsistence farming, while on a more commercial level, there is a trend towards increased amounts of rooibos farming with its growing popularity expanding its market, thus increasing demand. This has numerous potential impacts, such as increased abstraction of water from the surrounding fluvial systems along with natural vegetation removal for cultivation and harvesting of the rooibos. This might lead to lower averaged discharge rates and higher sediment yields making seasonally high discharges more effective and dramatic as well as opening the system to increased vulnerability to denudation by extreme events such as one in 50 or 100 year flooding events. The process of clearing, planting and harvesting fields, even wild rooibos, removes natural groundcover, thereby exposing, sometimes large expanses, of bare soil surface to potential erosion. The loss of topsoil is considered as serious alteration of a landscape as it leads to permanently degraded land. This exposure by agricultural practices, trampling of vegetation, growth of alien vegetation, etc. leads to a significant loss of topsoil and thin fynbos soils are particularly vulnerable to such soil erosion. The conditions to the east of the study site, towards the Tankwa Karoo, are harsh; soils are shallow and poor (not as fertile) and water sources are scarce. Figure 7.1 demonstrates that along the Tra-Tra River valley there is evidence of attempts at small-scale cultivation that has been abandoned or left derelict for a significant period of time. This scars the land surface and slopes, leaving patches of bare soil where no vegetation has re-established itself and essentially leads to increased soil erosion through sheetwash and rill erosion. This demonstrates that the conditions of the area are mostly inappropriate for cultivation. The aridity and lack of soil development within the basin, combined with variable and infrequent precipitation, are thought to be responsible for increasing the vulnerability of the landscape in the region to degradation. This also applies to Quaternary history as these same factors potentially attributed to past geomorphic changes. The northeastern Cederberg bordering the succulent Karoo has a generally low carrying capacity especially in terms of domestic stock, mostly sheep, although replaced by goats in some areas (Rutherford & Westfall, 1986). There is a history of soil erosion in the region due to overgrazing.

Figure 7.1: Patches of unvegetated soil exposed along the northern Tra-Tra River bank indicating scaring from attempts at small-scale agriculture

In this semi-arid region, water resources are especially precious. Some cultivation is relying on centre-pivot irrigation methods to irrigate fields, such as those noted in the Biedouw valley, potentially leading to far-reaching impacts on the environmental system as they may lower the water-table and exploit natural aquifers thus markedly affecting the sustainability of contemporary water resources. The abstraction pattern in the region shows a relatively high level of summer-season groundwater dependence, from whatever aquifer sources are locally available (DWAF, 2005). The groundwater resource has not been systematically explored and developed yet indicating that they may, in fact, be vulnerable to exploitation. Modelling of hydrological regimes can prove useful in future management of dryland river systems, such as the more major Tra-Tra and Moordenaarsgat rivers, and to a lesser degree, the Dassieboskloof River. Extraction may lower the subsurface water-table and reduce flow of water in springs and rivers, which in turn can endanger surface ecosystems that depend on water for sustainability, particularly when stressed during dry periods. There is very little quantitative knowledge of surface-groundwater interaction in this region, thereby raising further concerns with regards to the impact of groundwater abstraction on the ecosystems. Degradation of the natural environment can occur at an alarming rate as habitat destruction, alien vegetation introduction, loss of biodiversity, species-richness, and water loss transforms natural systems. With the introduction of cultivation, settlements and transport, alien vegetation is encroaching on the sensitive Cederberg region; the fynbos biome is severely invaded by mostly woody plant alien species, such as *Acacia cyclops* and *Acacia saligna* (Cullis *et al.*, 2007). Water extraction, through deep taproot systems, of groundwater destined to flow in rivers and streams culminates in the lowering of water levels along with degradation of the ecosystem. Invasive alien plant infestations are not yet significant in terms of water consumption in the area but occurrences should be controlled in accordance with good catchment management, as the riparian areas are prone to rapid invasion (DWAF, 2005). Prevention of infestation would provide the benefits of maintaining the base flows of the rivers.

During the meteorological record South Africa has experienced out of phase rainfall events and correlations on the scale of decades, ENSO scale and inter-annually (Tyson *et al.*, 2001). There has been an attempt to link the 10-12 and ~18 year rainfall oscillations in South Africa with single and double sunspot cycles and activity whilst more recently it has been found that the effects of the southern oscillation and ENSO on South African rainfall are modulated by solar activity. There is frequent substantial variability in annual precipitation leading to some years being considerably more arid than others. Taking these oscillations and variabilities into consideration along with proposed future aridification in the region, the northeastern region of the Cederberg may be increasingly vulnerable to droughts, such that have already occurred (Figure 7.2 clearly shows the effects of a recent drought on the southwestern Cape), and an increased frequency of high magnitude events. With this in mind, along with the type of rainfall (the erosivity of thunderstorms as opposed to gradual sheetwash from cyclonic rainfall), especially in areas of anthropogenically disturbed landscapes and cultivation, future potential areas of erosion or fluvial incision need to be highlighted. It is important to assess palaeofloods in the region with respect to predicting the effect of future floods on not only the safety of communities in the area, even if sparsely populated, but to mitigate the cumulative effects on the fluvial system, especially downstream, due to development along its banks and small-scale damming, which could induce more substantial discharge as well as increased resistance to flow resulting in more dramatic erosion. These factors demonstrate the potential vulnerability of the northeastern region of the Cederberg to even minor climatic variations coupled with anthropogenic forcing.

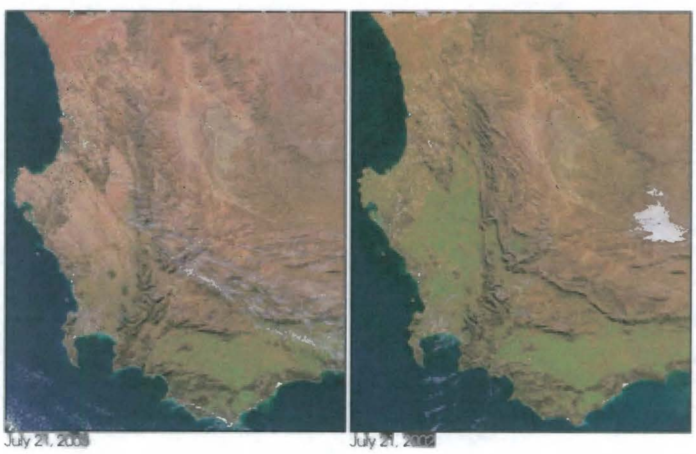


Figure 7.2: Comparative images of the southwestern Cape showing drought effects on land surface between 2002 and 2003 (NASA Visible Earth, 2003)

Conservation initiatives have developed through the formation of the Greater Cederberg Biodiversity Corridor, incorporating both the Cederberg Cape Floral Region and the Tankwa Karoo's succulent Karoo biomes, and the Tankwa Karoo National Park (SANP, 2006). The focus of these projects is to conserve biodiversity and limit the impact of human population growth and poverty in the region (Low *et al.*, 2004). This is relevant to the geomorphological environment, specifically the catchments and fluvial systems of the region, as rivers within the Western Cape, especially the middle and lower reaches, are poorly conserved resulting in increased levels of anthropogenic disturbance. The region bordering the margins of the Karoo, an environment dominated by fluvial landforms, incised drainage, badland gullying and pediplanation, including the Tankwa Karoo, are considered one of the most arguably vulnerable regions in South Africa to not only drought and accelerated anthropogenic influences, but also potentially to small-scale climatic changes (Holmes, 1998). It is interesting to note that the effect of the increasing release of bioshperic CO_2 has altered the $\delta^{13}C$ ratio of atmospheric CO_2 reflected in the tree-cellulose of the Clanwilliam cedar, exhibiting, in accordance with its chronology, a decline in $\delta^{13}C$ values along the same scale obtained for the northern hemisphere, thereby showing that the trend is not related to local ecophysiological influences but is rather representative of global $\delta^{13}C$ air values (February & Stock, 1999). This tracking of global changes demonstrates how so-called global fluctuations have localised effects and are mirrored even in relatively pristine areas. Thus, environmental change related to global factors has the potential to be manifested in even marginal regions. The contemporary significance of such changes to the study area is that it increases its vulnerability to future climate change, even if only subtle fluctuations; the marginal locality on the boundary of semi-arid and arid environments might induce an exaggerated effect on the landscape.

Given the sensitive nature and dynamic, "nonequilibrium" functioning of the semi-arid ecosystems and ecological variability within drylands, the recent intensification of agriculture and tourism within the Cederberg region might have more far-reaching effects beyond contemporary land degradation. Detailed process studies and the application of modern dating techniques continue to provide important data on rates of arid-land geomorphic process and their response to climate change on a variety of timescales, an understanding of which is vital in predicting how drylands will respond to future changes including the effects of global warming. Understanding the dynamics of fluvial systems provides a perspective for deciphering causes of the historic regional incision, ultimately allowing land managers to make more confident decisions (Joyal, 2003). An understanding of the nature of physical environments and the mechanisms, climate driven or otherwise, which induce change, is vital for human endurance while distinct geomorphic thresholds, in terms of processes and landforms, need to be identified within arid environments.

REFERENCES

Abell, P.I. and Plug, I. 2000. The Pleistocene/Holocene transition in South Africa: evidence for the Younger Dryas event. *Global and Planetary Change*, 26: 173-179.

Aitken, M.J. 1998. *An Introduction to Optical Dating*. Oxford University Press, Oxford.

