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Vegetational Change and Land Degradation in the Lake Baringo Area, Kenya, During the Late Holocene: Evidence from the Paleorecord and Remote Sensing

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VEGETATIONAL CHANGE AND LAND DEGRADATION IN THE LAKE
BARINGO AREA, KENYA, DURING THE LATE HOLOCENE: EVIDENCE FROM
THE PALEORECORD AND REMOTE SENSING

A Dissertation
Submitted to the Graduate Faculty of the
Louisiana State University and
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requirements of the degree of
Doctor of Philosophy

in

The Department of Geography and Anthropology

by

Lawrence Morara Kiage
B. Ed., Kenyatta University, 1995
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May, 2007
To Precious for the spirit she displayed
ACKNOWLEDGEMENTS

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I thank my colleague Jason T. Knowles for agreeing to temporarily suspend reason and engage the crocodiles and hippopotamus of Lake Baringo in coring exercises. I also thank my brother James N. Kiage for joining Jason and myself for fieldwork. John Omonywa is thanked for driving the research team on the treacherous road to Lake Baringo and around its drainage basin. His driving made us want to finish the exercise sooner. I thank my colleagues at the Global Paleoecology Laboratory; Terry McCloskey, Jennifer Hathorn, Jonathan Breaux, Tom Bianchette, and Yun Huang for making life in the laboratory livable. Special thanks to the entire Earth Scan Laboratory crew, especially Ric Haag and Shreekanth Balasubramanian for teaching me how to track satellites using the X-band antenna and look for an excuse to party.

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ABSTRACT

This dissertation investigates the history of vegetational changes and land degradation in the Lake Baringo area, Kenya, East Africa, during the Late Holocene. It uses $^{14}$C- and $^{210}$Pb-dated fossil pollen, fungal spores, microscopic charcoal, and stratigraphic record from lake-sediment cores to reconstruct the paleoenvironment in the East African region. More recent changes in the Lake Baringo are examined using remote sensing imagery. These data provide one the first high-resolution late Holocene pollen records from the semi-arid region of Kenya (and one of only a handful from East Africa in general).

Lake Baringo records a sedimentation rate of 1 cm yr$^{-1}$, which remained largely unchanged for both the prehistoric period, prior to European settlement in the Baringo ecosystem (AD 1890s), and after, including the present. That raised the possibility that background environmental conditions and not humans are the main drivers of land degradation in the ecosystem. A record from a 363-centimeter core retrieved from Lake Baringo reveals a largely dry environment that is punctuated by a succession of centennial- to decadal-scale wet and dry episodes, separated by rather sharp transitions, including two intense dry episodes at ca. AD 1650 and AD 1750 that led to drying of the lake. The two episodes are reflected by poor pollen preservation environment and by relatively high percentages of dry-indicator species such as Podocarpus, Olea, Euclea, Acacia, Balanitaceae, Gramineae, and Cyperaceae. However, climatic implications of the changes in pollen spectra in the Lake Baringo record are limited by the extremely high sedimentation rate in Lake Baringo that effectively constrains the chronological record of the cores to the last 300 $^{14}$C yr BP. Satellite imagery analyzed in this dissertation reveals
that deforestation and the resultant land degradation have contributed to increased sediment yield in the lake. Consequently, the lake surface area was reduced by over 10% and turbidity increased, which is confirmed by a statistically significant increase ($t = -84.699$, $p < 0.001$) in the albedo between 1986 and 2000. Although climatic variations account for some of the changes in the lake catchment most of the changes in land cover are inherently linked to mounting human and livestock population in the Lake Baringo catchment.
CHAPTER 1. INTRODUCTION

This dissertation is mainly a palynological investigation of paleoenvironmental change in East Africa. It is aimed at studying vegetational change and land degradation in the Lake Baringo area, Kenya, during the Late Holocene by means of the examination of fossil pollen and microscopic charcoal in lake-sediment cores. Remote sensing imageries are used to investigate more recent changes within the Lake Baringo drainage basin. Situated in the broad ecotone between savanna and steppe in East Africa, the Lake Baringo area has a rich biodiversity. It is a critical habitat for more than 500 avifaunal species and a source of livelihood for humans. Unfortunately, the rich biodiversity is threatened by land degradation as a result of mounting population pressure and climate fluctuations in East Africa.

In recent decades Lake Baringo is reported to have experienced drastic increases in sedimentation rates attributed to accelerated soil erosion, land use/land cover changes, and climatic fluctuations (Kurgat 2003; Aloo 2000). A combination of natural and anthropogenic factors, including increased sediment yield from soil erosion, evaporation, damming, and diversion of inflowing rivers has reduced the depth of the lake from a maximum of 9 m and a mean of 5.6 m in 1972 to the current maximum of 4 m and a mean of 2.5 m (Aloo 2002). According to a report by the United Nations Development Program (UNDP), should this alarming rate of land degradation in the Lake Baringo basin go unchecked, the 129 km² lake, an oasis in a semi arid environment, could fill up with sediments in the next 8 to 15 years (Kurgat 2003). This timing is interesting considering that, as part of a wider pattern of tropical climate change, the ice caps on Mount Kilimanjaro in East Africa are projected to disappear by around 2015 (Gasse
In this dissertation high-resolution paleoecological data is used to assess the role of natural versus anthropogenic causes for environmental changes in East Africa during the late Holocene. It also tests the validity of the postulation that Lake Baringo is “a dying lake”.

Palynology, the bedrock of this dissertation, is one of the sharpest tools for the reconstruction of environmental changes due to both natural and anthropogenic causes during geological and historical times (e.g., Peyron et al. 2000; Alin et al. 2002; Salzmann 2002; Darbyshire et al. 2003; Lamb et al. 2003; Behling et al. 2004). Evidences (biotic and abiotic) of past landscapes have been accumulated and archived in lakes, swamps, bogs, and even ice. These evidences provide an excellent opportunity for paleoecological studies through the use of proxy data such as pollen, spores, and charcoal. Pollen in lake sediment cores has been used to reconstruct past changes in vegetation and climate as well as human activities in different parts of the world (e.g., Hamilton 1982; Hamilton et al. 1986; Liu and Lam 1985; Burney 1987, 1993, 1997; McAndrews 1988; Behre 1988; Liu and Colinvaux 1988; Liu 1990; Edwards and MacDonald, 1991; Maley and Brenac 1998; DeMenocal et al. 2000; Gasse 2000; Salzmann 2000, 2002; Peyron et al. 2000; Wooller et al. 2000; Behling 2002; Darbyshire et al. 2003; Lamb et al. 2003, Behling et al. 2004). Though proven extremely useful, very few palynological studies have focused on land degradation and/or desertification, considered one of the most serious environmental problems in Africa (Mortimore 1998; Gonzalez 2001). This dissertation provides one of the first high-resolution Holocene pollen records from the semi-arid region of Kenya (and one of only a handful from East Africa in general). The markers of change (mainly introduced species) are investigated in
the context of changes in pollen composition and/or sedimentation rates in order to help identify the forcing mechanism responsible for the high rates of land degradation/aridification in the ecosystem.

This dissertation tracks the more recent changes in the Lake Baringo area by using remote sensing tools to examine land cover changes. The recent land cover changes are then linked to the increasing sedimentation rates in Lake Baringo. Over the years remote sensing techniques have proven useful for investigating phenomena such as drought, rainfall, deforestation, land degradation, and land cover change in many environments including semi-arid ecosystems (e.g., Kasai et al. 2005; King et al. 2005; Flügel et al. 2003; Servenay and Prat 2003; Wang et al. 2003; Nishida et al 2002; Khan et al. 2001; Anyamba et al. 2001, 2002; Asner and Lobell 2000; French et al. 2000; Gilabert et al. 2000; Metternicht and Zinck 1998; Nicholson 1998; Dwivedi et al. 1997; Tieszen et al. 1997; Mishra et al. 1994; Moran et al. 1994; Price 1993). The fine spatial resolution of Landsat Thematic Mapper (TM)/Enhanced Thematic Mapper plus (ETM+) used in this study is excellent for uncovering the complexity of ecological and social relationships. Remote sensing techniques help identify the hotspots of changes within the Lake Baringo area and link them to human, climate, and/or physical features.

Through a combination of palynological tools and remote sensing this dissertation attempts to achieve the following specific objectives:

- To reconstruct the paleoenvironmental history and land degradation of the Lake Baringo area during the late-Holocene, especially the past several centuries;
- To establish the role of fire in shaping the vegetation of the ecosystem;
• To identify the forcing mechanisms behind the high rates of land degradation/aridification in the ecosystem;

• To identify the markers of change (e.g., introduced species, charcoal peaks) in the core stratigraphy and relate them to changes in pollen composition and/or sedimentation rates;

• To evaluate the postulation that Lake Baringo is “a dying lake”.

Chapter 2 reviews the Late Quaternary paleoenvironmental changes in East Africa based on evidence from pollen, diatoms, microscopic charcoal, lake level records and associated proxies. The paleoenvironmental records discussed are derived from different complementary proxies to provide a more accurate and complete assessment of the paleoenvironmental changes in East Africa. This chapter basically contextualizes the themes that are discussed in the rest of the dissertation. It becomes apparent in Chapter 2 that most of the paleoenvironmental studies in Eastern Africa have focused on paleoclimatic reconstructions over long geologic time spans. There is an existing paucity of high–temporal-resolution studies (centennial-and decadal-scales) for the region, which this dissertation attempts to fill.

The following chapters focus largely on paleoenvironmental changes in East Africa since ca. AD 1650 as revealed in the Lake Baringo record. Chapter 3 provides the lithostratigraphic evidence of paleoenvironmental changes in the Lake Baringo area. Lithostratigraphic changes, especially sedimentation dynamics, offer one of the powerful approaches for investigating the timing and impact of human activities and/or climate change.
Chapter 4 investigates the history of vegetational change and land degradation using evidence from high-temporal-resolution analysis of pollen, fungal spores, and microscopic charcoal. This chapter highlights both the reality of climate variability in East Africa at all time scales and the increasing human impact on the environment. The vegetational changes in the Lake Baringo area presented in Chapter 4 highlights the challenge of separating purely background-level environmental and anthropogenically driven climatic change. Therefore, by employing a multiproxy approach this study aids in separating the natural and anthropogenically driven forcing mechanisms.

Chapter 5 addresses the issue of more recent environmental changes in the Lake Baringo area using remote sensing techniques. Specifically, satellite measurements are used to study vegetational change associated with deforestation and land degradation in the Lake Baringo area. Changes in land use/cover in the Lake Baringo drainage basin are ultimately linked to the sedimentation rates observed in the lake. Chapter 6 of this dissertation provides summaries of the major findings of this study and directions for future research.

1.1 Expected Significance

The Lake Baringo area constitutes an important habitat for many endemic avifaunal and aquatic species, many of which are endangered as a result of habitat changes. In addition, social stability and human livelihoods are in imminent danger from interruptions of their economic and resource bases. It is indeed timely, if not long overdue, for an in-depth study of vegetation change and land degradation in this critical environmental zone of East Africa.
This study provides one of the first high-resolution Holocene pollen records from the semi-arid region of Kenya and one of only a handful from East Africa in general. The contribution of this study to understanding the role of natural versus anthropogenic causes for environmental changes in East Africa during the late Holocene cannot be overstated. Its findings have significant policy implications on the conservation of biodiversity, lake restoration, land-use, and environmental management in the semi-arid regions of East Africa. In addition, this study offers a somewhat novel approach to historical paleoenvironmental reconstruction by combining multidisciplinary methodologies including remote sensing with palynological methodologies in resolving enigmatic environmental problems in tropical Africa. The study links long and short-term data sets in addressing a question of significant societal relevance – the impacts of human activities on geophysical processes and environmental change.

1.2 References


Gasse, F. 2000: Hydrological changes in the African tropics since the Last Glacial Maximum. *Quaternary Science Reviews* 19, 189-211.


CHAPTER 2. REVIEW OF LATE QUATERNARY PALEOENVIRONMENTAL CHANGES IN EAST AFRICA*

2.1 Introduction

Organic and inorganic sediments, archived naturally over millennia in environments such as lakes, swamps, bogs, and icecaps, provide an excellent source of data for paleoenvironmental reconstruction. In particular, pollen records from lake sediments can provide detailed records of past paleoecological and climatic changes (e.g., Liu and Colinaux, 1988; Maley and Brenac, 1998; deMenocal et al., 2000; Gasse, 2000; Salzmann, 2000, 2002; Peyron et al., 2000; Wooller et al., 2000; Behling, 2002; Darbyshire et al., 2003; Lamb et al., 2003, Behling et al., 2004). In addition to pollen, other proxies such as charcoal (Burney, 1987; Long et al., 1998; Whitlock and Millspaugh, 1996), phytoliths (Piperno 1988; Mworia-Maitima 1997; Piperno and Pearsall 1998), diatoms (Gasse et al., 1997; Barker et al., 2001), and grass cuticles (Mworia-Maitima 1997; Wooller et al., 2000) have proven to be very useful in paleoenvironmental reconstruction.

The use of multi-proxy approaches enables paleoecologists to more accurately and completely reconstruct the paleoenvironment (Boyd and Hall, 1998). Since the environment affects different proxies differently, the deficiencies in one proxy may be compensated for by another. For instance, prolonged droughts might compromise the preservation of biological proxies such as pollen due to the lowering of lake levels, yet paleoenvironmental conditions that prevailed during periods of low lake levels can still be recorded through geochemical and/or isotopic proxies such as total organic content.

*Note: This chapter has been previously published in Progress in Physical Geography. The citation is as follows: Kiage, L. M., and Liu, K-b. 2006: Late Quaternary Paleoenvironmental Changes in East Africa: Multi-Proxy Evidence from Palynology, Lake Sediments, and Associated Records. Progress in Physical Geography 30 (5), 633-658.
(TOC), hydrogen index (HI), $\delta^{18}$O, $\delta^{13}$C, and $\delta^{15}$N, among others.

We review different proxy evidences of climatic changes in East Africa from beyond 42,000 yr BP to the present millennium. Although we focus on the three East African countries of Kenya, Uganda, and Tanzania (Figure 2.1, Table 2.1), we include relevant information from tropical Africa as a whole. The region has considerable topographic relief, with stepped plateaus of 900-2000 m or more and high mountains such as Mounts Kenya, Elgon, Kilimanjaro, and the Ruwenzoris. We examine biological proxy records such as pollen, diatoms, and microscopic charcoal, as well as paleohydrological evidences for lake level fluctuations. Where appropriate, these records are complemented by chemical and isotopic proxies. Although we try to cover the entire breadth of the region we are limited by data gaps, especially from areas lacking suitable archiving environments. Ages referred to in the review are given in $^{14}$C years before present (BP), unless otherwise noted. Our review differs from that of Peyron et al. (2000), Marchant and Hooghiemstra (2004), or Battarbee et al. (2004) which, though excellent, are either too focused on a particular time period or too broad in regional coverage to give the complete picture of paleoenvironmental changes in East Africa through the period reviewed herein.

2.2 Modern Climate and Vegetation Patterns

The climatic patterns in East Africa are very complex and change rapidly over short distances. This is attributed to large-scale climatic controls that include major convergence zones which are superimposed upon regional factors associated with lakes, topography, and maritime influences (Nicholson 2000). Generally, East Africa has a semi-arid to sub-humid climate. However, the highlands and areas close to large lakes
Figure 2.1. Map showing East Africa and the neighboring countries. The approximate locations of most of the sites discussed in the review are also shown in the dots numbered 1 to 21, and cited in Table 2.1.
Table 2.1. Sites discussed in the text and plotted in Figure 1, with references to articles original data and interpretations.

<table>
<thead>
<tr>
<th>Site</th>
<th>Location</th>
<th>Elevation (m)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Lake Albert, Uganda</td>
<td>619</td>
<td>Beuning et al. 1997</td>
</tr>
<tr>
<td>3</td>
<td>Muchooya Swamp, Uganda</td>
<td>2256</td>
<td>Hamilton (1982), Taylor (1990)</td>
</tr>
<tr>
<td>4</td>
<td>Lake Mahoma, (Ruwenzori Mountains), Uganda</td>
<td>2960</td>
<td>Livinstone (1967)</td>
</tr>
<tr>
<td>5</td>
<td>Kamiranzovu, Rwanda</td>
<td>1950</td>
<td>Roche and Bikwemu (1989)</td>
</tr>
<tr>
<td>6</td>
<td>Kashiru, Burundi</td>
<td>2240</td>
<td>Bonnefille and Riollet (1988)</td>
</tr>
<tr>
<td>7</td>
<td>Lake Tanganyika, Tanzania</td>
<td>773</td>
<td>Vincens (1989; 1993)</td>
</tr>
<tr>
<td>8</td>
<td>Lake Rukwa, Tanzania</td>
<td>793</td>
<td>Butzer et al. (1972); Haberyan (1987), Talbot and Livingstone (1989)</td>
</tr>
<tr>
<td>9</td>
<td>Lake Masoko, Tanzania</td>
<td>770</td>
<td>Vincens et al. (2003)</td>
</tr>
<tr>
<td>10a</td>
<td>Mount Kilimanjaro, Tanzania</td>
<td>2650</td>
<td>Coetzee (1967)</td>
</tr>
<tr>
<td>10b</td>
<td>Mount Kilimanjaro, Tanzania</td>
<td>5893</td>
<td>Thompson et al (2002)</td>
</tr>
<tr>
<td>12</td>
<td>Pekington Bay, Lake Victoria, Uganda</td>
<td>1134</td>
<td>Talbot and Livingstone (1989); Stager et al. (2003)</td>
</tr>
<tr>
<td>14a</td>
<td>Sacred Lake, Mount Kenya, Kenya</td>
<td>2400</td>
<td>Coetzee (1967); Olago (1995); Street-Perrott and Perrott (1993); Street-Perrott et al. (1997)</td>
</tr>
<tr>
<td>14b</td>
<td>Simba Turn and Small Hill Turn (Mount Kenya, Kenya)</td>
<td>4595</td>
<td>Barker et al. (2001)</td>
</tr>
<tr>
<td>14c</td>
<td>Simba Turn and Small Hill Turn (Mount Kenya, Kenya)</td>
<td>4289</td>
<td>Barker et al. (2001)</td>
</tr>
<tr>
<td>15</td>
<td>Lake Bogoria, Kenya</td>
<td>990</td>
<td>Tiecelin et al. (1987)</td>
</tr>
<tr>
<td>16</td>
<td>Loboi Swamp, Kenya</td>
<td>1080</td>
<td>Ashley et al. (in press).</td>
</tr>
<tr>
<td>17</td>
<td>Cherangani Hills, Kenya</td>
<td>2900</td>
<td>Bakker (1964)</td>
</tr>
<tr>
<td>18</td>
<td>Lake Kimilili, Mount Elgon, Kenya</td>
<td>4150</td>
<td>Hamilton and Perrott (1978; 1979)</td>
</tr>
<tr>
<td>19</td>
<td>Laboot Swamp, Kenya</td>
<td>2880</td>
<td>Hamilton (1982)</td>
</tr>
<tr>
<td>20a, b</td>
<td>Lake Turkana, Kenya</td>
<td>360</td>
<td>Bonnefille (1976), Mohammed et al (1995)</td>
</tr>
</tbody>
</table>
tend to be wetter (Hamilton 1982; McGregor and Nieuwolt 1998; Nicholson 2000). The mean annual rainfall ranges from well over 1200 mm in Uganda and the highlands to less than 200 mm in the arid and semi-arid parts of Kenya (Figure 2.2) (c.f. Nicholson 2000). The climate of the region is governed by air streams and convergence zones that are linked to the seasonal shifts of the subtropical high pressure cells situated about 20°-30° north and south of the equator (McGregor and Nieuwolt 1998; Nicholson 1996; 2000). The air streams include the Congo air with westerly and southwesterly flow, the northeast monsoon (N.E. trade winds), and the southeast monsoon (S.E. trade winds) (Figure 2.3). The air streams are separated by two surface convergence zones, that is, the Inter-tropical Convergence Zone (ITCZ) and the Congo Air Boundary (CAB). The former separates the two monsoons, the latter, the easterlies and westerlies (Nicholson 2000).

Synoptically, the climate of East Africa is dominated by the interplay between the northeast monsoon in January and the southeast monsoon in July. Unlike the S.W. monsoon of Asia, the monsoons that affect East Africa are thermally stable, hence mostly dry. The northeast monsoon, coming from the Arabian Desert, is a dry wind. The southeast monsoon, coming off the Indian Ocean, constitutes the dominant air stream over East Africa in July. Although deep and moist, it is not associated with much rainfall because the winds are divergent (McGregor 1998). However, the Congo air, which originates in the south Atlantic, is humid, convergent, and thermally unstable. It is associated with rainfall in the northwestern part of the region (Boucher 1975; Hamilton 1982; McGregor and Nieuwolt 1998; Nicholson 2000).
Figure 2.2. The spatial distribution of rainfall in East Africa (Modified from Hamilton 1982).
Figure 2.3. Schematic of the general patterns of winds, pressure and convergence over Africa during the months of January and July. The dotted lines indicate the intertropical convergence zone (ITCZ), dashed lines, the Congo Air Boundary (CAB) (Modified from Nicholson 2000).
The vegetational distribution pattern in East Africa (Figure 2.4) closely mimics the rainfall and topographic conditions as well as the length of the dry season. Six major floristic regions can be delineated in East Africa (Lind and Morrison 1974; White 1983) (Figure 2.4): (1) Lowland and montane forests occur in patches in Uganda and Kenya. These semi-deciduous and evergreen forests are made up mainly of *Celtis* spp., Urticaceae, Myrtaceae, *Croton*, *Holoptelea*, *Prunus*, *Podocarpus*, *Ilex*, *Acacia*, and *Olea* among others. The Afroalpine belt occurs between 1600 and 3000 m and is characterized by *Prunus africana*, *Olea*, *Hagenia abyssinica*, *Juniperus procera*, and *Podocarpus*. The Ericaceous belt dominated by *Philipia* and *Erica* are found above 3000 m. (2) The drier savanna types (deciduous bushland and thicket), mainly consisting of *Acacia* and *Commiphora* are common in Kenya, Ethiopia and Somalia. (3) The grasslands, which often occurs in the same ecological regions with deciduous bushland are not widespread in East Africa. (4) The wetter savanna types, mainly made of Combretaceae and Acacia savannas, are widely distributed in Uganda, Tanzania, and Kenya (Figure 2.4). (5) In the southern parts of Tanzania the wetter Zambezian deciduous woodlands (mainly miombo) occur. These deciduous woodlands are basically a moist savanna type that is dominated by fairly tall (20 m) and densely-spaced leguminous trees belonging to such genera as *Brachystegia*, *Julbernardia*, *Isoberrlinia*, and *Uapaca*. (6) The semi-desert and desert vegetation occurs in northern Kenya and southern Ethiopia.

### 2.3 Pollen Records from East Africa

Investigations of East Africa’s paleoclimate and paleoecology commenced in the 1960s largely due to pioneering studies by Livingstone (1962; 1967), Bakker (1964) Coetzee (1964; 1967), Morrison (1968), and Kendall (1969), and have continued to the
Figure 2.4. The spatial distribution of the major vegetation types of East Africa (Modified from Lind and Morrison 1974).
present. Pollen analysis from the early studies provided the first evidence to reveal that
the highlands of equatorial Africa, like other parts of the world, were affected by
glaciation and that glacier retreat was roughly contemporaneous with those in Europe and
North America (Hamilton 1982). The longest continuous records retrieved from East
Africa are from Sacred Lake, Mount Kenya (Coetzee 1967; Olago 2001; Street-Perrott et
al., 1997) covering the period from 115,000 yr BP to the present. A record from two
swamps (Muchoya and Ahakagyezi) in the Rukiga Highlands, Uganda (Taylor 1990) has
a basal date of > 42,000 yr BP while those from Kashiru, Burundi (Bonnefille and Riollet
1988) and Cheshi, Zambia (Stager 1988) provide basal dates of ca. 40,000 yr BP.

The pollen records from the region have been drawn from diverse environments
(Figure 1, Table 1) including small lakes on mountains and mountain ranges e.g., Mount
Kilimanjaro (Coetzee 1967), Mt. Kenya (Coetzee 1967; Olago 2001; Street-Perrott et al.,
1997), Ruwenzori (Livingstone 1967), Cherangani (Bakker 1964), and Mount Elgon
(Hamilton and Perrott 1978; 1979), large lakes, e.g., Lake Victoria (Kendall 1969; Stager
1984; Stager et al., 1997), Lake Albert (Beuning et al., 1997), Lake Naivasha (Maitima
1991), Lake Turkana (Bonnefille 1979; Mohammed et al. 1995), Lake Tanganyika
(Vincens 1989; 1993), and Lake Rukwa (Butzer et al., 1972), as well as mires, e.g.,
Rukiga Swamp (Taylor 1990), Ahakagyezi Swamp (Hamilton et al., 1986; Taylor 1990),
and Laboot Swamp (Hamilton 1982), among others. However, most of the pollen records
only extend back to 30,000 $^{14}$C yr BP (Figure 2.5). Longer records, with a few
exceptions, tend to be characterized by hiatuses (Vincens 1991; Bonnefille et al., 1995;
The pollen records from this vast region reveal a remarkably high diversity of both
paleoclimate and paleovegetation types even over short distances, probably due to the presence of many mountainous areas (c.f., Peyron 2000). Pollen analysis from the different sites has illuminated the paleoenvironmental dynamics of the region beyond the LGM.

2.4 The paleoenvironment of East Africa

2.4.1 42,000 to 30,000 \(^{14}\text{C}\) yr BP

The period prior to ca. 42,000 \(^{14}\text{C}\) yr BP is characterized by warm climatic conditions similar to those that presently abound in East Africa. During this period the vegetation in Ahakagyezi and Muchoya Swamps, Uganda, was dominated by moist lower montane forest, where Urticaceae, Myrtaceae, Croton, Minulopsis, Ilex, Macaranga, Neoboutonia, Nuxia and Filcahoa occurred in the Ahakagyezi and Muchoya catchment at the same elevations as today (Figure 2.6) (Taylor 1990). Contemporaneous records from Sacred Lake, Mount Kenya (Coetzee 1967; Olago 2001) and Kamiranzovu Swamp (Hamilton 1982) confirm similar trends. Between 42,000 \(^{14}\text{C}\) yr BP and 30,000 \(^{14}\text{C}\) yr BP cold dry conditions engulfed East Africa. Consequently high altitude vegetation belts descended to lower elevations. Evidence from charred grass cuticles from Sacred Lake, Mount Kenya, shows a general increase in the percentages of Bambusoid/Arundinoid-type grasses as well as a Panicoid (Andropogonid)-dominated assemblage usually associated with mesophytic conditions. The source of the fire that left the charred particles at the Sacred Lake site is likely to have been natural considering that humans had not established themselves in the lake catchment (Wooller et al., 2000).

In the Rukiga Highlands the vegetation that constitutes the Ericaceous belt (e.g., Anthospermum, Stoebe, Artemisia and Cliffortia) typically at ca. 3300 m today was found
Figure 2.5. A generalized chronology of inferred environmental changes in East Africa for selected sites discussed in the text. The darker shades represent conditions that were colder or wetter than present conditions.
Figure 2.6  Fluctuations in % pollen and spores with depth in the Muchoya Swamp core MC2 (Modified from Taylor 1990).
in Muchoya Swamp at an elevation of 2260 m above sea level (Figure 2.6) (Taylor 1990). The present conditions in the Muchoya catchment suggests that it would have required a much colder and drier environment than present, equivalent to an altitudinal depression of up to 1000 m, for the Ericaceous belt to thrive. Similar observations are made from Mount Kenya (Olago 2001; Street-Perrott 1993), Lake Abiyata (Ethiopia) (Lezine 1982), and Kashiru Swamp (Burundi) (Bonnefille and Riollet 1988). In Kashiru Swamp there was a decline in forest taxa and an increase in Gramineae (Bonnefille and Riollet 1988; Aucour et al., 1994), also suggesting cold, dry conditions. The Mount Kenya record during this period shows *Hagenia* dominated vegetation coexisting with the Ericaceous belt components.

### 2.4.2 30,000 to 21,000 \(^{14}\)C yr BP

Cool and moist conditions prevailed in East Africa in the period leading up to the Last Glacial Maximum (LGM) period of the northern hemisphere (ca. 30,000 \(^{14}\)C yr BP to 21,000 \(^{14}\)C yr BP). There appears to have been a slight increase in temperature and precipitation given that upper montane forest replaced the ericaceous scrub in the Muchoya catchment (Figure 2.6) (Taylor 1990). Other montane sites in the region, including the Abadares, Kenya (Perrott and Street-Perrott 1982), Mount Kenya (Coetzee 1967; Olago 2001), and Cherangani Hills (Coetzee 1967), are characterized by more abundant arboreal pollen taxa. Charred grass cuticles from Mount Kenya show fire- and mesophytic related taxa with high percentages of *Phalaris*-type grasses and Pooid grass types that usually correspond to cool and moist conditions (Wooller et al., 2000). The cool wet conditions are consistent with those of West Africa inferred for the same period (Maley et al., 1990; Brenac 1988; Salzmann et al., 2002). However, the increase in
precipitation is not recorded at all sites in East Africa. For instance, Lake Albert was surrounded at this time by an open wooded grassland assemblage (Gramineae over 40%) (Beuning et al., 1997), suggesting a dry climate. The absence of semi-deciduous forest taxa, the abundance of Combretaceae, Chenopodium-type, and a general increase in herbaceous pollen types during that period further affirm the dry climate scenario. The temperatures were between 2 and 4.1°C colder than present (Coetzee 1967; Olago 2001).

