# HOLOCENE ENVIRONMENTAL AND HUMAN INTERACTIONS IN EAST AFRICA

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# Abstract

A multi-proxy approach analysing pollen, macro-charcoal, sediment characterisation and elemental profiles was used to develop palaeoecological records and reveal environmental changes since the late Pleistocene-Holocene transition period from Mau Forest and since the mid-Holocene from Amboseli. Mau Forest was characterised by diverse Afromontane forest taxa between ∼16,000 cal yr BP and ∼13,000 cal yr BP which decreased during the Younger Dryas. During the early Holocene, there was a slight increase in montane tree taxa and the main vegetation change noted during the Holocene was the increase in woody shrubs and herbs. The pollen, sediment characterisation and elemental profiles revealed that climatic variability was the main driver of forest composition change and periods of aridity and wetness were identified at ∼15,000, ∼13,400, ∼12,000 and ∼1200 cal yr BP where there was increased organic matter, sand, magnetic susceptibility with peaks in detrital elements suggesting periods of wetness.

Four new Amboseli records dating from the mid Holocene (∼5000 cal yr BP) revealed a predominantly dry environment characterised by localised wet and dry phases and fire activity. The spatial differences observed from the Amboseli records are attributed to hydrological variance as the swamps are all fed by ground water and the differential use by humans and wildlife. Kimana, Enkongu and Esambu swamps are Cyperaceae dominated; the pollen records indicate that Amboseli is a grassland savannah dominated by Poaceae, *Acacia, Commiphora* and *Euphorbia*. The pollen composition and abundance and charcoal concentration levels vary between the four Amboseli sites indicating localised drivers and controls of fire at each site. This long-term information is useful in the development of ecosystem management policies which are constantly being updated due to the evolving pressures caused by increasing populations and changing land use around the two ecosystems.

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# Abbreviations

| | |
|---|---|
| CAB | Congo Air Boundary |
| CONISS | Constrained Incremental Sum of Squares |
| ENSO | El Niño Southern Oscillation |
| IOD | Indian Ocean Dipole |
| ITCZ | Intertropical Convergence Zone |
| LGM | Late Glacial Maximum |
| LOI | Loss on Ignition |
| LSA | Late Stone Age |
| SST | Sea Surface Temperature |
| MCA | Medieval Climate Anomaly |
| MFC | Mau Forest Complex |
| NPP | Non Pollen Palynomorphs |
| PCA | Principal Component Analysis |
| PSA | Particle Size Analysis |
| RPM | Revolutions Per Minute |
| XRF | X-Ray Fluorescence |

# Author's Declaration

The candidate confirms that the work submitted is her own; where work has formed part of jointly authored publications the contribution of the candidate and the other authors to this work has been explicitly indicated below. The candidate confirms that appropriate credit has been given within the thesis where reference has been made to the work of others.

I declare that this thesis is a presentation of original work and I am the sole author. This work has not previously been presented for an award at this, or any other university. All citations are acknowledged in the bibliography. I was responsible for the research design, data collection, data analyses, interpretation and discussion of the results and drawing of the figures. The ITRAX analysis was carried out at the British Ocean Sediment Core Research Facility (BOSCORF) in Southampton and by myself at Aberystwyth University, Department of Geography and Earth Sciences. Sample processing was carried out at National Museums of Kenya, Palynology and Palaeobotany section and University of York, Environment Department with lab assistance from staff at both laboratories.

Parts of the literature in Chapter 2 of the thesis contribute into a publication analysing land cover in East Africa over the last 6000 years published in *Earth Science Reviews*. The work in chapters 3 and 4 are planned to be submitted to peer-reviewed science journals.

The Nyabuiyabuyi work presented in chapter 3 was performed by Esther Githumbi. **Work attributable to candidate:** Field work, preparation and analysis of the samples, data processing and interpretation and manuscript preparation. **Work attributable to others:** Mustaphi, C.J.C assisted with coring, macroscopic charcoal processing. Marchant, R. was the recipient of the field and lab work grant and provided assistance with editing of the manuscript. Some of the data from the Nyabuiyabuyi site was contributed to the manuscript (Subfossil statoblasts of *Lophopodella capensis* (Sollas, 1908) (Bryozoa, Phylactolaemata, Lophopodidae) in the Upper Pleistocene and Holocene sediments of a montane wetland, Eastern Mau Forest, Kenya) in *African Invertebrates* 57(1):39-52 May 2016.

The Amboseli work presented in chapter 4 was done by Esther Githumbi. **Work attributable to candidate:** Field work, preparation and analysis of the samples as required for the work, data processing and interpretation, manuscript preparation. **Work attributable to others:** Mustaphi, C.J.C assisted with coring and manuscript preparation. Muriuki, R. assisted with macroscopic charcoal, pollen and LOI sample processing of the Enkongu sediment core, Yun, K.J assisted with macroscopic charcoal processing in 2015 for his undergraduate research project, Muiruri, V. and Rucina, S. collected the core and processed the pollen in 2009 and Dr Marchant, was the recipient of the field and lab work grant and provided assistance with preparation and editing of the manuscript. A manuscript has been accepted as **Githumbi, E.N**., Mustaphi, C.J.C., Yun, K.J., Muiruri, V., Rucina, S.M., and Marchant, R.M. Wetland trangression and late Holocene vegetation and fire variability in the semi-arid Amboseli Landscape, Southern Kenya. Published in *AMBIO* (January 2018), DOI:10.1007/s13280-018-1014-2 A synthesis manuscript has been accepted as **Esther N. Githumbi**, Rebecca Kariuki, Anna Shoemaker, Colin J. Courtney Mustaphi, Maxmillian J. Chuhilla, Suzi Richer, Paul Lane, Rob Marchant. Pollen, People and Place: Palaeoenvironmental, Archaeological and Ecological Perspectives on Ecosystem change in the Amboseli Landscape, Kenya in *Frontiers in Earth Sciences* (January 2018), https://doi.org/10.3389/feart.2017.00113.

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# CHAPTER 1

# Introduction

## 1.1 Project overview

Rapidly growing East African populations depend on natural resources i.e. rain-fed agriculture, rangeland pastoralism and wildlife tourism for their livelihoods (Verschuren et al. 2011). The diversity of East African ecosystems caused by the interaction of major convergence zones, regional feedbacks and the varied topography highly limit the production levels necessary to sustain the population. Planning for the sustainable long-term use of these ecosystem services requires a historical perspective on the human-environment interactions (Marchant 2010, Marchant et al. 2010, Verschuren et al. 2011). Sedimentary, archaeological and historical archives record information about past environmental variability allowing us to reconstruct and interpret past environmental changes, understand the climate mechanisms and interpret how environmental conditions may have interacted with human agency. Climate and human interactions influence ecosystem state although the ability to differentiate, determine and quantify each components input is a major challenge in understanding ecosystem variability (Jackson & Stephen 2006). With such a range of processes, often operating in unison, it can be difficult to determine 'cause and effect' relationships.

Using a multi-proxy approach we can begin to disentangle natural variation from entangled anthropogenic impacts (Berkes et al. 2000, Gillson & Ekblom 2009, Heckmann 2011). Natural variability can be observed by looking at sedimentary records that document ecosystem changes under different climatic and land use states (Marchant 2010, Willis & Birks 2006). The longterm characterisation of ecosystem changes can be used to identify the drivers and direction of contemporary ecological change for ecosystem restoration (Willard & Cronin 2007). By placing the archaeological and historical records within the palaeoecological context it is possible to pair and examine historical human activity and climate changes with ecosystem changes. Several records from the savannah ecosystem in southern Kenya (Gillson 2006, Rucina et al. 2010), have applied historical knowledge to improve their interpretation of the human impacts on the vegetation as observed from the pollen records.

Understanding ecosystems, how they evolve over time and how the nature of the interaction changes is important in determining baselines for sustainable ecosystem management. Palaeoecological information is now being applied in management and restoration of degraded ecosystems (Seddon et al. 2014). Palaeoecological data can also be applied in a socioecological context to provide insight on ranges of ecosystem variability and their accompanying human land use interactions (Ekblom 2012). The development and implementation of land use policies and practices require an understanding of the controls on the ecosystems we observe today, and an understanding of these changes over multiple temporal (decadal-centennial-millennial) and across spatial (local-regional-global) scales.

Having a retrospective perspective is fundamental in providing observations of change under different conditions thus identifying or establishing thresholds through which ecosystems will change from one state to another (Gillson 2006). This would not be possible when using the recent past or present observational data as the records are too short and do not cover the range of variability. To achieve this aim palaeoecological, archaeological and historical data needs to be generated and synthesised in an applicable form that is easily understood by different user groups making it useful for conservation, restoration, policy and management (Gillson & Marchant 2014).

## 1.2 Thesis aims and objectives

High resolution palaeoenvironmental records from East Africa are relatively rare compared to other regions of the world and are needed to improve our understanding of the evolution of the different landscapes. Current East African landscapes are a result of climate change and anthropogenic utilisation over the ages. To improve our understanding of current ecosystems and the changes that have occurred over time more records improving our understanding of the East African environmental history are necessary. Several East African landscapes have been identified as important biodiversity areas with several ecosystems identified as conservation targets either for their having high numbers of endemic species, having huge numbers of megafauna that live in close proximity to humans and as cultural landscapes mainly due to the high number of archaeological finds documenting human evolution. Major shifts in climate have also been recorded through time such as the African Humid Period (AHP), the regional arid phase $\sim$ 4000 cal yr BP, the Medieval Warm Period (MWP) in the $10^{th}$ to $12^{th}$ century and the Little Ice Age (LIA) in the $14^{th}$ to $19^{th}$ century which have undoubtedly influenced human interactions with the environment.

Large scale human activities such as introduction of domesticated livestock, pastoralism, farming and trade combined with high levels of migration, population growth and large scale development. Although the timing of this events is known in some areas, the impacts on ecosystems are still not fully understood.

The study areas are important as both Mau forest and Amboseli have high biodiversity, contain endemic faunal and flora and they have both sustained human occupation and are now currently experiencing significant change due to increased human intervention and climate change. Complex climate patterns combined with varied topography and changing land use patters dominate the region and through this research project, sedimentary records recovered from a highland forest ecosystem (Eastern Mau) and a lowland savannah ecosystem (Amboseli) will provide insight into the environmental history, human interactions and socioecological context of two important East African ecosystems.

By developing new multi-proxy records of environmental history and synthesising these within existing palaeoecological, archaeological and anthropological records from East Africa, the patterns of environmental variability and the influence of natural ecosystem processes will be better understood. The primary aim of this study is to provide palaeoecological records that reconstruct environmental changes of Mau forest and across the Amboseli ecosystem using a multi-proxy approach. This will be achieved through:

- Developing a narrative of ecosystem development focusing on vegetation change, fire history and biophysical processes at Mau forest by following a high resolution multi-proxy approach; pollen, macro-charcoal and physiochemical records will be analysed.
- Expanding our understanding of vegetation composition, fire history, biophysical processes and human-landscape interaction across the Amboseli through time

and identify if there are differences in the timing of events and the signals between sites within a close distance using multi-site high resolution pollen, macro charcoal and physiochemical records.

- Using the insights about ecosystem history Mau forest and Amboseli, provide a synthesis of Holocene ecosystem change as well as assess how the information can be used to inform debates around current and future management of the study areas and land use management in general.

## 1.3 Layout of thesis

1. Chapter one provides a brief outline of the thesis aims and the rationale for the study.

2. Chapter two details the theoretical context and conceptual framework for understanding the East African environment and the human environment interactions recorded through time by providing a summary of environmental changes and the human history since the late Pleistocene-Holocene transition period.

3. Chapter three presents the results of the multi-proxy palaeoecological approach producing pollen, charcoal, elemental, and stratigraphy data including the radiocarbon dates at Nyabuiyabuyi wetland in Eastern Mau, Kenya since the Last Glacial period.

4. Chapter four describes the semi-arid Amboseli landscape since the mid Holocene presenting Esambu, Kimana, Ormakau and Enkongu with the aim of improving our understanding of Amboseli vegetation change, fire history as well as biophysical processes.

5. Chapter five synthesises the findings of the study and discusses the potential of combining multidisciplinary datasets to develop a comprehensive narrative of human-environment interactions within a landscape and the importance of palaeorecords in land use management.

# CHAPTER 2

## Understanding East African environment-human interactions through the Holocene

## 2.1 Overview

This chapter provides an outline of the information known about the East African environment with respect to climate, vegetation, soils and fires. A detailed summary describes the key changes that have occurred since the Late Glacial period focussing on ecosystems and human environmental interactions from the available palaeoecological, archaeological, historical and ecological studies. The aim of this chapter is to present the context for the thesis and provide an understanding of East African environments.

## 2.2 East African climate

Climate in tropical East Africa is defined by high temporal and spatial variability (Gelorini & Verschuren 2013) influenced by a variety of factors such as air streams and convergence zones, Sea Surface Temperature (SST) variation, monsoonal systems, large inland lakes and topographical variation (Scholz et al. 2007). Virtually all precipitation falling onto easternmost East Africa is derived from the Indian and Atlantic Ocean (Marchant et al. 2007, Tierney et al. 2010), the monsoonal winds transport moisture from the Indian, Pacific and Atlantic Oceans. Thus hydroclimatic patterns of East Africa have influenced the moisture balances of multiple major basins through a complex interplay of Indo-Pacific and Atlantic Ocean influences, alongside the effects of complex topographical features for example through influencing air mass and moisture development and movement (Ogallo et al. 1988, Omeny et al. 2008, Owiti & Zhu 2012).

The Intertropical Convergence Zone (ITCZ), (Gasse 2000, Trauth et al. 2003), El Niño-Southern Oscillation (ENSO), (Currie et al. 2013, Nicholson 2000, Tierney et al. 2015) and the Indian Ocean Dipole (IOD) (Marchant et al. 2007, Maslin et al. 2014, McHugh 2006) are the main influences on the rainfall patterns of the region. The monsoonal rainfall climate of East Africa is closely linked to the meridional passage of the ITCZ that is the primary control on the seasonal distribution of rainfall (Wolff 2011). The ITCZ is a low pressure zone extending from 5°N to 5°S (Figure 2.1) where the North East and South East trade winds converge, forcing warm air to rise and flow towards the poles; it is a major component of the global circulation system. Solar heating in the region forces air to rise through convection which results in precipitation; maximum solar input energy occurs in June in the Northern hemisphere and December in the Southern hemisphere influencing formation and movement of high pressure cells, the ITCZ position and thus the rainfall seasonality. The annual ITCZ migration over East Africa spans the widest latitudinal range about the equator in the world (Verschuren et al. 2009).

Annual passage of the ITCZ results in a bimodal rainfall pattern of two rainy seasons (March to May and October to December) and two dry seasons. Toward the subtropics the two rainy seasons merge. Kenya receives most of the rainfall during the months of March to May (long rains) and September to October (short rains). Southern Tanzania experiences a rainfall peak between December to February (Ogallo et al. 1988, Owiti & Zhu 2012). There is high inter annual and seasonal variability in rainfall resulting from interaction between atmosphere, SST, trade winds and diverse topography. Surface ocean warming intensifies and causes shifts of the ITCZ increasing East African precipitation (Schott et al. 2009, Weijers et al. 2007). The strong impact of topography is clearly reflected by the regional microclimates for example the southern flank of Mount Kenya receives about >2500 mm yr$^{-1}$ rainfall while the northern flank of the mountain is much drier receiving <1500 mm yr$^{-1}$ (Mizuno & Fujita 2014).

The El Niño-Southern Oscillation develops from air-sea interactions in the Pacific Ocean leading to increased SST thus general warming of the Indian Ocean (Currie et al. 2013). It is a fundamental ocean-atmospheric system with atypical warming of Eastern Pacific Ocean referred to as the El Niño while the Southern Oscillation refers to the circulation between the eastern and western Pacific Ocean (Ogallo et al. 1988). It recurs every seven years alternating between its two modal states of El Niño and La Niña (Rouke 2011). Largely through this teleconnection, the region is significantly influenced by ENSO during the October to December period, with positive (El Niño years typically being wetter from upwelling of warm tropical waters and La Niña occurring during the cold phase being atypically dry with onset varying across East Africa (Goddard & Graham 1999, Marchant et al. 2007, McHugh 2006, Ogallo et al. 1988).

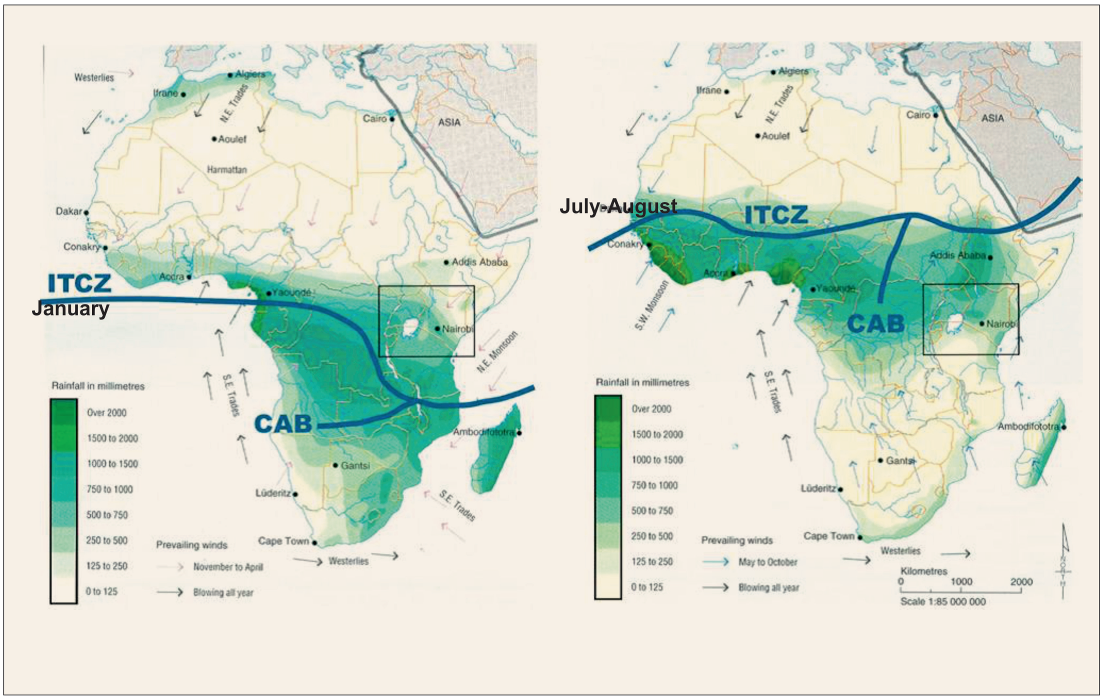



Figure 2.1: Location of the ITCZ and the CAB twice a year.
Modified from :`new.unep.org/dewa/africa/publications/aeo-1/032.htm`

Another important large scale circulation pattern is the Congo Air Boundary (CAB); this is a low pressure convergence zone that marks the confluence of the Indian Ocean air with the unstable Congo Basin air as it moves from west to east (Levin et al. 2009, Mapande & Reason 2005). The west to east hydroclimate pattern observed in tropical Africa relies on the behaviour of the CAB and its interaction with the meridional ITCZ movement (Figure 2.1). This moisture influx from the Atlantic Ocean is responsible for a rainfall peak between June-July-August mainly in Uganda and western Kenya (Barker & Gasse 2003, Bessems 2007, Colombaroli et al. 2014, Ogallo et al. 1988) during what is a dry season for most of Kenya.

In order to try and understand the interaction and influence of these hydroclimatic patterns on the regional climate, climate models have been developed. Climate modeling studies that simulate past, current and future climatic patterns as well as regional climatic models have projected increased precipitation across East Africa under global warming albeit with complex spatial heterogeneity (Endris et al. 2013, Indeje et al. 2000, Rowell et al. 2015). There is a current precipitation decrease trend observed particularly on the reduction of long rains leading to major droughts over the past 20-30 years with the worst drought experienced in the last 60 years experienced in 2011 (Owiti & Zhu 2012). This has been associated with decadal variability of the Pacific SST (Hastenrath 2007, Rowell et al. 2015), however, further studies are needed to improve the climate models.

## 2.3 East African vegetation

Within East Africa the landscape ranges from low-lying flat areas to high mountainous regions. Strong climate gradients, soil type and nutrient availability combine with biotic factors such as grazing type and density, to determine composition, distribution and structure of the prevailing ecosystem (Lehmann et al. 2014). Climate ranges from semi-arid areas having less than 250 mm $yr^{-1}$ of rainfall to wet regions experiencing more than 3500 mm $yr^{-1}$ of rainfall resulting in variation in vegetation across the landscape (Figure 2.2).

A large area of East Africa is covered by semi-arid savannah vegetation (Figure 2.2) which is characterised by a herbaceous layer dominated by grasses with varying densities of trees and shrubs (Knoop & Walker 1985). The lowlands of Kenya, south-eastern Ethiopia and southern Somalia are dominated by different types of savannah vegetation. This varied vegetation type is called the Somalia-Maasai deciduous *Acacia-Commiphora* bushland and thicket (White 1983), it forms a dense bushland of 3 to 5 metres heights with occasional trees up to 9 metres. *Euphorbia* woodlands can be found scattered throughout the savannahs (Lamb et al. 2003). Grasslands, consisting of grasses and occasionally herbs with less than 2 % canopy cover (Pratt et al. 1966) can be found in the same ecological zone as deciduous bushlands but are relatively rare in East Africa. Wooded grasslands are covered with grasses, herbs and woody plants covering between 10 to 40 % of the ground while woodlands are open tree stands at least 8 metres tall with a canopy cover greater than 40 % (Breugel et al. 2011, Pratt et al. 1966). To the south in Tanzania, and spreading farther southeast, Zambezian Miombo woodlands can be found that are dominated by *Brachystegia*, a

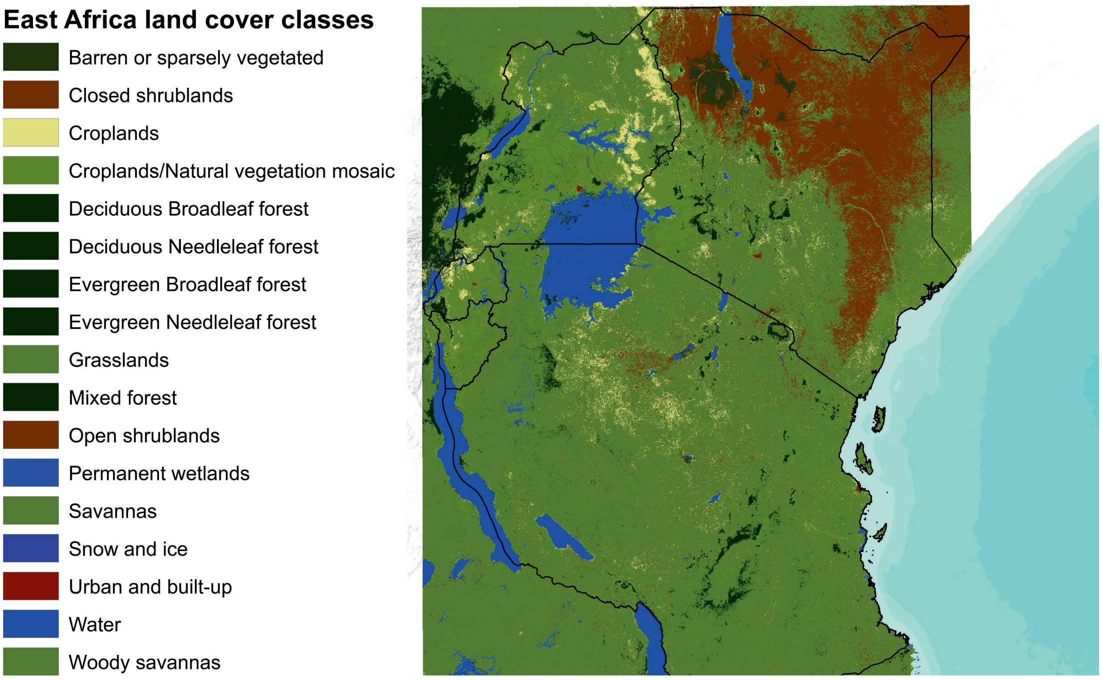



Figure 2.2: East African major vegetation cover types from MODIS12Q1 product type 1 IGBP (500m resolution)

10 to 20 metres tall tree that forms a nearly closed canopy, although the sub canopy remains relatively open due to the foliage of these trees. The drier woodlands have a height of less than 15 metres and feature a poorer floristic diversity (Desanker & Prentice 1994). This vegetation type is common in the humid to dry sub humid regions (Pratt et al. 1966). Woodlands have been in decline in several eastern, southern and central African areas (Dublin et al. 1990). Semi-desert and desert vegetation is present in northern Kenya and southern Ethiopia where highly drought tolerant taxa such as *Acacia, Aloe, Commiphora* and *Euphorbia* can be found.

Forests mainly occur between 1500 m and 2500 m (Kindt et al. 2011), montane forests can be found throughout East Africa on the uplands above 1500 m, with Afromontane evergreen forest dominated by *Juniperus, Olea* and *Podocarpus* (DeBusk 1997, Lamb et al. 2003). Afromontane forests can be classified as differentiated or undifferentiated depending on the presence of *Juniperus* and *Podocarpus* (Kindt et al. 2011). In places, such as the uplands around Lake Bogoria (Kenya), the forests are severely degraded (Vincens 1986) and occur as patches between montane grasslands e.g. around Lake Malawi (DeBusk 1997). The Ericaceous belt can be found above 3000 m, dominated by *Erica* and *Philipia* (Hamilton 1982, Herrick & Beh 2015, Lamb et al. 2003) with low-lying vegetation and sporadic and sparse cover into the alpine elevations dominated by *Seneccio*. East African forests mainly located in upland areas are the upper catchments of the main rivers that support key economic sectors such as hydropower generation, agriculture, tourism amongst others. However as demand for timber has led to widespread rapid deforestation (Carpenter et al. 2005), only around 2 % of East Africa is still covered by indigenous forest (Bussman 2001). Kenya's main forests form the upper catchments of all the main rivers and in total cover one million ha (Akotsi et al. 2006). The forests are undergoing deforestation at different rates with the Mau Forest Complex (MFC) experiencing the most rapid loss of indigenous forest cover where between 2000 and 2003, $\sim$6000 ha was lost (Akotsi et al. 2006). East African states have now drafted policies on management to monitor the health and well-being of marginal and threatened vegetation classes.

## 2.4 East African soils

East Africa has an extremely wide range of soil properties (Figure 2.3) that vary depending on the parent material (Maskall & Thornton 1996), climate and management inputs (Heitkamp et al. 2014, Kisinyo et al. 2005). The soil type has a strong influence on the vegetation and land use potential of landscapes. However nutrient mining which refers to loss of nutrients due to harvesting of agricultural crops in East Africa is among the highest in the sub-Saharan region which coupled with decreasing food production is a major source of concern Bekunda et al. (2004). This is highly important in East Africa considering that 98% of the calories consumed in Africa comes from the soil (Jones et al. 2013).

In the semi-arid to arid savannah zone the major soils are yermosols, xerosols, lithosols, regosols, solonetz which are salty soils with a high level of sodium (Tabor 2001) and solonchaks which are salty soils with little horizon development. Such soils are mainly sandy, rocky and calcareous or siliceous, with salt and gypsum deposits

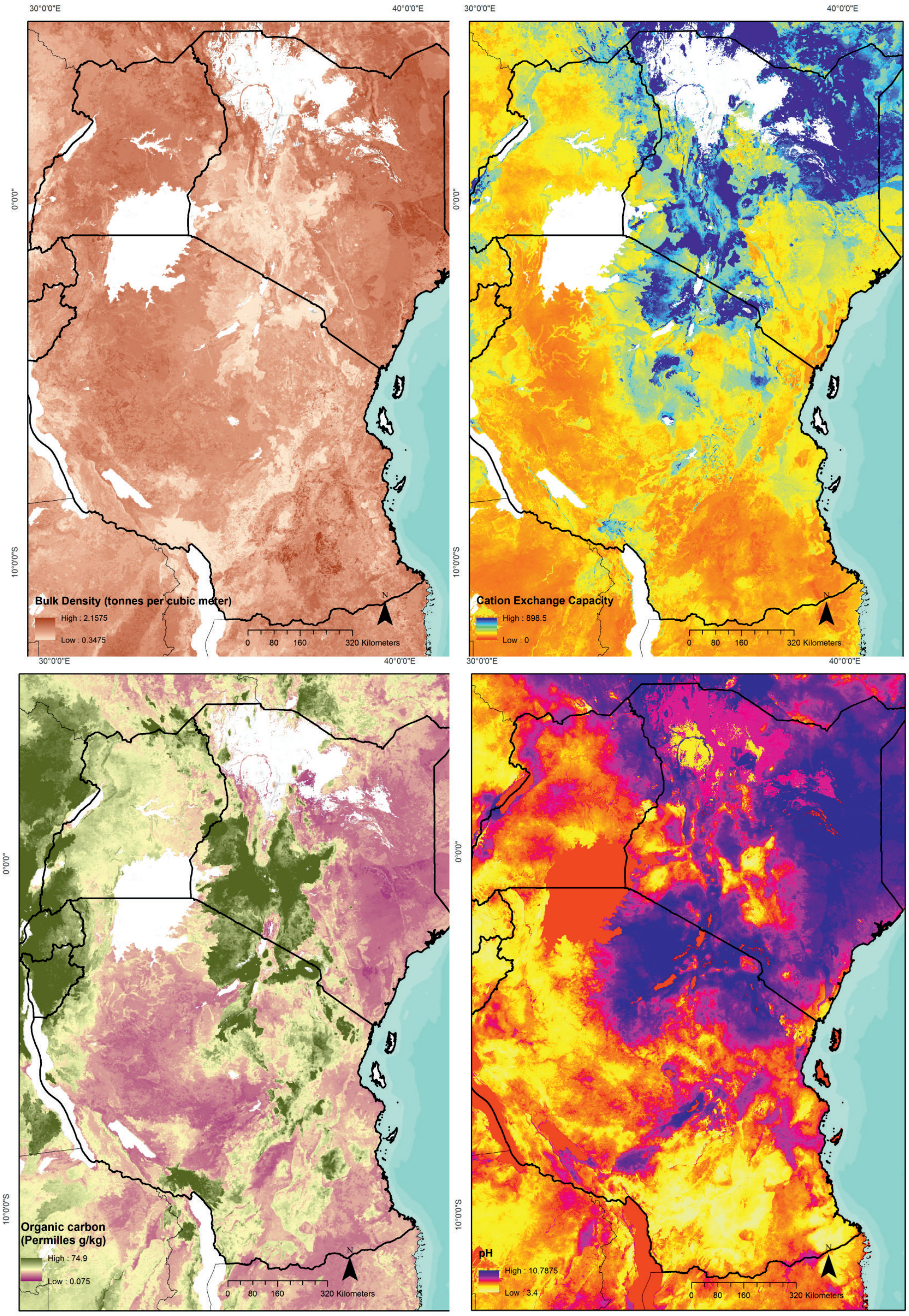



Figure 2.3: Extrapolated soil organic carbon, pH, bulk density and cation exchange capacity values calculated from 1950-2025 at 1km resolution. Data from the African soil database, November 2014.

occurring extensively. They are generally, weakly developed, low in fertility and very susceptible to wind erosion.

The soils of the semi-humid wooded to semi-arid savannah zone where the average rainfall ranges from 200 to 800 mm $yr^{-1}$ on level to near-level topography with isolated low-lying hills (inselbergs) with mainly short grass savannah are xerosols, luvisols, cambisols, arenosols, rendzinas and lithosols (Jones et al. 2013, Tabor 2001). Xerosols are for the most part shallow and difficult to use for crop production because of their stony and/or petrocalcic nature overlaying the deeper, medium-textured soils and associated alluvial soils can be used for the cultivation of cereals. The most important constraint which limits the ability of the soils of the zone to produce high yields of crops is moisture stress (Jones et al. 2013). Other restrictions include high susceptibility to wind erosion if vegetation has not developed, low fertility, salinity and alkalinity, deficiencies of iron and zinc, as well as the constraints listed above for similar groups occurring in the sub-humid wooded savannah zone.

The major soils of the humid to sub-humid wooded savannah zone are generally found on level to very gently undulating land interspersed with low-lying hills (inselbergs) and with a rainfall of 650 to 1400 mm $yr^{-1}$ under predominantly grassland vegetation. The soils are luvisols, ferralsols, arenosols, acrisols, nitisols, cambisols and lithosols. The upland soils are mostly well to moderately well drained, gravelly with a light-textured matrix which in some areas overlies an iron pan developed in situ at shallow depth. The organic matter content and fertility status of the soils within the zone are quite low compared with soils of the very humid to humid forest zone. These soils are used for the production of sorghum, maize, yam, millet and various leguminous crops including groundnuts, beans and cowpeas. Fluvisols, gleysols, and vertisols are all clay rich mostly non-gravelly with a reasonably moderate to high fertility status. Crops that can be successfully grown on these soils include rice, sugarcane and vegetables. They occur in low lying periodically flooded areas or areas where ground water comes close to the surface (Jones et al. 2013). Where vertisols and other lowland clayey soils are extensive, the main constraints are due to the very high clay content causing tillage difficulties, caused by stickiness when wet and hardness when dry, slow permeability, slow internal and external drainage, seasonal water-logging, flood risk and in certain cases, salinity and alkalinity.

Soils of the very humid to humid forest zone where the rainfall exceeds a mean of 1500 mm $yr^{-1}$ are ferralsols, nitosols, acrisols, arenosols, cambisols, and lithosols (ISRIC - World Soil Information 2015). Ferralsols, anitosols, acrisols and arenosols are very extensively developed and for the most part deeply weathered and gravelly, mostly acid to very acid in reaction but with a considerable amount of organic matter built up under natural conditions due to abundant leaf-fall and rapid decomposition. Cambisols are young soils lacking distinct soil horizons. Andosols which develop from ejected volcanic material are found along the Eastern Africa Rift Valley around old or active volcanoes such as Mt. Kenya and Mt. Kilimanjaro from porphyric rocks (Bush et al. 1998, Schüler 2012). Andosols develop a thick dark top soil through fixing of organic substances from weathered clay minerals by aluminium however they suffer from a strong phosphorous fixation due to high aluminium and iron levels (Jones et al. 2013). They tend to have high levels of organic matter and their as observed in figure 2.3 around the Rift Valley and water bodies.

## 2.5 Fires in East Africa

Fire is a global ecosystem process that influences ecosystem composition and structure (Whitlock et al. 2010). Africa is often referred to as the ”fire continent” owing to the regular and widespread occurrence of wild fires, and is characterised by $\sim$70% global burned area and 50% fire related carbon emissions (Andela & van der Werf 2014). Fire is an important driver of biogeographic change in transitional ecotonal environments in East Africa such as between the tropical forest and savannah (Colombaroli et al. 2014, Dublin et al. 1990).

It is suggested that in the absence of fire, woodlands or closed forests would cover most of the savannah area (Bond et al. 2005, Dublin et al. 1990). East African burning regimes are dependent on climatic and vegetation regimes as well as anthropogenic activity (Colombaroli et al. 2014, Tarimo et al. 2015, Thevenon et al. 2003). There is spatial heterogeneity (Figure 2.4) in the occurrence and frequency of fire which is dependent on hydroclimatic conditions and biomass availability (Colombaroli et al. 2014, Van Der Werf et al. 2008). In sub-humid regions such as the tropical forests where precipitation is high all year round, fire occurrence and frequency is limited by the moist conditions while in semi-arid and desert regions the occurrence and spread of fire is limited by biomass accumulated from previous rainfall seasons or complete lack of fuels in the case of deserts. Savannah woodland areas such as the Tanzanian Miombo experience regular large fires where 50% of the woodland area is affected annually (Tarimo et al. 2015).

Although lightning was one of the primary ignition sources of fires (Raison 1979), most fires today burn in fire-adapted ecosystems as a result of human ignition (Archibald et al. 2012). More than 100 million hectares of savannahs and grasslands burn annually in Africa south of the equator for comparison (FAO 2001). The most intense burning is concentrated in the subtropical belt which includes Angola, the southern Congo, Zambia, northern Mozambique and southern United Republic of Tanzania. During the year 2000 fire season, the area burn south of the equator may have reached more than 200 million hectares (FAO 2001).

Fire has been used as a land management tool for landscape transformation in East Africa (Colombaroli et al. 2014, Rucina et al. 2010, Tarimo et al. 2015, Thevenon et al. 2003). Fires were set and continue to be set on some communities to stimulate new and more nutritious pasture for livestock, extraction of resources such as harvesting honey/charcoal burning,hunting and for harvesting honey; fires were also set to kill pests, open up land to increase visibility, increase access to land for grazing, cultivation and settlement. However this traditional use of fires has had increasingly degrading consequences for example on Mt. Kenya, fires that occur in the highly flammable *Pinus* plantations damage the margins of the indigenous forest (Thijs et al. 2014). Fire has also been used as a violent political tool, for example there was deliberate application of fire during violence following the December 2007 Kenya general election in the Rift Valley (Anderson & Lochery 2008). On Mt. Kilimanjaro, recent fire activity has caused the upper timberline to decrease in elevation because reduced precipitation has changed the ecological conditions at those elevations (Hemp 2005). Evidence of fire has been detected from the Sacred lake sediment core from the LGM (Wooller et al.

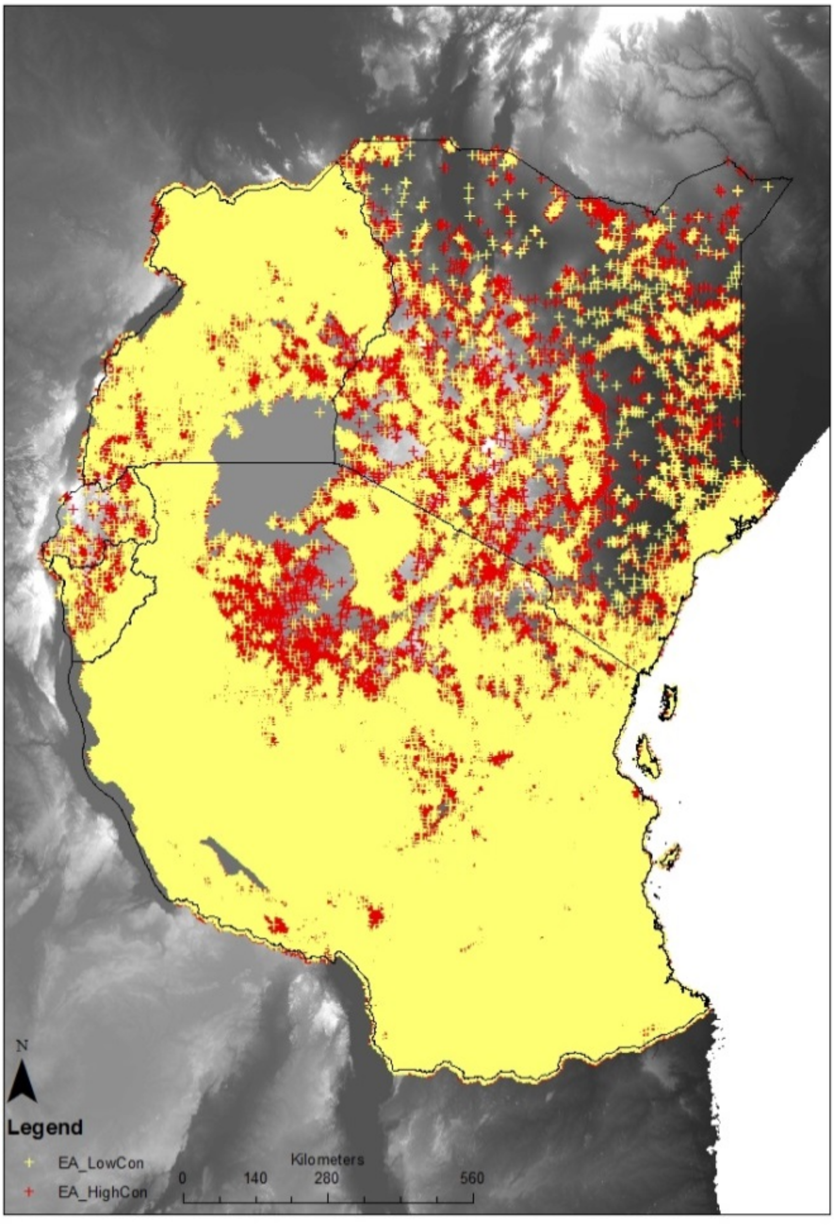



Figure 2.4: Fire occurrence in East Africa, red crosses indicate <50% confidence of fire occurrence while yellow crosses indicate >50% of fire occurrence between 2000 and 2012 based on satellite data (MODIS12Q1).

2000). Global biomass burning increased rapidly in the early Holocene, varied in the mid and late Holocene i.e. stable between 4000 and 3000 cal yr BP, increased from 3000 to 2000 cal yr BP and then declined from 2000 cal yr BP before a sharp increase during the industrial era which began to decline 50 cal yr BP (Marlon et al. 2013). East African palaeorecords have not been consolidated to provide a similar palaeofire summary.

The role of anthropogenic fire regimes in driving ecological and social systems is still under study as its continuous use has substantial current and future implications (Bowman et al. 2011). Studies of current use of fire are producing narratives that clarify the differences in fire management between sex, across seasons and through time in different ethnic communities (Kamau & Medley 2014), a study carried out in Chyulu Hills, Kenya in 2012 exploring how people use fire and comparing the perceptions of use of fire across the communities identified more than 22 uses of fire (Kamau & Medley 2014). Probability of fire occurrence across East Africa has high spatial heterogeneity (Figure 2.4), and most fires occur during the fire season which occurs between February to March and September to October. In Kenya, most protected forest e.g. Eastern Mau and savannah ecosystems e.g. Amboseli/ Tsavo National park declare February to March as the fire-free season forbidding any use of fire near their boundaries.

## 2.6 East African ecosystem change

The Last Glacial Maximum (LGM) is the last period with the greatest extension of ice sheets and its ending varies globally (Björck & Wohlfarth 2001). The period after the LGM referred to as the Late Glacial period is cool and dry up until the Younger Dryas. Despite evidence of climatic variability through the Holocene (Kiage & Liu 2009*b*, Tierney & Russell 2007, Trauth et al. 2003) and complex societal changes (Herskovits 1926, Mace 1993, Mercader et al. 2013, Waller 1985), East Africa still has relatively few multi-proxy records of environmental change (Figure 2.5) compared to other regions of the world. East African palaeorecords focus mainly on lake and wetland sediment, there are more wetland records than lake records as there is a higher number of wetlands dispersed across the landscape especially in the arid and semi-arid regions of East Africa (Ashley et al. 2004).

The sediment contains a range of proxies (pollen, charcoal, fungal spores, diatoms etc.) which when used together improve our understanding of different aspects of environmental history by increasing our accuracy in reconstructing different components of the environment. A wide range of palaeoecological proxies have proven useful in describing the environmental history for a range of different environments and have provided information on ecosystem composition and distribution (Gillson & Ekblom 2009, Jackson & Stephen 2006, Norström et al. 2012), human-land use interactions (Bussman 2001, Heckmann 2014, Lézine et al. 2013, Walsh et al. 2010), climatic changes (Bessems 2007, Gasse 2000, Nicholson 2001, Roubeix et al. 2014), erosion (Heckmann 2011, Kiage & Liu 2009*b*), sedimentation (McKee et al. 2005, Owen et al. 2009) and fires (Colombaroli et al. 2014, Thevenon et al. 2003, Tovar et al. 2014).

Wetlands accumulate deposits over thousands of years and sediment cores can be analysed for a range of microfossils such as pollen, phytoliths, fungal spores, charcoal, algae and inorganic components. Numerous studies have analysed wetland sediment to describe climate change, human-environment interactions and significant historical events recorded in the wetland stratigraphy. One of the main palaeoecological techniques used is palynology; this is the study of organic walled microfossils derived from flowering plants that can be identified and used to understand change in the presence of plant species within the surrounding catchment (Birks 1993). Palaeoecological records in East Africa have provided evidence of ecological variability, climate change and human activity.

Our understanding of East African palaeoecology varies with some areas having more information available such as the Rift Valley lakes. Some ecosystems, such as the Eastern Arc Mountains in Tanzania, have been described as relatively stable environments at certain time periods with relatively minimal vegetation shifts through observation of pollen at high resolution from ~50,000 years ago and the persistence of the forest indicating a mesic environment which begins to open up into an a dry forest type due to the regional aridity when the buffering ability was exceeded (Finch et al.

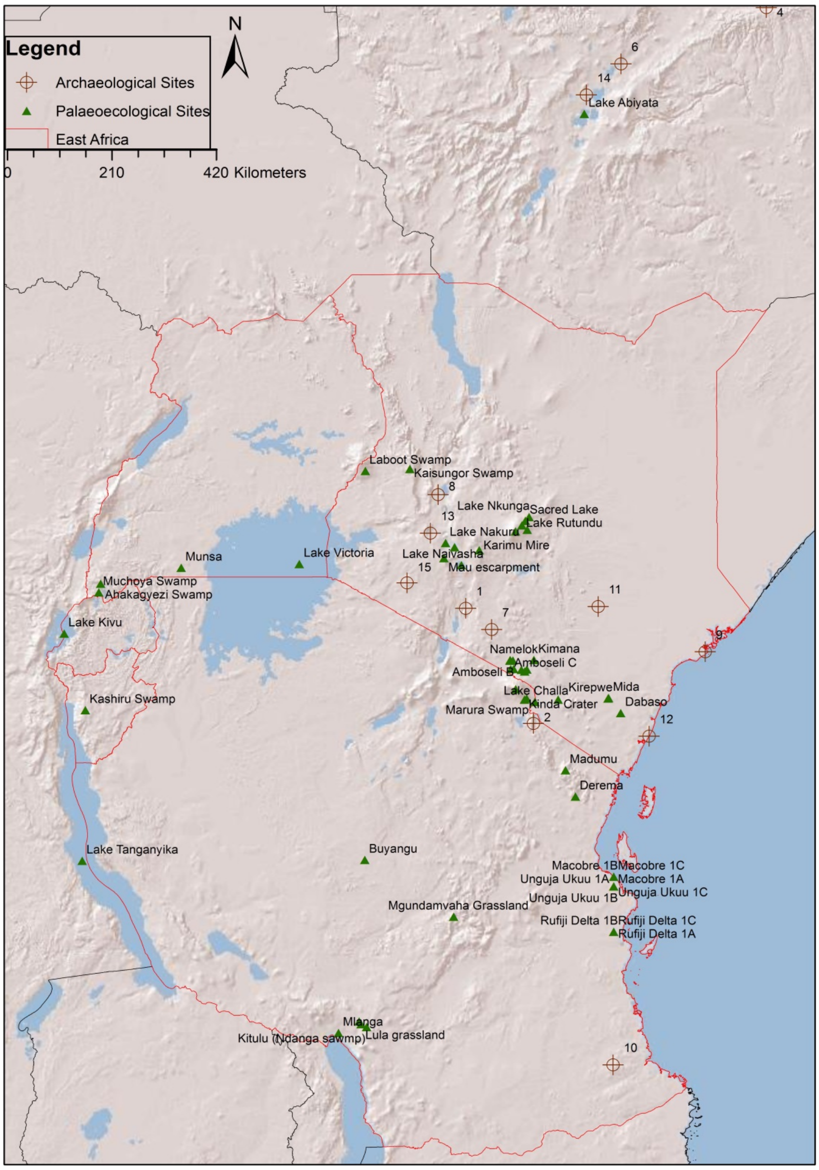



Figure 2.5: Examples of East African archaeological and palaeoecological sites in the York Institute for Tropical Ecosystems database; archaeological sites are marked by a cross within a circle while palaeoecological sites are marked by green triangles. Data points sourced from available published literature and plotted using ArcMap10.2.

2009, 2016, Marchant et al. 2007, Mumbi et al. 2008). East Africa experienced dramatic and sometime rapid hydrological fluctuations; these changes will be discussed within three broad time periods; the Late Glacial Period, the Holocene and the last 2000 years.

