THERMOCHRONOLOGY, LANDSCAPE EVOLUTION
AND HYDROGEOLOGY OF THE
KATONGA VALLEY IN SOUTH WEST UGANDA

VOLUME I

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DECLARATION OF AUTHENTICITY

I, John Graham Bradley, confirm that the work presented in this thesis is my own. Where information has been derived from other sources, I confirm that this has been indicated in the thesis.

John Hanning Speke, Source of the Nile (1863)

“We descended into the Katonga valley, where, instead of finding a magnificent broad sheet of water, as I had been led to expect by the Arabs' account of it, I found I had to wade through a succession of rush-drains divided one from the other by islands”
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**Dedication**

For Susan, who gave up her home in British Columbia, who took a leap into the unknown, who persevered through the daily grind year after year, who selflessly supported my research and who never lost faith in me, I can but make the small gesture of dedicating this work to you. Thank you. We are both grateful to our feline companions for maintaining our morale throughout the challenges of the last few years.
ABSTRACT

The reversed river systems of south west Uganda attracted geoscientists to study the geomorphology of the region in the mid 20th century. During the succeeding fifty years the population and GDP per capita have both risen between five and six fold with a consequent increase in water demand. This thesis aims to apply modern quantitative techniques and theoretical developments to improve our understanding of the landscape evolution of the Katonga Valley and contribute to the groundwater resource assessment of associated alluvial deposits.

Karoo-age glaciogenic strata were discovered filling the western valley. Subsequent apatite fission track analyses reveal that the currently exposed rocks were reheated to a temperature consistent with over 2 km of burial during the Mesozoic. Therefore, it is inferred that the western River Katonga has preferentially eroded a Gondwanan paleovalley exhumed from beneath the former sedimentary cover.

Electrical resistivity tomography of the valley fill has identified three broad cycles of erosion and deposition, including:

1) The Gondwanan palaeovalley with indurated glaciogenic strata;
2) The Neogene relict valley with fluviolacustrine sediments; and,
3) The late Quaternary channel and with recent wetland deposits.

Downwarping of the Victoria Basin in the east produced the first drainage divide on the originally westward flowing River Katonga during the middle Pleistocene. Downthrow of the George Basin in the west led to rejuvenation of the western landscape prior to the formation of a second drainage divide due to rift flank uplift. Sand and gravel associated with an old denuded landscape survives in terraces above the water table in the central valley. Variable climate and fluctuating lake levels led to the deposition of fluviolacustrine deposits in the eastern valley. Pumping tests in this silty sand indicate that the transmissivity is almost always adequate for village water use and is sometimes commensurate with town water supplies.
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1 INTRODUCTION

1.1 Overview

Uganda is situated within the upper drainage basin of the River Nile on the East African Plateau (EAP) (Figure 1-1). The Katonga Valley is located in south west Uganda, between the Western Rift of the East African Rift System (EARS) and Lake Victoria (Figure 1-2). The relief and physiography of south west Uganda inspired considerable interest in the landscape evolution of the region during the 20th century. In 1968, J. C. Doornkamp wrote ‘its rift valley, volcanoes, and its history of drainage reversal has given this area a world-wide reputation’ (Doornkamp, 1968a). E. J. Wayland (1931) first recorded that the westward pointing arrow-barb pattern of Lake Kyoga and the Katonga Valley tributaries (Figure 1-3) indicate that the eastward flowing rivers of south west Uganda, including the Kafu, Katonga and Kagera, once flowed west. The early geoscientists working in the region inferred that uplift parallel to the Western Rift initiated river reversal and the subsequent formation of Lake Victoria (Bishop and Trendall, 1967, Doornkamp and Temple, 1966). However, political instability in Uganda and growing dissatisfaction with qualitative approaches to landscape evolution embodied in Strahler’s (1952) call to place geomorphology on sound foundations for quantitative research caused interest in the region to wane during the 1970s and 1980s.

During the fifty years since those early studies of landscape evolution of south west Uganda, the population and gross domestic product per capita have both risen between five and six fold with a consequent increase in water demand. This prompted renewed interest in the geomorphology of Uganda in the 1990s by Taylor and Howard (1998, 1999a, 1999b, 2000) who applied their understanding of the landscape evolution of the region to explain the distribution of deeply weathered rock and improve predictions of groundwater resources in the regolith.

The early studies of landscape evolution in the south west Uganda did not consider the tectonic mechanisms. They were undertaken before the wide-spread acceptance of plate tectonic theory and the development of process based tectonic geomorphology. The development of new dating methods, including cosmogenic nuclide analysis, optically stimulated luminescence and low temperature thermochronology, together
with powerful modelling codes facilitated by increased computing speed during the 1990s and 2000s has led to renewed interest in quantitative landscape evolution studies. As Bishop (2007) states: “The re-emergence of geomorphology at the large spatial scale and long timescale has been made possible by the marriage of the approaches that Strahler (1952) advocated and the questions that Davis (1899) (and Du Toit (1937), Penck (1953), King (1962) and others) asked.”

Figure 1-1: Location of the Katonga Valley in Africa
Figure 1-2: Groundwater resource map of equatorial Africa showing the location of the River Katonga between the Western Rift and Lake Victoria (after Struckmeier, 2008)

Figure 1-3: Digital elevation model of the Katonga Valley catchment and location of the main study sites
Given the development of recent models and techniques, Bishop (2007) also argued that the post-orogenic landscapes of relatively stable continents like Africa, and the ancient landscape remnants proposed by mid 20th century geomorphologists are worth re-examining. In addition, the reconnaissance of the Katonga Valley for this study unexpectedly identified an outlier of indurated sedimentary rocks filling the western valley and terraces of sub-angular sand and gravel on the slopes of the central valley. These discoveries suggest that the history of the Katonga Valley is more complex than previous researchers have recognised.

About 60% (900 million) of the people of Africa live in rural areas and less than half of these people have access to a clean reliable water supply (JMP, 2004). The annual precipitation over all but the north east and south west margins of sub-Saharan Africa is greater than 500 mm/yr. The mean annual rainfall in Uganda is generally between 1,000 and 1,500 mm/a, which is greater than eastern England. However, low porosity and low permeability Precambrian igneous and metamorphic rocks occupy about 40% of the land surface of sub-Saharan Africa (Macdonald et al., 2005, MacDonald et al., 2008), including most of Uganda. There is little geological storage capacity in these areas, which facilitates run-off and together with high equatorial temperatures, promotes water loss to evaporation. Whilst a large proportion of Ugandans live within 100 km of lakes Victoria, Kyoga, Edward and Albert, engineered water distribution from these natural surface stores to dispersed rural communities is cost prohibitive.

The widespread distribution of Precambrian crystalline rocks in Uganda is reflected in the predominance of areas classified as ‘local and shallow aquifers’ with medium to low recharge shown on the groundwater resource map in Figure 1-2 (Stuckmeier, 2008). Despite these limitations, the occurrence of local stores of freshwater in transmissive parts of the weathered and near-surface fractured bedrock (Figure 1-4a) often represents the most cost effective means of providing dispersed water supplies for rural communities (MacDonald et al., 2008). Nevertheless, the quantity and quality of water available from deeply weathered and fractured rock is often difficult to predict and its distribution is sometimes stochastic in character. When present, the coarse-grained alluvial sediment often deposited in river valleys can represent important local sources of groundwater (Figure 1-4b). The prediction of alluvial aquifer distribution in
stable Holocene valleys can be relatively straightforward. However, the complex tectonic and climatic history of south west Uganda means that the distribution and characteristics of fill in the relict valleys of the area can be far from simple.

![Figure 1-4: Transmissive features of the African regolith (after Macdonald et al, 2005)](image)

**1.2 Research Aims**

The original motivation for this thesis was to update our knowledge of the landscape evolution of this classic locality in order to better understand the groundwater resource potential of the relict valley sediments of south west Uganda. A re-examination of the geomorphology and hydrogeology of the Katonga Valley is justified because of: 1) recent theoretical progress in tectonic geomorphology and historical climate research; 2) technological developments in relevant research tools; 3) the discovery of indurated sedimentary rocks in the western valley and sub-angular sand and gravel terraces in the central valley; 4) the availability of pumping test data held in the Uganda government archive; and, 5) increased demand for groundwater resources.

The overall aims of this work can be expressed in terms of the generic aims of science as summarised below.

**Describe** – the geometry, lithology and hydrogeological properties of the valley fill at selected research sites.

**Explain** – the observations of the valley form and fill through conceptual models of the history of burial, exhumation and landscape evolution.
Predict – the distribution of lithological units and hydrogeological properties in the Katonga Valley using the proposed conceptual models.

Apply – this new geological and geomorphological knowledge to assess the groundwater resource potential of the valley fill.

This thesis brings together a diverse range of techniques in geosciences and related sub-disciplines in order to synthesise the history of the Katonga Valley. The burial and exhumation history of the indurated sediments and the valley which they fill has been interpreted using apatite fission track analysis. The geometry of the valley form and fill has been investigated using digital elevation models, geomorphological field observations and electrical resistivity tomography. The lithology of the valley fill has been examined in field exposures, in thin sections, and in existing borehole logs. The transmissivity of the fluviolacustrine deposits has been interpreted using up-to-date pumping test analysis techniques. The specific research objectives and questions are presented in Section 2.4.

1.3 Thesis Structure

Chapters 1 to 4 set the context for the research presented in this thesis. Following this introduction, Chapter 2 describes the physiographic setting, the specific objectives, and the overall approach and methodology. Given the long history of the Katonga Valley and the wide range of disciplines which contribute to understanding this history, the literature review is divided into two chapters. Chapter 3 critically reviews the pre-African (> 120 Ma) geological context and controls on the current form of the Katonga Valley. It begins by summarising the Precambrian lithology and structure and their influence on the form of the Katonga Valley. It describes the Karoo (late Carboniferous to early Jurassic) geology of Africa, before presenting evidence that the indurated rocks of the western Katonga Valley are in fact glaciogenic sediment of Permo-Carboniferous age. Chapter 4 critically examines the post-Gondwana (< 120 Ma) geological and geomorphological history. It begins by describing our current understanding of the tectonic geomorphology and climatic variability of the region, before reviewing the literature concerning the palaeodrainage and Neogene sedimentary record of western Uganda.
Chapter 5, 6 and 7 contain the main body of new research. Chapter 5 presents the apatite fission track analyses and thermal history modelling used to elucidate the history of burial and exhumation of the palaeovalley valley in which the Gondwana-age sediment was deposited. Chapter 6 describes the results of the electrical resistivity tomography conducted at three strategic locations in the western, central and eastern Katonga Valley. Chapter 7 presents the results of pumping test analyses used to determine the transmissivity of the fluviolacustrine deposits in the eastern valley, and the weathered and fractured rock of the region for comparison.

Chapter 8 contains the overall synthesis and discussion, and addresses the objectives outlined in Section 2.5. It begins with a summary of the Phanerozoic history of burial and exhumation of the region and a broad conceptual framework for the overall phases of erosion and deposition in the Katonga Valley. The interpretation of the Neogene tectonic geomorphology and landscape evolution of the Katonga Valley is then presented. The chapter goes on to examine the compare the hydrogeological characteristics of the western valley and summarise the transmissivity characteristics of the fluviolacustrine deposits in the eastern valley. Finally, Chapter 9 presents the main conclusions of this thesis and recommendations for further research.
2 BACKGROUND

2.1 Introduction

This chapter provides background information which sets the context for the research presented in this thesis. The first three sub-sections describe the physical geographical context including: the regional physiography; the topographic characteristics of the Mpanga-Katonga Valley System; and the land use within the Katonga catchment. The final three sections discuss the research process itself including: the objectives; the approach; and the methodology.

2.2 Continental and Regional Topographic Setting

The shaded relief map of the whole of Africa shown in Figure 1-1 reveals a bimodal distribution with the lower peak in the range 400 to 600 m (green) and an upper peak in the range 800 to 1000 m (brown). The predominantly high region of southern Africa surrounds the Kalahari Basin, and the plateaux of East African and Ethiopia lie to the east of the large Congo and Chad basins. Figure 2-1a presents a digital elevation model (DEM) of equatorial Africa generated from 2-minute gridded global relief data (N.O.A.A., 2006). It is annotated with the northern swell of the Ethiopian Dome which is separated from the East African Plateau (also known as the Kenyan and Tanzanian Domes) by the Turkana Depression (also known as the Omo-Turkana lows) (Chorowicz, 2005).

Burke and Gunnell (2008) have suggested that the distinct continental hypsometry of Africa is related to the unique tectonic history of the continent since the start of the break-up of Gondwana 180 million years ago (180 Ma). During this period of ocean floor accretion around the African plate, they suggest the northward drift of Africa has been slow due to the existence of a low shear wave velocity province at the core-mantle boundary. The eruption of deep mantle plumes generated the Karoo and Afar Large Igneous Provinces (LIPs) in South Africa and Ethiopia respectively. Burke and Gunnell (2008) propose that the arrested continental motion due to these plate pinning events resulted in two episodes of shallow convection which produced the development of basin-and-swell topography. They conclude that the Karoo plate pinning event
ended about 130 Ma, and was followed by the break-up of west and east Gondwana. Between about 130 Ma and 30 Ma the African continent experienced relative tectonic quiescence when prolonged denudation and deep weathering produced a composite, low-lying surface of continental extent. The Afar plate pinning event began about 30 Ma and resulted in the active basin-and-swell structures including the East African Plateau on which Uganda sits today, bordered by the eastern and western branches of the East African Rift System as shown in Figure 2-1.

Figure 2-1: a) Digital elevation model of equatorial Africa and b) topographic section along the River Katonga in Uganda and River Ituri in the D.R. Congo

Figure 2-1b shows a vertically exaggerated west to east topographic section along the River Katonga on the East African Plateau (EAP) in Uganda and the River Ituri in the Congo Basin of the D.R. Congo. As discussed further in Section 4.3, this is the approximate proposed course of the westward flowing proto-River Katonga before the formation of the Western Rift. The first order topography of Africa is represented by the lower modal elevation (~0.5 km asl) of the Congo Basin, whilst the higher modal elevation (~1 to 1.25 km asl) is represented by the EAP on which Uganda is situated.
Second order topographic features include the high peaks of the Rwenzori Mountains; the rift valleys; the uplifted rift flanks; and the shallow Victoria Basin. The Rwenzori horst attains a maximum altitude of about 4 km above the floor of the Western Rift, and the deepest parts of the Albert and Edward sediment filled grabens to the north and south are about 4 km below the floor of the rift valley. It is useful to note that horizontal scale of 100s kilometres and the vertical scale of kilometres represented in Figure 2-1a and b is the scale under consideration when examining the history of burial and exhumation using apatite fission track analysis in Chapter 5.

2.3 Topography of the Mpanga-Katonga Valley System

Figure 1-3 presents a DEM image of the Katonga catchment based on shuttle radar topography mission (SRTM) data. A single valley system runs from Lake George in the west, through the papyrus wetland on the drainage divide, to Lake Victoria in the east. The River Katonga is located to the east of the divide and flows into Lake Victoria. To the west of the divide the surface water drains into Lake George via the Nyaitanga, Rusangwe and Mpanga rivers. The entire valley from Lake George to Lake Victoria is referred to in this thesis as the Mpanga-Katonga Valley System. Another striking feature of the Katonga catchment is the westerly flow component in the tributaries, including the rivers Nabakazi, Kibimba, Kyogya and Nabajuzi. A similar westward pointing arrow-barb pattern can be observed in the characteristic shape of Lake Kyoga, north of Lake Victoria. The ‘swamp divide’ and the westward pointing ‘arrow-barb’ pattern first led W.C. Simmons in the 1920s (reported in Wayland, 1931) to propose river reversal had occurred.

Figure 2-2 presents the longitudinal topographic profile of the Mpanga-Katonga Valley System at two different vertical scales shown beneath the DEM image at the same horizontal scale. The lower profile in Figure 2-2b shows the steep drop at the western end of the valley system from approximately 1,100 m asl above Mpanga Falls across the rift boundary fault into the George Basin at about 900 m asl. The current drainage divide at about 1,200 m asl is situated on the rift flank about 50 km along the valley system from the Mpanga Falls. The Katonga Valley itself slowly descends from the drainage divide to 1,133 m asl at Lake Victoria located 180 km along the valley to the
east. The upper profile in Figure 2-2b has an expanded vertical scale and reveals a break of slope in the Katonga Valley about 100 km east of the drainage divide (150 km on the profile).

Figure 2-2: a) Digital elevation model of the Katonga Valley filled to maximum lake level of 1190 m asl and b) longitudinal valley sections

As will be discussed in Section 4.3.3, high level strandlines on the margin of Lake Victoria and in the eastern Katonga Valley suggest that the lake level was once up to about 57 m higher relative to the local landscape today. The DEM and the upper profile in Figure 2-2 have been filled to 1,190 m asl to give a visual impression of the possible maximum extent of Lake Victoria. This elevation is approaching the Mpanga-Katonga Valley System drainage divide and in actual fact the lake level could not be much higher or else it would begin to overflow into several river valleys on its northern shoreline. The filled DEM in Figure 2-2a reveals that the valley narrows in the central section near Kyai. The parts of the Katonga Valley and its tributaries east of Kyai and...
below 1,190 m asl are up to several kilometres wide. Large parts of the western Katonga Valley and the Nabakazi Valley below 1,190 m asl are also a kilometre or more in width. The filled DEM gives an impression of the relative width of different sections of the Katonga Valley. However, Figure 2-2a should be interpreted with caution as strandlines have not been identified west of Kyai and tectonic movement that post dates the maximum lake level may have changed the relative elevations. Whilst it is likely that the valley east of Kyai was flooded during the maximum lake level, it is not clear if the narrow channel in the central valley facilitated flooding of the western valley all the way to the drainage divide at this time.

<table>
<thead>
<tr>
<th>Valley Reach</th>
<th>Average Gradient</th>
<th>Physiographic Characteristics of Reach</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Mpanga Falls</td>
<td>158m in 3km 1:19 (0.0527)</td>
<td>River channel on exposed bedrock in gorge descending into the Western Rift valley</td>
</tr>
<tr>
<td>2) Mpanga Falls to drainage divide</td>
<td>119m in 47km 1:394 (0.0025)</td>
<td>Discrete channel of Rusangwe and Mpanga Rivers on bedrock or alluvium with moderate gradient</td>
</tr>
<tr>
<td>3) Drainage divide to Nkonge</td>
<td>16m in 50km 1:3098 (0.0003)</td>
<td>Straight valley, 1km to 2km wide, with low gradient, linear profile and no discrete channel</td>
</tr>
<tr>
<td>4) Nkonge to Kyai</td>
<td>4m in 53km 1:13285 (0.0001)</td>
<td>Sinuous valley, less than 1km wide, with very low gradient, and no discrete channel</td>
</tr>
<tr>
<td>5) Kyai to Kisozi</td>
<td>9m in 21km 1:2385 (0.0004)</td>
<td>Narrow valley (≤0.5km), with low gradient, straight profile and discrete channel.</td>
</tr>
<tr>
<td>6) Kisozi to Lake Victoria</td>
<td>31m in 57km 1:1846 (0.0005)</td>
<td>Wide valley (2 to 5km), with low gradient, convex profile and no defined channel except near outlet.</td>
</tr>
</tbody>
</table>

Table 2-1: Internally consistent reaches of the Mpanga-Katonga Valley System

The Mpanga-Katonga Valley System has been divided into six internally consistent reaches as shown in Figure 2-2. The average gradient and main characteristics of each reach are summarised in Table 2-1. This thesis aims to explain the different characteristics of these reaches of the Mpanga-Katonga Valley System in terms of the pre-existing geology and the influence of tectonics and climate variation on the landscape evolution of the region.
Figure 2-3: Topographic profiles perpendicular to the western Mpanga-Katonga Valley System

The relative relief reduces from over 150 m in the narrow Mpanga Falls gorge to less than 30 m where the River Mpanga flows in a

Figure 2-4: Topographic profiles perpendicular to the eastern Mpanga-Katonga Valley System

Figure 2-3 shows representative profiles perpendicular to the western Mpanga-Katonga Valley System looking eastward. The relative relief reduces from over 150 m in the narrow Mpanga Falls gorge to less than 30 m where the River Mpanga flows in a
bedrock channel 8.5 km to the east. The floor of the broad Katonga Valley at Kabagole at 1,185 m asl is lower than the drainage divide at 1,200 m asl. Figure 2-4 shows representative profiles perpendicular to the eastern Mpanga-Katonga Valley System looking westward. The width of the valley reduces from nearly 6 km in the east to a narrow channel between Kyai and Kisozi, with a rise in elevation from about 1,150 m asl to about 1,170 m asl in 50 km. On the Kisozi profile, a broad terraced flood plain at about 1,170 m asl can be seen rising above the papyrus wetland at about 1,160 m asl, south of Kisozi Hill. Interestingly, the profile between Kyai and Kisozi in the eastern valley appears to show that the narrow channel has been cut into a former broad valley. The elevation of this apparent relict valley floor is about 1,195 m asl today, which is 15 m higher than the broad western Katonga Valley at Kabagole, and only 5 m lower than the modern drainage divide at Bihanga Station. The geomorphology of the central valley will be examined further in Section 8.6.

2.4 Catchment Land Use

Figure 2-5a shows a Landsat Enhanced Thematic Mapper (ETM) false colour image of the Katonga catchment close to the main valley. The ground truthed land cover classification interpreted from the remote sensing data produced for the United Nations (UN) Food and Agriculture Organisation (FAO) Africover Project is shown in Figure 2-5b (U.N.F.A.O., 2003).

The flat valley floors in the Katonga Valley and the eastern tributaries are classified into the different types of ‘natural aquatic’ vegetation shown in various shades of blue. The author has observed that the majority of these areas are occupied by papyrus wetland of low relief, with no continuous discrete river channel. Most of the central and western Katonga Valley is classified into different types of ‘natural terrestrial’ vegetation in various shades of green. The majority of these areas have varying amounts of tree cover and are used for cattle grazing. The raising of Ankole cattle is the traditional form of agriculture throughout the majority of the Katonga catchment. Areas with few trees or shrubs have been classified as ‘cultivated’ in various shades of brown. These areas tend to surround villages and town and whilst they include areas of
arable farming the author has observed that they also include some areas used for grazing.

Figure 2-5: a) Landsat ETM false colour image of the Katonga Valley, b) Africover land classification of the Katonga Valley

Figure 2-6 shows a typical scene in the western Katonga Valley near Kabagole, and the land use classifications used by the Africover Project.
2.5 Research Objectives and Questions

The overall aim of this thesis is to assess the exhumation history, landscape evolution and hydrogeology of the Katonga Valley. The specific objectives and associated research questions are summarised below.

Objective 1 – Describe the characteristics and explain the origin of the indurated sedimentary rocks identified in the western Katonga Valley.

- What was the environment of deposition of the sedimentary rocks?
- How old are the sedimentary rocks and the palaeovalley in which they reside?
- What is the regional context of the sedimentary rocks in the Katonga Valley?

Objective 2 – Develop a conceptual synthesis of the burial and exhumation history which resulted in preservation of sedimentary rocks in the western Katonga Valley.

- What was the maximum depth of burial of the indurated sedimentary rocks and the palaeovalley in which they reside?
- What is the exhumation history of the sedimentary rocks and the palaeovalley in which they reside?
Objective 3 – Determine the geometry and characteristics of the valley fill in the western, central and eastern reaches of the Katonga Valley.

- How many major depositional units and buried erosion surfaces are there?
- What are the grain-size characteristics of the main depositional units?
- What are the characteristics and environment of deposition of the sand and gravel identified forming terraces in the central valley at Kyai?

Objective 4 – Develop a conceptual synthesis of the landscape evolution of the Mpanga-Katonga Valley System which explains the current valley form and fill.

- What is the role of the Precambrian lithology and structure in determining the form of the modern valley system?
- What is the role of the palaeovalley, in which the indurated sedimentary rocks reside, in determining the form of the modern valley system?
- How have tectonic processes influenced the form of the valley system and the distribution of valley fill?
- How has climatic variability influenced the distribution and character of the valley fill?
- What is the history of formation and migration of the drainage divide?

Objective 5 – Assess the groundwater resource potential of alluvium in the Katonga Valley.

- What is the distribution of hydrostratigraphic units in the Katonga Valley?
- What are the hydrogeological properties (transmissivity and storage coefficients) of hydrostratigraphic units with groundwater resource potential?

These objectives cover a diverse range of geoscience disciplines. Objectives 1, 2 and 3 incorporate sedimentology, palaeontology, stratigraphy, thermochronology and geophysics. Objective 4 incorporates tectonics, geomorphology, climate history and Quaternary science. Objective 5 focuses on physical hydrogeology.
2.6 Approach

Philosophical assumptions underlie all research and it is important to be aware of our assumptions in order to know the limits of our assertions and respond to challenges. Appendix A presents a brief introduction to some of the philosophical issues pertaining to historical geoscience in general and the research presented in this thesis specifically.

The initial idea for this thesis grew out of the hypothesis that basal alluvial aquifers may occur in the relict river valleys of south west Uganda (Tindimugaya, 2008). Whilst the literature review did not falsify this hypothesis, it did not provide strong evidence to corroborate it either. Following examination of the digital elevation data, Landsat ETM images and Africover land use classification, potential research sites were selected at Kabagole, Nkonge and Kisozi. During reconnaissance visits to each site it was concluded that the Kabagole research site showed most promise and was conducive to further investigation. Therefore, 1D electrical resistivity surveys were conducted at Kabagole during the first field season in 2008 (Appendix J). The field observations and resistivity surveys unexpectedly revealed the presence of fine-grained sandstone in the western Katonga Valley. They also suggested that the unconsolidated valley fill at Kabagole was likely fine-grained sediment, which is not promising from a groundwater resource perspective.

Later examination of the sedimentary rocks at Bihanga Station on the drainage divide revealed lithofacies similar to the Lower Karoo (Permo-Carboniferous) glaciogenic rocks that occur on the Entebbe peninsular, 200 km to the east. In addition, reconnaissance of the central Katonga Valley crossing at Kyai identified terraces composed of sub-angular sand and gravel with sub-rounded cobbles. The indurated sedimentary rocks and sand and gravel terraces are important traces, or ‘smoking guns’, which provide evidence of past events and help to constrain the causal explanations considered during the conceptual synthesis of the landscape evolution of the Katonga Valley.

The reconnaissance investigations of the Katonga Valley, including the desk study, field observations and 1D resistivity surveys conducted during the 2008 field season revealed that the geological history and landscape evolution of the region is even more
complex than previous studies have recognised. These early observations led to the formulation of the research objectives and questions listed in Section 2.4. During the 2009 field season, 2D electrical resistivity tomography surveys were conducted at Kabagole, Kyai and Kisozi in collaboration with Nairobi and Makerere universities. During the 2010 field season, rock samples were collected with the assistance of the Ugandan Department of Geological Survey and Mines (DGSM) for apatite fission track analysis at the UCL-Birkbeck Thermochronometry Research Laboratory. In addition, pumping test data was compiled from the Ugandan Department of Water Resource Management (DWRM) data archive. The overall approach during this research is essentially inductive; however, some research questions may be framed in terms of the hypothetico-deductive method. Example hypotheses are listed below.

- **Hypothesis 1** – The sedimentary rocks at Bihanga Station are the same age and origin as the Lower Karoo strata at Entebbe.
  - Falsification/corroboration based on interpretation of facies observations.

- **Hypothesis 2** – The western Katonga Valley has been exhumed from beneath a significant cover of Phanerozoic sedimentary rocks.
  - Falsification/corroboration based on interpretation of apatite fission track observations.

- **Hypothesis 3** – The Katonga Valley contains coarse-grained depositional units.
  - Falsification/corroboration based on interpretation of borehole logs and electrical resistivity observations.

- **Hypothesis 4** – The flow direction in the River Katonga was reversed in a single event and the drainage divide has remained close to its present location ever since.
  - Falsification/corroboration based on geomorphological interpretation of digital elevation model, electrical resistivity survey and field observations.

- **Hypothesis 5** – The transmissivity of specific hydrostratigraphic units in the Katonga Valley alluvium is adequate to provide a practical groundwater resource.
- Falsification/corroboration based on interpretation of pumping test observations.

The general approach adopted during this research is summarised in the project stages listed below.

1. **Motivation** – Improve understanding of the landscape evolution of the Katonga Valley to facilitate assessment of the groundwater resource in associated alluvium.

2. **Desk Study** – Initial literature review and examination of DEM, Landsat images and Africover land use classification to select study sites for field reconnaissance.

3. **Reconnaissance** – Field observations at Kabagole, Nkonge, Kyai, Kisozi and Bihanga Station and 1D electrical resistivity surveys at Kabagole.

4. **Formulate Objectives** – Develop specific research objectives and questions based on the information acquired during the initial desk study and field reconnaissance.

5. **Geological Interpretation** – Assess the environment of deposition and age of the indurated sedimentary rocks in the western Katonga Valley and the sand and gravel terraces in central valley based on observations in the field and in thin section.

6. **Thermochronology** – Interpret the burial and exhumation history of the sedimentary rocks and the palaeovalley in which they reside using apatite fission track analysis.

7. **Geometry and Electrical Properties of the Valley Fill** – Determine the geometry of the depositional units with distinct electrical properties at Kabagole, Kyai and Kisozi using 2D electrical resistivity imaging.

8. **Hydrogeological properties** – Determine the transmissivity of hydrostratigraphic units with groundwater resource potential using pumping tests analyses.

9. **Conceptual Synthesis** – Use all of data, analysis and interpretation to produce conceptual synthesis of the burial and exhumation history, the landscape evolution and the hydrogeology of the Katonga Valley.
2.7 Methodology

The methodology adopted during the research presented here may be framed within the context of the inductive reasoning required to resolve an overarching inverse problem. The objective is to convert observable phenomenon into unobservable functions of the geoscientific system we are trying to understand. For example, whilst we cannot observe the landscape evolution directly, we can observe the traces left behind, and therefore infer the character and chronology of historical events. Whilst numerous methods of acquiring geoscientific observations have been used in the research presented here, the three main techniques are:

1. Apatite fission track analysis – to infer the thermochronology and interpret the burial and exhumation history;

2. Electrical resistivity tomography – to infer the geometry and relative porosity of the valley fill; and

3. Pumping test analysis – to interpret the transmissivity of the valley fill.

One feature that these diverse methods have in common is that they all attempt to resolve specific applications of the inverse problem. Forward modelling is an example of a well-posed direct problem, in which the system function \( S_f \) and input parameters \( I \) are known and the unique output parameters \( O \) can be calculated. This may be described schematically as:

\[ O = I \times S_f \]

The objective of inverse problems, such as those posed by interpretation of AFT, ERT and pumping test analyses are to determine either the system function or the input parameters, given that the other two parameters are known or assumed. The inverse problem may be described schematically as either

\[ S_f = O/I, \text{ or } I = O/S_f \]

In the case of AFT analysis (Chapter 5 and Appendix F) the system function \( ^{238}\text{U} \) decay rate and thermal annealing model) and output parameters (fission track density and length) are known and the input parameter (temperature history) is calculated. In
the case of ERT (Chapter 6 and Appendix I) the input parameter (electrical current) and output parameters (electrical potential and hence total resistance) are known and we attempt to determine the system function (ground resistivity distribution). In the case of pumping test analyses (Chapter 7 and Appendix K), once again it is the input parameter (pumping rate) and output parameters (change in hydraulic head) that are known and we attempt to determine the system functions (transmissivity, storage coefficients and flow geometry).

The inputs, outputs and system functions for each of the techniques employed during this study are summarized in Table 2-2. The interpretation of electrical resistivity surveys and pumping tests present somewhat analogous problems. Both involve the application of a current and the measurement of a potential to determine the spatial distribution of conductivity, or its inverse, the resistivity. However, the imposed geometries are quite different. In the case of ERT, the current is applied at two point sources to determine the two-dimensional planar geometry and resistivity, whereas in the case of pumping tests, the current is applied from a line source in order to determine the two-dimensional radial geometry and conductivity.

<table>
<thead>
<tr>
<th>Inverse Modelling Technique</th>
<th>Input Parameters (I)</th>
<th>Output Parameters (O)</th>
<th>System Functions (S_f)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Apatite fission track analysis</td>
<td>Thermal history (temperature versus age) below ≈110°C</td>
<td>Fission track count and length</td>
<td>238U decay rate and fission track thermal annealing model</td>
</tr>
<tr>
<td>Electrical resistivity tomography</td>
<td>Electrical Current (point sources)</td>
<td>Electrical potential (point measurement)</td>
<td>Geometry and resistivity (planar model)</td>
</tr>
<tr>
<td>Pumping test analyses</td>
<td>Flow rate (line source)</td>
<td>Groundwater head (line measurement)</td>
<td>Geometry, transmissivity and storativity (radial model)</td>
</tr>
</tbody>
</table>

Table 2-2: Summary of input/output parameters and system functions for inverse modelling techniques used in this study

AFT analysis applies the inverse modelling approach to the time domain rather than the spatial domain. In some cases, further assumptions may then be made to convert the temporal model (temperature versus time) to a one-dimensional spatial model (depth
versus time). In this case we assume that the system function, represented by the $^{238}$U decay rate and the thermal annealing model, is known and we calculate the input history based on the apatite fission track density and length.

Inverse problems are typically ill posed and the solution is usually non-unique. In other words, when the output parameters are complex, there may be several combinations of system functions or input parameters that are consistent with the data. Unless the system is simple, it is unlikely that only one model will be consistent with the input and output parameters. The production of inverse models therefore requires the analyst to use all the relevant information at their disposal to reduce the potential ambiguity (non-uniqueness) in the model. For both ERT and pumping test analyses this may include pre-existing information about the geology derived from surface outcrops and/or borehole logs. For AFT analyses it may include information about periods of denudation represented by unconformities in the stratigraphic record.
3 THE PRE-AFRICAN GEOLOGICAL CONTEXT AND CONTROLS

3.1 Introduction

The term ‘pre-African’ is adopted here to denote the vast period of time between the initial craton formation about 3,500 Ma during the Archaean Eon, and the final break-up of Gondwana at about 120 to 130 Ma during the Cretaceous Period. The aim of this chapter is to review the Precambrian geology which influences the form of the Katonga Valley, and the Gondwanan geology which influences our interpretation of sedimentary rocks in the Katonga Valley together with their burial and exhumation history. It begins with a synopsis of the regional geological setting and an overview of the Precambrian thermotectonic events and structural belts. The influence of Precambrian lithology and structure on the geomorphology of the Katonga Valley is then assessed. The chapter goes on to review the Karoo (late Carboniferous to early Jurassic) geology of Africa, focussing on the Karoo-age strata of Uganda, Tanzanian and the D.R. Congo. Finally, having reviewed the relevant literature, the sedimentary rocks of the Katonga Valley are examined in detail and reinterpreted. The information presented here is used in later chapters to interpret the apatite fission track analyses and discuss the influence of pre-existing structures on the landscape evolution of the Katonga Valley.

3.2 Geological Setting

The area of the globe now occupied by the Katonga Valley in south west Uganda has evolved over the last 3.6 billion years. During this time it has resided at different locations relative to the prevailing landmasses, plate margins and tectonic events. The geological setting cannot be defined by the current political borders and instead we must consider the geology of equatorial Africa as a whole as shown in Figure 3-1. The strategic events that lead to the distribution of the thermotectonic and stratigraphic units shown in Figure 3-1 are summarised below.

- **Craton Formation** (Archaean) – The geological history of the region begins with the formation of the regional craton, which incorporates the Congo, Kasai and Tanzanian Cratons (Schlüter, 2006), centred southwest of Uganda. The Archaean terrane on its north-east and eastern margins include the Gneissic-Granulitic
Complex of Uganda, which outcrops to the north and probably also immediately south of the Katonga Valley, and the Tanzanian Craton (Goodwin, 1996) which outcrops south of Lake Victoria.

• **Early Plate Tectonic Cycles** (Palaeoproterozoic and Mesoproterozoic) – The first evidence of modern style plate tectonic cycles (Wilson Cycles) of depositional basins, orogenesis and continental accretion is represented by the Palaeoproterozoic Kibalian Belt and its equivalents, including the Rwenzori Fold Belt (Petters, 1991) which is today traversed by the Katonga Valley. The Kibaran Belt represents a later orogenic cycle that includes the Karagwe-Ankolean System, which outcrops 60 km south of Katonga Valley.

• **Formation of Gondwana** (Neoproterozoic) – The Mozambique Belt of north east Uganda and eastern Kenya and Tanzania was formed during the Pan-African orogenic cycle which is associated with the closure of east and west Gondwana (Goodwin, 1996, Petters, 1991). Given that the sediments and structures associated with this event are located approximately 400 km east of the Katonga Valley (Schlüter, 2005) they have no direct influence on its current form.

• **Platform Cover Sediments** (Neoproterozoic) – Platform cover sediments overlie the Congo Craton and today outcrop around the periphery of the overlying Phanerozoic basin sequence. They include the Lindien Supergroup, exposed 200 km west of the Katonga Valley, and the Bukoban Supergroup which is exposed 90 km to the south. Similar age sediments in Uganda include the Bunyoro and Mityana Series which occur as outliers between 15 and 200 km north of the Katonga, and the Singo Series which at one location (Kisori) outcrops on the north side of the Katonga Valley (Petters, 1991, D.G.S.M., 1962).

• **Formation of the Congo Basin** (Neoproterozoic and Palaeozoic) – Subsidence of the Congo Basin appears to have began following failed Neoproterozoic rifting (Daly et al., 1992). Significant thicknesses of the Neoproterozoic and Palaeozoic sediments that may have once covered the Katonga region are now preserved only in the Congo Basin.
Figure 3-1: Regional geological map (annotated on CCGM and UNESCO map, 1986)
• **Gondwana Glaciation and Terrestrial Deposition** (Palaeozoic and Mesozoic) – Gondwana drifted towards the southern pole during the Palaeozoic which resulted in the Ordovician and Carboniferous glaciations. A period of exhumation, ending with glacial erosion likely removed existing sedimentary strata and influenced the relative topographic relief of the region. Karoo-age deposits, beginning with glaciogenic sediments, occur in the Congo Basin and fill depressions in the Precambrian surface of the eastern D.R. Congo (Cahen, 1954, Giresse, 2005). Three small Karoo exposures are found north of Lake Victoria in Uganda (Schlüter et al., 1993) and extensive Karoo deposits are preserved in the Mesozoic grabens and basins of south and east Tanzania and Kenya (Schlüter, 1997).

• **Formation of Africa and Marine Deposition** (Mesozoic) – Transcurrent movement along two faults occurred between the middle Jurassic and early Cretaceous, and created the eastern margin of Africa. Contemporaneously, seafloor spreading began to open up the North Atlantic during the middle Jurassic (Petters, 1991). The South Atlantic began to open in the early Cretaceous and by the late Cretaceous, Africa and South America separated. To the east of the Katonga Valley, Mesozoic sediments are restricted to the coastal margins of East Africa. To the west, they occur in the Congo Basin where they were deposited in lacustrine basins and marine lagoons (Giresse, 2005).

### 3.3 The Precambrian Geological Context

#### 3.3.1 The structural belts and thermotectonic events

Prior to around 750 Ma the uncertainty in plate positions is too large (> 45°) to construct palaeogeographic maps (Scotese, 2004) and therefore, the Precambrian geological context of Africa is usually described in terms of the thermotectonic events and structural belts. These are summarised here and described in more detail in Appendix B. Greater attention is given to those rocks which underlie or are adjacent to the Katonga Valley. Precambrian rocks outcrop over 80% of Uganda for which palaeontological (relative) dating methods are unavailable. The stratigraphic relationships are therefore interpreted using radiometric (absolute) methods. It is necessary to distinguish between the ages of deposition and subsequent deformation,
which can be particularly problematic where repeated tectonic activity has imposed superficial similarity on stratigraphically diverse formations (Schlüter, 1997). Therefore, the Precambrian rock units are described in terms of the structural belts to which they belong and the tectonothermal events which they have been influenced by. Table 3-1 summarises the stratigraphy, lithology, tectonothermal events and structural belts of Uganda.

As can be seen in Figure 3-2, the majority of central and northern Uganda is underlain by the Palaeoarchaean Gneissic-Granulitic Complex, previously known as the ‘Basement Complex’. A small outcrop of the Neoarchaean Nyanzian System occurs in south east Uganda where it is comprised of a greenstone belt assembly including quartzites, graphite shales, banded ironstones and volcanics subsequently metamorphosed to leptites and amphibolites. This system has no role in the history of south west Uganda and is not discussed further.

The Katonga Valley in south west Uganda is largely underlain by the Palaeoproterozoic Buganda-Toro System which is comprised of low to medium grade metasediments within the Rwenzori Fold Belt. The Mesoproterozoic Karagwe-Ankolean System outcrops in the far south of Uganda and is comprised of low grade metasediments within the Kibaran Belt. Small outcrops of Neoproterozoic sediments including the Bukoban System are found in south and central Uganda.

During the Neoproterozoic (1000 to 542 Ma) the Pan-African orogenic event assembled the supercontinent of Gondwana. It can be seen within the African plate as an anastomosing pattern of tectonic structures and mobile belts (Bumby and Guiraud, 2005) which include the Mozambique Belt in north east Uganda.
<table>
<thead>
<tr>
<th>ERA (age range)</th>
<th>STRATIGRAPHIC UNIT (System/Series/Period)</th>
<th>LITHOLOGY</th>
<th>PRECAMBRIAN STRUCTURAL BELTS</th>
<th>THERMOTECTONIC EVENT [Area of Uganda] (approx. age)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cenozoic (0 to 65 Ma)</td>
<td>Neogene and Quaternary (possibly limited Palaeogene)</td>
<td>Mudstone, siltstone, sandstone, conglomerate, extrusive volcanics</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>Palaeozoic and Mesozoic (65 to 542 Ma)</td>
<td>Permo-Carboniferous (possibly Triassic and early Jurassic)</td>
<td>Laminate, mudstone, siltstone, sandstone, diamictite</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>Neoproterozoic (542 to 1000 Ma)</td>
<td>Bunyoro-Kyoga, Mityana and Singo Series, Bukoban System</td>
<td>Conglomerates, sandstones, arkoses</td>
<td>Mozambique Belt (Karasuk Group)</td>
<td>Mozambique Event [NE] (≥650 Ma)</td>
</tr>
<tr>
<td>Mesoproterozoic (1000 to 1600 Ma)</td>
<td>Karagwe-Ankolean System</td>
<td>Shale, sandstone, quartzite and volcanics</td>
<td>Kibaran Belt</td>
<td>Kibaran Event [S] (≈1000 Ma)</td>
</tr>
<tr>
<td>Palaeoproterozoic (2500 to 1600 Ma)</td>
<td>Buganda-Toro System</td>
<td>Shale, mica-schist, phyllites, quartzites and volcanics</td>
<td>Rwenzori Fold Belt (Kibalian Belt)</td>
<td>Rwenzori Event [SW] (2500 to 1850 Ma)</td>
</tr>
<tr>
<td>Mesoarchaean and Neoarchaean (3200 to 2500 Ma)</td>
<td>Nyanzian System</td>
<td>Metamorphosed volcanics, lepiite and amphibolite, BIFs, quartzite and shale</td>
<td>Nyanzian Greenstone Belt</td>
<td>Aruan Event [N] (≈2600 Ma)</td>
</tr>
<tr>
<td>Palaeoarchaean (3600 to 3200 Ma)</td>
<td>Gneissic Granulitic Complex</td>
<td>Gneisses, granulites, amphibolites, schists, quartzite and marbles</td>
<td>Archaean Craton</td>
<td>Watian Event [N] (≈2900 Ma)</td>
</tr>
</tbody>
</table>

Table 3-1: Stratigraphic, structural and thermotectonic divisions of the Ugandan geology (modified after Schlüter, 1997)
No pre-Carboniferous Phanerozoic rocks have been identified in Uganda and it is therefore likely that erosion dominated the early Palaeozoic history of this part of Gondwana. Permo-Carboniferous clastic sediments have been identified preserved within three minor grabens, including one in Entebbe. This chapter ends with final break-up of Gondwana during the Jurassic Period. Chapter 4.0 of this thesis goes on to
discuss the post-Gondwana intracontinental uplift and rifting that occurred during the Cenozoic and resulted in formation of the East African Plateau and the East African Rift System (EARS). The Palaeozoic and Mesozoic sedimentary rocks of Uganda are insignificant in terms of their current extent and for this reason have perhaps been overlooked. This thesis examines the possibility that they were once much more extensive.

3.3.2 The Katonga Line

3.3.2.1 Structural and lithological evidence

To the east of the Mubende Granite the geological map of south west Uganda (Figure 3-3) shows a major contact located never more than 10 km south of the Katonga Valley and in some locations along the valley itself. North of the contact, the rocks are predominantly cleaved sandstone, slate, phyllite, schist and gneiss of the Buganda Group of the Buganda-Toro System. To the south of the contact the rocks are predominantly acid and basic gneisses and foliated granites (Johnson in King and Swardt, 1967). As previously discussed, although radiometric dating and structural considerations place this area to the south of the Katonga Valley in the Rwenzori Fold Belt, from a lithostratigraphic perspective it has been assigned to the Gneissic-Granulitic Complex (Schlüter, 2006).

A line of discontinuous ridges (Figure 3-3) occurs between the Goru Peninsula at the mouth of the Katonga to beyond Nkonge Hill over a distance of approximately 120 km to the west (Johnson, 1960). Bisset (1942) described these as massive quartz veins and possibly altered quartzites, which he called the Katonga quartzites. Johnson (in King and Swardt, 1967) makes the conjecture that the contact and associated Katonga quartzites represent a major tectonic feature which he calls ‘The Katonga Line’. He also suggests that the line has determined the course of the River Katonga. Schists in Gomba to the north of the Katonga Valley and gneisses in Masaka to the south of the Katonga Valley have an east to west trend. Although this is oblique to the main contact, Johnson (in King and Swardt, 1967) suggests the disparity becomes less when closer to the Katonga Line, thus indicating possible reorientation of existing structures.
Figure 3-3: Geological map of the Mpanga-Katonga drainage basin (annotated on Geological Survey of Uganda maps, 1962)
Figure 3-4: Key for 1:250,000 Uganda geological map (D.G.S.M, 1962)

At Kafuka Hill, west of the southern tributary known as the River Nabajuzi, the vein-like white or pink, jointed quartz is associated with fine-grained siliceous mylonitic rocks and sandstones of the Buganda-Toro System (Johnson, 1960). The fine-grained sandstones are sheared and contain small, conjugate, monoclinal folds with no preferred orientation.

Adjacent to Lake Victoria, the quartz body that outcrops on the Goru peninsula, is complex and contains evidence of successive episodes of quartz infilling, fracturing and replacement (Johnson, in King and Swardt, 1967). The main body is oriented east-west (90°), but a major subsidiary direction of wide banded quartz veins is 120°. Other cross-cutting vein directions are 15° and 140°. Adjacent gneisses show a strong schistosity trending from 40° to 80° due to shearing, which is later than the original mineral foliation.

The Katonga Line cuts across lithologies which is inconsistent with the early interpretation that it was a sedimentary contact (Bisset, 1942). Unlike the large quartz
bodies to the west and north of the Buganda-Toro Series, many of the Katonga quartzites do not appear to be of sedimentary origin (Johnson, in King and Swardt, 1967). The quartz veins appear to have developed in relation to the directions of shear and tension. Johnson (1960) proposes that Katonga Line represents a major dextral transcurrent fault, although his evidence for the direction of displacement is unclear. He also suggests that the prevalence of schists to the north, show downward movement on that side of the line. However, this interpretation should be viewed in the light of his later statement that ‘it occurs in such poorly exposed terrain that its significance is a matter of conjecture’ (Johnson in King and Swardt, 1967). There appears to have been little pertinent additional geological data acquired since and therefore this statement remains valid.

3.3.2.2 Seismic evidence

The Katonga Line does not appear to cut across any features younger than Precambrian in age (Figure 3-3). Nevertheless, recent seismic evidence has been associated with the Katonga Line. Figure 3-5 shows the distribution of earthquakes in Uganda between 1912 and 1973 (Maasha, 1975, Båth, 1975). Recent compilations appear to have been less comprehensive as fewer epicentres are shown in the vicinity of the Katonga Valley (Twesigomwe, 1997, Midzi et al., 1999). The area of the Western Rift, and the Rwenzori horst in particular, experienced the greatest number of earthquakes. The region of the Rwenzori Fold Belt and Lake Victoria experienced moderate seismicity. These events appear to be associated with WNW-ESE trending faults shown on the geological map (Maasha, 1975).

One of the largest earthquakes recorded in Uganda occurred in 1945 (Maasha, 1975) with the epicentre located near Nkozi on the equator less than 5 km from Lake Victoria (≈0.0°N 32°E). The Buganda Group/Gneissic-Granulitic Complex contact and a quartzite body associated with the Katonga Line occur on the Goru peninsula only 5 km south of this epicentre. The earthquake had a magnitude of 6.0 on the Richter scale with numerous aftershocks. Much damage was reported in Masaka and five people were killed. This single event appears to have led to the perception among some geologists in Uganda that the Katonga Line, also called the Katonga Break (Twesigomwe, 1997, Mavonga, 2007), is an active structure.
Figure 3-5: Distribution of earthquakes recorded in Uganda from 1912 to 1973 (after Maasha, 1975)

Despite the significance of the 1945 Nkozi earthquake, the pattern of seismicity shown in Figure 3-5 does not reflect the position of the Katonga Line, although this may be a function of the relatively small dataset. Recent wide-area seismic mapping of eastern and southern Africa has confirmed Maasha’s (1975) proposed zone of seismic activity extending along the Rwenzori Fold Belt, through Lake Victoria, to the Kavirondo rift in Kenya (Midzi et al., 1999). However, contouring of Ugandan seismic data for hazard mapping purposes does not emphasise the Katonga Line specifically (Twesigomwe, 1997). It is concluded that the single 1945 Nkozi event provides inadequate evidence to establish with reasonable certainty that the Katonga Line is an active structure.

3.3.3 Precambrian geological control on the Katonga Valley form

3.3.3.1 Regional context

At a regional scale (Figure 3-1) the Western Rift can be seen to change direction on the border of south west Uganda. The Lake Albert rift approaches from the north and
passes the Rwenzori Mountains to the west, whilst the Lake Edward rift approaches from the south and passes the mountains on the east. The normal faults either side of the Rwenzori Mountains are associated with uplift of the intervening horst block. Koehn et al (2008) propose that large scale structure of the East African Rift System (EARS) including the curvature of the of the Western Rift appears to be controlled by the rigid cratonic lithosphere. It is interesting to note that the Palaeoproterozoic Rwenzori Fold Belt, of which the Plio-Pleistocene Rwenzori horst block is comprised, appears to form the hinge around which realignment of the Western Rift has occurred. The Katonga Valley is located close to the axis of this hinge.

Figure 3-6 presents a sketch of the Rwenzori Fold Belt showing the main structural alignments (modified after Tanner, 1970, and Cahen et al., 1984). The overall structure of the Rwenzori Fold Belt is that of a broad complex syncline with a gently plunging axis (Schlüter, 1997). Its trend is WNW-ESE in the east where it is located approximately 25 km south of the River Katonga, and curves to a NW-SE trend in the west where it crosses the Katonga close to the drainage divide. North of the axis, the dominant strike of the Buganda-Toro System forms a broad sweep with the trend changing from SW-NE near Jinja, to W-E near Kampala and WNW-ESE near Mubende, to WSW-ENE along the Katonga Valley. To the south of the axis the structural trend in the Toro Supergroup and the Gneissic-Granulitic Complex curves to become almost N-S.
Figure 3-6: Sketch map of the main structural alignments in the Rwenzori Fold Belt. 1) Phanerozoic cover rocks; 2) Gneissic-Granulitic Complex 3) granite; 4) Karagwe-Ankolean (Kibaran); 5) Rwenzori Fold Belt (after Schlüter, 1997, Tanner, 1970 and Cahen et al., 1984)

The geological map in Figure 3-3 shows that the Katonga Valley is located closer to the lithological contact associated with the Katonga Line than to the structural axis shown in Figure 3-6. It is interesting to note that the Katonga tributaries to the north including the Nabakazi and Kibimba rivers are approximately aligned with the SW-NE structural trend, and the Katonga tributaries to the south including the Nabajuzi and Kyogya rivers are aligned more closely to the N-S structural trend. This potential structural influence on the drainage pattern could be of significance to the strength of argument for a westward flowing River Katonga based on the apparent westward pointing arrow-barb drainage pattern. However, further evidence for the arrow-barb drainage pattern, independent of structure is provided by the geomorphology of the north-west shore of Lake Victoria and the shape of Lake Kyoga. The possibility arises that the drainage pattern may itself have influenced interpretation of the regional structure shown in Figure 3-6.

East of the Mubende Granite (Figure 3-3), only about 35 km of 135 km of the Katonga Valley follows the contact between the Buganda Group and the Gneissic-Granulitic
Complex. The majority of the Katonga Valley east of the Mubende Granite is situated on the Buganda Group to the north of the contact. The majority of the Katonga quartzites are located south of the Katonga Valley, with a notable exception being east and west of the Nkonge where the Katonga Valley cuts through the quartzite ridge.

The geological map in Figure 3-3 shows several WNW-ESE trending faults dislocating the Mubende Granite. These have a similar orientation to the Katonga Line and can be seen to influence the course of the upper River Muzizi. However, it is unclear if the existence of these faults was inferred from the linear elements of the drainage pattern. As we have seen, although the lithological boundary and occurrence of quartzite bodies along the Katonga Line is well established, the interpretation of these features as an active transcurrent fault must be viewed as conjecture at present and no fault is marked on the geological maps (D.G.S.M., 1962). The general impression given by Figure 3-3 is that rather than the Katonga Line forming a more easily eroded lineament, the resistant quartzites have influenced the valley form by creating a barrier to southward migration of the river. Near Nkonge, the Katonga Valley has cut through the quartzite ridge, suggesting the occurrence of superimposed drainage.

3.3.3.2 Local context

This section summarises the local potential relationships between the Mpanga-Katonga Valley System and the Precambrian geology shown on the geological map in Figure 3-3 for each of the reaches identified in Table 2-1.

1. Mpanga Falls (gradient = 1m:0.2km)

The Falls of the River Mpanga occur in an area of undifferentiated gneisses immediately adjacent to the Western Rift boundary fault. The Precambrian geology appears to have negligible influence on location and form of the gorge.

2. Mpanga Falls to Drainage Divide (gradient = 1m:0.4km)

There is a suggestion that between the confluence with the River Rusangwe and the Mpanga Falls, the River Mpanga may adopt a preferential course on the Buganda-Toro System schists rather than on the gneisses. In contrast, the River Rusangwe east of the confluence remains on the gneisses, but follows the regional SW-NE structural trend for about 15 km before abruptly turning SE. The valley system continues NE via the
Nyaitanga wetland to the drainage divide. It is concluded that the regional SW-NE trend may have influenced the course of the main valley system west of the drainage divide, but subsequent tectonic events appear to have altered the relative importance of the various tributaries within the drainage network.

3. Drainage divide to Nkonge (gradient = 1m:3.1km)

At the drainage divide, the Katonga Valley appears to continue towards the west, whilst the Mpanga-Rusangwe-Nyaitanga valley system meets the Katonga Valley from the south west. Between the drainage divide and the Mubende Granite the Katonga Valley trends W-E cutting across the SW-NE trending Precambrian lithological boundaries and structures. For 12 km west of the Mubende Granite the valley continue west-east along the contact between predominantly schists of the Buganda Group to the north and the predominantly gneisses of the Gneissic-Granulitic Complex to the south.

4. Nkonge to Kyai (gradient = 1m:13.3km)

Approximately 5 km west of Nkonge Granite the Katonga Valley turns ENE passing through the quartz ridge shown in Figure 3-3, which is located on the lithological boundary between the schists and the gneisses. The resistant quartz body appears to have diverted the valley to the north where a branch of the main valley resumes a west-east orientation within the schists parallel to the elongate form of the Nkonge Granite. The main branch of the Katonga Valley cuts back towards the south east across both the Nkonge Granite and the quartz body resuming its course along the schist/gneiss contact. Given the limited exposure, the exact locations of the geological contacts should be treated with caution.

The area between Nkonge and Kyai appears both geologically and geomorphologically complex. From west to east, the River Katonga bends to the south along the lithological contact before bending back to the north where it is joined by the River Nabakazi and its tributary the River Katabulungu. The main river continues to bend until it resumes its south east trend sub-parallel to the schist-gneiss contact which is located no more than 10 km to the south. According to the geological map (D.G.S.M., 1962), Pleistocene deposits which may include ‘sands, clays, grits and gravel’ occur on the high ground between the Nkonge Granite, the River Katonga and its tributaries.
It is concluded that whilst the geological map shows the Katonga Valley following the schist/gneiss contact on its southerly bend the relationship with the Precambrian geology in this reach is complex and elsewhere the presence of quartz bodies outcropping at the schist/gneiss contact have diverted the valley to the north. The small branch of the Katonga Valley to the north of the Nkonge Granite suggest that the valley may have once adopted a straighter course, and the alignment and orientation of the Nabakazi Valley suggests it may even have formed been the previous course of the main valley.

This reach has an almost negligible gradient and the gradient remains less than one third that of adjacent reaches even after removing the influence of valley sinuosity. The reason for this is unclear and may be related to lithology, structure or precedent geological history. The role of the Pleistocene deposits seems pertinent but enigmatic.

5. Kyai to Kisozi (gradient = 1m:2.4km)

East of Kyai the Katonga Valley returns to a slightly higher but still low gradient. The valley remains relatively narrow and the drainage adopts a discrete channel. The geological map (D.G.S.M., 1962) shows the valley to be underlain by ‘cleaved fine sandstone, slates, phyllites and schists’ of the Buganda Group. The map reveals little evidence of structural or lithological control until approximately 10 km west of the Singo Series outcrop at Kisozi where the valley alignment changes from WNW-ESE to follow a NW-SE trending fault and crosses a band of ‘muscovite-biotite gneisses and subordinate schists’ until it reaches the contact with the Gneissic-Granulitic Complex (Johnson, 1955b). The fault is drawn beneath the recent alluvium and was likely inferred from the alignment of the valley and the cross-cutting relationship with the lithology. The Katonga Valley passes the Singo Series quartzitic sandstones on the south side before making a sharp bend to the north.

The relationship between the detailed route of the Katonga Valley and the Precambrian geology in this reach is once again complex. Although the valley appears to remain parallel and within 10 km of the schist-gneiss contact and quartz bodies, the Katonga Line does not appear to exert a simple deterministic control. The reason for the change in valley alignment to pass south, rather than north, of the Kisozi Hill Singo Series
outcrop is unclear and the inferred fault shown on the geological map appears to be a reasonable conjecture.

6. Kisozi to Lake Victoria (gradient = 1m:1.8km)

East of the Kisozi Hill Singo Series outcrop, the Katonga Valley immediately bends north east, cutting across the Buganda Group gneisses and returning the schists. The width of the valley rapidly increases to 2 km and after bending to the southeast increases again to between 3 and 5 km.

East of Kisozi Hill the River Nabajuzi joins from the south east where it follows the schist-gneiss contact for about 5 km. The sharp bend to the north in the Katonga Valley at this confluence resembles the bend in the Nabakazi Valley before its confluence with the Katonga Valley. This raises the possibility that the sharp bend in the Katonga Valley reflects a previous pattern of channel dominance in the drainage system.

Whilst the main section of wide Katonga Valley in this reach is once again aligned parallel and about 10 km north of the schist-gneiss contact there is little evidence of direct lithological or structural control. As shown in Figure 3-3, within 10 km of the outlet of the discrete channel of the River Katonga into Lake Victoria, the wide valley splits. The southern branch of the valley finally cuts across schist/gneiss contact west of the Goru peninsula. North of the Katonga outlet, the swamps of Lugungwa, Buvumba and Mabamba occupy the valleys of a relict tributary to the Katonga drainage system. These swamps form embayments on the northwest shore of Lake Victoria that appear to follow the general structural trend but cut across outcrops of both schist and gneiss.

In summary, the SW-NE trend of the Mpanga, Rusangwe and Nyaitanga valleys west of the drainage divide appears to conform to the general structural trend in the Toro Supergroup. The W-E aligned section of the Katonga Valley between the drainage divide and the Mubende Granite appears discordant to the Precambrian lithology and structures. East of the Mubende Granite, the overall location and WNW-ESE alignment of the Katonga Valley is sub-parallel to the Katonga Line, but on closer examination the relationship is complex. Those local lithological controls that can be established, such as the quartzite bodies, appear to hinder rather than facilitate valley
erosion. In some cases structural controls, such as the fault west of Kisozi Hill have been inferred.

3.4 The Gondwanan Geological Context

The initial reconnaissance identified indurated sedimentary rocks filling the western Katonga Valley. These are interpreted to be of Permo-Carboniferous age in Section 3.4.6. The aim of this section is to facilitate the interpretation of these rocks by examining the Gondwanan geological context, including the palaeogeography, palaeoclimate, tectonics, deposition and denudation. The Phanerozoic geology can be related to the changing palaeogeography including, from the late Carboniferous onwards, the break-up of Gondwana which generally occurred through the reactivation of Pan-African structures (Bumby and Guiraud, 2005). Extensional tectonics created basins at the margins and within the African plate. The fill preserved within these basins reflects the changing palaeoclimate as Africa, in its central position within Gondwana, drifted across the South Pole and northwards to its current location.

3.4.1 The palaeogeographic and palaeoclimatic history of Gondwana

Palaeogeographic and palaeoclimatic reconstructions are used here to provide the narrative context to the Phanerozoic geology of Africa. The lines of evidence used for palaeogeographic reconstruction include: 1) ocean floor linear magnetic anomalies; 2) palaeomagnetic orientation and inclination; 3) hotspot and igneous province tracking; 4) ocean floor tectonic structures; 5) climatic indicators in the geological record; and 6) lithological and structural features associated with plate tectonics (Scotese, 2004). The lithological indicators used in palaeoclimatic reconstruction include coals, evaporates, bauxites and tillites (Scotese et al. 1999).

Palaeogeographic maps should be viewed as hypotheses that explain the historical distribution of terrestrial and marine deposition as well as lack of deposition and denudation. Likewise, the hypothesised palaeoclimatic reconstructions are used to explain the distribution of climatic indicators in the geological record. These reconstructions are summarised here before subsequent sections go on to describe the specific lithostratigraphic units of relevance to this thesis. Scotese (2004) estimated that uncertainty in the plate positions increases from less than ±2° of latitude at the start of the Cenozoic Era (65.5 Ma), to ±10° at the start of the Mesozoic Era (251 Ma), and
±30° at the start of the Palaeozoic (542 Ma). Reconstruction of the palaeogeography before the formation of Gondwana at about 650 Ma is impractical due to the high level of uncertainty. The Mozambique Belt formed during the Pan-African orogenic cycle, which is associated with the formation of Gondwana, provides the link between the palaeogeographic narrative adopted here and the context of structural belts and therмотectonic events adopted in Section 3.3.

It is necessary to consider the palaeoclimatic conditions in the context of two components: 1) secular climate change due to plate movement through climatic zones; and 2) world-wide climate change due to the prevailing global conditions and external forcing. The boundaries of Gondwana remained over or close to the South Pole from its formation in the Neoproterozoic until its break-up in the Cretaceous (Scotese et al., 1999). Gondwana’s northern boundary extended beyond the equator and therefore different parts of the supercontinent simultaneously experienced the polar, temperate, arid and tropical climate zones.

Global climate change due to both internal and external forcing (e.g., thermo-haline oceanic circulation and Milankovitch cycles respectively) is superimposed on the pattern of secular climate change. During the 500 Ma that Gondwana existed, the global climate alternated between warm periods and cool periods as shown in Figure 3-7 (Rohde, 2009, based on Veizer et al., 1999). During the warm periods there were no permanent ice sheets at the poles and the average global temperatures may have been as high as 18°C to 22°C. During the cool periods, one or more of the poles was covered by permanent ice sheets and the average global temperatures were as low as 12°C to 14°C. For comparison, the average global temperature today is approximately 14°C (U.S. National Climate Data Centre).
Figure 3-7: Summary of global climate change during the Phanerozoic (from Rohde, 2009, based on Veizer et al., 1999)

Gondwana formed (650 Ma) during the protracted Vendian Ice Age, which was followed by warmer conditions in the early Palaeozoic as shown in Figure 3-7. Cool conditions returned briefly during the late Ordovician and early Silurian. This was followed by warmer global temperatures until the onset of the Permo-Carboniferous glaciations which is of particular significance to this thesis. With the exception of a relatively short period of moderate global cooling during the Late Jurassic and Early Cretaceous, the global temperatures remained high until the Miocene when gradual global cooling culminated in the Pleistocene glaciations.

Figures 3-8 and 3-9 show six geographic and climatic reconstructions at strategic times during the Palaeozoic and Mesozoic Eras (Scotese, 2002). During the Cambrian Period the African plate was rotated over 90° anticlockwise compared to today and located in the interior of Gondwana at a mid-latitude in the southern hemisphere (Figure 3-8a). During the Ordovician, the plate drifted south and by the middle Silurian the South Pole was located beneath what is now the edge of north-west Africa (Figure 3-8b). Late Ordovician tillites occur throughout North Africa, thus providing evidence of glacial conditions at that time. During the Devonian, the plate rotated clockwise and began its
northward journey towards the equator. Throughout the Palaeozoic Era central and East Africa were located in the continental interior and marine deposition was confined to the north and far south of the African plate. As a consequence, no early Palaeozoic rocks are found in Uganda today. By the late Carboniferous the South Pole was located on the southern edge of Africa, however, global cooling resulted in widespread glaciation across much of the continent (Figure 3-8c). Its maximum northern extent is indicated by the presence of tillites in Sudan and on the Arabian peninsula (Scotese, 2002). Glacial erosion likely shaped the late Palaeozoic landscape of east and central Africa prior to glacial and pro-glacial deposition. The oldest surviving Phanerozoic rocks of Uganda were deposited in a cold terrestrial environment during the late Carboniferous or early Permian (Schlüter et al., 1993).

Rift basins formed in the region now occupied by eastern Africa during the Permian and Triassic (Figure 3-9a). These likely represent early failed events in the break-up of Gondwana and are today filled by a large thickness of upper Palaeozoic and Mesozoic deposits (Schlüter, 1997). Contemporaneous deposition also continued in the subsiding Congo Basin (Giresse, 2005).
Figure 3-8: Palaeogeographic reconstructions for: a) Late Cambrian; b) Mid Silurian; and c) Late Carboniferous (after Scotese, 2002)
Figure 3-9: Palaeogeographic reconstructions for: a) Early Triassic; b) Early Jurassic; and c) Mid Cretaceous (after Scotese, 2002)
The actual break-up of Gondwana began during the Jurassic with the opening of the North Atlantic around 185 Ma. This was followed by the initial development of the Mozambique Channel and Indian Ocean around 160 Ma (Figure 3-9b), which was accompanied by marine deposition on the coastal margin of current day Kenya and Tanzania (Schlüter, 1997). The South Atlantic finally began to open in the middle Cretaceous at about 130 Ma and by the late Cretaceous the newly isolated continent of Africa had began to resemble its current form (Figure 3-9c).

The northward drift of the African continent slowed during the Mesozoic and this arrested continental motion has continued into the Cenozoic. Burke and Gunnell (2008) have proposed that the mantle plumes associated with the Karroo and Afar Large Igneous Provinces (LIPs), which erupted at approximately 183 Ma and 30 Ma, are responsible for plate pinning. Warm global climate persisted throughout most of the Mesozoic with moderate cooling in the late Jurassic and early Cretaceous (Scotese et al., 1999). The slow plate movement has resulted in different parts of the continents remaining within the same climatic zones for long durations.

3.4.2 Overview of the Karoo sedimentary basins

The term ‘Karoo’ is used in African geology to refer to a stratigraphic supergroup of continental deposits ranging in age from Upper Carboniferous to Lower Jurassic (Schlüter, 1997). The Karoo Supergroup is named after the area of south east South Africa where the type succession is found. Karoo-age sediments were deposited in tectonic basins widely distributed across Africa as shown in Figure 3-10 (Petters, 1991, Schlüter et al., 1993). The Karoo basins have varied origins which are related to the tectonic regimes associated with different regions of the African plate (Catuneanu et al., 2005):

1. Southern Africa – The main Karoo Basin developed under compression and accretion associated with closure of the Panthalassan Ocean along the southern margin of Gondwana within a retro-arc foreland setting adjacent to the Cape Fold Belt;

2. Eastern Africa – The rift basins which include those of Tanzania, Malawi and Kenya developed under extensional stresses propagated into the continent from the divergent margin of the Tethys Ocean; and
3. Western and northern Africa – The Karoo basins of northern and western Africa, including the Congo Basin, are interpreted to have formed due to intra-cratonic thermal sag.

The pattern of global climate change is superimposed on the tectonic control in the stratigraphic record. There is a general shift from cold semi-arid conditions in early Karoo times to warmer and eventually hot conditions with fluctuating precipitation in the rest of Karoo period (Catuneanu et al., 2005). The type succession of the main Karoo Basin is divided into the five groups based on their lithological characteristics, known as the: Dwyka Group; Ecca Group; Beaufort Group; Stormberg Group; and Drakensberg Group. The type stratigraphy of South Africa sets the context for understanding deposition of a similar age in central Africa closer to the Katonga Valley, and is summarized in Table 3-2.

The Permo-Carboniferous rocks at the western end of the Katonga Valley lie approximately 200 km from both Karoo-age rocks in the Entebbe graben and those infilling the Ituri valley in the D.R. Congo. To facilitate later interpretation of the Katonga beds, the Karoo and Mesozoic sediments within the basins of western and eastern Africa are described in more detail in the following sections. Table 3-3 shows the stratigraphic age range of the east and central African Karoo basins in comparison to the type sequence from the main Karoo Basin of South Africa.
<table>
<thead>
<tr>
<th>PERIOD</th>
<th>AGE (Ma)</th>
<th>GROUP</th>
<th>FILL PHASE</th>
<th>LITHOLOGY AND ENVIRONMENT OF DEPOSITION</th>
</tr>
</thead>
<tbody>
<tr>
<td>Early Jurassic</td>
<td>179</td>
<td>Drakensberg</td>
<td>Over filled Karoo Basin (non-marine)</td>
<td>Sedimentation ended with the outpouring of basalt lava flows of the Karoo large igneous province.</td>
</tr>
<tr>
<td></td>
<td>183</td>
<td></td>
<td></td>
<td>Initial cool conditions accompanied the production of a coarse-grained fluvial molasse, deposited by high-energy braided rivers fed from the Cape Fold Belt to the south.</td>
</tr>
<tr>
<td>Triassic</td>
<td>230</td>
<td>Stormberg</td>
<td></td>
<td>The onset of arid conditions during the early Jurassic resulted in freshwater lake deposits to the north being overlain by cross-bedded aeolian sandstone as well as playa lake, sheet flood and ephemeral stream deposits.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(hiatus)</td>
<td>Basin filling resulted in the deposition of sandy lithofacies by low to high sinuosity rivers in high-energy fluvial environments on the southern basin margins. Mudstones accumulated in inter-channel floodplains and lacustrine environments in the centre and northern areas of the basin.</td>
</tr>
<tr>
<td>Permian</td>
<td>237</td>
<td>Beaufort</td>
<td>Filled (shallow marine)</td>
<td>Warming climate resulted in transgression by an epicontinental sea resulting in the initial deposition of fossiliferous, phosphatic shales. Extensive alluvial coal swamps occurred along the northern margin and deltaic sediments prograded into the basin where flysch deposits and turbidites accumulated in deeper water.</td>
</tr>
<tr>
<td></td>
<td>255</td>
<td></td>
<td></td>
<td>Up to six upward fining glacial/interglacial cycles of diamictite (tillites), fine-grained cross-laminated sandstone (outwash), siltstones and laminites (glacio-lacustrine varves) overlie striated surfaces and infilled palaeovalleys.</td>
</tr>
<tr>
<td>Late Carboniferous</td>
<td>289</td>
<td>Ecca</td>
<td>Under filled Karoo Basin (glacial to deep marine)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>300</td>
<td>Dwyka</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 3-2: Stratigraphy of the Karoo Supergroup in the main Karoo Basin of South Africa (summarised from Catuneanu et al., 2005, Petters, 1991)
<table>
<thead>
<tr>
<th>Period</th>
<th>South Africa Karoo Supergroup</th>
<th>Malawi Tanzania</th>
<th>Tanzania</th>
<th>Kenya Tanzania</th>
<th>Uganda</th>
<th>DRC</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Group</td>
<td>Env. of Deposition</td>
<td>Galula, Tukuya &amp; Rukwa Basins</td>
<td>Ruhuhu Basin</td>
<td>Luwegu Basin</td>
<td>Tang and Mombasa Basin</td>
</tr>
<tr>
<td>Jurassic</td>
<td>Drakensberg</td>
<td>Volcanic</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Stormberg</td>
<td>Aeolian and fluvial</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Triassic</td>
<td>Beaufort</td>
<td>Fluvial and lacustrine</td>
<td></td>
<td>K8</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ecca</td>
<td>Flysch and deltaic</td>
<td></td>
<td>K7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Permian</td>
<td>Dwyka</td>
<td>Glacial</td>
<td></td>
<td>K6</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>K5</td>
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<td>K4</td>
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<td>K3</td>
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<td></td>
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<td></td>
<td></td>
<td>K2</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>K1</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 3-3: Comparative stratigraphic range of Karoo Basins (modified after Schlüter, 1997)
3.4.3 Karoo rift basins of East Africa

The Karoo rift basins of Kenya and Tanzania lie approximately 900 km east and south of the Katonga Valley on the opposite side of the Tanzanian Craton (Figure 3-10). The basal sediments of these basins have similar characteristics to the small Karoo outcrops.
in Uganda (Schlüter et al., 1993). An understanding of the structure and stratigraphy of the Karoo grabens to the east and south of the Tanzanian craton assists in placing the Karoo deposits of Uganda into the regional context.

3.4.3.1 Large northeast-southwest trending basins

As shown in Figure 3-10, the large Karoo basins of East Africa occupy a NE-SW trending corridor extending from Kenya on the Indian Ocean to Lake Malawi, with continuations in the Luangwa Basin of Zambia and the Metangula Basin of Mozambique (Verniers et al., 1989). Tanzanian Karoo deposits rest unconformably on the Mesoproterozoic Ubendian and Usagaran Systems. Wopfner and Diekmann (1996) suggest that the pre-Karoo surface was one of low relief due to the long preceding period of erosion, and tectonic modification was required to facilitate erosion and filling of depositional basins. They speculate that a heat anomaly within the insulated mantle beneath Pangaea (Veevers, 1990) resulted in doming which initiated riftting. Glacial erosion accentuated the palaeorelief and biostratigraphic evidence suggests sedimentation started in the Gzhelian (304 to 299 Ma), which is the last Stage of the Carboniferous Period. In East Africa, the Lower Karoo glaciogenic sediments are best exposed in south west Tanzania where they have been elevated along the rift shoulders of the Ruhuhu, Songwe-Kiwira and Galula Basins (Wopfner and Diekmann, 1996).

Stockley and Oats (1931, cited in, Schlüter, 1997) divided the sediments in the Ruhuhu Basin (Figure 3-10) into eight units (K1 to K8), which have been given the formation names shown in Table 3-4. These provide the reference sequence for the Tanzanian Karoo deposits. Comparison of Tables 3-2 and 3-4 reveals that despite the different tectonic settings, the climatic control has resulted in some broad similarities between the lithostratigraphy of East Africa and that of the main Karoo Basin in South Africa. The Dwyka Group and Idusi Formation (K1) are both glacial in origin, and the overlying Ecca Group and Mchuchuma Formation (K2) each contain coal measures. Whilst the detailed stratigraphy varies according to the local topography and tectonics, the global climatic conditions during the late Permian and early Triassic appears to have facilitated continental fluvial and lacustrine deposition in both the Beaufort Group and units K3 to K8 of the Ruhuhu basin in Tanzania.
<table>
<thead>
<tr>
<th>Period</th>
<th>SA Groups</th>
<th>Tanzanian Formations</th>
<th>Lithology and Environment of Deposition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Triassic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>K8</td>
<td>Manda (≈1000m)</td>
<td>Fifty cycles of coarse to medium-grained feldspathic sandstone becoming finer and bioturbated before being overlain by red and green mudstones and siltstones. <em>Deposited by meandering rivers with rift flank uplift.</em></td>
</tr>
<tr>
<td></td>
<td>K7</td>
<td>Kingori (≈380m)</td>
<td>Thick coarse-grained and current bedded quartz arenites. <em>Deposited in high-energy braided rivers initiated by renewed tectonism and humid climate.</em></td>
</tr>
<tr>
<td></td>
<td>K6</td>
<td>Usili (≈300m)</td>
<td>Lower limestone bands and fossiliferous nodular horizons are overlain by argillaceous sediments. <em>Deposited on playas with elastic sediments supplied by flooding.</em></td>
</tr>
<tr>
<td></td>
<td>K5</td>
<td>Ruhuhu (400-600m)</td>
<td>Frequent grey-green mudstones, calcareous siltstones and marls with rare limestones. <em>Deposited in large shallow lakes during warm and humid climate.</em></td>
</tr>
<tr>
<td></td>
<td>K4</td>
<td>Mhukuri (≈315m)</td>
<td>Black to grey carbonaceous mudstones and siltstones with coal beds and a few sandstone beds. <em>Deposited in a low energy flood plain environment and localised coal swamps without recognisable cyclothems.</em></td>
</tr>
<tr>
<td></td>
<td>K3</td>
<td>Mbuyura (200-300m)</td>
<td>Comprised of three units. Scarp Sandstone Member: coarse-grained arkose with thin lenses of coal and shale. Green Beds: alternating green fine-grained sandstone and red mudstone &amp; siltstone. Red Beds: red arkose with mud cracks, palaeosols and gypsum pseudomorphs. <em>Depositional conditions evolved from braided rivers to semi-arid flood plain playa environment.</em></td>
</tr>
<tr>
<td></td>
<td>Ecca Group</td>
<td>K2 Mchuchuma (110-280m)</td>
<td>Subdivided into three lithofacies, including basal sandstones, middle coal-bearing facies and upper coal/shale facies. <em>Deposited by meandering rivers in warming climate with luxuriant vegetation.</em></td>
</tr>
<tr>
<td></td>
<td>Dwyka Group</td>
<td>K1 Idusi (0-250m)</td>
<td>Up to 8m diamictite described as greenish grey, hard, pelitic matrix with sub-rounded to rounded clasts up 60cm. Overlain by 40m of well bedded, slightly calcareous siltstones and rythmites with deformation. <em>Deposited by valley glaciers and in pro-glacial lakes.</em></td>
</tr>
</tbody>
</table>

Table 3-4: Stratigraphy of the Karoo Supergroup in the Ruhuhu Basin of Tanzania (after Stockley and Oats, 1931, Kreuser and Markwart, 1989, cited in Schlüter, 1997)
The oldest sediments in the Luwegu Basin (Figure 3-10) are the Late Permian feldspathic sandstones and black shales deposited in a deltaic lacustrine environment. As shown in Table 3-4, the overlying sequence of continental clastic deposits is of equivalent age to the Triassic and Lower Jurassic Beaufort and Stormberg Groups of South Africa. This thick sedimentary sequence is representative of various environments of deposition within the rift, including alluvial fans, braided rivers, deltas, floodplains and lakes.

The Tanga and Mombasa Basins of the Tanzanian and Kenyan coast contain Karoo sediments dating from the Early Permian to Jurassic, together with a fringe of younger marine Mesozoic and Neogene strata (Schlüter, 1997). The lower unit of the basal Taru Formation is a tillite-like deposit of unsorted, angular, faceted, polygenetic clasts up to 5 cm across in a matrix of medium-grained angular phenoclasts and fine-grained silt and clay. The greatest thickness appears are preserved in depressions in the surface of the underlying Neoproterozoic rocks. Varve-like deposits are also recorded. These glacial deposits are overlain by arkose grits and carbonaceous shales and conglomerates which appear to have been deposited in alluvial fans. The periods of carbonaceous deposition were apparently too short to facilitate the formation of even thin coal seams. Nevertheless, these lower units of the Taru Formation can be seen to have some similar facies characteristics to the Dwyka and Ecca Groups of South Africa.

The upper unit of the Taru Formation has an increasingly calcareous character and was likely deposited in a deep lacustrine environment. The overlying Triassic and early Jurassic formations contain a thick sequence of shallow lacustrine, fluvial and deltaic clastic deposits. The upper unit appears to indicate the onset of aeolian conditions before the continental Karoo sedimentation ended in the Mombasa Basin with the Middle Jurassic marine transgression associated with the development of the Mozambique Channel (Schlüter, 1997).

3.4.3.2 Small Northwest-Southeast Trending Basins

Several small basins associated with zones of long lived crustal weakness, including the Galula, Tukuya, and Rukwa Basins, extend northwards from Lake Malawi towards the south east margin of Lake Tanganyika. Sedimentary basins in this region first formed during the Mesoproterozoic in the Palaeoproterozoic Ubendian belt on the western edge
of the Tanzanian Craton (Delvaux, 2001). They were reactivated during three Phanerozoic episodes of extension and basin filling, including: the Karoo rifting event; a late Mesozoic-Cenozoic phase; and the Late Cenozoic-Recent event associated with Western Rift of the East African Rift System.

As shown in Table 3-3, the Karoo deposits of the Tukuya and Rukwa Basins are limited to the Upper Carboniferous and Permian (Schlüter, 1997). The Karoo sediments are approximately 2,000 m thick in the Galula Basin and are overlain by about 7,000 m of Cenozoic sediments in the Rukwa Basin. In the Tukuya Basin, Wopfner and Diekmann (1991, cited in Schlüter, 1997) identified a morainal angular conglomerate (tillite) which can be seen interfingering pro-glacial sandur facies comprising conglomerates overlain by sandstone and then varvites. Wopfner and Diekmann (1996) suggest that erosion during the Carboniferous resulted in 300 to 500 m of relief with a NW-SE alignment. In the Rukwa Basin, the K1 unit (Dwyka equivalent) comprising poorly sorted conglomerates overlain by laminated siltstones and very fine sandstones rests non-conformably on often highly weathered basement. The Permian K2 to K5 units are all represented in the group of north west to south east trending basins, but the Triassic K6 to K7 have not been observed.

3.4.4 The Congo Basin

The Karoo sediment in the Ituri River valley on the eastern margin of the Congo Basin lie only 200 km north west of the Katonga Valley. Therefore, the history of deposition in the Congo Basin may be of greater relevance to understanding the origin of Phanerozoic sediments in the Katonga Valley than the history of deposition in the more distant rift basins on the opposite side of the Tanzanian Craton. The deep structure and stratigraphy of the Congo Basin provides evidence of periods of compression as well as subsidence. The age and character of these episodes of deformation and erosion recorded in the structures and unconformities of the Congo Basin therefore set the historical context to the denudation history of the region.

Unfortunately, decades of political instability in the D.R. Congo have prevented many recent studies and much of the available geological information pre-dates independence. The summary provided here is largely based on papers by Daly et al. (1992) who reviewed the tectonic structure in the light of contemporary seismic and
borehole data, Cahen and Lepersonne (1981) who summarised the character of the Karoo deposits based largely on their own pre-independence fieldwork (Cahen, 1954), and Giresse (2005) who compiled and rationalised the state of knowledge of the Mesozoic and Cenozoic history of the Congo Basin.

3.4.4.1 Basin Structure

The Congo Basin is surrounded by the West Congolian fold-belt to the west, Proterozoic and Archaean crystalline basement to the north, the Kibaran structural belt to the east and the Proterozoic Lukoshien basement and sediments of the Lufilian Arc to the south. There are two sub-basins which include the Kwango Basin in the southwest and the Cuvette Central closer to Uganda in the northeast. Based on four deep borehole logs and 2,900 km of seismic reflection surveys conducted between 1973 and 1981, Daly et al (1992) identified six seismically defined sequences in the Cuvette Centrale as follows:

Z – Proterozoic-Archaen basement.

1 – Infra-Cambrian carbonate-evaporite sequence.

2 – Cambrian sequence of fining upward marine clastic sediments.

3 – Ordovician marine sandstones and shales, overlain by Devonian arkosic sandstones.

4 – Permo-Carboniferous glacial deposits.

5 – Triassic to Cenozoic series of continental clastic rocks.

Based on their interpretation the seismic reflection surveys using these sequences they postulate the structural evolution of the Cuvette Centrale summarised in Figure 3-11. Daly et al. (1992) suggest that the basin was originally formed by late Proterozoic NW to SE oriented rifting and post-rift thermal subsidence and sediment loading. However, this is speculative as although potential rift fill sediments have been identified, no structure can be defined on the seismic surveys below sequence 1.
Early Palaeozoic compression associated with a late stage of the Pan-African thermotectonic event resulted in deformation of the ‘Infra-Cambrian’ (i.e., immediately preceding the Cambrian) and Cambrian sediments. According to the model of Daly et al. (1992) subsidence resumed after the Cambrian and the basin was filled by the Ordovician marine and Devonian continental clastic sediments of sequence 3. Deposition continued with the Lower Karoo sediments of sequence 4. On the eastern margin of the Congo Basin, these Permo-Carboniferous glacial deposits directly overlie the Precambrian rocks. Due to their structural and seismic continuity, sequence 3 and 4 are shown together in Figure 3-11.

Daly et al. (1992) suggest the occurrence of both a late Palaeozoic (Permo-Triassic) compressional event and a more significant mid-Jurassic regional uplift event. The evidence for the late Palaeozoic event is relatively subtle with mild fault reactivation suggested by small truncations below sequence 5. The characteristics of the post-Carboniferous stratigraphy of the Congo Basin is summarised in Table 3-5. Giresse (2005) states there are no Upper Karoo sediments in the Congo Basin and therefore only recognises a single unconformity between the Permian and Jurassic rocks.

**Figure 3-11: Structural evolution of the Cuvette Centrale in the Congo Basin**
### Periods and Groups

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Neogene</strong></td>
<td></td>
<td>Pliocene-Pleistocene sands and gravels. Infilling valleys.</td>
<td><strong>Kalahari System:</strong> Sables Ocre. 120m Fine sands without stratification or silicification. Often ochre coloured. Both rounded and angular grains, likely fluvial.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(Unconformity: dipping erosion surface)</td>
<td><strong>Kalahari System:</strong> Grès polymorphes. 80m. Basal conglomerate containing draikanter, overlain by pale sands and sandstones. Thick to finely bedded and cross bedding. Diatomaceous limestones and sand textures suggest arid climate pans and eolian deposition.</td>
</tr>
<tr>
<td><strong>Palaeogene</strong></td>
<td></td>
<td><strong>Kalahari System:</strong> Kamina Series. Sand, sandstone and gravel.</td>
<td>(Unconformity: dipping erosion surface)</td>
</tr>
<tr>
<td></td>
<td>Kwango Fmtn</td>
<td><strong>Kwango Series:</strong> Nsélé Group. 100m. Coarse-grained varied deposits with evidence of eolian deposition.</td>
<td><strong>Kwango Series:</strong> Inzia Group. Up to 200m. Basal conglomerate with arkosic matrix overlain by fine silty sand with marine fossils.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(Unconformity: slight uplift, deformation and erosion)</td>
<td><strong>Loia Fmtn</strong></td>
</tr>
<tr>
<td><strong>Cretaceous</strong></td>
<td></td>
<td><strong>Lualaba Series:</strong> Stanleyville Group. 350m. Conglomerate on uneven surface, overlain by continental bitumen sandstone, shale and marls</td>
<td>(Unconformity: irregular erosion surface)</td>
</tr>
<tr>
<td></td>
<td>Bokungu Fmtn</td>
<td><strong>Lukuga Group:</strong> Claystone, sandstone, and shale swathe coal beds and conglomerate at the base. Corresponds to the lower Beaufort and Ecca on the basis of mega and microflora.</td>
<td><strong>Stormberg Group:</strong> Haute Lueki Fmtn</td>
</tr>
<tr>
<td><strong>Jurassic</strong></td>
<td></td>
<td><strong>Upper Lukuga Group:</strong> Claystone, sandstone, and shale swathe coal beds and conglomerate at the base. Corresponds to the lower Beaufort and Ecca on the basis of mega and microflora.</td>
<td>(hiatus)</td>
</tr>
<tr>
<td></td>
<td></td>
<td><strong>Lower Lukuga Group:</strong> Tillites, with sandy or sandy clay matrix and large disseminated polygenetic clasts. Feldspathic sandstones, conglomeratic beds, shales and varve-like sediments with contorted beds. Plant fossils.</td>
<td><strong>Beaufort Group</strong></td>
</tr>
<tr>
<td></td>
<td></td>
<td>(hiatus)</td>
<td>(deformation)</td>
</tr>
<tr>
<td><strong>Triassic</strong></td>
<td></td>
<td><strong>Lukuga Fmtn</strong></td>
<td><strong>Lukuga Fmtn</strong></td>
</tr>
<tr>
<td></td>
<td></td>
<td><strong>Lukuga Fmtn</strong></td>
<td><strong>Lukuga Fmtn</strong></td>
</tr>
<tr>
<td><strong>Permian</strong></td>
<td></td>
<td><strong>Lukuga Fmtn</strong></td>
<td><strong>Lukuga Fmtn</strong></td>
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<td></td>
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<td><strong>Lukuga Fmtn</strong></td>
<td><strong>Lukuga Fmtn</strong></td>
</tr>
<tr>
<td><strong>Carboniferous</strong></td>
<td></td>
<td><strong>Lukuga Fmtn</strong></td>
<td><strong>Lukuga Fmtn</strong></td>
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<td></td>
<td></td>
<td><strong>Lukuga Fmtn</strong></td>
<td><strong>Lukuga Fmtn</strong></td>
</tr>
</tbody>
</table>

Table 3-5: Stratigraphy of the Karoo and Mesozoic Deposits in the Congo Basin (Daly et al., 1992, Giresse, 2005, Cahen and Lepersonne, 1981)
The mid-Jurassic regional uplift resulted in the Kiri High and the Lonkonia High on the line of the original northwest to southeast rift fault axis of the Cuvette Centrale. As shown in Figure 3-11, the pre-Triassic rocks have been completely eroded in the areas occupied by these highs. Post mid-Jurassic subsidence accommodated sedimentation of the Mesozoic sediments above this unconformity (Daly et al., 1992). Giresse (2005) identifies three additional but less significant unconformities in the Mesozoic stratigraphy (Table 3-5). Evidence from the Congo Basin therefore suggests significant uplift and denudation in the late Palaeozoic or early Mesozoic, following deposition of the Lower Karoo glacial deposits. The significance of this event in the exhumation of Karoo sediments further east in the Katonga Valley will be assessed in the context of apatite fission track analysis.

3.4.4.2 Karoo Deposits

The Karoo-age rocks of the Congo Basin belong to the Lukuga Group (Table 3-5). Their distribution on the eastern margin of the basin is shown in Figure 3-12 and extends intermittently from the Ituri Valley approximately 50 km west of the Ugandan border at the southern end of Lake Albert, to the Lubudi Valley approximately 100 km from the Zambian border in the south (Cahen, 1954, Cahen and Lepersonne, 1981).

The type section is in the valley of the River Lukuga which forms the outflow from Lake Tanganyika. The Lukuga Group is divided into two subgroups. The Upper Lukuga comprises reddish to dark grey and black claystones, shales and sandstones with coals beds and a conglomerate at the base. On the basis of palaeontological evidence, it corresponds to the lower Beaufort and Ecca of South Africa. The glaciogenic sediments of the Lower Lukuga are often assumed to correspond to the Dwyka of South Africa although the scarcity of fossil evidence makes this difficult to confirm. In the Lukuga Valley an unconformity occurs between the upper and lower subgroups and elsewhere the lower subgroup is often directly overlain by post-Permian rocks.
In the eastern D.R. Congo, the Lower Lukuga occupies the buried glaciated landscape of the underlying Precambrian terrane which according to Boutakoff (1948) includes U-shaped valleys, hanging valleys, and cirques. It rests on striated surfaces and exhumed roches moutonnées. In the Lukuga Valley the Lower Lukuga is between 100 and 300 m thick and forms three units (Cahen and Lepersonne, 1981) as follows:

- **Upper, mainly glacial beds** – Tillite with matrix of grey sandy clay with large disseminated polygenetic clasts, grading downwards into red, yellowish-grey clayey sandstones, often laminated with: disseminated rounded clasts; sandstones with plant remains; some rocks are contorted or overturned rocks; cemented fissures.
Mainly periglacial and interglacial beds – Black, sometimes grey or purple shales, locally unstratified with disseminated small to large clasts; black to greenish grey shales locally with calcareous nodules and related cone-in-cone structures; sandy to conglomeratic layers; contorted beds and scouring at the top; also infrequent plant remains.

Lower, mainly glacial beds – Tillite with matrix of grey or red sandstone or sandy clay with very large clasts; grey or reddish coarse feldspathic sandstones and conglomeratic beds containing large clasts; interstratified red or purple shales and coarse-grained sandstones with contorted laminae.

In the southern Shaba region of the D.R. Congo, up to five glacial/interglacial cycles of tillites and shales with a total thickness up to about 300 m can be recognised (Cahen, 1954). Approximately 450 m of Lower Lukuga glacial/interglacial sediments, including three tillites, have also been recognised below 715 m depth in an oil exploration borehole at Dekese (3°27´S, 21°24´E) in the central area of the Congo Basin (Cahen and Lepersonne, 1981). North of Lukuga in the Maniema, Kivu and Ituri regions only one tillite can be recognised at the base. The Lower Lukuga of the River Epulu valley (1°25´S, 29°05´E) in the Ituri region, 220 km northwest of the Katonga beds, is described by Sluys (1946, cited in Cahen and Lepersonne, 1981):

- Dark grey and bluish grey shales with fine laminations; overlying

- Tillite comprised of yellowish arkose with sandy beds carrying disseminated clasts which are more abundant at the base.

The megaflora of the Upper Lukuga is characterised by the Permian gymnosperm genus Glossopteris, without Gangamopteris, whilst the Lower Lukuga is characterised by the Permo-Carboniferous genus Gangamopteris without Glossopteris. Noeggerathiopsis occurs in both. Neither megaflora or palynological assemblages can be used to accurately date the Lower Lukuga, although it can be demonstrated that the Upper Lukuga is lower Permian (Bose and Kar, 1978, cited in Cahen and Lepersonne, 1981).

The facies characteristics of the Lower Lukuga clearly attest to its glacial, fluvioglacial and glacio-lacustrine origins. Cahen and Lepersonne (1981) state that the distribution
of glacial valleys and striated surfaces suggest that glaciers flowed down from a north-south oriented mountain range between 1°30´S, 28°15´E in Walikale and 6°00´S, 28°45´E in Lukuga (Figure 3-12). However, the strength of evidence in favour of this apparently fortuitous preservation of the ice divide on the eastern margin of the Karoo outcrop is unclear. Wopfner (1991) places the western margin of the proposed highland area in a similar location to Cahen’s and Lepersonne’s (1981) ice divide. Nevertheless, they both agree that the majority of the preserved palaeovalleys are on the western flank of the proposed range and flow west towards the Congo Basin. The considerable thickness in the Dekese borehole supports the interpretation that the mountain glaciers coalesced to form large piedmont glaciers at lower altitudes (Boutakoff, 1948, cited in Cahen and Lepersonne, 1981).

3.4.4.3 Mesozoic Deposits

It is unclear if the post-Palaeozoic rocks of the Congo Basin are contemporaneous with any part of the strata preserved in the river valleys of western Uganda. The Mesozoic and Cenozoic stratigraphy is briefly described in Table 3-5 and the environments of deposition are summarised here. Giresse (2005) summarises the palaeoclimatic history of the Congo Basin based on facies and mineralogical characteristics. From the early to late Permian and Triassic, the climate changed from glacial and semi-arid to tropical, and remained tropical throughout the Jurassic. Hot and slightly arid conditions accompanied by the development of extended lacustrine and lagoonal environments occurred during the early Cretaceous, and were followed by more humid conditions in the middle Cretaceous. A drying trend in the Late Cretaceous and Palaeogene led to aeolian deposition. Feldspars are generally abundant in the Mesozoic Lualaba and Kwango Series (see Table 3-5), but become rare in the Palaeogene Grès Polymorphe Series (Cahen, 1954). The Neogene Sable Ocres series are fluvial deposits which broadly correspond to the onset of a wetter climate.

During the Jurassic and Cretaceous the lacustrine or lagoonal depositional environments experienced relatively short intermittent marine incursions. Giresse (2005) suggests the first marine incursions occurred during the late Jurassic (Kimmeridgian) as evidenced by marine fossil fish in the Stanleyville Group of the Lualaba Valley in the northeast Congo Basin. Marine fossils of this age are absent on the western margin of the basin and therefore Giresse suggests this transgression
occurred from the east and was associated with the opening of Somali and Mozambique rifts. This conclusion is of potential relevance to the geological history of adjacent Uganda. However, given the absence of marine sediments of a similar age between the Congo Basin and the Indian Ocean, this hypothesis appears speculative. Late Jurassic/Early Cretaceous marine bivalves in the Kwango Series of south-central Congo Basin suggest a second incursion from a similar direction to Kimmeridgian transgression. Younger marine fauna in the Kwango Series of the northern and west-central Congo Basin indicate two later marine incursions during the middle and late Cretaceous, which Giresse (2005) suggests occurred via a northwest connection to the Tethys.

3.4.5 Previously recorded Karoo sedimentary rocks of Uganda

Three small exposures of Karoo age strata have previously been identified on the northern margin of Lake Victoria in Uganda at Entebbe, Dagusi Island and Bugiri (Pallister, 1959, Schlüter et al., 1993, Schlüter, 1997). They are labelled E, D and B in Figure 3-12. In addition, to these well established Karoo outcrops, there are several other potential occurrences of Karoo-age rocks in Uganda. These include subsurface strata in the Western Rift (Schlüter, 1997, Schlüter, 2006) and small outliers of indurated sedimentary rocks in the southwest river valleys (Wayland, 1925, Pargeter, 1948, Bishop, 1969).