2.4.3 21,000 to 12,500 \(^{14}\)C yr BP

Between ca. 21,000 and 12,500 yr BP East Africa’s environment was characterized by cool conditions, punctuated by episodes of prolonged desiccation (c.f. Kendall 1969; Street and Groove 1979; Beuning et al., 1997). During this period most sites are characterized by poor pollen preservation, probably due to low lake levels and subsequent oxidation in these environments (although in some sites pollen was well preserved). In the Rukiga highlands the Ericaceous belt became reestablished in the Muchoya Swamp catchment until ca. 15,700 yr BP, which is coeval with a substantial reduction in temperature and precipitation in the region (Taylor 1990). Lake level records from Lake Victoria and Lake Albert identify at least two separate episodes of extreme aridification during this period (Talbot and Livingstone 1989; Beuning et al., 1997). Two cores (Ibis 1 and Ibis 3) obtained from Lake Victoria at different depths, 32 m and 66 m respectively, reveal two distinct discontinuities that correspond to low lake levels (Talbot and Livingstone 1989). The physical properties of Core Ibis 1 mark the first discontinuity at 8.2 m (~11,710 +/-120 \(^{14}\)C yr BP) and the second at 8.85 m (~15,390 +/-220 \(^{14}\)C yr BP) (Figure 2.7). The two episodes are concurrent with prominent layers of the littoral gastropod Bellamya unicolor, which is consistent with desiccation and
Figure 2.7 Summary of Ibis core 1 from Lake Victoria, showing changes in the stratigraphy especially sections indicative of significant desiccation periods (Modified from Talbot and Livingstone 1989).
oxidation of sediments during low lake levels (c.f., Stager et al., 1986; Talbot and Livingstone, 1989). Traces of desiccation cracks and rootlets or burrows extend ca. 25 cm down into the mottled sediments below the color change (Figure 2.7) suggesting exposure at these levels (c.f., Stager et al., 1986).

The Lake Albert record is based on a 9.2 m core (core F) collected from 46 m below water level yielding a record for the last 30,000 $^{14}$C yr BP (Beuning et al., 1997). The core consists of predominantly dark grey to black, diatomaceous mud with a thin sandy zone at 6.6 m overlying drier, silty sediments to the base of the core (9.2 m) (see details in Beuning et al., 1997). The key to the sedimentation history of the core site is provided by the sequence between 8.4 and 6.6 m where low water contents and presence of rootlet traces suggest terrestrial conditions. The sequence is associated with calcite nodules crumb structure, and vertical and oblique, clay lined cracks are typical pedogenic features of terrestrial environments (Beuning et al., 1997). Such features are associated with cotton clays, a soil type that develops on floodplains and lake margins in warm, seasonally contrasted climates (c.f., Blodgett 1985).

Further scrutiny by Beuning et al. (1997) revealed that there were, in fact, two such levels in the Lake Albert stratigraphy (similar to core Ibis 1, Talbot and Livingstone 1989), one at 7.25 to 7.9 m and the other at 6.6 to 6.9 m (Figure 2.8) separated by a zone containing diatoms, Botryococcus, ostracods, fish fragments and rare pollen grains indicating more persistent aquatic conditions (Beuning et al., 1997). Therefore, the low stands must have been separated by at least one period of lake transgression.

The desiccation in East Africa must have been so severe that it led to a lowstand of at least 46 m below present levels at the core site in Lake Albert. The timing of this
Figure 2.8  Geochemical data from the core (Core F) from Lake Albert. TOC=total organic content (wt%), HI= Hydrogen Index. All determinations on bulk organic matter (from Beuning et al., 1997).
period of maximum aridity in Lake Albert is consistent with dated records of prolonged minimal lake levels throughout East Africa and most of the African continent from ca. 20,000 to 12,500 $^{14}$C yr B.P (e.g., Livingstone 1967, 1980; Kendall 1969; Talbot and Livingstone 1989; Street-Perrott and Robert 1983; Taylor 1990). Rising water in Lake Albert at 12,500 $^{14}$C yr BP was coincident with severe flooding along the main Nile (Said 1993). The Lake Albert record show that deep-water conditions were restored soon after 12,500 $^{14}$C yr B.P although there is an apparent change in sedimentation rate between 8,000 and 3,400 $^{14}$C yr B.P. During the latter period there was an interruption of sedimentation at the coring site that could be another, briefer period of low water, due to dry conditions in the catchment (Beuning et al., 1997). This dry episode may well be part of the mid Holocene low precipitation episode that is recorded throughout the tropics at ca. 4,000 yr $^{14}$C yr B.P.

The pollen record from Lake Naivasha pollen shows a brief moist period between 17,000 and 15,000 $^{14}$C yr BP characterized by somewhat increased percentages of forest taxa, *Podocarpus, Olea, Celtis,* and *Hagenia* (Maitima 1991). However, the Lake Naivasha record, though supported by diatom and ostracod data that suggest high lake levels, are inconsistent with all the other records from the region, which show arid conditions during this period (Bakker 1964; Coetzee 1967; Hamilton 1982). The anomalous Naivasha record is probably a result of dating problems. There was a slight amelioration of climate toward more humid conditions after 12,500 yr BP as evident by an expansion of montane forest in the Cherangani Hills and also on Mount Kenya (Coetzee 1967, 1987; Hamilton 1982). Evidence from Sacred Lake, Mount Kenya, suggests that the maximum aridity phase occurred just before 13,500 $^{14}$C yr BP.
During this period the charred grass cuticle record from Mount Kenya is characterized by a stable abundance of Bambusoid/Arundinoid-type grasses reflecting the consistent presence of montane forest grasses (Wooller et al., 2000).

2.4.4 12,500 to 10,000 $^{14}$C yr BP

The period after 12,500 $^{14}$C yr BP and into the onset of the Holocene basically marks the terminal phase of the last glaciation. This transition period was marked by an increase in temperature and moisture, as evidenced by an establishment of lower montane forest as indicated by a reduction of *Hagenia* and Urticaceae and increase in *Podocarpus* pollen at Muchoya Swamp (ca. 11,000 $^{14}$C yr BP) and Ahakagyezi (ca. 10,600 $^{14}$C yr BP) (Taylor 1990). However, a major climatic fluctuation involving an abrupt return to arid conditions occurred at several sites just before 10,000 $^{14}$C yr BP (Figure 2.9), probably reflecting the Younger Dryas climatic reversal (c.f. Coetzee 1967; Hamilton 1982; Gillespie et al., 1983; Gasse et al., 1989; Roberts et al., 1993; Beuning et al., 1997; Johnson et al., 2000; Olago 2001). Further evidence for the Younger Dryas event affecting East Africa comes from the Burundi highlands and the Aberdare Mountains (Barker et al., 2004). In the Burundi highlands the abrupt aridity was marked by a shift to grassland pollen types (Bonnefile et al., 1995) while the Aberdare record shows cessation of plant fibers growth (Street-Perrott and Perrott 1990). At the Lake Albert site there was an abrupt return to drier conditions (ca. 11,400 to 9,900 $^{14}$C yr BP), as reflected by a 40% increase in grass pollen contemporaneous with an independent reduction of all forest taxa percentages to near zero (Beuning et al., 1997). Moist conditions resumed soon after ca. 9,900 $^{14}$C yr BP but were again interrupted by dry arid conditions at 8,000 $^{14}$C yr BP as evidenced by a change in sedimentation rate.
Figure 2.9  Enlarged subset of the pollen diagram of core F from Lake Albert. Percentages of pollen types are based on percent total of all identified, non-aquatic pollen grains. Selected taxa most clearly depict the 1500-year arid interval around the lake from 11,400 to 9,900 $^{14}$C yr BP. This arid interval is divided into three zones with “P” (11,400 to 10,900 $^{14}$C yr BP) and “B” (10,400 to 9,900 $^{14}$C yr BP) being the most arid and a slightly moister phase during zone “A” (10,900 to 10,400 $^{14}$C yr BP). Calculations of arboreal pollen percentages excluding Gramineae from the pollen sum produce similar reductions (from Beuning et al., 1997).
Open forest pollen types began to appear in Lake Tanganyika at ca. 12,000 $^{14}$C yr BP and attained maximum diversity at ca. 10,000 $^{14}$C yr BP (Vincens 1989a). The Zambezi woodlands expanded and diversified while the forests and Afro-alpine vegetation belt in the Lake Tanganyika catchment abruptly retreated upslope to levels similar to those of the present at the onset of the Holocene (Vincens 1989b). These changes occur during a regime of rising temperatures and are observed in many sites within the region, albeit with a few exceptions. For instance, the Lake Naivasha record (Maitima 1991) does not show evidence of vegetation change around this time (10,000 $^{14}$C yr BP). Similarly there is no evidence of temperature change in the Ruwenzori record (Livingstone 1967). Generally, there is no agreement among the researchers as to whether the vegetation shifts evident during this period represents a temperature or moisture signal: it is probably a combination of both.

2.4.5 Early to Middle Holocene

Pollen data from the East African region generally suggest that the early Holocene was warm and moist. In most sites forest pollen types became dominant concurrent with a reduction in Gramineae and an increase in Pteridophytes (Coetzee 1967; Bakker and Coetzee 1972; Owen et al., 1982; Maitima 1991; Beuning et al., 1997; Olago 2001). Apparently there was montane forest and thicket taxa in the Lake Turkana basin during this period (c.f., Bonnefille 1976; Owen et al., 1982). The Lake Turkana basin presently lies in an arid environment (Mohammed et al., 1995) and these pollen types may have been transported into the basin by rivers flowing from the Ethiopian Highlands, which suggests an increase in precipitation. It is also likely that the highland forests were more extensive at that time (Owen et al., 1982). At Lake Victoria forest taxa became more
abundant after a brief decline at 10,000 $^{14}$C yr BP, followed by a shift from an evergreen to a more deciduous type between ca. 7,000 and 6,000 $^{14}$C yr BP (Kendall 1969). Although humans had established themselves in the region by that time, their impact is assumed to be minimal meaning that the changes were probably driven by climate. The Lake Naivasha record reveals that the changes towards dry conditions occurred around 6500 $^{14}$C yr BP (Maitima 1991), while in Kashiru forest extension was pronounced between 10,000 and 5,000 $^{14}$C yr BP and declined afterwards (Roche and Bikwemu 1989).

Oxygen isotope record from diatomic silica confirms that the early to middle Holocene in East Africa was warm and moist, albeit with rapid climatic changes. Between ca. 11,100 $^{14}$C yr BP and 5,000 $^{14}$C yr BP two major negative shifts in $\delta^{18}$O$_{diatom}$ are evident in data from two sites on Mount Kenya (Barker et al., 2001). The first occurred from 11,100 to 8,600 $^{14}$C yr BP, and the second shift between ca. 6,700 $^{14}$C yr and 5,600 $^{14}$C yr BP. Both correspond with enhanced graminoid, shrub, and forest taxa pollen percentages, consistent with wetter and/or warmer conditions (Barker et al., 2001). The negative shifts in $\delta^{18}$O$_{diatom}$ are contemporaneous with high lake levels throughout the region (Street and Grove 1976; Street-Perrott and Perrott 1990). Further evidence of a warm and wet early Holocene in the region is adduced from ice core records from Mount Kilimanjaro (Thompson et al., 2002). The period between ca. 11,000 and 4,000 BP is characterized by isotopic ($\delta^{18}$O) enrichment (indicating warmer conditions) and reduced concentrations of major aerosol species ($\text{Ca}^{2+}$, $\text{Mg}^{2+}$, $\text{SO}_4^{2-}$, and $\text{NO}_3^{-}$, indicating wetter conditions). Within the early and middle Holocene the Kilimanjaro ice core record documents two serious drought events for the region, one at ~8,300 BP and the other at
5,200 BP (Thompson et al., 2002). The Kilimanjaro ice core record parallels the high-resolution diatom record from Lake Victoria that identifies similar trends at ~8,200 and ~5,700 yr BP (Stager et al., 2003). However, except for the Lake Naivasha records (Maitima 1991), these abrupt changes are not conspicuous in pollen records from the region.

2.4.6 Middle to Late Holocene

There was a general reduction in moisture in East Africa from the middle Holocene to the present. Interestingly, it is during this period that human impact on the environment begins to register in the regional pollen sequences. The anthropogenic impact on the environment in the tropics during the mid-Holocene complicates paleoenvironmental reconstructions by providing signals that are sometimes difficult to differentiate from the climate signals. For instance, drought events are often accompanied by increases in Gramineae pollen or high frequencies of microscopic charcoal. Human activities such as farming and/or pastoralism register similar signatures (c.f., Merchant et al., 2002). However, most sites in East Africa are unlikely to have experienced extensive human disturbance especially during the mid-Holocene due to the delayed adoption of iron technology (Hamilton et al., 1986).

A recent pollen record from Lake Tanganyika documents increasingly arid conditions after 5,000 BP with the probable replacement of forest by open grassland in parts of the lake’s catchment (Msaky et al., 2005). The Lake Bogoria pollen sequence shows an abrupt disappearance of high altitude forest pollen such as Hagenia abyssinica, Hypericum, Stoebe, and Ericaceae, and increases in more drought-adapted taxa such as Podocarpus, Juniperus, Acacia and Dodonaea after 4,500 $^{14}$C yr BP (Vincens 1986).
There was a sharp increase in *Podocarpus* after 4,500 $^{14}$C yr BP in Mount Kenya (Street-Perrott and Perrott 1993) and a similar rise in Mount Elgon at ca. 3,500 $^{14}$C yr BP (Hamilton 1982), which signify the onset and establishment of dry conditions in the region. Of the four species of *Podocarpus* found in East Africa, only two are widely distributed—*Podocarpus gracilior* and *P. malanjianus*, with the former being more typical of dry montane forests than the latter (Hamilton 1982). During this same period there was a rise in dry forest components such as *Olea, Macaranga, Pygeum, Celtis, Syzygium, Neoboutonia*, and *Galiniera* (Coetzee 1967; Street-Perrott and Perrott 1993; Olago 2001).

The sharp increase in *Podocarpus* at ca. 4,500 $^{14}$C yr BP at the Mount Kenya site (Street-Perrott and Perrott 1993) and at ca. 3,500 $^{14}$C yr BP at Mount Elgon (Hamilton 1982) corresponds with the third abrupt climate event recorded in the Mount Kilimanjaro ice core record (Thompson et al., 2002). In the Kilimanjaro ice core record the desiccation episode is marked by a visibly distinct (30 mm thick) dust layer containing high concentrations of other chemical species ($\text{F}^-$ and $\text{Na}^+$), accompanying a possible hiatus in ice accumulation. Lake level records from Lake Rukwa, Tanzania (Talbot and Livingstone 1989), and Lake Cheshi, Zambia (Stager 1988) show an abrupt dry episode at ca. 3,000 $^{14}$C yr BP. The abrupt dry event during ca. 4,500 –3,500 $^{14}$C yr BP has now been recognized across tropical Africa (Hamilton 1982; Talbot 1982; Street-Perrott and Harrison 1984, 1985; Street-Perrott and Perrott 1993; Elenga et al., 1994; Maley and Brenac 1998; Bonnefille and Chalié 2000; Barker et al., 2004). This abrupt aridity has been associated with decline in the base flow of the White Nile (Talbot and Brendeland 2000) and has also been linked to the collapse of the Old Kingdom in the Nile Valley (c.f. Hassan 1997; Barker et al., 2004). The abrupt shift to a drier and more seasonal
environment at ca. 4,000 $^{14}$C yr BP has also been recorded at different sites in the island of Madagascar, which lies to the southeast of the study region (Burney 1993).

From 2,500 $^{14}$C yr BP there was progressive degradation of arboreal cover and a concurrent increase in Gramineae pollen in the northern basin of Lake Tanganyika, which is probably a consequence of both increasing aridity and human interference. There is evidence of human activities documented in the pollen sequence, such as the presence of the rare *Elaeis guineensis* (wild oil palm) in the Masoko record (Vincens 1989a). Pollen and charcoal records from Ahakagyezi Swamp show evidence of deforestation associated with agricultural activities before ca. 4,800 $^{14}$C yr BP (Hamilton *et al*., 1986). Unfortunately most pollen of cultivated plants has not been definitely identified (c.f., Msaky *et al*., 2005). Environmental changes associated with deforestation appeared in Muchoya Swamp at ca. 2,500 yr BP and contemporaneous changes are evident in the Batongo and Katenga mires (Morrison and Hamilton 1974; Taylor 1990). At Lake Tanganyika there is a consistent trend of a decrease in grass pollen, and increase in pteridophyte and forest indicator pollen during the last few centuries (Msaky *et al*., 2005). This trend suggests an increase in soil erosion and recycling of abundant arboreal pollen following with land clearance for agriculture. Increase in sedimentation rates during this period provides further evidence for increased soil erosion in the Lake Tanganyika catchment (Cohen *et al*., 2005; Msaky *et al*., 2005; Palacios-Fest *et al*., 2005).

### 2.4.7 The Last Millennium

Verschuren *et al*.(2000) obtained a 6 m sediment core from the partially submerged Crescent Island Crater (CIC) basin in Lake Naivasha, Kenya, to reconstruct
the record of rainfall and drought in equatorial East Africa during the past 1,100 years. Lake Naivasha is a freshwater lake with subsurface outflow and a short (<10 yr) water-residence time (Ojiambo and Lyons, 1996). Strong evaporation and highly variable river inflow cause its surface elevation to fluctuate in response to changing climate regimes (Verschuren et al., 2000). Although the lake level record for Lake Naivasha is shorter than that for Lake Victoria and Lake Albert it provides high-temporal-resolution information concerning recent climate history, occasionally corroborated by oral tradition.

A combination of sediment-inferred depth reconstruction and diatom composition and distributions revealed that Lake Naivasha experienced a long saline lowstand from ~AD 1000 to 1270 interrupted by one freshwater interval in the early part of the millennium (Figure 2.10). This lowstand was followed from ~AD 1270 to 1550 by a mostly positive water balance and the establishment of freshwater conditions, except for one recurrence of saline conditions dated around ~AD 1380 to 1420. The Lake Naivasha record consistent with the Lake Tanganyika record for this period, which shows a peak in total inorganic content (TIC) and a rapid decline after 1,100 AD concurrent with a rise in charcoal mass accumulation rates (Cohen et al., 2005). A second pronounced lowstand in Lake Naivasha is dated to ~1560 to 1590, followed by a highstand that lasted from between 1670 and 1770 with lake level continuously above the historical maximum reached in 1894. During the highstand period (1670-1770) the diatom flora was dominated by *Aulacoseira ambiguа*, which, compared to *A. granulata*, thrives in higher light and lower nutrient environments.
Figure 2.10. A plot of lake level, and salinity record for Crescent Island Crater (Lake Naivasha) compared with the decadal record of atmospheric $^{14}$CO$_2$ production as a proxy for solar radiation. The grey bars are the severe drought events as recounted from oral tradition. Bars A, B, and C represent the Wamara, Nyarubanga, and Lapanarat-Mahlatule droughts, respectively (Modified from Verschuren et al., 2000).
Lake-level and salinity fluctuations were superimposed on a long-term freshening trend in the lake that started around AD 1270; evident in the gradual disappearance of salt-tolerant diatom and chironomid taxa and the expansion of stenotrophic freshwater taxa such as *Psectrocladius viridescens*. The parallels in the different proxy records are strikingly apparent. The fluctuations in lake levels are indicative of reduced precipitation and extensive drought periods. This is confirmed by oral traditions (c.f., Webster, 1979; Verschuren et al., 2000). Evidence for drought-induced famine, political unrest, and large-scale migration of indigenous peoples is concentrated in three periods around AD 1390-1420 (Wamara drought), AD 1560-1625 (Nyarubanga drought), and AD 1670-1840 (Lapanarat-Mahlatule drought) that match the reconstructed sequence of lowstands (Verschuren et al. 2000).

The Lake Naivasha record corresponds with the findings from Lake Victoria for the same period. Stager et al. (2003) used a diatom record from Lake Victoria to reconstruct the paleoclimate of the region during the Holocene. Although they observe a general aridification trend over the last 3,000 $^{14}$C yr BP, abrupt lake level decline is registered between 1200 and 600 $^{14}$C yr B.P due to reduction in precipitation.

2.5 Synthesis

Only a handful of paleoenvironmental records from East Africa extend beyond 42,000 yr BP (Coetzee 1967, Taylor 1990; Street-Perrott et al., 1997; Olago et al., 1999). The few available (mostly pollen records), however, present useful data for the reconstruction of East Africa’s paleoenvironment beyond the LGM. The period prior to ca. 42,000 $^{14}$C yr BP is characterized by warm climatic conditions similar to the present. The shifting of high altitude vegetation belts to lower elevations between 42,000 $^{14}$C yr
BP and 30,000 \(^{14}\text{C}\) yr BP marks a change to cold dry conditions. Much of this evidence is obtained from pollen and diatom data as well as charred grass cuticles and is verified by carbon isotope (\(\delta^{13}\text{C}\)) record from the region (Olago 2001). Cool conditions prevailed in East Africa in the period leading to the LGM with temperatures being between 2 to 4.1\(^{\circ}\text{C}\) colder than present (Coetzee 1967; Olago 2001). Paleoenvironmental proxies show that between ca. 21,000 and 12,500 \(^{14}\text{C}\) yr BP East Africa’s environment was generally cool, punctuated by episodes of prolonged desiccation. (Kendall 1969; Street and Groove 1979; Beuning et al., 1997).

The Lake Albert records documents two extended episodes of lowstands of at least -46 m below the present lake levels during the Last Glacial Maxima (LGM) ca. 18,000-20,000 and at 12,500 \(^{14}\text{C}\) yr BP. Core Ibis 1 from Lake Victoria also documents two episodes of lowstand, at \(\sim 15,390 \pm/\pm 220\) \(^{14}\text{C}\) yr BP and \(\sim 11,710 \pm/\pm 120\) \(^{14}\text{C}\) yr BP. The aridity episode in Lake Victoria must have reduced the lake level to at least-66 m below present at the site of core Ibis 3.

It is likely that the two lakes are recording the same desiccation episodes, albeit with differences in dating control. The dating should be interpreted with caution because low lake levels are likely to be associated with increased mixing and bioturbation that could interfere with both the integrity and the time resolution of climate-proxy records. Whereas the first episode at 18,000-20,000 \(^{14}\text{C}\) yr BP or \(\sim 15,390\) (?)\(^{14}\text{C}\) yr BP (for Lake Albert and Lake Victoria respectively) could be synchronous with the LGM the second aridity event recorded in the two lakes (12,500 \(^{14}\text{C}\) yr BP, 11,710 \(^{14}\text{C}\) yr BP) could be associated with the Younger Dryas interval. Palynological and other proxy evidence from Lake Albert (Fig. 3) further provides clear evidence of significant and abrupt drying of
climate in the Lake Albert region at ~12,500 $^{14}$C yr BP. The terminal phase of the last glaciation, from 12,500 $^{14}$C yr BP into the onset of the Holocene was marked by an increase in temperature and moisture. However, more work needs to be done before the occurrence of the Younger Dryas in East Africa can be confirmed.

Different proxies e.g., pollen, diatoms, charred grass cuticles, and oxygen isotope records from diatomic silica, suggest that East Africa was characterized by warm and moist conditions amid rapid climatic changes during the early to middle Holocene period. The ice core record from Mount Kilimanjaro (Thompson et al., 2002) documents two significant and abrupt drought events in the region, one at ca. 8,300 BP and the other at 5,200 BP, which appear to correspond with an apparent change in sedimentation rate at Lake Albert sometime between 8,000 and 3,400 $^{14}$C yr BP (Beuning et al., 1997). However, these drought events are not recorded in the pollen sequence. This emphasizes the importance of proxy sensitivity and sampling resolution as it affects paleoenvironmental data interpretation. The ice core record provides higher-temporal resolution data than the lake sediment and pollen records can, and therefore is able to detect the two abrupt drought events. The longer and more extensive drought event at ca. 4,000 $^{14}$C yr BP (c.f. Marchant and Hooghiemstra 2004) was registered in both the pollen and diatom records at different sites (Street-Perrott and Perrott 1993; Hamilton 1982; Maitima 1991; Thompson et al., 2002).

Lake level records from Lake Naivasha emphasize the significance of high-resolution stratigraphic records in revealing decadal and century-scale paleoclimatic changes. The Naivasha lake level record suggests that the climate of equatorial East Africa was generally drier than today during the Medieval Warm Period (MWP; AD
1000-1270), and that fairly wet conditions during the Little Ice Age (LIA; ~AD 1270-
1850) were interrupted around AD 1380-1420, 1560-1620 and 1760-1840 by episodes of
persistent aridity, which seem to be more severe than those of more recent times
(Verschuren et al., 2001; Verschuren 2004). Again these subtle decadal and century-scale
events are missing in the pollen record from the lake (c.f., Maitima 1991), further
emphasizing the role of proxy sensitivity in revealing details of paleoenvironmental
conditions.

That the oscillations in the lake levels and changes in pollen and other proxy
records reflect changes in the climate and vegetation of the region is not in doubt.
However, the challenge remains in the understanding of the forcing mechanism behind
paleoenvironmental changes in a tropical region. Although the Milankovich cycles have
been advanced as a possible explanation for the changes (Short et al., 1991; Tyson 1999;
Tyson et al., 2000; Olago 2001; Stager et al., 2003; Verschuren et al., 2004), they
probably are not responsible for the centennial-scale and decadal-scale changes such as
those evident from the Lake Naivasha records. Therefore more studies and thoughts
ought to be given to possible causal mechanisms.

A number of hypotheses have been offered identifying these mechanisms. One
explanation is based on the intensity of the southeast summer monsoon and the Congo air
stream that influence the amount and seasonality of rainfall in the East African region
(c.f., Hamilton, 1982; Nicholson, 1996, 2001; Vincens et al., 2003). Since the southeast
summer monsoon acts in tandem with the seasonal movement of the Intertropical
Convergence Zone (ITCZ), their character must have dramatically changed during the
periods of intense aridity recorded in the East African lakes.
It is possible that the summer monsoon was very weak during the periods of intense aridification (ca. 20,000-18,000 $^{14}$C yr BP, 12,500 yr BP, and after ~4,000 $^{14}$C yr BP), resulting in intense droughts that reduced the lake levels to those registered in Lakes Albert, Victoria, and Naivasha. Indeed, records from many sites in terrestrial Asia and the Arabian Sea, which correspond with those from terrestrial sites in Africa, suggest the existence of significant century-scale monsoon weakening events especially during the early to middle Holocene (Van Campo et al., 1982; Bryson 1989; Street-Perrott and Perrott 1990; An et al., 1993; Van Campo and Gasse 1993; Gasse and Van Campo 1994; Overpeck et al., 1996). These variations in the summer monsoon may explain some of the abrupt climatic changes evident in the paleoecological records from East Africa.

Changes in the summer monsoon are linked to changes in global circulation patterns, especially temperature changes associated with events such as the deglaciation at the end of the LGM, and Younger Dryas episodes in the mid-latitude regions of the Northern Hemisphere as well as sea surface temperatures (SST). There is also increasing evidence linking changes in the monsoon and the ITCZ to SST changes (Gasse and Van Campo 1994; Gasse 2002) as well as the character of the thermohaline circulation (THC) (Nyberg et al., 2002).

The position of the ITCZ has been shown to be closely linked to the changes in the sea surface temperatures over the Atlantic Ocean (Gasse and Van Campo 1994; Sirockoo et al., 1996; Marshall et al., 2001) and the El Nino Southern Oscillation (ENSO) events (Nicholson 1994; 1996; Hastenrath 2002). Periods of lower precipitation in Africa have been associated with warm SST in the northern Atlantic Ocean and colder SST in the southern Atlantic Ocean. Gasse and Van Campo (1994) specifically link the weak
monsoons at ca. 4000 $^{14}$C yr BP to contemporaneous decreases in the North Atlantic SST. Decrease in the latter would reduce moisture in the parts of Africa where monsoonal airflow brings water vapour from the southern sub-tropical Atlantic anticyclone (Marchant and Hooghiemstra 2004).