### 2.6.1 Late Glacial period to Early Holocene (∼16,000-∼11,700 cal yr BP)

The Last Glacial period was characterised by significantly cooler, more arid conditions and lower atmospheric carbon dioxide concentrations than present; Lake Malawi which is the southernmost of the largest East African Rift lakes was 3.5°C cooler inferred from the $TEX_{86}$ proxy record and a lake level lowering of ∼100m inferred from comparing seismic profiles to geochemical and biostratigraphic profiles to map out fluctuations and increased $c_4$ vegetation was observed during periods of aridity (Castañeda et al. 2009, Powers et al. 2005, Scholz et al. 2007). Diatom analysis from Lake Rukwa infer low lake levels (Barker & Gasse 2003) until ∼16200 cal yr BP but not total desiccation unlike Lake Victoria which was desiccated until 15,000 cal yr BP from ∼20,000 cal yr BP (Gasse 2000). Lake Kivu and Tanganyika records indicate high sedimentation rate implying smaller, shallower lakes between 14,000 and 9500 cal yr BP (Haberyan & Hecky 1987). Geochemical and palynological evidence from the Lake Albert (Uganda) records infer arid conditions/prolonged desiccation and formation of a palaeosol with poor pollen preservation during the Late Glacial period. However speleothem records from the nearby Matupi cave indicate a wooded grassland around 14,000 cal yr BP (Beuning et al. 1997). The Rumuiku swamp record from Mt. Kenya records development of a montane forest during the LGM that underwent significant reorganisation reflecting the change from a cool, moist climate dominated by Ericaceous and upper montane taxa around 14,000 cal yr BP followed by a reduction in several dominant taxa such as *Junipers, Olea* and *Podocarpus* during the Younger Dryas (Rucina et al. 2009). A record from the Bale Mountains in Ethiopia indicates a grass dominated steppe like vegetation cover with shrubs (Amaranthaceae-Chenopodiaceae) inferring an arid climate characterised by the low pollen influx indicative of sparse vegetation cover from ∼16,700 cal yr BP to ∼13,400 cal yr BP.

There was a sharp decrease of shrub taxa and increased amounts of Cyperaceae (Umer et al. 2007) at a time corresponding to the Younger Dryas around 13,400 to 11,200 cal yr BP in the Bale record from Ethiopia. Sacred Lake and Lake Nkunga on Mt. Kenya record low lake levels and desiccation between ∼13,000 yr BP and 10,500 yr BP (Olago et al. 1999). The Lake Masoko pollen record does not record the Younger Dryas lake level drop, instead it indicates increased arboreal taxa between 16,000 and 11,700 cal yr BP derived from a semi-deciduous forested environment with low grass and herb taxa frequency. This abruptly changed to taxa indicating tolerance to long dry seasons (deciduous woodland), as well as more open vegetation cover while the magnetic susceptibility data indicate strong seasonal fluctuations and a high stand (Garcin et al. 2007, Vincens et al. 2007). After ∼13,000 cal yr BP there was an increase in tree cover across East Africa with many taxa reaching maximum abundances and diversities from 12,100 cal yr BP mirroring the Lake Rukwa pollen sequence (Vincens et al. 2005). In upland areas montane forests expanded to higher and lower elevations (Taylor 1990) whereas in the lowlands there was expansion of wetter *Miombo* woodlands.

### 2.6.2 The Early to Late Holocene (∼11,700 to ∼2000 cal yr BP)

Across northern and eastern Africa, the Younger Dryas period was followed by rapid regional warming, lake basins refilling (Gasse 2000) and vegetation of the Sahel becoming more verdant (Hoelzmann et al. 1998). The early to mid-Holocene is characterised by increased temperature and precipitation compared to the Younger Dryas. The Sahel supported several perennial lakes (Berke et al. 2012, Gasse 2000, Renssen et al. 2006), the Makdgadikgadi basin experienced a mega basin event (Burrough et al. 2009), when the large inter-connected lacustrine basins in the Kalahari were filled and joined to form a mega basin which at its largest extent encompassed 66, 000 $km^2$. There was also expansion of humid vegetation types (Olago 2001), with arboreal pollen replacing non-arboreal pollen across East Africa pollen records.

A pollen derived rainfall and temperature reconstruction from a Lake Tanganyika sediment core indicates synchronous increase in temperature and precipitation from

12,700 cal yr BP onwards (Vincens et al. 1993). while pollen records indicate extensive Afromontane forests as well as expansion of the Ericaceous belt in the Ethiopian Bale Mountains indicating warmer wetter conditions (Umer et al. 2007). A diatom record derived from Lake Victoria's Pilkington Bay and sampled at ∼25 years, reveals an early Holocene humid phase with maximum rainfall and lake mixing from c. 10,200-9500 cal yr BP (Stager et al. 2003). During this early Holocene humid phase, diatom data from Lakes Kivu and Tanganyika imply increased lake levels (Haberyan & Hecky 1987) from 10000 cal yr BP; the Kivu drainage opened at 9500 cal yr BP and overflowed toward Lake Tanganyika (Gasse 2000). The Suguta Basin in the northern Kenya Rift Valley until ca. 6700 cal yr BP, held a 300 m-deep lake which overflowed northward into Lake Turkana (Garcin et al. 2012, Junginger et al. 2014). Significantly wetter conditions have also been inferred for southern East African sites such as Lake Tanganyika (Tierney et al. 2008) and Lake Rukwa which experiences a humid lacustrine optimum (Barker et al. 2002, Thevenon et al. 2002, Vincens et al. 2005). High burning frequency was inferred in sub Saharan Africa during the early Holocene and the speleothem records imply sufficient effective moisture trends (Marlon et al. 2013) were analysed together with the charcoal records.

The termination of the Holocene humid period was attributed to a strengthening of the African monsoon due to gradual orbital increases in summer season insolation (Huber et al. 2004, Stager et al. 2003, Whitlock et al. 2010) which caused a low pressure area over North Africa leading to inland cyclonic flow of moist winds and precipitation coupled with vegetation-albedo feedback (Claussen et al. 1999) and increased surface ocean temperatures ended the rainfall maximum. There was reduced duration and intensity of precipitation with the transition occurring at different time periods from 8000 to 7000 cal yr BP (Stager et al. 2003), and as late as 5000 cal yr BP with a record from Lake Turkana identifying the termination of the African Humid Period around c. 5000 cal yr BP (Berke et al. 2012).

A mangrove record of sea-level change from the east coast of Africa indicates sea-level rise and fall during the Holocene; changes in mangrove abundance indicate an increase in sea-level around ∼7000 cal yr BP from the high occurrence and inward migration of *Rhizophora mucronata*, during the mid-Holocene there is a decrease in *R. mucronata* and increase in *Sonneratia alba* indicating a lower sea level (Punwong

2013). Significant environmental changes have occurred since c.6000 cal yr BP in East Africa characterised by pronounced decadal and century-scale rainfall variability superimposed on smaller inter-annual fluctuations (Wolff et al. 2011). Changes in incoming solar radiation induced changes in latitudinal thermal gradients and seasonality that affected the characteristics of the African monsoon during the second half of the Holocene (Hély et al. 2009, Tierney et al. 2011). Precipitation reduced across the landscape and this led to a reduction in humid vegetation types and an increase in grassland taxa; for example a steady increase in $^{13}C$ values from 6500 cal yr BP indicates opening up of the local swamp vegetation in Deva-Deva Swamp on the Lukwangule Plateau Tanzania (Finch et al. 2009). Lake levels began to decrease after 6000 cal yr BP particularly falling from 4000 cal yr BP between 10°and 22.5°N (Pachur & Hoelzmann 1991) ending what was a favourable period in the Sahara for human inhabitation. Rainfall and inferred lake levels decreased and the early Holocene wetter montane forest components were replaced by taxa that were more drought resistant (Herrick & Beh 2015).

In Tigray (northern Ethiopia), the climate became dryer after 5500 and 4000 cal yr BP (Dramis et al. 2003) with additional evidence of a shift to a relatively dry period apparent after 5000 cal yr BP from geomorphologic evidence (Moeyersons et al. 1999). Lake Tilo in Ethiopia experienced a low stand at c 4500 cal yr BP responding to reduced precipitation levels (Telford & Lamb 1999), rainfall level and inferred lake level declined after 5800 cal yr BP from the Pilkington Bay record (Stager et al. 2003). Local vegetation changes often reflect the regional trends; grasslands expanded and replaced wooded areas in numerous regions. Rumuiku swamp on Mt. Kenya, records the climate becoming warmer from 6500 cal yr BP to 4000 cal yr BP (Rucina et al. 2009) as evidenced by increases of *Macaranga* Poaceae, *Podocarpus* and *Polyscias*, and significant change in the fire regime (increased large fires). Forests around Lake Naivasha became less extensive as grassland expanded from 6000 cal yr BP until it reached a similar distribution to that of today around 4000 cal yr BP (Maitima 1991). Small gradual increases in dry Afromontane taxa were recorded around 6500 cal yr BP accompanied by reduction of *Erica* around 4500 cal yr BP (Umer et al. 2007). During the transition to the arid Late Holocene (Marchant & Hooghiemstra 2004), hydrothermal activity at Lake Kivu (c. 5000 cal yr BP) increased its salinity indicating a drier climate than previously (Haberyan & Hecky 1987). Lake Turkana water levels decreased significantly from 5000 cal yr BP and a closed basin status was achieved around 4200 cal yr BP (Garcin et al. 2012).

Around c. 5000 cal yr BP different sites across East Africa show an increase in abundance of Poaceae at the expense of arboreal taxa in the lowlands, coupled with increases in montane forest taxa indicative of drier climatic conditions, such as *Juniperus, Podocarpus* and *Olea* (Olago et al. 1999, Taylor 1990, Vincens 1986, Vincens et al. 2005) as well as the retreat of the Ericaceous belt in the Ethiopian highlands (Umer et al. 2007). Pollen data from Lake Victoria show an increase in Poaceae pollen starting from 4000 cal yr BP (Kendall 1969) and isotopic data at Lake Malawi show an increase in $C_4$ vegetation initiated again around 4000 cal yr BP (Castañeda et al. 2009). A significant change in climate is centred around 4000 cal yr BP and is most strongly recorded within the South American and African tropics (Marchant & Hooghiemstra 2004) and in East Africa it manifests as lake levels start decreasing (Haberyan & Hecky 1987, Umer et al. 2007) with the closing of the Lake Kivu basin c. 3500-1500 cal yr BP and Lake Rukwa experienced a low stand at c.3000 cal yr BP (Thevenon et al. 2002). Pollen records indicate decline in *Podocarpus* in the Ethiopian highlands (Umer et al. 2007). Increased temperatures and a decrease in amount and frequency of precipitation led to the drying up of the Sahara to its present state as a desert and this regional aridity was also experienced further south. For example at Lake Masoko, Tanzania the pollen record shows the continued presence of wetter Zambezian woodland, albeit, in association with abundant grasses, until 1650 cal yr BP when it was gradually replaced more open vegetation cover (Vincens et al. 2003).

The mid Holocene dry phase is documented clearly in swamps in southwest Uganda which show an increase in grass pollen, sediment $\delta^{13}$C values and charcoal after 3800 cal yr BP, and a hiatus in the Enkapune ya Muto record until 3300 cal yr BP. Sedge and grass pollen begun increasing c. 2700 cal yr BP indicating a shallowing trend from the increased sedimentation and grassland expansion at Pilkingtons Bay, Lake Victoria (Stager et al. 2003). In the Rukiga Highlands of Uganda, forests began to significantly decrease c. 2150 cal yr BP (Taylor 1990). This vegetation change may be the result of human disturbance since other sites in the region have shown intense human presence around this time. Selective declines over the past 2000 years of favoured timber species such as *Caesaria, Celtis, Podocarpus, Prunus* and *Psychotria* in the Eastern Arc Mountains of Tanzania represents a strong indicator of human impact (Finch et al. 2009, 2016, Mumbi et al. 2008). A low lake level stand was noted at c. 2700-2500 cal yr BP and 1200-600 cal yr BP at Lake Victoria (Stager et al. 2003) while temperature records (Scott et al. 2003, Thompson 2002, Tierney et al. 2008) show temperatures dropped at a steady rate between 1450 and 1000 cal yr BP from a $TEX_{86}$ Lake Victoria record.

### 2.6.3 The last 2000 years

The last 2000 years are characterised by warming across East Africa (Nicholson et al. 2013). A record from the Ethiopian Highlands (Umer et al. 2007) recorded a decline in *Podocarpus* with increased *Juniperus* which was represents continued aridity or increased preferential logging from 2000 cal yr BP. Farther south in Uganda, the Lake Albert record showed an expansion of grassland communities from the previous semi-deciduous composition, this signal could also be due to aridity or increased anthropogenic influence (Beuning et al. 1997). A brief warming around 1350 to 950 cal yr BP is followed by brief cooling, and then an abrupt temperature increase starts an extended warm period from 850 to 550 cal yr BP which partly overlaps with, but appears to slightly lag, the Medieval Climate Anomaly (MCA) of temperate latitudes (Graham et al. 2011, Nicholson et al. 2013).

A synthesis of temperature data mainly $TEX_{86}$ Lake Malawi and Lake Tnganyika indicates that temperatures fell abruptly by 1°C ∼500 cal yr BP, with more variable temperatures extending to the late 19$^{th}$ and early 20$^{th}$ century. A marked warming since 100 cal yr BP is apparent, suggesting that modern temperatures are the warmest of the last 1500 years with ”Lake surface temperatures have risen from 23°C to 25.8°C in the last two centuries, a change that is distinctly anomalous in comparison with any period over the last 1500 years” (Nicholson et al. 2013). A record from Lake Baringo documents the dry environment in the region from 300 cal yr BP that led to the complete drying of the lake at 300 and 230 cal yr BP, punctuated by dry and wet episodes at centennial-decadal scales. It provides evidence that land degradation began before the colonial period through pollen, fungal spores and microscopic charcoal analysis (Kiage & Liu 2009*b*).

Human impact is picked up in the records approximately 300 years ago with the introduction of typical crop pollen such as *Zea mays* (Finch et al. 2016, Lamb et al. 2003, Lejju et al. 2005). A desiccation event is observed from three lakes situated on both sides of the East African plateau in Kenya (Lake Baringo) and Uganda (Lakes Kanyamukali and Chibwera) (Bessems et al. 2008); lithologic evidence and magnetic susceptibility indicate a prolonged arid period ~200 cal yr BP across the East African plateau. Several swamps such as Loboi swamp in central Kenya (Ashley et al. 2004) and Ziwani swamp in Tsavo in southern Kenya (Gillson 2006) also record this event during the late 18$^{th}$ and early 19$^{th}$ century with significant changes in pollen, diatom and lithological composition. Gillson (2006) utilised charcoal and pollen from swamp sediments to identify major disturbance events and was able to identify a major disturbance event from the increased Cyperaceae:Poaceae ratio, increased charcoal concentration as well as increased gravel and carbonate content slightly later than the vegetation and charcoal changes. This arid period was followed by a high stand in the Kenya Rift Valley lakes (Ase et al. 1986, Johnson & Malala 2009), equatorial and southern parts of East Africa saw their highest lake levels of the past century in the mid-1960s (Nicholson 1999, Sene & Plinston 1994, Sutcliffe & Parks 1999), during a time when Ethiopia, remarkably, experienced a major drought-induced famine (Devereux 2000). Over the past few decades, climatic warming has increased as recorded by Indian Ocean SST leading to drier conditions during the March to June rainy season in East Africa with significant implications for food production and security (Funk et al. 2008).

Currently, global sea levels are increasing; however, regional sea level changes are poorly constrained in the Indian Ocean and further studies are required. Along the Kenyan coast, tide gauge stations show decreasing sea levels in the western Equatorial portion of the Indian Ocean since 1961-2008 (Han et al. 2010). On the Indian Ocean coast of Kenya, mangrove forests that had existed since at least the mid Holocene with minor distribution changes driven by minor sea-level fluctuations have recently been reduced in extent for paddy field cultivation at small scale (Punwong 2013). The east coast of Kenya and Tanzania may see little or no sea-level rise under projected climate change (Han et al. 2010); although, atmospheric warming may have significant impacts on circulation. This could have important implications for East Africa precipitation variability through ocean-terrestrial hydroclimatic connectivity (Giannini et al. 2003, Hoerling et al. 2006, Marchant et al. 2007) and has crucial implications for the future of development and water accessibility in East Africa.

## 2.7 East African human-environment interactions

Human-environment interactions have occurred throughout human history as humans use the environment to meet their daily needs (Maslin et al. 2014). The interaction of natural and anthropogenic factors on the landscape is compounded by temporal and spatial scales making it challenging to decipher the specific input of each of the driving factors (Lane 2011). Human land use activities have transformed landscapes since the late Pleistocene and this long term interaction has led to evolution of anthropogenic landscapes (Ellis et al. 2013) that dominate East Africa today. East Africa has been populated by hunter-gatherer-fishers (~10,000 cal yr BP), pastoralists (~6000 cal yr BP) and agriculturists, ranging from highly mobile communities to sedentary forest dwellers who respond and adapt to significant climate driven dynamics making the distinction between cultural and natural landscapes difficult (Gelorini & Verschuren 2013).

The diversity of climate, landscapes and cultural practises means that resource exploitation in East African culture cannot be described as purely pastoral, foraging or cultivation (Wright 2007). This is evidenced by the diversity of tools, ceramics, faunal and floral remains recovered from various sites (Wright 2007). The resources utilised and cultures practised by prehistoric people can be evidence of climatic and ecological conditions however evidence of human subsistence is limited by sample size, methods, lack of comparative data and poor preservation of sites due to ecological conditions that degrade remains, our understanding of human technological development is thus limited to samples discovered and analysed (Ambrose 1998, Mercader et al. 2013, Steele 2012, Steele & Klein 2009). The links between societies and sustainable use of landscapes is crucial as the current ecosystems are now a consequence of human management as they are of climate change; human-environmental narratives are informed by a combination of the study of humans as well as the environment. For example a review of zooarchaeological evidence from North Africa by Steele (2012) focusses on human subsistence during the Late Pleistocene, while a study the East African Rift System uses phytolith from archaeological sites analysis to understand the environment in which the Middle Stone Age hominin lived (Mercader et al. 2013).

Similar to palaeoenvironmental evidence, information on human-environmental interactions will be presented in three periods; the Late Glacial Period, the Holocene and the last 2000 years. *Eburran* is a term referring to the long lived lithic industry local to the central Rift of Kenya, it occurred in five stages each associated with a change in ceramics and domestic animals from $\sim$13,000 to $\sim$3000 cal yr BP, while *Elementian* refers to the lithic industry local to Lake Elementeita characterised by large microlithic blades, *Remnant Ware* ceramics and finished tools from $\sim$3000 to $\sim$1300 cal yr BP (Gifford-Gonzalez 1998).

### 2.7.1 The Late Glacial Period ($\sim$16,000 to $\sim$11,700 cal yr BP)

Rapid climate change from the Late Glacial period to the extreme aridity of the Younger Dryas around 13000 and 11700 cal yr BP and then the wetter conditions of the African humid period would have impacted on the size and migration patterns of the early modern human populations (Scholz et al. 2007). Acute ecological decline causes desertion of settlement areas (Mercader et al. 2013), when the precipitation levels were extremely low during the Younger Dryas communities would abandon their homes and move closer to water bodies and up highlands. There are few records available detailing human-environment interaction during the Late Glacial period, the only Late Stone Age (LSA) as well as Eburran 1 (dated $\sim$12000 cal yr BP) archaeological sequences are recorded from Lake Victoria, Lake Turkana, Rift Valley as well as Southern Kenya (Gifford-Gonzalez 1998).

### 2.7.2 The Holocene (∼11700 to ∼2000 cal yr BP)

Hunter-gatherers impacted the environment through deforestation and the use of wood for fire. Pottery was introduced into the hunter-gatherer material assemblage with the Kansyore pottery tradition from ∼8200 cal yr BP and there was burning of the grass environments of the Rift Valley to increase visibility of medium-sized mammals which were preferred by some groups (Ambrose 2001). Herding as a form of food production in arid eastern Sahara has been dated to ∼8000 cal yr BP (Chritz et al. 2015). There is evidence, for an early integrated exchange system and pastoral development with different domesticated animals such as cows, goats, sheep and donkeys from ∼8000 cal yr BP (Wright 2005) and in the Sudanese valley before 7000 cal yr BP (Bower & Nelson 1978). Phytolith evidence of bananas in Cameroon was dated to 2500 cal yr BP (Mbida et al. 2006) while finger millet was domesticated either in the Ethiopian Highlands or in East Africa by 1000 cal yr BP (Fuller & Hildebrand 2013).

A lake orientated LSA occupation "Eburran 2" is evidenced by bone harpoons and ceramics on both shores of Lake Turkana around 9000 cal yr BP (Gifford-Gonzalez 1998) prior to the arrival of domesticates. Between 8000 and 6000 cal yr BP, LSA and Eburran three and four archaeological sequences are recovered from the eastern shore of Lake Victoria, Lake Turkana, Rift Valley as well as Southern Kenya (Gifford-Gonzalez 1998, Robertshaw 1991) yielding microlithic artefacts, Kansyore pottery, fish bones and a range of mammals. Between 6000 and 4000 cal yr BP lake levels were quite low, Kansyore pottery and LSA artefacts associated with wild fauna as well as Eburran 5 and Nderit pottery sequences are recorded (Gifford-Gonzalez 1998, Robertshaw 1991, Smith 1992, Wright 2007), Eburran 5 hunter-gatherers are described as highly mobile evidenced by the sourcing of raw materials for their tools (Wright 2007). The mid to late Holocene was a period of climatic and resource stress within East Africa which could have resulted in a subsistence existence exploiting riverine resources (Wright 2007, Wright et al. 2011). This situation was observed along the Galana river with artefacts dating back to around 6000 cal yr BP and showing occupation from ∼6000 to 1000 cal yr BP (Wright 2007); adoption of exploiting different resources would ensure survival during abrupt changes in resource availability. The identification of Narosura ceramics not previously identified suggests cultural connections with Narosura tradition located west and south of Tsavo (Wright 2007).

There is a gap in the archaeological sequence between 6000 and 3000 cal yr BP in Kenya and Northern Tanzania highlands spanning the transition from foraging to food production. Herding cultures of eastern and southern Africa originated from Eastern Sahara (Wright 2014) and the first evidence of pastoralist communities in East Africa occurs about 4500 cal yr BP (Ambrose 1998, Gifford-Gonzalez 2000, Lane 2013, Smith 1992) with domestic species from Northern African and Southwest Asia interacting with local communities who then acquired and adopted herding techniques (Mutundu 2010). Pottery (4860 cal yr BP) and faunal remains (mainly closed habitat species) from Eburran phase 4 have been found at Enkapune ya Muto, Naivasha, Kenya (Ambrose 1998, Gifford-Gonzalez 1998).

There are several archaeological sites (Figure 2.5) such as Enkapune ya Muto location (Ambrose 1998), Narosura and Mau escarpment having both wild and domesticated faunal remains (Gifford-Gonzalez 2000). Rock shelters such as Enkapune ya Muto are thought to have been occupied during periods of drier climate (Ambrose 1998, 2001). These sites have been described as pastoral and hunter-gatherer sites due to the mix of artefacts found; there is difficulty in distinguishing new practices from migrants and evolution of economies. The arrival and spread of pastoralists to Central Kenya occurred between 4500 and 4200 cal yr BP, southern and western Kenya between 4400 and 4150 cal yr BP, and by about 3800 to 3500 cal yr BP reach Tsavo and the middle Sabaki River (Lane 2013). The spread of early Pastoral Neolithic people is hypothesised to have been driven by either precipitation with hyper-aridity forcing communities to stay close to water bodies and pastures or cattle diseases prevalent in forested areas acting as barriers to movement (Wright 2007).

Archaeofaunal remains of livestock have been identified in Neolithic sites from mid $\sim$5000 and 3000 cal yr BP with the earliest dates $\sim$5000 cal yr BP in the south of the Lake Turkana Basin (Gifford-Gonzalez 1998). *Barthelme* sites contained Nderit and Ileret ceramics dated between 4500 and 4000 cal yr BP which were excavated alongside livestock, fish and aquatic vertebrae remains along Lake Turkana (Gifford-Gonzalez 1998, Wright 2007). Most Elementeitan dates range between 3000 to 1300 cal yr BP, and Elementeitan lithics are characterised by large blade blanks and finished tools with the obsidian raw material sourced from nearby sources. Their diet incorporates plant as well as animal food (Gifford-Gonzalez 1998) as observed from the Gogo Falls record which are composed of cattle, sheep and fish dating back 2000 cal yr BP (Chritz et al. 2015).

The Pastoral Neolithic are described as the herding communities who relied heavily on domesticated stock, used pottery and employed LSA technologies inhabited East Africa from c. 4500 to 1300 cal yr BP with the appearance of stone tool technology and pottery (Bower 1991, Collett & Robertshaw 1983, Wright 2007). Some texts suggest earlier specialised pastoralism as well as employment of diverse economic strategies flourished in southern Kenya (Chritz et al. 2015), fishing and use of wild herbivores and birds supplemented pastoralism livelihoods. The final phase of the lower stone age is also dated to c.4400 cal year BP (Lane 2011). Unfavourable climate (regional hyper aridity), and epizootic diseases (Gifford-Gonzalez 2000) are some of the reasons for a slow migration of pastoralism past the Turkana basin between 4000 and 3000 cal yr BP. Increased aridity and cultural phenomena are suggested to be correlated by the evidence of exploitation of aquatic species by the pastoralists along the Galana River (Wright 2007). The Bantu are believed to have moved east from the Congo area in two movements, one due east across the lands north of Lake Victoria and the second south-easterly towards the coast (Hellenthal et al. 2014, Russell et al. 2014). The group that headed in the south easterly direction dispersed into two further groups, one spread north along the coast remaining in the Nyika groupings around Mombasa and the Pokomo along the River Tana banks and the other went further inland and established them around the Mt. Kenya region (Russell et al. 2014).

Emergence of diverse communities across East Africa practising pastoralism, cultivation and foraging could have established around 3000 cal yr BP when the modern bimodal rainfall pattern established coinciding with Elementeitan and Savannah Pastoral Neolithic occurrence (Gifford-Gonzalez 1998). Pastoralism as a specialised strategy is thought to have evolved around 3000 cal yr BP (Bower 1991, Chritz et al.

2015, Gifford-Gonzalez 1998, Wright 2007). Desiccation of the Sahara coupled with increased population resulted in a general southward movement of pastoralists from farther north and the Nile river valley to the savannah habitats (Bower 1991, Gifford-Gonzalez 2000, Smith 1992, Wright 2014). This migration could have been made possible due to the mountainous terrain providing the only marginal available area for food production and the expansion of grazing land on the edges of draining lakes and river floodplains evidenced by endoaquatic artefacts (Gifford-Gonzalez 1998, Robertshaw 1991, Wright 2007).

A synthesis of radiocarbon dates suggests that farming spread to East Africa from the west from 3000 to 2500 cal yr BP and rivers and coasts were important dispersal corridors (Russell et al. 2014). Demographically, early farming communities participated in cultivation and iron smelting and spoke proto-Bantu languages while cultures that dominantly relied on domestic livestock spoke proto-Cushitic or Southern Nilotic languages (Lane 2004). The use/cultivation of tubers would barely be visible in the genotypic and phenotypic record, as these reproduce through vegetative propagation, and linguistic research indicates that Proto-Great Lakes Bantu speakers grew yams before entering the area (de Maret 2013). Yams can grow in various environments, including moist forest conditions forests if the canopy is broken up (Fuller & Hildebrand 2013). There is little direct archaeological evidence of subsistence strategies of Urewe ware users apart from sorghum and finger millet pollen from the Gisagara region in Rwanda (Van Grunderbeek 1992). Work close to Lake Victoria has recently shown that people relied on various subsistence modes, not purely on hunting, gathering, herding or agriculture (Lane et al. 2007). Slash-and-burn was the most likely mode of agriculture, thus contributing to the deforestation of the Great Lakes area. As shown by Heckmann (2014), the most probable environmental impact of agriculture and iron working is deforestation followed by subsequent erosion and land degradation. Agriculture remains an important aspect to many people's lives in East Africa and is a conspicuous aspect of the landscape.

### 2.7.3 The past 2000 years

Forest clearance for cultivation and metal working spread almost 2000 years ago (Håkansson 2004) with evidence of iron age tools, Urewe and Kwale pottery (de Maret 2013, Gifford-Gonzalez 1998). Evidence of the export of agricultural goods and animal products, notably tortoise shell and ivory, suggest human impacts on the landscape in the interior of East Africa by c. 2000 cal yr BP. Global and regional trade systems have influenced the natural environment, ecology and human resource use in East Africa particularly from the ~18th century (Hâkansson & Widgren 2008, Håkansson 2004).

Archaeological, historical and ethnohistorical information provide insight into the extent of trade; at the end of the first millennium Arab and Persian ships were already working along the East African coast as ancient coastal trade engaged Graeco-Roman ships who exchanged dyed clothing, metals, wine, and crafts for the export or re-export of cinnamon, fragrances, and animal parts (Mathew 1963). Indian Ocean trade routes were likely developed by Persian and Sassanian traders in East Africa (Mathew 1963). There was expansion of overseas trade with exports to Europe and Asia between ~950 and ~400 cal yr BP while the Swahili speakers had already established themselves ~850 cal yr BP with the greatest expansion around 500 cal yr BP (Håkansson 2004). This expansion led to a regional decimation of elephant populations in eastern and southern Africa by the end of the 19th century (~200 cal yr BP) due to the expansion of the western Europe and United States market with increased market prices (Håkansson 2004, Sutton 1984).

Medieval sources such as the major trade articles and oral traditions recorded by contemporary historians mention cloth and elephant populations before and around 1000 cal yr BP, glass beads, cowrie and reworked ivory dating back to the 600 cal yr BP have been excavated from Bunyoro in Uganda and around 500 cal yr BP in Engaruka Tanzania (Håkansson 2004). The increased caravan trade prompted the slave trade with local chieftains selling dependants, criminals and prisoners raided from neighbouring communities prompting migration (Håkansson 2004). Along the Taita Hills and nearby Kisigau, rock shelters have been interpreted as refuges from slave traders from the 17$^{th}$ to 19$^{th}$ century (~400 to 200 cal yr BP) (Wright 2007).

Elephant populations have been found to have significant impacts on ecosystem composition; high elephant numbers maintain open grasslands which rapidly transform to woodlands when the elephant numbers drop as observed at Queen Elizabeth Park in the 1970s (Håkansson 2004). Thus elephants have been significant in maintaining open grasslands favouring pastoralism by providing grazing and a tsetse fly free environment ((Håkansson 2004). The 19$^{th}$ century ($\sim$300-200 cal yr BP) estimated to be the peak of ivory trading saw high elephant mortality rates with (Giblin 1986, Hâkansson & Widgren 2008, Håkansson 2004) estimating about 7000 elephants were killed in 1848 to supply the 342,050 kg of ivory exported from Zanzibar.

The East African local economy was based on the trade of grains, root crops, salt, medicine, specialist services and livestock with cattle as a common value base where cattle ownership secured wealth and kinship (Håkansson 2004) with variation in value and exchange rates between the interior and the coast set by demand and supply. Local economies were influenced by the influx of coastal goods as most people engaged in trade so as to increase their herds thus supporting the maintenance and growth of pastoralism (Håkansson 2004). The great Rinderpest epidemic c. 1982 decimated up to 95% of the cattle in many East African areas (Håkansson 2004), this is theorised to have intensified ivory trade again as the communities rebuilt their herds. Regional trade enabled agro-pastoralists and pastoralists to settle in harsh environments that were suitable for large herds and trade buffered the impact of cyclical droughts. This is because instead of storing grain for years, people used cattle and other currency to obtain food during food shortages (Håkansson 2004).

Slash-and-burn was the most likely mode of agriculture c.700 years ago and evidence for irrigation channelling of water and terracing to maintain soil fertility for farming is recorded form Engaruka, Tanzania (Sutton 1984, Westerberg et al. 2010) and Sonjo (c.500 years ago) as well as to the east near Kilimanjaro, Pare and Taita Hills (Sutton 1984). Historical land use in East Africa tends to assume that landscape degradation through human impacts occurred after colonisation; however indicators of land degradation such as loss of forest species have been identified earlier in several palaeoecological, explorer and colonial records. Use of aerial photographs provides a visual representation of land cover which is important when comparing with remote sensing images and oral history (Borjeson 2009). There are ethnographic records of clashes between the Maa speakers and the Bantu and Nilotes during drought periods for pasture in the central highlands and Lake Victoria region respectively (Hâkansson & Widgren 2008, Rutten 1992, Waller 1976). The Nilotes and Nilo-Hamites came from a northerly direction spreading from dispersal centres in the Nile valley region (Bower 1991, Wright et al. 2011).

Missionaries had been active in East Africa since their penetration into the interior from the 1840s (Lane 2011, Waller 1976). National Parks were established by the colonial government as hunting reserves for example Tsavo East National Park was established in 1948 (Wright 2007). Many aspects of the environment have now been impacted by human beings (Kareiva et al. 2007) and East African ecosystems have experienced major human interactions during the late Holocene such as the establishment of pastoralism as a significant part of the East African savannahs for the last three thousand years (Gifford-Gonzalez 2000, Lane 2013, Mace 1993, Smith 1992).

## 2.8 East African human-environment interactions

Humans have a very long history of interacting with East African environments and the spatiotemporal variability of these ecosystems has been important to the physical and social evolution and adaptation of humans. Relationships between humans, climate, herbivory and fire are dynamic and a multi proxy approach to try and disentangle them is needed to provide information on human land use and its manifestation on the landscape. Climatic impacts, anthropogenic impacts, herbivory, fires etc are heterogeneous through time and the implications occur at different spatial scales, for example tropical savannahs account for $\sim$40% of pyrogenic $CO_2$. According to the Millennium Ecosystem Assessment, humans have rapidly and extensively changed the ecosystems to meet growing demands than in any comparable time in history (Kasperson et al. 2011) and global temperatures have increased by 0.6°C.

During the Holocene, rapid land cover changes and land-use patterns evolved that altered many physical, chemical, and biological processes (Kaplan et al. 2011) and intensified through time (Bowman et al. 2011, Ellis et al. 2013, 2010, Foley et al. 2013). The use of tools and domestication of plants influenced human-environment relationships by the expanding use of metal agricultural equipment, irrigation channelisation, and multiple land-use practices. Transformation of landscapes to increase suitability for different purposes such as channelisation and terracing to improve agricultural production has been in practise at several sites with the earliest sites recorded on Mt. Kilimanjaro (Håkansson 2004, Sutton 1984, Westerberg et al. 2010). Faunal/pottery evidence of change in livelihood from foraging to food production has been dated to c. 4000 cal yr BP (Ambrose 1998).

Anthropogenic influences from population growth, trade expansion and ecosystem conversion amongst others on the global landscape have accelerated over the past five centuries, resulting in rapid transformations of many ecosystems and modifications to the landscape (Ellis et al. 2010, Hurtt et al. 2011). The introduction of exotic tree species altered the forested and farmland landscapes across East Africa and have implications for changes in soil, plant biodiversity and socioeconomic development pathways (Colombaroli et al. 2014, Conte 2007, Neil Sampson 2005). The biggest impact of exotic tree species across East Africa for example on Mt. Kilimanjaro, Kenya, the Aberdare ranges and the Mau Forest Complex is that the exotic trees have replaced large swathes of indigenous forest (Majule et al. 2009), accidentally by outcompeting indigenous species or on purpose through plantations.

During the late 19$^{th}$ century in the North Pare Mountains, the establishment of some exotic tree species plantations began to reduce soil erosion into fluvial systems (Heckmann 2014). Introduced trees can also host exotic pest species carried to East Africa and in 1968, pine aphids were found in Kenya and outbreaks occurred in 1969 (Odera 1969). Although, encroachment on indigenous forest remains a significant threat to biodiversity in Kenya, micro-forestation efforts in intensive farming systems increase forest cover and diversity at a landscape scale (Legilisho-Kiyiapi 2002). At fine scales, some human activities may result in reduced degradation of some aspects of ecosystems for example conservation strategies such as protecting threatened ecosystems and species is crucial for recovery.

Despite the many managed protected areas, (Figure 2.6) threats to sustainable ecosystem management continue. Some of the threats experienced include; ecosystem conversion into agricultural and urban areas, political-tribal mediated insecurity; ineffective governance; different use of resources by different ethnic groups; division of labour along gender and age lines; poverty and inability to diversify resources; traditions and neglect of traditional ecological knowledge (African Wildlife Foundation & Maasai Mara National Reserve 2009, Amboseli Ecosystem Stakeholders 2009, Macharia et al. 2010). Across East African landscapes, compensation, benefit sharing, market related schemes and outreach efforts (Salerno et al. 2014) have been introduced to encourage community participation in sustainable conservation and management (Bulte et al. 2008). In Kenya, the main protected areas are managed by government bodies such as the Kenya Wildlife Service (KWS) and Kenya Forestry Service (KFS). With contributions from international funding bodies the government has set up ecological monitoring programmes whose main purpose has been to understand the ecological composition, structure and functions for better conservation (African Wildlife Foundation & Maasai Mara National Reserve 2009, Amboseli Ecosystem Stakeholders 2009).

Placing current and ongoing research efforts in the region into a longer environmental and human history can be useful for long-term understanding of ecosystem dynamics which informs decision making at management level for the maintenance of ecosystem services and biodiversity. Acknowledgement that humans are an integral part of the earth involves recognition of archaeological sites as environmental archives, the combined use of archaeological and palaeoecological archives to interlink palaeoclimate, palaeoenvironment and human development to understand long-term perspectives in human-environment systems. For example some records document change in ecosystems with little human impact such as the $\sim$17,000 cal yr BP pollen record from Bale Mountains in Ethiopian (Umer et al. 2007) which provides the longest record of vegetation change above 3000m, improving our understanding of highland forest ecosystem change through the Holocene and justifying conservation efforts. Other

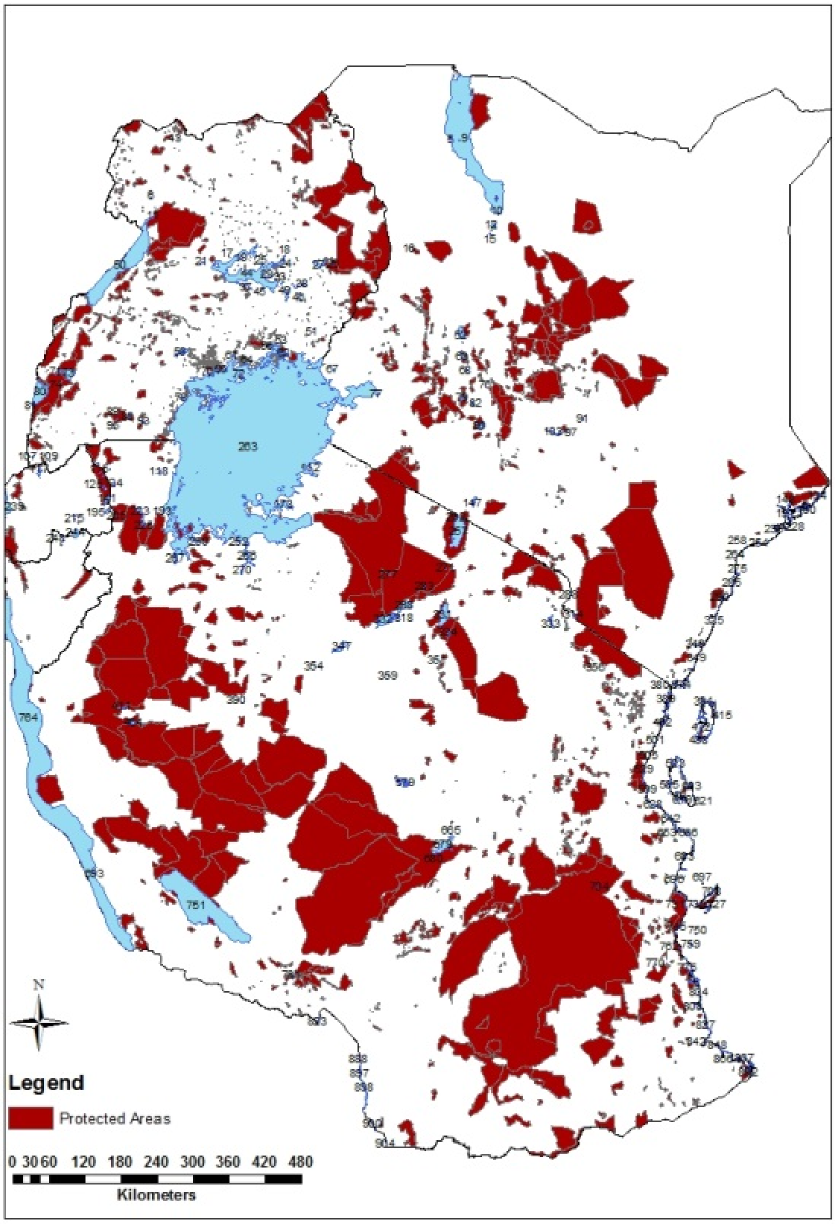



Figure 2.6: East African Protected areas under different management i.e. government protection, community run and private management.

studies attempt to discern interactions of different cultural groups within the same landscape; Mutundu (2010) utilises ethnoarchaeological observations to distinguish archaeological sequences associated with different socio-economic strategies during the Pastoral Neolithic in East Africa. Continued multi-proxy and multi-disciplinary studies should be encouraged to ensure we are able to capture as many aspects as possible providing a more accurate context for our interpretations of change.

## 2.9 Conclusion

The East African landscape has experienced a high degree of climatic variability as well as major developments in human technology and behaviour. The Late Glacial period cooling followed by the consistent warming as well as variable precipitation accompanied by a mainly mobile human society that has turned sedentary reflects a varied landscape. The arrival of pastoral economies $\sim$5000 cal yr BP along with new animal and plant species which were domesticated as well as interaction with local hunter-gatherer communities led to the development of new ethnic and socio-economic strategies. Additional records on environmental and human history are needed to improve our spatial and temporal understanding of these extensive and pervasive changes, this is mainly because most studies already carried out provide local interpretations and some records especially archaeological records can have conflicting intepretations. Understanding the longer environmental and human history provides insights into trends, trajectories, thresholds and legacies of change is crucial as we implement long-term sustainable and adaptive land use strategies.

# CHAPTER 3

## A multi-proxy Late Glacial sediment record documenting vegetation change and fire regime shifts from Eastern Mau, Kenya.

## 3.1 Overview

This chapter provides a description of the sediment coring, analysis and data collection of a 537 cm sediment core covering the last ∼16,000 cal yr BP retrieved from the Nyabuiyabuyi wetland, Eastern Mau, Kenya. Details are given of the regional setting, climate, hydrology, vegetation and land use to set the background for the palaeoecological data collected. This study provides a multi-proxy data set from which we can understand the interactions between vegetation change, fire regimes and wetland development since the Late Glacial period. Such a record improves our understanding of forest development through the Holocene as well as providing a comparison with other highland forest sites in East Africa such as Mt. Kenya, the Aberdare Ranges, Mt. Elgon and Mt. Kilimanjaro.

A high-resolution analysis of ecosystem change since the the Late Pleistocene-Holocene transition at Mau Forest Complex through the analysis of pollen, macroscopic charcoal, particle size analysis, loss on ignition, X-RAY fluorescence and magnetic susceptibility is presented. The chapter will highlight the main drivers of vegetation change within Mau forest and identify periods of significant change. X-ray fluorescence (XRF) elemental counts and magnetic susceptibility are used to trace amounts of terrigenous input which occur during periods of high sedimentation as a result of either increased erosion in wet periods or open ecosystems during arid periods. Periods of increased human activities are identified by the appearance and increase of chemical elements not detected previously and can be correlated with historical records. The pollen data provides an overview of the long-term vegetation patterns key for understanding the palaeoclimatic patterns while the lithological information from the particle size and loss on ignition values of organic content provide an understanding of the physical changes occurring in the sediment. The macro charcoal analysis extends our understanding of the long term variations in fire occurrence which we can use to explore the relationships between the causes of fires important for improving our interpretations of past fire regimes.

## 3.2 Introduction

The Late Glacial period is characterised by abrupt hydrological and temperature changes indicating a complex interaction between the atmosphere, land, ocean conditions and orbital forcing (Gasse 2000). A wetting/warming phase between 17,000 and 16,000 cal yr BP (Gasse 2000, Weijers et al. 2007) across the African tropics was followed by two similar transitions around 15,000 and 14,500 cal yr BP as recorded at Lake Victoria (Johnson et al. 2000), Lake Bosumtwi (Miller & Gosling 2014) and Lake Albert (Beuning et al. 1997). These transitions were characterised by dramatic rapid hydrological fluctuations culminating in a maximum precipitation-evaporation balance during the early Holocene resulting in a wet and green Sahara.

These generally wet and warm conditions were interspersed by major dry spells around 8400 to 8000 cal yr BP and 4200 to 4000 cal yr BP (Gasse 2000, Marchant & Hooghiemstra 2004). Between 3000 and 2000 cal yr BP several lake sediment records exhibit signs of increased evaporation indicating a general aridity trend across the landscape; pollen, diatom, elemental profiles as well as stable isotopes from Lake Tanganyika, Lake Naivasha, Sacred Lake, Lake Albert amongst others reveal that the aridity was regional (Beuning et al. 1997, Olago 2001, Verschuren et al. 2000).

The last 2000 cal yr BP have been very variable with some lakes experiencing extreme fluctuations, for example Lake Naivasha has experienced several fluctuations between low stands (shallow and pond like) and high stands (large and deep lake) (Mergeay et al. 2011, Verschuren et al. 2000). During the same period the Lake Simbi record infers that the natural vegetation around Simbi and Lake Victoria was open grassland with episodes of woody savannah during wet periods (Colombaroli et al. 2016).

The ecological history of the Mau Forest Complex (MFC) is relatively unknown with the exception of geological studies carried out by (Williams 1991). This upland ecosystem is a species rich evergreen Afromontane forest which thrives under cool moist conditions characterized by a two season rainfall pattern with an annual rainfall of $\sim$2000 mm yr$^{-1}$ and temperatures decreasing with increase in altitude (Kinyanjui

2011, Spruyt 2011). East African mountains and highlands support large tracts of forest and host the headwaters of multiple large watersheds that are crucial to providing a range of ecosystem services to the surrounding populations as well as a habitat to endangered mammals such as the yellow-backed duiker (*Cephalophus sylvicultor*) and the African golden cat (*Felis aurata*) (Sang 2001). The MFC is located on the Mau Escarpment with elevations ranging from 1200-3000 m asl; it is the largest forest block in Kenya and the largest single block of closed canopy forest in East Africa at 400,000 ha (Akotsi et al. 2006, Nkako et al. 2005, Spruyt 2011). It comprises of seven blocks i.e. Mau Narok, Maasai Mau, Eastern Mau, Western Mau, Southern Mau, South West Mau and Transmara (Boitt et al. 2016, Olang & Kundu 2011, Were et al. 2014) of which only the Maasai Mau is not officially declared a protected ecosystem (Sang 2001). Mau is the source of numerous rivers (Table 3.1) such as Ewaso Ng'iro, Sondu, Mara and Njoro which drain into Lakes Turkana, Victoria, Nakuru, Baringo, Magadi and Natron (Olang & Kundu 2011) making it the single most important water catchment in the Rift Valley (Akotsi et al. 2006, Boitt et al. 2016, UNEP 2006).

Table 3.1: Rivers flowing from Nyabuiyabuyi swamp.

| Hydronym | Notes |
|---|---|
| Kiptunga | Flows South East then South; a tributary of Nyabuiyabuyi |
| Makalia | Tributary of Lake Nakuru |
| Mara | Tributary of Lake Victoria |
| Marishoni | Flows North |
| Molo | Tributary of Baringo |
| Naishi | Tributary of Lake Nakuru |
| Nderit | Tributary of Lake Nakuru |
| Njoro | Tributary of Lake Nakuru |
| Nyabuiyabui | Flows South West and then South |
| Rongai | Flows East then North |
| Sondu | Tributary of Lake Victoria |
| Songi | Flows West and then South West |
| Southern Ewaso Ng'iro | Tributary of Lake Natron |

The MFC is one of the five largest water towers i.e. MFC, Mt. Elgon, Cherengani Hills, Aberdare ranges and Mt. Kenya with more than 18 water catchment areas classified as water towers in Kenya (MEMR 2012, Nkako et al. 2005). Water towers are forested upper water catchment areas that moderate water cycles and provision of water to ecosystems especially in the surrounding lowlands (Gasse 2000, Marchant & Hooghiemstra 2004).

The ability of sedimentary records from small wetlands to archive signals of climate dynamics, ecosystem change and anthropogenic influences has been essential for understanding palaeoenvironments (Ashley et al. 2004, Gillson 2006, Rucina et al. 2009). Palaeoenvironmental evidence is not available from the MFC thus limiting understanding of forest development and the interacting impacts on the forest. For example, fire is an essential part of this ecosystem but no information exists about the long-term relationship between fire, vegetation and climate. This study provides a narrative of the ecosystem change and the conditions under which the forest has developed since the Last Glacial period.