3.4.5.1 The Entebbe, Dagusi Island and Bugiri Karoo outcrops of Uganda

The Karoo outcrop at Entebbe is about 1.5 km wide and extends approximately east to west across the peninsula. The relationship with the Buganda Series schists to the north and the gneisses to the south are obscured by the superficial cover of lateritic gravels, beach deposits and weathered bedrock. It is assumed that the contacts are faults, and this appears to be confirmed by later drilling and a gravimetric survey conducted in 1950 which suggested a thickness of approximately 470 m in the east, and 415 m in the west (Brown, 1950, cited in Schlüter et al., 1993). The floor of the interpreted graben is inferred to rise slightly in a southerly direction.

Variegated rhythmite clays, assumed to be varves, have been reported near the western tip of the peninsula (Davies, 1948, Harris, 1952). Schlüter et al. (1993) describe an exposure in a 10 m high lakeshore cliff to the east of the Karoo outcrop. The lower
section consists of well-bedded, greyish blue mudstones and siltstones with pink, yellow and white laminae also present. Above about 5 m, the rocks become increasing weathered and are capped by laterite. The author has also observed interlaminated grey and yellow rhythmites with soft sediment deformation on the foreshore near this location.

A borehole sunk to 357 m in the 1920s encountered white, red and buff shales becoming grey to black at depth with pyritic horizons and traces of carbonaceous material. Some sandy and calcareous beds were also encountered, but the base of the sedimentary strata was not reached. A borehole drilled in 1965 to a similar depth encountered equivalent deposits and three thin diamictites, assumed to be tillites, at 77, 79 and 80 m below ground level (bgl) (Dixon, 1969).

Five boreholes were drilled on the 6 km² Dagusi Island, 3 km off the north east shore of Lake Victoria. Drilling showed the Karoo deposits to be 36 m thick in the centre and approximately 170 m thick on eastern side of the island. Buff to dark grey shales were found in the upper half of both successions becoming sandy and then gritty with depth (Davies, 1954). They rest on granites.

A small outcrop of Karoo deposits, identified by the presence of Glossopteris, also occurs at Bugiri approximately 35 km east of Jinja. Several boreholes were drilled. One first encountered laterite and then buff coloured shales from 11.3 to 21.3 m bgl and another encountered laterite and then sandstones from 2.4 to 23.8 m bgl (Dixon, 1969). They rest on Nyanzian metasediments.

Similarly to the Karoo-age rocks in the Congo Basin, the Uganda Karoo deposits include the genera Glossopteris, Noeggerathiopsis and Gangamopteris (Seward, 1924 and Du Toit, 1932, cited in Schütler et al., 1993). Gangamopteris is an index fossil of the Upper Carboniferous to Lower Permian. Noeggerathiopsis is recorded in the Ecca Group of South Africa. Glossopteris has a wider stratigraphic range and is therefore less useful as a zone fossil.

Schütler et al. (1993) appear to suggest an inconsistent correlation between the Uganda Karoo, the Ecca Group of South Africa, the Idusi Formation of Tanzania and the Lower Lukuga of the Congo Basin. These final two units have previously been correlated with
the Dwyka Group rather than the Ecca Group, as shown in Tables 3-4 and 3-5. Elsewhere in Africa, tillites are confined to the Dwyka Group and its equivalents. The purported tillites in the Entebbe borehole logs contain quartzite and gneiss pebbles and cobbles embedded in banded gritty black shale and are associated with varve-like laminites (Dixon, 1969) (cited in Schlüter et al., 1993). Unfortunately they are each only of the order of a metre thick and no striated clasts were reported making their interpretation uncertain. The plant debris and coal partings suggest sufficiently warm and humid climate to support abundant vegetation which elsewhere in Africa is characteristic of the Ecca Group and its equivalents (Schlüter et al., 1993). Gangamopteris, which was only identified in the Lower Lukuga (Dwyka equivalent) in the Congo Basin, was found in the Entebbe Karoo succession. The mixed evidence would suggest that the Uganda Karroo could correlate with either the Dwyka or lower Ecca Groups.

Like previous authors (Brown, 1950, Dixon, 1969), Schlüter et al (1993) interpret the Karoo deposits of Uganda to be preserved in grabens even though no fault contacts have been observed directly. The reported dip on the bedding does not appear consistent with a synclinal fold structure and the depth to width ratio of almost 1:3 of the Entebbe structure appears higher than might be expected for valley fill. At Entebbe the contacts strike east-west, and at Bugiri they strike southwest-northeast.

Schlüter et al. (1993) argue in favour of similarities between the Uganda Karoo grabens and those of Tanzania on the basis of such evidence as similar orientation of the Buguri outcrop, steeply dipping sediments and asymmetric structures. They ignore the small size of the Ugandan exposures and the inconsistent east-west orientation of the Entebbe outcrop. As shown in Figure 3-10, the well established Uganda Karoo exposures are located between the East African Karoo rift basins, located 700 to 900 km east and south, and the Karoo deposits on the eastern margin of the Congo Basin, located 400 to 500 km to the west. The rift basins to the south and south east are also located on the opposite side of the Tanzanian Craton and may therefore be representative of a different tectonic regime.
3.4.5.2 Potential Karoo deposits in the Western Rift

Schlüter (1997, 2006) has suggested that seismic and drilling data indicate subsurface Karoo strata in the Western Rift (Figure 3-10). However, despite considerable commercial drilling for oil the published data are still unclear as to whether Karoo and Mesozoic rocks underlie the well established Neogene strata of the Western Rift.

Hopwood and Lepersonne (1953, cited in Harris et al., 1956) describe a local exposure, apparently underlying the Miocene sediments in the Semliki Basin of the D.R. Congo to the west of the Rwenzori mountains in the Western Rift. Although much disturbed, they provisionally considered this series to be Karoo-age due to its general resemblance to the Lukuga series 40 km west in the Irumu Basin.

Harris et al. (1956), summarise the early evidence of oil in the Western Rift and include several borehole logs from the eastern shore of Lake Albert. The deepest and most northerly borehole, Waki-B1 well, was drilled in 1938 to a total depth of 1,232 m bgl to the northeast of Lake Albert, near Butiaba north of Kaiso (Figure 4-35). The upper 1,012 m bgl are ascribed to the Neogene Kisegi and Kaiso Formations (see Appendix E). Between 1,012 m bgl and the crystalline basement encountered at 1,222 m bgl, the log shows a sequence of oil shales, shales, clays and rare sandy beds with ferruginous sand and scattered pebbles near the base. It is suggested that this lower sequence may be Miocene. A brief unpublished report by the Ugandan Department of Geological Survey and Mines (D.G.S.M., 2005) shows an reinterpretation of the Butiaba Waki-B1 well with the sequence from 1,012 and 1,222 m bgl labelled as the Jurassic Waki Formation. It is described as ‘fluvio-lacustrine clastics possibly equivalent to Upper Jurassic Stanleyville formation of the Cuvette Congolaise’ and includes bituminous shale.

The unpublished D.G.S.M. report also presents a schematic cross-section based on preliminary interpretation of seismic reflection data which shows up to 1.5 km of Karoo-age strata, and 1 km of Cretaceous strata, in the western side of the Albertine Graben. Their combined thickness is shown to reduce eastwards to less than 1 km where they form the hanging wall of the westerly dipping rift margin fault. They are shown to be absent to the east of the Waki-B1 borehole where Tertiary strata sit directly on the rift margin fault. However, without peer review of up-to-date information, this
entire interpretation can only be regarded as historical speculation. A recent conference abstract by representatives of Tullow Oil (Ovington and Burden, 2009) states that oldest sediments encountered by exploratory drilling in the Albertine Graben so far have been Upper Miocene in age. The reservoir interval around northwest Lake Albert is Upper Pliocene and the reservoir interval around southwest Lake Albert is Upper Miocene to Pliocene.

3.4.5.3 Indurated strata of uncertain age in Nkusi and Muzizi Valleys

Wayland (1925) first identified ‘a comparatively soft sandstone’ exposed on the Fort Portal-Hoima road next to the River Muzizi, located about 80 km north of the Katonga Valley. These were initially assigned to the Kaiso Series, but later reassigned to the older Kisegi Series (Pargeter, 1948). Pargeter (1948) describes sandstones which are considered equivalent to the Muzizi Sandstone in the Nkusi Valley, which is the western part of the Nkusi-Kafu Valley System, located about 110 km north of the Katonga Valley. The outcrop which is between about 1.6 km wide extends about 13 km from the rift escarpment from northwest to southeast. About 8 km from the escarpment, the River Nkusi turns sharply to the northeast, whilst the outcrop widens to about 4.8 km and continues southeast.

The Nkusi deposits are uniformly fine-grained laminated sandstones with occasional rounded pebbles, and rare lenses of grit with a maximum thickness of 1.3 cm. They are described as extremely hard with little jointing. The dips are less than 5° except near the valley sides where depositional dips towards the valley can reach 10° to 15°. Ripple marks are common, but current bedding is rare. Weathering has resulted in cylindrical holes and ovoid structures which appear to be associated with concretions. No fossils were found and therefore the age of the Nkusi Sandstone could not be determined. Almost all the characteristics of the Nkusi Sandstone described by Pargeter (1948) are identical to those observed by the author for the Muzizi Sandstone near the outcrop identified by Wayland (1925). Towards the escarpment at Nkusi, the sandstones become much coarser and reddish in colour. The sediments extends to the river bed at the base of the scarp from which Pargeter (1948) estimates the thickness to be about 240 m. A 6 m thick conglomerate occurs about 30 m from the top of the slope.
3.4.6 The sedimentary rocks of the Katonga Valley

3.4.6.1 Kabagole Sandstone

The initial motivation to re-examine the sedimentary rocks of the Katonga Valley was the discovery of fine-grained sandstone adjacent to the eastern gate of the Katonga Wildlife Reserve near Kabagole (Figure 1-3) during reconnaissance of the site. The sandstone forms a mound in the centre of the Katonga Valley, and the subsurface geometry was later investigated during the ERT survey (Section 6.6).

The outcrop at Kabagole is shown in Figure 3-13. It is described in outcrop and hand specimen as a reddish yellow to greyish brown, thinly bedded to laminated, moderately strong to strong, fine-grained sandstone, with weathering related iron oxide Liesegang banding. The dip angle varies between 6° and 12°, and the dip direction varies between 180° (south) and 236° (south west). The author has observed that the Kabagole Sandstone has a similar lithology, strength and structure to the Muzizi Sandstone which forms a similar outlier in the Muzizi Valley 80 km to the north. Figure 3-14 shows a thin section of the fine-grained sandstone in plane polarised light and under crossed polars. Examination reveals that over 25% of clasts are feldspar and it is therefore classified as an arkose. Other clasts are comprised of quartz, mica, clay and heavy minerals. The texture in thin section is fine to very fine (0.25 to 0.063 mm), moderately well sorted, grain supported, low to high sphericity, angular to subangular grains. The cement minerals were not identified, but probably include clays and possibly an iron precipitate such as siderite.

Although the Kabagole Sandstone is moderately well sorted, the angular to subangular grains and abundant feldspar suggest it is a relatively immature sedimentary rock. In general, this implies limited transport and small duration between erosion and deposition, and the sediments are unlikely to have been re-worked. The gently undulating to flat, thin beds and fine sand sized grains indicates that it was deposited in a relatively low energy environment, but with some flow rather than static water conditions. The abundance of unaltered feldspar suggests little weathering occurred during erosion, transport and deposition, which implies the prevailing climatic conditions were either arid and/or subject to low temperatures.
Figure 3-13: *In situ* fine-grained arkose adjacent to Katonga Wildlife Reserve gate near Kabagole

Figure 3-14: Thin section of fine-grained arkose from near Kabagole in a) plane polarised light and b) crossed polars

No evidence of pressure dissolution was observed in thin section and only minor compaction of mica flakes was identified in one or two cases. Thin sections were impregnated with epoxy resin mixed with blue dye to reveal a moderately high porosity of about 20% as shown in Figure 3-15. These characteristics suggest that Kabagole Sandstone has not been buried to a great depth.
Figure 3-15: Thin section of fine-grained arkose from near Kabagole injected with blue dye stained epoxy resin in plane polarised light

There is little evidence to assist age determination. However, having observed both the Miocene Kisegi Formation and the fine-grained arkose at Kabagole, the author’s general impression is that the Katonga Sandstone is considerably more indurated and older in appearance than the Kisegi Formation.

3.4.6.2 Katonga Beds at Bihanga Station

Outcrop location and form

Examination of the 1:250,000 geological map (D.G.S.M., 1962) revealed an outcrop of sedimentary rocks, named the Katonga Beds, had previously been identified 27 km west of Kabagole at the drainage divide (Figure 3-3). The recorded outcrop of the Katonga Beds, which is about 10 km long and 1 to 1.5 km wide, has been reproduced and expanded on the topographic map shown in Figure 3-16. The surface water divide occurs within the region of the valley occupied by the Katonga Beds. About 4 km from the eastern end of the outcrop, the Nyaitanga Valley turns towards the south west, whilst the outcrop of the Katonga Beds continues towards the west. A similar discordance between the outcrop and the present day topography was also noted for the Nkusi Sandstone 110 km to the north. Both outcrops appear related to an older valley form. The Nyaitanga Valley joins the Rusangwe Valley to the south west and eventually forms a tributary of the River Mpanga. The now disused Kampala-Kasese railway, built in the 1950s to service the Kilembe copper mine in the eastern foothills of
the Rwenzori, continues westward beyond the Katonga Beds until it crosses the Mpanga River where the topographic expression of the Katonga Valley peters out.

![Figure 3-16: Outcrop of the Katonga Beds from the Geological Survey of Uganda 1:250000 map, 1962](image)

**Lithology and facies**

During five days of field work in the area, the main exposure was identified in the railway cutting shown in Figure 3-16. Another exposure was found in the bed of the murrum road near the western limit of the indicated outcrop, and boulders were found at the eastern limit of the indicated outcrop. No evidence of fault control on the extent of the outcrop was observed. Despite a comprehensive search, no other exposures were found, which raises the possibility that the extent of the outcrop shown in the published geological map is based on the interpolation between these three small exposures only.

The Katonga Beds at Bihanga Station and the Muzizi Sandstone on the Kyenjojo to Hoima road are described together on the 1:250,000 geological map (Figure 3-3) as ‘conglomerates, sandstones and siltstones’ of ‘?Miocene’ age (D.G.S.M., 1962).
The main outcrop in the railway cutting 2 km east of Bihanga Station is shown in Figure 3-17. Three main lithologies were identified at this location, including:

- **Rhythmite** – Interlaminated yellowish buff and grey siltstone and mudstone with sandy layers, and rare pebbles (Figure 3-18 showing fractured laminae, Figure 3-21b with soft sediment deformation);

- **Sandstone and siltstone** –
  - Orangey brown to yellowish buff, interbedded and interlaminated, fine to coarse-grained sandstone, with some reaction to 10% HCL (Figure 3-21a showing small scale contemporaneous brittle and ductile deformation with sediment injection);
  - Grey, fine-grained sandstone and siltstone with black carbonaceous plant fragments, and no reaction to 10% HCL; and

- **Diamictite** – Grey, massive, fine to medium-grained sandstone, with dispersed sub-rounded to rounded small pebbles to large cobbles, with some calcareous cement and veins which have moderate to vigorous reaction with 10% HCL (Figure 3-19 and 3-20).

In addition to these lithologies, the exposure in the road at the western end of the outcrop closely resembled the thinly bedded fine-grained sandstone identified at Kabagole in the Katonga Valley and near the Kyenjojo-Hoima road in the Muzizi
Valley. The boulders at the eastern end of the outcrop were comprised of diamictite. Due to the limited exposure and the structural relationships discussed below, no stratigraphic relationship could be established between the different lithologies.

The rhythmites of the Katonga Beds are similar to those described in the Karoo sediments at Entebbe (Schlüter et al., 1993), and the Lower Lukuga Group of the D.R. Congo (Cahen and Lepersonne, 1981). The author has also observed their similarity to Entebbe Karoo-age rhythmite samples in the Department of Geological Survey and Mines museum and is situ rocks exposed on the Lake Victoria foreshore south of the Geological Survey offices in Entebbe.
Figure 3-19: Diamictite from near Bihanga Station a) *in situ* at the railway cutting and b) in hand specimen

Figure 3-20: Thin section of diamictite from near Bihanga Station in a) plane polarised light and b) crossed polars

Figure 3-20 shows a thin section of the diamictite in plane polarised light and under crossed polars. The thin section reveals very poor sorting, with granules (2 to 4 mm) and sand (2 to 0.63 mm) sitting in a silty matrix (0.63 to 0.004 mm). The grains are mainly angular to sub-angular and some larger grains appear sub-rounded. The visible mineralogy is diverse, although the majority of sand grains are monocrystalline quartz and feldspar with some biotite grains. The mineralogy of the light brown fine-grained matrix cannot be resolved, but likely includes micas, clays and iron minerals. Some polycrystalline and lithic clasts are also present, including the granule-sized igneous clast in the bottom left-hand-side of Figure 3-20a and 3-20b, mainly comprised of quartz and muscovite. The large rounded lithic clast on the right-hand-side of Figure 3-
20a and 3-20b is formed from very fine-grained sandstone with angular grains similar to the Kabagole Sandstone shown in Figure 3-14. This indicates that fine-grained sandstones similar to the Kabagole Sandstone predate the diamictite. The rough surface texture suggests it was perhaps relatively weakly cemented at the time of re-working.

The massive character and range of grain size and shapes from silt, to angular to sub-angular sand, and to rounded pebbles and cobbles of the diamictite seen in outcrop and thin section (Figure 3-19 and 3-20) is indicative of a depositional environment where little sorting has occurred. Diamictites may be deposited by moving or stagnant glaciers (lodgement and ablation till) and in debris flows. These types of deposits are not necessarily mutually exclusive and mass flow can occur in saturated glacial deposits. The well-rounded character of the occasional pebbles and cobbles indicates fluvial transport prior to incorporation into the angular sandy matrix. The diamictites of the Katonga Beds are again similar to rocks described in the Karoo sediments at Entebbe (Schlüter et al., 1993), and the Lower Lukuga Group of the D.R. Congo (Cahen and Lepeironne, 1981).

**Syndepositional structures and glaciotectonics**

Most structures within the Katonga Beds show evidence of syndepositional origin. At the scale of metres to 10s metres large blocks have been overturned. For example, the laminae and bedding of the sandstone and siltstones shown in Figure 3.17 are sub-vertical and juxtaposed against the adjacent massive diamictite. Although the exact location of the contact is poorly defined, it is clear that the entire outcrop is in situ. Evidence from the Congo Basin suggests that post-Karoo deformation in equatorial Africa is unlikely to have produced locally overturned strata (Figure 3-11). Also, given that the sediments fill the valley, any regional structures resulting in dipping beds would need to have deformed the Precambrian rocks too. Given the lithofacies characteristics, a more likely explanation is that the sediments have been upturned as a result ice movement at the glacier margin.
Evidence of glaciotectonics in a proglacial setting also exist at the scale of centimetres to 10s centimetres. The structures shown Figure 3-21a include both brittle and ductile failure of the interbedded sandstone and siltstone. A set of imbricate micro thrust faults can be seen in the upper right-hand-side, and below them another micro fault can be seen to have dislocated a fold axis, indicating contemporaneous brittle and ductile deformation. Close examination reveals that grains from adjacent beds have been smeared along planes of movement. In addition, reddish brown silty sand has been injected along the main sub-vertical discontinuity in the centre of Figure 3-21. All these features are consistent with glaciotectonics in saturated sediment before lithification. Glaciotectonic displacement and loading of rhythmites with a high water content provides a likely explanation for the convolute bedding and laminations associated with soft sediment deformation shown in Figure 3-21b.

**Palaeontology**

Following an extensive search, two relatively poor quality plant fossil fragments were found. Unfortunately, neither was found in situ. A small woody stem fossil (Figure 3-22a) was found in a boulder adjacent to the exposure in the railway cutting, and a fossil leaf fragment (Figure 3-22b) was found in a pile of building stone at Bihanga Station derived from near the railway cutting.
Figure 3-22: Fossil plant fragments from near Bihanga Station including a) woody stem, and b) leaf resembling Permo-Carboniferous gymnosperm Noeggerathiopsis

Figure 3-23: Permo-Carboniferous gymnosperms a) Cordaites and b) Noeggerathiopsis (from McLoughlin and Drinnan)

Following communication with a palaeobotanist from Rhodes University in South Africa familiar with Karroo-age plant fossils (Prevec, 2010) it was concluded that the bark-like texture on the woody stem points towards a gymnosperm, rather than sphenopsid (horsetail). The general texture of the leaf fragment is also consistent with
a Karoo-age gymnosperm, although the small fragment and lack of overall leaf morphology prevent a definitive identification. However, the parallel venation is suggestive of the Permo-Carboniferous gymnosperm Noeggerathiopsis (Figure 3-23b), which is also found in the Upper and Lower Lukuga groups of the D.R. Congo (Cahen and Lepersonne, 1981) and the Karoo-age rocks of Uganda (Schlüter et al., 1993). Noeggerathiopsis is a Lower Gondwana gymnosperm, with similar leaf morphology to the Upper Gondwana gymnosperm Cordaites (Figure 3-23a), found in North America (1982, McLoughlin and Drinnan, 1996). Cordaites are seed-bearing gymnosperms with strap-like leaves that were ancestors to the modern conifers. Although Noeggerathiopsis is Cordaites-like, Pant (1982) states that no cordaitean stems or fructifications had been found in association with Noeggerathiopsis leaves at that time.

The presence of plant fossils in association with glaciogenic sediment may be explained by the proglacial sediment characteristics suggesting deposition on the ice margins, together with climate variability during the late Palaeozoic glaciation. Despite the poor quality of the fossils found, their similarity to known Permo-Carboniferous gymnosperms found in association with the Upper Lukuga/Ecca (Permian) and Lower Lukuga/Dwyka (late Carboniferous) appears too fortuitous to be easily dismissed.

### 3.5 Summary

#### 3.5.1 Precambrian

The Katonga Valley lies within the central part of the thermotectonically defined Rwenzori Fold Belt, which appears to form the eastern extension of the Palaeoproterozoic Kibalian Belt of the north east D.R. Congo. The valley is orientated sub-parallel to the WSW-ENE trending structural axis. To the north of the Katonga Valley, the rocks are predominantly cleaved sandstone, slate, phyllite, schist and quartzite of the Palaeoproterozoic Buganda-Toro System. To the south, the rocks are predominantly acid and basic gneisses and foliated granites of the Archaean Gneissic-Granulitic Complex. The late Palaeoproterozoic Mubende Granite is intruded north of the western Katonga Valley. West of the Mubende Granite the structure is complex and recent work has suggested that area is comprised of stacked slices of Proterozoic and Archaean rocks (Link et al., 2010). East of the Mubende Granite the contact between the low to medium and high grade metamorphic rocks is located never more than 10 km south of the Katonga Valley and in some locations along the valley itself.
A line of discontinuous quartzite ridges occurs along the contact. This feature has been named the Katonga Line (King and Swardt, 1967) and it has been speculated that it may represent a major transcurrent fault (Johnson, 1960). It has been suggested that the Katonga Line has controlled the development of the Katonga Valley (King and Swardt, 1967). However, closer examination reveals that rather than forming a more easily eroded lineament, the resistant quartzite ridges along the Katonga Line have created a barrier to southward migration of the river.

The relationship between the valley form and Precambrian geology represented on the geological maps for the main reaches of the Mpanga-Katonga Valley System may be summarised as follows:

- Mpanga Falls – no apparent relationship;
- Mpanga Falls to drainage divide – the Mpanga-Rusangwe-Nyaitanga valley system follows the SW-NE structural trend of interleaved Proterozoic and Archaean rocks;
- Drainage divide to Nkonge – west of the Mubende Granite, the Katonga Valley cuts across the Precambrian contacts, and east of the granite it follows the E-W contact between the low to medium and high grade metamorphic;
- Nkonge to Kisozi – whilst the quartzite ridges appear to have limited southward migration of the river, south of Nkonge it crosses the quartzite therefore suggesting superimposed drainage; and
- Kisozi to Lake Victoria – a NW-SE trending fault appears to have led the valley south of the resistant Singo Series before it turns towards the north onto the low to medium grade metamorphic rocks where it broadens as it approaches Lake Victoria.

### 3.5.2 Karoo

The Karoo Supergroup of Africa ranges in age from Upper Carboniferous to Lower Jurassic. The Karoo-age rocks are comprised of a sequence of terrestrial deposits which record a change from glacial, to humid fluvial and lacustrine, to arid aeolian environments of deposition as global climate changed and the continent of Africa drifted north. They were deposited in a retro-arc basin in South Africa, in the rift basins of Tanzania, Malawi and Kenya to the south east of Uganda, and in the Congo Basin to the west of Uganda. The geology of Uganda is dominated by Precambrian
rocks and the significance of the small Lower Karoo-age (Carboniferous and Permian) outliers has largely been overlooked. Whilst the Palaeozoic age of sedimentary outcrops at Entebbe, Dagusi Island and Bugiri has previously been recognised, the indurated sedimentary strata found filling small reaches of the Nkusi, Muzizi and Katonga valleys were previously thought most likely to be of Miocene age. However, the discovery of fine-grained sandstone at Kabagole in the Katonga Valley led to re-examination of the Katonga Beds at Bihanga Station. This produced strong evidence of their Permo-Carboniferous glaciogenic origin and their similarity to other Lower Karoo strata in Africa, including the Upper and Lower Lukuga groups of the D.R. Congo. This evidence has been categorised into lithofacies, structural and palaeontological as listed below.

- Glacial and proglacial lithofacies, including:
  - diamicite – grey, massive, unsorted, fine to medium-grained sandstone with angular to subangular grains, and dispersed sub-rounded to rounded small pebbles to large cobbles (tillite?); and
  - rhythmite – interlaminated yellowish buff and grey siltstone and mudstone with sandy layers, and rare pebbles (varvite?).

- Glaciotectonic structures, including:
  - overturned rhythmite blocks on a scale of metres to 10s metres, juxtaposed against in situ massive diamicite;
  - contemporaneous brittle and ductile deformation of sandstones and siltstones with grains smeared and injected along discontinuities on a scale of centimetres to 10s centimetres; and,
  - convoluted bedding in rhythmites characteristic of soft sediment deformation

- Plant fossils, including:
  - woody stem fragments with bark-like texture resembling Permo-Carboniferous gymnosperm; and,
  - leaf fragments with parallel venation resembling the Permo-carboniferous gymnosperm Noeggerathiopsis, also found in the Upper and Lower Lukuga groups of the D.R. Congo and the Karoo-age rocks of Uganda
Previous research on the Karoo strata of Uganda has looked towards English speaking East Africa and assumed the Ugandan Karoo deposits occur in small graben similar to the larger graben of Kenya and Tanzania, located over 1000 km from Bihanga Station on the opposite side of the Tanzanian Craton (Schlüter et al., 1993). However, examination of the French language literature from the central Africa (Cahen, 1954) reveals that Lower Karoo strata are found filling the exhumed Palaeozoic landscape of the eastern D.R. Congo. The closest outcrop, in the Ituri Valley, is located only 200 km north west of Bihanga Station. Given the lack of evidence for fault boundaries to the Katonga Beds it is concluded that, similarly to the Lower Karoo deposits of the D.R. Congo, they too are filling an exhumed palaeovalley. Indeed, the straight and wide character of the western Katonga Valley is reminiscent of a glacial valley. The history of burial and exhumation of the Katonga Beds and adjacent rocks is examined further in Chapter 5.0 using apatite fission track analysis.

Whilst it has been confirmed that the near-horizontally bedded, fine-grained arkose sandstone discovered at Kabagole in the Katonga Valley is similar to the western exposure of the Katonga Beds and the sandstones previously identified in the Muzizi and Nkusi valleys, its age and relationship to the glaciogenic rocks is uncertain. All these sandstones show considerably greater induration and generally older appearance and characteristics to the Miocene Kisegi Formation in the Western Rift. Indeed, thin sections of the diamictite from the Katonga Beds appear to include lithic clasts of similar appearance to the fine-grained sandstone suggesting that it pre-dates the glaciogenic rocks. It seems likely that both the glaciogenic rocks and the fine-grained sandstone are Karoo in age, although the exact nature of their relative relationship requires confirmation by further evidence.
4 THE POST-GONDWANAN TECTONICS, CLIMATE AND LANDSCAPE

4.1 Introduction

The term ‘post-Gondwanan’ is adopted here to denote the period of time between the final-break up of Gondwana about 130 Ma during the Cretaceous Period, and the present day. It is the period during which Africa has existed as a separate continent. The aim of this chapter is to review the literature of relevance to the landscape evolution of the Katonga Valley and the deposition of associated alluvial sediment. It begins by examining the tectonic development of the East African Plateau and Rift System, before discussing the tectonic geomorphology of the Western Rift in particular. The climatic history of East Africa is then discussed. The chapter goes on to consider the historical studies of landscape evolution in south west Uganda in the context of the recent tectonic and climatic research. Finally, the available records of the Quaternary sediments in the river valleys of south west Uganda and Lake Victoria are reviewed. The information presented here is used in later chapters to interpret the electrical resistivity tomography and discuss the influence of contemporaneous tectonics and climate variability on the landscape evolution and hydrogeology of the Katonga Valley.

4.2 The Tectonic Context

This section describes the tectonic context of the Katonga Valley on the East African Plateau (EAP) adjacent to the western branch of the East African Rift System (EARS). The geological history, regional lithospheric structure, local fault geometry and kinematics all influence the tectonic geomorphology, including the isostatic and flexural processes operating in the region occupied by the Katonga Valley.

4.2.1 Tectonic geology of the East African Plateau and Rift System

4.2.1.1 Geometry

The East African Rift System (EARS) runs from the Afar triangle in northern Ethiopia to the River Zambezi in Mozambique. Figure 4-1 shows the location of the main faults associated with the EARS, and the two branches which divide around the margins of the East African Plateau. The eastern branch runs for 2200 km from the Afar triangle in the north, to northern Tanzania in the south and is known as the Gregory or Kenya
Rift in Kenya. The western branch, known as the Western Rift, runs 2100 km from northern Uganda to southern Botswana (Chorowicz, 2005). In southern Tanzania, the reactivated Karoo-age Rukwa and Ruaha rifts form a triple junction with the Malawi Rift which forms the southern extension of the EARS (Schlüter, 1997).

Figure 4-1: Structural map of the East African Rift System (after Chorowicz 1989)

The EARS is comprised of a chain of graben and half-graben basins, often about 100 km in length, which are generally bordered by high relief of up to 1000s metres, comprising uplifted margins, adjacent plateaux and volcanic massifs. The Gregory Rift is generally 50 to 80 km wide whilst the Western Rift is about 40 to 50 km wide. The internal drainage of the Kenya Rift currently results in shallow alkaline soda-lakes, whilst the currently deeper lakes of the Western Rift are fresh water and have overflows. Figure 4-2 shows the Katonga Valley located to the east of a structurally
complex region of the Western Rift. The Lake Edward basin to the south is orientated NNE-SSW and ends to the east of the Rwenzori Mountains, whilst the Lake Albert basin to the north is orientated NE-SW and adjoins the similarly oriented Semliki basin to west of the Rwenzori Mountains. The Rwenzori themselves form a horst block with a maximum elevation of 5,111 m asl. The floor of the Semliki Valley to the west of the Rwenzori is about 900 m asl and Lake George to the east is about 920 m asl. The maximum elevation of the rift flanks either side of the Rwenzori in this section is between 1,400 and 1,600 m asl, and the average relative relief adjacent to the George Fault, west of the Katonga Valley, is about 300 to 400 m.

![Structural map of the Western Rift adjacent to the Katonga Valley](after Chorowicz 1989)

4.2.1.2 Chronology and volcanism

Cenozoic volcanism in the EARS is generally more common in the north and on the eastern branch (Figure 4.3) (Chorowicz, 2005). Ebinger (1989) has interpreted the onset of volcanism to be prior to or concurrent with rift faulting, and therefore
radiometric dating of the volcanic rocks has been used to constrain the chronological development of the EARS. This is worth considering further as it also provides a time frame for tectonic geomorphology of the Katonga Valley on the eastern flank of the Western Rift.

Figure 4-3: Structure and volcanism associated with the EARS (after Chorowicz, 2005)
It has been proposed that the earliest trap volcanism began in southern Ethiopia during the Eocene as early as 45 to 37 Ma (Ebinger et al., 1993, Ebinger and Sleep, 1998), although Hofmann et al. (1997) suggest an early Oligocene age (30 Ma) for the Ethiopian flood basalts. Figure 4-4 shows the onset age of volcanism against latitude prepared by George et al. (1998). It suggests a southward progression of the onset age in the eastern branch of the EARS, with volcanism beginning at 35 to 30 Ma in northern Kenya, 15 Ma in central Kenya and 8 Ma in Tanzania (George et al., 1998, Nyblade and Brazier, 2002). Braile et al (1995) suggested there is also a progression from north to south in the Western Rift, with volcanism beginning about 12.6 Ma in the Virunga province, about 10 Ma at Kivu, and 7 Ma at Rungwe in southern Tanzania. George et al. (1998) maintain that the apparent difference in ages of volcanism within the Western Rift are negligible compared to those within the Kenya Rift (Figure 4-4). Biostratigraphic evidence from the Kakara Formation near Lake Albert suggest the onset of the northern Western Rift adjacent to the Katonga region occurred between 9 and 12 Ma (Pickford et al., 1993). The volcanism of the Toro-Ankolean Province in Uganda began only about 2 to 1 Ma.

Nyblade and Brazier (2002) point out the Western Rift appears to have begun to open just as the Eastern Rift reached the edge of the Tanzanian craton. They propose that at this time, some of the east-west extensional stress was transmitted across the cratonic lithosphere to cause rifting in the weaker mobile belt lithosphere on the west side of the craton. It is unclear if uplift of the East African Plateau is contemporaneous with the
southward propagation of the EARS, or if it was at least in part an older topographic feature as suggested by Doucouré and de Wit (2003). The apparent southward propagation of rifting is commonly explained by inferring that the African continent has drifted northward over a mantle plume (e.g., George et al., 1998). However, this model is still uncertain and Burke (1996) has argued that the African plate has remained essentially stationary during the past 30 Ma.

4.2.1.3 Lithospheric structure of the East African Plateau and Kenya Rift

The tectonic geomorphology, including isostatic and flexural processes that have influenced the Katonga Valley and Lake Victoria basin are directly influenced by the deep lithospheric structure of the EAP and Western Rift. The lithospheric structure may also provide clues as to if and how the uplift of the EAP is related to the formation and timing of the EARS. Many gravity surveys have been undertaken in East Africa since Bullard (1936) conducted the first in the 1930s. Ebinger et al. (1989) used gravity data to calculate the lithospheric thickness beneath low-lying stable cratonic areas of between 64 and over 90 km, and beneath uplifted plateaux of 43 to 49 km. They estimated the Kenya, Western and Ethiopian Rifts had lithospheric thicknesses to be between 21 and 36 km. Large scale seismic refraction surveys along the Kenyan Rift between Lake Naivasha and Lake Turkana revealed up to 6 km of low velocity rift infill (2.5 to 5.2 km/s), underlain by the upper Precambrian basement to between 10 and 15 km (6.0 km/s) (KRISP, 1991, Schlüter, 1997). The higher velocity (6.5 to 7.1 km/s) lower crust extends down to about 35 km below Lake Naivasha in the south, but the crust thins to only about 20 km beneath Lake Turkana to the north (KRISP, 1991). Seismic refraction studies have placed the top of the low velocity asthenosphere between 35 and 65 km in the vicinity of the Kenya Rift, compared to the normal lithospheric thickness of 150 to 200 km.
Simiyu and Keller (1997) provide an integrated analysis of lithospheric structures based on both gravity surveys and seismic studies. Figure 4-5 presents a gravity map of East Africa with the Bouguer correction applied to remove the influence of topography. It shows the 1200 km wide negative gravity anomaly (-150 ±20 mGal) over the EAP and Tanzanian Craton. Smaller negative gravity values (about -70 mGal) are associated with the Congo basin to the west and the Lake Turkana region and Mesozoic Anza graben to the north. Positive anomalies are seen along the East African coast and may be associated with thinning of the less dense crust close to the continental margin. The Katonga Region is on the edge of the gravity minima associated with the Western Rift. Simiyu and Keller (1997) showed that as the minimum wavelength filter is increased a broad gravity minima is identified centred on Lake Victoria. They suggest this is due to a deep mantle anomaly, which may be related to the topography of the East African Plateau and the deep structure of the Tanzanian craton.

Focussing mainly on the Tanzanian Craton, south of Lake Victoria, Weeraratne et al. (2003) used Rayleigh wave seismic records to show that shallow shear wave velocities...
in the lithosphere are higher than average until they decrease sharply below about 140 km depth. They suggest this is evidence for an upper mantle plume centred beneath Tanzanian craton. S-wave velocity modelling conducted by Nyblade et al. (2000) showed higher than average velocities to between 200 and 250 km beneath the craton. Relatively shallow low velocity S-waves occur beneath the adjacent rifted mobile belts and extend beneath the high velocity cratonic lithosphere below 200 to over 400 km depth. Phase transitions in the mantle produces discontinuities in the P and S waveforms at about 410 and 660 km depth. Nyblade et al. (2000) showed a broad depression in the 410 km discontinuity occurs beneath the Tanzania craton, but no depression can be detected in the 660 km discontinuity. They interpret this as evidence for a deep mantle plume located towards the eastern side of the EAP in southern Tanzania.

Whilst the presence of the low-velocity, low-density plume can explain Cenozoic uplift and volcanism its relationship to rifting remains uncertain. Wichura et al (2010) demonstrated the existence of significant relief on the East African Plateau prior to rifting by examining the slope angle required to maintain flow of the 13.4 Ma Yatta phonolite lava flow for 300 km in Kenya. They combined an empirical viscosity model with the estimated cooling rate to calculate that the pre-rift slope angle was at least 0.2°. Given the length of the lava flow, they estimate that it must have originated at an elevation of about 1400 m. Therefore, as shown in Figure 4-6 they propose that the high topography of the East African Plateau existed prior to 13.5 Ma and likely formed between around 17 to 16 Ma in response to a mantle plume (Burov and Cloetingh, 2009).
Figure 4-6: Chronology of Cenozoic uplift and vegetation changes (after Wichura et al, 2010)

4.2.1.4 Lithospheric structure of the Western Rift

Figure 4-7a shows a contour map of the free-air gravity across the Western Rift to the north and west of the Katonga Valley. A deep negative anomaly can be seen beneath Lake Albert with a reduced negative anomaly associated with the eastern rift flank and a strong positive anomaly associated with the western rift flank. The difference between the rift flanks is largely due to the higher elevation of the western margin. Watts (2001) used Bullard’s (1936) original data, and applied isostatic corrections to account for both surface topography and crustal roots to show that the positive anomalies from the flanks of the Albertine Rift are replaced by negative values. The Bouguer gravity contours (i.e., topographic correction only) shown in Figures 4.7b indicate that the central area of the Rwenzori Fold Belt around the Katonga Valley appears to be associated with a local gravity minima.
Figure 4-7: a) Free-air and b) Bouguer gravity maps of the Western Rift north west of the Katonga Valley (after Karner et al, 2000, and Upcott et al, 1996)

Upcott et al. (1996) used aeromagnetic and gravity data to interpret the basin depths to the north and south of the Rwenzori Mountains. The north-west side of Lake Albert is bordered by the Bunia fault (Figure 4-2) with high rift flanks up to 1300 m asl whilst the south-east side of the lake is bordered by the lower relief of the Toro-Bunyoro fault scarp. The interpreted depth to the magnetic basement, supported by the large negative Bouguer anomaly, increases westward from less than 0.5 km close to the Toro-Bunyoro fault to greater than 5km in the central area of the Bunia fault. The Bouguer anomaly is less negative in the George basin compared to the Albert basins and Upcott et al. (1996) suggests that the sedimentary strata may only be about 1 km thick adjacent to the George fault west of the Katonga Valley (Figure 4-8). Such a moderate depth of sediment may be expected to produce moderate isostatic loading and hence limited flexural uplift of the adjacent rift flank.
Simiyu and Keller (1997) produced large scale parsimonious interpretive cross sections of the lithosphere and upper mantle structure with the aid of a gravity modelling program. The results of deep seismic refraction surveys, teleseismic studies, lithospheric isostasy calculations, igneous petrology and density measurements were used to constrain the models. Figure 4-9 shows the equator profile no more than 25 km south of the Katonga Valley. The model tries to combine the deep broad feature beneath the plateau with narrow shallower features beneath the rifts interpreted by previous researchers. It shows the EAP to be underlain by low-density asthenosphere below about 120 km depth. The deep low-density feature in the model is limited in extent, but presumably is considered to be attached to even deeper continuous asthenosphere. Two upwelling arms of the low-density (asthenospheric) mantle plume penetrate the lithosphere almost to the base of the crust beneath each branch of the EARS. The deeper and broader negative anomaly beneath the Kenya Rift is inferred to be associated with a larger and shallower plume head. There is modest crustal thinning beneath each branch.
Figure 4-9: Regional earth model and Bouguer gravity profiles along the equator from 25°E to 41°E longitudes. (after Simiyu and Keller, 1997)

Figure 4-9 shows the western Katonga Valley located above the negative gravity minima inferred to be caused by the shallow western arm of the plume, whilst the eastern reaches of the valley are located above a reduced negative anomaly associated with the deeper central plume. This model also assumes a crustal suture between the Archaean Tanzanian craton beneath Lake Victoria and the Proterozoic mobile belts exploited by the Western Rift. The modelled shallow mantle anomaly beneath the Western Rift deepens before it dies out to the north whilst continuing further north beneath the Kenya Rift (Simiyu and Keller, 1997).

There has been a resurgence of research in the Ugandan portion of the Western Rift in the last five to ten years. These contemporary studies are motivated by oil exploration and the desire to understand the relationship between tectonic geomorphology and Neogene climate change. Most of the oil industry investigations are commercial-in-confidence with occasional information released in papers or conference abstracts. The ongoing interdisciplinary RiftLink project, funded by the German Research Foundation is currently investigating the dynamics of the Western Rift including links between uplift and climate change has recently acquired new data and produced new hypotheses with regards to the rift and lithospheric structure adjacent to the Katonga Valley. Link
et al. (2010) have examined and compared the petrology of the ultra-basic Toro-Ankole volcanic rocks of Uganda and the basic to intermediate volcanic rocks of the Virunga province to the south in Rwanda. They conclude that the carbonate rich Toro-Ankole volcanic rocks are characteristic of an oxidized mantle lithospheric source at a depth of 120 to 180 km. However, the Virunga lavas are characteristic of melting in the amphibole stability field at less than 100 km depth. On the basis of the igneous geochemistry they propose that lithosphere-asthenosphere boundary is located at a shallow depth in the Virunga region (≈80 km) and deepens to the north in the Toro-Ankole region (>140 km) adjacent to the Katonga Valley. Link et al. (2010) infer that the Western Rift propagated northward from the region of thinner lithosphere beneath the mobile belt to where it now abuts the thick cratonic lithosphere of the Gneissic-Granulitic Complex in Uganda, interleaved with the Proterozoic Buganda-Toro System. Although the general pattern of increasing thickness of lithosphere beneath the rift towards the north is consistent with the gravity data, it can be seen from Figure 4-9 that Simiyu and Keller (1997) interpreted the depth of the asthenospheric plume head to be about 60 km at the equator in the vicinity of the Toro-Ankole volcanics. The depth suggested by Link et al. (2010) on the basis of geochemical evidence is consistent with the depth of the main plume beneath the EAP rather than the shallow arms beneath the rifts, inferred from gravity data.

Another RiftLink study recently published by Wölbern et al (2010) examined the seismic velocity structure of the crust and upper mantle in the region immediately west and north of the Katonga Valley using a temporary seismic network. As shown in Figure 4-10, the structure along the eastern flank of the Western Rift including the western Katonga Valley is relatively simple with a uniform crustal thickness of about 30 km. The study reveals a thinner crust beneath the Rwenzori Mountains of 20 to 28 km with no apparent crustal root. They also indicate reduced S-wave velocity at about 15 km beneath the Lake Edward and Lake George basins which they suggest may be attributable to partial melt associated with recent volcanic activity.
Wallner and Schmeling (2010) point out that specific setting of the Rwenzori horst mountains between the encircling rift segments of the Semliki Basin to the north west and the George Basin to the south east is likely related to the mechanism that resulted in their high altitude (5,199 m asl). They hypothesise that rising plumes beneath the two encircling rifts facilitates horizontal spreading between the lithospheric mantle and the crust beneath the Rwenzori horst in a process which they term ‘rift induced delamination’. Rapid uplift of the overlying buoyant crust occurs following separation from the dense mantle lithosphere. Wallner and Schmeling (2010) initially assumed that the source of buoyancy was a thick crustal root, but Wölbern et al (2010) showed that the crust is actually thinner beneath the Rwenzori horst and so Wallner and Schmeling (2010) suggest that the source of buoyancy may be a zone of partial melt. The fact that the measured relative crustal thickness beneath the Rwenzori horst and the adjacent rifts is opposite to that predicted by Wallner’s and Schmeling’s (2010) model does call into question its validity. This serves to remind us of the limitations of our current understanding of the connections between the crustal structure, tectonics and

Figure 4-10: Map presenting the crustal thickness in the vicinity of the Rwenzori Mountains and eastern flank of the Western Rift (after Wölbern et al., 2010)
large scale geomorphology in this complex region. Nevertheless, given that surface uplift of the Rwenzori Mountains is thought to have begun during the Neogene after 20 Ma (Bauer et al., 2010) and possibly as recently as 2.3 Ma (Ring, 2008) it is apparent that significant changes in the crustal structure have occurred on the same timescale as the tectonic geomorphic events that have left their mark on the current landscape.

4.2.1.5 Plate kinematics

![Figure 4-11: Relative motions along plate boundaries (after Stamps et al. 2008)](image)

Stamps et al. (2008) use GPS data and earthquake slip vectors to produce the kinematic model of the crustal extension created by the EARS shown in Figure 4-11. The large-scale structure of the Western and Kenya Rifts appear to be controlled by the geometry of the Tanzanian craton and regional extension has produced brittle fracture in the weaker mobile-belts surrounding the craton (Koehn et al., 2008). The overall extension rate across the EARS is estimated to be 3 to 6 mm/year, which if constant for 10 Ma would produce an overall extension of about 30 km. Since this is twice the maximum extension (15 km) estimated from fault geometries (Ebinger, 1989), the extension rate
must have initially been slower. The current extension rates decrease from to north to south along the Kenya Rift (Figure 4-11) and increase from north to south along the Western Rift. The combined extension rates from both rift branches are less variable which is consistent with termination of the Kenya Rift against the Tanzanian craton and stress transmission across the strong cratonic lithosphere to initiate rifting in the weaker mobile-belt lithosphere on the west side of the craton (Nyblade and Brazier, 2002, Stamps et al., 2008). The extension rate in the vicinity of the Western Rift to the west of the Katonga Valley is estimated to be 2.1 mm/year (2.1 km/Ma) towards the WSW. The rate across the Kenya Rift parallel to the direction of extension is between 2.0 and 3.2 mm/year.

4.2.1.6 Western Rift fault structures

The Western Rift contains many deep lakes (Figure 4-1) and seismic reflection studies in Lakes Malawi, Tanganyika, and Kivu have indicated that the narrow rift system is segmented into a series of about 80 to 100 km long basins with half-graben structures showing alternating directions of asymmetry along the rift (Karner et al., 2000). The initial brittle deformation led to growth of major normal border faults with associated antithetic/synthetic faults in the hanging wall. Later fault reactivation appears to be associated with flexural uplift of the rift flanks (Upcott et al., 1996). The Albertine Rift north west of the Katonga Valley (Figure 4-2) is typically asymmetric, with the major bounding Bunia fault system along the western side of the lake. The collapsed hangingwall on the eastern side of the lake is characterized by the large antithetic Toro–Bunyoro fault system, which itself is associated with a topographic escarpment.

A conceptual model of rift border faults (Figure 4-12) has been developed based mainly on work from around Lake Tanganyika (Rosendahl, 1987, Rogers and Rosendahl, 1989). The main border faults form broad crescent shapes which means that the movement changes from dip-slip at the centre to strike slip at the end creating an accommodation zone. Antithetic and synthetic faults occur in the basin parallel to the main border fault. The Seismicity up to 30 km depth suggests that border faults may attain great depths within the crust (Ebinger, 1989), but their curved nature means the dip reduces with depth and they are unlikely to penetrate the mantle (Schlüter, 1997).
Figure 4-12: Conceptual model and terminology of rift border faults (after Rogers and Rosendahl, 1989).

The relationship between the Semliki and South Toro-Bunyoro border faults to the north of the Rwenzori (Figure 4-2), and the Lake Edward and Lake George border faults to the south of the Rwenzori have some similarity to the two half grabens shown as case A and B in Figure 4-12. However, propagation of the Semliki fault from the north east and the Lake George fault from the south has produced what may be viewed as an unusual version of case C in which antithetic faulting has produced the intervening Rwenzori horst. Structural and seismological data indicates that the Rwenzori block is rotating clockwise (Koehn et al., 2010). A 20 km long transsection fault appears to be currently detaching the Rwenzori micro-plate on its north-east margin from the larger Victoria plate (Tanzanian craton). It is already detached in the south. Koehn et al (2008, 2010) have attempted to reproduce these observations by modelling rift propagation around the Rwenzori block. Their numerical model consists of an elasto-plastic sheet representing the upper crust overlying a viscous sheet representing the lower crust. Tension is applied with a resultant constant velocity of 2.1 mm/year which is the extension rate estimated from GPS measurements (Stamps et al., 2008) for 14 Ma assuming the rift faults first developed in the mid Miocene. Faults
develop when the yield strength of the elastic crust is exceeded. However, the model is two dimensional and does not reproduce uplift.

![Diagram of rift evolution stages]

- **Stage I**: Initial faults development at rift intersection.
- **Stage II**: Slow propagation of rift faults and rotation of the basement block.
- **Stage III**: Merging of rift segments and capture of the basement block.

**Figure 4-13: Numerical model of rift evolution and capturing of basement block in 3 stages. (after Koehn et al. 2010)**

During Stage I (Figure 4.14) the rift segments form within 2 Ma and propagate parallel to each other. Once the parallel overlap has reached twice their separation the system enters Stage II and the propagation slows. The rifts continue to open and the captured block begins to rotate because the corners are still connected to the main plates. After another 4 Ma the two rifts connect at one end of the block and the system enters Stage III. During Stage III which lasts 8 Ma in the simulation, the rotation may slow and the block only slowly detaches completely.

The model appears consistent with the seismological and field observations of the structural geology of the region. Although the model cannot simulate uplift, Koehn et al. (2010) suggest that significant uplift and tilt along the western fault of the Rwenzori block can only occur after the rifts are connected. They further suggest that their model is consistent with the formation of Lake Obweruka between the Albertine and Edward rifts proposed by Pickford et al (1993) based on faunal assemblages from about 8 Ma, and lake bifurcation due to uplift of the Rwenzori from about 4 Ma. Sedimentation in the George basin could not begin until after rift formation and the time required for sediment loading to approach the current depth of 1 km would create a delay in the rift.
flank uplift due to flexure with consequent repercussions for reversal of the River Katonga.

4.2.1.7 Formation of East African Rift System

Some early accounts of the EARS incorrectly suggested that it had formed due to compression forces (Wayland, 1921, Holmes, 1965) on the basis of reverse faults which later turned out to be Precambrian in age (Schlüter, 1997). Continental rifts are now known to be product of extensional processes arising from global plate movements (Watts, 2001). It is currently unclear if the process of continental break-up along the EARS will evolve into a mid-ocean ridge system or whether the EARS represents the ‘failed arm’ of the successful Red Sea Rift. Another issue, perhaps of more concern for tectonic geomorphology, is whether the EARS can be explained by active rifting, passive rifting, or a combination of both. Active rifting occurs when the ascent of a relatively large volume of hot, low-density, sub-lithospheric mantle (i.e., a hot spot) is the primary cause of extension (Carter, 2010b). This model accounts for pre-rift uplift and volcanism. Passive rifting occurs when extension is driven by tension associated with remote plate boundaries. A relatively small volume of hot, low-density sub-lithospheric mantle forms as a consequence of extension. This model accounts for the late development of partial melt (magmatic underplating) and volcanism after rifting has commenced.

Macdonald et al. (1994) reviewed the evidence available at that time for and against active and passive mechanisms in the formation of the EARS. Whilst some improvements have been made in our understanding of the internal structure of the lithosphere since their review, it is still apparent that our interpretation of the rift evolution and lithosphere-asthenosphere interaction is poorly constrained. Macdonald et al. (1994) conclude that the case for active rifting is supported by: 1) the long wavelength gravity and topographic anomalies which must be dynamically maintained by asthenosphere convection (Ebinger et al., 1989); 2) the short-wave length geophysical anomalies beneath the rift axes suggests active plume emplacement (Simiyu and Keller, 1997); 3) the volume of igneous and volcanic rock which suggests the upwelling of a large mantle plume; 4) the tendency for volcanism and faulting to become concentrated along the rift axis with time is consistent with a rising plume; 5) the different structural styles and magma compositions along the axis suggests a non-
uniformly rising plume (KRISP, 1991). The case for passive rifting is supported by: 1) continent-scale correlation between Cenozoic magmatism and rifting, with older structural weaknesses which suggests reactivation by external extensional forces rather than chance impingement by mantle plumes (Bailey, 1992); and 2) the boundary fault architecture (Rosendahl, 1987) appears consistent with externally driven tensional forces rather than plume upwelling under the rift axis.

Macdonald et al. (1994) proposed that a combination of active and passive mechanisms have formed the EARS. The African plate is envisaged to extend in response to external tension at the same time as a hot mantle plume rises (KRISP, 1991). Geoffroy (2005) proposes that rift initiation may in general be more accurately represented by a combination of active and passive models. He suggests that the global distribution of volcanic passive margins indicates a relationship between the locations of mantle melting and zones of strain concentration within the lithosphere. The relative components of active and passive mechanisms in the formation of the EARS have implications for the tectonic geomorphology of the region. As shown in Figure 4-11, the extension rate in the vicinity of the Western Rift to the west of the Katonga Valley is estimated to be 2.1 mm/year (2.1 km/Ma) towards the WSW. The rate across the Kenya Rift parallel to the direction of extension is between 2.0 and 3.2 mm/year. This gives a total extension of about 4.7 mm/year. Whilst much of this extension may be accommodated by westward movement of the Nubian Plate and eastward movement of the Somalian Plate, the response of the Victoria Plate will depend on whether the rifting is active or passive. An active component of rifting in the western and eastern branches would produce compressive stress in the Victoria Plate which could be accommodated by flexure of the lithosphere, resulting in uplift and/or downwarping of the Victoria Plate. In general terms, the long duration, large-scale geomorphology of the Katonga Valley may therefore need to consider the tectonics of the wider region and not just the Western Rift flank processes alone.

### 4.2.2 Tectonic geomorphology of the Western Rift

#### 4.2.2.1 Rock uplift, surface uplift and isostasy

Tectonic geomorphologists view changes to the landscape as the redistribution of surface loads. Unloading or loading of the Earth’s surface results in a rise or fall in the
lithosphere. Rock uplift is defined as the displacement of rock relative to the gravity equipotential surface (geoid) that best fits mean sea level. Surface uplift is the displacement of the Earth’s surface relative to sea level (England and Molnar, 1990). Exhumation is the displacement of rock relative to the Earth’s surface. The reduction of the Earth’s surface is known as denudation. It mainly occurs due to erosion but can also be produced by tectonic means, especially where there is extension. Rock uplift can also occur due to tectonics, especially where there is compression.

Isostasy is the height adjustment of the lithosphere in response to loading or unloading in order to maintain equilibrium (Carter, 2010a, Watts, 2001). In simple terms, the low-density rigid lithosphere (crust and upper mantle) is envisaged floating on the high-density ductile asthenosphere. According to Archimedes principle, the weight of the low-density crustal block is equal to the weight of the hypothetical smaller volume of high-density asthenosphere which it displaces, and therefore for a given area:

\[ h \rho_{\text{crust}} = r \rho_{\text{mantle}} \]  
\[ \text{Equation 4-1} \]

or

\[ \frac{r}{h} = \frac{\rho_{\text{crust}}}{\rho_{\text{mantle}}} \]

where,

\[ h \] – crustal height from Earth’s surface to compensation level in the mantle (m)
\[ r \] – crustal root from theoretical mantle free-surface to compensation level (m)
\[ \rho_{\text{crust}} \] – density of the continental crust (kg/m\(^3\))
\[ \rho_{\text{mantle}} \] – density of the mantle (kg/m\(^3\))

The compensation level is the depth in the mantle where the pressure exerted by the mass above under the influence of gravity is equal in all areas. From Equation 4.1 it can be seen that the ratio of the root thickness to the total thickness of the continental crust is determined by the density of the crust and the mantle. For typical crustal and mantle densities of 2,800 kg/m\(^3\) and 3,300 kg/m\(^3\) respectively, this ratio is 0.85. Assuming isostatic equilibrium, this ratio must remain constant and therefore for every 1.00 m initial reduction in the crustal height due to erosion, there must be 0.85 m rise of the crustal root such that the final reduction in elevation at the Earth’s surface is only
0.15 m. The relatively small density contrast between the crust and the asthenosphere means that the rebound after erosion is substantial (5/6ths), and this denudational isostasy is therefore important in maintaining relief (Bishop, 2007). We know that most rebound of the lithosphere beneath the ice sheets of the last Pleistocene glaciation has occurred in less than 10,000 years and therefore the isostatic response is rapid on geological timescales (Carter, 2010a).

The isostatic response is modified by the flexural strength of the elastic lithosphere (Carter, 2010a, Watts, 2001). At low temperatures rock behaves elastically under stress until with increasing stress the elastic limit is exceeded and brittle failure occurs. At high temperature rock becomes ductile and deforms plastically. The strain in response to a given stress therefore changes with depth in the lithosphere. Although the rock properties suggest that under the applicable stress and temperature the continental crust should become ductile at about 15 km or less, the empirical evidence from topography and gravity data suggest that the lithosphere is often elastic up to about 50 km or more. The full range that has been estimated at different locations is between 5 km to 100 km (Carter, 2010a). The amount of isostatic adjustment is inversely proportional to the lithospheric strength. The elastic thickness and flexural rigidity are largely determined by the overall thickness and temperature profile of the lithosphere. Flexural rigidity is generally greater in old cold lithospheric crust such as the Tanzanian craton. It decreases towards rifts where such as we have seen for the EARS, the lithosphere is thinner and hotter (Simiyu and Keller, 1997).

4.2.2.2 Mechanisms of rift flank uplift

Uplifted rift flanks with a wavelength of 50 to 200 km are commonly observed adjacent to continental rifts (Karner et al., 2000). The various mechanisms that have been proposed to explain rift flank uplift may be grouped as follows:

1. surface processes (including loading and unloading due to erosion and deposition);  
2. thermal processes (in the mantle and crust); and  
3. variable strain processes (including elastic, brittle and ductile deformation).

The conceptual models proposed to explain rift flank uplift often combine these mechanisms. Whilst the first mechanism is widely accepted, the relative contribution
of the other mechanisms is less clear and depends on the conditions at particular rift margins. Assessment of the potential role of respective mechanisms is often based on thermomechanical numerical modelling. Given the limited availability, accuracy and resolution of geophysical observations in the deep crust and upper mantle in the vicinity of continental rifts it is often difficult to falsify or corroborate these alternative models.

**Surface processes**

The simplest mechanism of rift flank uplift in which we have most confidence is the flexurally modified isostatic adjustment of the lithosphere in response to unloading and loading due to erosion and deposition in the vicinity of rift margins. The magnitude of lithospheric flexure can be examined by analogy to a semi-infinite beam. The approach is adapted from civil engineering where the flexure of elastic beams has been studied extensively. A summary of the approach is provided here and the details are given in Watts (2001). The amount of isostatic adjustment is inversely proportional to the lithospheric strength. Flexural rigidity is a measure of the force required to bend the rigid lithospheric plate to a unit curvature. It is related to the stress and strain behaviour of the lithosphere as follows:

\[
D = \frac{ET_e^2}{12(1-v^2)}
\]  
Equation 4-2

Where,

- \(D\) - flexural rigidity (Nm)
- \(E\) - Young’s modulus (tensile stress/tensile strain, N/m\(^2\))
- \(T_e\) - elastic thickness (m)
- \(v\) - Poisson’s ratio (transverse strain/axial strain)

Since Young’s modulus and Poisson’s ratio have relatively little spatial variation at the scale of the lithosphere, the flexural rigidity is largely determined by the elastic thickness. This depends on the overall thickness and temperature profile of the lithosphere. The flexural rigidity will determine the wavelength and amplitude of flexure, which is defined by the parameter, lambda (\(\lambda\)), as follows:

\[
\lambda = \left(\frac{(\rho_m-\rho_{inf})(\frac{1}{\theta})}{4D}\right)^{1/4}
\]  
Equation 4-3
\( \lambda \) - wavelength and amplitude parameter (m\(^{-1}\))

\( \rho_m \) - density (kg/m\(^3\)) of inviscid fluid (mantle) below the beam (lithosphere)

\( \rho_{\text{infill}} \) - density (kg/m\(^3\)) of fill (sediment) above the deflected beam (lithosphere)

\( g \) - acceleration due to gravity (m/s\(^2\))

**Figure 4-14: Flexure of a semi-infinite beam by a line load \( P_b \)/unit width (from Watts, 2001)**

The overall effect of end loading a semi-infinite beam is to produce a large downward deflection of the free end and a small upward deflection with wavelength and amplitude defined by lambda at some distance from the free end as shown in Figure 4-14. If the beam represents the lithosphere, then loading is produced by sedimentation in the rift and the upward deflection is created on the rift flank. In general, cooler margins have larger elastic thickness and therefore greater rigidity, which produces longer wavelength but lower amplitude uplift.

The total vertical deflection at a horizontal distance along the beam (i.e., away from the sediment filled rift basin) is given by:

\[
y = \frac{2P_b \lambda}{(\rho_m - \rho_{\text{infill}})g} e^{-\lambda x} \cos \lambda x
\]

Equation 4-4

\( y \) - total vertical deflection (m)

\( P_b \) - load per unit width on the free end of the semi-infinite beam (N)

\( x \) - distance from the free end of the semi-infinite beam (m)
In general, cooler margins have larger elastic thickness and therefore greater rigidity, which produces longer wavelength but lower amplitude uplift. The process of rift flank uplift is self-sustaining and focused erosion drives further isostatic rebound of the rift flanks which maintains the topography as rocks underlying the flanks are exhumed (Carter, 2010b). The lithospheric flexure due to loading of the George and Victoria Basins and its potential influence on the profile of the rift flank and the Mpanga-Katonga Valley System are discussed further in Section 8.5.3.

Burov and Cloetingh (1997) examine the potential influence of surface processes on processes occurring within the deeper crust using a combined thermomechanical model. As depicted in Figure 4-15, increased loads in rift basins lead to inelastic weakening of the lithosphere, whilst erosional unloading of the rift flanks leads to local strengthening and rebound. The induced pressure gradients are sufficient to produce flow towards the rift flanks in the ductile lower crust, promoting basin subsidence and possibly even additional rift flank uplift. If rift flank erosion ceases then the load on the rift margins can produce ductile flow in the lower crust back towards the basin. Whilst it is often assumed that the rate of lithospheric compensation is dependent on the asthenosphere properties Burov’s and Cloetingh’s (1997) modelling showed that it is controlled by the properties of the lithosphere itself.

Figure 4-15 shows that a uniform elastic thickness prior to rifting is initially decreased beneath the loaded basin. In this model, a further zone of decreased elastic thickness is produced beneath the outer slope of the raised rift flank. For young lithosphere the variations in the elastic thickness due to loading can be over 20% (Burov and Cloetingh, 1997). Whilst this variation may be significant across an individual rift it is a relatively small variation compared to those observed between different rift settings.
Figure 4-15: The influence of rift flank unloading and basin loading on elastic thickness of the upper crust and flow within the ductile lower crust (after Burov and Cloetingh, 1997)

It is clear that complex feedback mechanisms exist between surface processes and lithospheric processes at continental rifts such as the EARS. The current profile of the Katonga Valley is not simply the result of a passive response to independent tectonic forces. The surface processes have themselves influenced the tectonics. The flexurally moderated isostatic balance is dependent on the rate of loading and unloading at any given time. The relative rate of unloading of the Katonga flank and loading of the adjacent George Basin are subject to change. These can come about due to climate change, active tectonics or changes in sediment routing. For example, the eroded sediment from the rift flank may be redistributed in a variety of proportions between the George, Edward and Victoria basins. Another possibility is that rapid uplift and erosion of the Rwenzori could result in rapid loading in the George Basin compared to unloading of the Katonga flank. Given the long and complex history of rift flank uplift adjacent to the Western Rift it seems likely that the history of the Katonga Valley, including the flow direction and location of the water divide are also complex.
**Thermal processes**

Rift flank uplift can be influenced by thermal processes operating in the mantle and/or the crust. Section 4.2.1 discussed the evidence for a low density plume beneath the Western Rift (Simiyu and Keller, 1997, Wölbern et al., 2010, Link et al., 2010). Schmeling (2010) recently examined the processes of lithospheric thinning and asthenosphere ascent with decompressional (adiabatic) melting during continental rifting using a two dimensional numerical model. Passive rifting is modelled by imposing constant extension rates between 2.5 and 40 mm/year across half of the rift-flank system. The model solves the equations for the conservation of mass, momentum and energy in the crust and mantle for both solid and liquid-melt phases. It predicts that the deep thermal processes due to passive extension produce pinch and swell topography of 50 to 100 m relief during the first 1 Ma. As necking (thinning and extension) of the lithosphere progresses between 1 Ma and 3 Ma the rift deepens considerably with rift flank uplift reaching a maximum of about 200 to 250 m. As rifting continues the basin deepens to between 3.5 and 4 km depth after 6 Ma. This is still 1.5 to 2 km less than would be expected for the isothermal isostatic topography (Schmeling, 2010). The differences between the modelled topography and purely isostatic topography are due to dynamic forces associated with thermal buoyancy and ductile flow. The modelling conducted by Schmeling (2010) shows that the thermal effects are more significant in maintaining the elevation at the base of the rift than in modifying the rift flank topography. The topography of the rift flanks with melting is about 100 to 125 m higher than without. After extension stops, melting and convection remains active for 50 to 100 Ma.

Thermal effects can also have significant effects in the upper crust by influencing the elastic thickness. The filling of depositional basins leads to thermal insulation and temperature increases of 50 to 100 °C at about 15 km depth (Carter, 2010b). Stretching and thinning due to extensional forces results in increased geothermal gradients which decrease the elastic thickness and flexural rigidity. If the depth of necking is greater than the isostatically compensated depth of the sedimentary basin, the hot thin lithosphere can expand to produce a domal structure (Kooi, 1991, in Carter, 2010b). The asthenosphere also rises below the thinned crust, adding to the heat flow. This process can lead to uplift of 100s of metres (Carter, 2010b). This process of thermal...
uplift is applicable to young rifts like the EARS. It is a transient process, and cooling beneath older rifts would be expected to result in thermal subsidence. It is important to note that comparison of thermal uplift and mechanical uplift using thermomechanical models shows that flexural uplift of rift flanks is much larger than thermal uplift.

**Variable strain**

A common class of variable strain models produce rift flank uplift by redistributing the vertical proportions of the lithosphere and asthenosphere beneath the rifts under the action of necking. Initial models of extension during rifting assumed that stretching occurred at the same rate in the crust and mantle (McKenzie, 1978). Later models proposed non-uniform, depth dependant, discontinuous stretching. The discontinuous models required a marked difference in strain behaviour at the crust-mantle boundary and a mechanism to detach the crust from the mantle which resulted in space problems. Rowley and Sahagian (1986) presented a continuous depth dependant extension model to get over the space problems and strain discontinuities. Uplift occurs in these models because stretching is greater in the lithospheric mantle than in the crust. As the asthenosphere rises to replace the lithospheric mantle it undergoes adiabatic heating which causes its density to decrease and the rift flanks, outside of the region of crustal extension, to rise.

![Figure 4-16: Rift flank uplift due to continuous non-uniform stretching of the crust and mantle lithosphere (after Rowley and Sahagian, 1986)](image)

Most conventional mechanical models assume rift flank uplift occurs due to flexure caused by external loading. Sachau and Koehn (2010) present an alternative variable strain model in which rift flank uplift occurs due to reduced elastic strain caused by near surface brittle fracture. As shown in Figure 4-15, the extensional stress regime results in stretching (elastic strain) of the lithosphere, but fault formation (brittle
fractures) reduces the stress on the near-surface rocks resulting in elastic recovery and subsequent thickening. The material at the base of the faults remains stretched and thinned. It is the vertical gradient in the elastic strain that results in local uplift on the rift-flanks adjacent to the boundary faults. The standard flexural model results in larger rift flank uplift when the elastic thickness of the crust is thinner due to its lower rigidity compared to thick crust. However, fault depth will be greater when brittle failure occurs in thick elastic crust and so Sachau’s and Koehn’s (2010) model results in similar rift flank uplift for thin and thick crust.

Figure 4-17: Rift flank uplift due to reduced elastic strain caused by near surface brittle fracture (after Sachau and Koehn, 2010)

4.2.2.3 Coupled models of the Western Rift

Ebinger et al. (1991), Upcott et al. (1996), and Karner et al. (2000) all adopted a similar approach to kinematic and flexural modelling of the Western Rift first developed by Weissel and Karner (1989). The model geometry is based on the border fault geometry of the Western Rift described in Section 4.2.1. As shown in Figure 4-18, extension brings about brittle deformation in the upper crust and plastic deformation in the lower crust. Faulting reduces the thickness of the crust which produces isostatic rebound and flexural uplift of the foot wall rift flank. The passive rise of the asthenosphere into the thinned lower crust produces thermal buoyancy and dynamic uplift of the hanging wall rift flank. Subsequent erosion of the rift flanks and deposition in the rift basin produces the flexural response described in the context of surface processes.
Figure 4-18: Three stage model of crustal extension and rift flank uplift for the Albertine rift system (after Karner et al, 2000)

Using a numerical implementation of this conceptual model Ebinger et al. (1991) were able to simultaneously match the topography and free-air gravity across six different basins of the Western Rift. They showed that basin depths greater than 8,000 m and uplift greater than 1,000 m can be achieved with extensions of at least 10 km. They suggest that flexural rebound accounts for most of the rift flank topography of the Western Rift. Discrepancies occur between the model results and observed topography near some volcanic provinces, possibly due to unmodelled processes such as small scale convection, rheological variations in the crust, and local asthenospheric plumes. They estimate elastic thicknesses between 17 and 38 km which is lower than beneath unfaulted regions of the EAP where it has been estimated to be greater than 66 km (Ebinger et al., 1989). Nevertheless, the crust across the Western Rift appears to retain significant rigidity.
As gravity and bathymetry data became available during the 1990s, Upcott et al. (1996) were able to produce new model calibrations for the northern Western Rift, including the Albert and Edward basins. They showed that the basin and flank morphology could be largely explained by 6 to 9 km of extension with an elastic thickness of 25 km. Although flexural rebound could explain about 90% of the rift flank topography it was unable to match the higher topography on the western margin of Lake Edward and northern Lake Albert. Upcott et al. (1996) suggest the occurrence of unmodelled surface and thermal processes including local melt generation to explain these misfits.

Karner et al (2000) took advantage of improved calibration datasets and model capabilities to once again examine the tectonic evolution of the Albert Rift. They produced good matches between the modelled and observed topography and gravity data by simulating main boundary fault displacement together with synthetic and antithetic faulting in the hanging wall (Figure 4-19a). The total extension predicted by their models varied between 6 and 16 km and is not limited to the width of Lake Albert. Their models suggest that the elastic thickness of the crust varies between 24 and 30 km (Figure 4-19b), once again indicating significant flexural strength across the Albert Rift.

Figure 4-19: a) Comparison of model and observed topography and b) variation of the basin geometry, effective elastic thickness, and the total extension for profiles of the Albert Rift (after Karner et al, 2000)

4.2.2.4 Tectonic geomorphology of the Rwenzori Mountains

The Katonga Valley occupies a unique position on the eastern flank of the small George Basin, east of the highest peaks of the Rwenzori Mountains. This horst block
occupies a location at the junction of the NE-SW trending Albert Basin and the NNE-SSW trending Edward Basin. It is interesting to speculate that an active component of rifting would tend to produce compression in the vicinity of the obtuse angle between these two basins at the location of the Rwenzori Mountains and the Katonga flank. North of the George Fault where the Precambrian rocks outcrop between the George and Albert Basins, the Rwenzori horst is still joined to the Victoria Plate. It seems possible that the unique three dimensional tectonic geomorphology of this region may have influenced the topographic development of the Katonga rift flank.

Ring (2008) proposes the three phase conceptual model for development of the Rwenzori horst presented in Figure 4-18. His structural analysis of the Rwenzori fault and fault slip orientation concluded that the main Rwenzori border faults suggest early SE/SSE directed extension, whilst the faults adjacent to the high topography of the central Rwenzori suggest later E/ENE extension. Biostratigraphic evidence indicates surface uplift commenced about 2.3 Ma (Pickford et al., 1993). The central region of the Rwenzori Mountains is up to about 5 km asl and thermochronological evidence indicates that total erosion since the start of uplift is likely less than 2 km (MacPhee, 2006). Ring (2008) suggests that the border fault gouge thickness may indicate maximum displacements in excess 6 km. Therefore, he estimates that the total rock uplift between 2.3 Ma and about 300 ka was between 3 and 4 km along the border faults (Figure 4-20a). At about 300 ka the surface uplift was adequate to facilitate glaciation. The development of secondary faulting around the central Rwenzori area was accompanied by a kinematic change to SE directed extension. Ring (2008) speculates that the accelerated erosion rates promoted by glaciation of this central fault block may have produced the isostatic rebound which lead to its extreme elevation (Figure 4-20b). Volcanism began east of the Rwenzori at about 50 ka and appears related to rock uplift, although it is unclear if it is associated with the cause or effect (Figure 4-20c).
Figure 4-20: Simplified conceptual model of the uplift history of Rwenzori Mountains (after Ring, 2008)
4.3 The Climatic Context

Climate parameters influence the character and rate of weathering and erosion, with consequent effects on denudation and landscape evolution. Whilst increased temperature enhances chemical reaction rates, water is required to facilitate weathering and transport both disintegrated and decomposed material. Climate may be considered an allogenic process acting on the landscape but it is itself subject to forcing mechanisms which act at various scales (Maslin and Christensen, 2007). In the context of landscape evolution, we are primarily interested in the thousands, through millions to hundreds of millions of year temporal scales (10^3 to 10^8 year) and regional to global spatial scale (10^3 to 10^7 km). At these scales it is external forcing by tectonics and Milankovitch cycles that dominate the climate variability. Over hundreds of millions of years, plate tectonics can alter the position of the continents which changes the global albedo and ocean circulation resulting in globally warm or cool climates.

Coupled models (Sepulchre et al., 2006) have shown that on a time scale of tens of millions to millions of years uplift of the East African Plateau resulted in a rain shadow effect and reduced precipitation on the plateau and to the east. Modest climatic influences on the landscape at the century scale may be influenced by anthropogenic and solar forcing. Internal mechanisms of the ocean-atmosphere system such as the El Nino Southern Oscillation can influence short-term climate on a decadal to annual scale, and below this scale the processes become chaotic and are defined as weather.

The evidence of historical climate conditions is considered here at the temporal scales of the Cenozoic, Neogene, and Quaternary. To assist comparison of climate at these different scales, the global climate records compiled by Rohde (2009) are presented together in Figure 4-21 as follows: a) Cenozoic - 65 Myr (Zachos et al, 2001), b) Pliocene and Pleistocene - 5.5 Myr (Lisiecki and Raymo, 2005), and c) late Pleistocene and Holocene (450 Ka) (Petite et al, 1999, and Lisiecki and Raymo, 2005). These are referred to in the following sections.
Figure 4-21: Global climate records compiled by Rohde (2009) for the a) Cenozoic - 65 Myr (Zachos et al, 2001), b) Pliocene and Pleistocene - 5.5 Myr (Lisiecki and Raymo, 2005), and c) late Pleistocene and Holocene (450 Ka) (Petite et al, 1999, and Lisiecki and Raymo, 2005)
4.3.1 Cenozoic global climate history

4.3.1.1 Mechanisms of Cenozoic global climate change

The oscillations between glacial and interglacial climates of the Pleistocene and Holocene are thought to be primarily due to the Earth’s orbital parameters (Hayes et al., 1976) although there is rarely a direct correlation due to internal feedback mechanisms. The three main orbital parameters (Maslin and Christensen, 2007) are described below.

1. **Eccentricity** describes the change in the Earth’s orbit around the Sun from nearly circular to elliptical. Each cycle is approximately 100,000 years, with the maximum eccentricity varying over four cycles, creating a longer 400,000 years cycle (Zachos et al., 2001). However, the maximum variation in total annual insolation is only 0.03% and therefore the major effect is to work in conjunction with the precessional effects to influence the severity of the seasons.

2. **Obliquity** describes the tilt of the Earth’s axis relative to the plane of the elliptic. The tilt varies between 22.1° and 24.5° over a 41ka cycle. The main effect is to alter the seasonal contrast (Zachos et al., 2001).

3. **Precession** describes two different orbital parameters (Maslin and Christensen, 2007):

   a. **Axial** precession, which describes the wobble of the Earth’s rotation axis around a fixed axis (as can often be observed for a spinning top) every 27,000 years; and

   b. **Orbital** precession, which describes the precession of the Earth’s orbit, as the entire orbital path changes position in space with a cycle of 105,000 yrs.

Combining the axial and orbital precessions produces a period of 23,000 years. Combining the eccentricity and axial precession produces a period of 19,000 years. These two periodicities combine to produce the precession of the equinoxes which results in the summer solstice in each hemisphere coinciding with the perihelion (smallest distance from the Sun) about every 27,700 years. Combining eccentricity, obliquity and precession produces the complex pattern of solar insolation. The calculated minimum and maximum northern hemisphere summer solar insolation in the
last 600,000 years is approximately 450 Wm\(^{-2}\) and 550 Wm\(^{-2}\) respectively. To provide an oversimplified illustration, this is equivalent to a change in latitude from 65°N to 77°N, which would bring the current glacial limit from mid-Norway to Scotland (Maslin and Christensen, 2007).

It is generally thought that low summer insolation allows the build up of ice which creates a positive feedback due to increased albedo and suppression of local temperatures. An initial increase in northern hemisphere ice also disrupts the coupled ocean-atmospheric circulation system which ultimately results in a reduction in the amount of warm water moving northwards in the North Atlantic Drift and further ice sheet expansion (Maslin and Christensen, 2007). There is also growing evidence that atmospheric chemistry, including the abundance of ‘greenhouse’ gasses (from natural and anthropogenic sources) play a significant role in climate change. Following an initial increase in solar insolation several of the feedback mechanisms appear to operate quicker in reverse during deglaciation.

During the 1920s and 1930s, Wayland recognised elevated lake deposits and river terraces in Uganda which he associated with periods of increased rainfall that he called ‘pluvials’ (Wayland, 1930, Wayland, 1934b). He suggested, albeit with some reservations, that these may be correlated with the four northern hemispheres glacial periods then recognised. Today we recognise that the links between global or high latitude climate and the climate of East Africa are much more complex than Wayland envisaged. The global atmospheric carbon dioxide levels can influence vegetation in the tropics and the glacial cycles of slow cooling and rapid warming may also have an influence on the African climate (Maslin and Christensen, 2007). The available body of evidence is growing that precessional forcing directly influences precipitation in the East Africa (Trauth et al., 2005). Modelling by Clement et al. (2004) suggests that precession can cause significant shifts in tropical seasonality and variation of precipitation up to 180 mm/year, which is of a similar magnitude to variations due to the global glacial-interglacial cycle. The interaction between the direct influence of orbital forcing and the global climatic context on the regional climate of East Africa is only just beginning to be understood (Maslin and Christensen, 2007).
4.3.1.2 History of Cenozoic global climate change

Although the Cenozoic climate history has been interpreted using a wide variety of sources of information, the main evidence comes from high-resolution deep sea oxygen and carbon isotope records. Zachos et al. (2001) collated oxygen and carbon isotope data from the Deep Sea and Ocean Drilling Programs and presented them relative to the standard geomagnetic polarity time scale for the Cenozoic. The isotope ratios are expressed using the standard delta notation:

\[
\delta^{18}O \left( ^{\circ}/_{1000} \right) = \left( \frac{^{18}O/^{16}O_{\text{sample}}}{^{18}O/^{16}O_{\text{standard}}} - 1 \right) \times 1000
\]

\[
\delta^{13}C \left( ^{\circ}/_{1000} \right) = \left( \frac{^{13}C/^{12}C_{\text{sample}}}{^{13}C/^{12}C_{\text{standard}}} - 1 \right) \times 1000
\]

The values are reported relative to the Vienna Pee Dee Belemnite (VPDB) standard. The oxygen and carbon isotope ratios in calcite are derived from benthic foraminifera. In general, \( \delta^{18}O \) increases as temperature decreases (0.25\(^{\circ} /_{1000}/^{\circ}C \)) or the mass of continental ice increases (0.1\(^{\circ} /_{1000}/10 \text{ m sea level change} \)). The \( \delta^{13}C \) reflects changes in the dissolved inorganic carbon (DIC) ratio of sea water, which in turn reflects the changes in carbon supply and removal from the global organic and inorganic carbon reservoirs (Zachos et al., 2001).

It can be seen from Figure 4-21a that there was a warming trend in the early Cenozoic (65 Ma to 50 Ma) which culminated in a thermal optimum in the late Palaeocene and early Eocene. The increasing \( \delta^{18}O \) indicates a cooling trend in the middle to late Eocene (50 Ma to 34 Ma). The rise and fall of \( \delta^{18}O \) at the start and end of Oligocene (34 Ma to 24 Ma) is associated with expansion and contraction of the Antarctic ice sheet. A warm phase occurred during the early Miocene (24 Ma to 14 Ma), and was followed by gradual cooling and re-establishment of the major Antarctic ice sheet by 10 Ma. Cooling continued until about 6 Ma, and was followed by an approximately stationary average trend until about 3.2 Ma. This is followed by a further cooling trend and the onset of northern hemisphere glaciation.

As well as long term trends, the isotope records also indicate the occurrence of brief climate anomalies that stand out from the background. The three largest occurred at about 55, 34 and 23 Ma and are all close to the Epoch boundaries associated with
changes in the biosphere. It is less clear if these climate aberrations have had lasting effects on the landscape.

### 4.3.2 Neogene regional climate history

The primary sources of evidence for changes in temperature and precipitation during the Neogene are the palaeobotanical, geochemical and sedimentological records. Specific data sets include the isotope chemistry and organic remains obtained from marine and lake sediment cores.

#### 4.3.2.1 Evidence from the palaeobotanical and carbon isotope record

Today, the natural vegetation type for the majority of the western and central Katonga Valley is characterised by mixed woodland and savannah, with a more arid region around Kisozi, and mixed forest and savannah close to Lake Victoria (Bonnefille, 2010). However, much of the region is actually used for grazing cattle with arable agriculture around the settlements and close to Lake Victoria (Figure 2-5). The historic vegetation cover not only provides clues as to the climate at the time, but may also have directly influenced geomorphological processes.

Early Miocene (17.8 Ma) fruit and seed flora from Rusinga Island indicate lowland wet forest in the area now occupied by Lake Victoria. However, Pickford (1992) has suggested arid conditions prevailed to the west in the area now occupied by the D.R. Congo, citing reports of dune desert sand up 150 m thick. This seems to contradict the interpretation of the stratified fine sands of the Neogene Sables Ocres Series as having a fluvial origin (Giresse, 2005). Pickford (1992) also points out gypsum in mid Miocene Kisegi Beds found on the uplifted northern toe of the Rwenzori horst. These gypsum horizons, observed by the author, are syndepositional and represent evaporite accumulations under arid conditions. The current evidence, some of which appears contradictory, therefore points towards spatial and temporal variations in the region surrounding the Katonga Valley and by the late Miocene there appears to have been an apparent diversity of ecosystems.

The method of carbon fixation adopted by different plant families result in characteristic carbon isotope composition in soil carbonates. The palaeobotanical evidence indicates the occurrence of wide-spread forests and woodland characterised
by C₃ carbon fixation in the early Cenozoic (Bobe, 2006). C₄ carbon fixation evolved during the Oligocene (32 to 25 Ma) but did not become ecologically significant until the late Miocene (7 to 6 Ma) (Osborne and Beerling, 2006). C₄ plants have a competitive advantage under dryer conditions. Today they represent about 1% of species and 5% of biomass, but they include almost 50% of grass species. The relative abundance of C₃ to C₄ in soil and sediment from the late Miocene onwards provides a broad indicator of the change from wet to dry conditions. The marine record suggests a general early expansion of C₄ grassland in East Africa in the late Miocene between 10 and 6 Ma (Bonnefille, 2010, Sepulchre et al., 2006).

A total of 23 taxa of fossil fruits have been identified in the late Miocene to early Pliocene (6 to 4.5 Ma) Nkondo Formation of the Kaiso area on the eastern shore of Lake Albert in the Western Rift (Pickford et al., 1993, Deschamps et al., 1992). The plant associations change from drier Acacia woodland at about 6 Ma, to wet forest and woodland from 5 to 4 Ma, to the wettest forest at about 4.2 to 4.0 Ma. In general, the Nkondo Formation plant fossils in the Western Rift suggest the presence of both a swamp forest and a dense shady wet forest. They include mixed deciduous and evergreen plants such as those found in the Cameroon forests today where seasonal precipitation is 1000 to 1200 mm/yr. The mid Pliocene fossil assemblages from Hadar in northern Ethiopia have been used to estimate annual precipitation between 750 and 1200 mm/yr which is two to three times greater than today. Although global average temperature were higher in the early Pliocene, the local mean annual temperature at Hadar was estimated to be 21 to 22°C, which is slightly cooler than today (Bonnefille, 2010).

Carbon isotope records and pollen assemblages contained in marine core from the Deep Sea Drilling Project (DSDP) site 231 located downwind of East Africa in the Gulf of Aden has been examined (Bonnefille, 2010) for evidence of when extensive grasslands were established in East Africa. Absolute ages were provided by correlation of tephra layers of known age in the Turkana Basin using geochemical analysis. Samples older than 3.4 Ma indicate woodland C₃ vegetation with some fluctuation. From 3.4 to 2.7 Ma the isotopic values remain stable and again indicate densely wooded vegetation. Between 2.7 and 2.5 Ma there is a change towards greater aridity and the expansion of C₄ grassland, coincident with the Northern Hemisphere Glaciation (NHG). Although
C4 grassland expansion occurred in the late Pliocene, Cerling (1992) suggests that the development of the purely C4 Serengeti-type grassland may not have developed until the mid Pleistocene (1 Ma).

Sepulchre et al (2006) used numerical modelling to simulate the influence of regional uplift on the climate and vegetation of East Africa. Prior to 8 Ma the reduced topographic barrier permitted zonal circulation associated with strong precipitation. Tectonic uplift associated with the EARS led to reorganisation of atmospheric circulation resulting in regional aridification and associated vegetation changes. As shown in Figure 4-22 the soil carbonate isotope record indicate a shifts from C3 (generally trees and shrubs) to C4 (generally grasses) vegetation during the Pliocene and Pleistocene (Trauth et al., 2007).

4.3.2.2 Evidence from lacustrine sediment and diatomite

The global Pliocene and Pleistocene temperature history based on oxygen isotope ratios is shown in Figure 4-21. The regional climate history of East Africa over the equivalent time period has been investigated using the sedimentary record of a series of currently small alkaline lakes which first developed during the Pliocene in the eastern branch of the EARS (Trauth et al., 2007, Trauth et al., 2005). The interpretation of prominent lake periods and soil carbonate isotope records for East Africa is shown in Figure 4-22. The lakes are smaller than in those the western branch, but contain an often long and subaerially exposed rich sedimentary record intercalated with volcaniclastic deposits which facilitate $^{40}\text{Ar}/^{39}\text{Ar}$ age calibration of lake-level high stands. For lakes to form there must be a suitable basin present and a sustained positive precipitation-evaporation balance. As previously discussed, rifting which facilitated lake basin formation began in Ethiopia between 20 and 14 Ma, propagated through Kenya between 12 and 2.6 Ma, and occurred in Tanzania between 5 and 1.2 Ma (Trauth et al., 2005). Synchronous changes in the size and depth of several lakes inferred from sediment characteristics and diatom assemblages are attributable to climate change. The indicators of large, deep freshwater palaeolakes include the presence of pure diatomite with freshwater, planktonic diatom assemblages. The indicators of shallow alkaline lakes included mixed clastic/diatom strata with alkaline tolerant benthic-epiphytic diatom assemblages with phytolith and sponge spicules.
The lake records generally suggest periods of increased precipitation superimposed on the overall Pliocene-Pleistocene drying trend (Figure 4-22). Prior to 2.7 Ma the preservation of lake sediments is patchy, but there is limited evidence to suggest lake phases with approximately 400 ka intervals at 4.7-4.3 Ma, 4.0-3.9 Ma, ~3.4-3.3 Ma and ~3.20-2.95 Ma (Figure 4-20). Younger lake phases have been more clearly identified with approximately 800 ka intervals at 2.7-2.5 Ma, 1.9-1.7 Ma and 1.1-0.9 Ma. Trauth et al (2007) suggest that the early lake phases correlate with maxima in the 400 ka component of the Earth’s eccentricity cycle, whilst the three later lake phases correlate
with global climatic transitions. These include the onset of Northern Hemisphere Glaciation (NHG, 2.7-2.5 Ma), the development of Walker Circulation associated with the southern hemisphere trade winds (1.9-1.7 Ma) and the mid Pleistocene Revolution (i.e. the transition from~40ka to~100 ka glacial-interglacial cycles, at 1.0 to 0.7 Ma). After the onset of NHG, Trauth et al. (2007) suggest increased pole to equator gradients and compression of the Intertropical Convergence Zone (ITCZ) were required to make East Africa moisture availability sensitive to maxima in eccentricity and thus changes in precession. Rapid stratigraphic transitions in the lake sediment record suggest precessional forcing caused lakes to appear rapidly, remain stable for thousands of years, and then disappear rapidly.

Although coupled models (Sepulchre et al., 2006, Kaspar et al., 2010) have shown that on a time scale of tens of millions to millions of years, uplift of the East African Plateau resulted in a rain shadow effect and reduced precipitation, Maslin and Christensen (2007) point out that the exact timing, location and altitude of surface uplift in East Africa is unclear. The influence of sub-regional scale topography on the local climate within East Africa has not yet been examined. However, the RiftLink research group is currently examining the history of surface uplift of the Rwenzori Mountains to the west of the Katonga Valley and Kaspar et al (2010) have indicated that they intend to produce nested regional climate models to examine the effects or orographic forcing on the climate of western Uganda. The overall pattern of regional climate change during the Pliocene and early Pleistocene which is characterized by an overall drying trend with periods of increased precipitation may therefore suggest that discharge via the westward flowing River Katonga system after the development of the Western Rift but before reversal was also likely to follow a generally reducing trend but with with periods of increased discharge.

4.3.3 Quaternary regional climate history

The limitation of available dating techniques places restrictions on our current knowledge of the Pleistocene climate of East Africa. For example, the upper age limit of radiocarbon dating is about 45,000 years, and the application of luminescence and cosmogenic dating techniques is restricted to specific environments (Walker, 2006). This section focuses on evidence of climate change in the lake records. It may be compared against the northern hemisphere Quaternary temperature and ice volume
reconstructions shown in Figure 4-21c. Whilst uplift of the East African Plateau has been shown to produce an overall regional Neogene drying trend, the tectonically initiated Pleistocene river reversal in south west Uganda led to the formation of the Lake Victoria. Once formed, the lake developed a self-sustaining local climate where 90% of the water output is via evaporation and 80% of the water input is via direct precipitation (Talbot and Lærdal, 2000).

4.3.3.1 Evidence from Lake Victoria

The current mean depth of Lake Victoria is relatively shallow at about 40 m, with a maximum depth of 68 m (Johnson et al., 2000). Seismic reflection surveys of Lake Victoria indicate a total maximum thickness of lake sediments of about 60 m. According to Johnson et al (2000) a crude calculation based on an estimated average sedimentation rate of 150mm/1000 years indicates the lake origin at about 400 ka. Three erosion surfaces can be recognised separating seismic sequences which indicate low stands which Johnson et al (2000) suggest may be related to the post Pleistocene Revolution 100 ka glacial-interglacial cycles (Figure 4-21b).

The base of the uppermost seismic sequence varies between 0 and 9 m below the lake bed and is underlain by a widespread palaeosol with vertical cracks and preserved rootlets traces (Talbot and Lær dal, 2000) indicating that Lake Victoria completely dried out in the late Pleistocene. A review of the geophysical evidence (Stager et al., 2002) shows that the lake shrank, perhaps drying out, between 18 and 17 ka and complete desiccation occurred between about 15.9 and 14.2 ka (Talbot and Lær dal, 2000, Johnson et al., 2000, Stager et al., 2002, Stager and Johnson, 2008). Whilst this period of desiccation has been disputed by some biologists on the basis of endemic fish diversity, Stager and Johnson (2008) argue that the biological evidence must be reconciled with the incontrovertible geological evidence and not the other way around. They point out that increased aridity in the late Pleistocene is well supported by other research based on pollen records in the Rwanda/Burundi highlands and diatom records from lakes Tanganyika and Albert (Beuning et al., 1997, Gasse, 2000, Stager and Johnson, 2008).
Stager et al (2002) identified a correlation between the composite components of variation in chemical indicators from a Greenland ice core and the record of Lake Victoria water levels derived from abundance of shallow water diatom taxa (Figure 4-23b). They speculate about the apparent link between low boreal radiation intensity and Lake Victoria low stands whilst pointing out uncertainty in the internal and external climate forcing mechanism and direction of causation. The late Pleistocene Lake Victoria low-stand appears to coincide with the Heinrich-1 ice rafting event (H1 in Figure 4-23a), although it is unclear if this event was caused by external solar forcing or by an internally forced binge-purge mechanism (Stager et al., 2002). It is interesting to note that the data appears to indicate the opposite trend to Wayland’s simplistic link between glacial high latitudes and pluvial low latitudes (Wayland, 1930) and the reverse model of ‘cool north-dry tropics’ may be better supported (Stager et al., 2002).

De Gasse (2000) has pointed out that most African palaeohydrological records suggest dry conditions during the Last Glacial Maximum (23-18 ka). This observation can be
reproduced by coupled ocean-atmosphere models which predict reduced precipitation in tropical Africa associated with a general weakening of the hydrological cycle and the Afro-Asian monsoon in particular due to global cooling. Lake Victoria’s recovery is synchronous with the onset of the northern European Bølling-Allerød warm interstadial between the cold Oldest and Younger Dryas stadials (Johnson et al., 2000, Stager et al., 2002).

Multiple core samples reveal that biogenic silica concentration indicative of diatom abundance increases rapidly above the palaeosol and reaches a maximum concentration between 11.9 and 11.2 ka. The overlying biogenic silica concentration then drops and reaches a minimum between 9.8 and 7.5 ka. Johnson et al (2000) suggest that the reduction in silica corresponds to lake overflow which allowed dissolved silica to be removed by the river as well as diatoms.

Stager et al. (2003) examined a high-resolution diatom record from Pilkington Bay near Jinja. They used correspondence analysis to identify a principle component that mainly reflects relative abundance of shallow and deep water taxa and thus approximates to lake depth and precipitation/evaporation ratio. Since evaporation is less variable than precipitation they concluded that a rainfall maximum occurred between 8.8 and 8.3 ka, following which precipitation became more seasonal and decreased between 8.2 and 5.7 ka. Four century-scale increases were identified at about 8.5, 7.0, 5.8 and 4.0 ka. This final increase correlates with the age of the lowest strandline at 3m above the current level of Lake Victoria (Temple and Doornkamp, 1970). Conditions after 2.7ka remained similar to today, with the exception of periods of significant drought between 1200 and 600 yr BP during Europe’s medieval warm period.

4.3.3.2 Evidence from Lake Albert and Lake Edward

Beuning et al. (1997) examined the sedimentological, geochemical and palynological characteristics of two cores, 9.2 m and 10.6 m long, taken from 46 m below the water level of northern Lake Albert. Radiocarbon dates were obtained from the pollen-lignin-charcoal fraction of each core. The evidence suggests low lake levels between about 35 ka and 21 ka with nearby open wooded grassland and abundant C_4 vegetation. Between 21 and 15 ka the lake levels declined and two soil horizons can be recognised. This corresponds well with the late Pleistocene age of the palaeosol recognised in Lake
Victoria. It is not known if Lake Albert dried out completely at this time. Shortly after 15 ka the lake level rose at the core site and relatively high clastic input, presumed from enhanced river discharge, continued until about 9 ka. Although Lake Albert is today linked to Lakes Kyoga and Victoria via the Victoria Nile, Beuning et al. (1997) suggest it likely that the contemporaneous late Pleistocene lake level rise is due to a regional shift to more humid conditions rather than the arrival of the Victoria Nile in Lake Albert. This is evidenced by a change from arid grasslands to semi-deciduous forest communities. There is a re-expansion of grasslands between 13 and 11 ka before a return to greater semi-deciduous forest after about 11 ka. There are no deposits preserved at the core location with ages between about 8.9 and 3.6 ka which may have been caused by a low water stand towards the end of this period (Beuning et al., 1997). Since 3.6 ka the accumulation rates have remained low with an expansion of grassland communities.