Augustin, L., Barbante, C., Barnes, P., Bamola, J.M., Bigler, M., *et al.* 2004. Eight glacial cycles from an Antarctic ice core. *Nature*, 429: 623-628.

Avery, D.M. 1990. Holocene climatic change in Southern Africa: the contribution of micromammals to its study. *South African Journal of Science*, 86: 407-412.

Avery, D.M. 1993. Last interglacial and Holocene altithermals in South Africa and Namibia: micromammalian evidence. Proceedings of Symposia II-4 and II-6, XIIIth INQUA Congress, Beijing, August 1991. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 101: 221-228.

Baker, V.R. and Kochel, R.C. 1988. Flood sedimentation in bedrock fluvial systems. In: Baker, V.R. and Kochel, R.C. *Flood Geomorphology*. John Wiley & Sons, New York. pp. 123-137.

Bailey, R.M. 2000. The interpretation of quartz optically stimulated luminescence equivalent dose versus time plots. *Radiation Measurements*, 32: 129-140.

Barrable, A., Meadows, M. and Hewitson, B. 2002. Environmental reconstruction and climate modelling of the Late Quaternary in the winter rainfall region of the Western Cape, South Africa. *South African Journal of Science*, 98: 611-616.

Bell, M. and Walker, M.J. 1996. *Late Quaternary Environmental Change – Physical and Human Perspectives*. Longman, Harlow.

Boardman, J., Holmes, P.J., Rhodes, E.J. and Bateman, M.D. 2005. Colluvial fan gravels, depositional environments and luminescence dating: a Karoo case study. *South African Geographical Journal*, 87: 73-79.

Boelhouwers, J.C. 1998. Environmental controls on soil frost activity in the Western Cape mountains, South Africa. *Earth Surface Processes and Landforms*, 23: 211-221.

Boggs Jr., S. 1995. *Principles of Sedimentology and Stratigraphy* (2nd Ed). Prentice Hall, New Jersey.

Bond, G., Showers, W., Cheseby, M., Lotti-Bond, R., Almasi, P., deMenocal, Priore, P., Cullen, H., Hajdas, I., Bonani, G., 1997. A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates. *Science*, 278: 1257–1266.

Bond, W.J., Stock, W.D. and Hoffman, M.T. 1994. Has the Karoo spread? A test for desertification using carbon isotopes from soils. *South African Journal of Science*, 90: 391-397.

Boonzaier, E., Malherbe, C., Berens, P. and Smith, A. 2000. *The Cape Herders – A history of the Khoikhoi of southern Africa.* David Philip, Cape Town.

Boshoff, P., Kovacs, Z., Van Bladeren, D. and Zawada, P. 1993. Potential benefits from palaeoflood investigations in South Africa. *Journal of the South African Institution of Civil Engineering*, 35: 25-26.

Briggs, D., Smithson, P., Addison, K. and Atkinson, K. 1998. *Fundamentals of the Physical Environment* (2nd Ed). Routledge, London.

Bristow, C.S., Skelly, R.L. and Ethridge, F.G. 1999. Crevasse splays from the rapidly aggrading, sand-bed, braided Niobrara River, Nebraska: effect of base-level rise. *Sedimentology*, 46: 1029-1047.

Bull, W.B. 1991. *Geomorphic Responses to Climatic Change*. Oxford University Press, Oxford.

Bullard, J.E. 1997. Vegetation and dryland geomorphology. In: Thomas, D.S.G. (Ed). *Arid Zone Geomorphology : Process, Form and Change in Drylands* (2nd Ed). John Wiley & Sons, Chichester, pp. 109-134.

Byers, S., Mills, E. and Stewart, P. 1978. A comparison of methods of determining organic carbon in marine sediments, with suggestions for a standard method. *Hydrobiologia*, 58: 43-47.

Campbell, I.A. 1997. Badlands and badland gullies. In: Thomas, D.S.G. (Ed). *Arid Zone Geomorphology: Process, form and change in drylands* (2nd Ed). John Wiley & Sons, Chichester.

Cartwright, C. and Parkington, J. 1997. The wood charcoal assemblages from Elands Bay Cave, Southwestern Cape: Principles, procedures and preliminary interpretation. *South African Archaeological Bulletin*, 52: 59-72.

Carr, A.S., Thomas, D.S.G. and Bateman, M.D. 2006. Climatic and sea level controls on Late Quaternary aeolian activity on the Agulhas Plain, South Africa. *Quaternary Research*, 65: 252-263.

Carr, A.S., Thomas, D.S.G., Bateman, M.D., Meadows, M.E. and Chase, B.M. 2006. Late Quaternary palaeoenvironments of the winter-rainfall zone of southern Africa: Palynological and sedimentological evidence from the Agulhas Plain. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 239: 147-165.

Carver, R.E. 1971. *Procedures in Sedimentary Petrology*. Wiley-Interscience, New York.

Chase, B.M. and Meadows, M.E. 2007. Late Quaternary dynamics of southern Africa's winter rainfall zone. *Earth-Science Reviews*, 84: 103-138.

Chase, B.M. and Thomas, D.S.G. 2006. Late Quaternary dune accumulation along the western margin of South Africa: distinguishing forcing mechanisms through the analysis of migratory dune forms. *Earth and Planetary Science Letters*, 251: 318-333.

Chase, B.M. and Thomas, D.S.G. 2007. Multiphase Late Quaternary aeolian sediment accumulation in western South Africa: Timing and relationship to palaeoclimatic changes inferred from the marine record. *Quaternary International*, 166: 29-41.

Chorley, R.J. 1973. The Drainage Basin as the Fundamental Geomorphic Unit. In: Chorley, R.J. (Ed). *Introduction to Fluvial Processes*. Methuen & Co, Suffolk.

Clotet-Perarnau, N, Gallart, F. & Balasch, C. 1988. Medium-term erosion rates in a small scarcely vegetated catchment in the Pyrenees. In: Harvey, A.M. & Sala, M. (Eds). *Geomorphic Processes in Environments with Strong Seasonal Contrasts: selected papers of the "Regional Conference on Mediterranean Countries"*. Catena, Supplement 13, Catena Verlag, Cremlingen-Destedt.

Compton, J.S. 2004. *The Rocks and Mountains of Cape Town*. Double Storey Books, Cape Town.

Cornell, K-J. 2001. *Late Quaternary Sediments in the Pakhuis Basin, Northern Cederberg, Western Cape, South Africa*. Honours Dissertation, UCT.

Cornish-Bowden, G. 2005. *Environmental Reconstruction Through Sediment Analysis of the Wilderness Lake System, South Africa*. Honours Dissertation, UCT.

Cotter, E. 2000. Depositional setting and cyclic development of the lower part of the Witteberg Group (Mid- to Upper Devonian), Cape Supergroup, Western Cape, South Africa. *South African Journal of Geology*, 103: 1-14.

Cowling, R.M., Richardson, D.M. and Pierce, S.M. (Eds). 1997. *The Vegetation of Southern Africa*. Cambridge University Press, Cambridge.

Cowling, R.M., Cartwright, C.R., Parkington, J.E. and Allsopp, J.C. 1999. Fossil wood charcoal assemblages from Elands Bay Cave, South Africa: implications for Late Quaternary vegetation and climates in the winter-rainfall fynbos biome. *Journal of Biogeography*, 26: 367-378.

Cullis, J., Gorgens, A. and Marais, C. 2007. A strategic study of the impact of invasive alien plants in the high rainfall catchments and riparian zones in South Africa on total water surface yield. *Water SA*, 33: 35-42.

Daniels, M. 2003. Floodplain aggradation and pedogenesis in a semiarid environment. *Geomorphology*, 52: 225-246.

Deacon, J. and Lancaster, N. 1988. *Late Quaternary Palaeoenvironments of Southern Africa*. Clarendon Press, Oxford.

Dunkerley, D.L. and Brown, K.J. 1997. Desert Soils. In: Thomas, D.S.G. (Ed). *Arid Zone Geomorphology: Process, form and change in drylands* (2nd Ed). John Wiley & Sons, Chichester.

Dunwiddie, P.W. and LaMarche, V.C. 1980. A climatically responsive tree-ring record from *Widdringtonia cedarbergensis*, Cape Province, South Africa. *Nature*, 286: 796-797.

Dupont, L.M., Kim, J-H., Schneider, R.R. and Shi, N. 2004. Southwest African climate independent of Atlantic sea surface temperatures during the Younger Dryas. *Quaternary Research*, 61: 318-324

Elenga, H., Peyron, O., Bonnefille, R., Jolly, D., Cheddadi, R., *et al.* 2000. Pollen-based biome reconstruction for southern Europe and Africa 18,000 yr BP. *Journal of Biogeography*, 27: 621-634.

Ergenzinger, P. 1988. Regional erosion: rates and scale problems in the Buonamico Basin, Calabria. In: Harvey, A.M. & Sala, M. (Eds). *Geomorphic Processes in Environments with Strong Seasonal Contrasts: selected papers of the "Regional*

Conference on Mediterranean Countries". Catena, Supplement 13, Catena Verlag, Cremlingen-Destedt.

February, E. and Stock, W. 1999. Declining Trend in the 13C/12C Rtaio of Atmospheric Carbon Dioxide from Tree Rings of South African *Widdringtonia cedarbergensis*. *Quaternary Research*, 52: 229-236.