## 3.3 Study site: Nyabuiyabui Swamp

The Nyabuiyabui Swamp (Figure 3.1) is located along the broad ridge of the Kiptunga Forest Block and is managed by the Kenya Forest Service. It covers an area of 29,000 ha with an average elevation of 2865 m asl. Nyabuiyabuyi swamp expands and contracts in response to local hydroclimatic conditions and has minor ephemeral inflows and a single outflow to the southwest, a tributary of the Mara River. *Nyabuiyabui* is an Ogiek word meaning "spongy" (Spruyt 2011) or "marshy, bog-like" and supports a floating vegetation mat during wetter periods. The wetland covers 122 ha and the current wet surface area is ~6 ha. Logging roads have been constructed to the north, west and south sides of the wetland. The swamp is shallow with most water depths (>50 cm) above the surface during March-April 2014 and lower still (>20 cm) above the surface during April 2015.

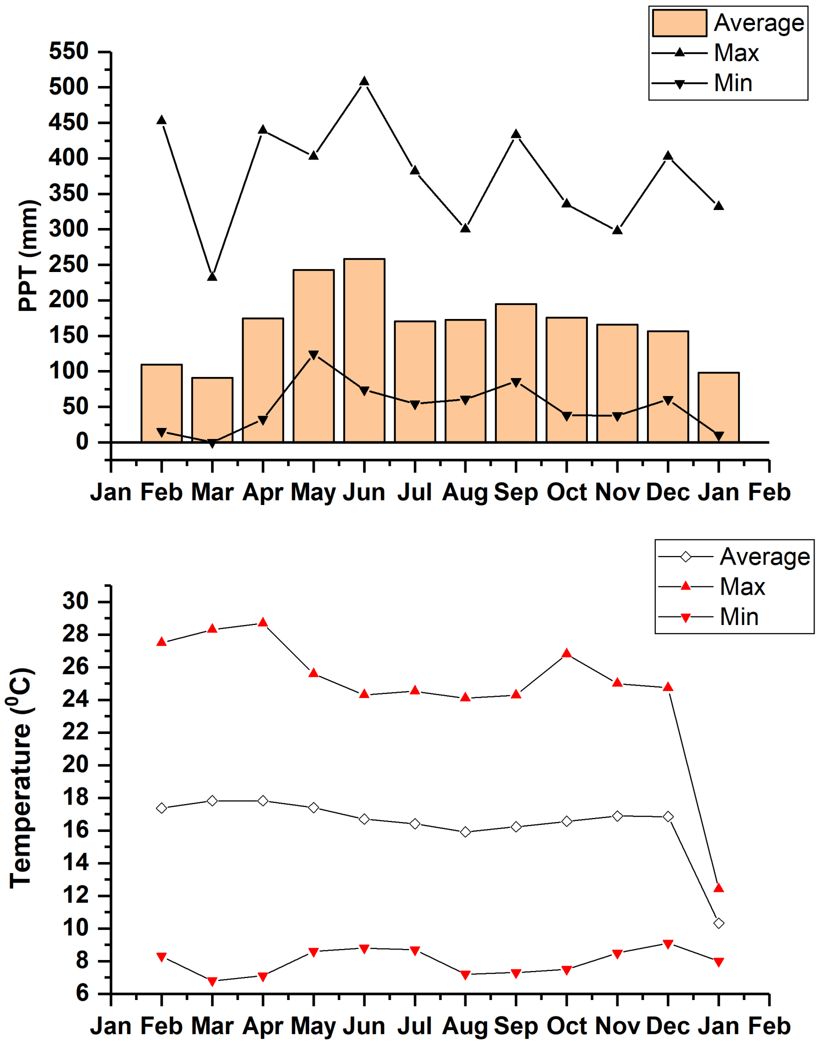



Figure 3.1: Monthly average, maximum and minimum precipitation values in mm from 17974 to 2009 and monthly average, maximum and minimum temperature values in °C from 1968 to 2011 graphed from Kericho weather station data (0.35 °S 35.28 °E, 2002m asl) which is 58km away from the Nyabuiyabuyi wetland.

The water level has been decreasing recently with some locals mentioning there was open water within the swamp and a channel to the road bridge prior to 1972. Evidence of recent human modifications of the wetland include the bridge construction over an outlet channel, a cut line running along the west margin to keep grass fires from threatening the forest, an abandoned, colonial government built pumping station immediately to the east of the bridge that used to supply water to the Kiptunga Forest Station. Cattle graze along the dry margin and further impact the hummocky ground and morphology of the grass and Juncaceae tussocks. The swamp is continuously covered by Cyperaceae-Poaceae dominated vegetation and surrounded small patches of indigenous forest and monoculture tree plots of varying age with planting dates from 1935 to 2006 with the bulk dating from the 1960s (Mustaphi-Courtney et al. 2014, Photomap 2008).

### 3.3.1 Geology and climate

The Mau Escarpment forms the western face of the Great Rift Valley. The geology is characterised by undifferentiated Tertiary volcanics overlain by Tertiary eutaxitic welded tuffs, which are evident in streambeds and occasional outcrops. The surficial geology consists of an extensive, thick mantle of Upper Pleistocene Mau ashes with basal tuffs (Jennings 1971, Williams 1991), possibly sourced from Londiani or Kilombe. The major soils are Andosols, Planosols and Vertisols (Olang & Kundu 2011, Were et al. 2014) with the slopes ranging from 2% in the plains to 30% in the foothills (Olang & Kundu 2011). The relatively young soils are moderately productive containing volcanic tephra (Jones et al. 2013), the texture ranges from silty clay loam to clay loam and clay with a pH range of ~5.5 to 6.4 (Olang & Kundu 2011). The climate varies depending on altitude and topography however the region is relatively mesic experiencing bimodal rainfall peaking during the long rains (March-May) with an annual average of 2000 mm $yr^{-1}$ (Nkako et al. 2005, Olang & Kundu 2011, Were et al. 2014). The monthly rainfall average experienced in Mau ranges from 80 to 280 mm while the average temperature is ~15°C (Figure 3.2).

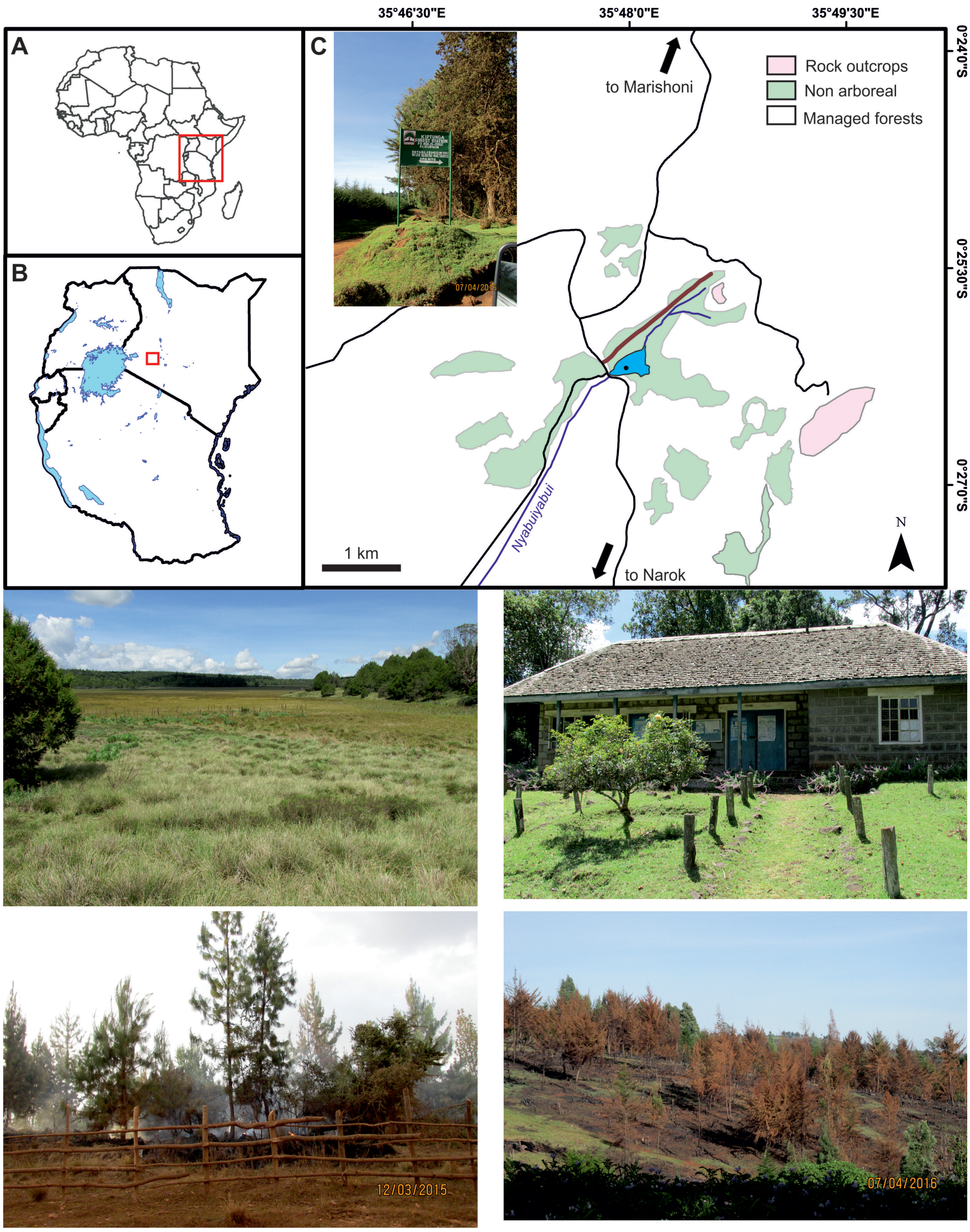



Figure 3.2: Map of the Nyabuiyabuyi within the study area, East Africa and East Africa within Africa. Photographs of: the wetland, Kiptunga forest block administration office and impacts of fires across the landscape.

### 3.3.2 Vegetation

Highland vegetation zones are largely controlled by orographic climate and vary from closed canopy forests in the hilly uplands to shrubland, scattered trees and grasslands at lower altitudes. The major vegetation zones (Figure 3.3) along a mountain/highland slope include deciduous open woodland below 2000 m, evergreen deciduous forest including (*Olea, Juniperus,* and *Podocarpus*) between 2000 and 3000 m, evergreen montane bamboo and then evergreen ericaceous with altotropical moorland species such as *Carex* and *Lobelia* between 3000 and 4000m. The Eastern Mau vegetation zones match vegetation zones from other East African highlands but at lower altitudes (summarized in Kratz (1993)).

| Kratz 1993 - Eastern Mau | | Hedberg 1951-Main East African mountains | | Mizuno and Fujita, 2014-Mt. Kenya | |
|---|---|---|---|---|---|
| Summit | Afroalpine belt: Tussock grassland, Giant Senecio '*Senecio keniensis* and *S. keniodendron*' and Lobelia '*Lobelia keniensis*' | Summit | Afroalpine belt: Tussock grassland, Giant Senecio '*Senecio keniensis* and *S. keniodendron*' and Lobelia '*Lobelia keniensis*' | Summit | Afroalpine belt: Tussock grassland, Giant Senecio '*Senecio keniensis* and *S. keniodendron*' and Lobelia '*Lobelia keniensis*' |
| <2600m | Open glades and grassland, Giant Senecio and Lobelia | ±3550-4100m | Ericaceous belt | 2900-3600m | Ericaceous Belt |
| 2300-2600m | Thick mature forests | Upto 3300m | Hagenia-Hypericum Zone | >2400m | Montane Forest Belt |
| 2100-2300m | Dense forest with glades | | Montane Forest Belt: Bamboo zone | | |
| >2100m | Open bushy forest | | Montane Rain Forest Zone | | |

Figure 3.3: Eastern Mau vegetation zonation from < 2100m compared alongside a composite of East African mountain vegetation by (Hedberg 2011, Mizuno & Fujita 2014).

Table 3.2: Tree plantations surrounding Nyabuiyabuyi swamp Photomap (2008)

| Species | Family | Origin | Name | Area(ha) |
|---|---|---|---|---|
| *Pinus Patula* | Pinaceae | Mexico | Pine | 46.5 |
| *Cupressus lusitanica* | Cupressaceae | Guatemala | White Cedar | 303.8 |
| *Corymbia maculata* | Myrtaceae | E.Australia | Spotted Gum | 5.2 |
| *Juniperus procera* | Cupressaceae | Native | Pencil Cedar | 36.6 |
| Unknown taxa | | | | 8.6 |
| Other local taxa | | | | 28.8 |
| Total | | | | 429.5 |

Between 1973 and 2011, significant land cover changes have been recorded, the forest-shrubland area has reduced from 1067 to 639 km$^2$, grasslands have reduced from 589 to 331 km$^2$, cropland has increased from 293 to 953 km$^2$, built up land has increased from 4 to 28 km$^2$ while bare land has reduced from 4 to 2 km$^2$ (Were et al. 2013). Currently the forest vegetation is partitioned into plantations (Table 3.2) with cyclic harvesting; plantation taxa include the indigenous Juniperus procera, and increasingly exotic taxa, which include *Corymbia maculata (syn. Eucalyptus maculate)*, *Cupressus lusitanica ,Grevillea robusta* and *Pinus patula* (Courtney Mustaphi et al. 2016, Mustaphi-Courtney et al. 2014, Photomap 2008).

Stresses on the plantation forests include pest outbreaks, such as rats and moles, monkeys that damage the apical buds of *Cupressus* and *Pinus*, and fires during dry periods. Scattered minor pockets of remnant indigenous broadleaf forests include *Croton, Dombeya, Hagenia, Juniperus, Olea* spp., *Podocarpus* and *Prunus*. Eastern Mau was declared Crown Land in the 1930s and made a Natural Reserve in the 1940s and officially gazetted in 1954 as a Forest Reserve under the Forest Act (Sang 2001). The Mau Forest hosts wildlife populations of birds, antelopes, gazelles, small primates and hyenas, it is classified as an important biodiversity and bird area (IBA) due to the rich bird population and rare mammals and an endemic butterfly (*Capys cupreus*) (Bird Life International 2015).

### 3.3.3 Human history and resource harvesting within Mau

The region is inhabited by the Ogiek community who traditionally practiced hunter-gatherer livelihoods but are increasingly practicing cattle keeping and much of the lower elevations have been converted to agriculture (Spruyt 2011). Believed to be the last forest dwellers in Kenya, the Ogiek communally owned and lived in the forests (Sang 2001). Mau is thought to have been occupied by hunter-gatherers and agro-pastoralists up to the early 1900s (Hakansson 1994), when the British colonial government signed land agreement with the Maasai who did not own the land occupied by the Ogiek, the first forceful eviction from Mau occurred between 1911 and 1914 and again in 1918 to Narok (Sang 2001).

The Mau highlands are divided into seven forest blocks; Eastern Mau Forest being the smallest and due to the fragmented nature of the forest has been identified as the most susceptible to negative impacts caused by anthropogenic land use/cover changes and global climatic change (Kinyanjui 2011). In Eastern Mau more than half the land has been converted into small arable plots with poor farming techniques ( such as monocropping, no terracing, over grazing) and thus the majority of the land experiences soil erosion (Olang & Kundu 2011). The forest has experienced vast logging to clear land for settlement and agriculture (Were et al. 2013) in the lower slopes, and commercial logging and planting of faster growing species in the higher belts under the management of the local government (both colonial and current) by the 1930s (Taylor 1962) just like the rest of the East African natural forests (Bussman 2001). Current human activities include burning and clearing for settlement, agriculture, charcoal burning and commercial timber sales. These activities have affected the water quality and quantity (river and lake levels), hydrological regime, temperature distribution and biodiversity (Boitt et al. 2016, Were et al. 2014). Land ownership varies with the government owning the protected areas i.e. the forest and national park, subsistence farmers own small farms and commercial farmers/ranchers leasing large scale farms adjacent to the forest (Were et al. 2014).

The highly valued timber species are *Albizia gummifera, Olea capensis, Polyscias kikuyuensis, Podocarpus spp, Pouteria spp, Prunus africana* and *Strombosia spp* with *P. kikuyuensis* being endemic to central Kenya. These species found in the Mau Forest Complex are highly sought after with legal and illegal logging taking place targeting them. The remaining forests are managed by Kenya Forest Service as an industrial forest for timber production. Neighbouring Mau forest blocks have experienced varying degrees of anthropogenic disturbance that impacted on ecosystem services such as biodiversity, soil geochemistry and soil seed bank resources, but they have ecological recovery potential (Kinyanjui et al. 2013). Cultivated land and plantation have been identified as the major causes of forest depletion over the last ~30 years (Boitt et al. 2016, Olang & Kundu 2011). The major land cover types (forests-shrublands, grasslands, croplands, built-up lands, bare lands and water bodies) have undergone tremendous changes over the last few decades with a decrease in the natural vegetation accompanied by an increase with the built up land and loss of water bodies (Were et al. 2013).

## 3.4 Methods

### 3.4.1 Field methods

Site selection was based on the depth and availability of a sedimentary archive. Vegetation surveys, sediment coring and profiling were undertaken in April 2014 (Figure 3.4). A suitable coring spot was determined by probing with fiber glass rods to locate the thickest sediment accumulation. A 537 cm core was recovered from S 00°26.188' E 035°47.978' at an elevation of 2919 m near the swamp centre using a hand pushed Russian D-shaped corer (Jowsey 1966) in 50 cm drives with ten cm overlapped sections. Cores were transferred to PVC tubes, wrapped in plastic film and aluminium foil, shipped to the University of York, UK, and refrigerated at 4°C.

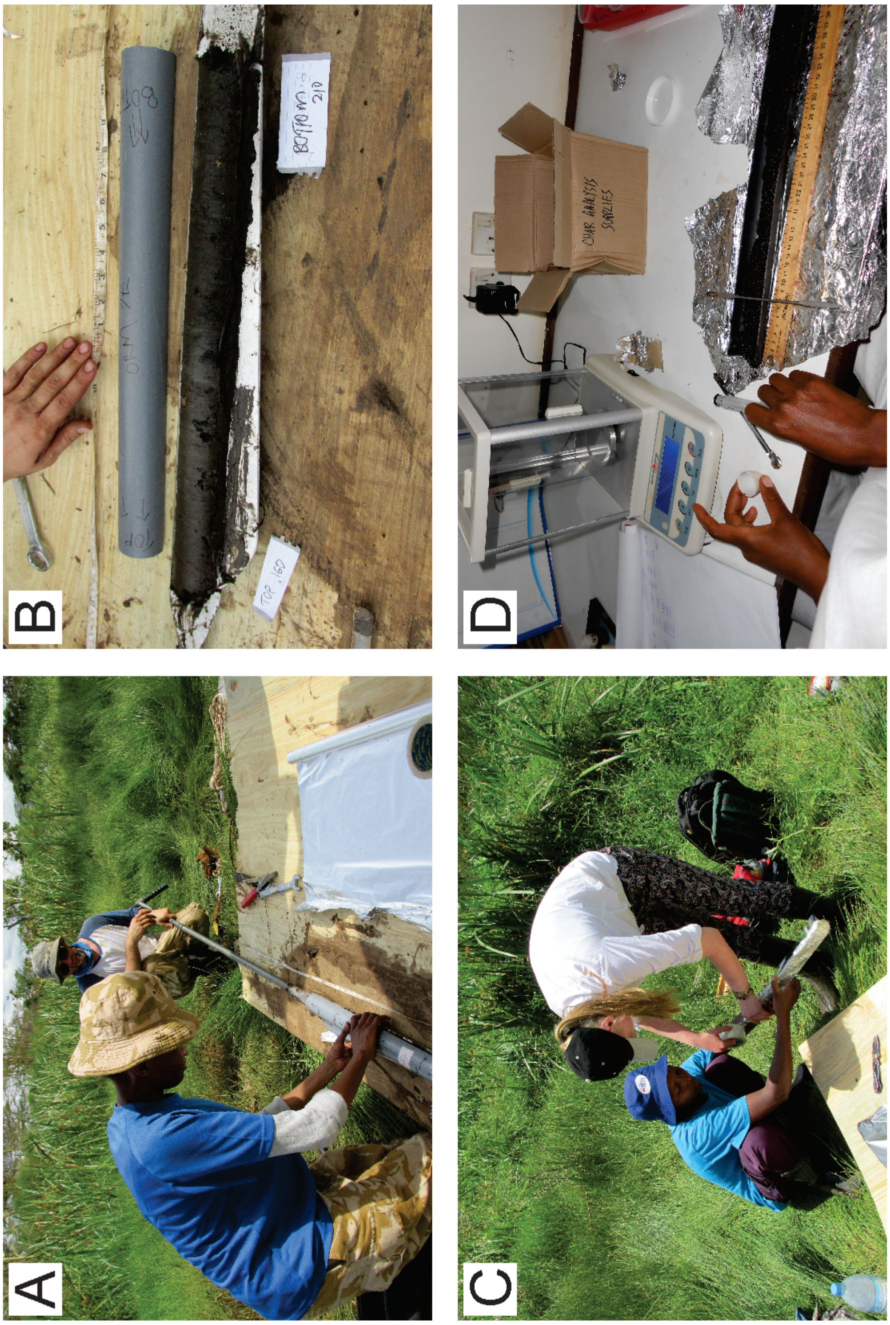



Figure 3.4: Field and laboratory work showing: (A) Sediment core removal from corer to plastic tubes for transport, (B) Photograph of core section for later description, (C) Core section wrapping for transport and storage and subsampling, (D) Subsampling for analysis.

### 3.4.2 Laboratory methods

#### 3.4.2.1 Radiocarbon dating

Four bulk sediment sub-samples were AMS radiocarbon dated at Queens University $^{14}$CHRONO lab, three picked organic fragments were AMS radiocarbon dated at Direct AMS and one was prepared to graphite at the NERC Radiocarbon Facility-East Kilbride and passed to the SUERC AMS Laboratory for $^{14}C$ analysis, thus a total of eight radiocarbon dates were acquired. The samples were selected from levels with visible stratigraphic changes such as colour and textures that could be an indication of change in the sedimentation process. The IntCal13 curve (Reimer et al. 2013) was used to calibrate the dates which were presented in calibrated year BP (AD 1950). An age-depth model was developed from the eight radiocarbon dates using BACON. BACON uses bayesian statistics to reconstruct accumulation histories over time taking into account variability (Blaauw 2010, Blaauw & Christen 2013, R Development Core Team 2011), in this case it excludes the outliers which were three dates around an age reversal. The minimum depth was set as 0cm and maximum depth as 537 cm, a run of cal BP and calibrated c-14 dates were used to run the model with 108 sections, acc.shape:1.5, acc.mean:20, mem.strength:4 and mem.mean:0.7.

#### 3.4.2.2 ITRAX-XRF scanning, optical and X-radiographic images and magnetic susceptibility

The core faces were cleaned and scanned with a Cox Analytical Systems ITRAX$^{TM}$ core scanner at the Department of Geography and Earth Sciences, Aberystwyth University, UK. Optical imagery of the core face was collected with a RGB camera, magnetic susceptibility was measured with a Bartington MS2E sensor at 1 centimetres intervals and air corrected between measurements. Twenty two elements were examined at 0.05 centimetre intervals through x-ray fluorescence (XRF) with a 3kW water-cooled Mo anode x-ray tube operating at 60 kV, 35 mA, 200 ms exposure, 10s dwell time every 500 microns giving readings in thousands of counts per second (kcps).

Elements measured are predetermined by the tube in the XRF scanner, the molybdenum (Mo) tube provides a scan of elements with high atomic numbers while the Chromium (Cr) tube provides a scan of elements with a low atomic number (Arnaud et al. 2014). The geochemical data represent a semi-quantitative measurement of elemental composition of the sediment and is influenced by potential x-ray absorption and/or scattering across the core due to variability in water content, particle size distributions, mineralogy and surface roughness of core face (British Ocean Sediment Core Research Facility 2012, Croudace et al. 2006, Croudace & Rothwell 2010). The results are normalized against aluminium (Al) which is the most abundant metal on earth, least reactive element and routinely used in normalising XRF data (Arnaud et al. 2014, Löwemark et al. 2011, Thomson et al. 2006). Normalization using detrital divisor can distinguish terrigenous or productivity origin.

The elemental profiles when combined with particle size analysis and organic matter content data provide a history of erosion and sedimentation events which may have several drivers such as climate change and catchment use as well as the palaeohydrology (Burrows et al. 2016, Thevenon et al. 2002). The information can also be useful for understanding sediment source: for example Al, Ti and Fe are terrigenous elements whose concentration levels increase during high accumulation events. K and Rb are commonly associated with detrital clay and K can be used for identifying moisture fluctuations and may be enhanced in turbidite muds, while Fe is mobilized during redox-related diagenesis and elevated Fe is commonly seen in oxic, sediment (Thomson et al. 2006). Si is an important terrigenous or productivity indicator (Croudace et al. 2006). Peaks in Ti, Fe, Si, Sr, and Rb which are detrital elements can be used to identify precipitation peaks (Burrows et al. 2016) when accompanied by increased silt-clay levels. Peaks in total organic carbon (TOC) correspond to minima in Al and Ti (Thomson et al. 2006). Sharp Cu peaks are largely of diagenetic origin, commonly an indicator of pyrite which may be detrital or authigenic in origin (Croudace et al. 2006).

The magnetic susceptibility profile was constructed using a MS2 meter with MS2C sensor every 500$\mu$m. Readings expressed in $\kappa$ or volume susceptibility, $\kappa$ representing the ratio of magnetization in samples (per unit volume) to the magnetic field created by the sensor, $\kappa$ is dimensionless with a scale of $10^{-5}$in SI units (Burrows et al. 2016). Magnetic susceptibility is how 'magnetisable' a material is and reveals the iron bearing minerals, calculate their concentration, process of formation and/or transportation, this is important for identifying periods of high erosion and sedimentation within the sediment when peaks and drops in the magnetic susceptibility are accompanied by peaks and drops of some of the elements. The values are useful as a concentration proxy for the input of the denser terrigenous minerals derived from the catchment providing a direct measure of changes in the wetland hydrology.

#### 3.4.2.3 Macroscopic charcoal analysis

Subsamples of 1 $cm^3$ of sediment were extracted at 1 cm intervals from the wet core face and soaked in sodium hexametaphosphate solution and a drop of hydrogen peroxide to disaggregate the samples and aid in the separation of the organic material and the clay particles (Bamber 1982, Schlachter & Horn 2010, Whitlock et al. 2010). Samples were wet sieved through a 125 $\mu$m mesh and the retained charcoal were identified by visual inspection and probed with a metal needle and pieces were tallied under a Zeiss Axio Zoom V16 microscope at 10 to 40X magnifications. Counts were converted to charcoal concentration values i.e. number of particles per unit of volume (pieces/$cm^{-3}$) and charcoal concentration rates (number/$cm^2$/$yr^{-1}$).

#### 3.4.2.4 Pollen and NPP analysis

A 1 $cm^3$ sub-sample was obtained every ten centimetres for pollen and spore analysis following the standard protocol (Faegri & Iversen 1950, Moore et al. 1991). An exotic marker (*Lycopodium* spores) was added prior to pollen analysis to aid in calculation of absolute concentrations (Bonny 1972, Stockmarr 1971). The sediment cores were subsampled every ten centimetres and 1 $cm^3$ was sampled into 100 ml beakers together with the one *Lycopodium* tablet with a known number of spores (9666 spores) added to each sample. 10 ml of HCl (to remove the calcium carbonate in the samples) was added to the samples, vortexed to encourage mixing of the sample with the acid and placed on a hot water bath for 2 minutes. Samples were then centrifuged at 3000 revolutions per minute (rpm) for 3 minutes and the liquid decanted. 7 ml of distilled water was added to each sample, vortexed, centrifuged and the liquid decanted. This was done twice to remove all the HCl.

20 ml of 10% KOH (to digest organic matter) was added to the samples which were then gently boiled while mixing on a hot plate to facilitate peptisation for five minutes and then left to cool. The samples were sieved (to remove large unwanted particles) through 250$\mu$m mesh screens into 15 ml polypropylene centrifuge tubes and centrifuged for 1 minute at 2000rpm and the liquid decanted. Samples were then washed twice with deionized water by adding 2 to 3 ml deionized water, thoroughly shaking the samples, topping up with water and centrifuging at 2000 rpm for 1 minute. If carbonates were still present 1 ml ethanol and 1 ml $H_2O$ were added and the sample vortexed. Seven ml of 96% Glacial acetic acid was added and carefully mixed into the mixture and left for 8 to 12 hours. Samples were then washed with deionised water to prepare for acetolysis. During acetolysis samples were washed twice with acetic acid i.e. 4 ml Acetic acid was added to the samples and centrifuged at 2000 rpm for one minute and the clear liquid decanted. Acetolysis mixture 96% $H_2SO_4$: Acetic anhydride at a ratio of 1:9 was added to the samples which were heated in aluminium block heater to 100°C. The acetolysis mixture digests the cellulose covering the pollen making the exine features distinct. The tubes were vortexed and heated further for 10 minutes. Samples were centrifuged and decanted to remove acetolysis mixture and washed twice in deionised water.

Heavy liquid separation using Sodium polytungstate ($3NaWO_4.9WO_3.H_2O$ with d=2) was carried out to separate the pollen from the remaining organic material. Three ml of the heavy liquid was added to samples and vortexed, water was carefully added using a glass rod to prevent mixture of the two liquids, the mixture was centrifuged at 3000 rpm for 1 minute and an organic suspension layer appeared at the boundary between the heavy liquid and water before the organic suspension was then carefully transferred to another test tube and samples were washed twice in deionised water. Prepared samples were transferred to residue tubes using 96% alcohol, centrifuged and decanted. Glycerine (same volume as residue) was then added to the sample and the tubes left to evaporate in stove at 60°C. Pollen samples were mounted onto the pollen slides where the identification and enumeration of the pollen, spores and micro charcoal was carried out at a magnification of 400 to 1000 and from each slide a minimum of 300 pollen grains, excluding Poaceae and Cyperaceae, using a Leica DM4000B. The pollen grains were identified using images and descriptions from the African Pollen Database and published atlases (Hamilton 1976, Hamilton & Street-Perrott 1980). Pollen were grouped into Afromontane, Trees, shrubs, herbs and aquatics. Afromontane and trees were differentiated so that we could observe if there were changes in forest taxa type.

The NPP were identified using images and descriptions from Van Geel (Van Geel 2001, Van Geel et al. 2011). More than 40% of the spores and NPP observed could not be identified even with the use of the references. The NPP and spore curves were analysed and interpreted as one curve as a general indicator of herbivore numbers.

#### 3.4.2.5 Sediment Properties

Organic matter and carbonate content in sediment may reflect the degree of soil erosion from the surrounding landscape by in washing caused by either a cold period limiting ground plant cover or human disturbance (Heiri et al. 2001). Loss on ignition and carbonate analysis provides the organic matter, carbonate (inorganic carbon) and siliciclastic content of sediment and the procedure involves weighing the fresh samples and then again after drying at different temperatures.

A sufficient number of clean and dry crucibles were numbered and weighed. The empty crucible weights and numbers were recorded as crucible weight (CW). 1 $cm^3$ of wet sample taken every 5 cm along the sediment core was weighed and recorded as wet weight (WW). The samples were dried for 14 to 24 hours at 105°C. Samples were removed from the oven and added to the desiccator to cool to room temperature for 25 to 30 minutes then reweighed for dry weight and recorded as dry weight after 105°C (DW105). After weighing the samples were put in the muffle furnace at 550°C for 4 hours. When cooled enough, the crucibles were removed using tongs and put in the desiccator to reach room temperature (45 minutes). Samples were reweighed for ignition weight DW550 and recorded.

1. The organic content was calculated as LOI 550 (organic content) = [(DW105-CW)-(DW550-CW)/ (DW105-CW)]*100. After weighing DW550 the sample was put back into the muffle furnace at 950°C for 2 hours and timing only started once temperature was reached. Once the furnace cooled significantly, samples

were removed using tongs and put in the desiccator. Once cooled to room temperature (at least 45 minutes) samples were weighed to obtain carbonate (inorganic carbon) weight DW950.

2. The carbonate content was calculated as LOI 950 = (((DW550-CW)-(DW950-CW))/(DW105-CW))*100

#### 3.4.2.6 Particle Size Analysis

PSA was determined using the Malvern Laser granulometer (MEH/MJG180914), before each run, cleaning and calibration was carried out to ensure integrity of the results. The cleaning procedure was carried out four times before each run; twice with tap water, then twice with deionised water. This involved adding 700 ml tap water to a one L beaker and putting it in the measurement position on the laser granulometer. Running the pump at 4000 for 20 seconds then lifting the sample unit to drain position and allowing tubes to drain. The calibration involved running a sample of known measurement (standard operating procedure) to compare the results with those provided in the log and check if the granulometer is working correctly. If the soil sample contained >3.5% organic matter it was pre-treated with 30% hydrogen peroxide ($H_2O_2$).

Samples were taken at every 5 cm and 10 ml of 30% $H_20_2$ was added to 10 g of the dry sample into a 500 ml beaker in a fume cupboard and left for one hour. After an hour, another ten ml of 30% $H_2O_2$ was added to the sample in a fume cupboard and warmed for 2 hours at 50°C on a hot plate whilst maintaining approximate volume by adding deionised water. After 2 hours it was briefly brought to boil then left to cool before starting the measurements. If the soil sample contained <3.5% organic matter, the hydrogen peroxide treatment was skipped and the runs started right after cleaning and calibration. At a pump speed of 1500, 1 to 2 grams of sample was added until obscuration was 4%. The granulometer took three measurements and calculated an average result (Malvern Instruments Ltd 2007). A total of three measurements were made for each sample and the average used as the final value. The distribution abundances within the 0.2-2000 $\mu$m grain size range were calculated by the Malvern software. All sediments were classified to three classes: clay (particles with a diameter < 2 mm), silt (2 to 60 mm) and sand (20 to 200 mm).

Grain size analysis of sediment is crucial for understanding the energy of the environment thus the processes behind the movement of the particles. In high energy environments, there's high kinetic energy thus only large particles are deposited with the smaller ones being carried away. In general, deposition occurs in low-energy environments and erosion in high energy environments. The particle size of aquatic sediment reflects the amount of water movement with fine particles deposited in still water.

### 3.4.3 Statistical analysis

The aim of the study was to document ecoystem change over time by focussing on the vegetation, geochemical and other sediment properties.The data collected were analysed and the results graphed for visual display. The radiocarbon dates were analysed using BACON, a statistical package in R that utilises bayesian statistics to develop age depth models. Bayesian statistics combine new data ( $C^{14}$ dates) with prior observations (Blaauw & Christen 2013, Blaauw et al. 2007) and a calibration curve, this results in an age-depth model where all the depth levels in the sediment record now have an interpolated calibrated age.

The raw pollen counts were converted to pollen percentages and presented in a pollen diagram drawn using the C2 Version 1.7.6 software program (Juggin 2011). The primary vertical axis represents the depth of the core in metres (m) and the secondary vertical axis represents the chronology of the core in calendar years before present (cal. yr BP). The horizontal axis indicates the relative percentages of the identified and counted taxa in the first pollen diagram and the pollen accumulation rates in the second pollen diagram. The percentages and rates were based on a pollen sum in which Poaceae, aquatic taxa (Cyperaceae, *Ludwigia, Nymphaea* and *Typha*), spores, NPP, micro-charcoal and indeterminable grains are excluded.

A broken stick (b-stick) analysis was carried out to determine the number of zones selected and then a Constrained Incremental Sums of Squares cluster analysis (CONISS) which groups samples sharing similar values in a hierarchical order. Both b-stick and

CONISS are components of the Rioja package in R to define and display the significant vegetation zones.

Ordination analysis (multivariate analysis) were carried out to search for underlying relationships between the variables, the principal components analysis (PCA) and exploratory factor analysis (package "nFACTORS") were selected to explore the relationships between the ITRAX-XRF observations, sediment properties and particle size analysis, this was carried out in R. A chi-square statistic was used to test the hypothesis whether the number of components identified using the PCA and factor analysis was sufficient.

Pearson's correlation coefficient values were calculated to measure the strength of the linear relationships between the variables, the values range from -1 to 1 with values $<1$ indicating a negative association, values $>1$ indicating a positive association and 0 indicating no association. Values that are $\leqslant$-0.5 and $\geqslant$0.5 indicate strong negative and positive association respectively.

Peak analysis was carried out on univariate data i.e the charcoal and magnetic susceptibility data to enhance visualisation and description. The local maximum method is used for every five points across each record with the 2nd derivative set as the mode and Adjacent averaging as the smoothing method. This was carried out using ORIGIN-PRO2017. Graphing of the climate data, pollen counts, radiocarbon dates, elemental profiles, charcoal counts and sediment properties was carried out in C2 and R with final formatting in CorelDraw and ORIGIN.

## 3.5 Results

### 3.5.1 General lithology description and age-depth model

Six lithological units are recognised from the Nyabuiyabuyi sediment core, dark organic detritus between 0 and 8 centimetres comprised of plant material at different stages of decay, a dark organic layer between 8 and 84 centimetres which changed into organic rich clay with coarse sand between 84 and 274 centimetres. At 274 to 435 centimetres the dark brown mud gets darker and a thick layer is visible at 386 centimetres. From 435 to 517 centimetres the sediment changes colour to dark grey and light coloured laminations and concretions appear. The bottom section from 517 to 537 centimetres is comprised of grey mud with a lot more light coloured concretions that are medium to coarse sized sand particles (Figure 3.5). The transitions in sediment colour and type are rapid and distinct except in the bottom one meter where laminations are visible. The radiographic images (Figure 3.5) are positives, with higher density sediments appearing darker and lower density sediments appearing lighter (Croudace et al. 2006, Kylander et al. 2012).

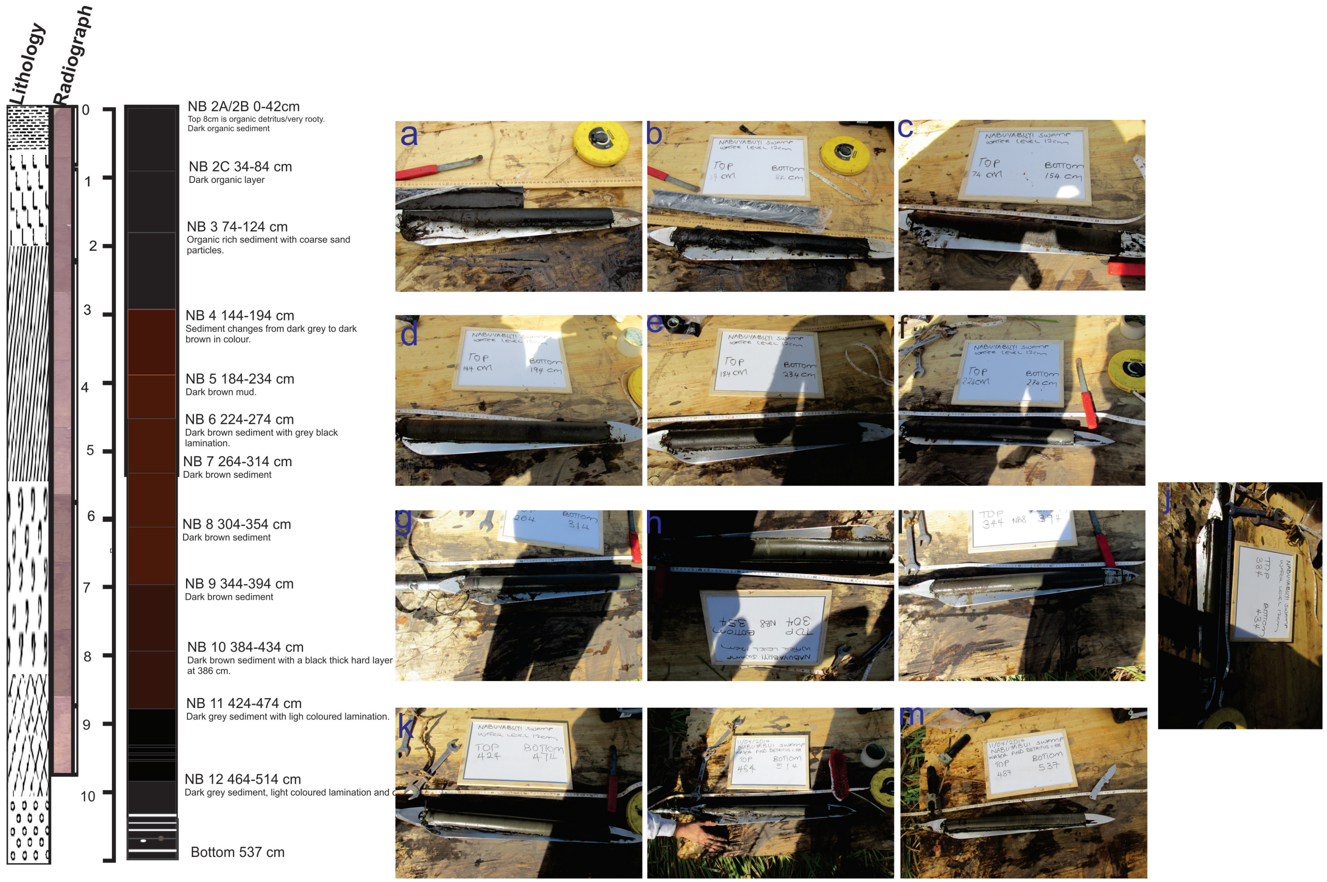



Figure 3.5: Nyabuiyabuyi lithology, radiograph image and photographs of core sections taken after core extrusion and used description of the stratigraphic changes in Nyabuiyabuyi sediment core, the depths of each core section is on the white board next to the picture.

### 3.5.2 Radiocarbon dating

Eight radiocarbon dates (Table 3.3), provide the basis of the BACON (Blaauw & Christen 2013). Non radiocarbon dates (top date which is the coring date) are presented as cal yr BP with the modern reference date taken as 1950 AD. There is an age reversal where the sample from 100 centimetres was older (10721$\pm$47) than the samples from two lower depths, as a result three dates (50 centimetres: 2449$\pm$35, 100 centimetres: 10721$\pm$47 and 128 centimetres: 7616$\pm$33) were noted as outliers the model did not pass through them (Figure 3.6).

Radiocarbon dating of wetland sediment has several disadvantages and the major ones are the occurrence of several outliers and hiatuses in the record that are difficult to interpret. The lack of annually resolved wetland records results in large chronological uncertanities that impact on the interpretation of proxy data. Due to budget limitations, not all levels can be dated thus interpolations are made by the model with limited information for the undated levels taken into consideration. Even with sophisticated models and many data points, accumulation histories of non-annually layered deposits cannot be exactly known. The age reversals from Nyabuiyabuyi wetland could be due to a potential hiatus, sediment reworking or root growth bringing younger material to lower sediment.

The cause of the potential hiatus during the early and middle Holocene ( 8000 to 4000 cal yr BP) is unknown but typically occurs in variable environments when there is a shift from lake to wetland conditions or a discontinuity in the accumulated material e.g. when floods erode part of the sedimentary deposit. Sediment reworking can be caused by increased sedimentation into the wetland during periods of flooding where sediment from elsewhere is introduced to the sedimentary record; this material could be older or younger than the sediment it settles on. Wetland ecosystems that sustain large wildlife also experience sediment reworking by animals that regularly tread along/across the wetland during watering and grazing and also those who utilise wetlands as habitats e.g. hippos. They are responsible for constant lateral reworking of the sediment, Nyabuiyabuyi wetland currently supports small cattle numbers with pasture and water.

Table 3.3: Selected samples for radiocarbon dating, sample and lab details.

| Depth (cm) | Age ($^{14}$C years) | pMC | Material | LabID | Age (cal BP) |
|---|---|---|---|---|---|
| 0 | -64 | | | Top of core | -64.1 |
| 30-31 | 201±23 | 97.53±0.28 | Plant remains | D-AMS-009664 | 956.1 |
| 50-51 | 2449±35 | 73.72±0.32 | Bulk sediment | SUERC-57340 | 2221.7 |
| 74-76 | 1865±37 | 17.58±0.13 | Charcoal | UBA-27553 | 2325.9 |
| 100-101 | 10721±47 | 26.33±0.15 | Bulk sediment | UBA-26117 | 8100.8 |
| 128-129 | 7616±33 | 38.75±0.16 | Plant remains | D-AMS-009663 | 8475.6 |
| 230-231 | 9837±42 | 29.39±0.15 | Bulk sediment | UBA-26118 | 11526.4 |
| 315-316 | 13963±60 | 18.19±0.19 | Bulk sediment | UBA-27554 | 14711.1 |
| 536–537 | 13692±83 | 18.19±0.19 | Bulk sediment | UBA-26116 | 17094.1 |

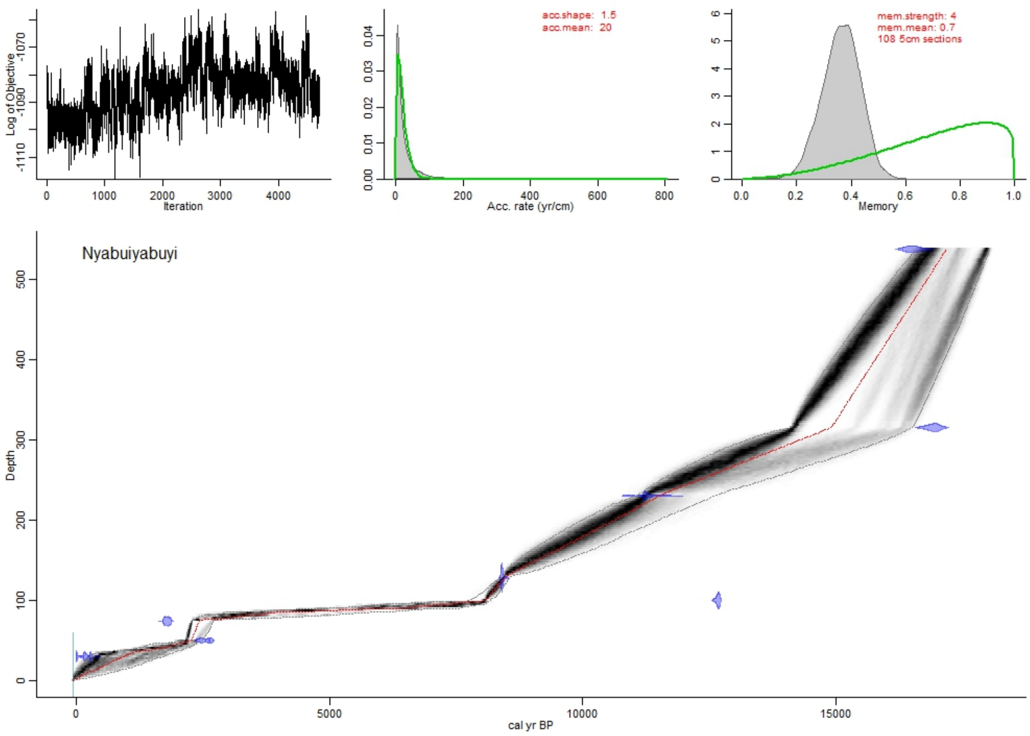



Figure 3.6: Nyabuiyabuyi wetland BACON age-depth model where three outliers are identified and not weighed by the model.

### 3.5.3 Magnetic susceptibility ($\kappa$)

The magnetic susceptibility varied across the sediment record from -4.14 to 28.99; the average magnetic susceptibility of the sediment core was 2.7±0.12. Six peaks are identified from the magnetic susceptibility curve (Figure 3.7), the peaks identified occur at: the top of the core = -2.07, at 114 cm (∼6872 cal yr BP) = -4.135, at 234 cm (∼11356 cal yr BP) = -3.15, at 267 cm (∼11971 cal yr BP) = 13.36, at 349 cm (∼ 13372 cal yr BP) = 28.99 and at 440 cm (∼14956 cal yr BP) = 13.46. The magnetic susceptibility has low linear correlation both negative and positive with charcoal concentration and the elements measured. The strongest positive correlation is with potassium (K) and iron (Fe) at 0.27 and 0.21 respectively while the strongest negative correlation is with copper (Cu) and chlorine (Cl) at -0.147 and 0.145 respectively, the magnetic susceptibility shows a weak negative correlation with charcoal concentration at -0.03 (Figure 3.7 and Table 3.4).