The current sedimentological and geochemical data available from Lake Edward only extend back to 11 ka BP (Russell et al., 2003), but does include information about the period missing from the Lake Albert record described above. Organic, diatomaceous mud indicates shifts in wind intensity and stratification within an overall early Holocene wet phase. However, the appearance of authigenic calcite at 5.2 ka indicates the onset of aridity and a period of falling lake levels that culminated in a lowstand between 4 ka and 2 ka. The lake level then rose rapidly, attaining modern positions by 1.7 ka.

4.4 Historical Studies of Palaeodrainage

4.4.1 Palaeodrainage patterns

The passive margins which surround most of Africa have facilitated relative stability and subdued tectonics, which has tempted researchers to interpret the long-term historical drainage patterns established by rivers eroding the land surfaces described in the previous section. The description of palaeodrainage patterns in Africa, and southern Africa in particular, became fashionable in the 1940s and 50s (King, 1944, Dixey, 1955). These were largely based on geomorphological and geological evidence of superimposed and antecedent drainage. Superimposed drainage is that which developed on softer overlying strata and now retains the same pattern on underlying
more resistant strata. Antecedent drainage is that which developed prior to the activation of structures such as folds and faults which it now crosses. Geomorphological evidence for increases or decreases in stream power such as river capture and erosion surfaces, or relict valleys and aggradation may be explained by either climate or tectonic-driven changes in base level. It is often difficult to determine which mechanism was the cause. In reviewing the palaeofluvial geomorphology of southern Africa, Dollar (1998) suggests that older (early to mid Pleistocene) drainage modifications have usually been attributed to a tectonic mechanism, whereas younger (late Pleistocene to Holocene) drainage modifications are usually attributed to climate changes. This is likely because of the longer duration often required for tectonic modification of the landscape compared to climatic alteration of the hydrology. Our models of late Pleistocene to Holocene climatic variation have a much higher temporal resolution compared to Neogene tectonic models.

Wayland (1931) reports that W. C. Simmons first noted the influence of tectonics on the drainage pattern in western Uganda. The arrow-barb pattern of Lake Kyoga and tributaries to the Katonga point westward, which together with the western ‘swamp divides’, suggest the original direction of flow was towards the west. Wayland (1931) also records that Combe interpreted furrows on the bed of Lake Victoria as the channels of the former westward flowing rivers. These early researchers therefore came to the conclusion that the rivers of western Uganda, principally the Kafu, Katonga and Kagera originally formed the headwaters of the Congo drainage system prior the development of the Western Rift (Taylor and Howard, 1998).
Figure 4-24: Interpretation of pre Lake Victoria drainage pattern after: a) Cooke and b) de Heinzelin (from Temple, 1966)

Most early researchers, like Cooke (1958, cited in Temple, 1966) assumed that the westward flowing rivers extended across the area now occupied by Lake Victoria (Figure 4-24a). In examining the drainage evolution of Kenya, Ojany (1971) suggested that the proto-Katonga and proto-Kagera joined headwater streams now on the eastern shore of Lake Victoria in western Kenya. In contrast, de Heinzelin (1959, cited in Pickford et al., 1993) suggests a radially convergent drainage pattern existed in the northern Victoria basin with an outlet from the southern shore (Figure 4-24b). These interpreted drainage patterns are not necessarily contradictory, and may have existed at different times. Both Cooke and de Heinzelin appear to assume the proto-Katonga flowed down the same valley as the current Mpanga into the George basin. De Heinzelin also suggested the river continued down the current Kazinga Channel towards the Edward basin. De Swardt and Trendall (1969) point out that the Kazinga Channel between Lake George and Lake Edward is 30 to 90 m deep and was likely cut by a powerful stream in direct line with the Mpanga and Rusangwe, and therefore may have once formed a continuation of the westward flowing River Katonga system. Cooke (1958, cited in Temple, 1966) further speculates that the proto-Katonga joined the proto-Kagera and flowed north towards the area now occupied by the Beni Gap through the western flank of the Western Rift (Figure 4-24a).

Taylor and Howard (1998) support the view that the development of the South Atlantic in the mid-Cretaceous prompted the development of the Congo drainage system which
extended across current day equatorial Africa all the way to western Kenya. However, Reeves et al. (2004) have suggested that passive margin uplift occurred on the rift flanks following opening of the South Atlantic. The current Congo River cuts through highlands parallel to the coast in a deeply incised gorge (Goudie, 2005) which suggests current drainage direction to the South Atlantic may have recently been established following river capture. Previous authors have suggested such diverse late Mesozoic to Early Cenozoic outlets for the River Congo as Lake mega-Chad to the north (Moguedet et al., 1990, cited in Giresse, 2005) and the Indian Ocean via the River Lualaba and Rufiji to the east (Stankiewicz and de Wit, 2006). Based on either offshore sedimentological or onshore geomorphological evidence, the time at which the River Congo began to drain to the Atlantic has previously been estimated to have occurred anywhere between the Eocene and the Pleistocene (Anka et al., 2010). However, Anka et al. (2010) have reported recent offshore seismic reflection data and flexural modelling of the Congo deep sea fan which suggests that the River Congo’s outlet has remained stable since the late Cretaceous after all.

Turning our attention back to the eastern Katonga Valley, Temple (1966) examined the location of topographic channels and troughs on the bed of Lake Victoria which he interpreted from early Admiralty bathymetry charts (Figure 4-25a). The average depth of Lake Victoria is approximately 40 m and the maximum depth is about 68 m (Johnson et al., 2000). As shown in Figure 4-25b, Temple (1966) attempted to identify the remnants of fluvial channels on the bed of Lake Victoria, which he classified as less than 30.5 m (100 ft) deep (near-shore), between 30.5 and 61.0 m (100 to 200 ft) deep, and over 61.0 m (200 ft) deep. The charts also indicated the existence of a wide trough over 76.2 (250 ft) deep in the east central part of the lake. Bishop and Trendall (1967) conducted a similar exercise and concluded that pre-lake drainage was masked by sediments immediately offshore from the mouths of the rivers Katonga and Kagera. They suggest that channels can be identified close to the northern shoreline where sediment supply from the small catchment to the north is limited and a SSW-NNE trench near Jinja may be one of the main tributaries of the proto-Katonga. The deep WSW-ENE trending trough in the east central part of the lake appears contiguous with the Kavirondo Gulf which perhaps indicates a tectonic origin. Temple (1966) suggests a pre-lake drainage pattern that connects the isolated channels to the present day mouths of the River Kagera and Katonga. This is a reasonable hypothesis but his
interpretation appears influenced by his preconception that such a pattern may be expected to exist.

Figure 4-25: a) Lake Victoria bathymetry and b) interpretation of pre-lake drainage by Temple (1966)

Seismic surveys of Lake Victoria conducted during the 1990s (Johnson et al., 1996, Scholz et al., 1998, Johnson et al., 2000, Stager and Johnson, 2008) indicated lacustrine sediments of several 10s metres thick and a maximum thickness of about 60 m. Three buried erosion surfaces were identified on the seismic sections and the most recent occurs beneath up to about 9 m of sediment. The surveys revealed that the lake floor is smooth over most of its extent and the only significant relief appeared to be associated with submerged inselbergs (Scholz et al., 1998). However, there is considerable relief on the basement topography underlying the sediment. High resolution echo sounding did detect a submerged valley near the mouth of the River Kagera (Scholz et al., 1990), but there was no evidence of its continuation across the lake. Therefore although the pattern of the lake shore, particularly around the mouth of the Katonga, provides ample evidence of submerged river valleys, the seismic and echo sounding surveys appear to cast doubt on the strength of evidence for the location of palaeodrainage channels beneath the deeper parts of the lake. If there are indeed submerged channels on the lake
bed as Temple (1966) suggests then question remains, do they reflect the original headwater valleys of the Katonga and Kagera rivers eroded in the bedrock surface, or are they features produced by a pattern of internal drainage eroded in the lake sediment during the last desiccation event?

Figure 4-26: Interpreted drainage evolution of Uganda since the upper Miocene after Pickford et al. (1993)

As shown in Figure 4-26, Pickford et al (1993) attempt to reconstruct the drainage evolution of Uganda since the Miocene based on the previous work discussed above, palaeo-biogeography, and their investigation of sediments in the Western Rift. They suggest that uplift first occurred on the eastern margin of the East African Plateau and established a hydrographic connection between Uganda and the Congo drainage basin. They propose that a shallow downwarp in the region of the Western Rift developed during the mid to late Miocene, presumably coincident with uplift of the western margin. Initially the influx of fluvial sediment kept pace with basin subsidence and the
rivers continued to flow westward (Figure 4-26a). Sedimentological evidence indicates the formation of a large lake, named Lake Obweruka by Wayland, which occupied the Western Rift from the late Miocene to late Pliocene (Figure 4-26b). Pickford et al (1993) indicate that western molluscs in Lake Turkana suggest a hydrographic connection to Lake Obweruka throughout the Pliocene. They show Lake Wembere (also called Lake Manonga) which formed on the East African Plateau to the south of the Katonga-Kagera river basins, draining towards the Indian Ocean at this time. They also indicate that palaeontological evidence suggest Lake Obweruka was connected to the Congo palaeogeographic district and therefore propose that an overflow continued to the west in the area now west of the Rwenzori Mountains, known as the Beni Gap. Pickford et al (1993) suggest the westward outflow continued throughout the Pliocene and into the late Pleistocene, even after regression of Lake Obweruka and initial development of the Rwenzori Mountains (Figure 4-26c). They state that Nilotic fauna did not become evident in the Albertine faunas until the middle Pleistocene. On the basis of the elevation correlation of mollusc associations, they suggest a regional tectonic tilt occurred in the late Pleistocene and this left the Beni outflow perched above the lake and led to the development of the northern outflow into the Nile (Figure 4-26d). They appear to suggest a late date of 12 ka for river reversal and development of Lake Victoria which is inconsistent with the evidence presented below. It is interesting to note that while Pickford et al (1993) suggest the outflow to the White Nile may not have been initiated until the Pleistocene, recent oil drilling in the Victoria Nile delta play have encountered reservoirs at 1000 m depth suggesting a similar environment of deposition has existed in this area since the upper Pliocene (Ovington and Burden, 2009)

Working in Rwanda to the south, on the eastern shoulder of the Western Rift, Holzforster and Schmidt (2007) were able to take advantage of a well exposed and detailed Neogene fluvio-lacustrine sedimentary record to reconstruct the history of the palaeo-Nyabarongo River. They showed that river reversal at this location was initiated by a middle Pleistocene lava flow associated with the Virunga volcanic region, and infer that the major reversal from drainage towards the rift to drainage away from the rift occurred over approximately 300-350 ka. Unfortunately, the fluvial sedimentary record is often poorly preserved and/or exposed in the valleys of western Uganda. However, the overall drainage pattern and the tectonic regime strongly support the
hypothesised river reversal in this region. Nevertheless, many of the palaeodrainage patterns proposed by previous researchers appear quite speculative. Whilst the palaeodrainage reconstructions are inevitably drawn in plan it should be remembered that the river channels were also descending by 10s to 100s of metres during the denudation of the current landscape. Therefore, the interplay between uplift and erosion, stream power and direction during the Neogene is potentially quite complex.

4.4.2 Drainage reversal and the surface water divide

Early geomorphologists working in East Africa soon recognized the influence of tectonics on the current landscape. Gregory (1921) envisaged Lake Victoria occupying low lying ground between two extended north-south arches. He believed the rift valleys were similar to keystones which had collapsed due to lateral tension during the foundering of Gondwana. In contrast, Wayland suggested compression (1934b) was responsible for the formation of the arches and downward forcing of the rift valleys. His observations of alluvial deposits located over 30 m higher than Lakes Victoria, Albert, Edward and George led him to suggest they represented raised beaches and/or deltaic deposits. He went on to give a predominantly climatic rather than tectonic explanation, involving high lake levels due to wet ‘pluvial’ periods (Wayland, 1930, Wayland, 1934b). Wayland (1934b) first published the view that the westward pointing arrow-barb pattern of tributaries to the eastward flowing Kafu and Katonga rivers suggested that they had been reversed ‘by earth-movements with which the formation of the rift valley was associated’.
Figure 4.27: Stages in landscape evolution of southern Uganda (after Doornkamp and Temple, 1966)
Figure 4-27 presents Doornkamp’s and Temple’s (1966) classic model of the landscape evolution of south west Uganda based around the upland, lowland and infill landscapes described in Appendix C. The main phases described by Doornkamp and Temple (1966) are summarised below.

A) Upland erosion cycle – this probably represents the Buganda Surface at a time when the main drainage was towards the west.

B) Regional uplift starts lowland erosions cycle – the base level change initiated river erosion and the drainage pattern became partly antecedent.

C) Rift depression and adjacent upwarp with river incision – the upland landscape elements become isolated and antecedent drainage through the rift adjacent upward is exaggerated.

D) Upwarp, river reversal, lake formation and infill cycle – further upwarp overcomes erosion, ponding creates Lake Victoria and aggradation begins east of upwarp.

E) Recent warping leads to Lake Victoria regression – strandlines and lacustrine deposits are exposed by warping before lake levels fall.

The upland landscape currently slopes from about 1,300 m asl around Masaka in the east, to 2,000 m asl on the rift margin, 160 km to the west (Temple, 1966). The lowland landscape rises from about 400 m asl to 700 m asl over the same distance. The upland landscape therefore rises by up to 700 m while the lowland landscape rises by up to 300 m, thus indicating that land surface deformation began while the rivers flowed towards the west thus creating antecedent drainage.

This broad conceptual model is largely based on geomorphological mapping and careful observation of the current landscape. It only considers exhumation within the scale of the current relief and does not consider the mechanisms of tectonics or whether these are likely to be continuous or episodic. It is only concerned with the local surface expression of tectonic processes which actually act at great depths and are influenced by regional forces. Consequently the tectonic processes are viewed in terms of their influence on the topography rather than on the rock mass, and terms such as ‘upwarp’, ‘downwarp’ and ‘tilting’ are used in relation to land surface deformation only. Doornkamp and Temple’s (1966) conceptual model, like others of the time, does not
incorporate isostasy and lithospheric flexure, and assumes that the process of tectonic uplift acts independently of denudation.

Figure 4-28: Major tectonic features of Uganda (after Bishop and Trendall, 1967)

Bishop and Trendall (1967) combined the study of erosion surfaces with their own work on the sedimentary record, including that preserved in the Western Rift and the Kagera Valley in particular. In attempting understand the Neogene history and regional correlations they constructed the map of major tectonic features shown in Figure 4-28. Bishop and Trendall (1967) envisage an axis of uplift parallel to the Western Rift, and an axis of downwarp through the centre of Lake Victoria, the Victoria Nile outlet at Jinja and the eastern part of Lake Kyoga. On the western side of Lake Victoria, they also drew a line between the ‘zone of drowning’ and the ‘zone of late river reversal’. The location of this line is determined by the western limit of wide lacustrine sediment
filled valleys, and a change of gradient from 0.9 m/km to 1.74 m/km on the lower River Kagera between Nsongezi and Nyabusora (Figure 4-29). Bishop and Trendall (1967) speculate that the lower 3 m of coarse sediment consisting of boulders and gravel at Nsongezi was deposited when the Kagera flowed westward. No direct dating is possible, but they indicate that early African bush elephant molars (Loxodonta africana Arambourg) from similar deposits at Kikagati have been previously ascribed to the early Pleistocene. They also suggest that the overlying 15 m of mainly sand is indicative of post-reversal lacustrine deposits. A layer within the sand member contains stone tools of mainly late Acheulian design with Sangoan influences near the top. They suggest that tentative correlation with radiocarbon dating for similar assemblages at the southern end of Lake Tanganyika indicates they are likely around 50 ka. On the basis of this speculative interpretation and wide age range they propose that river reversal occurred ‘by the end of mid-Pleistocene times’.

Figure 4-29: Longitudinal profile of the Kagera Valley (after Bishop and Trendall, 1967)

Wayland (1929, 1934b) and many subsequent researchers interpreted the ‘axis of upwarp’ to coincide with the current drainage divide which they assumed was the location at which the original river reversal occurred. Doornkamp (1970) pointed out that this is not necessarily the case and the location of the ‘axis of upwarp’ could be derived in several different ways, including:

a. the maximum difference in altitude between his upland and lowland landscapes;
b. the maximum elevation of his upland landscape; and,

c. the maximum elevation of the lowland landscape.

Figure 4-30: Stream directions south of the Mpanga- Katonga Valley System (after Doornkamp, 1970)

Doornkamp’s (1970) axes lie to the west of the line formed by Wayland’s ‘swamp divides’. Figure 4-30 shows that the drainage divide is close to the axis of upwarp in the south. However, the River Rusangwe catchment lies entirely east of Doornkamp’s axis of upwarp and discharges into the River Mpanga which flows westward through the axis of upwarp towards Lake George. The Mpanga-Rusangwe valley system is connected to the Katonga valley at the Bihanga Station ‘swamp divide’. Doornkamp (1970) describes how knick points would be established in initially westward flowing rivers as the upwarp progresses. He pointed out that the deepest incision will occur around the axis of upwarp but the knick point may lie east of the axis as has occurred on the Mpanga-Katonga Valley System. The previous large catchment of the westward flowing western River Katonga, southward flowing River Mpanga, and northward flowing River Rusangwe, provided enough power to maintain bedrock erosion through the axis of upwarp. Eventually uplift overcame erosion upstream of the Mpanga and Rusangwe confluence, and the drainage divide was formed at the head of the Katonga
Valley. Section 8.5 discusses the possibility that the original drainage divide on the Mpanga-Katonga Valley System was located much further east.

![Diagram](image)

Figure 4-31: Changes in the drainage pattern of southwest Uganda as a result of tectonics (after Doornkamp and Temple, 1966)

As shown in Figure 4-31, the Ruizi and Kibale river systems located south of the Katonga drainage basin have had a similarly complex reversal history. Prior to upwarp and river reversal, Doornkamp (1970) suggests that the River Ruizi was a tributary to the westward flowing River Kagera. Local north-south water divides occurred between the rivers Ruizi, Orichinga and Kibale which were all tributaries of the Kagera. Today the Ruizi-Kibale basin drains eastward and discharges into Lake Victoria. Given the pattern of north-south drainage divides between the main Katonga-Kagera divide, it can
be envisaged that the interplay of uplift and erosion which causes east-west migration of the main north-south divide will result in the discrete capture of local drainage basins by the main east or west flowing rivers.

As well as the upwarp parallel to the Western Rift and downwarp along the north-south axis of Lake Victoria, several researchers have suggested an overall north to south tilting of the Ugandan landscape. This interpretation is based on the observed change in altitude of correlated erosion surfaces, and correlated stratigraphy in the Western Rift. Lepersonne (1949a, cited in Pickford et al., 1993) pointed out that the top of the Semliki Series fell by 300 to 450 m between Lake Edward and Lake Albert, and the top of the Kaiso Series fell by 300 to 550 m over the same distance. Pickford (1990, 1993, 2010) presents evidence that sedimentary strata in the Western Rift containing correlated Pliocene mollusc associations also show a general decline in altitude from south to north of between 350 and 800 m. Pickford (1993) implies that this may be associated with ongoing uplift of the East African Plateau, but also warns that the blocks of sediments for which these elevations are recorded are themselves subject to local rift related tectonics.

Wayland (1925) described a break in slope on the rift escarpment which he names the ‘hanging base line’. He infers that the lower, steeper facet is associated with renewed fault movement during the Pleistocene. Additionally he noted that the ‘hanging base line’ is 130 m above the lake in the north and rises to over 700 m in the south. Pickford et al. (1993) state that this further suggests a north-south component of tilt, but do not recognise that it could be produced by differential uplift of the rift flank or down throw of the rift floor in the north. They also conclude that the eastern rift flank at Nyabusosi and Kaiso, south and east of Lake Albert respectively, actually displays three facets. They suggest that the initial faulting which created the middle facet is associated with a change from the deposition of fine-grained lacustrine sediment to coarser grained sediment derived from the uplifted rift flank at the end of ‘Kaiso’ times around 2.3 Ma (see Appendix D). However, as discussed further in Section 4.5.1, the relationship between sediment accommodation and supply cycles with the tectonics is complex and simple correlations between slope facets and stratigraphic boundaries must therefore be regarded as speculative over interpretation. They go on to propose that Wayland’s lower ‘hanging base line’ is associated with a younger faulting episode which resulted...
in river reversal and the formation of Lake Victoria about 12,000 year ago. They do not appear to appreciate that uplift of the rift flank may occur due to the gradual process of flexural loading within the rift, rather than due to independent tectonic episodes. The reason for proposing such a late date for river reversal is unclear and as will be shown below appears to be inconsistent with the depth and rate of sedimentation in the Victoria Basin.

4.4.3 The origin of Lake Victoria and strandline development

The evidence for dating the origin Lake Victoria originally presented by Temple (1966) may be updated and summarised as follows:

- the lake formation may associated with development of the northern section of the Western Rift and therefore is likely younger than 8 Ma (Nyblade and Brazier, 2002);

- the considerable thickness of sediment in the Western Rift suggests the source region was larger than the current rift flanks and therefore the main drainage reversal likely occurred some considerable time after the initial rifting (Doornkamp and Temple, 1966);

- the modern fish faunas of Lake Edward and Victoria are similar with a few exceptions, and Beadle (1981) has suggested a late Pleistocene connection via the River Katonga between the two lakes;

- the late-Acheulian artefacts in the lacustrine sands at Nsongezi on the River Kagera have been dated by analogy with C14 dated artefacts in southern Tanzania at 60 ka; and

- given that the maximum thickness of lacustrine sediments determined from seismic surveys is about 60 m (Johnson et al., 2000), a crude calculation based on an estimated average sedimentation rate of 150 mm/1000 years indicates the lake origin at about 400 ka.

Therefore, it would appear that the modern Lake Victoria dates from the middle to late Pleistocene. Interestingly, river reversal may have occurred earlier in Tanzania further south, where palaeontological evidence from lacustrine sediments suggests a Pliocene age for a shallow lake basin located in the valleys of the present day Manonga and Wembere Rivers, known as Lake Manonga by Stewart (2001). While palaeontological
evidence suggests the early Pliocene lake may have drained into the Indian Ocean, two new fish species appear in the later Pliocene indicate it may have been connected to the Kagera-Katonga drainage basin prior to the existence of Lake Victoria (Stewart, 2001).

Following its initial formation, Lake Victoria appears to have attained a much greater extent than it is at present. The lakes maximum extent was at least 100 km west of the current shoreline as demonstrated by the lacustrine sands at Nsongezi on the River Kagera (Bishop, 1969). Evidence of the former lake high stands were once again first recorded by Wayland (1934b). Subsequent studies were conducted by the Uganda Geological Survey in the 1950s (Pallister, 1953, Johnson, 1955a, Bishop, 1959). However, the most extensive survey were conducted by Temple (1966) who undertook 160 km of levelling along the north west and south east shoreline of Lake Victoria. Evidence of former lake levels include raised beach deposits, wave-cut platforms, sand bars and raised swamps. On the basis of these surveys, Temple assigned elevations to interpreted raised beach deposits and wave-cut features.

![Figure 4-32: Elevations of raised beaches along the north shore of Lake Victoria (after Temple 1966)](image_url)

The elevations of features related to former lake levels identified in four areas on the north shore of Lake Victoria are shown in Figure 4-32. It can be seen that there is a
large scatter, perhaps suggesting numerous minor still stands during lake regression. However, Temple (1966) suggested that the elevation of the lower raised beaches cluster around 10 feet (3 m) above Lake Victoria. There is another cluster at about 45 feet (13.7 m) above Lake Victoria. The elevation of higher raised beaches appears more scattered, but Temple draws an average line at 60 feet (18.3 m) above Lake Victoria. Temple (1966) also identified two higher strandlines at about 80-85 feet (25 m) and 180-185 feet (55 m) above Lake Victoria which he named the Kiwala and Kayugi strandlines after type localities west of Lake Nabugabo, south of the Katonga mouth. The elevations of raised strandlines interpreted by Temple and previous researchers are summarised in Table 4-2. Despite the scatter of elevations for the raised strandlines shown in Figure 4-34, there appears to be relative consistency between the elevations selected by different researchers for the dominant features.

<table>
<thead>
<tr>
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</tr>
</thead>
<tbody>
<tr>
<td>S1: ‘Kilwala’ (tilted)</td>
<td>51.8</td>
<td>48.8</td>
<td>57.9-62.5</td>
<td>61.0-67.1</td>
<td>54.9-56.4</td>
</tr>
<tr>
<td>S2: ‘Kayugi’(tilted)</td>
<td>38.1</td>
<td>33.5-35.1</td>
<td>33.5-35.1</td>
<td>24.4-25.9</td>
<td></td>
</tr>
<tr>
<td>S3: ‘60-65 feet’</td>
<td>21.3</td>
<td>18.3-19.8</td>
<td>18.3-19.8</td>
<td>18.3-19.8</td>
<td></td>
</tr>
<tr>
<td>S5: ‘10-12 feet’</td>
<td>4.6</td>
<td>3.0-4.6</td>
<td>5.2</td>
<td>3.0-3.7</td>
<td></td>
</tr>
</tbody>
</table>

Note: Elevations from Doornkamp and Temple (1970) above average lake level converted from feet to metres.

Table 4-1: Strandline elevations (m above Lake Victoria) measured on northwest shore

Temple (1966) suggests five possible causes for changes in the level of Lake Victoria:

1. Changes to climate parameters including precipitation and evaporation;
2. Catchment change, due to river capture for example;
3. Isostasy and lithospheric deformation;
4. Regional tectonics (‘tilting’ and ‘warping’); and
5. Reduction in the elevation of the lake outlet due to erosion.
While the interpreted lower three strandlines remain at similar elevation around the lake shore, the upper two strandlines appear to increase in elevation towards the west (Bishop and Trendall, 1967, Temple and Doornkamp, 1970), indicating continued tilting of the land surface after their formation. For example, the Kayugi strandline increases from about 25 m above the lake near Lake Nabugabo to up to 35 m above the lake in the lower Katonga Valley 25 km to the west. It is interesting to note that the higher strandline is about 1,195 m asl which is a similar elevation to the drainage divide at Bihanga Station in the western Katonga Valley. Johnson (1955a) states that the Kayugi and Kilwala strandlines are uplifted by about 18 m and 23 m respectively at their western limit. Bishop and Trendall (1967) show that the upper surface of the lacustrine deposits in the Kagera valley (Kafunzo flat), which is about 37 m above Lake Victoria at Nsongezi, runs into a strandline at about 30 m above the lake, which Temple and Doornkamp (1970) suggest is the Kayugi strandline of the Lake Nabugabo area.

Temple and Doornkamp (1970) interpret the tilting of the high level strandlines to indicate that the initial regression of Lake Victoria occurred as a result of tectonic deformation of the land surface (‘warping’). However, they suggest that the lower strandlines which are not affected by tectonics were formed due to phased incision of the lake outlet via the Victoria Nile at Jinja. Neither explanation relies on changes in climate, although this was occurring at the same time.

If Kayugi strandline is equivalent to the Kafunzo Flat then it must be younger than the late-Acheulean artefacts dated at 60,000 yrs BP and Bishop and Posnansky (1960) have suggested a date of 25,000 yrs BP. The lowest strandline at 3 m above the lake in Hippo Bay near Entebbe has been dated using radiocarbon at 3,720 ± 200 yrs BP. The 14 m and 18 m strandlines may therefore be expected to be very late Pleistocene in age.
Figure 4-33: Three stages in the evolution of the Katonga Mouth: 1) Pre-lake, 2) maximum lake extent 3) present drainage (Temple and Doornkamp, 1970)

Temple and Doornkamp (1970) examined the recent evolution of the wide eastern section of the Katonga Valley downstream of the Kisozi crossing. Their ambitious reconstruction of the pre-lake drainage pattern reconstructed using the shape of the current shoreline is presented in Figure 4-33a. The maximum extent of Lake Victoria
shown in 4-33b shows the area of lacustrine deposition extending to beyond the current crossing at Kisozi. Major spits and bars of white quartz sand and gravel run parallel to the present shoreline and appear to have blocked the old mouth of the River Katonga south of the Goru peninsula, causing it to be diverted to north of the peninsula (Bishop, 1959). Older sand bars of yellow-red sands extend across the mouths of the northern tributaries parallel the main Katonga Valley. Johnson (1955a) originally noted that they accumulated from east to west and he proposed they were river bank shoals from when the River Katonga flowed westward, but Bishop (1959) indicates it is far more likely that their direction of accumulation is attributable to the prevalent wind and wave direction in a lacustrine environment and they were deposited during the lake highstand. The current form of the eastern Katonga Valley and north west lake shore is shown in Figure 4-33c.

When interpreting the raised strandline it is important to note that if the lake level fell and rose again between strandline formations there would be no record of the event. Therefore it is not known if the stages of lake transgression occurred in one direction followed by stages of lake regression in the other direction, or if there were intermediate stages of lake transgression and regression. Given the evidence presented in Section 4.3.3, it would seem likely that at least the higher tilted strandlines formed before Lake Victoria last dried out between about 16 and 14 ka.

4.5 The Neogene Sedimentary Record

The sedimentary record preserved in the sedimentary basins of the Western Rift, the relict river valleys of western Uganda and the Victoria Basin is significant for our understanding of the Neogene history of the Katonga region. The sedimentary deposits in the Katonga Valley itself are poorly exposed and little has been written about them. However, the history of the Katonga Valley is intimately connected to the history of sedimentation in the rift valley at its western end and the Victoria basin at its eastern end. In addition, gold prospecting during the 1920s in the Kafu Valley to the north, and archaeological investigations during the 1930s and 1950s in the Kagera Valley to the south, has led to sedimentological observations of potential relevance. This section examines each of the available sedimentological records for evidence of the Neogene landscape evolution which influenced the Katonga Valley.
4.5.1 The Western Rift

The primary motivations for examining the sediments in the Western Rift are: 1) to determine if they include indurated rocks of similar character to the fine-grained sandstone identified at Kabagole and Bihanga Station in the Katonga Valley; and, 2) to examine the Pleistocene record for vertical variations in sediment character which might indicate changing stream power and sediment supply from the rivers of south west Uganda due to changing climate and/or relief on the rift margins. This section describes those studies conducted after 1990. A summary of the historical studies primarily undertaken in the 1950s and 1960s is included in Appendix D.

The location of the sedimentary basins in the Western Rift, between Lake Edward and Lake Albert is determined by the Neogene tectonic evolution of the region. Bauer et al (2010) describe the tectonic history in five stages represented on a west to east section through the Rwenzori Mountains in Figure 4-34 and summarised below.

1. ~20 to 18 Ma – Pre-rift early faulting and subsidence.
2. ~12 Ma – Virunga volcanism and development of a downwarp with fluvial sedimentation.
3. ~10 to 8 Ma – Tectonic episode in the Albert Basin with synrift sedimentation.
4. ~8.5- 8 Ma – 1st major episode of rifting and development of Lake Obweruka
5. ~3- 2 Ma – 2nd major episode rifting with uplift of the rift shoulders and the Rwenzori.

The main sedimentary basins that exist today in the Western Rift adjacent to the Katonga Valley are shown in Figure 4-35 and listed below together with the maximum estimated age of sediments logged in outcrop.

- Albert Basin
  - Kisegi-Nyabusosi area – mid Miocene to Holocene (15 Ma)
  - Kaiso-Nkondo area – late Miocene to Holocene (6 Ma)
- Semliki Basin – mid Pliocene to Holocene (4 Ma)
- George and Edward Basins (3 Ma)
Figure 4-34: Stages in tectonic evolution of the Western Rift shown on a west to east section through the Rwenzori Mountains (after Bauer et al., 2010)
The maximum age of sediment encountered during drilling in the Albert Basin is disputed, with recent reports suggesting it is no older than the upper Miocene (Ovington and Burden, 2009) in the northern Albert Basin, while previous interpretations have suggested Cretaceous and possibly even Karoo sediments at the base of the rift (D.G.S.M., 2005). A borehole drilled in 1938 northeast of Lake Albert, near Butiaba encountered fluvio-lacustrine sediments to 1,222 m bgl (Harris et al., 1956). Seismic reflection surveys for oil exploration appear to indicate sediments up to 2,000 m thick on the eastern side of the Albert Basin and up to 4,000 m thick on the western side of the basin (D.G.S.M., 2005). Oil exploration wells have recently been drilled up to 3,500 m bgl (Roller et al., 2010), but the results are not yet publically available.

The Semliki Basin is located on the western side of the Rwenzori horst, and the Albert Basin is located north of the region of Achaean and Paleoproterozoic rocks which
connect the Rwenzori horst to the eastern rift flank (Link et al., 2010). The George and Edward Basins are located immediately south west of the Mpanga- Katonga Valley System. The sedimentary archive in this downstream location may therefore be expected to contain the most relevant information about the landscape evolution of the Katonga region. However, the oldest and most complete sequence of sediments occurs in the Albert Basin, which have also been the subject of the most recent investigation. These recent studies are described here, whilst the historical studies in all of the basins are summarised in Appendix D.

Recent studies of the Western Rift sediments are considered here to have begun in the mid 1980s with the Uganda Palaeontological Expedition (Pickford et al., 1993) and continue today with the ongoing RiftLink project (Roller et al., 2010) and oil exploration by Tullow Oil plc and Heritage Oil plc. Unfortunately, with the exception of one or two government reports and conference papers (e.g., Ovington and Burden, 2009, Lirong et al., 2004) few results from oil exploration have yet been published. While Pickford et al. (1993) focus on the palaeobiology they do construct a more detailed stratigraphy of the Kisegi and Kaiso areas than proposed previously. They criticise much of the early research conducted in Uganda for imprecise biostratigraphic classification and attempts to force observations into the preconceived conceptual framework of pluvial and interpluvial periods influenced by global Pleistocene climatic cycles. Pickford et al. (1993) point out that sediments of Miocene to Pleistocene in age were all previously grouped into the Kaiso Series (Bishop, 1969) mainly because they all contained ironstones. However, these ironstones can be categorised into six different types as follows:

1. Fossiliferous pisolithic and oolitic ironstones – precipitated at the surface;
2. Massive to shaley onion skin ironstones – precipitated during diagenesis
3. Scoriaceous ironstones – bioturbated oolitic or shaley ironstones;
4. Fossiliferous sand and grit with iron cement – post deposition iron precipitation
5. Ironstone nodules from clay horizons – diagenetic origin; and,
6. Pedogenetic iron rich horizons – surficial formation of murrum and ferricrete.
<table>
<thead>
<tr>
<th>EPOCH</th>
<th>MOLLUSC ASSOC.</th>
<th>KISEGI AREA</th>
<th>KAIKO AREA</th>
</tr>
</thead>
<tbody>
<tr>
<td>Holocene</td>
<td>G8 (recent)</td>
<td>Semliki Plains – Recent Alluvium</td>
<td>Lake Shore Alluvium</td>
</tr>
<tr>
<td></td>
<td></td>
<td><em>Unconformity</em></td>
<td><em>Unconformity</em></td>
</tr>
<tr>
<td>Late Pleistocene to Holocene</td>
<td></td>
<td>Wasa Beds (high plains alluvium)</td>
<td>Kaiso Plains Alluvium</td>
</tr>
<tr>
<td></td>
<td></td>
<td><em>Unconformity (river reversal?)</em></td>
<td><em>Unconformity (river reversal?)</em></td>
</tr>
<tr>
<td></td>
<td>G7 (0.5 Ma)</td>
<td>Rwebishengo Beds</td>
<td>Muisega Beds</td>
</tr>
<tr>
<td></td>
<td></td>
<td><em>Unconformity (river reversal?)</em></td>
<td><em>Unconformity (river reversal?)</em></td>
</tr>
<tr>
<td>Mid Pleistocene</td>
<td>G6 (1.8 Ma)</td>
<td>Nyabusosi Formation</td>
<td>Kagusa Member (1.8 Ma tuff)</td>
</tr>
<tr>
<td></td>
<td></td>
<td><em>Unconformity (river reversal?)</em></td>
<td>Museta Beds (1.5 Ma)</td>
</tr>
<tr>
<td>Early Pleistocene</td>
<td>GX’ (2.3 Ma)</td>
<td>Behanga Member</td>
<td>Kaiso Village Beds (2.3 Ma)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Kaiso Village Formation</td>
<td>Hohwa Beds (2.6 Ma)</td>
</tr>
<tr>
<td>Pliocene to Pleistocene</td>
<td>GX (2.6 Ma)</td>
<td>Katorogo Formation</td>
<td>Kyeoro Formation (3 Ma)</td>
</tr>
<tr>
<td>Late Pliocene</td>
<td>G5 (3 Ma)</td>
<td>Nyakabingo Formation (3 Ma)</td>
<td>Warwire Formation (3.45-3.6 Ma tuffs)</td>
</tr>
<tr>
<td></td>
<td></td>
<td><em>Unconformity</em></td>
<td><em>Unconformity</em></td>
</tr>
<tr>
<td>Mid Pliocene</td>
<td>G4c (3.4 Ma)</td>
<td>Nyaburogo Formation (5-3.5 Ma)</td>
<td>Kaiso Village Beds (2.3 Ma)</td>
</tr>
<tr>
<td></td>
<td>G4b (3.7 Ma)</td>
<td></td>
<td>Hohwa Beds (2.6 Ma)</td>
</tr>
<tr>
<td></td>
<td>G4a (4.3 Ma)</td>
<td></td>
<td>Kaiso Village Beds (2.3 Ma)</td>
</tr>
<tr>
<td>Early Pliocene</td>
<td>G3b (4.9 Ma)</td>
<td></td>
<td>Kyeoro Formation (3 Ma)</td>
</tr>
<tr>
<td></td>
<td>G3a (6.2 Ma)</td>
<td></td>
<td>Warwire Formation (3.45-3.6 Ma tuffs)</td>
</tr>
<tr>
<td>Late Miocene</td>
<td>G2b (7.1 Ma)</td>
<td>Oluka Formation (8-6 Ma)</td>
<td>Nyaweiga Member (5 Ma)</td>
</tr>
<tr>
<td></td>
<td>G2a (7.5 Ma)</td>
<td></td>
<td>Warwire Formation (3.45-3.6 Ma tuffs)</td>
</tr>
<tr>
<td></td>
<td>G1 (11-10Ma)</td>
<td>Kakara Formation (12-9 Ma)</td>
<td>Nkondo Member (7-6 Ma)</td>
</tr>
<tr>
<td>Mid Miocene</td>
<td>G0 (15-14Ma)</td>
<td>Kisegi Formation (mid Miocene)</td>
<td></td>
</tr>
</tbody>
</table>

Table 4-2: Biostratigraphy of Kisegi and Kaiso areas (after Pickford et al., 1993, Pickford, 2010)
### KISEGI AREA

<table>
<thead>
<tr>
<th><strong>Semliki Plains</strong> - Recent to modern alluvium</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Wasa Beds</strong> - Cover the Toro Plains. Similar in appearance to the Kaiso Plains alluvium</td>
</tr>
<tr>
<td><strong>Rwebishengo Beds</strong> - Clays and silts infilling valleys in the Nyabososi.</td>
</tr>
<tr>
<td><strong>Kagusa Member</strong> - Clays and silts with ironstones with thin tuff at the base (20 m)</td>
</tr>
<tr>
<td><strong>Makondo Member</strong> - Clays and silts with ironstones with Oldowan artefacts (20 m)</td>
</tr>
<tr>
<td><strong>Katorogo Formation</strong> - Poorly sorted pebbly clays and calcareous nodules (20-100 m)</td>
</tr>
<tr>
<td><strong>Nyakabingo Formation</strong> - Clays, silts, sands and microconglomerates with gypsum below ironstones near the top (60 m)</td>
</tr>
<tr>
<td><strong>Nyaburogo Formation</strong> - Clays, silts and rusty ironstones (120 m)</td>
</tr>
<tr>
<td><strong>Oluka Formation</strong> - Includes green/grey 0.5m silicified sandstone underlain by ironstone with late Miocene fossils (60 m)</td>
</tr>
<tr>
<td><strong>Kakara Formation</strong> - Onset of rifting. Red floodplain sediments. Upper Miocene fossils. Ironstone at top and rounded quartz pebbles (40 m)</td>
</tr>
<tr>
<td><strong>Kisegi Formation</strong> - Pale sands, silts and microconglomerates with gypsum, few fossils (110 m)</td>
</tr>
</tbody>
</table>

### KAISO AREA

<table>
<thead>
<tr>
<th><strong>Lake Shore Alluvium</strong> - Currently accumulating in river valleys and along the present lake shore, especially in back beaches and lagoons</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Kaiso Plains Alluvium</strong> - Silts and sands, grading towards the escarpment into sheet conglomerates and fanglomerates with artefacts. Characterised by palaeosols, murrum and kunkar layers</td>
</tr>
<tr>
<td><strong>Muisege Beds</strong> - Local wedge of sediment, comprising light brown fluvial fine muddy sand</td>
</tr>
<tr>
<td><strong>Kaiso Village Beds</strong> - Clays and silts with ironstone horizons (13 m)</td>
</tr>
<tr>
<td><strong>Hohwa Beds</strong> - Clays and silts with oolitic fossiliferous ironstones and tuff at the top (4 m)</td>
</tr>
<tr>
<td><strong>Kyeoro Formation</strong> - Sands and silts with minor but fossiliferous ironstones near top (over 8 m)</td>
</tr>
<tr>
<td><strong>Warwire Formation</strong> - Silts, sands and microconglomerates with limited ironstones. Thin limestone in the middle. Tuffs present (70 m)</td>
</tr>
<tr>
<td><strong>Nyaweiga Member</strong> - Oolitic ironstone overlain by ironstone boudins. Clays and silts with ironstone horizons with many fossils (up to 30 m)</td>
</tr>
<tr>
<td><strong>Nkondo Member</strong> - Clays and silts with oolitic and sandy ironstones. Fossil mammals indicate ‘latest’ Miocene (over 40 m)</td>
</tr>
</tbody>
</table>

**Table 4-3:** Summary descriptions of Kisegi and Kaiso strata after Pickford et al. (1993)

Figure 4-36: Geological map of the Kisegi area (after Pickford et al., 1993)

Figure 4-37: Geological map of the Kaiso area (after Pickford et al., 1993)
Pickford et al. (1993) used the biostratigraphic classification together with lithofacies to split the previously grouped ‘Kaiso Series’ into nine subdivisions in the Kisegi area south of Lake Albert (Figure 4-36), and six subdivisions in the Kaiso area east of Lake Albert (Figure 4-37). Their biostratigraphy and summary descriptions of the stratigraphic units in the Kisegi and Kaiso areas are summarised in Table 4-3 and 4-4. The discovery of tuffs in the Albert Basin succession facilitated biostratigraphic correlations and comparison with radiometric dates from across East Africa.

When considering this stratigraphy it is useful to keep in mind that the range of estimated dates for river reversal associated with rift flank uplift and the formation of Lake Victoria. Johnson et al. (2000) suggested reversal occurred about 400 ka based on the depth of sediment and rate of deposition in Lake Victoria. This is consistent with the general qualitative estimates of mid Pleistocene river reversal suggested by several earlier researchers (Wayland, 1934b, Bishop and Trendall, 1967, Doornkamp and Temple, 1966). If this date is correct then the Rwebishengo and Muisega Beds may have been deposited after river reversal occurred. However, these beds are characterised by clayey silts and sands and Pickford et al. (1993) suggest river reversal occurred after their deposition and as recently as 12 ka, prior to deposition of Wasa Beds and Kaiso Plains Alluvium (Table 4-4). Therefore, they assume that all stratigraphic units below G7 Gautier mollusc association were deposited prior to river reversal.

While Pickford et al. (1993) have created a systematic biostratigraphy with good age control, the Uganda Palaeontological Expedition was not focused on sedimentological features and facies. Sedimentological investigations in connection with the RiftLink project have recently been conducted in the Kisegi area (Roller et al., 2010), also visited by the author. The aim of their investigation is to examine the sedimentary archive for evidence of denudation characteristics, sediment flux, tectonic activity and climate change during the Neogene. Roller et al. (2010) estimate the maximum thickness of Neogene sediments in the Kisegi area to be about 600 m of which they logged 350 m in nineteen sections with at least one logged section in each biostratigraphic unit. They produced detailed logs of grains size, sedimentary structures, fossil content, gypsum, iron, ooid and calcareous content. From these descriptions they defined nineteen lithofacies, including: five fine-grained (clay, silt);
five medium-grained (sand); six coarse-grained (gravel); and three others which are volcaniclastic, iron ooliths and iron pisoliths. At a larger scale they also defined eight architectural elements, including: channel fill; levee; crevasse splay; floodplain fines; coastal mudflat; beach; delta foreset; and lake. The lithofacies and architectural elements indicated that the sedimentary succession was deposited under alternating fluvial and lacustrine conditions with transitional settings. These were classified into five main depositional environments which are: alluvial plain; delta plain; delta front; nearshore; and lacustrine.

Figure 4-38 shows the sedimentological summary of the Kisegi area. Roller et al. (2010) used the changing depositional environments to define depositional cycles. Deposition during these cycles is either accommodation controlled which results in coarse-grained shallow water sediments, or sediment-supply controlled which results in fine-grained deep water sediments (Cross and Lessenger, 1998). Retrogradational base-level rise (upright triangles in Figure 4-38) culminates in lacustrine environments of deposition and progradational base-level fall (upside down triangles in Figure 4-38) culminates in alluvial plain environments of deposition. Roller et al. (2010) suggest that meso-scale base level cycles likely represent autocyclic sedimentary processes, whilst macro-scale base level cycles likely represent allocyclic processes, such as the tectonic processes associated with the graben setting. Their interpretation of small and large base level cycles are shown in Figure 4-38.

The overall results summarized in Figure 4-38 show evidence a much more changeable environment of deposition than that interpreted by early researchers, who divided the entire sequence into the fluvial Kisegi Series and the lacustrine Kaiso Series, with intermediate Passage Beds (Pargeter, 1948, Bishop, 1969). Whilst there are generally a lower proportion of lacustrine deposits in the Kisegi Formation, than in later formations, the sedimentology indicates a far greater degree of complexity than a simple change from fluvial to lacustrine deposits.
Environment of Deposition

- lacustrine
  - delta front/delta plain
  - alluvial plain

alluvial plain

- lacustrine
- nearshore
  - delta plain/delta front
  - alluvial plain
  - nearshore

nearshore

alluvial plain

lacustrine

lacustrine
  - delta plain

alluvial plain

nearshore

lacustrine

coastal

delta front

delta plain

alluvial plain

**Figure 4-38: Sedimentology of the Kisegi area (Roller et al., 2000)**

**Key**

Sections: widths $\propto$ grain size

Colour codes: black – lacustrine, dark grey – transitional, light grey – alluvial

Base level cycles: thin – fine-grained, deep, thick – coarse-grained, shallow water
Roller et al. (2010) recognize five macro base level cycles. They point out that the coarse conglomerate overlying the Precambrian rocks at the base of the Kisegi Deposits must have been deposited in a high energy environment with pronounced relief, thus suggesting the onset of rift related faulting. They also suggest that the high fraction of feldspar and disseminated gypsum in the first macro cycle (Kisegi Formation) indicates a semiarid climate, whilst the appearance of iron impregnations and ooliths in the second cycle suggests a change to wet tropical climate. The reappearance of feldspar and decrease in iron impregnated beds at the top of the sequence in the Nyabusosi Formation suggests a return to more arid conditions during the Pleistocene.

The interpreted macro base level change cycles suggests there were five occasions on which the system become accommodation limited (alluvial plains) and five occasions on which it became sediment-supply limited (lacustrine). It is important to note that while downward displacement of graben floor may create more accommodation space it may also create greater relief on the rift flanks and hence increase the sediment supply. The relationship between the accommodation and supply cycles with the tectonics is therefore complex. Roller et al. (2010) do suggest that the high sedimentation rate of 100 m/Ma in the upper Pliocene and Pleistocene, compared to 20-45 m/Ma in the older strata, may be a result of increased supply from the Rwenzori, and/or decreased transport of sediment out of the basin. The youngest, unconformable alluvium of poorly sorted angular to sub-rounded gravel with clayey matrix is ascribed to episodic deposition in a large alluvial fan. They suggest that it may be representative of high relief and climate variability in the late Pleistocene and Holocene.

4.5.2 The Kagera Valley

The River Kagera is the principle river flowing into Lake Victoria and contributes the equivalent of about 30% of the total outflow to the Victoria Nile. The modern river begins about 25 km from Lake Tanganyika and flows northwards for about 400 km, eventually forming the border between Rwanda and Tanzania before turning eastward at Kikagati on the border between Tanzania and Uganda. It forms a fast flowing river for about 12 km between Kikagati and Nsongezi where it occupies a gorge cut perpendicular to quartzites of the Karagwe-Ankolean System (Figure 4-39). The Kagera is about 90 m above Lake Victoria at Nsongezi where it turns toward the southeast and follows the outcrop of argillaceous meta-sediments about 4 km wide,
between two quartzites bands (Figure 4-40). About 40 km downstream of Nzongezi at Nyabusora the river bends eastward onto the Buganda-Toro System and as shown in Figure 4-29 the gradient decreases from about 1.8 m/km to about 0.3 m/km (Bishop, 1969). As shown in Figure 4-39, it bends southeast again and finally northeast becoming sluggish in the lower flood plain before entering Lake Victoria.

Figure 4-39: Former maximum extent of Lake Victoria (after Bishop, 1969)
The most extensive investigation of the sedimentary record in the Kagera Valley has been undertaken at Nsongezi in the terraces at the confluence with the Orichinga Valley, 110 km west of Lake Victoria. In the 1930s and mid 1950s Wayland (1937) supervised extensive archaeological excavations including many trenches and 86 pits. In total, 9000 tonnes, equivalent to 4850 m$^3$, of material was removed manually by local labourers (Bishop, 1969). A set of lithological descriptions from an excavation which Wayland name ‘paddock 1’ are given in Table 4-5. The descriptions for the upper part of the excavation, shown in italics, were re-logged by Bishop (1969) whilst the lower part are Wayland’s (1937) original descriptions. Bishop (1969) groups Nsongezi Series deposits into three members: I) Boulder Gravel Member; II) Sand Member and; III) Clay Silt Member. Figure 4-41 shows the relationship between these lithological units on a generalised section across the northern margin of the Kagera Valley at Nsongezi.
### LITHOLOGICAL CLASSIFICATION

<table>
<thead>
<tr>
<th>III. Clay / Silt Member</th>
<th>ORIGINAL DESCRIPTIONS OF ‘PADDOCK 1’ EXCAVATION</th>
<th>Above River (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.4 to 10.4 m thick</td>
<td>‘Soil and creep’</td>
<td>30</td>
</tr>
<tr>
<td></td>
<td>‘Sandy-brown buff clay’</td>
<td>28.8</td>
</tr>
<tr>
<td></td>
<td>‘Buff to cream silt and fine sand. Ferruginous mottling and concretions. Numerous flake artefacts near the top’</td>
<td>26.7</td>
</tr>
<tr>
<td></td>
<td>‘Grey to white sandy clay with bands of small gravel seen’</td>
<td>25.9</td>
</tr>
<tr>
<td></td>
<td>‘Slumped and not seen’</td>
<td>24.4</td>
</tr>
<tr>
<td></td>
<td>‘Grey-brown silts’</td>
<td>22.7</td>
</tr>
</tbody>
</table>

Pale grey to buff clays mottled loams. Microlith and Levallois-type stone tools and silts with some grit horizons. Weathers to ‘Soil and creep’ to ‘Sandy-brown buff clay’ to ‘Grey to white sandy clay with bands of small gravel seen’.

<table>
<thead>
<tr>
<th>II. Sand Member</th>
<th>‘Grey to white sticky clay in 1” and 2” beds plus pockets of ferruginous sand’</th>
<th>21.8</th>
</tr>
</thead>
<tbody>
<tr>
<td>10.7 to 15.2 m thick</td>
<td>‘Ferruginous coarse sands.’</td>
<td>21.0</td>
</tr>
<tr>
<td></td>
<td>‘Complex pebble and artefact MN horizon. Small boulder bed and thicker stiff sandy clay with thin grit bands’</td>
<td>20.3</td>
</tr>
<tr>
<td></td>
<td>‘Clay with some sandy bands’</td>
<td>19.4</td>
</tr>
<tr>
<td></td>
<td>‘Grit bands in clay’</td>
<td>18.3</td>
</tr>
<tr>
<td></td>
<td>‘Sand false bedding to S’</td>
<td>17.8</td>
</tr>
<tr>
<td></td>
<td>‘Coarse grits in clay’</td>
<td>16.9</td>
</tr>
<tr>
<td></td>
<td>‘Coarse sand with some grit bands’</td>
<td>15.2</td>
</tr>
<tr>
<td></td>
<td>‘Coarse grit with clay’</td>
<td>14.2</td>
</tr>
<tr>
<td></td>
<td>‘Clay with intermediate bed of sandy clay’</td>
<td>13.6</td>
</tr>
<tr>
<td></td>
<td>‘Fine argillaceous sand’</td>
<td>12.8</td>
</tr>
<tr>
<td></td>
<td>‘Grit with some pebbles’</td>
<td>11.7</td>
</tr>
<tr>
<td></td>
<td>‘Clay with thin grit band’</td>
<td>10.4</td>
</tr>
<tr>
<td></td>
<td>‘Grit sand and clay inter- and cross bedded’</td>
<td>9.7</td>
</tr>
<tr>
<td></td>
<td>‘Grit and gravel with some large boulders and clays’</td>
<td>9.0</td>
</tr>
</tbody>
</table>

Sandy beds with often clean, ferruginous and showing false and slump bedding within the generally parallel major beds which are interbedded with more clayey and silty horizons. The rubble-artefact ‘MN’ occurs between 18.3 to 20.4 m above the river.

<table>
<thead>
<tr>
<th>I. Boulder Gravel Member</th>
<th>‘Boulder bed with individuals up to 8” ’</th>
<th>7.8</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.9 to 3.0 m thick</td>
<td>‘Boulder bed with individuals up to 3” and many fragments of phyllite (grows coarse with depth)’</td>
<td>7.6</td>
</tr>
</tbody>
</table>

Basal boulder bed grading from II.

| Karagwe-Ankolean System | ‘Phyllite bed rock. ’                                           | 4.8  |

Table 4-4: Classifications and descriptions of the terrace deposits at Nsongezi (after Wayland, 1937, Bishop, 1969)
Table 4-5 shows that Wayland describes individual clasts in the Boulder Gravel member up to between 3 and 8 inches which is equivalent to 76 to 203 mm and would today be classified as cobbles rather than boulders (>256 mm) on the Wentworth scale. Bishop (1969) claims that the bottom 3 m likely represents river cobbles and gravel dating from when the Kagera flowed westward. However, no indicators of current direction have been recorded at Nsongezi and this conclusion appears to be based only on the assumption that that following river reversal the stream power would be too low to transport course grained material. No sphericity, angularity, surface textures or provenance are recorded and therefore it is difficult to assess the distance over which this bed load has been transported. However, he does mention a section logged by Solomon (1939, cited in Bishop and Trendall, 1967) which was exposed during construction of a hydroelectric plant at Kikagati which contained false bedding the direction of which indicated a westerly flowing river. If correct, this appears to be the only sedimentological evidence of a westerly current direction prior to river reversal in Uganda and has apparently been overlooked. Fossil elephant molars from Loxodonta africana Arambourg were also found in the deposits at Kikagati and these have been ascribed to the early Pleistocene.

Bishop (1969) interprets the Sand Member to have been deposited by an extensive lacustrine phase in the Kagera Valley following river reversal. The MN artefact horizon recorded by Wayland (1934b) and Lowe (1952) is contained in the upper third of the Sand Member and represents a period of relatively low lake level when the surface sediments were emerged from the water. Bishop (1969) describes the artefact assemblage as ‘evolved Acheulian’ with ‘marked Sangoan influence’ in its upper part. He suggests they are equivalent to similar assemblages at Kalambo Falls in Tanzania (Clark, 1962, cited in Bishop and Trendall, 1967) where radiocarbon dating has dated evolved Acheulian assemblages as 57,000 yrs BP and Sangoan assemblages as 40,000 to 43,000 yrs BP. Bishop (1969) therefore suggests a date of 50,000 yrs BP for the top of the MN horizon, and 30,000 to 25,000 yrs BP for artefacts contained in the overlying Clay/Silt Member which he interprets to mark the end of open-water conditions.
Figure 4-41: Generalised section across the northern margin of the Kagera Valley at Nsongezi (after Bishop, 1969)

Figure 4-41 shows a cross section of the geometry of these sediments on the north side of the Kagera Valley at Nsongezi. About 30 m of sediment are shown overlying a bedrock bench in the Karagwe-Ankolean phyllite. The base of the Boulder Gravel Member is about 5 m above the River Kagera and 90 m above Lake Victoria. The upper surface of the terrace known as the Kafunzo Flat is about 30 m above the river and 120 m above Lake Victoria. Base level reduction associated with uplift has allowed the modern river to cut to below the Nsongezi Series deposits. However, the presence of a lower level terrace beneath the Kishen Flat about 8 m above the river indicates a period of aggradation during the overall long-term reduction in base level. The terrace contains artefacts derived from the MN horizon in the Sand Member above. It is comprised of sand and gravel thus indicating that coarse-grained material has previously been transported by the easterly flowing river. None of the researchers at Nsongezi mention groundwater discharge in the excavations, even from the likely transmissive Boulder Gravel Bed. Given the geometry shown in Figure 4-41 and the potentially free draining coarse sediments at the base of the terrace, it therefore seems likely that Nsongezi deposits all occur above the water table.
Figure 4-42: Levelled profiles of the Kagera Valley (after Bishop, 1969)

Figure 4-42 shows the section at Nsongezi in comparison to sections located 2 km to the north in the Orichinga Valley, 18 km downstream at Nkurungu, and 40 km downstream at Nyabusora as constructed by Bishop (1969). It can be seen that the upper surface of the lacustrine series descends from around 120 m (400 ft) above Lake Victoria at Nsongezi, to about 90 m (300 ft) at Nkurungu, to about 60 m (200 ft) at Nyabusora. The River Kagera is consistently about 30 to 35 m below the terrace at each location. As mentioned above, a gravel terrace occurs at a lower elevation to the lacustrine series beside the River Kagera at Nsongezi. However, at Nkurungu a gravel terrace is also shown above the lacustrine series (Lowe, 1952). Bishop (1969) describes this as ‘a continuous spread which mantles the hill-top to its maximum height of 260 ft AR [79 m above the river]. Gravel ranges upward from 200 ft [61 m] above the river but some slope-creep may have occurred.’ He indicates that their relationship with the Nsongezi deposits is unknown. Bishop (1969) also suggests that they may relate to former lacustrine conditions and the 200 ft (61 m) strandline of Lake Victoria.
Fluvio-lacustrine sediments similar to those at Nsongezi extend 110 km eastward to where the River Kagera flows into Lake Victoria. Spurr (1939, included in Temple and Doornkamp, 1970) logged these sediments (Figure 4-40) between 30 and 40 km southwest of Nsongezi at Nyabusora and Nyakanyasi. He recorded ‘boulders and cobbles’ near the base of several sections. Bishop (1969) points out that these sections, located close to the decrease in topographic gradient, include a greater proportion of deep water sediments including silts, silty clays and some diamictites near the top of the logged sequences. The lower Kagera to the east of the change in gradient which Bishop names the ‘hinge line’ (Figure 4-39), appears to have been extensively flooded during high stands of Lake Victoria. Whilst, Spurr attempts to correlate a low stand in the form of an unconformity between sections, in general it is difficult to correlate individual horizons. The interbedded sands, silts and clays shown in Spurr’s sections indicates locally variable spatial and temporal patterns of deposition, which suggests autochthonous sedimentary processes as well as external climatic and tectonic influences at work. Google Earth images show channels descending from the quartzite escarpments on either side of this section of the Kagera Valley and it seems likely that axial as well as longitudinal deposition has occurred. It appears that following uplift, the River Kagera has cut into the logged fluvio-lacustrine deposits which now lie above the water table to the west of the Bishop’s ‘hinge line’. During the maximum extent of Lake Victoria, deeper water lay to the east. However, near-shore coarse-grained deposits were likely deposited during lake transgression and regression. Arenaceous fluvio-lacustrine deposits with groundwater resource potential likely occur in the lower section of the Katonga Valley.

In his overall interpretation, Bishop (1969) proposes that the River Kagera once flowed westward across what is today a drainage divide in the Rutuha Swamp at a time when the Boulder Gravel Member was deposited. Subsequent uplift in western Uganda during the mid Pleistocene led to the formation of Lake Victoria, which at its maximum extended westward at least as far as Nsongezi and led to the deposition of the lacustrine Sand Member, and deep water deposits further east in the Kagera Valley. The western extension of Lake Victoria continued until the late Pleistocene with some oscillations indicated by the 50,000 year old MN artefact horizon. Bishop (1969) states that ‘renewed uptilting’ turned the finger lake into swamp which led to the deposition of the Clay/silt Member. Continued uplift led to progressive incision by the River Kagera.
He suggests that the upper surface of the Kafunzo Flat, which is 120 m above Lake Victoria at Nsongezi, declines towards the east until at about 30 m above Lake Victoria it becomes part of the higher strandlines along the western shoreline.

Whilst the overall drainage pattern and the presence of the swamp on the drainage divide strongly suggest that the drainage system once flowed west, the evidence that the basal Boulder Gravel Member was deposited at this time appears inconclusive. The attempted age assignment to deposits based on correlation of stone tools with other sites also contains significant uncertainty. It is clear that the detailed variability in the unconsolidated deposits downstream of Nzongezi and the existence of gravel terraces both above and below the Nsongezi deposits suggests a more complex history of base level changes than contained in the general conceptual framework provided by Bishop.

### 4.5.3 The Kafu and Muzizi Valleys

The southwest flowing River Nkusi and the northeast flowing River Kafu form a continuous valley system between Lake Kyoga and Lake Albert (Figure 4-43). The arrow-barb pattern of Lake Kyoga suggests this entire drainage system once flowed westward prior to uplift of the eastern shoulder of the Albertine Rift. The 140 km long River Kafu flows sluggishly towards Lake Kyoga through a broad valley of low relief containing papyrus wetland up to about 1 km wide. In 1924 Wayland identified potential evidence of gold in the Kafu Valley alluvium at the bridge on the Kampala to Hoima road. Hirst (1926) conducted extensive investigations of the alluvium using a Isler hand-boring rig mounted on a timber platform over the papyrus wetland, and trial pits in the adjacent terraces. Sixty seven boreholes and ninety pits were sunk along six lines of section at the locations shown in Figure 4-43.
As can be seen on the sections shown in Figure 4-44, the sediments are relatively shallow and rarely exceed 2 to 3 m thick. The relief on the underlying weathered schists and gneisses of the valley sides is between about 15 and 60 m. Hirst’s (1926) lithological descriptions of the unconsolidated deposits can be divided into three classes as follows:

**Figure 4-43: Location of the Kafu-Nkusi Valley and former extent of Lake Kyoga (after Bishop, 1969)**
1. Terrace Gravels
   a. ‘Gravels of 175 ft terrace’
   b. ‘Gravels of 50 ft terrace’
   c. ‘Gravels of Kafu Flat’

2. Sand
   a. ‘Coarse sand with 0.5 to 1 inch pebbles’
   b. ‘Medium-grained sand’
   c. ‘Fine white to light grey sand’

3. Clay
   a. ‘Sandy yellow clay overlying gravels of Kafu Flat’
   b. ‘Light grey clay often somewhat sandy’

The terrace gravels occur beneath parts of the floodplain and on the valley sides (Figure 4-44). Sandy deposits underlie the central valley floor and clayey sediment overlies the sand and parts of the Kafu Flat gravels. The general geometry suggests initial valley erosion, followed by deposition of extensive gravels in a high energy fluvial environment, before being eroded from the central part of the valley to create terraces on the valley sides. Later deposition of sandy deposits occurred in the central valley bottom before the wetland environment was established and accompanied by deposition of clayey sediment. It is noted that these early descriptions do not include ‘silt’ which presumable is contained in the sand and clay deposits. Neither do they include descriptions of angularity, sphericity and surface texture for the gravels which might provide evidence of the environments of deposition and distance of transport.

<table>
<thead>
<tr>
<th>Hirst (1926)</th>
<th>Bishop (1969)</th>
</tr>
</thead>
<tbody>
<tr>
<td>245 ft peneplain (75 m above river)</td>
<td>Ferruginous conglomerates</td>
</tr>
<tr>
<td>225 ft terrace gravels (69 m above river)</td>
<td>Not identified</td>
</tr>
<tr>
<td>175 ft terrace gravels (53 m above river)</td>
<td>Kinogo Terrace</td>
</tr>
<tr>
<td>50 ft terrace gravels (15 m above river)</td>
<td></td>
</tr>
<tr>
<td>Kafu Flat gravels (&lt; 6 m above river)</td>
<td>Flood plain and alluvial channel</td>
</tr>
<tr>
<td>Sub-flat gravels</td>
<td></td>
</tr>
</tbody>
</table>

**Table 4-5: Summary of Kafu Valley terraces identified by Hirst (1926) and Bishop (1969)**
Figure 4-44: Simplified lithological sections of the Kafu Valley alluvial deposits (after Hirst, 1927)

Table 4-6 summarises the sequence of terraces identified by Hirst (1926) and Bishop (1969). Hirst suggested that the terraces decreased in height and thinned to the east with an accompanying coarsening grain size which he used as evidence that they were deposited when the river flowed westward. However, Bishop resurveyed the terraces and did not observe a change in elevation or grain size. Hirst (1926) originally proposed a complex sequence of tilting in opposite directions to create the terraces, while in the same volume, Wayland proposed an explanation based on climatic pluvial and interpluvial periods which he assumed would correlate with European glaciations. Bishop (1969) suggests that the arrow-barb pattern of Lake Kyoga indicates a single comparatively recent reversal. He proposes that Lake Kyoga formed following initial rift shoulder uplift but the lake continued to overflow via the Nkusi Valley. The high level gravels were deposited by the westward flowing drainage and uplift caused down-
cutting by the River Kagera which created the terraces. He suggests that the sand was deposited following river reversal when Lake Kyoga was at its maximum extent (Figure 4-43). The lake level began to fall after it began to drain via the lower Victoria Nile and the papyrus wetland developed in the Kafu Valley and clayey sediment was then deposited.

The River Muzizi is located between the Kafu-Nkusil Valley System and the Mpanga-Katonga Valley System and flows westward into the southeast corner of Lake Albert. In 1919 Wayland identified thick gravels overlying the Muzizi Sandstone where the Fort Portal to Hoima road crossed the River Muzizi (Wayland, 1925). He noted that the width of the gravels led him ‘to believe that at one time the Muzizi was a much greater river than it is at present’, and speculates that it may have formed a branch of Lake Albert prior to of development of the present relief on the rift flank. He also proposes that the Muzizi gravels are contemporary to gravel found at 300 ft above Lake Victoria based on stone stools, although the strength of evidence seems unclear. Whilst Wayland’s interpretations are speculative his notes provide a useful observation of another fluvial gravel deposit in the region. Unfortunately, like many early researchers in Uganda he does not provide a detailed lithological description.

4.5.4 The Katonga Valley

The sedimentary fill in the Katonga Valley is poorly exposed and there have been no systematic archaeological or prospecting investigations like those in the Kagera and Kafu valleys. Evidence for the character of sedimentary fill in the Katonga Valley comes from a variety of sources including Uganda Geological Survey reports and mapping, railway construction investigations, water supply boreholes and personal observations. Figure 4-45 shows the locations of evidence for the character of the sedimentary deposits in and around the Katonga Valley.
Figure 4-45: Locations of unconsolidated sedimentary deposits recorded in the Katonga Valley and its tributaries
<table>
<thead>
<tr>
<th>TYPE</th>
<th>LOCATION</th>
<th>SUMMARY DESCRIPTION</th>
<th>REFERENCE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Interfluve &amp; Terraced Sand</td>
<td>Musozi Station, Nabakasi Valley</td>
<td>Angular quartz grits, gravels and water worn pebbles with distinct bedding. Cemented by a red earthy material or kaolin. Purple to grey gritty mudstones. Over 6 m thick. Between about 60 and 115 m above Lake Victoria (LV)</td>
<td>(Johnson, 1958)</td>
</tr>
<tr>
<td>Sand &amp; Gravel</td>
<td>Nabakasi /Katonga Interfluve</td>
<td>Variable gravels and water worn pebbles, overlain by gritty mudstones and angular quartz gravels with red earthy cement. Up to 14 m thick. ‘Pleistocene to Holocene’ ‘Sands, clays, grits and gravels’ Approximately 60 m to 120 m above Lake Victoria.</td>
<td>(Johnson, 1958)</td>
</tr>
<tr>
<td></td>
<td>Nabajusi Tributary</td>
<td>Similar deposits to Nabakasi Valley near River Katonga on Kampala/Masaka road and in a tributary of the Nabajusi between Kalungu and Villa-Maria.</td>
<td>(Johnson, 1958)</td>
</tr>
<tr>
<td></td>
<td>Kyai Crossing, Central Katonga</td>
<td>Outcrop log: light grey, weakly to moderately cemented, sub-angular occasionally sub-rounded, coarse sand to medium pebbles with layers of coarse pebbles and cobbles. Some beds dipping 8-17 ° to S. 57 to 77 m above LV</td>
<td>Field notes</td>
</tr>
<tr>
<td>Wetland Deposits</td>
<td>Kabagole, Western Katonga</td>
<td>Outcrop log: moist, grey-brown mottle orange, firm, silty Clay with mica and occasional fine to coarse quartz sand. Over 3 m thick at about 47 m above LV Auger logs: generally ‘clay’, ‘clayey sand’ and ‘fine to medium sand’ Over 3 m thick at about 47 m above Lake Victoria</td>
<td>Field notes (Waldron, 1952)</td>
</tr>
<tr>
<td></td>
<td>Kisozi, Eastern Katonga</td>
<td>Borehole log: ‘Clay loam’ and ‘clay soil’, probably clayey Silt, and silty Clay Approximately 3 to 6 m thick. Top at about 36 m above Lake Victoria</td>
<td>(China-Zhonghao, 2006)</td>
</tr>
<tr>
<td>Raised Beaches</td>
<td>North side, Eastern Katonga</td>
<td>Poorly developed strandline between 58 and 76 m above Lake Victoria marked by coarse quartz grits and gravels.</td>
<td>(Johnson, 1960)</td>
</tr>
</tbody>
</table>

Table 4-6: Summary of the available sediment descriptions in the Katonga Valley and surrounding area
On the basis of these observations, the sediments in the Katonga Valley can be divided into four broad categories: 1) interfluve and terrace grits and gravels; 2) wetland deposits; 3) fluviolacustrine deposits; and 4) raised beaches. The lithological descriptions of each of these sediment types are summarized in Table 4-7.

**Interfluve and terrace sand and gravel**

The interfluve and terrace sand and gravel is perhaps the most enigmatic deposit of the Katonga Valley and surrounding area. The three main sources of information about these deposits are: 1) the 1:250,000 geological map Sheet N.A. 36-14 (Kampala); 2) an unpublished Department of Geological Survey and Mines field report on the Nabakasi Valley prepared by Johnson (1958); and 3) the author's own observations of the terrace deposits at Kyai in the Katonga Valley.

The geological map (D.G.S.M., 1962) categorises the deposits shown on the interfluves around the confluence of the Katonga, Nabakasi and Katabalang rivers (Figure 4-45) as ‘*Pleistocene to Holocene*’ ‘sands, clays, grits and gravels’. Unfortunately, the author was unable to locate the interfluve deposits in the short time available when travelling through the area. They appear to be related to the deposits identified by Johnson (1958) during investigations along the route of the Kampala-Kasese railway. The two main areas where these sediments are found are in the railway sections between the River Nabakasi and River Katonga and in the vicinity of Musozi Station, further north in the Nabakasi Valley (Figure 4-45). Johnson (1958) describes the quartz-grits and gravels as widespread around Musozi Station. He identifies three different quartz grits and gravels which despite their superficial resemblance appear to have three different modes of origin as follows:

**In situ weathering** – distinguished by no sorting, remnants of quartz veins, books of mica, overlain by stone line, contains unrounded tourmaline crystals and bands of country rock.

**Colluvium** – Similar to above but subject to soil movement which can produce bedded appearance. Generally angular pebbles may have subangular facets.

**Alluvium** – Distinct bedding, angular quartz-grits and water worn pebbles, cemented by red earthy material. Frequent rounded tourmaline crystals. Mauve, pink, purple or
grey gritty mudstone occurs in many areas and quartz-grit with black haematite matrix is locally present.

The alluvium in the Nabakasi Valley is largely unexposed and so its extent is unknown. Near Musozi Station it appears to be at least 3 to 6 m thick and occurs between about 60 and 115 m above Lake Victoria. A sketch by Johnson (1958) shows it present near the floor of the Nabakasi Valley, but missing from the lower third of the slope, before reappearing on a bench and outcropping towards the top of the slope. Given that the deposits on the spurs occur up to about 120 m above Lake Victoria (approximately 1250 m asl) and are absent from the valley bottom, they appear older than recent valley erosion (Johnson, 1958). The base of the deposits on the interfluves around the confluence of the three rivers is at least 15 m above the valley bottoms. Johnson (1958) also suggests that similar deposits occur further east at the head of a southern tributary to the Katonga Valley, known as the Nabajusi River (Figure 4-45). Johnson (1958) speculates that these sands and gravels may have been deposited when Lake Victoria was at a higher level, and may be related to the 60 m strandline above Lake Victoria. However, the terrace and interfluve deposits occur up to 120 m above Lake Victoria. Their maximum elevation is up to 50 m higher than the current drainage divide of the Katonga Valley and higher than drainage divides to the north. Therefore, it seems unlikely that the shoreline of Lake Victoria could ever have been high enough to account for these deposits. On the basis of their general characters, Johnson (1958) speculates that they may be late Pliocene to early Pleistocene in age.

Photographs of the bedded sand and gravel in the terraces at Kyai are shown in Figure 4-46 and 4-47, and their location is indicated on Figure 4-45. The southern terrace is better exposed than the northern terrace on the west side of the murrum road crossing the Katonga Valley. The southern terrace is about 15 m thick with, with elevation at the top and bottom of approximately 1200 and 1185 m asl respectively. Phyllite outcrops between the base of the terrace and the valley bottom at about 1175 m asl, which is occupied by papyrus wetland.
Figure 4-46: Sand and gravel terraces in the central Katonga Valley at Kyai

Figure 4-47: Sand and gravel beds on southern terrace dipping between 8° and 17° towards the south east and south west into the valley side

As shown in Figure 4-47 and Figure 4-48a, the sediment in the terraces is predominantly compact, weakly to moderately cemented, bedded, light grey, poorly to moderately sorted, angular to sub-angular sand and gravel, with sub-angular to sub-rounded cobbles in some beds. Weathering near the top of the terrace has produced a moderately strong cement of clay and iron oxides, as shown in Figure 4-46 and Figure
4-48b. It is difficult to break by hand, but breaks easily with light hammer blows. The sand and gravel is over 90% quartz, with some igneous lithic clast, and rare but distinctive rounded blueish black pebbles.

Figure 4-48: Angular to sub-angular sand and gravel including a) lower unweathered deposits and b) upper weathered deposits

Figure 4-49: a) angular to sub-angular sand and gravel with sub-rounded cobbles and b) colluvium above, sand, gravel and cobbles, resting on phyllite

Some beds exposed in the lower part of the southern terrace contain considerable quantities of sub-angular to sub-rounded cobbles, as shown in Figure 4-49a. Whilst the cobbles are water-worn, they are often matrix supported, and there is no evidence of
imbrication as might be expected if deposited from a consistently strong current. Bedding is on the scale of 10s of centimetres and some horizons show possible upward gradation from cobbles and gravel, to sand and gravel, or sand. In the upper part of the southern terrace, the beds are sub-horizontal (Figure 4-46). However, gullying of the lower part of the terrace has exposed the dipping beds shown in Figure 4-47. It was hoped that the dip on the beds would provide an indication of the flow direction in the River Katonga at the time of deposition. However, they were found to dip between 8° and 17° towards the south east and south west into the valley side, rather than parallel to the thalweg. The sand, gravel and cobbles directly overlie the in situ highly weathered phyllite of the Buganda-Toro System at the base of the terrace as shown in Figure 4-49b, and the irregular surface of the phyllite dips towards the valley bottom.

The generally coarse grain size indicates that the sediment at Kyai was deposited in a high energy environment. However, the angularity of the clasts suggests that they were not transported far. The poor to moderate sorting and lack of imbrication provides evidence that the strong current was not sustained over long periods. Although dipping beds were observed the multiple foresets characteristic of fluvial cross-bedding was not, although this could be due to poor exposure. The cobbles appear to be restricted to the lower part of the southern terrace and beds in the upper part of the terrace are sub-horizontal. It is concluded that the deposits at Kyai are unlikely to have been deposited by a permanent, consistent, and powerful river. Whilst the topographic setting is not one generally associated with alluvial fans, the facies are characteristic of deposition by low frequency, high magnitude, events as might be expected in an arid or semi-arid environment. The dips that were observed in the southern terrace are towards the south and into the valley side, which is inconsistent with a flow direction parallel to the Katonga Valley. The coarse grained sediment may therefore have been transported perpendicular to the valley sides from the north, and it may be hypothesised that they were once contiguous with the interfluve deposits shown on the geological map (Figure 4-45). Whilst the evidence for the direction of sediment transport remains uncertain, we can have confidence that the deposits once filled Katonga Valley at Kyai. The exposed slopes on the weathered phyllite facilitated the transport of cobbles to the base of the original valley, and gradual filling produced the observed dips on the sand and gravel beds. However, as the valley filled and the slope angle reduced, the beds
became sub-horizontal and the available energy decreased so that cobbles were no longer transported and deposited.

Sometime after deposition, base level changes appear to have renewed erosion and the modern Katonga Valley cut through the sand and gravel to create the terraces that we see today. The presence of the angular to sub-angular coarse grained deposits and their subsequent erosion to create the river terraces suggest that the landscape evolution of the Katonga Valley is likely more complex than the historical conceptual framework provided by previous researchers (e.g., Doornkamp and Temple, 1966, Bishop and Trendall, 1967).

**Wetland deposits**

The wetland (paludal) sediments are the most recently deposited. Descriptions are available from near Kabagole in the western Katonga Valley and Kisozi in the eastern Katonga Valley. The author’s observations of an excavation for a cattle watering hole adjacent to the south side of the papyrus wetland about 2.3 km east of the Katonga Valley crossing at Kabagole, revealed the near-surface sediment to be composed of moist, grey-brown mottle orange, firm, silty clay with mica and occasional fine to coarse quartz sand (Figure 4-50).

The only record found of an investigation focused on the near-surface deposits on the floor of the Katonga Valley is contained in an unpublished field report of the Ugandan Geological Survey (Waldron, 1952). Eight hand driven augers were used to investigate sediments up to depths between 1.5 and 2 m bgl during the search for suitable construction materials for the Kampala-Kilembe railway. The investigation was conducted about 1.6 km east of the Kabagole crossing on the south side of the papyrus wetland (Figure 4-45). The auger logs from this investigation are shown in Table 4-8 and Table 4-9 (Waldron, 1952). They reveal a general increase in grain size from the ground surface to between 2 and 3 m bgl from clay, to clayey sand, to fine to medium-grained sand.
Figure 4-50: Silty clay wetland deposits near Kabagole Station

Waldron (1952) does not consider the environment of deposition. The upper black clay and silty clay are similar to that which might be expected to be deposited in a papyrus wetland similar to the conditions that prevail today. However, the slightly coarser clayey sand and fine to medium-grained sand might suggest periods of historically less vegetation and an associated reduction in the impedance to surface water flow.

<table>
<thead>
<tr>
<th>Auger Hole 1</th>
<th>Auger Hole 2</th>
<th>Auger Hole 4</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Base (m bgl)</strong></td>
<td><strong>Lithology</strong></td>
<td><strong>Base (m bgl)</strong></td>
</tr>
<tr>
<td>2.03</td>
<td>‘Black and grey clay and silty clay’</td>
<td>0.46</td>
</tr>
<tr>
<td>&gt;3.05</td>
<td>‘Light grey and buff very slightly clayey sand’</td>
<td>1.63</td>
</tr>
<tr>
<td>2.24</td>
<td>‘Buff fine to medium-grained very slightly clayey sand’</td>
<td>1.83</td>
</tr>
</tbody>
</table>

Table 4-7: Auger logs 1, 2 and 3, 300 m south of papyrus wetland, 1.6 km east of Kabagole ERT survey (after Waldron, 1952)
<table>
<thead>
<tr>
<th>Auger Hole 5</th>
<th>Auger Hole 6</th>
<th>Auger Hole 7</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Base (m bgl)</strong></td>
<td><strong>Lithology</strong></td>
<td><strong>Base (m bgl)</strong></td>
</tr>
<tr>
<td>0.41</td>
<td>‘Black clay soil’</td>
<td>0.61</td>
</tr>
<tr>
<td>0.76</td>
<td>‘Grey clay’</td>
<td>1.42</td>
</tr>
<tr>
<td>1.52</td>
<td>‘Light grey and buff clayey sand and sandy clay’</td>
<td>1.98</td>
</tr>
<tr>
<td>1.98</td>
<td>‘Light grey very slightly clayey sand’</td>
<td>&gt;2.24</td>
</tr>
<tr>
<td>&gt;2.24</td>
<td>‘Buff sand’</td>
<td></td>
</tr>
</tbody>
</table>

Table 4-8: Auger logs 5, 6 and 7, 300 m south of papyrus wetland, 1.6 km east of Kabagole ERT survey (after Waldron, 1952)

In the eastern Katonga Valley, two water supply boreholes (Kikuumaddungu Boreholes 1 and 2) were identified adjacent to the south side of the papyrus wetland either side of the Kisozi crossing road. Borehole 2 is located approximately 100 m east of the road and Borehole 3 is located approximately 350 west of the road. The driller’s logs are shown in Table 4-10. The area where the boreholes are located currently sits about 10 m higher than the present papyrus wetland. Nevertheless, the upper 6 m of deposits consist of clay and loam soils which likely represent former wetland sediments deposited when the relative water level was higher.

<table>
<thead>
<tr>
<th>Kikuumaddungu Borehole 2</th>
<th>Kikuumaddungu Borehole 3</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Depth (m bgl)</strong></td>
<td><strong>Notes from driller’s log</strong></td>
</tr>
<tr>
<td>0 to 6</td>
<td>‘Clay loam’</td>
</tr>
<tr>
<td>6 to 12</td>
<td>‘Loam soil’</td>
</tr>
<tr>
<td>12 to 16</td>
<td>‘Sandy clay soil’</td>
</tr>
<tr>
<td>16 to 26</td>
<td>‘Quartz mixed with sand’</td>
</tr>
<tr>
<td>26 to 32</td>
<td>‘Reddish sand’</td>
</tr>
<tr>
<td>32 to 54</td>
<td>‘Quartz mixed with sand’</td>
</tr>
<tr>
<td>54 to 60</td>
<td>‘Weathered rock’</td>
</tr>
<tr>
<td>60 to 96</td>
<td>‘Competent rock’</td>
</tr>
</tbody>
</table>
Table 4-9: Summary of driller’s logs from Kikuumaddungu water supply boreholes at Kisozi crossing

**Fluvialacustrine deposits**

There are numerous water supply boreholes in the lacustrine deposits close to Lake Victoria, however, the two boreholes at Kisozi crossing (Table 4-10) provide the only logs of the western deposits within the Katonga Valley itself. The driller’s logs require some interpretation. ‘Quartz mixed with sand’ likely refers to quartz gravel and sand. Although silt is not mentioned, it is likely present in the clay and loam deposit and as will be shown using pumping test interpretation in Section 6.0, is also likely present in the sand too. The combination of clay and sand with some gravel suggests possible lacustrine or fluvial overbank environments of deposition with occasional higher energy currents. The surface soil characteristics and topography suggest that original flood plain extends to about 2 km south of the papyrus wetland. The boreholes indicate that the alluvium is about 55 m deep adjacent to the papyrus wetland.

**Beach deposits**

Johnson (1955a, 1960) recognized the strandlines on the south eastern slopes of spurs on the north side of the wide eastern Katonga Valley (Figure 4-45) at about 35 (Kayugi strandline) and 60 m (Kilwala strandline) above Lake Victoria. He points out that the higher strandline is not always well developed, but is marked by raised beach deposits, including coarse quartz grits and gravels. Sand bars of yellow-red sands extend across the mouths of the northern tributaries parallel the main Katonga Valley (Figure 4-45). They are contiguous with the Kayugi (35 m) strandline (Johnson, 1960) and their direction of accumulation is appears to be attributable to the prevalent wind and wave direction in a lacustrine environment (Bishop, 1959).

**4.5.5 Lake Victoria**

Information about the lacustrine sediments deposited in Lake Victoria is available from onshore water supply boreholes logs, offshore seismic surveys and some shallow offshore sediment core.
4.5.5.1 Borehole records

The fluviolacustrine deposits identified in the boreholes at Kisozi are contiguous with those that occur further east near the mouth of the Katonga Valley and around the northwest shore of Lake Victoria as shown in Figure 4-51. The descriptions from driller’s logs for boreholes in the lacustrine sediments at the locations shown in Figure 4-51 are presented in Tables 4-11 and 4-12. Table 4-11 includes those boreholes with groundwater abstraction permits for downhole electronic pumps, and Table 4-12 includes boreholes with hand pumps which are listed in the DWRM database. Unfortunately, the driller’s descriptions are not systematic. Silt is never mentioned but is presumably present, and the relative proportions of different grain size constituents are unclear. Since the presence of only about 10% fines can reduce the transmissivity of sand and gravel to that of clays and silts, the lack of systematic descriptions considerable reduces confidence in the utility of these logs for hydrogeological interpretation. Coordinates are recorded for the boreholes at Bukakata and Kagera (Table 4-11), but only the predicted lithology is available.
Figure 4-51: Geological map of the northwest Lake Victoria shoreline showing the distribution of lacustrine deposits and borehole locations

The borehole at Serinya (Table 4-12) located on the south side of the Katonga Valley, 15 km east of the Kisozi boreholes (Table 4-10). The log shows what is likely fluviolacustrine sediments between 7 and 43 m bgl which are mainly comprised of sand, with some layers also containing clay or gravel. Anomalously, ‘laterite’ is recorded between 0 and 7 m bgl, and there is no log available below 43 m bgl.
<table>
<thead>
<tr>
<th>Depth (m bgl)</th>
<th>Description</th>
<th>Depth (m bgl)</th>
<th>Description</th>
<th>Depth (m bgl)</th>
<th>Description</th>
<th>Depth (m bgl)</th>
<th>Description</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 to 2</td>
<td>‘top soil’</td>
<td>0 to 3</td>
<td>‘top soil’</td>
<td>0 to 3</td>
<td>‘top fine sand brown’</td>
<td>0 to 3</td>
<td>‘black sandy soil’</td>
<td>TD not recorded. Bedrock at 80. No log. Predicted: ‘Lake deposits consisting of clay lenses, silt, sand, gravels and probably boulders at the base’</td>
</tr>
<tr>
<td>2 to 4</td>
<td>‘khaki clay with water’</td>
<td>3 to 6</td>
<td>‘yellowish khaki clay’</td>
<td>3 to 6</td>
<td>‘clay black’</td>
<td>3 to 6</td>
<td>‘brown clay’</td>
<td>TD = 90.8 Sediment depth not recorded.</td>
</tr>
<tr>
<td>3 to 12</td>
<td>‘streaky clay/khaki clay’</td>
<td>6 to 18</td>
<td>‘dark grey shale with sticky clay’</td>
<td>6 to 9</td>
<td>‘fine sand’</td>
<td>6 to 21</td>
<td>‘brown clayey sand’</td>
<td>No log.</td>
</tr>
<tr>
<td>12 to 18</td>
<td>‘dark grey clay with H₂S’</td>
<td>18 to 21</td>
<td>‘light grey clay with few gravels’</td>
<td>9 to 15</td>
<td>‘brown clay’</td>
<td>21 to 24</td>
<td>‘brown clay’</td>
<td>No log. Predicted: ‘Lake deposits consisting of silt, fine sand and some black clay in minor proportions’</td>
</tr>
<tr>
<td>18 to 48</td>
<td>‘gravel and bedrock’</td>
<td>21 to 45</td>
<td>‘medium to coarse gravels’</td>
<td>15 to 93.6</td>
<td>‘coarse sandy clay - medium-grained’</td>
<td>24 to 36</td>
<td>‘coarse-grained sand’</td>
<td></td>
</tr>
<tr>
<td>48 to 61</td>
<td>‘gneissic rock’</td>
<td>45 to 61</td>
<td>‘gneissic bedrock’</td>
<td>93.6 to 118</td>
<td>‘sand medium to coarse’</td>
<td>36 to 62</td>
<td>‘sandy clay and quartz pebbles’</td>
<td></td>
</tr>
</tbody>
</table>

Table 4-10: Summary of driller’s logs from boreholes in lacustrine sediments with abstraction permits
<table>
<thead>
<tr>
<th>Serinya (WDD8306)</th>
<th>Kirinya, Lukaya (DWD16278)</th>
<th>Kirinya, Lukaya (DWD16279)</th>
<th>Mwota, Lukaya (DWD16280)</th>
<th>Nkozi (DWD11676)</th>
<th>Nkozi (DWD11676)</th>
<th>Nkozi (DWD11676)</th>
<th>Nkozi (DWD11676)</th>
<th>Nkozi (DWD11676)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 to 7 ‘laterite’</td>
<td>0 to 6 ‘top soil’</td>
<td>0 to 1 ‘loamy top soil’</td>
<td>0 to 2 ‘top soil’</td>
<td>0 to 9 ‘sandy soil’</td>
<td>0 to 0.5 ‘black top soil’</td>
<td>0 to 1 ‘black top soil’</td>
<td>0 to 0.5 ‘black top soil’</td>
<td>0 to 8 ‘sandy soil’</td>
</tr>
<tr>
<td>7 to 9 ‘sand and clay’</td>
<td>6 to 18 ‘clay’</td>
<td>1 to 9 ‘ferruginised clay soil’</td>
<td>2 to 18 ‘sandy clay’</td>
<td>9 to 12 ‘black clay’</td>
<td>0.5 to 3 ‘sandy soil’</td>
<td>1 to 10 ‘sand’</td>
<td>0.5 to 9 ‘sand’</td>
<td>8 to 12 ‘black clay’</td>
</tr>
<tr>
<td>9 to 13 ‘coarse sand and quartz’</td>
<td>18 to 34 ‘sandy clay’</td>
<td>9 to 21 ‘sandy clay’</td>
<td>18 to 27 ‘ferruginised’</td>
<td>12 to 21 ‘grey clay’</td>
<td>3 to 12 ‘black clay’</td>
<td>10 to 15 ‘black clay’</td>
<td>9 to 12 ‘black clay’</td>
<td>12 to 15 ‘grey clay’</td>
</tr>
<tr>
<td>13 to 43 ‘fine sand’</td>
<td>34 to 45.5 ‘quartz boulder’</td>
<td>21 to 36 ‘slightly weathered rock’</td>
<td>27 to 44 ‘completely weathered granite’</td>
<td>21 to 40 ‘clay and gravel’</td>
<td>12 to 24 ‘grey clay’</td>
<td>15 to 21 ‘grey clay’</td>
<td>12 to 14 ‘grey clay’</td>
<td>15 to 17 ‘clay and gravel’</td>
</tr>
<tr>
<td>43 to 114 no log</td>
<td>45.5 to 66 ‘sandy clay’</td>
<td>36 to 51 ‘moderately weathered rock’</td>
<td>44 to 66 ‘weathered granite’</td>
<td>40 to 45 ‘weathered granite’</td>
<td>24 to 52 ‘clay and gravels’</td>
<td>21 to 48 ‘clay and gravels’</td>
<td>14 to 15 ‘clay and gravels’</td>
<td>17 to 61 ‘weathered rock’</td>
</tr>
<tr>
<td>66 to 90.6 ‘sandy clay, very fresh granite’</td>
<td>51 to 117.8 ‘fresh hard rock’</td>
<td>66 to 90.8 ‘grey granite’</td>
<td></td>
<td>52 to 61 ‘weathered granite’</td>
<td>48 to 61 ‘weathered granite’</td>
<td>15 to 64 ‘highly weathered granite’</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 4-11: Summary of driller’s descriptions from boreholes in lacustrine sediments adjacent to eastern Katonga Valley
There are several borehole records available around Nkozi on the north side of the Katonga Valley. They appear to have been drilled through sediment in a tributary valley (Figure 4-51). It can be seen from Tables 4-11 and 4-12 that these sediments vary between 15 and 50 m thick. Above about 20 to 25 m bgl they appear to be dominated by clay with some sandy layers. All five boreholes listed in Table 4-12 shows ‘clay and gravels’ towards the base of the sediments, although it is unclear if the gravel is dispersed or forms layers in the clay and silt. The two Nkozi boreholes listed in Table 4-11 show between 20 and 30 m of gravel at the base of the sediment.

Several borehole records are also available from around Lukaya on the south side of the broad mouth of the Katonga Valley. The approximate depth of each of the dominant lithologies in four boreholes near Lukaya is summarised by Tindimugaya (2008) and reproduced in Table 4-13. This summary suggests that clay is dominant up to between 23 and 42 m bgl. An underlying layer dominated by sand is between 18 and 44 m thick. It is unclear if the ‘sand’ also contains hydrogeologically significant amounts of silt and/or clay.

<table>
<thead>
<tr>
<th>Dominant Lithology</th>
<th>CD1806</th>
<th>WDD5398</th>
<th>DWD15303</th>
<th>DWD14304</th>
</tr>
</thead>
<tbody>
<tr>
<td>Top soil/murrum</td>
<td>0 to 2</td>
<td>0 to 2</td>
<td>0 to 1</td>
<td>0 to 2</td>
</tr>
<tr>
<td>Clay</td>
<td>2 to 39</td>
<td>2 to 42</td>
<td>1 to 26</td>
<td>2 to 23</td>
</tr>
<tr>
<td>Sand</td>
<td>39 to 67</td>
<td>42 to 60</td>
<td>26 to 64</td>
<td>23 to 67</td>
</tr>
<tr>
<td>Weathered/fractured bedrock</td>
<td>67 to 115</td>
<td>60 to 95</td>
<td>n/a</td>
<td>n/a</td>
</tr>
</tbody>
</table>

**Table 4-12: Summary of borehole logs near Lukaya (from Tindimugaya, 2008)**

Table 4-12 includes boreholes from Kirinya and Mwota near Lukaya but there exact coordinates are not recorded. The DWRM database lists clays and sand sized particles in these logs, but it is not clear whether these boreholes penetrate sediment or weathered rock. Table 4-11 list boreholes in the villages of Kalangala and Bulingo located a few miles south of Lukaya. The logs indicate fine-grained clayey soils to between 15 and 24 m bgl, with underlying sand, sandy clay and gravel. The borehole log for DWD17021 indicates that sandy clay and gravel extends to 63 m bgl, whilst log for DWD17020 indicates that medium to coarse sand extends to 118 m bgl. This
second depth is particularly large and, if correct, might suggest the presence of sediment filled channel perhaps associated with the River Katonga. The logs do not record the bedrock surface and given the quality of many logs it is unclear if this is intentional or an omission. No logs could be found for the boreholes at Bukakata and Kagera, although the results of the pumping tests (Chapter 7.0) suggest that they penetrate coarse-grained transmissive sediments.

4.5.5.2 Seismic investigations

Lake Victoria straddles the equator, and appears to have formed in a tectonic sag basin, between the eastern and western arms of the EARS. As previously discussed, the drainage pattern and exposed lacustrine sediments to the west of the lake reveal its origin as a result of uplift along the rift margin and consequent river reversal. The late-Pleistocene to Holocene palaeoclimatic interpretation of data from shallow sediment cores were discussed in Section 4.3.3. This section focuses on the overall geometry and history of sedimentation in the Victoria Basin determined mainly from about 2,000 km of seismic reflection profiles which were conducted during 1995 as part of the International Decade for East African Lakes (IDEAL) program (Scholz et al., 1998). The location of all of the survey lines are shown in Figure 4.52a and the only published profiles, named V95-2P, V95-16A and V95-22, are shown in Figure 4.52b.

It is interesting to compare the results of the bathymetric survey produced by IDEAL shown in Figure 4.52a, with that shown in Figure 4.52b (Temple, 1966) based on British Admiralty charts of 1902 to 1906. The presence of a deep trough immediately north of V95-2P and V95-22A is disputed by Johnson et al (2000), who maintain that ‘The floor of Lake Victoria is as flat as a billiard table, reflecting the sedimentary infilling of an irregular underlying morphology’. The current mean depth of Lake Victoria is relatively shallow at about 40 m, with a maximum depth of 68 m (Johnson et al., 2000).
Figure 4.52: Seismic reflection survey of Lake Victoria a) track line locations and b) interpretation of surveys (after Johnson et al., 1996)
The seismic reflection surveys of Lake Victoria indicate a maximum thickness of lake sediments of about 60 m. Five separate units with erosional boundaries and named A to E were identified by the seismic reflection surveys. Unit A is 0 to 9 m thick and is highly stratified but with low reflectivity (Scholz et al., 1990, Johnson et al., 1996). Piston core samples reveal that it consists of dark, organic rich diatomaceous mud. It overlies a strong reflector at the top of Unit B which has been identified as a palaeosol and dated at 14.5 ka. Units B and C are also strongly stratified which leads Johnson et al (1996) to suggest that they are probably similar in composition to the overlying sediment, despite their higher reflectivity. Units B and C are up to about 23 m thick. Unit D is also stratified and has high reflectivity, but as can be seen in V95-16A (Figure 4.52c) it is up to 35m thick in the western part of the lake but absent in the eastern part of the lake. In contrast, the overlying units are thickest in the eastern parts of the lake. This suggests uplift to the west relative to the east. The underlying Unit E is quite different in character. The reflectors are sometimes discontinuous and sometimes wavy. Johnson et al (1996) suspect that this substantially older unit could correspond to pre-Lake Victoria Miocene lacustrine, fluvial and volcanioclastic rocks that outcrop in the northeast part of the lake. Profile V95-16A in Figure 4.52c shows that Unit E itself appear to be underlain by a deeper sequence dipping to the southwest, but this is not mentioned by earlier authors.