At the centennial timescale, the lake level fluctuations could also be a result of the solar activity and atmospheric radiocarbon ($^{14}$C) production (c.f., Jouzel et al., 1993; Olago 1995; Street-Perrott et al., 1999). The latter phenomena are negatively linked because solar activity (as measured by reduced sunspot numbers) increases cosmic ray bombardment in the upper atmosphere, which enhances $^{14}$C production. Indeed a comparison of the Lake Naivasha record with that of reconstructed atmospheric $^{14}$CO$_2$ production reveal that the inferred MWP African aridity and all the low lake-level episodes were coeval with high solar activity, with the intervening highstands contemporaneous with low solar activity (c.f., Street-Perrott et al., 1997; van Geel et al., 2000; Olago 2001; Verschuren et al., 2000 Thompson et al. 2002). For instance, the highest rainfall inferred from Lake Naivasha over the past millennia was contemporaneous with the ‘Maunder Minimum’ of solar radiation (Verschuren et al., 2000). Therefore, variation in solar activity may have contributed to the changes in the paleoclimate of East Africa.

Another hypothesis for the paleoenvironmental changes in East Africa is the possible teleconnection between high and low latitudes during the early and mid-Holocene (Stager et al., 2003). This could be the forcing mechanism behind the abrupt dry episodes at ~ 8,300 and 5,200 BP (Thompson et al., 2002) that were synchronous with major meltwater and ice rafting phenomena in the North Atlantic (Bond et al.,
1997). However, the 8,200 BP meltwater event was based on calendar years (first detected in ice core records). Firm conclusions regarding the teleconnections hypothesis are not possible due to uncertainties surrounding the radiocarbon chronologies (c.f., Stager et al., 2003).

Human-induced vegetation changes in East Africa, especially during the Holocene, cannot be ruled out. During this period humans progressed from being relatively minor influences, restricted to a few ecosystems, to becoming a major force across even remote and inhospitable environments (Hamilton 1982). The records from Ahakagyezi (Hamilton et al., 1986; Taylor 1990), Muchoya (Taylor 1990; Marchant and Taylor 1998), Sacred Lake (Street-Perrott and Perrott 1997), Lake Tanganyika (Cohen et al., 2005; Msaky et al., 2005), and Masoko (Vincens et al., 2003) show reduction in arboreal taxa, and an increase in fire episodes (charcoal peaks) after 5,000 $^{14}$C yr BP that have been interpreted as human-induced. These interpretations are based on the presence of high counts of microscopic charcoal (suggesting fire occurrence) concurrent with forest clearance. In some cases (e.g., Vincens et al., 2003) forest clearance is accompanied by increased pollen frequencies of cultivated crops (e.g., *Elaeis guineensis*), Gramineae, and disturbance markers (e.g., *Ricinus communis* that may indicate clearance for agricultural and/or pastoral activities).

Rapid vegetation changes due to human activities have the potential of producing remarkable changes in land surface albedo capable of inducing feedback processes that could result in changes in climate (c.f., Kutzbach et al., 1996; Doherty et al., 2000). Climate models have shown that vegetation-albedo changes provide positive feedback effects that cause African climate to switch abruptly between wet and dry conditions.
(Claussen 1997; Claussen et al., 1999). It is possible that the mid-Holocene aridification may have been stimulated by changes in land-surface conditions (c.f., Lamb et al., 1989; de Menocal et al., 2000). However, the fundamental problem remains to separate purely background-level environmental or climatic change from that caused by humans. Since most indigenous crop plants in East Africa (e.g., cereals, cassava, banana, and legumes) have not left identifiable traces in the pollen record, the palynological identification of agricultural influence in the region becomes difficult. Another problem is that natural drought events are often accompanied by increases in grass pollen or high frequencies of microscopic charcoal. Human activities such as farming and/or pastoralism register similar signatures, further complicating the interpretation process. Therefore, whenever anthropogenic disturbance is suspected in paleoenvironmental changes, doubt can only be eliminated by thorough multi-proxy evidences including archaeological findings.

Deciphering the Holocene anthropogenic fire-related vegetation changes should be treated with caution. Fire episodes in the East African records appear as early as ~30,000 $^{14}$C yr BP (Wooller 2000) and may have no anthropogenic link. Records from the island of Madagascar (latitude 12°S to 26°S) (Burney 1987a), similar to East Africa in terms of climate and vegetation communities, document fire episodes well before the presumed arrival of humans ca. 1,500-2,000 $^{14}$C yr BP (c.f. Burney 1987b; 1993; 1997). Clearly, not all paleoecologically recorded fires were induced by humans, since even in the absence of humans, wild fires occur when necessary conditions of low fuel moisture and a natural ignition source (for instance lightning and volcanic activities) are met. However, changes in fire regimes and intensity may point to the human agency, as evident in the Masoko records (Vincens et al., 2003). Unfortunately most of the records
from tropical Africa have temporal resolutions that are too coarse to fully investigate the anthropogenic effect on the vegetation.

Unequivocal interpretation of the various sedimentological, biological, and geochemical climate proxy data extracted from the lake sediments with respect to past climatic variations is an extremely complex and challenging exercise. Whereas this review offers a contribution to our understanding of paleoenvironmental changes in East Africa, it also highlights what is unknown. For instance much of the accepted paradigms are based on coarse-temporal-resolution analyses, which may miss subtle, yet important, changes in vegetation and climate. Although humans and their ancestors are assumed to have interacted with East African environments longer than in most other regions in the world, very little is known about their impacts on the environment especially during the Holocene, let alone the earlier periods. There is a need for more higher-temporal-resolution palynological studies with an emphasis on multi-proxy methodologies in order to improve our knowledge of the human/climate impact in the region. We still need to devise methods to separate purely climatic and human-driven environmental changes. Identification of pollen of indigenous cultivated crops is certainly part of the solution.

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CHAPTER 3. STRATIGRAPHIC EVIDENCE OF CLIMATE CHANGE AND LAND DEGRADATION IN THE LAKE BARINGO AREA, KENYA, EAST AFRICA, OVER THE PAST 350 YEARS

3.1 Introduction

Climate-sensitive lakes of the East African rift system offer the best archives for paleoenvironmental record in the African tropics. They archive both short- and long-term regional climate dynamics and integrate anthropogenic influences and other environmental signals in their sedimentary records (c.f., Johnson 1993; Russell et al. 2003). Lake Baringo (Figure 3.1), one of the lakes within the eastern arm of the East African Rift Valley, is auspiciously positioned to archive local and regional environmental changes. Well-dated lithostratigraphic sequences from the lake have a great potential for offering high resolution record of paleoenvironmental change in the tropics where there are few high-temporal-resolution climate and vegetation histories.

Changes in the lake stratigraphy and sedimentation rates can play a complementary role to other proxy records in the reconstruction of anthropogenic influences on paleoenvironments. Where human impact on the watershed varies through time, changes in sedimentation rates can provide a means of identifying the timing and magnitude of deforestation and/or land degradation. In order to tap this potential, analysis of the quantity of sediment discharge into the lake needs to be determined. Lake Baringo is of great ecological interest because it has recently undergone dramatic change in surface area and volume that threatens the rich biodiversity within its ecosystem (UNDP 2002; Kurgat 2003).

Paleoecological investigations of sediment cores from East African lakes, swamps, and mires, already outlined in chapter two of this dissertation, have established
Figure 3.1. Geographical position of Lake Baringo in East Africa. The lake is one of the lakes within the Eastern (Gregory) Rift Valley.
the outline of paleoenvironment since the Last Glacial Maximum (LGM) (e.g., Livingstone 1962, 1967; Bakker 1964; Coetzee 1967; Bonnefille and Riollet 1988; Taylor 1990; Street-Perrott and Perrott 1993; Olago 1995; Street-Perrott et al. 1997; Gasse 2000; Kiage and Liu 2006). These studies largely offered a broad synopsis of the East African paleoenvironment at millennial time-scales with little emphasis on centennial- and/or-decadal scales. Unfortunately, to date very few lake-based records contain the temporal resolution and dating control required for centennial- or decadal-scale climate variability (Thompson et al. 2002).

Higher resolution studies, where available, have illuminated sub-millennial oscillations in African paleoclimatic conditions akin to the Holocene Bond Cycle (Bond et al. 1997), including the Younger Dryas (Roberts et al. 1993; Beuning et al. 1997; DeMenocal et al. 2000; Barker et al. 2001), the Medieval Warm Period (MWP), and the Little Ice Age (LIA) (Marchant and Taylor 1998; Verschuren et al. 2001; Lamb et al. 2003). The Late Holocene climate of the African tropics, especially the last 2000 years was previously assumed to be climatically stable (e.g., Livingstone 1975, 1980; Hamilton 1982) mainly due to poor dating control and low temporal resolution that obscured decadal- and century-scale climate variability. This erroneous assumption was further enhanced by failure to recover the soft surface muds that contain the 20th century record (c.f., Verschuren et al. 2004). The use of AMS technique in radiocarbon dating coupled with $^{210}$Pb and $^{137}$Cs has led to the revision of the paradigm.

Increasing evidence (e.g., Crossley et al. 1984; Lamb et al. 2003; Verschuren 2004) now suggests that climatic conditions throughout tropical Africa over the last millennia may have been quite unstable, characterized by moisture-balance fluctuations
that must have seriously disrupted indigenous civilizations. Considering that these fluctuations occurred at all time scales the challenge currently facing the paleoenvironmental researchers is to structure their studies to a temporal resolution and precision that can elucidate the causes and pacemakers of decadal to century-scale environmental changes (Verschuren 2004). Fine-temporal-resolution studies have the potential for discriminating anthropogenic influences from climatic and background dynamics and their effect on the environment.

This chapter presents high-resolution stratigraphic data from a 350-year sediment record from Lake Baringo based on four sediment cores recovered from the lake. A comparison of the Lake Baringo stratigraphic record, especially its sedimentation rate, to those from other lakes in the East African region and tropical Africa is made to identify similarities and unique features with regard to climate and human signals. In this dissertation I define sedimentation rates (in contrast to accumulation rates which express the weight of sediment per unit area) as the thickness of the solid, pore free sediment deposited during a given time frame (c.f., Einsele and Hinderer 1998). The comparison of sedimentation rates is a necessary ingredient for a holistic picture of paleoenvironmental records from the African tropics, which is key to understanding past changes in the climate system as well as the sensitivity of tropical regions to future climate change (Thompson et al. 1998; Barker et al. 2001). Where possible the stratigraphic record is linked to rainfall records within the Lake Baringo watershed.

3.2 Background

Lake Baringo (latitude 0°30′N and 0°41′N and longitude 36°00′E and 36°10′E) (Figure 3.1) is located in the Rift Valley province of Kenya. The lake is part of the
Gregorian Rift Valley system lakes and is one of only two lakes in the eastern fork of the Rift Valley that have fresh water. The eastern Rift extends northwards from Lake Malawi through Tanzania, Kenya, and Ethiopia, and is considered to be continuous with the valley that contains the Red Sea, and the Jordan valley. The earth-movements which preceded the formation of the Rift Valleys began in the early Mesozoic and were accompanied by volcanic activity as evident by numerous craters on the valley floor (Renaut et al. 1999, 2000; Tiercelin et al. 1987). The lavas that were discharged during the volcanic activity were rich in soda but poor in lime, which is reflected in the chemistry of the waters within the Rift Valley lakes (c.f., Renaut et al. 1999, 2000; Tiercelin et al. 1987; Gregory 1921). Thus all the lakes within the Rift Valleys, including Lake Baringo (pH=8.6), are alkaline, their alkalinity being mainly due to sodium carbonate (10.1 meq 1⁻¹), and not calcium carbonate (1.7 ± 0.1 meq 1⁻¹) as the case with most natural waters (Beadle 1932; Barton et al. 1987; Oduor et al. 2003).

Lake Baringo has an area of about 129 km² and lies at an altitude of 970 m above sea level (Renaut et al. 2000). The lake is presently shallow with a mean depth of 2.5 m, and the deepest part in the northern zone measures 3.8 m (Figure 3.2) (Aloo 2002; Hickley et al. 2004). It is situated in a semi-arid zone with low annual rainfall averaging 600 mm falling within two rainy seasons controlled by the movement of the Intertropical Convergence Zone (ITCZ). Much of the rainfall comes in sporadic downpours within a few days in the months of April/May and October/November. These downpours are highly erosive of the clay and clay loams that characterize the area, especially in the context of sparse vegetation cover. Variation in rainfall in Lake Baringo occurs on a range of time scales, including El Nino and La Nina periods.
Figure 3.2  Bathymetric map of Lake Baringo (August 2003) modified from Hickley et al. 2004. The contour lines progress at 0.5 m (beginning at 1.5 m) and the dots numbered 1-7 are the coring sites. Contour lines less than 3 m around the islands are mostly omitted. The deepest points in the lake are also indicated. The Bathymetric map offers broad generalization of the depth and some sections of the lake may not fall within the scheme.
i.e., every 5-7 years (LaVigne and Ashley 2002). The average temperatures are fairly high (25-30 °C) and the potential evapotranspiration exceeds 2500 mm (c.f., LaVigne and Ashley 2002; Driese et al. 2004). Longer-term climatic variations have been noted in the region including ca. 1500 year dry-wet climate cycles (Mohammed et al. 1995; Verschuren et al. 2000; Lamb et al. 2003) and Milankovich forcing at ca. 20 ka, 41 ka, and 100 ka that respectively correspond with orbital precession, obliquity, and eccentricity cycles (Pokras and Mix 1987; Gasse et al. 1989; de Menocal and Bloemendal 1995; Sirocko 1996).

Lake Baringo is fed by four main rivers, namely Endau, Ewaso Nanyokie (the red river, formed by rivers Perkerra and Molo), Ol Arabel, and Mukutan (Figure 3.3). Other small rivers include Sandalo, Bargera, Tigeri, and Marigato. All these rivers, with the exception of Ewaso Nanyokie and Mukutan, which receive their supply from the Elgeyo Escarpment and from the north end of the Mau escarpment, are seasonal (i.e. they flow only during the wet season). To the east and west of Lake Baringo the land rises abruptly as a series of distinct fault-steps that form the Laikipia Escarpment and Tugen Hills fault-block, respectively (Renaut et al. 2000). A number of hot springs on the largest volcanic island (Kokwa) also discharge into the lake. The water from the hot springs has a higher alkalinity (>10.6 meq l⁻¹) than that of the lake water (~9.4 meq l⁻¹) (Beadle 1932; Oduor et al. 2003).

Although Lake Baringo is classified as fresh-water it has no active surface outlet. It is hypothesized that the lake must have an underground outlet that ensures its comparatively low alkalinity in spite of the inflow of the soda springs. It is probable that the present underground outlet feeds the springs of Kogore Valley some 10 kilometers to
Figure 3.3  Map of Lake Baringo showing the rivers that feed the lake. All the rivers with the exception of rivers Molo and Perkerra are ephemeral.
the north of the lake (c.f., Beadle 1932; Barton et al. 1987). Kogore Valley is 30 m below the level of Lake Baringo.

The littoral vegetation community around much of Lake Baringo is very poorly developed except for some marshes on the southern shores. Although the lake is located within the vegetation zone classified by White (1983) as deciduous bushland and thicket (drier savanna types) (Figure 3.4) its terrestrial vegetation, away from the littoral zone is highly heterogeneous with high species diversity and significant intra-specific variations in physiognomic characteristics along the elevational gradient.

Over forty-six percent of the land around Baringo is either too steep or too dry to support agriculture (Hickley et al. 2004). The main occupation of the local population is livestock herding, with flocks consisting mainly of goats which are able to browse in the steep slope environment. However, intensive agriculture is practiced in the highlands (Tugen Hills and the Mau Escarpment) where the principal crop is maize (corn). The widespread transformation of vegetated areas to degraded sites and bare surfaces, consisting of grazing lands, farms, and human settlements, is readily visible (Snelder and Bryan 1995; Mwasi 2001; Johansson and Svensson 2002).

3.3 Methods

Eleven cores were raised from Lake Baringo on different dates over two field seasons in the months of January 2004 and 2005, using a modified Livingstone corer (c.f., Wright et al. 1984) from different water depths ranging from 2.5 m to 3.5 m (Figure 3.5). All the cores were preserved in tubes that were carefully sealed on both ends at the field collection sites. The cores were then transported to Louisiana State University the laboratory, each core was longitudinally split into two equal halves, photographed,
Figure 3.4 Map of East Africa showing the major vegetation communities. The approximate location of Lake Baringo is shown by the red dot.
and macroscopically described.

The bathymetric map of Lake Baringo (Figure 3.2) was modified from Hickley et al. (2004) and was used to determine core sites during fieldwork. The map was then digitized using geographic information system software (ArcGis/ArcMap, ESRI), registered to a Landsat 7 ETM+ image of January 28 2000, and subsequently used to calculate corresponding lake area and volume. Additional information for geo-referencing and geo-correction of the bathymetric map and the satellite image was obtained from the topographic map of Lake Baringo area (Survey of Kenya 2002).

Several 1-cm\(^3\) samples were taken consecutively throughout the cores at 1 cm intervals and subjected to loss-on-ignition analysis. Weight loss was measured after drying the samples overnight at 105°C, and after combustion at 550°C and 1000°C to establish the water, organic, and carbonate content respectively (c.f., Dean 1974; Heiri et al 2001; Boyle 2004). The data were entered onto a spreadsheet and the graphs were plotted using TILIA and TILIAGRAPH version 2.0.2 (Grimm 2004).

Three samples from the longest cores (LB-5 and LB-6) were sent to Beta Analytic Inc. for AMS \(^{14}\)C analysis to establish the chronology. Once the chronology for the oldest sediments was established the younger sediments were sampled at 20 cm intervals and analyzed for \(^{210}\)Pb dating at the Coastal Studies Institute, LSU. Samples for radionuclide measurements were dried for 24 hours at 60°C, ground with a porcelain mortar and pestle, and sealed in air-tight petri dishes. Weighed masses were then counted for 24 hours on Canberra low-background planar gamma detectors; correction for self absorption of \(^{210}\)Pb was done using the method of Cutshall et al. (1983). Total \(^{210}\)Pb was determined by measurement of the 46.5-KeV \(^{210}\)Pb gamma peak. Supported \(^{210}\)Pb from
Figure 3.5  North-South transect of the bathymetry of Lake Baringo showing the different water depths under which the cores were collected.
the decay of $^{226}$Ra within the lakebed was determined by measurement of the
granddaughters of $^{226}$Ra, $^{214}$Pb (at 295 and 352-KeV) and $^{214}$Bi (at 609-KeV).

The material that was sampled for the AMS $^{14}$C analysis was part of the hard
clayey plant fibers material at the base of core LB-5 between 343 and 363 cm. The AMS
$^{14}$C dates were complemented by nineteen $^{210}$Pb dates taken at 20 cm intervals throughout
core LB-5. All the cores were correlated primarily by prominent lithostratigraphic
characteristics and especially the sedimentary changes. Figure 3.6 shows the Age versus
Depth plot based on core LB-5 and LB-6. The main features of note in depth-age profiles
are:

1) A relatively uniform sedimentation rate of 0.83 cm/yr for the past 100 years based
on $^{210}$Pb chronology.

2) Sedimentation rate of 0.99 cm/yr over the past ca. 350 calendar years based on
AMS $^{14}$C chronology.

3.4 Results

Below we present the lithostratigraphic and geochemical analyses based on four
of the longest cores from Lake Baringo beginning from the longest to the shortest.

3.4.1 Core LB-5

LB-5 was collected from under 352 cm water depth in the southwestern part of
the lake (Figure 3.2). The core was raised in three separate segments of varying lengths
i.e., 126 cm, 130 cm, and 107 cm respectively. All the core segments were in excellent
condition at recovery. The lithostratigraphic features of LB-5 are summarized in figure
3.7. Gross lithological differences which, with a few exceptions, are in phase with
changes in the water content obtained from the LOI procedure have been used to
Figure 3.6 Age versus depth plot based on core LB-5 and LB-6. Sediment accumulation rates for the top 100 cm are based on 210 Pb and the rest of the core is estimated based on two AMS 14C radiocarbon dates.
Figure 3.7  Results of loss-on-ignition for core LB-5. The zones I through V in the diagram were based on the lithostratigraphic characteristics of the core.
subdivide the core into seven major units (zones): VII (363-343 cm), VI (343-302 cm), V (302-260 cm), IV (260-163 cm), III (173-130 cm), II (130-54 cm), and I (54-0 cm).

Zone VII, from 363-343 cm, is composed of hard dark-gray, clay with crumbly appearance containing plant microfossils. At the bottom of the unit well-preserved rootlets and seeds, probably of plants of marsh/wetland environment, were recovered and constituted part of the organic material that was submitted to Beta Analytic Inc. for AMS $^{14}$C analysis. The coring process was terminated in this zone because the sediments were so hard that further penetration by the modified Livingstone corer became impossible. The bottom of this unit was dated to approximately 300 ± 50 $^{14}$C years BP (Cal AD 1650) (Beta Analytic Laboratory Inc. Number 207988). The results of LOI show that the sediments in this unit consist of approximately 36-40% water, 10-12% organic, and less than 2% carbonate content. Although the carbonate content is low throughout the core, zone V consistently records the lowest amounts.

Zone VI (343-302 cm) consists of soft dark-grey mud devoid of any form of lamination that dominate upper potions of the core. The organic content in this zone remains above 10% as in the preceding zone (VII) while the carbonate content is higher by about 1%. The water content in this zone is slightly over 60%, the highest record in the entire core.

Zone V (302-260 cm) is characterized hard dark grey clayey material with a crumbly structure. This zone differs from VI by having significantly lower water content very similar to those recorded in zone VII. It has a similar lithological composition to zone VII except having slightly higher carbonate contents (~2.7%).
Zone IV (260-163 cm) is dominated by brown-dark brown clay sediments that are fairly well laminated. It is the longest portion of core LB-5. The water content is fairly uniform (45-50%) while the other constituents of LOI do not show significant change from zone V.

Zones III (163-130 cm) is characterized by a band of dark clay with no laminations caped by ~5 cm of dark hard crumbly clay devoid of laminations while zone II (130-54 cm) is similar to zone IV. However, in zone II the organic content is consistently above 8% while zones IV and III had an average of 5-6%.

The sediments in zone I (54-0 cm) are made of dark-brown clayey material that has a crumbly structure, especially towards the top of the core. The crumbly structure is most pronounced in the top ten centimeters that are also characterized by fairly low water content (33-35%). The rest of the zone consists of brown to dark-brown soft clay sediments that had no apparent laminations which contrasts sharply with the zones beneath. The organic and carbonate contents in this section are similar to those of zone II with average values of 7-8% and 2-3%, respectively.

3.4.2 Core LB-6

Core LB-6 (Figure 3.8) was collected approximately three kilometers to the northwest of the location of core LB-5 (Figure 3.2) from under 259 cm of water. The core, which measured 228 cm, was raised in two segments of 123 cm and 105 cm respectively. The sediments at the bottom of the core were so hard that coring could not be continued beyond 228 cm. Core LB-6 appears to match the characteristics of the other cores described in this chapter, especially those of core LB-5. Gross lithostratigraphic differences were used to subdivide the core into five zones.
Figure 3.8 Results of loss-on-ignition for core LB-6. The zones I through V in the diagram were based on the lithostratigraphic characteristics of the core.
The bottom of core LB-6, zone V (228-187 cm), is made of very hard clay. The basal hard clay layer with crumbly structure is topped by swamp soil and plant microfossil which are very similar to those of zones VII and V in core LB-5. The presence of traces of rootlets and plant microfossil is confirmed by LOI analysis which shows a high organic content (15-25%). A sample taken from 216 cm for AMS $^{14}$C established a date of 210 year BP (AD 1660). Based on that date and lithostratigraphic features, while taking into account the margin of error, it is likely that zone V in LB-6 corresponds with zone V of core LB-5 that has a calibrated date between ca. AD 1715 and 1750. All the other portions of core LB-6 have organic contents of 6-7%. The carbonate content in this zone has average values of 1.5-2% while the water content varies between 40% and 65%. The rest of zone V is made of sticky dark grey mud with no lamination. Fine laminations start appearing at the base of zone IV (187-139 cm) and are well defined throughout the zone. The laminations are largely made of light brown and dark brown clay bands and are consistent throughout the remainder of the core albeit with differences in breadth. The water content in zone IV have an average of 42-45% while the average for organic and carbonate content was 6.5% and 2.5%, respectively.

Zone III (139 - 110 cm) of core LB-6 is band of dark brown-grey clay. The sediments in this section have no apparent laminations, unlike the zones sandwiching it. LOI results show that the sediments in zone III have fairly uniform water (33-35%), organic (6-7%), and carbonate (3%) contents.

The sediments in zone II (110 cm-50 cm) have features that are similar to those that characterize zone III. However the former has broad laminations wider that become less apparent towards we approach 50 cm. The sediments in zone I (50-0 cm) are made
of stiff and largely light brown clayey material with crumbly structure, devoid of any form of laminations, similar to those encountered earlier at the top of core LB-5. The crumbly structure of the sediments is most pronounced upwards of 16 cm. The results of LOI for zones II-I reveals little change in the organic (6-7%), and carbonate contents (2.5-3%). However, the water content in zone II averages 55-65% while in zone one the average is 35-40%.

3.4.3 Core LB-4

The location of the coring site for core LB-4 was approximately two kilometers south of the core LB-5 site (Figure 2). The core was collected from under 213 cm of water in two segments of varying lengths, i.e. 121 cm and 91 cm for the first and second segments, respectively. The lithostratigraphic features of core LB-4 (Figure 3.9) closely resembled those of core LB-6 probably because both cores were collected from approximately the same depth. Based on the main stratigraphic features and the results of LOI analysis core LB-4 has been subdivided into five main lithostratigraphic zones viz. V (212-192 cm), IV (192-140 cm), III (140-120 cm), II (120-42 cm) and I (42-0 cm).

Zone V is characterized by presence of a layer of very hard clay containing rootlets at the bottom, which is succeeded by hard dark-grey clay. The hard basal clay has a crumbly structure but becomes soft towards the top of the zone. The results of LOI analysis show that the sediments in zone V consist of 40-45% water. The organic content in this zone is 8-10% while the carbonate content is 2.5%. Both the organic and carbonate contents do not change much throughout the rest of the core. The dark-grey clay material is succeeded by a continuous unit (zone IV) of softer clay material with fine laminations. The laminations, just like those already described in the preceding cores are
Figure 3.9  Results of loss-on-ignition for core LB-4. The zones I through V in the diagram were based on the lithostratigraphic characteristics.
made of light and dark clay bands and dominated much of the core. Besides the fine laminations, zone IV is characterized by a high water content (60-70%). The organic and carbonate contents are fairly uniform yielding values of 7-8% and 2.5%, respectively.

Zone III is made of a 20-cm-thick band of dark-brown clay that is devoid of laminations. The clay band is soft in the lower section but becomes hard and light-colored and crumbly upwards of 128 cm. Although the organic and carbonate contents remained the same as zone IV, the water content decreased to 34-40%, among the lowest recorded in the core.

Zone II is dominated by light-and dark-brown clay laminations very similar to those described earlier in zone IV. However, the laminations are wider and less apparent towards the top of the zone and disappear altogether after 42 cm to the core top (zone I). The results of LOI analysis showed that the sediments in this unit consisted of approximately 45-55% water, 7-8% organic, and less than 2% carbonate content. The sediments in top eleven centimeters of the core are light-brown with a crumbly structure devoid of laminations, and with low water content (31-37%).

**3.4.4 Core LB-3**

Core LB-3 (Figure 3.10) was collected from the southern part of Lake Baringo, approximately two and a half kilometers from the lake shores, and near the mouth of river Ewaso Nanyokie, about three and a half kilometers south of LB-4 core site. Core LB-3 is subdivided into three zones based on the lithostratigraphic characteristics. Zone III (206-162 cm) is characterized by hard dark-grey clay material that is devoid of laminations. The bottom seven centimeters in zone III have a crumbly structure. However, unlike the other cores already described there are no plant fibers or traces of rootlets at the base of
Figure 3.10  Results of loss-on-ignition for core LB-3. The zones I through III in the diagram are based on the lithostratigraphic characteristics.
the core. LOI results shows that the sediments in zone III have lower water content (40-50%) when compared to the succeeding zone. The organic content is 10-18%, the highest recorded from all the cores while the carbonate content has values below 2%.

Although the laminations in zone II (162-32 cm) appear to be no different from those in the other cores the main constituents of the sediments are soft lake mud with a higher water content (65-80%) than other cores. The organic content in this zone declined to 7-8% from a maximum of 18% in zone III while the carbonate content recorded values above 3%. In zone I (35-0 cm) the sediments are devoid of laminations and not as soft as those in zone II and appear crumbly towards the top of the core with signs of mechanical disturbance.

3.5 Stratigraphic Correlation of the Cores

The four cores from Lake Baringo, already described in the preceding section, appear to correlate well with one another when placed next to each other albeit with minor differences. Figure 3.11 shows that all the major stratigraphic transitions or zones that are represented in the four cores. It is evident from figure 3.11 that core LB-5, which is the longest, has two stratigraphic horizons which do not feature in the other cores. The two include zone VII, which was characterized by hard clay with crumbly structure and traces of rootlets and plant microfossil that have been dated to ca. AD 1650. The other is zone VI which consists of soft dark-grey muds with no laminations. Zone VI was deposited some time between AD 1650 and ca. AD 1750. The other cores were not long enough to reach the two horizons.