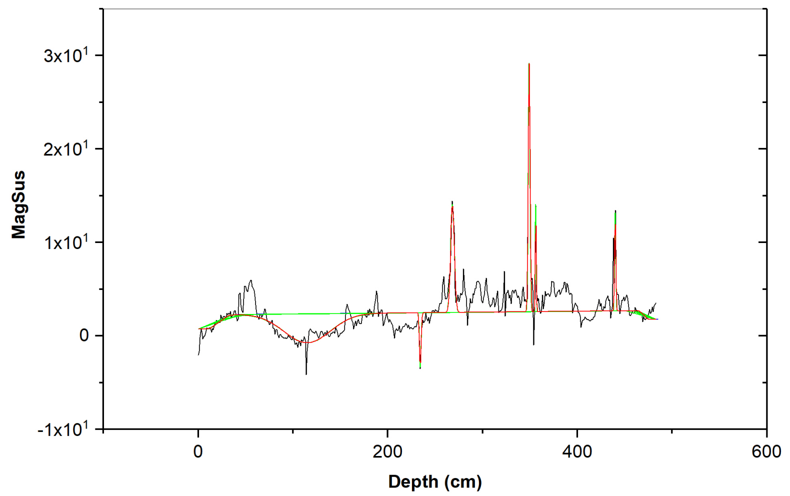
### 3.5.4 Geochemical properties using XRF ITRAX core scanning

A total of 22 elements (Table 3.4) were detected using the ITRAX core scanner, no absolute measures of geochemistry were performed as the ITRAX results represent semi-quantitative information of the relative variations of the elements within the sediment matrix as counts per second (kcps). Elemental composition was dominated by Fe, manganese (Mn), zirconium (Zr), yttrium (Y), rubidium (Rb) and titanium (Ti), with counts ranging from 5 to over 35,000 counts per second. K, mercury (Hg), lead (Pb), zinc (Zn), bromine (Br) and strontium (Sr) have counts ranging from zero to ∼318 counts per second. I, Cl, P, Ni, Cu, S, Hg, Ar, Si and Ca have the lowest counts of <100 counts per second. I, Ni, and Cl have the lowest values throughout the sediment core at a mean of 0.01, 0.45± 0.07 and 0.69±0.03 kcps respectively while Fe, Zr and Y have the highest values with a mean of 19234.22±258.17, 1348.85±8.79 and 535.65±5.19 respectively. Fe counts are extremely high throughout the sediment core because the base rock is felsic volcanic ash (Williams 1991).

All the elements except Cu, I and Pb are strongly positively correlated with Pearson's correlation coefficients ranging between 0.4 and 0.8 however Cu, I and Pb are negatively correlated (Figure 3.8). An exploratory factor analysis and PCA both agree

**Peak Analysis**

Data Set:[Magsus]Magsus!D Date:30/11/2017
BaseLine:End Points Weighted
Chi^2=2.06197E+000 Adj. R-Square=6.72902E-001 # of Data Points=483
SS=9.58817E+002 Degree of Freedom=465




Fitting Results

| Peak Index | Peak Type | Area Intg | FWHM | Max Height | Center Grvty | Area IntgP |
|---|---|---|---|---|---|---|
| 1 | Gaussian | -188.43314 | 57.19668 | -3.09496 | 114.96451 | -53.19483 |
| 2 | Gaussian | -12.79462 | 2.06809 | -5.812 | 234.2842 | -3.61193 |
| 3 | Gaussian | 57.18456 | 4.65674 | 11.53625 | 268.18723 | 16.14325 |
| 4 | Gaussian | 63.8781 | 2.25405 | 26.62295 | 349.07283 | 18.03284 |
| 5 | Gaussian | 14.59852 | 1.19562 | 11.47853 | 356.0665 | 4.12117 |
| 6 | Gaussian | 17.34311 | 1.55376 | 10.48606 | 439.93325 | 4.89598 |

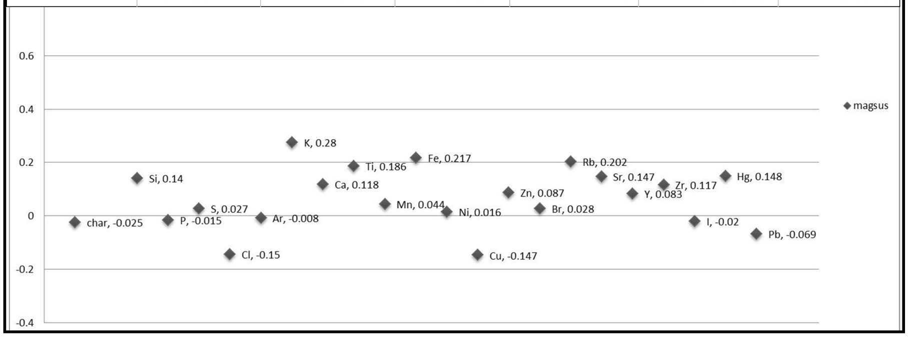



Figure 3.7: Nyabuiyabuyi Magnetic Susceptibility and the six main peaks identified using a Gaussian peak analysis.

that the first three components are sufficient to explain the relationship, from the principal component analysis (PCA) the first component explains 65.8% of the variance, the second component 12.3% and the third component 4.1% (Figure 3.8). A chi square analysis to test whether 3 factors are sufficient agrees ($\chi^2$=2966.95, df=150, p-value$\gtrsim 0.05$).

Zn shows the highest positive correlation with charcoal concentration of 0.1 while Cu and Pb show the highest negative correlation with charcoal concentration at -0.13 and -0.11 respectively. K, Fe, Rb and Ti show the highest correlation with magnetic susceptibility at 0.28, 0.22, 0.20 and 0.19 respectively while Cu, Cl and Si show the strongest negative correlation at -0.147, -0.145 and -0.141 respectively (Table 3.4). ITRAX1 is at the bottom of sediment core from 484 to 294 centimetres between 16,000 and ~12,600 cal yr BP; this is then followed by ITRAX2 which is further subdivided into ITRAX2A and ITRAX2B at ~3500 cal yr BP.

I, Ni and Cl have the lowest mean counts of all the elements detected, Fe, Zr and Y have the highest and Cu, I and Pb significantly increase towards the top. ITRAX1 covering the Late Glacial period ~16,000 cal yr BP to the Younger Dryas at ~12,600 cal yr BP displays the highest counts and fluctuations of all the elements. In this zone the mean counts per second of I and Ni are higher than the record means while Cl lower in this this zone than the mean of the complete record. Fe, Zr and Y are significantly higher in this zone (8272.8, 495.9 and 203.8) compared to the mean of the complete record (5968.9, 405.5 and 163.1) respectively. Cu is almost twice as high in this zone as the mean of the record while Pb is significantly lower in this zone than the record mean.

The second zone is divided into two sub zones i.e. ITRAX2A (~12,500 cal yr BP to ~3500 cal yr BP) and ITRAX2B(~3500 cal yr BP to the top of the core). The end of the Younger Dryas to the late Holocene (ITRAX2A) is characterised by reduced elemental counts compared to the previous zone, I and Cl reduce from the previous zone while Ni increases. Fe, Zr and Y reduce from 8272.8 495.9 and 203.8 to 4336.7, 339.3 and 146.1 respectively, Cu and Pb also reduce in this sub zone. In sub zone ITRAX2B, all the elements increase except Y which reduce to 105.5 from 146.1, Cu and Pb show the increase with Cu increasing from 0.26 to 0.79 and Pb increasing from 1.58 to 14.45.

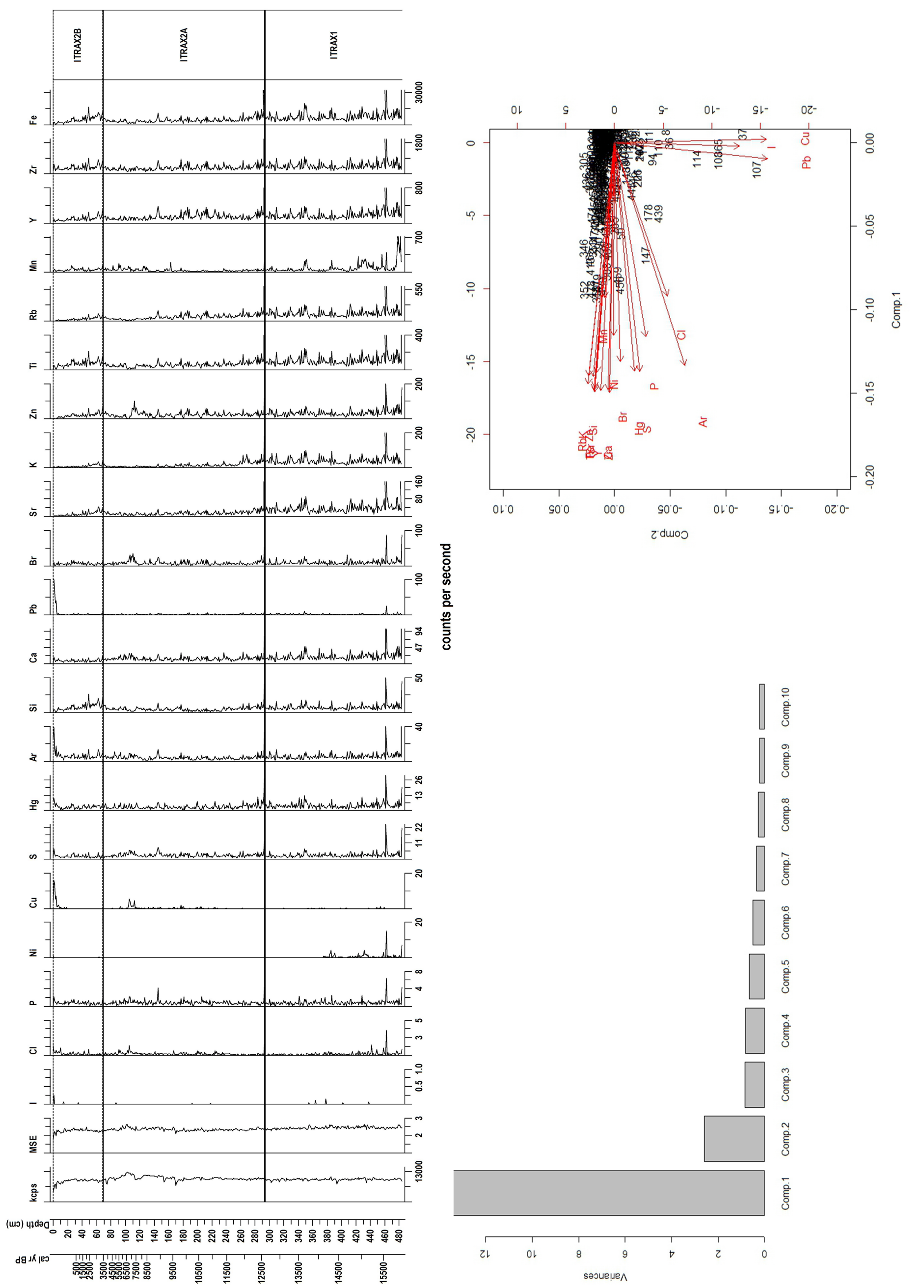



Figure 3.8: Nyabuiyabuyi elemental profile from ~16,000 cal yr BP divided into two significant zones, PCA graph showing the change in the proportion of variance contributed by the first 10 components and a PCA biplot showing the relationship between the elements and the components.

### 3.5.5 Macroscopic charcoal record

The mean charcoal concentration and charcoal accumulation rate (CHAR) from the Late Glacial period to the present is 217±6.4 pieces/cm$^3$ and 10.28±0.34 pieces/cm$^2$/yr$^{-1}$ respectively. There are several charcoal concentration peaks with the largest one at ~1600 cal yr BP of 1198 pieces/cm$^3$ while the highest CHAR peak is at ~11,400 cal yr BP 50.75 pieces/cm$^2$/yr$^{-1}$. The charcoal concentration and CHAR curves show 26 and 27 peaks respectively within the last 4~16,000 cal yr BP (Figure 3.9).

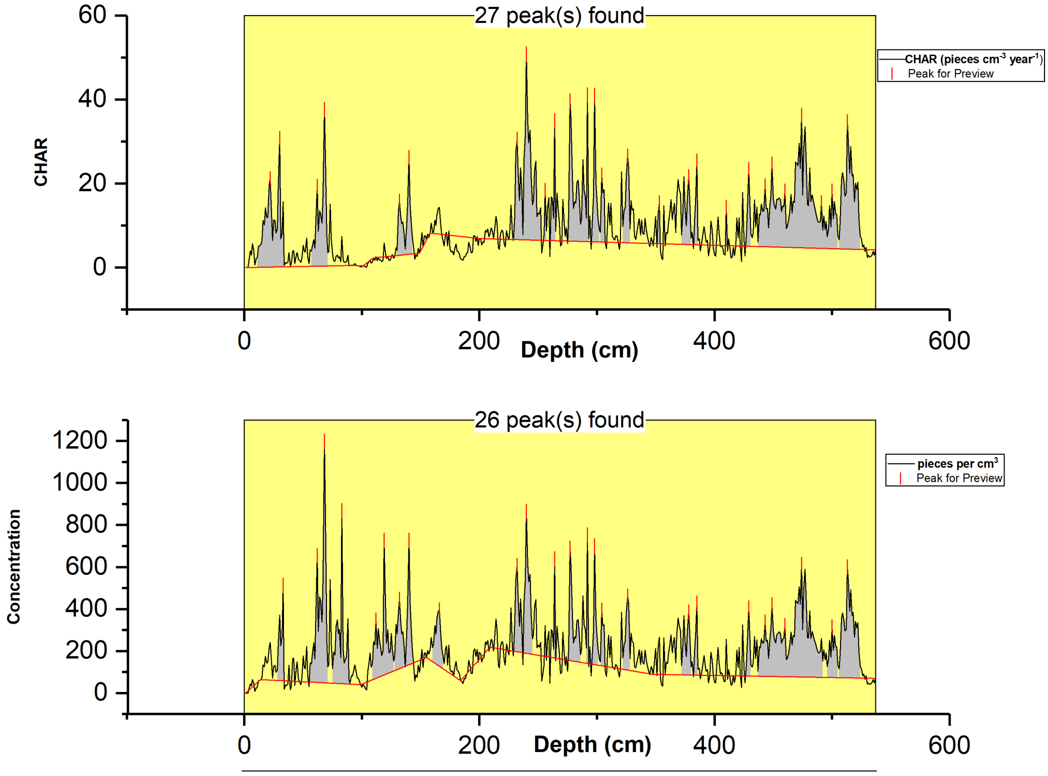



Figure 3.9: Charcoal concentration and charcoal accumulation rate (CHAR) values with the peaks identified.

Table 3.4: Pearson's correlation coefficients between macro charcoal, magnetic susceptibility and geochemical elements, numbers in bold highlight the significant relationships.

| | Char | Magsus | Si | P | S | Cl | Ar | K | Ca | Ti | Mn | Fe | Ni | Cu | Zn | Br | Rb | Sr | Y | Zr | I | Hg | Pb |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Char | 1 | -0.03 | 0.05 | -0.04 | -0.03 | -0.06 | -0.06 | 0.06 | 0.03 | 0.06 | 0.05 | 0.07 | -0.01 | -0.13 | 0.10 | - 0.003 | 0.07 | 0.04 | 0.07 | 0.05 | -0.06 | -0.02 | -0.12 |
| Magsus | -0.03 | 1 | 0.14 | -0.02 | 0.03 | -0.15 | -0.01 | **0.28** | **0.12** | 0.19 | 0.04 | 0.22 | 0.02 | -0.15 | 0.09 | 0.03 | 0.20 | 0.15 | 0.08 | 0.12 | -0.02 | 0.15 | -0.07 |
| Si | 0.05 | 0.14 | 1 | 0.63 | 0.72 | 0.45 | **0.78** | 0.83 | 0.82 | 0.92 | 0.51 | 0.92 | 0.60 | -0.12 | 0.82 | 0.70 | 0.83 | 0.85 | **0.84** | 0.91 | -0.02 | **0.80** | **-0.02** |
| P | -0.04 | -0.02 | 0.63 | 1 | **0.79** | 0.52 | **0.71** | 0.55 | 0.72 | 0.67 | 0.40 | 0.67 | 0.48 | 0.16 | 0.65 | 0.73 | 0.63 | 0.62 | **0.72** | 0.73 | 0.11 | 0.67 | 0.10 |
| S | -0.03 | 0.30 | 0.72 | **0.79** | 1 | 0.58 | 0.82 | 0.70 | 0.88 | 0.80 | 0.57 | **0.81** | 0.63 | 0.13 | 0.78 | 0.86 | **0.77** | **0.79** | **0.85** | 0.84 | 0.09 | 0.79 | 0.12 |
| Cl | -0.06 | -0.15 | 0.45 | 0.52 | 0.58 | 1 | **0.65** | 0.42 | **0.53** | 0.49 | 0.22 | 0.48 | **0.57** | 0.24 | 0.47 | 0.51 | 0.45 | 0.49 | 0.55 | 0.54 | **0.09** | **0.58** | **0.21** |
| Ar | -0.06 | -0.01 | [0.78 | **0.71** | 0.82 | **0.65** | 1 | 0.70 | 0.80 | 0.80 | 0.50 | 0.80 | **0.65** | 0.34 | 0.72 | 0.75 | 0.73 | 0.77 | 0.79 | 0.84 | 0.27 | **0.82** | 0.44 |
| K | 0.06 | 0.28 | 0.83 | 0.55 | 0.70 | 0.42 | 0.70 | 1 | **0.88** | **0.94** | 0.47 | 0.92 | 0.69 | **-0.14** | 0.84 | **0.68** | **0.96** | 0.92 | **0.87** | 0.89 | -0.02 | 0.82 | -0.02 |
| Ca | 0.03 | 0.12 | 0.82 | 0.72 | 0.88 | **0.53** | 0.80 | **0.88** | 1 | **0.92** | **0.70** | 0.90 | 0.72 | -0.02 | **0.88** | 0.85 | 0.90 | 0.94 | **0.92** | **0.90** | 0.01 | 0.82 | 0.01 |
| Ti | 0.06 | 0.19 | 0.93 | 0.67 | 0.80 | 0.49 | 0.80 | **0.94** | **0.92** | 1 | 0.56 | **0.98** | **0.72** | -0.13 | 0.89 | 0.77 | 0.95 | **0.95** | 0.94 | 0.097 | -0.02 | 0.86 | -0.01 |
| Mn | 0.05 | 0.04 | 0.51 | 0.40 | 0.57 | 0.22 | 0.50 | 0.47 | **0.70** | 0.56 | 1 | 0.57 | **0.47** | -0.03 | 0.51 | 0.58 | 0.47 | 0.62 | **0.53** | 0.48 | -0.01 | 0.43 | -0.01 |
| Fe | 0.07 | 0.22 | 0.92 | 0.67 | **0.81** | 0.48 | 0.79 | 0.92 | 0.90 | **0.98** | 0.57 | 1 | **0.69** | **-0.13** | 0.86 | 0.78 | 0.94 | 0.93 | 0.93 | 0.95 | -0.03 | 0.86 | -0.02 |
| Ni | -0.01 | 0.02 | 0.60 | 0.48 | 0.63 | **0.57** | **0.65** | 0.69 | 0.72 | **0.72** | **0.47** | **0.69** | 1 | -0.03 | 0.66 | 0.56 | 0.67 | **0.77** | 0.69 | **0.66** | -0.02 | 0.61 | 0.06 |
| Cu | -0.13 | -0.15 | -0.11 | 0.16 | 0.13 | 0.24 | 0.34 | **-0.14** | -0.02 | -0.13 | -0.03 | **-0.13** | -0.03 | 1 | -0.10 | 0.05 | -0.15 | -0.11 | -0.09 | -0.04 | 0.51 | **0.09** | 0.74 |
| Zn | 0.10 | 0.09 | 0.82 | 0.65 | 0.78 | 0.47 | 0.72 | 0.84 | **0.88** | 0.89 | 0.51 | 0.86 | 0.66 | -0.10 | 1 | 0.76 | **0.88** | 0.87 | 0.91 | **0.90** | -0.05 | 0.78 | -0.05 |
| Br | -0.03 | 0.03 | 0.70 | 0.73 | 0.86 | 0.51 | 0.75 | **0.68** | 0.85 | 0.77 | 0.58 | 0.78 | 0.56 | 0.05 | 0.77 | 1 | 0.74 | 0.77 | 0.83 | 0.81 | 0.01 | 0.69 | 0.01 |
| Rb | 0.07 | 0.20 | 0.83 | **0.63** | 0.77 | 0.45 | 0.73 | **0.96** | 0.90 | 0.95 | 0.47 | 0.94 | 0.67 | **-0.15** | **0.88** | 0.74 | 1 | **0.94** | 0.95 | **0.94** | -0.04 | 0.84 | -0.03 |
| Sr | 0.05 | 0.15 | 0.85 | 0.62 | **0.79** | 0.49 | 0.77 | 0.92 | 0.94 | **0.95** | 0.62 | 0.93 | **0.77** | -0.11 | 0.87 | 0.77 | **0.94** | 1 | **0.92** | 0.91 | -0.03 | 0.81 | -0.02 |
| Y | 0.07 | 0.08 | 0.83 | **0.72** | **0.85** | 0.55 | 0.79 | **0.87** | **0.92** | 0.94 | 0.53 | 0.93 | 0.69 | -0.09 | 0.91 | 0.83 | 0.95 | **0.92** | 1 | 0.97 | -0.02 | 0.84 | **-0.01** |
| Zr | 0.05 | 0.12 | 0.91 | 0.73 | 0.84 | 0.54 | 0.84 | 0.89 | **0.90** | 0.97 | **0.48** | 0.95 | **0.66** | -0.04 | **0.90** | 0.81 | **0.94** | 0.91 | 0.97 | 1 | 0.02 | 0.87 | 0.04 |
| I | -0.06 | -0.02 | -0.02 | 0.11 | 0.09 | **0.09** | 0.27 | -0.02 | 0.01 | -0.02 | -0.01 | -0.03 | -0.02 | 0.51 | -0.05 | 0.01 | -0.04 | -0.03 | -0.02 | 0.02 | 1 | 0.06 | 0.63 |
| Hg | -0.02 | 0.15 | 0.80 | 0.67 | 0.79 | **0.58** | **0.82** | 0.82 | 0.82 | 0.86 | 0.43 | 0.86 | 0.61 | **0.09** | 0.78 | 0.70 | 0.84 | 0.81 | 0.84 | 0.87 | 0.06 | 1 | **0.18** |
| Pb | -0.11 | -0.07 | -0.02 | 0.10 | **0.12** | **0.20** | 0.44 | -0.02 | 0.01 | -0.01 | -0.01 | -0.02 | 0.06 | 0.74 | -0.05 | 0.01 | -0.03 | -0.02 | **-0.01** | 0.04 | 0.63 | **0.18** | 1 |

### 3.5.6 Pollen and Non Pollen Palynomorph Production (NPP)

Pollen preservation varied down the Nyabuiyabuyi sediment core with >70 pollen types observed and recorded. The sample with the highest diversity of taxa had 68 pollen types identified while the sample with the lowest pollen taxa count had 18 pollen types identified with average diversity of 52 pollen taxa in each sample. For ease of interpretation and discussion they were grouped into vegetation types i.e. Afromontane, trees, shrubs, herbs and aquatics. Aquatic species include *Ludwigia, Nymphaeae*, Cyperaceae and *Typha*. Throughout the sediment core Afromontane pollen accounted for 29.8±1.04%, tree pollen accounted for 13.64±0.72%, shrubs accounted for 26.58±0.86%, herbs for 29.27±0.88% and unknown was <1%. No individual pollen taxa accounted for more than 8% as the percentage range of all the pollen identified was 0 to 8% (Figure 3.10). The NPP counts ranged from 10 to 582 throughout the sediment core with an average count of 112±17.36.

The pollen diagram was divided into three pollen zones using a b-stick and clustered hierarchical analysis (CONISS) (Figure 3.10). The lowermost pollen zone NBPOLL1 occurring between ∼16,500 and ∼12,700 cal yr BP had the highest species composition; Afromontane taxa accounted for 32.01±1.39% with *Apodytes, Celtis, Podocarpus* and *Prunus* dominating and *Hagenia* constantly present around 8%. The tree pollen accounted for 15.8±0.91%, shrubs were 27±0.86, herbs were 24±0.76 while unknown pollen was >1%. The Poaceae to aquatics ratio was 1:3. NPP counts were lowest in this zone ranging from 10 to 54 per $cm^3$ and averaging 30±3.08 per $cm^3$.

The second pollen zone NBPOLL2 (∼12,700 to ∼5200 cal yr BP) was further zoned into NBPOLL2A (∼12,700 to ∼10,200 cal yr BP) and NBPOLL2B (∼10,200 to 5200 cal yr BP). The average Afromontane, tree and shrub pollen coverage reduced from 32%, 15% and 24% to 29%, 11% and 24% respectively while the herbaceous pollen increased from 24% to 34%. The Poaceae to Cyperaceae ratio increased from 1:3 to 2:5 while the NPP counts significantly increased from an average of 30 per $cm^3$ to 102 per $cm^3$ ranging from 24 to 284 counts per $cm^3$.

NBPOLL2A was marked by a sharp decrease in the pollen abundance across all the vegetation types. The Afromontane taxa dominating (>10%) this time period were *Cordia, Croton, Draceana, Juniperus* and *Olea*. *Cordia, Juniperus* and *Olea* were constant while the other Afromontane taxa peaked and dropped. *Allophylus, Annona, Maerua, Monoetes, Myrsine, Rhus* and *Schefflera* peaked in this zone between 10 and 20%. Most of the shrub and herbaceous taxa from the previous zone decreased or disappeared except Asteraceae, *Cissus, Cyathula, Erica, Justicia*, Liliaceae, Malvaceae, *Narcissus*, Solanaceae and Urticaceae which also significantly increased from <5% to 10% and 20%. The aquatic taxa significantly increased from between 5% and 10% in NBPOLL1 to 10-20%, Poaceae also drastically increased. This decrease in composition and abundance indicates a dry period that did not allow a diverse array of shrub and herbaceous taxa also reducing fuel availability thus fewer, smaller fires. NBPOLL2B (∼10,200 to 5200 cal yr BP), showed a marked increase in pollen, spore, NPP composition and abundance. The increase in composition and abundance was constantly present through this humid early to mid-Holocene period which gradually started to decrease. There was a decrease in Cyperaceae and Poaceae levels with little variation until a peak at ∼6400 cal yr BP.

The third zone NBPOLL3 (∼5200 cal yr BP to present), was significantly drier than the rest of the sediment core, terrestrial taxa including Afromontane trees, woody shrubs reduced in composition and abundance to <5%. The average Afromontane pollen taxa reduced to 25% from 29% in the previous zone, tree pollen increased from 11% to the previous levels of 15%, shrubs were at their highest coverage here at 29% compared to 27% in NBPOLL1 and 24% in NBPOLL2 while the shrubs reduced to 28% from 34% which was still higher than 24% at NBPOLL1. The NPP counts increased to their highest level of an average of 343±37 ranging from 201 to 582 pieces/$cm^3$. This zone was dominated by *Cordia, Croton, Cupressus, Ficus, Juniperus, Olea* and Pinus between 5 and 10%. *Cupressus* and *Pinus* appeared towards the top of the sediment core with a high abundance as they were recently introduced taxa at commercial scale. Acanthaceae, Asteraceae, Brassicaecae, *Tarchonanthus* and Malvaceae dominate at 8 to 10% while Cyperaceae, *Typha* and Poaceae experienced their highest abundance.

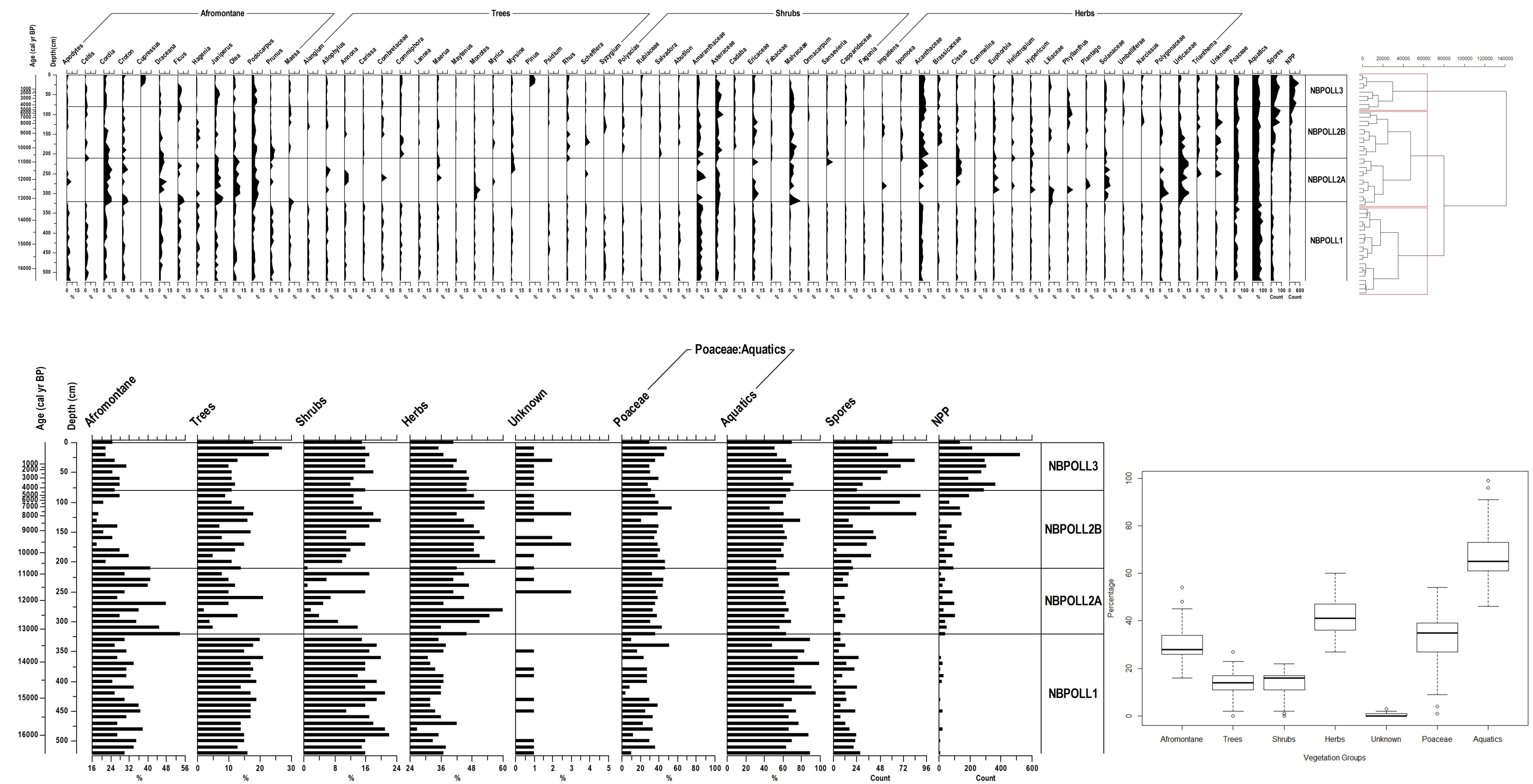



Figure 3.10: Vegetation change since ∼16K cal yr BP zoned into three significant zones using a constrained hierarchical cluster analysis, the top graph includes all taxa identified, while the bottom graphs the vegetation groups for a better comparison, pollen taxa values are % concentration while the spore and NPP curves are counts. A boxplot illustrates the abundance level of each vegetation group.

### 3.5.7 Sediment properties

The LOI results show that the sediment organic matter ranges from 0 to100% with an average organic matter content of ∼21%, the carbonate content ranges from 0 to 40% with an average of 64% while the water content ranges from 34 to 90% with an average of ∼60%. A stratigraphically constrained cluster analysis divided the record into three significant zones (Figure 3.11).

Zone LOI1 with a bottom date of ∼16,600 cal yr BP at 530cm and a top date of ∼11,500 cal yr BP at 240 cm was composed of ∼55% water, and the rest of the dry sediment was ∼12% organic matter, five percent carbonate content and 81% siliciclastic, it also showed the lowest organic matter and carbon content variation with a standard deviation of 4.6% and 5.9% respectively. Loss on ignition values from zone LOI2 (∼411,300 cal yr BP to ∼7900 cal yr BP at 120 cm) show an increase in average organic matter, carbonate content and water content to 20% and 8.7% and 66% respectively while also experiencing the highest variability of carbonate content. The siliciclastic content drops to 71%. The top zone LOI3 from ∼7500 cal yr BP to present has the highest average organic matter content at 42% while experiencing the highest variability (standard deviation of 27%), the carbonate content reduces to 4.8% while the water content and siliciclastic component further reduce to is at 63% and 53%.

The water content, organic matter content and carbonate content in the sediment core were positively correlated. Organic matter and water content have the strongest positive correlation of r = 0.55, the water content and carbonate matter have a weaker positive correlation of r=0.34, and the organic matter and carbonate content have the weakest positive correlation of r=0.04, the siliciclastic content shows a strongly negative correlation with water content (r=-0.63), organic matter (r=-0.79) and carbonate content (r=-0.65) (Table 3.5). A PCA analysis shows that the first two components account for ∼88% of the variance and the organic matter, water content and carbonate content are negatively correlated to the first components (Figure 3.11).

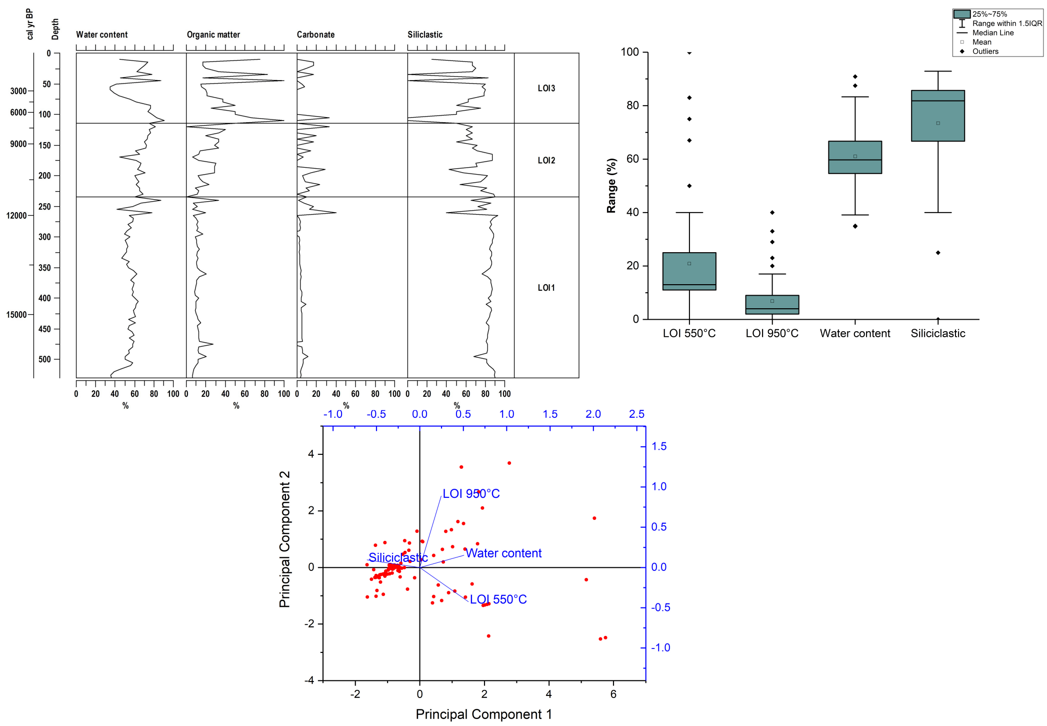



Figure 3.11: Sediment properties values zoned into three significant zones using a constrained hierarchical cluster analysis.

Table 3.5: Pearson's correlation coefficients of the water, organic matter, carbonate and siliciclastic content at 5 cm resolution, significant values are in bold.

| | Water content | Organic matter | Carbonate | Siliciclastic |
|---|---|---|---|---|
| Water Content | 1 | **0.551** | **0.344** | **-0.631** |
| Organic matter | **0.551** | 1 | 0.434 | **-0.791** |
| Carbonate | **0.343** | 0.043 | 1 | **-0.646** |
| Siliciclastic | **-0.632** | **-0.791** | **-0.646** | 1 |

### 3.5.8 Particle Size Analysis

Nyabuiyabuyi sediment is mainly composed of silt- sand-clay; the silt content is the highest with an average of 72.44±1.13% ranging from ∼34% to 86%, followed by sand with an average of 17.84±1.13% ranging between 2% and ∼63% and finally clay with an average of 9.63±0.39% and ranging between ∼1.6% and 19% (Figure 3.12). The silt component experiences the highest variance throughout the sediment core. The smaller sized soil components i.e. clay, very fine silt and fine silt all with a diameter <0.016$\mu$m are strongly positively correlated with Pearson's correlation coefficients between 0.57 and 0.92 (Table 6) and negatively correlated with all the other components with a larger diameter than 0.016$\mu$m.

The bottom zone (PSA1) extends from ∼16,600 cal yr BP at 530 cm to ∼10,300 cal yr BP at 195 cm and is comprised of an average of 73±1.37% silt which is the highest average of the three zones, 16.48±1.63% sand and 9.6±0.43 clay%. The silt content ranges between ∼42% and 85% while the sand ranges between two percent and ∼55% and the clay which ranges between ∼3% and ∼18% and the highest variance is experienced in the sand levels. The highest sand peak is experienced around ∼16,500 cal yr BP with a peak of 55%, the highest silt peak is around ∼13,800 cal yr BP at 86% and the highest clay peak is experienced around ∼13,300 cal yr BP at 18%.

Table 3.6: Pearson's correlation coefficients of the particle size analysis, significant values are in bold.

| | Clay | V. fine silt | Fine silt | Medium silt | Coarse silt | V. fine sand | Fine sand | Medium sand | Coarse sand | V. coarse sand | V. fine pebbles |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Clay | 1 | **0.92** | **0.57** | -0.10 | **-0.64** | **-0.73** | **-0.55** | **-0.39** | -0.16 | -0.11 | -0.08 |
| V. fine silt | **0.92** | 1 | **0.82** | 0.11 | **-0.66** | **-0.86** | **-0.68** | **-0.52** | **-0.23** | -0.17 | -0.15 |
| Fine silt | **0.57** | **0.82** | 1 | **0.60** | **-0.35** | **-0.90** | **-0.85** | **-0.70** | **-0.35** | **-0.25** | **-0.21** |
| Medium silt | -0.10 | 0.11 | **0.60** | 1 | **0.49** | **-0.46** | **-0.72** | **-0.67** | **-0.42** | **-0.30** | **-0.25** |
| Coarse silt | **-0.64** | **-0.66** | **-0.35** | **0.49** | 1 | **0.47** | **-0.01** | **-0.20** | **-0.26** | **-0.21** | -0.17 |
| V. fine sand | **-0.73** | **-0.86** | **-0.9** | **-0.46** | **0.47** | 1 | **0.83** | **0.50** | 0.11 | 0.04 | 0.04 |
| Fine sand | **-0.55** | **-0.68** | **-0.85** | **-0.72** | -0.01 | **0.83** | 1 | **0.84** | **0.58** | 0.38 | 0.30 |
| Medium sand | **-0.39** | **-0.52** | **-0.70** | **-0.67** | **-0.20** | **0.50** | **0.84** | 1 | **0.58** | **0.89** | **0.62** |
| Coarse sand | -0.16 | **-0.23** | **-0.35** | **-0.41** | **-0.25** | 0.11 | **0.29** | **0.58** | 1 | **0.89** | **0.62** |
| V. coarse sand | -0.11 | -0.17 | **-0.25** | **-0.30** | **-0.21** | 0.04 | 0.16 | **0.38** | **0.89** | 1 | **0.88** |
| V. fine pebbles | -0.08 | -0.15 | **-0.21** | **-0.25** | -0.17 | 0.04 | 0.15 | **0.30** | **0.62** | **0.88** | 1 |

The second significant zone PSA2 from $\sim$10, 000 cal yr BP to $\sim$5800 cal yr BP experiences a reduction in silt and clay percentages to $\sim$70% and $\sim$7% respectively however there is an increase in the sand content to $\sim$23%. The silt in this zone ranges from $\sim$34% to $\sim$83%, while the clay content ranges between two percent and $\sim$13% and the sand ranges between seven percent and $\sim$63%. The highest sand peak in this zone is around $\sim$7000 cal yr BP of 63%, the highest silt peak is around $\sim$9000 cal yr BP of 84% while the highest clay peak is around $\sim$9300 cal yr BP of around $\sim$13%.

The final zone PSA3 covers the period from the mid Holocene from $\sim$5300 cal yr BP at 90 cm to the top of the core is comprised of $\sim$69% silt, $\sim$17% sand and 13% clay, this is a significant increase in the clay content from 6% to 12%. Over the last $\sim$5300 cal yr BP the silt content ranges between $\sim$51% and 81%, sand content ranges between 6% and 41% while the clay content ranges between $\sim$6% and $\sim$19% and experiences less variation than the silt and sand content. The highest sand peak in this zone occurs around the year 2004 at 41% while the highest silt peak occurs around $\sim$4100 cal yr BP with 81% and the highest clay peak around $\sim$1800 cal yr BP with 19% clay content. The clay and silt particles are negatively correlated to the first component of the PCA and positively correlated to the second component (Figure 3.12).

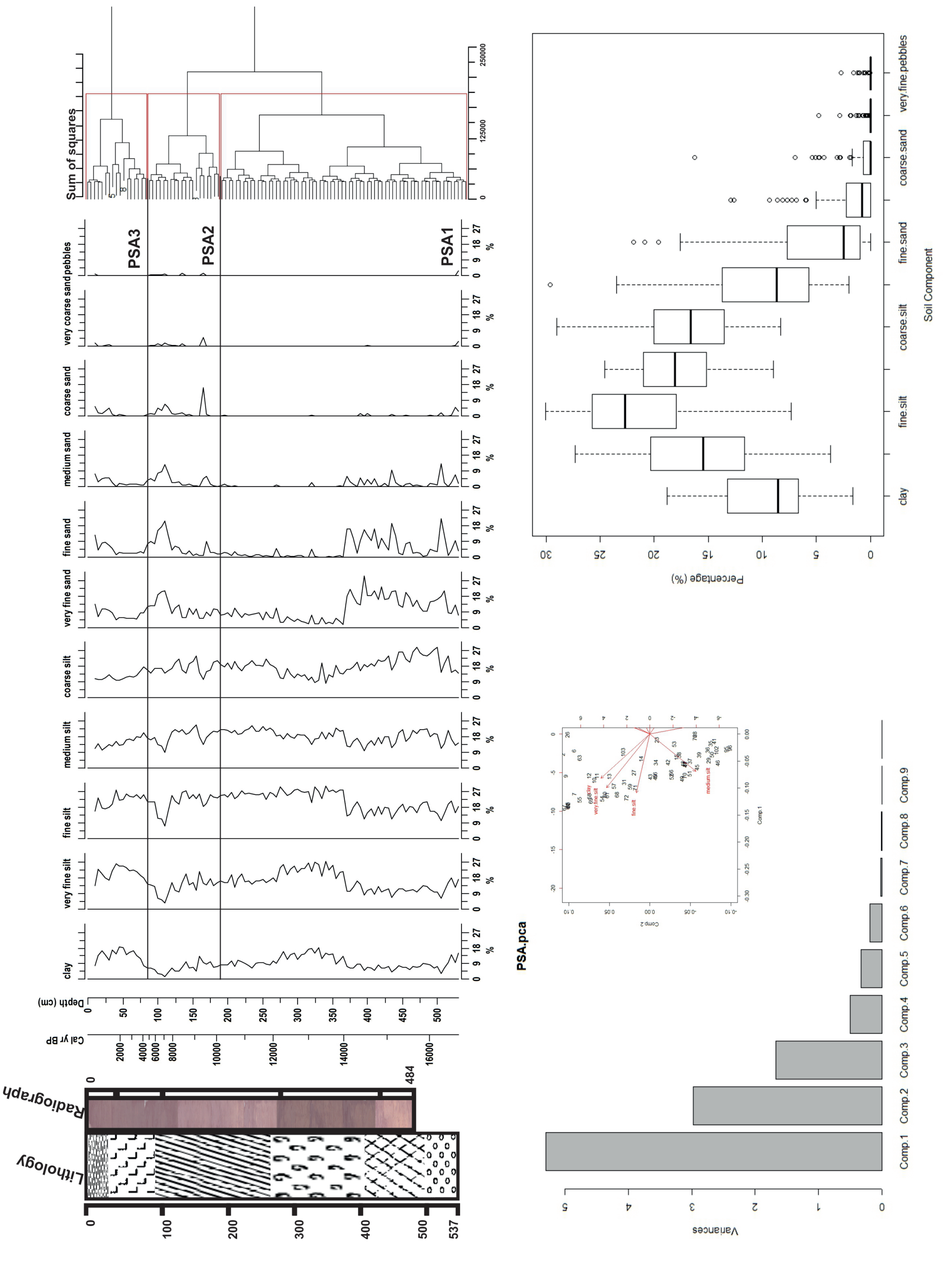



Figure 3.12: Particle size distribution throughout the sediment core and a principle component analysis of the proportion of variance explained by 10 components with the first two explaining 75% of the total variance.

### 3.5.9 Synthesis of the proxies

The PCA scores from the proxies i.e ITRAX, magnetic susceptibility, pollen, macro charcoal, loss on ignition and particle size are are displayed as a biplot (Figure 3.13), each of the proxies cluster analysis grouping is displayed to enable comparison of the zones and the alignment is based on the sediment depth To judge the sources of sediments and distinguish the natural and anthropogenic inputs on environmental data three factors were found to explain ∼55% of the variance identified. Factor 1 which accounts for 30% is characterised by high positive loadings of Rb, Ti, Fe, K, Sr and magnetic susceptibility, with high negative loadings of water content, P, Cl and organic matter. The association of the major elements with metals and magnetic susceptibility indicates the detrital origin of the terrigenous sources. Factor 2 accounts for 14% and is characterised by high positive loadings of Pb, Ar, I and Cu and high negative loadings of P, Br, S and Y. This factor could represent anthropogenic origin and hence the high positive loadings of Pb, Ar, I and Cu associated with increased human activity. Factor 3 accounts for 11% and is characterised by high positive loadings of sand, Ni, Mn and Ca and high negative loadings with Al, Zr, silt and clay. The association of sand and the elements shows their implies their similar detrital origin, while the association of the association of the clay and silt is just a function of size, however the association of Al and Zr shows their affinity to fine sediment.

Across the six proxies studied, each was divided into three zones, the first zone from the bottom of the sediment core broadly covers the period ranging from the Late Glacial to the Younger Dryas (∼16,600 cal yr BP at 530 cm to ∼12,400 cal yr BP at 292 cm), the second zone ranges from ∼12,400 at 291 cm to ∼5700 cal yr BP at 104 cm and the final zone ranges from ∼5600 cal yr BP to the present (Figure 3.13). The pollen, charcoal and elemental zones are very similar with the first significant zone ending at 13,000±500 cal yr BP and the second significant zone ending at ∼3000 cal yr BP. The particle size analysis and loss on ignition zones were closely delineated, the first significant zone ended around ∼10,300 cal yr BP at 195 cm and ∼11,500 cal yr BP at 240 cm while the second significant zone ended at ~5200 cal yr BP at 90 cm and around ~7500 cal yr BP at 115 cm.