Galbraith, R.F., Roberts, R.G., Laslett, G., Yoshida, H. and Olley, J.M. 1999. Optical dating of single and multiple grains of quartz from Jinmium rock shelter, northern Australia, part 1, Experimental design and statistical models. *Archaeometry*, 41: 339-364.

Gale, S.J. and Hoare, P.G. 1991. *Quaternary Sediments*. Belhaven Halsted Press, New York.

Gallart, F. & Clotet-Perarnau, N. 1988. Some aspects of the geomorphic processes triggered by an extreme rainfall event: The November 1982 flood in the eastern Pyrenees. In: Harvey, A.M. & Sala, M. (Eds). *Geomorphic Processes in Environments with Strong Seasonal Contrasts: selected papers of the "Regional Conference on Mediterranean Countries"*. Catena, Supplement 13, Catena Verlag, Cremlingen-Destedt.

Goudie, A. 1994. *Geomorphological Techniques* (2nd Ed). Routledge, London.

Greenbaum, N., Schick, A. and Baker, V. 2000. The palaeoflood record of a hyperarid catchment, Nahal Zin, Negev Desert, Israel. *Earth Surface Processes and Landforms*, 25: 951-971.

Harvey, A.M. 1997. The role of alluvial fans in arid zone fluvial systems. In: Thomas, D.S.G. (Ed). *Arid Zone Geomorphology : Process, Form and Change in Drylands* (2nd Ed). John Wiley & Sons, Chichester, pp. 231-260.

Heine, K. 2005. Holocene climate of Namibia: A review based on geoarchives. *African Study Monographs*. Suppl. 30: 119-133.

Heiri, O., Lotter, A. and Lemcke, G. 2001. Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results. *Journal of Palaeolimnology,* 25: 101-110.

Helgren, D. 1979. *River of Diamonds.* University of Chicago, Research Paper 185, Chicago.

Hiroki, Y. and Terasaka, T. 2005. Wavy lamination in a mixed sand and gravel foreshore facies of the Pleistocene Hosoya Sandstone, Aichi, central Japan. *Sedimentology,* 52: 65-75.

Holland, R.M. 2000. *An Assessment of Land Degradation in the Headwater Catchment of the Klein Zeekoei River, Great Karoo, South Africa.* Masters Thesis. Unpublished, UCT.

Holmes, P.J. 1998. *Late Quaternary Geomorphic Palaeoenvironments of Some Central and Marginal Great Karoo Uplands, South Africa.* PhD Thesis. UCT.

Holmes, P.J., Boardman, J., Parsons, A.J. and Marker, M.E. 2003. Geomorphological palaeoenvironments of the Sneeuberg Range, Great Karoo, South Africa. *Journal of Quaternary Science*, 18: 801-813.

Holmgren, K., Karlen, W., Lauritzen, S.E., *et al.* 1999. A 3000-year high-resolution stalagmite based record of palaeoclimate for northeastern South Africa. *The Holocene,* 9: 295-309.

Holmgren, K., Lee-Thorp, J., Cooper, G., Lundblad, K. *et al.* 2003. Persistent millennia-scale climatic variability over the past 25,000 years in southern Africa. *Quaternary Science Reviews,* 22: 2311-2326.

Itzstein-Davey, F. 2004. An early Holocene palaeoenvironmental record from Two Mile Lake, South-Western Australia. *Australian Geographer,* 35: 317-332.

Joyal, T. 2003. Fluvial geomorphic response to Holocene climate change on the Mogollon Rim, Arizona. *Rocky Mountain - 55th Annual Meeting.* Paper No. 16-9.

Klein, R.G. 1991. Size variation in the Cape dune molerat and Late Quaternary climatic change in the southwestern Cape Province. *Quaternary Research*, 36: 243-256.

Knighton, D. 1996. *Fluvial Forms and Processes*. Edward Arnold, London.

Knox, J. 2000. Sensitivity of modern and Holocene floods to climate change. *Quaternary Science Reviews*, 19: 439-457.

Kochel, R., Baker, V. and Patton, P. 1982. Palaeohydrology of Southwestern Texas. *Water Resources Research*, 18: 1165-1183.

Kochel, R.C. and Baker, V.R. 1988. Palaeoflood analysis using slackwater deposits. In: Kochel, R.C. and Baker, V.R. *Flood Geomorphology*. John Wiley & Sons, New York. pp. 357-376.

Krauskopf, K. and Bird, D. 1995. *Introduction to Geochemistry* (3^{rd} Ed). McGraw-Hill, Inc., New York.

Lamy, F., Hebbeln, D., Rohl, U., Wefer, G., 2001. Holocene rainfall variability in southern Chile: a marine record of latitudinal shifts of the Southern Westerlies. *Earth and Planetary Science Letters*, 185: 369–382.

Lancaster, N. 1996. Desert Environments. Chapter in: Adams, W.M., Goudie, A.S.and Orme, A.R. (Eds). *The Physical Geography of Africa*. Oxford University Press, Oxford.

Lee-Thorp, J.A. and Talma, A.S. 2000. Stable light isotopes and environments in the southern African Quaternary and Late Pliocene. In: Partridge, T.C. and Maud, R. (Eds). *The Cenozoic of Southern Africa*. Oxford University Press, Oxford, pp.236-251.

Leeder, M.R. 1991. *Sedimentology: process and product*. Harper Collins Academic, London.

Lioubimtseva, E. 2004. Climate change in arid environments: revisiting the past to understand the future. *Progress in Physical Geography*, 28: 502-530.

Long, A. 2001. Mid-Holocene sea-level change and coastal evolution. *Progress inPhysical Geography*, 25: 399-408.

Longley, P., Goodchild, M.F., Maguire, D.J. and Rhind, D.W. 2001. *Geographic information systems and science.* John Wiley & Sons, Chichester.

Low, A.B., Mustart, P. and Van der Merwe, H. 2004. *Greater Cederberg biodiversity corridor: provision of biodiversity profiles for management.* Prepared for the project management unit, Greater Cederberg biodiversity corridor.

Lowe, J.J. and Walker, M.J.C. 1997. *Reconstructing Quaternary Environments* (2nd Ed). Longman, London.

Mabote, M.E. 1997. *Sedimentological and Geochemical Evidence for Late Quaternary Environmental Changes in Southern Africa: A case study of the Mudbelt Deposits off Namaqualand.* Masters Thesis. Unpublished.

Makaske, B., Smith, D. and Berendsen, H. 2002. Avulsions, channel evolution and floodplain sedimentation rates of the anastomosing upper Columbia River, British Columbia, Canada. *Sedimentology*, 49: 1049-1071.

Marker, M. 1995. Late Quaternary environmental implications from sedimentary sequences at two high altitude Lesotho sites. *South African Journal of Science*, 91: 294-303.

Marker, M. 1998. Cenozoic climate change in southern Africa: the evidence from geomorphology, 1967-1996, a personal view. *South African Geographical Journal*, 80: 1-8.

Marker, M. 2003. The Knysna Basin, South Africa: geomorphology, landscape sensitivity and sustainability. *The Geographical Journal*, 169: 32-42.

Martin, A.R. 1959. The Stratigraphy and History of Grownvlei, a South African Coastal Fen. *The Australian Journal of Botany*, 7: 142-167.

Martin, A.R. 1967. Pollen Analysis of Groenvlei Lake Sediments, Knysna (South Africa). *Review of Palaeobotany and Palynology*. Elsevier Publishing Company, Amsterdam.

Meadows, M. 1988. Landforms and Quaternary Climatic Change. In: Moon, B.P. and Dardis, G.F. (Eds). *The Geomorphology of Southern Africa*. Southern, Johannesburg, pp. 296-316.

Meadows, M. 2001. The role of Quaternary environmental change in the evolution of landscapes: case studies from Southern Africa. *Catena*, 42: 39-57.

Meadows, M. and Baxter, A. 1999. Late Quaternary Palaeoenvironments of the southwestern Cape, South Africa: a regional synthesis. *Quaternary International*, 57/58: 193-208.

Meadows, M., Baxter, A. and Parkington, J. 1996. Late Holocene Environments at Verlorenvlei, Western Cape Province, South Africa. *Quaternary International*, 33: 81-95.

Meadows, M.E. and Baxter, A.J. 2001. Holocene vegetation history and Palaeoenvironments at Klaarfontein Springs, Western Cape, South Africa. *The Holocene*, 11: 699-706.

Meadows, M. and Holmes, P. 2000. *Late Quaternary Sediments in the Pakhuis basin, Northern Cederberg, South Africa*. SASQUA, unpublished.

Meadows, M. and Sugden, J.M. 1991. A vegetation history of the last 14 000 years on the Cederberg, south-western Cape Province. *South African Journal of Science*, 87: 34-43.

Meadows, M. and Sugden, J.M. 1993. The Late Quaternary Palaeoecology of a floristic kingdom: the southwestern Cape, South Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 101: 271-281.

Miall, A.D. 1987. Recent Developments in the Study of Fluvial Facies Models. In: Ethridge, F., Flores, R. and Harvey, M. (eds). *Recent Developments in Fluvial Sedimentology.* Society of Economic Palaeontologists and Mineralogists Special Publication No. 39, Tulsa.

Millington, A.C. and Pye, K. 1994. *Environmental Change in Drylands: Biogeographical and Geomorphological Perspectives.* John Wiley & Sons, Chichester.

Moon, B.P. and Dardis, G.F. (Eds). 1988. *The Geomorphology of Southern Africa.* Southern, Johannesburg.