Zone V which had similar characteristics to those already described for zone VII, is represented in three out of the four cores. Visual examination of the cores showed that
Figure 3.11 Stratigraphic correlations of cores LB-5, LB-6, LB-4, and LB-3 from Lake Baringo. The cores are arranged from the longest to the shortest while the delineation of the different zones (1-7) is based on the lithostratigraphic characteristics of the cores.
the amount plant microfossil in zone V was scanty compared to zone VII. Unfortunately, it is not possible to measure the full extent of zone V in LB-6 and LB-4 because the entire horizon was not recovered during coring and it is not represented in LB-3.

Zone VI, which consists of soft clay with clear laminations, is represented in most of the cores and has very similar character to zone II. It is possible that the sediments in the two zones were deposited under similar climatic conditions. Zone III, which is a band of hard clay sediments devoid of laminations, features in all the cores albeit with minor differences. In cores LB-5 and LB-4, zone III is characterized by hard clay sediments devoid of laminations capped by 2-4 cm of hard crumbly sediments. In core LB-6 the crumbly structure is not as pronounced in the other cores. Core LB-3 was slightly different from all the other cores because it has more extensive zones II and III.

Zone I varies in breadth from 35 cm to 56 cm among the different cores. The sediments in this zone are light brown in color and become harder and crumbly towards the top of core. The exception is core LB-3 in which the sediments are very soft or soupy. The difference in zone I of core LB-3 could be ascribed to mechanical disturbance considering that it was collected from under shallower water that is not beyond the reach of marauding hippopotamus or crocodiles in the lake.

3.6 Interpretation and Discussion

3.6.1 Evidence of Droughts in the Stratigraphic Record

The key to the paleoenvironmental and sedimentation history of Lake Baringo is provided by the core LB-5 profile between 363 and 343 cm (zone VII) and 280-260 cm (zone V). The low water content, crumbly appearance, presence of traces of rootlets, and well preserved seeds suggests terrestrial environments, probably marsh conditions. The
sediments in these sections of the core are made of hard clay with crumb structure, which is a typical pedogenic feature. The crumb structure is a unique characteristic of *vertisols* (*black cotton soil*), often associated with flood plains, lake margins, and marsh environments (c.f., Blodgett 1985) subject to subaerial weathering processes. This interval probably results from periodic wetting and drying of the lake floor. AMS $^{14}\text{C}$ analysis of the well preserved seeds that were recovered along with the plant fibers at the bottom of core LB-5 yielded a date of $300 \pm 50$, calibrated to approximately AD 1650. That date places the peak of the first dry episode at Lake Baringo at some time before ca. AD 1650 and the second episode at ca. AD 1750-1820. The two episodes appear to coincide with the Little Ice Age (LIA) period in Europe (AD 1560-1850) that appear to have affected even the tropical regions of Africa (Driese 2004; Cohen et al. 2005; Palacios-Fest et al 2005).

A comparison of the lithostratigraphic features from the all the cores recovered from Lake Baringo (Figure 3.11) reveals that the sediments at the bottom of core LB-5 (zone V) have no equivalents in the other cores collected from the lake. Whereas most of the cores have evidence of the second drought episode only LB-5 was long enough to reach the first drought episode. It is not possible to establish the exact duration of the first dry episode that resulted in the lake drying up at the LB-5 core site in the period leading to ca. AD 1650 because the plant fibers and hard clay could not be penetrated further by the coring equipment. However, based on the thickness of the marker horizon for the second dry event (280-260 cm) it can be posited that the AD 1650 drought must have persisted for over a decade. The second dry episode lasted for between six and eight years.
Considering that core LB-5 was collected from under 352 cm of water and was 363 cm long it can be posited that at the peak of the first drought episodes at ca. AD 1650 the lake level fell by as much as -6 m. That drop in lake level and subsequent conversion of the lake surface into a marsh environment is likely to have reduced the lake surface to less than 1/3 of its present size, thereby probably confining the lake to the current northern part of the basin (Figure 3.12). Probably extensive areas in the southern part of the lake dried out and were colonized by marshes in the period leading to ca. AD 1650. The stratigraphy of core LB-5, LB-4, and LB-6 provide evidence for a second intense drought episode at Lake Baringo at ca. AD 1750-1820. All these cores have a horizon in their stratigraphy that is characterized by swamp soil preceded by stiff dark-grey clay suggesting subaerial weathering, a feature of vertisols described earlier.

Synchronicity of the Baringo record with records from most of the lakes in the region is impossible because fine-temporal-resolution is missing in the other records. Perhaps the exception is Lake Naivasha, Lake Malawi, Lake Chilwa, and Lake Tanganyika (Verschuren et al. 2000; Lamb et al. 2003; Verschuren 2004; Crossley et al. 1984). Lake Malawi’s annually-laminated record of diatom silica suggests presence of a well-defined low lake-level period between ca. AD 1570 and 1850 (Johnson et al. 2004; Owen et al. 1990). The duration of the Lake Malawi lowstand appears to be part of a gradual but extended period of aridity that encompasses the two distinct dry episodes in the Lake Baringo record. Considering that Lake Malawi’s diatom inferred record shows the most pronounced lowstand between AD 1780 and AD 1830 (c.f., Johnson et al. 2004; Verschuren 2004; Owen et al. 1990) it may be reasonable to assume the possibility of two or more decadal scale intense dry episodes within the extensive period of aridity between
Figure 3.12  Map of Lake Baringo showing the extent of the lake at the peak of the first drought episode at ca. AD 1650. Much of the lake was colonized by marsh environments.
ca. AD 1570 and 1850. Crossley et al. (1984) put the peak of the aridity in the Lake Malawi record between the late 1700s and early 1800s which is consistent with the second drying episode at Lake Baringo. At the peak of Lake Malawi’s drying episode large tropical trees grew in what is now part of the lake surface (c.f., Crossley et al. 1984). Lake Chilwa, situated at the southern end of the Gregorian rift valley and to the south of Lake Malawi also records an extended lowstand some time between AD 1750 and 1850 (Crossley et al. 1984; Lancaster 1981).

Lake-level and salinity record from Lake Naivasha shows what must have been a period of intense drought around AD 1560-1625 (Verschuren et al. 2000). Although the timing of the AD 1560-1625 Lake Naivasha drought episode is consistent with the first intense drought episode in the Lake Baringo record that ended in ca. AD 1650 there is neither written document nor reliable archaeological survey from that locality with which to compare. A survey of an archaeological find of a burial complex at Kampi ya Samaki (on the western shore of Lake Baringo) was not conclusive and has no established date (c.f. Robertshaw 1984). Also, documentary records of climate data from the region are rare and instrumental records where available are only limited to the last 120 years in spite of the region’s long and rich cultural history (Nicholison 2001; Nicholson et al. 2001). The AD 1650 intense aridity episode in Lake Baringo which corresponds to the Naivasha lake-level record is generally coeval with pre-colonial history of East Africa that records drought-induced famine (*Nyarubanga drought*) and socio-political unrests that match the periods of lowstand at Lake Naivasha (Verschuren 2004; Verschuren et al. 2000). Anderson (2002) recounts from oral tradition that droughts in the Baringo region
were often accompanied by famine, livestock diseases, and locust invasions which adversely affected local populations.

The two episodes of aridity in the Lake Baringo record were evidently more severe than any drought recorded in the twentieth century. Even the much remembered series of droughts that affected the Baringo ecosystem in the 1920s and 1930s (c.f., Anderson 2002; Kettel 1984) did not result in the drying up of the lake. It must have taken a combination of low or failure of precipitation, and/or high evapotranspiration within the drainage basin for significant portions of Lake Baringo’s surface area to dry up.

The drying up of Lake Baringo is not without precedent in the region. Longer paleorecords from other African lakes suggest the presence of even more intense periods of desiccation at different times during the Holocene that led to the lakes drying up (Talbot and Livingstone 1989; Riehl and Meitin 1979; Talbot and Delibrias 1977). Geochemical evidence and lake level records indicate that Lake Rukwa, southern Tanzania, has dried out for significant periods on at least three occasions since 4000-3000 yr BP (Talbot and Livingstone 1989). The first drying out at Lake Rukwa occurred some time prior to 3000 BP while the latest episode, whose exact date remains unknown, occurred during sub-historical times that are contemporaneous with the ca. AD 1750-1850 lowstand in Lake Malawi (Talbot and Livingstone 1989; Crossley et al. 1984) which is coincident with the Lake Baringo record.

The lithostratigraphy of core LB-5 (zone VI) reveals what we interpret as a brief period of high precipitation after ca. AD 1650 and before the second period of intense drought that ended around 1850. This section of the core is characterized by soft dark-
grey mud that is devoid of any horizontal lamination that characterize much of the rest of the core. The amount of precipitation during that time period was probably higher than any other period covered by core LB-5 and the other cores. It is likely that the first intense drought episode was succeeded by a period of heavy precipitation that was probably accompanied by flooding. Presence of well preserved seeds in zone V may suggest that the material may have been buried rapidly resulting in the excellent preservation. The Lake Chilwa record shows evidence of lake level peak (9 m above modern mean level) some time between AD 1650 and 1750 (Crossley et al. 1984).

Applying modern evapotranspiration rates to the AD 1650-1750 peak at Lake Chilwa would translate to approximately 35-40% increase in precipitation (Lancaster 1981). Increase in precipitation around that period is consistent with tree ring record obtained from baobab trees at Nchisi Island near the western shore of Lake Chilwa (Crossley et al. 1984). Unfortunately most of the paleorecords from the East African region have low temporal resolution and/or have poor chronological control so that centennial to decadal scale events are not recorded (c.f., Verschuren 2004). Very few climate- and anthropogenic-proxy records have good enough centennial- or decadal-scale resolution and age control that cover the last 500 years. Therefore apart from the information gained from oral tradition East Africa’s paleorecord of the last five centuries is rather blank.

After the second drying out episode at Lake Baringo at ca. AD 1750 the climate of the region appears to have stabilized to modern conditions characterized by high seasonal variability in precipitation. This is reflected in the stratigraphic features in nearly all the cores characterized by a dominance of horizontal laminations made of dark-
and light-brown layers of varying thickness. The laminations (light-and dark-brown/gray couplets) are not “varves”. Varves represent annual couplets. An investigation that involved counting of the laminations in the Lake Baringo cores showed that they were not annual. We hypothesize that the light and dark couplets correspond to alternating wet and dry periods on an interannual basis. It is possible that the laminations are related to the precipitation character that could be related to the El Nino/South Oscillation. In zone IV (260-163 cm) of core LB-5 the laminations were finer in the lower portion of the core and become less defined towards the top of the core. A gradual transition in the sediments to course light-brown mud devoid of lamination occurred between 163 and 130 cm (zone III). The transition in the sediments begun in the late AD 1800s and ended before AD 1900. This pattern is reflected in all the cores. We interpret these changes in the stratigraphy to be indicative of a decrease in lake-level related largely to climate-driven shifts in the pattern of precipitation from fairly wet to arid conditions with the consequent decrease in river discharge. The resultant shallow lake conditions broke the thermal stratification in the lake; permitting resuspension of sediments and complete mixing, thereby terminating the lamination process.

The sediment transition that occurred before AD 1900 can be correlated with the pre-European history of the Baringo region as recounted from oral tradition. After the second intense drought episode that ended around AD 1750 the Lake Baringo area, especially the Il Chamus (also known as Njemps) villages to the immediate south of the lake experienced a continuous era of prosperity. During the era of prosperity the Il Chamus villages became synonymous with fortified settlements, sophisticated networks...
of irrigation, and agricultural success during an era of Maasai pastoralist expansion in the rift valley (Spencer 1998; Little 1992; Anderson 1989).

By AD 1850 the Lake Baringo area was well known to the Arab traders and early European explorers as an area rich in ivory and an important source of food provisions for trade caravans moving to the north and west from the East African coast (Anderson 2002; Ambler 1988; Gregory 1896; Thomson 1887). However, European missionaries and explorers during the AD 1880s found Baringo in the middle of a serious drought. For instance Joseph Thomson could not find food for his caravan in the Njemps villages in AD 1883. The irrigation networks had collapsed following a succession of years with no rainfall, and the local people (the Njemps) were largely dependent on hunting and gathering wild roots and berries for food (Thomson 1887). Another European expedition led by Gregory in AD 1893 also recounted experiences of drought and famine in Baringo which had not received rains for two consecutive years and the area was characterized by dry riverbeds (Huxley 1935; Gregory 1896). The pre-colonial histories of the Baringo area recounted from oral tradition testify that the late 19th century experienced a series of droughts and famines, as well as violent contacts between pastoralists and agriculturalists (Ambler 1998). All these accounts attest to the presence of droughts in the region beginning ca. AD 1880 that led to a severe lake-level reduction in Baringo. However, it is likely that the severity of this drought event was more localized and less intense compared to the ca. AD 1650 and 1750 droughts that resulted in the drying out of the lake because there is no sedimentary record of it in the other lakes from the East African region.
That cessation of laminations in the Lake Baringo sediments is consistent with reduction in lake level is not in doubt. This is supported by the fact that the more recent sediments in the lake (the top portions of the cores) have no laminations. All the cores show that the top 20-50 cm, which represents sediments deposited over the past 40-50 years, are missing laminations that resumed after the AD 1880-1900 series of droughts. This change in stratigraphy is attributed to a combination of low lake levels due to the present high evapotranspiration rates in the region and decreasing precipitation as well as increased land degradation. The mean depth of the lake was approximately 7.5 m in 1932 (Beadle 1932), 5.6 m in the 1960s (Ssentongo 1995), and had declined to 2.5 m by 2004 (Hickley et al. 2004). The present shallow lake conditions limit thermal stratification; encourage mixing, and frequent resuspension of sediments, which coupled with the inflow of high sediment load from heavy erosion in the catchment have caused lamination to cease in the lake.

### 3.6.2 High Sedimentation Rates

One of the most conspicuous features in the Lake Baringo record is the high sedimentation rates (0.84 - 1 cm/yr). Whereas most of the major climate effects associated with intense droughts in the Lake Baringo area are manifested as near-synchronous events in other lake records from the East African region, the sedimentation rates tend to vary from one locality to the other (Einsele and Hinderer 1998; Cohen 1989). The Lake Baringo record showed that between AD 1650 (calibrated from 300 ± 50 AMS $^{14}$C date) and 2005 approximately 360 cm of sediments had accumulated at the LB-5 core site, which translates to a sedimentation rate of approximately 1 cm/year. Seven excess $^{210}$Pb-based dates that were sampled from the top 100 cm of core LB-5
provided a sedimentation rate of 0.84 cm/year. Another set of seven excess $^{210}$Pb-based
dates from the top of core LB-7, eight kilometers to the north of LB-5 core site, gave a
sedimentation rate of 0.83 cm/yr. Thus sedimentation rates based on $^{210}$Pb chronology
from the two cores compare well with the AMS $^{14}$C based rate, both showing a high
sedimentation rate over the past ca. 350 calendar years. Assuming uniform sedimentation
rates of 1 cm/yr throughout the lake and lake surface area of 129 km$^2$ we roughly
estimate that over 1290000 m$^3$ of sediments is washed into Lake Baringo every year.
This indicates that if all factors are constant and the sedimentation rates continue
unchecked Lake Baringo will fill up with sediments within 250 years. These estimates
are based on the current mean depth of 2.5 m. However, more dating control is needed
for these rates to be ascertained.

Although many water bodies in East Africa experience high sedimentation rates,
those from Lake Baringo are certainly the highest in Africa (Table 3.1). When compared
to lakes of comparable size the enormity the sedimentation rates at Lake Baringo
becomes obvious. The high sedimentation record in the lake could be explained in terms
of environment of deposition, especially high rates of land degradation in the catchment
areas, relief, climate conditions, lithology, and anthropogenic factors.

Surrounded by a landscape that features high relief, steep slopes, and low
vegetation, we hypothesize that transport of solid matter by inflowing rivers is probably
the most important source of the sediments in Lake Baringo. Indeed, the extreme
turbidity that has come to characterize Lake Baringo is largely due to soil erosion in its
catchment (Plate 3.1). Soil erosion has been exacerbated by deforestation and
overgrazing, both of which are linked to increasing human and livestock populations.
Table 3.1: Sedimentation rates in select African lakes

<table>
<thead>
<tr>
<th>Lake</th>
<th>Location</th>
<th>Surface Area (km²)</th>
<th>Sedimentation rate (cm/year)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Albert, Uganda</td>
<td>0°40’ S, 29°50’ E</td>
<td>2325</td>
<td>0.14</td>
<td>Russell et al. 2003; Beuning et al. (1997)</td>
</tr>
<tr>
<td>Naivasha, Kenya</td>
<td>0°45’ S, 36°20’ E</td>
<td>130</td>
<td>0.12</td>
<td>Mworia-Maitima (1991)</td>
</tr>
<tr>
<td>CIC, Naivasha</td>
<td>0.75° S, 36.37° E</td>
<td>6400</td>
<td>0.66</td>
<td>Lamb et al. (2003)</td>
</tr>
<tr>
<td>Malawi, Malawi</td>
<td>10°16’ S, 34°19’ E</td>
<td></td>
<td>0.02-0.05</td>
<td>Johnson et al. 2002; (2004)</td>
</tr>
<tr>
<td>Cheshi, Zambia</td>
<td>9°05’ S, 29°45’ E</td>
<td>100</td>
<td>0.02-0.07</td>
<td>Stager (1987)</td>
</tr>
<tr>
<td>Rukwa, Tanzania</td>
<td>8°00’ S, 32°25’ E</td>
<td>2600</td>
<td>0.27</td>
<td>Talbot and Livingstone (1989)</td>
</tr>
<tr>
<td>Victoria, East</td>
<td>0°18’ N, 33°20’ E</td>
<td>69000</td>
<td>0.06-0.17</td>
<td>Stager et al. (1997); Stager et al. (2003)</td>
</tr>
<tr>
<td>Baringo, Kenya</td>
<td>0°30’ N - 0°41’ N,</td>
<td>129</td>
<td>0.84-1</td>
<td>Kiage and Liu (in this study)</td>
</tr>
<tr>
<td></td>
<td>36°00’ E - 36°10’ E</td>
<td></td>
<td></td>
<td>Muhammed et al. 1995; Halfman et al. (1994)</td>
</tr>
<tr>
<td>Turkana, Kenya</td>
<td>4°00’ N, 36°05’ E</td>
<td>7560</td>
<td>0.3-0.8</td>
<td></td>
</tr>
<tr>
<td>Lake Tanganyika,</td>
<td>3° 20’ - 8° 48’ S</td>
<td>32600</td>
<td>0.01</td>
<td>Scholz et al. (2003)</td>
</tr>
<tr>
<td>Tanzania</td>
<td>29° 5’ - 31° 15’ E</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Plate 3.1 (a) Picture of gully erosion at Njemps flats near Lake Baringo (b) The base of one of the tributaries of Molo River showing sediments that are transported to lake Baringo.
Figure 3.13  Population of Baringo District in Rift Valley, Kenya between 1915 and 2004. Between 1915 and 1940 the population size was estimated by British colonial government based on hut tax. The first census was performed in 1948 (source: Kenya Ministry of Planning and Development, Central Bureau of Statistics 1989, 1994, 1997, 2004)
Figure 3.14 Livestock population of Baringo District in Rift Valley, Kenya between 1910 and 1995. Estimates for 2003 are not included (source: Kenya Ministry of Planning and Development, Central Bureau of Statistics 2002)
The population of Baringo District now stands at 302,245; it quadrupled between 1948 and 2004 (Figure 3.13) and is growing at the rate of 2.65 % per annum (Kenya Ministry of Planning and National Development 2005). The livestock numbers are correspondingly large; approximately 900,000 goats; 200,000 sheep, and 300,000 cattle in 2003 (Figure 3.14) (c.f., Hickley et al. 2004). Such a remarkable increase in human and livestock numbers has put immense pressure on the fragile vegetation cover in the catchment making land degradation inevitable.

The ecological effects of the high sedimentation rates at the lake are already being felt. Hickley et al. (2004) observed that increased deforestation and soil erosion in the Lake Baringo catchment and the consequent high turbidity in the lake have led to near extinction of submerged macrophytes. The result is a lakebed virtually devoid of benthic fauna. (Hickley et al. 2004; Aloo 2002) Over time the impacts of the ecological changes on the local population whose sustenance is dependent on the lake could be catastrophic.

The high sedimentation rates in Lake Baringo are accentuated by the steep slopes that constitute over 46% of the Baringo landscape (c.f., Hickley 2004). The slope of the land and the soil properties often combine to enhance the erodibility (i.e., soil resistance to detachment and transport) of any given landscape (Vrieling 2006; Lal 2001; Morgan 1995). In the Lake Baringo area land degradation is compounded by the soil properties of clay, clay loams, and silty clay that prominently feature in the area (Hickley 2004; Johansson and Svensson 2002). Clay soils have high erodibility because they seal soil pores, reduce infiltration, and encourage large surface runoff, thereby increasing the possibility of high rates of erosion (c.f., Lal 2001; Morgan 1995).
The climate of Baringo, especially the rainfall pattern, compounds the land degradation problem. The amount of rainfall received in the area is low (less than 600 mm per annum) and erratic, falling in heavy downpours within a few days in April/May/June and October/November (c.f., Johansson and Svensson 2002). Low rainfall can only support sparse vegetation cover in a region that experiences high evapotranspiration rates (<2500 mm per year) while the sporadic nature of the rainstorms has high erosive power on bare grounds or low vegetation cover. There is close relationship between vegetation density and soil loss especially in environments that are characterized by abrupt rainstorms. A study by Snelder and Bryan (1995) in Njemps flats, adjacent to Lake Baringo, recorded soil loss values of over 80 g m\(^{-2}\) for 30 minute storms and 30 g m\(^{-2}\) for 60 minute storms during single rainstorm events. Thus, storm duration and intensity greatly influence soil erosion rates. The rainfall pattern in Baringo appears to encourage soil erosion even in the context of limited anthropogenic influence. All these factors- mounting population pressure, relief features, soil properties, and the climate- have interacted in such a way as to render the Lake Baringo area highly susceptible to high denudation rates. It is, therefore, no wonder that almost 90% of the catchment (6820 km\(^2\)) is considered degraded (c.f., Onyando et al. 2005; Hickley et al. 2004; Johansson and Svensson 2002).

However, it is somewhat surprising that the Lake Baringo record shows no change in the high sedimentation rates for both the prehistoric period, prior to European settlement in the Baringo ecosystem (AD 1900), and after, including more recent times. This study expected to find a significant difference in sedimentation rates between the two time periods based on the widely held view that land degradation followed European
settlement and colonization. During the colonial period the European settlers deprived
the indigenous pastoralists of their dry season grazing lands by occupying, and fencing
off the well-watered highlands to the south and east of Lake Baringo (c.f., Anderson
2002), thereby confining the local pastoralists to the climatically harsh and ecologically
fragile lowlands. Before the colonial period the Baringo lowlands were part of a wider
production system, developed out of adaptation to the variable semi-arid conditions, that
involved transitory use of grazing resources and was attuned to the fragile semi-arid
environment, which ensured that drought generated damage was minimized. By taking
possession of the well-watered highlands the European settler community severed the
seasonal grazing areas and watering points from the African herders and condemned
them to the permanent use of the fragile lowlands, which quickly became degraded (c.f.,
individuals took possession of previously held European lands and maintained the status
quo. Thus, land degradation came to be identified with the Baringo lowlands.

There are two possible explanations for the high pre- and post-colonial or modern
sedimentation record at Lake Baringo. First, it is possible that land degradation in the
Lake Baringo area commenced much earlier than it has been previously documented, and
the rates of pre-historic land degradation are similar to those experienced in modern time.
If that assertion is true then the romanticized hypothesis of the invariably
environmentally friendly use of grazing resources by the indigenous pastoralists and their
indigenous knowledge of conservation will need to be revised. Although the pre-colonial
population in the Lake Baringo area was much lower compared to the present, their effect
on the landscape was as profound as that witnessed in modern times. Thus, the
indigenous pastoralists and farmers were not simply victims of colonialism but were always active agents and shapers of the degraded landscape presently evident in the region.

The second explanation for the high and unchanged sedimentation rates at Lake Baringo is that humans may not be the main drivers of land degradation in the area, and the high sedimentation rates at the lake. This hypothesis is reasonable considering that change in population and increased use of vegetation resources in the Lake Baringo area over time has not affected the sedimentation rates. It is likely that the steep landscape, sparse vegetation cover, clay and volcanlastic material, and the climate of the Lake Baringo area interact to yield the high sedimentation rates independent of anthropogenic impacts. Indeed the soil properties of clay, clay loams and silty clay in the context of steep slopes, sparse vegetation, and heavy sporadic rainstorms may hold the key to understanding the high sedimentation record in Lake Baringo. Einsele and Hinderer (1998) have shown from studies elsewhere in East Africa that areas with crystalline rocks and denser vegetation (e.g., around Lakes Tanganyika and Kivu) experience lower sedimentation rates compared to areas with volcanlastic material and sparse cover (e.g., Lake Turkana).

The Lake Turkana region is similar to Lake Baringo in many ways (c.f., Muhammed et al. 1995; Halfman et al. 1994); however the latter has more steep landscapes, high relief, and experiences slightly higher precipitation. Therefore, although both lakes are characterized by relatively high sedimentation rates, those for Lake Baringo are much higher (Table 3.1) and the difference can be accounted for by features lacking in the Lake Turkana landscape. That underscores the significance of high relief,
steep landscapes, soil properties, and climate in heightening denudation and subsequent high sedimentation rates in Lake Baringo.

3.7 Conclusions

A number of findings have emerged in this Chapter that have increased our understanding of paleoenvironmental changes in East Africa while concurrently challenging previously held views. During the past 350 years the climate of East Africa has experienced variability at decadal-to centennial scales. The stratigraphic record from Lake Baringo has revealed presence of two abrupt and severe dry episodes at ca. AD 1650 and AD 1750 in East Africa that led to drying up of the lake. At the peak of the first and second drought episodes the lake level fell by as much as -6 m at the LB-5 core site. The two abrupt intense dry episodes were punctuated by a period of heavy precipitation that is documented in other high-temporal-resolution records from East Africa. The Lake Baringo record shows evidence of a third period of desiccation at ca. AD 1880 which resulted in lowering of lake level and which is corroborated by oral tradition from the area.

The Lake Baringo record documents sedimentation rates of 0.84-1 cm/yr, one of the highest in Africa. These sedimentation rates are attributed to high rates of soil erosion and land degradation within the Lake Baringo drainage basin. Presently over 1290000 m$^3$ of sediments is washed into Lake Baringo every year, which if not checked will see the lake filled with sediments in the next 250 years. The effects of the high sedimentation are evident in the near extinction of submerged macrophytes and benthic fauna in the lake. The Lake Baringo record seems to shows no difference in the sedimentation rates for the pre-colonial and post colonial period. However, more dating
control is required in order to confirm this finding which is presently based on just two AMS $^{14}$C dates and seven $^{210}$Pb dates. If confirmed, this finding would challenge the current understanding of the assumed origins and causes of land degradation and soil erosion in the Lake Baringo area. Although human activities exacerbate the land degradation process in the area they only serve as a catalyst. The main forces behind the land degradation in the ecosystem lie in the relief features, soil properties, and the nature of rainfall experienced in Lake Baringo. The unchanged sedimentation rates for pre-colonial and post-colonial periods also implies that indigenous pastoralists and farmers were not simply victims of colonialism but active agents and shapers of the degraded landscapes in the ecosystem.

3.8 References


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CHAPTER 4. PALYNOLOGICAL EVIDENCE OF CLIMATE CHANGE AND LAND DEGRADATION IN THE LAKE BARINGO AREA, KENYA, EAST AFRICA, OVER THE PAST 350 YEARS

4.1 Introduction

Palynological records from the African tropics are key to understanding past changes in the climate system as well as the sensitivity of tropical regions to present and future climate changes. Unfortunately few high-temporal-resolution records of climate and vegetation are available from the tropics, yet changes in this region have strong influence on the global climate system (Thompson et al. 2002; Thompson et al. 1998; Barker et al. 2001). Lake Baringo (Figure 3.1) is well positioned geographically to archive both local and regional environmental changes in the tropics. Therefore, biological proxies from the lake can provide continuous records of vegetational and climate history for the East African region. Investigating the history of vegetational change using proxy data such as pollen, spores, microscopic charcoal, and other microfossils is useful for analyzing the trends in land cover, climate, human impacts, and water resources through time.