According to Johnson et al (2000) a crude calculation based on an estimated average sedimentation rate of 150mm/1000 years indicates the lake origin at about 400ka. This assumed sedimentation rate can be compared against an average rate for the Holocene of 250mm/1000 years and a rate for the most recent 1ka of 700mm/1000 yrs (Talbot and Lærdal, 2000). Johnson et al (2000) suggest that the three lows stands associated with erosional surfaces may be related to the post Pleistocene Revolution 100ka glacial-interglacial cycles. More than have of the Quaternary sedimentary sequence (Unit D), possibly representing over 200 k yrs of deposition, appears to have occurred prior to the tectonic movement which moved the depositional centre further east.
4.6 Summary

4.6.1 Tectonics

Today, the surface of the George Basin is about 930 m asl, the water divide on the Mpanga-Katonga Valley System is about 1,200 m asl, and the bed of Lake Victoria is about 1,070 m asl. Uplift of the EAP most probably began in the mid Miocene (about 17 Ma) (Wichura et al., 2010) due to the upwelling of a low-density (asthenosphere) mantle plume (Weeraratne et al., 2003). Calibrated geophysical models suggest plume heads occur beneath the rift valleys on either side of the EAP. Radiometric dating of associated volcanic rocks indicates that the EARS developed from north to south in Kenya between the early Oligocene (30 Ma) and mid Miocene (15 Ma) (Nyblade and Brazier, 2002). The Western Rift between northern Uganda and Tanzania began to develop in the later Miocene (10 Ma). The crustal extension rate perpendicular to the rift, west of the Katonga Valley is about 2.1 mm/yr, whilst the extension rate of the Kenya Rift at equivalent latitude is about 3.2 mm/yr (Stamps et al., 2008).

The central part of the EAP, occupied by Lake Victoria has undergone relative downwarping. Possible explanations include compression between the opposing active rifts, thermal sag associated with changes in the deep plume, or relative uplift of the margins associated with the shallow plume heads proposed to occur beneath the rifts (Simiyu and Keller, 1997). Once initiated the downwarp would have been accentuated by isostatic compensation due to water and sediment loading of the Victoria Basin.

To the west of the Katonga Valley during the late Miocene and Pliocene, the George rift basin propagated northwards and the Semliki rift basin propagated southwards until the south end of the intervening crustal block was detached and uplift of the Rwenzori horst commenced about 2.5 Ma (Koehn et al., 2008). During the Pliocene and Pleistocene the George Basin continued to fill with up to 1 km of sediment (Upcott et al., 1996), which created isostatic depression of the basin and associated flexural uplift of the rift flank (Watts, 2001). The location of the Katonga Valley to the east of the Rwenzori horst at a pivotal location between the SSW-NNW trending Edward Basin and the SW-NE trending Albert Basin may also have influence the tectonic geomorphology.
4.6.2 Climate

Global temperature reached a maximum in the early Eocene (52 Ma), before decreasing towards the start of the Oligocene (34 Ma) when Antarctic glaciation commenced. The temperature increased again at the start of the Miocene (23 Ma) with an optimum in the mid Miocene (15 Ma), prior to cooling again and Antarctic reglaciation (Zachos et al., 2001). The temperature continued to fall gradually during the late Miocene to the end of the Pliocene (2.6 Ma) when Arctic glaciation commenced (Lisiecki and Raymo, 2005). The average global temperature began to decrease more rapidly in the early Pleistocene (2.6 Ma to 1 Ma) albeit with large temperature variations influenced by the Earth’s orbital parameters on a 41 kyr cycle. From about 1 Ma the temperature variations became more pronounced and changed to the 100 ky cycle associated with significant northern hemisphere ice expansion and contraction (Petit et al., 1999).

In East Africa, an increase in plants using the C\textsubscript{4} carbon fixation method compared to C\textsubscript{3} method occurred from the late Miocene (7 Ma) and especially during the Pleistocene (<2.5 Ma). This indicates a gradual change to drier average conditions (Bonnefille, 2010); however, grassland savannah may not have been established until the mid Pleistocene (1 Ma) (Cerling, 1992). The drying trend is likely due to the reduction in the atmospheric energy budget associated with global cooling and regional uplift of the EAP and rift flanks which created topographic barriers and disrupted zonal circulation (Sepulchre et al., 2006). The diatom record from alkaline lake sediment in the Kenya Rift indicates large climate fluctuations within this overall drying trend and there is evidence of significant lake periods in both the Pliocene (4.7-4.3 Ma, 4.0-3.9 Ma, 3.4-3.3 Ma and 3.20-2.95 Ma 2.7-2.5 Ma) and the Pleistocene (1.9-1.7 Ma and 1.1-0.9 Ma) (Trauth et al., 2005). A crude calculation based on sedimentation rate suggests that Lake Victoria likely formed in the mid Pleistocene (about 400 ka). Seismic reflection data identified three erosion surfaces which indicate late Pleistocene lake desiccation events, with the last occurring between 15.9 and 14.2 ka. Johnson et al. (2000) suggest these may be related to the 100 ka climate cycles identified in the northern hemisphere.

The wetter and warmer climate of the mid Miocene likely represents the period of optimum development of deep weathering in the Precambrian rocks of Uganda. Although the overall late Neogene regional drying trend may have reduced the average
discharge in the River Katonga, this is likely less significant for understanding the geomorphology than the profound variations about this trend. Whilst the sedimentary record of the Western Rift may have preserved the influence of Pliocene and early Pleistocene climate fluctuation, it seems likely that overwriting of earlier events means that only the late Pleistocene climate fluctuations identified in the Lake Victoria sediment records will be preserved in the Katonga Valley sediment.

4.6.3 Palaeodrainage

Wayland (1931) first recorded that the westward pointing arrow-barb pattern of Lake Kyoga and Katonga Valley tributaries indicate that the eastward flowing rivers of southwest Uganda, including the Kafu, Katonga and Kagera, once flowed west. Without considering the mechanism, several mid 20th century geomorphologists inferred that uplift parallel to the Western Rift initiated river reversal and the subsequent formation of Lake Victoria (Bishop and Trendall, 1967, Doornkamp and Temple, 1966). Whilst the detailed timing of river reversal remains elusive even today, on the basis of the sedimentological evidence summarised below, most researchers have assumed that it occurred in the mid Pleistocene. Doornkamp (1970) pointed out that changes in elevation parallel to the rift margin produces discrete capture of tributary catchments rather than continuum migration of the main north-south drainage divide.

Several researchers have attempted palaeodrainage reconstruction, and whilst some confidence can be placed in their interpolations over 10s of km between modern tributaries and drowned valleys on the lake margins, some workers indulged in more speculative extrapolation across Lake Victoria to the east and the rift valley to the west (Cooke, 1958, Temple, 1966, Ojany, 1971). Given that seismic reflection surveys have since shown that the Victoria Basin contains up to 60 m of sediment (Johnson et al., 2000), the validity of the relationship between purported lake bed channels and the pre-lake drainage pattern is called into question. On the basis of palaeobiological evidence, Pickford et al (1993) propose that outflow continued from the Western Rift into the Congo drainage basin until regional tectonic tilt in the late Pleistocene led to the development of the northern outflow into the Nile.
Temple (1966) showed the elevations of strandlines around the mouth of the River Katonga beside Lake Victoria are clustered. He infers that the upper strandlines (65-55 and 35-25 m above the lake) were formed during a wetter climate and have been gently tilted towards the basin since their formation. He proposes that the lower strandlines (20-18, 14-12, 4-3 m) formed due to episodic downcutting of the Victoria Nile outlet at Jinja. Given that the 35-25 m strandline has been tentatively correlated with lake deposits of less than 60 Ka in the Kagera Valley, it seems plausible that the high tilted strandlines formed during the wetter climate associated with a late stage of the last interglacial (Ipswichian in the UK). The lowest strandline has been radiocarbon dated at 3,720±200 yrs BP, and it seems likely that all three lower strandlines post-date the late Pleistocene lake desiccation event.

4.6.4 Sediments

The sediment in the George Basin west of the Katonga Valley is poorly exposed. However, the sediment at the southern end of the Albert Basin have been uplifted by the Rwenzori horst and provides a detailed record dated back to mid Miocene (15 Ma) using fifteen mollusc associations (Pickford et al., 1993). Roller et al. (2010) recognised five macro base level cycles during which the environment of deposition changed between accommodation-limited alluvial plains and sediment supply-limited lacustrine phases. The relationship between tectonics, relative relief and sediment supply is complex and the stratigraphic record provides few clues to the size and character of the catchment and the timing of river reversal. However, Roller et al. (2010) suggest that generally higher sedimentation rates in the late Pliocene and Pleistocene may be a result of increased supply due to the new relief of the Rwenzori Mountains and/or decreased transport of sediment out of the basin.

In the Kagera Valley 120 km south of the Katonga Valley, the sediment has been investigated in detail at Nsongezi by Wayland and others (Lowe, 1952, Bishop and Posnansky, 1960). Bishop (1969) proposed that a basal gravel and cobble bed was deposited when the river flowed westward. He suggested that the overlying sand was deposited following river reversal in an arm of the newly formed Lake Victoria. An early report also claims to identify false bedding suggesting westerly flow in sediment dated to the early Pleistocene about 12 km west of Nsongezi (Solomon, 1939). Whilst
the general trend of upward decreasing grain size has been observed in other sections between 30 and 40 km east of Nsongezi (Spurr, 1955) there is significant local variation, and a gravel terrace was identified 19 km east and 41 m higher than the base of the Nsongezi deposits (Bishop, 1969). The Kagera Valley sediment appears to be influenced by a complex history of allochthonous and autochthonous processes and the evidence that they record the transition from a westward to eastward flowing river system appears inconclusive.

The sediment in the Kafu Valley 130 km north of the Katonga Valley records a sequence of high ferruginous conglomerates between about 55 and 75 m above the river and lower terrace gravel a few metres thick and up to 15 m above the river (Bishop, 1969, Hirst, 1926). The gravels have been eroded from the centre of the valley and replaced by sandy deposits overlying clayey deposits up to about 3 m thick. Bishop speculates that the high terrace gravels were deposited prior to river reversal, but once again there is no strong evidence to support this assertion.

Quaternary sediment in the Katonga Valley is poorly exposed and there have been no detailed investigations. A set of auger logs recorded in the western valley near Kabagole suggest an increase in grain size from clay near the surface to sandy deposits at 2 to 3 m bgl (Waldron, 1952). In the eastern valley near Kisozi two borehole logs record up to 6 m of clayey overbank and/or paludal sediment overlying fluvial lacustrine silty sand with gravel to a depth of 55 m (China-Zhonghao, 2006). Coarse beach deposits associated with the high level strandlines also occur north of the wide eastern section of the Katonga Valley (Johnson, 1960). The sand and gravel that occurs as terraces in the central Katonga Valley, the northern Nabakazi Valley, and the intervening interfluves is perhaps the most enigmatic deposit. The poorly to moderately sorted, angular to sub-angular sand and gravel, with sub-angular to sub-rounded cobbles logged at Kyai in the Katonga Valley are characteristic of deposition by low frequency, high magnitude, events as might be expected in a arid environment. They appear to have once filled the valley prior to the base level and climate change that initiated renewed erosion and formed the terraces that exist today.
5 APATITE FISSION TRACK ANALYSIS – BURIAL AND EXHUMATION OF A GONDWANAN PALAEOVALLEY

5.1 Introduction

The three small outliers of sedimentary rocks identified in the Katonga and Muzizi valleys are shown in Figure 5-1. As discussed in Section 3.4.6, the Katonga Beds at Bihanga Station have been reinterpreted as Permo-Carboniferous glaciogenic sediments. The indurated character of these rocks leads to the intriguing possibility that the River Katonga has exhumed a once buried palaeovalley and a thick sedimentary cover sequence once overlaid the entire Katonga region at some time between the late Palaeozoic and late Cenozoic. This hypothesis has been tested using the thermochronological technique known as apatite fission track (AFT) analysis. Thermochronology has been described as the ‘quantitative study of the thermal histories of rocks using temperature sensitive radiometric dating methods’ (Lisker et al., 2009). The interpreted thermal histories of the apatite contained in rock samples from the eastern flank of the Western Rift adjacent to Lake George are used to infer the Phanerozoic history of burial and exhumation of the Katonga Valley and the surrounding region.

Following this introduction, this section describes the basic principles of AFT analysis before documenting the sample collection and preparation. A more detailed description of AFT analysis theory and methods is provided in Appendix F. The section then goes on to present the results of the fission track age calculations and thermal history modelling.

5.2 Basic principles

Radiometric dating is based on the comparison of the abundance of a naturally occurring radioactive isotope of known half-life and its decay products. Fission track analysis is similar to other radiometric dating techniques except that the daughter product is damage to the crystal lattice rather than another isotope (Gallagher et al., 1998b, Lisker et al., 2009). The decay of natural trace radionuclides produces linear tracks of disrupted atoms in insulating solids, reflecting atomic scale damage to the crystal lattice. The unenhanced tracks are generally between 3 and 14 nm wide and up
to 20 μm long (Paul and Fitzgerald, 1992). Whilst there is continued debate over the exact mechanism of track formation, the ‘ion explosion spike’ model is widely accepted (Fleischer et al., 1965, Tagami and O'Sullivan, 2005). According to this model, the track forms in three steps:

1. The rapidly moving positively charged decay particle strips lattice electrons, leaving an array of positive ions;
2. The resultant cations repel each other creating vacancies in the lattice structure; and,
3. The stressed region relaxes elastically, straining the initially undamaged lattice.

Due to its relative abundance, spontaneous tracks found in minerals are almost exclusively produced by the decay of $^{238}\text{U}$. Approximately 99.3% of natural uranium is $^{238}\text{U}$ with a half-life of 4.5 billion years, and the remaining 0.7% is largely $^{235}\text{U}$ with a half-life 0.7 billion years. Fortunately the concentration of $^{238}\text{U}$ in the commonly occurring phosphate mineral apatite is in the appropriate range to ensure that the track count on geological timescales is not too low or too high to facilitate measurement. The generic formula for the composition of apatite is $\text{Ca}_5(\text{PO}_4)_3(\text{F,Cl,OH})$, where the relative proportion of chloride, fluoride and hydroxide varies. Thus the $^{238}\text{U}$ concentration and fission track density in apatite may be used to interpret the fission track age.

Fission tracks in apatite have the additional useful characteristic of annealing over geological time scales, which leads to track shortening. The theoretical factors that cause annealing of fission tracks include: 1) temperature; 2) time; 3) pressure; 4) chemical solutions; and 5) ionising radiation (Gallagher et al., 1998a). However, temperature and time are the primary factors that influence the annealing rate of apatite in a geological setting. Therefore, the track length distribution contains information about the thermal history of the rock. Although limited annealing occurs in apatite at near-surface temperatures over long geological timescales, significant partial annealing begins at temperatures of around 60 °C and complete annealing is achieved over temperatures of about 110 °C. This range is often referred to as the partial annealing zone (PAZ). However, the actual temperature range in which significant partial annealing occurs varies with the prevailing conditions and apatite composition.
Laboratory experiments have been conducted to produce quantitative empirical models of the reduction in track length with time for given temperatures (Laslett et al., 1987, Ketcham et al., 1999, Ketcham et al., 2007b). The statistical distribution of fission track length and single grain ages therefore contains information about the thermal history of the apatite grains. Appendix F contains details of the fission track age calculation and the theory of thermal history modelling.

If the historical geothermal gradient of a region is constant and known with reasonable confidence, it may be assumed that the temperature history correlates with exhumation and/or burial history. Typical geothermal gradients away from plate margins are often 25 °C to 30 °C, and therefore the nominal PAZ for apatite is approximately 2 km to 4 km depth. AFT analysis is applicable to examining the thermal history over the duration which rocks have been exhumed from such depths. It is known that the Lower Karoo strata of the Katonga Valley were exposed at the surface during deposition in the Permo-Carboniferous. Therefore, perhaps the primary question that AFT analysis can answer is whether significant track shortening occurred due to burial of the Karoo strata following deposition to below the upper depth of the PAZ.

5.3 Sample Collection and Preparation

5.3.1 Collection

Rock samples were collected for AFT analysis along two 100 km long transects in order to put the thermochronological history of the Lower Karoo rocks from Bihanga Station into the regional context. One transect is orientated approximately west to east, parallel to the Katonga Valley and perpendicular to the rift flank. The other is orientated approximately south to north between the sedimentary outliers at Bihanga Station and Muzizi (Figure 5-1). Given that the rocks in this region are often weathered to a depth of 10s of metres, there were a limited number of localities where samples could be collected. Samples were therefore limited to rocks resistant to weathering which are located close to roads.
Figure 5-1: Location of sedimentary outliers and apatite fission track samples
<table>
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<th>Northing UTM</th>
<th>Elev. (m asl)</th>
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Table 5-1: Summary of rock samples collected for apatite fission track analysis

Figure 5-1 shows the location and reference ID for AFT samples collected during this study, together with the location of samples collected by Bauer et al (2010), with which the results of this study will be compared. The UTM coordinates, rock name and sample mass are summarised in Table 5-1. Due to the strength of unweathered granites and gneisses and smooth character of the outcrops, in several cases it proved impractical to obtain samples direct from in situ exposures. Therefore, it was necessary to obtain samples from detached boulders using a sledgehammer. Nevertheless, care was taken to obtain samples of similar character to the in situ rock based on proximity and lithology. Given the limited size of all exposures in relation to the scale of the study and the lack of mechanisms or evidence of long-distance transport at the sample sites, this non-ideal field procedure is not thought to introduce significant uncertainty at the scale of investigation. The main outcrop of the Katonga Beds at Bihanga is exposed in a railway cutting and boulders blasted from the cutting lie on the adjacent flat ground within a few 10s of metres of the outcrop. Once again, due to the rock strength, it was necessary to obtain samples from excavation debris in the form of large boulders resting on adjacent ground comprised of recent silty alluvium. Samples were
acquired from fresh relatively unweathered boulders. Given that the alternative was to obtain no samples from the Karoo strata, the risk of sampling from the excavation debris was considered acceptable.

Following the sample preparation described below, the suitability of apatite present for fission track analysis was assessed. Those samples with little apatite or unsuitable track definition are highlighted in Figure 5-1 and Table 5-1. Samples from nine out of fourteen sampling locations were suitable for AFT analysis, and fortunately these covered a wide spread of the sampled transects. Two samples of the Karoo diamictite (GB13 and GB14) were collected and analysed for comparison.

5.3.2 Preparation

Apatite crystals or crystal fragments commonly occur as an accessory mineral in igneous, metamorphic and sedimentary rocks. It is only slightly magnetic and has a density of 3.15 to 3.20 g/cm³, which makes it a heavy mineral compared to common rock forming minerals such as quartz, feldspar and calcite which have a density less than 2.9 g/cm³. Apatite usually amounts to less than 1% of the rock volume. It occurs as fine sand sized grains, typically < 300 µm. A minimum diameter of > 50 µm is usually required to facilitate AFT analysis (Donelick et al., 2005). Given the sparsity of apatite in rock samples it must be separated and concentrated before its availability can be determined and fission track analysis performed. The procedure used during this study for separating the apatite grains from the whole rock sample is summarised below.

- Disaggregate a rock sample weighing between 1 and 2 kg by first putting it through a jaw crusher and then a disk mill, being careful to thoroughly clean equipment between samples.
- Sieve the crushed sample using a 300 µm mesh size, and then hand wash the < 300 µm fraction with tap water until the water runs clear, before drying under a low heat overnight.
- Conduct first stage heavy liquid separation by filling large (1000 ml) conical separatory funnel to the wide section with bromoform (density ≈ 2.9 g/cm³) and
adding < 300 μm sample. Stir to encourage separation whilst draining heavy fraction with bromoform through the stop cock at the bottom. Collect solids in filter paper whilst draining bromoform. Wash sample thoroughly in acetone before drying under heat lamp.

- Conduct magnetic separation using Frantz Magnetic Barrier Laboratory Separator (Model LB-1) by pouring the dried heavy mineral fraction of the sample into the top of the chute inclined at a slope of about 25° with a tilt of about 10°. The rate at which the grains travel down the chute is controlled by adjusting the vibration frequency whilst the magnetic portion is retained above the barrier, as the non-magnetic portion falls below the barrier.

- Conduct second stage heavy liquid separation by filling small (100 ml) conical separatory funnel to wide section with diiodomethane (density ≈ 3.3 g/cm³) and adding non-magnetic sample fraction. Stir to encourage separation whilst draining heavy fraction with diiodomethane through the stop cock at the bottom. Retain the floating fraction (containing apatite) and flush separately onto filter paper using acetone and then dry under heat lamp.

- If concentrate sample is large and still contains a significant non-apatite fraction, it may be desirable to conduct final sieve with < 150 μm mesh cloth.

The apatite grains are then mounted on glass slides with the sample identity scratched on the reverse side. A small amount of epoxy resin is applied to the slide and the apatite concentrate is poured onto the resin and gently stirred to homogenize the slide whilst being careful not to create air bubbles before placing on a hot plate to set. The initial horizontal polished surface is cut manually by circulating the slide on P180 (82 μm) and/or P500 (30.22 μm) aluminium oxide (Al₂O₃) wet and dry paper. The final polishing is conducted using a mechanical polishing machine first with 3.0 μm alumina slurry and finally with 0.3 μm alumina slurry. The quality of the mount is assessed using a binocular microscope.

In order to measure the track density and length using a high magnification optical microscope it is necessary to enhance their visibility by etching the surface of each apatite mount with nitric acid (HNO₃). The tracks are etched by dipping them in an
aqueous solution of H\textsubscript{3}NO\textsubscript{4} with a concentration of 5 mol/L (31\%) for 20 seconds at 20 °C (room temperature), after which they are immediately washed in distilled water and left to dry in open air. Once etched, the feasibility of measuring the fission track parameters is assessed under a high magnification optical microscope (x1200). If suitable, the slides are cut to fit into a standard irradiation tube and prepared to determine the uranium concentration as discussed below in Appendix F.

5.4 AFT Age, Length and Apatite Composition for All Samples

5.4.1 Fission track density and central ages

Table 5-2 presents the results of the fission track density counting and the calculated central age statistics. The age calculations are based on the number of fission tracks counted in 20 to 24 grains per rock sample. As recommended, the total fission track count in the detector, exceeded 4000 (see Appendix F.1.1). All spontaneous fission track counts are over 1000 except for GB19 (Bisyoro gneiss) which is 640. All induced fission track counts are over 700 except again for GB19 (Bisyoro gneiss) which is 350. All spontaneous fission track densities are higher than induced fission track densities except in the case of GB03 (Muzizi granulite).
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<th>$Ns$</th>
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<th>$Ni$</th>
<th>$\rho_i$</th>
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<th>RSE (%)</th>
<th>$P\chi^2$ (%)</th>
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<th>$\bar{x}$ Length ±1 SE (µm)</th>
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<tr>
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<td>Muzizi granulite</td>
<td>20</td>
<td>4380</td>
<td>1.053</td>
<td>1138</td>
<td>2.438</td>
<td>1262</td>
<td>2.713</td>
<td>158.3±7.6</td>
<td>8.3</td>
<td>21.4</td>
<td>65</td>
<td>12.29±0.25</td>
<td>2.04</td>
</tr>
<tr>
<td>GB05</td>
<td>Muzizi granite</td>
<td>20</td>
<td>4380</td>
<td>1.053</td>
<td>1442</td>
<td>1.833</td>
<td>1242</td>
<td>1.565</td>
<td>203.5±8.9</td>
<td>6.0</td>
<td>28.3</td>
<td>83</td>
<td>11.9±0.19</td>
<td>1.86</td>
</tr>
<tr>
<td>GB07</td>
<td>Kyenjojo granite</td>
<td>20</td>
<td>4380</td>
<td>1.053</td>
<td>1026</td>
<td>1.540</td>
<td>636</td>
<td>0.934</td>
<td>279.6±20.5</td>
<td>22.3</td>
<td>0.4</td>
<td>71</td>
<td>11.66±0.22</td>
<td>1.83</td>
</tr>
<tr>
<td>GB13</td>
<td>Bihanga diamicrite</td>
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<td>4380</td>
<td>1.053</td>
<td>1548</td>
<td>2.041</td>
<td>991</td>
<td>1.333</td>
<td>273.2±13.4</td>
<td>9.9</td>
<td>18.5</td>
<td>98</td>
<td>12.21±0.17</td>
<td>1.78</td>
</tr>
<tr>
<td>GB14</td>
<td>Bihanga diamicrite</td>
<td>20</td>
<td>4380</td>
<td>1.053</td>
<td>1593</td>
<td>2.304</td>
<td>1217</td>
<td>1.755</td>
<td>228.9±10.2</td>
<td>7.2</td>
<td>28.4</td>
<td>124</td>
<td>11.69±0.16</td>
<td>1.83</td>
</tr>
<tr>
<td>GB15</td>
<td>Mpanga gneiss</td>
<td>22</td>
<td>4380</td>
<td>1.053</td>
<td>1280</td>
<td>1.662</td>
<td>883</td>
<td>1.163</td>
<td>256.6±13.5</td>
<td>11.1</td>
<td>23.1</td>
<td>58</td>
<td>12.09±0.26</td>
<td>2.05</td>
</tr>
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<td>GB17</td>
<td>Mpanga granulite</td>
<td>20</td>
<td>4380</td>
<td>1.053</td>
<td>1327</td>
<td>1.374</td>
<td>732</td>
<td>0.767</td>
<td>314.7±15.4</td>
<td>3.2</td>
<td>37.1</td>
<td>115</td>
<td>12.02±0.16</td>
<td>1.73</td>
</tr>
<tr>
<td>GB19</td>
<td>Bisyoro gneiss</td>
<td>20</td>
<td>4380</td>
<td>1.053</td>
<td>640</td>
<td>0.681</td>
<td>350</td>
<td>0.370</td>
<td>318.5±25.1</td>
<td>17.0</td>
<td>16.3</td>
<td>103</td>
<td>12.48±0.16</td>
<td>1.64</td>
</tr>
<tr>
<td>GB20</td>
<td>Mubende granite</td>
<td>20</td>
<td>4380</td>
<td>1.053</td>
<td>1770</td>
<td>2.570</td>
<td>1135</td>
<td>1.625</td>
<td>271.6±15.9</td>
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<td>0.10</td>
<td>100</td>
<td>12.03±0.19</td>
<td>1.89</td>
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<td>GB22</td>
<td>Busegele granite</td>
<td>20</td>
<td>4380</td>
<td>1.053</td>
<td>1307</td>
<td>0.832</td>
<td>666</td>
<td>0.429</td>
<td>335.6±19.0</td>
<td>11.3</td>
<td>31.6</td>
<td>95</td>
<td>12.36±0.17</td>
<td>1.69</td>
</tr>
</tbody>
</table>

**Table 5-2: Summary of fission track analytical data and results of central age calculations**

(i) Track densities are (x10^6 tr cm^-2) numbers of tracks counted (N) shown in brackets; (ii) Analyses by external detector method using 0.5 for the 4π/2π geometry correction factor; (iii) Ages calculated using dosimeter glass CN-5; (apatite) $\zeta_{CN5} = 339±5$; calibrated by multiple analyses of IUGS apatite and zircon age standards (Hurford1990); (iv) $P\chi^2$ is probability for obtaining $\chi^2$ value for $v$ degrees of freedom, where $v$ = no. crystals - 1; (v) Central age is a modal age, weighted for different precisions of individual crystals (Galbraith, 2005)
The calculated central ages (quoted with one standard error) are relatively old and vary between 158 ±8 Ma (late Jurassic) and 336 ±19 Ma (middle Carboniferous). The central ages for the Bihanga Station diamictite are 273.2 ±13.4 Ma (middle Permian) and 228.9 ±10.2 Ma (middle Triassic) which are generally consistent with late Palaeozoic exhumation and deposition. As expected, samples with older ages and smaller fission track counts have larger standard errors on the central age. The standard errors vary between 8 Ma and 25 Ma. The ±2SE range, which may be thought of as the 95\% confidence range on the geometric mean age, is therefore between 32 My and 100 My. The largest ±2SE range is associated with GB19 (Bisyoro gneiss) which has the second oldest central age and lowest track density counts. The ±2SE range is a useful indication of precision when attempting to explain associated exhumation in terms of independently proposed geological events. Figure 5-2 shows that the trend on the ±2SE range increases exponentially from about 30 My when the central age is 150 Ma (early Jurassic) to about 80 My when it is 325 Ma (mid Carboniferous).

![Figure 5-2: ±/±2SE range versus central age](image)

### 5.4.2 Same-sample single grain age error and dispersion

Figure 5-3 shows a set of histograms of same-sample grain ages for all ten samples. Due to the decrease in precision with age, histograms are generally regarded as a poor representation of AFT age data. Nevertheless, they do give a first impression of the dispersion of same-sample grain ages, which is between about 200 Myr and 400 Myr.
Figure 5-3: Histograms of single grain ages (Ma)
Figure 5-4: Histograms of standardised log grain ages (z scores)
The dispersion in the same-sample grain ages shown in Figure 5-3 occurs due to several reasons including variability in apatite composition, measurement error and the stochastic nature of the radioactive decay process. Figure 5-4 shows histograms of the standardised log grain ages (z scores) together with the normal distribution curve. A chi-squared ($\chi^2$) test was used to show that at the 95% confidence level there is no significant difference between the distribution of the log same-sample grain ages and the normal distribution. Therefore it is considered acceptable to use parametric statistics to estimate average sample AFT age (see appendix Section F.2).

![Figure 5-5: Radial plots of single grain age distribution for samples GB03, GB05, GB07 and GB17](image)

Figure 5-5: Radial plots of single grain age distribution for samples GB03, GB05, GB07 and GB17
Figure 5-6: Radial plots of single grain age distribution for samples GB13, GB14, GB15, GB19, GB20, GB22
Figure 5-5 and 5-6 show the radial plots (see Appendix F.2.3) of same-sample grain ages for all ten samples produced using Radial Plotter (Vermeesch, 2009). The $D_{par}$ value are indicated using a colour scale. The radial plots provide a means of visualising the ±2SE on the central age and the individual grain ages at the same time. Since y is calculated relative to the central age, the horizontal lines of $y = 0$ and the ±2SE of $y$ extrapolated to the arc scale, give the central age and ±2SE on the central age. Since the arc scale is logarithmic, care should be taken when comparing the distance between the ±2SE of $y$ bars on different plots. Samples with smaller central ages will have ±2SE of $y$ bars further apart than samples with larger central ages, but an equivalent standard error.

The relative standard error (RSE) on individual grain ages is shown as a percentage on the x axis. Rock samples GB05, GB17, GB14 and GB20 all have two or less single grain ages with an RSE of greater than 25%. These samples contained apatite grains with consistently high track counts in each grain. Samples GB07 and GB19 have a significant number of individual grain ages with RSE over 25%. These samples have the lowest overall track counts and a significant number of grains with particularly low counts. Particular care should be taken not to over-interpret the central ages ascribed to these samples. If the same-sample single grain ages on an individual radial plot have the same variance and belong to a single normal distribution then we would expect 95% of all data points to lie within two standard errors of $y$. Given that there are between 20 and 24 data points, only one or two should lie outside of the horizontal two standard errors of $y$ bars. This criterion is met by all samples except GB07, GB20.

Given that the ratio of $^{238}$U to $^{235}$U is assumed to be constant and the track density depends on the uranium concentration, the ratio of spontaneous to induced tracks should also be similar. The results of a chi-squared ($\chi^2$) test to examine if there is a significant difference between the ratios of spontaneous to induced fission tracks determined for same-sample singles grains is presented both on the radial plots and in Table 5-2. The null hypothesis that there is no significant difference between the spontaneous and induced fission track ratios is rejected if the probability ($p$-value) of the calculated $\chi^2(n-1)$ exceeding the $\chi^2$ statistic is less than 5% (0.05). Samples GB07, GB20 fail this criterion. This suggests that care should be taken not to over-interpret
these particular results. Given that the fission track age is proportional to $\rho_s/\rho_i$, those samples with several single-grain ages outlying the 95% confidence limits on the central age, may be expected to fail the chi squared test at the 5% significance level. The overall dispersion of the age estimates has been assessed using the relative standard deviation of the log-normal distribution of $\rho_s/\rho_i$. Rock samples GB07, GB19 and GB20 all have dispersion greater than 15% (Figure 5-5 and 5-6).

### 5.4.3 Fission track length statistics

The mean spontaneous fission track length and associated standard error on the mean and standard deviation of the sample are presented in Table 5-2 together with the samples size. The track length distributions are based on between 58 and 115 track lengths. Figure 5-7 shows the histograms of observed track length and c-axis corrected track lengths for all ten rock samples (see Appendix F.3.2). Several of the observed track length distributions are negatively skewed, most likely due to accelerated annealing at increasing angles to the c-axis.

Since the statistics quoted in Table 5.2 assume that the data is normally distributed, they should be interpreted with caution for skewed samples. In all cases, the histograms of c-axis projected lengths give the visual impression of being closer to a normal distribution. The only track length distribution which appears poorly defined is GB15 (Mpanga gneiss). Not surprisingly, this sample is the smallest and has the largest standard error and standard deviation. The mean observed track lengths and associated standard error varies between 11.69 ±0.16 $\mu$m and 12.48 ±0.16 $\mu$m. Given that the mean $D_{pat}$ for all samples is 2.05 $\mu$m, the observed initial mean track length is estimated to be 15.92 $\mu$m. This indicates that there has been significant annealing since track formation, with an average track length reduction of between 4.23 $\mu$m and 3.44 $\mu$m. This will be explored further using thermal history modelling.
Figure 5-7: Histograms of observed and c-axis projected track-in-track horizontal confined fission track lengths (μm)
5.4.4 Apatite composition and uranium content

Table 5-3 presents summary statistics for the estimated $^{235}\text{U}$ concentration and $D_{\text{par}}$ measurements for each rock sample. The concentrations are estimated from the ratio of the density of induced fission tracks in the sample to the density of induced fission tracks in the glass standard, multiplied by the $^{235}\text{U}$ concentration in the glass standard. The current day ratio of $^{235}\text{U}$ to $^{238}\text{U}$ in nature is approximately 0.00725 and since $^{238}\text{U}$ is 138 times more abundant than $^{235}\text{U}$, it is largely responsible for spontaneous fission track formation. The results of $^{235}\text{U}$ concentration in individual grains show a wide variation from about 2 ppm to about 62 ppm. The mean concentration (with one standard deviation) in samples varies between about 5 ±2 ppm and 31 ±15 ppm. Given that apatite grains in the diamictite may have a diverse provenance, it might be expected that the $^{235}\text{U}$ concentration could show greater variability than that estimated for apatite formed in the igneous and high grade metamorphic rocks. The $^{235}\text{U}$ concentration range for the two diamictite samples (GB13 and GB14) is indeed the first and third largest out of all samples.

Table 5-3 shows that the value of $D_{\text{par}}$ (etch figure diameter parallel to the c-axis, see Appendix F.3.1) for individual grains generally varies between 1.5 $\mu$m and 3.0 $\mu$m with the mean value for rock samples in the range of 1.7 to 2.3 $\mu$m range. The range and distribution of $D_{\text{par}}$ for same-sample grains was such that all can be regarded as belonging to the same population.
<table>
<thead>
<tr>
<th>I.D.</th>
<th>Rock Name</th>
<th>No. grains</th>
<th>( ^{235}\text{U} ) Concentration (ppm)</th>
<th>( D_{\text{par}} ) (μm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Mean</td>
<td>SD</td>
</tr>
<tr>
<td>GB03</td>
<td>Muzizi granulite</td>
<td>20</td>
<td>30.9</td>
<td>14.8</td>
</tr>
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<td>GB05</td>
<td>Muzizi granite</td>
<td>20</td>
<td>17.8</td>
<td>4.3</td>
</tr>
<tr>
<td>GB07</td>
<td>Kyenjojo granite</td>
<td>20</td>
<td>10.6</td>
<td>3.5</td>
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<td>GB13</td>
<td>Bihanga diamicrite</td>
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<td>14.2</td>
</tr>
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<td>Mpanga gneiss</td>
<td>22</td>
<td>13.2</td>
<td>9.0</td>
</tr>
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<td>Mpanga granulite</td>
<td>20</td>
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<td>3.0</td>
</tr>
<tr>
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<td>Bisyoro gneiss</td>
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<td>4.2</td>
<td>1.7</td>
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<td>GB20</td>
<td>Mubende granite</td>
<td>20</td>
<td>18.5</td>
<td>3.6</td>
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<td>GB22</td>
<td>Busegele granite</td>
<td>20</td>
<td>4.9</td>
<td>1.8</td>
</tr>
</tbody>
</table>

Table 5-3: Summary statistics for \( ^{235}\text{U} \) concentration and \( D_{\text{par}} \)
5.5  AFT Age and Thermal History Models for Bihanga Diamictite

5.5.1 Apatite fission track age

Samples GB13 and GB14 are both diamictite samples, interpreted to be of Lower Karoo (Permo-Carboniferous) glaciogenic origin on the basis of geological evidence. They were sampled from two boulders within 10s of metres of each other and the railway cutting from which they were excavated during the 1950s. Given that it is a sedimentary rock, the specific provenance of the secondary apatite is uncertain. Whilst the rock characteristics suggest that individual clasts have not been transported far (Section 3.4.6), it is possible that the apatite grains may have more diverse history and composition than the primary apatite in the granite and gneiss samples. The central age and associated standard error for GB13 is 273.2 ±13.4 and for GB14 it is 228.9 ±10.2. The ±2SE range for GB13 is 246 Ma to 300 Ma and for GB14 it is 209 Ma to 249 Ma. There is therefore 44 Myr difference in the central ages and the ±2SE range only overlaps by 3 Myr at the start of the Triassic. The two ±2SE ranges encompass 92 Myr from end of the Carboniferous to end of the Triassic.

Examination of the arithmetic histograms of same-sample grain ages in Figure 5-3 shows that the wider age range for GB13 may be influence by a high single-grain age outlier. The mean of the observed track lengths of GB14 (11.69 ±0.16 mm) is slightly shorter than GB13 (12.21 ±0.17). A Student’s t test indicates that there is no significant difference at the 95% confidence level (p = 0.10) in the distribution of Dpar between samples GB13 and GB14. This suggests that the range of apatite composition is similar in both rock samples. Table 5-2 reveals that whilst there is only a 3% difference in spontaneous track density between GB13 and GB14, there is a 19% difference in the induced track density. This indicates that the difference in estimated central ages is more strongly influenced by the measured induced track density (proportional to uranium content) than the spontaneous track density (proportional to age and uranium content). The difference in estimated central ages may be explained by a difference of induced track density measurements in four out of twenty four grains, which highlights the sensitivity of the age calculation.
5.5.2 Thermal history modelling

The thermal history modelling theory is described in Appendix F.4, and the particular approach adopted during this study is described in Appendix F.5. The results of the thermal history modelling for the Bihanga diamictite (GB13 and GB14) samples are presented in Figures 5-8a and 5-8b. The horizontal axes cover almost the entire Phanerozoic Eon from 500 Ma to the present. The vertical axes cover the temperature range from 0°C to 160°C. The nominal boundaries of the partial annealing zone (PAZ) are shown at around 60°C and 110°C. The general constraint boxes used to encourage inflexions are shown as thin blue lines. The present day temperature constraint of 20°C is indicated by a large yellow dot. The yellow box between 0°C to 20°C represents the additional surface temperature constraint on the Permo-Carboniferous sediments.

The central ages with associated ±2SE have been annotated as vertical black lines on the plots to facilitate comparison. The divergence between the central ages calculated for GB13 and GB14 and the small degree of overlap in the ±2SE range is immediately apparent. The ‘unconstrained’ area of the time-temperature plot corresponds to the region which is earlier than the oldest fission track population that is not completely annealed according to the ‘best fit’ model. The oldest fission track age is estimated to be 390 Ma (middle Devonian) for GB13 and 336 Ma (middle Carboniferous) for GB14. Given that the thermal history modelling code, HeFTy, selects models according to their goodness of fit to calculated ages as well as track length data, and we have already seen that the higher average induced fission track density produces a younger central age for GB14, it is unsurprising that HeFTy also predicts a younger ‘oldest age’ for GB14.
Figure 5-8: Annotated thermal history models of AFT data sets for the Bihanga diamictite
The ‘acceptable’ time-temperature (t-T) paths shown in green represent those paths where there is a less than 5% chance that a random sample, of size $n$, drawn from the continuous modelled length distribution would have a greater separation from it than the measured length distribution, also of size $n$ (Appendix F.4.2). They may be thought of as representing those t-T paths for which there is greater than 95% confidence of no significant difference. The ‘good’ t-T paths shown in pink represent those paths where there is less than 50% chance that a random sample, of size $n$, drawn from the modelled length distribution would have a greater separation from it than the measured length distribution, also of size $n$. In other words, we can have confidence that the green t-T paths produce modelled distributions that are not significantly different from the measured distributions, but the pink t-T paths produce modelled distributions that have greater similarity to the measured distribution. Given the overall variability and uncertainty inherent in thermal history modelling, no attempt is made to draw attention to those t-T paths with apparently greater than 50% chance of similarity or the ‘best fit’ track itself. It is immediately apparent from Figures 5-8a and 5-8b that many ‘good’ t-T paths are consistent with the constraint of Permo-Carboniferous surface temperatures. The AFT data therefore supports the conclusion from the geological evidence that the Bihanga diamictite is of Lower Karoo-age.

No ‘good’ or ‘acceptable’ t-T paths occur in the lower right part of the time-temperature plots. The upper boundary of this area is determined by the lower tails of the track length and age distributions. Younger tracks have experienced less annealing and therefore there are no ‘good’ or ‘acceptable’ t-T paths in this region of the plot consistent with the data. Since fission tracks shorten rather than grow after initial formation, there is no equivalent upper boundary on the region occupied by ‘good’ and ‘acceptable’ t-T paths. The t-T paths that rise above the PAZ may compensate for the initial history of slow annealing by re-entering the PAZ, thus enhancing the later annealing rate. The ‘good’ and ‘acceptable’ paths therefore fill all of the time-temperature space above the minimum-time/maximum-temperature (min-t/max-T) boundary line. The general shape of the min-t/max-T boundary is well defined. It enters the PAZ at a steep gradient and then adopts a monotonic trend between high temperature boundary of the PAZ and low temperature boundary of the PAZ at the present day.
Given that the min-t/max-T boundary line is well defined, it provides a useful means of comparing thermal history models for different samples. The min-t/max-T boundary line for GB13 (Figure 5-8a) is compared against all other models produced during this study. Figure 5-8b shows that whilst there is a good match between the base of the ‘acceptable’ matches produced by GB14 and GB13 at times younger than the GB14 central age, at older times there are ‘acceptable’ GB14 t-T paths which are younger than the GB13 min-t/max-T boundary. This difference in character between the thermal history models appears to be a direct result of matching the age distribution for GB14 based on fewer grains with low induced track densities but similar spontaneous track densities to GB13. The general relationship between the central ages and the thermal history will be examined in the context of other samples, where no Permo-Carboniferous low temperature constraint has been imposed.

When viewing the time-temperature plots shown in Figure 5-8, it is tempting to hypothesise information-laden complex thermal histories. Therefore it is important to remember that the information content of the thermal history models cannot be greater than the information content of the simple age and length histograms against which the models are calibrated. Examination of Figure 5-8a and 5-8b reveals the key characteristics of the t-T paths listed below.

- ‘Acceptable’ t-T paths fill all of the time-temperature space above the min-t/max-T boundary, except for a small region above the modern temperature constraint of 20°C.
- ‘Good’ t-T paths enter and exit the late Palaeozoic temperature constraint box over the entire age range of the Permo-Carboniferous.
- All ‘good’ t-T paths experience increased temperatures after exiting the Permo-Carboniferous surface temperature constraint box and most appear to re-enter the PAZ.
- The ‘good’ t-T paths are clustered above and below low temperature boundary of the nominal PAZ during the Mesozoic, with central temperature of the ‘good’ t-T path cluster increasing from about 80°C to 60°C during the Mesozoic.
- The central temperature of the cluster of ‘good’ t-T paths increases from about 60°C to 20°C during the Cenozoic, such that the average rate of cooling appears higher in the Cenozoic than the Mesozoic.

- Some ‘good’ t-T paths attain low near-surface temperatures during the Mesozoic before the temperature increases again.

5.5.3 Examination of reheating and cooling history

The apparent trend of post Permian t-T paths re-entering the PAZ was explored further by forcing the t-T paths to remain at a temperature below 60°C following the Permian. It was found that no ‘good’ or ‘acceptable’ t-T paths occurred within 100,000 realisations. This result strongly suggests that the Bihanga diamictite has been buried to a depth that facilitates significant annealing. When the high temperature constraint is increased to 70°C there were 9 ‘good’ t-T paths in 100,000 realisations. Since the annealing rate is still relatively low between 60°C and 70°C, the ‘good’ t-T paths had to remain within the PAZ without episodic cooling during the Mesozoic in order to achieve the observed track shortening. Multiple episodes of cooling and reheating only start to occur when the high temperature constraint is increased to 80°C, at which point there were 89 ‘good’ t-T paths in 100,000 realisations. With no high temperature constraint, there were 118 ‘good’ t-T paths in 100,000 realisations.

The reliability of these conclusions is strongly dependant on the long-duration, low-temperature annealing behaviour represented in the Ketcham et al (2007b) annealing model incorporated into the HeFTy code. As discussed in Appendix F.3.3, the low temperature annealing behaviour described by the similar Ketcham et al (1999) model was corroborated by Spiegel et al (2007a) using apatite in deep sea sediment with low accumulation rates. Their data shows that apatite fission tracks in sediment acquired from the Caribbean Ocean (ODP/DSDP Leg 165, site 999) estimated to be 120 Ma with historical temperature not exceeding 38°C had an average track shortening of 7.5%. If this same rate of track shortening is extrapolated linearly to the central ages calculated for the Bihanga diamictite, it would suggest possible long-duration, low-temperature mean track shortening of 17.1% for GB13 (273 Ma) and 14.3% for GB14 (229 Ma). If we assume an initial track length of 16.05 µm based on an average $D_{par}$ for GB13 and
GB14 of 2.18 μm (Carlson et al., 1999), the estimated mean track shortening is 23.9% for GB13 and 27.2% for GB14. Therefore, the track shortening is greater than might be expected if the historical temperature had remained below the nominal lower temperature boundary of the PAZ, further suggesting that the samples have been significantly reheated, most likely due to burial.

It can be seen in Figures 5-8a and 5-8b that the central cluster of the ‘good’ t-T paths reaches the low temperature boundary of the PAZ in the middle to late Cretaceous (100 Ma to 65 Ma) and, on average, there appears to be an increased rate of cooling during the Cenozoic. However, given that cooling during the Cenozoic occurs at temperatures lower than the nominal PAZ temperature range where annealing is considered negligible, there is no information in the track length data to constrain the actual character or timing of cooling during the Cenozoic. The fact that the change of gradient in the cluster of ‘good’ t-T paths occurs where they exit the low temperature boundary of the PAZ raises the possibility that this is an artefact of the modelling process rather than the data. This late-time increase in the rate of cooling is similar to the behaviour predicted by the superseded annealing model of Laslett et al (1987) which has been shown to under-estimate the amount of low temperature annealing over geological time-scales (Ketcham et al., 1999). Given that the low temperature annealing behaviour represented in the updated Ketcham et al (2007b) model has been corroborated (Spiegel et al., 2007a), another potential cause of this behaviour is an over-estimate of the initial track length. When the initial track length is long, the amount of track shortening required to match the track length data is also more. Hence the t-T paths have to spend longer within the PAZ. If they remain within the PAZ for an extended period then they will need to adopt a steeper rate of cooling below the lower boundary of the PAZ to achieve the present day temperature.

Given that we expect the form of the thermal history to be dependent on the exhumation rate, which is independent of the initial track length, we do not expect the initial track length to cause a change in gradient in the thermal history. The initial track length based on a mean $D_{par}$ of 2.56 μm (Carlson et al., 1999) used during the thermal history modelling of GB13 is 15.97 μm, and the equivalent c-axis projected track length is 16.32 μm. Scoping models showed that altering the initial track length to
15.75 \( \mu m \) resulted in many t-T paths achieving the surface temperature of 20°C in the early Cenozoic. An initial track length of 5.85 \( \mu m \) results in a monotonic trend in the central cluster of ‘good’ t-T paths arriving at 20°C at the present day without a significant change in gradient as the t-T paths exit the low temperature boundary of the PAZ. The time at which the ‘good’ t-T paths exit the low temperature boundary of the PAZ, and the subsequent cooling rate below 60°C is therefore sensitive to the initial track length and a difference of only 0.12 \( \mu m \) makes an appreciable difference to form of the thermal history. This demonstrates one of the many limitations in attributing the detailed form of the thermal history model to the actual thermotectonic events.

### 5.6 Comparison of Thermal History Models for All Samples

#### 5.6.1 Relationship between thermal history models and central ages

Figure 5-9 presents the results of the thermal history modelling for all AFT data sets. It can be seen that the ‘good’ and ‘acceptable’ t-T paths fill the time-temperature space above the min-t/max-T boundary. The region older than the oldest fission track population that is not completely annealed is identified as ‘unconstrained’. The boundary of the unconstrained region is related to the age at which the t-T paths enter the base (high temperature) boundary of the PAZ. The central age \( \pm 2SE \) determined from the fission track density data is also shown on the thermal history plots. Given that the central age is intended to define the age of the oldest preserved spontaneous fission tracks, it might be expected that it should coincide with the period where the modelled t-T paths enter the base of the PAZ. Examination of Figure 5-9 reveals that the central age \( \pm 2SE \) lines based on the fission track density alone, actually occur to the younger side of where the modelled ‘good; and ‘acceptable’ t-T paths enter the PAZ. This apparent discrepancy occurs because the age equation assumes that no annealing has taken place and the equivalent isotropic track length is the same as the mean track length. This assumption is necessary as the annealing history is unknown. If annealing has occurred then the density of spontaneous fission tracks will appear less than the actual density when sampled on a plane, and the age equation will underestimate the time over which the tracks have been forming (Galbraith, 2005).
Figure 5-9: Thermal history modelling of all data sets
One of the simplest and most robust uses of thermal history modelling is to estimate the potential error (age reduction) in the central age ±2SE due to the assumption of length isotropy (Green et al., 1989, Foster and Gleadow, 1996). In order to compare the ‘measured’ age distribution with the modelled age distribution predicted by any individual t-T path, HeFTY introduced a density reduction which is proportional to the track length reduction calculated using the annealing model for each time step (Ketcham, 2005). The age at which t-T paths enter the base of the PAZ in the thermal history model takes into account the annealing behaviour captured by the length distribution and therefore reflects an upper bound on the start of spontaneous fission track preservation compared to the central age calculation, which does not.

<table>
<thead>
<tr>
<th>I.D.</th>
<th>Rock Name</th>
<th>Central Age +2SE (Ma)</th>
<th>Central Age -2SE (Ma)</th>
<th>Oldest FT Age (Ma)</th>
<th>Youngest t-T path Entering the PAZ (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GB03</td>
<td>Muzizi granulite</td>
<td>174</td>
<td>143</td>
<td>197</td>
<td>165</td>
</tr>
<tr>
<td>GB05</td>
<td>Muzizi granite</td>
<td>221</td>
<td>186</td>
<td>267</td>
<td>220</td>
</tr>
<tr>
<td>GB07</td>
<td>Kyenjojo granite</td>
<td>321</td>
<td>239</td>
<td>364</td>
<td>310</td>
</tr>
<tr>
<td>GB13</td>
<td>Bihanga diamictite</td>
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<td>246</td>
<td>390</td>
<td>290</td>
</tr>
<tr>
<td>GB14</td>
<td>Bihanga diamictite</td>
<td>249</td>
<td>209</td>
<td>336</td>
<td>265</td>
</tr>
<tr>
<td>GB15</td>
<td>Mpanga gneiss</td>
<td>284</td>
<td>230</td>
<td>335</td>
<td>295</td>
</tr>
<tr>
<td>GB17</td>
<td>Mpanga granulite</td>
<td>346</td>
<td>284</td>
<td>399</td>
<td>355</td>
</tr>
<tr>
<td>GB19</td>
<td>Bisyoro gneiss</td>
<td>369</td>
<td>268</td>
<td>384</td>
<td>330</td>
</tr>
<tr>
<td>GB20</td>
<td>Mubende granite</td>
<td>303</td>
<td>240</td>
<td>376</td>
<td>300</td>
</tr>
<tr>
<td>GB22</td>
<td>Busegele granite</td>
<td>374</td>
<td>298</td>
<td>435</td>
<td>375</td>
</tr>
</tbody>
</table>

Table 5-4: Central age range and age range of t-T paths entering the PAZ from thermal history modelling

Table 5-4 presents a comparison of the central age ±2SE and the age range between the oldest fission track and youngest t-T path entering the PAZ derived from the thermal history modelling. The age range at which spontaneous fission track preservation began predicted by the thermal history modelling is older for all samples. Whilst the age range for the youngest samples (GB03 and GB05) remains mainly within the Jurassic and Triassic, the revised age ranges for most samples is generally shifted from
Carboniferous/Permian to Devonian/Carboniferous. The older age range for the Bihanga diamictite samples (GB13 and GB14) is more consistent with exhumation through the PAZ prior to the start of Lower Karoo-age (Permo-Carboniferous) deposition.

Given that the min-t/max-T boundary describes the line of 95% confidence that there is no significant difference between the measured and modelled age, it is related to the central age -2SE at the base of the PAZ by the amount of adjustment to the spontaneous fission track density due to annealing. Since the central age +2SE approximately coincide with the min-t/max-T boundary for several samples included in the present study, the age adjustment due to annealing appears to be similar to ±2SE range. In other words, the central age ±2SE, uncorrected for length anisotropy, lies immediately to the younger side of min-t/max-T boundary (Figure 5-9). Some relatively small discrepancies in this relationship occur because the goodness of fit was assessed using the Kuiper's Statistic, which equalises sensitivity between the tails and the central tendency. For the AFT data analysed by this study, it therefore appears that increasing the central age ±2SE by two standard errors gives an age for the start of spontaneous fission track formation which is more consistent with the thermal history modelling. This potential underestimate in the central age should be taken into account when comparing central ages from different datasets which have undergone different amounts of thermal annealing.

Comparison of the calculated central ages with the thermal history modelling, shows that we can only really define the most recent likely time at which spontaneous fission track preservation began. Spontaneous track preservation could always have started earlier and the tracks subsequently lost due to protracted annealing within the PAZ. The larger cumulative annealing, shorter recorded lengths and greater uncertainty in the contributing track density for early time steps means there is greater relative uncertainty on the older t-T paths, until they become completely unconstrained by the data. This can be seen on the results of the Bayesian QTQt thermal modelling presented in appendix Figure G-1, where the probability decreases towards the oldest track boundary of the model. Appendix G presents a corroboration of HeFTy thermal history modelling, using the QTQt software, which adopts a Markov chain Monte Carlo
approach to t-T path selection. The concept of a central age on the oldest preserved spontaneous fission tracks is only really valid if it is reasonable to assume that monotonic cooling has occurred through the PAZ at a rate that facilitates track preservation.

### 5.6.2 Comparison of thermal history models

This section highlights the differences and similarities between thermal modelling results for different rock samples. The spatial relationship between the AFT ages is discussed in the next section. The interpretation presented here will adopt a conservative approach, nevertheless for completeness, the statistics used to compare the modelled ‘best fit’ t-T path (the thin black t-T paths on Figure 5-9) with the data distribution are presented in Table 5-5. It should be remembered that the ‘acceptable’ (green) t-T paths in Figure 5-9 correspond to a ‘goodness of fit’ (GOF) based in Kuiper’s Statistic of greater than 0.05, and ‘good’ t-T paths correspond to a value greater than 0.5. The GOF reported for the ‘best fit’ model are all much greater than 0.5, and all but two are above 0.9.

<table>
<thead>
<tr>
<th>ID</th>
<th>Rock Name</th>
<th>HeFTy Age Statistics</th>
<th>HeFTy C-axis Length Statistics</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Model (Ma)</td>
<td>Measured (Ma±95%)</td>
</tr>
<tr>
<td>GB03</td>
<td>Muzizi granulite</td>
<td>158</td>
<td>159+15/-14</td>
</tr>
<tr>
<td>GB05</td>
<td>Muzizi granite</td>
<td>204</td>
<td>203+18/-17</td>
</tr>
<tr>
<td>GB07</td>
<td>Kyenjojo granite</td>
<td>282</td>
<td>281+32/-29</td>
</tr>
<tr>
<td>GB13</td>
<td>Bihanga diamictite</td>
<td>273</td>
<td>272+26/-23</td>
</tr>
<tr>
<td>GB14</td>
<td>Bihanga diamictite</td>
<td>229</td>
<td>229+20/-19</td>
</tr>
<tr>
<td>GB15</td>
<td>Mpanga gneiss</td>
<td>253</td>
<td>253+25/-23</td>
</tr>
<tr>
<td>GB17</td>
<td>Mpanga granulite</td>
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<td>315+33/-30</td>
</tr>
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<td>GB19</td>
<td>Bisyoro gneiss</td>
<td>317</td>
<td>317+46/-40</td>
</tr>
<tr>
<td>GB20</td>
<td>Mubende granite</td>
<td>271</td>
<td>272+24/-22</td>
</tr>
<tr>
<td>GB22</td>
<td>Busegele granite</td>
<td>341</td>
<td>340+36/-33</td>
</tr>
</tbody>
</table>

**Table 5-5**: Summary statistics for the HeFTy ‘best fit’ model
It can be seen from Figure 5-9 that for most samples, the ‘good’ t-T paths enter the base of the PAZ in the latter half of the Palaeozoic, and continue to cluster between the middle temperature and low temperature boundaries of the PAZ during the Mesozoic, before cooling during the Cenozoic. However, whilst only the models for the Bihanga diamictite (GB13 and GB14) have low temperature constraints imposed during the Permo-Carboniferous, all models include some ‘good’ t-T paths with episodic low surface temperatures. All samples with Palaeozoic fission tracks are therefore consistent with late Palaeozoic to Mesozoic reburial as supported by the geological control provided by the Karoo-age Bihanga diamictite. The exception is the Muzizi granulite sample (GB03) acquired from 80 km north of the Katonga Valley, which in general produces lower fission track ages. The thermal history modelling for this sample suggests monotonic cooling since entering the base of the PAZ at the beginning of the Mesozoic. This is consistent with this sample residing at a lower depth than other samples during late Palaeozoic re-burial and only being exhumed through the PAZ during the Mesozoic.

Given both theoretical and practical difficulties in comparing the central tendency, or weighted average, in the ‘good’ t-T paths, the min-t/max-T line is used here to compare boundary of the time-temperature space occupied by ‘acceptable’ t-T paths. The solid black lines in Figure 5-9 show the min-t/max-T boundary for individual samples and the dashed line shows the min-t/max-T boundary for GB13 to facilitate comparison with the thermal history model for the Bihanga diamictite. In general, the min-t/max-T boundary has an initial steep gradient into the base of the PAZ, and somewhere between the base and the middle of the PAZ there is a quick gradient reduction. Following the rapid gradient reduction, the min-t/max-T boundary then gradually rises towards the present day. The horizontal shift in the min-t/max-T boundary depends on the ‘measured’ distribution of same-sample single grain ages, and corresponds approximately to the central age +2SE where the min-t/max-T enters the base of the PAZ.

The vertical shift in the min-t/max-T boundary depends on the rate of track shortening required to match the ‘measured’ length distribution. The rate of track shortening was calculated by subtracting the mean c-axis projected track length from the initial c-axis
projected track length (based on Dpar, Carlson et al., 1999) and then dividing by the central age. The track shortening rate for each sample was then divided by the track shortening rate for GB13 for comparison. This confirmed that those samples with the lowest track shortening rates relative to GB13, including GB17 (89%), GB19 (80%), GB22 (74%), also have min-t/max-T boundaries closer to the top of the PAZ than GB13. GB07 (103%), GB15 (109%) and GB20 (100%) have similar track shortening rates to GB13. GB03 (164%) and GB05 (135%) have track shortening rates higher than GB13, but these samples also have lower fission track ages. Given that GB14 is also Bihanga diamictite its track shortening rate which is 134% of the rate for GB13 appears anomalously high and inconsistent with the relatively good agreement between the thermal history modelling. The rate for GB14 appears high due to the low central age, which is appears to be influenced by a few grains with lower induced fission track counts. The influence of the lower central age is masked in the thermal history modelling by the imposition of the Permo-Carboniferous constraints box.

Generally speaking the thermal history modelling suggests that samples GB17 (Mpanga granulite), GB19 (Bisyoro gneiss) and GB22 (Busegele granite) possibly entered the PAZ earlier than the apatite contained in the Bihanga diamictite and were not buried as deeply by subsequent sedimentation. GB07 (Kyenjojo granite), GB15 (Mpanga gneiss) and GB20 (Mubende granite) possibly experienced similar thermal histories to the Bihanga diamictite apatite. Eighty kilometres north of the Katonga Valley, GB05 (Muzizi granite) and GB03 (Muzizi granulite), possibly entered the base of the PAZ later than the Bihanga diamictite, and GB03 was probably below the PAZ during the episode of Karoo reburial.

5.7 Interpretation and Discussion

5.7.1 Regional context of Katonga Valley AFT ages

A review of previous low temperature thermochronology studies in East Africa is presented in Appendix H. The Kohn et al. (2005) constructed the contoured image of AFT age data for eastern Africa shown in Figure 5-10. Some care is required when interpreting this image due to the differences in data density and quality, depending on the number of samples and the age of the work and shown in appendix Figure H-1.
Representations of the AFT ages calculated by this study and the recent Rwenzori study (Bauer et al., 2010) have been added to the region north west of Lake Victoria for which Kohn et al. (2005) had no data. It can be seen that the new data is broadly consistent with the existing pattern of AFT ages previously acquired in the region.

**Figure 5-10: Katonga Valley and Rwenzori (Bauer et al, 2010) AFT ages shown on regional AFT age contours (after Kohn et al, 2005)**

Figure 5-10 shows the central region of the EAP around Lake Victoria, from the Kenya Rift (Wagner et al., 1992) to the Katonga Valley and south onto the Tanzanian Craton (Noble, 1997, Noble et al., 1997) to be comprised of rocks that were exhumed through the PAZ during the Palaeozoic (250 Ma to 400 Ma). This central region is bordered by rocks that were exhumed at a later date during the early to middle Mesozoic (100 Ma to 250 Ma). If uplift of the EAP was most active in the central region, then it would be expected that denudation would have uncovered rocks with younger AFT ages adjacent to Lake Victoria. The general pattern of older AFT ages in the central region of the EAP compared to the younger AFT ages at the margins of the plateau suggests that the areas adjacent to the Western Rift and Kenya Rift have experienced greater uplift and erosion. The distribution of AFT ages is therefore consistent with uplift of the rift margins and a central sag basin currently occupied by Lake Victoria.
To the east, the early to middle Mesozoic ages extend from the Cherangani Hills, north of Eldoret (Foster and Gleadow, 1993), in Kenya, through the central region of the Kenya Rift (Foster and Gleadow, 1992) and down towards the inland eastern region of Tanzania (Noble et al., 1997). The AFT ages acquired towards the east of Kenya (Foster and Gleadow, 1996) and near the coast of Tanzania (Noble, 1997, Noble et al., 1997) indicate late Mesozoic to early Cenozoic (60 Ma to 100 Ma) AFT ages.

To the west, the early to middle Mesozoic AFT ages calculated for the Rwenzori (Bauer et al., 2010) are consistent with those determined in eastern Rwanda and northern Burundi (van den Haute, 1984). The area of southern Burundi to the east of northern Lake Tanganyika with extremely old AFT ages is based on two value of 390 Ma and 423 Ma obtained by the early study of van den Haute (1984). No AFT ages are available for the eastern flank of the Western Rift in Uganda south of the Katonga Valley and the Mesozoic ages shown in this region are extrapolated from Rwanda. Given that the AFT ages for the Katonga Valley are generally of Palaeozoic age, it is likely that contouring based on the new data would result in expansion of the blue region towards the Western Rift flank in south west Uganda. The region of late Mesozoic AFT age between Lake Edward and Lake Kivu is based on three values calculated by van den Haute (1984) no more than 30 km north-east of Lake Kivu.

5.7.2 Burial of the Gondwanan palaeovalley and regional exhumation

AFT thermal history may be related to: a) thermal disturbances relating to tectonics; b) local magmatism; c) geothermal fluid flow or d) denudation of the land surface and consequent exhumation. Kohn et al. (2005) suggest that the thermal effects of tectonics are probably negligible within 10 km of the surface, and the effects of magmatism and geothermal fluids are likely to be local. Therefore most cooling in the near surface crust (< 5 km) will be dominated by denudation. The geothermal gradients contoured across East Africa by Kohn et al. (2005) are shown in Figure 5-11. This is based on data in the U.S. National Geophysical Data Centre global heat flow data set compiled by Pollack et al (1993). The global heat flow dataset suggests that the range of continental geothermal gradients in Precambrian and Palaeozoic terranes across the globe is between about 20°C/km and 30°C/km. Figure 5-11 shows that there is a high variability in the near-surface geothermal gradient in East Africa between about
The geothermal gradient is shown as about 25°C/km in the region of the Katonga Valley, but since there is no local data, this is simply the average expected value. Where there is a lot of data around the Kenya Rift in central Kenya, there appears to be steep horizontal geothermal gradients between about 20°C/km and 35°C/km over 10s km.

Figure 5-11: Contours of present geothermal gradient

The main findings of the thermal history modelling of the Bihanga diamictite AFT data are that it is consistent with the geological interpretation of a Permo-Carboniferous age, and that significant re-heating occurred during the post-Permian thermal history. If re-heating of the Bihanga diamictite occurred primarily due to the deposition of overlying sediment and consequent burial then we can estimate a maximum depth of burial, based on the modelled temperature increase and an assumed geothermal gradient. The broadest possible assumptions, based on a wide range of geothermal gradients (15°C/km to 45°C/km), surface temperatures (0°C to 20°C), and maximum burial temperatures (60°C to 110°C), give a maximum possible range of burial depths from 0.9 km [(60°C – 20°C)/45°C/km] to 7.3 km [(110°C – 0°C)/15°C/km]. A more likely range of geothermal gradients (20°C/km to 30°C/km), surface temperatures (10°C to
20°C) and burial temperatures (80°C to 100°C) gives a probable range of burial depths between 2 km [(80°C – 20°C)/30°C/km] and 4.5 km [(100°C – 10°C)/20°C/km]. These estimates are based on the assumption that the geothermal gradient has not changed in the last 300 Ma. Whilst this seems unlikely, the historical geothermal gradient is unknown, and it is perhaps reasonable to assume that current regional variation captures the range of historical geothermal gradients.

The AFT ages presented in Figure 5-10 are based on values calculated from fission track density measurements alone. As previously discussed, these tend to underestimate the age at which the rocks entered the PAZ by a few 10s of million years when there has been significant fission track annealing. To try and get around this, Kohn et al. (2005) modelled the thermal history of fission track data acquired by Foster and Gleadow (1996) and Noble (1997) for East Africa, based on the annealing model of Laslett et al (1987), using consistent match parameters, approach and constraints. The Laslett et al (1987) model is known to underestimate the low temperature annealing rate and is based on an initial track length of 16.3 μm. To compensate for this issue, Kohn et al. (2005) used a relatively low initial track length of 14.5 μm, based on the typical spontaneous track lengths from geological settings with little thermal disturbance.

Assuming that temperature can be used as a proxy for depth, Kohn et al. (2005) use the geothermal gradient shown in Figure 5-11 to construct maps of the quantity of denudation at various ages between the early Jurassic and middle Cenozoic. Since the current geothermal gradients in East Africa are likely associated with the Neogene tectonic setting the regional denudation maps for the Mesozoic and Palaeogene are probably unduly influenced by recent events and must be treated with caution. Nevertheless some broad inferences can be tentatively suggested. Figure 5-12 shows the estimated amount of denudation since 180 Ma and 120 Ma estimated by Kohn et al. (2005). The lower AFT ages and sometimes higher geothermal gradients in central Kenya and the Tanzanian coastal region suggest up to 6 km of denudation since the mid Cretaceous (120 Ma). Given the older AFT ages and lower geothermal gradients associated with the central area of the EAP, Kohn et al. (2005) estimate that 2 to 4 km of denudation has occurred in this region since the middle Jurassic (180 Ma), and 1 to 2 km since the mid Cretaceous (120 Ma). Whilst the precision on these estimates is
relatively low, it is interesting to note that this range of exhumation in the central region of the East African Plateau since the middle Mesozoic is broadly consistent with the estimated depth of burial of the Katonga palaeovalley in the late Palaeozoic and early Mesozoic.

![Figure 5-12: Contours of cumulative denudation since early Jurassic and middle Cretaceous (after Kohn et al, 2005)](image)

### 5.7.3 Horizontal distribution of the rift flank and Rwenzori AFT ages

The AFT ages calculated by Bauer et al (2010) for the 15 samples acquired from the central and northern Rwenzori Mountains are compared here with the data from the eastern flank of the Western Rift acquired by this study along north-south and east-west transects.
Figure 5-13: Central age +/-2SE versus Northing, for the eastern rift flank and Rwenzori (Bauer et al. 2010) AFT data

Figure 5-14: Age range of initial track preservation from thermal history modelling versus Northing, for the eastern rift flank and Rwenzori (Bauer et al. 2010) AFT data
The location of all samples is shown on the geological map in Figure 5-1. Figure 5-13 shows the central age ±2SE versus northing (UTM 36N) for the eastern rift flank data (this study) and the Rwenzori Mountains data (Bauer et al., 2010). This transect is orientated sub-parallel to the Western Rift. Since the Rwenzori data were collected further west than the rift flank data, the data is effectively projected onto the same transect. The central ages ±2SE for the Katonga Valley have a relatively large range from Carboniferous, through Permian, to Triassic. The central age for the Kyenjojo granite, sampled 50 km north of the Katonga Valley, is also Permian. It is unclear if the scatter in the Katonga Valley central ages is real or due to unquantified errors in the age calculation. Potential explanations include: real local differences in uplift and denudation due to faulting; differences in apatite composition not identified by Dpar; exaggerated differences in annealing behaviour due to complex thermal history; or, a combination of factors. The two samples collected in the Muzizi Valley area, about 75 km north of the Katonga Valley have younger central ages which are late Triassic and Jurassic.

The central ages for the central Rwenzori samples located further west (Figure 5-1) range between Jurassic and Cretaceous with the majority being of early Cretaceous age. The younger AFT age of the rocks from the Rwenzori Mountains in comparison to the rift flank indicates that they have been uplifted by a greater amount since the Mesozoic. The youngest AFT age for the Rwenzori Mountains occurs 40 km north of the main sampling area on the western edge of the low elevation ‘nose’ of the Rwenzori horst (Figure 5-1). However, the oldest AFT age for the Rwenzori samples, which is early Jurassic, is located only about 15 km further north on the eastern edge of the ‘nose’. Interestingly, this older age is comparable to those central ages corresponding to similar latitude on the eastern rift flank in the area of the Muzizi Valley. Caution should be adopted when making inferences on the basis of only three data points. It is possible that the similar ages reflect similar tectonic history. As shown in Figure 5-1, they all lie to the north of Buganda-Toro Formation and southern limit of the Albert Basin and associated border fault. Several WNW to ESE trending faults can also be seen to displace the Mubende Granite south-east of the Muzizi sample locations (GB03 and GB05). This raises the possibility that the area to the north has been up-thrown, exposing rocks, following denudation, with younger AFT ages. However, assuming
that the granite pluton increases in size with depth, the displacement shown on the geological map suggests that it has been down-thrown to the north.

As previously discussed, the protracted cooling history of most samples has likely led to an underestimate of the central ages relative to the start of spontaneous fission track preservation. Figure 5-14 therefore presents the age range of initial track preservation from thermal history modelling versus northing. These AFT ages plotted in Figure 5-14 are those presented in Table 5-4. For the Rwenzori data, the intermediate data point shown on the age ranges in Figure 5-14 is the point where the ‘best fit’ model enters the high temperature boundary of the PAZ, whereas the intermediate point on the rift flank age ranges is simply the central value. However, it is emphasised that interpretation of this data should focus on the age ranges, rather than the intermediate data point. As expected the maximum AFT ages predicted by the thermal history modelling shown in Figure 5-14 is older than the central ages shown in Figure 5-13. The thermal history modelling suggests that the rocks in the region of the Katonga Valley passed into the base of the PAZ during the Devonian to Carboniferous. This is consistent with exhumation prior to Lower Karoo deposition of the Bihanga diamictite. The revised age range for the majority of central Rwenzori samples is Jurassic. There is less difference in the revised age for the northern most sample (Muzizi granulite) and so there is a greater age range (Permian to Jurassic) for the northern samples.
Figure 5-15: Central age +/-2SE versus Easting, for the eastern rift flank and Rwenzori (Bauer et al. 2010) AFT data

Figure 5-16: Age range of initial track preservation from thermal history modelling versus Easting, for the eastern rift flank and Rwenzori (Bauer et al. 2010) AFT data
Figure 5-15 shows the central age ±2SE versus easting. The eastings for the Rwenzori sample locations are in the UTM 35N region, and so it was necessary to convert them to an equivalent UTM 36N easting so that they could be plotted with the data from the Katonga Valley on the eastern rift flank. The west-east transect is approximately perpendicular to the Western Rift and the Rwenzori horst and so this orientation provides a better representation of the influence of the rift and horst tectonics on the pattern of AFT ages than the north-south orientation. The discordant shift in the central age ±2SE between the rift flank and the horst block is consistent with the intervening fault movement. No AFT age is of course available for the bedrock underling the intervening downthrown George Basin, which may be equivalent to, or older than, the AFT ages determined for eastern rift flank. The northern samples can be seen to have central ages which are intermediate to those of the Rwenzori and Katonga Valley. The elevation of the Rwenzori samples generally increases towards to west (Figure 5-1). Although a linear regression can be fitted to the Rwenzori data which suggests an increase in AFT age towards the west the scatter in the data is large and the coefficient of determination ($r^2 = 0.22$) is small.

Figure 5-16 shows the age range of initial spontaneous track formation determined from thermal history modelling versus easting. The shift towards older ages compared to the central ages is again apparent. Whilst there is a clear difference between the AFT ages determined for the Rwenzori horst and the eastern rift flank, the pattern of AFT ages on the rift flank with distance from the rift boundary fault parallel to the Katonga Valley is ambiguous. If differential uplift has occurred on the rift flank perpendicular to the boundary fault at a scale consistent with the resolution of the AFT data (greater than about 500 to 1000 m) then this should be reflected in the Katonga Valley data. The overall difference in elevation between the base of the rift and the top of the highest peaks on the eastern flank is about 500 m and the relief on the fault boundary escarpment is about 200 m (Figure 2-4). Unfortunately the sample collected from near the base of the escarpment in the Mpanga Gorge (GB16) was unsuitable for AFT analysis and the relative relief between the AFT samples collected above the fault escarpment is 228 m. The pattern of AFT age perpendicular to the flanks of other tectonic rifts has previously been used to infer the mechanism of rift flank uplift.
It can be seen from Figure 5-15 that a linear regression can be drawn through the Katonga Valley data with younger ages towards the rift boundary fault, as might be expected due to greater displacement closer to the fault. However, there is a wide scatter in the data and the coefficient of determination on the relationship between central age and easting is only 0.21. In addition, the central age measured for GB17 (Nkongoro granulite) sampled closer to the rift boundary fault is Carboniferous, which is similar to those samples taken furthest to the east. The intervening samples collected either side of the water divide at Bihanga (GB13 and GB14) are Permian to Triassic in age. This raises the possibility that the pattern of AFT ages has been influenced by denudation following flexure with a wavelength of about 40 km. The polynomial drawn through the middle age determined from thermal history modelling shown in Figure 5-16 does indeed show a minimum value close to the water divide, consistent with maximum uplift and denudation. The coefficient of determination on the polynomial fit is 0.71. However, given the overlap in the age ranges, little confidence can be placed in the significance of this apparent relationship. As previously discussed, the variation in the AFT ages for the Katonga Valley is likely associated with local tectonics or un-quantified errors.

5.7.4 Vertical distribution of the rift flank and Rwenzori AFT ages

Further information about the timing of rock uplift is often available by examining the relationship between AFT age and elevation. For example, Foster and Gleadow (1996) examined the thermochronology of the Proterozoic crystalline rocks adjacent to the eastern arm of the EARS of central Kenya (Appendix H). They showed that the AFT age data from the Samburu District on the eastern flank between the Kenya Rift and Anza Rift, and data from the Cherangani Hills on the western flank north of Eldoret can be combined if corrected due to intervening fault offsets, to produce the consistent profile of AFT ages versus elevation shown in Figure 5-17. The profile reveals three distinct intervals of about 1200 m, 600 m and 600 m height, over an elevation range of 1000 m asl to 3500 m asl, where there is little change in the AFT age. The central ages corresponding to these intervals are approximately 180 Ma, 115 Ma and 65 Ma.
Figure 5-17: Composite AFT central age versus elevation for central Kenya (after Foster and Gleadow, 1996)

Figure 5-18: Elevation versus central age for eastern rift flank and Rwenzori (Bauer et al, 2010) AFT data sets
As previously discussed, the assumption of length isotropy in the age equation tends to lead to an underestimate of the age at the start of spontaneous fission track preservation. Therefore thermal history modelling was used to calculate ages of ≥220 Ma (mid Triassic), 140-120 Ma (early Cretaceous) and 70-60 Ma (late Cretaceous to early Palaeogene) for the start of cooling through the PAZ which takes annealing into account. Rapid changes in age between these intervals occur at about 2,200 to 2,300 m asl and 2,800 m asl. Foster and Gleadow (1996) interpret the periods of concordant ages to be related to relatively rapid denudation separated by slower periods of denudation. The isochronous intervals are regionally consistent if it is assumed that they have been offset by faults of over 1 km displacement, up to 100 km east of the Kenya Rift.

A similar plot of central age versus elevation is shown in Figure 5-18 using the AFT data from the eastern rift-flank (this study) and the Rwenzori horst (Bauer et al., 2010). The rift-flank rocks samples were collected between about 1,150 and 1,430 m asl and have relatively old AFT ages between middle Palaeozoic and early Mesozoic. The Rwenzori rock samples were collected between 1,070 and 4,800 m asl and have younger AFT ages between middle and late Mesozoic. Figure 5-18 clearly illustrates that the rocks of the Rwenzori horst have experience significant rock uplift which has revealed rocks with younger AFT ages compared to those exposed on the adjacent rift-flank. However, there is no consistent pattern in the relationship between AFT age and elevation within the dataset acquired from the Rwenzori Mountains even though they cover a similar age range and larger elevation range to those sampled in central Kenya (Figure 5-17). Foster and Gleadow (1996) were only able to construct the age versus elevation relationship for central Kenya by including an estimated offset due to the regional faults that occur between the sample locations. Bauer et al (2010) also infer that differential rock uplift between adjacent fault blocks is the likely cause of the lack of consistent relationship between AFT age and elevation in the Rwenzori data (Figure 5-18). However, there is insufficient geological evidence to quantify the displacement and reconstruct a corrected profile in the manner of Foster and Gleadow (1996). Therefore, it is not currently possible to identify periods of slow and rapid exhumation with reasonable precision (10s Ma) for the rocks of the Rwenzori Mountains.
5.8 Conclusions

The conclusions of the AFT central age calculations are listed below.

- All but one sample taken from adjacent to the Katonga Valley (GB17, GB15, GB13, GB19, GB20 and GB22) have Carboniferous to Permian AFT central ages (256.6 ±13.5 Ma to 335.6±19.0 Ma).

- The Bihanga diamictite samples (GB13 and GB14) have Carboniferous and Triassic AFT central ages (273.2 ±13.4 Ma to 228.9 ±10.2 Ma). It is noted that the younger estimated age (GB14) is influenced by the smaller induced fission track count measured in as few as 4 grains.

- The sample collected approximately 50 km north of the Katonga Valley at Kyenjojo has a Permian AFT central age (279.6 ±20.5 Ma) similar to the Katonga Valley samples.

- The two samples collected approximately 75 km north of the Katonga Valley in the vicinity of the Muzizi Valley have late Triassic (203.5 ±8.9 Ma) and Jurassic (158.3 ±7.6 Ma) central ages.

The observed spontaneous fission track lengths of between 11.69 ±0.16 μm and 12.48 ±0.16 μm suggest track shortening of between 3.4 and 4.2 μm indicating protracted cooling with significant annealing. Several fission track length data sets are negatively skewed. This indicates that the assumption of length isotropy in the age equation is likely invalid and the central ages probably underestimate the age at which the rock entered the base of the PAZ and spontaneous fission track preservation began.

The thermal history modelling of the Katonga Valley and Kyenjojo samples, calibrated against both length and age data, suggests that the age range at which exhumation occurred, and the preservation of spontaneous fission tracks began, was Devonian to Carboniferous. This is consistent with exhumation prior to deposition of Lower Karoo sediments. The AFT age and length data for the Bihanga diamictite samples (GB13 and GB14) is also consistent with Lower Karoo (Carboniferous to Permian) surface temperatures during the time at which the geological evidence suggests these glaciogenic rocks were deposited. In addition, the thermal history modelling indicates that Mesozoic reheating is required up to temperatures equivalent to those of the
nominal PAZ to reproduce the observed track shortening. Given the modelled range of maximum reheating to between 80°C and 100°C and the likely historical geothermal gradients between 20°C/km and 30°C/km, the estimated maximum thickness of Karoo (Permian to Jurassic) cover sediments is 2 km to 4.5 km.

Unconstrained thermal history modelling of all samples (except the Muzizi granulite, GB03, which has a Jurassic AFT age) include time-temperature paths that are consistent with surface or near-surface temperatures in the late Palaeozoic, followed by early Mesozoic re-burial prior to final exhumation. However, the thermal history modelling is unable to determine if the rocks experienced other episodes of cooling and reheating during the Mesozoic similar to that indicated by the presence of Lower Karoo sediments containing apatite with evidence of reburial. The granulite sample collected from the Muzizi valley (GB03), approximately 75 km north of the Katonga Valley appears to have remained below the PAZ during deposition of the Karoo cover sediments in the Permian and Triassic, before experiencing cooling from the Jurassic onwards. The annealing required for matching the track length data for this sample within a shorter time period suggests that episodic cooling and heating is less likely than for those samples with older AFT ages.

Since the time-temperature paths for all samples must spend time within the nominal PAZ during the Mesozoic, the AFT data is inconsistent with the preservation of Mesozoic land surfaces as suggested by King (1962) and others. Most time-temperature paths exit the top of the nominal PAZ in the late Mesozoic, suggesting that the rocks which form the current land surface was likely still covered by up to 1 km or more at the start of the Cenozoic. However, since Cenozoic cooling occurs at temperatures lower than the nominal PAZ, the AFT analysis cannot determine the rate or timing of Cenozoic cooling.

The oldest AFT ages calculated for samples collected adjacent to the Katonga Valley occur furthest west near the rift boundary fault (GB17, 314.7±15.4 Ma) and furthest east (GB22, 335.6±19.0 Ma), and the youngest occurs near the water divide (GB14, 228.9±10.2 Ma). Whilst it is tempting to infer lithospheric flexure as the cause of this distribution, the relatively small number of samples and large overlap in the age ranges lends little confidence to this conclusion. The AFT ages derived for the eastern rift flank (Devonian to Carboniferous) are substantially older than the AFT ages recently
calculated for the central Rwenzori Mountains (Bauer et al., 2010) (mainly Jurassic). This is consistent with post Mesozoic rock uplift of the horst block revealing rocks with AFT ages younger than those determined for the adjacent rift flank. The regional history of burial and exhumation is discussed further in Section 8.2.
6 ELECTRICAL RESISTIVITY TOMOGRAPHY – PHASES OF EROSION AND DEPOSITION IN THE KATONGA VALLEY

6.1 Introduction

The objective of the electrical resistivity tomography described here is to assess the sub-surface geometry of the valley and fill perpendicular to the valley axis at representative locations. The geometry of the subsurface lithology provides clues as to the major phases of erosion and deposition in the valley. Given the practical requirement to be able to traverse the full width of the valley, including the papyrus wetland, it was necessary to select investigation sites with existing murrum road causeways across the valley. The crossing at Kabagole in the broad western Katonga Valley was selected for the initial 1D reconnaissance survey. The 1D survey results proved the efficacy of electrical resistivity methods for investigating the Katonga Valley fill. During the 1990s, 1D electrical resistivity survey methods were largely replaced by 2D electrical resistivity tomography (ERT) in high income countries. The ERT surveys presented in this chapter were conducted using up-to-date commercial ERT equipment in collaboration with Department of Geology at the University of Nairobi, during July 2009.

During the reconnaissance, additional Katonga crossings were visited at Nkonge, Kisozi and Bugomola. It was established that the narrow crossing at Nkonge was unsuitable for ERT due to the presence of prominent concrete culverts. The Bugomola crossing is in the wide reach close to Lake Victoria and therefore it seemed likely that the valley fill within the depth of investigation would be dominated by fine-grained lacustrine deposits rather than any deeper coarse-grained basal fluvial deposits. Therefore, the crossings at Kabagole and Kisozi were selected for the ERT (Figure 1-3). Following completion of these two surveys, a reconnaissance visit was made to the crossing at Kyai (Figure 1-3), where the sand and gravel terrace was identified overlying weathered phyllites. A third short survey was subsequently conducted on the causeway through the papyrus wetland in the valley bottom at Kyai.

The electrical resistivity survey theory and methods are documented in Appendix I and the results of the 1D electrical resistivity reconnaissance survey at Kabagole is
presented in Appendix J. This section describes the field procedures, analysis and interpretation of the ERT surveys at Kisozi, Kabagole and Kyai.

### 6.2 Field procedures

Almost 6 km of 2D resistivity surveys were carried out along four survey lines at three research sites as summarised in Table 6-1 and shown in Figures 6-1.

<table>
<thead>
<tr>
<th>Site</th>
<th>Length</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kabagole</td>
<td>3240 m</td>
<td>Reach C in Table 2-1. Same location as 1D reconnaissance survey in a wide straight section of western Katonga Valley</td>
</tr>
<tr>
<td>Kyai</td>
<td>360 m</td>
<td>Reach D in Table 2-1. Survey across papyrus wetland in the base of a narrow central Katonga Valley.</td>
</tr>
<tr>
<td>Kisozi</td>
<td></td>
<td>Reach F in Table 2-1. Two surveys where valley begins to widen in the eastern Katonga Valley.</td>
</tr>
<tr>
<td>North</td>
<td>a) 1440 m</td>
<td>a) Survey across papyrus wetland (borehole data available)</td>
</tr>
<tr>
<td>South</td>
<td>b) 900 m</td>
<td>b) Survey on valley slope between alluvium and laterite.</td>
</tr>
</tbody>
</table>

**Table 6-1: Summary of ERT survey locations**

The sites were selected to be amenable to the surveys and to be representative of the western, central, and eastern Katonga Valley (Table 2-1). The valley at Kabagole and Kisozi was also considered most likely to contain any potential basal aquifer sand and gravel. Two boreholes were identified adjacent to the papyrus wetland at Kisozi crossing, and following a month of enquiries whilst in Uganda, the borehole records were obtained from the Directorate of Water Development (DWD) in Kampala. The borehole logs provided 'ground truth' data and the pumping tests indicated the occurrence of a local groundwater resource. The site at Kyai was visited on the final day of the 2009 fieldwork. Following discovery of a sand and gravel terrace overlying weathered schist on the valley side it was decided to conduct a third short ERT survey across the causeway over the papyrus wetland in the base of the valley. The field procedures and analysis description were similar for the Kabagole, Kyai and Kisozi sites and are described together below. The interpretations of the inversion models produced for each site are considered in three separate sections.
Figure 6-1: Plan view of ERT surveys at: a) Kabagole; b) Kyai; and c) Kisozi

All surveys were carried out with four fieldworkers, including; myself, a geophysicist from Nairobi University Geology Department, a hydrogeologist from the Uganda Directorate of Water Resource Management (DWRM) and a fieldwork assistant from Makerere University. The field procedure began at each site by laying out the four cables and hammering each of the 72 steel electrodes into the ground. The GPS coordinates were recorded for each electrode. The outside cables were connected together and inner two cables were connected to the Syskal R1 Plus automated resistance meter placed in the centre of the array. An initial test was then conducted to check the electrical conduction between each adjacent pair of electrodes. Electrodes with unusually high resistance were assumed to have a poor connection with the ground. In such circumstances the connection was improved by moving the electrode, hammering it more deeply into the ground and pouring water onto the electrode and adjacent ground. Given the difficulties with obtaining a good connection in the often hard dry murrum beside the road or lateritic soils, it became standard practice to pour water over each electrode. Once the ground connections were satisfactory, the pre-programmed switching sequence was run to inject the current and measure the potential differences.

The switching sequence was pre-programmed by Nairobi Geology Department using the Electre II software provided by Iris Instruments and uploaded to the Syscal R1 Plus. The switching sequence was based on a Schlumberger array, with the current electrode spacing increased in 10 m intervals between 15 m and 65 m, 20 m intervals between 65 m and 165 m, 20 m intervals again between 180 m and 260 m, with a return to 10 m
intervals between 265 m and 355 m. The electrical potential electrode spacing \((a)\) was set at 5 m for current electrode spacing between 15 m to 165 m, 10 m between 180 and 260 m, 15 m between 265 and 300 m, and 30 m between 305 and 355 m. The range of the median depths of investigation with continuous coverage is 2.7 m to 49.9 m. Due to the decreasing number of potential measurement points along the fixed cable length with increasing current electrode spacing, below 50 m pseudo-depth the data forms inverted triangles with the apex at the maximum theoretical depth of investigation of 68.2 m (see Appendix I.1.3).

The main switching sequence was used to inject current and measure the potential difference for all combinations of electrode spacing for the expanding Schlumberger array. This took about 1.5 hours. Once the first recording sequence was complete, one 90 m cable (18 electrodes) was moved from the trailing edge of the survey to the advancing edge of the survey and the automatic resistance meter was moved to the next central point. Since only one cable was moved, the next roll-along switching sequence is only required to measure a subset of the electrode combinations in order to maintain complete coverage along the survey section. This took about 40 minutes to complete each time. The roll-along procedure was repeated as many times as was necessary to complete the profile. The data was downloaded to the field computer each evening and a field analysis was undertaken to check the data quality.

6.3 Analysis Procedure

6.3.1 Error assessment

As can be seen in Figure 6-1, due to practical constraints, it was not possible to conduct ideal linear surveys. Prior to conducting the analysis, an estimate was made of the potential error due to curvature in the survey lines. The ratio between the linear distance and the distance along the survey line was used to estimate the potential error as described for the 1D electrical resistivity profiling in appendix Section J.1.2. The potential error is summarised in Table 6-2. The lower estimated percentage error assumes both the current and potential electrode spacing are reduced proportionally due to the line curvature, whilst the higher estimated error assumes only current electrode spacing is reduced. Since the survey line generally followed a smooth curve beside the
murrum road it is likely that the low estimated error is applicable. Overall it may therefore be considered conservative to assume the maximum error is applicable. It should be noted that the error will change depending on the change in curvature along the survey line. The interpretation of specific areas of the apparent resistivity pseudo-sections should therefore consider the linearity of the survey line at that location.

<table>
<thead>
<tr>
<th>Site</th>
<th>Survey Distance (m)</th>
<th>Linear Distance (m)</th>
<th>Distance Difference (%)</th>
<th>Resistivity Error$^1$ (%)</th>
<th>Maximum Errors $\rho_a=10, 100, 1000\Omega m$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kabagole</td>
<td>3240</td>
<td>3067</td>
<td>94.7</td>
<td>5.3 to 11.6</td>
<td>1.2, 10.2, 102</td>
</tr>
<tr>
<td>Kyai</td>
<td>360</td>
<td>327</td>
<td>90.1</td>
<td>9.9 to 21.1</td>
<td>2.1, 21.1, 211</td>
</tr>
<tr>
<td>Kisozi</td>
<td>a) 1440</td>
<td>1407</td>
<td>97.7</td>
<td>2.3 to 5.1</td>
<td>0.51, 5.1, 51</td>
</tr>
<tr>
<td>North</td>
<td>b) 900</td>
<td>869</td>
<td>96.6</td>
<td>3.4 to 7.5</td>
<td>0.75, 7.5, 75</td>
</tr>
<tr>
<td>South</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Note: $^1$ Low error assumes $L$ and $a$ reduced proportionally, high error assumes only $L$ reduced

Table 6-2: Potential errors in calculated apparent resistivity due to line curvature

It can be seen in Table 6-2 that the magnitude of the maximum potential error is about 5% for Kisozi-North, 7.5% for Kisozi-South, 10% for Kabagole and 20% for Kyai. Given that the apparent resistivity varies over a log scale whilst the error due to the line curvature varies linearly, the calculated potential errors do not undermine the objectives of the ERT. Nevertheless, caution is required not to over-interpret the apparent resistivity variations within the potential error at a given order of magnitude.

6.3.2 Inversion modelling

Due to the available borehole ground-truth data, the model calibration and sensitivity analyses were focussed on the Kisozi-North ERT survey. The initial setting for the inversion modelling software, RES2DINV (see appendix Section I.3.3), were modified until the model was consistent with existing knowledge of the surface and borehole geology before conducting a comprehensive sensitivity analysis to determine the apparent range of subsurface interpretations. A total of 59 inversions were conducted for the Kisozi-North survey. The final preferred modelling options are presented in Table 6-3.
<table>
<thead>
<tr>
<th>Parameter Group</th>
<th>Settings</th>
<th>Preferred settings</th>
</tr>
</thead>
<tbody>
<tr>
<td>Inversion methods</td>
<td>Initial model</td>
<td>Homogenous</td>
</tr>
<tr>
<td></td>
<td>Data inversion constraint</td>
<td>Robust</td>
</tr>
<tr>
<td></td>
<td>Model inversion constraint</td>
<td>Smooth and blocky</td>
</tr>
<tr>
<td></td>
<td>Jacobian matrix</td>
<td>Optimised, fast, Jacobian calculation</td>
</tr>
<tr>
<td></td>
<td>Inversion calculation</td>
<td>Incomplete Gauss-Newton</td>
</tr>
<tr>
<td></td>
<td>Resistivity limits</td>
<td>No</td>
</tr>
<tr>
<td>Discretisation</td>
<td>Layer parameters</td>
<td>First layer half electrode spacing, increases by 10% per layer, no increase in model depth</td>
</tr>
<tr>
<td></td>
<td>Block parameters</td>
<td>All same width, not extended beneath survey sides, do not exceed data points</td>
</tr>
<tr>
<td>Damping factors</td>
<td>Vertical flatness ratio</td>
<td>Flatness ratio = 0.5 (horizontal bias)</td>
</tr>
<tr>
<td></td>
<td>Damping factors</td>
<td>Manual, initial factor = 0.3, minimum factor = 0.1, depth increase factor = 2</td>
</tr>
<tr>
<td></td>
<td>Resistivity limits</td>
<td>No</td>
</tr>
<tr>
<td>Progress settings</td>
<td>Convergence criteria</td>
<td>RMSE = 5%, RMSE change = 1%</td>
</tr>
<tr>
<td></td>
<td>Maximum iterations</td>
<td>10</td>
</tr>
<tr>
<td>Forward modelling</td>
<td>Mesh parameters</td>
<td>Finite element</td>
</tr>
<tr>
<td>Topographic model</td>
<td>Damping factor</td>
<td>Medium damping factor = 0.75</td>
</tr>
</tbody>
</table>

**Table 6-3: Preferred model settings used for analysis of Kabagole, Kyai and Kisozi ERT**

Once the preferred modelling routines were selected for the Kisozi-North model, the same options were selected for the other surveys to ensure consistency and facilitate comparison between surveys. Up to 10 additional sensitivity analyses were conducted for the other sites to ensure any site-specific artefacts did not influence the overall interpretation. A consistent contour interval for displaying the model output was selected after examining and optimising the selected interval for all sites simultaneously. Two alternative versions of the final models were produced, one using the smooth model inversion constraint and another using the blocky model inversion constraint.

Given some noise in the data, the length of the survey lines, and the complexity of the subsurface geology it was difficult to achieve the absolute RMS error criterion of 5%.
The noise is likely due to the hard dry ground beside the murrum road which occasionally produced difficulties in achieving a good electrical contact. In most cases the iteration cycles stopped when the change in the RMS error between successive iterations was less than 1%. The standard data inversion constraint often produced absolute RMS errors of greater than 10%, and therefore the robust data inversion constraint was applied to achieve RMS errors between 5 and 7% as presented in Table 6-4. These errors are considered acceptable provided that care is taken not to over interpret detailed variations in the final models.