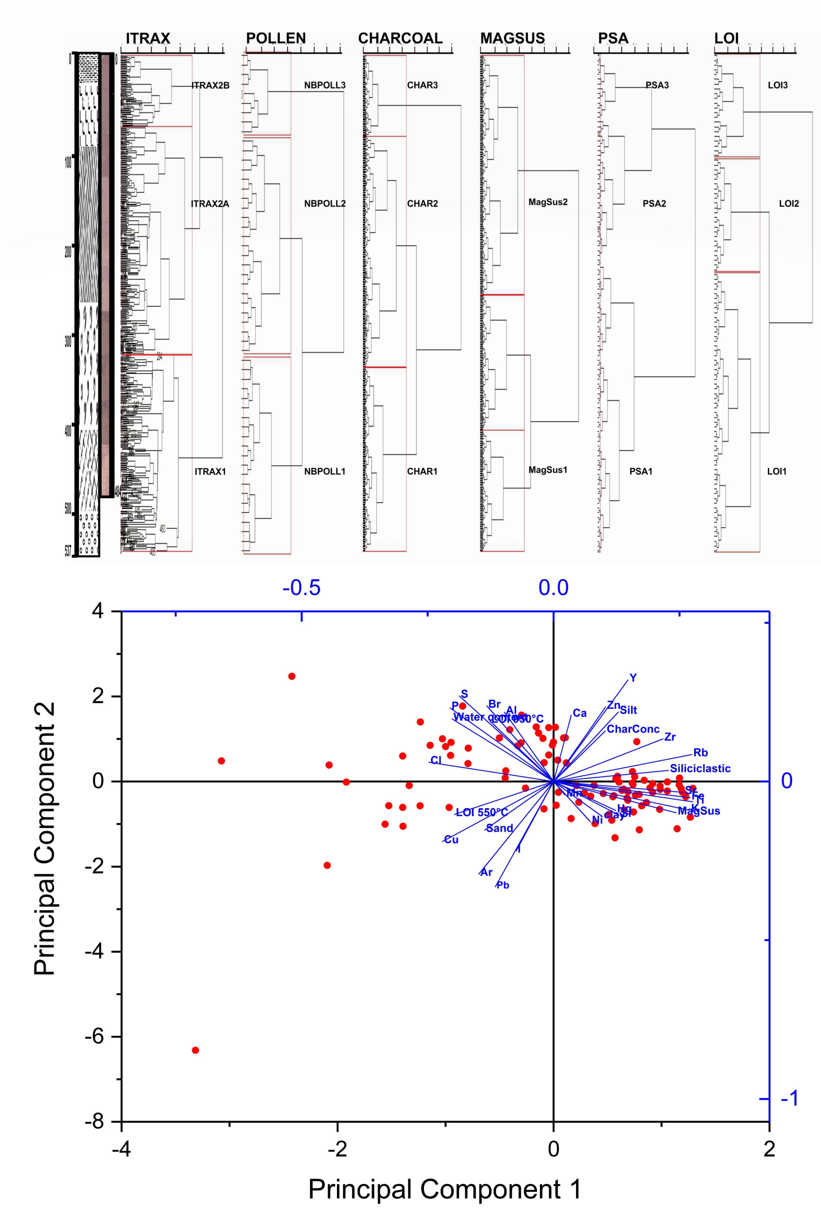



Figure 3.13: ITRAX, pollen, charcoal, MagSus, PSA and sediment properties CONISS zones aligned by the depth to provide a visual comparison of the zones across the proxies. A biplot of the PCA scores of the proxies.

# 3.6 Discussion

The Nyabuiyabuyi Swamp record covers the period from the Late Glacial (∼16,000 cal yr BP) to the present providing insight into the dynamics of the swamp as well as the wider Eastern Mau forest. A potential hiatus is experienced during the Holocene, a trend which is quite common in East African palaeorecords (Finch et al. 2014, Street-Perrott et al. 2007, Verschuren 1999). The changes in stratigraphy can be used to identify the occurrence, timing and magnitude of events e.g. deforestation or major climatic events such as drought (Cohen et al. 2007, Kiage & Liu 2009*b*, Lejju et al. 2005). The changes in the elemental profiles, magnetic susceptibility, pollen, charcoal, PSA and LOI in the Nyabuiyabuyi record are discussed in three time periods: the Late Glacial period, the Early to Mid Holocene and the Late Holocene.

## 3.6.1 Late Glacial Period to Early Holocene development of the Eastern Mau (∼16,600-11,700 cal yr BP)

Most of East Africa experienced cool dry conditions during the Late Glacial period between ∼21,000 cal yr BP and ∼12,000 cal yr BP (Olago 2001, Sonzogni et al. 1998). The Nyabuiyabuyi sediment in this zone exhibits the lowest organic matter content (∼12±0.5%), carbonate content (∼5±0.25%) and water content (∼54±0.9%). There is a higher sand component which decreases towards ∼13,000 cal yr BP accompanied by an increase in the clay and silt levels, the sediment in this level is laminated indicating little disturbance during and after sedimentation.

At Nyabuiyabuyi, the Late Glacial period is the wettest period with the highest arboreal cover while the herbs and grasses are at minimal frequencies. The vegetation is highly diverse during the Late Glacial evidenced by the number of pollen types and abundance. *Apodytes, Celtis, Draceana, Hagenia* and *Podocarpus* dominate and there is a turnover around ∼13,000 cal yr BP to an Afromontane ecosystem dominated by *Cordia, Croton, Ficus, Juniperus* and *Olea* thus changing from a predominantly cold, dry Afromontane mosaic similar to Rukiga Highlands, Ruwenzori, Kashiru Swamp, Lake Albert, Mt. Elgon as well as Mt. Kenya (Beuning et al. 1997, Bonnefille & Riollet 1988, Hamilton 1987, Taylor 1990). This period experiences the most significant fires with high macro charcoal concentration implying a continuous connected source of fuel. *Juniperus*, a pioneer species invading gaps after a fire thus leading to forest regeneration and establishment of broad-leaved species such as *Olea* as has been observed in Mt. Kenya and the South Aberdare Range (Bussman 2001) and in Nyabuiyabuyi its peak lags behind *Hagenia* peaks. These findings indicate high moisture availability adequate to sustain a richly abundant Afromontane forest. The palaeoemagnetic signals are highest around 13,400 cal yr BP at a time of high sedimentation (Figure 3.7).

A record from the Middle Atlas in Morocco was dominated by a herb-grassland assemblage characteristic while in Central Sahara maximum aeolian activity correlated with aridity peaks around this period (Olago 2001) of low moisture. Western and Central African palaeo records are dominated by signals inferring cool, dry conditions such as grassland expansion, forest taxa changes to semi-deciduous forest taxa and a drop in lake levels as well as low diatom content which is also observed in Eastern African records (Chalié & Gasse 2002, Mumbi et al. 2008, Van Zinderen Bakker & Coetzee. J.A. 1988). Globally there was a reduction in sea level rise during the Late Glacial period however abrupt changes in climate and sea levels inhibited coastal settlement (Pope & Terrell 2008). The Lake Challa record indicates increased precipitation due to the intensification of the South-Easterly Indian Ocean Monsoon from $\sim$16,500 cal yr BP to the early Holocene and only interrupted by the Younger Dryas between $\sim$13,300 cal yr BP and 11,700 cal yr BP (Verschuren et al. 2009) which contrasts with the other records. The Rumuiku record correlates with the Empakai record indicating a regional response (Rucina et al. 2009), Rumuiku stratigraphic changes as well as increases in Cyperaceae, Poaceae and *Typha* indicate low water levels.

The Late Glacial period up to $\sim$12500 cal yr BP show enriched trends in all the geochemical elements compared to the rest of the record, a dry phase between $\sim$12,500 cal yr BP and 11,700 cal yr BP is observed and macro-charcoal concentration increases while micro charcoal counts decrease, this implies increased local fire compared to fire from outside the swamp catchment. This is accompanied by a decrease in pollen diversity and abundance with a lot of trees and shrub taxa occurring at >5% and an increase in Poaceae thus a much more open forest. This is similar to the Mt. Kenya sites (Sacred Lake, Lake Rutundu and Small Hall Tarn) where there was an expansion of the grassland (Street-Perrott et al. 2007). There is increased sedimentation of silt and clay with a drop in sand as well as higher magnetic susceptibility. Peaks in magnetic susceptibility coincide with peaks in Si, Ti, Fe, Rb and Sr, which are detrital elements and peak with silt as well. This may indicate higher surface run off conditions after episodes of heavy or continuous rainfall. Throughout the Nyabuiyabuyi record, the macrocharcoal shifts and the pollen zone shifts have a close correspondence however the shifts do not occur at the same time (Figure 3.9 and Figure 3.10), as the charcoal zonation breaks occur before the pollen ones, the fire regime during the Late Glacial Period is characterised by a high charcoal concentration and charcoal accumulation rate that drops during the Younger Dryas.

At Nyabuiyabuyi, the Late Glacial Period leading up to the Younger Dryas can be described as cool and dry with periods of high intensity rainfall that lead to peaks in sedimentation detrital elements, silt and sand, experiencing a much higher a lot more fire with a charcoal concentration average higher than the average of the whole sediment core. During the Younger Dryas, there is a significant drop in pollen abundance (Figure 3.10) as it gets dryer; there is a drop in the concentration of detrital elements as well as less sand in the sediment composition.

### 3.6.2 The Holocene (from ∼11,700cal yr BP until the potential hiatus from about ∼8000 cal yr BP to ∼3000 cal yr BP)

The post glacial period is marked by wet conditions in East Africa likely reflecting increased moisture transport as well as increased sea surface temperatures from the Indian Ocean (Tierney et al. 2008, 2011). This post-deglacial interval is systematically warm and wet across the majority of northern, tropical, and southern Africa, associated with a major transgression in East African rift lakes (Gasse 2000, Haberyan & Hecky 1987, Trauth et al. 2003) and vastly wetter conditions in the Sahara (Hoelzmann et al. 2004) and Kalahari (Burrough et al. 2009). Lake Rukwa experiences a high stand until 7000 cal yr BP evidenced by reduced magnetic susceptibility and magnesium (Mg) content and a low stand from 7000 cal yr BP to 3000 cal yr BP where the was an increase in magnetic susceptibility as well as Mg and increased grain size (Thevenon et al. 2002). Lake Kivu and Tanganyika increased in volume experiencing maximum lake levels between 10,000 cal yr BP and 7000 cal yr BP similar to Lake Rukwa and become more dilute with a transition to higher salinity around ~5000 cal yr BP as water level reduced (Haberyan & Hecky 1987). The early Holocene high stand at Lake Abiyata in Ethiopia was determined by decreased conductivity and conductivity together with a diatom flora that prefer deep water (Chalié & Gasse 2002).

At Nyabuiyabuyi the low Ti, Fe and Mn levels (Fe and Mn indicate redox) between ~11,700 cal yr BP and ~10, 000 cal yr BP indicate increasing dryness (Burrows et al. 2016). There is a marked decrease in pollen taxa identified and abundance until ~10,000 cal yr BP (Figure 3.10) while the Poaceae pollen increases implying a warm and dry period. The forest becomes increasingly open and there are fewer fires within the area as the micro charcoal decreases, the macro charcoal concentration levels increase which could be because there are more fires further within the landscape. After ~10, 000 cal yr BP, there is a slight increase in all the taxa observed. The subalpine forest association of *Hagenia-Juniperus* is interpreted as an early succession after disturbance (Bussman 2001) such as fire as observed around ~8500 cal yr BP. There is an increase in *Apodytes, Celtis, Olea Podocarpus* and *Erica* as well as organic matter content. On Mt. Kilimanjaro the warm and wet climate enables development and expansion of the Afromontane forest (Schüler et al. 2014). There are fewer fires within the Nyabuiyabuyi swamp catchment implied by lower macro charcoal concentration, the magnetic susceptibility fluctuates but always below eight. This period experiences the highest increase in organic matter and decrease in carbonate content, accompanied by an increased silt and clay content as well as increase bulk density implying high sedimentation. The increased organic matter content peaking with sulphur content could indicate that the increased sedimentation is anoxic (Tierney & Russell 2007). High C/N ratios (>20%) which gradually decrease indicate the autochthonous source of sediment in the swamp (Burrows et al. 2016). The detrital elements (Si, Ti, Fe, Rb and Sr), reduce and plateau out with increase in organic matter content. The decline in elemental counts may reflect the ecosystem drying up; less precipitation to cause sedimentation through runoff would account for the very low counts of TI, Fe, Mn and elevated Br levels (Burrows et al. 2016). An increase in Ti, Fe and Zn reflect a much wetter period than previously, this elements peak and drastically reduce implying a wet period that rapidly ends. The dark brown sediment from 350 cm to the top have no laminations, this could indicate sediment reworking, resuspension or benthic fauna mixing up the sediments as observed in the Lake Tanganyika sediment (Haberyan & Hecky 1987).

There is a likely hiatus in the Nyabuiyabuyi sedimentary record between ~8000 cal yr BP and ~3,000 cal yr BP, this covers the African Humid Period in which several East African lakes reached their highest levels for example the Zwai-Shalla palaeolake in Ethiopia reached its overflow level between ~9500 cal yr BP and ~8000 cal yr BP (Telford & Lamb 1999) followed by the arid period recorded from ~4000 cal yr BP (Van Zinderen Bakker & Coetzee. J.A. 1988). Climate in East Africa was generally warm and wet (Bonnefille & Riollet 1988) until ~ 4000 when it gets drier and there is more open grassland (Garcin et al. 2012, Msaky & Davis 2005). Many East African sites are characterised by sedimentary hiatuses around this period increasing difficulty in understanding the changes in the ecosystem occurring in those gaps.

### 3.6.3 The Late Holocene

During the Late Holocene, East African highland ecosystems such as Mt. Kenya experienced pronounced ecosystem shifts (Rucina et al. 2010) such as increased arid conditions punctuated with dramatic lake level fluctuations. A continent wide arid period (Marchant & Hooghiemstra 2004), is noted beginning ~4000 cal yr BP and East African lake levels fluctuate revealing periods of rainfall and aridity (Cohen et al. 2005, Gasse 2000, Verschuren et al. 2000). In the Nyabuiyabuyi record the increasing arid conditions begin from 100 cm and there is continued replacement of Afromontane taxa with other arboreal taxa as well as increased grassland cover (Figures 3.12 and 3.15), the intense loss of Afromontane and arboreal taxa could be a signal of the widespread clearance recorded in several montane forest sites such as Rukiga Highlands in Uganda and Rumuiku Swamp on Mt. Kenya (Mumbi et al. 2008, Rucina et al. 2009, Van Zinderen Bakker & Coetzee. J.A. 1988) attributed to farming activities across Eastern and Central Africa. Transformations of the landscape are hypothesised to have reached significant levels globally ~3000 cal yr BP (Ellis et al. 2013), data from East Africa indicates a shift in the adoption of livelihood patterns, for example as indicated by Wright (2007) while exploring resource exploitation among Neolithic hunters and herders. The last ~3500 cal yr BP show an increasing trend in all the elements with spikes in Cu, S, Hg, Ar and Pb over the last ~200 cal yr BP.

There was an increase in aquatic taxa and Poaceae indicative of swamp development due to water levels within the last ~3000 cal yr BP. The NPP consistently increase up the core suggesting an intensification in the use of the wetland by herbivores which would occur during an arid period (Gelorini et al. 2012). Decline in high-altitude forest taxa observed from a Lake Bogoria record (Kiage & Liu 2006), increase in drought tolerant taxa on Mt. Kenya and Mt. Elgon (Hamilton 1982, Vincens 1986) indicated the establishment of arid conditions. Progressive tree cover degradation could be due to increasing aridity coupled with human impact which was become severe over the last century (Bessems et al. 2008, Gelorini et al. 2012, Kiage & Liu 2009*b*). Forest composition was characterised by high levels of *Cordia, Cupressus, Hagenia, Pinus* and *Podocarpus* with increased levels of Asteraceae, Acanthaceae and *Vernonia* that could be due to increased disturbance. The charcoal concentration and accumulation rate reduce within the last ~3000 cal yr BP. The carbon and nitrogen ratios fluctuate over the last 3000 cal yr BP, nitrogen levels remain low >2% however there are C:N ratio peaks above 20% indicating influx of organic matter from terrestrial sources or emergent taxa (Haberyan & Hecky 1987, Mumbi et al. 2008).

Several severe arid events are recorded within the between ~2500 cal yr BP and ~1800 cal yr BP from diatom and leaf wax evidence with lakes Turkana, Tanganyika, Naivasha experiencing low levels (Cohen et al. 2005, Verschuren et al. 2000). At Nyabuiyabuyi wetland there was a drop in all the elemental units except Cl, Ar, Cu, Hg and Pb as a result of reduced input of terrigenous material. Increased Cl is often as a result of precipitated chloride from NaCl during dry conditions from lake sediment (Kristen 2010).

## 3.7 Conclusions

Using a multi-proxy palaeoecological approach, this study described the vegetation history since the Late Glacial period and analysed the geophysical factors related and driving the vegetation changes (i.e. soil physical and geochemical factors and fire regime). The results of the present study reveal variations in sediment properties with time and support the use of geochemical in addition to magnetic susceptibility parameters on wetlands to understand changing environmental conditions of recent past. The analysis revealed that the Nyabuiyabuyi catchment has undergone significant changes in forest composition from a highly diverse Afromontane to a more open forest ecosystem indicative of increased aridity in the region. The MFC Late Glacial record closely agrees with other upland East African records covering the same period such as the Mt. Kenya, Mt. Elgon, Rukiga highlands and Ruwenzori mountains. The pollen record does not provide an indication of recent human activity despite the deforestation known to have occurred since 1890s. Thus climate variations have been the major driver of changed in vegetation and fire regime and the human activity signal is not very explicit.

The major limitations were the uncertainties in the radiocarbon dates observed and further dating of the chemical extract such as humin/humate for comparison would increase confidence in the age depth model developed. Destruction of pollen traps left in the field meant that a modern pollen analysis for the establishment of a modern analogue for comparison could not be achieved. The Mau Forest Complex hold a great deal of information as it occupies a vast geographical area experiencing different local climate and use by local communities. Further studies on other wetlands located along the complex would greatly improve our understanding of this ecosystem. Distinguishing climate driven and human induced changes was difficult, anthropogenic deforestation hypothesised to have been signifcant ~3000 cal yr BP cannot be observed in the pollen records however that does not mean it did not occur.

The analyses suggest that climatic change mainly precipitation drives the change in vegetation composition through time and additional effects of fire are do not seem to have a significant impact in the Mau Forest Complex. The increased loss of Afromontane taxa and fire activity is attributed to first the occupation of this ecosystems, where use of fire is an important land management technique both for clearing land to settle on and also making the landscape suitable for livestock. However within the MFC, increased logging is recorded during the colonial period during for commercial purposes. Use of large trucks to transport the felled trees to the milling sites could be a potential source of the increased copper and lead content in the sediment record.

# CHAPTER 4

## Ecosystem dynamics across the Amboseli savannah since the mid Holocene.

## 4.1 Overview

This chapter provides a summary of the palaeoecological work undertaken in the Amboseli basin. The semi-arid landscape of Amboseli, Southern Kenya, East Africa, is punctuated with intermittent groundwater-fed wetlands that record past ecosystem changes in the sedimentary geoarchives. Pollen, NPP and macroscopic charcoal analyses were performed on four radiocarbon-dated sediment cores collected from Esambu, Kimana, Ormakau and Enkongu Narok swamps to provide an environmental history since 5000 cal yr BP.

The chapter follows the same format used in the previous chapter and include a description and background of the study site, details of sample collection and analysis, core description including lithostratigraphy, construction of age models and the proxy composition of each sediment record. The discussion will provide a temporal and spatial overview of the Amboseli landscape which can be used to understand the drivers of change and provide information to inform sustainable management.

The aim of this chapter is to develop new palaeoecological records that inform us how the vegetation composition, fire history and biophysical process have changed over time, whether there are significant changes within the landscape and if the changes are observed across the landscape (different sites as well as the timing of the changes between the sites. This is particularly important as the Amboseli landscape is an important wildlife habitat that supports migratory wildlife which one of the main sources of income for the Kenyan and Tanzanian governments. Understanding the responses of the wetlands to changes in the driving factors is necessary to inform sustainable wetland management that provide crucial dry season grazing refuges for the wildlife.

## 4.2 Introduction

Ecosystems are dynamic with interacting abiotic and biotic components combining over multiple spatiotemporal scales to determine the composition, structure and distribution (Gillson & Marchant 2014). Rapid ecosystem changes across the globe, particularly during the Anthropocene, are primarily attributed to anthropogenic modifications and are superimposed on long-term climatic and landscape-scale changes (Dearing et al. 2010, Foley et al. 2013, Smith & Zeder 2013, Young 2014). To understand and uncouple the various abiotic and biotic factors, it is crucial to have a long-term perspective on ecosystem change and understand how entangled interactions between the environment, ecosystems and humans influence current ecosystem state and possible future trajectories (Gillson & Marchant 2014, Marchant & Lane 2014).

Savannah ecosystems cover 70 to 80% of East Africa and are characterised by temporal and spatial variations in availability of water and vegetation resources. The savannahs are of great conservation value due to their ability to support vast numbers of large herbivores that are of significant ecological and economic benefit to the local economies. Large herbivores impact on wetland morphology (Deocampo 2002), the interaction between grassy and woody vegetation cover ratios influence herbivore type, number and migration across the Amboseli basin (Dublin et al. 1990, Western & Maitumo 2004). It has been suggested that removal of elephants and fire causes a high grass biomass that leads to hot fires, which kill tree seedlings. Elephants are able to remove the few establishing tree seedlings thus maintaining the grassland while in a woodland state trees shade the grass and remove water, thus reducing grass biomass and fuel for hot season fires (African Wildlife Foundation & Maasai Mara National Reserve 2009).

Fires are an important controlling factor within savannah ecosystems with their impact being largely controlled by the interaction between climate variability, rates of primary productivity, distribution of fuels, and human land use (Andela & van der Werf 2014, Colombaroli et al. 2014, Van Der Werf et al. 2008). Fire occurrence is limited by fuel availability (Andela & van der Werf 2014); in savannah ecosystems this is driven by precipitation patterns that interact with fuel accumulation and fuel distribution. East

African seasonal fires within the savannah were mainly controlled by fuel production and distributions, dry season variability, and ignitions before the influence of humans. Burning has been documented as a traditional land management practice globally to stimulate vegetation for grazing, clear land for cultivation, kill disease causing vectors such as ticks and clear bushes to improve plant biodiversity (Kamau & Medley 2014). Currently fires in the Amboseli area occur during July and are likely human-caused ignitions and limited by fuel abundances, conditions and connectivity.

There have been relatively few palaeoecological studies in southern Kenya, particularly due to the extensive coverage of xeric savannah ecosystems, woodland and scrub vegetation with limited suitable sedimentary basins (Gillson 2004*b*, 2006, Rucina et al. 2010). Within the Amboseli area, there are several perennial wetlands intermittently distributed across the predominantly semi-arid landscape. These groundwater-fed wetlands, recharged from orographic precipitation falling on Mount Kilimanjaro, form an important wildlife refuge and provide a series of 'stepping stones' for animal migrations between Amboseli National Park and neighbouring Tsavo and Chyulu Hills National Parks. Thus, these wetlands enhance ecological resilience in the landscape and wider ecosystem functioning especially as human population pressure increases in the area.

Ephemerally wet areas of arid ecosystems at wetland margins can have higher biodiversity than nearby permanent water bodies and their resilience is impacted to an unknown degree when impacted by land use changes such as agriculture (Casanova & Powling 2014). Unlike seasonally rain-fed wetlands, that are inundated during the wet season and dry during the dry season, the Amboseli wetlands sustain locally high water-tables that support grazing lawns and long grasses and enable peat accumulation at small pools, which persist through drought periods. The more ephemerally-moist fringes of the wetlands and the surrounding ecosystems are sensitive to hydrologic and climatic variability, and are drastically responsive to changing human land use practices and land cover modifications (Casanova 2012, MEMR 2012).

The current savannah ecosystem is characterised by a high interaction of climatic variability, intense human land use systems and wildlife utilisation making it difficult to decipher signals of climatic versus human disturbance, particularly as palaeoecological sites are rarely located near archaeological sites that would provide spatiotemporal correlation of evidence (Lejju et al. 2005). The Amboseli sites under study will provide data points, filling in the spatial gap in our understanding of the palaeoenvironment however archaeological evidence is still needed to improve our interpretation of human activities responsible for impacts identified.

## 4.3 Study region

The Amboseli is approximately 3000 km$^2$ (Western 1975) and is characterised by a semi-arid savannah experiencing bimodal rainfall distribution caused by the movement of the ITCZ (Hulme 1996, Swift et al. 1996) (Figure 4.1). The Amboseli Wetland Basin is located in Kajiado South County (formerly Loitokitok District) along the northern base of Mt. Kilimanjaro and borders Tanzania to the southwest, Kajiado District to the north, Kibwezi District to the east and Taveta District to the south (Figure 4.1). The area is generally arid to semi-arid with limited variations in its agro-ecological zones.

### 4.3.1 Climatic, geologic and hydrologic setting

Climate in Amboseli is characterised by high average annual temperatures 23°C with little annual variability and a bimodal precipitation regime characterised by short rains falling from October to December and long rains falling from March to May (Figure 4.2). November and April are the two wettest months with an average of 66 and 110 mm of rainfall respectively, July is the driest month with an average of 1 mm. Average annual precipitation is 586 mm year$^{-1}$ although with considerable intra-annual variation (range 226 to 990 mm year$^{-1}$) from 1979 to 2009 (Figure 4.2).

The hydroclimate of East Africa is tightly linked to the meridional passage of the ITCZ, the seasonally-migrating zone of maximum solar insolation where northern and southern hemisphere trade winds converge. Moisture in East Africa is largely derived from the Indian Ocean (Marchant et al. 2007) and rainfall patterns are teleconnected with the El Niño, with positive El Niño years typically being wetter than non El Niño years (Goddard & Graham 1999). Locally, the precipitation distribution is moderated

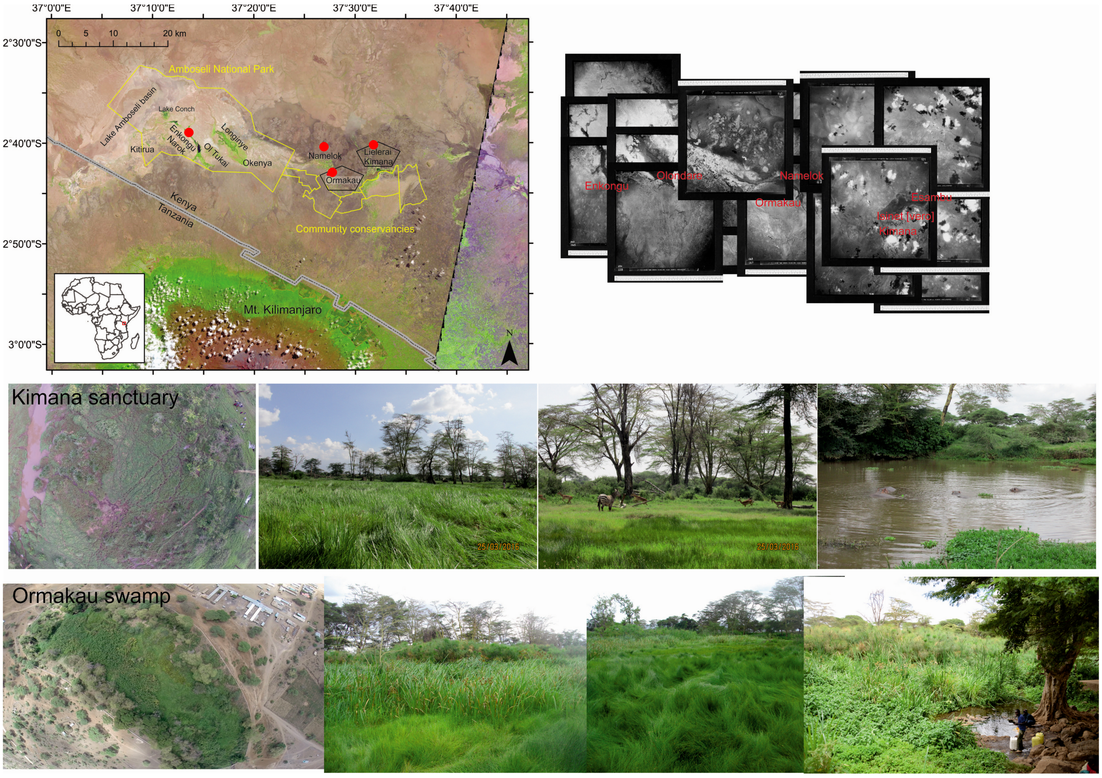



Figure 4.1: A. Location of the Amboseli ecosystem in southwestern Kenya , within the greater Amboseli ecosystem the four wetlands cores are marked by red dots while the community conservancies (yellow delimitations) create a corridor westward towards Amboseli National Park. Base LandSAT images LC81680622014LGN00. B and C. Orthomosaic images of the aerial views of Kimana and Ormakau wetlands respectively taken using an unmanned aerial vehicle (March 2014). Vegetation cover shown as relative abundances of the 25 plant species observed and their location on the wetlands and aerial images obtained from 1690 and 2017.

by the topography of Mount Kilimanjaro that orographically uplifts air masses resulting in localized precipitation, often falling as snow above 3000 m. Precipitation falling on Mount Kilimanjaro recharges groundwater flows into the lowlands of Amboseli (Altmann et al. 2002, Meijerink & Wijngaarden 1997, Sarkar 2006). Surface runoff percolates into the basaltic bedrocks with the resulting groundwater flowing towards the Amboseli Basin that maintain the wetlands via perennial springs during dry periods; floodwater from ephemeral channels also recharge the wetlands during the rainy

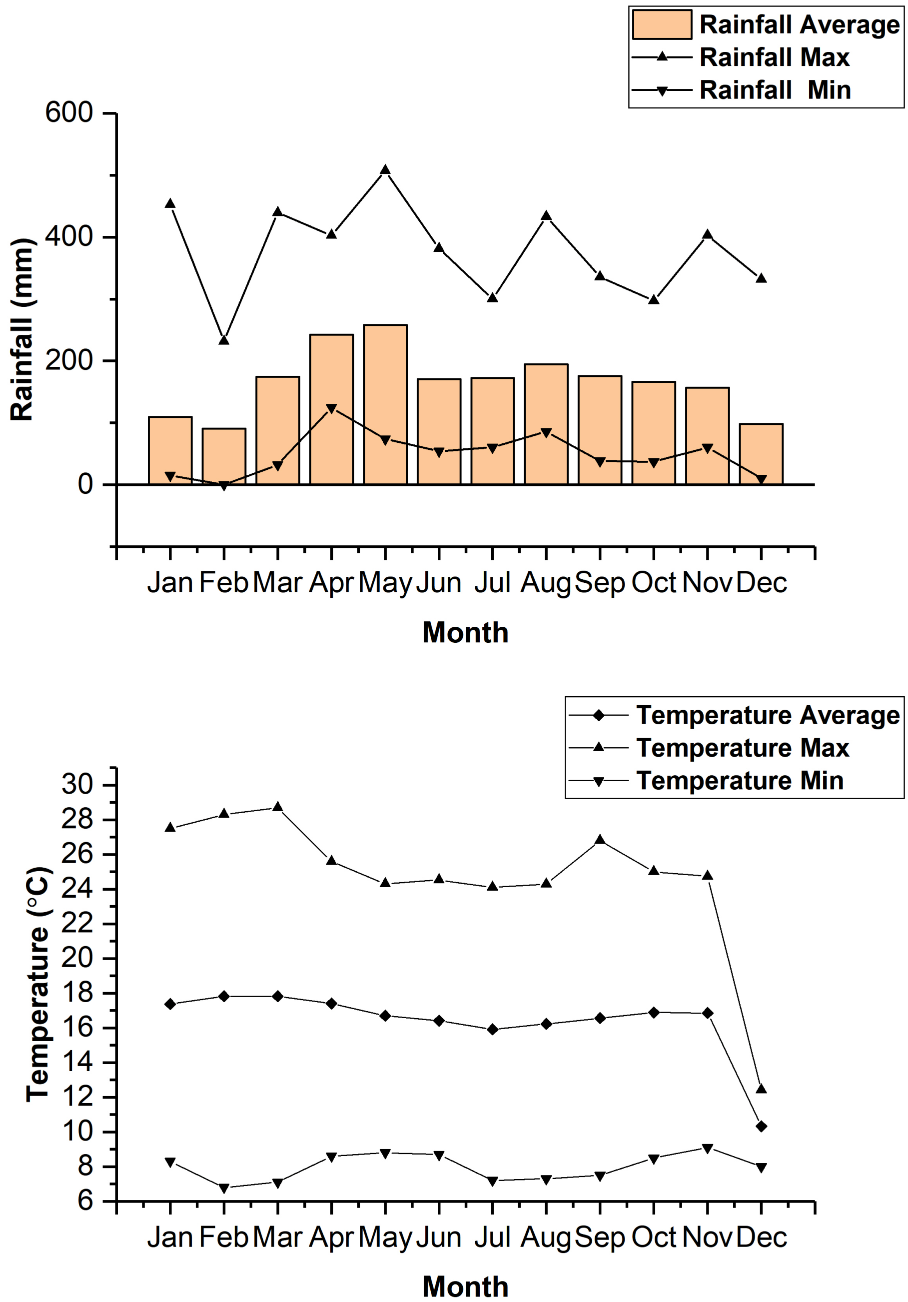



Figure 4.2: Average, minimum and maximum monthly precipitation and temperature values from 1979-2009 recorded from Makindu meteorological station (Kenya Meteorological Department), which is ∼60 km west of the study site.

seasons. To the east of the area is Lake Amboseli (1125 m a.s.l.) that floods during extreme wet seasons although it has remained relatively dry since the 1960s.

The arid to semi-arid Amboseli landscape (Figure 4.1) is characterised by five main components. The northern gently undulating basement plains on deep, acidic, well drained soils. The central palaeolacustrine basin and swamps, which is a depression zone overlaying strongly alkaline, poorly drained, clayey soils dominated by short grasses such as *Sporobolus* spp. The volcanic foot slopes of Mount Kilimanjaro with deep, well drained neutral soils of primarily Pleistocene age (Meijerink & Wijngaarden 1997) that receives more precipitation and supports denser tree cover than the surrounding areas. The east and northeast are characterised by volcanic landscapes with abundant small volcanic cones of lapilli to block sized, reddish brown scoria pyroclasts and shallower soils with semi-arid woodland and scrub. Further eastwards towards the Chyulu Hills there is denser woodland savannah on very young basaltic parent material. Vegetation in the region is characterised by sparse shrubland savannah and *Acacia-Commiphora* dominated dry woodland-lowland savannah covering an area of 8500 $km^2$ (Western 1976*a*,*b*). Riparian wooded areas follow along semi-permanent and seasonal channels and groundwater and surficial flows drain north from Mount Kilimanjaro and diverge westward into the Amboseli Basin and eastward to converge with the Galana River.

The Amboseli ecosystem includes the Amboseli basin to the South and the wetlands to the East with water feeding the wetlands coming from aquifers, except for Esoitpus that is fed by Lolterish River. Hydrological studies carried out on the basin suggest Lake Amboseli and Ol' Tukai wetlands overlay Pleistocene lacustrine and fluviatile deposits, whereas the Enkongu Narok, Longinye, Kimana, Namelok, Lenkir and Esoitpus wetlands overlie volcanic rock, primarily basalt (Irungu 1992). These wetlands have critical functional roles for local and regional communities as well as the expansive wildlife community.

### 4.3.2 Study sites

Within the Amboseli landscape, there are around 7 swamps i.e. Enkongu, Olondare, Ormakau, Namelok (also known as Engumi), Isinet, Kimana and Esambu within an area of $\sim$618km$^2$. Enkongu which is the farthest to the west at an altitude 1136 m a.s.l and Esambu farthest to the east at an elevation of 1196 m a.s.l. Olondare and Enkongu (both located inside Amboseli National Park), Isinet, and Esambu swamps were cored in 2009 by the National Museums of Kenya paleobotany section team. The remaining two sites Kimana and Ormakau were thus selected as the two remaining study sites well as analysing the Esambu and Enkongu sites. The area around Isinet swamp, around 540 km$^2$ was sub divided in1994 and the swamp has been completely drained for cultivation.

#### 4.3.2.1 Esambu Swamp

Esambu Swamp (1191 m a.s.l.) is a $\sim$0.4 km$^2$ wetland associated with the Kikarankot River within the Lielerai-Kimana wetland complex that is used to supply water to an area of 12,000 km$^2$ with a population of $\sim$100,000 people (Wetlands International 2009) (Figure 4.1 and 4.3). Kikarankot is one of two permanent rivers in the Amboseli landscape, flowing eastward from Kimana wetland through Esambu and Ilchalai swamps (Okello & Kioko 2011) toward the south of Chyulu Hills. Esambu supports a 1.16 km$^2$ irrigation scheme while Ilchalai supports a 4.0 km$^2$ irrigation scheme (Amboseli Ecosystem Stakeholders 2009). The vegetation cover within Esambu Swamp is dominated by Cyperaceae-Poaceae with *Cyperus rotundus, C. papyrus* and *Typha. Acacia xanthophloea* dominates the riverine arboreal taxa. Farther from the wetland the semi-arid woodland is sparsely vegetated with *Acacia spp., Balanites, Commiphora, Cynodon* and *Euphorbia*.

Large areas of the wetland have been cultivated Figure 4.3 (B) to (E) as water is extracted for furrow irrigation of fruits and vegetables. Pastoralist populations with historical formal and informal land tenure have recently reorganised into more sedentary group ranches whereby some groups and individuals maintain pastoral livelihoods on an increasingly fragmented landscape under a more formal legal framework. People have migrated into the region and the increased population have undertaken varying degrees of modification to the Amboseli wetlands (MEMR 2012, Sarkar 2006). Land cover and land uses include commercial agriculture, particularly expansive in and around the wetlands, pastoralism across the landscape, rural urbanisation and infrastructure development, and various protected area management types, including National Parks, community and private game reserves.

#### 4.3.2.2 Lielerai Kimana Swamp Sanctuary

The Kimana wetlands are part of the larger Amboseli wetland system. Kimana Swamp is located at 2°44.930' S, 37°30.922' E at an elevation of 1221 m asl, covering 6400 ha and is channel fed by the Kimana River that is hydrologically sourced by natural springs recharged from the Chyulu Hills to the east and the northern slopes of Mt. Kilimanjaro 40 km to the south-southwest (Figure 4.1). The surface runoff percolates into the basaltic bedrocks and the groundwater flow toward the low elevation Amboseli Basin are the sources for the springs.

The Kimana basin is flat with depressions scattered across it, Kimana River flows south to north through the sanctuary and floods during the short rains and merges hydrologically with Marura Swamp. Local soils are Inceptisols, suborder tropepts. Lielerai, derived from *Olerai*, is a Maa word meaning the *Acacia xanthophloea* tree and Kimana means 'continuous circle' or 'something going around'. Immediately surrounding the wetlands is open woodland of trees, woody shrubs and grasses (Table 4.1). Kimana wetlands are dominated by Cyperaceae and Poaceae. Kimana swamp functions as an important dry season refuge and wildlife dispersal region connecting animals to other migration corridors outside the boundaries of Amboseli National Park. Lielerai Kimana Swamp supports a number of large mammals such as elephants, giraffes, large cats, zebras, ungulates, jackals, and elands, and a resident breeding population of hippopotami. The swamp has been the focus of various conservation and tourist industry claims since 1996 when the United States Agency for International Development financed the Kimana Community Wildlife Sanctuary.

Table 4.1: Plant species identified at Kimana.

| Grouping | Family, subfamily | Species |
|---|---|---|
| Trees | Fabaceae | *Acacia xanthophloea* |
| | Convovulaceae | *Ipomoea arborescens* |
| | Arecaceae | *Phoenix roebelenii* |
| Shrubs | Apocynaceae | *Tabernaemontana elegans* |
| | Malvaceae | *Abutilon mauritanium* |
| | Fabaceae | *Senna* spp. |
| Herbs | Lamiaceae | *Ocimum aratissimum* |
| Grasses | Poaceae | *Cynadon dactylon* |
| | Poaceae | *Digitaria ciliaris* |
| | Poaceae | *Pennisetum clandestinum* |
| Sedges | Cyperaceae | *Cyperus laevigatus* |
| | Cyperaceae | *Cyperus papyrus* |
| | Cyperaceae | *Cyperus rotundus* |
| | Juncaceae | *Juncus* sp. |

#### 4.3.2.3 Ormakau Swamp

Ormakau Swamp is a Cyperaceae-Poaceae dominated swamp located at 2°43.166’ S, 37°27.329’ E, at an elevation of 1170 m asl, in Namelok, Kajiado South County (Figure 4.1 and Table 4.2). Ormakau Swamp currently occupies an area of 0.95 $km^2$, which is entirely enclosed by a stone wall erected in early 2014 with financial assistance from the African Development Bank and Ministry of Agriculture. The wall is designed to deter livestock and wildlife but also any unsanctioned human activity from occurring within the boundary to protect the water that is used to irrigate the surrounding plots. The swamp is sustained by five springs opening from the basaltic bedrock with local inceptisol soils.

There are numerous small gullies within the steeper hypsographic gradients near the swamp and there is one ephemeral river channel entering from the east side of the swamp. The springs are fed from sources on the northern slopes of Mt Kilimanjaro, located 40 km to the south-southwest. Ormakau is a Maa word meaning ”hippopotamus” and Namelok means ”sweet place”. Currently, the two main swamps in Namelok are

Ormakau and Engumi, named after two perennial rivers. Engumi and Ormakau Rivers flow from Mount Kilimanjaro and drain eastwards towards the Chyulu Hills. A sediment study at Engumi Swamp based on a core that provided a 3100 year record showed that the vegetation in the area had responded to changes in climate and land use activities by humans and herbivores (Rucina et al. 2010). As of 2003, over 50% of the original Namelok swamp had been converted to agriculture.

Table 4.2: Plant species identified at Ormakau.

| Grouping | Family, subfamily | Species |
|---|---|---|
| Trees | Fabaceae | *Acacia xanthophloea* |
| | Fabaceae | *Acacia brevispica* |
| | Fabaceae | *Acacia mearnsii* |
| | Fabaceae | *Acacia torilis* |
| | Fabaceae | *Albizia* spp. |
| | Fabaceae | *Senegalia senegal* |
| | Sterculiaceae | *Dombeya burgessiae* |
| Shrubs | Apocynaceae | *Tabernaemontana elegans* |
| | Solanaceae | *Solanum incanum* |
| | Myrtaceae | *Syzygium* spp. |
| | Malvaceae | *Abutilon mauritanium* |
| Herbs | Amaranthaceae | *Amaranthus hybridus* |
| | Amaranthaceae | *Achyranthes aspera* |
| | Asteraceae | *Aspilia pluriesta* |
| | Lamiaceae | *Leonitis mollissima* |
| | Lamiaceae | *Ocimum aratissimum* |
| | Lamiaceae | *Plectranthus barbatus* |
| | Malvaceae | *Pavonia patens* |
| | Salvadoraceae | *Azima tetracantha* |
| Grasses | Poaceae | *Cynadon dactylon* |
| | Poaceae | *Pennisetum clandestinum* |
| Sedges | Cyperaceae | *Cyperus laevigatus* |
| | Cyperaceae | *Cyperus papyrus* |
| | Cyperaceae | *Cyperus rotundus* |
| | Juncaceae | *Juncus* sp. |
| | Polygonaceae | *Polygonum* spp. |
| | Polygonaceae | *Rumex usambarensis* |
| | Typhaceae | *Typha latifolia* |

#### 4.3.2.4 Enkongu Narok

Enkongu Narok is a sedge dominated wetland located at 2°42’ 16.8” S, 37°15’38.8” E at 1151m asl in Namelok, Kajiado South County (Figure 4.1 and Table 4.3) and occupies an area of ∼1.6 km$^2$. Enkongu Narok means ’eye of the waters’, it is located within the Amboseli National park close to the base of the Amboseli Baboon Research Project which has been running since 1971. It is a spring fed permanent swamp (Meijerink & Wijngaarden 1997). It overlays deep well drained and neutral soils from lava flows and partly reworked pyroclastic deposits of Pleistocene age. It is surrounded by bushed grassland dominated by various *Acacia* spp.

Table 4.3: Plant species identified at Enkong’u Narok.

| Grouping | Family, subfamily | Species |
|---|---|---|
| Trees | Fabaceae | *Acacia xanthophloea* |
| | Fabaceae | *Acacia brevispica* |
| | Fabaceae | *Acacia mearnsii* |
| | Fabaceae | *Acacia tortilis* |
| | Salvadoraceae | *Salvadora persica* |
| Shrubs | Acanthaceae | *Barleria* spp. |
| | Malvaceae | *Abutilon mauritanium* |
| | Malvaceae | *Grewia* spp. |
| | Verbanaceae | *Lantana camara* |
| Herbs | Lamiaceae | *Ocimum aratissimum* |
| Grasses | Poaceae | *Cenchrus ciliaris* |
| | Poaceae | *Pennisetum clandestinum* |
| Sedges | Cyperaceae | *Cyperus laevigatus* |
| | Cyperaceae | *Cyperus papyrus* |
| | Cyperaceae | *Cyperus rotundus* |
| | Juncaceae | *Juncus* sp. |

### 4.3.3 Historical and current land use

Competition for land tenure and water resources in Kajiado district has a long history. In 1891, the Maasai herds were almost wiped out by the rinderpest pandemic and, in the following year, there was a severe smallpox epidemic(Gillson 2006). The disruption caused by heavy stock and population losses was completed by the consequent outbreak of widespread inter-sectional raiding (Waller 1990). In 1902 the abundance of wildlife in Amboseli in conjunction with colonial 'tribal' containment/land grabbing policies led to the creation of the Southern Maasai Reserve, which was expanded in 1911 to 38,000 $km^2$ to accommodate Maasai who were being resettled from the abolished Northern Reserve.

Within the Southern Reserve, land was communally owned by Maasai and administered by Kajiado County Council, the Amboseli landscape is located within the Southern Reserve. While disputes did arise over access to grazing areas, particularly during droughts, water was not a majorly contested resource as the Amboseli area provided large, year-round open sources in the forms of swamps. Pressures directing people toward more organised land ownership structures came from the 1945 National Parks Ordinance that established within the Southern Reserve the 3,260 $km^2$ Amboseli National Park in 1974. It become the Amboseli Game Reserve for wildlife conservation in 2005 after President Mwai Kibaki transferred control from the Kenya Wildlife Service to the Olkejuedo County Council and its residents, the Maasai tribe. This is still being contested in the courts because of its implications that could jeopardize Kenya's other national parks. Post-Independence settlement of Kikuyu and Kamba cultivators on the lower slopes of Kilimanjaro and around the swamps of Amboseli fuelled Maasai land annexation anxieties. This led to the division into group ranches from as early as the 1950s, and was operating by the late 1960s (Rutten 1992). The sub-division of group ranches in Kajiado District is a complex issue by rising populations, migration, and ultimately expansion of cultivation and settlement down the ecological gradient of Mt. Kilimanjaro and onto wetter rangeland areas. Thus, an overall trend of landscape fragmentation centred on water abundant areas is evident. During drought periods in 1960 to 1961, 1973 to 1976, 1980, and 1984, pressure was put on Maasai, who saw huge reductions in herd numbers, to join or sell/lease plots of land to non-Maasai groups.

A survey carried out across the Amboseli wetlands to compare water quality along land-use classes identified the main land use types having the highest negative impact on the wetlands (Githaiga et al. 2003). Expanding agriculture (irrigation) was found to have the highest negative impact on water quality and quantity through competition with other water users and chemical contamination from fertilizer/pesticide use. Namelok was identified as representing the highest level of human impact from irrigated agriculture and wildlife impacts while Kimana was identified as experiencing intermediate levels of impact from irrigation, livestock and wildlife.

Conservation and management of wetlands has become an emerging priority due to recent intense drought years such as the 2009 drought and rapid increases in human population and land use pressures (Courtney Mustaphi et al. 2014, MEMR 2012, Rutten 1992, Waller 1985). Many of these wetland ecosystems have experienced management initiatives with mixed outcomes such as defaunification or conservation, pastoral use agreements, water resource infrastructure construction, abandonment and removal, drainage and conversion to agriculture. A ten year management plan was developed through participation between Amboseli Basin stakeholders that builds on the previous management plan (1991 to 1996) that did not fully consider wildlife dispersal corridors and a longer-term perspective (Amboseli Ecosystem Stakeholders 2009). Understanding how ecosystems have evolved under these changing interactions of climate variability, human land use and land cover, wildlife use, and fire activity through time and across space is vital to providing an informed context for management. This study presents palaeoecological evidence of vegetation and fire regime changes during the late Holocene. These data provide evidence of climatic and hydrological variability influences on vegetation and disturbances within the Esambu wetland ecosystem.

# 4.4 Methods

## 4.4.1 Field Methods

Vegetation surveys, sediment depth profiling, sediment coring and placement of pollen traps were undertaken at Lielerai Kimana Swamp on 29 March and 1 April, 2014, Ormakau Swamp from 30 to 31 March and 2 April. Two cores were retrieved from Ormakau swamp ORM 1 which was 121cm and ORM 2 which was 304cm, the Lielerai Kimana swamp core was 384 cm long. The Esambu and Enkongu Narok cores were retrieved in 2009; the Esambu core was 247.5 cm and the Enkongu Narok core was 195 cm.

A hand-pushed, D-shaped Russian corer was used to core the sediment in 50 cm sections from proximal, overlapping, parallel boreholes through a small amount of standing water before being transferred to plastic pipes and wrapped in aluminium foil. Cores were transported and refrigerated at the Palynology and Palaeobotany Section, National Museums of Kenya (NMK), Nairobi.