In particular, quantitative analyses of fossil pollen and microscopic charcoal assemblages in lake sediments provide an excellent basis for investigating the timing and nature of shifts in an ecosystem brought about by climate and/or human impact (e.g., Cohen et al. 2005; Msaky et al. 2005; Salzmann 2000, 2002; Wooller et al. 2000; Burney 1995; Gasse 2000; DeMenocal et al. 2000; Maley and Brenac, 1998; Edwards and MacDonald 1991; Hamilton 1982; Hamilton et al. 1986). Pollen is considered excellent for paleoenvironmental reconstruction for a number of reasons. First, pollen in lake sediments provides a record of past vegetation, and vegetation is acknowledged as an
accurate indicator of past climatic environment of an area (c.f., Gasse 2000; Maley and Brenac 1998; Fredlund and Tieszen 1994). Second, sites and media from which pollen could be obtained can be found in different regions (Birks and Birks 1985). Third, the pollen is sensitive to environmental variables and can be obtained in sedimentary sequences, which allows for reconstruction of vegetation (and climate) change over time. Fourth, pollen can be well dispersed, and this enables both local and regional vegetation to be represented.

The interpretation of pollen assemblages and other biological proxies found in lake sediment cores is largely based on uniformitarian principles that were first formulated by James Hutton in 1788, but more fully defined and discussed by Charles Lyell in 1830. Birks and Birks (1980) note that the philosophical foundation of uniformitarianism can be stated as “the present is the key to the past”. The effectiveness of pollen as a paleoenvironmental proxy depends on the following assumptions:

a) The modern ecology of plants represented in the pollen assemblage is well understood.

b) The present plant distributions are in equilibrium with their controlling variables.

c) Former plant distributions were in equilibrium with their controlling variables.

d) The former plants have modern analogues.

e) The ecological tolerance and preference of plants have remained unchanged.

f) The fossil assemblage is a true representation of the death assemblage.

g) The taphonomy of pollen in the sediments can be established.

h) Fossil remains can be identified to low taxonomic levels.
The palynological record like all fossil data has certain limitations. First, many useful details may be missing, perhaps never to be found. Second, sequential events may be collapsed together, inverted, or mixed with evidence from other times and generally very few relevant parameters can be measured directly (Burney, 1999). Despite these limitations palynological records still provide reliable records for reconstructing paleovegetation and paleoenvironments.

This chapter presents the history of vegetational changes and land degradation in the Lake Baringo area, Kenya, East Africa, during the late Holocene through the examination of fossil pollen, spores, and microscopic charcoal in lake-sediment cores. It complements the findings based on the lithostratigraphic record that are presented in chapter three. The high-temporal-resolution paleoecological data presented here are derived from a core (LB-5) collected from Lake Baringo. Palynology is used to investigate the role of natural versus anthropogenic causes for environmental changes in the Lake Baringo area during the late Holocene. This investigation was conducted with the realization that studying human-vegetation and climate interactions in East Africa can be a daunting task. This is because humans and their ancestors have probably interacted with east African landscapes for longer than any other ecosystems in the world (Burney, 1996a). That is compounded by the fact that tropical climates including those of East Africa are not as stable as previously assumed; they fluctuate on various time scales e.g., interannual, decadal, centennial, and millennial (Verschuren 2004; Verschuren et al. 2000; Lamb et al. 2003; Gasse 2002; Marchant and Taylor 1998; Crossley et al. 1984; Talbot and Delibrias 1980). However, the use of multiple proxies permits a more accurate reconstruction of the paleoenvironment (c.f., Boyd and Hall, 1998).
4.2 Background

Fossil pollen and other palynomorphs are ubiquitous components of many lake sediment cores and are excellent for tracking paleoenvironmental changes because they serve as a link to past vegetation and human activities. In East Africa where documentary and instrumental records of climate and human activities are limited to the past 120 years palynology provides the most reliable means of reconstructing the paleoenvironment. Unfortunately the published records (e.g., Ryner et al. 2006; Beuning et al. 1997; Stager et al., 1997; Street-Perrott et al., 1997; Mohammed et al. 1995; Olago 1995; Taylor 1990; Maitima 1991; Vincens 1989, 1993; Bonnefille and Riollet 1988; Stager 1988; Livingstone 1967, 1962; Bakker 1964; Coetzee 1964, 1967) from the region have mostly focused on reconstructions over long geologic time spans. Very few data on centennial-and/or decadal time scales are available for the Holocene period, especially the past three thousand years when humans became a major force in shaping many ecosystems including remote and inhospitable environments (c.f., Burney 1997; 1993; 1987; Hamilton et al 1986; Hamilton 1982).

Generally there is paucity of high-temporal-resolution paleoecological studies in East Africa. Of those available, only a few studies have focused on vegetation and climate change and the role humans play in shaping past and present landscapes (Cohen et al 2005; Msaky et al 2005; Verschuren et al 2000, 2004; Lamb et al 2003; Hamilton et al 1986). However, during the Late Holocene humans have significantly impacted the vegetation of East Africa through the expansion of agriculture, pastoralism, and fire; in some cases severely degrading ecosystems. A study by Hamilton et al. (1986) based on examination of pollen and microscopic charcoal provided evidence for agricultural
development, punctuated by varying episodes of deforestation in East Africa as early as 4800 yr BP. Verschuren et al (2000) showed that during the past millennium the climate of East Africa has alternated between contrasting moisture conditions over decadal- and centennial time-scales. Throughout the millennium, human impact on the vegetation tended to mirror the prevailing climatic conditions, with human impact increasing with time (Lamb et al. 2003). Considering that rapid vegetation changes due to human activities could result in changes in climate (c.f., Kutzbach et al., 1996; Doherty et al., 2000), the importance of investigating the history of vegetational change in East Africa cannot be overemphasized. In the Lake Baringo record we focus on the climate, vegetational change, and human impact over the last 350 years.

The Geographic location and physical environment of Lake Baringo (Figure 3.1) has already been discussed in detail in section 3.2. However, details on vegetation around the lake need to be highlighted. The littoral vegetation community around much of the Lake is very poorly developed. Although the lake is located within the vegetation zone classified by White (1983) as deciduous bushland and thicket (drier savanna types), (Figure 2.4) its plant communities, away from the littoral zone, are highly heterogeneous with high species diversity and significant variations in physiognomic characteristics along the elevational gradient.

The present vegetation in the immediate surrounding of the lake is dominated by several Acacia and Commiphora species with little undergrowth (Plate 4.1), but as the elevation increases a wide variety of woody plant species combine with acacias to form the main vegetation layer with an under-storey vegetation of moderate to dense perennial forbs and grasses. The natural vegetation on the highlands that constitute the lake
catchment is characterized by a diverse woodland consisting of Celtis spp., Urticaceae, Myrtaceae, *Croton, Holoptelea, Prunus, Podocarpus, Acacia*, and *Olea* among others. Some papyrus (*Cyperus papyrus*) and *Typha* grow in the mouths of rivers, and especially in the wetlands in the southern part of the lake. Patches of remnant natural forests are found within the hills and along perennial water bodies.

**4.3 Materials and Methods**

Over two field seasons in January 2004 and 2005, thirteen lake sediment cores were collected from Lake Baringo from different water depths throughout the lake. The cores were collected using a modified Livingstone corer (c.f., Wright et al. 1984) from a platform mounted between two boats. The corer was constituted of a rubber piston that was fitted tightly inside a 50 mm-transparent polyvinyl chloride (PVC) pipe that had a wall thickness of about 5 mm. The piston was attached to a narrow cable that ran through a head piece attached at the top of the PVC pipe. The head piece was made in a way that allowed for attachment of extension rods and also makes room for movement of water out of the top of the tube during coring. The number of the extension rods (152 cm) used was dependent on the depth from which the individual cores or core segments were collected. A sharp-toothed, cylindrical steel cutting shoe was attached to the bottom end of the 150 cm long PVC pipe to aid the penetration of the corer through the sediments.

The coring process involved pushing the core tube into the sediments while the piston cable was securely tied off. The pushing of the tube into the sediments ceased whenever the piston jammed against the core head piece indicating the filling up of the tube with sediments or when further pushing was impossible due to hardness of the sediments. During retrieval of the cores, the core tube and piston cable were removed.
Plate 4.1 Typical vegetation of the lowland plains in the vicinity of Lake Baringo dominated by Acacia species with little or no undergrowth.
together. The piston provides suction that hold the sediments intact inside the tube. The cores were then carefully labeled, described, and the core-tube ends tightly sealed using a rubber stopper and a PVC cap secured with duct tape to preserve the moisture and keep the core intact. The cores were then transported to Louisiana State University Biogeography and Quaternary Paleoecology Laboratory for processing and analysis. In addition to raising cores from the lake, 24 surface soil samples were collected from different environments in the vicinity of Lake Baringo. These surface samples consisted of soils from different landscapes, such as livestock (goat, sheep, and cattle) enclosures (including dung), savanna bushland/woodland, upland forests, riverbeds, and farmland. The pollen, spores, and microscopic charcoal examined in this study were processed from Core LB-5. The core was raised from Lake Baringo on January 5, 2005 from under 2.5 m of water and was preserved in the transparent PVC tube in which it was collected. Three samples were sent to Beta Analytic Inc., for $^{14}$C analysis to establish the chronology of the sediments. Chronology for the upper section of the core was established through $^{210}$Pb analysis, at Dr. Sam Bentley’s Laboratory, Coastal Studies Institute, Louisiana State University. Sediment samples for pollen analysis were taken at 5 cm intervals throughout the core. The sediments were chemically treated to concentrate pollen and microscopic charcoal following the standard procedure of Faegri and Iversen (1989): dissolving carbonates and silicates with dilute HCl (10%) and cold HF (70%), respectively. This was followed by removal of colloidal silica with warm diluted HCl, and removal of humic acids by dilution in KOH (10%) solution. The residue obtained was then suspended in silicone oil. Before the chemical treatment two tablets containing
a known quantity of exotic spores of *Lycopodium* sp. were added to each sample to aid in the calculations of pollen concentration and influx (c.f., Stockmarr 1971).

After the samples were processed and mounted onto slides, counts were made of pollen types, spores, and microscopic charcoal abundance at each level. Viewing and counting was done on a Nikon HFX-II compound microscope mostly at 400X magnification. For each sample counting ceased when at least 300 identifiable pollen, fungal spores, and pteridophyte spores or 1000 *Lycopodium* spores were counted. The identifications were based on the reference collection at the Louisiana State University Biogeography and Quaternary Paleoecology Laboratory, and on specialized publications relevant to East Africa’s pollen morphology (Bonnefille 1971a, b; Hamilton 1976; Bonnefille and Riollet 1980). Other pollen references that proved useful for pollen identification included publications of pollen from the tropical regions (e.g., Willard et al. 2004; Heusser 1971). Publications by van Geel (2001; 1986) and van Geel et al. (2003) were heavily relied upon for fungal spore identification. The total count included unknown types (well preserved but unmatched to any in the reference type), which were included in the pollen sum for the calculation of percentages. The counts were recorded by hand and then entered into a spreadsheet, sorted, and plotted using Tilia and Tilia-Graph software (Grimm 2004).

### 4.4 Results and Discussion

Pollen in core LB-5 was generally well preserved except for a few stratigraphic sections, namely 363–343 cm, and 295-260 cm. These sections of poor preservation corresponded with periods of low lake levels already discussed in Chapter 3. Totally 105 pollen and spore taxa were identified. The Appendix provides a description and
photographs of the major pollen types found. Pollen profiles (percentages) of principal pollen taxa in core LB-5 are presented in Figure 4.1. Overall, aquatic and emergent marsh taxa were generally rare, despite the shallowness of the lake, reflecting very limited development of marshland or littoral vegetation. Gramineae pollen dominated the spectra with percentages averaging 20-40%, typical of regional African pollen assemblages (c.f., Msaky et al. 2005). Low percentages of arboreal pollen (less than 20%) were observed throughout the core. Visual inspection of the lithostratigraphic signature, discussed in the previous chapter, was used in conjunction with pollen percentage curves to divide the sequence into five zones (I-V).

Although the pollen abundance is high in zone V (363 – 343 cm), the preservation was not good; many pollen and spores were corroded and difficult to identify. Despite the poor preservation, zone V still registered high frequencies of Gramineae, Cyperaceae, Combretaceae, Podocarpus, Olea, Tarchonanthus, Phyllanthus, Acacia, and Euclea. Other well represented pollen types in the zone include Artemisia, Hypoestes, and Balanitaceae. Microscopic charcoal counts were lower in this zone than in succeeding zones. Figure 4.2 shows that zone V had low counts (<100) of fungal and fern spores and other spores of terrigenous origin. Fungal spores of Sordaria-type, Cercophora type, and Glomus type accounted for over 5% of the sum. Polypodiaceae, and other Pteridophytes (including indeterminable or unknown types grouped under the category “others”), recorded 20% of the count.

The pollen in zone IV (343-295 cm) were well preserved compared to zone V although the pollen sum was higher. There was a general increase in the percentages of Macaranga, Phoenix, and Gramineae. The pollen of Acacia, Sesbania,
Figure 4.1  Pollen percentage diagram showing the results of the analysis based on core LB-5 from Lake Baringo over the past 350 calendar years.
Figure 4.2 Fungal and Fern spore diagram based on analysis of sediment samples from Lake Baringo.
*Euclea*, and *Phoenix* showed an increase while *Olea, Tarchonanthus, Phyllanthus*, and Combretaceae decreased. The microscopic charcoal count increased to an average of 100, double that in the preceding zone. All fungal spore taxa recorded significant increases in percentage with *Sordaria* type, *Spormiella* type, and *Cercophora* registering the highest increases.

In zone III (295-260 cm) there was an increase in the percentages of *Acacia, Euclea, Balanitaceae, Chenopodiaceae and Amaranthaceae, Typhaceae, Justicia, and Phyllanthus*. Although the zone recorded a general decrease in pollen sum there was an abrupt and significant decrease in the percentages of Gramineae and Cyperaceae pollen from > 20% to < 2%. This zone did not record much change in the frequency of microscopic charcoal compared to zone IV. The pattern of fungal spore assemblage in zone III was similar to that displayed by pollen, characterized by significant decreases in the percentages of all major taxa. However, there was a significant increase in the frequency of the unknown or indeterminable spores.

Zone II (260-35 cm) shows a recovery of Gramineae and Cyperaceae pollen, especially at ca. 260-200 cm, and after 150 cm. However at 200-150 cm there was a decrease in the percentages of Gramineae and Cyperaceae to levels similar to those observed in zone III. This zone showed a gradual increase in the percentages of *Podocarpus, Juniperus, Combretaceae and Acacia* albeit with minor oscillations. The percentages of most of the other pollen types were also marked by oscillations at different points within zone II, but remained largely below 10%. The pollen of *Pinus* appears for the first time in the core at 92 cm, and increases towards the top of the core. The microscopic charcoal sequence shows a peak in the frequency at ca. 200 cm with counts
of up to 500, followed by an abrupt drop to an average of 200. Fungal and other spores show oscillations throughout the zone. At ca. 200-140 cm there was a drop in the percentages of Chaetomium, Spormiella, Sordoria, Gelasinospora, and Cercophora to less than 5%, while the unknown and/or indeterminable spores recorded high percentages (40-60). Despite the changes in the frequency of different spore taxa their sum showed a gradual increase towards the top.

Zone I generally presents pollen and spore assemblage patterns similar to zone II. What is conspicuous in zone I is the low representation of Acacia pollen, despite the dominance of Acacia woodland vegetation in the savanna landscape at present. There appears to be a general decrease in arboreal pollen especially Podocarpus, Oleaceae, and Celtis. The zone is characterized by the absence or low representation of pollen of Euclea, Dodonaea, Macaranga, Myrtaceae, Phoenix, and Sesbania, a trend that commenced in zone II. Pollen of Gramineae and Cyperaceae accounted for over 80% of the pollen sum in zone I while fungal and fern spores showed a general decrease with significant dips in the frequency of Glomus, Cercophora, Polypodiaceae, and other Pteridophytes. However the spores of Sordaria, Spormiella, and Chaetomium recorded an increase in their percentages when compared to the preceding zone. The frequency of microscopic charcoal showed a general downward trend before recovering slightly at the top of the core.

4.4.1 Vegetation Response to Climate Variability

An important factor in interpreting the Lake Baringo pollen profiles is the mode of transport for different pollen types. The mode of transport plays a significant role in determining pollen deposition characteristics and can be important for interpretation.
Most of the pollen in the Lake Baringo profile results from fluvial deposition, and to a lesser extent wind transport. Considering that most of the rivers flow from the south of the lake (Figure 3.3) it is likely that most of the pollen in Lake Baringo record are drawn from the plains and highlands within the lake’s drainage basin to the south. Therefore, the record represents both local and regional vegetation.

The material at the bottom of core LB-5 was probably deposited following very arid conditions that led to the drying of Lake Baringo at that site. Evidence of desiccation during this period was provided by such macrofossils as seeds and rootlets, recovered at the base of the core. A period of prolonged aridity could explain the relatively poor state of pollen preservation during that period. Pollen in Zone V (ca. AD 1650), though very poorly preserved, was characterized by relatively high percentages of Gramineae, Cyperaceae, Combretaceae, *Podocarpus*, *Olea*, *Tarchonanthus*, *Acacia*, and *Euclea*.

*Euclea* species belong to the Ebenaceae family and is often well established in bush, dry forest margins, thornscrub, and open woodlands of East Africa (Vincens et al. 2006; Bussmann et al. 2006; Sharam et al. 2006; Bonnefille and Riollet 1980; Bonnefille 1971a, b). It is usually associated with *Acacia* species, also growing on anthills and river banks in dry areas below 900 m. Therefore, presence of *Euclea* and *Acacia* suggests an expansion of savanna woodland and scrub vegetation, consistent with an increasingly arid environment. The relative abundance of *Podocarpus*, Combretaceae, and *Olea* in zone V further suggests the prevalence of a drier, and probably cooler, climate in the Lake Baringo area than at present. *Olea* pollen is found in many pollen diagrams from East Africa (e.g., Rayner et al. 2006; Vincens et al. 2006; Lamb et al. 2003; Mworia-Maitima
1991; Hamilton 1982), since members of the Oleaceae family are widely distributed in the dry forests. The species represented in the Lake Baringo sediments is probably *Olea hochstetteri*, an important component of the dry forests often occurring in abundance in association with *Podocarpus* (c.f., Olago 2001; Hamilton and Perrott 1981). The other members of the Oleaceae family *O. Africana* and *O. welwitschii* tend to be found in moister montane forests between 2000 and 3000 m (Hamilton 1982). Although the presence of *Olea hochstetteri* pollen often corresponds to dry conditions, accurate interpretation of the climate signal is hampered by the fact that various species of *Olea* with very similar pollen grow together in a wide spectrum of forest types. The presence of *Podocarpus* and *Olea* in the Baringo record should, therefore, be interpreted with caution due to taxonomic problems with the latter and the wide range of habitats within which the former can thrive (c.f., Beuning et al 1997; Livingstone 1993). The two common species of *Podocarpus* (*P. falcatus* and *P. latifolius*) in the Kenyan highlands have distinct ecological preferences, yet their pollen grains are identical (c.f., Lamb et al 2003; Beentjee 1994; Lind and Morrison 1974). *P. falcatus* tends to be associated with drier climates with rainfall between 1000-1500 mm yr-1 while *P. latifolius* thrives in much wetter conditions. Although the pollen assemblage during ca. AD 1650 offers evidence for aridity, the ambiguity surrounding *Podocarpus* and *Olea* limits our ability to make definitive statements regarding temperature around Lake Baringo at that time.

The Lake Baringo area recovered rapidly, albeit briefly, from the period of desiccation that ended in ca. AD 1650. The plant assemblage in Zone IV (ca. AD 1650 - 1750), point to an increase in moisture as evidenced by the relatively high percentages of *Juniperus, Macaranga, and Gramineae*. There was also an increase in the pollen
percentage of *Phoenix* which requires relatively moist conditions (c.f., Dransfield 1986). During this period, there was a slight decrease of dry-indicator plants belonging to Combretaceae, Chenopodiaceae and Amaranthaceae. There was also a dip in the frequency of Cyperaceae pollen, probably due to expansion of the lake surface to occupy habitats that had been colonized by swamps during the intense drought period that ended in ca. AD 1650.

At 295 cm (ca. AD 1750) in zone III (260-295 cm), there was an abrupt return to drier conditions, similar to those of the period ending ca. AD 1650. The drier conditions were marked by an increase in dry-indicator plants such as *Acacia*, *Euclea*, Balanitaceae, Chenopodiaceae and Amaranthaceae, *Justicia*, and *Phyllanthus*. Although pollen percentages of *Podocarpus* and *Olea* were relatively high there was a general reduction in the total pollen count, accompanied by a reduction in the pollen percentages of grasses (Gramineae) and sedges (Cyperaceae), from more than 20% to less than 5%. Low pollen count is often consistent with aridity and unfavorable pollen preservation conditions in a lake following a drop in lake level (c.f., Faegri and Iversen 1989). The poor pollen preservation environment during that period is supported by an increase in the number of indeterminable and unidentifiable pollen. Lithostratigraphic features during that period (see Chapter 3 for details) were very similar to those preceding ca. AD 1650, characterized by crumbly appearance and presence of traces of rootlets, which is consistent with terrestrial environments. Although we hypothesize that intense desiccation at ca. 1750 led to low water levels in Lake Baringo, and subsequent drying up of some parts of the lake, intense drought conditions alone may not fully explain the dramatic reduction in pollen of Gramineae and Cyperaceae in the Lake Baringo record.
It is possible that the intense drought during that period was accompanied or followed up by a series of locust invasions. Historical records and oral tradition from the Lake Baringo region provide evidence which suggests that locust invasion often preceded period of intense droughts (c.f., Anderson 2002). For instance, some of the worst drought episodes in the lowlands of southern and central Baringo: Kiplel Kowo (1927-1929), Talamwei and Kipkoikoio (1931-1934), were all accompanied by locust invasions (Anderson 2002). Although desert locusts (*Schistocerca gregaria*) eat nearly all leafy vegetation, accounts from oral tradition indicate that they have preference for pasture grasses and sedges, as well as such crops as millet, sorghum, and corn. A combination of intense droughts and locust invasion around AD 1750 may have contributed to the low pollen count and low percentages of Gramineae and Cyperaceae pollen.

For the most part, the pollen assemblage in zone II (260-35 cm) which covers the period after AD 1750, reflects a series of short-term vegetational changes. It is within this period (at ca. AD 1800) that *Zea mays* (maize) pollen first appears in the Baringo record. The presence of *Zea mays* pollen suggests arable farming; however in the case of the Baringo pollen record it was not accompanied by any significant changes in vegetation, with the probable exception of a slight increase in the percentage of *Ricinus communis*, an indicator of human disturbance. (c.f., Vincens et al. 2003, Lamb et al. 2003). There was high variability in the pollen percentages of major taxa which was probably a response to regional climate variability on decadal time scales rather than human disturbance. Most pollen indicators in zone II generally point to increasing aridification of the area. There was a short-lived rebound of pollen percentages of
Gramineae, Cyperaceae, Balanitaceae, and *Phyllanthus* in the period leading to ca. AD 1800, after which the percentages plummeted. The pollen of *Acacia*, *Tarchonanthus*, and Combretaceae showed a similar pattern.

Although high percentages of *Acacia* pollen in zone II denote the prevalence of dry conditions after ca. AD 1800, the decrease in the percentages may not necessarily represent amelioration of regional climatic conditions. The low percentages of *Acacia* upwards of 112 cm and especially towards the top of core LB-5 are probably an indication of browsing pressure from both small and large mammals rather than climate change. This view is supported by the fact that percentages of other pollen taxa already identified as dry-indicators such as Balanitaceae, *Phyllanthus*, *Podocarpus* and *Juniperus* remain largely unchanged. This trend continues into zone I that includes the present time. There was an obvious change in the percentages of arboreal pollen towards the top of zone II (early 20th century) and into zone I, marked by significant decreases in the percentages of *Podocarpus* and *Olea*, accompanied by the near absence of *Celtis* and *Macaranga*. These vegetation changes are likely due to deforestation and changes in agricultural practices that accompanied the colonial rule in East Africa. An important indicator of European influence in the pollen record is provided by the presence of the pollen of *Pinus* and other introduced species.

At 98 cm, which is consistent with the early part of the 20th century, *Pinus* pollen appeared for the first time in the Lake Baringo record, and by 1930s it accounted for 5-7% of total pollen. Pine trees were introduced by Europeans who established plantations of mainly *Pinus patula* and *Pinus radiata* in the Rift Valley highlands in the 1910 and 1920s. Therefore, the appearance of pine pollen in the Lake Baringo records marks the
establishment of European and/or colonial influence in the East African environment since most pine species take between 7 and 15 years to produce viable seeds (Richardson et al. 1994). Increase in the percentage of pine pollen was coincident with an increase in Juniperus pollen reflecting the contribution from Cupressus lusitanica, which was introduced along with pine. The increase in the percentage of pine and Juniperus pollen appears to occur about the same time as the decrease in Acacia and Hypoestes pollen. This was probably due to pressure from pastoralists who were confined to permanent use of lowland grazing resources after being displaced by European settlers from the highlands which formerly formed dry season grazing lands.

4.4.2 Human Impact on Vegetation

The pollen profile from Lake Baringo does not seem to provide clear evidence of human influence in the ecosystem, probably because the typical dominant woody taxa of the savanna vegetation do not produce abundant wind pollinated pollen (c.f., Lamb et al. 2003). However, evidence of human impact on the Lake Baringo area over the past 350 years can be inferred from the fungal spore record, as fungal remains are valuable anthropogenic indicators. Such phenomena as deforestation, soil erosion, grazing, and crop cultivation are usually associated with distinct fungal assemblages relative to undisturbed natural ecosystems (c.f., van Geel et al 2003). Therefore, fungal spores can be useful pointers of anthropogenic influence in environments, such as those of East Africa, where indigenous cultivated crops (e.g., cereals, cassava, banana, and legumes) do not leave identifiable pollen. (c.f., Vincens et al 2003; Taylor 1990; Hamilton et al. 1986). It is against this background that fungal spores as proxy for human impact on the Lake Baringo area were examined.
Figure 4.2 shows the percentages of principal fungal spore taxa in core LB-5 that are of relevance for the investigation of anthropogenic and environmental changes. During the period preceding ca. AD 1650 (zone V, 363 cm and 343 cm) the sequence was characterized by a low count of spores across all the major taxa with the exception of *Glomus*-type. The low count (less than 100) of identifiable fungal spores was probably due to poor preservation environment following low lake level resulting from prolonged desiccation. *Glomus*-type, which exhibit extreme variability in size (17-137 µm), are usually found below soil surfaces because they are endomycorrhical fungi that occur in the roots of a variety of host plants (van Geel et al. 2003). Their presence in zone V is an indication of soil erosion. The near absence of *Sporormiella*-type fungal spores suggest that the ca. AD 1650 drought was probably so severe that only a very small number of herbivores survived in the ecosystem. *Sporormiella* is an ascomycete fungus that thrives primarily in dungs of domestic herbivores as well as wild herbivores, including elephants (c.f., Davis and Shafer 2006; Ebersohn and Eicker 1997). The other possible explanation for the absence is that the intense aridity inhibited growth of the dung fungus.

The period after AD 1650 and before 1750 (zone IV, 343-252 cm) showed signs of remarkable recovery from aridity that preceded this period. The recovery is evidenced by increases in coprophilous fungal spore counts. The assemblage in zone IV which is dominated by *Sordaria*-type fungal spores is also characterized by high percentages of *Sporormiella*-type, *Chaetomium*-type, *Cercophora*-type, and *Glomus*-type especially during ca. AD 1700-1750 (300-252 cm). The presence of *Glomus*-type fungal spores could be an indication of increasing soil erosion and probably land degradation due to human impact on the ecosystem. This is because the increase in *Glomus*-type spores is
coincident with increased percentages in the spores of *Sporormiella*-type, *Chaetomium*-type, and *Sordaria*-type. Both *Sordaria*-type and *Cercophora*-type fungal spores are coprophilous taxa that belong to the *Sordaria* genus which is common in human settlement sites (van Geel et al 2003; Willemsen et al 1996). The latter often thrives on decaying wood, herbaceous leaves and stems often occurring in combination with *Sporormiella*-type and other indicators of dung from herbivores (c.f., Buurman et al 1995; Davis 1987).

Fungal spores of *Sporormiella*-type also live in dung from wild herbivores (c.f., Davis and Shafer 2006). Therefore, the presence of *Sporormiella*-type in the Lake Baringo area at ca. AD 1700-1750, though suggestive, does not necessarily provide unequivocal evidence for human impact in an environment probably teeming with large numbers of wild herbivores including zebras, gazelles, antelopes, rhinoceros, hippopotamus, and elephants. However, because *Sporormiella*-type is coincident with *Sordaria* and *Chaetomium* species, there is a strong case for human influence in the ecosystem. Also, whereas soil erosion and/or land degradation at ca. AD 1700-1750 may have been due to trampling of the soil from the increasing population of wild game, the presence of *Chaetomium* ascospores provides compelling evidence for human input. *Chaetomium* species are saprophytic cellulose-decomposers that occur on leather, bones, feathers, cloth, and decaying herbaceous stems (van Geel et al 2003). Thus, soil erosion was probably due to a combination of human impact due to pastoralism and large numbers of wild herbivores following an increase in wetness. There was also an increase in the percentages of fungal spores classified under the category of “others” which further confirm the amelioration of climate after ca. AD 1650. Although low percentages of
Polypodiaceae were recorded, the percentages of other Pteridophytes were much higher than in the preceding zone.