<table>
<thead>
<tr>
<th>ERT Survey Site</th>
<th>RMS Error (%)</th>
<th>Smooth Model</th>
<th>Blocky Model</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kisozi-North</td>
<td>6.4</td>
<td>6.5</td>
<td></td>
</tr>
<tr>
<td>Kisozi-South</td>
<td>6.3</td>
<td>6.8</td>
<td></td>
</tr>
<tr>
<td>Kyai</td>
<td>5.9</td>
<td>6.6</td>
<td></td>
</tr>
<tr>
<td>Kabagole</td>
<td>5.7</td>
<td>5.5</td>
<td></td>
</tr>
</tbody>
</table>

Table 6-4: RMS errors for smooth and blocky models of Kisozi, Kyai and Kabagole ERT

Due to the length of the ERT surveys and the large number of data points (up to 15768), the optimised, fast Jacobian matrix and incomplete Gauss-Newton calculation were used facilitate a reasonable processing time of 10s of minutes. Flatness ratios of 1 or more produced vertical features which were difficult to reconcile with the prior conceptual understanding of the geology. A flatness ratio of 0.5 which biased the subsurface geometry to horizontal features produce an inversion consistent with the expected geometry of the fine-grained channel fill beneath the papyrus wetland in the Kisozi-North survey and the zone of expected laterite soils beneath the hill in the Kisozi-South survey. The sensitivity analysis examined the relationship between the damping factors and the RMS error. Increasing the damping factor reduced the apparent anomalous artefacts in the inversion due potentially to noisy data, but also increased the RMS error. Decreasing the damping factors reduced the RMS error, but accentuated potential artefacts. A compromise was achieved between RMS errors no greater than 7% and the damping factors shown in Table 6-3. No limits were placed on the range of estimated resistivity. The finite element method was selected for forward modelling of the inversion response to calculate the apparent resistivity. A moderate
damping factor of 0.75 was used to model the influence of topography on the resistivity with depth.

The results of the ERT inversion modelling are presented in Figures 6-3 to 6-5 (Kisozi-North), 6-7 to 6-9 (Kisozi-South), 6-11 to 6-13 (Kabagole) and 6-15 to 6-17 (Kyai). Consistent contour intervals and colour shading have been selected to facilitate comparison between the ERT results for different locations. A user defined logarithmic contour scale was used with the minimum contour set at 10 Ωm, and the contour increase factor set at 1.47 (6 contours per decade). Those figures which present three pseudo-sections together show: (a) the contoured measured apparent resistivity at the top; (b) the forward modelled apparent resistivity in the middle; and (c) the inverse modelled resistivity at the bottom of the page. The cell geometry, rather than contouring is shown for the resistivity model. The first two figures for each ERT survey show the results of the modelling for the smooth and blocky model inversion constraints respectively. The third figure in each series presents the inversions for both the smooth (a) and the blocky (b) models with the contoured results and topography. The horizontal and vertical scales are those selected automatically by the software to optimise the fit to the display. Therefore, the scales are not the same for each site.

6.4 Kisozi-North ERT Lithological Interpretation

Figure 6-2 shows two photographs of the road beside which the Kisozi-North ERT survey was conducted, including: a) looking south west over the raised flood plain; and, b) looking north east across the murrum causeway. Figures 6-3 and 6-4 present the results of the ERT model with the smooth and blocky inversion constraints respectively for the Kisozi-North survey across the papyrus wetland. Figure 6-5 presents the inversions for both the smooth and the blocky models with the contoured results and the topography. Given the large number of blocks used in the 2D inversion model compared the number of layers used in the 1D model discussed in Appendix J, the difference between the apparent resistivity and resistivity models is more subtle for the 2D models. The overall effect of the inversion is to discretise the resistivity and produce more distinct boundaries compared to the ‘smeread’ values and gradational boundaries of the apparent resistivity data and model.
Figure 6-2: Location of the ERT survey at Kisozi-North a) on the fluviolacustrine terrace and b) on the causeway across the papyrus wetland

Interpretation of the inverse models should be conducted in consideration of other sources of geological information. These include geological maps, literature review, field observations and borehole logs. As discussed in Section 4.5.4, two boreholes were identified adjacent to the south side of the papyrus wetland at Kisozi crossing, named Kikuumaddungu boreholes 2 and 3, after the nearest village to the south. Kikuumaddungu borehole 2 is located approximately 100 m east of the ERT survey and Kikuumaddungu borehole 3 is located approximately 350 west of the ERT survey. The lithology is generally described 3 to 6 m of ‘clay loam’ overlying sands and clays with ‘quartz’ horizons, resting on weathered rock and phyllites at about 55 m bgl. The description of ‘quartz’ in the logs likely refers to the presence of quartz gravel.

The geomorphology in the vicinity of the boreholes and the presence of clay and sand with rare gravel suggests possible lacustrine and fluvial overbank environments of deposition with occasional input from a higher energy fluvial or beach environment. Silts are not recorded but are likely present. The surface soil characteristics and topography suggest that original lake-arm and later River Katonga flood plain extended to about 2 km south of the papyrus wetland. The boreholes indicate that the alluvium is about 55 m deep adjacent to the papyrus wetland. The Kisozi-South ERT investigates the alluvium/weathered rock boundary on southern edge of the terrace flood plain.
Figure 6-3: Kisozi-North smooth inversion: a) measured $\rho_a$; b) modelled $\rho_a$; c) modelled $\rho$

Figure 6-4: Kisozi-North blocky inversion: a) measured $\rho_a$; b) modelled $\rho_a$; c) modelled $\rho$
Figure 6-5: Kisozi-North a) smooth and b) blocky resistivity inversion models with topography
Weathered rock is noted to 60 m in Borehole 2. In Borehole 3, the description of ‘phyllites’ from 56 m bgl to 72 m bgl, overlying ‘granite’ to 84 m bgl suggests that the borehole may have been drilled through the Katonga Line, and the Gneissic-Granulitic Complex may underlie the Buganda-Toro System at this location. However, given the poor quality of other lithological descriptions in the log this conclusion is tentative.

Given that the depth of the ERT investigation with continuous coverage is about 50 m and the maximum depth is about 65 m it seems unlikely that the ‘competent bedrock’ observed below 60 m bgl in Borehole 2 will be detected by the ERT survey.

Based on the maps, field observations and borehole logs, the following geological information was used to guide the ERT interpretation:

- Singo Series quartz sandstone outcrops north east of the papyrus wetland and exposed on Kisozi Hill;
- The papyrus wetland is likely directly underlain by very fine-grained (silty clay) paludal deposits;
- Silty sand with occasional thin gravel layers underlie the area south of the papyrus wetland up to about 55 m bgl; and
- Weathered rock likely underlies the alluvium below 55 m bgl. This may comprise weathered rocks of the Buganda-Toro System or the Gneissic-Granulitic Complex.

Comparison of the smooth and blocky models shown in Figure 6-5 reveals some of the ambiguity in the modelled resistivity. It is important to remember that the survey configuration only provides continuous coverage to a pseudo-depth of about 50 m bgl. Below this depth the data points become increasingly sparse. The effect of this appears greater in the blocky model (Figure 6-5b) where the contours below 50 m bgl are vertical and no horizontal features can be differentiated. The smooth model (Figure 6-5a) has captured the slight increase in apparent resistivity measured at depth to the south of the survey, shown in Figures 6-3a and 6-4a. Nevertheless, given the reduced coverage and increased uncertainty in the inversion models between 50 and 68 m bgl, in hindsight, it would probably have been preferable to select a maximum pseudo-depth of 50 m bgl when displaying the results.
Table 6-5 summarises the main features of the inversions shown in Figure 6-5, including the lithological interpretation. The model is divided into regions with similar resistivity characteristics and the lithological interpretation is made based on the prior knowledge of the geology listed above and the comparison with the resistivity range of common rock types presented in appendix Figure I-1. The distance along the ERT survey line and the pseudo-depth are given for each region to assist identification.

<table>
<thead>
<tr>
<th>Region of model</th>
<th>Distance (m) / depth (m bgl)</th>
<th>Resistivity range (Ωm)</th>
<th>Lithological interpretation and comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>North (brown-purple)</td>
<td>0-500/0-65</td>
<td>320 to 1496 (high)</td>
<td>Singo Series quartz sandstone. Unsaturated near surface. Artefacts due to poor contacts.</td>
</tr>
<tr>
<td>Middle shallow and south - shallow (blue)</td>
<td>500-960/0-25, 960-1435/0-5</td>
<td>22 to 69 (low)</td>
<td>Silty clay in channel beneath papyrus wetland and shallow layer beneath flood plain to the south.</td>
</tr>
<tr>
<td>Middle and south - deep (yellow-green)</td>
<td>500-960/25-65, 960-1435/5-65</td>
<td>69 to 321 (medium)</td>
<td>Silty sand beneath channel fill and flood plain. Artefact due to possible man-made feature at 1000 m along survey.</td>
</tr>
</tbody>
</table>

**Table 6-5: Summary interpretation of Kisozi-North ERT survey**

In general, it can be seen that to the north of the papyrus wetland the Kisozi-North ERT survey is underlain by a high resistivity region (the brown-purple region in Figure 6-5) which likely coincides with the Neoproterozoic Singo Series quartz sandstones. A distinctive channel with a maximum depth of about 25 m can be seen beneath the papyrus wetland. The channel fill has a low resistivity (blue) and is interpreted to be silty clay paludal deposits. The low resistivity material forms a thin layer, no more than 5 m deep, over the south of the model. The region below the channel and to the south of the papyrus wetland is underlain by relatively uniform medium resistivity material (yellow-green). Based on the borehole logs and the resistivity range, this is likely dominated by silty sand.

The ground to the north of the papyrus wetland was generally very hard and dry, but a short intense rain shower improved the ground contact during the survey. The variable
quality of the ground contact may account for some of the resistivity variations in the northern region, but some variation may be due to genuine variation in lithology and weathering characteristics.

The near-surface blocks with highest resistivity to the north of the model are immediately underlain by blocks with low resistivity compared to the average in this region. To the south of the model, the near-surface blocks with lowest resistivity are immediately underlain by blocks with high resistivity compared to the average in this region. It seems likely that these underlying contrasting features are artefacts and unrepresentative of the actual ground conditions.

Immediately south of the papyrus wetland is a small high resistivity feature. This is in the location between the water supply boreholes near to the pump control house. It seems likely that this represents a man-made low resistivity feature such as a concrete culvert beneath the road containing water pipes. An anomalous inverse ‘shadow’ artefact can once again be seen underlying this interpreted man-made feature.

6.5 Kisozi-South ERT Lithological Interpretation

Figure 6-6 shows a photograph of the road beside which the Kisozi-South ERT survey was conducted where the topographic gradient increases at the southern margin of the apparent flood plain. Figures 6-7 and 6-8 present the results of the ERT model with the smooth and blocky inversion constraints respectively for the Kisozi-South survey. Figure 6-9 presents the inversions for both the smooth and blocky models with the contoured results and the topography. The Kisozi-South ERT location was selected to investigate the ground conditions at the southern edge of the low lying ground which is interpreted as a relict raised flood plain (terrace) of the River Katonga.
No borehole logs are available in the immediate vicinity of the Kisozi-South survey. However, based on the maps and field observations the following geological information was used to guide the ERT interpretation:

- Yellowish silty soils were observed covering the low lying ground to the north (Figure 6-6);
- Sands, silts and clays with occasional thin gravels of about 55 m thick underlie the low lying ground adjacent to the papyrus wetland 2 km to the north; and
- Reddish lateritic soils were observed from the base of the hill southward towards the high ground (Figure 6-6), suggesting underlying deeply weathered bedrock; and
- The geological map shows the entire ERT survey to be underlain by ‘*undifferentiated gneisses, and some granites, including migmatised Buganda Series*’ (D.G.S.M., 1962).

Table 6-6 summarises the main features of the inversions shown in Figure 6-9. The shallow low resistivity layer (blue) above the low lying ground to the north of the Kisozi-South ERT, appears to be a continuation of the 5 m deep low resistivity layer on the south side of the Kisozi-North ERT survey (Figure 6-5).
Figure 6-7: Kisozi-South smooth inversion: a) measured $\rho_a$; b) modelled $\rho_a$; c) modelled $\rho$

Figure 6-8: Kisozi-South blocky inversion: a) measured $\rho_a$; b) modelled $\rho_a$; c) modelled $\rho$
Figure 6-9: Kisozi-South a) smooth and b) blocky resistivity inversion models with topography
### Table 6-6: Summary interpretation of Kisozi-South ERT survey

Beneath the hill to the south of the Kisozi-South ERT survey is a region of variable low to medium resistivity (yellow-brown) of up to 30 m thick. Some of this variability may be due to noise created by the difficulty in obtaining good electrical contacts in the hard dry ground. Nevertheless, it is likely that some variation is a genuine response to the ground conditions and is likely associated with the presence of deeply weathered bedrock, suggested by the reddish lateritic soils at the surface.

The deeper part of the ERT survey below an elevation of about 1160 m asl (Figure 6-9) is characterised by a region of relatively uniform medium resistivity of between 148 and 471 Ωm (brown). No horizontal change in resistivity can be seen at depth beneath the base of the hill at about 450 m on the survey line. On the basis of the geological map and the red lateritic soils it is likely that weathered gneisses continue at depth beneath the higher ground. However, it is not known if the weathered bedrock continues beneath the lower lying ground to the north or is replaced by silty sandy alluvium of a similar resistivity which underlies the southern region of the Kisozi-North ERT survey.
6.6 Kabagole ERT Lithological Interpretation

Figure 6-10 shows a photograph of the route of the ERT survey across the wide western Katonga Valley at Kabagole. The photograph foreshortens the survey route and makes it look more curved than the route in plan-view shown in Figure 6-1. The interpretation of the Kabagole ERT survey was constrained by several sources of geological information, including maps, field observations, a borehole log and auger logs. The geological map shows the Mubende Granite outcropping on either side of the ‘swamp deposits and alluvium’ in the Katonga Valley. The author’s observations of the Mubende Granite outcrop on the northern slope of the valley reveal it to be a grey and white, coarse-grained, granite with large pink phenocrysts of orthoclase feldspar. In contradiction to the geological map, the author’s observations of the outcrops on the south slope of the Katonga Valley reveal it to be highly weathered gneiss, now decomposed to white and red-brown clay with sand sized grains, and occasional slightly weathered foliated quartzite bands.

![Figure 6-10: Location the resistivity profile and ERT at Kabagole crossing](image)

An outcrop on the northern slope of a topographic mound in the centre of the Katonga Valley was observed to be composed of reddish yellow to light grey-brown, thinly
bedded to laminated, moderately weak to strong, fine-grained sandstone. This is not recorded on the geological map and was described in Section 3.4.6. In thin section the sandstone is described as a grain supported, angular to sub-angular, moderately sorted, fine to very fine-grained arkose. The literature review revealed that the outcrop of arkosic sandstone at Kabagole is similar to outcrops at Bihanga Station in the Katonga Valley and other outcrops further north in the Muzizi and Nkusi Valleys. One borehole log was available from a water supply well drilled in the Katonga Wildlife Reserve. This borehole is located 850 m west of the ERT survey on the north side of the papyrus wetland. The notes on the driller’s log are reproduced in Table 6-7 and appear to suggest that the Katonga Wildlife Reserve borehole was drilled into the Mubende Granite to a depth of 113 m with a zone of weathered rock between 30 and 52 m. It is unclear from the log whether the zone above 30 m depth is also decomposed weathered granite, disintegrated weathered sandstone, or alluvium.

<table>
<thead>
<tr>
<th>Depth (m bgl)</th>
<th>Notes from driller’s log</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 to 1</td>
<td>‘Black loamy topsoil’</td>
</tr>
<tr>
<td>1 to 4</td>
<td>‘Brown moist clay with gravel’</td>
</tr>
<tr>
<td>4 to 30</td>
<td>‘Reddish sandy clay’</td>
</tr>
<tr>
<td>30 to 52</td>
<td>‘Brown pice (sic) weathered rock’</td>
</tr>
<tr>
<td>52 to 55</td>
<td>‘Grey unweathered fractured granite’</td>
</tr>
<tr>
<td>55 to 100</td>
<td>‘Grey unweathered hard granite’</td>
</tr>
<tr>
<td>100 to 106</td>
<td>‘Grey unweathered fractured hard granite’</td>
</tr>
<tr>
<td>106 to 112.89</td>
<td>‘Grey unweathered hard granite’</td>
</tr>
</tbody>
</table>

Table 6-7: Summary of driller’s log from Katonga Wildlife Reserve borehole

Observations of an excavation for a cattle watering hole adjacent to the south side of the papyrus swamp, 2.3 km east of the Kabagole ERT survey, revealed the near surface sediment to be composed of moist, grey-brown mottle orange, firm, silty clay with mica and occasional fine to coarse quartz sand. As discussed in Section 4.5.4, an investigation for construction materials for the Kilembe mine railway which was constructed through the Katonga Valley was conducted 1.6 km east of the Kabagole ERT survey on the south side of the papyrus wetland in 1952. The auger logs reveal a
general increase in grain size from the ground surface to between 2 and 3 m bgl from clay, to clayey sand, to fine to medium-grained sand (Waldron, 1952).

Figures 6-11 and 6-12 present the results of the ERT model with the smooth and blocky inversion constraints respectively for the Kabagole ERT survey. Figure 6-13 presents the inversions for both the smooth and the blocky models with the contoured results and the topography. Table 6-8 summarises the main features of the inversions shown in Figure 6-13. The Mubende Granite can be seen to form a region of high resistivity (brown-purple) to the north of the Kabagole ERT survey. Lower resistivity near the surface suggests that weathering may extend to a depth of around 15 m. A medium to high resistivity region (yellow-red) underlies the southern 350 m of the model. This is interpreted to be the weathered gneiss.

The arkosic sandstone forms a large region of uniform resistivity (yellow) in the centre of the model. The sandstone underlies 1 km of the 2.5 km wide valley bottom, and appears to be overlain by a layer of lower resistivity (blue) fine-grained sediment up to 15 m deep over much of its extent. The location where the medium resistivity (yellow) region of the model intercepts the surface coincides with the location of the sandstone outcrop identified in the field.

Very low resistivity regions (black-dark blue) occur to a depth of up to 30 m beneath the low lying grazing land and papyrus wetland either side of the sandstone and appear to be underlain by low resistivity regions (light blue). The auger logs recorded 1.6 km to the east of the ERT survey suggested that shallow sediments in this region may include sands below 2 to 3 m bgl. However, the resistivities in this region are between <7 and 69 Ωm (black-blue) which is less than those of between 69 and 321 Ωm (yellow-green) recorded on the Kisozi-North ERT survey close to the Kikuumaddungu boreholes. Since these borehole logs record generally sandy soils with clay, it is likely that the lower resistivities (black-blue) at Kabagole indicate even finer grained soils with predominantly clays and silts. If the unconsolidated valley fill at Kabagole was sand then it might be expected to have a similar medium resistivity (yellow) to the sandstone underlying the central region of the model. Nevertheless, it is acknowledged that there is significant uncertainty remaining as to the character of the sediments beneath the papyrus wetland in the southern half of the section shown in Figure 6-13.
It is possible that the resistivity range from 50 to 75 Ωm (light blue) also includes sand or sandstone.

Figure 6-11: Kabagole smooth inversion: a) measured $\rho_a$; b) modelled $\rho_a$; c) modelled $\rho$

Figure 6-12: Kabagole blocky inversion: a) measured $\rho_a$; b) modelled $\rho_a$; c) modelled $\rho$
Figure 6-13: Kabagole a) smooth and b) blocky resistivity inversion models with topography
### Table 6-8: Summary interpretation of Kabagole ERT survey

Below about 40 m pseudo-depth the contacts between the bedrock and the fine-grained alluvium appear to be very steep, and this is particularly true of the blocky model. However, it should be remembered that the base of the continuous coverage provided by the electrode array is about 50 m pseudo-depth. In hindsight, it appears that the increase in the damping factor with depth may be too large. The smooth model suggests an area of medium resistivity (yellow) beneath 50 m depth in the centre of the southern channel, but the reliability of the inversion models at this depth should therefore be treated with caution.

Some of the resistivity variation in the upper 10 to 15 m of the model is likely to be due to noise caused by variable quality electrical contacts with the ground. However, some variation may indicate weathering of the bedrock surfaces and variability in the composition of the alluvium.

#### 6.7 Kyai ERT Lithological Interpretation

Figure 6-14 shows a photograph of the route of the ERT survey across the narrow central Katonga Valley at Kyai. Figures 6-15 and 6-16 present the results of the ERT
model with the smooth and blocky inversion constraints respectively for the Kyai ERT survey. Figure 6-17 presents the inversions for both the smooth and the blocky models with the contoured results and the topography.

Figure 6-14: Location of the ERT survey at Kyai crossing

The Kyai ERT survey was not originally planned and was conducted on the final day of the fieldwork after discovering terraces composed of coarse-grained sediment on the side slopes of the central Katonga Valley which is only about 0.5 km wide at Kyai. As discussed in Section 4.5.4 the terraces are generally composed of light grey, weakly to moderately cemented, sub-angular, coarse sand to medium gravel with coarse gravel and cobble layers. The terraces rest directly on weathered phyllite. They occur on both the north and south slopes and the coarse sediments appear to have once filled the valley. In the base of the valley, the phyllite is overlain by silty soils and a murrum causeway crosses the papyrus wetland which is only about 200 m wide.

Table 6-9 summarises the main features of the inversions shown in Figure 6-17. Little horizontal variation can be seen in the model and therefore it has only been categorised into a low resistivity (blue) shallow region (<20 m) and a medium resistivity (yellow-brown) deeper region (20 to 65 m). Since weathered phyllite was observed immediately south of the survey, the very low resistivity region (dark blue) to the south is likely weathered phyllite. However, it appears to have a similar resistivity to the paludal silts and clays which likely underlie the papyrus wetland in the centre of the model.
There is little information upon which to base the interpretation of the underlying medium resistivity (brown) layer. Given that highly weathered phyllite outcrop on the side of the valley and cannot be underlain by alluvium it seems likely that this medium resistivity layer may also be schists, but only moderately weathered.
Figure 6-17: Kyai a) smooth and b) blocky resistivity inversion models with topography

<table>
<thead>
<tr>
<th>Region of model</th>
<th>Description</th>
<th>Distance (m) / depth (m bgl)</th>
<th>Resistivity range (Ωm)</th>
<th>Lithological interpretation and comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shallow (blue)</td>
<td>0-355/0-20</td>
<td>47-101 (low)</td>
<td>Silty clay beneath papyrus wetland and highly weathered schist</td>
<td></td>
</tr>
<tr>
<td>Deep (yellow-brown)</td>
<td>0-355/20-65</td>
<td>101-321 (medium)</td>
<td>Likely moderately weathered schist</td>
<td></td>
</tr>
</tbody>
</table>

Table 6-9: Summary interpretation of Kyai ERT survey

The interpretation of the ERT survey of the Katonga Valley at Kyai is supported by field observations alone. The main conclusions that can be drawn from the ERT survey is that the low resistivity fine-grained deposits beneath the papyrus wetland are probably no more than 20 m thick and appear most likely to directly overlie weathered bedrock. The sand and gravel deposits that form terraces on the valley sides do not occur in the base of the valley.
6.8 Conclusions

6.8.1 Lithological interpretation

This chapter has presented the first known application of ERT in Uganda. A total of six kilometres of roll-along ERT surveys were conducted at Kabagole, Kyai and Kisozi, located in the western, central and eastern Katonga Valley respectively. The ERT surveys of the Katonga Valley have demonstrated the utility of ERT as a tool for geomorphologists and hydrogeologist to assess the variation of lithology in two dimensions.

The main conclusions which can be drawn from the Kisozi-North ERT with regard to the geometry of lithological variations are:

- There is a steep contact between the alluvial deposits beneath the northern edge of the papyrus wetland and the bedrock to the north;
- Medium resistivity alluvial silty sands of the order of 55 m thick extend from below the papyrus wetland to the north to below the raised flood plain to the south;
- A channel beneath the present location of the papyrus wetland, cut into the medium resistivity silty sand, contains low resistivity silty clay sediment; and
- A layer of low resistivity fine-grained sediment up to 6 m thick covers the raised flood plain to the south and overlies the medium resistivity alluvial silty sand.

The main conclusions which can be drawn from the Kisozi-South ERT with regard to the geometry of lithological variations are:

- Low resistivity fine-grained sediment up to 10 m thick covers the raised flood plain up to the break of slope at the edge of the higher ground;
- Variable resistivity, deeply weathered bedrock occurs beneath the higher ground immediately south of the flood plain; and
- Due to a low resistivity contrast between the weathered gneiss beneath the higher ground and the silty sandy alluvium identified near the papyrus wetland, the Kisozi-
South ERT survey has not been able to determine if the medium resistivity alluvium extends 2 km south to the edge of the raised flood plain.

The main conclusions which can be drawn from the ERT survey at Kabagole with regard to the geometry of lithological variations are:

- High resistivity crystalline rocks underlie the valley sides to the north and south;
- The higher resistivity of the Mubende Granite to the north suggests that it is less weathered at depth than the gneiss to the south of the valley;
- The valley contains a medium resistivity fine grained sandstone outlier, which outcrops in the central valley where it forms a topographic mound; and
- The papyrus wetland and the low lying grazing land to the north and south of the sandstone outcrop are underlain by over 50m of low resistivity fine-grained clay and sandy silt.

The main conclusion which can be drawn from the ERT survey at Kyai with regard to the geometry of lithological variations is that the low resistivity fine-grained deposits beneath the papyrus wetland are probably no more than 20 m thick and appear most likely to directly overlie weathered bedrock. The sand and gravel deposits that form terraces on the valley sides do not occur in the base of the valley.

6.8.2 Discussion of phases of erosion and deposition

A general sequence of erosion and deposition may be interpreted from the geometry of the resistivity and lithological variations at each of the study sites. The different reaches of the Katonga Valley appear to have been influenced by a different combination of erosion and deposition events. The proposed geomorphological history of each study site is described here from west to east, beginning with Kabagole and ending with Kisozi.

The Kabagole ERT survey reveals the following general sequence of events in the western Katonga Valley:

- A palaeovalley was cut into the Proterozoic crystalline rocks which outcrop on the valley sides to the north and south;
The palaeovalley was then filled with clastic sediment that subsequently underwent diagenesis and lithification thus suggesting a significant depth of burial;

- The buried landscape was subsequently exhumed, and the River Katonga has exploited the softer sedimentary rocks within the palaeovalley;

- The now relict valley of the River Katonga was subsequently filled with predominantly fine-grained paludal or lacustrine deposits.

The resistivity colour coding shown in Figures 6-13 gives the impression of an ‘island’ of (yellow) sandstone between two channels filled with (blue) silty clay. This geometry would require deep (10s metres) fluvial channels to have been eroded on both sides of the sandstone mound. Given that processes of fluvial erosion would likely result in channel migration, the sequence of events required to result in channels either side of a sandstone ‘island’ would seem unlikely. This lends support to the possibility that the area of medium resistivity (light blue) beneath the southern channel may be a continuation of the indurated sedimentary rocks, represented in outcrop by the fine-grained arkose. The true lithological succession beneath the papyrus wetland can only be confirmed by direct observation during a drilling investigation.

The Kyai ERT survey in combination with field observations reveals the following general sequence of erosion and deposition in the central Katonga Valley:

- The Katonga Valley was initially formed by fluvial erosion of the Precambrian crystalline bedrock;

- Sub-angular sand and gravel with dipping beds was deposited during high energy, short duration events and eventually came to fill the valley;

- The sand and gravel were subsequently eroded from centre of the valley, exposing bedrock at the base and leaving terraces high on the valley sides; and

- Fine-grained paludal sediments were deposited in the base of the valley.

The Kisozi ERT survey reveals the following general sequence of erosion and deposition in the eastern Katonga Valley:
A valley of 2.75 km wide, with a maximum depth of at least 55 m deep was initially eroded in the Precambrian crystalline bedrock;

Fluctuating lake levels produced a combination of fluvial and lacustrine deposition which filled the valley with silty sand and occasional gravel;

A subsequent fall in the lake level left the terrace flood plain above the river level and resulted in channel erosion; and

Rising lake levels once again created a sluggish paludal environment resulting in the channel being filled by fine-grained silt and clay beneath the present day papyrus wetland.

This chapter has largely considered the interpretation of each individual ERT surveys in isolation. Sections 8.3 and 8.4 presents a simultaneous interpretation of the ERT surveys from all three study sites in order to construct a conceptual synthesis of the major phases of erosion and deposition preserved in the Katonga Valley.
7 PUMPING TEST ANALYSES – WATER RESOURCE POTENTIAL OF THE FLUVIOLACUSTRINE SEDIMENT

7.1 Introduction

The original motivation for this thesis was to update our knowledge of the landscape evolution of the Katonga Valley in order to better understand the groundwater resource potential of the relict valley sediment. The discovery of late Palaeozoic sedimentary rocks in the western Katonga valley led to the inference of a more complex geological history than previously understood. The ERT surveys of the Neogene relict valley indicates that a bed rock high occurs in the central Katonga Valley at Kyai. The ERT and borehole logs at Kisozi indicate that the eastern Katonga Valley is filled with fluviolacustrine sediment comprised of silty sand with occasional gravel. The results of the ERT at Kabagole indicate that the western Katonga Valley, cut into the Palaeozoic palaeovalley, is filled with finer grained sediment (possibly clayey silt) than the fluviolacustrine sediment in the eastern valley. Although a hand pumped well was observed sunk through the Karoo sediments at Bihanga Station, it has not been possible to acquire the borehole log and pumping test data for this well. Pumping test data from Katonga borehole DWD25847 near Kabagole also fails to provide hydrogeological properties for the Karoo sediments. This section therefore focuses on the interpretation of pumping tests conducted in the fluviolacustrine sediment in the eastern Katonga Valley and around the north-west shore of Lake Victoria. So as to place the transmissivity of the fluviolacustrine sediments into the wider context, pumping test analyses have also been conducted on data from wells drilled in the deeply weathered regolith and fractured crystalline rocks of south west Uganda.

Appendix K provides a summary of the pumping test analysis theory employed during this study. Following this introduction, Section 7.2 gives an overview of the analysis approach adopted during this study. Section 7.3 describes the interpretation of pumping tests in weathered and fractured rock and Section 7.4 describes pumping tests in the fluviolacustrine sediment. Section 7.5 presents the conclusions. Further discussion of the hydrogeology of the fluviolacustrine hydrostratigraphic unit is presented in Section 8.7.
7.2 Analysis Procedure

The locations of all boreholes for which data have been analysed are shown in Figure 7-1. Data from Katonga borehole DWD25847 were obtained directly from Samadhura Tech Ltd (Figure 7-2b). Data for Kikuumaddungu boreholes 2 and 3 (Figure 7-2-a), located on President Yoweri Museveni’s ranch, were obtained from the Ugandan Directorate of Water Development (DWD). All other data were obtained from the Ugandan Directorate of Water Resource Management (DWRM) license application records for abstraction wells with an electronic submersible pump. Most data has not been analysed previously and it was necessary to input data from hard copy documents.

![Figure 7-1: Location of wells with available pumping test data](image)

Figure 7-1: Location of wells with available pumping test data
All tests were analysed using the type curve method (Kruseman and de Ridder, 1990) implemented in the commercially available AQTESOLV software. The detailed analysis theory and methods are described in Appendix K. Given that all data was obtained from pumping wells, and none from observation wells, it was not possible to obtain reliable estimates of storativity. The delayed yield response characteristic of release from unconfined storage was not observed in any test, and therefore, neither was it possible to estimate the specific yield. Although the shape of the drawdown and recovery during the transition from wellbore storage to the aquifer response is controlled by both the borehole skin and the aquifer storage coefficient, it is not possible to determine both values independently (see appendix Section K.4). In many cases the borehole skin is likely to be significant and therefore this study adopts the same approach used to analyse data from production wells in the petroleum industry. A consistent value of storativity is assumed for all tests and the corresponding borehole skin is calculated for individual wells. The assumed storativity value of 1E-4 (-) is based on the analyst's experience and the following scoping calculation:
\[ s = \phi C_t h \rho g = 0.2 \times 10^{-9} \times 50 \times 10 \times 1000 = 1 \times 10^{-4} \]  
\hspace{1cm} \text{Equation 7-1}

Where

\[ s \] – storativity (-)
\[ \phi \] – porosity (0.2)
\[ C_t \] – total compressibility (1E-9 Pa\(^{-1}\))
\[ h \] – interval length (50 m)
\[ \rho \] – water density (1000 kg/m\(^3\))
\[ g \] – acceleration due to gravity (10 m/s\(^2\))

In theory, the storativity could vary between about 5E-3 in highly porous compressible sediment and about 5E-5 in low compressibility fractured rock (Kruseman and de Ridder, 1990). However, given that there are examples of tests with both positive and negative skin, and the analysis of one particularly good quality dataset (Kagera borehole Reg. No. 307) gives a value zero skin when the storativity value is assumed to be 1E-4, this value seems reasonable. Nevertheless, it should be remembered that test results with apparent positive skin may also be indicative of storativity < 1E-4 and tests with negative skin may also be indicative of storativity > 1E-4. Due to these uncertainties, greater emphasis is placed on the reliability of transmissivity data derived from analysis of the pumping well data.

The analysis procedure adopted by this study using the AQTESOLV software is summarized below.

- Input the casing radius (\(r_c\)) and the well radius (\(r_w\)).
- Input all flow rates (L/min) and durations (min) recorded during the test.
- Input all times (min) and displacement (m) for both drawdown and recovery.
- View the displacement versus time log-log diagnostic plot for the drawdown data and derivative.
- Select the appropriate analytical model, which in all cases is the Dougherty and Babu (1984) solution for a pumping test in a confined aquifer with wellbore storage and skin.
- Set the storativity to an assumed value of 1E-4 (-) and adjust the transmissivity (m²/day) and skin (-) to find the best match to the drawdown data.

- View the displacement versus Agarwal equivalent time log-log diagnostic plot for the recovery data and derivative.

- If necessary make small adjustments the effective casing radius to change the effective wellbore storage to account for unrecorded time offsets in the data.

- Fine tune the transmissivity and skin to obtain the best overall match to both the drawdown and recovery data simultaneously.

- Where appropriate add a single constant head boundary to reproduce a rapid late-time decrease in the derivative observed in two data sets, and adjust to boundary distance to obtain the best fit.

7.3 Interpretation of Pumping tests in Weathered and Fractured rock

The available pumping tests in weathered and fractured crystalline rocks have been analysed to facilitate a comparative assessment of the transmissivity calculated for pumping tests in the fluviolacustrine sediments. The results of the pumping test analyses in weathered and fractured crystalline rocks are summarised in Table 7-1 and Figures 7-3 to 7-12.

Six of the ten test results listed in Table 7-1 show negative skin. This is consistent with a connection between the borehole and the wider formation through a high transmissivity feature, and could be explained by a higher than average transmissivity fracture connected to an average transmissivity fracture network. Boreholes with positive skin are connected to the wider formation through a low transmissivity feature and care should be taken when interpreting the transmissivity as it will be unrepresentative of the well yield.
<table>
<thead>
<tr>
<th>Borehole ID</th>
<th>Location (District)</th>
<th>Depth (m bgI)</th>
<th>Summary Lithology</th>
<th>Q (m³/hr)</th>
<th>T (m³/d)</th>
<th>Skin' (-)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>DWD 25847²</td>
<td>Katonga (Kyenjojo)</td>
<td>0 to 52</td>
<td>Weathered rock</td>
<td>0.55</td>
<td>0.22</td>
<td>-1.58</td>
<td>Possible vertical fracture flow. No radial flow stabilisation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>52 to 113</td>
<td>Fractured granite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DWD 16272</td>
<td>Busujja (Wakiso)</td>
<td>0 to 21</td>
<td>Weathered rock</td>
<td>11.92</td>
<td>7.83</td>
<td>-2.85</td>
<td>Poor flow rate control and no clear radial flow stabilisation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>21 to 91</td>
<td>Fractured gneiss</td>
<td>10.11</td>
<td>9.97</td>
<td></td>
<td></td>
</tr>
<tr>
<td>DWD 16275</td>
<td>Busujja (Wakiso)</td>
<td>0 to 36</td>
<td>Weathered rock</td>
<td>3.00</td>
<td>3.90</td>
<td>0.30</td>
<td>Good flow rate control and clear radial flow stabilisation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>36 to 70</td>
<td>Fractured gneiss</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Reg. No. 370</td>
<td>Kasangati (Wakiso)</td>
<td>0 to 43</td>
<td>Weathered rock</td>
<td>step 1: 3.00</td>
<td>5.88</td>
<td>0.00</td>
<td>Matched to two step test. No clear radial flow stabilisation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>43 to 50</td>
<td>Fractured gneiss</td>
<td>step 2: 5.22</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Reg. No. 371</td>
<td>Kasangati (Wakiso)</td>
<td>0 to 33</td>
<td>Weathered rock</td>
<td>step 1: 1.50</td>
<td>1.43</td>
<td>-2.4</td>
<td>Matched to three step test. No clear radial flow stabilisation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>33 to 50</td>
<td>Fractured gneiss</td>
<td>step 2: 3.00</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>step 3: 4.50</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DWD 7986</td>
<td>Kyadondo (Mpigi)</td>
<td>0 to 47</td>
<td>Weathered rock</td>
<td>1.60</td>
<td>2.91</td>
<td>1.13</td>
<td>Good flow rate control, but stopped before radial flow stabilisation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>37 to 87</td>
<td>Fractured gneiss</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DWD 15296</td>
<td>Lukuuku (Masaka)</td>
<td>0 to 37</td>
<td>Weathered rock</td>
<td>3.0 (31 hrs)</td>
<td>3.93</td>
<td>-3.53</td>
<td>Good flow rate control and clear radial flow stabilisation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>37 to 88</td>
<td>Fractured gneiss</td>
<td>4.5 (0.52 hrs)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DWD 16584</td>
<td>Nakatake (Masaka)</td>
<td>0 to 48</td>
<td>Weathered rock</td>
<td>10.59, 11.80</td>
<td>16.75</td>
<td>1.05</td>
<td>Flow rate fluctuations, but reasonable radial flow stabilisation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>48 to 103</td>
<td>Fractured gneiss</td>
<td>11.83, 11.43</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DWD 16561</td>
<td>Rwakishakizzi (Mbarara)</td>
<td>0 to 36</td>
<td>Weathered rock</td>
<td>2.04 to 2.11</td>
<td>1.59</td>
<td>-2.70</td>
<td>Reasonable flow rate control and radial flow stabilisation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>36 to 106</td>
<td>Fractured gneiss</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DWD 16277</td>
<td>Mpwege (Wakiso)</td>
<td>0 to 42</td>
<td>Clay, sand texture</td>
<td>5.95 to 6.00</td>
<td>2.94</td>
<td>-4.55</td>
<td>Reasonable flow rate control. Possible fracture flow then radial flow stabilisation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>42 to 61</td>
<td>Weathered gneiss</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Note: 1) Storativity assumed = 1E-4 and skin is allowed to vary. 2) Katonga well fitted with hand pump after pumping test, all others contain motorised pumps.

Table 7-1: Summary of pumping test analyses for boreholes mainly in weathered and fractured rock
7.3.1 Katonga borehole

Katonga borehole DWD25847 is the only well fitted with a hand pump after the pumping test. It was selected for analysis due to its location on the north side of the Katonga Valley in the Kyenjojo District about 850 m west of the Kabagole ERT survey (Figure 7-1). The driller’s lithological log is shown in Table 6-7. Soft ‘reddish sandy clay’ is recorded between 4 and 30 m bgl, but it is unclear if these drill cuttings were derived from the Quaternary alluvium, the Palaeozoic-Mesozoic sedimentary rocks identified in the western valley, or the weathered Proterozoic granite. Hard ‘brown pice (sic) weathered rock’ is shown between 30 and 52 m bgl, overlying unweathered fractured granite. A gravel pack is placed with plain and slotted screens between 33 and 64.2 m bgl, with open below to 112.89 m bgl. The inflow is recorded in the fractured rock between 53 and 102 m bgl.

During the pumping test, the relatively low flow rate of 0.55 m$^3$/hr produced a maximum drawdown of 38.41 m after only three hours. The log-log diagnostic plot in Figure 7-3a is suggestive of some fluctuation in the early-time flow rate. Nevertheless, the overall sub-parallel positive slopes on the displacement versus time and derivative which approaches 2:1 is characteristic of 1D flow geometry. This could be interpreted as a high transmissivity vertical fracture or channel flow on a horizontal fracture plane. In this case, it has been modelled using negative skin (-1.6). The 1D response is also
seen in the recovery data shown in Figure 7-3b. It is unclear if the sharp reduction in the gradient in the late-time drawdown data is due to a reduced pumping rate or is indicative of a transition to radial flow. No radial flow stabilisation is present in the data and the estimated 2D transmissivity of 0.22 m²/d is unreliable and perhaps represents a maximum likely value.

It is therefore concluded that this hand pumped, low yield well is drawing water from a vertical fracture or channelized flow and the maximum sustainable yield cannot be estimated on the basis of this pumping test. In order to determine if the near-wellbore fracture is connected to a transmissive fracture network or overlying transmissive regolith capable of providing a useful sustainable yield, it would be necessary to conduct a longer term test with a reduced flow rate.

### 7.3.2 Busujja boreholes

Busujja is located near the Hoima road in the Wakiso District, about 30 km north west of central Kampala (Figure 7-1). The pumping test in Busujja DWD16272 provides an example of a poorly controlled flow rate. Each change in flow rate recorded during the test has been entered into AQTESOLV and used to calculate the type curve, but the discrepancy between the type curve and the data in Figure 7-4 suggests that the rate may not always have been accurately recorded. It appears that the initial flow rate of approximately 12 m³/hr was too high and this has been reduced to approximately 10 m³/hr when the maximum drawdown of 28.7 m approached the pump inlet. The flow rate was changed mid-time during the onset of the radial flow stabilisation. The recovery data is anomalous, and although this may be partly attributed to the change in flow rate, it appears inconsistent with the drawdown data. Nevertheless, the vertical match on the derivative for both the drawdown and recover appears generally consistent and this provides some confidence in the estimated transmissivity of 7.8 m²/d.
Figure 7-4: Busujja DWD16272 (weathered / rock) log-log data and derivative type curve matches for a) drawdown and b) recovery

During the pumping test in Busujja borehole DWD16275 the flow rate of 3 m$^3$/hr was relatively well controlled as can be seen from the smooth displacement and derivative data curves in Figure 7-5a. The maximum drawdown after 72 hours was 25.6 m. A good match can be obtained between the drawdown data and type curve with a small positive skin of 0.3 and a transmissivity of 3.9 m$^2$/d (Figure 7-5a). However, the apparent start of the radial flow stabilisation on the derivative of the recovery data is lower than the drawdown data (Figure 7-5b). The recovery data is therefore matched with a larger positive skin of 6.1 and higher transmissivity of 5.9 m$^2$/d than the
drawdown data. The reason for this discrepancy is difficult to assess without further information. The parameters derived from the drawdown data are quoted in Table 7-1.

### 7.3.3 Kasangati boreholes

Kasangati is located near the Masindi road in the Wakiso District, about 12 km north of central Kampala (Figure 7-1). A two step pumping test was conducted in Kasangati Reg. No. 370 with flow rates of 3 m$^3$/hr and 5.2 m$^3$/hr corresponding to maximum drawdowns of 12.7 m and 25.3 m respectively. A three step pumping test was conducted in Kasangati Reg. No. 371 with flow rates of 1.5 m$^3$/hr, 3 m$^3$/hr and 4.5 m$^3$/hr corresponding to maximum drawdowns of 10.9 m, 22 m, and 39.3 m respectively. All pumping steps were 1.5 hours each.

AQTESOLV only facilitates analysis of the recovery as a separate flow period using Agarwal equivalent time, and therefore all drawdown steps must be analysed together. A log scale on the time axis is inappropriate when matching distinct steps of equal duration and therefore Figures 7-6a and 7-7a show the linear plots of displacement versus time. The recovery period is shown on the usual log-log diagnostic plot.

![Figure 7-6: Kasangati Reg. No. 370 (weathered regolith) a) linear plot for step drawdown data and b) log-log plot for recovery data](image-url)

Figure 7-6: Kasangati Reg. No. 370 (weathered regolith) a) linear plot for step drawdown data and b) log-log plot for recovery data
The pumping test analysis for Kasangati Reg. No. 370 shows a generally good match to both the drawdown and recovery with zero skin and a transmissivity of 5.9 m$^2$/d. However, there is a poor match to the end of the second drawdown period. The discrete increase in drawdown after about 140 minutes is more dramatic than would be expected due to a formation response and is most likely due to an unrecorded increase in flow rate.

The pumping test analysis for Kasangati Reg. No. 371 shows a generally good match to the first drawdown step and the recovery period with a negative skin of -2.4 and a transmissivity of 1.4 m$^2$/d (Figure 7-7a and 7-7b). However, Figure 7-7a shows that these parameters produce a poor match to the second and third drawdown steps. This increased drawdown at higher flow rates relative to the type curve is consistent with non-linear well losses (see appendix Section K.4.2). The third flow period appears better matched with a positive skin of about 4 and a higher transmissivity of about 4 m$^2$/d. The apparent change in skin is consistent with fracture flow. As the flow rate is increased, the high velocity flow via the fracture becomes turbulent thus creating non-linear well losses. A longer constant rate flow period and recovery period is required to establish greater confidence in the large scale transmissivity. This test provides a good example of the contradictory aims of a pumping test. The relatively short duration
steps provide information about the borehole efficiency, but a longer term constant rate test is required to investigate the large scale aquifer properties.

7.3.4 Kyadondo borehole

Kyadondo is located about 2 km west of Kasangati, also in the Wakiso District and 12 km north of central Kampala (Figure 7-1). The smooth drawdown response shown in Figure 7-8 indicates there was good flow rate control during the pumping test. The flow rate of 1.6 m³/hr was maintained for 5 hours and produced a maximum drawdown of s with a final drawdown of 18.1 m. However, both the drawdown and recovery periods were stopped before radial flow was fully established. Nevertheless, sufficient data was recorded to give reasonable confidence in the estimated transmissivity of 2.9 m²/d, especially since the skin (1.1) and transmissivity are consistent between the drawdown and recovery periods. Unusually, the recovery data appears shifted to the right and a larger casing radius (and hence wellbore storage) was therefore required to obtain an acceptable time match. This may be due to an error in the time record.

![Figure 7-8: Kyadondo DWD7986 (weathered regolith) log-log data and derivative type curve matches for a) drawdown and b) recovery](image)

7.3.5 Lukuuka borehole

Lukuuka is located about half way between Sembabule and Masaka (Figure 7-1) in the Masaka District. The relatively smooth drawdown response shown Figure 7-9a indicates that there was reasonable flow rate control at 3 m³/hr for the full 31 hours,
which produced a drawdown of 12.46 m. Given the higher available drawdown, the flow rate was increased to 4.5 m$^3$/hr for the final 32 minutes, which increased the drawdown to 18.97 m. The lack of a hump in the early to mid-time derivative for both the drawdown and recovery (Figures 7.9a and 7.9b) is consistent with a large negative skin (or high storativity) which is calculated to be -3.5. The long test duration, good radial flow stabilisation, and consistency between the drawdown and recovery data give a high degree of confidence in the transmissivity which is calculated to be 3.9 m$^2$/d.

Figure 7-9: Lukuuka DWD15296 (bedrock) log-log data and derivative type curve matches for a) drawdown and b) recovery

7.3.6 Nakatate borehole

Nakatate is located about 24 km south west of Masaka, in the Masaka District (Figure 7-1). The rough shape of the drawdown derivative shown in Figure 7-10 is indicative of fluctuations in the flow rate between about 10.6 and 11.8 m$^3$/hr. Pumping continued for 72 hours, by which time the maximum drawdown was 26.64 m. A positive skin of 1.1 was consistent with the drawdown and recovery data. The long test duration, good radial flow stabilisation, and consistency between the drawdown and recovery data give a high degree of confidence in the transmissivity which is calculated to be 16.8 m$^2$/d.
Figure 7-10: Nakatate DWD16584 (weathered / rock) log-log data and derivative type curve matches for a) drawdown and b) recovery

7.3.7 Rwakishakizi borehole

Rwakishakizi is located 14 km west of Mbarara in the Mbarara District (Figure 7-1). The flow rate during the pumping test was maintained between about 2 and 2.1 m$^3$/hr for 24 hours, resulting in a maximum drawdown of 26.66 m and the good radial flow stabilisation seen in Figure 7-11a. The parallel separation of the early-time drawdown and derivative in Figure 7-11a suggests a vertical fracture response and this together with the lack of hump on the derivative during the transition has been matched using a negative skin of -2.7. There is close agreement in the level of the radial flow stabilisation seen on the drawdown and recovery data (Figures 7.11a and 7.11b) giving confidence of the calculated transmissivity of 1.6 m$^2$/d. However, there is a time offset in the recovery data perhaps due to a recording error. In addition, the recovery response resembles that expected due to positive rather than negative skin. It is unclear if this is an artefact of measurement error or a real change in the skin during pumping.
Figure 7-11: Rwakishakizi DWD16561 (weathered / rock) log-log data and derivative type curve matches for a) drawdown and b) recovery

7.3.8 Mpwege borehole

Mpwege is located about 4 km south of the Busujja boreholes, on the Hoima road in the Wakiso District, approximately 30 km north-west of central Kampala (Figure 7-1). During the pumping test, the flow rate was maintained between 5.95 and 6 m$^3$/hr for 72 hours, with a maximum drawdown of 27.02 m, and a good radial flow stabilisation as seen in Figure 7-12a. The separation between the early-time drawdown and derivative suggests a vertical fracture connected to the borehole, and this behaviour has been modelled with a large negative skin of -4.55. There is a particularly large separation between the drawdown and derivative curves in the first ten minutes of drawdown (Figure 7-12a) and this same behaviour appears to be accentuated in the first few 10s of minutes of the recovery data (Figure 7-12b). The mid-time drawdown data is indicative of a good radial flow stabilisation corresponding to a transmissivity of 2.9 m$^2$/d. However, the overall shape of the recovery data is inconsistent with the drawdown data and the cause of this discrepancy cannot be assessed without further information.
Figure 7-12: Mpwege DWD16277 (sediment / weathered) log-log data and derivative type curve matches for a) drawdown and b) recovery

7.4 Interpretation of Pumping Tests in Fluviolacustrine Sediment

The results of the available pumping tests conducted mainly in the fluviolacustrine sediments of the eastern Katonga Valley and northwest shore of Lake Victoria are summarised in Table 7-2 and Figures 7-13 to 7-20. It can be seen from Table 7-2 that the maximum flow rates are all greater than 3.6 m$^3$/hr which is the minimum rate that Chilton and Foster (1995) suggest is usually required to maintain a motorised pump (see Section 8.7.3).
<table>
<thead>
<tr>
<th>Borehole ID</th>
<th>Location (District)</th>
<th>Depth (m bgl)</th>
<th>Summary Lithology</th>
<th>$Q$ (m$^3$/hr)</th>
<th>$T$ (m$^2$/d)</th>
<th>Skin$^1$ (-)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kisozi BH2</td>
<td>Kikuumaddungu (Sembabule)</td>
<td>6 to 54</td>
<td>Silty sand, gravel Gneiss</td>
<td>2.45 to 3.06 to 4.08</td>
<td>1.60</td>
<td>-3.08</td>
<td>Flow rate fluctuations with less than ideal radial flow stabilisation.</td>
</tr>
<tr>
<td>Kisozi BH3</td>
<td>Kikuumaddungu (Sembabule)</td>
<td>3 to 57 57 to 84</td>
<td>Silty sand, gravel Phyllite, gneiss</td>
<td>4.24 then 6.00</td>
<td>7.55</td>
<td>7.3</td>
<td>Two step flow rate with non ideal radial flow stabilisation. Positive skin during recovery.</td>
</tr>
<tr>
<td>DWD 11679</td>
<td>Nkozi (Mpigi)</td>
<td>2 to 48 48 to 61</td>
<td>Clayey silt, gravel Gneiss</td>
<td>9.95, 11.81, 5.91, 5.95</td>
<td>13.05</td>
<td>12.63</td>
<td>Poor flow rate control with less than ideal radial flow stabilisation.</td>
</tr>
<tr>
<td>DWD 11680</td>
<td>Nkozi (Mpigi)</td>
<td>3 to 45 45 to 61</td>
<td>Clayey silt, gravel Gneiss</td>
<td>9.00 to 6.90</td>
<td>6.79</td>
<td>0.90</td>
<td>Flow rate fluctuations with less than ideal radial flow stabilisation.</td>
</tr>
<tr>
<td>Reg. No. 305</td>
<td>Bukakata (Masaka)</td>
<td>0 to 80</td>
<td>Not available (silt, sand, gravel)</td>
<td>~12.00</td>
<td>45.64</td>
<td>-3.50</td>
<td>Reasonable flow rate control and radial flow stabilisation.</td>
</tr>
<tr>
<td>DWD 17020</td>
<td>Bukulula (Masaka)</td>
<td>6 to 118</td>
<td>Silty clay, med. to coarse sand</td>
<td>7.98 to 6.75 to 6.99</td>
<td>11.44</td>
<td>4.13</td>
<td>Flow rate fluctuations, but reasonable radial flow stabilisation.</td>
</tr>
<tr>
<td>DWD 17021</td>
<td>Mukono (Masaka)</td>
<td>6 to 62</td>
<td>Clayey sand, coarse sand, cobbles</td>
<td>25.05, 24.29, 24.36, 23.63</td>
<td>175.10</td>
<td>9.23</td>
<td>Poor flow rate control with non ideal radial flow stabilisation. Possible constant head boundary in recovery data (70 m).</td>
</tr>
<tr>
<td>Reg. No. 307</td>
<td>Kagera (Rakai)</td>
<td>0 to 90.8</td>
<td>Not available (silt, sand, gravel)</td>
<td>8.50</td>
<td>17.78</td>
<td>0.00</td>
<td>Good flow rate control and clear radial flow stabilisation. Possible constant head boundary in drawdown data (350 m).</td>
</tr>
</tbody>
</table>

Note: 1) Storativity assumed = 1E-4 and skin is allowed to vary.

Table 7-2: Summary of pumping test analyses for boreholes mainly in fluviolacustrine sediment
7.4.1 Kikuumaddungu boreholes

The Kikuumaddungu boreholes are located about 3 to 4 km north east of Kikuumaddungu village, in the Sembabule District close to the Katonga papyrus wetland and Kisozi crossing (Figure 7-1) where the ERT survey described in Section 6.4 was conducted. Borehole 2 is located approximately 100 m southeast of the murrum road and Borehole 3 is located about 350 northwest of the road. The driller’s logs for Kikuumaddungu boreholes 2 and 3 are shown in Table 4-9 and discussed in Section 4.5.4. As shown in Table 7-2 the lithology is predominantly silty sand, overlying gneiss, and most flow appears to be coming from a thin gravel horizon near the base of the sediment at 55 m bgl.

The rough character of the derivative calculated for Kikuumaddungu borehole 2 (Figure 7-13a) is indicative of poor flow rate control during the drawdown. Unfortunately, the flow rate was gradually increased from 2.5 to 4.1 m$^3$/hr over a period of 36 hours with the maximum drawdown eventually attaining 44 m. Despite the poor rate control, the early-time log-log diagnostic plots for the drawdown (Figure 7-13a) and recovery (Figure 7-13b) are both consistent with a negative skin of -3.1. In addition the overall type curve match and consistency between the drawdown and recovery data provide confidence in the calculated transmissivity of 1.6 m$^2$/d.

The character of the derivative calculated for Kikuumaddungu borehole 3 (Figure 7-14a) is again indicative of poor flow rate control during the drawdown, with the flow rate being increased in two main steps from about 4.2 m$^3$/hr to about 6 m$^3$/hr after 1.7 hours. The maximum drawdown after 60 hours was 46.8 m. The changes in flow rate mask the diagnostic character of the derivative during the drawdown (Figure 7-14a). The large drop in the derivative towards the end of the drawdown (Figure 7-14a) and the recovery period (Figure 7-14b) is indicative of a large positive skin, suggesting an increase in transmissivity with distance from the borehole. There is no clear radial flow stabilisation and therefore the estimated large scale transmissivity of 7.6 m$^2$/d is based on the derivative at the end of the drawdown. Given the proximity of the borehole to the papyrus wetland it is possible that the rapid decrease in the derivative is due to the presence of a nearby constant head boundary. Thus, as shown in Figure 7-14b,
although the near-field transmissivity is only of the order of $1 \text{ m}^2/\text{d}$, the well is capable of maintaining a sustainable yield of $6 \text{ m}^3/\text{hr}$, with a large drawdown of 47 m, because of the higher far-field transmissivity ($> 7.6 \text{ m}^2/\text{d}$) or nearby constant head boundary.

Figure 7-13: Kikuumaddungu BH 2 (sediment) log-log data and derivative type curve matches for a) drawdown and b) recovery

Figure 7-14: Kikuumaddungu BH 3 (sediment) log-log data and Derivative type curve matches for a) drawdown and b) recovery
7.4.2 Nkozi boreholes

Nkozi is located on the equator in the Mpigi District about 2 km west of the Kampala-Masaka road (Figure 7-1). The boreholes are drilled into clayey silt and gravel in a tributary of the Katonga Valley near its current outlet to Lake Victoria. The flow rate during the pumping test in DWD11679 was initially increased from 10 m$^3$/hr to 11.8 m$^3$/hr and then reduced to about 5.9 m$^3$/hr which resulted in the poor diagnostic quality of the log-log drawdown plot shown in Figure 7-15a. The final drawdown after 72 hours was 39.1 m. The slope of the early-time drawdown (Figure 7-15a) and recovery data (Figure 7-15b) is anomalous and suggests possible measurement error. However, the mid-time data for both the drawdown and recovery is indicative of a large positive skin, estimated to be 12.6. Although there isn’t a clearly define radial flow stabilisation, the overall match to the late-time drawdown and recovery data suggests a large scale transmissivity of the order of 13 m$^2$/d.

![Figure 7-15: Nkozi DWD11679 (sediment) log-log data and derivative type curves matches to a) drawdown and b) recovery](image)
The pumping test records for DWD11680 indicate that the pumping rate was reduced from 9 m$^3$/hr to 7.2 m$^3$/hr after 18 minutes. However, the rough character of the derivative shown in Figure 7-16a suggests that further fluctuations in the pumping rate occurred during the test. After an increased rate of drawdown, the displacement reaches a maximum of 33.3 m after 9 hours and remains constant until the end of the drawdown at 40 hours. This anomalous behaviour is likely an artefact of the test control and records rather than the aquifer properties. Unfortunately the recovery period was also stopped before the onset of the radial flow stabilisation. Despite these problems an overall match to both the drawdown and recovery period produces consistent values for the skin and transmissivity of 0.9 and 6.8 m$^2$/d, respectively.

### 7.4.3 Bukakata borehole

Bukakata is located in fluviolacustrine sediment on a peninsular beside Lake Victoria in the Masaka District, 32 km east of Masaka, and 8 km west of Bugala Island. During the pumping test in Bukakata borehole Reg. No. 305 the flow rate was maintained at about 12 m$^3$/hr for 24 hours and the maximum drawdown was only 5.4 m. A wide separation in the early-time displacement and derivative curves followed by a mid-time radial flow stabilisation is observed in both the drawdown (Figure 7-17a) and recovery (Figure 7-17b) data. This response has been matched with a skin of -3.5 and a transmissivity of 45.6 m$^2$/d.
Figure 7-17: Bukakata Reg. No. 305 (sediment) log-log data and derivative type curves matches to a) drawdown and b) recovery

7.4.4 Bukulula and Mukono boreholes

Bukulula and Mukono are located between 1.5 and 5 km south of Lukaya in the Masaka District on the edge of the fluviolacustrine sediment (Figure 7-1). The flow rate during the pumping test in Bukulula borehole DW1720 initially reduced from about 8 m$^3$/hr to 6.75 m$^3$/hr with a final pumping rate of about 7 m$^3$/hr. The final drawdown after 72 hours was 29.9 m. A consistent match was achieved to both the drawdown data (Figure 7-18a) and recovery data (Figure 7-18b) which using a large skin of 4.1 and a transmissivity of 11.4 m$^2$/d.
The flow rate during the pumping test in Mukono borehole DW1721 varied between 25.1 m$^3$/hr and 23.6 m$^3$/hr producing the large fluctuations in the derivative seen in Figure 7-19a. The final drawdown after 48 hours was 9.6 m. Unfortunately, there is no clear radial flow stabilisation in either the drawdown (Figure 7-19a) or the recovery (Figure 7-19b) data, but a consistent match can be achieved to both with a positive skin of 9.2 and a large transmissivity of 175.1 m$^2$/d. Given the relatively large far-field transmissivity and large positive skin, further development of this well could potentially result in a higher yield.
Figure 7-19: Mukono DWD17021 (sediment) log-log data and Derivative type curves matches to a) drawdown and b) recovery

Figure 7-19b shows an apparent rapid reduction in the late-time derivative during the recovery period. This feature has been matched with a constant head boundary located only 70 m from the well. However, this behaviour is not seen in the drawdown data although it could be masked by the changing flow rate. A better controlled test and further local hydrological and geological information is required to improve the hydrogeological interpretation.

7.4.5 Kagera borehole

Kagera borehole Reg. No. 307 is located near the mouth of the River Kagera about 2 km west of Lake Victoria in the Rakai District. The flow rate during the pumping test was maintained at 8.5 m$^3$/hr for 24 hours and the maximum drawdown was 15.45 m. The early-time data during the drawdown (Figure 7-20a) and recovery (Figure 7-20b) is indicative of zero skin and a well constrained transmissivity of 17.8 m$^2$/d. There is a sharp increase in the derivative during the late-time recovery data which appears anomalous. In contrast, the smooth downturn in the derivative during the late-time drawdown period resembles a natural response and has been modelled using a constant head boundary located 350 m from the well. Whilst the discrepancy in the late-time data recorded during the drawdown and recovery makes the boundary conditions uncertain, the agreement between the early and mid-time data provides confidence in the well performance and aquifer transmissivity.
Figure 7-20: Kagera Reg. No. 307 (sediment) log-log data and derivative type curves matches to a) drawdown and b) recovery

7.5 Conclusions

This section has shown that well test analysis techniques originally developed by the petroleum industry to interpret results obtained from production wells may be successfully applied to water supply boreholes in Africa to get as much information as possible about the hydrogeological unit from data of mixed quality. The early to mid-time data in all of the datasets analysed here were matched with a relatively simple model incorporating wellbore storage and skin, followed by radial flow. No pumping tests in the weathered/fractured rock or the fluviolacustrine sediments produced a classic delayed yield response, characterised by a trough in the mid-time derivative, and indicative of an unconfined or double porosity system. However, this does not necessarily mean that the pumped hydrogeological units are confined by classic aquicludes. It may simply mean that the volume of the cone of depression is small compared to the volume influenced by the test as a whole. The primary zone of inflow may be deep below the water table and the vertical hydraulic gradient, accentuated by lower transmissivity layers, ensures that there is insignificant drawdown of the water table itself.

Two tests (Mukono DWD17021 and Kagera Reg, No. 307) showed late-time recharge, although this was not consistently observed during the drawdown and the recovery. Taken at face value, this could imply either horizontal recharge from a vertical constant
head boundary, or vertical recharge from an horizontal leaky aquifer above or below. A constant head boundary has been used to provide an aesthetic match to the late time data in these cases. To investigate these recharge conditions further, it is recommended that a repeat test is carried out so as to attempt to obtain consistent drawdown and recovery data. Further local geological and hydrological information is also required to assess the most likely form of recharge boundary.

Most of the pumping tests interpreted here were previously unanalyzed. The hard copy data was transcribed prior to analysis. From the license application files, it appears that the yield selected for abstraction permits is frequently assessed on a qualitative basis without estimating specific well losses and aquifer properties. Nevertheless, this section has shown that the archived pumping test data held by DWRM provides a valuable resource for assessing the overall pattern of borehole performance and hydrogeological properties of the most transmissive rocks in Uganda. This valuable archive also provides a basis for DWD and DWRM to make recommendations to private water supply contractors for improving pumping test procedures to facilitating better assessment and management of Ugandan groundwater resources. The transmissivity datasets are summarised and compared further in Section 8.7.3.
8 SYNTHESES AND DISCUSSION OF THE GEOLOGICAL HISTORY AND HYDROGEOLOGY OF THE KATONGA VALLEY

8.1 Introduction

This chapter brings together the diverse AFT, ERT and pumping test data sets and synthesises the exhumation history, landscape evolution and hydrogeology of the Katonga Valley. The discussion begins on a large spatial and temporal scale by first considering the Phanerozoic thermotectonic history and continental setting of the Katonga Valley. It then focuses on the geometry of the valley fill and discusses episodes of erosion and deposition within the valley itself. Evidence from the DEM, ERT, borehole logs, lake bathymetric and seismic surveys are included in a discussion of the Neogene tectonic geomorphology of the rift flank location of the Katonga Valley. Finally, the synthesis focuses on the fluvio-lacustrine sediments deposited in the eastern Katonga Valley and examines their groundwater resource potential.

8.2 Regional History of Burial and Exhumation from AFT Analysis

The conclusions of the rift-flank AFT analysis and thermal history modelling (Chapter 5.0) are here combined with evidence from the Rwenzori AFT analysis (Bauer et al., 2010) and observations of the Karoo-age outliers (Chapter 2.0) to construct a conceptual model of the thermotectonic history of the region adjacent to the Western Rift, including the Katonga Valley. The thermotectonic history may be divided into five main periods as illustrated in Figure 8-1 and described below.

a. Middle to Late Palaeozoic: exhumation

Thermal history modelling of the Katonga Valley AFT data indicated that the Precambrian rocks of the Katonga Valley first entered the base of the partial annealing zone (PAZ, nominally about 4 km depth) during the Devonian and Carboniferous (416 to 399 Ma). Since apatite in the Bihanga diamictite has the same fission track age as the Precambrian rocks and was likely deposited during the late Carboniferous or early Permian glacial event it can be inferred that there was about 4 km of denudation in less than 120 Ma. As shown in Figure 8-1a the late Palaeozoic was therefore characterised
by overall exhumation. For comparison, the late Palaeozoic net denudation rate appears greater than the denudation rate on the central East African Plateau and similar to the denudation rate in central Kenya and coastal Tanzania during the last 120 Ma (see Figure 5-12). This period of exhumation ended with glacial erosion and the formation of a landscape of high relief in the Precambrian rocks now exposed once again in the eastern D.R. Congo and western Katonga Valley (Boutakoff, 1948, Cahen, 1954, Cahen and Lepersonne, 1981).

b. Late Palaeozoic to Early Mesozoic: burial

The thermal history modelling also indicated that the fission track shortening observed in the Bihanga diamictite required post-depositional reheating consistent with burial depths between about 2 and 4.5 km. Evidence from the Congo Basin suggests that deposition continued throughout the Permian (Lower Karoo Lukuga Group) although Giresse (2005) and Daly (1992) disagree on whether Triassic to early Jurassic (Upper Karoo) sediments are also represented in the basin succession. Giresse (2005) reports the presence of a major angular unconformity above the Permian strata indicating a period of deformation prior to post-mid Jurassic deposition. In Tanzania, terrestrial deposition continued in the Karoo grabens until the late Triassic (Schlüter, 1997).

On the basis of this evidence it therefore seems likely that a considerable thickness of sediments were deposited above the Bihanga diamictite between the early Permian and the Jurassic. However, the depth of burial was insufficient to cause complete closure of apatite fission tracks in the vicinity of the Katonga Valley. Since the base of the sediments attained a maximum thickness within the depth range of the PAZ, the thickness of underlying Precambrian rocks that retained Palaeozoic AFT ages, and are exposed today, must therefore be less than the total thickness of the PAZ (< 2 km) as shown in Figure 8-1b.
Figure 8-1: Conceptualisation of thermotectonic history leading to current distribution of Apatite Fission Track ages

e. **Horst and Rift-Flank Uplift**
   From Rwenzori and Katonga Valley AFT central ages and thermal history modelling. Sediment loading and rift-flank uplift.

   d. **Plateau Uplift and Rift Grabens**
   Cenozoic exhumation from AFT thermal history modelling. Semliki and George grabens propagate from north and south.

   c. **Exhumation**
   From Rwenzori AFT central ages and Katonga Valley thermal history modelling (multiple exhumation/burial possible).

   b. **Burial**
   From Karoo Outcrop and Katonga Valley AFT thermal history modelling.

   a. **Exhumation**
   From Katonga Valley AFT central ages and thermal history modelling.
In the vicinity of the Muzizi Valley, located about 70 km north of the Katonga Valley, the rocks contain apatite with Triassic and Jurassic fission track ages. Therefore, the rocks with Palaeozoic AFT ages have been lost in this northern region due to either greater early Mesozoic burial and reheating, or post-mid Mesozoic denudation.

c. Middle to Late Mesozoic: exhumation

The central ages calculated for the Rwenzori AFT data (Bauer et al., 2010) indicated that the rocks that now form the Rwenzori Mountains first entered the base of the PAZ during the Jurassic to early Cretaceous. This suggests a period of net exhumation during the middle to late Mesozoic as shown in Figure 8-1c. However, due to the non-unique nature of solutions to the inverse problem, the thermal history modelling conducted as part of this study is unable to determine if Mesozoic cooling was continuous or episodic. As discussed in Section 5.7.4, the presence of offset fault blocks in the Rwenzori Mountains also prevents the AFT age versus elevation data from being used to identify periods of enhanced exhumation during the middle to late Mesozoic. Given the extensive outcrop of Karoo-age sediments that exist today on the eastern margin of the Congo Basin (Cahen and Lepersonne, 1981) and the small outliers identified by this study on the EAP, the Mesozoic uplift and denudation removed less than 2 to 4.5 km of rock. Allowing for continued denudation during the Cenozoic before and after uplift of the EAP it seems likely that the region retained a more extensive cover of Karoo-age sediments at the end of Mesozoic than exists today.

d. Miocene to Pliocene: plateau uplift and rift grabens

The timing of the initial uplift of the EAP remains uncertain although several researchers have assumed that it is related to propagation of the EARS (Ebinger and Ibrahim, 1994, Chorowicz, 2005). Volcanism associated with the eastern branch of the EARS began at about 35 to 30 Ma in northern Kenya and 15 Ma in central Kenya. Volcanism associated with the northern part of the Western Rift began about 12.6 Ma in the Virunga Province and 10 Ma at Kivu (Nyblade and Brazier, 2002). Therefore it
is estimated that uplift of the order of 1 km on the EAP began between the early Oligocene and middle Miocene.

Currently, there is no evidence to quantify any potential increase in the erosion rate of the central EAP following uplift. Neogene erosion rates on the EAP were such that the weathering zone has been able to retain an average to maximum thickness of 30 to 100 m until the present day. Kohn et al (2005) showed that the rocks of the central EAP around Lake Victoria have Palaeozoic AFT ages whilst those on the margins in Kenya and Rwanda have Mesozoic AFT ages (Figure 5-10). This suggests that on a large scale, greater uplift and denudation occurs closer to parts of the rift flanks. However, on a smaller scale, it was shown in Section 5.7.3 that there is no significant trend in AFT ages away from the rift flank parallel to the Katonga Valley (Figure 5-13 and 5-14).

As discussed in Section 4.2.1, biostratigraphic evidence indicates that rifting of the Albert basin north of the Katonga Valley, began between 12 and 9 Ma (Pickford et al., 1993). Koehn et al (2010) used modelling to show that given the measured crustal extension rate of 2.1 mm/yr, southward and northward propagating rift faults bordering the Semliki Basin and the George Basin would take between 6 and 14 Ma to detach the intervening basement block allowing uplift of the Rwenzori horst (Figure 4-13). Figure 8-1d illustrates Neogene uplift of the EAP. The Karoo-age sediments are reduced to a thin cover. Whilst rocks with Palaeozoic AFT ages may have been exposed at this time in the region of the Katonga Valley, those with Mesozoic AFT ages were not. In accordance with the numerical modelling conducted by Koehn et al (2010), Figure 8-1d shows lakes within the propagating rift grabens either side of the yet to be detached Rwenzori basement block. This is different to previous interpretations which ignored the likely precursor structural conditions prior to uplift of the Rwenzori horst and proposed the existence Lake Obweruka extending from the Albert Basin to the Edward Basin (Pickford et al., 1993). Given that the relative stresses during rift propagation from the north and south, it seems likely that an upward rather than downward force would be exerted on the intervening basement block. However, significant uplift of the Rwenzori horst could only have occurred after the Semliki Basin connected to the Edward Basin and the southern end of the basement block was detached (Koehn et al.,
Figure 8-1d therefore shows no significant vertical displacement on the intervening block prior to detachment.

e. Quaternary: Rwenzori horst and rift-flank uplift

Figure 8-1e shows a simplified representation of the present day distribution of AFT ages. Following partial detachment, uplift of the Rwenzori horst commenced in the early Pleistocene (Ring, 2008). As discussed in Appendix H, MacPhee (2006) concludes that no more than 1.7 km of denudation could have accompanied 5 km of Cenozoic surface uplift of the Rwenzori, suggesting rock uplift of no more than 6.7 km. Bauer et al. (2010) also determined some Oligocene to Miocene apatite (U-Th)/He ages which is consistent with late Neogene rock uplift. However, surface denudation could not keep up with rock uplift and therefore rock samples taken from the Rwenzori Mountains have Mesozoic AFT ages. As shown in Figure 8-1e, rock with Cenozoic AFT ages remains buried.

Figure 8-1e shows the sediment which has been eroded from the Rwenzori Mountains and rift flanks infilling the Semliki and George Basins. Unloading of the rift flanks produces isostatic rock uplift, whilst sediment loading in the basins produces flexural uplift of the rift flanks. Recent denudation has removed almost all evidence of the former Karoo-age sedimentary cover from the EAP with only one or two small outliers remaining in the exhumed Gondwanan palaeovalleys. More extensive Karoo-age deposits remain on the edge of the Congo Basin to the west. Rocks with Palaeozoic AFT ages are exposed on the rift flanks, although the period of late Palaeozoic to early Mesozoic burial implies that these are likely underlain by rocks with Mesozoic AFT ages at a relatively shallow depth (< 2 km).

8.3 Geometry of Valley Erosion and Deposition from ERT Surveys

8.3.1 Integrated interpretation of ERT surveys and field observations

This section presents a integrated interpretation of the ERT surveys from the three study sites at Kabagole, Kyai and Kisozi (Figure 1-3) in order to construct a conceptual framework for the major phases of erosion and deposition preserved in the Katonga
Valley. Figure 8-2 shows the four ERT surveys from the three research sites at the same scale and using the same resistivity contour spacing and colour scheme. The topography of the valley sides are shown continuing beyond the edge of each ERT survey. The location of the sand and gravel terrace above the ERT survey is also shown at Kyai. In addition, the approximate maximum elevation of Lake Victoria (1,190 m asl) and current lake level (1,133 m asl) are shown on all three cross sections.

The variation in the width and depth of the Katonga Valley at the three study sites becomes apparent when all ERT sections are presented on the same scale. In particularly, it can be seen that the central valley is much narrower at the Kyai crossing compared to the western valley at Kabagole and the eastern valley at Kisozi.

The ground resistivity models created from individual ERT surveys presented in Chapter 6 are used here to identify and classify common erosion surfaces and depositional units at each of the research sites. A parsimonious interpretation has been applied to produce a first-pass synthesis of those parts of the history of erosion and deposition preserved in the Katonga Valley. For example, it is assumed that the overlying very low resistivity sediment may be categorised into a single near-contemporaneous depositional unit at all sites. Three main erosion surfaces have been interpreted and are named A, B and C. The depositional units immediately overlying each erosion surface have been given a numerical suffix. A description and interpretation of each of the erosion surfaces and depositional units annotated in Figure 8-2 is provided in Table 8-1. The interpretation presented in Table 8-1 is not necessarily definitive and should be viewed as a hypothesis to be corroborated or falsified as further data become available.
Figure 8-2: Episodes of erosion and depositional interpreted from ERT inversion models
<table>
<thead>
<tr>
<th>EVENT</th>
<th>DESCRIPTION</th>
<th>INTERPRETATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Erosion Surface A is cut into the Precambrian rocks in the western valley at Kabagole and is overlain by a sedimentary outlier.</td>
<td>This erosion surface is associated with the original topography of the Palaeozoic palaeovalley which was eroded prior to deposition of the Lower Karoo age sediments (see A1) and is likely of Carboniferous age.</td>
</tr>
<tr>
<td>A1</td>
<td>Depositional Unit A1 in the western valley at Kabagole comprises brownish grey to reddish yellow, thinly bedded, moderately strong, fine-grained sandstone.</td>
<td>Rocks with similar characteristics in hand specimen to Depositional Unit A1 are found in association with diamicites and rhythmites at Bihanga Station. These have been interpreted to be of Lower Karoo age on the basis of glaciogenic facies characteristics and the presence of a plant fossil resembling the Permo-Carboniferous gymnosperm, Neoggerathia.</td>
</tr>
<tr>
<td>B</td>
<td>Erosion Surface B has exploited earlier Erosion Surface A in the western valley and cut into the Precambrian rocks in the eastern and central valley where it is overlain by silty sand at Kisozi and sand and gravel terraces at Kyai.</td>
<td>Erosion Surface B is associated with the topography of the Neogene relict valley formed prior to deposition of fluvial and lacustrine deposits. Whilst exhumation of the Karoo age sedimentary cover may have continued since the Mesozoic, the final erosion surface is likely of Quaternary age.</td>
</tr>
<tr>
<td>B1</td>
<td>Depositional Unit B1 in the central valley at Kyai is comprised of light grey, weakly cemented, sub-angular to sub-rounded, planar to cross-bedded sand and gravel, with cobbles.</td>
<td>Depositional Unit B1 is exposed in terraces on the valley side and the ERT suggests it is absent from the valley bottom. It rests directly on weathered phyllite. The large grain size and bedding structure suggest it was deposited in a high energy environment, but the sub-angular clasts indicate limited transport. These Pleistocene deposits once filled the relict valley at Kyai.</td>
</tr>
<tr>
<td>B2</td>
<td>Depositional Unit B2 in the eastern valley at Kyai comprises silty sand with occasional gravel.</td>
<td>The lithology of Depositional Unit B2 is interpreted from driller’s logs of the Kikuumaddungu boreholes at Kisozi. It likely represents fluvial and/or near-shore lacustrine deposits filling the relict valley in an arm of Lake Victoria during the late Pleistocene.</td>
</tr>
<tr>
<td>EVENT</td>
<td>DESCRIPTION</td>
<td>INTERPRETATION</td>
</tr>
<tr>
<td>-------</td>
<td>----------------------------------------------------------------------------</td>
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</tr>
<tr>
<td>C</td>
<td>Erosion Surface C is cut into: a) sandstone at Kabagole; b) sand and gravel and Precambrian rocks at Kyai; and c) the silty sand at Kisozi. It is overlain by clayey silt.</td>
<td>Although the upper erosion surface at Kabagole, Kyai and Kisozi is cut into different substrates, it is categorised as a near-contemporaneous feature because the ERT suggests it is overlain by similar clayey silt deposits (see C1) at all three sites. It is interpreted as an episode of recent fluvial erosion along the entire valley.</td>
</tr>
<tr>
<td>C1</td>
<td>Depositional Unit C1 is present along the entire valley and has been described at Kabagole as grey-brown mottle orange, firm, silty clay with sand</td>
<td>Stagnation occurred following the final episode of channel erosion (C) resulting in the formation of Holocene wetlands and associated fine-grained clayey silt paludal deposition.</td>
</tr>
</tbody>
</table>

Table 8-1: Summary of major episodes of erosion and deposition identified from ERT

8.3.2 Relationship between valley fill and Lake Victoria water levels

The maximum water level historically attained by Lake Victoria based on strand lines is about 1,190 m asl (Johnson, 1960, Temple, 1966). As discussed in Section 4.4.3, tilting of the upper two strandlines by up to 10 m within 35 km of Lake Victoria suggests post-deposition relative uplift towards the west, or downwarp towards the east. However, in the absence of collated evidence further west, this differential uplift has not been accounted for when placing the maximum level of Lake Victoria at 1,190 m asl on the sections in Figure 8-2. Nevertheless, the influence of changes in ground elevation as well as lake elevation is considered during interpretation.

As shown in Figure 8-2, the approximate maximum elevation of Lake Victoria lies about 15 m above the flood plain at Kisozi in the eastern Katonga Valley. It intersects the slope to the south of the terrace flood plain investigated during the Kisozi south ERT survey. This is consistent with the view that there were periods of the middle to late Quaternary when the eastern Katonga Valley was flooded and formed an arm of Lake Victoria (Temple and Doornkamp, 1970).

In the central valley at Kyai, the maximum Lake Victoria elevation lies about 10 m above the low resistivity channel but near the base of the sand and gravel terraces.
(Figure 8-2). It is feasible that inundation during maximum lake levels may have extended through the central valley as shown in Figure 2-4. Both lacustrine and paludal environments would have facilitated fine-grained deposition in the base of the valley. The sub-angular sand and gravel terraces at Kyai are higher than the uppermost lakeshore deposits (1,190 m asl) identified closer to Lake Victoria (Johnson, 1960). As discussed in Section 4.5.4, similar deposits have been described between about 1,190 and 1,250 m asl in the Nabakasi Valley and on the interfluves to the north (Johnson, 1958). Their character and elevation suggest that they have a different origin to the beach deposits further east. Their coarse clast size and sub-angular clast shape are evidence of relatively short transport distance in a high energy environment and the once likely filled the high level valley.

In the western valley at Kabagole, the maximum Lake Victoria elevation occurs about 5 m above the papyrus wetland and close to the same level as the grazing land underlain by low resistivity fine-grained sediments on either side of the valley (Figure 8-2). On the valley slopes, the low resistivity material extends to an elevation of about 20 m higher than 1,190 m asl. However, it is unclear if this low resistivity material is a continuation of the fine-grained alluvium of the valley floor or the deeply weathered crystalline rock of the valley sides. As shown in Figure 2-4, it is possible that the Katonga Valley was inundated almost as far west as the current water divide during periods when Lake Victoria attained its maximum elevation. If the low resistivity material above 1,190 m asl is indeed lacustrine or paludal deposits then this would suggest that the ground surface has risen since Lake Victoria attained its maximum elevation.

Figure 8-2 shows that the current level of Lake Victoria (1,133 m asl) at Kisozi in the eastern Katonga Valley occurs about 5 m below the deepest part of the low resistivity channel and 15 m above the deepest part of the bedrock surface. The arrow-barb pattern of the eastern Katonga drainage system was formed after removal of the Karoo-age sedimentary cover when the early Neogene river flowed westward. Given that the average depth of the closed basin containing Lake Victoria is currently only about 40 m, it seems unlikely that significant vertical erosion would have continued following drainage reversal and therefore the deepest part of Erosion Surface B in the bedrock for
likely formed prior to river reversal. Although the arrow-barb river pattern indicates the original drainage was towards the west, the Katonga Valley broadens in the east towards Lake Victoria (Figure 2-4). Sagging of the Victoria Basin likely contributed to preservation of the original pre-reversal broad Katonga Valley in the east. The valley width may also have been increased by post-reversal lateral erosion. The sediments in the eastern Katonga Valley were likely deposited by both fluvial and lacustrine processes as the lake surface elevation fluctuated between current and maximum levels.

At Kyai, the current level of Lake Victoria occurs about 20 m below the low resistivity channel cut into the medium resistivity bedrock (Figure 8-2). The narrow character of the central Katonga Valley appears to have been created when Erosion Surface C cut through the old valley floor (Erosion Surface B), creating the sand and gravel terraces, and continued cutting down into the underlying weathered Proterozoic rocks.

At Kabagole, the current elevation of Lake Victoria is lower than the maximum depth of the resistivity model (Figure 8-2). Extrapolation of the resistivity contrast on the sides of the valley (Erosion Surface B/A) suggest that the base of the palaeovalley is lower than the current level of Lake Victoria (1,133 m asl). As discussed in Section 6.6 there is some uncertainty regarding the interpretation of the low resistivity material (< 68.7 Ωm shown in blue in Figure 8-2) filling the valley on either side of the medium resistivity sandstone mound (68.7 to 101 Ωm shown in yellow in Figure 8-2). If the low resistivity material comprises unconsolidated lacustrine or paludal deposits it would appear that the base of the Neogene to late Quaternary valley (Erosion Surfaces B and C) are lower than both the current level of Lake Victoria and the bedrock surface at Kyai. However, it is difficult to envisage a plausible sequence of fluvial erosion that would result in the preservation of a central sandstone mound between two relatively deep (> 40 m) channels within the western Katonga Valley. It is possible that the region with resistivity between 46.7 and 68.7 Ωm (shown in light blue in Figure 8-2) to the south of the sandstone mound is a fine-grained continuation of the Karoo-age sedimentary rocks, rather than fine-grained Quaternary deposits. Such an interpretation would suggest that the bedrock surface beneath the south side of the Katonga Valley as Kabagole occurs at about 1160 m asl similarly to that at Kyai. However, the lower resistivity material (<46.7 Ωm) at depth to the north of the central sandstone mound
indicates that the bedrock surface to the north is deeper than the bedrock surface at Kyai. A drilling investigation is required to establish the depth of the unconsolidated material either side of the sandstone mound with greater confidence. However, based on the ERT surveys it appears that there is at least one, if not two, Neogene channels in the western Katonga Valley at Kabagole that are deeper than the bedrock surface in the central Katonga Valley at Kyai. The fact that there is not a continuous fall in elevation of the valley floor towards the east, and the occurrence of a bedrock high at Kyai, implies that the water divide may have once occurred around Kyai, approximately 100 km east of its present location.

8.4 Conceptual Framework for Phases of Erosion and Deposition

Landscapes evolve in response to both internal forcing by site-specific features, events and autogenic processes, and external forcing by regional, or global allogenic processes. There are many challenges in attempting to correlate the geomorphological record of erosion and deposition with external forcing by climate and tectonics. In general, as the duration under consideration increases from thousands to millions of years, the relative scale and quantity of climatic variations which may have influenced the landscape also increase. However, the older the event, the less likely it has of being preserved in the geomorphological record. In general, the geomorphological record is less sensitive to climate change than the detailed proxy records (isotopes, diatoms, pollen etc) used to reconstruct the climate history (see Section 4.3). Whilst we may expect the landscape to respond to climate variability and tectonics, the possibility of deciphering a correlation is often problematic because:

- the older the geomorphological event the more likely its record has been overwritten by subsequent events;
- the smaller the geomorphological event the more likely its record has been overwritten by subsequent events
- due to the incomplete record and spatial dispersion it is often difficult, if not impossible, to determine the relative age of geomorphological features;
• geomorphological features are often less amenable to absolute dating than the sensitive proxy datasets selected specifically to reconstruct climate history; and,
• the geomorphological record often has a lower temporal resolution than the proxy datasets used to reconstruct climate history over the equivalent period.

The problem is therefore one of attempting to correlate a palimpsest of relatively low resolution, crudely dated, spatially dispersed, poorly correlated erosion surfaces and depositional units with a reconstructed relatively high resolution climate history, and/or lower resolution tectonic history. These constraints place significant limitations on the reconstruction of a temporally continuous chronology of erosion and deposition in the Katonga Valley. Instead, the data provided by field observations, AFT analysis, ERT surveys and borehole records are here synthesised to provide a very broad conceptual framework for evolution of the Katonga Valley and associated fill. A parsimonious synthesis of the data from all three study sites suggest that the evidence of erosion and deposition preserved in the Katonga Valley can be categorised into three phases of diminishing duration as summarised below.

**Phase I: The Gondwanan Palaeovalley**

A – Glacial erosion of the Carboniferous valley

A1 – Terrestrial deposition of the late Palaeozoic to early Mesozoic sedimentary rocks

**Phase II: The Neogene Relict Valley**

B – Late Mesozoic to Palaeogene denudation removes most of the sedimentary cover. The proto River Katonga continues exploiting the softer sediment in the western palaeovalley and begins eroding a new valley in Precambrian rocks to the east

B1 – Deposition of sub-angular sand and gravel in the central valley under the action of high energy, short transport fluvial processes such as those associated with alluvial fans

B2 – Deposition of silty sand with gravel in the eastern valley under the action of fluvial and near-shore lacustrine processes in an arm of Lake Victoria
Phase III: The Late Quaternary Channel

C – Erosion of a channel in the local substrate at each site, including the silty sand at Kisozi, the sand and gravel and Precambrian rocks at Kyai and the Karoo-age sandstone at Kabagole

C1 – Stagnation of the river flow and deposition of fine-grained sediment in the channel

Figure 8-3 presents a conceptual framework for evolution of the Katonga Valley through each of the three main phases leading up to the current geometry of the valley fill interpreted mainly from the ERT surveys. A vertically exaggerated west to east profile of the Mpanga-Katonga Valley System is shown at the top of Figure 8-3 and schematic representations of the evolution of the erosion surfaces and depositional units are shown beneath the profile approximately centred on the locations of Bihanga Station, Kabagole, Kyai and Kisozi.

Phase I of the historical conceptual model shown in Figure 8-3 begins during the Carboniferous with erosion of a glacial valley (Erosion Surface A). The form and location of this Gondwanan palaeovalley exerts an influence over the western part of the modern Katonga Valley between the water divide and Nkonge (Figure 1-3). This section of the valley is generally wide and straight, and indurated terrestrial sediment has been observed at Bihanga Station and Kabagole. The thermal history modelling indicated that the Karoo-age terrestrial sediments (Depositional Unit A1) attained a thickness of between 2 and 4.5 km during the Mesozoic.
Figure 8-3: Historical conceptual framework for erosion and deposition in the Katonga Valley fill
No record remains of the many tens of millions of years during the Mesozoic and Palaeogene between Phases I and II when the Gondwanan palaeovalley lay beneath the regional cover of sedimentary rocks. The weakly cemented sandstone Kisegi Formation was deposited in the area now occupied by the Western Rift before significant downthrow had occurred in the middle Miocene (Bishop, 1969, Roller et al., 2010). This indicates that the regional land surface approached the current level of exhumation by the early Neogene. The evidence for sediments of Pleistocene age in the valleys of south west Uganda (Hirst, 1926, Johnson, 1958, Bishop, 1969) and lack of evidence for earlier valley-fill, suggests that the base of the Katonga Valley (Erosion Surface B) may have begun to approach its current elevation during the Pliocene or early Pleistocene. Sub-cycles of deposition and erosion may have occurred between the late Miocene and early Pleistocene but left no trace in the current valley geometry and sedimentary record. In the west, the Neogene valley preferentially exploited the softer Karoo-age sediments in the Gondwanan palaeovalley compared to the adjacent Proterozoic crystalline rocks. In the east, where all evidence of the former Palaeozoic sediments and buried landscape has been removed, the River Katonga began to erode directly into the Precambrian rocks, and its southward migration appears to be limited by the quartzite ridges along the Katonga Line (Section 3.3.2).

The Neogene Relict Valley (Figure 8-3) formed during Phase II was originally eroded by a westward flowing river under conditions of greater stream power than exist today. No alluvial sediments observed in the base of the Katonga Valley are interpreted to have been deposited prior to river reversal. The bedrock high near Kyai suggests that early drainage reversal may have occurred about 100 km west of the current surface water divide. Given the location of the bedrock high far from the rift flank, the initial river reversal appears to be associated with sagging of the Victoria Basin. The thickness of lacustrine deposits and modern rates of sedimentation in Lake Victoria have been used to suggest that Lake Victoria originally formed around 400 ka (Johnson et al., 2000). The River Katonga west of Kyai continued to flow westward and erode the sediment within the Gondwanan palaeovalley. Downstream sedimentation and loading in the Western Rift likely produced lithospheric flexure, which together with the other potential mechanisms discussed in Section 4.2.2, resulted in rift flank uplift. As discussed further in Section 8.5, this led to a new surface water divide near its current location. The reduced gradient in the western Katonga Valley may have
initiated early deposition of fine-grained lacustrine and paludal sediments in the western valley. Solomon (1939, cited in Bishop and Trendall, 1967) reported cross bedded sediments indicating westward flow in the Kagera Valley at Kikagati associated with early Pleistocene fossil elephant molars. No further evidence of the chronology of river reversal has been identified in recent years and therefore the best estimate for the time interval during which the water divides developed and changed locations in the Katonga Valley is between about 1.2 and 0.4 Ma during the middle Pleistocene.

Given that the sandstone at Kabagole and the fluvial lacustrine deposits at Kisozi have medium resistivity (68.7 to 101 Ωm) and the valley fill either side of the sandstone mound at Kabagole has low resistivity (<68.7 Ωm), it is hypothesised that the fine-grained deposits continue to depth on the north and south side of the valley at Kabagole (Figure 8-3). However, it is also possible that the low resistivity region at depth on the south side of the valley at Kabagole may be a continuation of fine-grained Karoo-age rocks. In either case, the cross sectional geometry at Kabagole and Kyai is indicative of the complex interplay between tectonics, erosion and isostasy in the western Katonga Valley during the Pleistocene.

In contrast to the western valley, the geomorphology of the eastern Katonga Valley is more strongly influenced by Quaternary climate variability and fluctuations in the level of Lake Victoria. The sub-angular sand and gravel (Depositional Unit B1) observed at Kyai was likely deposited during an generally drier climate. Irregular but intense precipitation would have produced flash floods, facilitating erosion and creating high energy, short distance sediment transport. An arid climate would also have reduced vegetation cover and further enhanced erosion of the already deeply weathered regolith. The weak ferruginous cement suggests precipitation from groundwater and may have occurred at a time when the sand and gravel filled the valley. It probably filled the landscape for some considerable time before the terraces formed and it became free draining. The overall appearance and characteristics give the impression that the sand and gravel is probably at least older than late Pleistocene. The suggestion that it is indicative of arid environment deposition at least implies that the level of the early Lake Victoria was probably low at this time. As the climate turned wetter, the high energy river flowing towards the initially low base level in the Victoria Basin transported and deposited a thin layer of quartz gravel and cobbles immediately above
the bedrock at Kisozi. The eastward flowing river may have begun to erode the sub-
angular sand and gravel filling central valley. As the lake level rose and fluctuated
between its current and maximum elevation, a sequence of silt and sand fluviolacustrine
sediments were deposited around Kisozi and further east (Depositional Unit B2).

The start of Phase III shown in Figure 8-3 is associated with erosion of the Late
Quaternary Channel (Erosion Surface C) identified by the ERT surveys. The channel
may have formed after a period of aridity when the base level was still low in the
Victoria basin but increased precipitation created adequate stream power to erode the
fluviolacustrine sediment at Kisozi and cut through the sand and gravel into the bedrock
at Kyai. The most recent major fluctuation of climate, base level and stream power
occurred near the end of the Pleistocene when evidence suggests Lake Victoria
completely dried out between about 15.9 and 14.2 ka (Stager and Johnson, 2008).
Erosion through the sand and gravel at Kyai produced the terraces and reconnected the
east and west Katonga Valley. Once the surface of Lake Victoria approached its
current elevation, the surface water flow became slow and dispersed as the modern
papyrus wetlands were established during the Holocene and the former channel was
filled with fine-grained paludal deposits. Occasional raised lake levels appear to have
inundated the higher ground at Kisozi leaving behind a layer of low resistivity fine-
grained deposits between 3 and 6 m thick.

8.5 Tectonic Geomorphology in Cross-section

8.5.1 Evidence from DEM, ERT, bathymetry and lake seismic surveys

Figure 8-4 presents the vertically exaggerated cross-section of the Mpanga-Katonga
Valley System to assist understanding of the landscape evolution. Although, it is not a
detailed representation of the actual geology it is intended to be generally consistent
with the DEM, ERT, Lake Victoria bathymetry and seismic surveys. The upper surface
of the brown-filled area represents the topography of the Western Rift, Rwenzori
Mountains and the interfluves north of the Mpanga-Katonga Valley System taken from
a DEM constructed from SRTM data. The topography of the valley system itself,
shown in light blue, is also taken directly from the DEM. The representative depth of
the valley fill is interpreted from the ERT data acquired for this thesis (Chapter 6). The
representative depth of Lake Victoria is taken from the bathymetry data (Temple, 1966,
Crul, 1995), and the representative depth the sediment beneath Lake Victoria is interpreted from the seismic surveys (Johnson et al., 1996).

The extreme elevation of the Rwenzori horst is immediately apparent to the west of the section. The next highest elevations are to be found on the rift flanks. The highest peak north of the Mpanga Valley is the resistant quartzite of the Kabuga Hill Singo Series, and the main elevated area north of the Katonga Valley is associated the Mubende Granite. The water divide in the Mpanga-Katonga Valley System occurs between these two elevated regions to the north. The bedrock high at Kyai is shown about 50 m above the base of the sediments in the western and eastern Katonga Valley on either side. The eastern Katonga Valley fill is contiguous with the lacustrine sediment in the Victoria basin. The maximum lake depth and sediment thickness are each shown to be of the order of 60 m.

Figure 8-4: Simplified longitudinal geometry of the Mpanga-Katonga Valley System based on DEM, ERT, bathymetry and lake seismic surveys

In the absence of evidence for localised uplift around Kyai the bedrock high is explained by fluvial erosion away from a drainage divide towards both the east and the west as indicated by the arrows in Figure 8-4. The present day divide occurs at Bihanga Station and the westward flowing Mpanga River in the west is cut into bedrock south of Kabuga Hill. The original drainage divide at Kyai was also likely formed due to downwarp of the Victoria Basin but later rift flank uplift formed the
modern drainage divide 100 km to the west. Uplift of the rift margin would have isolated the western Katonga Valley between the old and new drainage divides and caused it to fill with fine-grained lacustrine and paludal sediment. The medium energy environment of later rivers in the eastern valley flowing into the Victoria basin facilitated the deposition of silt and sand with occasional gravel. The two stage reversal of the River Katonga proposed here was previously unrecognised by Doornkamp and Temple (1966) in their classic description shown in Figure 4-27. However, Bishop (1967) did recognise a ‘hinge line’ passing through the Kyai area with downwarping to the east and uplift on the rift flank (Figure 4-28).

8.5.2 Evidence from depth of weathering

The geomorphological evidence examined so far suggests that surface uplift of the rift flank controls the current location of the drainage divide. However, it does not indicate the magnitude of rock uplift on the rift flank during the Neogene. This section attempts to examine the general magnitude of rock uplift using information about the thickness of the weathered mantle. It begins by adopting the traditional paradigm that the weathered landscape is essentially a relict feature, before going on to consider the implications of dynamic equilibrium and transient conditions.