Morgan, R.P. 1995. *Soil erosion and conservation.* Longman, Harlow.

Mulock-Houwer, A. 2001. *Late Quaternary Environmental Reconstruction and Climate Modelling in the Winter Rainfall Region of the Western Cape, South Africa.* Masters Thesis. Unpublished, UCT.

Murray, A.S. and Olley, J.M. 2002. Precision and accuracy in the optically stimulated luminescence dating of sedimentary quartz: a status review. *Geochronometria*, 21: 1-16.

Murray, A.S. and Wintle, A.G. 2000. Luminescence dating of quartz using an improved single-aliquot regenerative-dose protocol. *Radiation Measurements*, 32: 57-73.

Nash, D.J., Thomas, D.S.G. and Shaw, P.A. 1994. Timescales, Environmental Change and Dryland Valley. In: Millington, A.C. and Pye, K. 1994. *EnvironmentalChange in Drylands: Biogeographical and Geomorphological Perspectives.* John Wiley & Sons, Chichester.

Nash, D.J. 1997. Groundwater as a geomorphological agent in drylands. In: Thomas, D.S.G. (Ed). 1997. *Arid Zone Geomorphology : Process, Form and Change in Drylands* (2nd Ed). John Wiley & Sons, Chichester, pp. 319-350.

Nash, D.J. 2000. Arid Geomorphology. *Progress in Physical Geography*, 24: 425-443.

Parkington, J., Cartwright, C., Cowling, R.M., Baxter, A. and Meadows, M. 2000. Palaeovegetation at the last glacial maximum in the Western Cape, South Africa: wood charcoal and pollen evidence from Elands Bay Cave. *South African Journal of Science*, 96: 543-546.

Partridge, T.C., Avery, D.M., Botha, G.A., Brink, J.S., Deacon, J., *et al.* 1990. Late Pleistocene and Holocene climatic change in Southern Africa. *South African Journal of Science*, 86: 302-305.

Partridge, T.C., 1997. Evolution of landscapes. In: Cowling, R.M., Richardson, D.M. and Pierce, S.M. (Eds). *The Vegetation of Southern Africa.* Cambridge University Press, Cambridge, pp. 5-18.

Partridge, T.C., Scott, L. and Hamilton, J.E. 1999. Synthetic reconstructions of southern African environments during the Last Glacial Maximum (21-18 kyr) and the Holocene Altithermal (8-6 kyr). *Quaternary International*, 57/58: 207-214.

Partridge, T.C., Scott, L. and Schneider, R. 2004. Between Agulhas and Benguela: responses of southern African climates of the late Pleistocene to current fluxes, orbital precession and the extent of the circum-Antarctic vortex. In: Batterbee, R.W. *et al.*, (Eds). *Past Climate Variability through Europe and Africa.* Springer, Dordrecht, pp. 45-68.

Paton, T.R., Humphreys, G.S. and Mitchell, P.B. 1995. *Soils: A new global view.* UCL Press Limited, London.

Pelletier, J. 2004. Signatures of climatically-driven hillslope evolution in river andalluvial-fan terraces in the southwestern U.S. *2004 Denver Annual Meeting*. Paper No. 164-5.

Plummer, C.C. and McGeary, D. 1998. *Physical Geology* (7th Ed). Wm. C. Brown, Dubuque.

Preston-Whyte R.A, and Tyson, P.D. 1993. *The Atmosphere and Weather of Southern Africa.* Oxford University Press, Cape Town.

Pye, K. (Ed). 1994. *Sediment Transport and Depositional Processes.* Blackwell Scientific Publications, Edinburgh.

Ramsay, P. and Cooper, J. 2002. Late Quaternary sea-level change in South Africa. *Quaternary Research*, 57: 82-90.

Rancourt, D. and Dang, M-Z. 2005. Absolute quantification by powder X-ray diffraction of complex mixtures of crystalline and amorphous phases for applications in the Earth Sciences. *American Mineralogist*, 90: 1571-1586.

Reid, I. and Frostick, L.E. 1997. Channel form, flows and sediments in deserts. In: Thomas, D.S.G. (Ed). 1997. *Arid Zone Geomorphology : Process, Form and Change in Drylands* (2nd Ed). John Wiley & Sons, Chichester, pp. 205-230.

Reid, D.L., Rogers, J. and Minter, W.E.L. 2001. *The Geology of the Western Cape.* Excursion Guidebook, University of Cape Town.

Reineck, H.-E. and Singh, I.B. 1980. *Depositional Sedimentary Environments* (2nd Ed). Springer-Verlag, Berlin.

Rendell, H. 1997. Tectonic Frameworks. In: Thomas, D.S.G. (Ed). *Arid Zone Geomorphology : Process, Form and Change in Drylands* (2nd Ed). John Wiley & Sons, Chichester, pp. 13-24.

Ritter, D., Kochel, R. and Miller, J.R. 1995. *Process Geomorphology* (3rd Ed). Wm. C. Brown, Dubuque.

Rollinson, H. 1998. *Using geochemical data: evaluation, presentation, interpretation.* Longman Group, Harlow.

Romero-Diaz, M.A., Lopez-Bermudez, F., Thornes, J., Francis, C. and Fisher, G. 1988. Variability of overland flow erosion rates in a semi-arid Mediterranean environment under Matorral cover, Murcia, Spain. In: Harvey, A.M. & Sala, M. (Eds). *Geomorphic Processes in Environments with Strong Seasonal Contrasts: selected* *papers of the "Regional Conference on Mediterranean Countries"*. Catena, Supplement 13, Catena Verlag, Cremlingen-Destedt.

Rutherford and Westfall. 1986. *Biomes of Southern Africa*. Memoirs of Botanical Survey of South Africa, Botanical Research Institute.

Santisteban, J., Mediavilla, R., López-Pamo, E., Dabrio, C., Zapata, M., García, J., Castaño, S. and Martínez-Alfaro, P. 2004. Loss on ignition: a qualitative or quantitative method for organic matter and carbonate mineral content in sediments? *Journal of Palaeolimnology*, 32: 287-299.

Saint-Laurent, D. 2004. Palaeoflood hydrology: an emerging science. *Progress in Physical Geography*, 28: 531-543.

Saynor, M. and Erskine, W. 1993. Characteristics and implications of high-level slackwater deposits in the Fairlight Gorge, Nepean River, Australia. *Australian Journal of Marine and Freshwater Research*, 44: 735-747.

Schumann, W. 1998. *Minerals of the World.* Sterling Publishing, New York.

Scott, L. 1994. Palynology of late Pleistocene hyrax middens, southwestern Cape Province, South Africa: a preliminary report. *Historical Biology*, 9: 71-81.

Scott, L., Steenkamp, M. and Beaumont, P.B. 1995. Palaeoenvironmental conditions in South Africa at the Pleistocene-Holocene transition. *Quaternary Science Reviews*, 14: 937-947.

Scott, L. and Vogel, J.C. 2000. Evidence for environmental conditions during the last 20 000 years in Southern Africa from ^{13}C in fossil hyrax dung. *Global and Planetary Change*, 26: 207-215.

Scott, L and Lee-Thorp, J.A. 2004. Holocene climatic trends and rhythms in southern Africa. In: Batterbee, R.W. *et al.*, (Eds). *Past Climate Variability through Europe and Africa*. Springer, Dordrecht, pp. 69-91.

Scott, L. and Woodborne, S. 2007. Pollen analysis and dating of Late Quaternary faecal deposits (hyraceum) in the Cederberg, Western Cape, South Africa. *Review of Palaeobotany and Palynolonogy,* 144:123-134.

Shaw, A.I. 1997. Palaeoenvironments in the Headward Vicinity of Dias Beach, Cape Point. Honours Dissertation, UCT.

Shaw, A.I. and Goudie, A.S. 2004. Chemical sediments of the palaeo and active Okavango alluvial fan, Botswana: evidence for variations in the processes of formation. *Z. Geomorph. N.F.*, 136: 89-112.

Shaw, P.A. 1997. Africa and Europe. In: Thomas, D.S.G. (Ed). *Arid Zone Geomorphology: Process, Form and Change in Drylands* (2nd Ed). John Wiley & Sons, Chichester, pp.467-486.

Shi, N., Dupont, L., Beug, H-J. and Schneider, R. 2000. Correlation between vegetation in southwestern Africa and oceanic upwelling in the past 21 000 years. *Quaternary Research,* 54: 72-80.

Slingsby, P. 2003. *Cederberg – The Map (2nd Ed).* Baardskeerder, Muizenberg.

Slingsby, P. 2006. *Cederberg – The Map (6th Ed).* Baardskeerder, Muizenberg.

Small. J. and Witherick, M. 1995. *A Modern Dictionary of Geography* (3rd Ed). Edward Arnold, London.

Smith, M. 1992. Holocene palaeoclimatic trends from palaeoflood analysis. *Global and Planetary Change,* 5: 235-240.

Smuts, R. 2007. *Peregrinations – The Long Road to the Heart of the Tanqua Karoo.* Veld and Flora, Botanical Society of South Africa, June 2007.

South African Committee for Stratigraphy (SACS), 1980. *Stratigraphy of South Africa– Handbook 8. Part 1: Lithostratigraphy of the Republic of South Africa, Namibia and the Republics of Bophuthatswana, Transkei and Venda* (Comp. L.E. Kent). Department of Mineral and Energy Affairs, Pretoria.