The analysis of fungal and fern spores from surface samples collected from 24 sites within the Lake Baringo catchment clearly demonstrates that Sporormiella-type, Chaetomium-type, and Sordaria spores are associated with human activities. The samples were divided into 6 main groups based on the land cover/use and vegetation of their source areas. The fungal and fern spore data derived from the surface samples are shown in Figure 4.3. Group 1 (samples 1-4) corresponds to fungal and fern spore assemblages that were processed from soil and dung samples collected from goat sheds near the lake. The group 2 (samples 5-8) assemblages were derived from soils taken from enclosures for cattle in the vicinity of Lake Baringo. The results show that Groups 1 and 2 are dominated by coprophilous fungi taxa especially Sporormiella-type, Chaetomium-type, and Sordaria spores which account for 80-90% of the total count. This assemblage is typical of sites associated with human settlement and livestock (c.f., Davis and Shafer 2006; van Geel 2003).

Group 3 (samples 9-13) consisted of those that were processed from soil samples collected from the forested area in the Tugen Hills highlands that constitute one of the catchment areas for the lake. Figure 4.3 shows that Group 3 is dominated by Cercophora type (10-20%), Glomus type (10-20%), and fern spores (Polypodiaceae and other Pteridophytes), which account for up to 30% of the total sum. The spores of Sordaria-type, Sporormiella-type, and Chaetomium-type were very poorly represented (mostly less than 5%) in Group 3 because apparently there is minimal or no grazing in upland forests.

The results also show that Group 4 (samples 14-19) which included fungal and fern spores
Figure 4.3  Fungal and fern spores that were analyzed from surface samples that were collected from 24 sites within the Lake Baringo drainage basin. Groups 1 and 2 were collected from livestock enclosures. The other sites included forest (Group 3), savanna bushland/woodland (Group 4), farms (Group 5) and riverbeds (Group 6).
derived from the lowland plains that were covered mainly by acacia bushland/woodland landscape, had a near equal representation of different fungal and fern spores, probably due to widespread pastoralism in these sites.

The Group 5 (samples 20-22) assemblage was processed from soil samples that were collected from farms on the Tugen Hills. This group was dominated by spores of *Sordaria*-type (20-30%), *Cercophora*-type (10-15%), and *Glomus*-type (15-25%), again linked to human activities. Group 6 (samples 22-24) consisted of spores that were processed from two soil samples collected from dry riverbeds about 5 kilometers from the lake. Although the spores of *Sordaria*-type dominate this group (ca. 20%), spores from other species are well represented, with each type accounting for ~10% of the total count.

The presence of significant numbers of spores of *Sordaria*-type, *Cercophora*-type, *Chaetomium*-type, and *Spormiella*-type in the groups 4 and 5, and to a limited extent in Group 3, correlate to the ubiquitous livestock herding in those sites. Indeed Group 4 has the greatest representation of the coprophilous fungal spores outside of Groups 1 and 2, since the woodland plains are important grazing areas.

Fungal spore taxa data derived from these six groups were subjected to discriminant analysis in SPSS to validate the groups. Discriminant analysis is an excellent tool for evaluating prior classification of surface data into groups (e.g., Liu and Lam 1985). Figure 4.4 shows a plot of the groups against Discriminant Functions 1 and 2 in statistical space. A total of 23 out of 24, or 92 % of the samples, were correctly classified into their respective groups. The validity of the classification was further confirmed through the cross-validation technique of discriminant analysis. In the cross
Figure 4.4 A plot of the six groups against Canonical Discriminant Functions 1 and 2.
validation technique, discriminant analysis basically removes sample individually and test the probability that the sample will be classified back into its original group. These analyses therefore demonstrate that the six groups are valid and representative of the spores from their respective sites. Each group yields a distinct assemblage of spores and coprophilous fungal spores that are consistent with human settlements and increased livestock activity.

There was a general decrease in the percentages of major fungal spore taxa at 252-165 cm (ca. AD 1750-1840s), zone III. One of the most conspicuous characteristics of that period was the near absence of *Sporormiella*-type fungal spores. The paucity in the *Sporormiella*-type may imply a drought-driven drop in the population of herbivores when compared to the preceding period. The decrease in the population of herbivores, especially domesticated animals, in the region is likely to have been accompanied by a drop human in population due to famine and/or migration from the area. The percentages of spores of *Cercophora*-type and *Sordaria* dropped to 5% from 15-30%, and 15% from 40%, respectively. However, there was no reduction in the percentages of *Glomus*-type spores which suggests that soil erosion and/or land degradation continued unabated during that period. It is likely that wind erosion was the primary agent of soil erosion in an environment characterized by intense aridity. During that period other spores, mostly unknown and poorly preserved fungal spores of terrigenous origin, recorded an increase in percentages when compared to zone IV. Again, the poorly preserved spores and pollen during that period could be viewed as evidence of desiccation in the Lake Baringo area. Further evidence for aridity could be adduced from the microscopic charcoal record which shows a peak with counts in excess of 500 particles at 200 cm. During droughts
any fire incident is likely to have become either more widespread and/or more intense, thereby accounting for the high charcoal count.

The reappearance of *Sporormiella*-type spores, accompanied by an increase of *Chaetomium*, and *Sordaria* species after 165 cm (ca. AD 1850s), is consistent with increased anthropogenic activity within the Lake Baringo catchment. This anthropogenic signal in the Baringo record is coeval with the era of Maasai expansion in the Rift Valley lowlands, and large settlements at Il Chamus (also known as Njemps) flats in the vicinity of Lake Baringo in the mid 19th century (c.f. Anderson 2002). Oral histories of several Maa-speaking and Kalenjin-speaking peoples who reside within the Rift Valley, some of which are augmented by written accounts of European travelers, confirm that the period leading to the mid 19th century was a time of great success in pastoralism (c.f., Waller 1978; Krapf 1860; Wakefield 1870). It is therefore reasonable to argue that after ca. AD 1830 the area around Lake Baringo was probably occupied by pastoralists and some agriculturalists who must have been actively impacting on the ecosystem.

The increase in microscopic charcoal in the Lake Baringo record after ca. AD 1830 can be interpreted as evidence of increasing human disturbance in the ecosystem. The Il Chamus were renowned for their agricultural activities that included sophisticated irrigation techniques around Lake Baringo (Anderson 2002; Little 1992, 1996). The charcoal peak in the record could be associated with their farming practices that probably included slash and burn. The other possible source of the charcoal in the lake is the honey harvesting practice in Lake Baringo that involves smoking the bees. Informal interviews with local residents during fieldwork revealed that the practice of smoking the bees
sometimes lead to fire accidents that become widespread, especially during the dry season.

The Baringo record shows that the human impact on the Lake Baringo area that commenced in the mid 19th century has persisted to the present time albeit with minor fluctuations. Throughout that period the percentages of *Glomus* species that are markers for soil erosion remained high suggesting that land degradation persisted throughout that period, increasing with time. The slight reduction in the *Glomus* percentages towards the top of the core, correlating to recent times, is somewhat confusing. This is because the decrease in the percentage of *Glomus* implies a slight amelioration in land degradation which suggests that the conservation efforts that were first introduced in the 1930s may not have been entirely in vain. However, evidence from Satellite imagery, introduced and discussed in Chapter 5, clearly show massive deforestation and land degradation since 1986. However, the percentages of the spores of *Spormiella* and *Sordaria* increased towards the top of the core, indicating increasing livestock activities and human influence in the Lake Baringo area.

### 4.5 Conclusion

This chapter has described and discussed a fossil pollen and spore record from Lake Baringo in terms of environmental change and human activities in the East African region over the past 350 years. The record shows dominance of *Podocarpus, Olea, Acacia*, Balanitaceae, Gramineae and Cyperaceae throughout much of core LB-5 which indicates prevalence of generally dry conditions throughout the period. Even in the context of a largely dry environment the pollen profile showed oscillations between wetter periods and more arid conditions over decadal and even sub-decadal time scales.
which are well marked through subtle vegetational changes. Although grazing and
browsing by domestic animals, as well as deforestation, could be partially responsible for
the vegetational changes evident in the record, the general pattern of the pollen profiles
strongly indicate climate variability.

Overall, there were no marked changes in the pollen profiles (with the exception
of introduced species; Zea and Pinus) that can be directly linked to human impact.
Although the 20th century was marked by changes, especially in the percentages of
arboreal pollen, which are consistent with increasing deforestation, those changes are
somewhat muted. A partial explanation could be the fact that woody taxa in the savanna
landscape do not produce abundant pollen, and many (like Acacia) are not wind
pollinated. It is also possible that human impacts on vegetation were less important
compared with climate induced changes and, therefore, were masked by the latter.
However, interpretation of this study’s pollen record is both difficult and complex
because the Lake Baringo catchment from which pollen enters the lake is large and
covers diverse altitudinal extents ranging from less than 900 m to beyond 2200 m above
sea level. Two episodes of intense droughts, at ca. AD 1650 and AD 1750-1800, that led
to drying up of large portions of Lake Baringo are well marked in both the pollen and
spore record. The two desiccation episodes were characterized by poor pollen
preservation due to low lake levels.

Although the anthropogenic influence on the vegetation of Lake Baringo is
largely masked by climate induced changes the fungal spore record helps to illuminate
the history of human impacts on the ecosystem. High percentages of spores of
Spormiella, Chaetomium, Sordaria and Cercophora species, which are common
indicators of livestock and other human activities, correlate well with vegetational
changes associated with climate variability in the Lake Baringo area. Spores of *Glomus*
type, which are a proxy for soil erosion, recorded high percentages at different points in
the historic time prior to European influence in the East African region. Thus, the high
rates of soil erosion and land degradation presently evident in the Lake Baringo area have
been part of the Baringo landscape at least for the past 350 years, albeit which
oscillations whose pace appears to be set by climate variability.

4.6 References

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5.1 Introduction

Most of East Africa’s land-cover is in a state of permanent flux at a variety of spatial and temporal scales, due to both climatic variability and human activities. The fluxes are most pronounced in the transitional zone between forest and savanna particularly in the semi-arid environments. The Lake Baringo area in East Africa (Figure 5.1) presents a good example of environments experiencing remarkable land cover change due to multiple causes. Changes that are of great interest to ecologists and resource managers are those that are ultimately linked to human activities such as deforestation, and land clearing for agriculture or pastoralism. Satellite remote sensing can play a crucial role in providing information on land-cover/use modifications on local, regional, and even global scales, especially where aerial photographs are missing or outdated. The ability of any remote sensing system to detect and monitor such fluxes in land-cover depends on its capability to adequately deal with the reference data base while simultaneously accounting for both short term variability (e.g., seasonal) and longer-term secular change.

Satellite remote sensing data provide important information that is potentially useful for estimating levels and rates of deforestation, land degradation, habitat degradation, and other landscape changes. To date, satellite monitoring techniques have proven excellent for detecting land-cover/use change through time, thereby providing

Figure 5.1 Map of East Africa showing the approximate location of the lake Baringo catchment.
These techniques have been applied extensively in monitoring change in a variety of natural environments (e.g., Wulder et al. 2006; Jin and Sader 2005; Cohen et al. 2003; Dowson et al. 2003; Townshend 2002; Wilson and Sader 2002; Gemmell et al. 2001; Boyd et al. 1996; Collins and Woodcock 1999; Skole and Tucker 1993).

Both optical remote sensing and synthetic aperture radar platforms have been successfully used for monitoring and assessing such phenomena as drought, land degradation, and land cover change in semi-arid ecosystems (e.g., Servenay and Prat 2003; Wang et al. 2003; Nishida et al. 2002; Asner and Lobell 2000; Gilabert et al. 2000; Sujatha et al. 2000; Metternicht and Zinck 1998; Dwivedi et al. 1997; Tieszen et al. 1997; Tripathy et al. 1996; Mishra et al. 1994; Moran et al. 1994; Nemani et al. 1993; Price 1993). For instance, Servenay and Prat (2003) used unsupervised algorithms to delineate four stages of soil erosion in Mexico using multitemporal SPOT data. Wang et al. (2003) used Landsat TM data to study the relationship between soil classes and erodibility, demonstrating that Landsat TM data was excellent for reproducing spatial variability of erodibility in a number of different landscapes. Other studies (e.g., Flügel et al. 2003; Khan et al. 2001; French et al. 2000; Mati et al. 2000) have used vegetation indices matched with change detection techniques to study land degradation processes. French et al. (2000) used a combination of Normalized Difference Vegetation Index (NDVI) and thermal infrared emissivity to discriminate bare soil surfaces from senescent vegetation.

The remote sensing change detection techniques used for monitoring change in land-cover/use are based on the assumption that areas which have changed will exhibit greater differences between dates than will areas that have not changed (Radke et al.
2005; Kasischke et al. 2004). Various studies (e.g., Muttitanon and Tripathi 2005; Kiage et al. 2005; Lotsch et al. 2003; Tucker et al. 1985; Ehrlich et al. 1995; Munyati 2000; Maselli and Rembold 2001; Anyamba 2001, 2002, 2003; Lupo et al. 2001) have demonstrated the versatility of change detection techniques in unraveling varied climatic and vegetation/environmental change problems. Although change detection techniques are fairly straightforward they can be complicated by “false changes” that may be unrelated to actual land-cover change. These include the effects of sun angle, viewing geometry, atmospheric conditions, variability in climate, and phenology (Muttitanon and Tripathi 2005; Kasischke et al. 2004; Varjo 1996; Olsson 1995). In particular, atmospheric effects can have significant effects on data, and inability to correct for them can influence classification results and vegetation indices (c.f. Vogelmann et al. 2001). However, most of these complications can be eliminated through radiometric image preprocessing techniques and/or selecting “anniversary” images for the change detection procedure (Kasischke et al. 2004; Vermote et al. 1997).

One of the principal goals of this study was to use satellite measurements to study vegetational change associated with land degradation in the Lake Baringo area, Kenya, and link these changes to sedimentation rates in the lake. The Lake Baringo area, a critical habitat for more than 500 avifaunal species and a source of livelihood for humans, is threatened by land degradation as a result of mounting population pressure. A combination of natural and anthropogenic factors, including deforestation in the lake catchment, increased sediment yield from soil erosion, evaporation, damming, and diversion of inflowing rivers has reduced the depth of the lake from a maximum of 9 m and a mean of 5.6 m in 1972 to the current maximum of 4 m and a mean of 2.5 m.
Reports have recently appeared in the media in East African quoting the United Nations Development Programme (UNDP) authorities that project the lake to fill up with sediments in the next 15 years (Kurgat 2003). These environmental problems are ultimately linked to land-cover/use changes. It is therefore important to investigate land-cover/use changes in order to monitor environmental change and develop sustainable resource management plans for fragile ecosystems such as Lake Baringo.

This article examines the land-use/cover change in the Lake Baringo catchment over a fourteen-year period using Landsat Thematic Mapper/Enhanced Thematic Mapper Plus (TM/ETM+) imagery. Our focus was largely on the changes in the forest and bare ground cover, which are linked to different forms of soil erosion, a proxy for land degradation. Change-related images are generated from change detection techniques including post classification image differencing, Kauth-Thomas tasseled cap transforms, and the Normalized Difference Vegetation Index (NDVI) differencing. Although land-use and land-cover are often used interchangeably, the two concepts are not identical. The latter measures the physical attributes or characteristics of the earth surface while the former describes how land-cover is utilized (c.f. Seto et al. 2002). Without additional information, remote sensing detects land cover, not land use—though in most cases one can derive land use from land cover. However, most remote sensing change detection studies use the two concepts interchangeably (e.g., Muttitanon and Tripathi 2005; Heikkonen and Varfis 1998; Dimyati et al. 1996) because land-use often corresponds to land-cover. For instance, in pastoral environments pasture could describe land-use type while at the same time it is a form of land-cover. In such scenarios the two concepts
could be considered synonymous. In this article the two terms are used interchangeably unless otherwise specified

5.2 The Study Site

The focus of this study is the main catchment for Lake Baringo estimated to be 8655 km\(^2\) (Hickley et al. 2004), and the immediate vicinity of the lake (Figures 5.1 and 5.2). The entire catchment lies within the eastern fork of the Gregorian Rift Valley in East Africa. It extends from approximately 0°43’N to 0°40’ S, and 35°18’E to 36°20’E. A variety of environmental gradients are represented in the catchment ranging from semi-arid areas with bare or sparsely vegetated ground on the lowlands to evergreen forests on the highlands and upper reaches of the catchment. These variations are mainly related to altitude, which ranges from < 900 m in the central and northern parts of the catchment around Lake Baringo to well above 2200 m at the Tugen Hills and the highlands on the western and southern regions (Figure 5.2).

The inhabitants of the savanna and semi-arid lower reaches of the catchment are nomadic pastoralists with livestock under communal grazing arrangements. Both large and small scale agriculture is practiced in the lush environments that characterize the upper reaches of the catchment. The spatial distribution of rainfall in the catchment correlates well with the topography: less than 600 mm/yr in the lowlands and over 1200 mm/yr in the highlands. The major land-covers in the catchment are forest (deciduous and evergreen), water (mainly the lake), bushland/woodland/scrub (dominated by different Acacia species), wetland, bare ground (little or no vegetation), and pasture/farmland.
Figure 5.2 The Lake Baringo catchment delimited with white line within the Landsat Enhanced Thematic Mapper Plus image of 28 January 2000 (bands 5, 4, 3 displayed as red, green and blue) that was used in the study. The names of the main forests (all situated on high elevation, above 1950 m above sea level) and lakes are provided. Lake Baringo is at the top part of the image and looks different from the other lakes because of its high reflectance (due to high turbidity).
Table 5.1. Attributes of the Landsat TM/ETM+ imagery used in the study.

<table>
<thead>
<tr>
<th>Acquisition date</th>
<th>Sensor</th>
<th>Spatial Resolution (meters)</th>
<th>No. of Bands*</th>
<th>Sun elevation</th>
<th>Sun azimuth</th>
<th>RMS.error (GCP No.)</th>
<th>Radiometric Correction</th>
</tr>
</thead>
<tbody>
<tr>
<td>January 27 1986</td>
<td>TM 5</td>
<td>30</td>
<td>6</td>
<td>53.58</td>
<td>122.50</td>
<td>0.503 (14)</td>
<td>yes</td>
</tr>
<tr>
<td>January 28 2000</td>
<td>ETM+</td>
<td>15†, 30</td>
<td>7</td>
<td>48.93</td>
<td>118.89</td>
<td>Reference</td>
<td>yes</td>
</tr>
</tbody>
</table>

* Thermal bands were excluded in the classification and change detection process
†Panchromatic band resolution is 15 m.
5.3 Methods

5.3.1 Data Acquisition and Initial Processing

A crucial requirement for change detection is availability of relatively cloud-free satellite image at each date. However, obtaining anniversary cloud free images from the tropics is a daunting task because the region tends to be cloudy throughout the year. One has to take advantage of cloud free windows such as the dry season. Two anniversary (27\textsuperscript{th} January 1986 and 28\textsuperscript{th} January 2000) satellite images were acquired using the Landsat 5 TM and Landsat 7 ETM+ sensors, thereby limiting exogenous effects. Table 5.1 provides the characteristics of the images used. Although slight differences in spectral band width, position, and calibration exist between TM and ETM+ sensors (Teillet et al. 2001; Vogelmann et al. 2001), they are not significant enough to affect outcome of the analyses used in this study (c.f. Cohen et al. 2003). The scene location was based on the Landsat worldwide reference system Path 169 and Row 60, covering the entire Lake Baringo catchment.

The steps involved in the methodology for this study are summarized in Figure 5.3. After data acquisition, the ETM+ image was further georeferenced to the 1:50 000 Universal Transverse Mercator (UTM) topographic maps (Clarke 1880 spheroid, UTM 37) prepared by the Kenya Ministry of Lands, Department of Survey. Polynomial transformation model in ERDAS-Imagine 9.0 (ERDAS 2005) was then used in the image to image registration with the already georectified ETM+ image as the reference. Radiometric calibration (discussed later in this section) was performed before the images were subset using the Lake Baringo catchment (drainage basin) as the area of interest (Figure 5.2).
Figure 5.3. Summary of the change detection procedure.
5.3.2 Change Detection Techniques

In almost all studies involving the use of satellite imagery for monitoring environmental change, imagery from one date is compared with another image from a different date. Within this paradigm of analyzing images as endpoints, numerous methods have been developed (Yang and Liu 2005; Kasischke et al. 2004; Hobbs 1990; Coppin and Bauer 1994). Some methods are suited for measuring specific kinds of change while others are more general in their applicability. Kasischke et al. (2004) identified three broad steps involved in change detection: (1) radiometric preprocessing, (2) data transformation, and (3) mapping change. However, not all the steps are necessary for change detection studies, and generally the relative importance of the first two steps is dependent on the method recruited.

5.3.2.1 Radiometric Preprocessing

The quality of information derived from Landsat imagery depends on such factors as image quality, analysis techniques, and an array of temporal/phenological considerations. Even cloud-free Landsat TM/ETM+ data often contains noise from atmospheric interference with propagation of electromagnetic energy, changes in illumination geometry, and some noise from the instrument. It is often helpful to remove exogenous effects from multidate imagery to eliminate error and facilitate effective mapping of vegetational or environmental change (Muttitanon and Tripathi 2005; Kasischke et al. 2004; Varjo 1996). Radiometric preprocessing of TM/ETM+ data is one of the most effective ways of doing so. Vogelmann et al. (2001) and Teillet et al. (2001) provide a good summary of the significance and issues involved in the calibration of TM and ETM+ data.
Kasischke et al. (2004) and Olsson (1995) group radiometric preprocessing techniques for change detection into two broad categories; relative and absolute correction. The former involves an attempt to match the dates of image acquisition so that such factors as sun angle and the atmosphere are constant between the dates. Usually one date (the subject image) is calibrated against the other date (the master image) using unchanging features within the image. The effect of this calibration is that the two dates appear as if they were acquired under the same illumination and atmospheric conditions. These corrections are relative because the effects are not actually removed from the imagery. The effects of relative corrections are very similar to choosing “anniversary” images. In absolute radiometric calibration the original brightness values in the images are converted into surface reflectance using a number of atmospheric correction and calibration equations (Lillisand and Kiefer 2004; Vermote et al. 1997). Reliable and accurate conversion factors have been developed for a number of satellite sensors including TM and ETM+ (Huang et al. 2001; Masek et al. 2001). Commonly used radiative transfer codes for atmospheric corrections include MODTRAN and 6S (Vermote et al. 1997).

The anniversary images used in this study enabled us to largely sidestep the effects of sun angle, and phenology. However, to further minimize errors we also performed radiometric calibrations, by converting the original brightness values into surface reflectance thereby enabling us to compare the TM and ETM+ images.

5.3.2.2 Data Transformation Methods

Transformation techniques provide an intermediate step in the change detection processes; the results from this step were used for mapping change. The primary goal of
data transformation is to help reveal changes in surface reflectance. A number of
different techniques were used in this study, including post classification image
differencing, Tasseled Cap transformation, and NDVI differencing.

Image differencing is where one image (usually from an earlier date) is simply
subtracted from the other (usually the most recent date) to produce a difference image.
The net effect of differencing is a product where the no-change pixels are centered on the
value of zero. Thus, when the change image is displayed in grey scale, areas that exhibit
the brightest tones are associated with greater change. In its rudimentary form this
technique may not require atmospheric correction since while atmospheric and other
exogenous effects shift the mean away from zero, they do not affect the image content
(Kasischke et al. 2004). However, Nielsen et al. (1998) note that differencing non-
transformed Landsat data may provide spurious results due to data noise and variable
sensitivity of individual sensors.

Simple differencing of NDVI images was incorporated in this study to circumvent
the errors that may accrue from differencing of individual standardized spectral bands.
The NDVI is a technique that separates green vegetation from other surfaces since the
chlorophyll of healthy green vegetation absorbs red light for photosynthesis and reflects
in the near infrared (NIR) wavelengths due to scattering caused by internal leaf structure
(Wilson and Sader 2002). Dense vegetation shows up very strongly in the imagery, and
areas with little or no vegetation are also clearly identified. Thus NDVI is an excellent
tool for change detection studies. The index is calculated using equation 1:

\[ \text{NDVI} = \frac{\text{NIR} - \text{RED}}{\text{NIR} + \text{RED}} \]  \tag{1}
where NIR is the reflectance radiated in the near-infrared waveband (760-900 nm) and RED is the reflectance radiated in the visible red waveband (630-690 nm) of the satellite radiometer. NDVI values range between -1 and 1, with values above 0.6 indicating dense vegetation and values below 0 indicating no vegetation. Bare grounds have values of 0-0.1, grasslands 0.2-0.3, while negative values correspond with water and ice surfaces. The simplicity of utilizing NDVI in detecting vegetation or the lack of it, plus the ease in calculation from various satellite data and in interpretation, makes it the most widely used vegetation index (Muttitanon and Tripathi 2005; Kasischke et al. 2004; Wilson and Sader 2002).

Tasseled Cap Transformation (TCT) also known as Kauth-Thomas tasseled cap (TC) was initially developed for crop development surveys (Kauth and Thomas 1976). It is basically a guided and scaled Principal Components Analysis (PCA) that transforms the six Landsat ETM/TM bands into three orthogonal planes or components of known characteristics (Huang et al. 2002; Collins and Woodcock 1994; Fung 1990). TC-component one (BI) is a measure of brightness or albedo derived from the responses of all Landsat bands except the thermal band (Armenakis et al. 2003). TC-component two (GI) is a measure of greenness obtained by comparing the NIR with the visible bands. TC-component three (WI) contrasts the sum of the visible and NIR bands with the longer infrared bands to determine the amount of moisture being held by the vegetation or soil (Cohen et al. 1998; Crist and Cicone 1984). It is referred to as the wetness index because it is sensitive to soil and plant moisture (Wilson and Sader 2002; Crist and Cicone 1984). The different components of TC for this study were calculated by using the following formula adopted from Price et al. (2005) (c.f. Jensen et al. 1996; Crist et al. 1986):
BI = 0.2909(TM1) + 0.2493(TM2) + 0.4806(TM3) + 0.5568(TM4) + 0.4438(TM5) + 0.1706(TM7)  \hspace{1cm} (2)

GI = -0.2728(TM1) - 0.2174(TM2) - 0.5508(TM3) + 0.7221(TM4) + 0.0733(TM5) - 0.1648(TM7) \hspace{1cm} (3)

WI = 0.1446(TM1) + 0.1761(TM2) + 0.3322(TM3) + 0.3396(TM4) - 0.6210(TM5) - 0.4186(TM7) \hspace{1cm} (4)

The TC-components were used in this study because a number of studies (e.g., Parmenter et al. 2004; Huang et al. 2002; Collins and Woodcock 1994) have shown that they can account for over 97% of spectral variability present in any given scene. We specifically used the BI components to calculate albedo change in the Lake Baringo waters as a proxy for increased sediment yield and turbidity.

**5.3.2.3 Classification**

Image classification has been used in many ways and for a variety of purposes in monitoring landscape changes (e.g., Muttitanont and Tripathi 2005; Yang and Liu 2005; Nishida et al 2002; Seto et al 2002; Asner and Lobell 2000). We displayed the images using false colour composites by assigning the red, blue and green color guns to bands 5, 4, and 3, respectively. Supervised classification was found to be inadequate in this study because of the complexity of vegetation and topography in the study area. We used the unsupervised classification by applying the Interactive Self-Organizing Data Analysis (ISODATA) algorithm in ERDAS-imagine 9.0 (ERDAS 2005) to identify and classify land-cover/use classes in the satellite images. The ISODATA algorithm was implemented without assigning predefined signature sets as starting clusters which enabled us to avoid the impacts of sampling characteristics. Each image was grouped into 70 spectral clusters when the convergence reached 0.99.
The output of the ISODATA clustering was disintegrated into five broad land-cover classes in order to simplify the interpretation of change in the landscape. Each spectral cluster was assigned to one of the six land-cover/use classes using visual inspection of original images, reference data, and familiarity of the study area. These six classes were forest, water, bushland/woodland/scrub, wetland, bare ground, and pasture/farmland. Table 5.2 provides a descriptive summary of the land cover/use classes. The classification and determination of the land-cover classes was done independently for each image (see figure 5.4). The reference data included seven 1:50,000 topographic maps covering the catchment, and groundtruthing in the Lake Baringo catchment, involving the collection of global positioning system (GPS) points and description of the associated land-cover/use types. Fieldwork was conducted during the months of December 2003/2004 and, again in January 2004/2005, coinciding seasonally with the acquired images. A total of 150 GPS points covering the catchment were collected.