## 4.4.2 Laboratory Analysis

### 4.4.2.1 Radiocarbon Dating

Bulk sediment samples were picked from each of the cores for radiocarbon dating at Direct AMS, the NERC Radiocarbon Facility-East Kilbride and the SUERC AMS Laboratory for $^{14}C$ analysis. Bulk sediment samples were selected when there were no other macrofossils present to pick for dating except one *Acacia* spp. wood fragment from the Esambu sediment. The samples were selected from levels with visible stratigraphic changes such as colour and textures that could be an indication of change in the sedimentation process. Six samples were selected from the Esambu sediment core, seven samples were picked from the Kimana core, six from the Ormakau core and six from the Enkongu site. The IntCal13 curve (Reimer et al. 2013) was used to calibrate the dates and presented in calibrated year BP (AD 1950) with the year 1950 chosen as it was around the time the first 14C dates were obtained (Birks et al. 2012). An age-depth model was developed for each site using BACON (Blaauw 2010, Blaauw et al. 2007).

The priors set were different for each age-depth model, for Kimana the minimum depth was set as 0 cm and maximum depth as 384 cm, the accumulation mean (acc.mean) was changed to 2yr/cm as suggested by BACON during the analysis, a run of cal BP and calibrated c-14 dates were used to run the model with 77.5 sections, acc.shape:1.5, mem.strength:4 and mem.mean:0.7 to produce the age-depth model. Enkongu priors were set as minimum depth = 0 cm, maximum depth = 195 cm, acc.shape:1.5, acc.mean:20, mem.strength:4 and mem.mean:0.7. Ormakau priors were set at minimum depth = 0 cm, maximum depth = 304 cm, acc.shape:1.5, acc.mean:10, mem.strength:4 and mem.mean:0.7 and 61.5 sections using a a run of cal BP and calibrated c-14 dates. Esambu age-depth model was developed with minimum depth = 0 cm, maximum depth = 247.5 cm, acc.shape:1.5, acc.mean:20, mem.strength:4 and mem.mean:0.7 and 49 sections using a a run of cal BP and calibrated c-14 dates. An age-depth model was developed for Namelok, a wetland within the Amboseli ecosystem which has been studied so as to confirm the age estimates for comparison with the new sites, the Namelok age-depth model priors were; minimum depth = 0 cm, maximum depth = 400 cm, acc.shape:1.5, acc.mean:10, mem.strength:4 and mem.mean:0.7 and 75 sections using a a run of cal BP and calibrated c-14 dates.

#### 4.4.2.2 X-ray Fluorescence (XRF) scanning, optical and X-radiographic images and magnetic susceptibility

The core faces were cleaned and scanned with a Cox Analytical Systems ITRAX$^{TM}$ core scanner at the British Ocean Sediment Core Facility (BOSCORF) in Southampton, UK. Optical imagery of the core face was collected with a RGB camera. Magnetic susceptibility was measured with a Bartington MS2E sensor at 1 centimetres intervals and air corrected between measurements. 36 elements were examined at 0.05 centimetres intervals through x-ray fluorescence (XRF) with a 3kW water-cooled Mo anode x-ray tube operating at 45 kV, 35 mA, 200 ms exposure, 25s dwell time every 500 microns giving readings in thousands of counts per second (kcps).

Elements measured are predetermined by the tube in the XRF scanner, the molybdenum (Mo) tube provides a scan of elements with high atomic numbers while the Chromium (Cr) tube provides a scan of elements with a low atomic number (Arnaud et al. 2014). The geochemical data represent a semi-quantitative measurement of elemental composition of the sediment and is influenced by potential x-ray absorption and/or scattering across the core due to variability in water content, particle size distributions, mineralogy and surface roughness of cleaned core face (British Ocean Sediment Core Research Facility 2012, Croudace et al. 2006, Croudace & Rothwell 2010). The results are normalized against aluminium (Al) which is the most abundant metal on earth, least reactive element and routinely used in normalising XRF data (Arnaud et al. 2014, Löwemark et al. 2011, Thomson et al. 2006). The magnetic susceptibility profile was constructed using a MS2 meter with MS2C sensor every 0.5 centimetres. Readings expressed in '$\kappa$ or volume susceptibility, $\kappa$ representing the ratio of magnetization in samples (per unit volume) to the magnetic field created by the sensor, '$\kappa$' is dimensionless with a scale of $10^{-5}$in SI units (Burrows et al. 2016). The values are useful as a concentration proxy for the input of the denser terrigenous minerals derived from the catchment providing a direct measure of changes in the wetland hydrology.

#### 4.4.2.3 Macroscopic charcoal analysis

Subsamples of 1 $cm^3$ of sediment were extracted at 1 centimetre intervals from the Kimana and Ormakau cores, every 0.5 centimetres from the Enkongu Narok core, and between 1 and 5 cm from the Esambu core. The samples were soaked in sodium hexametaphosphate solution and a drop of hydrogen peroxide to disaggregate the samples and aid in the separation of the organic material and the clay particles (Bamber 1982, Schlachter & Horn 2010, Whitlock et al. 2010). Samples were wet sieved through a 125 $\mu$m mesh and the retained charcoal were identified by visual inspection and probed with a metal needle and pieces were tallied under a Zeiss Axio Zoom V16 microscope at 10 to 40X magnifications. Counts were converted to charcoal concentration values i.e. number of particles per unit of volume (pieces/$cm^{-3}$) and charcoal concentration rates (number/$cm^2$/$yr^{-1}$).

#### 4.4.2.4 Pollen and NPP analysis

A 1 $cm^3$ sub-sample was obtained every twenty centimetres for pollen and spore analysis from Kimana and every 2.5 cm from Esambu following the standard protocol (Faegri & Iversen 1950, Moore et al. 1991). An exotic marker (*Lycopodium* spores) was added prior to pollen analysis to aid in calculation of absolute concentrations (Bonny 1972, Stockmarr 1971). The sediment cores were sub-sampled every ten centimetres and 1 $cm^3$ was sampled into 100 ml beakers together with the one *Lycopodium* tablet with a known number of spores (9666 spores) added to each sample. 10 ml of HCl (to remove the calcium carbonate in the samples) was added to the samples, vortexed to encourage mixing of the sample with the acid and placed on a hot water bath for 2 minutes. Samples were then centrifuged at 3000 revolutions per minute (rpm) for 3 minutes and the liquid decanted. 7 ml of distilled water was added to each sample, vortexed, centrifuged and the liquid decanted. This was done twice to remove all the HCl.

20 ml of 10% KOH (to digest organic matter) was added to the samples which were then gently boiled while mixing on a hot plate to facilitate peptisation for five minutes and then left to cool. The samples were sieved (to remove large unwanted particles) through 250$\mu$m mesh screens into 15 ml polypropylene centrifuge tubes and centrifuged for 1 minute at 2000rpm and the liquid decanted. Samples were then washed twice with deionized water by adding 2 to 3 ml deionized water, thoroughly shaking the samples, topping up with water and centrifuging at 2000 rpm for 1 minute. If carbonates were still present 1 ml ethanol and 1 ml $H_2O$ were added and the sample vortexed. Seven ml of 96% Glacial acetic acid was added and carefully mixed into the mixture and left for 8 to 12 hours. Samples were then washed with deionised water to prepare for acetolysis. During acetolysis samples were washed twice with acetic acid i.e. 4 ml Acetic acid was added to the samples and centrifuged at 2000 rpm for one minute and the clear liquid decanted. Acetolysis mixture 96% $H_2SO_4$: Acetic anhydride at a ratio of 1:9 was added to the samples which were heated in aluminium block heater to 100°C. The acetolysis mixture digests the cellulose covering the pollen making the exine features distinct. The tubes were vortexed and heated further for 10

minutes. Samples were centrifuged and decanted to remove acetolysis mixture and washed twice in deionised water.

Heavy liquid separation using Sodium polytungstate ($3NaWO_4.9WO_3.H_2O$ with d=2) was carried out to separate the pollen from the remaining organic material. Three ml of the heavy liquid was added to samples and vortexed, water was carefully added using a glass rod to prevent mixture of the two liquids, the mixture was centrifuged at 3000 rpm for 1 minute and an organic suspension layer appeared at the boundary between the heavy liquid and water before the organic suspension was then carefully transferred to another test tube and samples were washed twice in deionised water. Prepared samples were transferred to residue tubes using 96% alcohol, centrifuged and decanted. Glycerine (same volume as residue) was then added to the sample and the tubes left to evaporate in stove at 60°C. Pollen samples were mounted onto the pollen slides where the identification and enumeration of the pollen, spores and micro charcoal was carried out at a magnification of 400 to 1000 and from each slide a minimum of 300 pollen grains, excluding Poaceae and Cyperaceae, using a Leica DM4000B. The pollen grains were identified using images and descriptions from the African Pollen Database and published atlases (Hamilton 1976, Hamilton & Street-Perrott 1980).

#### 4.4.2.5 Sediment properties

1 $cm^3$ of wet sample taken every 5 centimetres along the Kimana sediment core and every 0.5 centimetre along the Enkongu Narok core, weighed and recorded as wet weight (WW). The samples were dried for 14 to 24 hours at 105°C. Samples were removed from oven and added to desiccator to cool to room temperature for 25 to 30 minutes then reweighed for dry weight and recorded as dry weight after 105°C (DW105). After weighing the samples were put in the muffle furnace at 550°C for 4 hours. When cooled enough, the crucibles were removed using tongs and put in the desiccator to reach room temperature (45 minutes). Samples were reweighed for ignition weight (DW550) and recorded. After weighing (DW550) the sample was put back into the muffle furnace at 950°C for 2 hours and timing only started once temperature was reached. Once the furnace cooled significantly, samples were removed using

tongs and put in the desiccator. Once cooled to room temperature (at least 45 minutes) samples were weighed to obtain carbonate (inorganic carbon) weight DW950.

1. The organic content was calculated as LOI 550 (organic content) = [(DW105-CW)-(DW550-CW)/ (DW105-CW)]*100.

2. The carbonate content was calculated as LOI 950 = (((DW550-CW)-(DW950-CW))/(DW105-CW))*100.

#### 4.4.2.6 Particle Size Analysis

Particle size was determined using the Malvern Laser granulometer, before each run, cleaning and calibration was carried out to ensure integrity of the results. Samples were taken at every 5 centimetres from the Kimana sediment core and 10 ml of 30% $H_20_2$ was added to 10 grams of the dry sample into a 500 ml beaker in a fume cupboard and left for 1 hour. After an hour, another 10 ml of 30% $H_2O_2$ was added to the sample in a fume cupboard and warmed for 2 hours at 50°C on a hot plate whilst maintaining approximate volume by adding deionised water. After 2 hours it was briefly brought to boil then left to cool before starting the measurements. If the soil sample contained <3.5% organic matter, the hydrogen peroxide treatment was skipped and the runs started right after cleaning and calibration. At a pump speed of 1500rpm, 1 to 2 grams of sample was added until obscuration was 4%.

The granulometer took three measurements and calculated an average result (Malvern Instruments Ltd 2007). A total of three measurements were made for each sample and the average used as the final value. The distribution abundances within the 0.2 to 2000 $\mu$m grain size range were calculated by the Malvern software. All sediments were classified to three classes: clay (particles with a diameter < 2 mm), silt (2 to 60 mm) and sand (20 to 200 mm).

### 4.4.3 Statistical analysis

Statistical methods are similar to ones used in chapter 3.

### 4.4.4 Illustrative Landscape Model

The results from the pollen, spore and charcoal analysis, combined with current observations of the wetland landscape, were used to infer a system-wide visual representation of the wetland and surrounding ecosystem components. CONISS zones from the pollen spectra were used to create the time steps but dates have been rounded for graphical representation. Both planar schematic and cross-sectional views were created to represent the changes in the system over the late Holocene. The varying relative importance of the three major controls on fire regimes at centennial to millennial temporal scales and at local to landscape scale using a fire triangle diagram (Whitlock et al. 2010). Broad plant functional types were used to group the swamp core taxa into aquatic taxa, emergent taxa and riparian components. Although not an absolute scale, the relative sizes of the areas have been inferred from the plant functional type abundance sums and observations of the modern wetland. The human developments illustrated apply to contemporary and recent history, while the summary of the continuum of livelihoods is derived from archaeological work from Amboseli (Foley 1981, Lane 2013)

## 4.5 Results

### 4.5.1 General Lithology

Three lithological units are recognised from the 384 cm Kimana sediment core (Figure 4.4), a dark organic rich sediment layer from the top to 103 cm with dark brown laminations at 29 cm to 30 cm and 52 cm to 53 cm from103 cm, this abruptly changes to a dark brown mud (mainly silt and clay) up to 380 cm with dark grey laminations at 240 cm to 243 cm, 260 cm to 261 cm and two sand layers at 348 cm to 351 cm and 377 cm, the bottom 4 centimetres changes into dark grey with light coloured sand laminations.

There are two lithological units recognised from the 304 cm Ormakau sediment core (Figure 4.4), a dark organic rich layer from the top to 130 cm. Through out the top 130 cm there is organic matter such as rootlets. The top 1 metre of the Ormakau record is highly mixed as evidenced by the presence of fresh organic matter around

90 cm. The dark organic layer grades into a dark brown mud to 210 cm and there is a sand layer around 150 cm. Between 200 cm and 250 cm, it changes back into dark organic sediment with a clay layer and finally back into a dark brown silty clay sediment. Between 120 cm and 170cm, a hard sandy layer is encountered, between 200 cm and 250 cm a two hard clay layers are encountered.

The 195 cm long Enkongu Narok sediment core (Figure 4.4) is all peat with differences occurring in the colour and rate of decomposition. The minor colour and decomposition changes within the Enkongu sediment can be described as: brown, decomposed fibrous root peat from the top to 20cm, it gets darker and the fibrous roots are less decomposed until 37 cm. Between 37 cm and 78 cm, the fibrous material is more decomposed and the peat a darker brown, 79cm to 90 cm the the peat is dark grey, which browns out again between 91 cm and 100 cm. 101 cm to 131 cm the peat is dark grey, 131 cm to 134 cm it is brown, 134 cm to 136 cm it turns grey again. From 136 cm to 161 cm the peat is dark grey which lightens to a light grey from 161 to the bottom of the core. Between 171 cm and 195 cm (bottom of the core) the light grey peat displays lighter laminations as well as small crumbly concretions of clay.

The 247.5 cm Esambu sediment core is characterised by two main lithological units, dark brown silty peat from the top to 210 cm that grades into grey-brown clay. From the top to 92cm, it is composed of decomposed dark brown peat, between 92 cm and 100 cm the dark brown peat contains fibrous roots. Between 100 cm and 183cm, the dark brown peat has dark laminations as well as concretions. 183 cm to 186 cm was mainly a large *Acacia* sp. fragment. From 186 cm, the dark brown peat surrounding the *Acacia* sp. fragment starts grading into grey clay. From 195 cm to the bottom of the core ~248 cm, the sediment can be described as grey silty clay.

### 4.5.2 Radiocarbon Dating

The radiocarbon dates (Table 4.4) provide the basis of the BACON age depth models (Blaauw & Christen 2013); the seven dates from Kimana when modelled give the basal date of Kimana core to be ~1200 cal yr BP. The model recognises the age from 310 cm as an outlier as it is younger than the depths from the two levels above it. At Ormakau, the six dates are modelled to give a basal date of ∼2700 cal yr BP with one date not modelled (118±22 from 150 cm) which was also recognised as an outlier due to being older than the two dates above it (Figure 4.5).

The Enkongu core has a basal date of ∼2000 cal yr BP modelled from the six radiocarbon dates (Figure 4.5, Table 4.4). There is an age reversal where the sample from 100 centimetres was older (10721±47) than the samples from two lower depths, as a result three dates (50 centimetres: 2449±35, 100 centimetres: 10721±47 and 128 centimetres: 7616±33) were noted as outliers and not modelled (Figure 4.5). The age reversal could be due to a potential erosional or depositional hiatus, sediment reworking or root growth bringing younger material to lower sediment. Radiocarbon dates are presented as cal yr BP with the modern reference date taken as 1950 AD.

Table 4.4: Kimana, Ormakau, Enkongu Narok and Esambu samples for radiocarbon dating showing labID, site, age, error and depth. All samples were bulk sediment except ***Esambu at 180 cm**

| LabID | Site | $^{14}C$ Age | Error | Depth(cm) | cal yr BP |
|---|---|---|---|---|---|
| SUERC-64044 | Kimana | 578 | 37 | 20.5 | 147 |
| SUERC-6404 | Kimana | 526 | 37 | 40.5 | 336 |
| SUERC-57339 | Kimana | 424 | 37 | 50 | 362 |
| UBA-26122 | Kimana | 540 | 24 | 100 | 534 |
| D-AMS 009667 | Kimana | 946 | 25 | 260 | 908 |
| D-AMS 009668 | Kimana | 455 | 24 | 310 | 1085 |
| UBA-26122 | Kimana | 1458 | 31 | 370 | 1316 |
| SUER-57341 | Ormakau | 102 | 3 | 50 | 102 |
| UBA-26120 | Ormakau | 150 | 23 | 100 | 263 |
| D-AMS 009669 | Ormakau | 118 | 22 | 150 | 512 |
| D-AMS 009670 | Ormakau | 1071 | 24 | 200 | 987 |
| D-AMS 009671 | Ormakau | 1825 | 25 | 255 | 1763 |
| UBA-26119 | Ormakau | 2700 | 26 | 304 | 2781 |
| D-AMS 011827 | Enkongu Narok | 161 | 26 | 38 | 168 |
| D-AMS 0011828 | Enkongu Narok | 354 | 24 | 71 | 336 |
| D-AMS 0011829 | Enkongu Narok | 325 | 22 | 104 | 431 |
| D-AMS 0011830 | Enkongu Narok | 457 | 26 | 133 | 520 |
| D-AMS 0011831 | Enkongu Narok | 974 | 27 | 162 | 849 |
| UBA-26125 | Enkongu Narok | 2122 | 29 | 192 | 1949 |
| UBA-27555 | Esambu | 146 | 23 | 40 | 79 |
| UBA-27556 | Esambu | 296 | 24 | 110 | 313 |
| *UBA-276124 | Esambu | 394 | 24 | 180 | 489 |
| D-AMS 009665 | Esambu | 3299 | 29 | 200 | 2774 |
| D-AMS 009666 | Esambu | 2296 | 29 | 218 | 294 |
| UBA-26123 | Esambu | 4001 | 27 | 242 | 4296 |

The Esambu age depth record reveals a hiatus. Linear interpolation of six radiocarbon dates show that the sediment record spans from 4974 cal yr BP to present. A single outlying bulk sediment date was not included in the age-depth model (Table 4.4, Figure 4.4); the outlying date at 200 to 201 cm is an older date that could be the result of older carbon entering the depositional environment as it transitioned from semi-arid to a more hydric and eventually inundated wetland ecosystem. The potential hiatus may have been produced during the transition from dominantly semi-arid conditions to a regime of increased moisture and expansion of the wetland vegetation at the coring location, but does not impact the overall interpretation of the ecosystem transitions.

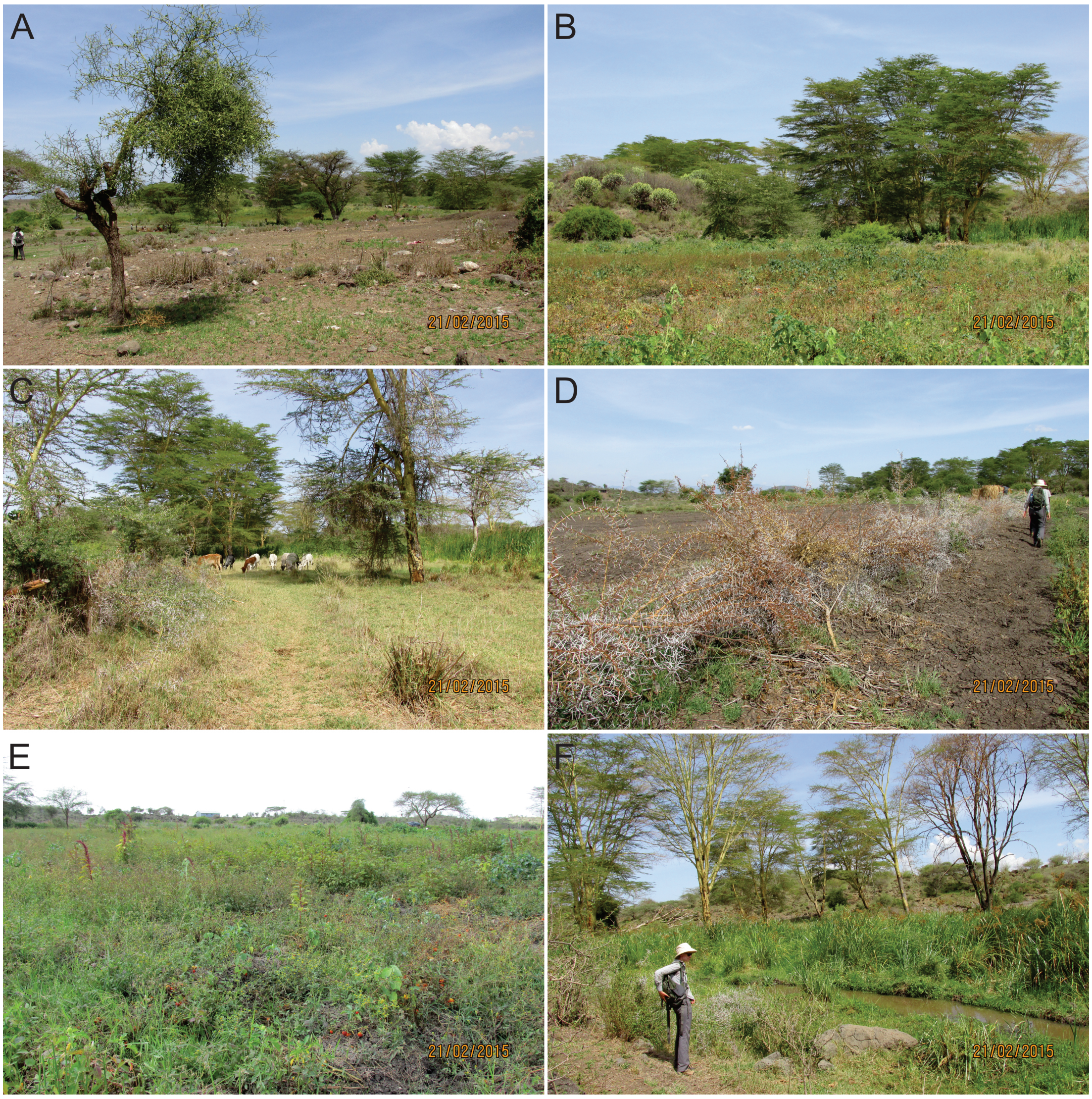



Figure 4.3: Landscape views around Esambu wetland showing the vegetation type and land use. A. Common mix of *Commiphora* and *Acacia* trees, grasses and barren patches surrounding Esambu wetland. B. Typical wooded savannah and a volcanic mound on the left hosting xerophyillic taxa, such as young *Euphorbia*. C. Cattle grazing within the savannah around the wetland. D. An *Acacia* fence used to control grazing access to moister sections of the wetland. E. Small scale tomato farming within ruderal herbaceous, woody and grass taxa surrounding the wetland. F. Emergent aquatic taxa of the hydric sediments within the wetland dominated by Cyperaceae. Photographs: Esther Githumbi.

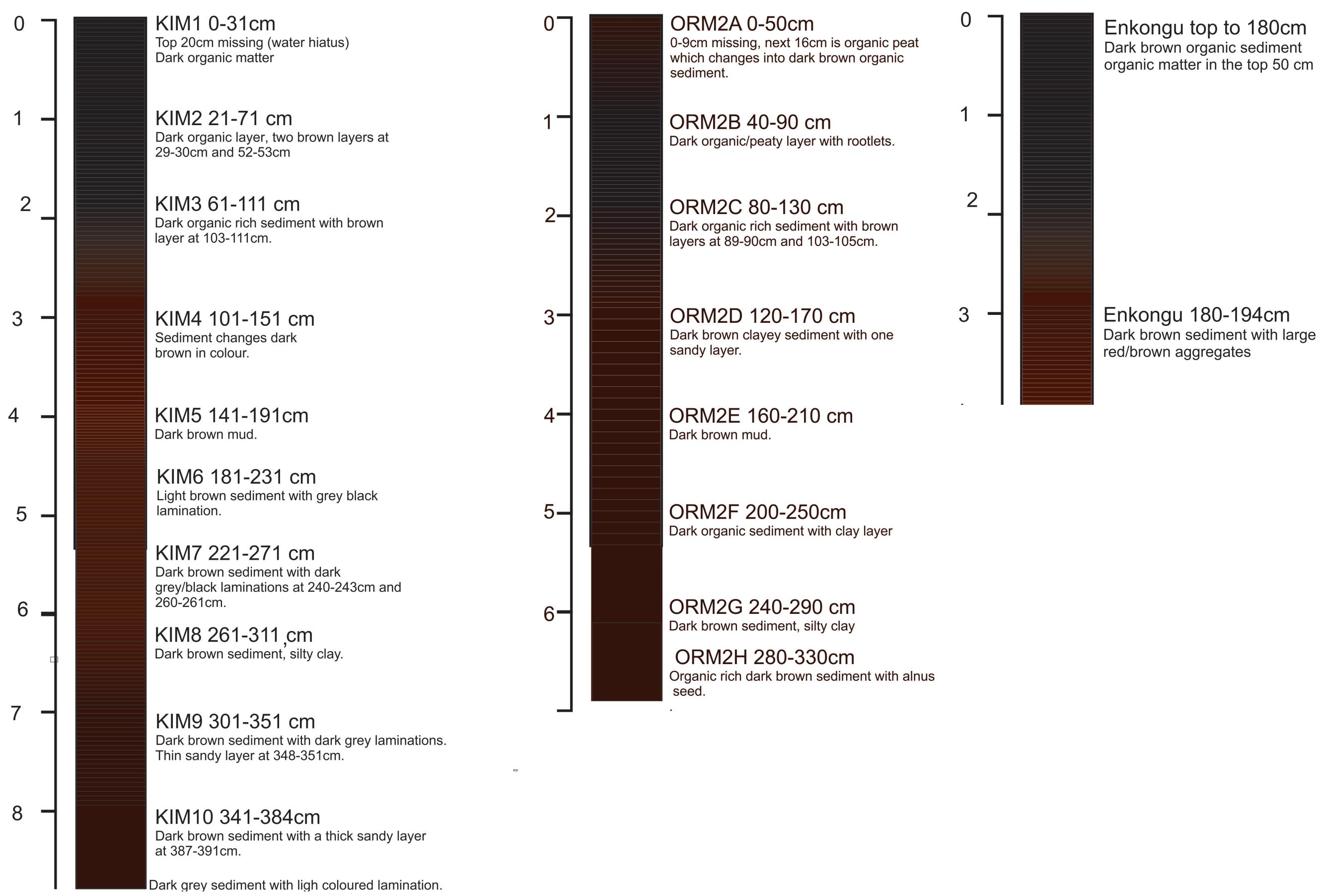



Figure 4.4: Stratigraphic descriptions of the Kimana, Ormakau and Enkongu Narok sediment cores.

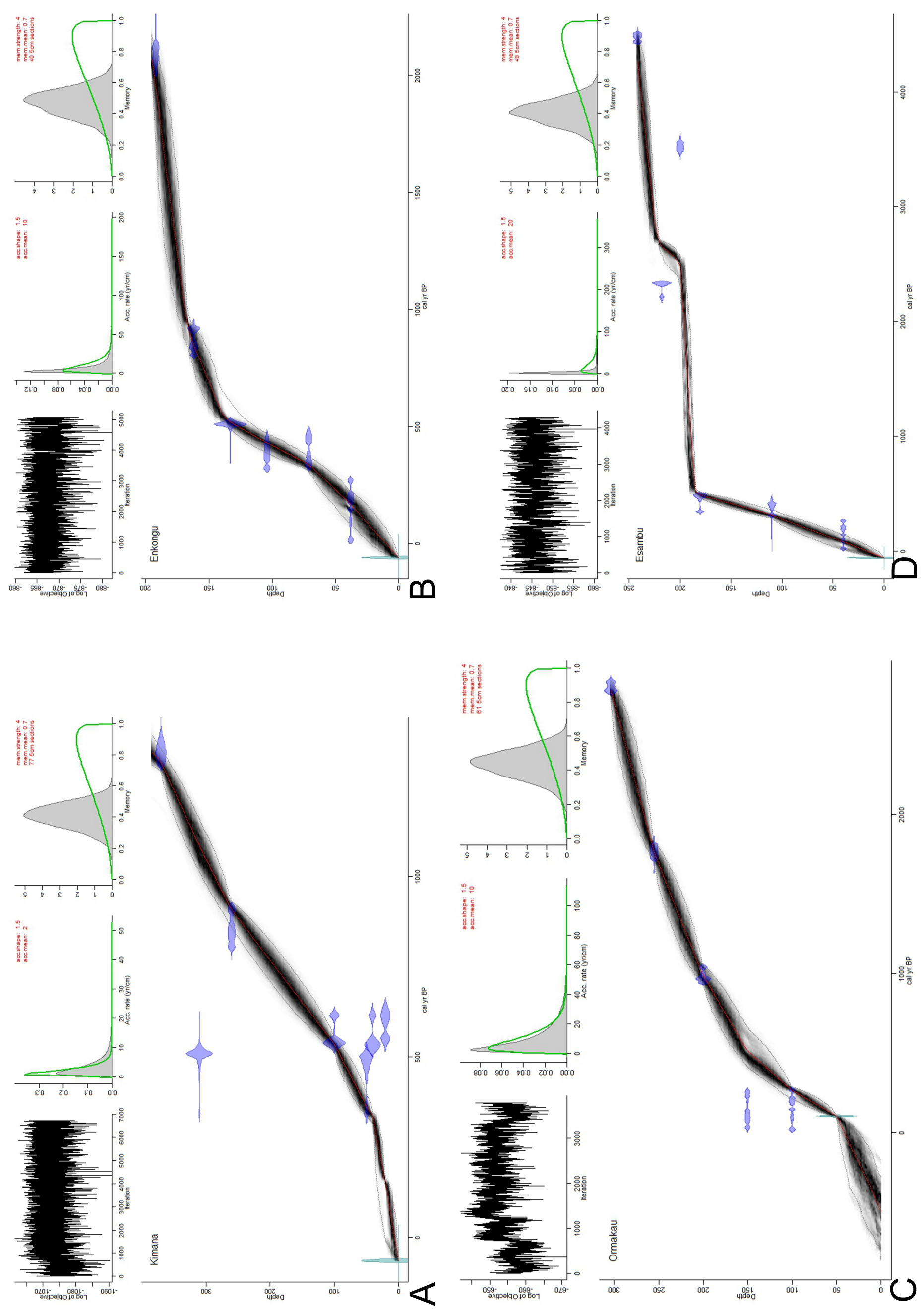



Figure 4.5: Kimana, Ormakau and Enkongu and Esambu BACON age -depth models.

### 4.5.3 X-ray Fluorescence (XRF) of the elemental properties

Only the newly cored sites Kimana and Ormakau wetland were analysed using the ITRAX core scanner. A total of 36 elements were detected using the ITRAX XRF core scanner and the data provided is relative variation of the various elements within the sediment matrix as counts per second. The Ormakau record has several sections lacking the geochemical analysis due to technical difficulties experienced and so only the Kimana record will be presented and discussed.

The Kimana record was dominated by Fe, Ti, Zr, Sr and Ca all with mean counts per second >2000 while Cs, Sc, Ge, Cl and As had mean counts below 50. Al, P, Ta, W, Ni and Br have the strongest positive correlation with Pearson's correlation values above 0.5, Rb, Zn and Tb have the strongest negative correlation with Pearson's correlation values between 0.4 and 0.5 (Figure 4.7 ). A b-stick and CONISS analysis delineated the record into three zones. KimITRAX1 is from ∼2700 cal yr BP to ∼570 cal yr BP while KimITRAX2 if from ∼570 cal yr BP to ∼500 cal yr BP and KimITRAX3 covers the last 500 cal yr BP. A principle component analysis shows that the first three components are sufficient to explain 81% of the variance.

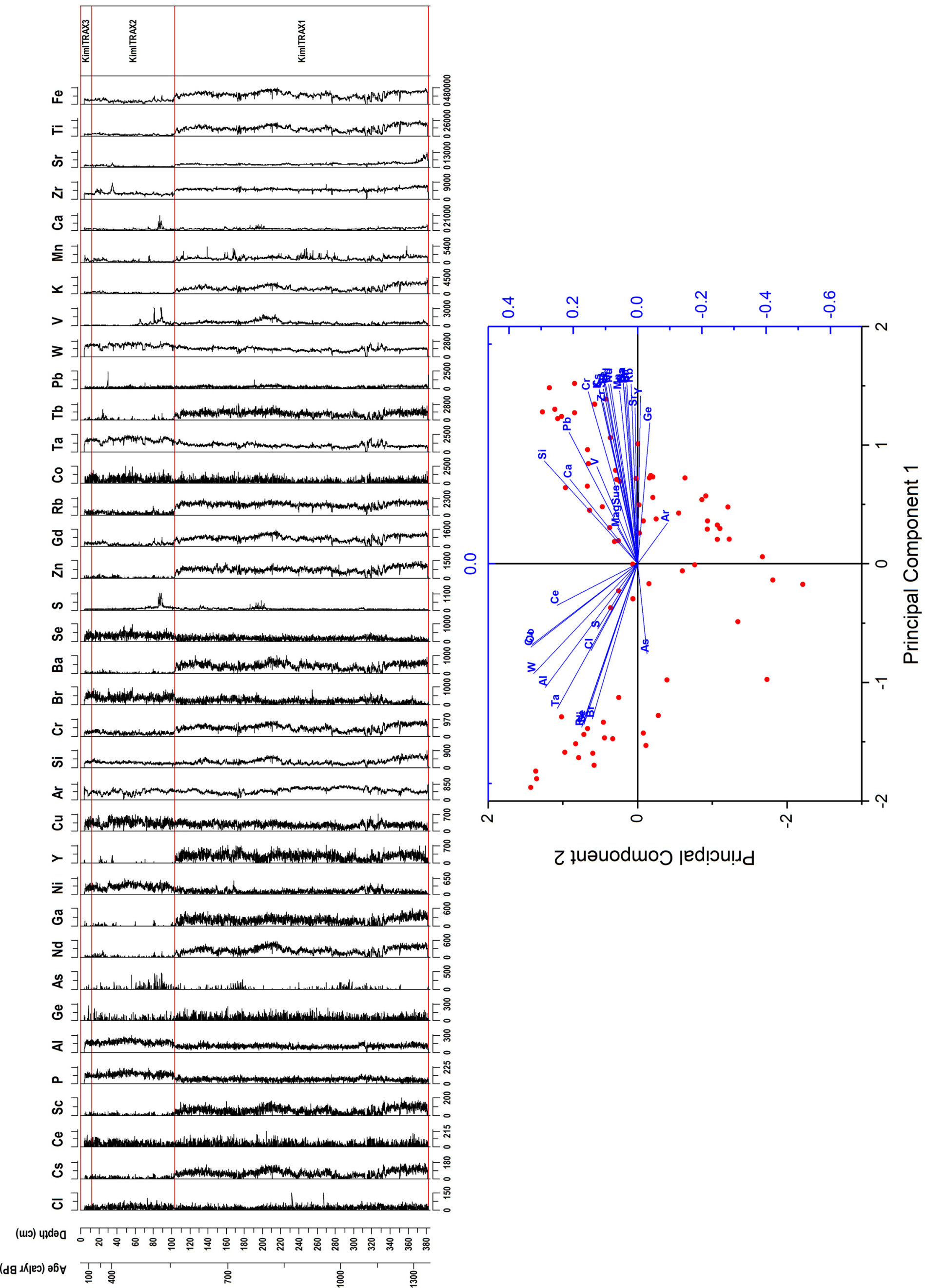



Figure 4.6: Elemental change through time at Kimana as well as a PCA biplot of the spread of the elements along the two main components which explain ~81% of the variance.

### 4.5.4 Magnetic susceptibility

The Kimana magnetic susceptibility is three times a high as the Ormakau susceptibility, however the Ormakau magnetic susceptibility shows higher variation. The Kimana magnetic susceptibility (Figure 4.8) varied from 3.90 to 617.41 and the core had a mean magnetic susceptibility of 193.73±5.543 displaying 23 peaks. Ormakau magnetic susceptibility ranged from zero to 943.37 with a mean magnetic susceptibility of 100.92±6.752, the Ormakau record showed 5 peaks.

### 4.5.5 Macroscopic charcoal

The maximum charcoal concentration at Esambu was 6384 pieces/cm$^3$ at ∼ 1300 cal yr BP and the minimum charcoal concentration was 0 pieces/cm$^3$ at ∼4200 cal yr BP. The mean charcoal concentration for the whole record was 352.87 pieces/cm$^3$ while the mean charcoal accumulation rate was 137.11 pieces/cm$^2$/yr$^{-1}$.

The Kimana record mean macroscopic charcoal concentration and charcoal accumulation rate over the last ∼1300 cal yr BP was 326 pieces/cm$^3$ and 82.16 pieces/cm$^2$/yr$^{-1}$ respectively. The charcoal concentration significantly increased over the last 500 cal yr BP with a mean concentration of 979.31±101.06 pieces/cm$^3$, which was three times the mean macroscopic charcoal concentration of the whole record. The charcoal accumulation rate also increased to 221.89±27.50 pieces/cm$^2$/yr$^{-1}$, twice the macroscopic charcoal accumulation rate of the whole record.

The Ormakau mean charcoal concentration and accumulation rate over the last ∼2700 cal yr BP was 186.33±11.70 pieces/cm$^3$ ranging from zero to 1238 pieces/cm$^3$ and a very low charcoal accumulation rate of 0.82 pieces/cm$^2$/yr$^{-1}$.The last ∼700 cal yr BP record a decrease in the charcoal concentration to 157.78±15.87 pieces/cm$^3$ from the mean macroscopic charcoal concentration while the charcoal accumulation rate remains at a similar level of 0.89±0.08 pieces/cm$^2$/yr$^{-1}$.

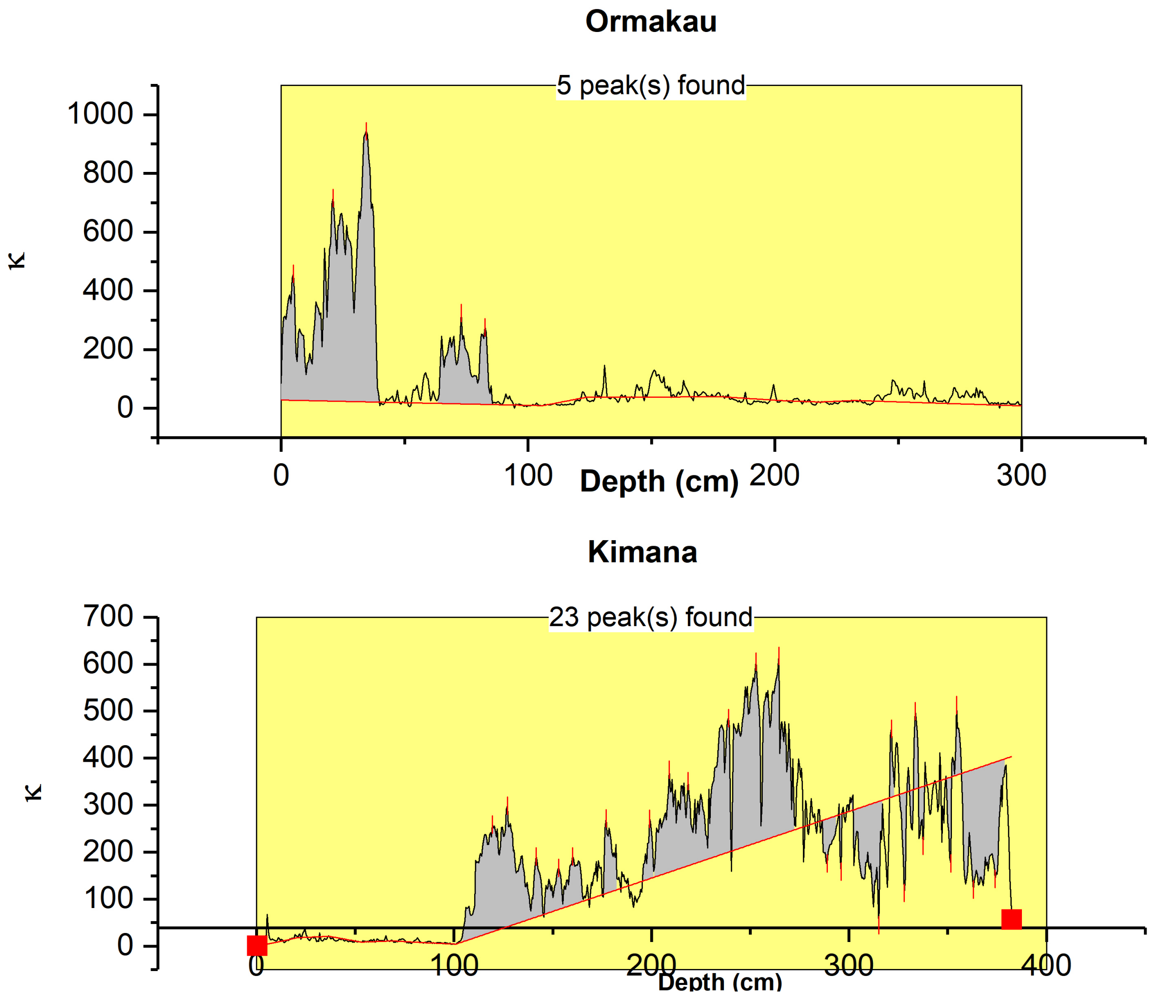



Figure 4.7: Magnetic susceptibility profiles of the Kimana and Ormakau sediment records and a box plot of the mean values in each site.

The Enkongu Narok sediment macro charcoal record covers the last ∼2000 cal yr BP, the mean charcoal concentration over that period has been 170.494±7.176 pieces/cm$^3$ ranging between 12 and 994 pieces/cm$^3$. The mean charcoal concentration reduces again over the last ∼500 cal yr BP to 137.61±5.59 pieces/cm$^3$ from the mean macroscopic charcoal concentration ranging between 12 pieces and 572 pieces/cm$^3$.

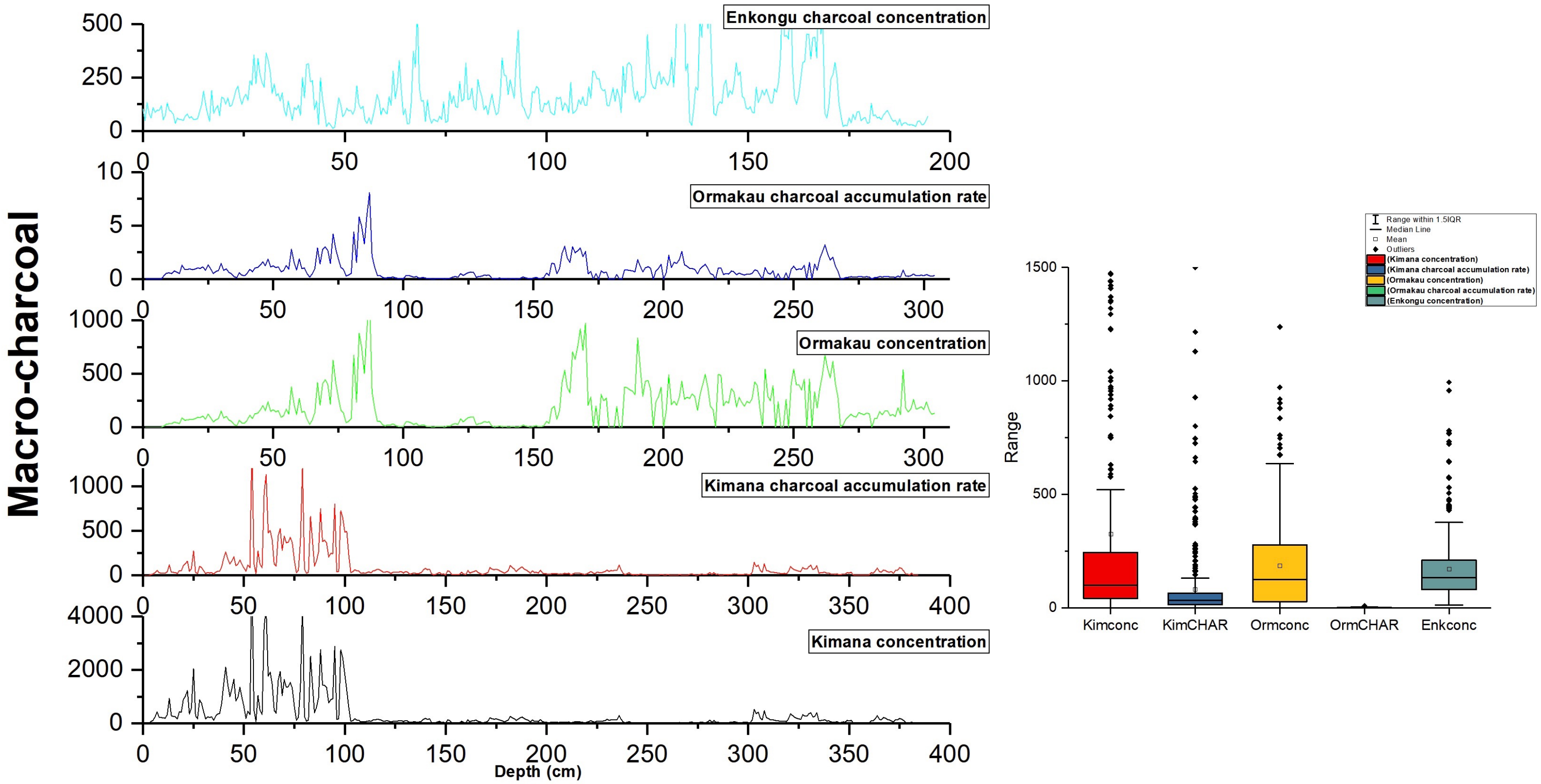



Figure 4.8: Macroscopic charcoal concentration and charcoal accumulation rates from Esambu, Kimana, Ormakau and Enkongu sediment records.

### 4.5.6 Pollen

Over sixty pollen types were identified from the Esambu, Enkongu and Kimana pollen records (Table 4.5).

#### 4.5.6.1 Esambu pollen and NPP

A total of 93 pollen taxa and 74 NPP taxa were identified throughout the core. Pollen assemblage zones were delineated by CONISS to interpret the pollen and NPP result within a land cover change context. Only two pollen zones were statistically significant, separating the deeper sediments of ESAM1 representing the semi-arid woodland taxa and the smaller Cyperaceae-dominated wetland from ESAM2 showing an expanded wetland and increased sedimentation rates (Figure 4.9). Pollen zone ESAM2 was further divided by visual differences observable in the pollen spectra (ESAM2a-c). CONISS zones in the NPP assemblages did not show significant transitions concomitant with pollen (Figures 4.10 and 4.11). The pollen zones are described further below through each zone stratigraphically and the zones are visually represented in Figure 4.10 to summarise.