Taylor and Howard (1998) propose that climatic conditions in the early Miocene favoured deep weathering prior to development of the Western Rift. If negligible deepening of the weathering profile has occurred since the development of the rift flanks (i.e., the depth of weathering is a relict feature) then the shape of the base of weathering will reflect the degree of uplift. In addition, any reduction in the thickness of the weathering zone will reflect the variation in erosion rates across the rift flank.

Information about the depth of weathering is available from the national borehole database held by DWRM, which contains a field for the ‘depth to bedrock’. Most boreholes were drilled for village water supplies and therefore the density of boreholes is biased towards habitable regions of the landscape where the population density is also greatest. Whilst it is acknowledged that different drillers and field engineers may have identified the base of the regolith using different criteria, in the absence of overlying sediment this dataset still provides the best available approximation of the depth of weathering. A database query revealed 372 borehole records for Districts
adjacent to the Mpanga-Katonga Valley System. Of these records, 243 recorded ‘depth to bedrock’. Figure 8-5 shows a histogram of the ‘depth to bedrock’ for those boreholes with available data. It can be seen that the distribution has a positive skew with a minimum of 0 m bgl, a modal value of about 25 m bgl, a mean of 42 m bgl, and a maximum measured depth of 116 m bgl. The reason for the skewed distribution has not been examined. It may be related to the relative abundance of parent material and/or local hydrological and microclimatic conditions which promote differing susceptibility to weathering in different parts of the landscape. It may also be related to the spatial distribution of erosion. Another possibility is that the shape of the histogram is a transient characteristic related to a specific climate and erosion history. For example, the areas with deepest weathering could reflect relict features related to a time with optimum weathering conditions and little erosion, whereas the average thickness may be more closely related to current conditions.

![Histogram of weathering depth adjacent to the Mpanga-Katonga Valley System](image)

**Figure 8-5: Histogram of weathering depth adjacent to the Mpanga-Katonga Valley System**

Whilst there are 243 boreholes with the ‘depth to bedrock’ recorded, the location of most boreholes is given by the village name only. UTM coordinates are recorded for just 41 boreholes. Figure 8-6 shows the west to east cross section of the Mpanga-
Katonga Valley System with the depth and thickness of the deeply weathered regolith shown for these 41 boreholes (vertical red bars). Of those boreholes with both UTM and ‘depth to bedrock’ data, 30 occur adjacent to the eastern Katonga Valley and 11 occur adjacent to the western Katonga and Mpanga Valleys in the region influenced by rift flank uplift.

Figure 8-6: Profile of weathering depths on the eastern rift flank adjacent to the Katonga Valley

Given the small amount of data, care must be taken not to place undue confidence on any attempted inference. Nevertheless, the average weathered zone thickness of 53 m in the east is greater than the average weathered zone thickness of 26 m in the west as would be expected if greater erosion had occurred closer to the rift flank. The horizontal green bars indicate where the ground surface would have been if the regolith was originally 100 m thick and recent weathering can be ignored. The lower red line in Figure 8-6 was drawn to reflect the lowest parts of the weathered mantle and the upper red line reflects the potential top of the weathered mantle prior to rift flank uplift. Together, they give an indication of the amount of rift flank uplift that may have occurred assuming negligible increase in the depth of weathering since uplift began. It can be seen that the relief on the base of weathering is similar to the relief on the current land surface topography. Therefore, assuming that the recent increase in depth of weathering is negligible compared to the rate of erosion, it appears unlikely that
relative rock uplift due to both the flexural loading of the George Basin and the isostatic unloading of the rift flank was several hundred metres and it is more likely that it was only 10s of metres more than surface uplift. At the very least we can conclude that interplay between flexural uplift, erosion and isostasy on the rift flank has not occurred at a rate that facilitated complete removal of the weathered mantle on the interfluves. However, increased erosion was adequate to reduce the thickness of the weathered mantle on the rift flank interfluves and removed the weathered mantle completely in the channel of the Mpanga River close to the rift boundary fault.

The discussion so far has adopted the traditional geography-based paradigm that the depth of weathering is primarily a relict feature. An alternative view assumes that there is a state of dynamic equilibrium between the rate of erosion at the surface and the rate of propagation of the weathering front at depth. This second approach has been examined by geology-based researchers working at large spatial and temporal scales (Hilley and Porder, 2008, Hilley et al., 2010). From this perspective, the internal characteristics of the weathering zone and their origin via monocyclic or multicyclic processes, which are important to hydrogeologists (Dewandel et al., 2006), is regarded as a second order phenomenon. The difference between the two perspectives is partly a question of scale, and each can inform the other.

The thickness of the weathering zone may be expected to depend on factors such as permeability, changes in permeability due to mineral disintegration and decomposition, temperature, biological activity, and erosional processes. However, there is currently no process based theory to predict this relationship. Hilley and Porder (2008) propose that the supply rate of minerals to the weathering zone is proportional to the downward propagation rate of the weathering zone, which is dynamically coupled to the rate at which erosion removes material from the surface. This is necessarily the case since a long-term imbalance in these rates would lead to either the ubiquitous exposure of unweathered rock or the development of an infinitely thick weathering zone. Based on an exponential weathering zone model and constrained by available data, Hilley et al (2010) propose the relationship between weathering zone thickness and erosion rate shown in Figure 8-7. The data include observations from the Himalaya and New Zealand’s Southern Alps which show that at erosion rates greater than 2 mm/yr (2 km/Myr) the weathering zone is less than 0.5 m thick. Additional constraints are
provided by published estimates of soil thickness and erosions rates from the Southern Appalachians, Puerto Rico and the Amazon Basin. Due to the small number of sites with available data, there is considerable uncertainty, but Hilley and Porder (2008) suggest this is accounted for by the generous upper and lower bounds.

Figure 8-7: Concept of weathering depth and equilibrium or transient conditions applied to the Katonga Valley (modified after Hilley et al., 2010)

It can be seen from Figure 8-7 that the maximum depth of weathering observed on the East African Plateau adjacent to the Katonga Valley of 116 m is only consistent with optimum weathering conditions and very low erosion rates similar to those observed in the Amazon Basin. In contrast, the current modal weathering zone thickness adjacent to the Katonga Valley of about 25 m is consistent with a large combination of weathering conditions and erosion rates.

The weathering zone model of Hilley et al. (2010) has been annotated in Figure 8-7 to demonstrate hypothetical equilibrium and transient states of weathering zone thickness adjacent to the Mpanga-Katonga Valley System. Prior to uplift of the EAP and
development of the Western Rift, the depth of weathering may have been similar in the areas now occupied by the east and west Mpanga-Katonga Valley System. Point A represents a hypothetical common mean thickness of 65 m that may have existed under more favourable warm humid conditions in the early Miocene as shown in Figure 4-21a (Zachos et al., 2001). Assuming that weathering conditions are less favourable today, we would expect the weathering zone thickness to decrease, even if the erosion rate remains the same. The current weathering zone thickness of 53 m observed in the western Katonga Valley could represent a transient state on the way to a new dynamic equilibrium represented by point B. Closer to the rift flank the erosion rate may also have increased and therefore the average weathering zone thickness of 26 m could represent a transient state on the way to a new dynamic equilibrium represented by point C. In reality, there is currently inadequate data to draw the lines of dynamic equilibrium for different part of the landscape of south west Uganda and therefore we do not know if the currently observed weathering zone thicknesses is in a transient state or dynamic equilibrium. Nevertheless, this conceptualisation demonstrates that the old paradigm of relict weathering zone thickness is an end member of the dynamic equilibrium model that requires low erosion rates relative the timescale under consideration.

It is interesting to note that based on Figure 8-7 and assuming conditions of dynamic equilibrium, the maximum possible erosion rate (under optimum weathering conditions) in the eastern Katonga Valley is about 1 m/Myr (1E-3 mm/yr), and the maximum possible erosion rate in the western Katonga Valley is about 60 m/Myr (6E-2 mm/yr). The former rate is extremely low and the latter rate is perhaps consistent with erosion within the scale of the current landscape relief on a Pleistocene timescale. In order to compare these observations with the exhumation history presented in Section 8-2, it is worth noting that an intermediate erosion rate of 10 m/Myr (1E-2 mm/yr) would be required to remove the 2 km of Karoo-age sedimentary cover in 200 Myr.

### 8.5.3 Lithospheric flexure due to sediment loading

The potential magnitude of lithospheric flexure in response to sediment loading of George Basin and Victoria Basin has been examined by analogy to a semi-infinite beam. The approach is adapted from civil engineering where the flexure of elastic
beams has been studied extensively. A summary of the approach was given in Section 4.2.2 and the details are given in Watts (2001). The lithospheric flexure predicted using the semi-infinite beam model has been estimated independently for the George Basin and the Victoria Basin using the parameters listed in Table 8-2. It is intended to be a simple scoping exercise, rather than a reliable calculation of the actual lithospheric flexure. Reasonable values of basin half-width, sediment and water thickness, and density were used to estimate the end load on a semi-infinite beam of unit width. The approach was to select an effective lithospheric elastic thickness for each basin which produces a wavelength of flexure consistent with the topographic profile between the George Basin and the Victoria Basin shown in Figure 8-8.

<table>
<thead>
<tr>
<th>PARAMETER</th>
<th>GEORGE BASIN</th>
<th>VICTORIA BASIN</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lithosphere Young’s modulus (GPa):</td>
<td>65</td>
<td>65</td>
</tr>
<tr>
<td>Lithosphere Poisson’s ratio (-):</td>
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<td>0.25</td>
</tr>
<tr>
<td>Mantle bulk density (kg/m3):</td>
<td>3,300</td>
<td>3,300</td>
</tr>
<tr>
<td>Sediment bulk density (kg/m3):</td>
<td>2,400</td>
<td>2,400</td>
</tr>
<tr>
<td>(S.G. = 2.65, porosity = 0.15)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Basin half width (km):</td>
<td>12.5</td>
<td>125</td>
</tr>
<tr>
<td>Sediment thickness (m):</td>
<td>1,000</td>
<td>50</td>
</tr>
<tr>
<td>Acceleration due to gravity (m/s²):</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>Depth of water (m):</td>
<td>n/a</td>
<td>50</td>
</tr>
<tr>
<td>Water density (kg/m3):</td>
<td>n/a</td>
<td>1,000</td>
</tr>
<tr>
<td>Calculated end load (Nm):</td>
<td>3.00x10¹¹</td>
<td>2.13x10¹⁰</td>
</tr>
<tr>
<td>Selected elastic thickness, ( T_e ) (km):</td>
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<td>30</td>
</tr>
<tr>
<td>Calculated flexural rigidity, ( D ) (Nm)</td>
<td>5.78x10²¹</td>
<td>1.56x10²³</td>
</tr>
<tr>
<td>Calculated lambda, ( \lambda ) (m⁻¹)</td>
<td>2.49x10⁻⁵</td>
<td>1.09x10⁻⁵</td>
</tr>
<tr>
<td>Maximum bulge (m):</td>
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<td>3.5</td>
</tr>
<tr>
<td>Distance to maximum bulge (km)</td>
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<td>220</td>
</tr>
</tbody>
</table>

Table 8-2: Summary of assumed and calculated parameters used to estimate lithospheric flexure
Figure 8-8: Comparison of eastern rift flank topography and simple model of lithospheric flexure due to basin loading

Although the calculations are based on two separate semi-infinite beams, the wavelength of flexure and the distance between basins is such that it is considered reasonable to combine the graphical results for ease of comparison with the actual topography as shown in Figure 8-2. An elastic thickness of 10 km is required to achieve a similar wavelength to the observed rift flank topography in the west, and an elastic thickness of 30 km is required to achieve the observed wavelength of the Victoria Basin in the east. These values are consistent with the lower elastic thickness expected in areas of increased thermal gradients and reduced lithospheric thickness. The overall trend of larger elastic thickness near the centre of the East African Plateau is consistent with lower geothermal gradients in the centre of the Tanzania craton (Kohn et al., 2005). However, the estimated elastic thickness of 10 km calibrated against the rift flank topography is less than the crustal thickness of 32 km estimated by Wölbern et al. (2010) from receiver-function analysis of seismic data in the same region (Figure 4-11).

Whilst care must be taken not to over-interpret the results, the scoping calculations presented here do show that calibration of the semi-infinite beam model against the topography between the George and Victoria Basins can be achieved with reasonable estimates of loading and elastic thickness. Given that the loading in George Basin is
based on the maximum likely sediment thickness of 1 km (Upcott et al., 1996) and the elastic thickness of 10 km calibrated against the topographic wavelength is close the minimum reasonable value, the estimated rift flank uplift of 111.5 m will be close to maximum probable amplitude due to flexure alone. It can be seen from Figure 8-8, that the predicted flexural bulge on the rift flank is of a similar order of magnitude to the current topographic relief above the rift boundary fault itself. This is consistent with the conclusions drawn from the depth of weathering data presented in Section 8.5.2. The bulge due to loading of the Victoria Basin of 3.5 m is negligible. Nevertheless, calibration of the semi-infinite beam model (Figure 8-4) against the topography and loading of the Victoria Basin itself does give a reasonable value for the elastic thickness of 30 km.

8.5.4 The influence of tectonics and erosion on the water divide

Figure 8-9 presents a conceptual model of the tectonic geomorphology between the Western Rift and the Victoria Basin. The projected surface elevations and the implied deflections and denudation are intended to give a general indication of their likely trends and order of magnitude rather than detailed quantitative representations. Given that the highest features of the ‘Buganda Surface’ formed from the Buganda-Toro System quartzites immediately north of Lake Victoria are about 1300 m asl and the mid-Miocene plateau elevation estimated by Wichura et al. (2010) based on the Yatta lava flow in Kenya is 1400 m asl, the best estimate for the maximum former plateau elevation in south west Uganda is between 1300 m asl and 1400 m asl (Figure 8-9, annotation 1). The age of the Yatta lava flow, 13.5 Ma, suggests that uplift of the EAP occurred before the mid-Miocene; Wichura et al. (2010) have suggested that it took place between 17 and 16 Ma.
Figure 8-9: Conceptualisation of the tectonic geomorphology along the Mpanga-Katonga Valley System between the rift flank and Victoria basin

The bedrock high and break in the valley slope at Kyai suggests the water divide on the Mpanga-Katonga Valley System was initially formed in this area, most likely due to early downwarping of the Victoria Basin (Figure 8-9, annotation 2). Whilst the semi-infinite beam calculation indicated that loading of the Victoria Basin is sufficient to produce significant deflection, the initial downwarping that facilitated sediment and surface water accumulation must have been related to other tectonic processes. The cause of this downwarping is unclear but could be related to thermal sag and the large scale dynamics of an underlying mantle plume. Another possibility could be compression of the EAP following the development of active extensional rifting to both the west and east of the Tanzanian Craton. Based on sediment thickness in the Victoria Basin of 60 m and an average sedimentation rate of about 150 mm/1000 yrs, Johnson et al. (2000) estimated that the Lake Victoria is about 400 Ka. This suggests that downwarping of the central area of the Tanzanian Craton occurred long after uplift of the EAP in the mid-Miocene. The difference in elevation between the bedrock surface beneath sediment at Kyai (∼1160 m asl) and in the centre of Lake Victoria (∼1020 m asl) is about at 140 m. The semi-infinite beam model predicts that sediment and surface water accumulation in the Victoria Basin could have enhanced deflection by up to 50 m, which is a little over one third of the total estimated downwarping.
The relatively deep valley between Bihanga Station and Nkonge to the west of Kyai was likely partly, if not wholly, cut into the Karoo-age sediments filling the palaeovalley before downwarping of the Victoria Basin, at a time when the entire River Katonga system east of the Western Rift continued to flow westward. Erosion could have been enhanced by increased stream power due to a low base level in the Western Rift. Sediment loading of the George Basin would eventually have produced lithospheric flexure, which together with other mechanisms in this complex tectonic setting, created rift flank uplift. Surface uplift facilitated erosion which created further isostatic uplift and drove a positive feedback system supplying additional sediment to the George Basin. Sometime after creation of the Victoria Basin in the mid-Pleistocene, the rate of rift flank uplift overtook the rate of erosion and a surface water divide was created at Bihanga Station. The western Katonga Valley filled with fine-grained lacustrine sediment and a lake may have formed at this time. Surface water then overflowed towards the Victoria Basin.

8.6 Landscape Evolution of the Katonga Valley

Global tectonics converts geothermal energy into potential energy during uplift and solar energy drives the climate that facilitates denudation. Material is eroded from the upland landscape and stored in the lowland landscape producing unloading and loading which drives flexurally modified isostatic adjustment of the lithosphere. This section considers the combined influence of tectonics and climate on the evolution of the Mpanga-Katonga Valley System by examining the changing drainage pattern in plan-view as shown in Figure 8-10. However, it is important to remember that the landscape is actually evolving in four dimensions. In other words, the relative vertical landscape distribution and the plan-form are both evolving through time as depicted in the block diagrams shown in Figure 8-11.
Figure 8-10: Evolution of the Mpanga-Katonga Valley System in plan-view
Figure 8-11: Cartoon of the evolution of the Mpanga-Katonga Valley System

a) Westward flowing river with stable base-level and denuded landscape

b) Victoria basin downwarps create eastern drainage divide. Western Rift downthrow initiates river capture and rejuvenates the western landscape.

c) Rift flank uplift creates the western drainage divide and erosion continues west of divide. Reduced gradient east of the divide initiates paludal/lacustrine deposition. Raised lake level in eastern valley initiates lacustrine deposition.

d) Erosion continues west of drainage divide. Western valley filled with fine grained sediment and papyrus wetlands established. Lake regression exposes fluvial lacustrine sediment in eastern valley.
8.6.1 The westward flowing River Katonga

Whilst the structural trends in the Rwenzori Fold Belt may exert some influence over the direction of tributaries to the River Katonga (Figure 3-6), on balance, the westward pointing arrow-barb pattern of the Katonga, Kagera and River Kafu systems continue to provide strong evidence for the initial westward flow direction first recognised by Simmons in the 1920s and reported by Wayland (1931). Figure 8-10a shows the likely pattern of the westward flowing River Katonga on a DEM image of the catchment area, prior to downwarping of the Victoria Basin and rift flank uplift. A cartoon of the topography and drainage pattern of the westward flowing River Katonga is shown in Figure 8-11a. In the east, the pattern of swamps and embayments along the north shore of Lake Victoria suggest that the proto-Katonga drainage system continued eastward across the area now occupied by the lake. However, seismic evidence indicates that sediment accumulation up to 60 m thick occurred following initial downwarping of the Victoria Basin (Johnson et al., 1996). This would have obscured evidence of the early drainage pattern and casts doubt on Temple’s (1966) interpretation of purported lake-bed channels.

To the west of the current water divide at Bihanga Station, the Katonga Valley and Karoo outlier within the valley continue towards the west, whilst the Mpanga-Rusangwe-Nyaitanga valley system meets the Katonga Valley from the south west (Figure 3-16). Today, the westward extension of the Katonga Valley comes to an end where it meets the north-south valley of the River Mpanga. However, the geomorphology and geology suggests that prior to rift flank uplift it may have continued westward, most likely south of the Singo Series quartzite of Kabuga Hill, along the approximate route of the Kampala-Kasese railway line. Westward river flow likely continued even after the formation of the Western Rift, before significant rift flank uplift had occurred. It is important to note that although the interpreted drainage pattern is superimposed on the current landscape in Figure 8-10a, the elevation of the proto-Katonga Valley west Kyai was likely related to a higher landscape that has since been eroded, as can be seen comparing Figure 8-11a and b.
8.6.2 The denuded landscape and high level Pleistocene deposits

Examination of the digital elevation model of the central Katonga Valley shown in Figure 8-12 reveals a smooth landscape to the east and south of Kyai, compared to a rough landscape adjacent to the Nabakazi Valley and the western Katonga Valley.

Figure 8-12: Landscape features of the central valley and section locations

Figure 8-13: Sections through the central Katonga Valley
Figure 8-13 presents five north-south profiles across the Katonga Valley at the locations shown in Figure 8-12. Sections 3, 4 and 5 are drawn through the smooth landscape and reveal the maximum elevation (excluding the resistant Singo Series quartzite of Kisozi Hill) to be between about 1,225 and 1,270 m asl, which is higher than the maximum historical level of Lake Victoria (1,190 m asl) based on strandline elevations (Temple, 1966). Therefore whilst the smooth landscape coincides with the region of extensive valley fill adjacent to Lake Victoria, we can exclude the possibility that the highest surfaces were once drowned beneath the lake. The smooth characteristics of the eastern landscape suggest that it represents late-stage denudation in response to a long-established base level compared to the rougher western landscape which formed in response to a more recent base level change in a westward draining catchment. It can therefore be inferred that the smooth denuded landscape to the east also likely formed when the River Katonga flowed west. It is interesting to note that this newly identified denuded eastern landscape coincides with the area of the Buganda Surface identified by previous authors (see Figure C-2, Appendix C).

The sands, grits and gravels identified in the valleys and interfluves of the central Katonga and Nabakasi rivers (Johnson, 1958) between about 1,185 and 1,270 m asl remain one of the most enigmatic features of the landscape of south west Uganda. It is perhaps significant that these deposits appear to occur on the western margin of the older eastern denuded landscape (Figure 8-11). Johnson (1960) also described coarse quartz grit and gravel between 1,190 and 1,209 m asl on the north side of the wide eastern section of the Katonga Valley, which he suggested are raised beach deposits marking the highest but poorly developed lake strandline. Although Johnson (1960) observed both the gravels in the Nabakasi Valley and the high strandline deposits in the eastern Katonga Valley, he did not infer a connection between them.

Sand and gravel deposits and terraces have also been described in the Kafu (Hirst, 1926, Wayland, 1930, Bishop, 1969), Muzizi (Wayland, 1925), and Kagera (Wayland, 1934b, Lowe, 1952, Spurr, 1955, Bishop, 1969) valleys. Unfortunately, the facies characteristics such as angularity, sphericity, surface texture and composition of the gravel were not recorded at these locations which make it difficult to tell if they have the same origin as the deposits which form the terraces at Kyai in the Katonga Valley. The highest gravel identified adjacent to the Kagera Valley occurs on a hill top at
Nkurungu between 1,265 to 1,274 m asl and 61 to 79 m above the river (Bishop, 1969). This is at least 41 m higher and 18 km east of the Boulder Bed at Nsongezi, which occurs at about 1,224 m asl (Figure 4-42). The occurrence of the Nkurungu hill top gravel shows that the interpretation of the Boulder Bed at Nsongezi as having been deposited before river reversal (Bishop, 1969) on the basis of grain size alone is overly simplistic. The history of deposition of the sand and gravel deposits in south west Uganda appears more complex than previously recognised and further evidence is likely required before a satisfactory explanation can be provided.

Johnson (1958) describes the alluvial deposits in the Nabakasi Valley as distinctly bedded, angular quartz grits and water worn pebbles, cemented by red earthy material. Similar deposits were observed by the author forming terraces in the Katonga Valley at Kyai where they are described as light grey, weakly to moderately cemented, sub-angular to sub-rounded, coarse sand to medium gravel with coarse gravel and cobbles (Section 4.5.4). Early researchers generally assumed that the gravels of south west Uganda were deposited during a wetter climate at a time when the rivers flowed towards the west and stream power was greater (Hirst, 1926, Wayland, 1930). However, at Kyai it is noted that although the coarse grain size does indicate they were transported in a high energy environment, the sub-angular shape also suggests they were not transported far. These characteristics are often associated with flash flood-like deposits. Whilst the topographic setting is not one generally associated with alluvial fans, it seems plausible that the sub-angular sand and gravel was deposited at a time when the general climate was drier, during high intensity, low frequency rainfall events. The lower density vegetation cover associated with an arid climate would have reduced cohesion of the surface soils and facilitated erosion of the already weathered regolith. Fine-grained silt and clay was washed out, leaving only the coarse-grained material behind. The southerly dip direction observed in the terrace on the south side of the Katonga Valley at Kyai suggests the material was transported from the north. Therefore, it may have been contiguous with the interfluve deposits, and together they once blanketed the landscape and filled the central Katonga Valley. However, it is noted that since the interfluve deposits were not observed during this study, it has not been possible to assess the reliability of the assumption made by Johnson (1958) that they are related to the terraces at Musozi Station in the Nabakazi Valley. The weathered and weakly
cemented character of the sand and gravel at Kyai gives the general appearance of pre-dating the Late Pleistocene (>100 ka?). However, the characteristics of the sand and gravel also make them unsuitable for dating by optically stimulated luminescence and no archaeological evidence has been found that dates the central Katonga or Nabakasi deposits.

Although the detailed age and origin of the high level coarse-grained sediments is uncertain, the sand and gravel in the central Katonga Valley and elsewhere, provides strong evidence of significant environmental change during the late Neogene. The Neogene in East Africa is characterised by an overall drying trend, albeit with high climate variability about this trend (Section 4.3.2). Coupled ocean-atmosphere models have indicated that reduced precipitation in tropical Africa is associated with a general weakening of the hydrological cycle, and the Afro-Asian monsoon in particular, due to global cooling. De Gasse (2000) has pointed out that most African palaeohydrological records suggest dry conditions during the Last Glacial Maximum. It is interesting to compare this modern interpretation with Wayland’s (1930) simplistic link between glacial high latitudes and pluvial low latitudes and the reverse model of ‘cool north-dry tropics’ may be better supported (Stager et al., 2002). Thus, if the sub-angular sand and gravel is indeed associated with a more arid climate, they may well have been deposited during one or more Pleistocene glacial periods, when the equatorial climate was drier.

8.6.3 Victoria Basin downwarp and initial river reversal

The north-south profiles of the central Katonga Valley presented in Figure 8-13 show that the narrow and deep central Katonga Valley on section 4 (~25 m deep between Kyai and Kisozi) and section 3 (~50 m deep near Kyai) cuts into a broad relict high-level valley floor. The narrow and deep central valley appears to be a continuation of the channel observed by the ERT below the papyrus wetland at Kisozi (Figure 8-2). The broad high-level valley floor appears to be a continuation of the terrace flood plain at Kisozi (section 5 in Figure 8-13). The western Katonga Valley is about 50 m lower than the former high-level valley floor at Kyai. The ERT survey also indicated that the bedrock beneath the floor of the modern, narrow, valley at Kyai is higher than the bedrock beneath the floor of the western Katonga Valley at Kabagole, and the eastern Katonga Valley at Kisozi. The central narrow valley is cut deep enough to create a
small but continuous eastward gradient towards Lake Victoria, although the longitudinal profile does show a break of slope near Kyai (Figure 2-2b). The interfluve deposits north of Kyai, shown on section 3 in Figure 8-13, sit on the high-level surface. The long wavelength of the eastern denuded landscape shown north of the Katonga Valley on section 4 can be compared with the short wavelength western rejuvenated landscape shown in section 5.

The presence of the former broad high-level valley and the apparent bedrock high beneath the deep, narrow valley at Kyai, together with the break of slope in the longitudinal profile all suggest that the elevation of the Victoria Basin has decreased relative to the central Katonga Valley. It is proposed that the relative downwarp of the Victoria Basin created the initial drainage divide in the originally westward flowing River Katonga as shown in Figure 8-10b and Figure 8-11b. The lower elevation of the western Katonga Valley relative to the former high-level central valley is explained by continued erosion by the westward flowing river system to the west of the initial drainage divide at Kyai.

Following drainage reversal, the water level in the Victoria Basin began to rise. Lateral erosion by the eastward flowing River Katonga widened the eastern valley and the mouth of the Katonga took on the characteristics of a freshwater estuary during times of high lake levels. Based on sediment thickness beneath Lake Victoria and rate of deposition, fluviolacustrine sediment accumulation appears to have began around the 400 ka in the mid Pleistocene (Johnson et al., 1996). Climate variation associated with high latitude glacial cycles caused the lake level to fluctuate. Johnson (2000) suggested that the three interruptions in sedimentation identified from seismic reflection surveys may be related to the post Pleistocene Revolution 100 ka glacial-interglacial cycles.

### 8.6.4 George Basin downthrow, river capture and rejuvenation

The valley form around the modern drainage divide strongly suggests that the River Katonga was captured from the south west. It seems plausible that northward propagation of the George Basin resulted in greater downthrow along the rift boundary fault to the south. This reduced the local base level in the George Basin and increased the stream power on the rift flank in the south-westward flowing Mpanga-Rusangwe-Nyaitanga river system, which in turn was able to capture the River Katonga near the
modern location of Bihanga Station (Figure 8-10b). This in turn led to rejuvenation of the Katonga catchment to the west of the initial drainage divide near Kyai created by downwarp of the Victoria Basin (Figure 8-11b). This explains the increased landscape roughness and stream order observed on the DEM in the western tributaries of the Katonga Valley and in the Nabakazi Valley in particular shown in Figures 8-10b and 8-12. It may also have played a role in producing the superimposed drainage that dissect the quartzite ridges. The rejuvenated westward flowing river likely exposed the relatively unweathered surface of Karoo sediments in the western valley at this time. The fact that the interfluve deposits appear to have been dissected by the rejuvenated River Katonga is evidence that they were deposited before river reversed had occurred in this part of the valley. The knickpoints in the rejuvenated landscape migrated eastwards (Figure 8-10a) towards the initial drainage divide on the smooth denuded landscape (Figures 10b). One such knickpoint may have cut into the former high-level valley and begin to create the terraces observed in the central valley near Kyai.

8.6.5 Rift flank uplift and late Quaternary climate fluctuations

Sometime after formation of the Victoria Basin in the mid Pleistocene, rift flank uplift created a new surface water divide approximately 50 m east of the rift boundary fault and 100 km west of Kyai, near its current location at Bihanga Station as shown in Figure 8.10c and Figure 8-11c. West of the new drainage divide, the Mpanga River became the main tributary from the north and the Rusangwe River became the main tributary from the south. The Katonga Valley around Kabagole may have become a finger lake at this time which gradually filled with fine-grained deposits until paludal conditions were established (Figure 8-11c).

Although uncertainty remains about the exact origin of the subangular sand and gravel at Kyai, it is clear that at some point the River Katonga was able to erode a new channel through these coarse-grained deposits. One possible explanation is that increased precipitation occurred immediately following an arid climate cycle when the base level in the Victoria Basin was still low. This could have provided adequate stream power to erode the sand and gravel and create the terraces we see today. This reconnected the west and east Katonga Valley and facilitated drainage towards Lake Victoria as shown in Figure 8-10c and Figure 8-11c. Fluctuating lake level during the Late Pleistocene
resulted in the deposition of fluviolacustrine deposits in the eastern Katonga Valley (Figure 8-11c). The last lake low-stand between 15.9 and 14.2 ka (Stager and Johnson, 2008) resulted in the erosion of the channel in the fluviolacustrine sediments identified by the ERT at Kisozi. The return of a more humid climate and higher lake levels during the Holocene resulted in the papyrus wetlands being established and the onset of fine-grained paludal deposition throughout the Katonga Valley. Figure 8-11d shows a cartoon of the Katonga Valley today with extensive papyrus wetlands and exposed fluviolacustrine deposits in the eastern valley.

8.7 Hydrogeology of the Katonga Valley

8.7.1 Hydrogeology of the western and central valley

Deeply weathered regolith and near-surface fractured rock have long been known to provide an important source of locally available fresh water for widely dispersed rural communities in sub-Saharan Africa (Clark, 1985, Jones, 1985, Chilton and Foster, 1995). Taylor and Howard (1999a, 2000) recognised that the tectonic geomorphology of south west Uganda influences the spatial distribution of erosion rates and hence the current thickness of the weathered mantle. They showed that the Rusangwe catchment near the rift valley to the west of the Mpanga-Katonga surface water divide has a reduced weathered rock thickness with consequently smaller transmissivity, basin storage and recharge/run-off ratio compared to the Aroca catchment to the east of the Nkusi-Kafu surface water divide north west of Lake Kyoga. Even so, the yield and water quality accessible from an individual well in deeply weathered and near-surface fractured rock is often difficult to predict and the distribution sometimes appears stochastic in character. If present in the Katonga Valley, coarse-grained alluvial sediment could represent local sources of groundwater with a predictable distribution.

Tindimugaya (2008) suggested that transmissive alluvial deposits may exist in the relict river valleys of south west Uganda. The initial motivation for this thesis was therefore to investigate the geomorphology of the Katonga Valley and determine the groundwater resource potential of associated alluvial deposits. However, rather than a simple alluvial aquifer, the complex geological history and landscape evolution of the region has led to a spatially variable sequence of erosion surfaces and depositional units.
It has been shown that the western Katonga Valley is exploiting an exhumed Gondwanan palaeovalley containing an outlier of Karoo-age sedimentary rocks. At Bihanga Station near the water divide, these comprise mudstone, siltstone, sandstone and diamictites of unknown overall proportions. Although a hand pumped well was observed sunk through the Karoo sediments at Bihanga Station, it has not been possible to acquire the borehole log and pumping test data for this well. Unfortunately, pumping test data from the borehole DWD25847 in the Katonga Wildlife Reserve near Kabagole also failed to provide the hydrogeological properties of the Karoo-age sediments. Therefore, whilst it appears that the Karoo-age rocks have been used as a source of water for at least one local community in the western Katonga Valley it has not been possible to quantify their resource potential during this study. Nevertheless, on the basis of field observation, the ERT survey and some reasonable hydrological assumptions it has been attempted to construct an initial hydrogeological conceptual model of the Katonga Valley at Kabagole, as shown in Figure 8-14.

Figure 8-14: Cartoon of the hydrogeological conceptual model at Kabagole

Figure 8-14 shows infiltration via transmissive pathways through the upper clayey part of the weathered regolith in the gneiss and the granite on the valley sides to the north and south. Porous flow occurs through channels comprised of disintegrated granular material at the base of the saprolite towards the Katonga Valley, where it enters the
sandy silt alluvium. Limited groundwater flow may also occur in transmissive fractures through the bedrock and directly into the Karoo-age fine-grained sandstone.

The electrical conductivity of the alluvium that fills the Neogene relict valley at Kabagole is higher than both the Karoo-age fine-grained sandstone at Kabagole and the fluviolacustrine silty sand at Kisozi. This suggests that it likely has a higher silt and clay content than the moderate transmissivity fluviolacustrine deposits at Kisozi. Although this inference reduced the perceived cost effectiveness of a hydrogeological drilling investigation in the alluvium at Kabagole for this project, it remains to be corroborated or falsified by future drilling at the site. For the purpose of the conceptual model shown in Figure 8-14 it is assumed that the hydraulic conductivity of the fine-grained sandstone is greater than the sandy silt alluvium.

Although the longitudinal topographic gradient is low in the western valley, surface water can be observed flowing downstream through the papyrus wetland. The conceptual model assumes that the surface water flow rate is greater than the groundwater flow rate through the fine-grained sandstone. The geometry and continuity of the sandstone is unknown downstream. It may be bounded by lower hydraulic conductivity alluvium or crystalline bedrock as the valley becomes shallower towards the central Katonga Valley. If any groundwater flows downstream in the fine-grained sandstone it must therefore eventually enter the weathered crystalline rocks or emerge as surface water flow in the central Katonga Valley upstream of Kyai. Another potentially significant water output from the western Katonga Valley is evapotranspiration from the papyrus wetland.

As we have seen, sand and gravel has been identified in the central Katonga Valley at Kyai. However, its origin is uncertain and its sub-angular character, southward dipping beds, and proximity to high level gravels on the interfluve to the north, suggest that it may be associated with high intensity, low frequency floods rather than a continuously flowing powerful river system. A later channel has been eroded through these deposits and they only survive as terraces above the water table. Consequently they have no known groundwater resource potential in the Katonga Valley. The research presented here finds that there is no current evidence for a substantial basal sand and gravel
aquifer in the Katonga Valley as less detailed analyses had previously speculated (Tindimugaya, 2010).

8.7.2 Hydrogeology of the eastern valley and fluviolacustrine sediment

It has been shown that the eastern Katonga Valley contains fluviolacustrine silty sand, likely deposited under conditions of late Quaternary climate variability and fluctuating lake levels. The groundwater resource in the fluviolacustrine deposits in the eastern valley and adjacent to Lake Victoria have been exploited for village and town water supplies for several decades. Previous investigations have focused on the short term yield achievable by individual wells. This section first proposes a hydrogeological conceptual model of the Katonga Valley at Kisozi, going on to characterise the transmissivity of this fluviolacustrine hydrostratigraphic unit as a whole and compare it with that of weathered and fractured rocks.

The perimeter of the topographic surface that indicates the aerial extent of the fluviolacustrine hydrostratigraphic unit has been annotated on the DEM of the eastern Katonga Valley in Figure 8-15. The perimeter of the papyrus wetland identified by slightly lower topography is also shown. Quaternary climate variation and ongoing tectonism likely resulted in intermittent transgression and regression of Lake Victoria within this broad and low lying region of the Katonga Valley and its eastern tributaries. The top of the fluviolacustrine deposits at Kisozi attain an elevation between 1170 and 1180 m asl (Figure 8-2), which is between 37 and 47 m above the lake. This raised floodplain or terrace surface appears synonymous with the Kayugi strandline identified by Temple (1966) and descends to about 1158 m, asl or 25 m above Lake Victoria, to the west of Lake Nabugabo. The surface of the papyrus wetland is about 1165 m asl at Kisozi and descends eastward towards the lake level at 1,133 m asl.

Figure 8-16 shows a photographic view from the Singo Series quartzites of Kisozi Hill, looking south west across the papyrus wetland towards the fluviolacustrine terrace and the distant interfluve underlain by weathered gneiss. The wetland, terrace and interfluve surfaces identified on the DEM at Kisozi (Figure 8-11) are all visible in Figure 8-16.
Figure 8-15: Areas of fluviolacustrine sediments and locations of groundwater abstraction wells
Figure 8-16: Geological features of the Katonga Valley shown on a photograph taken from Kisozi Hill

Figure 8-17: Geological features and distribution groundwater yield determined by ERT and borehole logs at Kisozi

Figure 8-17 presents the inversion model produced from the ERT survey between points A and B on Figure 8-16, annotated with a summary of the groundwater yield distribution obtained from drilling logs and lithological logs of Kikuumaddungu Boreholes 2 and 3. These boreholes are located 100 m east and 350 m west of the survey respectively, on the south side of the wetland. The total recorded yield is 2 m$^3$/hour from Borehole 2 and 5 m$^3$/hour from Borehole 3 and the transmissivity calculated for each well is 1.6 m$^2$/day and 7.6 m$^2$/day respectively. Due to adjustments
in the flow rate during the pumping tests the detailed flow geometry could not be characterised and it was assumed to be infinite acting radial flow. The driller’s logs which were reproduced in Table 4-12 are not detailed and require additional interpretation (see Section 4.5.4). As shown in Figure 8-17, both ERT and borehole logs identify a surface layer of fine-grained sediment up to 6 m thick which is contiguous with the low resistivity sediment filling the channel beneath the papyrus wetland. The ERT survey is able to differentiate the fluviolacustrine sediments from the paludal sediment and the Singo Series bedrock, but due to the reduction in resolution with depth is unable to identify the interface between the sediment and the weathered rock identified in the borehole logs.

About 30% of the yield from the two boreholes is derived from the main body of the silty sand which is about 49 m thick. Between 10 and 20% of the yield is derived from the underlying 30 to 40 m of weathered rock. However, between 50 and 60% of the yield is derived from close to the interface between the sediment and the weathered rock. The borehole logs both record ‘quartz’ at the base of the sediment, which is assumed to mean quartz gravel or cobbles. The underlying bedrock is described as ‘weathered rock’ in Borehole 2 and ‘phyllites’ in Borehole 3. Given that decomposed phyllite will contain a high proportion of clay minerals, it seems likely that the enhanced yield near the bedrock interface is derived from the overlying coarse-grained sediment. Nevertheless, due to the quality of available data, there is uncertainty in this interpretation.

The geometric mean transmissivity derived from pumping tests in the two boreholes is 3.5 m²/day. The detailed vertical distribution of the transmissivity throughout the fluviolacustrine sediments is unknown. If the majority of the yield is derived from a 5 cm thick layer, the hydraulic conductivity of that layer would be 70 m/day, which is commensurate with coarse sand and gravel. However, if the yield is dispersed throughout 50 m of sediment, the equivalent hydraulic conductivity of the whole hydrostratigraphic unit would be 0.07 m/day, which is commensurate with silt. The range of hydrogeological properties determined for the fluviolacustrine sediments is discussed further in the next section.
Figure 8-18: Cartoon of the hydrogeological conceptual model at Kisozi

Figure 8-18 presents a simplified hydrogeological conceptual model of the Katonga Valley at Kisozi. Similarly to the conceptual model at Kabagole, recharge is proposed to occur via transmissive pathways through the upper clayey part of the weathered crystalline rocks on the valley sides. Some infiltration may also occur through floodplain deposits. Groundwater flow towards the Katonga Valley occurs through channels comprised of disintegrated granular material at the base of the saprolite and limited flow may also occur in transmissive fractures through the bedrock. The Singo Series quartzite to the south is less susceptible to weathering and groundwater likely occurs through joints and fractures. The internal architecture of the fluviolacustrine silty sand is largely unknown, although the borehole logs suggest that a thin layer of transmissive gravel occurs near the base. Other transmissive layers or lenses may be distributed within the fluviolacustrine hydrostratigraphic unit.

Surface water can be observed flowing downstream through the papyrus wetland. The conceptual model assumes that there is also a moderate downstream groundwater flux near the interface between the fluviolacustrine deposits and the underlying weathered bedrock, probably predominantly within the coarse-grained sediment. The downstream flux is the cumulative rate of groundwater entering the fluviolacustrine aquifer upstream, minus any water loss to evapotranspiration via the papyrus wetland.
8.7.3 Groundwater resource potential of the fluviolacustrine deposits

Chapter 7.0 presented the first consistent sets of pumping test analyses for the fluviolacustrine hydrostratigraphic unit and the weathered and fractured bedrock of south west Uganda. The results are used here to assess the transmissivity range of the fluviolacustrine sediment and place it the context of the transmissivity of common lithologies and the requirements of local groundwater supplies. It is noted that the boreholes in which the tests were conducted were not designed to test the sediment and the weathered and fractured bedrock separately. Those boreholes which have been classified at representative of weathered and fractured bedrock were not overlain by sediment (Table 7-1). However, those boreholes which have been classified as representative of the sediment were generally also sunk into underlying bedrock (Table 7-2). Where data are available, the majority of the borehole yield was associated with the sediment and therefore the calculated transmissivity is assumed to be representative of the sediment. This is not ideal, but it does enable a first pass examination of the relative transmissivity range.

The average transmissivity and range of transmissivity for the pumping tests conducted in weathered/fractured rock and the fluviolacustrine sediment are summarised in Tables 8-4 and 8-5 respectively. It should be remembered that these results are based on tests in boreholes selected for groundwater abstraction using motorised electric pumps. Therefore, the statistics are not necessarily representative of the distribution of hydrogeological properties for these units as a whole, and are likely biased towards the more transmissive ‘tail’ in the distribution.

The effective hydraulic conductivity has been calculated by dividing the transmissivity estimate for each test by the maximum thickness of the potentially contributing lithological unit in either weathered and fractured rock, or fluviolacustrine sediment. Hydraulic conductivity is based on the concept of an ideal homogeneous porous medium. In most cases, the actual flow will originate from a few relatively thin fractures or layers. Therefore, the hydraulic conductivity should be regarded as a theoretical average only, integrated over the entire thickness of the hydrostratigraphic unit for purposes of normalising the transmissivity per unit thickness and facilitating comparison.
Table 8-3: Summary of results for boreholes mainly in weathered and fractured rock

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Log T (log m²/d)</th>
<th>T (m²/d)</th>
<th>Log K (log m/d)</th>
<th>K (m/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arithmetic mean of log T</td>
<td>0.60</td>
<td>3.94</td>
<td>-1.01</td>
<td>0.10</td>
</tr>
<tr>
<td>Geometric mean of T</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Standard Error</td>
<td>0.11</td>
<td></td>
<td>0.11</td>
<td></td>
</tr>
<tr>
<td>Standard Deviation</td>
<td>0.33</td>
<td></td>
<td>0.33</td>
<td></td>
</tr>
<tr>
<td>Minimum</td>
<td>0.16</td>
<td>1.43</td>
<td>-1.38</td>
<td>0.04</td>
</tr>
<tr>
<td>Maximum</td>
<td>1.22</td>
<td>16.75</td>
<td>-0.52</td>
<td>0.30</td>
</tr>
<tr>
<td>Count</td>
<td>9</td>
<td></td>
<td>9</td>
<td></td>
</tr>
</tbody>
</table>

Table 8-4: Summary of results for boreholes mainly in fluviolacustrine sediment

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Log T (log m²/d)</th>
<th>T (m²/d)</th>
<th>Log K (log m/d)</th>
<th>K (m/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arithmetic mean of log T</td>
<td>1.16</td>
<td>14.30</td>
<td>-0.36</td>
<td>0.43</td>
</tr>
<tr>
<td>Geometric mean of T</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Standard Error</td>
<td>0.21</td>
<td></td>
<td>0.23</td>
<td></td>
</tr>
<tr>
<td>Standard Deviation</td>
<td>0.60</td>
<td></td>
<td>0.64</td>
<td></td>
</tr>
<tr>
<td>Minimum</td>
<td>0.20</td>
<td>1.60</td>
<td>-1.48</td>
<td>0.03</td>
</tr>
<tr>
<td>Maximum</td>
<td>2.24</td>
<td>175.10</td>
<td>0.66</td>
<td>4.61</td>
</tr>
<tr>
<td>Count</td>
<td>8</td>
<td></td>
<td>8</td>
<td></td>
</tr>
</tbody>
</table>

High transmissivity features usually have a lower density than low transmissivity features and this often creates a large positive skew to the transmissivity distribution measured in many rock types. In fact, the statistical distribution of both transmissivity and hydraulic conductivity frequently has a log-normal distribution. The geometric mean of the transmissivity and hydraulic conductivity (i.e., the arithmetic mean of log transmissivity and log hydraulic conductivity) is therefore used to provide a representative values in Tables 8-4 and 8-5. Figure 8-19 and Figure 8-20 present histograms of the logarithm of hydraulic conductivity and the logarithm of transmissivity for the fluviolacustrine sediment and the weathered/fractured rock.
Figure 8-19: Comparison of hydraulic conductivity for abstraction wells in weathered/fractured rock and fluvio-lacustrine sediment

Figure 8-20: Comparison of transmissivity for abstraction wells in weathered/fractured rock and fluvio-lacustrine sediments
Given that there are only nine tests in the weathered and fractured rock and eight in the fluviolacustrine sediment, care must be taken not to place too much confidence in any inferences that are drawn. Nevertheless, the normal distribution of the log hydraulic conductivity histogram presented in Figure 8-19 suggests that the test results capture the central tendency for the fluviolacustrine sediment. On the other hand the small range of high hydraulic conductivity values derived from pumping tests in the weathered and fractured rock suggests that the low hydraulic conductivity tail of the distribution has been severely censored. Therefore, the geometric mean hydraulic conductivity of 0.4 m/day for the fluviolacustrine sediment shown in Table 8-5 appears to be a reasonable estimate of the average value for the hydrostratigraphic unit. However, the geometric mean hydraulic conductivity of 0.1 m/day for the weathered and fractured rock shown in Table 8-4 is only the average of those boreholes suitable for installation of submersible electronic pumps. The average hydraulic conductivity is likely to lie far to the left of the data shown Figure 8-19. Plotting hydraulic conductivity, rather than transmissivity allows direct comparison with the expected parameter range for common lithologies (Kruseman and de Ridder, 1990) as shown in Figure 8-19. The range of hydraulic conductivity measured in the fluviolacustrine sediment from 0.03 to 4.6 m/day lies within the expected range for silt and sand, and the geometric mean lies around the boundary of hydraulic conductivity expected for the two grain sizes. From a hydrogeological perspective, the fluviolacustrine deposits therefore have the average properties expected for silty sand.

In many parts of the world silty sand would be regarded as a relatively low productivity aquifer. However, in areas where the bedrock geology is dominated by weathered crystalline basement it may still provide a much needed local source of fresh water. It is useful to place the transmissivities shown in Table 8-5 and Figure 8-20 into the context of the parameters required to sustain hand pumped wells for rural water supplies and motorised pumps for town, agricultural and industrial water supplies in Africa. Macdonald et al. (2005) point out that an average rural water supply for 250 people, each requiring 20 L/day, would need a sustainable yield of 5,000 L/day or \( \approx 0.21 \, \text{m}^3/\text{hr} \). They estimate that with a drawdown of 10 to 15 m this requires a minimum transmissivity of about 1 m\(^2\)/day. Figure 8-20 shows that all of the results of pumping tests analysed for this study were greater than 1 m\(^2\)/day. The final pumping
rate during each of the tests conducted in the fluviolacustrine aquifer varied between about 4 and 25 m$^3$/hr. This is consistent with a survey of well yield in sub-Saharan Africa undertaken by Chilton and Foster (1995), who determined that hand pump wells usually have yield $\geq 0.72$ m$^3$/hr and wells with motorised pumps usually have yields $\geq 3.6$ m$^3$/hr.

Examination of the results from a small sample of pumping tests in weathered and fractured rock of south west Uganda suggests that the average transmissivity of randomly selected well locations is likely to be lower than the minimum value of 1 m$^2$/d required to sustain an electronically pumped well. Expert knowledge of the local geology and geomorphology is therefore particularly important for siting wells in the crystalline basement rocks, and even then, the discovery of a transmissive feature may be subject to chance. However, the results of the small sample of pumping tests in the fluviolacustrine hydrostratigraphic unit suggests that almost all wells sunk in these deposits will have a transmissivity and yield suitable for village water use at least. In additions, some boreholes will encounter much high transmissivity sediments such as the well at Mukono near Lukaya which had a transmissivity of 175 m$^2$/d and a yield of 25 m$^3$/hr.
9 CONCLUSIONS

9.1 Overview

Research on large-scale landscape evolution fell out of favour among geography-based geomorphologists in the 1970s and 1980s and was largely replaced by small-scale process-based studies. This occurred due to a perception that historical qualitative research was less rigorous than quantitative experimental approaches and the need to understand coastal, fluvial and hill slope processes from an engineering perspective. This change in emphasis caused some geomorphologists to lament their lack of ‘enchantment’ with ‘timeless, theoretical and utilitarian problems’ (Baker and Twidale, 1991) and call for geomorphology to reconnect with the landscape. Renewed interest in landscape evolution began in the 1990s among geology-based geomorphologists doing research on tectonic geomorphology and thermochronology. Working within the context of plate tectonic theory, they focused on the role of large-scale denudation and deposition in producing flexurally modified isostatic adjustment of the lithosphere. The recognition of links between tectonic, climatic and geomorphological processes led to the adoption of the Earth system paradigm. Whilst some geography-based geomorphologists have perceived these developments as a threat to their discipline (Church, 2005), others have seen them as an opportunity (Summerfield, 2005). The dialogue continues between those geomorphologists who promote ‘local environmental systems science...concerned with process interaction’ (Richards and Clifford, 2008) and others who believe that ‘large-scale geomorphology is on the threshold of another “golden” era to match that of the first half of the 20th century’ (Bishop, 2007).

This thesis presents the most comprehensive examination of the landscape evolution of the Katonga Valley since the mid 20th century heyday of geomorphological research in the region. It seeks to integrate qualitative data acquired using traditional methods with quantitative data produced using modern research techniques. It incorporates recent theoretical developments, and adopts the place-specific perspective of historical geoscience to understand the modern landscape and achieve the utilitarian goal of improving our understanding of groundwater resources in the alluvial sediment. It incorporates the first low temperature thermochronological study of the eastern margin of the Western Rift between Lake George and the north west shoreline of Lake
Victoria. The field work includes the first recorded electrical resistivity tomography conducted in Uganda using commercial multi-electrode resistivity imaging equipment. In addition, the interpretation includes the first statistical characterisation the transmissivity of the fluvialacustrine hydrostratigraphic unit located in the eastern Katonga Valley and north west shoreline of Lake Victoria.

The research reported here has led to new knowledge about the geology, geomorphology and groundwater resources of south west Uganda. Perhaps the most significant conclusions that can be inferred in each of the three main research areas are:

**Thermochronology** – The Precambrian rocks of south west Uganda were covered by between 2 and 4.5 km of Karoo-age sediment during the early Mesozoic.

**Landscape Evolution** – Reversal of the originally westward flowing River Katonga occurred in two stages. Downwarp of the Victoria Basin created an initial drainage divide about 150 km east of the Western Rift before rift flank uplift later created the present day drainage divide about 50 km east of the Western Rift.

**Hydrogeology** – The complex Neogene tectonics and climatic variability of the region means that the groundwater resource potential of each valley reach should be assessed in the context of its unique history. The fluvialacustrine sediment in the eastern Katonga Valley and north west shoreline of Lake Victoria has a transmissivity range that is usually suitable for village water supply and is sometimes commensurate with town water use.

The following sections discuss the conclusions in terms of the five main objectives and research questions set out in Section 2.5.

### 9.2 Former Karoo Sedimentary Cover

The discovery of fine-grained arkose sandstone at Kabagole in the Katonga Valley led to re-examination of the Katonga Beds at Bihanga Station. Previously, this small outlier of sedimentary rocks (2×10 km) which fills the Katonga Valley at the drainage divide had been described on the 1962 geological map as ‘conglomerates, sandstones and siltstones’ of ‘?Miocene’ age (D.G.S.M., 1962). However, the geological
observations acquired during this study provide strong evidence of their Permo-
Carboniferous glaciogenic origin and their similarity to the Lower Karoo age rocks
outcropping on the Entebbe peninsular and the Lukuga Group of the D.R. Congo.

The lithofacies evidence of the glacial and proglacial origins of the Katonga Beds
includes an unsorted diamictite with angular sandstone matrix and sub-rounded cobbles
resembling a tillite, and rhythmically interlaminated siltstone and mudstone resembling
varves with occasional drop stones. The beds also contain glaciotectonic structures
including: overturned rhythmite blocks on a scale of metres to tens of metres,
juxtaposed against in situ massive diamictite; contemporaneous brittle and ductile
deformation of sandstones and siltstones on a scale of centimetres to tens of
centimetres; and, convoluted soft sediment deformation of the rhythmites.

Palaeontological evidence of the age of the Katonga Beds includes woody stem
fragments with bark-like texture resembling gymnosperms; and, leaf fragments with
parallel venation resembling the Permo-Carboniferous gymnosperm Noeggerathiopsis.

Whilst the Karoo-age rocks of Kenya and Tanzania occur in large grabens located over
1000 km from Bihanga Station on the opposite side of the Tanzanian Craton (Schlüter
et al., 1993) the Lower Karoo strata of the eastern D.R. Congo are found filling an
exhumed Palaeozoic landscape (Boutakoff, 1948, Cahen, 1954, Cahen and Lepersonne,
1981). The closest outcrop, in the Ituri Valley, is located only 200 km north west of
Bihanga Station. Given the lack of evidence for fault boundaries to the Katonga Beds it
is concluded that, similarly to the Lower Karoo deposits of the D.R. Congo they too are
filling an exhumed palaeovalley. Indeed, the straight and wide character of the western
Katonga Valley is reminiscent of a glacial valley.

Exhumed erosional landforms of the late Palaeozoic glaciation have been identified in
Africa as far apart as Ethiopia (Bussert and Schrank, 2007, Bussert, 2010) and Namibia
(Green and Duddy, 2010). In northern Ethiopia, numerous palaeolandforms including
chatter marks, striations, crescent gouges, and roche moutonnées and whalebacks have
been exhumed and indicate the region was occupied by a northward flowing wet-based
continental glacier. Glaciolacustrine claystones and siltstones with drop stones are
found filling glacial valleys several kilometres wide and tens of metres deep (Bussert,
2010). Minor diamictites are interpreted as various types of tillites (Bussert and
Schrank, 2007). In north west Namibia, another exhumed late Carboniferous glacial landscape is partly covered by a few hundred metres of Lower Karoo glaciogenic sediments including tillites, which is in turn overlain by Cretaceous basalt (Green and Duddy, 2010).

Whilst it has been confirmed that the near-horizontally bedded, fine-grained arkose sandstone discovered at Kabagole in the Katonga Valley is similar to the western exposure of the Katonga Beds and the sandstones previously identified in the Muzizi and Nkusi valleys, its age and relationship to the glaciogenic rocks are uncertain. All of these sandstones show considerably greater induration and generally older appearance and characteristics to the Miocene Kisegi Formation in the Western Rift. Indeed, thin sections of the diamictite from the Katonga Beds appear to include lithic clasts of similar appearance to the fine-grained sandstone suggesting that it pre-dates the glaciogenic rocks. It seems likely that both the glaciogenic rocks and the fine-grained sandstone are Karoo in age, although the exact nature of their relative relationship requires confirmation by further evidence. The small Karoo-age outcrops of Uganda have previously been neglected (Schlüter et al., 1993) but when viewed together with the continental distribution of Karoo strata (Figure 3-10) suggest that prior to the break-up of Gondwana there was a wide-spread sedimentary cover across much of the area of the Earth’s crust now occupied by the African continent.

9.3 Burial and Exhumation History

The apatite fission track analyses indicate that gneiss and granite samples collected adjacent to the Katonga Valley have Carboniferous to Permian AFT central ages (256.6 ±13.5 Ma to 335.6±19.0 Ma). The Bihanga diamictite samples have Carboniferous and Triassic AFT central ages (273.2 ±13.4 Ma to 228.9 ±10.2 Ma). However, the negatively skewed fission track length data indicates that central ages probably underestimate the age at which the rock entered the base of the partial annealing zone (~4 km). The observed spontaneous fission track lengths of between 11.69 ±0.16 µm and 12.48 ±0.16 µm suggest track shortening of between 3.4 and 4.2 µm indicating protracted cooling with significant annealing.
The thermal history modelling of the Katonga Valley rock samples, calibrated against both length and age data, indicates that the age range when exhumation occurred through the PAZ was Devonian to Carboniferous. This is consistent with exhumation prior to deposition of the Lower Karoo sediments. Thermal history modelling of the Bihanga diamictite samples also reveals that Mesozoic reheating is required at temperatures up to and equivalent to those of the nominal PAZ to reproduce the observed fission track shortening. Given the modelled range of maximum reheating to between 80°C and 100°C and the likely historical geothermal gradients between 20°C/km and 30°C/km, the estimated maximum thickness of Karoo (Permian to Jurassic) cover sediments is between 2 and 4.5 km. Since the time-temperature paths spend time within the nominal PAZ during the Mesozoic, the AFT data are inconsistent with the preservation of Mesozoic land surfaces as suggested by King (1962). Most modelled time-temperature paths exit the top of the nominal PAZ in the late Mesozoic, suggesting that the rocks which form the current land surface were likely still buried at depths of up to 1 km or more at the start of the Cenozoic. However, since Cenozoic cooling occurs at temperatures lower than the nominal PAZ, the AFT analysis cannot determine the rate or timing of Cenozoic cooling.

The sample collected approximately 50 km north of the Katonga Valley at Kyenjojo has a Permian AFT central age (279.6 ±20.5 Ma) similar to the Katonga Valley samples. However, the two samples collected approximately 75 km north of the Katonga Valley in the vicinity of the Muzizi Valley have late Triassic (203.5 ±8.9 Ma) and Jurassic (158.3 ±7.6 Ma) AFT central ages. Thermal history modelling indicates that the sample with the younger AFT central age likely remained below the PAZ during deposition of the Karoo cover sediments, before experiencing cooling from the Jurassic onwards.

The AFT ages derived for the Katonga Valley samples (Devonian to Carboniferous), on the eastern rift flank are substantially older than the AFT ages recently estimated for the central Rwenzori Mountains (mainly Jurassic) (Bauer et al., 2010). This is consistent with post Mesozoic rock uplift of the horst block revealing rocks with AFT ages younger than those determined for the adjacent rift flank. The thermotectonic history
of the region now occupied by the Western Rift, Rwenzori horst and the Katonga Valley as presented in Figure 8-1 is summarised below.

- **Middle to Late Palaeozoic exhumation** – The Precambrian rocks of the Katonga Valley region first entered the base of the PAZ during the Devonian and Carboniferous. Of the order of 4 km of denudation occurred in less than 120 Ma from the time when these rocks entered the PAZ to the deposition of Lower Karoo sediment. This period of exhumation ended with glacial erosion and the formation of a landscape of high relief, including the palaeovalley exposed again today in the western Katonga Valley.

- **Late Palaeozoic to Early Mesozoic burial** – Apatite fission track shortening in the Bihanga diamictite indicates post-depositional reheating consistent with burial depths between about 2 and 4.5 km. Evidence from the Congo Basin suggests that deposition continued throughout the Permian and possibly also in the Triassic and early Jurassic (Daly et al., 1992, Giresse, 2005).

- **Middle to Late Mesozoic exhumation** – The rocks that now form the Rwenzori Mountains first entered the base of the PAZ during the Jurassic to early Cretaceous (Bauer et al., 2010). This suggests a period of net exhumation during the middle to late Mesozoic, although it is not possible to determine if Mesozoic cooling was continuous or episodic.

- **Miocene to Pliocene plateau uplift and rift grabens** – Surface uplift of the EAP of the order of 1 to 1.5 km likely began between the Oligocene and middle Miocene (Wichura et al., 2010). The rocks of the central plateau have Palaeozoic AFT ages whilst those on the margins in Kenya and Rwanda have Mesozoic AFT ages (Kohn et al., 2005). Therefore, on a large scale greater uplift and denudation appears to have occurred closer to the margins of the EAP, although at a smaller scale no significant trend was observed in AFT ages parallel to the Katonga Valley.

The Karoo-age sediments were likely reduced to a thin, discontinuous cover by this time, and rocks with Palaeozoic AFT ages were exposed in the Katonga Valley. The Semliki Basin began to propagate south from the Albert Basin, and the George
Basin began to propagate north from the Edward Basin either side of the Rwenzori basement block. Whilst hydraulic connections via lakes and rivers may have allowed aquatic species migration, the compressional tectonic setting between the propagating basins appears to preclude the existence of the single large Lake Obweruka suggested by previous researchers (Pickford et al., 1993).

- **Quaternary Rwenzori horst and rift flank uplift** – Following partial detachment at the southern end, uplift of the Rwenzori horst commenced in the early Pleistocene (Ring, 2008). Thermochronological evidence indicates that less than 1.7 km of denudation accompanied 5 km of Cenozoic surface uplift (MacPhee, 2006) and the rocks of the Rwenzori all have Mesozoic AFT ages (Bauer et al., 2010). Denudation on the rift flank reduced the thickness of rocks with Palaeozoic AFT ages and removed almost all of the Karoo sedimentary cover apart from one or two small outliers remaining in the exhumed Gondwanan palaeovalleys. Sediment loading in the George Basin produced flexural uplift of the eastern rift flank.

### 9.4 The Katonga Valley Fill

The ERT images presented in Chapter 6 have been interpreted with the field observations and inferences based on the AFT analysis to produce a parsimonious synthesis of the Katonga Valley fill in the western, central and eastern valley at Kabagole, Kyai and Kisozi respectively. The depositional units and erosion surfaces are categorised into three phases of diminishing duration including the Gondwanan palaeovalley (Phase I), the Neogene relict valley (Phase II) and the late Quaternary channel (Phase II) as presented in Figure 8-3 and summarised in the following sections.

#### 9.4.1 Gondwanan palaeovalley

The Gondwanan palaeovalley was formed over 300 Ma when the area of the Earth’s crust now occupied by south west Uganda was part of a large landmass close to the South Pole. Given that the palaeovalley contains glaciogenic sediment it seems likely that it experienced glacial erosion and therefore had the characteristics of a broad, ‘U’ shaped, low-sinuosity glacial valley. The general shape of the western Katonga Valley is consistent with these characteristics. However, Gondwanan palaeovalley should not be regarded as a proto-Katonga Valley, but rather an ancient feature of the Palaeozoic
landscape that following exhumation from beneath several kilometres of Karoo-age rock has been exploited by the modern River Katonga.

The initial motivation to re-examine the sedimentary rocks of the Katonga Valley was the discovery of fine-grained sandstone that forms a mound in the centre of the valley near Kabagole. In outcrop, it appears similar to the Muzizi Sandstone located 80 km to the north and is described as a reddish yellow to greyish brown, sub-horizontal, thinly bedded to laminated, moderately strong to strong, fine-grained sandstone. In thin section it can be seen to be moderately well sorted, grain supported, low to high sphericity, angular to subangular, fine to very fine (0.25 to 0.063 mm), comprising over 25% potassium feldspar grains (arkose). The dominant lithologies of the Katonga Beds at Bihanga Station are:

- **Rhythmite** – Interlaminated yellowish buff and grey siltstone and mudstone with sandy layers, and rare pebbles (varve);
- **Sandstone and siltstone** –
  - Orangey brown to yellowish buff, interbedded and interlaminated, fine to coarse-grained sandstone, with some reaction to 10% HCL;
  - Grey, fine-grained sandstone and siltstone with black carbonaceous plant fragments, and no reaction to 10% HCL; and
- **Diamictite** – Grey, massive, fine to medium-grained sandstone, with dispersed sub-rounded to rounded small pebbles to large cobbles, with some calcareous cement and veins which have moderate to vigorous reaction with 10% HCL.

There is no record of the many tens of millions of years during the Mesozoic and Palaeogene when the Gondwanan palaeovalley was slowly exhumed from beneath the regional cover of Karoo-age sedimentary rocks.

### 9.4.2 Neogene relict valley

The Neogene relict valley was originally eroded by a westward flowing river under conditions of greater stream power than exist today. There is no clear evidence of sediments remaining in the Katonga Valley today prior to river reversal. However, the geological map (D.G.S.M., 1962) indicates ‘Pleistocene to Holocene’ ‘sands, clays,
grits and gravels’ on the interfluves which appear to have been dissected by the confluence of the Katonga, Nabakasi and Katabalang rivers. These deposits appear to be related to alluvial sand and gravel in the Nabakasi Valley (Johnson, 1958) and in the terraces of the central Katonga Valley observed during this study. As discussed further in the next section, these high level coarse-grained deposits occur on the western edge of an apparently older denuded landscape. Landscape erosion to the west appears to have been reinvigorated, possibly by downthrow of the George Basin. Hence, the interfluve sediment, and possibly the terrace sand and gravel, may have been deposited prior to rejuvenation of westward flowing River Katonga.

The terrace sediment at Kyai is predominantly compact, weakly to moderately cemented, bedded, light grey, poorly to moderately sorted, angular to sub-angular sand and gravel, with sub-angular to sub-rounded cobbles in some beds. Some beds exposed in the lower part of the southern terrace contain considerable quantities of sub-angular to sub-rounded cobbles. The generally coarse grain size indicates that the sediment at Kyai was deposited in a high energy environment. However, the angularity of the clasts suggests that they were not transported far. The poor to moderate sorting and lack of imbrication provides evidence that the strong current was not sustained over long periods. Although beds were observed dipping between 8° and 17° into the north facing valley side, the multiple foresets characteristic of fluvial cross-bedding was not seen. It is concluded that the deposits at Kyai once filled the central Katonga Valley and were likely deposited by low frequency, high magnitude events as might be expected in an arid or semi-arid environment. Sometime after deposition, base level changes renewed erosion and the River Katonga cut through the sand and gravel to create the terraces that exist today.

The western Katonga Valley, which has partly exploited the Gondwanan palaeovalley, is situated between the original drainage divide near Kyai and the modern drainage divide near Bihanga Station (see Section 9.5). Following rift flank uplift, the reduced gradient in the western Katonga Valley may have initiated deposition of fine-grained lacustrine and paludal sediments in the Neogene relict valley. Shallow auger logs from near Kabagole (Waldron, 1952) indicate an increase in grain size from clay, to clayey sand, to fine to medium-grained sand between 2 and 3 m bgl. However, the
unconsolidated sediment at Kabagole has a very low resistivity (<68.7 Ωm) which is consistent with fine-grained clayey silt to over 50 m depth.

The fill in the eastern Katonga Valley is strongly influenced by Quaternary climate variability and fluctuations in the level of Lake Victoria. Two boreholes sunk into the 2 km wide terrace which stands about 10 m above the papyrus wetland near Kisozi indicate the presence of up to 6 m of clayey silt overlying about 50 m of silty sand with occasional gravel towards the base, resting on phyllite and gneiss. These sediments are interpreted as overbank or paludal deposits overlying fluviolacustrine deposits which are contiguous with sediment along the north west margin of Lake Victoria.

9.4.3 Late Quaternary channel

The ERT images from Kabagole, Kyai and Kisozi all indicate the presence of channel filled with very low resistivity material (<68.7 Ωm) cut into the pre-existing substrate including the Karoo-age rocks at Kabagole, the phyllite and sand and gravel at Kyai, and the fluviolacustrine deposits at Kisozi. The channel surface is generally several metres lower than the adjacent valley floor. It was likely cut by the eastward flowing Katonga Valley when the base level in the Victoria Basin was lower than it is today. Seismic reflection surveys have indicated that Lake Victoria dried out at least three times since its formation and the last such event occurred between about 15.9 and 14.2 ka (Stager and Johnson, 2008). A return to higher lake levels during the Holocene facilitated the formation papyrus wetlands and the slow and dispersed surface water flow within the modern Katonga Valley. The channel subsequently filled with the fine-grained paludal deposits. Occasional raised lake levels appear to have inundated the higher ground at Kisozi leaving behind the low resistivity fine-grained deposits overlying the fluviolacustrine sediment.

9.5 Landscape Evolution of the Katonga Valley

The scientific basis of landscape evolution, like forensic science, rests on the identification of a ‘smoking gun’. This is a trace that indicates one of the competing histories, or hypotheses, provides a better causal explanation of all available traces (Appendix A). Whilst many of the previously recognised traces were discussed in Chapters 3 and 4, the new traces recognised during this study include:
• Karoo-age strata filling the Gondwanan palaeovalley;
• apatite fission tracks in the Karoo rocks indicating significant Mesozoic reheating;
• low electrical resistivity sediment filling the western Neogene relict valley;
• sand and gravel terraces and interfluve deposits in the central Katonga Valley;
• break in slope and bedrock high in the central Katonga Valley;
• eastern denuded landscape and western rejuvenated landscape
• broad terrace comprised of medium electrical resistivity fluviolacustrine deposits in the eastern valley;
• late Quaternary channel containing low resistivity sediment beneath the papyrus wetland in the eastern and central Katonga Valley; and
• modern channel cut into old broad elevated valley in the central Katonga Valley.

Whilst forensic science often seeks the standard of evidence required to establish criminal responsibility ‘beyond all reasonable doubt’ (perhaps > 95% probability) researchers working in historical geoscience are often prepared to provisionally accept the ‘preponderance of evidence’ (perhaps > 50% probability) required in civil law. The standard of evidence used to reconstruct the proposed landscape evolution of the Katonga Valley is associated with varying degrees of certainty. For example, whilst there is strong evidence that river reversal has occurred, the explanation of when and under what conditions the interfluve deposits and terrace sand and gravels were deposited is less clear. The conceptual synthesis of the landscape evolution of the Katonga Valley presented here is based on a parsimonious interpretation of the available data. It is proposed as a basis for understanding the origins of the current landscape. The main difference between applied forensic science and academic historical geoscience is that judgement in academic science is provisional. The hypothetico-deductive method is an iterative process, and it is acknowledged that as new data becomes available the hypothesised history proposed below will likely be refined and possibly modified considerably.
1. **Westward flowing River Katonga**

The westward pointing ‘arrow barb’ pattern of the tributary valleys and the north shore of Lake Victoria provide strong evidence of the initial westward flow direction of the River Katonga (Wayland, 1931). The Nyaitanga Valley enters the ‘swamp divide’ on the Mpanga-Katonga Valley System from the south west, but the Katonga Valley and the Karoo outlier continue westward, therefore suggesting that the River Katonga also originally flowed due west. The Lower Karoo lithofacies and the broad, straight form of the valley indicate that the western River Katonga has preferentially exploited an exhumed Carboniferous glacial palaeovalley. In the eastern Katonga Valley, a line of discontinuous quartzite ridges on the contact between the Proterozoic schists and Archaean gneiss, called the Katonga Line (Johnson, 1960), has formed an impediment to southward river migration.

2. **Development of denuded landscape with high level Pleistocene deposits**

The digital elevation model reveals a smooth denuded landscape adjacent to the eastern valley. Pleistocene deposits occur on the interfluves and high-level sand and gravel terraces occur in the Nabakazi and Katonga valleys on the western margin of this smooth landscape. Their association with the denuded landscape suggests that they were deposited before rejuvenation modified the landscape to the west. The terrace sediment appears to have once filled the original high-level central Katonga Valley and the poor sorting and angular lithofacies characteristics indicate it was likely deposited by low frequency, high magnitude events, as might be expected in an arid environment.

3. **Victoria Basin downwarp and formation of eastern drainage divide**

The profiles perpendicular to the valley indicate that the modern narrow valley in the central Katonga Valley has cut into a former high-level broad valley floor. In addition, the ERT survey indicates that even following erosion of the modern low-level and narrow central Katonga Valley, the elevation of the bedrock surface remains higher than the bedrock surface in both the western and eastern valley. This implies that erosion occurred both to the east and the west of the central valley and provides further evidence that the original drainage divide formed in this region
near Kyai, approximately 150 km to the east of the Western Rift. The longitudinal profile of the modern Katonga Valley contains a break of slope near Kyai in the central valley, with the steeper gradient towards the Victoria Basin. Thus, the evidence suggests that the initial drainage divide was created by downwarp of the Victoria Basin, which appears to have commenced prior to rift flank uplift. Based on the depth of sediment and the rate of deposition, the Victoria Basin has been estimated to have formed in the mid Pleistocene about 400 ka (Johnson et al., 2000).

4. **George Basin downthrow, river capture and landscape rejuvenation**

The valley form around the modern drainage divide strongly suggests that the River Katonga was captured from the south west. It seems plausible that northward propagation of the George Basin resulted in greater downthrow along the rift boundary fault to the south. This reduced the local base level in the George Basin and increased the stream power on the rift flank in the south-westward flowing Mpanga-Rusangwe-Nyaitanga river system, which in turn was able to capture the River Katonga near the modern location of Bihanga Station. This in turn led to rejuvenation of the Katonga catchment to the west of the initial drainage divide near Kyai. This explains the increased landscape roughness and stream order observed on the DEM in the western tributaries of the Katonga Valley and in the Nabakazi Valley in particular. It may also have played a role in producing the superimposed drainage that dissects the quartzite ridges and Pleistocene deposits on the interfluves. The rejuvenated westward flowing river likely exposed the relatively unweathered surface of Karoo sediments in the western valley at this time and may have cut a knickpoint in the high level central valley close to the drainage thus beginning terrace formation in the sub-angular sand and gravel.

5. **Rift flank uplift and formation of western drainage divide**

Deposition in the George Basin produced isostatic loading and associated flexural uplift of the rift flank. Given that the Rwenzori horst is still attached to Victoria plate only 20 km north of the Katonga Valley, it also seems plausible that the complex tectonic regime at the junction of the Albert and Edward Basins may also have influenced surface uplift in this region. Whatever, the exact combination of
mechanisms, it is clear that surface uplift in the vicinity of Bihanga Station eventually produced a second drainage divide only 50 km east of the Western Rift. This created low relief east of the new drainage divide and the western Katonga Valley near Kabagole filled with fine-grained lacustrine and/or paludal sediment.

6. Quaternary climate cycles and deposition of fluviolacustrine deposits

Seismic reflection data identify three desiccation surfaces in the Lake Victoria sediment which may be related to the 100 ka climate cycles of the late Pleistocene (Johnson et al., 2000). The lake level was likely higher during the wetter periods which probably coincided with northern hemisphere interglacial periods. The initial return to a wetter climate when the base level in the Victoria Basin was still low may have provided the stream power required to continue erosion of the central valley and reconnect the western and eastern Katonga Valley. Fluviolacustrine sediment was deposited in the eastern Katonga Valley during periods when the lake level was high, and formed the broad terrace flood plain observed today at Kisozi. The eastward migration of the depocentre in the Victoria Basin is consistent with the observed tilting of the high level terraces (Temple and Doornkamp, 1970, Johnson et al., 2000). This indicates relative vertical movement on the central area of the East African Plateau although it is unclear if this involved uplift to the west or downwarp to east.

7. Late Pleistocene Lake Victoria desiccation and channel formation

The last desiccation event in the Victoria Basin occurred between 15.9 and 14.2 ka (Stager and Johnson, 2008). Once again, a return to a wetter climate with an initially low base level would have facilitated erosion, which created the channel observed by the ERT survey in the fluviolacustrine deposits at Kisozi.

8. Holocene high lake level and development of papyrus wetlands

Once the contemporary climate and high lake level was re-established during the Holocene, the papyrus wetland observed today began to develop and the onset of fine-grained paludal deposition within the late Quaternary channel commenced throughout the Katonga Valley.
9.6 Hydrogeology of the Fluvialacustrine Sediment

The complex landscape evolution of south west Uganda has fascinated geoscientists for almost a century but has also produced a spatially variable pattern of erosion and deposition. Unfortunately, this means that there is a lower probability of identifying the kind of wide-spread alluvial aquifers that can be found in stable lowland settings. The groundwater resource potential of each valley reach must be assessed on an individual basis, although this is aided by a detailed understanding of the landscape evolution across the region.

The western Katonga Valley is exploiting an exhumed Gondwanan palaeovalley containing an outlier of Karoo-age sedimentary rocks. These consist of mudstone, siltstone, sandstone and diamictites of unknown overall proportions. Whilst a hand-pumped well has been sunk through these strata at Bihanga Station it was not possible to locate the drilling logs, installation details and pumping test data for this borehole and so the hydrogeological properties of the Karoo rocks remains unquantified. The very low electrical resistivity of the alluvium that fills the Neogene relict valley at Kabagole is similar to the paludal deposits that fill the late Quaternary channel at Kisozi, and has a lower resistivity than both the adjacent fine-grained sandstone and the fluvialacustrine silty sand at Kisozi. This suggests that the alluvium in the western Katonga Valley has a higher silt and clay content than the moderate transmissivity fluvialacustrine deposits at Kisozi. However, this hypothesis remains to be tested by future drilling at the site. The sand and gravel terraces in the central valley are located above the water table and have no groundwater resource potential, and the base of the valley at Kyai is once again filled with very low resistivity paludal deposits.

Quaternary climate variability has led to fluctuations in the level of Lake Victoria and the deposition of fluvialacustrine silty sand with occasional gravel in the eastern Katonga Valley. Relative tectonic movement has raised the western deposits to form a tilted terrace and moved the depositional centre in Lake Victoria towards the east. The fluvialacustrine sediment is contiguous with that on the north west shore of Lake Victoria and has been exploited for village and town water supplies for several decades. However, previous investigations have focused on the short term yield achievable by individual wells. Pumping test data was analysed for eight well installed in the
fluviolacustrine deposits and, for comparison, nine wells installed in the weathered and fractured bedrock. All wells were installed with downhole submersible electronic pumps and therefore exclude the low transmissivity tail of the statistical distribution. These are small data sets the results should be treated with caution. They suggest that the transmissivities sampled in the fluviolacustrine hydrostratigraphic unit have captured the central tendency, whilst those acquired from the weathered and fractured bedrock are only representative of the high transmissivity tail of the statistical distribution.

The flow rate control during the pumping tests varied from good to poor, but it was possible to obtain a reasonable transmissivity estimate in almost all cases. The effective hydraulic conductivity has been calculated by dividing the transmissivity by the maximum potentially contributing thickness of the relevant hydrostratigraphic unit. The hydraulic conductivity range of the fluviolacustrine deposits is 0.03 to 4.61 m/day, with a geometric mean of 0.43 m/day. These fall within the expected range for silt and sand. In comparison, the sampled hydraulic conductivity range for the weathered and fractured bedrock is 0.04 to 0.3 m/day. This is equivalent to silt and the average representative transmissivity of this hydrostratigraphic unit may be even less.

In many parts of the world silty sand would be regarded as a relatively low productivity aquifer. However, in areas where the bedrock geology is dominated by weathered crystalline basement it may still provide a much needed local source of fresh water. Macdonald et al. (2005) estimate that an average rural water supply for 250 people requires a minimum transmissivity of about 1 m²/day. Given that the wells analysed during this study were installed with electronic pumps, it is perhaps unsurprising that all pumping test analyses gave transmissivities greater than 1 m²/day. The final pumping rate during each of the tests conducted in the fluviolacustrine aquifer varied between about 4 and 25 m³/hr. The pumping tests in weathered and fractured rocks suggest that the average transmissivity of randomly selected well locations is likely to lower than 1 m²/d. Expert knowledge of the local geology and geomorphology is therefore particularly important for siting wells in the crystalline bedrock, and even then, the discovery of a transmissive feature may be subject to chance. However, the results of pumping tests in the fluviolacustrine hydrostratigraphic unit suggests that
almost all wells sunk in these deposits will have a transmissivity and yield suitable for at least village water supply. In addition, some boreholes will encounter much higher transmissivity suitable for town water supplies such as the well at Mukono near Lukaya which had a transmissivity of 175 m²/d and a yield of 25 m³/hr.

9.7 Recommendations

9.7.1 Thermochronology and the Karoo sedimentary cover

This study and that of Bauer et al. (2010) have started to fill the gap in our knowledge of low temperature thermochronology of the north west region of the EAP identified by Kohn et al. (2005). However, there are considerable areas north of Lake Victoria and along the western shoreline for which there are as yet no available apatite fission track ages. The compilation of a completed map of fission track ages across the EAP will lead to improved understanding of the variation in the large-scale rock uplift and denudation between the central plateau and the rift flanks. This could potentially provide insights into the deep tectonic and thermal processes in the region.

Bauer et al. (2010) showed that there is no apparent relationship between Mesozoic AFT ages and elevation in the Rwenzori Mountains. This indicates relative displacement between adjacent fault blocks and provides little information on the timing of tectonic uplift and quiescence. Further, information about the timing of surface uplift of the Rwenzori horst and the rift flanks may be obtained if it is possible to obtain samples for detrital AFT analysis of the rift basin sediment. Sediment deposited prior to surface uplift of the Rwenzori horst may be expected to have uniform Palaeozoic AFT ages, whilst those deposited after uplift of the Rwenzori may be expected to have mixed Palaeozoic Aft ages (from the rift flanks) and Mesozoic AFT ages (from the Rwenzori horst).

This study has shown that the Karoo-age rocks of Africa may have been thicker and more widely distributed during the Mesozoic than recognised previously. Whilst the Karoo strata have been partly mapped (Schlüter, 2006) there has been no systematic attempt to interpret the continental variation in their former thickness. It is recommended that consistent thermal history modelling is conducted on AFT data collected from Karoo strata across Africa to determine the extent and degree of
reheating which these rocks have experienced. This could potentially provide information on the thickness and extent of the late Gondwanan sedimentary cover prior to the formation of the African continent.

### 9.7.2 Landscape evolution and tectonic geomorphology

This study is largely based on the direct interpretation of observations. Whilst an initial attempt has been made to connect the landscape evolution to underlying tectonic mechanisms, there are large uncertainties remaining about the nature of these mechanisms. The observations suggest that relative downward displacement of the Victoria Basin commenced before local uplift occurred on the George Basin rift flank. The potential mechanisms that could be responsible for this include thermal sag associated with the evolution of a mantle plume beneath the Tanzanian Craton (Nyblade et al., 2000, Weeraratne et al., 2003), compression between active rifts (Stamps et al., 2008), or long wavelength uplift of the plateau margins associated with shallow arms of the deep seated plume (Simiyu and Keller, 1997). Coupled numerical modelling of the mechanical and thermal processes may be able to test the ability of these mechanisms to reproduce the observed surface displacement on the EAP.

Given the location of the western Katonga Valley at the northern end of the George Basin and due east of the highest peaks of the Rwenzori horst, it seems likely that knowledge of the causes and timing of rift flank uplift in this area may be related to overall understanding of the complex tectonic regime at the junction between the Edward Basin and the Albert Basin. This is the subject of ongoing research by the RiftLink project.

The contrast between the denuded landscape adjacent to the eastern Katonga Valley and the rejuvenated landscape of the western Katonga catchment was identified late in the research process presented here. Systematic and quantitative examination of the DEM is required to map the extent of, and relationship between, these landscape elements. In addition, the relationship between the Quaternary interfluve deposits (1,230 to 1,270 m asl) and the sub-angular sand and gravel terraces in the Nabakazi and central Katonga Valley (1,190 to 1,250 m asl) remains unconfirmed and would benefit from further field observations. Whilst it is likely that the interfluve deposits are
associated with the denuded landscape, the temporal relationship between the eastern
denuded landscape and the sub-angular sand and gravel found within the terraces is less
clear, and would also benefit from further examination.

The Uganda borehole database provides a valuable resource for characterising the
distribution of the depth to bedrock in boreholes across Uganda. Recent preparation of
the Water Supply Atlas 2010 and associated GIS may help to locate many of the
boreholes. It is recommended that an attempt is made to map the regolith thickness for
the purposes of understanding both the geomorphology and the groundwater resource
distribution. Further information on the age of horizons within the deeply weathered
regolith may be gleaned from studies of the palaeomagnetism (Taylor and Eggleton,
2001) and could be used to test the hypothesis that the dominant thickness of the
weathering profile observed today was initially formed in the mid Miocene (Taylor and
Howard, 1998).

Whilst it is acknowledged the maximum potential temporal resolution of
geomorphological features observed in the Ugandan landscape will likely always be
much less than the resolution of Neogene climate variability determined using sensitive
proxy data sets, it would still be interesting to identify the broad age-range of numerous
relict surfaces and deposits. The recent developments in Quaternary dating methods,
including optically stimulated luminescence (OSL) and cosmogenic nuclide (CN)
dating may prove useful for this task. The application of OSL to date the sand and
gravel terraces at Kyai was considered during this study, but the material was thought
unsuitable. Given the potential of these techniques to provide quantitative data to
constrain the still largely relative chronology of the landscape evolution of south west
Uganda, it is recommended that future research is directed towards identifying
appropriate sediment, rock surfaces, and geomorphic environments. However, given
that some of the critical events in the landscape evolution of the region, including the
downwarp of the Victoria Basin and rift flank uplift occurred as gradual processes,
likely over 100 ka, this recommendation is tempered by the possibility that given the
limitations of OSL and CN dating they may eventually prove unsuitable for this task.
9.7.3 Hydrogeology and groundwater resources

The Uganda Water Supply Atlas 2010 (M.W.E., 2011) was recently prepared ‘to provide stakeholders with good knowledge and information on matters concerning the current safe water supply coverage, functionality and distribution of water’. It provides a comprehensive database of the functional and non-functional groundwater and surface water collection and distribution points. However, it does not provide information on the distribution of groundwater resources. Such information is derived from the interpretation and compilation of the transmissivity and storage coefficients of the hydrostratigraphic units from which the groundwater is being drawn, together with data on the water quality. This study has shown that the existing government archive contains valuable information about the transmissivity distribution. It also contains information about the groundwater chemistry. Previous researchers have recommended the establishment of groundwater monitoring networks and databases to identify changes in groundwater storage (Tindimugaya, 2008). This study also recommends the preparation and compilation of transmissivity estimates and hydrogeochemistry data from existing sources to facilitate the national groundwater resource assessment.

This study has shown that modern pumping test analysis techniques, including the semi-log derivative and recovery analysis using Agarwal time are useful tools for diagnosing the flow geometry and transmissivity from pumping tests with varying degrees of flow rate control and quality of water level measurements. However, these techniques are only recommended for trained analysts when computer-based analysis software is available. In most cases, the Theis recovery and Jacob’s straight line method of analysis will continue to provide results of adequate quality. Nevertheless, it is recognised that continued professional development training in pumping test diagnostic behaviour does facilitate the identification of poor flow rate control and low quality data, together with the appropriate flow geometry models for analysis. It is suggested that all applications for groundwater abstraction licenses are accompanied by the results of a step test and constant rate pumping test analysis for entry into the national database.

The pumping test data acquired from combined regolith/bedrock boreholes and analysed during this study have highlighted the difficulty in identifying the distribution
of transmissivity and storage in deeply weathered and fractured bedrock low-productivity aquifers. Previous studies (Jones, 1985, Chilton and Foster, 1995) have proposed a conceptual hydrogeological model of such systems which includes high storage in the decomposed saprolite and high transmissivity in the fractured and disintegrate saprock (see appendix Figure D-3). Unfortunately, whilst the underlying high transmissivity fractures may be used to collect water from the overlying high porosity saprolite (or sediment) the low transmissivity of the overlying regolith will results in a low overall well yield. A number of studies (e.g., Marechal et al., 2004, Dewandel et al., 2006, Holland and Witthüser, 2011) have interpreted geological and hydrogeological data and produced conceptual models of low-productivity crystalline rock aquifer systems; however, they have not implemented numerical representations. It is recommended that exploratory modelling of porous medium overlying discrete fracture network systems is undertaken, and constrained by the data, in order to better understand the hydrogeological behaviour of these low-productivity aquifer systems which form an important source of water to rural communities in sub-Saharan Africa.

At a local level, it is recommended that the hydrogeological properties of the Karoo sedimentary rock in the western Katonga Valley are assessed. However, given the limited extent of the outcrop and the low population density in the vicinity of the Katonga Wildlife Reserve it is acknowledged that this hydrostratigraphic unit may have limited value from a groundwater resource perspective. As a minimum, it is recommended that the borehole information from the unidentified well at Bihanga Station is located and examined to obtain an order of magnitude estimate of the transmissivity and thickness.

Given the limited financial resources available to this study, it was not considered cost effective to commission a drilling investigation of the relict valley sediment near Kabagole in the western Katonga Valley. The ERT survey suggested that the low electrical resistivity sediment likely contains a higher proportion of fines than would be expected in a transmissive unit suitable for groundwater supply. However, this conclusion remains to be tested, and from a strategic perspective further direct observations obtained from a drilling investigation at Kabagole is recommended. This would provide ground truth information about the observed electrical resistivity and
facilitate interpretation of future investigations of other relict valley sediment. This study has confirmed that fluviolacustrine deposits adjacent to Lake Victoria are significant for local groundwater supply. Therefore, it is recommended that the groundwater resource potential of other fluviolacustrine deposits associated with former Pleistocene lakes on the EAP south of Lake Victoria (e.g., Lake Wembere) are also examined.
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THERMOCHRONOLOGY, LANDSCAPE EVOLUTION
AND HYDROGEOLOGY OF THE
KATONGA VALLEY IN SOUTH WEST UGANDA

VOLUME II: APPENDICES

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A RESEARCH PHILOSOPHY

The geoscience research presented in this thesis was conducted with UCL Department of Geography and Birkbeck Department of Earth and Planetary Sciences. Geography departments often comprise multidisciplinary research environments with individuals working in the humanities, social sciences and/or natural sciences. Researchers working in different fields sometimes adopt dissimilar philosophical perspectives. Interdisciplinary research is perhaps necessary if we are to address the complex web of cultural and physical problems that come together in development topics such as water security in the global south. Philosophical assumptions underlie all research and we should be aware of our assumptions in order to know the limits of our assertions and respond to challenges. Gandy (2008) has called for a form of ‘critically reflexive naturalism’ but also recognises that some epistemological strategies adopted in different disciplines will remain philosophically incommensurate. This appendix presents a brief introduction to some of the philosophical issues pertaining to the historical sciences and the research presented in this thesis in particular.

Whilst metaphysical questions regarding the nature of reality are beyond the scope of this thesis, it is still necessary to adopt several ontological assumptions. Whilst I remain cognisant of the philosophical challenges posed by radical scepticism (Grayling, 2008) and the strength of instrumentalist and pragmatist arguments (Bird, 1998, Chalmers, 1999), for the purposes of this research I have adopted a realist stance. Other ontological commitments inherent in this research include the correspondence theory of truth, continuously connected forward causality, parsimony, and the uniformity of nature. Having outlined the ontological assumptions about the nature of reality, it is necessary to consider the epistemological position regarding what can be known about this reality. Social science researchers often adopt the relativist position that knowledge is socially constructed. Historically, it is common for natural scientists to adopt the positivist position that a proposition is only true if it is verifiable by observation. However, Popper (2005) pointed out that it is impossible to verify a universal statement, which would require infinite observations, however, it is possible to falsify a universal statement with a single counter-observation.
During the course of this research I have adopted a combination of the three main forms of scientific reasoning: induction; deduction; and abduction. In other words, I have inferred hypotheses because they are consistent with the observations (Baconian inductivism) or I have rejected hypotheses because they are inconsistent with the observations (Popperian hypothetico-deductive method). However, when there are several explanations that are consistent with the observations, it is necessary to use a priori criteria to make an ‘Inference to the Best Explanation’ (Peirce’s abduction). Typical scientific selection criteria, also known as non-empirical theoretical virtues, include: accuracy, resolution, consistency, scope, parsimony, fruitfulness etc.

This research is conducted within the context of contemporary geoscientific paradigms (Khun, 1996) and whilst qualified commitment is necessary to explore a paradigm’s full potential, it is occasionally useful to question the underlying assumptions. Although some critics of science have claimed that the paradigm-relative nature of science makes it non-rational and entirely socially constructed, Khun (2000) contended that incommensurability between paradigms is partial and paradigm choice based on reasonable shared criteria is rational. With regard to the progressive claims of science, he stated: ‘Conceived as a set of instruments for solving technical puzzles in selected areas, science clearly gains in precision and scope with the passage of time. As an instrument, science undoubtedly progresses’.

Experimental scientists often adopt the hypothetico-deductive method, by first making predictions and then testing whether they are corroborated or falsified under controlled conditions in the laboratory. In contrast, historical scientists often attempt to explain observable phenomenon in terms of their past causes using inductive reasoning and abductive inference (Cleland, 2001, Cleland, 2002). For example, the research presented in this thesis attempts to explain the current character of the Katonga Valley fill in terms of the influence of past tectonic and/or climatic events. Since historical hypotheses cannot be tested in the laboratory, it is sometimes claimed that historical science is inferior to experimental science. In response, some historical researchers have adapted the method of falsification by testing multiple working hypotheses (Battarbee et al., 1985). However, Cleland (2001) suggests that most historical scientists continue to search for positive evidence in the form of a ‘smoking gun’. This is an individual trace which identifies one of the competing hypotheses as providing a
better causal explanation of all available traces. The laboratory studies of historical scientists are generally aimed at clarifying traces rather than directly testing hypotheses under experimental conditions. The computer models of historical scientists can determine the consequences of specific hypotheses under the explicitly represented conditions, but they cannot determine which of these hypothetical conditions exist in reality.

Cleland (2001, 2002) proposes that the different methods of experimental and historical science selectively exploit different consequences of a pervasive feature of nature known as the asymmetry of overdetermination (Lewis, 1979). Localised events tend to be causally connected in an asymmetric manner. For example, it is often easy to establish that a volcanic eruption has occurred in the recent geological past due to the abundance of evidence that is left behind. However, volcanic eruptions are notoriously difficult to predict due to the large number of relevant conditions which are difficult to measure. Cleland (2001) suggests that the asymmetry of overdetermination implies that only a small sub-collection of traces is required to confer reasonable probability of occurrence on an historical causal event. However, the causal underdetermination of future events means that experimental scientists must carefully control their experiments to rule out potentially auxiliary causal factors. As a consequence, Cleland (2001, 2002) argues that the claim sometimes made that historical science is inferior to experimental science cannot be sustained.

Turner (2005) draws an important distinction between the ontological thesis about the asymmetry of overdetermination proposed by Lewis (1979) and the epistemological conclusions drawn by Cleland (2001, 2002). He argues that whilst the asymmetry of overdetermination may be a pervasive feature of nature, it is also compatible with local epistemic underdetermination in the historical sciences. He argues that there are many cases in the historical sciences where alternative hypotheses are equally supported by empirical evidence. Two incompatible hypotheses are considered empirically equivalent in a weak sense if they are equally supported by the current available evidence. However, they are considered empirically equivalent in a strong sense if they are equally supported by all empirical evidence that can ever be available. For example, many may consider alternative hypotheses regarding the colour of dinosaurs to be empirically equivalent in the strong sense. Turner (2005) points out that whilst
philosophers are often interested in the implications of global underdetermination of theory by evidence, historical scientists focus on tractable problems where underdetermination appears to be local. In other words, they confine their research to those historical puzzles where they hold out reasonable hope of discovering a ‘smoking gun’.

Turner (2005) gives several examples of local underdetermination in the historical sciences, including the biological classification of partial fossil remains, the assignment of trace fossils to extinct species, and the adaptationist hypotheses of evolutionary biologists. He goes on to point out that the geological record is usually incomplete and many processes such as weathering, erosion and transport destroy historical information. These observations are used to support his argument that local underdetermination problems are more pervasive in the historical sciences compared to the experimental sciences. Although historical scientists can develop new technologies for identifying and studying historical traces, as Turner puts it: ‘when they go looking for smoking guns, they are, to a large degree, at nature’s mercy’. For example, during the reconnaissance for this thesis an historical trace was discovered in the form of the indurated sedimentary rocks in the western Katonga Valley. It is only due to this discovery, and the use of previously developed thermochronological techniques, that it was possible to infer the Phanerozoic history of burial and exhumation of the Gondwanan palaeovalley.

Turner (2005) argues that whilst background theories involving information destroying processes tend to limit the epistemic ambitions of historical scientists, different background theories tend to assist and support, rather than hinder, experimental science. Nevertheless, he does not regard the matter of whether historical science is epistemically inferior to experimental science as settled. Neither does he consider any particular historical theory necessarily less well confirmed than one supported by experimental observations. But he does maintain that local underdetermination of theory by evidence is a more pervasive problem in historical science. It is perhaps just as important to understand the inherent uncertainty associated with an explanation as it is to propose the explanation itself. It is attempted to consider the issue of uncertainty throughout the research presented in this thesis.
B THE PRECAMBRAIN ROCKS OF SOUTH WEST UGANDA

This appendix provides supplementary description and discussion about the Precambrian rocks of Uganda. It refers back to tables and figures presented in the main volume.

B.1 Archaean Gneissic-Granulitic Complex

During the Archaean, the Earth’s heat flow reduced from approximately three to two times the current levels, volcanic activity was considerably higher than today and the tectonic style was likely different (Stanley, 2005). The earliest evidence of life is found during the Palaeoarchaean although oxygenation of the atmosphere did not occur until the early Proterozoic. The lack of Archaean rocks world-wide indicates that the proto-continents that did form at this time were relatively small.