South African Committee for Stratigraphy (SACS), 1994. *Lexicon of South African Stratigraphy*. Council for Geoscience, Geological Survey of South Africa, Pretoria.

South African National Parks (SANP). 2006. *Tankwa Karoo National Park – Park Management Plan.*

Spoenemann, J. and Hagedorn, J. 2000. Morphotectonics in southwest Africa between Heidelberg and Luderitz. *Gottinger Geographical Papers,* 107: 100-109.

Steffen, D., Schlunegger, F. and Preusser, F. 2007. A climatic fingerprint recorded in fluvial terraces and alluvial fans, Valley de Pisco, Peru. *Geophysical Research Abstracts,* 9: 03322.

Stokes, S. 1997. Dating of desert sequences. In: Thomas, D.S.G. (Ed). *Arid Zone Geomorphology: Process, Form and Change in Drylands* (2nd Ed). John Wiley & Sons, Ltd, Chichester, pp. 607-638.

Stokes, S., Thomas, D.S.G. and Washington, R. 1997. Multiple episodes of aridity in southern Africa since the last interglacial period. *Nature*, 388: 154-158.

Stuut, J.B.W., Prins, M.A., Schneider, R., Weltje, G., Jansen, J. and Postma, G. 2002. A 300-kyr record of aridity and wind strength in southwestern Africa: inferences from grain-size distributions of sediments on Walvis Ridge, SE Atlantic. *Marine Geology*, 180: 221-233.

Sugden, J.M. and Meadows, M. 1990. History of the Clanwilliam cedar. *South African Forestry Journal*, 153: 64-71.

Summerfield, M.A. 1993. *Global Geomorphology.* Longman Scientific and Technical, Harlow.

Sutherland, R.S. 1998. Loss-on-ignition estimates of organic matter and relationships to organic carbon in fluvial bed sediments. *Hydrobiologia*, 389: 153-167.

Syvitski, J. (Ed). 1991. *Principles, methods, and application of particle size analysis.* Cambridge University Press, Cambridge.

Talma, A.S. and Vogel, J.C. 1992. Late Quaternary palaeotemperatures derived from a speleothem from Cango Caves, Cape Province, South Africa. *Quaternary Research,* 37: 203-213.

Thomas, D.S.G. (Ed). 1997. *Arid Zone Geomorphology : Process, Form and Change in Drylands* (2nd Ed). John Wiley & Sons, Chichester.

Thomas, D., O'Connor, P., Bateman, M., Shaw, P., Stokes, S. and Nash, D. 2000. Dune activity as a record of Late Quaternary aridity in the Northern Kalahari: new evidence from northern Namibia interpreted in the context of regional arid and humid chronologies. *Palaeogeography, Palaeoclimatology, Palaeoecology,* 156: 243–259.

Thomas, D.S.G. and Shaw, P.A. 2002. Late Quaternary environmental change in central southern Africa: new data, synthesis, issues and prospects. *Quaternary Science Reviews*, 21: 783-797.

Thomas, M.F. 2004. Landscape sensitivity to rapid environmental change - a Quaternary perspective with examples from tropical areas. *Catena*, 55: 107-124.

Tooth, S. and McCarthy, T.S. 2004. Anabranching in mixed bedrock-alluvial rivers: the example of the Orange River above Augrabies Falls, Northern Cape Province, South Africa. *Geomorphology*, 57: 235-262.

Tyson, P.D. 1999. Late Quaternary and Holocene palaeoclimates of southern Africa: A synthesis. *South African Journal of Geology*, 102: 335-349.

Tyson, P., Karlen, W., Holmgren, K. and Heiss, G.A. 2000. The Little Ice Age and medieval warming in South Africa. *South African Journal of Science*, 96: 121-126.

Tyson, P., Odada, E. and Partridge, T. 2001. Late Quaternary environmental change in southern Africa. *South African Journal of Science*, 97: 139-150.

Van der Merwe, H. 2007. *Floral Spectacle in the Succulent Karoo.* Veld and Flora, Botanical Society of South Africa, June 2007.

Vogel, J.C. (Ed). 1984. *Late Cenozoic Palaeoclimates of the Southern Hemisphere.* A.A. Balkema, Rotterdam.

Walker, M. 2006. *Quaternary Dating Methods.* John Wiley & Sons, Chichester.

Watson, A. and Nash, D. 1997. Desert crusts and varnishes. In: Thomas, D.S.G. (Ed). 1997. *Arid Zone Geomorphology : Process, Form and Change in Drylands* (2nd Ed). John Wiley & Sons, Chichester, pp. 69-108.

Westaway, R., Bridgland, D. and Mishra, S. 2003. Rheological differences between Archaean and younger crust can determine rates of Quaternary vertical motions revealed by fluvial geomorphology. *Terra Nova*, 15: 287-298.

White, R.E. 1997. *Principles and Practice of Soil Science – The Soil as a Natural Resource* (3rd Ed). Blackwell Science, Oxford.

Whitten, D.G.A. and Brooks, J.R.V. 1972. *The Penguin Dictionary of Geology.* Penguin, London.

Williams, M., Dunkerley, D., De Deckker, P., Kershaw, P. and Chappell, J. 1998. *Quaternary Environments* (2nd Ed). Arnold, London.

Zawada, P., Hattingh, J. and Van Bladeren, D. 1996. Palaeoflood hydrological analysis of selected South African rivers. *Rep. Water Res. Commn.*, Pretoria, 509/1/96: 209.

Websites:

CapeNature. *Launching the Greater Cederberg Biodiversity Corridor.* http://www.capenature.org.za/index.php?fArticleId=403

Climate Data for Calvinia

http://www.weathersa.co.za/Climat/Climstats/CalviniaStats.jsp

Cluster Analysis. 2004.
http://www.mathworks.com/access/helpdesk/help/toolbox/stats/multiv14.html
http://www2.chass.ncsu.edu/garson/pa765/cluster.htm

CNdV Africa., 2005. *The Western Cape - Provincial Spatial Development Framework.* http://www.capegateway.gov.za/Text/2005/12/4a_wc_pages_51-114_web.pdf.

Cullis, J., Gorgens, A. and Marais, C. 2007. *A strategic study of the impact of invasive alien plants in the high rainfall catchments and riparian zones of South Africa on total surface water yield.* http://www.wrc.org.za/downloads/watersa/2007/Jan%2007/2044.pdf.

Department of Environmental Affairs and Tourism (DEAT), 2000. http://www.environment.gov.za

Drees, R. 1997. *Fabric and Mineral Analysis.* http://www.ces.ncsu.edu/plymouth/programs/fab.html

Department of Water Affairs and Forestry (DWAF), 2005. Olifants/Doorn Water Management Area. www.dwaf.gov.za/Documents/Other/WMA/17/OlifantsDoornISPFeb05p8.pdf

Geology of the Cape Peninsula
http://web.uct.ac.za/depts/geolsci/cape.htm

Google Earth
http://earth.google.com/

International Symposium on Gully Erosion under Global Change
http://www.soilerosion.net/cost623/strasbourg_writeup.doc

LANDSAT Imagery

https://zulu.ssc.nasa.gov/
http://geo.arc.nasa.gov/sge/landsat/daccess.html
http://landsat7.usgs.gov/index.php

McKee, C. XRD Facility, New Mexico Bureau of Geology and Mineral Resources.
http://geoinfo.nmt.edu/labs/x-ray/

MODIS Rapid Response System (imagery)
http://rapidfire.sci.gsfc.nasa.gov/

Nasa Visible Earth and Imagery
http://visibleearth.nasa.gov/
http://www.nasa.gov/galley/photo/
http://glcf.umiacs.umd.edu/data

Nordstrom, K., Psuty, N., Gares, *et al.* 1986. *Living with the New Jersey Shore*. Duke University Press Durham, North Carolina.
http://www.njmsc.org/gofish.html

Principal Component Analysis. 2006.
http://www.fon.hum.uva.nl/praat/manual/PCA.html
http://www2.chass.ncsu.edu/garson/pa765/factor.htm

Rossouw, R. 2006. CSIR XRD Laboratory.
http://www.csir.co.za/plsql/ptl0002/PTL0002_PGE001_XRD

SPOT Imagery
http://www.spotimage.fr/home/
http://www.spaceimaging.com/
www.sac.co.za

StatSoft Electronic Textbook. 2004. *Cluster Analysis*.
http://www.statsoft.com/textbook/stcluan.html#h

StatSoft Electronic Textbook. 2003. *Principal Components and Factor Analysis.*
http://www.statsoft.com/textbook/stfacan.html#