Table 4.5: Pollen identified from Esambu, Enkongu and Kimana sediment cores in more than one sample (X-Present, x-Absent)

| Family | Species | Grouping | Esambu | Enkongu | Kimana |
|---|---|---|---|---|---|
| Cyperaceae | *Cyperus* spp. | Sedge | X | X | X |
| Typhaceae | *Typha latifolia* | Herb | X | X | X |
| Cupressaceae | *Juniperus* sp. | Tree | X | X | x |
| Oleaceae | *Olea* sp. | Tree | X | X | x |
| Podocarpaceae | *Podocarpus* spp. | Tree | X | X | x |
| Primulaceae | *Maesa* sp. | Tree | X | X | x |
| Fabaceae | *Acacia* spp. | Tree | X | X | X |
| Burseraceae | *Commiphora* sp. | Tree | X | X | X |
| Sapotaceae | *Pouteria* sp. | Tree | x | X | x |
| Proteaceae | *Faurea* sp. | Tree | x | x | X |
| | *Neoboutania macrocalyx* | Tree | x | x | X |
| Myrtaceae | *Syzygium* sp. | Tree | X | x | x |
| Rosaceae | Rosaceae | Tree | x | X | x |
| Celastraceae | *Maytenus* sp. | Tree | x | x | X |
| Zygophllaceaea | *Balnites* sp. | Tree | X | X | X |
| Capparaceae | *Capparis* sp. | Tree | x | x | X |
| Celastraceae | Cassine sp. | Tree | x | x | X |
| Malvaceae | Abutilon mauritanium | Tree | x | x | X |
| Amaranthaceae | *Achyranthes aspera* | Shrub | X | X | X |
| Alangiaceae | *Alangium* sp. | Shrub | x | x | X |
| Asphodelaceae | *Aloe* spp. | Shrub | x | x | X |
| Asparagaceae | *Asparagus* spp. | Shrub | x | x | X |
| Asparagaceae | *Sansevieria* sp. | Shrub | x | x | X |
| Ericaeceae | *Blaeria* sp. | Shrub | X | X | X |
| Euphorbiaceae | *Euphorbia* sp. | Shrub | X | X | X |
| Asteraceae | *Helichrysum* sp. | Shrub | X | X | X |
| Asteraceae | *Artemisia* sp. | Shrub | X | x | x |
| Asteraceae | *Stoebe* sp. | Shrub | X | X | x |
| Acanthaceae | *Justicia* sp. | Shrub | X | X | X |
| Verbenaceae | *Lantana* sp. | Shrub | x | x | X |
| Araliaceae | *Polyscias* sp. | Shrub | x | x | X |
| Capparaceae | *Cadaba* sp. | Shrub | x | X | x |
| Capparidaceae | Capparidaceae | Shrub | X | X | x |
| Fabaceae | *Indigofera* sp. | Shrub | x | X | x |
| Zygophyllaceae | *Tribulus* sp. | Shrub | x | X | X |
| Rubiaceae | Rubiaceae | Shrub | x | X | x |
| Aizoacaeae | *Trianthema* sp. | Herb | x | x | X |
| Apocynaceae | *Tabernaemontana elegans* | Shrub | X | X | X |
| Solanaceae | *Solanum incanum* | Shrub | x | X | X |
| Asteraceae | *Aspilia pluriesta* | Herb | x | X | x |
| Asteraceae | *Guizotia* sp. | Herb | x | x | X |
| Asteraceae | *Vernonia* sp. | Herb | x | x | X |
| Amaranthaceae | *Gomphrena* sp. | Herb | x | x | X |
| Amaranthaceae | *Cyathula* sp. | Shrub | X | X | X |
| Balsaminaceae | *Impatiens* sp. | Herb | x | x | X |
| Boraginaceae | Boraginaceaea | Herb | x | x | X |
| Brassicaceae | *Farsetia* sp. | Herb | x | x | X |
| Convulvulaceae | *Ipomoea* spp. | Herb | X | X | x |
| Cyathecaceae | *Cyathea* sp. | Herb | x | x | X |
| Onagraceae | *Ludwigia* spp. | Herb | X | X | x |
| Araliaceae | *Hydrocotyle* sp. | Shrub | x | x | X |
| Malvaceae | *Pavonia patens* | Herb | X | X | X |
| Salvadoraceae | *Azima tetracantha* | Herb | x | X | x |
| Polygonaceae | *Polygonum* spp. | Herb | X | X | X |
| Polygonaceae | *Rumex usambarensis* | Herb | x | x | X |
| Phyllanthaceae | *Phyllanthus* sp. | Herb | X | X | X |
| Umbelliferae | Umbelliferae | Herb | x | X | X |
| Urticaeae | *Laportea* sp. | Herb | x | x | X |
| Poaceae | Poaceae | Grass | X | X | X |

Pollen zone ESAM 1: 247.5-210.5 cm (c. 5000 to 2000 cal yr BP). Amaranthaceae-Chenopodiaceae (20-84%) and Cyperaceae (56-100%) dominated this zone, *Typha* 56-100% and *Tapura* ranged between 0 -72% peaking at 72% at 3887 cal yr BP. *Acacia, Aloe*, Asteraceae, *Capparis* and Poaceae were consistently present. *Acalypha* and *Juniperus* appeared towards the top of the zone (around 220 cm; Figure 4.10). There were 18 spore types identified in this zone, *Chaetomium* (0-30%), *Coniochaeta cf. Ligniaria* (0-64%), Sporormiella (0-20%), T. HdV-1020 (0-17%) and *cf. Valsaria sp.* (0-13%) dominated (Figure 4.11). Pteridophytes appeared at 4800-3000 cal yr BP. The charcoal concentration and accumulation rate was quite low in this zone with a maximum of 1.5 pieces $cm^{-2}$ $yr^{-1}$ around 2300 cal yr BP.

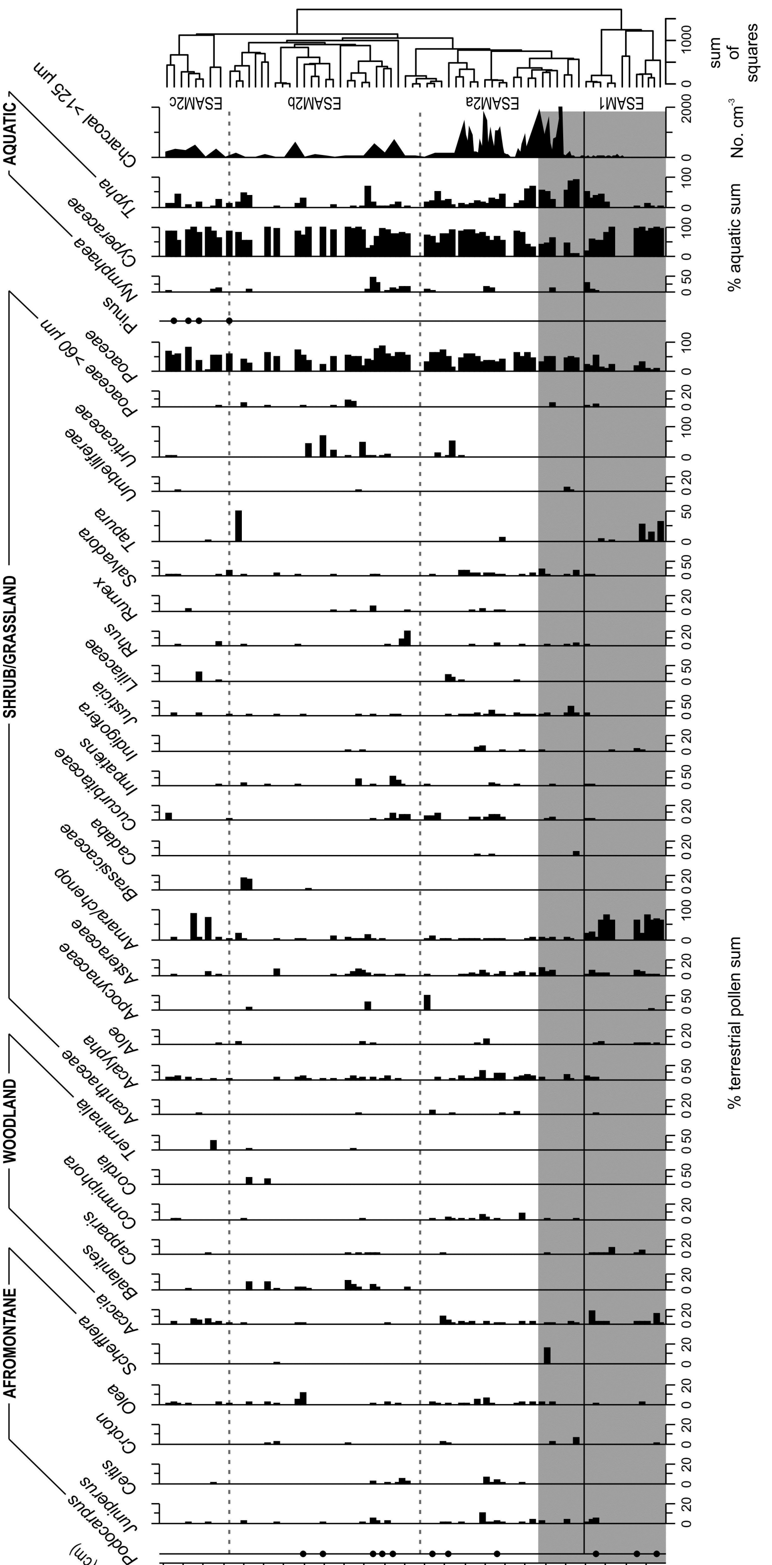



Figure 4.9: Esambu relative abundance of selected pollen taxa divided into two significant zones and a grey shaded area highlighting the radiocarbon period of uncertainty

Pollen zone ESAM 2a: 210.5-131.5 cm (2000-400 cal yr BP). Cyperaceae (10-92%), Poaceae (17-64%), *Salvadora* (0-30%) and *Typha* (8-90%) dominated this zone; Acalypha (3-34%), Asteraceae (0-10%) and Justicia (0-30%) appeared consistently. *Nymphaea* abruptly appeared at 1700 cal yr BP and fluctuated between 0.5-32%. There were more arboreal taxa present in this zone: *Celtis, Commiphora, Croton, Juniperus, Olea, Podocarpus* and *Schefflera* were commonly present. *Commiphora* and *Podocarpus* were present throughout the zone between 3-10%. A total of 32 spore types were identified in this zone, *Coniochaeta cf. Ligniaria* drastically reduced to 0-27% in this zone. *Cf. Sordaria* (0-68%), *Cercophora* type (0-54%) and T.HdV-1010 (0-554%) appeared consistently and dominated this zone. *Chaetomium* occurred at moderately high levels 0-78% in most levels 17% while *cf. Valsaria* sp. (12% at one level only) and *Sporormiella* reduced to very low numbers between 1-10%. The charcoal concentration and accumulation rate rapidly increased at this level having several charcoal peaks at 130 pieces $cm^{-2}$ $yr^{-1}$ around 1300 cal yr BP, 1074 pieces $cm^{-2}$ $yr^{-1}$ around 444 cal yr BP, 1730 pieces $cm^{-2}$ $yr^{-1}$ around 433 cal yr BP and 1386 pieces $cm^{-2}$ $yr^{-1}$ around 423 cal yr BP (Figures 4.10 and 4.11).

Pollen zone ESAM 2b: 131.5-27.5 cm (400 to 100 cal yr BP). Cyperaceae, Poaceae and *Typha* dominated this zone. Liliaceae increased drastically from 2% in the previous zone and fluctuated up to 30% before disappearing at 270 cal yr BP. *Acalypha*, Asteraceae, *Impatiens* and *Rhus* ranged between 0-20% and spiked at 330 cal yr BP. *Balanites* and *Celtis* appeared at 330 cal yr BP before completely disappearing at 203 cal yr BP. *Juniperus, Olea, Podocarpus* all appeared at levels below 5% in this zone. *Celtis* and *Podocarpus* disappeared at 200 cal yr BP and 240 cal yr BP and respectively. A total of 27 spore types were identified in this zone; *Cercophora* type (8-77%), *Clastesporium* sp. (0-72%) and T.HdV-112 (0-45%) dominated, whereas *Usculina deusta* and *Sporormiella* occurred at low levels (>10%). The charcoal concentration and accumulation rate reduced again with only two peaks in this zone of 666 pieces $cm^{-2}$ $yr^{-1}$ around 370 cal yr BP and 185 pieces $cm^{-2}$ $yr^{-1}$ around 230 cal yr BP.

Pollen zone ESAM 2c: 100 to -46 cal yr BP (27.5-2.5 cm). Cyperaceae (53-98%), Poaceae (21-76%) and *Typha* (1-39%) dominated this zone. *Nymphaea* reappeared at 12.5 cm and fluctuated up to 6%. Amaranthaceae/Chenopodiaceae (24%), *Salvadora*

(1-3%) and Poaceae grains >60 $\mu$m (0.5-2.5%) appeared consistently throughout this zone at low levels and Asteraceae appeared at the top (1-5%). *Terminalia* appeared once throughout the core in the recent past at 5 cm at the same time as *Acacia* (0.5-1.5%) and *Podocarpus*. *Pinus* appeared at four levels but at very low counts. *Acalypha, Balanites, Commiphora* and *Cordia* occasionally appeared at the top while *Celtis* (2%), Juniperus (1-2.5%) and Olea (1-12%) were observed once or twice. Of the 17 spore types identified in this zone, *Chaetomium* (0-59%), Usculina deusta (0-10%), T.HdV-112 (0-58%) and T.HdV-121 (0-44%) dominated. *Sporormiella* (0-11%) occurred consistently in this zone (Figure 4.11). This zone had the lowest charcoal concentration levels with the last peak of 90 pieces $cm^{-2}$ $yr^{-1}$ centred around 20 cal yr BP. The macroscopic charcoal occurred at very low levels throughout the core except at a distinct high concentration zone between 148.5-198 cm where the average number of charcoal pieces per $cm^{-3}$ was 800 with one level (197.5 cm) having 6384 pieces. The macroscopic charcoal peaks corresponded to increases in *Typha* peaks.

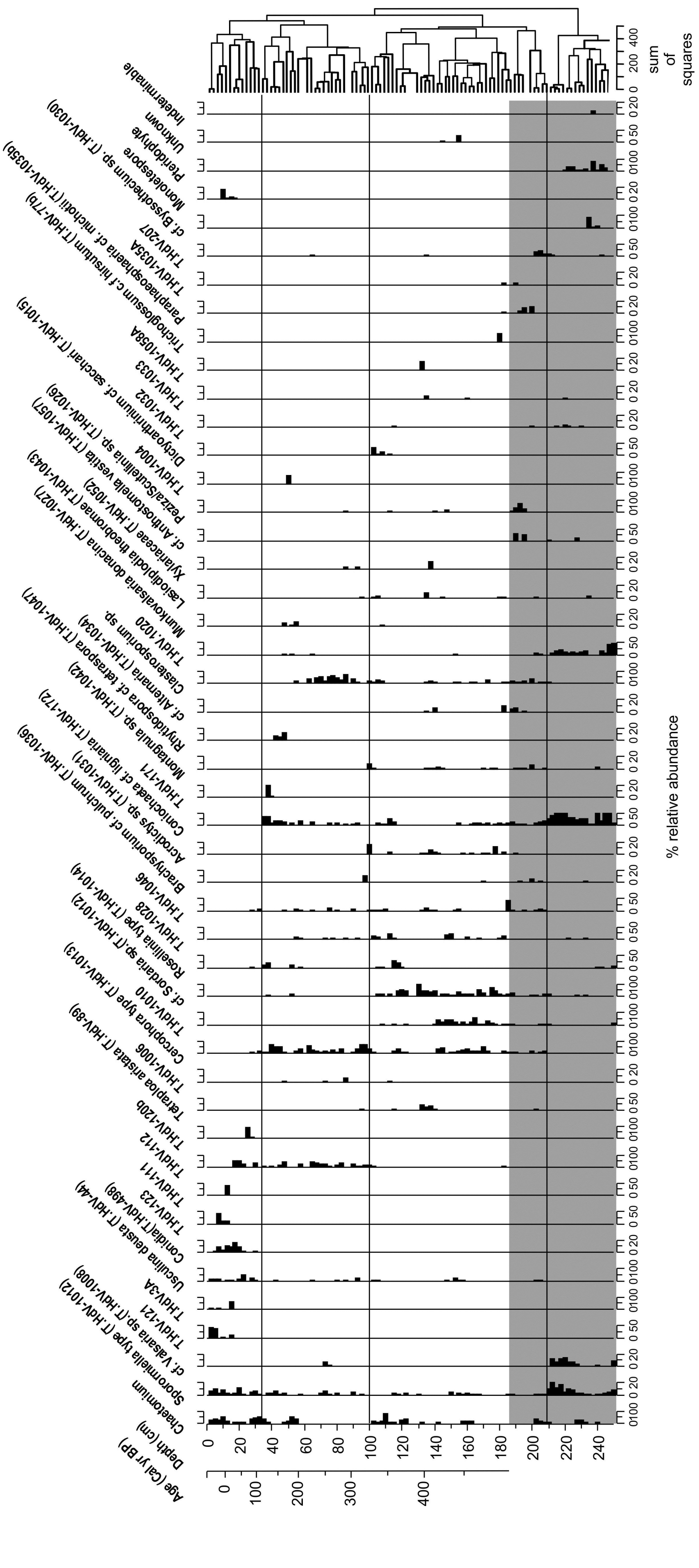



Figure 4.10: Relative abundance of spore types from the Esambu core with the four zones.

#### 4.5.6.2 Kimana Pollen and NPP

Thirty seven pollen types were identified from the Kimana sediment core, the pollen diagram was divided into three zones (Figure 4.12). The lower zone KimPol1 covered the period from $\sim$1200 cal yr BP to $\sim$900 cal yr BP, the second zone (KimPol2) covered from $\sim$900 cal yr BP to $\sim$500 cal yr BP and the third zone (KimPol3) covered the last $\sim$500 cal yr BP. The individual pollen taxa occurred at very low percentages and have been grouped for ease of interpretation into local trees, shrubs, herbs, non-local trees, Poaceae and aquatics.

The sediment core was dominated by herbaceous taxa with a mean percentage of 48.11$\pm$0.98%, this is followed by the shrub taxa at 42.96$\pm$0.97% and local trees at 8.92$\pm$0.39. Poaceae had a mean of 35.33$\pm$1.3% and aquatics were 64.67$\pm$1.3%. The Poaceae and aquatics are not part of the total pollen sum and were represented as grasses: aquatics. The mean percentage was 35.33$\pm$1.3%:64.67$\pm$1.3%(1.90$\pm$0.1) and it increased through time in the sediment record. The non-local trees were Afromontane and montane taxa which could either be from Mt. Kilimanjaro or the Chyulu hills and their mean accounted for 19.21$\pm$0.633%. The NPP count ranged between 40 and 50 within the core with a mean of 52%.

KimPol1covers $\sim$1200 cal yr BP to $\sim$900 cal yr BP, the composition did not vary much from the mean composition with the herbaceous: shrub: local trees taxa comprising 48.20$\pm$1.82%, 42.89$\pm$1.73% and 8.91$\pm$0.81% respectively. The non-local tree taxa were at 16.87$\pm$1.47%, while the grasses to aquatics were 36.92$\pm$3.41% to 63.08$\pm$3.41% (1.81$\pm$0.27). This remains the same in KimPol2 ($\sim$900 cal yr BP to $\sim$600 cal yr BP) and the only change noted was the slight increase in non-local tree taxa from $\sim$16% to 20.14$\pm$0.74% as well as the aquatics to grasses ration slightly increased. Over the last $\sim$600 cal yr BP, the vegetation taxa composition remained similar to the previous zone except a slight decrease in grasses and increase in aquatics at 33.09$\pm$1.39% and 66.91$\pm$1.39% (2.06$\pm$0.36).

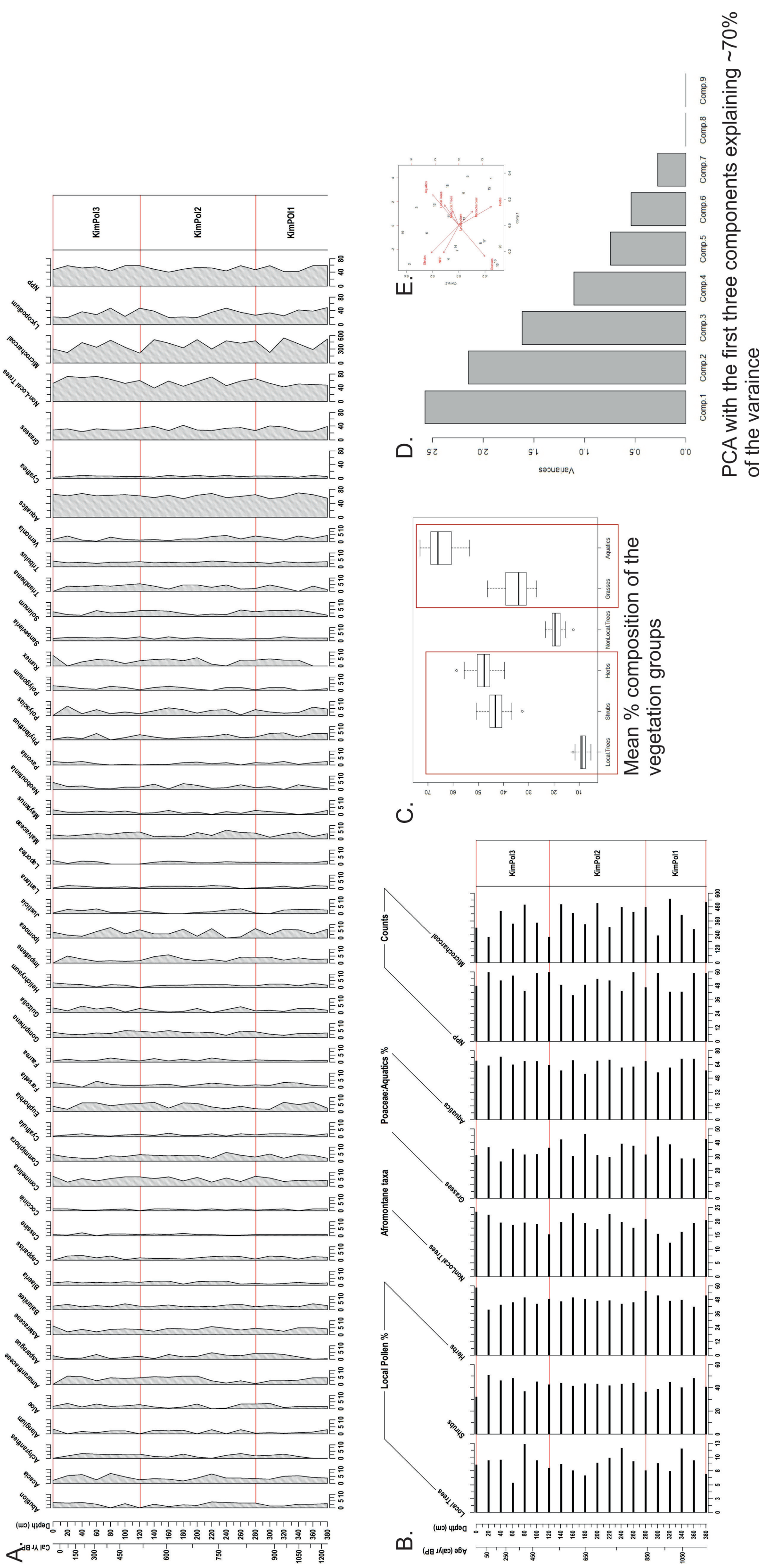



Figure 4.11: Kimana pollen composition over the last 1200 cal yr BP with a PCA graph and biplot showing the spread of the vegetation groups along the driving components.

#### 4.5.6.3 Enkongu pollen and NPP

The Enkongu pollen record displayed poor preservation of pollen especially for the last 20 cm of the sediment core, 38 pollen taxa were identified and three pollen zones delineated using CONISS (Figure 4.13). Overall pollen was categorised as local (immediate to the wetland and surrounding hydric soils on the margin), extra-local from the short grass savannah semi-arid shrubland and woodland, and regional coming from neighbouring highlands carrying montane forest and Ericaceous taxa. Thus the pollen composition was grouped into aquatic, herbaceous, savannah, shrubland-woodland and montane taxa. The minimum taxa identified from any sample was 19 while the maximum was 24 with an average of 16.25±2.69. The Shannon-Weiner indices range from 0.3 to 1,8, with 15 of the samples having an abundance 1 and 9 having an abundance 1.

The pollen was dominated by aquatic taxa (Cyperaceae, *Hydrocotyle, Ludwigia* and *Typha*), with the aquatic taxa composing on average ∼80% of the pollen identified ranging from 43% to 96%. The Cyperaceae to Poaceae ratio varies along the core with an average of ∼27%. A PCA shows that the first three components explain ∼81% of the variance. A b-stick and CONISS analysis delineated the pollen records into three zones. Enkpoll1 covered ∼2000 to ∼1500 cal yr BP (183 to 195 cm), the Enkongu pollen record composition consisted of aquatic taxa at a mean ∼88%, savannah taxa with a mean of ∼8%, montane, shrubland-woodland and herbaceous taxa at ∼2%, ∼1% and 1% respectively. The Cyperaceae to poaceae ratio in this zone was at ∼18%. The dominant taxa in this zone included Amaranthaceae/Chenopodiaceae at ∼3.2%, Asteraceae at ∼3.5% and *Maesa* sp. at ∼1.15, other taxa occurred at less than 1%.

Enkpoll2 covered the period between ∼1500 and 890 cal yr BP (166 to 182 cm), the Enkongu pollen composition in this zone was aquatic taxa with a mean of ∼83% a 5% drop from the previous zone, savannah taxa with a mean of ∼7.6%, montane taxa at ∼6.1% and the herbaceous and shrubland woodland savannah at 2% each. The cyperaceae to poaceae ratio significantly increased to ∼32%. The most dominant taxa was *Maesa* sp. at ∼4.9%, Asteraceae at ∼2.9%, Amaranthaceae/Chenopodiaceae at ∼2.5% and Solanum at ∼1.2% and the other taxa occurred at 1%.

Enkpoll3 covered the last ~890 cal yr BP (from the top to 165 cm), the aquatic taxa reduced to 76% in this zone down from 83%, the savannah increased to ~13.5%, the shrubland-woodland taxa increased to ~4.16%, the herbaceous increased to ~3.1% while the montane taxa reduced to ~2.3%. The cyperaceae to poaceae ratio slightly reduced to ~31%. The dominating taxa were Amaranthaceae/Chenopodiaceae at ~5.35%, Poaceae at ~3.84%, *Commelina* at ~2.22%, Solanum at ~2.3% and Asteraceae at 1.79% while *Maesa* sp. which was dominating in the previous zone reduced to 1%. All the other taxa occurred at 1% except *Acacia* spp. at 1.5% and *Juniperus* sp. ~1.2%.

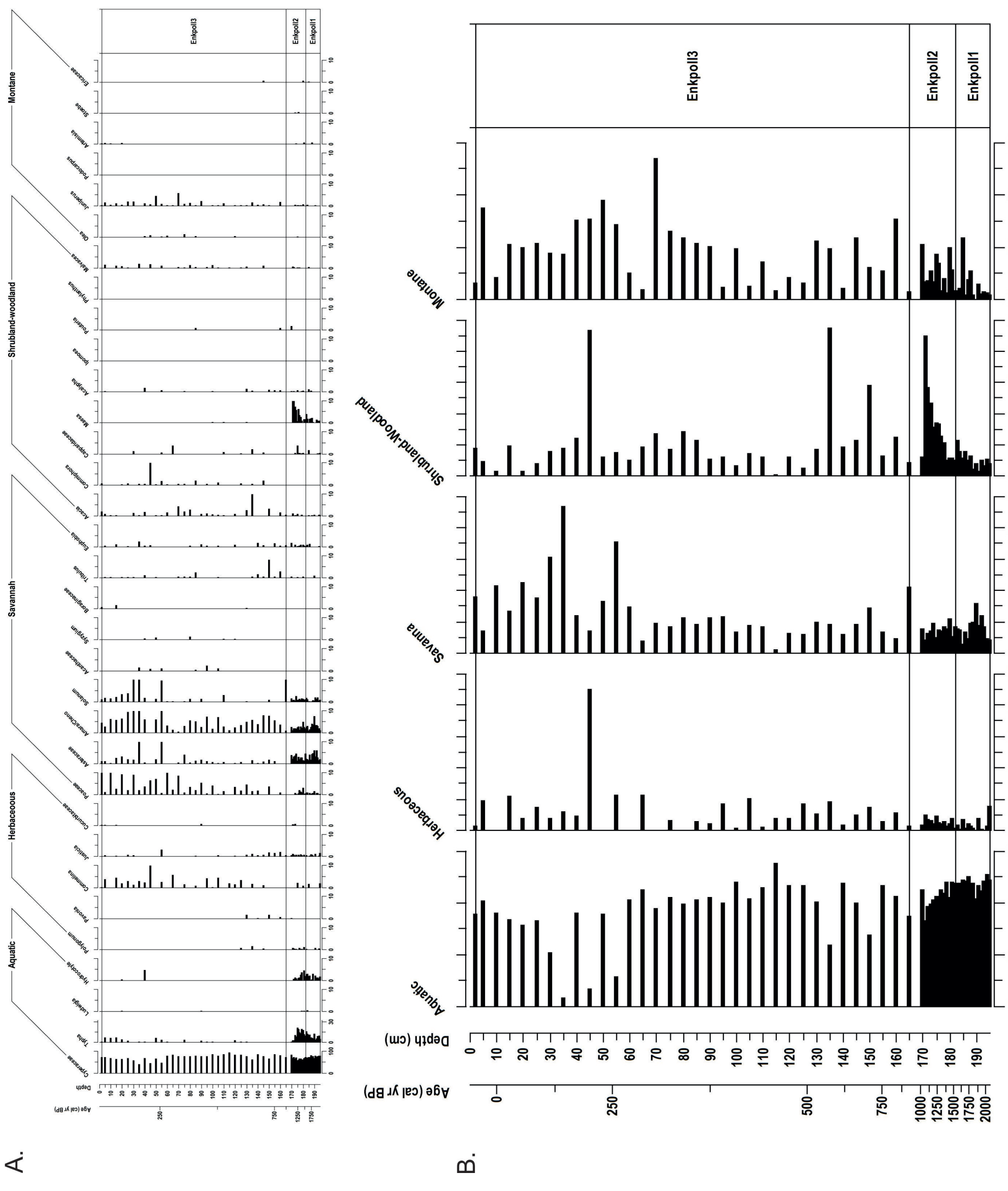



Figure 4.12: Enkongu Narok pollen composition over the last 2000 cal yr BP with a boxplot comparing the pollen taxa percentage levels.

### 4.5.7 Sediment properties

The loss on ignition analysis show that the average organic matter content in the Kimana core is 28.348±2.707% and ranges from ∼5% to 90%, the mean carbonate content is ∼5% and ranges from ∼1% to ∼69% while the siliciclastic content ranges from 10% to 92% with a mean of ∼67%. The mean water content throughout the core is 54% ranging from 12% to 80%. A PCA analysis of the components reveals that the first two components account for 89% of the variance. A CONISS analysis divides the record into three significant zones (Figure 4.14). KimLOI1 from ∼1200 cal yr BP to ∼1000 cal yr BP, KimLOI2 from ∼1000 cal yr BP to ∼500 cal yr BP and the KimLOI3 covering the last 500 cal yr BP.

KimLOI1 (Figure 4.14) has a mean water content of 36% ranging from 21% to 43%, the organic carbon in this level is very low at ∼9% ranging from 5% to 11% while the carbonate content is ∼4% ranging between 2% and 4% the siliciclastic content is high at 87% ranging between 86% and 91%. KimLOI2 (Figure 4.14) experiences an increase in water content to 50% ranging between 12% and 60%, organic matter to 17% ranging between 9% and 61% and carbonate at 6% however the siliciclastic content reduces to 80% ranging between 37% and 86. % last ∼500 cal yr BP, KimLOI3 (Figure 4.14) has the highest water content at 72% ranging 41% and 81%, the organic carbon also significantly increase to a mean of 61% ranging between 23% and 89%. The carbonate content reduces down to 3% and the siliciclastic content also significantly reduces to 35% ranging between 10% and 68%.

### 4.5.8 Particle Size Analysis

The Kimana swamp sediment mean composition is silt, sand and clay at 61%, 27% and 12% respectively, very fine pebbles occurred at 1%. A stratigraphically constrained analysis of particle size data divides the record into three significant zones (Figure 4.14). The bottom zone KimPSA1 covers ∼500 cal yr BP between ∼1200 cal yr BP and ∼700 cal yr BP, the clay content in this zone is at 16% ranging from 5% to 21%, the silt content is at 65% ranging from 36% to 77% while the sand is at 19% ranging from 6% to 58%. The clay content reduces in the second zone KimPSA2 (Figure 4.14) to 11%is change into the second zone while the silt content remains at 65%, the sand

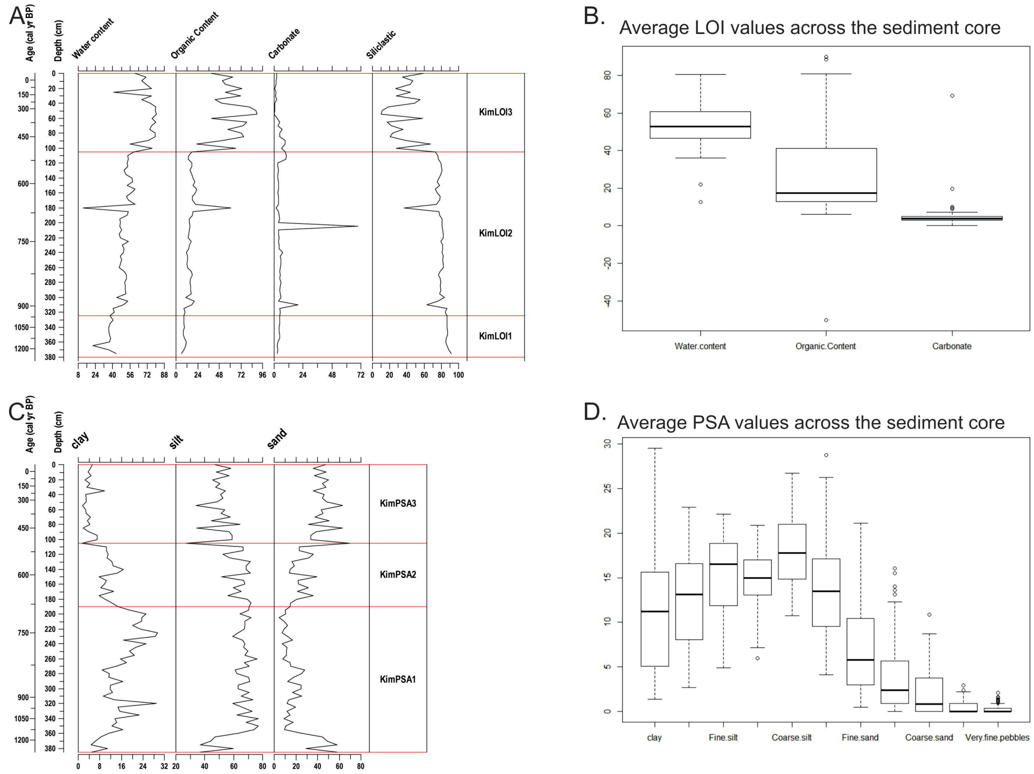


Figure 4.13: Loss on Ignition and Particle Size analysis profiles of the Kimana record divided into zones using CONISS. A) Changes in the water, organic, carbonate and silicate content along the Kimana sediment profile. B) The mean value of the water, organic and carbonate content of the sediment core. C) Clay, silt and sand distribution throughout the Kimana and D) Box plot showing the mean distribution of the different particle class sizes.

content increase to a mean of 23%. In the top zone (KimPSA3, Figure 4.14)) which is the last ∼500 cal yr BP, the mean clay and silt content both reduce to 4% ranging between 15 and 10% and 50% ranging between 27% and 645 respectively. The sand increases to 45% ranging between 31% and 70%.

### 4.5.9 Landscape interpretation

The pollen, NPP and charcoal data, combined with information on the recent development history of the wetland, was used to produce a schematic model (Figure 4.15) of the landscape transitions based on pollen zonation (Figure 4.10). ESAM1 can be summarised as a predominantly semi-arid landscape with woodland and scrub taxa with little fuel connectivity and a relatively constrained Cyperaceae dominated wetland. ESAM2a was a period of increased moisture regime in the wetland and increased fuel connectivity linked to Cyperaceae and *Typha* expansion at the wetland margin. ESAM2b was similar to ESAM2a and was characterised with increased occurrences of standing water. ESAM2c experienced the intensification of human modifications to the channels and wetland area for agriculture. The aquatic and riparian areas may have increased over time (Figure 4.15); change in land use around Esambu swamp has led to the conversion of the lawn areas into agricultural production (*shambas* (allotments) and commercial farming) as marked by the influx of crops and ruderal taxa.

The Kimana ITRAX record identifies the last ∼500 cal yr BP as an important time period, this is accompanied by the lowest magnetic susceptibility values experienced in Kimana after a peak at ∼600 cal yr BP. The Ormakau magnetic susceptibility increases to its highest values over the last ∼500 cal yr BP. The macrocharcoal concentration in the Enkongu, Ormakau and Esambu records is significantly lower than the previous zones with this time period experiencing the lowest mean charcoal concentration over the period of each record. In the Enkongu record the macrocharcoal concentration reduces by half over the last 500 cal yr BP, in the Ormakau record it reduces by almost half over the last ∼700 cal yr BP while in the Esambu record it reduces by ∼65%. However the Kimana macrocharcoal concentration significantly increases (by over 10 times) over the last ∼500 cal yr BP.

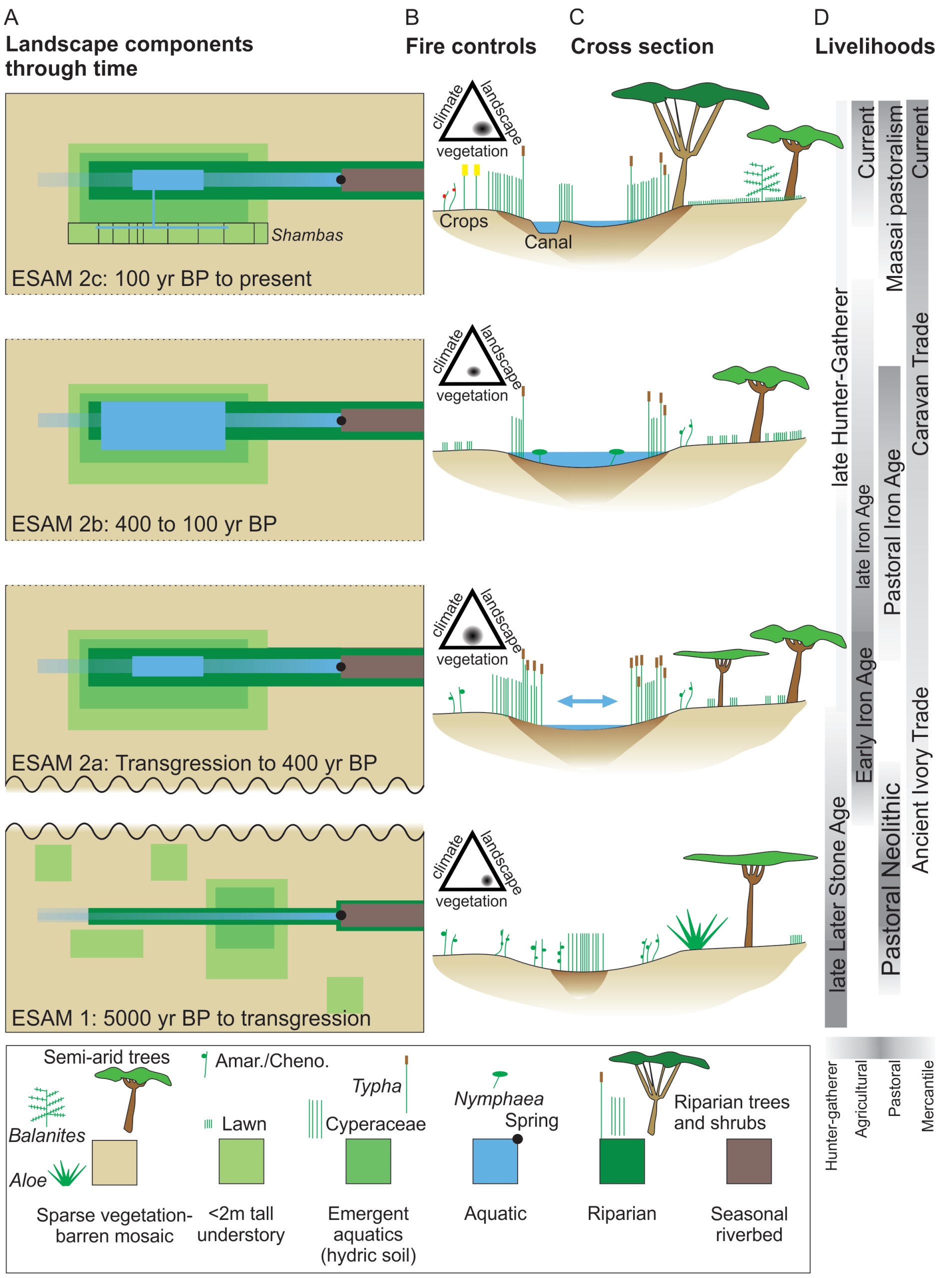



Figure 4.14: Conceptual interpretation of the Esambu catchment based on the CONISS delineation of the pollen, charcoal and modern history data.

# 4.6 Discussion

The Amboseli records cover a period of regional aridity within East Africa evidenced by lowering of lake levels, expansion of dry montane forest types, change in vegetation composition to more open landscapes (Finch et al. 2016, Kiage & Liu 2009*a*, Marchant & Hooghiemstra 2004, Trauth et al. 2003, Verschuren et al. 2000). An increase in human impacts on the East African landscape is also expected with increased population growth and change of livelihoods due to arrival of food crops and technology transitions with the onset of the Late Iron Age (Hamilton et al. 1986, Iles et al. 2014), this has continually increased over the last several centuries following permanent settlement and intensive cultivation. Pastoralism spreading south of the Sahara in this period believed to driven by the increasing aridity as well as diversification of livelihoods with faunal, ceramic and stone artefacts identified from Lake Turkana (Ambrose 2001). and there is, however there is little archaeological evidence of the pastoralism as they reside in areas for short periods of time and then move with all possessions (Gifford-Gonzalez 1998, McDonald 1998). The pastoral system transformed due to diversification of livelihoods and specialised commodity trading (Gifford-Gonzalez & Hanotte 2013).

## 4.6.1 Amboseli vegetation history

Significant environmental changes occurred within the Amboseli as reflected by the changing ecosystem character over the last 5000 cal yr BP. Mid to late Holocene changes in the composition of the ecosystem reflect a combination of shifts in hydroclimate, herbivory and human interaction. The sites are particularly representative of local ($\leqslant$10 km$^2$) vegetation assemblage variability due to the small size of the wetland (Sugita 1994) and distinct land cover components spread across the landscape.

The Esambu, Kimana and Enkongu Narok pollen records were typical of semi-arid grassland savannahs with Afromontane input from Mt. Kilimanjaro and the Chyulu Hills. From c. 5000 cal yr BP to 2000 cal yr BP, the Esambu local ecosystem was open with Amaranthaceae/Chenopodiaceae and Acacia as the dominant vegetation followed by Asteraceae and *Capparis* during the arid period (Figure 4.10). The only aquatic and semi-aquatic taxa identified were Cyperaceae, *Nymphaea, Tapura* and *Typha* that indicate the Esambu wetland was much more restricted at this time. There is a high level of spore abundance and the most dominant species (*Chaetomium, Coniochaeta cf. Ligniaria, Sporormiella*, T. HdV-1020 and cf. *Valsaria* sp.) are indicators of high herbivore density (Gill et al. 2013), most likely as wild herbivores were concentrated within a relatively small area to access water and grazing. Herbivore densities were likely much higher prior to the heightened defaunation pressures of intense ivory trading of the last millennium (Håkansson 2004) further driven by European and Asian markets.

The Maasai have several stories about their origin and past, that more or less accord with these assumption and by the mid 19th century (200 cal yr BP) occupied a large extent from northern Kenya to Central Tanzania (Berntsen 1980) however at low scattered populations (Waller 1985). Archaeological excavations have revealed evidence of livestock domestication (Mutundu 2010) while linguistic evidence and the oral traditions of the Maasai point to the Nile Valley, somewhere in the Sudan-Uganda border area as the original birthplace of the Maasai (Mol 1978). Because of pressures building up in the area the group decided to move southwards into the Rift Valley, probably around A.D.1400 ($\sim$600 cal yr BP). During their migration they displaced, by-passed and absorbed several peoples inhabiting the plains before them. After a temporary residence west of Lake Turkana in the Kerio Valley, the Maasai reached the Serengeti Plains in northern Tanzania probably by the 17th Century (400 cal yr BP).

During the late Holocene, vegetation change is associated with increased temperatures and human impacts (McGlynn et al. 2013), ecosystems perceived as stable such as the Eastern Arc Mountains, Mt. Kenya, Mt. Kilimanjaro and other upland forests experienced a loss of upper montane forest and montane forest taxa which is led to lower montane species shifting upslope and opening up of the landscape (Finch et al. 2016, Mumbi et al. 2008, Rucina et al. 2009, Schüler 2012, Street-Perrott et al. 2007). The vegetation from Kimana swamp (Figure 4.12) records an arid environment over the last $\sim$1200 cal yr BP dominated by annual herbaceous taxa found within the local catchment of the swamp and the river flowing through it. The shrub and local tree taxa combined account for $\sim$50% of the taxa recorded indicating a woody savannah dominated by *Acacia, Balanites, Commiphora* and *Euphorbia*. The herbaceous and shrub taxa identified are indicators of disturbed shrub land which is expected due to the high number of wildlife that occupy the Amboseli landscape and utilise the wetlands and areas around the wetland during periods of low rainfall. Identifying signals of human impact is not possible from the Kimana record as there is no significant change that was observed.

Between approximately 2000 and 400 cal yr BP there is an increase in pollen diversity as the ecosystem surrounding Esambu Swamp became more mesic with an increase in woodland (*Balanites* and *Cordia*) and aquatic taxa (Cyperaceae and *Nymphaea*); Afromontane forest (*Celtis, Commiphora, Croton, Juniperus, Olea* and *Schefflera*) on the adjacent highlands also expanded at this time (Figure 4.9). Variability between woody arboreal taxa and Poaceae relative abundances reflected similar variations in moisture during this more mesic phase (Ssemmanda et al. 2014). This increase in arboreal cover is likely due to an increase in local moisture regimes and an increase in convectional rainfall on Mount Kilimanjaro causing an expansion in forest cover. From c. 400 cal yr BP the number of Afromontane and woody-shrubland taxa reduces, this could be as a result of human impact as forest on the mountain slopes were cleared and developed into the mixed Agroforestry systems typified by the Chagga home gardens on Kilimanjaro. This is also a period of more xeric environmental conditions: several severe drought lasting decades have been recorded in history and observed in sediment records (Gillson 2006, Hulme 1992, Verschuren et al. 2000). The coprophilous spore taxa are still consistently present however the source of the dung could be derived from wild ungulates or domesticated species (Davis & Shafer 2006).

The Kimana pollen composition and abundance does not record significant changes, the proximity of the swamp to the permanent river might be buffering the impacts of changes in local moisture availability, however the Cyperaceae to Poaceae ratio can be interpreted as signal of water levels as has been the case with study of other Amboseli wetlands such as those in Tsavo National Park and Namelok.. Increase in Cyperaceae would be expected during extremely low water periods such as drought due to the exposure of sediment providing habitat for marginal plants like Cyperaceae and *Typha* sp. (Gillson 2004*b*). The Cyperaceae to Poaceae ration increases from $\sim$ to present at Kimana, the highest Cyperace to Poaceae ratio is recorded $\sim$ 200 cal yr BP; a time period recorded as experiencing the most severe drought in Maasai records and also observed in a Tsavo pollen record (Gillson 2004*b*). This significant increase in Cyperace to Poaceae ratio is also observed in the Esambu record.

The landscape around Enkongu is currently arid with very few taxa and this is reflected in the record. From $\sim$2000 to $\sim$ 800 cal yr BP the pollen abundance is higher however, compared to the period after $\sim$800 cal yr BP, it can be interpreted as mesic and is dominated by Cyperaceae, Poaceae, Asteraceae, Amaranthaceae, *Typha, Hydrocotyle, Solanum* and *Maesa* sp. The period between $\sim$800 and $\sim$300 cal yr BP is the most arid; having the lowest pollen diversity and counts as well as the highest cyperaceae:poaceae ratio.

Cyperaceae, *Typha*, Amaranthaceae/Chenopodiaceae, Poaceae, Asteraceae, *Solanum, Commelina* and *Maesa* dominate the Enkongu pollen record. *Maesa lanceolata* is a shrub or small-sized tree often occurring in woodlands, forest margins and streams (Dharani 2011, Dharani et al. 2009). The Enkongu pollen record (Figure 4.13) is difficult to decipher due to the poor preservation and low pollen counts in some levels. The period between $\sim$1300 and $\sim$700 cal yr BP could be reflecting a more mesic phase in this landscape, the low spore count could indicate less use of the swamp or lack of access to the deeper parts of the swamp by herbivores due to the high water level. The ecosystem is dominated by herbaceous taxa, this becomes more xeric between $\sim$700 and $\sim$400 cal yr BP indicated by the significant drop in the herbaceous taxa and accompanied by a significant increase in aquatic taxa. The drying of the swamp during this period increases the surface area for Cyperaceae and *Typha* growth. There is a significant increase in spore counts as well as more animals utilise the swamp for drinking and grazing/foraging. The youngest part of the record indicates more mesic conditions than the previous time period, the Cyperaceae:Poaceae ratio reduces and the spore count significantly reduces.