The Palaeoarchaean Tanzanian Craton, which outcrops to the south of Lake Victoria (Figure 3-1), is a distinct tectonic unit which has formed a rigid block since the Archaean. It existed in the late Archaean during the formation of the Nyanzian greenstone belt of south-east Uganda, western Kenya and Tanzania. Approximately 60% of Uganda is underlain by the Palaeoarchaean Gneissic-Granulitic Complex (Figure 3-1), consisting of gneisses, granulites, schists, quartzites and marbles (Davies, and Bisset, 1948).

The Gneissic-Granulitic Complex outcrops across extensive areas of central and northern Uganda. Schlüter (1997) has suggested that in northern Uganda it forms the north east part of the Congo Craton and in south west Uganda where it emerges from beneath the Buganda-Toro System it forms part of the Tanzanian Craton. However, large parts of the Gneissic-Granulitic Complex have been imprinted by post-Archaean thermotectonic events and its cratonic status is questionable.

Some sequences of the Gneissic-Granulitic Complex may be described using igneous terminology whereas others appear layered and include quartzite, which suggest a sedimentary origin (Hepworth, 1970). Since its initial formation, the Gneissic-Granulitic Complex has become a palimpsest of subsequent tectonothermal events that have affected different regions of Uganda. The approximate average ages of these
events have been calculated using K:Ar, Rb:Sr and U:Pb radiometric dating by various researchers (Cahen and Snelling, 1966, Leggo, 1974) and summarized by Vail (1976), Cahen (1984), Schlüter (1997). As shown in Table 3-1, they include the Watian (≈2900 Ma), Aruan (≈2600 Ma), Rwenzori (2500 to 1800 Ma), Kibaran (Karagwe-Ankolean) (≈1000 Ma) and Mozambique (≈650 Ma) events (Hepworth and Macdonald, 1966, Vail, 1976).

The 1962 geological map of Uganda (D.G.S.M., 1962) identifies the Toro System (i.e., the Buganda-Toro System) north and south west of the Katonga Valley and the Basement Complex (i.e., the Gneissic-Granulitic Complex) southeast of the Katonga Valley. However, the regional imprint of the Rwenzori tectonothermal event that occurred around 1800 Ma, has sometimes led to confusion in the stratigraphic classification of the region. There appear to be Archaean stratigraphic units within the Rwenzori Fold Belt and, therefore, it should not be equated with the Palaeoproterozoic Buganda-Toro System.

B.2 Palaeoproterozoic Buganda-Toro System and the Rwenzori Fold Belt

The beginning of the Proterozoic Eon is marked by the evolution of photosynthetic organisms and consequent oxygenation of the atmosphere (Stanley, 2005). Approximately 20% of Uganda is underlain by Proterozoic rocks. The Palaeoproterozoic Buganda-Toro System outcrops to the north and south west of the Katonga Valley (Figure 3-3). This system is comprised of two lithostratigraphic groups of metasediments which include the Buganda Group to the east and the Toro Supergroup to the west (Schlüter, 1997). The basal series of the Buganda Group is comprised of massive quartzites with common evidence of shearing or mylonitisation (McConnel 1959) and occasional preservation of sedimentary structures including, bedding, ripple marks, pebble bands and conglomerates. The quartzites are overlain by shales, slates and phyllites, which are succeeded by basic volcanics, amphibolites and tuffs (King 1959, King and Swardt 1970). Pallister (1954) (1959) has estimated the Buganda Group to be less than 1000 m thick around Jinja near its eastern limit, and up to 7000 m thick in the central area of the outcrop west of Kampala.
The Toro Supergroup appears to form a continuous extension to the south and west of the Buganda Group (King 1959). It likely represents higher members of the Buganda-Toro System with generally higher grade metamorphism and more complex structures (Cahen and Snelling, 1966). The stratigraphic succession of the Toro Supergroup is similar to the Buganda Group and includes quartzites and conglomerates, tholeiitic intrusions, pelites, schists, amphibolites and marbles (Tanner, 1970, Tanner, 1973). Schists, quartzites, and volcanics of the Toro Supergroup also form the highest peaks of the Rwenzori Mountains which form a horst block that appears to have been recently uplifted during the Quaternary.

The regional structure of the Buganda-Toro System can be regarded as a broad complex syncline with a gently plunging WSW axis. Barth and Meinhold (1974) assume that the Buganda-Toro System was deposited within a geosyncline, whereas Tanner (1973) states that the development of the Rwenzori Fold Belt provides the earliest evidence in the region of a tectonic margin. Petters (1991) describes the Rwenzori Fold Belt as the northern termination of the Ubendian Belt. The forearc basin sediment of the Buganda-Toro System and the subsequent orogenic event that produced the Rwenzori Fold Belt (Schlüter, 1997) provide Paleoproterozoic evidence of modern plate tectonic processes.

Cahen et al (1984) consider the Buganda-Toro System to be encompassed by the age range 2500 to 1850 ±40Ma, based mainly on radiometric dating of gneisses and granites. Schists often appear to have been influenced by the later Kibaran tectonothermal event. Cahen and Snelling (1966) determined the large Mubende granite (Figure 3-3) to have been intruded into the Buganda Group at 1807 ±60 Ma.

Mapping of the area to the southeast of the Katonga Valley is incomplete (Figure 3-3). Although it is thought to lie within the tectonothermal influence of the Rwenzori Fold Belt (Tanner, 1970), classifications based on radiometric dating have perhaps misinterpreted this area as belonging to the lithostratigraphic based Buganda-Toro System (Cahen et al., 1984). More recently, Schlüter (2006) has classified the area southeast of the Katonga Valley as part of the Gneissic-Granulitic Complex. Recent structural mapping and geochronological studies have begun to shed light on the complex geology of the western Katonga region. Link et al. (2010) have proposed that the area east and south of the Rwenzori Mountains consists of a west-south west to
east-north-east trending fold belt where stacked slices of Proterozoic (Buganda-Toro System) and Archaean (Gneissic-Granulitic Complex) have been thrust onto the craton from the south. The presence of Archaean units is supported by U-Pb age determinations (2637 to 2584 Ma) and repeated Palaeoproterozoic stratigraphy is demonstrated using a basal quartzite marker horizon. The thrust belt is overlain by a Neoproterozoic nappe formed from the Karagwe-Ankolean rocks of the Kibaran Belt to the south.

### B.3 Mesoproterozoic Karagwe-Ankolean System

The Mesoproterozoic (1600 to 1000 Ma) Kibaran Belt is one of the major geological features of central and eastern Africa (Figure 3-1). Large outcrops occur around the southern end of Lake Tanganyika, the Shaba and Kivu provinces of western D.R. Congo through Burundi and Rwanda into northern Tanzania and southwest Uganda where they are known as the Karagwe-Ankolean System (Schlüter, 1997). The northern boundary of the Karagwe-Ankolean System with the Buganda-Toro Sequence and the Gneissic-Granulitic Facies is complex and often unclear (Figure 3-2). In Uganda, it has been grouped into three divisions using quartzite marker horizons. The upper division is comprised of mudstones, sandstones, grits and occasional conglomerates. The middle division is mainly sandstone underlain by mudstones and phyllites, and the lower division is largely muscovite schists, phyllites and quartzites. The Karagwe-Ankolean System of Uganda has been intruded by granites which form prominent geological and geomorphological features that have been called ‘arenas’ (Wayland, 1920). Temple (1967) proposed that the granites intruded into, and altered, the underlying Buganda-Toro Series. He suggests that the overlying Karagwe-Ankolean Series in the centre of the dome was then eroded to expose the low elevation ‘arena’ granite, surrounded by the remaining Karagwe-Ankolean outcrop. The River Kagera flows through the outcrop of the Karagwe-Ankolean Series (Figure 3-2) and the Katonga Valley lies at a minimum distance of 60 km to the north.

### B.4 Neoproterozoic Sedimentary Rocks

The Bukoban System of north-west Tanzania (Figure 3-1) is comprised of sub-horizontal, un-metamorphosed conglomerates, sandstones, quartzites, shales, limestones and basalts. It is thought to be correlated with the predominantly
arenaceous Bunyoro, Singo and Mityana Series of Uganda, based on their Neoproterozoic age, unfolded character and absence of plutonic intrusions (Schlüter, 1997).

The Bunyoro Series outcrops north of the River Kafu to the east of Lake Albert (Figure 3-2). It is comprised of a lower Conglomeratic Member, dominated by conglomerates and arkosic grits, overlain by the Argillaceous Member dominated by mudstones and shales (Davies, 1938, cited in Schlüter, 1997). The presence of a tillite-like rock suggests this series may have been deposited during the Vendian glacial period.

The Mityana Series sandstones and conglomerates outcrop around Lake Wamala approximately 30 km north of the Katonga Valley where they form a shallow basin dipping towards the lake. It sits unconformably on the Buganda-Toro System and the Singo Series (Schlüter, 1997).

Outcrops of the Singo Series occur on Kabuga Hill located 15 km west of the drainage divide on the Mpanga-Katonga Valley System at Bihanga Station, and on the Kisozi Hill north of the eastern Katonga Valley (Figure 3-3). In general, the Singo Series consists of conglomerates and sandstones with rare shales deposited on a mature surface following erosion of the Buganda-Toro Series (Schlüter, 1997). Johnson (1960) specifically describes the Kisozi Hill outcrop as uncleaved, pink, white or buff, often banded, sandstones and mudstones. They are locally conglomeratic with ripple marks and current bedding. The cement is silica and they may be classified as quartzites, especially to the west of the outcrop. Mudstones are more common to the east. The Singo Series of Kisozi Hill forms a west-northwest to east-southeast trending syncline with beds dipping between 5° and 50°. It is cut by two north-west trending faults which appear to have downthrown each end of the hill.
The success of W. M. Davis’s 19th century concept of the erosion cycles of youth, maturity and old age led many 20th century geomorphologists to identify numerous final-stage erosion surfaces preserved in landscapes of southern and eastern Africa, including Uganda (Wayland, 1933, King, 1948, Bishop and Trendall, 1967, Doornkamp, 1968b, Brosh and Gerson, 1978, McFarlane, 1989, Ollier, 1992). The historical study of erosion surfaces sets the traditional context for understanding exhumation and landscape evolution of the Katonga region, against which the modern understanding based on thermochronology and tectonic geomorphology may be compared. The main protagonists, together with the areas in which they were interested and the names they gave to erosions surfaces are summarised in Table C-1. In general they recognised between three and five main surfaces. The surfaces are associated with broad elevation ranges which were generally explained by post-formation regional modifications by tectonics. There is broad agreement in the altitude categories and the general age ranges that have been assigned, although the strength of evidence for the correlations and ages is arguably often tenuous.

Wayland (1933) first interpreted three erosion surfaces in Uganda (Table C-1), although his reporting is less rigorous than later workers and the reasons for his inferences are often far from clear. In keeping with Davis’s paradigm, Wayland describes the highest recognised erosion surface, Peneplain I (PI) which reaches over 1,490 m, as the only ‘true’ peneplain, i.e. originally reduced to sea level in extreme ‘old age’. He describes the lower regolith covered surfaces, PII up to 1,400 m asl and PIII up to 1,200 m, as ‘etched’ plains. He observed that given the existence of higher surface they do not represent regional reduction of the landscape to a new base level, although he continues to call them peneplains.

Working in southern Africa, King (1948, 1953) rejected Davis’s notion of peneplains formed by fluvial processes in favour of what he termed pediplains, formed by scarp retreat. However, he retained the notion of erosion cycles and proposed three major cycles to explain the current African landscape (King, 1948). Writing before the wide-scale acceptance of continental drift and development of plate tectonic theory, he
hypothesised that an originally extensive ‘Gondwana Surface’ (Table C-1) formed prior to Mesozoic rifting of the continental margins. He suggested that ‘outward tilting’ of the continental margins resulted in the Gondwana Surface being preserved at a lower elevation near the coast. The later ‘African Surface’ was exhumed following the base level change associated with rifting and the resulting pediments progressed inland via scarp retreat. He proposed that later erosion surfaces were established following mid-Tertiary uplift of southern and eastern Africa. Although the elevation of surfaces associated with each cycle may vary due to later tectonics and their location on the scarp or pediment, King assumed that correlation was possible over large distances, and even between continents (King, 1956, 1962). Rather than interpret landscapes on the basis of inductive inference from local data, he often appears to have attempted to force observations into the context of his grand theory. King reduced the falsifiability of this overall deductive hypothesis by assuming that the original surfaces of the major erosion cycles may be locally modified and covered by sediments of any age younger than the original surface (King, 1953). By accepting the maximum possible extent and oldest possible age, King concluded that a large part of the current African landscape is of Mesozoic to mid-Cenozoic age. He later further divided the continental landscape by adding post-Gondwana and post-African surfaces which have been applied to Uganda (McFarlane, 1989). Dixey (1937, 1945), working in the previous decade, had identified surfaces corresponding to those of King, but described them as peneplains, and gave them the age based names shown in Table C-1.
### Table C-1: Summary of the main erosion surfaces identified by historical landscape evolution studies in Uganda

<table>
<thead>
<tr>
<th>Region of Interest:</th>
<th>Uganda</th>
<th>Southern &amp; E Africa</th>
<th>Southern &amp; E Africa</th>
<th>Uganda</th>
<th>Central &amp; N. Uganda</th>
<th>Southwest Uganda</th>
<th>Southern Uganda</th>
<th>Central &amp; N. Uganda</th>
<th>Southwest Uganda</th>
<th>Africa</th>
</tr>
</thead>
<tbody>
<tr>
<td>&gt;1490</td>
<td>Peneplain I</td>
<td>Jurassic</td>
<td>Gondwana</td>
<td>Ankole</td>
<td></td>
<td></td>
<td></td>
<td>Buganda</td>
<td></td>
<td>African</td>
</tr>
<tr>
<td>1400 to 1490</td>
<td>Peneplain II</td>
<td>Cretaceous</td>
<td>Post-Gondwana</td>
<td>Koki</td>
<td></td>
<td></td>
<td></td>
<td>Buganda</td>
<td></td>
<td>Jurassic to Mid-Cretaceous</td>
</tr>
<tr>
<td>1250 to 1400</td>
<td></td>
<td>Mid Tertiary</td>
<td>African</td>
<td>Buganda</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Buganda</td>
<td>African</td>
</tr>
<tr>
<td>1070 to 1250</td>
<td>Peneplain III</td>
<td>End Tertiary</td>
<td>Victoria Falls / Post African I</td>
<td>Tanganyika</td>
<td>Kyoga (N) Kasubi (C)</td>
<td>Lowland Landscape (partly filled)</td>
<td>Lowland Surface and Valley-floor</td>
<td>African</td>
<td>African (modified on rift flank)</td>
<td>n/a</td>
</tr>
<tr>
<td>910 to 1070</td>
<td>n/a</td>
<td>Lower Pleistocene</td>
<td>Post African II</td>
<td>Acholi</td>
<td>Karamoja (N)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The table provides a summary of the main erosion surfaces identified by historical landscape evolution studies in Uganda, categorized by region and indicative elevation. Each surface is described by its geological characteristics and the authors who have documented them. The table highlights the complexity of erosion processes in Uganda, with various surfaces dating from different geological eras, including the Peneplain, which represents significant landscape evolution over millions of years.
C.2 Research between 1960 and 1970

Bishop and Trendall (1967) point out the difficulty in correlating the erosion surfaces of Uganda with those in other areas of eastern and southern Africa and therefore conclude the use of local names is to be preferred. They go on to review the surfaces identified by earlier workers and are particularly influenced by the work of McConnell (1955) and Pallister (1959, 1960) who’s proposed erosion surfaces are shown in Table C-1. The highest surface defined by McConnell (1955, cited in Bishop and Trendall, 1967), named the Ankole surface, occurs in southern Uganda. Bishop and Trendall (1967) indicate it has previously been associated with Wayland’s PI surface and King’s Gondwana surface (Table C-1).

The most pronounced surface in central Uganda has been named the Buganda Surface (McConnell, 1955, cited in Bishop and Trendall, 1967, Pallister, 1960). It has previously been correlated with Wayland’s PII surface and King’s African surface (Table 4-1). These laterite-capped often flat hill-tops occur with an elevation of about 1,300 m asl around Kampala and up to 1,400 m asl further west. The surface is best preserved on the water divide between the Katonga and Kyoga drainage basins. Bishop and Trendall (1967) state that this surface has been modified by later tectonics and the laterite cap, which is up to 9 m thick, may suggest significant lowering due to solution weathering. McConnell (1955, cited in Bishop and Trendall, 1967) also defined the higher Koki surface, which reaches up to 1,450 m asl in the south, but is confined to hills formed of the Karagwe-Ankolean System. As was common practice by mid 20th century geomorphologists working on landscape evolution, Bishop and Trendall (1967) made a large inference that the Buganda and Koki surfaces were originally formed at the same time and once constituted a continuous plain. They proposed that the present difference in elevation is due to differences in post-origin alteration rates on different rock types. This inference demonstrates the different paradigms observed by geologists and geomorphologists working at that time. Geologists often assumed that over short geological timescales the rock’s resistance to erosion determined the dominant landscape form, and used this relationship for reconnaissance mapping of the geology. Geomorphologists often assumed that over long geographical timescales the rock’s resistance was less important than the regional base level to which the landscape is reduced, and used this relationship to correlate erosion surfaces. These different
paradigms actually constitute end members of a continuum and their relative significance will depend on the local geology and denudation rates.

Bishop and Trendall (1967) locally define the flat-topped spurs about 80 m below the Buganda Surface in the Kampala area as the Kasubi Surface. They indicate that this is probably equivalent to their Kyoga Surface, which is preserved beneath early Miocene (~19 Ma) volcanic rocks in eastern Uganda. Both the Kasubi and Kyoga surfaces are considered equivalent to McConnell’s (1955, cited in Bishop and Trendall, 1967) and Pallister’s (1959) Tanganyika Surface. Bishop and Trendall (1967) also state that the Karamoja plain in northern Uganda, and identified by McConnell (1955, cited in Bishop and Trendall, 1967) as the Acholi Surface, is likely a modified remnant of the Kyoga Surface.

Holmes (1965) summarised the early view of Ugandan landscape evolution in stating the region was reduced to a pediplain by the Miocene (corresponding to the African or Buganda surface) with subsequent upwarp of the East African Plateau and consequent renewed erosion. He describes ‘swells with deep rifts at their crests’ superimposed on the plateau. He seems to have believed the ‘swells’ were formed contemporaneously with or even before the rifts, rather than after rifting due to rift flank tectonics as described in Section 4.2.2.

Given the problems of understanding the origins and correlations between apparent surfaces Doornkamp and Temple (1966) adopted a morphological based classification. Working in south west Uganda (Figure C-1), they categorised the morphological elements into the:

1. Upland Landscape;
2. Lowland Landscape; and,
3. Infill Landscape.

Doornkamp went on to review the relationship between these morphological categories and previously defined erosion surfaces (Doornkamp, 1968b) and recognised the broad relationships shown in Table C-1. He also applied these broad categories to geomorphological mapping of the Mbarara area (Doornkamp, 1970) during which he defined many more minor surfaces of local significance.
Doornkamp (1968b) suggests the Upland Landscape broadly corresponds to the Koki and Buganda surfaces of McConnell and Pallister (1955, and 1960, cited in Bishop and Trendall, 1967). Doornkamp and Temple (1966) mainly confined their studies to the area south of the Katonga Valley. In the area immediately southeast of the Katonga Valley around Masaka, the Upland Landscape survives as flat-topped hills with laterite caps (Figure C-1). However, they state that the presence of unresistant granites, gneisses and schists over much of the Katonga drainage basin has not favoured the preservation of the upland landscape. The surviving remnants of the Upland Landscape are mainly concentrated in southern Ankole, where the Karagwe-Ankole shale and quartzite outcrops. Numerous minor surfaces in the Upland Landscape are separated by up to 120 m in elevation. Although the Upland Landscape has a polycyclic form, Doornkamp (1968b) maintained that the overall impression is of continuity of this major erosional feature across southern Uganda.

The Lowland Landscape broadly corresponds to the Tanganyika Surface of McConnell and Pallister (1955, and 1960, cited in Bishop and Trendall, 1967) and the Kasubi and Kyoga surfaces of Bishop and Trendall (1967). It forms the majority of the Katonga drainage basin, excluding the flat valley bottoms. It is also comprised of minor surfaces which Doornkamp suggests is due to intermittent river incision. Occasional
granite inselbergs and quartz ridges rise above the Lowland Landscape (Doornkamp, 1968c) as the author has observed south of the Katonga Valley. Doornkamp and Temple (1966) state that regional uplift led to valley incision and the superimposition of pre-existing upland drainage on lower structures. Doornkamp (1968b) infers from Bishop and Trendall’s (1967) dating of the Kasubi/Kyoga surface as early Miocene (~19 Ma) that the Lowland Landscape was initiated prior to the Miocene. However, given that archaeological material has dated alluvial sediments at Nsongezi in the Kagera Valley as mid Pleistocene, he also points out that incision of the Lowland Landscape likely continued after the Miocene. Therefore, he suggests that the correlation between the Kasubi and Kyoga surfaces may be incorrect or the Kyoga surface represents a higher stage of the Lowland Landscape which continued to deepen between the mid Miocene and mid Pleistocene.

The third element in Doornkamp and Temple’s (1966) morphological landscape classification is the Infill Landscape. This includes aggraded valley floors, the alluvial margins of Lake Victoria and the sediments of the Western Rift. They are equivalent to Pallister’s (1960) Acholi surface. These sediments are discussed further in Section 4.5.

De Swardt and Trendall (1969) recognised three local cycles of erosion associated with the change in base level brought about by the Western Rift. The eastern limit of the first cycle is defined by the surface water divide between steep sided valleys to the west and flat-bottomed valleys to the east. They assume that the current water divide generally defines the approximate location of a tectonic axis of upwarp, except in a few locations where the headwaters of local valleys appears to lie beyond the regional axis. They suggest that a break in slope described by Combe (1934) in Bushenyi, south of the Katonga provides evidence of a second erosion cycle initiated by a rift related base-level change. Evidence for a recent third rift cycle comes from the deep gorges such as that cut by the Mpanga into the base of the rift scarp.

C.3 Research between 1970 and 2000

Brosh and Gerson (1978), working on soil characteristics in the valleys and interfluves north of Nsongezi, identified three morphological elements, the Rwampara, Gayaza and Sanga surfaces (Table C-1) with laterite duricrust which appear to be equivalent to Bishop and Trendall’s (1967) Buganda Surface and Doornkamp and Temple’s (1966)
Upland Landscape. Their lowland surface and valley floor also appear equivalent to Doornkamp and Temple’s (1966) Lowland Landscape and Infill Landscape (Table C-1). The multiple minor surfaces of Brosh and Gerson, and Doornkamp mark the culmination of the ‘splitters’. Later workers are perhaps better described as ‘lumpers’. Ollier (1992) returns to three surfaces, Taylor and Howard (1998) essentially describe two surfaces with local modification adjacent to the Western Rift, and Burke and Gunnell (2008) define a single composite surface.

Ollier (1992) presents a broad conceptual model of the relationship between the three main erosion surfaces in Uganda. Similarly to Bishop and Trendall (1967) he names the high, sometimes flat-topped, hills of southern and central Uganda, the Buganda Surface. Figure C-2 shows that in south west Uganda this broadly correlates with Doornkamp and Temple’s (1966) Upland Landscape. Given that the lower planation surface is of pre-lower Miocene age (Bishop and Trendall, 1967) McFarlane (1989) considers it appropriate to name it the African Surface. Ollier (1992) also names the lower weathered surface the African Surface, which has led to confusion since this is
the term previously associated with the higher Buganda Surface (Bishop and Trendall, 1967). Similarly to Pallister (1959, 1960) Ollier names the rock-cut erosion surface of northern Uganda the Acholi Surface. Ollier (1992) states that the deep saprolite is best preserved on the Buganda Surface in the south and is progressively stripped to the north. The central African surface cuts across the saprolite and the northern Acholi Surface is cut into the rock. Ollier also describes how the deep saprolite has been stripped adjacent to the Western Rift since the Miocene, and speculates that the deep saprolite formed in the Mesozoic or earlier and therefore the original unweathered Buganda Surface was Paleozoic age.

Taylor and Howard (1998) point out that the final separation of Africa and South America (Figure 3-9c) coincided with a climatic optimum in the mid-Cretaceous (Figure 3-7) which was conducive to deep weathering. They suggest a Mesozoic origin for the upper surface and name it the Jurassic/mid-Cretaceous Surface, which is equivalent in elevation to Ollier’s (1992) Buganda Surface. Like McFarlane (1989) and Ollier (1992) they name the lower denuded surface the African Surface. While ferricrete does not underlie the early-Miocene volcanic rocks on Taylor and Howard’s (1998) African Surface (equivalent to Bishop and Trendall’s (1967) Kyoga/Kasubi Surface), it does underlie sediments in the northern part of the Western Rift (Bishop, 1965). It is suggested that this ferricrete must have developed on the African Surface before the down-faulting of the Western Rift at about 9 to 10 Ma (Nyblade and Brazier, 2002). Taylor and Howard (1998) suggest that the mid Miocene climatic optimum (Figure 4-21a) could have facilitated ferricrete development on the African surface during the intervening period of up to 10 Myr. They cite Pickford et al. (1993) in proposing the transition from fluvial to fine-grained lacustrine sediments during the late Miocene (~8Ma) marked the initial development of a deep graben. Like de Swardt and Trendall (1969) they draw attention to the subsequent locally eroded surface on the rift flank and name this the Modified African Surface.

C.4 Research after 2000

The interest in long-term landscape evolution, embodied by the study of erosion surfaces, slowed in the late 20th century. The growing dissatisfaction with qualitative approaches and speculative inferences was embodied in Strahler’s (1952) call to place geomorphology on sound foundations for quantitative research. This eventually led to a
focus on functional, morphometric and process-based approaches to geomorphology (Bishop, 2007). Misguided studies of erosion rates, based on areas of anthropogenic disturbance and ignoring isostasy (Bishop, 2007), also suggested that landscapes were relatively young and probably no older than the Pleistocene (Thornbury, 1969). Bishop (2007) suggests the main new theoretical development in landscape evolution during this period was perhaps Hack’s (1975) theory of dynamic equilibrium. Contrary to Davis (1899), Hack argued that vertically exhumed landscapes attain equilibrium between lithological resistance and gradient and do not evolve through cycles leading to erosional plains. At the same time, Crickmay (1975) hypothesised that fluvial erosion actually becomes concentrated in valley bottoms, which leads to landscapes of increasing relief. Recent numerical simulations have suggested that Hack-type dynamic equilibrium models are appropriate in settings where stream power and sediment flux keep up with erosion, and Davis-type declining gradient models are appropriate in settings where there are low rates of rock uplift. These findings suggest possible different modes of current landscape evolution on the low relief to the east of the Katonga region and the rift flanks to the west of the Katonga region.

Recently, Burke and Gunnel (2008) returned to the theme of the African erosion surface, first named by King (1948) sixty years earlier, and attempted a continent-scale synthesis of the geomorphology, tectonics and environmental change over the past 180 Myr. They first point out that erosion surfaces may be associated with different types of ages, including:

1. Initial age – the time when lowering of the base level first initiated onset of erosion;
2. Local age – time-bracket given by dateable surface deposits, sedimentary or volcanic; and,
3. Terminal age – the change in base level that terminates stability (initial age of next cycle).

Burke and Gunnel (2008) also define a polygenetic or composite surface as one which has evolved over $10^7$ to $10^8$ years during which second-order environmental changes such as climate or local tectonics may have influenced the surface development. Therefore local surfaces at different elevations but blanketed by continuous laterite, or related to different lithological resistances, or influenced by site-specific
geomorphological processes, may all be part of the same regional composite surface. They state that local surfaces in Africa may separated by steps of typically up to 100 m but rarely more than 200 m. Given these definitions and considering that the break-up of Pangaea to the east of Africa occurred at about 180 Ma (mid Jurassic), they propose a new definition for the ‘African Surface’ as: ‘the composite erosion surface that dominated Afro-Arabian scenery ca. 30Ma as the outcome of up to 150 m.y. of continental denudation’

A second initial age is provided by initiation of the South Atlantic at about 125 Ma (mid Cretaceous). The terminal age is defined by the rise of Africa’s ‘swells’ which Burke and Gunnell (2008) propose occurred about 30 Ma and include the East African Plateau. As shown in Table C-1 they consider Bishop and Trendall’s Buganda Surface (1967) and all higher surfaces to be equivalent to their ‘African Surface’. They recognise that Ollier (1992) named the surface below the Buganda surface as his African surface (Table C-1), but nevertheless get confused as they appear to indicate that the early Miocene volcanic rocks of eastern Uganda dated by Bishop (Bishop and Trendall, 1967) rest on their ‘African Surface’ when they actually rest on Bishop’s Kyoga surface, equivalent to Ollier’s (1992) and Taylor and Howard’s (1998) lower defined African Surface.
D THE FORM AND PROCESSES OF DEEP WEATHERING

D.1 Weathering profiles

Mechanical and chemical weathering results in the disintegration and decomposition of rock in situ under the action of external agents such as water, temperature changes, plants and bacteria. Deep weathering of over 100 m depth can occur in the tropics where fast reaction rates and abundant biota are facilitated by high temperature and large rainfall totals. A variety of terms have been used to describe weathering profiles, as shown in Figure D-1 taken from Taylor and Eggleton (2001).

Figure D-1: Terms used to describe a weathering profile that may relate to the character of a specific profile or the author’s preference (after Taylor and Eggleton, 2001)

The terms fresh rock, saprock, saprolite and soil are most commonly used in the Ugandan literature. In many cases the boundary between fresh rock and saprock is associated with the oxidation of iron compounds from the ferrous (Fe$^{2+}$) state to ferric (Fe$^{3+}$) state. The weathering front moves down into the rock and therefore the material
in the most advanced stage of weathering is closest to the surface. The fresh rock/saprock interface is also often associated with increased fracture density and/or aperture. Saprock is used to describe the zone above the fresh rock that contains both weathered and unweathered material, which often contains core stones, and retains the original rock fabric. The overlying saprolite is more altered than saprock. Weathering resistant material remains in the saprolite but weathered material is wholly or partially replaced by clays, oxides and oxyhydroxides so that the original bulk fabric, but not the original material is often maintained. A considerable proportion of the material may be locally transported (Taylor and Eggleton, 2001). The original rock fabric is completely lost above the saprock. Percolating water moves soluble and finer particles downward, leaving coarse grains of low solubility material such as quartz behind. In some areas, the processes above the saprolite may be similar to soil formation. Therefore, some authors have coined the term pedolith and use the classifications O, A, E, B and C horizons for organic, leached, accumulate and parent material respectively which are used to classify temperate soils.

<table>
<thead>
<tr>
<th>Thickness (m)</th>
<th>Aroca</th>
<th>Rusangwe</th>
</tr>
</thead>
<tbody>
<tr>
<td>2 to 4</td>
<td></td>
<td>absent</td>
</tr>
<tr>
<td>5 to 14</td>
<td>0 to 21</td>
<td></td>
</tr>
<tr>
<td>&gt;3 to &gt;9</td>
<td>7 to &gt;15</td>
<td></td>
</tr>
<tr>
<td>Total:</td>
<td></td>
<td>Total:</td>
</tr>
<tr>
<td>&gt;12 to &gt;21</td>
<td>7 to &gt;27</td>
<td></td>
</tr>
</tbody>
</table>

**Figure D-2: Stratigraphy, texture and geochemistry of deep weathering in the Aroca and Rusangwe catchments (after Taylor and Howard, 1999)**

Taylor and Howard (1999b) conducted a detailed study of the lithological profile, texture, mineralogy and geochemistry of deeply weathered profiles of two catchments in Uganda. Six boreholes were drilled in the Aroca catchment, which is located in the
Apac District north of Lake Kyoga to the east of the uplifted rift flank. Four borehole were drilled in the Rusangwe (also called Nyabisheki) catchment, which is located immediately south of the western Katonga Valley and drains into the Mpanga River which flows westward in a valley incised through the uplifted rift flank. Taylor and Howard (1999b) identify three broad zones in the weathering profiles. As shown in Figure D-2 there is a progressive decrease in sand sized particles up the profile which is consistent with increased decomposition. A sandy zone occurs beneath the water table with biotite and plagioclase altered to mixed-layer clays and kaolinite respectively. Above the water table, both primary minerals and already altered material are further weathered to produce kaolinite and goethite. The profiles in Aroca are capped by massive or pisolithic laterite. Only two boreholes were drilled to relatively unweathered bedrock at approximately 7 and 12 m bgl, and these were both located in the Rusangwe catchment. All other boreholes were completed in the sandy saprolite and so the total depth of weathering is not known. Although the deepest borehole was also in the Rusangwe catchment (27 m bgl), it is clear that if originally present, the overlying laterite (ferricrete) has now been removed by erosion, and in one case, the clayey saprolite zone has been removed too.

The advanced decomposition in the upper saprolite leads to a reduction in permeability compared to the early stages of decomposition near the saprock/saprolite interface (Jones, 1985). Although the transmissivity and storativity of the lower regolith is often variable and difficult to quantify (Taylor and Howard, 2000), the utility of this zone for rural water supply in many areas of Uganda is commonly accepted (Macdonald et al., 2005). Figure D-3 shows a conceptual model of the hydrogeology of a deeply weathered regolith developed by Chilton and Foster (1995).
D.2 The relationship between laterites and land surfaces

If erosion is slow relative to the weathering rate then the saprolite evolves towards a lateritic profile in which the near surface layer becomes enriched in hydrated iron oxides. McFarlane (1970) describes the 19th and early 20th century view that lateritic duricrust is the residual material remaining after removal of the soluble minerals and reduction in that land surface which was supported by Wayland (1934a) and revived by Trendall (1962). However, the most common explanation for duricrust formation is the seasonal rise and fall in the water table. During the wet season, reducing conditions develop and ferrous iron is dissolved and together with primary quartz is leached from the weathered material, leaving behind a kaolin rich zone. Due to its light colour this has been termed the pallid zone. During the dry season, the solution is drawn into the capillary fringe, above the water table, where the water evaporates leaving behind ferric iron (Taylor and Eggleton, 2001). A hard duricrust forms in the zone of evaporation and precipitation. Decaying plants and other organisms within the regolith can create local variations in the redox conditions and produce a mottled zone between the duricrust and the pallid zone. Taylor and Eggleton (2001) note that the term ‘laterite’ has been used historically to describe both the duricrust and the entire duricrust/mottled

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Figure D-3: A typical weathering profile and commonly observed vertical distribution of hydrogeological characteristics (after Jones, 1985 and Chilton and Foster, 1993)
zone/pallid zone profile. They therefore state the term ‘laterite’ can be confusing and suggest using ferricrete to describe the near-surface zone of cemented iron oxides. Duricrusts may also be formed from other minerals including aluminium oxides (bauxite) or silica (silcrete), although these are less common in Uganda.

McFarlane (1970) conducted a detailed study of the geometry and character of weathering profiles to elucidate the relative chronology of the Buganda Surface. She refers to duricrust formed by water table fluctuations as ‘ground-water laterite’. If formed by precipitation rather than surface reduction, the presence of laterite implies that the associated erosion cycle has reached completion. McFarlane (1970) describes four forms of laterite on the Buganda surface east of Kampala, including one pedogenetic laterite and three ground-water forms; vermiform; spaced pisolithic; and packed pisolithic laterites. The pedogenetic laterite rests on the ground-water laterites and appears to have formed from them. The vermiform laterite often forms resistant tops of mesa-like hills and is characterised by well developed pipe-like structures, containing dark brown goethite surrounded by a pale yellowish-red matrix. The pipe orientation is vertical at the top of the profile, but becomes irregular with depth and horizontal at the base. As the name implies, pisoliths are well defined pea sized structures. They are generally dark red. Larger, softer, irregular concretions also occur in the spaced pisolithic laterite. They are surrounded by a kaolin and quartz matrix. This form breaks down easily and produces pisolithic rubble on exposure. It does not support mesa-like hills. The pisoliths in the packed form are regular, small and hard with a reddish yellow surface layer. This form is generally hard, and also gives rise to mesa-like hills (McFarlane, 1970).
McFarlane (1970) identified both detrital and primary laterite on the Lowland Landscape (Doornkamp, 1968b) exposed beside the Kampala to Jinja, Mbarara, Mityana, Mubende, Bombo and Soroti roads. From her careful observations, she proposed that the packed pisolithic laterite forms through a process of residual accumulation from the underlying spaced pisolithic laterite. It tends to be thicker in surface depressions. Vermiform laterites are associated with near-level topographies. McFarlane (1970) proposes a processes whereby spaced pisolithic laterites are first developed on degrading slopes, associated with long-term lowering of the water table and more oxidising conditions (Figure D-4). With time, a residual accumulation of packed pisolithic laterite forms above the spaced pisolithic laterite. Eventually, vermiciform laterites are developed from the packed pisoliths on level landscapes with stable but oscillating water tables. The duration required for their formation must be consistent with the longevity of the Lowland Landscape on which they occur.

**Figure D-4: Conceptual model of laterite development in relation to land surface development (after McFarlane, 1970)**
McFarlane (1970) suggests that the occurrence of vermiform laterites on at least two separate surfaces in the upland landscape, including the Buganda Surface, indicates two periods of water table stability associated with two planation surfaces. The conceptual model depicted in Figure D-5 shows the development of a deep weathering profile and vermiform laterite on an old planation surface, which eventually comes to form the highest elevation of the Upland Landscape. Subsequent lowering of the base level and water table results in development of spaced pisolithic laterite and dissection of the land surface. A lower vermiform laterite forms on the lower surface, although remnants of the older, higher, surface and associated vermiform laterite remain. McFarlane (1970) has named the lower vermiform laterite in the Lugazi area, 30 km east of Kampala, the Ntenga Surface. Detrital packed pisolithic laterite forms on and at the base of the slopes. Finally, another reduction in the base level resulted in the formation of the Lowland Landscape. While the vermiform laterite is associated with the planation surfaces and the spaced pisolithic laterite is associated with the slopes, the detrital packed pisolithic laterite occurs on both. The base of the weathered rock forms a complex pattern dependent on the rock susceptibility and the history of exhumation.
McFarlane (1970) warns that if laterite sheets on the flat-topped hills are polycyclic, then so are the breaks of slope, which is unfortunate when previous interpretations have used them for considerable diagnostic purposes. She emphasises the utility of defining planation cycles based on the development of vermiform laterite, but also stresses the difficulty with attributing altitudinal variations to lithological controls, tectonics or recent settlement. If the Lowland Landscape, including the Kasubi/Kyoga Surface formed prior to lower Miocene volcanism (~19 Ma) in eastern Uganda then all laterite surfaces in the Upland Landscape, including the Buganda Surface, must be even earlier. Furthermore, all else being equal, McFarlane (1970) suggests that the greater development of vermiform laterite on the Upland Landscape suggests a longer duration for their development than that on the lowland landscape.

### D.3 Etchplanation

More recently, a number of authors (Thomas, 1989a, Thomas, 1989b, Twidale, 2002) have returned to the idea of ‘etchplanation’ first attributed to Wayland (1934a) and Willis (1934, in Twidale, 2002) and the associated ideas of Büdel (1957). This theory explicitly incorporates the process of deep weathering into a two-stage landscape development. The first stage is the development of a deeply weathered mantle, which may incorporate differential weathering of susceptible geological features. The second stage is the stripping of the regolith to expose the bedrock form (Twidale, 2002). This process has been used to explain inselbergs and large surface boulders as resistant core stones. Twidale (2002) points out that a two-stage development implies periods when weathering is dominant followed by periods when erosion is dominant. This paradigm was applied by Taylor and Howard (1998) in their conceptual model of the post-Paleozoic evolution of Ugandan land surfaces by tectonically controlled deep weathering and stripping. Thomas (1989a, 1989b) has linked etchplanation to Quaternary environmental changes in West Africa. However, he has also noted that where development of the weathering front and erosion occur at similar rates then the surface may be in dynamic equilibrium. While most authors assume that surface processes dominate the erosion stage of etchplanation, McFarlane (1989) appears to contend that subsurface dissolution is the primary process by which material is lost. Near-surface leaching results in collapse of the saprolite and formation of a colluvial mantle.
D.4 Dating weathering profiles

The task of dating the time of weathering itself is a difficult challenge which has not yet been attempted in Uganda, although palaeomagnetic techniques have been used elsewhere. These methods assume the weathering front moves downwards and the Earth’s polarity is recorded by formation of iron oxides as it does so. The magnetostratigraphy may be compared against the pattern of polar reversals which is well understood back to about 100 Ma. The apparent polar wander paths (APWP) created by continental drift have also been used in a similar way. Iron oxide precipitates developed on the Deccan Traps appear to show they began to weather soon after being erupted 68 to 60 Ma (Schmidt et al., 1983, in Taylor and Eggleton, 2001). The first use of APWP was conducted by Idnum and Senior (1978) on Cretaceous and inset Tertiary sediments in Queensland, Australia. They showed weathering ‘events’ in the late Cretaceous (145 to 65 Ma) and Miocene (23 to 5 Ma). The oldest weathering profile dated in this way is of Carboniferous (359 to 299 Ma) age in the Ordovician rocks of New South Wales, Australia (Pillans, 1998, in Taylor and Eggleton, 2001). The weathered surface is incised by a palaeochannel which itself contains weathered sediment of Oligocene (34 to 23 Ma) or Miocene (23 to 5 Ma) age. These results have shown that very old weathering profiles do exist, although Taylor and Eggleton (2001) warn that the results of this work has been inappropriately extrapolated to similar weathering surfaces, without confirmatory dating, and this has led to erroneous interpretation of landscape development over large parts of Australia.

Théveniaut and Freyssinet (1999) examined the palaeomagnetism in a 54 m thick lateritic profile in French Guiana. They discovered five magnetic reversals in the saprolite profile with the youngest at a depth of 8.8 m. This suggests an average rate of saprolite formation of 11.3 ±0.5 m/Ma. For comparative purposes only, this figure would correspond to a minimum age for the mean (4 2 m) and maximum (116 m) depths of weathering in the Katonga region of 3.7 Ma and 10.3 Ma respectively. This appears relatively young in comparison to other evidence and ignores the feedback between the rate of development of the weathering front at depth and erosion from the top of the profile in determining the depth of weathering at the present time.

Vasconcelos et al. (1994) use direct K/Ar and $^{40}$Ar/$^{39}$Ar dating of potassium and manganese oxides to establish the chronology of laterite formation in Precambrian
rocks of Brazil. They showed an almost continuous spread of ages between 72 and 20 Ma. Thus it can be shown there is much evidence from around the world for a wide range of potential ages for well developed tropical weathering profiles. The depth and stage of evolution of weathering profiles in Uganda therefore appears to be a poor indicator of the absolute chronology of their dynamic history of development. Without direct palaeomagnetic or radiometric dating it is not possible to assign an absolute age to the development of units within the deeply weathered profiles of Uganda and all we can say is that they are older than overlying biostratigraphically dated sediments or radiometric dated volcanic rocks.
E HISTORICAL STUDIES OF THE WESTERN RIFT SEDIMENTS

E.1 The Albert Basin

Wayland (1925) initially classified the sediments of the Albert Basin into the arenaceous Kisegi Beds and the overlying argillaceous Kaiso Beds. He identifies both sequences near the Kisegi River on the northern tip of the Rwenzori block where they have been uplifted by the Rwenzori horst and tilted to the north. He describes the Kisegi Beds as largely arenaceous and sometimes conglomeratic, with false bedding and frequently gypsiferous, and interprets them to be deposited in shallow water during the initial Miocene subsidence of the rift. He describes the Kaiso Beds, which outcrop above the northward dipping Kisegi beds and 100 km to the northeast near the village of Kaiso on the eastern side of Lake Albert, as being of considerable thickness and typically arenaceous sediments with intercalations of sands and ironstones containing many fossils, including molluscs, fish and mammals. Wayland (1925) suggests that the Kaiso beds are associated with deposition in Lake Obweruka which developed in the late Miocene and extended from the current Albert Basin to the Edward Basin prior to uplift of the Rwenzori horst. In later reports Wayland (1934b) envisaged that the fossiliferous horizons represented death assemblages due to stages of lake desiccation during ‘interpluvial periods’.

Pargeter (1948, and Harris et al., 1956) suggested that fine-grained sandstone outliers identified in the Nkusi and Muzizi valleys (also observed by the author in the Katonga Valley) were of equivalent age to the Kisegi Beds. However, the author has observed that the Kisegi beds are considerably more friable than the indurated sandstones of the river valleys which give the general impression of having undergone more significant diagenesis due to greater age and/or depth of burial. Pargeter (1948) split the Kaiso Beds into a lower and upper sequence which he ascribes to the Pliocene and Pleistocene respectively. He describes the Lower Kaiso beds as predominantly argillaceous, and the Upper Kaiso beds as argillaceous. Adjacent to the fault scarp they are overlain by late Pleistocene high level beach and river gravels, which are in turn overlain by what he terms ‘red earth’. Finally, he describes the recent lake beach gravels.
Figure E-1: Sketch map showing areas of accessible outcrops in the Western Rift, west of the Katonga Valley (after Roller et al., 2010)

Bishop (1965, 1967) returned to the area around the Kisegi and Wasa rivers between the Rwenzori Mountains and Lake Albert and describes the following sequence from top to bottom:

- Recent swamp deposits and low-level strand lines of Lake Albert

Tectonic activity

- **Wasa Beds and Plateau Gravels** – ‘White to drab clayey sands with gravelly horizons’ 30m (100 ft)

Tectonic activity

- **Kaiso Series** – ‘Parallel-bedded, poorly consolidated, drab, gray to greenish clays...interbedded with buff micaceous silts and fine sands, and contain local rusty ironstone bands’ 823 m (2700 ft)

- **Passage Beds** – ‘Sandstones...interstratified with thick beds of dense clay that range from dark red to blue and greenish gray’ 198 m (650 ft)
• **Kisegi Series** – “Coarse grits, hard and soft sandstones locally showing current bedding, and subordinate clay horizons with local bands of lignite or gypsum, together with some bands of ferruginous or calcareous nodules” 274 m (900ft)

Bishop (1965) estimates the total thickness of exposed sediments to be about 1,325 m, although Pickford et al. (1993) have more recently stated that the total is only 660 m. Current bedding in the Kisegi Series suggests the sediment source was to the northwest. The dip on the bedding gradually decreases from about 12° in the southwest of the outcrop to between 7° and 5° in the northeast. He describes the Kisegi Series as similar to the Mohari Beds located about 30 km to the north west in the D.R. Congo, which have been dated by the presence of Miocene fauna. He indicates a gradual disappearance of the coarse sandstones and grits to the east of the Kisegi River, and names these intermediate strata the Passage Beds. Within the Kaiso Series, Bishop (1965) describes the fossiliferous ferruginous horizons as discontinuous, frequently sandy and sometimes flaggy or oolitic. Towards the escarpment they ‘tend to thicken and become first a grit and then a medium-grained quartz gravel, cemented by limonite’. Bishop (1965) proposes that the rhythmic sedimentation is due to cycles of deposition and subsidence of the rift floor rather than climatic change as proposed by Wayland (1930). He interprets the ironstone bands to be the sites of shallow basins that occasionally dried out and were later rapidly inundated providing conditions suitable for fossil preservation (Bishop, 1969).

Several researchers (Gautier, 1970, Pickford et al., 1993, Taylor and Howard, 1998) have suggested that the change from arenaceous to argillaceous sediment and the increase in ferruginous horizons indicates a change in the source region from an environment dominated by erosion to one dominated by deep weathering and iron saturated run-off. Taylor and Howard (1998) support the hypothesis of enhanced weathering at this time by citing evidence of humid forests in late Miocene sediment of the Kaiso area (Deschamps et al., 1992, Pickford et al., 1993) and laterite (ferricrete) underlying the fluvialacustrine Namiska Beds in the Murchison Falls area north of Lake Albert (Bishop, 1965). The Namiska Beds are comprised of ‘pebble gravels false-bedded sands and grits, which are commonly highly kaolinitic, to fine sands and gritty clays’ over 120 m thick with laterite developed on their upper surface (Bishop, 1965). Bishop (1965) states ‘The age of the Namiska beds is unknown, but their base is
tentatively assigned to the lower Miocene, in view of their similarity to the Kisegi Series and the Mohari Beds’. Whilst he acknowledges that their similar character may simply indicate a similar environment of deposition he also points out ‘the time required for duricrust formation, the deposition of succeeding sediments and the general tectonic history established for the Albert rift suggests that they are probably as old as Miocene’. Recent oil exploration in the Murchison Falls area has encountered upper Pliocene reservoirs at 1000 m bgl in the Victoria Nile delta play (Ovington and Burden, 2009) 15 km west of the area described by Bishop (1965), but unfortunately the geological relationship between drilling area and the Namiska Beds has not been reported.

Further south, Wayland (1925) described a break in slope on the rift escarpment east of Lake Albert which he names the ‘hanging base line’. Bishop (1965) speculates that the subdued topography above the break in slope is associated with the sluggish rivers which drained into the low energy lacustrine environment in which the Kaiso Series was deposited. He further speculates that the fault movement which brought about the steeper lower escarpment, and increased the relief on the rift flank, was responsible for the change in sedimentation to the younger and coarser sediments of the Wasa Beds and the Semliki Series. The Plateaux Gravels appear to have been deposited by streams from the Rwenzori Mountains and the main rift scarp following erosion of the underlying Kaiso Series. They grade into the upper horizons of the fluviolacustrine Wasa Beds, which contain Middle Stone Age artefacts of late Pleistocene age.

Several of the hypotheses described in these historical studies such as: the continuity of Lake Obweruka from the Albert Basin to the Edward Basin; the correlation between iron deposition and deep weathering; and, the association between the break in slope on the rift shoulder with late Pleistocene rift flank uplift and renewed arenaceous deposition together form a consistent story. However, each inference is non-unique and caution is required not to assign an unwarranted level of certainty to the overall interpretation. Many of these interpretations would be regarded as over-simplistic by modern geoscientists.
E.2 The Semliki Basin

The Semliki Basin lies to the west of the Rwenzori Mountains in the D.R. Congo. The Semliki River flows northward between Lake Edward and Lake Albert. Given its location adjacent to the opposite rift flank to the Katonga region the sedimentary record in the Semliki Basin is only considered briefly here. Whilst acknowledging the work of many early researchers in this area (e.g., Hooijer, 1963, Lepersonne, 1970, Gautier, 1970, cited in Pickford et al., 1993) this summary focuses on the accounts of Verniers and Heinzelin (1990) and Makinouchi et al. (1992) of the geology of the upper (i.e., near Lake Edward) and lower (i.e., near lake Albert) Semliki valley respectively. The location of the study sites is shown in Figure E-1 and the stratigraphy and lithological summaries are presented in Table E-1.

The Lusso Beds (Table E-1) outcrop next to Lake Edward near the exit of the Semliki River (Figure E-1). Verniers and Heinzelin (1990) describe these beds as mainly clay, silts and fine sands with ironstones. They are likely of late Pliocene age based on the presence of the G5 mollusc association (Pickford et al., 1993). Verniers and Heinzelin (1990) suggest that the Lusso facies are similar to the Kaiso Beds in Uganda and the Sinda Beds in the Lower Semliki and that they were likely deposited in a lacustrine environment at various depths. However, they also point out that it has not been substantiated that all of the Pliocene lacustrine sediments were deposited in the single large Lake Obweruka, first postulated by Wayland and supported by Pickford et al. (1993). The overlying Semliki Beds seem to indicate a change from sediment supply controlled to more accommodation controlled conditions during the early Pleistocene with lateral facies varying from lacustrine to high energy fluvial with northward current directions. Successive base level reductions have created a series of terraces in the late Pleistocene and Holocene fluvial sands and gravels. Ash from the Katwe crater field across the border in Uganda dated as 6,890 ±75 years BP covers the lower terrace.
### Table E-1: Summary descriptions of sediments in the Upper Semliki valley (after Verniers and Heinzelin, 1990) and Lower Semliki valley (after Makinouchi et al., 1992)

<table>
<thead>
<tr>
<th>EPOCH</th>
<th>UPPER SEMLIKI</th>
<th>LOWER SEMLIKI</th>
</tr>
</thead>
<tbody>
<tr>
<td>Holocene</td>
<td><strong>Recent Deposits</strong> – Including gully infill</td>
<td><strong>Recent Deposits</strong> – sand, gravel and cobbles of Sinda and Mohari Rivers and tributaries</td>
</tr>
<tr>
<td></td>
<td><strong>Katwe Ash</strong> – 6,890 ±75 years BP (8 to 30 m)</td>
<td></td>
</tr>
<tr>
<td>Late Pleistocene</td>
<td><strong>Kabale 2</strong> (3.5 m above river) – Gravel and sands, lower fluvial &amp; upper colluvial gravels</td>
<td><strong>Lower Terrace</strong> – Sands and gravels a few metres above current river</td>
</tr>
<tr>
<td></td>
<td><strong>Lower Terraces</strong> (8 to 13 m above river) – (22 to 12 ka) Fine gravel and sand</td>
<td></td>
</tr>
<tr>
<td></td>
<td><strong>Kabale 4</strong> (18 to 20 m above river) – Gravels</td>
<td><strong>Middle Terrace</strong> (10 m above river) – Brown fine to coarse sand and scarce gravel. Possibly equivalent to Rwebishengo Beds (~10m thick)</td>
</tr>
<tr>
<td></td>
<td><strong>Upper Terraces</strong> (27 to 31 m above river) – e.g., Kasaka terrace : fine gravel, medium sorted, grading upwards into poorly sorted clay with silt, sand and calcrete (0.4 to 5 m)</td>
<td></td>
</tr>
<tr>
<td>Early Pleistocene</td>
<td><strong>Semliki Beds</strong> – Three lateral facies: 1) North flowing wide fluvial channel deposits with lower gravelly sand, middle clayey sand and upper hard clayey palaeosols; 2) North flowing high energy cross-beded fluvial deposits; 3) Lacustrine clay deposits with medium to coarse sand throughout, and Oldowan artefacts (total over 32m)</td>
<td><strong>Higher Terrace</strong> – Gravels, cobbles and boulders ‘several metres thick’.</td>
</tr>
<tr>
<td>Pliocene</td>
<td><strong>Lusso Beds</strong> (G5) – Mainly clay, smectite, micaceous, light-coloured fine silts and some sands with (oolitic) ironstone beds. Mammal fossils indicate older than 2.1ma BP Similar to Sinda Beds and Kaiso Beds. Not substantiated if these belong to the same Lake Obweruka (&gt; 52 m).</td>
<td><strong>Sinda Beds: Upper Member</strong> (G5/GX) – 20 m of weakly consolidated sand and silt</td>
</tr>
<tr>
<td></td>
<td></td>
<td><strong>Sinda Beds: Middle Member</strong> (G3/G4) – 30 m partly consolidated alternating sand and silt with basal gravel.</td>
</tr>
<tr>
<td>Miocene</td>
<td></td>
<td><strong>Sinda Beds: Lower Member</strong> – 100 m of mainly clayey, white coarse sand with much granule gravel and quartz grains. Upper part: Ongoliba. Lower part: Edo Joh (also known as Kabuga Beds and Mohari)</td>
</tr>
</tbody>
</table>
The base of the Lower Semliki succession (Table E-1) begins with a coarse sand horizon at the base of the Sinda Beds Lower Member, named the Edo Joh by Makinouchi et al., (1992) but previously named the Kabuga Beds and the Mohari (Lepersonne, 1970, Hooijer, 1963, cited in Pickford et al., 1993). It is unfossiliferous and thought to be Miocene to early Pliocene in age. The overlying upper part of the Sinda Beds Lower Member, previously known as the Ongoliba contains similar fauna to the early Pliocene Nkondo Member in the Albert Basin (Pickford et al., 1993). The overlying Sinda Beds Middle Member and Upper Member contain a faunal sequence from early Pliocene to early Pleistocene Fauna (G3 to GX). As a whole the Sinda Beds are predominantly fine-grained clays and silts with sand horizons. Similarly to the Upper Semliki, successive base level changes have produced a sequence of terraces in the overlying Pleistocene and Holocene fluvial sands and gravels.

In general, the broad geological history of the Semliki Basin appears to be consistent with that of the Albert Basin, with the initial late Miocene downwarp and accommodation limited basin conditions, followed by the development of more supply-limited basin conditions and lacustrine sediments in the Pliocene. Later fluviatile Pleistocene sediment has been eroded to form terraces. Recent researchers do not appear to have explicitly considered the potential causes of the overall reduction in base level during the Pleistocene in the Semliki Basin. The apparent regional tilt discussed in the Section 4.4.1 provides one explanation, although climatic change and downcutting at the exit of the White Nile are other possible explanations. Normal faults occur north and south of the Lower Semliki study area adjacent to the western rift flank, and these may be locally responsible for changing base level during the Pleistocene (Makinouchi et al., 1992).

**E.3 The Edward and George Basins**

Figure E-2 shows the geological map (D.G.S.M., 1962) of the area west of the Katonga Valley, including the George Basin and the Kazinga Channel which connects Lake George to Lake Edward. The sediments exposed in the rift valley are all assigned a Pleistocene to Holocene age and categorised into two lithologies. ‘Sands, clays, grits and gravels’ are shown northeast of Lake Edward adjacent to the Kazinga Channel and around the perimeter of the George Basin. Younger ‘swamp deposits, alluvium and lacustrine deposits’ are shown overlying the coarser grained sediment in the centre of
the George Basin. This demonstrates that the current elevated relief on the rift flanks is not sufficient by itself to result in widespread arenaceous sedimentation in the rift. While it is likely that high relief on the rift flanks will result in coarse-grained axial deposition (alluvial fans/colluvium), the occurrence of basin wide arenaceous deposition depend on the maintenance of stream power (high gradients and discharge) over a large axial distance and/or in the longitudinal direction. The development of accommodation limited or sediment supply limited basin conditions depends on a complex relationship between stream power, tectonics and climate. The basin system is dynamic and the character of the sedimentary record depends on the relative rates of vertical movement, erosion and sedimentation, rather than on any simple relationship with rift flank relief.

Figure E-2: Geological map of the George Basin (D.G.S.M, 1962)
Figure E-3: Geological sketch map of northeast Lake Edward and the Kazinga Channel area (after Bishop, 1969)

Figure E-4: Geological section of the Kazinga Channel area (after Bishop, 1969)

Figure E-3 shows the geology around northeast Lake Edward and the Kazinga channel according to Bishop (1969). The lithological descriptions of Bishop (1969) and Pickford et al. (1993) are summarised in Table E-2. Bishop (1969) identifies the fine-grained deposits with ironstones immediately east of Lake Edward as equivalent to the ‘Kaiso Series’ of the Albert Basin. He describes the Edward Basin ‘Kaiso Series’ as
Pleistocene, however, Pickford et al. (1993) name deposits from the same area the Kazinga Beds and state that they contain mollusc association G3b, which would give them an early Pliocene age (4.9Ma). Bishop (1969) states that the ‘Kaiso Series’ dip to the east under current bedded sand and gravel which he identifies as equivalent to the Semliki Series already described in the Upper Semliki area across the border in the D.R. Congo. Pickford et al. (1993) name the deposits overlying the Kazinga Beds, the Bushabwanyama Beds and state that they contain mollusc association G5 which indicates they are upper Pliocene (3Ma). This biostratigraphic classification suggests they are equivalent to the upper Lusso Beds in the Upper Semliki, the Sinda Beds Upper Member in the Lower Semliki. Pickford et al. (1993) appear to dismiss Bishop’s (1969) assignment of the Kazinga area sediments to the Semliki Series although they do not describe the eastern part of the outcrop shown in Figure E-3.

<table>
<thead>
<tr>
<th>N. EDWARD AND S. GEORGE BASINS, AND KAZINGA CHANNEL</th>
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<tbody>
<tr>
<td>(after Bishop, 1969)</td>
</tr>
<tr>
<td>Mainly erosion with local deposits</td>
</tr>
<tr>
<td><strong>Gully fill</strong> – Re-worked heterogeneous deposits</td>
</tr>
<tr>
<td><strong>River gravels</strong> – coarse cross-bedded gravels with</td>
</tr>
<tr>
<td>limonite staining (up to 30 m above Lake Edward).</td>
</tr>
<tr>
<td>Contains Acheulian artefacts on Mweya peninsula.</td>
</tr>
<tr>
<td><strong>Kagwa and Bunyaruguru Volcanics</strong> – coarse</td>
</tr>
<tr>
<td>agglomerates and up to 20 feet of fine ash ‘8 to 10 ka’</td>
</tr>
<tr>
<td><strong>Semliki Series (Katanda Formation)</strong> – horizontal</td>
</tr>
<tr>
<td>current bedded fluvio-lacustrine sands and gravels.</td>
</tr>
<tr>
<td>Base indicated by coarsening grade and lack of</td>
</tr>
<tr>
<td>ferruginous horizons. Bishop estimated *Mid to</td>
</tr>
<tr>
<td>beginning of Later Pleistocene* (post dated by</td>
</tr>
<tr>
<td>Acheulian and artefacts from Mweya peninsula)</td>
</tr>
<tr>
<td><strong>Kaiso Series</strong> – Pale buff to grey, interbedded,</td>
</tr>
<tr>
<td>fine-grained micaceous sands, silts and clays.</td>
</tr>
<tr>
<td>Ferruginous fossiliferous horizons. Dips beneath</td>
</tr>
<tr>
<td>Semliki Series. Bishop estimated <em>early Pleistocene</em></td>
</tr>
<tr>
<td>(after Pickford et al., 1993)</td>
</tr>
<tr>
<td><strong>Lion Bay Beds</strong> – silts and flaggy limestones</td>
</tr>
<tr>
<td>with terrestrial molluscs and modern fauna.</td>
</tr>
<tr>
<td>Tuffs from Katwe and Bunyaruguru Volcanics</td>
</tr>
<tr>
<td><strong>Hima Limestones</strong> – volcanic origin, with clays</td>
</tr>
<tr>
<td>and silts</td>
</tr>
<tr>
<td><strong>Buyaruguru Volcanics</strong> – sub-aerial tuffs with</td>
</tr>
<tr>
<td>leaf impressions suggesting forest on rift flank</td>
</tr>
<tr>
<td><strong>Bushabwanyama Beds</strong> – Upper Pliocene (G5)</td>
</tr>
<tr>
<td>Includes ironstones (&gt;20 m)</td>
</tr>
<tr>
<td>Appears to be similar location to base of Bishop’s</td>
</tr>
<tr>
<td>‘Semliki Series’</td>
</tr>
<tr>
<td><strong>Kazinga Beds</strong> – Lower Pliocene (G3b), micaceous</td>
</tr>
<tr>
<td>silts and clays with several ironstone horizons (&gt;15m)</td>
</tr>
<tr>
<td>Appears to be similar location to Bishop’s ‘Kaiso</td>
</tr>
<tr>
<td>Series’</td>
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</tbody>
</table>

Table E-2: Summary descriptions of sediments in the area between the Edward and George Basins

Pickford et al. (1993) consider the Kazinga Channel sequence important because it demonstrates that the Pliocene fauna underwent similar evolutionary history to that in the Lake Albert area and south of Lake Edward. They use this as supporting evidence for Wayland’s (1934b) single large Lake Obweruka which was of a similar size to the
present day Lake Tanganyika and Lake Malawi. It is not clear why they preclude the possibility of molluscan and other fauna moving between lakes via connecting rivers.

Pickford et al. (1993) describe a series of silts and flaggy limestone containing Holocene fauna overlying the Bushabwanyama Beds which they name the Lions Bay Beds. These beds contain tuffs derived from the Katwe and Bunyaruguru Volcanic Fields. Figure E-4 (Bishop, 1969) shows a layer of ‘wind blow ash’ covering the area. Leaf impressions in the Bunyaruguru tuffs suggest forested rift margins and valley floor grasslands as early as the late Pleistocene.

Bishop (1969) describes unconsolidated river gravel with limonite staining and concretions up to 30 m above the current Lake Edward. The Kazinga Channel itself is incised by 30 to 90 m into the surrounding surface between Lake George and Lake Edward. The terrace sequences in the Semliki Valley and incision of the Kazinga Channel would suggest a relatively recent base level reduction throughout the Edward, Semliki and Albert Basins. Russell et al. (2003) use the presence or absence of evaporite minerals among other lake sediment parameters to suggest changes in the level of Lake Edward during the Holocene. They propose a period of aridity in the late Pleistocene, followed by wetter climate in the early Holocene which they suggest could correlate with raised shorelines 14.5 m above the present Lake Edward, Kazinga Channel and the Semliki River. They present lithological evidence for the development of a sub-lacustrine fan associated with entrenchment of the Kazinga Channel which they estimate began about 8600 yr BP. They suggest the base level lowering occurred due to local tectonics around the Semliki outlet. The onset of mid-Holocene aridity resulted in a minimum recent low stand of about 14 m below the current lake level between 4,000 and 2,000 yr BP. Russell et al. (2003) state that the current level of Lake Edward was established about 1,700 yr BP.
F APATITE FISSION TRACK ANALYSIS THEORY AND METHODS

F.1 Parameter Definition and Measurement

There are four parameter types required for AFT analysis (Donelick et al., 2005):

1. Induced fission track densities – used to determine relative Uranium concentration and calculate ages;
2. Spontaneous fission track densities – used to calculate ages;
3. Fission track lengths – used to constrain cooling history; and
4. Fission track annealing kinetic parameters – used to estimate annealing behaviour and constrain cooling history.

The following sections describe the measurement of each of these parameter types.

F.1.1 Induced fission track density

The external detector method was used to determine the relative \( ^{238} \text{U} \) concentration in apatite during this study. In this method, fission is induced in \( ^{235} \text{U} \) by irradiating the sample with low-energy thermal neutrons and the resultant fission tracks produced in an external detector are counted. High-energy neutrons induce fission from \( ^{238} \text{U} \) and \( ^{232} \text{Th} \) and are therefore to be avoided. The ratio \( ^{238} \text{U}/^{235} \text{U} \) is constant in nature and therefore the concentration of \( ^{238} \text{U} \) can be estimated (Gallagher et al., 1998a). The external detector is a low-uranium fission-track free muscovite mica cleavage sheet, which is placed in contact with the mounted apatite grains. One corner of the mounts and the mica sheets is cut so that they can be aligned in their original orientation for comparison once separated. The practical procedure for measuring track density is the same as for spontaneous tracks which is described in the following section.

During this study, the mount-mica pairs were irradiated using the Forschungsreaktor Heinz Maier-Leibnitz (FRM II) facility at the Technische Universität München. It is difficult to determine the thermal-neutron flux applied directly and therefore silicate glass standards of known uranium concentration with their own external detectors are placed at each end of the stack of apatite grain mounts (Donelick et al., 2005). Uranium-doped Coning Glass CN-5 was used as the dosimeter in this study. The number of induced fission tracks \( (N_d \text{ in Equation F-3}) \) is counted on an area of the
external detector in contact with glass standard to determine the induced fission track density ($\rho_d$ in Equation F-2). A total count of over 4000 induced tracks is reasonable to exceed the number of spontaneous tracks in most samples and not limit the precision of the age calculation. If there is a significant gradient in the neutron flux then the induced fission track density will be different in each glass standard and the corresponding fission track density must be interpolated along the stack for each sample position. No neutron flux gradient was detected during this study and therefore no interpolation was necessary.

The mica detectors were etched in 48% hydrofluoric acid (HF) for 25 minutes at room temperature to reveal the induced fission tracks. The number of induced fission tracks ($N_i$ in Equation F-3) is counted on an area of the external detector for a specific apatite grain to determine the induced fission track density for that grain ($\rho_i$ in Equation F-1). The external detector method is used to determine the relative uranium concentration by comparing the density of induced fission tracks for a grain to the density of induced fission tracks for the glass standard with known $^{235}$U concentration. It will be discussed further in Section F.2.1 that these relative values are related to the $^{238}$U concentration in the apatite grains through the use of an analyst specific calibration factor (zeta, $\zeta$ in Equation F-2) which is established using the same glass standards and samples of known age.

**F.1.2 Spontaneous fission track density**

Fission tracks form in random orientations, but the measured track density is based on a two-dimensional sample of tracks intersecting the polished surface of the grain (Gallagher et al., 1998b). Grains are chosen that are cut parallel to the crystallographic c-axis which is more resistant to etching (low bulk etch rate) and retains some polish scratches. Such orientations provide the closest match to a geometry factor of 0.5 based on the $4\pi/2\pi$ ratio of sample to external detector. Those faces perpendicular to the c-axis are often pitted in appearance. Since the age calculation assumes that fission tracks have intersected the polished surface from above or below, the density should not be measured within 10 $\mu$m of the edge of the grain. Fission tracks close to the grain edge are not only formed by intra-grain decay particles which are restricted to inside the grain boundary, but may also be formed by inter-grain decay particles which have
entered from outside the grain boundary. The area to be counted should be free of imperfections and crystallographic defects.

Fission track density and length measurements were undertaken during this study with a Zeiss axioplan microscope fitted with an automated XY stage and digitising tablet under computer control using the counting software FT Stage 4.04 written by Trevor Dumitru of Stanford Fission Track Thermochronology Laboratory. Measurements were undertaken with a magnification of 1250× (100× objective, 1.25× condenser, 10× ocular). The ocular lens is fitted with a graduated scale known as an eyepiece graticule. The number of spontaneous fission tracks ($N_s$ in Equation F-3) is counted over a selected area for a specific apatite grain to determine the fission track density for that grain ($\rho_s$ in Equations F-1 and F-2) (Donelick et al., 2005). The area is determined by counting the number of whole graticule squares ($N_g$) over which the number of fission tracks is counted.

All fission track counts and measurements for this study were undertaken by Dr. Andrew Carter, an experienced AFT analyst and Head of UCL and Birkbeck Thermochronometry Research Laboratory. Fission track analysts are faced with many decisions when counting the track density and measuring the track length (Donelick et al., 2005). For example, when counting spontaneous tracks, they need to decide which features are fission tracks, which features in the border area should be included. When counting induced tracks, they also need to decide which area of the external detector corresponds to the counted area of the apatite grain. The automated microscope stage and connected software allows the orientation of the apatite mounts and external detectors to be coordinated through triangulation. To minimise counting bias, the analyst was unaware of the sampling locations and geology, predetermined criteria were used to select fission tracks on the boundary of the counting area and the running totals were not viewed during counting.

**F.1.3 Fission track lengths**

Tracks which intersect the grains surface are truncated and therefore the original track length is unknown. The convention is therefore to measure the full etched lengths of near horizontal tracks which have both ends confined within the apatite grain. Therefore, the etchant must have reached the confined track via other open features that
intersect both the surface of the grain and the confined etched track. Donelick et al. (2005) described three categories of such features as follows:

- **TINT** (Track-IN-Track) – the etchant has reached the confined track via another track;
- **TINCLE** (Track-IN-CLEavage) – the etchant has reached the confined track via cleavage;
- **TINDEF** (Track-IN-Defect) – the track has reached the confined track via a crystallographic defect.

TINCLEs and TINDEFS have been shown to be statistically longer than TINTs (Barbarand et al. 2003b, Jonckheere and Wagner, 2000) perhaps due to an original reduced resistance to the fission particle or post-formation resistance to annealing. Therefore, only TINTs should be used to determine the track length distribution. Since only confined tracks are included, the fission tracks used for length measurement are drawn from a different sample to those used to determine the track density, which all intersected the surface of the grain. Horizontally confined tracks can be identified by the strong back scattering that they produce in reflected light. However, their full length should be measured in transmitted light. Similarly to during track density counting, only apatite grains parallel to the crystallographic plain are used. In addition, only confined tracks within approximately 10° from horizontal are measured, which gives a maximum length error of less than 2%. In practice, this means selecting tracks with clearly visible ends, without adjusting the focus. A small red dot visible in the eye piece is placed at one end of the track and the position is recorded in the software before the moving the red dot to the other end of the track. The procedure used during this study was to place the red dot outside of the track but touching the end for the first measurement and inside the track but touching the end for the second measurement. Track length can be measured to within ±0.2 μm and typically 50 to 150 individual track lengths are measured.

The mean length of fission tracks in apatite varies according to their orientation relative to the crystallographic long c-axis and therefore it is important to record the fission track orientation relative to the c-axis. Although the long axis of the crystal is not always apparent in the shape of the apatite grain, the orientation of the c-axis on the
grain surface, itself selected parallel to the c-axis, can be determined using the preferential orientation of etch pits. In reflected light, etched geometric figures of hexagonal appearance can be seen where the fission tracks intersect the polished surface and these etch figures are elongated in the direction of the c-axis. The measurement grid is aligned parallel to the elongated etch pits and the c-axis orientation is recorded by selecting two widely spaced points on the same grid line. This is done for each grain before any length measurements are taken and the software automatically calculates the orientation of each fission track measured subsequently relative to the c-axis orientation.

F.1.4 Fission track annealing kinetic parameters

The annealing behaviour of apatite fission tracks is influenced by the relative proportions of chloride, fluoride and hydroxide in the host apatite composition (Donelick et al., 2005). Electron probe microanalysis is sometimes used to measure the percentage weight of chloride (Cl wt%) directly. However, it is currently more common to use an empirical measurement based on the etch pit geometry known as $D_{\text{par}}$ (Donelick, 1993) as an indicator of the apatite composition. $D_{\text{par}}$ is the mean maximum etch figure diameter parallel to the c-axis. In other words, it is the distance between the tips of an elongated hexagonal-like etch pit. During this study, $D_{\text{par}}$ varied between about 1.6 and 2.6 $\mu$m. $D_{\text{par}}$ is positively correlated with Cl wt% and OH wt% and negatively correlated with F wt%. Donelick et al. (2005) recommend that where possible the average $D_{\text{par}}$ is calculated for each grain on the basis of at least four $D_{\text{par}}$ measurements.

F.2 Fission Track Age Assessment

F.2.1 Single grain age calculation

The induced fission track densities measured in the external detector and the dosimeter and the spontaneous fission track density measured in the apatite, are related to the fission track age for each individual apatite grain using Equation F-1 (Price and Walker, 1963, Naeser, 1967, Fleischer et al., 1975, Galbraith, 2005).

\[
t = \frac{1}{\lambda_d} \ln \left( \frac{\lambda_d}{\lambda_f} I \sigma_f \frac{\rho_s}{\rho_i} g + 1 \right)
\]  

Equation F-1
Where

t – single grain fission track age (Ma)
λd – α particle decay constant for $^{238}$U ($\approx 1.55 \times 10^{-4}$ Ma$^{-1}$)
λf – spontaneous decay constant for $^{238}$U ($\approx 7 \times 10^{-11}$ Ma$^{-1}$)
I – $^{235}$U/$^{238}$U isotopic ratio in nature ($7.25 \times 10^{-3}$)
σf – fission cross section of $^{235}$U ($580.2 \times 10^{-24}$ cm$^2$)
Φ – neutron fluence during irradiation (neutrons/cm$^2$) where $\Phi \sigma_f$ is set $< 10^{-5}$
ρs – spontaneous fission track density
ρi – induced fission track density

In general, it can be seen from Equation F-1 that the greater the proportion of spontaneous fission tracks to induced fission tracks ($\rho_s/\rho_i$), the greater the age. A geometry factor ($g$) is required to account for the fact that the charged particles that created the fission tracks in the detector could only have come from the direction of the sample, whereas the particles creating the fission tracks in the sample could have originated from above or below the polished surface. $^{238}$U may decay through alpha particle emission or spontaneous decay, and whilst alpha emission decay rate ($\lambda_d$) is much larger than the spontaneous decay rate ($\lambda_f$), both are taken into account. Since the external detector method measures the relative abundance of fission tracks induced by $^{235}$U decay, but spontaneous fission tracks occur due to $^{238}$U, the age equation also includes the ratio $^{235}$U/$^{238}$U ($I$). The induced $^{235}$U fission track density depends on the neutron fluence ($\Phi$) during irradiation and the fission cross section of $^{235}$U ($\sigma_f$).

The original age equation relies on the experimental determination of an empirical constant ($B_c$) determined using metal activation monitors to relate the fission track density in the dosimeter ($\rho_d$) to the neutron fluence during irradiation, such that $\Phi = B_c \rho_d$. However, Hurford and Green (1983) showed that unaccounted for errors in the response of metal activation monitors, and the stochastic response of uranium glass standards, results in significant errors in $B_c$ when determined for a small number of irradiations. Therefore, they proposed the zeta (ζ) calibration approach which circumvents the requirement to estimate the fluence directly. Zeta is an empirical constant which is defined such that $\zeta \rho_d = I \sigma_f \Phi / \lambda_f$ (Hurford and Green, 1983, Galbraith, 2005). Thus the zeta factor incorporates the uranium isotope ratio, the
neutron fluence during irradiation, the fission cross section of $^{235}$U, and the spontaneous decay constant for $^{238}$U. Therefore, the fission track age equation becomes:

$$t = \frac{1}{\lambda_d} \ln \left( \frac{\lambda_d}{\rho_i} \rho_d \zeta g + 1 \right)$$  \hspace{1cm} \text{Equation F-2}$$

\( t \) – single grain fission track age
\( \lambda_d \) – \( \alpha \) particle decay constant for $^{238}$U
\( \rho_s \) – spontaneous fission track density
\( \rho_i \) – induced fission track density
\( \rho_d \) – fission track density in dosimeter (glass of known uranium concentration)
\( g \) – geometry factor
\( \zeta \) – calibration factor determined from >30 samples of known age & given dosimeter

Zeta is calculated by rearranging Equation F-2 to make it the subject. Each analyst trains over a period of months to produce consistent fission track density measurements, leading to a personal zeta factor. Therefore, the zeta method has the additional advantage of accounting for the inevitable counting biases inherent to individual analysts. Training involves undertaking 30 to 40 calibrations using apatite samples dated independently using K/Ar, $^{40}$Ar/$^{39}$Ar and Rb/Sr radiometric techniques and irradiated with the same glass standard dosimeter. Age standards are selected to have the following characteristics (Hurford and Green, 1983):

1. well documented and accessible parent geology with abundant age standard mineral;
2. first and single generation crystals of a homogenous age;
3. unambiguous, independent K/Ar and Rb/Sr ages consistent with known stratigraphy; and,
4. the fission track age must relate to the independent age of crystallisation, and not to inherited component or post-formational event.
A commonly used age standard is Durango apatite from the Cerro de Mercado open-pit iron mine, north of Mexico City. This apatite was precipitated in situ between the eruptions of two ignimbrites. McDowell et al. (2005) recently used $^{40}$Ar/$^{39}$Ar dating to give a reference age for the apatite of $31.44\pm0.18$ Ma (2σ). Once the analyst is able to reproduce consistent zeta values for repeated measurements it may then be used as the constant of proportionality in Equation F-2 to derive unknown fission track ages. The analyst specific zeta value used during this study was 338 and the standard error on zeta determined empirically during calibration is 5.

The relative standard error of the single grain age is given by:

$$\frac{se(t)}{t} = \left[ \frac{1}{N_s} + \frac{1}{N_i} + \frac{1}{N_d} + \left( \frac{se(\zeta)}{\zeta} \right)^2 \right]^{1/2}$$

Equation F-3

Where

$se(t)$ – standard error of the single grain age

$N_s$ – number of spontaneous fission tracks for given area of the apatite grain

$N_i$ – number of induced fission tracks for equivalent area of external detector

$N_d$ – number of induced fission tracks in dosimeter used to determine $\rho_d$ and $se(\zeta)$

$\frac{se(\zeta)}{\zeta}$ – relative standard error of zeta ($\zeta$)

In practice, for a single grain age, the final two terms in Equation F-3 are usually small compared to the first two terms and therefore the standard error depends on the number of measured spontaneous and induced fission tracks. The relative standard error on the single grain ages produced for this study varied between about 15% and 30%.

**F.2.2 Multiple grain average age calculation**

Single grain fission track ages were estimated for between 20 and 24 grains for all samples containing suitable apatite examined during this study. The average fission track age for a sample may be calculated using the pooled age, mean age or central age (Gallagher et al., 1998b). The pooled age is based on the ratio of all spontaneous counts to all induced counts for all measured grains and is calculated as follows:
Appendices

Equation F-4

\[ t_{\text{pooled}} = \frac{1}{\lambda_d} \ln \left( \lambda_d \frac{\sum_{j=1}^{n} N_{sj}}{\sum_{j=1}^{n} N_{ij}} \rho_d \xi g + 1 \right) \]

The **mean age** is based on the arithmetic mean of the ratio of spontaneous to induced counts for individual grains, where the number of measured grains is \( n \) (20 in this study) as follows:

Equation F-5

\[ t_{\text{mean}} = \frac{1}{\lambda_d} \ln \left( \lambda_d \frac{\sum_{j=1}^{n} N_{sj}}{n} \rho_d \xi g + 1 \right) \]

Both the pooled age and the mean age equations assume that the spontaneous and induced fission track densities measured for different grains are drawn from the same population, and effectively have the same ‘true’ fission track age. In actual fact, the ‘true’ age may itself vary between grains, due to differences in composition for example (Galbraith, 2005). The random effects model assumes that the measured values of ratios of spontaneous to induced fission track densities \( \rho_s/\rho_i \) for several grains have a log-normal distribution, such that \( \log(\rho_s/\rho_i) \) is a normal distribution with mean \( \mu \) and standard deviation \( \sigma \). The age corresponding to \( \mu \) is known as the **central age** \( (t_{\text{central}}) \). It corresponds to the geometric mean of \( \rho_s/\rho_i \) for individual grains. It is less influenced by outliers and non-Poisson (skewed) distributions sometimes observed in the distribution of multiple grain ages (Gallagher et al., 1998b). The central age is the multiple grain average age used throughout this study.

In practice, the values \( \mu \) and \( \sigma \) estimated from a sample of grains will differ from the true values because: 1) the grains are a sample of the total population of grains; and 2) the measured fission track densities are based on a sample of the total population of fission tracks in a given grain. Since \( \mu \) and \( \sigma \) cannot be calculated explicitly, the central age and age dispersion must be estimated from an iterative algorithm (Galbraith and Laslett, 1993, Galbraith, 2005), which is summarised below.

1. For \( j = 1, 2, \ldots n \) let:

   \[ m_j = N_{sj} + N_{ij} \]

   \[ p_j = N_{sj}/m_j \]
2. Set starting values of $\sigma$ and $\theta$ as follows

$$\sigma = 0.6 \times \text{standard deviation of (}z_1, z_2, \ldots, z_n\text{)}$$

$$\theta = \frac{\sum_{j=1}^{n} N_{sj}}{\sum_{j=1}^{n} m_j}$$

3. For $j = 1, 2, \ldots, n$ compute

$$w_j = \frac{m_j}{\{\theta(1 - \theta) + (m_j - 1)\theta^2(1 - \theta)^2\sigma^2\}}$$

4. Compute new values of $\sigma$ and $\theta$ as follows

$$\sigma = \sigma_{old}\left(\sum_{j=1}^{n} w_j^2 (p_j - \theta)^2 \right)^{\frac{1}{2}}$$

$$\theta = \frac{\sum_{j=1}^{n} w_j p_j}{\sum_{j=1}^{n} w_j}$$

5. Repeat steps 3 and 4 until the change in $\sigma$ and $\theta$ is insignificant. For this study, steps 3 and 4 were repeated 20 times for each multigrain average age calculation.

6. The final value of $\sigma$, denoted by $\hat{\sigma}$, is the estimated age dispersion and the final value of $\theta$, denoted by $\hat{\theta}$, is the estimate of $\rho_s/(\rho_s + \rho_i)$. The central age is then calculated from the age equation.

$$t_{central} = \frac{1}{\lambda_d} Ln \left( \lambda_d \frac{\theta}{1-\theta} \hat{\theta} \hat{\xi} g + 1 \right)$$  \hspace{1cm} \text{Equation F-6}$$

The relative standard error of the central age is given by:

$$\frac{se(t_{central})}{t_{central}} = \left[ \frac{1}{\hat{\theta}^2(1+\hat{\theta})^2 \sum_{j=1}^{n} w_j} + \frac{1}{N_d} \left( \frac{se(\xi)}{\xi} \right)^2 \right]^{\frac{1}{2}}$$  \hspace{1cm} \text{Equation F-7}$$

The process for calculating the central age and the relative standard error of the central age for all samples has been carried out during this study using an Excel spreadsheet.
F.2.3 Comparison of single grain ages, standard error and dispersion

Histograms of single grain ages calculated for an individual rock sample do not represent the error associated with each age estimate. In addition, larger age estimates tend to have larger standard errors which results in a positively skewed histogram. To overcome these limitations, Galbraith (1990) developed the radial plot method of displaying the single grain age data, as shown in Figure F-1. For single grain age estimates $z_1, z_2, ..., z_n$, with standard errors $\sigma_1, \sigma_2, ..., \sigma_n$, the x and y values on the radial plot are calculated as follows:

$$x_i = 1/\sigma_i \quad \text{and} \quad y_i = (z_i - z_0)/\sigma_i$$

Given that the standard error is often observed to increase as the age estimate increases, the logarithm of the single grain ages is often used to represent values of $z$. The central age (or its logarithm) is usually used as the value of $z_0$ so that the values of $y_i$ scatter around zero (Galbraith, 2005). Therefore, a radial plot is a scatter plot with standardised estimates of y plotted against precisions x. The slope of a line connecting coordinates 0, 0 with x, y on the radial plot is given by:

$$\frac{y}{x} = \frac{(z-z_0)/\sigma}{1/\sigma} = z - z_0$$

Therefore, all points on a radial line extending through the origin have the same value of $z-z_0$ and hence the same single grain age, $z$. As shown in Figure F-1, a scale of slopes showing single grain ages ($z$ values) can be displayed on an arc centred on the origin. If the logarithm of the single grain ages were used to represent values of $z$, then a logarithmic scale is used to represent the ages on the radial axis. Points further from the origin on the x scale have a higher precision and corresponding lower relative standard error. Since the y values have unit standard error, the distance on the y axis between ±2 standardised estimates of y can be placed on any data point to give the ±2 standard error values on the single grain age by extrapolating a radial line from the origin, through the ends of the error bar, to the arc scale (Figure F-1).
Figure F-1: Principles of the radial plot with estimates on a log scale (after Galbraith 1990)

The age dispersion ($\sigma$) calculated using the algorithm described in Section 5.2.2 and the radial plots themselves enable an assessment of whether the single grain ages are consistent with a common average age. If the single grain age estimates are distributed homoscedastically (have the same variance) about the central age then about 95% should lie between ±2 standardised estimates of $y$. If the sampled single grain age estimates are widely dispersed then they may be drawn from more than one age population. A chi squared ($\chi^2$) test has been applied during this study to assess if there is a significant difference between the ratios of spontaneous to induced fission tracks determined for single grains taken from the same rock sample. The null hypothesis that there is no significant difference between the spontaneous and induced fission track ratios was rejected if the probability ($p$-value) of the calculated $\chi^2(n-1)$ exceeding the $\chi^2$ statistic was less than 5%. The Java application Radial Plotter (Vermeesch, 2009) was used to produce all radial plots prepared for this study.
F.3 Characteristics of Fission Track Annealing

F.3.1 Kinetic parameters and the influence of apatite composition

Apatite is the name given to a group of phosphate minerals with different relative concentrations of hydroxide (OH$^-$), fluoride (F$^-$), and chloride (Cl$^-$) ions in the crystal. The general formula which incorporates the three end members is Ca$_5$(PO$_4$)$_3$(F,Cl,OH). The rate of fission track annealing in apatite is known to depend on chemical composition of the host apatite, however, the exact composition of individual grains is difficult to determine. Fortunately, the annealing behaviour has been shown to be correlated with so called kinetic parameters, the most frequently measured of which are chlorine content (Cl wt%) and the mean etch pit diameter parallel to the c-axis, known as D$_{par}$.

Electron probe microanalysis may be used to estimate the relative abundance of the halogens directly (Donelick et al., 2005). The results are commonly reported as Cl wt%. It has been shown that apatite grains with high values of Cl wt% (>1 to 2%) anneal relatively slowly (Carlson et al., 1999). However, apatite with low values of Cl wt% (<1 to 2%) usually, but not always anneal quickly. The presence of other elements, including Mn, Fe, as well as combinations of OH and Cl appear to influence the annealing rate. Whilst Cl wt% may be thought to have a direct theoretical relationship with annealing rate, Donelick et al. (2005) argues that the alternative empirical D$_{par}$ parameter is at least as effective, and more easily measured.

D$_{par}$ is the kinetic parameter employed during this study. It is important to note that it is not a proxy for Cl wt% (Donelick et al., 2005). However, as shown in Figure F-2 it is positively correlated with Cl wt% and OH wt%, and negatively correlated with F wt%, albeit with a large scatter and moderate coefficient of determination. In general, apatite with a low D$_{par}$ are slightly more common and generally faster annealing with a high F wt% composition, even if Mn and Fe is present. Conversely, apatite with high D$_{par}$, usually, but not always, anneal more slowly. D$_{par}$ is poorly correlated with annealing rate when D$_{par}$ is high but so is OH wt%.
Ketcham et al. (1999) showed that the reduced length of a fission track in any apatite can be related to the length of a fission track in apatite more resistant to an annealing by the equation:

\[
 r_{lr} = \left( \frac{r_{mro} - r_{mro_0}}{1 - r_{mro_0}} \right)^k 
\]

Equation F-8

Where

- \( r_{lr} \) – reduced length of fission track in apatite less resistant to annealing
- \( r_{lr} \) – reduced length of fission track in apatite less resistant to annealing
- \( r_{mro} \) and \( k \) – empirical parameters fitted to the data

Furthermore, Ketcham et al. (2007b) suggest that \( r_{mro} + k \approx 1.04 \) and \( r_{mro} \) is related to \( D_{par} \) and Cl wt% by the following empirical equations:

\[
 r_{mro} = 0.84 \left[ (4.58 - D_{par})/2.98 \right]^{0.21} \]

Equation F-9a

\[
 r_{mro} = 0.83 \left[ (Cl\text{ wt}\% -0.13)/0.87 \right]^{0.23} \]

Equation F-9b

Figure F-2: Correlation of the kinetic parameter \( D_{par} \) with a) Cl wt%, b) F wt% and c) OH wt% (after Donelick et al, 2005)
In this study, Equations F-8 and F-9a are used with measured track lengths and $D_{par}$ to normalise track lengths measured in different apatite grains relative annealing resistant apatite.

**F.3.2 The influence of crystallographic orientation**

The mean length of fission tracks in apatite varies with the crystallographic orientation (Donelick et al., 2005). Donelick (1991) was able to show using experimental data that this is because the rate of annealing increases as the angle between the fission track and the c-axis increases. These observations were later extended to show that beginning when track length is less than about 65% of the initial track length, fission tracks at high angles to the c-axis begin to undergo accelerated track shortening (Donelick et al., 1999). Since natural fission track samples show similar anisotropy to induced fission tracks, it is concluded that similar processes are responsible. Donelick et al. (1999) derived the empirical model represented in Figure F-3 to enable the length of any horizontal, confined fission track inclined at an angle to the c-axis to be converted to an equivalent fission track length parallel to the c-axis. The equations that constitute the Donelick et al. (1999) c-axis projection model are:

\[
I_a = e_1 l_c + e_0 \quad \text{Equation F-10a}
\]

\[
\phi_{alr} = s_0 \exp(s_1 l_c) \quad \text{Equation F-10b}
\]

\[
I_a = a_1 \phi_{alr} + a_0 \quad \text{Equation F-10c}
\]

Where

- $I_a$ – track length perpendicular to the c-axis
- $l_c$ – track length parallel to the c-axis
- $\phi_{alr}$ – angle of onset of accelerated length reduction
- $I_a$ – intercept of line segment connecting ellipse and crystallographic a-axis
- $e_0, e_1, s_0, s_1, a_1$ and $a_0$ are empirical parameters fitted to the data.

Equation F-10a relates the track lengths parallel and perpendicular to the c-axis for subpopulations for which an ellipse can be fitted. Equation F-10b relates the angle at which accelerated length reduction commences relative to the track length parallel to
the c-axis. Equation F-10c relates the intercept on the crystallographic a-axis to the angle at which accelerated length reduction commences.

![Figure F-3: General form of the apatite fission track c axis projection model (after Donelick et al, 1999)](image)

Although the calibrated model used in this study is slightly different to Figure F-3, it does portray the generic form of the relationship between track length and orientation relative to the c-axis. The outer ellipse in Figure F-3 is a circle such that the initial track length both parallel and perpendicular to the c-axis is shown as 17 μm. It can be seen that when the track length parallel to the c-axis has reduced to 14 μm, the equivalent track perpendicular to the c-axis has reduced to 12 μm. A fission track with an orientation approximately 30° to the c-axis will have an intermediate track length represented by point A on Figure F-3. The conversion is undertaken by following the ellipse until it intersects the c-axis. The accelerated shortening at high angles can be seen to begin when the track length perpendicular to the c-axis is less than about 10 μm. When a fission track parallel to the c-axis is 10.5 μm, the equivalent track perpendicular (90°) to the c-axis is only 1 μm. Point B shows an equivalent track with an orientation approximately 30° to the c-axis and a length greater than 1 μm. The
equivalent length parallel to the c-axis of 10.5 \( \mu m \) is calculated by following the straight line which characterises region of accelerated annealing to the ellipse which intersects the c-axis.

Ketcham (2003) showed that the original model of Donelick et al. (1999) did not take into account analyst bias (including under-reporting) and total annealing (loss) of fission tracks plotted in the of high angle short track length part of the polar coordinate plot. By taking into account the density of fission tracks measured in different areas of the polar coordinate plot during model calibration, they revised the fitted parameters and were able to improve the empirical model for c-axis projection. More recently, Ketcham et al. (2007a) reduced the number of empirically fitted parameters from 6 to 4 by replacing Equation F-10c with:

\[
I_a = I_{alr} \sin^2(\phi_{alr})
\]

Equation F-11

Ketcham et al. (2007a) also demonstrated that c-axis projection is effective in reducing analyst bias. This appears to occur due to the damping effect of taking measured track lengths in the range 0 to 10 \( \mu m \) and projecting them into a smaller range of about 9 to 13 \( \mu m \) (Figure F-3).

### F.3.3 Thermal annealing model

Laboratory experiments have been used to constrain apatite fission track annealing characteristics at temperatures from 95 °C to 425 °C, and for durations from 5 minutes to 500 days (Gallagher et al., 1998a, Ketcham et al., 2007b). Like chemical reaction rates, the relationship is described by an Arrhenius-type model, such that the track length reduction is dependent on the inverse of absolute temperature and the logarithm of time (Tagami and O'Sullivan, 2005). Laslett et al. (1987) first constructed a mathematical description of the annealing process, which they based on data presented by Green et al. (1986). As shown in Figure F-4a, they were able to produce a reasonable match (explaining 96.7% variation in the data) by fitting linear parallel contours of fission track length reduction on the Arrhenius plot. However, as shown in Figure F-4b, a slightly better fit was achieved (explaining 98% of the variation in the data) when the linear contours of equal track length reduction are slightly fanned. Using the results of variable temperature annealing experiments, Duddy, et al (1988)
were able to show that the annealing behaviour of fission tracks is independent of the prior conditions. Therefore, models derived from constant temperature experiments can be adapted to predict the result of temperatures which vary with time. Green et al. (1989) went on to extrapolate the experimentally derived models to geological timescales and corroborate the results with apatite fission track data from several geological situations where the thermal history is known with confidence.

Ketcham et al. (1999) developed a new empirical Arrhenius-type thermal annealing model based on an annealing resistant chloro-hydroxy apatite in the Carlson et al. (1999) data set. They recognised that it was not enough to base annealing models on experimental data alone and emphasised the importance of seeing how well they match observations at geological timescales. They established high and low temperature benchmarks, for which the thermal history of apatite could be reconstructed independently and with some degree of confidence and precision. The high temperature benchmark is used to assess the point at which fission tracks fully anneal. The low temperature benchmark is used to assess the long term annealing characteristics of apatite at near surface conditions.

Figure F-4: Arrhenius plots: a) linear parallel model and b) linear fanning model (after Laslett et al. 1986) and c) linear fanning model and d) curvilinear fanning model extrapolated to geological time scales (after Ketcham et al. 1999)
Ketcham et al. (1999) used a high-temperature benchmark based on an assessment of apatite fission tracks in samples taken from various depths in the Otway Basin, South Australia (Green et al., 1989). Strata in this basin have undergone monotonic heating due to burial and the thermal history is known with reasonable confidence. For samples taken from a deep borehole, fluoride rich apatite was shown to be completely annealed at 92°C whilst chloride rich apatite appeared to retain tracks to temperatures above 124°C. Ketcham et al. (1999) also point out that the original Laslett et al. (1987) model appears to underestimate the annealing rate at low temperatures. To achieve the observed track shortening, thermal history models based on Laslett et al. (1987) therefore tend to predict that the apatite remains at elevated temperatures until late in its history with consequent recent cooling commencing at about 40°C to 60°C. This can lead to the misinterpretation of a modelling artefact as recent exhumation from 1 km depth or more. Ketcham et al. (1999) use a low temperature benchmark based on apatite obtained from deep ocean sediments where the rate of deposition is small and temperatures have not exceeded 15°C to 20°C for over 100 My (Vrolijk et al., 1992).