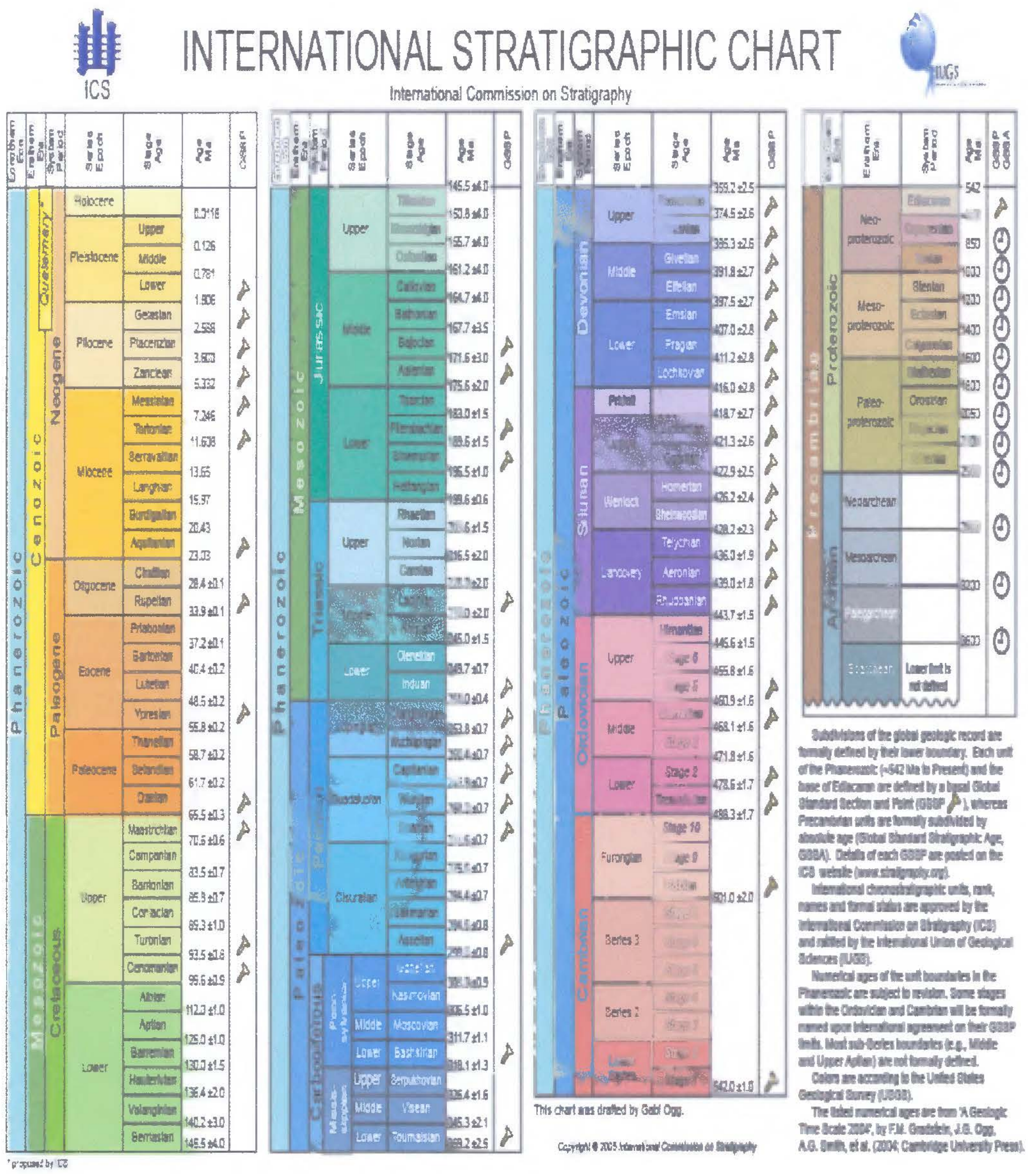
Tankwa Karoo National Park
http://www.sanparks.co.za/parks/tankwa/default.php
http://www.sabirding.co.za/birdspot/010337.asp

Technology of Materials, *X-Ray Diffraction Analysis.*
http://www.xraydiffrac.com/Soilandclaymineralanalysis.html

Wuppertal Homepage
http://www.wupperthal.co.za/

APPENDIX I

International Stratigraphic Chart used in study (International Commission on Stratigraphy, 2005).



APPENDIX II

Stratigraphic column for the Cape Supergroup, consisting of Table Mountain, Bokkeveld and Witteberg Groups (Theron & Thamm, 1990 in Reid *et al.*, 2001).

GROUP	SUB GROUP	FORMATION	THICKNESS (m)	AGE (~y)	
WITTEBERG	LAKE MENTZ SUBGROUP	WAAIPOORT	35	325 VISEAN 342	CARBONIFEROUS
		FLORISKRAAL	70		
		KWEEKVLEI	50	TOURNAISIAN 355	
		WITPOORT	310	FAMENNIAN	DEVONIAN
		SWARTRUGGENS	450	FRASNIAN	
		BLINKBERG	80		
		WAGEN DRIFT	70	375	
BOKKEVELD	BIDOUW SUBGROUP	KAROOPOORT	50		
		OSBERG	55	GIVETIAN	
		KLIPBOKKOP	170		
		WUPPERTAL	65		
		WABOOMBERG	200		
	CERES SUBGROUP	BOPLAAS	30		
		TRA-TRA	85	EIFELIAN	
		HEX RIVER	100		
		VOORSTEHOEK	115		
		GAMKA	135		
		GYDO	160	EMSIAN 390	
TABLE MOUNTAIN	NARDOUW SUBGROUP	RIETVLEI	150	PRAGIAN 410	
		SKURWEBERG	206		SILURIAN
		GOUDINI	120	438	
		CEDARBERG	120	HIRNANTIAN	ORDOVICIAN
		PAKHUIS	40		
		PENINSULA	1550		
		GRAAFWATER	150		
		PIEKENIERSKLOOF	390	510	

APPENDIX III

Summary table of study sites, position, location, altitude, etc. Distances and direction are measured relative to a benchmark, i.e. the fork in the road at the centre of Wuppertal; altitude is quoted in metres above sea level.

WT5	40	1.236	0.628	243	32.733	67.267	34.426	63.775	1.800	65.574	1.986	2.060	0.937	Moderately Sorted	-0.125	Coarse Skewed	0.941
WT5	90	3.517	1.328	1077	71.495	28.505	51.546	41.121	7.332	48.454	1.862	1.950	0.954	Moderately Sorted	-0.128	Coarse Skewed	1.014
WT5	120	1.905	0.999	2430	63.946	36.054	52.546	45.356	2.098	47.454	1.623	1.669	0.820	Moderately Sorted	-0.079	Symmetrical	1.000
WT5	190	3.389	2.600	3380	72.068	27.932	52.643	41.967	5.389	47.357	1.266	1.329	0.766	Moderately Sorted	-0.114	Coarse Skewed	0.969
WT5	240	2.483	1.450	3430	65.585	34.415	50.648	48.816	0.535	49.352	1.297	1.315	0.729	Moderately Sorted	0.009	Symmetrical	1.088
WT5	300	2.399	1.834	783	53.400	46.600	45.421	52.519	2.060	54.579	1.706	1.780	0.821	Moderately Sorted	-0.107	Coarse Skewed	1.057

SITE	DEPTH (cm)	Kurtosis Description	Sand Grade	Gravel %	Coarse Sand %	Medium Sand %	Fine Sand %	Very Fine Sand %	MUNSELL	Descrip-tion	CHAR-COAL	Nodules - frequency and size (mm)	Nodule texture	XRD	OSL Dates
TT1	surface	Mesokurtic	Medium Sand	0.66	38.39	57.18	1.05	1.17	10YR 7/3	very pale brown	Y	few, < 20	very coarse to fine		
DBK1	surface	Mesokurtic	Medium Sand	0.65	6.50	78.68	13.32	0.67	10YR 6/2	light brownish grey	Y	occ, < 5	medium		
MG1A	surface	Leptokurtic	Medium Sand	0.69	10.26	36.73	40.92	8.21	10YR 5/6	yellowish brown	N	some, < 10	coarse		
MG1B	surface	Mesokurtic	Medium Sand	0.26	27.62	64.23	5.79	0.75	10YR 7/4	very pale brown	N	-	-		
MG1C	surface	Leptokurtic	Medium Sand	0.41	7.16	43.53	42.63	5.56	10YR 5/4	yellowish brown	Y	sporadic, < 5	medium		
MG1D	surface	Mesokurtic	Medium Sand	0.65	2.00	44.87	50.47	2.01	10YR 6/4	light yellowish brown	Y	occ, < 5	medium		
MG2A	surface	Leptokurtic	Medium Sand	0.31	13.22	36.06	39.32	8.71	10YR 5/4	yellowish brown	N	numerous, < 10	coarse	A	
MG2B	surface	Leptokurtic	Fine Sand	1.22	0.35	41.00	52.59	4.63	10YR 6/4	light yellowish brown	Y	some, < 5	medium		
MG2C	surface	Mesokurtic	Fine Sand	0.52	0.92	11.86	51.50	34.87	10YR 5/4	yellowish brown	N	more numerous, < 20	very coarse to fine		
MG3A	surface	Mesokurtic	Medium Sand	0.46	31.69	58.76	4.07	0.83	10YR 7/3	very pale brown	Y	sporadic, < 5	medium		