To label the spectral clusters and assign classes, we made the ISODATA output largely transparent and overlaid it on the original image in false colour composite. This was done by assigning the targeted spectral cluster an opacity value of 1 and a bright colour (yellow) while the rest of the clusters were assigned a zero value thereby making the overlaid ISODATA output transparent except for the cluster being examined. Although individual image colour was used in determining class assignment, in some cases image elements such as association and site were used to eliminate possible classification error. Before the final land cover/use products were assessed they were subjected to 8 x 8 hectare clump, and 2 x 2 hectare eliminate filters to reduce noise
Table 5.2. Land cover types used in the classification of satellite-derived land cover.

<table>
<thead>
<tr>
<th>Land cover</th>
<th>Code</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Forest</td>
<td>1</td>
<td>High density of trees with little or no undergrowth. Dominated by such tropical trees as <em>Celtis</em> spp. <em>Urticaceae</em>, <em>Myrtaceae</em>, <em>Croton</em>, <em>Holoptelea</em>, <em>Prunus</em>, <em>Podocarpus</em>, and <em>Olea</em> among others. Restricted mainly to the upper reaches (highlands) of the catchment.</td>
</tr>
<tr>
<td>Bushland/woodland/scrub</td>
<td>2</td>
<td>Common in the escarpments and lowland plains especially near the lake. Typically dominated by several <em>Acacia</em> and <em>Commiphora</em> species with little or no undergrowth.</td>
</tr>
<tr>
<td>Pasture/farmland</td>
<td>3</td>
<td>Environments dominated mostly by grasses and herbaceous plants. Some sections (i.e. farms) are characterized and distinguished by regular, linear and rectangular shaped features.</td>
</tr>
<tr>
<td>Water</td>
<td>4</td>
<td>Mainly the lake waters, dams, and rivers.</td>
</tr>
<tr>
<td>Bare ground</td>
<td>5</td>
<td>Exposed soil surfaces, cultivated areas, and in some instances very scantly vegetated areas.</td>
</tr>
<tr>
<td>Wetland</td>
<td>6</td>
<td>Dominated by swamps vegetation mainly papyrus (<em>Cyperus papyrus</em>) and <em>Typha</em>.</td>
</tr>
</tbody>
</table>
Figure 5.4. The land cover maps generated from unsupervised classification of the 1986 TM and 2000 ETM+ images.
5.3.3 Accuracy Assessment

The method described by Congalton and Green (1999) was used in performing accuracy assessment once the final land cover/use classes were obtained. Classification accuracy assessment was performed on land cover/use maps from both the 1986 and 2000 images. We used seven 1:50 000 topographic maps from the Kenya Department of Survey, as well as data and GPS points collected during fieldwork in Lake Baringo as the reference information for accuracy assessment. The result of the accuracy assessment was a confusion matrix showing errors of omission (producer’s accuracy) and commission (user’s accuracy) and a Kappa coefficient (Table 5.3 and 5.4). The overall classification accuracy for each map was computed by dividing the sum of all the correctly classified pixels (diagonal of the confusion matrix) by the total number of pixels in the confusion matrix.

The user’s accuracy or reliability corresponds to a commission error, representing the probability that a pixel classified on the map actually represents that land cover type on the ground (Story and Congalton 1986). The producer’s accuracy corresponds to the omission error; the probability of a reference pixel being correctly classified. Overall classification accuracy makes use of the diagonal elements of the confusion matrix to measure agreement. Though simple to compute, it tends to overestimate classification accuracy since it overlooks the proportion of random agreement between datasets (Congalton and Green 1999). That limitation is taken care of by the Kappa coefficient which attempts to control for chance agreement by incorporating the off-diagonal elements of the confusion matrix (c.f. Verbyla 1995). The calculation of the Kappa
Table 5.3  Confusion matrix of the land cover classification map derived from the 2000 Landsat-7 ETM+ image

<table>
<thead>
<tr>
<th>Classified data (Land cover type)</th>
<th>Reference data</th>
<th>Classified total</th>
<th>Producer’s accuracy</th>
<th>User’s accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Forest</td>
<td>31 4 0 0 0 0</td>
<td>35</td>
<td>93.9</td>
<td>88.6</td>
</tr>
<tr>
<td>2. Bushland/scrub</td>
<td>2 33 0 0 0 0</td>
<td>35</td>
<td>78.6</td>
<td>94.3</td>
</tr>
<tr>
<td>3. Pasture/Farmland</td>
<td>0 3 29 0 0 3</td>
<td>35</td>
<td>87.9</td>
<td>82.9</td>
</tr>
<tr>
<td>4. Water</td>
<td>0 0 0 15 0 3</td>
<td>15</td>
<td>100</td>
<td>100</td>
</tr>
<tr>
<td>5. Bare Ground</td>
<td>0 2 1 12 0 0</td>
<td>15</td>
<td>100</td>
<td>80</td>
</tr>
<tr>
<td>6. Wetland</td>
<td>0 0 3 0 12 0</td>
<td>15</td>
<td>80</td>
<td>80</td>
</tr>
<tr>
<td>Reference total</td>
<td>33 42 33 15 12 15</td>
<td>150</td>
<td>Overall classification accuracy = 88%</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Overall Kappa statistics = 0.85</td>
<td></td>
</tr>
</tbody>
</table>
Table 5.4 Confusion matrix of the land cover classification map derived from the 1986 Landsat-5 TM image

<table>
<thead>
<tr>
<th>Classified data (Land cover type)</th>
<th>Reference data</th>
<th>Classified total</th>
<th>Producer’s accuracy</th>
<th>User’s accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Forest</td>
<td>33 2 0 0 0 0</td>
<td>35</td>
<td>91.7</td>
<td>94.3</td>
</tr>
<tr>
<td>2. Bushland/scrub</td>
<td>3 31 1 0 0 0</td>
<td>35</td>
<td>81.6</td>
<td>88.6</td>
</tr>
<tr>
<td>3. Pasture/Farmland</td>
<td>0 3 29 0 0 3</td>
<td>35</td>
<td>85.3</td>
<td>82.9</td>
</tr>
<tr>
<td>4. Water</td>
<td>0 0 0 15 0 0</td>
<td>15</td>
<td>93.8</td>
<td>100</td>
</tr>
<tr>
<td>5. Bare Ground</td>
<td>0 2 3 19 0 0</td>
<td>15</td>
<td>100</td>
<td>60</td>
</tr>
<tr>
<td>6. Wetland</td>
<td>0 0 1 0 0 14</td>
<td>15</td>
<td>82.4</td>
<td>93.3</td>
</tr>
<tr>
<td>Reference total</td>
<td>36 38 34 16 9 17</td>
<td>150</td>
<td>Overall classification accuracy = 87.3%</td>
<td></td>
</tr>
<tr>
<td>Overall Kappa statistics = 0.842</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
coefficient is similar to the Chi-square, and is computed using the following formula,

\[ K = \frac{\text{Overall Classification Accuracy} - \text{Expected Classification Accuracy}}{(1 - \text{Expected Classification Accuracy})} \]  

The Kappa coefficient is basically the proportion of agreements after chance agreement has been excluded. Its upper limit is 1.00 (total agreement). Complete agreement at a chance level has a coefficient of zero.

Table 3 presents the results of the confusion matrix for the 1986 land cover/use classification accuracy assessment. The overall classification accuracy was 87.3% with an overall Kappa statistic of 0.842. These results generally suggest that a good agreement exists between the classification and the actual land cover categories with few misclassifications occurring across nearly all categories. For instance, although the producer’s accuracy for farmland and/or pasture class was 85.3%, the user’s accuracy is slightly lower (82.9%). However, the bare ground category had a perfect producer’s accuracy (100%), but a very low user’s accuracy (60%). This means that although 100% of the bare ground class was correctly identified only 60% of the areas labeled bare were actually bare implying a significant misclassification of pixels in that category.

The confusion matrix for the 2000 land cover/use classification map (Table 4) was a slight improvement on the 1986 classification. It had an overall classification accuracy of 88% and a Kappa coefficient of 0.85. In this classification the reference data and the classified groups were largely in agreement probably because of the use of improved spatial accuracy due to improved methodology i.e. the use of GPS [2000] vs. topographic maps [1986].
5.4 Results and Discussion

The major output of this study is the compilation of change detection maps for the Lake Baringo catchment. Comparison of land cover/use classification maps (see figure 5.4) for 1986 and 2000 were the basis for the change detection output (Figure 5.5) obtained through simple image differencing. Table 5.5 is a summary of the change in land cover in the areas that could be described as hotspots of change in land cover within the catchment (labeled A, B, C, D, E, and F in Figure 5.5). The change detection map (Figure 5.5) shows that many areas in the catchment underwent change in land cover/use over the 14-year period. However the bulk of the changes are confined to a few hotspots, the most important being the highlands that define the upper reaches of the catchment, the lake surface, and the central plains south of the lake.

There was a significant reduction in the total surface area of Lake Baringo between 1986 and 2000. The land cover change detection analysis revealed that approximately 1518 hectares (~10.8%) of the lake surface was lost or converted to wetland over the fourteen year period. Much of the change (represented in red) occurred in the shallower parts of the lake (southern and eastern shores) (Figure 5.6). The reduction in the surface area of the lake may be attributed to one or a combination of three possible factors: (1) infilling by sediments brought in by rivers that feed the lake, (2) reduction in the volume of water in the lake as a result of a decrease in precipitation, and (3) an increase in evapotranspiration. The rainfall receipt in Baringo is consistent with the general East Africa rainfall pattern characterized by variations on annual and interannual timescales correlated to the phase of El Nino/South Oscillation (McHugh 2006; Mutai and Ward 2000; Ogallo 1988). Rainfall variability is most pronounced in the
Table 5.5 The results of change detection in the forest, bare ground, and wetland land cover types based on sites A to F in the change map (see figure 5.5).

<table>
<thead>
<tr>
<th>Site</th>
<th>Land cover type</th>
<th>1986 (Hectares)</th>
<th>2000 (Hectares)</th>
<th>Change (Hectares)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A. Lake</td>
<td>Water surface</td>
<td>14100</td>
<td>12582</td>
<td>-1518</td>
</tr>
<tr>
<td></td>
<td>Wetlands</td>
<td>1951</td>
<td>1230</td>
<td>-721</td>
</tr>
<tr>
<td>B. Njemps flats</td>
<td>Forest</td>
<td>751</td>
<td>453</td>
<td>-298</td>
</tr>
<tr>
<td></td>
<td>Bare grounds</td>
<td>633</td>
<td>5492</td>
<td>4859</td>
</tr>
<tr>
<td>C. Tugen hills</td>
<td>Forest</td>
<td>30753</td>
<td>17835</td>
<td>-12918</td>
</tr>
<tr>
<td></td>
<td>Bare grounds</td>
<td>3</td>
<td>79</td>
<td>76</td>
</tr>
<tr>
<td>D. Lembus and Londiani Forests</td>
<td>Forest</td>
<td>84529</td>
<td>72703</td>
<td>-11826</td>
</tr>
<tr>
<td>E. Mau forest complex</td>
<td>Bare grounds</td>
<td>17</td>
<td>511</td>
<td>494</td>
</tr>
<tr>
<td></td>
<td>Forest</td>
<td>86503</td>
<td>62789</td>
<td>-23714</td>
</tr>
<tr>
<td>F. Marmanet Forest</td>
<td>Bare grounds</td>
<td>39</td>
<td>1696</td>
<td>1657</td>
</tr>
<tr>
<td></td>
<td>Forest</td>
<td>25386</td>
<td>17214</td>
<td>-8172</td>
</tr>
<tr>
<td>Entire catchment</td>
<td>Bare grounds</td>
<td>66</td>
<td>168</td>
<td>102</td>
</tr>
<tr>
<td></td>
<td>Forest</td>
<td>227892</td>
<td>170994</td>
<td>-56898</td>
</tr>
</tbody>
</table>
Figure 5.5  The change detection map obtained from comparing the 1986 and the 2000 land covers. The sites labeled A to F represent areas that experienced significant change in land cover. More details are provided in Table 5.
Figure 5.6 Subset of change detection map highlighting change in lake surface area and the Njemps plains (south of the lake).
semi-arid environments such as those that define the Baringo lowlands where the long-term average shows a slight downward trend since the 1960s (c.f. Johansson and Svensson 2002) in keeping with the downward trend of precipitation since the 1960s as recognized throughout East Africa (Nicholson 1996).

Despite heightened rainfall variability, records from two weather stations within the Lake Baringo catchment (Figure 5.7) show no evidence of a dramatic reduction in precipitation for the period of interest, making it unlikely that the decrease in the lake surface area is an outcome of reduced precipitation. A more probable cause of the reduction in the surface area of Lake Baringo is infilling by sediments brought in by the rivers that feed the lake. Indeed, one of the lake’s most noticeable features is its extreme turbidity (Figure 5.2). The brightness output from the Kauth-Thomas Tasseled Cap transformation revealed that the albedo of the lake increased significantly between 1986 and 2000 (Figure 5.8), with the average brightness index values increasing from 0.248 and 0.316 for 1986 and 2000, respectively (Figure 5.9). To test the significance of change in brightness values for 1986 and 2000 we used SPSS software to perform a t-test on readings from 100 randomly selected points throughout the lake. The results of the t-test revealed that the brightness values were significantly different \( t = -84.699, \ p < 0.001 \) at the 0.05 significance level, indicating that the change in the TC-brightness index can not be explained by chance.

High turbidity is further confirmed by Secchi disk readings from the lake that were approximately 3 cm in August 2003 (Hickley et al. 2004). The increased turbidity of the lake can probably be best explained by an increase in wind driven re-suspension resulting from the decreasing depth of the lake. Similar observations about sediment
Figure 5.7  Rainfall data from two weather stations in the Lake Baringo catchment i.e. Marigat station (970 m above sea level) situated less five kilometers south of the lake and Nakuru (1915 m above sea level) approximately 120 kilometers south of Lake Baringo. The annual rainfall amount in the two stations between 1986 and 2000 does not deviate from the earlier period beginning 1960.
Figure 5.8  Tasseled cap brightness transform (proxy for albedo) images for Lake Baringo surface for 1986 (on the left) and 2000 (on the right) displayed in grey scale. The lake surface in the 1986 image is darker when compared to the 2000 image which is consistent with increase albedo of more turbid water in the latter.
Figure 5.9  Tasseled Cap Brightness values along a North-South transect in Lake Baringo. The TC Brightness values obtained from the 1986 TM image were significantly different from those obtained from the 2000 ETM+ image ($t = -84.699$, $p < 0.001$).
re-suspension and turbidity have been made in the Atchafalaya-Vermilion Bay region in Louisiana (Walker and Hammach 2000). Sediment cores collected from different parts of Lake Baringo under different depths show average sedimentation rates of up to 1 cm/year (Figure 3.6). These phenomenal sedimentation rates suggest that land degradation, especially soil erosion in the catchment area, is responsible for the high turbidity in the lake.

The results of the change detection procedure, NDVI analysis (Figure 5.10 and 5.11), and BI provide evidence for land cover change that may well indicate increased deforestation and land degradation in the Lake Baringo catchment (Table 5.5). Considering that the primary objective of this study was to investigate changes related to land degradation we focused on changes linked to forest loss or gain as well as increases in bare ground cover. Overall there was an increase in the size of the land cover class classified as bare. The bare ground class encompassed areas that were either entirely or nearly devoid of vegetation as characterized by NDVI values below 0.1 (Figure 5.10). The change detection map (Figure 5.5) revealed that the Njemps area (Figure 5.6) in the plains immediately south of the lake, was the most affected in terms of land degradation, as large areas that were classified as pasture/grassland or bushland in the 1986 image had been converted to the bare ground class in the 2000 image. These changes resulted in the bare ground class increasing from 633 hectares to 5492 hectares from 1986 to 2000.

Field surveys in the Njemps flats in December 2003 and January 2005 confirmed that most of the areas depicted as bare were indeed heavily degraded. Much of the area was not only bare or scantily vegetated but was scarred by different forms of soil erosion, especially gully erosion (Plate 5.1). Increasing human and livestock populations in the
Figure 5.10  Map of classification based on NDVI values for 1986 and 2000 shows an increase in bare ground cover (brown) and pasture/farmland/other cover type. However, most of the change has values of less than 0.3.
Figure 5.11 The NDVI difference between 1986 and 2000, only differences >0.3 are displayed. Although changes in NDVI were observed throughout much of the catchment on the previously forested sites and the lake shore recorded values above 0.3.
area subject the vegetation cover to intense pressure. For instance, the population of Baringo District was 203,762 in 1979 and 220,922 in 1989. By 2004 the population was estimated at 302,245 and was growing at a rate of 2.65% per annum. The corresponding livestock numbers have been consistently large; approximately 900,000 goats, 200,000 sheep, and 300,000 cattle (Kenya Ministry of Planning and National Development 2004; Hickley et al. 2004). The inevitable outcome of such mounting human and livestock population is pressure on the limited vegetation resources in the catchment. Immense pressure on vegetation in the form of overgrazing, and deforestation coupled with the soil properties of the clay loams and silt (c.f. Vrieling 2006; Lal 2001) that characterize the area, steep slopes, and the high intensity and sporadic rainfall in the fragile semi-arid environment is a perfect recipe for soil erosion and land degradation.

However, not all areas classified as bare ground (Figure 5.6) depicted in low NDVI values are necessarily degraded. For instance, the Perkerra irrigation scheme near Marigat, in the Njëmps plains was classified as bare grounds because the land had been ploughed for irrigation farming. This was also true for most of the bare ground in the Tugen Hills, and the southern plains near Nakuru where rain-fed agriculture is practiced. However, these misclassifications are of minor importance as the cultivated farms represent only a small proportion of the bare ground class and can easily be identified by their rectangular and linear shapes in the change detection map. It was also noted during fieldwork that many parts of the Baringo plains classified as pasture and bushland had suffered livestock-induced sheet erosion that did not appear in the change detection map. Apparently, the network of acacia bushes there remained largely unchanged over the
Plate 5.1    Large sections of the areas classified as bare in this study such as this picture of Njemps flats, the plains to the south of Lake Baringo (labeled B in figure 5), were also severely degraded.
14-year period preventing the land from more severe forms of soil erosion. The NDVI values also remained largely unchanged between 0.2 and 0.4 (Figure 5.10).

Although forests constitute less than 10% of the total land cover in the Baringo catchment they play a crucial role in the ecosystem. They are a biodiversity and genetic resource seen as key to poverty alleviation (Food and Agriculture Organization (FAO) 2003). Forests are an important natural resource that provides both material goods and environmental services such as soil protection and erosion control. Deforestation is a proxy for land degradation because it is often accompanied by an increase in soil erosion and sediment delivery from river systems (c.f. Kasai et al. 2005; King et al. 2005). Forest clearance in the Lake Baringo catchment between 1986 and 2000 has probably led to a large enough increase in fluvial input of sediments into Lake Baringo to significantly alter the albedo of the lake (Figure 5.8).

The rate of sediment delivery in the Lake Baringo catchment must have increased following deforestation in the upper reaches of the catchment leading to increased surface flow coupled with increased riverbank erosion (Plate 5.2). It is easy to identify areas of forest loss in satellite imagery because forest vegetation is associated with NDVI values above 0.56 (Figure 5.11) while non-forest classes have lower values. Change in NDVI values often points to change in land cover/use. The changes in forest cover were most pronounced in Marmanet forest (Table 5.5) which is the main catchment of Ol Arabel River where over 32% of forest cover was lost between 1986 and 2000. Significant decreases in forest cover were also evident in the Mau (27.4%) and Lembus and Londiani (14%) forests (Table 5.5), which are part of the highland chain that constitutes the main catchment for Molo and Perkerra Rivers, the only permanent rivers in the Lake
Plate 5.2 Picture of riverbank erosion on a section of (a) Endau and (b) Perkerra Rivers. The former dries out completely while the latter experiences remarkably low discharge during the dry season. Most of the eroded sediments end up in Lake Baringo.
Baringo catchment.

The most significant forest cover change was in the Tugen Hills (42%), another important catchment for the lake. The Tugen Hills are characterized by very steep slopes, some of which are severely eroded following deforestation but do not show up in the Landsat imagery, due to the limitation of the spatial resolution. Forest clearance in the Lake Baringo catchment is largely due to pressure to increase land for agriculture or pasture to accommodate the growing population. Although wood is the inhabitants’ primary energy source its collection does not contribute to land degradation due to its abundance, especially in lowlands. A much more important contribution to land degradation is logging for charcoal which is then sold to urban residents. Charcoal stacks abound in roadsides throughout the Baringo plains that are most severely degraded. Informal interviews with some inhabitants of the lower reaches of the catchment gave the impression that seasonality of the ephemeral rivers has increased, and that the dry season flow in the permanent rivers has decreased remarkably in recent years. Forests enhance infiltration capacity and wetness of soils, which coupled with higher water table under forests increase the ability of a forested catchment to support dry season flow in rivers (c.f. Sandström 1995). Deforestation has increased overland flow and decreased infiltration in areas previously under forest cover leading to decrease in ground water recharge and dry season discharge.

5.5 Conclusion

This study shows that Landsat TM and ETM+ can be used to delineate land cover change in ecosystems threatened by land degradation in semi-arid environments such as Lake Baringo. The different methodologies used in this study show that the nature of
land cover/use and associated impacts influence geomorphic processes associated with land degradation. Our analysis revealed that the TC-brightness (albedo) index can be useful for measuring changes in turbidity. Both the NDVI differencing and post classification comparison effectively depicted the hotspots of land degradation and land cover/use change in the Lake Baringo catchment. Change detection analysis showed that the forest cover was the most affected class, in some sections recording reductions of over 40% in a 14-year period. Deforestation and subsequent soil erosion or land degradation has increased the sediment yield in the lake. Land degradation coupled with the high evapotranspiration rates has led to reduction in lake surface area and increased turbidity as confirmed by the statistically significant increase in the albedo of Lake Baringo between 1986 and 2000.

Our results reveal that deforestation and land degradation are serious problems in the Lake Baringo catchment, which poses a threat to the rich biodiversity in this ecosystem. Although climatic variations may account for some of the changes in the lake catchment, most of the changes in land cover are inherently linked to mounting human and livestock populations in the Lake Baringo catchment. Urgent steps need to be undertaken to help prevent further damage to the Lake Baringo area and possible environmental catastrophe.

5.6 References


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CHAPTER 6. SUMMARIES AND CONCLUSIONS

The aim of this dissertation has been to reconstruct the vegetational history and land degradation in the Lake Baringo area and enhance our understanding of the paleoenvironment of East Africa during the Late Holocene. The different chapters of this dissertation have largely focused on paleoenvironmental reconstruction based on the lithostratigraphic record, pollen, spores, and microscopic charcoal record derived from lake sediment cores that were collected from Lake Baringo. More recent changes in the area have been provided by evidence from satellite imagery that covered the Lake Baringo drainage basin in its entirety.

To put the recent environmental changes in Lake Baringo area into context I provided a synoptic coverage of paleoenvironmental changes in East Africa during the Late Quaternary through a review of literature. Different proxy evidences offered a detailed summary of paleoclimatic changes in East Africa from beyond 42,000 yr BP to the present millennium. Although only a handful of these records from East Africa extended beyond 42,000 yr BP available records have presented useful data for the reconstruction of East Africa’s paleoenvironment beyond the Last Glacial Maxima (LGM). It is apparent that the period prior to ca. 42,000 $^{14}$C yr BP was characterized by warm climatic conditions similar to the present. The shifting of high altitude vegetation belts to lower elevations between 42,000 $^{14}$C yr BP and 30,000 $^{14}$C yr BP marks a change to cold dry conditions. Much of this evidence was obtained from pollen and diatom data as well as charred grass cuticles and is verified by carbon isotope ($\delta^{13}$C) record from the region. Cool conditions prevailed in East Africa in the period leading to the LGM with temperatures being between 2 to 4.1$^{0}$C colder than present. Between ca. 21,000 and
12,500 $^{14}$C yr BP East Africa’s environment was generally cool, punctuated by episodes of prolonged desiccation.

East Africa experienced two extended episodes of desiccation that are unprecedented during the Last Glacial Maxima (LGM). Evidence from two of the largest lakes in the region, Lakes Albert and Victoria, place the first episode at ca. 18,000-20,000 $^{14}$C yr BP and the second at ~15,390 +/-220 $^{14}$C yr BP and ~11,710 +/-120 $^{14}$C yr BP. The intense aridity during those two periods resulted in a decrease in water level at Lake Victoria to at least -66 m below present lake level. It has been postulated that the first and second episodes of aridity could be synchronous with the LGM and the Younger Dryas interval, respectively. However, more longer and continuous records are required before the hypotheses can be confirmed. The terminal phase of the last glaciation and the onset of the Holocene was marked by an increase in temperature and moisture. Evidence from different biological proxies, and oxygen isotope records from diatomic silica, suggest that during the early to middle Holocene period East Africa was characterized by warm and moist conditions amid rapid climatic changes.

Two significant abrupt drought events occurred in East Africa during the early to middle Holocene period, one at ca. 8,300 BP and the other at 5,200 BP. These episodes appear to correspond with an apparent change in sedimentation rate at Lake Albert sometime between 8,000 and 3,400 $^{14}$C yr BP. Although these drought events are well marked in the ice core record from Mount Kilimanjaro they are missing in the pollen records from the region. This emphasizes the importance of proxy sensitivity and sampling resolution as it affects paleoenvironmental data interpretation, and the need for a multi-proxy investigative approach. The longer and more extensive aridification event
at ca. 4,000 $^{14}$C yr BP was registered in both the pollen and diatom records at different sites in East Africa. Many records from the region generally show an aridification trend that began at ca. 4,000 $^{14}$C yr BP and continued to the present, albeit with oscillations.

Although most of the vegetational changes in East Africa prior to ca. 4,000 $^{14}$C yr BP can be attributed to climate changes the same cannot be said for the following period. The period after ca. 4,000 $^{14}$C yr BP saw the influence of humans progress from relatively minor and restricted to a few ecosystems, to becoming a major force across even remote and inhospitable environments. Records from different parts of east Africa during this period show reduction in arboreal taxa and an increase in fire episodes (charcoal peaks) that have been interpreted as human-induced. These interpretations are based on the presence of high counts of microscopic charcoal (suggesting fire occurrence) concurrent with forest clearance. In some cases (e.g., Vincens et al. 2003) forest clearance was accompanied by increased pollen frequencies of cultivated crops (e.g., *Elaeis guineensis*), Gramineae, and disturbance markers (e.g., *Ricinus communis* that may indicate forest clearance for agricultural and/or pastoral activities).

Given the reality of climate variability in East Africa at all time scales and the increasing human impact on the environment, the challenge for paleoecologists is reconstruct the paleoenvironment with a time resolution sufficient to illuminate the causal mechanism of vegetational change. The multi-proxy high-resolution record from Lake Baringo is a major contribution to our understanding of the more recent changes in the paleoenvironment of East Africa. The difficulty of separating purely background-level environmental or climatic change, from that caused by humans has been evident throughout our discussion.
The Lake Baringo record presented in this dissertation has revealed a largely dry environment that is punctuated by a succession of centennial- to decadal-scale wet and dry episodes, separated by rather sharp transitions. The pollen sequence from Lake Baringo showed a general dominance of grass pollen with percentages averaging 20-40% and lower percentages of arboreal pollen (less than 20%) throughout the core. The pollen assemblage basically suggests that the present savanna-like landscape prevailed in the ecosystem throughout the time covered by the cores. This finding implies that the hypothesis that the present environment is largely a consequence of deforestation may be exaggerated. However, the prevalence of high counts of microscopic charcoal throughout much of core LB-5 suggests that the presence of fire may have facilitated the prevalence of grasses and herbaceous vegetation in the Lake Baringo area at the expense of forests. The fires that supported the dominance of the savanna-like environment are likely to be of anthropogenic origin because lightning generated fires are presently rare. The fires may have been used to clear the landscape of ticks or accidental in the process of harvesting honey. This implies that over time humans promoted the savanna/bushland landscapes through activities that may not have necessarily involved deforestation. It is also possible that the climatic conditions in the Lake Baringo area throughout the time covered by the cores analyzed in this dissertation were incapable of promoting the growth and/or thriving of vibrant forest communities.

The lithostratigraphic record from the lake shows what appear to be abrupt intense drought episodes prior to ca. AD 1650 and AD 1750-1820. The two drought episodes were more severe than all the droughts that have been recorded in the region since the nineteenth century. These dry episodes in the Lake Baringo are marked by rather sharp
changes in the stratigraphy from lake mud to sediments that are characterized by very hard clay material with crumb structure and rootlets. Well preserved seeds and plant fibers were recovered along with the hard clay at the top of the horizon that marked the first dry episode. The crumb structure (a typical pedogenic feature), traces of rootlets, and plant fibers are consistent with the drying out of the lake at the core sites and subsequent recruitment of wetland vegetation communities. The drying of the lake at the core sites requires a lowering of lake level by at least -6 m at the peak of the first drought episodes at ca. AD 1650 and AD 1750. That drop in lake level and subsequent conversion of lake surface into marsh environment would have reduced the lake surface to 1/3 its present size.