Figure F-4c shows the Arrhenius plots for the linear fanning model similar to that proposed by Laslett et al. (1987) but extrapolated to geological timescales. It can be seen that although the high temperature benchmark (H) matches the model at about 40 Myr reasonable well, the low temperature model at about 100 Myr lies well within the total stability zone on the plot. As shown in Figure F-4d, Ketcham et al. (1999) found that a curvilinear fanning model was able to produce an equally good match to the experimental data at low durations and high temperatures, but a better match to the high and low temperature benchmark at large durations. The equation which describes the lines of constant track length transformation in the curvilinear fanning model shown in Figure F-4d takes the form:

\[
f = C_0 + C_1 \left[ \frac{\ln(t) - C_2}{\ln(1/T) - C_3} \right]
\]

Equation F-12

Where \(C_0, C_1, C_2\) and \(C_3\) are all empirical parameters derived by fitting the model to the data. In its simplest form a track length transform function was originally selected to produce a linear trend on an Arrhenius plot of \(\ln(t)/\ln(1/T)\). When introducing the fanning model, Laslett et al. (1987) modified the track length transform function to become:
Where

\[ g(r) = \frac{[1 - \frac{r^\beta}{\alpha}]^\alpha - 1}{\alpha} \]  \hspace{1cm} \text{Equation F-13a}

\( g(r) \) – transform function of relative track length reduction \( r \), where \( r = l/l_0 \)

\( \alpha \) and \( \beta \) are empirical parameters derived by fitting the model to the data

Ketcham et al. (2007b) proposed setting \( \beta \) to -1 so that Equation F-13a is simplified to:

\[ g(r) = \left( \frac{1}{r} - 1 \right)^\alpha \]  \hspace{1cm} \text{Equation F-13b}

Replacing the variables with the recommended model parameters, the final form of the Ketcham et al (2007b) curvilinear fanning annealing model used in this study to relate fission track length reduction to duration and temperature becomes:

\[ \left( \frac{1}{r_{c,mod}} - 1 \right)^{0.04672} = 0.39528 + 0.01073 \left[ \frac{\ln(t) + 65.12969}{\ln(1/T) + 7.91715} \right] \]  \hspace{1cm} \text{Equation F-14}

Where \( r_{c,mod} \) is the normalised reduced length of a fission track parallel to the c-axis. In order to use Equation F-14 each fission track must first be projected onto the c-axis, and normalised relative to annealing resistant apatite as described in Sections F.3.1 and F.3.2.

When using AFT to assess if sedimentary rocks have been heated due to burial as is the case in this project, it is particularly important to have confidence in the low temperature annealing rates predicted by empirical annealing models. Due to slow annealing rates at low temperatures they cannot be corroborated using laboratory experiments. In a similar approach to that adopted for the low temperature benchmark by Ketcham et al (1999), Spiegel et al. (2007a) studied fission tracks in apatite grains acquired from deep ocean cores in order to corroborate or falsify the annealing models discussed here. They examined volcanogenic sediments exposed to temperatures between about 10°C and 70°C over about 15 Ma to 120 Ma. The thermal histories were independently reconstructed using vitrinite reflectance measurements and/or numerical modelling. Their results showed that even when the temperature remains below the nominal apatite fission track partial annealing temperature for geological time scales, track shortening of between 4% and 11% had occurred. They showed that
when the temperature remained less than 30°C the annealing model of Laslett et al. (1987), based on laboratory data only, overestimates the mean track length of low chloride apatite by up to 0.6 μm. The model was in reasonable agreement for intermediate to high chloride apatite when temperatures exceeded 30°C. However, the annealing model of Ketcham et al. (1999) which is calibrated against a low temperature benchmark, was with few exceptions, consistent with fission track data for samples that remained below about 30°C. When the maximum temperature experienced by the samples was between 30°C and 70°C, the model of Ketcham et al. (1999) tended to overestimate track annealing for high chloride apatite. In other words, higher temperatures are actually required to produce the predicted track length. Therefore, in some circumstances, the model of Ketcham et al. (1999), could predict temperatures, and subsequently interpreted burial depths, that are too low. Whilst Spiegel et al. (2007a) tested the Ketcham et al. (1999), the conclusions are also applicable to the improved model of Ketcham et al. (2007b) used in this study.

F.4 Thermal history modelling

F.4.1 Forward modelling: temperature history to track length distribution

The thermal history modelling process described here is based on the approach implemented in the software HeFTy v.1.6.7 (March, 2009) developed by Richard Ketcham (Ketcham and Donelick, 2005) and used during this study.

When the temperature history is pre-defined we can use the thermal annealing equation (Equation F-14) to forward model the predicted fission track length distribution. As discussed in Section 2.7, forward modelling is an example of a well-posed direct problem, in which the input parameters ($I$) and system function ($S_f$) are known and the unique output parameters ($O$) can be calculated. This may be described schematically as:

$$ O = I \times S_f $$

The input parameters correspond to the thermal history, the transform function corresponds to the apatite annealing characteristics, and the output data corresponds to the fission track length distribution. The forward modelling process can be divided into
three stages involving separate algorithms (Crowley, 1993, Ketcham et al., 2003, Ketcham, 2005) as follows:

1. **Parsing**: subdivide the time-temperature (t-T) path into isothermal steps;

2. **Annealing**: calculate the amount of annealing for the population of fission tracks formed in each time step, and the annealing each population undergoes in subsequent time steps;

3. **Integration**: using an initial track length, calculate the present day mean and standard deviation of the track lengths for populations formed during each time step and sum them.

The more segments that the pre-defined t-T path is divided into the more accurate the solution will be. However, an excessive number of time-steps will require an unnecessary number of computations. HeFTy uses a t-T parsing algorithm based on the work of Issler (1996) who showed that time-steps should be shorter as total annealing is approached. Ketcham et al (2003) found that the precision is greater than 0.5% if there is no step greater than 3.5°C within 10°C of the total annealing temperature of F-rich apatite. In practice, the t-T path is subdivided into a large number of evenly spaced time-steps with additional time-steps added as necessary to ensure that the precision does not fall below 0.5%.

The evolution of the mean track length with time and temperature is calculated using Equation F-14, and the application of the equivalent time concept. This concept posits that the annealing behaviour of a fission track is dependent on its length and not its prior t-T history (Duddy et al., 1988). A population of fission tracks is assumed to form at the beginning of time-step 1 and the amount of annealing which occurs during time-step 1 is calculated from Equation F-14 (i.e., the curvilinear fanning model of the form shown in Figure F-4d). For time-step 2, the mean temperature is calculated from the pre-defined t-T path. The equivalent time required to achieve the amount of annealing calculated for time-step 1 using the temperature of time-step 2 is calculated from Equation F-14. This equivalent time is then added to the duration of time-step 2 to obtain the total time used to calculate the additional amount of annealing during time-step 2 once again using Equation F-14 (Ketcham, 2005).
When applied forward in time, a series of equivalent time calculations is required beginning with the new fission track population formed in each time-step. However, HeFTy reduces the number of iterative calculations required by performing the annealing algorithm backwards in time (Crowley, 1993). If the amount of annealing is calculated for the final time-step \((n)\) then this length can be used to calculate the equivalent time that can be applied to the previous time-step \((n-1)\). This allows the total annealing (during time-steps \(n\) and \(n-1\)) for the fission track population formed during time-step \(n-1\) to be calculated. The equivalent time for each time-step is applied to the one before so that the entire t-T path only needs to be traversed once in the reverse direction, rather than forward from each time-step (Ketcham, 2005).

The annealing algorithm calculates the relative reduced length \((r = l/l_0)\) and therefore the mean track length is derived by multiplying by an assumed initial track length \((l_0)\). Carlson et al (1999) showed that the initial track length is correlated with the kinetic variables and \(D_{par}\) was used in this study to estimates the initial track length projected onto the c-axis using the following empirical relationship:

\[
l_{0,c,mod} (\mu m) = 16.10 + 0.205 D_{par} (\mu m), \text{ where } r^2 \text{ (coefficient of determination) } = 0.81.
\]

The c-axis projected track length population is assumed to be normally distributed about the mean and the standard deviation is calculated as a function of track length using the empirical relationship defined by Ketcham (2003) as follows:

\[
\sigma_{c,mod} = 0.01 \times l_{c,mod} - 0.2827 \times l_{c,mod} + 2.501
\]

The track length distributions from populations beginning at all time steps must be combined so that the integrated distribution can be compared against measured fission track length distribution. However, before they can be compared, the modelled track length distribution must be modified to take into account the reduction in uranium concentration with time due to radioactive decay and the reduced likelihood of observing some tracks due to their short lengths and high angles (Ketcham, 2005). Ketcham et al (2000) describe a weighting factor to account for the reduction of uranium concentration. They also propose additional empirical factors to account for under-sampling of short tracks on the etched surface of the grain and the under-counting of short tracks at high angles to the c-axis.
An effective fission track age is derived from the modelled fission track length data so that it can be compared against the age derived from the measured fission track density. The modelled density reduction is calculated by assuming that it is proportional to the length reduction. The effective age is then calculated by summing the duration of each time step, and modifying this value by the apparent density reduction due to the track shortening during each time step, relative to an age standard (Willett, 1997, 2005). HeFTy also reports the age of the ‘oldest track’, which corresponds to the earliest fission track population that is not completely annealed. Earlier history is therefore unconstrained by the fission track data (Ketcham, 2005).

F.4.2 Inverse modelling: track length distribution to temperature history

Inverse modelling is an example of an ill-posed indirect problem. The inverse problem is commonly encountered in Earth science. Often we know the input data and the output data and the aim is to estimate the system function which transforms one to the other. The situation in thermal history modelling is slightly different. In this case we know the output data ($O$) which corresponds to the present day track length distribution, and the system function ($S_f$) which corresponds to the thermal annealing model and we wish to estimate the input data ($I$) which corresponds to the thermal history. This may be described schematically as:

$$I = O / S_f$$

The approach adopted by HeFTy is to assume fixed apatite thermal annealing characteristics and forward model numerous realisations of the candidate thermal histories to calculate their corresponding fission track length and age distributions. Statistical tests are then performed to assess the goodness of fit between the modelled and measured fission track length distribution and age distribution.

HeFTy allows the analyst to define various constraints on the candidate t-T paths. These are defined on a graphical interface of temperature against time using the procedure described below (Ketcham and Donelick, 2005).

- Set the present-day surface temperature.
- Set additional constraints by creating boxes through which the \( t-T \) path must pass, where the relative temperatures at the centre of adjacent boxes determines whether the overall temperature trend between constraints is heating or cooling.

- For each line segment between boxes:
  - Set the number of times which the segment is halved to create nodes where the slope of the \( t-T \) path may change;
  - Set the smoothing factor on the change in slope at the nodes; and
  - If required, set the maximum heating or cooling rate between constraints.

- Set the path between constraint boxes to be either:
  - ‘Monotonic consistent’, meaning the paths between nodes must be consistently heating or cooling (depending on the relative temperatures of the box centres); or
  - ‘Monotonic’, meaning that paths between nodes may be heating or cooling, providing that the overall path between constraint boxes is monotonic (depending on the relative temperatures of the box centres).

For example, a constraint box was used during this study to define a period of near-surface temperatures during deposition of the Karoo-age diamictite. Since the \( t-T \) path between constraints must be monotonic, the number of allowed inflexions in the \( t-T \) path is increased by increasing the number of constraint boxes and allowing the \( t-T \) paths between nodes to be either heating or cooling. Since the post-deposition thermal history of apatite in the Karoo-age diamictite is unconstrained by the geology, the approach adopted here was to introduce several constraint boxes with large temperature ranges so as to encourage HeFTy to explore a wide area of the time-temperature space with candidate thermal histories. Within the defined constraints, HeFTy creates nodes at random using a Monte Carlo simulation method to sample the time/temperature space. The goodness of fit between the modelled and observed fission track length and age distribution is conducted for each realisation.

The Kolmogorov-Smirnov (K-S) test may be used to evaluate the goodness of fit between the modelled and observed fission track length distributions (Ketcham, 2005). The form of the K-S test used in HeFTy assumes that the modelled track lengths form a
continuous known distribution, and the observed data are a sample (of size $n$) taken from an unknown population distribution. The K-S test is used to calculate the probability that a random sample (also of size $n$) taken from the modelled distribution would have a greater separation from it than the observed data distribution. For example, a K-S probability of 0.05 means there is a 5% chance that the distribution of a random sample (of size $n$) drawn from the modelled distribution has a greater separation from it than the observed distribution. This corresponds to the 95% confidence limit that there is no significant difference between the modelled and observed distributions. A Goodness of Fit (GOF) value of >0.05 represents an ‘acceptable fit’.

If the observed track length distribution is randomly sampled from the modelled track length distribution, then other randomly sampled distributions taken from the same modelled distribution would have a 50/50 chance of having greater separation from the modelled distribution than the observed distribution. A ‘good fit’ is therefore represented by a GOF of 0.5. A GOF above 0.5 implies that the observed distribution has less separation from the modelled distribution than most distributions of the same size that have been randomly sampled from the modelled distribution.

The K-S test is most sensitive to the median length in the distribution, and so HeFTy also provides the option of using a modified version, known as the Kuiper’s statistic which balances the sensitivity between the median and the tails (Ketcham and Donelick, 2005). An analogous scheme is used to evaluate the goodness of fit between ages calculated from the modelled and observed data (Ketcham, 2005). During this study, HeFTy was set to continue creating realisations of the candidate t-T paths until 100 ‘good’ paths (i.e., merit value > 0.5) were created.

As will be seen, due to the non-uniqueness of the inverse problem and the relative insensitivity of fission track thermochronology, the t-T paths which produce ‘acceptable’ and ‘good’ fits to the observed fission track length distribution fill a large portion of the time-temperature space. Whilst in theory it is possible to produce a best-fit thermal history, given the many uncertainties in the apatite fission track system and modelling process it is perhaps more appropriate to use thermal history modelling simply to rule out those histories which are inconsistent with the observed apatite fission track length distribution.
F.5 Modelling approach and procedures adopted by this study

The goal of the inverse modelling during this study was to define the region of the time-temperature space which is inconsistent with ‘good’ or ‘acceptable’ t-T paths. This approach is in contrast to one which attempts to define the ‘best fit’ model (e.g., Bauer et al., 2010), which would require significant reliance on the AFT data alone and the assumption of parsimony. These assumptions, when applied without geological controls lead to monotonic cooling histories. Green and Duddy (2010) have proposed that the cases where independent geological controls are available demonstrate that episodic thermal and burial/exhumation histories are more realistic than monotonic cooling/denudation histories. The data input parameters and procedures adopted for the thermal history modelling during this study using the HeFTy software (Ketcham and Donelick, 2005) are summarised below.

1. Enter the age data for each grain (20 to 24 grains), including:
   a. Spontaneous and induced track counts;
   b. $D_{\text{par}}$ (μm) for each grain;
   c. Zeta and standard error of zeta;
   d. Dosimeter, track count and track density;
   e. Select uncertainty mode ‘±95%’.

2. Enter the length data for each track (58 to 124 tracks), including:
   a. Length (μm) and angle from c-axis;
   b. $D_{\text{par}}$ (μm) set as mean value for each grain or all grains;
   c. Initial track length calculated from $D_{\text{par}}$ (Carlson et al., 1999);

3. Select the goodness of fit method ‘Kuiper’s Statistic’.

4. Select the Ketcham et al. (2007b) annealing model.

5. Select c-axis projection and use the Ketcham et al. (2007a) c-axis projection model.

6. Select $D_{\text{par}}$ kinetic parameter (all data assumed to belong to same population).

The HeFTy inverse modelling procedures adopted during this study are summarised below.
1. Set the present day (0 Ma) temperature at 20 °C for all models

2. To encourage multiple inflections and filling of the time/temperature space by t-T paths, define six or seven constraint boxes (depending on the oldest fission track age) with:
   a. Time intervals of 500-360 Ma (Cambrian to Carboniferous), 360-300 Ma (Carboniferous), 300 to 250 Ma (Permian), 250 to 200 Ma (Triassic), 200 to 145 Ma (Jurassic), 145 to 65 Ma (Cretaceous), and 65 to 0 Ma (Cenozoic);
   b. The low temperature was set at 0°C for all constraint boxes, except for the Cenozoic constraint box, where it was set to 10°C;
   c. The high temperature limit on the constraint boxes was reduced during iterative model runs so as to increase modelling efficiency without constraining the area occupied by ‘good’ or ‘acceptable’ t-T paths;
   d. The high temperature limit on the three Mesozoic constraint boxes was set to the same value, so that the central temperature was also the identical. This allowed the temperature to rise or fall between boxes.

3. For the Bihanga diamictite samples only (GB13 and GB14) additional surface temperature (0 to 20°C) constraint boxes were defined during the Carboniferous to Permian time period (360 to 260 Ma).

4. Set the properties of the t-T paths between constraint boxes as follows:
   a. Select one subdivision of the t-T path between constraint boxes to create nodes where the slope of the t-T path may change;
   b. Select the ‘intermediate’ smoothing factor on the change in slope at the nodes;
   c. Select the ‘monotonic’ constraint, meaning that paths between nodes may be heating or cooling, providing that the overall path between constraint boxes is monotonic (depending on the relative temperatures of the box centres).

5. Select the ‘Monte Carlo’ method and ‘random’ sub-segment spacing for defining nodes on the t-T paths within the set constraints for each realisation.

6. Set the ending condition to 100 ‘good’ paths (>0.5).
Typically between 40,000 and 180,000 realisations of random t-T paths within the set constraints were required to produce 100 ‘good’ paths.
The results of the thermal history modelling conducted using the HeFTy software (Ketcham and Donelick, 2005) have been compared against preliminary results obtained using the recently developed QTQt software (Gallagher, 2010). QTQt adopts a Bayesian approach, rather than the frequentist approach used by HeFTy for assessing the probability of alternative time-temperature histories. It also uses an inversion scheme based on the Markov Chain Monte Carlo (Gallagher et al., 2009) approach in which the complexity of the model (i.e., the number of time-temperature nodes) is inferred from the data, rather than set in advance. Each realisation of the time-temperature path is based on a random walk (the Markov Chain) through the time-temperature space. Each new node in the chain is conditional on the previous one, but independent of how former ones were arrived at. The user is required to define parameters which control the number of iterations of the sampling chain, the proposed magnitude of the time and temperature moves, and the birth and death rates for new time-temperatures nodes. As for the HeFTy modelling, additional constraints were added for the Permo-Carboniferous and present day surface temperature. The approach adopted by Bayesian statistics is to assume a prior probability, which is then updated in the light of new data. The Bayesian perspective assumes that the probability of a posterior model can only be defined relative the probability of the prior model in the light of new data, and no probability can be defined in absolute terms.

QTQt has not been used to conduct an extensive analysis of the data during this study, but rather to compare the alternative model output for the Bihanga diamictite sample GB14 only. Since both HeFTy and QTQt use the thermal annealing model of Ketcham et al (2007b) this comparison is simply used to test for consistency between the two different inversion schemes and approaches of assessing the thermal history probabilities. Figure G-1 presents the modelling results obtained from a) HeFTy and b) QTQt for sample GB14. The time-temperature plot for the QTQt model shows the colour coded conditional probability density of the thermal history. This is effectively the probability that the thermal history passes through a box of size 1°C × 1 My. Although the probability of passing through any individual box is low, it does provide an indication of the relative probability of the thermal history passing through any particular region of the time-temperature space. The preferred single model, known as
the expected model in Bayesian terminology, is shown as a solid white line. This is effectively a weighted mean, where the weighting is based on the posterior probability of each model. It can be seen from Figure G-1b that the expected model lies within the red region of maximum probability (>1.75%). The dashed white lines represent the 95% probability range on the expected model. These occur towards the edge of the green region of intermediate probability (>0.5%).

The initial QTQt model runs appeared to focus heavily on the parsimonious post-Palaeozoic monotonic cooling history at the expense of possible episodic cooling and heating histories. In the same way as inflection points were added to the HeFTy model to ensure that t-T paths explored the entire time-temperature space, the Markov chain parameters were adjusted to encourage the model to explore the entire time-temperature space. Even though a smaller constraint box was used for the Permo-Carboniferous surface temperature in the QTQt model, it can be seen that there is good agreement between the region of maximum probability calculated by QTQt and the cluster of ‘good’ t-T paths produced by HeFTy. Although, in theory the entire thermal history embodied in ‘good’ and ‘acceptable’ t-T paths produced by HeFTy should be considered together, the comparison with the high probability region of the QTQt plot suggests that regions where ‘good’ t-T paths cluster, also represent regions of the time-temperature space through which the thermal history is more likely to pass. Whilst QTQt appears to indicate that simple post-Permian heating followed by monotonic cooling history is most likely, this is again based on the assumption of parsimony. It should be remembered that the prior assumption of parsimony is an epistemic maxim, rather than an ontological necessity. It is noted that although the probabilities of the thermal history passing through any individual 1°C × 1 My box is low, there is only a factor of four difference between the centre of the red region and the edge of the green region. It is also interesting to note that like HeFTy, QTQt also predicts the increased rate of cooling during the Cenozoic. As discussed previously, it seems likely that this artifact is a function of the estimated initial track length, rather than the inversion scheme, or the data.
Figure G-1: Corroboration of HeFTy thermal history modelling with QTQt software using Markov chain Monte Carlo approach.
H THERMOCRONOLOGICAL STUDIES OF EAST AFRICA

This section reviews the thermochronological studies previously conducted in East Africa in order to place the results of this study into context. Kohn et al. (2005) report that at the time of their study, over 430 AFT ages had been determined across eastern Africa, and about 350 of these included track length data. The sample locations together with the references to the main published AFT research in the region (van den Haute, 1984, Wagner et al., 1992, Foster and Gleadow, 1992, 1993, 1996, Noble et al., 1997, Eby et al., 1995, Van der Beek et al., 1998) are shown in Figure H-1.

![Figure H-1: Locations and references for the main AFT data sets in eastern Africa (after Kohn et al, 2005)](image-url)
Up until 2010, there was no AFT data available for Uganda. The data acquired by this study from the eastern flank of the Western Rift and the recent RiftLink project in the Rwenzori Mountains (Bauer et al., 2010) has therefore filled a gap in our knowledge of the region. Prior to these studies, the only AFT data available from the Western Rift north of Tanzania was obtained by Van den Haute (1984) from samples taken in Rwanda and Burundi. He calculated AFT ages for 27 samples and track length distributions for 18 samples. Given the developments in AFT analysis since 1984, the reported ages should be treated with caution. Generally speaking they suggest a south to north pattern of age reduction from approximately 400 Ma to 300 Ma in southern Burundi, and ages between 300 Ma and 200 Ma in northern Burundi and eastern Rwanda. One sample 30 km east of Lake Kivu and two on the south-east side of Virunga Mountains in western Rwanda have AFT ages between 85 Ma and 75 Ma. Van den Haute (1984) recognises that some limited annealing has taken place, resulting in an apparent AFT age between what he refers to as the ‘pure cooling age’ and a later thermal event. Interestingly, he anticipates some of the findings of this study by speculating on the burial of basement rocks beneath Karoo-age sediments as the source of this later annealing event. However, he rejects this hypothesis on the basis of an unsupported assumption that “only a limited cover of probably a few hundred metres of continental sediments have been deposited in the studied area during this period” and because he expects to see a simple correlation between AFT age and track length reduction if this hypothesis were correct. This study was conducted before the development of the thermal annealing model based on laboratory studies conducted in the late 1980s (Green et al., 1986, Laslett et al., 1987, Duddy et al., 1988, Green et al., 1989) and so Van den Haute (1984) did not have the means to quantify the degree of reheating required to reproduce the observed track shortening. Nevertheless, he was able to conclude that the rock uplift and subsequent erosion associated with Cenozoic rifting was insufficient to bring rocks to the surface with AFT ages of the same age as the movements themselves.

Wagner et al. (1992) conducted AFT analyses on 43 samples from Kenyan basement rocks (Figure H-1). The oldest AFT ages of between 319 Ma and 289 Ma were calculated for samples from south west Kenya, in the Loita Hills between the Kenyan Rift and Lake Victoria. The pattern of AFT ages in central Kenya and between the Kenya Rift and the Anza Rift originally identified by Wagner et al. (1992) was later
examined in greater detail by Foster and Gleadow (1992, 1993, 1996). They showed that the amount of recent denudation on the shoulders of the Kenya Rift was not large enough (< 1 km) to expose rocks that passed into the nominal PAZ (~110°C to 60°C) during the Cenozoic. Therefore, AFT ages document the Mesozoic tectonothermal history of central Kenya. Foster and Gleadow (1992) point out that, all else being equal, the time intervals corresponding to episodes of rapid exhumation will have longer average fission track lengths as they will have experienced less annealing because of the shorter duration they have spent in the PAZ. The central Kenyan AFT data of fission track age against mean track length shown in Figure H-2a includes three periods with apparent peaks in the mean track lengths between about 60 to 70 Ma (late Cretaceous to early Palaeogene), 100 to 120 Ma (mid Cretaceous) and 160 Ma (late Jurassic). The greater annealing of some mean track lengths between 100 to 120 Ma suggests that the assumption of length isotropy in the age equation may have resulted in an underestimate of the start of mid Cretaceous exhumation. It is noted that the older two apparent peaks are based on relatively few samples with long mean track lengths.

Figure H-2b shows a similar plot of mean track length against central age for the Western Rift eastern flank AFT data produced during this study and that recently acquired by Bauer et al. (2010) for the Rwenzori Mountains. It is immediately apparent that the AFT ages for the Rwenzori are younger (Jurassic to Cretaceous) than those for the rift flank (Carboniferous to Triassic), which is consistent with the Rwenzori horst having experienced greater exhumation relative to the rift flank. Unlike the AFT data for central Kenya, there are no obvious peaks in the mean track length data and the track lengths measured in the area of the Western Rift in Uganda tend to be shorter (approximately 11 to 12.5 μm) than those measured near the Kenya Rift (approximately 12 to 14.5 μm). The overall lower track length could be attributable to different apatite composition (and hence different initial track length and/or annealing behaviour), or, perhaps more likely, greater annealing following the start of spontaneous fission track preservation (due to slower exhumation and/or reburial). The lack of peak fission track lengths possibly suggest consistent rather than episodic exhumation or possibly that the signal is simply lost in the noise due to a complex annealing history and/or local tectonics and apatite compositional differences.
Figure H-2: Mean track length versus fission track age for a) central Kenya apatite samples (after Foster and Gleadow, 1992) and b) the eastern flank of the Western Rift, and the Rwenzori horst (Bauer et al, 2010)

Foster and Gleadow (1992, 1993) conducted thermal history modelling on the data shown in Figure H-2a and concluded, as expected, that the start of spontaneous fission track preservation (corresponding to passage into the PAZ) occurred a little early than the age calculation suggests. The revised age estimates of the three main exhumation episodes are approximately and \( \geq 220 \text{ Ma} \) (late Triassic), 145 to 120 Ma (early Cretaceous) and 75 to 65 Ma (late Cretaceous). The ages of these exhumation events together with those proposed by other researchers in the region are summarised together with depositional and tectonic events in Table H-1. Foster and Gleadow (1993) suggest the exhumation episodes for central Kenya correlate with:
1. Permo-Triassic development of Karoo Basins, eventually culminating during Jurassic in the triple junction separating Africa and Madagascar, with the Anza Rift as the failed arm;

2. Early Cretaceous far-field accommodation and rift development in west and central Africa, through to the Anza Rift, due to opening of the South Atlantic (Fairhead and Binks, 1991);

3. Late Cretaceous change in the plate motions between South America and Africa resulting in further far-field accommodation and rift reactivation.

In general terms, the denudational episodes appear to be driven by plate tectonics resulting in rock uplift due to faulting. Whilst these explanations are less than certain, Foster and Gleadow (1993) maintain that they are more plausible than explanations based on mantle plumes, climate change or tectonic denudation.

Historical researchers have attempted to assign erosion surfaces to the upland areas of central Kenya studied by Foster and Gleadow (1993), in the same way as they did for concordant landscape features in Uganda (see Appendix C). King (1962) assigned peaks of the Mathews Range to his Jurassic African surface, and the Karisia Hills to his Tertiary surface. Saggerson and Baker (1965) suggested that the Karisia Hills preserved the late Cretaceous surface whilst the higher Mathews Range must be older. Bishop and Trendall (1967) and King (1962) thought the tops of the Cherangani Hills correlated with Cretaceous, early Tertiary, or mid Tertiary erosion surfaces. However, Foster and Gleadow (1993) showed that the rocks exposed on these surfaces were actually at depths of greater than 2 km until after the beginning of the Cenozoic. During the 1980s and 1990s, AFT analysis conclusively overturned the paradigm adopted by the earlier qualitative geomorphologists, and demonstrated the danger of correlating and assigning ages to concordant landscape features without stratigraphic and/or quantitative control.
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<th>Age (Ma)</th>
<th>Period</th>
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Table H-1: Comparison of cooling/exhumation, depositional and tectonic events

Spiegel et al (2007b) returned to similar localities (Figure H-1) to Foster and Gleadow (1992, 1996) to look at apatite (U-Th)/He thermochronology data. (U-Th)/He dating is
based on the accumulation of $^4$He during $\alpha$ disintegration of $^{238}$U, $^{235}$U and $^{232}$Th. In general terms, heating results in the diffusive loss of $^4$He. The nominal $^4$He Partial Retention Zone (PRZ) occurs between about 85°C and 40°C and is therefore sensitive to lower temperatures and smaller exhumation depths than the nominal PAZ (110°C and 60°C) for AFT analysis. Spiegel et al (2007b) state that thermal history modelling matched to the (U-Th)/He data and consistent with the AFT data from the rift flank samples suggests rapid late Cretaceous cooling from about 59°C to 57°C for the eastern flank of the Kenya Rift and about 62°C to 59°C for the western flank. Best fits are obtained when the onset of Neogene cooling occurs between about 7 and 4 Ma for the eastern flank and about 5 and 4 Ma for the western flank.

Noble et al (1997) conducted AFT analysis on samples collected from a 280 km transect orientated ESE to WNW in eastern Tanzania, from the Karoo-age Selous Basin, across the Mozambique Belt, to the Tanzanian craton (Figure H-1). On the basis of periods of increased mean track length and thermal history modelling, they conclude that the data reflect long periods of slow cooling punctuated by at least three relatively rapid cooling events in the early Cretaceous, late Cretaceous to Early Palaeogene, and the late Eocene to early Oligocene (Table H-1). Whilst the region was undergoing extension during the break up of Gondwana, they contend that high-angle block faults moved in response to intra-plate stress. Whilst denudation was occurring on the blocks, deposition was occurring in the Karoo grabens. They showed that the Tanzanian Craton has itself been subject to significant exhumation due to uplift and denudation during the Phanerozoic.

Van der Beek et al (1998) conducted AFT analyses for samples of basement rocks taken from the flanks of the Rukwa and Malawi Rifts in south west Tanzania and Malawi (Figure H-1). They describe three episodes of extension and basin filling in the region, including Karoo rifting (Permo-Triassic), a late Mesozoic to early Cenozoic event and the late Cenozoic rifting associated with the EARS. The age of the middle event is controversial with estimates dating from late Jurassic to Miocene. AFT ages for the region vary between 30 ±15 Ma (central age ±1SE) and 296±10 Ma. The relatively short mean track lengths of between 11.4 and 13.2 µm indicate a protracted cooling history.
The AFT ages calculated for samples taken from the Livingstone Mountains increase from about 130 Ma at about 500 m asl, to about 220 Ma at 1900 m asl, and then remain concordant to about 2200 m asl. There is also an apparent increase in the mean track length with elevation, between about 11.5 μm at 500 m asl to about 12.5 μm at 2000 m asl, perhaps suggesting quicker cooling of the higher concordant 220 Ma old samples. Thermal history modelling was consistent with cooling through the low temperature boundary of the PAZ at about 250 to 200 Ma (Triassic). This is clearly inconsistent with the proposition that the summit plateau of the Livingston Mountains represents the Jurassic-age Gondwana erosion surface as suggested by King (1962).

Samples taken from the western flank of the Malawi Rift have a wide age range from about 100 Ma to about 215 Ma between 500 m asl and 2000 m asl with little relationship to elevation. Samples immediately west of the rift record a late Jurassic to early Cretaceous cooling event at around 150 Ma. The AFT ages for the flanks of the Rukwa Rift cluster between 270 and 200 Ma at around 1000 m asl with a wide spread from about 300 to 150 Ma between about 1500 m asl and 2200 m asl. Van der Beek et al (1998) explain the spatial variation in ages as occurring due to block faulting and local denudation. They conclude that the flanks of the Malawi and Rukwa Rifts have experienced periods of accelerated regional cooling, and by inference exhumation, between about 250 and 200 Ma (Triassic to early Jurassic), and at about 150 Ma (late Jurassic to early Cretaceous) and at or less than about 50 Ma (Cenozoic). They point out that these are broadly consistent with the exhumation events identified by Foster and Gleadow (1993) at around 220 Ma, 140 to 120 Ma, and 60 to 70 Ma, suggesting the drivers were of regional scale significance. As shown in Table H-1 it is proposed that that the Triassic to early Jurassic event is associated with the separation of Madagascar and Africa (Wopfner, 1993), and the late Jurassic to early Cretaceous event is associated with opening of the South Atlantic. Although it has been suggested that the earliest Cenozoic event is associated with major plate reorganisation in the Indian Ocean, it also seems possible that the occurrence of apparent accelerated cooling during the Cenozoic is an artefact of the Laslett et al (1987) model used by Van der Beek et al (1998).

MacPhee (2006) did not conduct AFT analyses, but he did undertake (U-Th)/He and U-Pb analyses of apatite, titanite and zircon samples from the Rwenzori Mountains. As
expected, the U-Pb analyses gave Palaeoproterozoic ages (~1900 Ma) related to the Rwenzori Fold Belt (Kibalian) tectonothermal event itself. (U-Th)/He zircon and apatite ages ranged from >400 Ma to >70 Ma with no relationship to elevation. This indicates that insufficient denudation has occurred to expose low temperature thermochronological ages related Cenozoic tectonic events. Given the likely depth of the (U-Th)/He apatite PRZ and the altitude of the Rwenzori Mountains, MacPhee (2006) concludes that no more than 1.7 km of denudation could have accompanied 5 km of Cenozoic surface uplift, suggesting rock uplift of no more than 6.7 km. Given that fossil molluscs suggest the existence of extensive Lake Obweruka at 2.5 Ma (Pickford et al., 1993, Damme and Pickford, 1995) prior to uplift of the Rwenzori horst, the MacPhee (2006) suggests that the minimum required average uplift rate is at least 1.6 km/My.

The detailed distribution of AFT ages acquired by Bauer et al. (2010) from the Rwenzori Mountains (Figure H-1) is discussed in detail in Section 5.7.3 and 5.7.3, where it is compared against the AFT ages from this study. Bauer et al. (2010) obtained AFT central ages from 15 samples taken from the central and northern Rwenzori Mountains (Figure 5-1). The central ages (±1SE) vary between 85.3 ±5.3 Ma (late Cretaceous) and 195.0 ±5.3 Ma (early Jurassic), with the older sample also being the most northerly. They also obtained apatite (U-Th)/He ages for between 2 and 6 grains from 13 samples. The (U-Th)/He single grain ages ranged from 210.0 ±6.0 Ma to 24.9 ±0.5 Ma, with single grain ages within one sample ranging from 108.9 ±2.1 Ma to 52.9 ±1.1 Ma. The large age range is thought to contain un-quantified errors due to processes influencing the diffusion behaviour that has not been accounted for. No correlation between either AFT age or (U-Th)/He age, was observed with elevation.

Bauer et al. (2010) also conducted thermal history modelling on 11 samples using the HeFTy software. Whilst the inclusion of (U-Th)/He data in the modelling provides additional constraints on the t-T path within the nominal PRZ between 85°C and 40°C, their attempt to define the ‘best-fit’ t-T path is less conservative than the approach adopted by this study. They began by manually defining proposed individual inverse solutions of the t-T history and forward modelling the corresponding age and length distributions. They then entered constraint boxes around their manually defined ‘best-fit’ solution and ran 20,000 random realisation to try and refine the ‘best fit’ model and
obtain the pattern of ‘good’ and ‘acceptable’ t-T paths within the predefined constraints. Based on the results of the first model run, the constraint boxes were reduced in size and further random realisations were carried out. This approach forces cooling paths along trajectories imposed by the constraint boxes and leads to an overemphasis on monotonic cooling at the expense of episodic heating and cooling. It places significant reliance on the parsimonious solution whilst failing to acknowledge that the relative probability of this solution compared to episodic t-T paths may in fact be relatively small. Unsophisticated interpretation of the apparently well-defined ‘best fit’ t-T paths can lead to an unwarranted level of confidence in the derived thermal history. Whilst the probability of t-T paths entering the low-temperature region of the time-temperature domain may be under-represented, the age at which the t-T paths enter the high temperature boundary of the PAZ is likely consistent with more conservative approaches.

Bauer et al. (2010) conclude that the modelled t-T paths reflect protracted cooling histories with accelerated cooling during the Permo-Triassic and the Jurassic. Given the optimistic AFT modelling approach and inadequate exploration of alternative solutions and explanations, it is unclear to what extent the proposed accelerated cooling below PAZ temperatures during the Neogene can be relied upon. Nevertheless, given the lack of relationship between elevation and age, and some Oligocene to Miocene apatite (U-Th)/He ages the data is consistent with late Neogene rock uplift. Denudation could not keep up with rock uplift, leading to surface uplift of the Rwenzori Mountains.
I ELECTRICAL RESISTIVITY THEORY AND METHODS

I.1 Basic Theory

I.1.1 Overview and definitions

The basic method adopted by electrical conductivity surveys is to pass a current through the ground and measure the drop in electrical potential between a second pair of electrodes placed in line between the first pair (Griffiths and King, 1981). In uniform ground conditions the ground resistivity can be calculated directly, but in non-uniform ground only the ‘apparent resistivity’ can be calculated. With the exception of clays and metallic ores, the conduction of electricity in rocks occurs through the groundwater. In general, the higher the saturated porosity is the lower the resistivity is. An increase in salinity also results in a reduction in resistivity. The variability in porosity and mineral content means that resistivity varies over orders of magnitude and there is no precise correlation with lithology. Nevertheless, there is a general trend and resistivity tends to increase from clay, to sand and gravel, to limestone, to crystalline rocks as shown in Figure I-1.

Resistance is defined from Ohm’s law as:

\[ R = \frac{V}{I} \]  

where,

- \( R \) – resistance (ohm, \( \Omega \))
- \( I \) – current (ampere, amp)
- \( V \) – potential (volts, V)

Resistivity is defined as the resistance per unit length for a given cross-sectional area:

\[ \rho = \frac{RA}{l} \]  

where,

- \( \rho \) – resistivity (ohm metre, \( \Omega \)m)
- \( A \) – cross-sectional area perpendicular to current (m\(^2\))
- \( l \) – distance over which potential drop (V) used to calculate \( R \) is measured (m)
When a point source of current is applied to uniform ground, the flow lines will diverge in a radial pattern and the equipotential surfaces will form hemispheres which intersect the ground in circles around the electrode. This assumes that the current flows from the positive ‘source’ to a negative ‘sink’ which is so far distant that its effect can be neglected. The electrical potential at a radius \( r \) from the electrode is given by:

\[ V = \frac{l \rho}{2\pi r} \tag{Equation I-3} \]

If the potential difference \( \delta V \) is calculated by measuring the electrical potential at two points separated by a known radial distance \( \delta r \) the resistivity is given by:

\[ \rho = \frac{2\pi r^2}{l} \cdot \frac{\delta V}{\delta r} \tag{Equation I-4} \]

**I.1.2 Four electrode resistivity calculations**

It is not practical to have a negative ‘sink’ electrode placed a large distance from the positive ‘source’ electrode as assumed for Equations I-3 and I-4. A common
The arrangement is to have four collinear electrodes with the two current electrodes ($C_1$ and $C_2$) most distant and the potential measurement electrodes ($P_1$ and $P_2$) in between (Griffiths and King, 1981) as shown in Figure I-2a. The potential difference between $P_1$ and $P_2$ is given by:

$$V_1 - V_2 = \Delta V = \frac{\rho_a}{2\pi} \left( \frac{1}{r_1} - \frac{1}{r_2} - \frac{1}{R_1} + \frac{1}{R_2} \right)$$  

Equation I-5

where,

$V_1$ – measured potential at $P_1$ (V)

$V_2$ – measured potential at $P_2$ (V)

$\Delta V$ – potential difference between $P_1$ and $P_2$ (V)

$\rho_a$ – apparent resistivity (assuming in practice the ground is non-uniform)

$r_1$ – distance from $C_1$ and $P_1$ (m)

$r_2$ – distance from $P_1$ and $C_2$ (m)

$R_1$ – distance from $C_1$ and $P_2$ (m)

$R_2$ – distance from $P_2$ and $C_2$ (m)

Equation I-5 may be rearranged to calculate the apparent resistivity from the known current and measured potential difference as follows:

$$\rho_a = \frac{2\pi \Delta V}{l} \left( \frac{1}{r_1} - \frac{1}{r_2} - \frac{1}{R_1} + \frac{1}{R_2} \right)$$  

Equation I-6

When the distance between each pair of the four electrodes is equal such that $r_1=R_2=a$, and $r_2=R_1=a$, the configuration is known as a Wenner array (Figure I-2b) and Equation I-6 is simplified to:

$$\rho_a = -\frac{2\pi a \Delta V}{l}$$  

Equation I-7
Figure I-2: a) Generic four electrode array, b) Schlumberger array, c) Wenner array

Figure I-3: Vertical section showing distribution of lines of equipotential and current flow between two electrodes at the ground surface
In the Schlumberger configuration (Figure I-2c) the distance from \( C_1 \) to \( P_1 \) and \( P_2 \) to \( C_2 \) is equal and the distance from \( C_1 \) to \( C_2 \) is usually about ten times the distance from \( P_1 \) to \( P_2 \). This is the configuration of the electrode array used during both the 1D and 2D resistivity surveys carried out for this study.

In practice, the four electrode resistivity survey instruments normally give a resistance value and in the field this is converted into resistivity graphically using a pre-prepared nomogram. The apparent resistivity may be calculated using the ‘geometric factor’ \( k \) which is simply a combination of all the terms other than resistance \( (AV/I) \) on the right hand side of Equation I-6 as follows (Loke, 2009):

\[
\rho_a = kR
\]

Equation I-8

Where,

\[
k = \left( \frac{2\pi}{\frac{1}{r_1} + \frac{1}{r_2} - \frac{1}{r_1} - \frac{1}{r_2}} \right)
\]

Equation I-9

The geometric factor is commonly calculated using two terms related to the electrode configuration, known as the dipole length \( a \) and the dipole separation factor \( n \). For a Schlumberger array, the geometric factor is calculated as follows:

\[
k = \pi n(n + 1)a
\]

Equation I-10

where,

\[
n = \left( \frac{(L/2) - (a/2)}{a} \right)
\]

Equation I-11

where,

\( L \) – distance (m) between the current electrodes

\( a \) – distance between the potential electrodes

In other words, as shown in Figure I-2c the dipole separation factor \( n \) is the ratio of the distance between \( C_1 \) and \( P_1 \) or \( P_2 \) and \( C_2 \) \( ((L/2)-(a/2)) \) to the dipole length between \( P_1 \) and \( P_2 \) \( a \).
I.1.3 Depth of investigation

Figure I-3 shows the distribution of the lines of electrical equipotential and the flow lines for the electrical current between a pair of current electrodes. In uniform ground the potential gradient will decrease in a non-linear pattern with distance from the current electrodes. The potential gradient will also be higher between the electrodes compared to outside the electrodes.

Increasing the distance between the electrodes will generally result in the flow of current achieving greater penetration and a consequent increase in the depth of investigation. However, it should be kept in mind that even with a single current source, a thin high resistivity layer can be detected at depth by measuring the potential difference with closely spaced electrodes at successively increasing radial distance from the source (Griffiths and King, 1981). The relationship between current electrode spacing and the representative depth of investigation is therefore not straight forward.

![Figure I-4: Depth sensitivity function versus ratio of depth to electrode spacing (from Loke, 2009)](image)

Loke (2009) recommends calculating the pseudo-depth of investigation for the purposes of producing pseudo-sections using the depth sensitivity function. The depth
sensitivity function is calculated using a mathematical application known as the Fréchet derivative and for the 1D situation can be simplified to the form shown in Equation I-12:

\[ F_{1D}(z) = \frac{2}{\pi} \frac{z}{(a^2 + 4z^2)^{1.5}} \]  

Equation I-12

where

\[ F_{1D}(z) \quad - \quad 1D \text{ depth sensitivity function} \]
\[ z \quad - \quad \text{depth (m)} \]

Figure I-4 shows the 1D depth sensitivity function (Equation I-12) plotted against the ratio of depth to current electrode half spacing \((L/2)\) for the Wenner array configuration. It can be seen that the maximum sensitivity occurs when the depth is about one third of \(L/2\). Some authors have used the maximum sensitivity to represent the depth of investigation. However, Edwards (1977) suggests that the median depth of investigation \((z_e)\) is a more robust estimate. This is the depth which splits the area under the depth sensitivity curve shown in Figure I-4 in half, and is generally a little over one third of the current electrode spacing. The ground above this depth has the same influence on the measured potential as the ground below this depth. A similar calculation can be performed for the 2D and 3D situations.

For the Schlumberger array used during the 1D investigation, the median depth of investigation \((z_e)\) will depend on the ratio of the inter-electrode distances characterised by the dipole separation factor \(\langle n \rangle\) in Equation I-11) (Loke, 2009). The \(n\) values for the 1D investigation conducted for this study varied between 1.0 and 11.5 and the corresponding ratio of the median depth of investigation to the current electrode spacing \((z_e/L)\) varies between 0.173 and 0.192 (Loke, 2009). In other words, for current electrode spacing between 3 m and 240 m the median depth of investigation varies between approximately 0.52 m and 46.1 m (slightly greater than a third of \(L/2\)).

For the Schlumberger array used during the 2D multi-electrode investigation the \(n\) values varied between 1 and 16 and the corresponding ratio of the median depth of investigation to the current electrode spacing \((z_e/L)\) varies between 0.179 and 0.192 (Loke, 2009). For current electrode spacing between 15 m and 355 m the median depth of investigation varies between approximately 2.7 and 68.2 m.
The calculation of the median depth of investigation is strictly only valid for uniform ground conditions. Whilst the depth of investigation calculated in this way should be treated with caution, it does provide a convenient means of displaying and assessing the data. Large resistivity contrasts near the surface can result in particularly high errors in the estimated median depth of investigation.

The median depth of investigation (Edwards, 1977) is used here to ensure consistency between manual plotting of the 1D survey methods and the automated 2D pseudo-sections produced by the RES2DINV software (Loke, 2009). However, it is acknowledged that given the uncertainties and variability in the depth of investigation another common approach of simply plotting the pseudo-depth as one third of the current electrode half spacing (0.33 x \( L/2 \)) is not inappropriate (Griffiths and Barker, 1993).

I.2 1D Electrical Resistivity Survey Methods

I.2.1 Resistivity profiles

The one-dimensional survey methods described here were conducted using an Abem Terrameter SAS 1000 with four electrodes configured in a Schlumberger array. Steel electrodes approximately 30 cm long were hammered into the ground and connected to the Terrameter via protected copper cables using crocodile clips.

When conducting a resistivity profile, the distance between the current electrodes (\( L \)) and the distance between the potential electrodes (\( a \)) is kept constant. The appropriate current electrode spacing is chosen to obtain the desired median depth of investigation. Following each resistance measurement, all four electrodes are moved along a linear survey path until the central point of the array is located at the centre of the next measurement location. To provide reasonable coverage the central point should not be moved more than the \( L/2 \) distance.

Resistivity profiling is a quick and common method of identifying potential near surface electrical anomalies in the ground. For example, Batte et al (2008) have used parallel resistivity profiles to assess the location and orientation of potentially water-bearing low resistivity linear geological structures such as fault zones. Resistivity
profiling is often conducted first in order to assist in locating features to be examined further using vertical electrical sounding.

### I.2.2 Vertical electrical soundings (VES)

When applying the method of vertical electrical soundings (VES), the centre point of the electrode array remains fixed, whilst the spacing between the electrodes is increased in order to investigate the vertical variation in resistivity. For the Schlumberger array adopted during this study, the potential electrodes were placed either 1m or 10m apart and connected to the Abem Terrameter with copper cored cables. The current electrode cables were marked with the eleven $L/2$ distances between 1.5 and 120 m.

The measured resistance was converted to apparent resistivity in the field using a pre-prepared industry standard nomogram which presents the values on a log-log scale. It is common practice during water well siting in Uganda to interpret this hand prepared log-log graph of apparent resistivity versus current electrode half spacing ($L/2$) without further data processing. When developing groundwater for rural water supply, the VES curves may be compared directly against pre-prepared shapes such as those presented by Macdonald et al (2005). In the context of crystalline rock, boreholes are often sited where the VES profile displays a decrease in resistivity to 100 $\Omega$m or less, at a depth of a few metres to tens of metres, before increasing again at greater depths. This is the pattern expected for a weathered rock profile with unsaturated, decomposed, clay-dominated regolith near the surface, underlain by saturated, disintegrated, sandy, permeable regolith, resting on often low-permeability fractured crystalline bedrock (Chilton and Foster, 1995). Such an interpretation may be practical when limited resources area available, but it does not allow the analyst to extract the maximum information from the data.

During this study, the inversion and forward modelling code RES1D ver.1 (Loke, 2001) was used to interpret the VES data. The program attempts to match the smooth apparent resistivity versus electrode spacing ($L/2$) curve using discrete horizontal layers with differing electrical resistivities. For the 1D model the resistivity is assumed to vary with depth only and not horizontally.

The 1D modelling steps can be summarised as follows:
1. Read in data file, including the array type, apparent resistivity data, electrode spacing and instructions for automated or user defined number of layers;

2. Select settings, including damping factors, number of iterations and convergence limits; and,

3. Conduct inversion using the method of least squares and an iterative subroutine to modify the layer resistivity and geometry to reduce the difference between the calculated and measured apparent resistivity.

Following the production of each discrete layered inverse model of resistivity, a forward modelling routine is used to calculate the smooth apparent resistivity curve predicted by this inverse model. The calculated and measured apparent resistivity curves are compared and the inverse model is recalculated in an attempt reduce the difference following the next iteration. The iterative loop is stopped when the predefined convergence criterion is met.

When the electrical resistivity of the ground varies in a continuum it has been argued that it is better to allow the model to automatically adopt a large number of relatively thin layers. This method will inevitably lead to a good fit between the measured and calculated apparent resistivity. However, the resultant inverse model is often more complex than can be justified by the quality and character of the data. Without additional geological evidence to support such a complex model the solution cannot be considered parsimonious. Therefore, the approach adopted by this study was to select the smallest number of layers that produced a good fit for all datasets. It was found that all datasets could be matched with a three layer model.

I.3 2D Electrical Resistivity Survey Methods

I.3.1 Equipment and field methods

The two-dimensional electrical resistivity tomography (ERT) (Griffiths and Barker, 1993) was conducted with an IRIS Instruments Syscal R1 Plus Switch-72 automatic resistance meter. This is a multi-electrode resistivity imaging system. It automatically switches the output current and input voltage measurements between the 72 electrodes to perform predefined sets of resistance measurements. Consecutive surveys can be joined together using the roll-along capability to create surveys up to kilometres in
length. The output voltage of the current electrodes is automatically adjusted to optimise the measured voltage quality. The equipment includes four 90 m long heavy duty multi-core seismic cables with 18 electrode connections located 5 m apart on each cable.

![Diagram of electrode arrangements and measurement sequence]

**Figure I-5: Arrangements of electrodes for 2D survey and measurement sequence to create pseudo-section (from Loke, 2009)**

The general steps in the field methods are as follows:

1. Lay out the four cables, hammer 72 electrodes into the ground, connect cables with the Syscal R1 Plus automatic resistance meter and external battery in the centre location;
2. Use GPS to measure coordinates of each electrode location;
3. Perform check to ensure resistance between each pair of adjacent electrodes is within acceptable limits, and if necessary, move, deepen or wet electrodes to improve contact;
4. Upload switching sequence prepared with Electre II software to Syscal R1 Plus;
5. Run automated switching sequence and collect resistance measurements;
6. For roll-along procedure, move one 90 m cable and 18 electrodes from the trailing end of the survey to the advancing end, and check ground connections;
7. Perform automated roll-along data collection switching sequence; and
8. Repeat steps 5 and 6 as required.

Long surveys may be run over several days, with the data downloaded to a field computer each evening.

I.3.2 Data processing

The Prosys II software is used to process the data and create an input file for the analysis software, RES2DINV. The data processing steps are as follows:

1. Import data files to Prosys II software and stitch the data to create a single file;
2. Whilst erring on the side of few deletions, filter anomalously high resistance data points;
3. Perform apparent resistivity calculations based on resistance and electrode spacing;
4. Calculate electrode elevations using GPS coordinates and SRTM data set and add topographic information to the Prosys II file; and
5. Export data to RES2DINV file input format ready for analysis.

I.3.3 Analysis methods

The RES2DINV software, produced by Geotomo Software (Loke, 2004) was used to conduct the analysis. A bit lock is required to implement the full functionality, and this was kindly loaned by Allied Associates Geophysical Ltd. The analysis steps are summarised as follows:

1. Import data file with apparent resistivity, electrode spacing and array type to RES2DINV;
2. Visually examine data profiles for anomalously high values and, with caution, manually remove any remaining bad data points;
3. Select the inversion methods, discretisation options, damping factors, mesh parameters, inversion progress settings and topography processing methods;
4. Run the inversion routine to calculate the resistivity model and the forward modelling routine to calculate the associated apparent resistivity;
5. Assess the analysis quality by visual examination of the inversion model and consideration of the root mean squared error (RMS error %) of the difference between the measured and calculated apparent resistivity;

6. Attempt to reduce the RMS error and improve conformance with known geological features by adjusting modelling routines, mesh characteristics, damping factors or convergence factors, and repeating steps 3 to 5;

7. Once the model results are within a satisfactory range, perform a sensitivity analysis by adjusting inversion methods, discretisation options, damping factors, mesh parameters, and inversion progress settings by repeating steps 3 to 5;

8. Examine the final model with implementation of the topographic model and adjust topographic damping if necessary; and

9. Adjust model display characteristics, including contour intervals and scales to ensure consistency and facilitate comparison between different surveys.

The RES2DINV program allows many of the modelling options to be adjusted, including the inversion methods, discretisation options, damping factors, mesh parameters, inversion progress settings and topography processing methods. However, the program is designed to be run in an automatic and robust manner with a set of default parameters which in most cases give reasonable results (Loke, 2009). The most significant modelling parameters which may be altered are listed below.

1. Inversion methods
   a. Initial model (yes or no, homogeneous or approximate inverse)
   b. Standard or robust data inversion
   c. Smooth of blocky inversion method
   d. Jacobian matrix (quasi-Newton, Jacobian matrix - standard or fast calculation)
   e. Inversion calculation (standard or incomplete Gauss-Newton)

2. Inverse model discretisation options
   a. Layer parameters (first layer thickness, increase with depth, model depth)
   b. Block parameters (extended model, block widths, ratio of blocks to data)
3. **Inverse model damping factors**
   
   a. Vertical flatness ratio (emphasise horizontal boundaries or vertical layers)
   
   b. Damping factor (automatic or manual, initial, minimum, first layer, depth increase)
   
   c. Resistivity limits (yes or no, upper and lower bounds)

4. **Inversion progress settings**
   
   a. Convergence criteria (RMS %, change in RMS %)
   
   b. Maximum iterations (5 to 10)

5. **Forward modelling mesh**
   
   a. Mesh parameters (finite-element or finite-difference, mesh resolution)

6. **Topography processing methods**
   
   a. Model type (uniform, damped)

The inversion procedure attempts to find a model of the resistivity that gives an apparent resistivity response that is similar to the measured values. In RES2DINV, a cell-based initial model is modified in an iterative manner so that the difference between the model response and the measured data is reduced (Loke, 2009). The difference is characterised using the sum of the squares of the discrepancy vector between the model and measured apparent resistivity. The Gauss-Newton method (Lines and Treitel, 1984) is used to minimise the sum of the squares error by altering the model parameters. The change in the model response due to the change in model parameter is defined for each element of the ‘Jacobian’ matrix. Variations on the Gauss-Newton method are used to produce smooth (default) or blocky models. Further modifications may be implemented to vary the damping factors with depth or give preference to horizontal or vertical structures.

The model response may be calculated using either finite-difference of finite-element methods for comparison against the measured apparent resistivity, and calculation of the root mean squared error.
J 1D ELECTRICAL RESISTIVITY RECONNAISSANCE SURVEY AT KABAGOLE

Following an initial review of the DEM, Landsat and Africover images described in Section 2.3 and 2.4, it was decided to focus the initial reconnaissance on the wide reach between the drainage divide and Nkonge. There are three road crossings within this reach, including Bihanga Station at the western end, Kabagole in the middle, and Nkonge at the eastern end. An initial reconnaissance was conducted using established 1D electrical profiling and vertical electrical sounding (VES) survey methods and locally available equipment in collaboration with the Uganda Department of Water Resource Management (DWRM). Given its central location in one of the widest parts of the reach, the Kabagole crossing was selected for the reconnaissance investigation.

The analysis and interpretation of the reconnaissance survey presented in this section was conducted in 2008 following the first field season. No attempt has been made to alter the results based on the ERT conducted in 2009. The results of the reconnaissance survey are compared against the Kabagole ERT survey in Section J.4 to assess the efficacy of 1D survey techniques and cost effectiveness of the 2D techniques.

J.1 Field procedures

J.1.1 Resistivity profiles

Kabagole was selected for the site of the initial reconnaissance investigation due to the wide valley and ease of access. An electrical resistivity profile was first conducted across the entire valley from an elevation of 1,226.4 metres above sea level (m asl) on the Mubende Granite to the north, through a minimum elevation of 1,182.2 m asl in the papyrus wetland, to 1,226.6 m asl on a weathered gneissic hill to the south (Figure 1-3 and Figure 6-10).

It was impractical to conduct an ideal linear resistivity profile across roads, fences, the adjacent Katonga Wildlife Reserve, and the papyrus wetland. Therefore the actual profile conducted along the side of the murrum road includes a small curvature as shown in Figure J-1. The potential error due to the curvature is examined in the context of the analysis.
A Schlumberger array configuration was used with the potential electrode half spacing \( (a/2) \) of 5 m and the current electrode half spacing of \( (L/2) \) of 27.5 m. The centre point was moved 20 m between each successive measurement. The pseudo-depth of investigation based on the median depth of investigation (Edwards, 1977) for an \( L/2 \) of 27.5 is 10.6 m below ground level (m bgl). The coordinates of the centre point and current electrode positions was recorded using a Geographical Positioning System (GPS) for each measurement location.

### J.1.2 Vertical electrical soundings

The resistivity profile was used in combination with practical access considerations to select six locations for VES across the Katonga Valley. The VES locations shown in Figure J-1 are summarised below,

VES 1 and 2 – Perpendicular to the valley axis, on a causeway through papyrus wetland, in an area of low resistivity on the profile.
VES 3 and 4 – Parallel to the valley axis, on grazing land, in an area of increasing resistivity to the south.

VES 5 – Parallel to the valley axis, on grazing land near sandstone outcrop, in an area of moderate resistivity on the profile.

VES 6 – Parallel to valley axis, on grazing land near sandstone outcrop, in an area of moderate resistivity on the profile.

Given that it is expected that the depth of geological layers are likely to remain constant parallel to the valley axis and vary perpendicular to the axis, all VES should be orientated parallel to the valley. However, this was not possible in the papyrus wetland, and so VES 1 and 2 were conducted along the causeway, perpendicular the valley axis. The resultant 1D vertical profiles may therefore be influence by 2D changes in resistivity (Figure J-1).

The resistance between the potential electrodes was measured with the current electrodes placed at each of the $L/2$ distances shown on the industry standard nomogram of 1.5, 2.1, 3, 4.4, 6.3, 9.2, 13.2, 19, 27.5, 140, 58, 83 and 120 m. The distance between the potential electrodes ($a$) was set at 1 m for $L/2$ distances between 1.5 and 27.5 m, and at 10 m for $L/2$ distances between 13.2 and 120 m. The overlap facilitated a quality assurance check by comparing the calculated resistivity for the different potential electrode spacing in the middle range. For the current electrode spacing between 3 m and 240 m the calculated median depth of investigation varies between about 0.52 m and 46.1 m respectively. Once again, the coordinates of the centre point and current electrode positions was recorded using a Geographical Positioning System (GPS) for each measurement location.

J.2 Analysis

J.2.1 Resistivity profile

A total of 166 resistance measurements were made. The GPS coordinates were used to plot the location of the survey line shown in Figure J-2. The corresponding elevations were extracted from the DEM based on SRTM data and used to construct the topographic profile shown in Figure J-2.
Figure J-2: Resistivity profile across the Katonga Valley at Kabagole

The resistance, and electrode half spacing ($L/2$ and $a/2$) were input to an Excel spreadsheet and Equations I-8, I-10 and I-11 were used to calculate the resistivity. The resistivity profile and topographic profile are shown in Figure J-2. The linear distance from one end of the profile to the other of 2988 m is 94.5% of the actual distance along the profile of 3,160 m. This ratio can be used to make an estimate of the potential magnitude of error in the apparent resistivity due to the curvature of the survey line. Given that the current electrode spacing ($L$) along the profile is 55 m, the curvature may reduce the linear current electrode spacing to 51.98 m. The error calculated in this way is sensitive to whether it is assumed that the potential electrode spacing ($a$) also reduces proportionally. If it remains 10 m then the difference in the calculated apparent resistivity is 13.3%, but if the linear distance between potential electrodes is also reduced proportionally to 9.45 m then the difference in the calculated apparent resistivity is 8%. It may therefore be assumed that the potential error is of the order of 10%, and when the apparent resistivities are 10, 100 and 1000 Ωm, the associated potential errors are 1, 10 and 100 Ωm respectively. Given that the variation in resistivity due to lithology ranges over orders of magnitude and the resistivity profile is used assess the relative apparent resistivity, the potential error does not undermine the objectives of the resistivity profiling. Nevertheless, caution is required not to over-interpret the apparent resistivity within the potential 10% error.

J.2.2 Vertical electrical soundings

Analysis of the VES began in a similar fashion to the resistivity profile. For each VES, the set of measured resistance, and corresponding electrode half spacing ($L/2$ and $a/2$) were input to an Excel spreadsheet and Equations I-8, I-10 and I-11 were used to
calculate the resistivities. All VES results are shown on a combined graph of apparent resistivity versus $L/2$ in Figure J-3.

Figure J-3: VES results shown as measured apparent resistivity versus current electrode half spacing ($L/2$)
For each VES, a data file was created including the array type, apparent resistivity data, electrode spacing and instructions for automated or user defined number of layers. This was then read into the RES1D code. The default damping factors and maximum number of iterations (12) were used throughout. An iterative manual procedure was used to assess the optimum number of layers to achieve a good balance between best-fit and parsimony, whilst using the same number of layers for all models. It was found that in all cases a good visual fit could be achieved between the measured and modelled...
apparent resistivity using a three layer model. The RMS error for the three layer models was between 3 and 8%.

Figures J-4 and J-5 present the measured apparent resistivity, the layered inverse resistivity model and the forward modelled apparent resistivity curves for each of the six VES. The data is plotted on a log-log graph with the resistivity shown on the horizontal access and median depth of investigation (Edwards, 1977) shown on the vertical axis. The median depth of investigation rather than the \( L/2 \) is shown on the vertical axis so that the curves may be interpreted in a pseudo-depth section and compared against the 2D pseudo-sections produced using the RES2DINV software (Loke, 2009).

All inverse models except VES 6 (Figure J-5) show a near surface decrease in resistivity, followed by an increase in resistivity at depth. For VES 6, the resistivity initially increases and there is a small decrease in resistivity at depth. All inverse models can be seen to have achieved a good fit between modelled apparent resistivity and the trend in the measured resistivity. The occasional small divergence of individual measurements from the model may represent either measurement error or local variation beneath the scale of interest of this reconnaissance study.

### J.3 Interpretation

In the absence of any geological data from borehole logs, the interpretation was carried out by comparing all six VES on a single graph (Figure J-6) in order to identify common layer depths and resistivities. Based on comparison of the layered resistivity models and surface observations the resistivity versus depth relationship was divided into classes A, B, C and D as shown in Table J-1. A preliminary geological interpretation is also included in the table. All models except VES 6 include a near-surface high resistivity layer (Class A), which is likely to be indicative of unsaturated ground. This layer is also present in VES 1 and 2 conducted in the papyrus wetland because the surveys were conducted on the murrum causeway which is itself about 2 m high.
<table>
<thead>
<tr>
<th>Class</th>
<th>Applicable VES Models</th>
<th>Depth Range (m bgl)</th>
<th>Resistivity Range (Ωm)</th>
<th>Preliminary Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>1, 2, 3, 4, 5</td>
<td>0 to 3.5</td>
<td>70 to 500</td>
<td>Near surface unsaturated ground</td>
</tr>
<tr>
<td>B</td>
<td>1, 2, 3, 4, 6</td>
<td>0 to 50</td>
<td>10 to 40</td>
<td>Silt and clay</td>
</tr>
<tr>
<td>C</td>
<td>4, 5, 6</td>
<td>1 to &gt;50</td>
<td>40 to 100</td>
<td>Fine-grained sandstone, fine-grained sand and weathered rock</td>
</tr>
<tr>
<td>D</td>
<td>3, 5</td>
<td>30 to &gt;50</td>
<td>500 to 1000</td>
<td>Weathered or fractured crystalline rock</td>
</tr>
</tbody>
</table>

**Table J-1: Classification and interpretation of VES resistivity model layers**

Class B has the lowest resistivity and occurs beneath the wetland and adjacent grazing land. Examination of hand dug water collection ditches for cattle at the edge of the papyrus wetland revealed the sediments to be moist, grey-brown mottle orange, firm, silty clay, with occasional fine to coarse quartz sand. On the basis of these observations and the resistivity range, Class B is interpreted to be predominantly saturated silt and clay. On the basis of the resistivity, the saturated silt and clay extend to the surface in the vicinity of VES 6. This is consistent with the low lying damp ground observed at this location.

Class C represents moderate resistivity ground beginning between 1 and 20 m bgl and extending to depth. During the reconnaissance investigation a yellow brown, thinly bedded to laminated, fine-grained sandstone was identified on the topographic mound north of VES 5 (Figure J-6). The model layers for VES 5 and 6 which fall into Class C are therefore interpreted as fine-grained sandstone. VES 4 also contains a layer in Class C, but this is located 2km south of the sandstone outcrop on the opposite side of the papyrus wetland. In this case, Class C may be interpreted on the basis of the resistivity range as fine-sand or weathered rock.
Figure J-6: Classification of layers common to several VES resistivity inverse models

Class D represents the highest modelled resistivities, which occur below 30 m depth. During the reconnaissance, the Mubende Granite was observed to outcrop north of the Katonga Valley and weathered gneisses outcrop to the south. The contact must therefore occur at an unknown location beneath the valley fill. Fresh crystalline rocks generally have a resistivity greater than 1000 $\Omega$m (Loke, 2009). Since the resistivity range of this class is between 500 and 1000 $\Omega$m, it is interpreted to represent weathered or fractured granite or gneiss.

Figure J-7 presents the interpretation of a section perpendicular to the Katonga Valley based on the combined results of the reconnaissance survey. The horizontal scale is distance along the profile. There are two vertical scales. The upper profile is the topography, corresponding to the elevation shown on the vertical axis on the right. The
lower profile is the electrical resistivity, corresponding to the vertical log-axis on the left. The VES models are also shown on a log-linear scale and placed on the section, with their centre locations located on the profiles, and the pseudo-depth drawn at an equivalent vertical scale to the topography. The geological interpretation shown in Table J-1 has been used to draw lines between the class boundary depths on the VES models.

Figure J-7: Preliminary interpretation of section perpendicular to the Katonga Valley based on the 1D electrical resistivity reconnaissance

It should be remembered that the electrical resistivity profile shown on this section has a constant pseudo-depth of 10.6 m and cannot be related to the elevation axis. Several horizontal features can be identified on the profile, which are consistent with the VES interpretation. From south to north, the resistivity is high beneath the valley slope with the weathered gneiss outcrop, and falls to a minimum beneath the southern grazing land. There is a gradual increase in resistivity from south to north beneath the papyrus wetland. A line of moderate resistivity coincides with the low mound where the fine-grained sandstone outcrops in the centre of the Katonga Valley. To the north, there is another section of low resistivity, before it increases again beneath the valley slope where the Mubende Granite outcrops to the north.

When plotted on a linear scale the relatively shallow depth of the unsaturated ground (Class A) in the valley is evident. This 2D interpretation of the 1D VES suggests that the deepest part of the silt and clay (Class B) is located beneath the southern grazing
land rather than the papyrus wetland. The interpreted distribution of Class C suggests that the fine-grained sandstone may actually underlie a considerable volume of the valley. VES 6 suggests a second area of silt and clay fill (Class B) north of the fine-grained sandstone. The higher resistivities of Class D were only suggested by the inverse modelling of VES 3 and 5. These have been combined with the information from the resistivity profile to suggest a continuous higher resistivity layer at depth, associated with the crystalline bedrock. There is no evidence for the location of the contact between the underlying igneous rocks to the north and the metamorphic rocks to the south.

J.4 Comparison of 1D and 2D Resistivity Surveys as Kabagole

Given that multi-electrode ERT equipment (e.g. IRIS Instruments Syscal R1 Plus Switch-72) costs over four times as much as four electrode VES equipment (e.g. Abem Terrameter SAS-1000) and the latter is already commonly used in Uganda, it is useful to compare the efficacy of the two systems. Such an assessment is facilitated by this study because the interpretation of the 1D survey was conducted before the 2D survey along the same survey line at Kabagole. The interpretation of the 1D reconnaissance survey was not updated following the detailed ERT. In environments where the solid geology and weathering profile is consistent over a large area then it might be expected that 2D survey techniques would provide little advantage compared to the established 1D techniques commonly used in sub-Saharan Africa. However, in environments such as infilled valleys where the vertical profile is likely to vary perpendicular to the valley axis it might be expected that 2D techniques would facilitate a more accurate interpretation.
Figure J-8: Interpretation of 1D Survey overlain on ERT inversion model

Figure J-8 shows the zones of similar resistivity manually interpolated from the 1D inversion results (Figure J-7) superimposed on the 2D inversion model produced using the multi electrode ERT survey at Kabagole (Figure 6-13b). The orientation of the 1D interpretation is reversed compared to Figure J-7. Whilst the interpretation of the reconnaissance survey is based on six VES consisting of 13 data points each, making only 78 resistivity measurements in total, the ERT inversion model is produced using 15768 data points. Given that the reconnaissance survey is based on only 0.5% of the quantity of data obtained from the ERT survey there is a relatively close match between the two models. The VES and ERT derived models identify the same five regions with internally similar resistivity ranges on the vertical cross section, which are: 1) the high resistivity northern granite; 2) the low resistivity northern clayey silt paludal/alluvial deposits; 3) the medium resistivity central fine-grained sandstone; 4) the low resistivity southern clayey silt paludal/alluvial deposits; and 5) the high resistivity southern metamorphic rocks.

Whilst the modelled resistivity in the 2D ERT inversion was allowed to vary relatively smoothly within the constraints listed in Table 6-3, the 1D inversion models were required to fit the VES apparent resistivity data using only three layers. The colour scheme selected for the ERT visualisation draws attention to the boundary between yellow and light blue at the 69 Ωm contour. This contour was selected as the likely boundary between sand and silt/clay based on the ground truth information provided by the boreholes at Kisozi. However, the resistivity range selected for grouping the middle resistivity range of the 1D inversion models is 40 to 100 Ωm. The parsimonious three
layer 1D interpretation therefore implies that the lithological boundary at Kabagole may occur on the ERT inversion model between the 32 to 47 Ωm contours (between light and dark blue) rather than at the 69 Ωm contour (between yellow and the light blue). It is reasonable to conclude that the very low resistivity material (< 32 Ωm) represents unconsolidated silts and clay. However, the depth of the resistivity boundary coincident with the surface of the indurated sedimentary rocks beneath the papyrus wetland and low lying grazing land is uncertain. Given that it is unlikely for anastomosing fluvial channels to have eroded contemporaneous deep gorges on both sides of a resistant sandstone it seems unlikely that the alluvial fill extends below the maximum depth of investigation (>68 m bgl) on both sides of the sandstone outcrop. Nevertheless, even if the boundary between alluvium and sedimentary rocks occurs around the 32 Ωm contour, the maximum depth of the alluvium is still about 30 m bgl. If the indurated sedimentary rocks are contiguous below the papyrus wetland and low lying grazing land then lithological variations or differences in groundwater conductivity could be responsible for the slightly lower resistivity.

This comparison of results from the 1D and 2D resistivity surveys at Kabagole has shown that six VES were able to define the general distribution of the main subsurface resistivity units as the ERT survey. However, the efficacy of the 1D survey depended more heavily on the experience and judgment of the analyst in siting the surveys and interpreting the results than did the ERT survey. The understanding derived from producing parsimonious 1D inversion models and manual interpolation complemented the greater coverage and resolution obtained from the ERT survey. It is concluded that initial experience conducting a 1D survey can assist later interpretation of ERT. In general, VES surveys can be as accurate as ERT surveys, but have far lower resolution. Therefore, in environments where the geology varies horizontally, the ERT provides greater confidence that resistivity variations have been identified. The introduction of ERT to Uganda therefore offers the possibility of enhanced geological characterisation in specific circumstances. However, the use of multi electrode ERT equipment does not necessarily provide more valuable information than four electrode VES equipment in areas with horizontally consistent planar lithological profiles.
K PUMPING TEST ANALYSIS THEORY AND METHODS

K.1 Overview

Pumping test analysis is the third inverse modelling technique employed in this thesis. The principles of pumping test analysis are perhaps more easily understood as a pattern recognition problem (Gringarten, 1979, 1986). A known input signal (the pumping rate) is applied to an unknown system (the borehole, aquifer/formation and hydraulic boundaries) and the output response (change in head with time) is measured during the test. The objective of pumping test analysis is to identify the properties of the hydrogeological system knowing only the input and output signals and some relevant information about the borehole and the geology. As previously discussed, the solutions to inverse problems are often non-unique, and we must use the assumption of parsimony and our limited prior knowledge to select an appropriate solution. We do this by producing analytical models of the theoretical borehole-aquifer system and forward modelling the output response for the known input signal. The objective is to identify which analytical model best fits the actual output response observed during the pumping test. This is in contrast to the approach adopted during thermal history modelling in Chapter 5, where the limited complexity and information content of the fission track age and length histograms only facilitated the identification of regions of the time-temperature domain which were inconsistent with the data, and less emphasis was placed on the overall best fit model. Nevertheless, it should be noted that the best fit model adopted during pumping test analysis may sometimes give a false impression of the conceptual certainty, especially in a system of complex hydraulic geometry as may be expected in fractured rock-regolith systems.

Some of the analysis techniques adopted during this study were first developed for use in the petroleum industry. Whilst it may at first appear that the high technology approach of the petroleum industry is far removed from the intermediate technology required for African groundwater development, both sectors do have at least one thing in common; they rarely have observation wells at their disposal, and often require analysis of pumping well data. Section K.2 summarises the principles of log-log type curve analysis for unsteady-state flow towards a well. Section K.3 then briefly describes the semi-log approximations based on the straight line analysis methods which are recommended for rural water supply development where experienced
analysts and computer-aided techniques are unavailable (Macdonald et al., 2005). Sections K.4 to K.8 go on to describe the computer-aided log-log type curve analysis methods which incorporate wellbore storage and skin, derivative analysis, and Agarwal equivalent time to analyse drawdown and recovery data simultaneously. These methods help to reduce the local underdetermination of the inverse solution and are recommended here for town, industrial and agricultural water supplies where experienced hydrogeologists and computer-aided analysis techniques are available. A more detailed description of the pumping test analysis techniques employed here can be found in Kruseman and de Ridder (1990) and Horne (1995).

**Log-log Type Curve Analysis**

Theis (1935) first developed an analytical solution for unsteady-state flow (transient pressure conditions). He noted that the rate of decline of head, multiplied by storativity and summed over the area of influence equals the discharge under the following assumptions:

1. the aquifer is confined (elastic storage);
2. the aquifer has infinite areal extent;
3. the aquifer is homogeneous, isotropic and has uniform thickness;
4. prior to pumping the piezometric surface is horizontal;
5. the groundwater is pumped at a constant rate; and
6. flow towards the well is both horizontal and radial (cylindrical).

By analogy with the conduction of heat, Theis (with mathematical help from C. I. Lubin) derived the equation for unsteady-state flow as follows:

\[ s = \frac{Q}{4\pi KD} \int_{u}^{\infty} e^{-y} \frac{dy}{y} = \frac{Q}{4\pi KD} W(u) \]  

**Equation K-1**

\( s \) — drawdown measured in observation well at distance \( r \) from the pumping well (m)

\( Q \) — pumping rate (m³/day)

\( KD \) — hydraulic conductivity (m/day) \( \times \) aquifer thickness (m) = transmissivity (m²/day)
\[ u = \frac{r^2 s}{4 K D t} \quad \text{and consequently} \quad \frac{r^2}{t} = \frac{4 K D u}{s} \quad \text{Equation K-2} \]

\[ S \quad \text{storativity ( - )} \]

\[ t \quad \text{time since pumping started (days)} \]

\[ r \quad \text{radial distance from the pumping well (r)} \]

\[ W(u) = -0.5772 - \ln u + u - \frac{u^2}{2.2!} + \frac{u^3}{3.3!} + \frac{u^4}{4.4!} + \ldots \quad \text{Equation K-3} \]

Equation K-1 includes a special function called an exponential integral which in hydrogeology is known as the ‘well function of u’ (W(u)), where its argument, \( u \), is defined in Equation K-2. This function is required to describe the shape of the ideal curve of drawdown versus time. Equations K-1 and K-2 can be rearranged and transformed to separate out the \( W(u) \) and \( u \) as follows:

\[ \log s = \log \left( \frac{Q}{4\pi K D} \right) + \log(W(u)) \quad \text{Equation K-4 (from Equation K-1)} \]

\[ \log \left( \frac{r^2}{t} \right) = \log \left( \frac{4 K D}{S} \right) + \log(u) \quad \text{Equation K-5 (from Equation K-2)} \]

The transmissivity and storativity are then derived using the method of type curve matching. This analysis method is based on the fact that if \( s \) is plotted against \( \frac{r^2}{t} \) (the data curve) and \( W(u) \) against \( u \) (the type curve) on log-log axes, the data and the type curves will be the same shape but offset by the constants \( \frac{Q}{4\pi K D} \) and \( \frac{4K D}{S} \). The curves can be made to match, and since \( Q, s, t, r, W(u) \) and \( u \) are known for any arbitrary match point, they can be used to calculate transmissivity (KD) and storativity (S) from Equations K-4 and K-5. Since the data and type curves have the same shape, \( W(u) \) and \( u \) are known in the oil industry as ‘dimensionless pressure’ (\( P_D \)) and ‘dimensionless time’ (\( t_D \)).

The Theis type curve method is based on the ideal response identified in an observation well where the pumping well may be assumed to approximate to a line source. Since the early-time pumping well response will be influenced by the properties of the pumping well itself, including its finite storage volume and near-field disturbance/connections with the wider formation, the Theis type curves will usually be a poor fit to the early-time pumping well data.

**Semi-log Straight Line Analysis**
The simplest and most appropriate analysis methods for use by relatively inexperienced analysts without access to computer-aided techniques are based on the fact that if displacement versus time is plotted on a semi-log graph the period of radial flow towards the well is characterised by a constant slope which is inversely proportional to the transmissivity. Macdonald et al. (2005) point out that the best quality data obtained during a pumping test is often derived from the recovery period since the effect of small changes in pumping rate are smoothed out. Therefore, they recommend the Theis (1935) recovery method for rural water supplies. However, when the pumping rate is reasonably well controlled, and/or recovery data is unavailable, Macdonald et al. (2005) recommend Jacob’s method (Cooper and Jacob, 1946) for analysing the drawdown data. Figure K-1 show an example of Jacob’s method and Theis-recovery analyses for data collated during this study for Kagera borehole Reg. No. 307. This dataset was chosen as an example due to the well controlled flow rate and good quality displacement versus time data.

The straight line analysis methods are described here so that the results may be compared against the more sophisticated analysis methods used during this study. Jacob’s method (Cooper and Jacob, 1946, Kruseman and de Ridder, 1990) is based on an approximation of the Theis equation. It can be seen from Equations K-2 that $u$ decreases as time of pumping increases and the radial distance ($r$) decreases. For small values of $u$ ($<0.01$) Equation K-1 can therefore be approximated in terms of the base 10 logarithm as:

$$s = \frac{2.30Q}{4\pi KD} \log \frac{2.25KDt}{r^2S} \quad \text{Equation K-6}$$
When the flow rate, transmissivity and storativity are constant, a plot of drawdown against the logarithm of time will form a straight line. If the straight line is extended until the drawdown is zero, it intercepts the time axis at $t_0$ and since

$$s = 0 \quad \text{and} \quad \frac{2.30Q}{4\pi KD} \neq 0$$

it follows that $\log \frac{2.25KDt_0}{r^2S} = 0$ and $\frac{2.25KDt_0}{r^2S} = 1$

and therefore $S = \frac{2.25KDt_0}{r^2}$, Equation K-7

Since, the slope of the straight line $\frac{2.30Q}{4\pi KD} = \frac{\Delta s}{\log t/t_0}$ and $\log t/t_0 = 1$

it follows that $KD = \frac{2.30Q}{4\pi \Delta s}$ or $\frac{0.183Q}{\Delta s}$, Equation K-8

Where $\Delta s$ is the change in drawdown measured over one log cycle of time. Thus the transmissivity can be calculated from the slope of the straight line portion of the plot of drawdown versus log time, as shown for the example analysis of data from Kagera borehole Reg. No. 307 in Figure K-1a. Since the storativity calculation is based on the time delay ($t_0$) between the start of pumping and the arrival of the pressure signal, it is best derived from analysis of observation well data, and a reliable storativity cannot be derived from the pumping well data. However, since the transmissivity is derived from
the rate of change of head which is influenced by the formation beyond the measurement borehole, it can be derived from the pumping well data.

The Theis recover method (Theis, 1935, Kruseman and de Ridder, 1990) is based on a similar approximation to Jacob’s method. If the residual drawdown \( s’ \) is plotted against the logarithm of time since the start of pumping over the time since cessation of pumping \( \log t/t’ \) it can be shown that, similarly to Equation 7.8:

\[
KD = \frac{2.30Q}{4\pi \Delta s^2} \text{ or } \frac{0.183Q}{\Delta s^2} \tag{Equation K-9}
\]

Thus, the transmissivity can be calculated from the slope of the straight line portion of residual drawdown versus \( \log t/t’ \), as shown for the example analysis in Figure K-1b. The slope of the straight lines in Figures K-1a and K-1b are both equivalent to a transmissivity of 18 m\(^2\)/day. Given that the straight line is a better fit to the drawdown data than the recovery data, there appears to be a small inconsistency between the datasets. The departure from the straight line in the early-time data in Figure K-1a and K-1b is influenced by the properties of the pumping well itself and the late time data departs from the straight line due to the apparent influence of a hydraulic boundary. The causes of these effects are discussed next.

**K.4 Storage and Skin**

**K.4.1 Pumping well storage**

The Theis type curve method was developed for analysing pumping test data acquired from an observation well under conditions where it is appropriate to assume a line source and the pumping well properties do not influence the hydraulic response in the observation well. However, no observation well data are available for the pumping tests analysed during this study. The properties of the pumping well itself must therefore be considered as the inner boundary conditions for the analytical solution. These include the borehole volume (wellbore storage) and the zone of altered hydraulic properties (skin) in and around the borehole. In an open well, the wellbore storage is the volume per unit length of borehole and its effect can be understood by considering the difference in the downhole flow rate entering the open borehole and the flow rate measured at the surface (Gringarten, 1979, Dougherty and Babu, 1984). The moment that pumping commences and drawdown begins the initial surface flow occurs due to
the emptying of water from the wellbore (and a small amount of decompression) rather than water entering the borehole from the aquifer. As pumping proceeds the drawdown in the borehole increases and a greater proportion of the surface flow is derived from the aquifer. Wellbore storage therefore only influences the early-time pumping test data. Eventually the rate of drawdown in the borehole decreases as the condition of pseudo-steady state radial flow towards the well is approached. At this point the influence of wellbore storage is negligible and effectively all of the flow pumped at the surface is derived from the aquifer.

K.4.2 Well losses and the skin factor

The drawdown in a pumped well occurs due to head loss created by friction in the aquifer and head loss created by friction and turbulent flow in and around the well. Well losses include a linear component that remains constant during drawdown, and a nonlinear component that increases with drawdown due to the onset and increase of turbulent flow (Kruseman and de Ridder, 1990). For the drawdown in a pumped well, Jacob (1947, cited in Kruseman and de Ridder, 1990) defined the following relationship

\[
s_w = B(r_{ew}, t)Q + CQ^2
\]

**Equation K-10**

- \(s_w\) – drawdown in pumped well (m)
- \(B(r_{ew}, t)\) = \(B_1(r_w, t) + B_2\)
- \(B_1(r_w, t)\) – linear aquifer head loss coefficient
- \(B_2\) – linear well head loss coefficient
- \(C\) – non-linear well head loss coefficient
- \(Q\) – pumping rate (m³/day)
- \(r_{ew}\) – effective well radius (m)
- \(r_w\) – actual well radius (m)
- \(t\) – time since pumping started (days)

In groundwater resource studies, step drawdown tests are conducted to determine both the linear well loss and non-linear well loss coefficients. This enables selection of the optimum efficient yield for a specific well. However, most tests analysed during this study were constant rate and therefore the non-linear well loss coefficient could not be
calculated. All well losses are therefore assumed to be linear and are accounted for by a parameter known as the ‘skin’ factor which was originally introduced in the petroleum industry (Horne, 1995). The head loss coefficients in a pumping well may be related to the skin factor via the following equation (Ramey, 1982, cited in Kruseman and de Ridder, 1990):

\[ B_2Q + CQ^2 = \frac{1}{2\pi KD}(\text{skin} + C'Q)Q \]  

Equation K-11

Where

\[ C' = C \times 2\pi KD \quad \text{– high velocity well loss coefficient} \]

\[ S_w = B_2 \times 2\pi KD \quad \text{– skin factor} \]

Positive skin occurs when the interface between the wellbore and the formation appears less conductive than the formation. It can occur due to the development of a mud cake during drilling, improper screen size or silting up of the filter pack for example. In fractured rock, the presence of apparent positive skin can occur when the borehole is connected to a higher transmissivity fracture network through a lower transmissivity fracture. Negative skin occurs when the interface between the wellbore and the formation is more conductive than the formation. In fractured rock, it is most often seen when the borehole is connected to a lower transmissivity fracture network through a higher transmissivity fracture. Skin effects influence the shape of the pumping test response between the period dominated by wellbore storage and the period dominated by the properties of the aquifer itself.

**K.4.3 Wellbore storage and skin type curves**

Using the same principles as the Theis curve matching method, Gringarten et al (1979) working in the petroleum industry, developed a set of type curves of dimensionless pressure (\(P_D\)) versus dimensionless time over dimensionless wellbore storage (\(t_D/C_D\)). Each curve is characterised by a different value of \(C_D\) and is therefore related to the skin (s). A similar approach was adopted in the hydrogeological literature for the type curves developed by Dougherty and Babu (1984) and this model is implemented in the AQTESOLV pumping test analysis software used during this study.
The wellbore storage and skin type curves for the pressure (i.e., head) response are labelled in red on Figure K-2. It can be seen that the period of wellbore storage is characterised by a unit slope (1:1) on the log-log plot. The value of wellbore storage is usually determined directly from the physical borehole volume and is therefore dependant on the casing radius alone. The shape of the transition from wellbore storage in the borehole to the radial flow in the aquifer is influenced by both borehole skin and aquifer storage. Unfortunately, the effects of skin and aquifer storage on the pumping test response are similar. Lower storage coefficients and positive skin accentuate the inflection during the transition, whereas higher storage coefficients and negative skin subdue the inflection during the transition. In the petroleum industry, during the analysis of pumping well data it is usual to calculate the confined storage coefficient based on the porosity ($\phi$) × total compressibility ($C_t$) × test thickness ($h$) or assume a reasonable commonly observed value of around 1E-4. The analyst selects the $C_{De}^{2S}$ type curve which best fits the data and the match points between the type curve and data curve are used to calculate transmissivity and skin in a similar manner to the Theis analysis.
K.5 Other Aquifer Models and Boundary Conditions

Type curves have been developed for pumping test responses to many theoretical aquifer geometries, including:

1. Single fractures – early-time flow from planar (1D) vertical fracture e.g., Gringarten and Witherspoon (1972) and Gringarten et al. (1974)
2. Unconfined aquifers – mid-time delayed yield from the cone of depression e.g., Neuman (1974) and Moench (1997);
3. Double porosity – mid-time delayed yield from the matrix storage e.g., Warren and Root (1963) and Moench (1984);
4. Leaky aquifers – late-time contribution of vertical flow (3D) from storage above and/or below the pumped aquifer e.g., Neuman and Witherspoon (1969) and Moench (1985)
5. Generalised radial flow – flow from hydrogeological unit (e.g., fracture network) where rate of change in flow area of contributory conduits (i.e., the flow dimension) is not necessarily an integer (i.e., does not need to equal 1D, 2D or 3D) e.g., Barker (1988)

Given the observed responses and the limitations of the flow rate control, test duration, and available geological data for the pumping tests analysed here, the application of these solutions were in general either unnecessary or unjustifiable. In most cases the parsimonious solution simply includes wellbore storage and skin followed by infinite acting radial flow (Dougherty and Babu, 1984).

Two tests (Mukono DWD17021 and Kagera Reg. No. 307) show a quicker reduction in the late-time slope of displacement versus time than expected for infinite acting radial flow. This response may be modelled as a leaky aquifer (horizontal recharging), but in this case it was decided to model these responses using a constant head boundary (vertical recharging). The response of a constant head boundary resembles that which would be expected if the transmissivity were to approach infinity.

Aquifer boundaries may be modelled by introducing simulated ‘image wells’ into the existing analytical solution (Kruseman and de Ridder, 1990). The modelled discharge
or recharge from the simulated image wells is used to reproduce the same effect as a no-flow or constant head boundary respectively. Stallman (presented in Ferris et al., 1962) developed a curve matching method for one or more linear recharge or no-flow boundaries. If the distance between the real pumping well and an observation point is \( r \), and the distance between the image well and the same observation point is \( r_i \), the ratio of the distances is defined as \( r_i/r = r_i \). Therefore, from Equation K-2:

\[
\frac{r_i^2 S}{4 K_D t} = \frac{r_i^2 r^2 S}{4 K_D t} = r_i^2 u
\]

Equation K-12

For a single recharge boundary, from Equation K-1:

\[
s = \frac{q}{4\pi K_D} \left[ W(u) - W(r_i^2 u) \right]
\]

Equation K-13

Thus test-specific type curves can be constructed based on the modified Theis well function of \( u \) for a constant head boundary, represented by a simulated image well, at a given distance from the observation point. In this study, the observation point is the pumping well itself.

**K.6 Derivative analysis**

The development of downhole gauges capable of recording the pressure versus time and computer-aided analysis techniques were the catalyst for many developments in well test analysis in the petroleum industry during the early 1980s (Bourdet et al., 1983, Gringarten, 1986). These developments include analysis of the rate of change of pressure with respect to the natural logarithm of time (\( \delta s/\delta \ln t \)). Some 30 years after its initial development, electronic pressure gauges and data loggers are starting to be used to monitor and test boreholes for some town-water supplies in sub-Saharan Africa. However, with the increasing portability and accessibility of computer-aided analysis, it will be shown here that derivative analysis is a useful tool for pumping test diagnostics and type-curve matching even when there are only manual water level measurements available.

Derivative analysis was originally introduced in the petroleum industry (Bourdet et al., 1983, Bourdet et al., 1989) and later entered the groundwater literature (Spane and Wurstner, 1993), although to this day it continues to remain undervalued by many
hydrogeologists. The derivative is a useful diagnostic tool because it is more sensitive to small variations in the rate of pressure change than a simple plot of pressure versus time, and it also simplifies recognition of specific flow regimes. For example, since the onset of two dimensional radial flow is characterised by a constant slope on the plot of drawdown versus log time, the derivative on a semi-log plot is characterised a horizontal line. This removes the need to identify different slope angles, and instead the analyst can more easily identify the level of the horizontal stabilisation corresponding to a particular transmissivity. The horizontal derivative corresponding to the constant slope matched with a straight line for Jacob’s method is shown in Figure K-1a. In practice, the derivative is calculated as a weighted average of the slopes computed for data points on either side of the data point \(i\) (Duffield, 2007), as follows:

\[
\frac{\delta s}{\delta \ln T} = \frac{\Delta s_{i-1}/\Delta \ln T_{i-1} \Delta \ln T_{i-1} + \Delta s_{i+1}/\Delta \ln T_{i+1} \Delta \ln T_{i+1}}{\Delta \ln T_{i+1} + \Delta \ln T_{i-1}}
\]

Equation K-14

Where \(\Delta s_{i-1}/\Delta \ln T_{i-1}\) and \(\Delta s_{i+1}/\Delta \ln T_{i+1}\) are the two slopes.

The degree of smoothing used to reduce noise in the derivative depends on the value of \(l\) selected in Equation K-12. The Bourdet (1989) method is used in this study where the derivative at point \(i\) uses data points for the differential interval separated by between 0.1 and 0.5 log cycles.

Figure K-2 shows the \(C_{pS}e^{2S}\) type curves derived by Bourdet et al (1983) for the semi-log pressure derivatives labelled in blue on the log-log plot. Both the pressure versus time and the semi-log derivative have a 1:1 slope corresponding to wellbore storage in early-time. As expected, the mid to late-time derivative corresponding to radial flow, is a horizontal line. For a given storage coefficient, the shape of the intermediate transition corresponds to a particular skin factor. The lower the radial flow stabilisation on the derivative the lower, the slope on the semi-log plot and the higher the transmissivity. The higher the radial flow stabilisation on the derivative, the higher the slope on the semi-log plot and the lower the transmissivity. Consequently, it is intuitive to visualise a large hump on the transition as a large increase in transmissivity, corresponding to a large positive skin. In contrast, the lack of a hump in the transition suggests a decrease in transmissivity corresponding to a negative skin. Alternatively, if the skin is zero, then a low storage coefficient will produce a sharp inflection on the semi-log plot and a large hump in the derivative. In contrast, a high storage coefficient
will produce a gradual change in slope on the semi-log plot and a subdued hump or no hump on the derivative. For the analyses of pumping test data from the pumping well it is common to assume the storage coefficient and calculate the skin.

Figure K-3 shows an example log-log diagnostic plot of the drawdown versus time and the derivative for Kagera borehole Reg. No. 307. Throughout the pumping test interpretations presented in this thesis, ‘early-time’ refers to the period dominated by wellbore storage and the start of the transition, ‘mid-time’ refers to the period of transition and radial flow, and ‘late-time’ refers to either continued infinite acting radial flow or the onset of boundary conditions, as annotated on Figure K-3. It can be seen from Figure K-3 that the first data point recorded after 1 minute corresponds to the end of early-time where the response is beginning the transition out of the period dominated by wellbore storage. The storativity is assumed to be 1E-4, which in this case accounts for the observed hump in the derivative during the transition in the mid-time data without any skin effects (i.e., skin = 0). The horizontal two dimension (2D) radial flow stabilisation corresponds to a transmissivity of 18 m²/day, which is the same as the straight line match on the semi-log plot shown in Figure K-1a. The late-time decrease in the derivative has been matched using a constant head boundary. The overall match between the data and the model presented in Figure K-3 provides a more comprehensive interpretation of the data than the straight line analyses presented in Figure K-1 which increases confidence in the derived value of transmissivity. The sensitivity and diagnostic power of the log-log displacement versus time and derivative plots, combined with knowledge of the common borehole-aquifer responses, is particularly useful when there are fluctuations in the flow rate and/or limited recovery data.
Through a simple transformation of the time variable, Agarwal (1980) devised a procedure that enables the same solutions developed to analyse drawdown data to be used to analyse recovery data. This procedure has been used during this study to analyse all available recovery data. Where there are two different flow rates, the Agarwal equivalent time during the second flow period is calculated as follows:

\[ t_{equiv} = \left[ \left( \frac{t_1}{t_1 + t'} \right) \left( \frac{q_1}{q_1 - q_2} \right) \right] t' \]  

Equation K-15

Where

- \( t_{equiv} \) – Agarwal equivalent time
- \( t_1 \) – duration of first pumping period (days)
- \( t' \) – duration since pumping rate changed (days)
For those tests in which a single constant flow rate test has been applied prior to the recovery period, the pumping rate of the second flow period is effectively zero the Agarwal equivalent time is simplified to:

\[ t_{equiv} = \frac{t_p x t_r}{t_p + t'} \]  

Equation K-16

Where

- \( t_{equiv} \) – Agarwal equivalent time
- \( t_p \) – duration of pumping period (days)
- \( t' \) – duration since pumping stopped (days)

Unfortunately, either intentionally or unintentionally, the flow rate during some of the tests analysed here fluctuated during the drawdown. For multiple flow rate tests the Agarwal equivalent time (Agarwal, 1980) is calculated by AQTESOLV (Duffield, 2007) using the generic formula taking into account differing constant flow rates of variable duration for flow periods 1 to \( n \) as follows:

\[
t_{equiv} = \left[ \prod_{j=1}^{n-1} \left( \frac{t_{n-1} - t_{j-1}}{t'+t_{n-1} - t_{j-1}} \right) \right]^{q_j - q_{j-1}} \frac{q_{j-1} - q_j}{q_{n-1} - q_n} t' \]

Equation K-17

The recovery is calculated as follows:

\[ s_{recov} = s_p - s' \]

Where

- \( s_p \) – drawdown when pumping stopped (m)
- \( s' \) – residual drawdown (m)

To conduct the recovery analysis using the drawdown type curves, the recovery is plotted against the Agarwal equivalent time (also known as the time superposition function). For derivative analysis of the recovery data, the derivative is simply calculated with respect to the recovery and the natural logarithm of the Agarwal equivalent time.