MG4A	surface	Mesokurtic	Coarse Sand	0.50	71.22	8.05	0.69	0.16	10YR 7/4	very pale brown - dull yellow orange	N	-	-		
TT2	0	Leptokurtic	Medium Sand	0.38	1.88	55.67	36.85	5.11	10YR 6/4	light yellowish brown	Y	sporadic, < 2	fine		
TT2	27	Mesokurtic	Medium Sand	0.47	2.12	52.32	41.96	3.00	10YR 6/3	pale brown - dull yellow orange	Y	few, < 8	coarse		
TT2	47	Leptokurtic	Medium Sand	0.46	2.59	48.65	42.72	5.43	10YR 6/2	light brownish grey	Y	numerous, < 15	very coarse to fine	B	
TT2	67	Mesokurtic	Medium Sand	0.49	2.53	51.85	42.29	2.78	10YR 6/3	pale brown	Y	numerous, < 13	very coarse to fine		
TT2	87	Mesokurtic	Medium Sand	0.27	23.70	63.13	11.55	0.94	10YR 7/3	very pale brown	Y	numerous, < 8	coarse		
TT2	107	Mesokurtic	Medium Sand	0.46	38.50	54.38	5.12	0.81	10YR 7/2	light grey	Y	some, < 10	coarse		
TT2	137	Mesokurtic	Coarse Sand	0.42	59.55	35.23	1.40	0.63	10YR 8/2	white	Y	few, < 2	fine		
WT3	Lam	Mesokurtic	Coarse Sand	0.47	65.49	18.61	1.96	0.85	2.5Y 3/1	brownish black	N	occ, < 2	fine		
WT3	0	Mesokurtic	Medium Sand	0.86	24.55	59.84	12.62	1.39	7.5YR 7/1	light brownish grey	Y	sporadic, < 5	medium		
WT3	10	Mesokurtic	Medium Sand	0.33	20.71	60.87	15.73	1.77	2.5Y 6/1	yellowish grey	Y	-	-		
WT3	25	Mesokurtic	Medium Sand	0.81	24.95	52.13	17.71	2.39	10YR 6/1	brownish grey	N	occ, < 5	medium	C	
WT3	40	Mesokurtic	Coarse Sand	0.77	45.80	41.14	5.17	1.51	7.5YR 5/1	brownish grey	N	some, < 10	coarse		

WT4	340	Mesokurtic	Coarse Sand	0.51	61.41	32.19	0.89	0.59	10YR 7/2	light grey	Y	-	-		
WT4	355	Mesokurtic	Coarse Sand	0.44	66.57	10.48	1.74	1.29	10YR 7/2	light grey	Y	-	-		
WT5	0	Leptokurtic	Medium Sand	0.34	21.33	58.18	15.90	2.28	2.5Y 5/1	grey		some, < 5	medium	H	
WT5	40	Mesokurtic	Medium Sand	0.38	14.53	31.13	40.28	12.53	10YR 6/4	light yellowish brown		-	-		
WT5	90	Mesokurtic	Medium Sand	0.21	16.96	32.72	37.90	10.20	7.5YR 6/6	orange reddish yellow		-	-		
WT5	120	Mesokurtic	Medium Sand	0.61	18.79	44.35	30.33	3.91	10YR 7/6	yellow (yellow brown)		-	-		
WT5	190	Mesokurtic	Medium Sand	0.41	28.12	50.45	13.38	2.31	10YR 6/6	brownish yellow		-	-	I	
WT5	240	Mesokurtic	Medium Sand	0.95	28.70	53.76	11.61	3.08	7.5YR 6/6	orange reddish yellow		numerous, < 10	coarse	J	
WT5	300	Mesokurtic	Medium Sand	0.56	16.72	42.46	33.17	5.80	10YR 6/8	brownish yellow		-	-		

APPENDIX VI

Input data table in STATISTICA showing variables used for statistical analysis and correlation matrix from PCA.

SITE	DEPTH (cm)	Organic Carbon %	Inorga-nic Carbon %	Conduc-tivity µS/cm	Mean (ø)	Median (ø)	Sorting (ø)	Skew-ness (ø)	Kurtosis (ø)	Total Clay %	Total Silt %	Total Fines (Clay & Silt) %	Coarse Sand %	Medium Sand %	Fine Sand %	Very Fine Sand %	Total Sand %
TT1	surface	0.351	0.257	-999	1.093	1.117	0.532	-0.086	1.031	-999	-999	0.836	38.387	57.176	1.049	1.169	98.414
DBK1	surface	0.344	0.100	-999	1.569	1.600	0.450	-0.068	1.050	-999	-999	0.755	6.504	78.683	13.317	0.668	99.245
MG1A	surface	1.815	0.442	-999	1.923	1.981	0.884	-0.149	1.239	15.372	84.628	28.055	10.256	36.726	40.916	8.211	70.249
MG1B	surface	0.419	0.087	-999	1.249	1.268	0.581	-0.066	1.062	-999	-999	7.090	27.624	64.227	5.792	0.749	92.550
MG1C	surface	1.782	0.300	-999	1.920	1.971	0.650	-0.112	1.289	-999	-999	18.340	7.161	43.532	42.635	5.555	81.629
MG1D	surface	1.287	0.342	-999	1.999	2.028	0.469	-0.083	1.101	-999	-999	17.238	1.997	44.872	50.474	2.007	82.762
MG2A	surface	2.085	0.589	-999	1.899	1.956	0.899	-0.125	1.144	12.935	87.065	27.147	13.220	36.060	39.320	8.710	61.684
MG2B	surface	2.388	0.459	-999	2.063	2.084	0.490	-0.022	1.122	-999	-999	16.063	0.349	41.004	52.590	4.632	83.937
MG2C	surface	10.307	0.961	-999	2.719	2.741	0.665	-0.083	0.987	13.250	86.750	67.056	0.919	11.861	51.499	34.872	32.944
MG3A	surface	0.416	0.202	-999	1.141	1.170	0.589	-0.120	1.105	-999	-999	5.076	31.687	58.762	4.072	0.828	94.924
MG4A	surface	0.463	0.193	-999	0.382	0.366	0.499	0.072	1.021	-999	-999	4.090	71.225	8.050	0.689	0.160	93.077
TT2	0	2.205	0.415	-999	1.931	1.900	0.539	0.125	1.131	43.689	56.311	12.071	1.877	55.665	36.847	5.111	87.929
TT2	27	2.105	0.342	-999	1.939	1.933	0.510	0.036	1.080	55.146	44.854	11.024	2.117	52.317	41.960	2.996	88.976
TT2	47	2.381	0.621	-999	1.987	1.975	0.538	0.065	1.148	32.378	67.622	18.972	2.594	48.648	42.723	5.426	81.028
TT2	67	0.867	0.506	-999	1.949	1.937	0.497	0.038	1.101	31.213	68.787	9.802	2.525	51.847	42.294	2.785	90.198
TT2	87	0.487	0.406	-999	1.355	1.352	0.582	0.040	1.042	47.005	52.995	6.834	23.700	63.130	11.550	0.940	93.166
TT2	107	0.383	0.336	-999	1.148	1.136	0.556	0.034	1.077	39.896	60.104	4.956	38.499	54.376	5.116	0.809	94.961
TT2	137	0.186	0.362	-999	0.858	0.846	0.514	0.057	1.060	-999	-999	2.286	59.549	35.230	1.398	0.629	96.852
WT3	Lam	1.302	0.352	-999	0.595	0.562	0.591	0.102	1.089	-999	-999	10.746	65.490	18.610	1.960	0.850	88.834
WT3	0	0.717	0.190	121	1.361	1.385	0.628	-0.036	1.030	-999	-999	2.714	24.546	59.844	12.617	1.386	97.188
WT3	10	0.465	0.170	-999	1.443	1.450	0.634	0.001	1.068	-999	-999	2.583	20.710	60.870	15.732	1.769	96.923
WT3	25	0.633	0.168	110	1.394	1.400	0.740	-0.027	1.014	-999	-999	7.447	24.950	52.129	17.713	2.389	92.414
WT3	40	0.732	0.246	-999	0.970	0.961	0.708	0.049	1.074	-999	-999	10.139	45.800	41.140	5.170	1.510	89.580
WT3	60	0.715	0.149	122	0.968	0.957	0.732	0.045	1.048	-999	-999	11.019	45.299	40.224	5.955	1.519	88.025
WT3	80	0.745	0.192	-999	0.637	0.604	0.606	0.112	1.137	-999	-999	6.069	65.624	20.016	1.439	1.739	92.926
WT3	95	0.440	0.080	73	0.490	0.486	0.481	0.012	1.057	-999	-999	4.122	75.953	9.563	1.108	0.819	95.322
WT3	110	2.116	0.163	-999	0.883	0.822	0.776	0.152	1.035	-999	-999	9.409	50.660	31.150	7.310	1.690	90.169