Migration into the Amboseli landscape continued into the 21$^{st}$ century and by 1928 the penetration of Kikuyu into the Mau area of the Masai Reserve had assumed such significant proportions that the Masai Local Native Council passed a resolution forbidding Kikuyu to cultivate land outside the townships of Narok and Ngong' (Rutten 1992). 'The attitude of the Masai to this problem is not easy to understand or explain. As a tribe they have been emphatic in refusing to agree to penetration but as individuals they have welcomed the Kikuyu' (Berntsen 1980). The Amboseli vegetation composition indicates land degradation especially overgrazing (Majule et al. 2009). According to Jcobs (1975) until the early 1960s the majority of the Pastoral Maasai not only rejected the non-livestock based alternative modes of subsistence available to them, but they possessed strong prohibitions against the eating of agricultural and other non-pastoral foods. Over the last half a century, Maasai have handed over control of the resources they need to survive as pastoralists to farming and conservation (Majule et al. 2009, Rutten 1992). This movement within the Amboseli landscape could be the reason behind the changes observed in the pollen, charcoal, magnetic susceptibility and elemental profiles.Their movement across the landscape coupled with increased numbers of livestock and changes in mobility patterns would leave an environmental signal.

Relationships between NPP and biotic and abiotic conditions are similarly poorly resolved and difficult to interpret with much certainty; but further analyses are necessary to move from analyses of indicator taxa to taxonomic assemblage interpretations. PCA results suggests some possible relationships between taxa, such as Type 121 and Cruciferae and Brassicaceae, as well as *Valsaria* sp., Type 1030, 1020 and 172 with Amaranthaceae and Chenopodiaceae, which could be useful to explore in further studies. The relationship between *Sporormiella* and Cyperaceae suggests that the density of herbivores at the coring location was higher when Cyperaceae relative abundance increased and aquatic pollen diversity decreased when water level was lower. Aquatic taxa were less abundant but Cyperaceae abundances were maintained by growing in the hydric soils of the moist wetland margin.

### 4.6.2 Fire regimes at Amboseli

Fire occurs regularly in savannah ecosystems of eastern Africa, but its spread is dependent on availability of biomass as fuel (Colombaroli et al. 2014, Nelson et al. 2012). In the semi-arid landscape of Amboseli, the variable degree of fuel connectivity limits the spread of fire due to the sparse and patchy nature of the vegetation cover as a result of the arid environment and/or grazing pressure. The degree of vegetation connectivity varies through space and time (Gillson 2004*b*, MEMR 2012, Senici et al. 2015) and interact with broad scale variations in climate that influence vegetation dynamics, landscape geomorphological variation and herbivory (Norton-Griffiths 1979). Anthropogenic modifications also influence the patchiness of vegetation (Figure 4.3D), accumulation of dry fuels and ignitions, and these interactions have persisted on East African landscapes since at least the Pleistocene (James et al. 1989) but have intensified in the recent past (Archibald et al. 2012, Bowman et al. 2011). Barren ground patches (Figure 4.3A) limit the spread of fire in many ecosystems (Burrows et al. 1991, Pausas & Paula 2012, Senici et al. 2015) and grazing intensities can reduce fuel connectivity leading to short-stature lawns, especially during the dry seasons when conditions are more conducive to burning (Archibald et al. 2005, Hempson et al. 2015). Yet, the perennial wetlands support locally mesic and hydric soils conducive to continuous vegetation that provides fuels that may burn during dry periods. Fuel limitation as a bottom-up control on fire has become quite important in the East African savanna grasslands over last c. 6000 to 5000 cal yr BP (Nelson et al. 2012).

Anthropogenic fires are rife as people use fire to modify landscapes for their livelihoods. Although burning occurs as a significant traditional resource management practice and shows ecological advantages, human set fires are viewed by conservation authorities as destructive and a cause for environmental degradation (Kamau & Medley 2014). Low levels of biomass burning from 5000 to 1300 cal yr BP reflects absence or very low level of local burning, this is a dry time period and the pollen record reveals low abundances of woody vegetation and lack of understory vegetation. This late Holocene savannah ecosystem was likely characterised by sparse tree cover with reduced biomass and reduced fuel connectivity, especially within the xeric Amaranthaceae/Chenopodiaceae. Lack of continuous understory vegetation and abundant woody species means increased areas of open space that act as fire breaks for ground fires. The reduced grassy and Cyperaceae dominated wetland fringe may also have had reduced biomass abundances - thus bottom-up controls of fire were important.The Kimana, Ormakau and Enkongu Naork macrocharcoal records show a presence of fire throughout the records with increase over the last 500 cal yr BP, the Kimana and Ormakau record show presence of fires at low levels with several peaks with the Enkongu record shows a consistent presence of fire.

At Esambu, the period between 1300 to 400 cal yr BP has significantly high charcoal accumulation rates (average of 249 pieces $cm^{-2}$ $yr^{-1}$), this is accompanied by an increase in Afromontane, woody and understory vegetation which is interpreted as higher regional increases in moisture that increased fuel connectivity thus increased local burning (Ssemmanda et al. 2014). This is because higher precipitation allows for higher net primary production which provides the biomass for burning (Van Der Werf et al. 2008). The reduced abundances of Amaranthaceae/Chenopodiaceae increased Poaceae and increased woody taxa richness and abundances suggest modest increases in regional moisture, changing from a more xeric ecosystem to semi-arid savannah woodland. The increased moisture, likely from climatic mechanisms related to regional atmospheric circulation changes over the Horn of Africa (Tierney et al. 2015) and equatorward, increased the woody fuel loads of the landscape and increased the continuity of fuels from the Poaceae savannah and Cyperaceae-Poaceae-*Typha* wetland. This was followed by a period between 400 and 300 cal yr BP where the mean CHAR values remained low until present (65 pieces per $cm^{-2}$ $yr^{-1}$) due to reduced fuels amenable to burning. Published literature on the historical human use of fire in the region is scant and it is difficult to separate natural and human agency in influencing the temporal patterns in fire activity. Current populations near Chyulu Hills apply fire to biomass for a number of reasons between May and October (Kamau & Medley 2014) and MODIS satellite observations show relatively fires in the Esambu area that occur almost exclusively during July.

The Esambu, Enkongu and Ormakau macrocharcoal records show a significant decrease in the macrocharcoal concentration over the last $\sim$500 cal yr BP, while the Kimana record shows a significant increase. Esambu, Enkongu and Ormakau are all groundwater fed while Kimana is ground water fed as well as having the Kimana river flow right next to it. Thus the Esambu, Enkongu and Ormakau would signal local climate and landscape change faster than Kimana which is buffered by the river. The decrease in macrocharcoal concentration which is preceded by a peak could be a signal of increased fire in the landscape just as the Maasai arrive, this then reduces because in the arid landscape, there is no more fuel to burn. With increased burning, areas around the river as well as Kimana swamp would be the only regions of the landscape with enough fuel to burn and thus the increase in the macrocharcoal concentration.

### 4.6.3 Regional drivers of ecosystem change

East African savannah palaeoenvironmental records indicate a highly variable environment from the mid Holocene to present environment (Gillson 2004*a*, Lamb et al. 2003, Rucina et al. 2010, Thompson 2002, Verschuren et al. 2009). The pollen records are typical of a semi-arid savannah with the common savannah trees *Acacia, Balanites, Commiphora, Cordia* and *Terminalia*, grasses, local aquatic taxa (Cyperaceae, *Nymphaea* and *Typha*) combined with the deposition of well-dispersed montane taxa *Juniperus, Podocarpus* and *Olea*. This assemblage is similar to other sites within the region (Gillson 2006, Rucina et al. 2010); from c. 5000 to 2000 cal yr BP the Esambu record was dominated by Amaranthaceae-Chenopodiaceae, Poaceae and sparse woodland and shrub taxa that characterised the semi-arid Amboseli landscape and reflected a drier climate than the present day. Regional records show increase and domination of Poaceae pollen in Lake Turkana and Tanganyika, increase in *Acacia* and *Dodonea* (Olago 2001) with a decrease of high altitude taxa such as *Hagenia, Hypericum* and *Stoebe* from Lake Baringo between c. 5000 and 2500 cal yr BP (**?**) and a decrease in levels of Lake Abiyata, Ethiopia (Chalié & Gasse 2002).

Significant environmental changes have occurred over the past 5000 years within the Amboseli landscape, after the mid Holocene thermal maximum at about 5000 cal yr BP (Adger et al. 2013) temperatures in East Africa began to decrease (Kiage & Liu 2006). Insolation changes driven by orbital changes of the Earth's obliquity and precession affected latitudinal thermal gradients and seasonality that influenced the characteristics of the African monsoon during the second half of the Holocene (Hély et al. 2009). The East Africa environment shifted toward drier conditions characterised by reduced precipitation, increased evaporation, and/or an extension or intensification of the dry season (Marchant & Hooghiemstra 2004). Around 5000 cal yr BP different sites across East Africa showed an increase in abundance of Poaceae at the expense of arboreal taxa in the lowlands, coupled with increases in montane forest taxa indicative of drier climatic conditions, such as *Juniperus, Podocarpus* and *Olea* (Olago et al. 1999, Taylor 1990, Umer et al. 2007, Vincens 1986, Vincens et al. 2005). Pollen data from Lake Victoria show an increase in Poaceae pollen starting from 4000 cal yr BP (Kendall 1969) and isotopic data at Lake Malawi show an increase in $C_4$ vegetation, initiated again around 4000 cal yr BP, that is indicative of increasingly arid environmental conditions (Castañeda et al. 2009). A spike in eolian dust around 4000 cal yr BP also indicates regionally dry conditions contributing to dust deposition within the Kilimanjaro glacier (Thompson 2002). Determining the cause of the hiatus in the sedimentary record is problematic as it could be lack of sediment accumulation or erosion of the sedimentary archive during a period of increased flooding.

Archaeological studies from Tsavo show evidence of domesticates from faunal diagnostic finds identifying cattle *Bos taurus* (Wright 2007) after 3700 cal yr BP. At Esambu, the presence of Poaceae >60 $\mu$m which can be interpreted as cereal is also accompanied by increased presence of macroscopic charcoal which could be due to increased fuel availability as it is highly correlated to high *Typha* levels or increased human activity such as farming. Sustained human impact, in the form of intensive crop cultivation across East Africa, occurs much later with the introduction of *Zea mays* (Lamb et al. 2003, Lejju et al. 2005). Increased presence of *Acacia*, Amaranthaceae/Chenopdiaceae, *Balanites* and Poaceae in the uppermost samples from Esambu dating to the last 200 years are indicative of a drier environment although could also be a consequence of decrease in elephant population which have a major influence on vegetation composition, distribution and structure (Håkansson 2004, 2012).

After 2000 cal yr BP the Esambu record is characterised by high diversity of savannah and montane taxa; most likely a local signal of increased moisture, this signal is not recorded in the Kimana or Enkongu records. Between 2150 and 800 cal yr BP another record from the Amboseli Basin also shows a relatively moist environment with the increased presence of *Syzigium* a tree taxa that can grow within water-logged soils (Rucina 2011). After c. 800 cal yr BP the record from Namelok Swamp shows a drier environment with the increased presence of Amaranthaceae and Poaceae and decreased levels of *Syzigium* (Rucina 2011). A dry climate, with increasing signals of human-ecosystem interaction is observed from Lake Naivasha and Lake Challa (Lamb et al. 2003, Westerberg et al. 2010). The last 300 to 400 cal yr BP were characterised by a decrease in abundance and variability of woody taxa and an increase in the Cyperaceae-Poaceae ratio indicative of increased aridity, or increased human activity and the impact on the surround forests; the accompanied increase in charcoal abundance also might be due to increased natural and anthropogenic ignition.

## 4.7 Conclusions

The semi-arid landscape of Amboseli, Southern Kenya, East Africa, is punctuated with intermittent groundwater-fed wetlands that record past ecosystem changes in the sedimentary geoarchives. The dynamic nature of the local ecosystem indicates the interactions between the climate, ecosystem, herbivores and humans. Pollen taxa recorded come largely from within the catchment. Sparse woodland and shrub taxa characterised the semi-arid Amboseli landscape from c. 5000 to 1700 cal yr BP with a change from a drier environment with a small aquatic area and expands as regional climates become moist. Vegetation taxa diversity increases and montane forests on adjacent mountains expand, most likely to lower altitudes after 2000 cal yr BP.

Current land cover and land use impacts over the past centuries years have resulted in the conversion of a portion of the wetland to agricultural production as human activity becomes much more visible within the Amboseli landscape. Sedimentary hiati are inferred from the sedimentary record and thus caution is advised due to the chronological uncertainties. This limits interpretation and the complex spatial pattern of climate due to the local and regional factors can be problematic even for replicate reconstructions within the same region. Comparison between sites based on chronology is challenging due to the complexity associated with radiocarbon dating which can rarely be calibrated to decadal scale combined with the uncertainty of age-depth models where prior information is not available.

The research finds that the last ~500 cal yr BP, there has been a significant change in the Amboseli ecosystem observed in the pollen, fire and geochemical records. The timing of this changes occur around the time that the Maasai are believed to have arrived in the Amboseli ecosystem and so this change across the landscape could be a signal of that. The physiochemical signals indicate aridity interspersed with periods of increased moisture which could be increased local rainfall or increased water table levels caused by increased precipitation on Mt. Kilimanjaro. The timing of the vegetation, sedimentary and fire regime changes across the swamps indicates localised controls; this emphasises the differential responses ecosystems of each wetland to driving factors such as climate and human impacts.

The current Amboseli landscape has no analogue in the last ~5000 cal yr BP; this indicates the highly dynamic nature of wetlands as current conditions impact on the swamp leading to the emergence of novel ecosystems reflecting modern day pressures. Ecosystem-human-environment interactions are inherently dynamic and knowledge on the interactions is vital to inform adaptive management, particularly in a future characterised by increased impacts climate change and competing resource-use demands. Long term records that provide local understanding of these interactions are excellent sources of information for providing a foundation to inform sustainable environmental management decisions (Gillson & Marchant 2014). Such long term perspectives are crucial under predicted climate change and associated livelihood impacts (Ipcc 2001), so that suitable responses to ensure sustainable management practices can be developed.

# CHAPTER 5

## Synthesis and conclusion

## 5.1 Overview

Palaeoenvironmental, archaeological and historical archives record information about environmental variability and human activity and are used to reconstruct and interpret the changes that have occurred in the past. Analyses of the Nyabuiyabuyi (since ∼16,000 cal yr BP), Kimana, Ormakau and Esambu (since ∼5000 cal yr BP) sediment cores provides insights into how the environmental conditions have changed and how human agency impacted this landscapes through time and across space.

By summarising the findings, the research gaps which could be a focus for future studies are highlighted. The new records improve our understanding of the regional response of ecosystems to change i.e. the Nyabuiyabuyi record showing coeval changes with other highland sites (figure 5.1). The records also highlight the sensitivity of wetlands to record local and regional ecological change and hence the differences in pollen composition and abundance as well as the macro charcoal trend in the Amboseli swamps.The elemental profiles provide continuous records of elemental deposition through time, an important archive for understanding human impacts on the global cycle of elements. Magnetic susceptibility, elemental analysis and sediment component analysis indicate increased enrichment of heavy metals, organic matter and sand in the upper portions of all the cores suggesting increased sedimentation. The values of the different parameters analyzed, depend, on the autochthonous material as well as allochthonous whether due to anthropogenic or environmental impacts.

This chapter provides a summary of a synthesis of ecological change using the new data from the Mau and Amboseli landscapes as well as existing palaeoecological, archaeological and historical records that combine proxies and multidisciplinary datasets to develop a comprehensive narrative of human-environment interactions within East Africa. Understanding the ecological responses to external stressors through the study of long-term ecosystem change can improve our understanding of ecosystem responses to different kinds and rates of change, ecosystem resilience, biodiversity evolution and prediction of future scenarios.

The certainty of environmental information from palaeoecological data is increased by our understanding of the environment from which the sediment archive is sourced. Sediment accumulation controls and vegetation dynamics within the wetland influence environmental reconstruction. This can then have a practical application in programs that address challenges caused by changes in climate, population growth, economic development by informing biodiversity management, land use policies and other ecological efforts.

## 5.2 Mau and Amboseli environmental history synthesis

Large fluctuations in precipitation and temperature characterise the African tropics (Chalié & Gasse 2002), these fluctuations are recorded in palaeoecological records and help us understand changes in natural and anthropogenic drivers of change. Understanding socio-ecological change through time requires a combination of high quality representative palaeoecological records, well preserved archaeological artefacts and well preserved historical records ideally across the landscape; however theses three conditions rarely occur in unison. The multi-proxy approach (analysis of grain size, magnetic susceptibility, biogeochemistry, charcoal and biological remains) on wetland sediment from five sites, one from the Mau Forest Complex and four from the Amboseli landscape provides an understanding of the changes in composition between autochthonous and allochthonous sources, changes in sedimentation rate, reconstructing the vegetation change, fire regimes and climate variability.

A multidisciplinary approach provides additional context for analysing particular signals, such as environmental changes being due to climate change or human activity, that are impossible to disentangle with only one line of evidence. For example, if a pollen stratigraphic study reveals a major shift from closed forests to open grasslands, additional lines of evidence would be needed to understand the causal mechanisms of anthropogenic deforestation or climatic change.

Descriptive studies of paleoenvironments are useful for understanding past patterns of environmental variability; but, can often be difficult to robustly identify the mechanisms underpinning environmental changes. For example, sedimentological and paleoecological data derived from a geoarchive give insight about changing vegetation cover. Yet the mechanisms driving transitions from one ecotone to another may be ambiguous as to being climatically-driven, or caused by herbivory or human activity (Gelorini et al. 2012). It also may also be possible for a combination of these causes to produces the past variations. This highlights the need for high-resolution, multi-proxy, multi-site studies of environmental change and a rich combination of approaches to begin to resolve the varying relative importance of the major ecosystem processes interacting at multiple spatiotemporal scales (Dearing et al. 2010, Lejju et al. 2005, Lotter et al. 2008).

Evidence of use of fires to manipulate habitats has been recorded in the fossil record as old as 25,000 cal yr BP (Kamau & Medley 2014). Burning has been documented as a traditional land management practice globally to stimulate vegetation for grazing, clear land for cultivation, kill disease causing vectors such as ticks, clear bushes to improve plant biodiversity etc. (Kamau & Medley 2014). This study sought to understand and provide more records of high resolution data on how the wetlands has changed over time, develop the patterns of the vegetation and fire regime shift at different sites as well as processes such as erosion and deposition through the analysis and interpretation of the various multi proxy data sets.

The MFC record covers the last 16,000 cal yr BP, while the Amboseli records cover the last ~5000 cal yr BP. The Nyabuiyabuyi macrocharcoal record shows that fires occurred frequently with a high mean charcoal concentration and charcoal accumulation rate especially between ~16, 000 cal yr BP and ~13, 000 cal yr BP, within the last ~3000 cal yr BP there have been fewer larger fires than compared to the last ~16, 000 cal yr BP. The Amboseli macrocharcoal records signify very variable fire histories across the sites, Ormakau, Esambu and Enkongu record decreased fire concentration within the last ~500 cal yr BP while Kimana records a significant increase. Esambu records high charcoal concentration and macrocharcoal accumulation rates between ~ 450 to ~ 800 cal yr BP

The period covered by the records is recognised as a period in East Africa where the climate is known to have experienced large fluctuations ranging from a cold and wet Late Glacial period, a cold and dry Younger Dryas, a warm and wet early Holocene, progressive aridity from the mid Holocene (Marchant & Hooghiemstra 2004) and periods of precipitation (Cohen et al. 2005, Verschuren et al. 2000). The last millennium particularly is characterised by population growth coupled with changes in mobility and settlement patterns such as sedentary agriculture, arrival of the European explorers, ivory trade and colonial land use management practices (Håkansson 2012, Lane 2011, Malhi et al. 2013, Wright 2007).

Pollen, NPP, macroscopic charcoal, loss on ignition, particle size analysis and XRF geochemical analyses were performed on five radiocarbon-dated sediment cores from the Eastern Mau Forest and the Amboseli savannah providing us with one new data set of change since the Late Glacial period and four from the mid Holocene. Precipitation and moisture balance changes have been very important to the evolution of landscapes and, in particular, to livelihood practices, which remain important in the region. Global changes to atmospheric circulation driven by orbital changes have had important implications for the long-term positioning and seasonality of the Intertropical Convergence Zone. Global temperatures have varied in response to changes in orbital patterns, atmosphere and ocean circulation, and greenhouse gas compositions, and these processes have influenced the regional climate of East Africa.

A synthesis of the last $\sim$16,000 cal yr BP is presented in two time periods; the Late Glacial period including the Younger Dryas and the Holocene combining the Early, Mid and Late Holocene and where available, archaeological and historical information. A section on wider implications, applications and challenges of long-term studies is then followed by the conclusions.

### 5.2.1 The Late Glacial Period

African palaeovegetation and palaeoclimatological records that extend further than the LGM are characterised by hiatuses (Mumbi et al. 2008, Olago 2001) thus not giving continuous histories. The Nyabuiyabuyi record covers the last 16,000 cal yr BP of change within the Eastern Mau forest, and a potential hiatus is noted from the Early to Late Holocene. The change in vegetation type observed in the Nyabuiyabuyi record is similar to other East African highland palaeorecords such as the Rukiga highlands, Ruwenzori Mountains, Mt. Elgon and Mt. Kenya covering the Late Glacial Period (Figure 5.1). The records indicate an increase in amount of montane forest taxa following the late glacial conditions.

The Late Glacial period dating from ∼22,000 cal yr BP to 14,000 cal yr BP is described as cool and dry (Mumbi et al. 2008) with a migration of highland taxa to lower altitudes and expansion of grasslands at the Eastern Arc Mountains. Between 16,000 cal yr BP and 13,000 cal yr BP, the Nyabuyabuyi record is dominated by Afromontane taxa indicative of cool and dry conditions accounting for 32% of the composition. Within the Rukiga highland swamps (Ahakagyezi and Muchoya), *Potamogeton* and *Callitriche* and Ericaceous vegetation dominate the pollen record prior to ∼14000 cal yr BP with *Callitriche* and Ericaceous vegetation indicating cool, dry conditions (Figure 5.1).

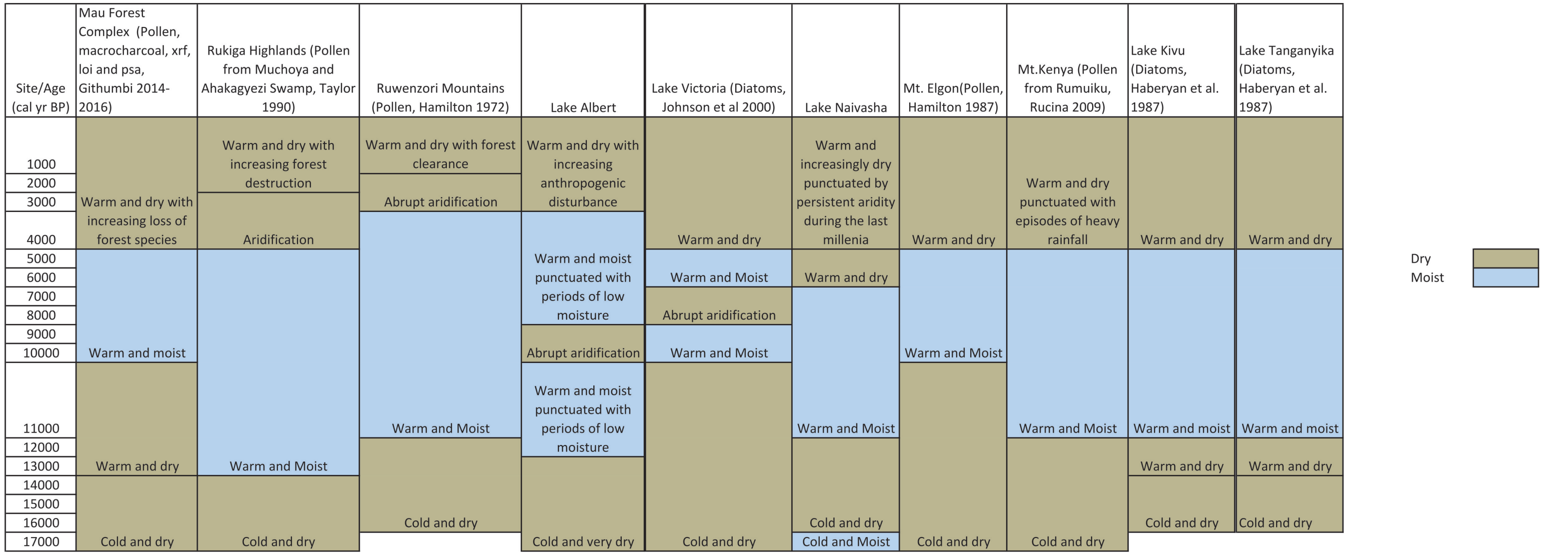



Figure 5.1: Late Glacial period environmental synthesis from 10 East African sites summarising the climatic changes from ∼17,000 cal yr BP.

A record from North Africa (Middle Atlas) indicates a herbaceous rich grassland from ∼18,000 to 8000 cal yr BP, while the Sahara records signal a peak in aridity and aeolian activity around 18,000 cal yr BP with the Younger Dryas marked by warm and cold episodes. In Central and West Africa, grasslands expanded and cold dry forests occurred in refugia which experienced a lowering of montane taxa replacing lowland forests (Olago 2001). Pollen records from East African highlands indicate dry montane scrub and high Poaceae levels coupled with the migration of upper montane taxa between ∼22,000 cal yr BP and ∼22,000 cal yr BP around the Burundi Highlands (Bonnefille & Riollet 1988), Rukiga highlands (Taylor 1990), Mt. Elgon (Hamilton 1987) and Mt. Kenya (Rucina et al. 2009). A diatom record from Lake Abiyata in Ethiopia is composed of mainly coarse sand and a low diatom content of mainly saline-alkaline species (Chalié & Gasse 2002) indicating low water levels. Low lake levels were recorded from Lake Kivu and Tanganyika between ∼16,000 and ∼14,000 cal yr BP. Lake Tanganyika shows a rise in lake level ∼13, 000 cal yr BP while Lake Kivu water levels increased ∼12, 500 cal yr BP (Haberyan & Hecky 1987, Hamilton 1982).

The Younger Dryas is recorded as a period of increased aridity within the Nyabuyiyabuyi record. The Mau Afromontane taxa reduced to 29% between 13,000 cal yr BP and 10,000 cal yr BP, Urticacae and Ericaceae while the shrub and herbaceous taxa reduced in composition and abundance indicating a dry period. From the Rukiga highlands, the record is dominated by *Erica-Myrica*, however there was an increase in *Hagenia* and Urticaceae and most of the pollen identified from this period indicated an altitudinal shift in the vegetation belt indicating warming up between ∼14,000 and 11,000 cal yr BP. Diatom records show a low diatom content with an assemblage similar to the Late Glacial Period (Chalié & Gasse 2002), and a lowering of Lake Malawi (Castañeda et al. 2009).

The North African record (Middle Atlas) indicated a spread of deciduous and pine forests while *Artemisia* and Poaceae record the transition from the Younger Dryas to the Holocene; the Central and Western African forests also experienced an expansion with a marked increase towards the Holocene (Olago 2001). Prior to the Holocene, a

Lake Victoria record does not record the presence of a forest and if present it was inferred as occupying a small extent (Kendall 1969). Vincens (1986) records an increase in herbaceous taxa and progressive increase of low land forest taxa accompanied by a decrease in upper montane taxa. Peak *Artemisia* values ~14,000 cal yr BP from Sacred Lake on Mt. Kenya indicates that peak aridity occurred during the Younger Dryas (Olago et al. 2000). The Rumuiku record from Mt. Kenya shows an increase in Cyperaceae, Poaceae, Asteraceae and Urticaceae and a decrease in *Hagenia, Polyscias* and *Schefflera* (Rucina et al. 2009). However some records indicate existence of wet conditions e.g. Lake Masoko (Garcin et al. 2007). The Nyabuiyabuyi pollen record shows an increase in pollen composition and abundance, a recovery from the Younger Dryas period. The dominating taxa at this time period indicate the development of warm and moist conditions

### 5.2.2 The Holocene

The environmental changes that occurred throughout the Holocene had significant impacts on the opportunities for human occupation and development in East Africa. During the Early Holocene when the African Humid Period occurred the Sahara Desert in North Africa was completely vegetated by grasses and shrubs (DeMenocal & Tierney 2012, Jolly et al. 1998), caused by increased Northern Hemisphere summer insolation leading to a rise of dry season precipitation and a reduction in rainfall seasonality in East Africa (Tierney et al. 2011). There was an increase in local tree cover across East Africa with many taxa reaching maximum abundances and diversities (Vincens et al. 2005). There was a general northward shift of vegetation zones resulting in a greener Sahara (Hoelzmann et al. 1998). This occurred in upland areas where montane forests expanded to higher elevations (Taylor 1990) and in the lowlands; notably, with the expansion of wetter *Miombo* woodlands (Vincens et al. 2005).

The Early Holocene is described as a rapidly fluctuating but generally wet and warm episode (Chalié & Gasse 2002, Schüler et al. 2014) with higher precipitation than evaporation. Within the Nyabuiyabuyi sediment record, a potential hiatus occurs from ~8000 cal yr BP to ~3000 cal yr BP. Other East African wetlands also record hiatus from the early Holocene; Lake Magadi, Rukwa, Victoria, Mubwindi swamps, Mt. Elgon are examples of records with this hiatus (Hamilton 1987, Stager et al. 2003, Vincens et al. 2005). This period is characterised by a slight increase in Afromontane taxa in the Nyabuiyabuyi record but not to levels experienced in the Late Glacial period, while in the Rukiga highlands there is an increase in pollen and spore taxa indicating presence of moist montane forest within the catchment (Taylor 1990).

Lake levels in tropical East Africa were high due to a strengthened monsoon caused by insolation patterns (Demenocal et al. 2000), for example, the Suguta Basin in the northern Kenya Rift Valley is currently dry but, until ~6000 cal yr BP, held a 300 m deep lake which overflowed northward into Lake Turkana (Garcin et al. 2012, Junginger et al. 2014). Significantly wetter conditions have also been inferred for southern East African sites such as Lake Tanganyika (Tierney et al. 2008) and Lake Rukwa (Barker et al. 2002, Vincens et al. 2005). This humid phase resulted from increased monsoon precipitation induced by orbital forcing, East African lakes experienced increases in volume with many of them reaching their highest stands for example in Ethiopia four lakes merged into a single lake reaching 112m above todays lake level (Chalié & Gasse 2002).

This humid phase might be the cause of the potential hiatus that is recorded from the Early Holocene at Nyabuiyabuyi. High precipitation and increased run off would mean the sediment record is always getting reworked by the new material eroded into the wetland. Diatom records show and increase in fresh water species as well as planktonic species due to the increase lake levels experienced (Chalié & Gasse 2002). Lake Tanganyika experiences increased rainfall inferred from fern spores and diatom assemblage changes (Barker et al. 2001). There was development and expansion of Afromontane forest on the slopes of Mt. Kilimanjaro (Schüler et al. 2014), increased forest taxa pollen in the Sacred Lake taxa (Coetzee. J.A. 1967, Van Zinderen Bakker & Coetzee. J.A. 1988).

Increased lake levels, reduced *Artemisia* and *Cedrus* and a *Pinus, Olea-lentiscetum* rise indicating increased moisture are observed from North African lake and marine cores (Olago 2001) until ~5000 cal yr BP. Forests established and replaced grasslands as observed from Lake Barombi Mbo, Lake Bosumtwi and the Niger Delta prior to a deterioration at ~4000 and ~2500 cal yr BP (Olago et al. 2000). An evergreen then deciduous forest is inferred from the Pilkington Bay record (Kendall 1969) which was not present before the Holocene. A steady increase in $\delta^{13}C$ values from 6500 yr BP indicates opening up of the local swamp vegetation in Deva-Deva Swamp on the Lukwangule Plateau (Tanzania; Finch et al., 2009). The Rumuiku stratigraphic and pollen record suggests that the swamp dried out during or soon after the Younger Dryas followed by a transition to a mixed montane pollen record with *Schefflera* and *Polyscias* replacing *Hagenia* and *Stoebe* (Rucina et al. 2009).

Climate abruptly became drier across East Africa around ~5000 cal yr BP, although the cause of this abrupt climate change has been difficult to interpret due to variable responses of hydroclimatic proxy records (Demenocal et al. 2000, Junginger et al. 2014, Kröpelin et al. 2008, Liu et al. 2007, Weldeab et al. 2014). The Mid Holocene is characterised by reduced precipitation leading to reduction in lake levels comparable to present day levels Maitima (1991), Olago (2001) and can thus be described as hot and dry. East African lakes tended toward negative water balances and created more isolated basins (Garcin et al. 2012, Olaka et al. 2010, Russell et al. 2003, Tierney et al. 2008).

The Afromontane taxa in MFC dropped to 25% during the last 5200 cal yr BP. The drop in Afromontane taxa was accompanied by a drop in upper montane and montane taxa and an increase in shrubs, herbs and grasses. The decrease in upper montane taxa does not recover during the Holocene however lower montane forest taxa increase. This implies the climate is warming and drying as the MFC wet montane forest becomes replaced by forest taxa suited to drier conditions. The sudden precipitation regime change experienced during the mid-Holocene is recorded by change in $C_3$ and $C_4$ vegetation on Mt. Kenya (Sacred Lake) and Lake Rutundu (Ficken et al. 2002, Wooller et al. 2000). Different sites across East Africa show an increase in abundance of Poaceae at the expense of arboreal taxa in the lowlands, coupled with increases in montane forest taxa indicative of drier climatic conditions, such as *Podocarpus, Juniperus*, and *Olea* (Olago et al. 1999, Taylor 1990, Umer et al. 2007, Vincens 1986, Vincens et al. 2005). In general, early Holocene wetter montane forest components were replaced by taxa that were more drought resistant (Kiage & Liu 2006). Local vegetation changes often reflected the regional trends. Grasslands expanded and replaced wooded areas in numerous regions.

The Amboseli swamps would have developed around this time period of aridity; the Amboseli swamps are all ground water fed and increased snow melt water from the Kilimanjaro glacier would increase the amount of water available and raise the water table causing formation of swamps in areas with springs. The four Amboseli sediment records cover the last ~5000 cal yr BP, Esambu swamp covers the longest period which is the last ~5000 cal yr BP, Ormakau covers the last ~ 2700 cal yr BP, Enkongu covers the last ~2000 cal yr BP while Kimana covers the last ~1200 cal yr BP. The Amboseli pollen records are typical of the current semi-arid savannah landscape with the common savannah taxa plus montane taxa from the surrounding highlands. From 5000 cal yr BP to 2000 cal yr BP, the Esambu pollen record was dominated by the semi-arid taxa (*Acacia, Capparis*, Amaranthaceae-Chenopodiaceae, Asteraceae and Cyperaceae), and low charcoal concentrations.

*Cordia, Croton, Cupressus, Ficus, Juniperus* , *Olea* and *Pinus* dominated the MFC pollen record during the last ~5000 cal yr BP ranging between 5% and 10%. This was accompanied by increased Poaceae % and NPP counts reflecting a more open forest ecosystem suitable for grazing and foraging by wildlife due to reduced undergrowth inferred from reduced shrub taxa. This reduction in pollen abundance was a significant drop from the earlier zones reflecting a drier conditions.

*Cupressus* and *Pinus* are exotic taxa that appear at the top of the record and dominated the pollen abundance, a signal of human impact through the introduction of new taxa. Within Rukiga highlands around 4000 cal yr BP, charcoal concentration increased and *Erica-Myrica* composition was replaced by the the aquatic pollen as the dominant pollen, there was a general increase in *Celtis* comp., *Olea africana, Olea capensis* and *Podocarpus* pollen reflecting increased dry forest extent. Charcoal concentration reduced towards the top due to human-induced burning in order to promote the regrowth of *Pycreus* which is used locally as thatch (Taylor 1990). This is because burning was now controlled to promote growth of *Pycreus* and the increased fire occurence would prevent growth of woody material thus reducing accumulation of woody material that would act as fuel.

Within the Nyabuiyabuyi record, the last ~3300 cal yr BP was delineated as a zone experiencing the least fires over the last ~16,000 cal yr BP with two peaks recorded at ~1600 cal yr BP and ~300 cal yr BP. The Esambu charcoal record indicated ~5000 to ~2000 cal yr BP was a period of low charcoal concentration. Pollen data from Lake Victoria show an increase in Poaceae pollen starting from 4000 yr BP (Kendall 1969) and isotopic data at Lake Malawi show an increase in $C_4$ vegetation initiated again around 4000 yr BP (Castañeda et al. 2009). Lowland Amboseli pollen data from Esambu indicate a dry period dominated by Amaranthaceae-Cheonpodiaceae and Cyperaceae. Lake Tanganyika and Lake Kivu diatom and salinity levels show a reduction in water level and increase in salinity as well as reduced diatom and ostracode diversity (Cohen et al. 2005, Haberyan & Hecky 1987). The increase and then decrease in charcoal concentration, with the largest peak around 2000 cal yr BP is similar to charcoal peaks observed in the Rukiga highland (Haberyan & Hecky 1987) and Eastern Arc Mountains (Finch et al. 2016). This could be a result of increased deforestation through burning after which the charcoal concentrations go down as there is little fuel in the environment to sustain previous burning levels.

Late Holocene vegetation is associated with increased warming as well as increased clearance of upper montane and montane forest taxa within the Eastern Arc Mountains (Mumbi et al. 2008). There is an unreversed decline in forest tree pollen and a concomitant increase in the pollen from plants locally associated at present with disturbed and degraded soils from Nyabuiyabuyi, Ahakagyezi and Muchoya within the last ~3000 cal yr BP. East African pollen data point to deforestation during the last two millennia (Taylor 1990, Van Zinderen Bakker & Coetzee. J.A. 1988). Pollen data from the Esambu swamp after ~2000 cal yr BP records an increase in pollen composition and abundance compared to the previous ~3000 cal yr BP. The pollen was dominated by Poaceae and *Typha* and increased abundance of wind-blown Afromontane forest taxa (*Celtis, Juniperus* and *Olea*) derived from the slopes of Mount Kilimanjaro and Chyulu Hills and local arboreal taxa (*Acacia, Commiphora* and *Salvadora*, they indicate a period of increased moisture concomitant with increased local biomass burning.

The four Amboseli records reflect the spatial heterogeneity within a landscape, the macrocharcoal records differences in the drivers of fires between the sites. The low charcoal counts occur at periods of low vegetation abundance during semi-arid phases, this also occurs at levels with high spore abundances which indicate high herbivore densities. This increased moisture signal was also recorded at Lake Masoko (Tanzania); here the pollen record showed the continued presence of wetter Zambezian woodland, albeit, in association with abundant grasses, until 1650 cal yr BP when it was gradually replaced more open vegetation cover (Vincens et al. 2003).

In the Rukiga Highlands of Uganda, forests began to be significantly reduced around 2200 yr BP (Taylor 1990). This vegetation change may be the result of human disturbance since other sites in the region have shown intense human presence around this time. Selective declines over the past 2000 years of favoured timber species, such as *Caesaria, Celtis, Podocarpus, Prunus* and *Psychotria*, in the Eastern Arc Mountains of Tanzania represents a strong indicator of human impact (Finch et al. 2009).

Definite, lasting human impact is picked up in the records much later, approximately 300 years ago with the introduction of typical crop pollen such as *Zea mays* (Lamb et al. 2003, Lejju et al. 2005). During the last few millennia, there is evidence of East African hydroclimatic variability at inter-annual to centennial time scales. Lake records from the region suggest that an intense, century-scale, arid event around ~2000 cal yr BP may have represented the driest conditions of the mid-to-late Holocene and occurred in both eastern and western portions of the East African Plateau (Verschuren & Charman 2008).

There is little direct evidence of human-environment interactions and early inhabitants were diverse hunter-gatherer-fishers, ranging from highly mobile communities to sedentary forest peoples (Kusimba & Kusimba 2005). Certain human land-use practices may have had larger scale implications for environmental change through hunter gathering practices and later once herds of animals were introduced to the region. It is thought that hunter-gatherers could have impacted the environment through deforestation and the use of wood for fire, especially after pottery was introduced into the hunter-gatherer material assemblage with the Kansyore pottery tradition (Kusimba & Kusimba 2005).

Expanding impacts of settlements and industries were observed at both sites; burning of large swathes of the forest for cultivation in MFC, grazing cattle within the swamps in both Mau and Amboseli, human-wildlife conflicts leading to injuries and death of both livestock and wildlife. The activities observed at both MFC and Amboseli impact on the vegetation, soils, water quality and the landscape at large. Human activities have altered several elemental global cycles.

Magnetic susceptibility combined with chemical analysis have been successfully utilised to explore the relationships between depositional conditions, sedimentation rates, sedimentation sources and as proxies for extreme weather events and human activities such as pollution (Arnaud et al. 2014, Singh et al. 2014). Cu, Zn, Fe, Ni, Cd, Cr, and Pb, are heavy metals that are stored in sediment and increase with increase in sedimentation (Daigneault et al. 2012). There is an increase in all the elements detected within the top 20 cm of the sediment core with the greatest increase in Pb, Cu, Ar and Fe, this is accompanied by an increase in the organic matter and sand content suggesting increased sedimentation and contamination. Pb and Cu have been identified as signals of pollution from mining and industrialisation leading to combustion of coal and gasoline (Arnaud et al. 2014, Kathiresan & Bingham 2001, Schillereff et al. 2014).

### 5.2.3 Wider implications and applications of palaeoecological studies.

The new records from Mau Forest Complex and the Amboseli landscape highlight the importance of new palaeoecological records. Under current rapid environmental change, human and wildlife populations experience numerous challenges. The MFC record of forest development since the Late Glacial record and the mid Holocene Amboseli records provide evidence of climatic change, ecological succession and anthropogenic impacts. Collaboration with archaeology to understand significant land use and settlement patterns has already been identified as an important area of study.

During this study, an archaeological survey was also being carried out within Amboseli to try and resolve the spatial issues that arise from combining palaeoecological and archaeological evidence from different catchments. However limitations of sediment for analysis and optimum archaeological sites were not available. Historical records have proven more useful wherever they are available . Records of migration, deforestation, game hunting and trade have played an important role in explaining changes in palaeoecological data for example selective logging accompanied by a reduction in certain tree taxa. Further extensive and intensive archaeological and historical research is needed to fill in the gaps about the movement of people and livestock and the decisions influencing their decision making.

Long-term studies face several challenges in their attempt to develop environmental histories. Disentangling the role of the different drivers in ecosystem change due to their long-term interactions causes complexities. Ecosystems where human impact is known to be present but there is no archaeological or historical evidence of their activities on the landscape causes a level of uncertainty during data interpretation. Distinguishing between extrinsic and intrinsic forcing is crucial for understanding causes of regime shifts and their chances of occurrence when interpreting whether shifts are driven by an abrupt external driver or internal processes.

A clear understanding of long-term change and interactions between society and the environment which can be gained from palaeorecords and the drivers of change can inform understanding future scenarios by simulating change under different climate or human conditions. This is now being attempted where species distribution under different climate conditions are now being simulated. Global efforts such as the Intergovernmental Panel for Climate Change (IPCC), the Millennium Ecosystem Assessment (MEA) compile information on trends, trajectories, frequencies of causes of change as well as future predictions of change.

This use of long-term empirical information is now possible for development of sustainable and adaptive land management policies and the more palaeorecords are developed the better the decision making process due to increased certainty. Given the rapidity of land cover change, there is increasing need for long-term records that can contextualise processes governing long-term ecosystem dynamics. In ecosystems known to have little or no human activity, palaeoecology provides critical information about the ecosystems response to climate conditions and disturbances. These conditions can then be taken into consideration when deliberating changes in the ecosystem especially if the new new land use plans include opening up the ecosystem to human activity. In ecosystems where human beings have been part of the landscape for a very long time, new management policies would have to provide a balance between preserving cultural landscapes and the ecosystem.

The rapid decline in biodiversity and habitat loss leading to a loss of essential ecosystem services has prompted the rise of conservation biology as a discipline to understand causes of this change. Study of long-term ecosystem change can provide critical information on vegetation and fire responses to human activity and climate change identifying conservation targets and developing sustainable and adaptive management plans.

Identifying the degree of past landscape changes can help determine the appropriate degree of conservation effort. Ecosystems developed differently and are altered differently by external factors and thus different localised conservation strategies are needed. The palaeo techniques can identify drivers leading to current ecosystem states and asses the effectiveness of conservation techniques.

Responses of ecological communities to past climate change can be used to anticipate future ecological community changes to climate. Palaeorecords can be used in conservation and restoration efforts when restoring a certain ecosystem to certain state is considered the best option. By identifying baselines, a conservation plan can modulate the conditions needed for an ecosystem to return to a certain state.

### 5.2.4 Conclusions

The main objective of this study was to produce new datasets that would improve our understanding of the Mau Forest Complex and the Amboseli basin; the ecosystem changes that occurred since the Late Pleistocene at Nyabuiyabuyi and the mid Holocene at Amboseli and combine the new information with existing palaeoecological, archaeological and historical data to improve our understanding of East African ecosystems and highlight the need and application for such records.

A multi-proxy approach to understanding palaeoecological systems is essential for providing multiple types of evidence of ecological change for current and future management of landscapes. The information obtained can tell us to what degree past climate regimes and environmental conditions played a role in the development and spread of current societies and how these developments are superimposed on natural ecosystem processes.

This multi-proxy reconstruction of past change and impacts offers the necessary frame to evaluate the changes through time (Gelorini & Verschuren 2013). Dealing with archaeological and palaeoenvironmental data brings up issues of scale where seasonal and annual community responses have to be resolved with decadal to centennial palaeoenvironmental data. East African landscapes have changed rapidly and in complex ways since the Late Pleistocene period. Hydroclimatic variability has been a major driver of ecosystem change and had significant implications for human land use patterns. Humans have a very ancient entanglement with East African landscapes and have influenced biotic and abiotic processes to varying degrees throughout the Quaternary.

Recent impacts of human activity have intensified and anthropogenically induced landscape changes are happening at unprecedented amplitudes, scales, and in novel ways that stress biodiversity from the genetic to population scales. Descriptive studies of palaeoenvironments are useful for understanding past patterns of environmental variability; but, can often be difficult to robustly identify the mechanisms underpinning environmental changes. This highlights the need for precisely-dated, high-resolution, multi-proxy, and multi-site studies of social and environmental change in studies focusing on the recent past and present, as well as longer-term studies focusing on the evolution and variability these interactions. With these types of data a more informed approach can be fostered for policy, land management, and conservations decisions.

Further research aimed toward disentangling the relative contributions of anthropogenic activity from natural earth system processes will allow for a more nuanced understanding of landscape-scale processes and lead to important constraints for realistic future projections of change. This type of fundamental research can inform development and sustainable land-use policy and appropriate plans for multi-stakeholder conservation efforts. Stakeholders from different disciplines (modellers, land use managers, policy makers, farmers) can begin to appreciate long-term information once presented in an easily understandable manner. Understanding how the landscape changes and respond to different uses can be inform scenario building.

The connections between climate and culture remain largely unclear and require case-by-case study with high-temporal resolution and precise chronologies. Understanding the human dimension within global change science is of immense societal, political and economic relevance. The long-term relationships between environment and human activity, both past and present, must be taken into account to adequately understand the evolution and future of landscapes. Critical information to do this type of conceptualisation and modelling of interactions between environments and cultures will emerge from site-specific studies that combine archaeology and palaeoecology in situ.

Remote sensing and satellite-based earth surface products offer very detailed land cover information over the latter half century and can complement integrated archaeological, anthropological, demographic, historical, and palaeoecological studies. For example, a satellite-image based land cover change analysis was combined with sedimentary charcoal records collected from small lakes in central Africa and used to interpret the impacts of human population increases and forest clearance over the past $\sim$100 years (Aleman et al. 2013). This led to much more insight on tropical fire regimes, including fire frequency, fire extents, and position on the forest-grassland-agricultural mosaicked landscape. The increasing accessibility of remotely sensed products, such as SPOT imagery and historical air photographs of East Africa, offer great potential for assessing recent land cover change and land use changes, spatial analyses of archaeological sites, and integration of multidisciplinary research.

Together, all of these lines of evidence will provide collaborating evidence about human-environment interaction processes and can be used to constrain quantified projections and scenarios for future landscape changes (Platts et al. 2012). Modeling approaches combining climate models with dynamic vegetation models can be used to examine projected vegetation cover changes and can be combined with models of human population pressure to examine potential degrees of recovery or degradation regionally. Other landscape-scale models could be used to understand future corridors of animal migration and distribution changes, erosion and sedimentation rates, hydrological changes, and policy change impacts. A combined approach between empirical, qualitative, and modeling approaches is best suited to understand the complexity of climate-human-ecosystems interactions (Gelorini et al. 2012) and to provide information to inform policy makers, land managers and conservationists at relevant spatial and temporal scales for appropriate developments. This calls for teams of researchers across multiple disciplines and multiple non-academic partners at all stages of the scientific inquiry and communication process.

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