WT3	125	0.412	0.150	59	1.254	1.275	0.695	-0.032	0.942	-999	-999	2.986	32.391	51.979	11.943	1.139	96.126
WT3	135	0.770	0.241	-999	1.209	1.198	0.577	0.075	1.263	-999	-999	4.341	31.791	58.618	7.238	0.957	95.612
WT3	145	0.438	0.080	54	1.066	1.100	0.500	-0.069	1.115	-999	-999	2.866	39.451	54.229	3.565	0.779	97.092
WT3	155	0.564	0.165	-999	0.881	0.895	0.567	-0.002	1.103	-999	-999	5.963	52.079	38.804	3.059	0.620	93.211
WT3	170	0.233	0.085	36	0.463	0.454	0.516	0.044	1.027	-999	-999	1.112	69.973	12.035	0.728	0.509	97.362
WT3	185	1.670	0.084	-999	1.149	1.122	0.632	0.096	1.195	-999	-999	11.568	39.467	49.830	6.393	2.364	88.191
WT3	190	0.369	0.077	45	1.260	1.262	0.564	-0.003	1.097	-999	-999	1.754	27.485	64.157	6.759	0.600	98.246
WT3	195	1.217	0.242	-999	1.257	1.259	0.692	-0.051	1.289	-999	-999	10.636	25.096	58.829	9.388	1.454	89.329
WT3	215	0.377	0.157	38	1.122	1.144	0.496	-0.057	1.079	-999	-999	1.711	36.002	59.307	3.054	0.409	98.289
WT3	230	3.810	0.428	-999	1.448	1.390	0.727	0.119	1.081	-999	-999	24.799	22.664	53.563	19.506	2.208	75.201
WT3	250	1.486	0.334	121	1.312	1.303	0.694	0.047	1.088	-999	-999	18.898	27.950	54.762	13.111	1.999	80.835
WT4	Lam1	1.765	0.601	159	1.362	1.367	0.620	-0.005	1.122	-999	-999	22.907	22.210	62.368	12.138	1.457	76.790
WT4	Lam2	1.913	0.776	-999	1.562	1.570	0.694	-0.040	1.139	-999	-999	35.307	15.955	57.763	21.742	1.986	64.611
WT4	0	0.763	0.239	121	1.341	1.365	0.669	-0.047	1.076	-999	-999	11.735	24.736	58.139	12.527	1.935	88.204
WT4	30	0.821	0.300	-999	1.338	1.338	0.630	0.019	1.088	-999	-999	6.478	24.845	60.204	12.073	1.369	93.522
WT4	55	0.861	0.236	105	1.314	1.322	0.654	-0.007	1.048	-999	-999	11.155	26.946	57.658	11.719	1.539	88.759
WT4	100	1.105	0.307	159	0.931	0.983	0.589	-0.113	1.134	-999	-999	9.985	44.624	45.204	2.958	0.759	88.309
WT4	130	0.641	0.234	-999	1.326	1.345	0.630	-0.039	1.097	-999	-999	8.327	24.318	61.669	10.465	1.239	91.280
WT4	155	0.682	0.176	70	1.382	1.419	0.644	-0.073	1.207	-999	-999	7.824	19.918	64.020	11.108	2.238	92.078
WT4	185	0.871	0.254	-999	1.477	1.457	0.642	0.088	1.088	-999	-999	8.927	19.330	60.870	16.432	2.429	91.029
WT4	205	1.221	0.233	289	1.567	1.607	0.548	-0.127	1.101	-999	-999	6.113	12.655	67.869	16.650	1.209	93.887
WT4	210	0.644	0.201	-999	1.282	1.301	0.630	-0.040	1.056	-999	-999	3.823	27.507	60.700	8.840	1.068	96.081
WT4	230	0.470	0.086	47	1.535	1.559	0.504	-0.060	1.050	-999	-999	1.915	11.988	71.771	14.714	0.908	98.068
WT4	250	1.464	0.229	-999	1.461	1.435	0.472	0.068	1.080	-999	-999	6.044	11.954	77.671	8.996	0.650	93.956
WT4	265	1.805	0.176	71	1.363	1.355	0.533	0.035	1.093	-999	-999	4.483	19.276	69.924	8.307	1.057	95.517
WT4	275	0.884	0.152	-999	1.123	1.179	0.716	-0.107	1.032	-999	-999	4.449	31.950	53.860	7.020	0.910	94.769
WT4	280	0.800	0.123	89	1.456	1.513	0.608	-0.182	1.231	-999	-999	5.705	14.792	67.804	12.886	0.858	87.895
WT4	285	0.800	0.146	-999	1.366	1.376	0.607	-0.048	1.218	-999	-999	5.847	18.604	66.856	10.256	1.168	79.274
WT4	300	0.336	0.082	37	1.144	1.171	0.480	-0.075	1.108	-999	-999	1.538	32.278	63.398	2.557	0.280	98.429
WT4	320	0.197	0.072	-999	0.746	0.776	0.569	-0.113	0.975	-999	-999	1.198	56.219	32.111	1.079	0.609	93.095
WT4	340	0.159	0.154	37	0.796	0.795	0.527	0.037	1.008	-999	-999	1.215	61.410	32.188	0.889	0.589	98.196

WT4	355	0.164	0.080	27	0.417	0.399	0.562	0.113	1.090	-999	-999	0.900	66.573	10.483	1.737	1.288	96.721
WT5	0	2.215	0.332	285	1.426	1.433	0.704	0.003	1.170	30.566	69.434	32.501	21.334	58.180	15.901	2.277	65.428
WT5	40	1.236	0.628	243	1.986	2.060	0.937	-0.125	0.941	32.733	67.267	34.426	14.531	31.126	40.281	12.526	63.775
WT5	90	3.517	1.328	1077	1.862	1.950	0.954	-0.128	1.014	71.495	28.505	51.546	16.960	32.720	37.900	10.200	41.121
WT5	120	1.905	0.999	2430	1.623	1.669	0.820	-0.079	1.000	63.946	36.054	52.546	18.787	44.349	30.329	3.907	45.356
WT5	190	3.389	2.600	3380	1.266	1.329	0.766	-0.114	0.969	72.068	27.932	52.643	28.118	50.454	13.380	2.307	41.967
WT5	240	2.483	1.450	3430	1.297	1.315	0.729	0.009	1.088	65.585	34.415	50.648	28.698	53.758	11.613	3.080	48.816
WT5	300	2.399	1.834	783	1.706	1.780	0.821	-0.107	1.057	53.400	46.600	45.421	16.720	42.460	33.170	5.800	52.519

Correlations of Full Statistical Dataset:

	Organic Carbon %	Inorganic Carbon %	Conductivity µS/cm	Mean (ø)	Median (ø)	Sorting (ø)	Skewness (ø)	Kurtosis (ø)	Total Clay %	Total Silt %	Total Fines (Clay & Silt) %	Gravel %	Coarse Sand %	Medium Sand %	Fine Sand %	Very Fine Sand %	Total Sand %
Organic Carbon %	1.00	0.54	0.13	0.60	0.59	0.34	-0.08	-0.07	0.48	0.48	0.78	-0.02	-0.41	-0.22	0.58	0.86	-0.75
Inorganic Carbon %	0.54	1.00	0.63	0.35	0.37	0.49	-0.24	-0.21	0.65	0.63	0.83	-0.08	-0.26	-0.11	0.39	0.37	-0.83
Conductivity µS/cm	0.13	0.63	1.00	-0.03	0.00	0.34	-0.26	-0.29	0.32	0.30	0.51	0.01	0.02	0.01	-0.04	0.00	-0.51
Mean (ø)	0.60	0.35	-0.03	1.00	1.00	0.27	-0.36	0.13	0.53	0.54	0.54	-0.08	-0.94	0.30	0.90	0.64	-0.52
Median (ø)	0.59	0.37	0.00	1.00	1.00	0.29	-0.42	0.12	0.54	0.55	0.55	-0.08	-0.94	0.30	0.89	0.64	-0.53
Sorting (ø)	0.34	0.49	0.34	0.27	0.29	1.00	-0.28	-0.09	0.43	0.42	0.65	0.00	-0.17	-0.16	0.30	0.38	-0.67
Skewness (ø)	-0.08	-0.24	-0.26	-0.36	-0.42	-0.28	1.00	-0.01	-0.13	-0.13	-0.27	-0.05	0.36	-0.20	-0.25	-0.23	0.31
Kurtosis (ø)	-0.07	-0.21	-0.29	0.13	0.12	-0.09	-0.01	1.00	-0.12	-0.11	-0.12	0.38	-0.23	0.25	0.09	-0.13	0.10
Total Clay %	0.48	0.65	0.32	0.53	0.54	0.43	-0.13	-0.12	1.00	1.00	0.68	-0.21	-0.40	-0.13	0.59	0.50	-0.68
Total Silt %	0.48	0.63	0.30	0.54	0.55	0.42	-0.13	-0.11	1.00	1.00	0.67	-0.21	-0.41	-0.14	0.60	0.52	-0.67
Total Fines (Clay & Silt) %	0.78	0.83	0.51	0.54	0.55	0.65	-0.27	-0.12	0.68	0.67	1.00	-0.05	-0.38	-0.21	0.58	0.67	-0.99
Gravel %	-0.02	-0.08	0.01	-0.08	-0.08	0.00	-0.05	0.38	-0.21	-0.21	-0.05	1.00	0.03	0.04	-0.09	-0.09	0.07
Coarse Sand %	-0.41	-0.26	0.02	-0.94	-0.94	-0.17	0.36	-0.23	-0.40	-0.41	-0.38	0.03	1.00	-0.57	-0.76	-0.38	0.36
Medium Sand %	-0.22	-0.11	0.01	0.30	0.30	-0.16	-0.20	0.25	-0.13	-0.14	-0.21	0.04	-0.57	1.00	-0.07	-0.35	0.22
Fine Sand %	0.58	0.39	-0.04	0.90	0.89	0.30	-0.25	0.09	0.59	0.60	0.58	-0.09	-0.76	-0.07	1.00	0.63	-0.56
Very Fine Sand %	0.86	0.37	0.00	0.64	0.64	0.38	-0.23	-0.13	0.50	0.52	0.67	-0.09	-0.38	-0.35	0.63	1.00	-0.66
Total Sand %	-0.75	-0.83	-0.51	-0.52	-0.53	-0.67	0.31	0.10	-0.68	-0.67	-0.99	0.07	0.36	0.22	-0.56	-0.66	1.00

APPENDIX VII

The following site abbreviations were used throughout the study, in the field, labelling of samples and subsequent laboratory analysis. They also appear in text during referrals and discussion.

Site	Abbreviation
Dassieboskloof River	DBK
Wuppertal main exposure	WT
Tra-Tra River Valley	TT
Moordenaarsgat River Valley	MG