The high–temporal-resolution palynological record from Lake Baringo confirms the stratigraphic changes. The pollen record shows variations in relative abundances of local and regional pollen taxa that track the changes in lake level fairly well. The periods culminating in the two dry episodes at ca. AD 1650 and 1750-1820 were characterized by relatively high percentages of Gramineae, Cyperaceae, Combretaceae, *Podocarpus*, *Olea*, *Tarchonanthus*, *Acacia*, and *Euclea* within a poor preservation environment. Such pollen assemblages are typical of dry environments in East Africa. Although high frequencies of Gramineae and Cyperaceae pollen are normally associated with dry environmental conditions, in this particular case they are probably a result of marsh colonization of large portions of dried lake surface. The poor preservation of pollen, supported the great abundance of unidentifiable and corroded pollen, is consistent with poor preservation due to low water levels and exposure to an oxidizing environment. However, the climatic implications of the changes in pollen spectra in the Lake Baringo
record are limited by the high sedimentation rates in the lake which limited the focus of this study to the last 300 $^{14}$C yr BP.

The results obtained in this study show that Lake Baringo records a sedimentation rate of 1 cm yr$^{-1}$, which is one of the highest sedimentation rates in tropical Africa. Assuming uniform sedimentation rates throughout the lake, then roughly over 1290000 m$^3$ of sediments are washed into Lake Baringo every year. The lake showed no change in sedimentation rates for both the prehistoric period, prior to European settlement in the Baringo ecosystem (AD 1900) and after, especially during the more recent times. One of the implications of this finding is that land degradation in the Lake Baringo area commenced much earlier than it has been previously documented, and the rates of pre-historic land degradation are similar to those evident in modern time. This then implies that the erstwhile romanticized hypothesis of the invariably environmentally friendly use of grazing resources by the indigenous pastoralists and their indigenous knowledge of conservation needs to be revised, and that the indigenous pastoralists and farmers were not simply victims of colonialism but were always active agents and shapers of the degraded landscape presently evident in the region. The alternative possibility is that humans are not the main drivers of land degradation the ecosystem, but that the steep landscape, sparse vegetation cover, clay and volcaniclastic material, and the climate with the Lake Baringo area interact to yield the high sedimentation rates independent of anthropogenic impacts.

Satellite imagery analyzed in this dissertation revealed a decrease in the surface area of Lake Baringo of up to 11% or 1518 hectares between 1986 and 2000. The decrease in the surface area is largely a consequence of increased sediment delivery into
the lake confirmed by a significant increase in the turbidity of the Lake Baringo waters. The high turbidity is linked to increased soil erosion and land degradation within the ecosystem due to deforestation and pressure on vegetation resources from mounting human population in Baringo area. Remote sensing data showed evidence of massive deforestation in the upper reaches of the Lake Baringo drainage basin and extensive degradation of the lowland plains in the vicinity of the lake which must be checked to alleviate imminent ecological disaster that threatens the rich biodiversity its ecosystem. Lake Baringo requires comprehensive, participatory, and catchment-wide management effort that integrates conservation and stakeholder priorities that are based on sound science.

6.1 Directions for Future Research

This dissertation research has contributed to our understanding of the paleoenvironmental changes in East Africa, during more recent times. The high-temporal-resolution analyses and the multi-proxy approach adopted in the study have unraveled decadal-scale climatic and vegetational changes that may have been missed in previous studies. The coarse-temporal-resolution of previous studies from the region probably contributed to these changes remaining unnoticed. The multi-proxy approach adopted in this dissertation research needs to be replicated in different sites from East Africa to corroborate the findings of the Baringo record. This will help answer the following question: What was the spatial extent of the decadal-and sub decadal-scale climatic and vegetational changes in Lake Baringo record, were the changes confined to the Baringo area or were they more regional? More high-temporal-resolution lake sediment studies from different sites in East Africa are needed to address that gap. One
of the potential sources of paleoenvironmental data of such fine resolution is ice core record from Mount Kilimanjaro. Already the groundwork has been laid by Lonnie Thompson and his team from Ohio State University. However, ice core studies must be done sooner because the icecaps in East Africa are receding fast and may disappear within the next two decades.

Whereas humans have contributed to vegetational changes in the Lake Baringo record, it is evident that climate variability was the prime mover of change. Unfortunately, the climatic implications of the Lake Baringo record are limited by the high sedimentation rates at the lake which limited the focus of this study to the last 300 $^{14}$C yr BP. More studies are needed to investigate the force behind the climate variability. There is also need for more higher-temporal-resolution palynological studies with an emphasis on multi-proxy methodologies in order to improve our knowledge of the human/climate impact in the region. We still need to devise methods to separate purely climatic and human-driven environmental changes.
APPENDIX A: POLLEN NOTES

*Justicia* [ACANTHACEAE] ~60 x 35

Monad, pollen grains isopolar and radiosymmetric; verrucate (verrucae variable in size); usually diporate but can also be heterocolpate, the pores are circular. *Justicia* is the largest genus in the family Acanthaceae with over 600 species and is widely distributed in the tropics and subtropics of both hemispheres. *Justicia* genus widely distributed in East Africa is mostly constituted of erect shrubs or subshrubs which gain heights of 1-3 m.

*Podocarpus* [PODOCARPACEAE] ~ 45 x 75 µ

Monad, heteropolar, bilateral; vesiculate, corpus provided with a pair of laterally and somewhat distally attached sacci; mostly psilate, sacci reticulate, grain length 53-86 µ, corpus length 38-50 µ, saccus width 34-50 µ. *Podocarpus* is a wind-pollinated genus of gymnosperms producing
winged pollen that is often abundant in pollen diagrams from East Africa. *Podocarpus* pollen cannot be confused with any of the other indigenous East African plant. The recently introduced *Pinus* (early 20th century) produces similar pollen but is much larger. Only two species of *Podocarpus* are widely distributed in East Africa, i.e., *P. gracilior* which is typical of drier montane forests, and *P. latifolius* which prefers wetter conditions.

*Pinus* [PINACEAE] ~ 90-120 × 50-80 µ

Monad, heteropolar, bilateral; vesiculate, two-saccate, sacci divergent, placed laterally; sacs coarse reticulate; corpus psilate-microreticulate. *Pinus* is an introduced species in East Africa; does not appear in the pollen record until after AD 1900.

AMARANTHACEAE ~ 16-23 µ

Monad, apolar; spheroidal; psilate, also scabrate-verrucate; periporate, pore number of the order 22-45, circular pores (or poroids) 2-4 µ diameter. Most members of this family are weedy herbs (mostly ephemeral) and have pollen that is virtually indistinguishable from Chenopodiaceae. Thus the two groups are mostly considered together in pollen diagrams. They are wind pollinated producing enormous amounts of pollen. Large representation in the pollen record from lake sediments from East Africa often signifies arid conditions. One of the readily identifiable species of the Amaranthaceae family is *Alternanthera*.
*Alternanthera* spp. [AMARANTHACEAE] ~ 22-30 µ

Monad, periporate, 14-16 pores, recessed in lacunae of a coarse-ridged reticulum, polyhedral. Members of this species are mostly shrubs. *Alternanthera* pollen was infrequent in the Lake Baringo record.

**CHENOPODIACEAE ~ 19-33 µ**

Monad, apolar; spheroidal; exine thin and granular, psilate or scabrate; periporate, pores slightly recessed, 1-4 µ in diameter, of the order of 14-α. The grains of this family are rather uniform and indistinguishable from those of the related family Amaranthaceae. Most of the members of Chenopodiaceae in East Africa are annual (mostly ephemeral) or perennial herbs with wide distribution. As in Amaranthaceae, large representation in pollen records from lake sediments is associated with prevalence of arid conditions.

*Typha* [TYPHACEAE] ~ 20 x 25 µ
Monad or tetrad, heteropolar, radio-symmetric; monoporate, outline of pore irregular; polyhedral; reticulate, heterobronchate. *Typha* species occupy wetland environments but not very well represented in the Baringo record.

*Juniperus* [CUPRESSACEAE] ~ 22-26 µ

Monad, apolar; inaperturate, spheroidal; exceedingly thin and transparent exine, easily ruptured (as in picture above). Only one species of the Cupressaceae family is native to East Africa i.e. *Juniperus procera*, commonly known as the East African Juniper. *Juniperus procera*, a medium-sized tree reaching 20-25 m (rarely 40 m) tall, is a constituent of the montane forest. It is the only juniper known to occur south of the equator, and is thought to be a relatively recent colonist of Africa; the species shows very little of the variability associated with a long period of evolution (Farjon 2005). Presence of Juniperus pollen in the pollen record suggests presence of dry montane forest. In the early 20th century other species of the Cupressaceae family were introduced into the East African environment by Europeans. The pollen of the introduced species is virtually indistinguishable from the native species.

*Phoenix* [PALMAE] ~16 x 18 µ

Monad, heteropolar; radiosymmetric; trichomonocolpate, colpi reaching close to the equator, colpus ~ 11 µ long, narrow but often splayed; subrectangular; elliptical circular, psilate. *Phoenix* is a component of the dry deciduous forest. The species is also found in moister savanna environments of east Africa.
CYPERACEAE ~ 10-34 x 15-41 µ

Aberrant tetrad, pseudo-mond; subtriangular, pear shaped; periporate, three lateral and one basal pore, lateral pores elongate-linear or indistinguishable; sometimes may appear inaperturate. Members of this family thrive in marshy and swampy environments.

Balanitaceae [BALANITACEAE] ~ 22 x 23 µ

Monad, oblate spheroidal; tricolporate; reticulate. The most common species in East Africa is *Balanites aegyptiaca* (commonly known as the desert date). This is an important thorny shrub or tree in the semi-arid ecosystems with many beneficial attributes. The species occurs often in association with *Acacia* and *Commiphora* species, and other dry-indicator species.

*Euclea* [EBENACEAE] ~ 20 x 23 µ

Monad, suboblate to oblate spheroidal; tricolpate; mostly psilate; subtriangular polar view, semielliptical equatorial view. The genus *Euclea* is widely distributed in the savanna environments of East Africa.

COMBRETACEAE ~ 20 x 26 µ
Monad; usually has up to six colpi, but also tricolporate; barrel-shaped (subprolate), subprolate or suboblate; psilate, also reticulate-polybrochate; sexine thinner than nexine. Combretaceae pollen is readily identifiable because of its shape and pattern especially from the polar view. Members of this family in East Africa are mostly trees or shrubs the most common species being those of *Combretum* and *Terminalia* in forest and savanna environments.

![Combretum and Terminalia pollen](image)

*Combretum* and *Terminalia* pollen are readily identifiable due to their distinctive shape and pattern.

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Monad, oblate spheroidal; usually triporate but can sometimes have between 1 and 7 pores or pseudopores (i.e., stephanoporate or periporate), pores irregularly arranged, longest axis 10-17 μ; usually psilate. There are over 25 Acalypha species (mostly herbs or shrubs) in East Africa, found in both forest and savanna environments. Some *Acalypha* species e.g., *A. brachystachya* and *A. psilostachya* can also be found on cultivated grounds (c.f., Agnew 1974).

![Acalypha pollen](image)

*Acalypha* [EUPHOBIAEAE] ~ 12 x 13 μ

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Monad, oblate spheroidal, amb circular; reticulate; tricolporate with poorly defined endoapertures, transversely elliptic aperture. *Macaranga* is a light loving member of the montane forest in its Africa often competing with *Hagenia* species. The presence of *Macaranga* pollen in the records usually signify human disturbance of forests or deforestation.

![Macaranga pollen](image)

*Macaranga* [EUPHOBIAEAE] ~ 14 x 13 μ
Phyllanthus [EUPHOBACEAE] ~ 16 x 18 µ

Monad, oblate spheroidal, amb circular (~18 µ); reticulate. Phyllanthus is a typical bush and/or savanna shrub that is widely distributed in the semi-arid ecosystems of east Africa. It often occurs in association with different species of Tarchonanthus, Acacia and Commiphora in the transition zone between forest and savanna.

Ricinus communis [EUPHOBACEAE] ~ 26 x 28 µ

Monad, suboblate to oblate spheroidal; triporate; fine reticulate. Ricinus communis (whose common name is Castor), may be found growing wild in many of the warmer regions of the world; the species probably originated in East Africa, but has become naturalized in many environments following early introductions (Perry 1972; Robellen et al. 1989; Salunkhe et al. 1992). The plant is valued for the pale yellow seed oil, which has long been used for medicinal purposes and is now of importance in a wide variety of technical applications due to its unique content of ricinoleic acid. Presence of Ricinus communis in the pollen from East Africa record is usually a marker for agricultural activities.

Acacia [MIMOSACEAE] ~ 74-96 x 62-72 x 43-50 µ
Polyad, in the order up to 64 individual grains, arrangement of grains lack order and symmetry; mostly inaperturate; psilate. Acacias are the dominant woody plants over much of the semi-arid environments of East Africa. They basically define the savanna environments with their characteristic exaggerated umbrella-shaped crowns which enable the tree to capture the maximum amount of sunlight, with the smallest possible leaves. Some species have deep tap root systems, allowing them to survive in semi-arid terrain along dry watercourses (they tap into the water table) while others have shallow but extensive splayed root systems to capture limited rainfall in dryer areas. Presence of large amounts of *Acacia* pollen in the pollen record signifies increased aridity of the environment.

Poaceae [GRAMINEAE] $\sim < 35 \mu$

Monad, spheroidal; psilate-scabrate; heteropolar, radiosymmetric; monoporate, pore circular with distinct annulus. Gramineae pollen is readily identifiable but difficult to subdivide usefully except for maize (*Zea mays*). The latter is distinguished from other grass pollen based on size, being mostly $> 50 \mu$ while other Gramineae pollen rarely exceeds $37 \mu$. Grasses are widespread and abundant in most vegetation types in East Africa. Large percentages of Gramineae in the pollen record often signify arid conditions.

*Zea mays* [GRAMINEAE] $> 50 \mu$

This species is readily distinguished from the other members of Gramineae family because of its enormous size.
**Celtis** [ULMACEAE] ~21 x 23 µ

Monad; usually triporate (most of those in Lake Baringo had 3 pores) but can have between 2 and 6 pores, pores easy to see under x 400 magnification, annulate; scabrate, suboblate to oblate spheroidal, polar axis 17-25 µ, equatorial axis 20-27 µ. Seven species of Celtis occur in east Africa i.e., *C. adolfi-fridericii*: a large canopy tree of lowland forest, *C. africana*: medium sized tree of lowland forest, riverine forest, and montane forests (extends higher than other Celtis species, up to 2400 m), *C. durandii*: medium sized tree of lowland and montane forests, *C. integrifolia*: medium sized tree that usually occurs in dry riverine forest and deciduous woodland including savanna, *C. mildbraedii*: large canopy tree of lowland forest, *C. wightii*: small to medium sized tree of lowland forest, and *C. zenkeri* large canopy tree or second-story tree of lowland forest.

**Myrica** [MYRICACEAE] ~ 27 x 30 µ

Monad, suboblate to oblate spheroidal; usually triporate; subtriangular in polar view; pores elongated towards the poles, protruding pores, sexine thickened around pores making them look slightly annulate; scabrate. *Myrica* pollen is readily recognizable. Only two species are widespread in East Africa i.e. *Myrica kandtiana* and *M. salicifolia*. The two have very similar pollen but thrive in different ecological conditions. *Myrica kandtiana* is a swamp plant found at relatively low latitudes while *Myrica salicifolia* is dryland species.
Sesbania [FABACEAE] ~ 30 x 26 µ

Monad, suboblate; tricolporate; scabrate; subrectangular polar view. Species of *Sesbania* are mostly annual or perennial herbs or trees that are widely distributed in East Africa. Some species are aquatic or semi aquatic growing along rivers and on lake shores. *Sesbania sesban*, a short-lived shrub or small tree up to 8 m tall, is the most common species in East Africa. Presence of Sesbania in the pollen record is difficult to interpret but is often associated with wetland communities.

MYRTACEAE ~ 19-22 x 26-32 µ

Monad, isopolar, radiosymmetric; oblate, amb triangular-subtriangular; psilate-scabrate; syncolporate, colpi fused across the polar areas in such a way as to delimit a triangular boundary. Myrtaceae pollen is readily identifiable by the triangular formation in the middle of the pollen. Members of the Myrtaceae family are mostly trees and shrubs that are widespread in East Africa. Important members of this family include *Hagenia abyssinica* (a light requiring species), a spreading tree of up to 25 m height. Its presence in pollen diagram indicate presence of montane forests or sometimes, human disturbance. *Eucalyptus*, an introduced species in East Africa belongs to the Myrtaceae family.

Prunus [ROSACEAE] ~ 21-24 µ
Monad, isopolar, radiosymmetric; tricolporate, colpi long, somewhat narrow; prolate, amb circular; striate. *Prunus* is an important component of the montane forest in East Africa. Pollen closely resembles others from Rosaceae family and may be difficult to isolate. High percentages of *Prunus* in the pollen record usually suggest presence of montane forests, but are very poorly represented in pollen record from Lake Baringo. *Prunus africana* (a tree that grows up to 30 m), an important constituent of the montane forests in East Africa, has recently been overexploited for its medicinal value (bark extracts) and is currently listed among the endangered species (Schippmann 2001; Acworth 1999).

**LORANTHACEAE ~ 25 x 50 µ**

Monad, isopolar, radiosymmetric; oblate, amb triangular with sides concave or straight; tricolpate, colpi long, narrow. Pollen of the Loranthaceae family is readily identifiable because its unique shape from the polar view. The Loranthaceae family is constituted of woody plants, mostly hemiparasites including the mistletoe species. They are rare in the pollen record from Lake Baringo and East Africa in general.

**Croton spp. [EUPHOBIAE] ~ 60-80 µ**

Monad, apolar; spheroidal; inaperturate; clavate, thick exine ~ 4-6 µ packed with clavae. Members of this species are shrubs or trees mostly in the dry forests. Pollen of *Croton* species was rare in the Lake Baringo record.
Olea spp. [OLEACEAE] ~

Monad, isopolar, radiosymmetric; heterobrochate, per-reticulate; prolate spheroidal, amb circular; tricolpate, colpi narrow and lengthy. *Olea* pollen is a readily identifiable type. There are three species of *Olea* that are widespread in East Africa i.e. *Olea africana* and *O. hochstetteri*, that thrive in drier montane forests, often occurring together, and *O. welwitschii* that tends to prefer wetter environments. High frequency of *Olea* in the pollen record is indicative dry and cool climatic conditions especially when coincident with high frequencies of *Podocarpus*.

**FERN SPORES**

Polypodiaceae [PTERIDOPHYTA] ~ 44-60 × 26-40 µ

Monad, heteropolar; bilateral or radiosymmetric; oval- or bean-shaped; monolete or trilete, laesura length ~ 16-21 µ; psilate or regulate. Generally these fern spores occurred infrequently in the Lake Baringo record.
Representatives fern spores (monolete and trilete) [PTERIDOPHYTA] in the Baringo record.

**FUNGAL SPORES**

*Sordaria*-type ~ $18-31 \times 10^{-17}$ μ

Ascospores ellipsoidal; one-celled; brown or dark-brown; thick and dark pore at apex (~1-2 μ); basal end often slightly flattened. The most common members of this species include two homothallic species, *S. fimicola* and *S. macrospora*, and one heterothallic species, *S. brevicollis* all of which thrive in the dung of herbivorous animals. *Sordaria* is an indicator of livestock activity in the drainage basin.
*Tripterospora*-type $\sim 18-25 \times 12-17 \mu$

Ascospore ellipsoidal; one-celled; mostly brown; truncated base, diameter $\sim 10 \mu$; often has hyaline basal pedicel; eccentrically placed 11.5 $\mu$ germ pore at the apex. This species is coprophilous often occurring with *Sordaria* spp.

*Sporormiella*-type $\sim 14-18 \times 12-17 \mu$

Ascospores rectangular or triangular shape; often occurs with lateral furrow; brown to dark-brown color; coprophilous. The dung fungus *Sporormiella* is an indicator of herbivore density, and has been shown to increase in historic times after the introduction of grazing animals or in environments with abundant magafauna.

*Chaetomium*-type $\sim 6-12 \times 4-9 \mu$

Ascomycete; mostly demataceous, that is, they produce dark spores; readily identifiable because of its lemon-shape; single germ spore, often flattened in one plane. *Chaetomium* species are saprophytic cellulose decomposers common on cloth, feathers, leather,
decaying vegetation and plant debris. Presence of ascospores of *Chaetomium* species in sediments often suggests human impact in the drainage basin.

*Cercophora*-type ~ 12-30 × 6-18 µ

Ascospores often slightly truncated at the basal end, apical end tapered; subapical pore ~0.5-0.8 µ; sometimes with one septum with no constriction; may or may not have hyaline appendage at basal side. *Cercophora*-type fungal spores are common in livestock dung but may also thrive on decaying wood or vegetation.

*Glomus*-type ~ 17-138 µ

Arbuscular mycorrhizal fungi; thick-walled, asexual fungal spore that is derived from a hyphal cell and can function as a resting spore; globose shape; always characterized by subtending hyphae; extremely variable in size (17-138 µ). Members of this species normally colonize plant roots, particularly in disturbed habitats. They often form symbiotic relationships with plant roots (mycorrhiza); the fungus mycelium is able to permeate the soil further than the plant's root hairs and so can collect minerals from a larger soil volume. These are made available to the plant while sugars from photosynthesis leak from the roots to the fungus. *Glomus* depends largely on mycorrhizal root debris. Presence of Chlamydomospores of *Glomus* in lake sediments is indicative of soil erosion in the catchment of the lake.
Valsaria-type ~ 30-40 × 11-23 µ

Ascospores are unisepate; has a velum which forms an equatorial ring around the septum. Valsaria spp are associated with eutrophic wet conditions.

Diporotheca-type ~ 47-52 × 18-26 µ

Ascospore ellipsoidal, one-celled with single pores on the basal and apical ends. Spores of Diporotheca sp. indicate eu-mesotrophic aquatic conditions.

Tetraploa-type ~ 35-50 µ

Monad, characteristic two or three forks at the basal side; readily identifiable. Tetraploa spp. Natural habitat includes leaf bases and stems just above the soil on many kinds of plants and trees. The species is often abundant in decaying plant tissue.
Glossary of palynological terms used in the pollen notes (modified from Willard et al. 2004; Faegri and Iversen (1998); (Erdtman 1971); Heusser (1971); Kremp (1965), refer to these texts for details).

Amb: The outline of a pollen grain or spore viewed from above one of the poles. The amb can be round or triangular.

Annulus (pl. annuli): A ring bordering a pore of a pollen grain. It is formed mostly by the thickening of the ektexine around the pore.

Aperture: Modification or an opening of the exine of a pollen grain or spore e.g., pores, colpus or laesura.

Apolar: lacking symmetry.

Baculate: Sculpture consisting of bacula, i.e. tiny rods (lacking thickening or thinning at either end).

Bisaccate: having two sacci.

Brochus (pl. brochi): The part of a reticulum consisting of a single lumen or opening and a portion of the surrounding ridge.

Clava (pl. clavae, adj. clavate): A club-shaped element of the sexine that is higher than 1 um, with diameter smaller than height and thicker at the apex than at the base.

Columella (pl. columellae) – A rod-like element of the sexine supporting the tectum.

Colporate: pollen grain with both colpi and pores often oriented along the equator.

Colpus (pl. colpi): Longitudinal furrow in the exine of pollen grains.

Diporate: Pollen with two pores and no colpi.

Echinate: Sculpture type consisting of echinae (spines).

Ektexine: The outermost layer of the exine.
Equator: Imaginary line connecting points midway between the polar extremities of symmetrical spores or pollen.

Equatorial diameter: A line in the equatorial plane, perpendicular to the polar axis and passing through it. In palynomorphs with bilateral symmetry, the longest line is usually taken as the equatorial diameter.

Equatorial view: The view of a pollen grain where the equatorial plane is directed towards the observer.

Ellipsoidal: Shaped like an ellipsoid.

Endoaperture: An aperture in the inner layer of the sporoderm (nexine).

Endoexine: Innermost layer of the exine.

Exine: The outermost, very resistant layer of pollen or spore.

Globose: Spheroidal or globular.

Granular: Synonymous with scabrate.

Heterobrochate: Reticulate sculpture in which the brochi of the reticulum are not of uniform size.

Heterocolpate: Pollen grains with long and narrow lacunae (pseudo colpi) symmetrically arranged on a colporate type.

Heteropolar: Pollen grain with dissimilar proximal and distal portions as to shape, aperture, and sculpture.

Inaperturate: Without clearly defined apertures.

Isopolar: Distal and proximal portions identical in shape, apertures, and sculpture.

Lacuna (pl. lacunae): Depressed area in the ektexine surrounded by ridges.

Laesura (pl. laesurae): Narrow opening representing an aperture in spores which can be triradiate (in trilete types) or linear (in monolete types).

Lumen (pl. lumina): One of the meshes in a reticulum.

Microechinate: Sculpture consisting of minute spines.

Monad: Pollen grain occurring singly, as opposed to pollen in tetrads or polyads.

Monoaperturate: With single aperture.

Monocolpate: possessing one colpus.

Monocolporate: With one colpus and one pore.

Monolete: Usually in spores with one linear opening.

Monoporate: With one pore.

Murus (pl. muri): A ridge that is part of the ornamentation that separates the lumina in a reticulate pollen grain or the striae in striate pollen grain.

Nexine: The inner, non-sculptured part of the exine which lies below the sexine.

Oblate: Shape of spore or pollen grain where the polar axis is shorter than equatorial diameter i.e. P/E ratio is 0.50-0.75.

Oblate spheroidal: Shape with P/E ratio of 0.88-1.

P/E ratio: The ratio of the polar axis (P) to the equatorial diameter (E).

Pericolpate: With more than three colpi positioned in such a way that some or all cross the meridians of the grain.

Pericolporate: same as preceding but colpi also contain pores.

Periporate: with more than three pores scattered over the surface.

Perprolate: Shape where the P/E ratio is >2.

Per-recticate: semitectate structural type with columellae fused above to form an open reticulum.
**Polar area:** The part of a pollen grain poleward from the ends of the colpi and associated structures.

**Polar view:** A view of a pollen grain in which the polar axis is directed towards the observer.

**Polyad:** Many grains united in multiples of more than four.

**Polybrochate:**

**Polyplicate:** Alternating meridional ridges and grooves.

**Pore:** Circular to slightly oval openings in the exine of a pollen grain.

**Prolate:** Shape of spore or pollen in which the polar axis is longer than the equatorial diameter.

**Pseudoaperture:** A thinning of the exine which, although superficially resembling an aperture, is not equivalent in function.

**Pseudocolpus (pl. pseudocolpi):** A colpus-like pseudoaperture.

**Pseudomonad:** Seemingly freely occurring pollen grain e.g., Cyperaceae, but which actually represents a kind of tetrad that has not separated but instead has remained intact.

**Pseudopore (pl. pseudopores):** A pore-like pseudoaperture.

**Psilate:** Sculpture type characterized by smooth walls of pollen or spores. The term applies also to exines with pits or reticular less than 1 µm in diameter.

**Radiosymmetric:** Shaped in such a way that any plane including the polar axis that passes through produces identical halves.

**Reticulate:** Sculpture type in which pollen or spore have a network of ridges (muri) forming a more or less regular netlike pattern.

**Saccate:** Having sacci.

**Saccus (pl. sacci):** Air sac or bladder, or wing characteristic of vesiculate pollen types.

**Scabrate:** Sculpture consisting of more or less isodiametric projections (scabrae), less than 1 µm in diameter.

**Sculpture:** The surface relief or topography of a spore or pollen.

**Sexine:** The outer, sculptured layer of the exine, which lies above the nexine.

**Spheroidal:** The shape of a spore or pollen grain in which the equatorial axis and the polar axis are more or less equal i.e. circular, P/E ratio of 0.88-1.14.

**Striate:** Pollen sculpture with multiple, more or less parallel ridges or grooves in the exine.

**Structure:** Texture or makeup and distribution of the component parts of the exine.

**Subspheroidal:** Shape of spore or pollen grain with P/E ratio of 0.75-1.33.

**Suboblate:** Pollen shape with P/E ratio of 0.75-0.88.

**Subprolate:** Shape with P/E ratio of 1.44 - 1.33.

**Subtriangular:** An amb view that has an imperfect triangular shape.

**Supra-reticulate:** Structural type with a reticulum arranged distally on the tectum.

**Syncolp(or)ate:** Describing a pollen grain with two or more colp(or)i the ends of which anastomose at the pole.

**Tectum (pl. tecta, adj. tectate):** The layer of sexine which forms a roof, supported by the columellae.

**Tetrad:** Pollen or spore type with four components.

**Trichotomocolpate:** Describing pollen with a three-armed colpus.

**Tricolpate:** Having three colpi.
**Tricolporate:** With three colpi, each containing a pore.

**Trilete:** Spores with triradiate tetrad scar.

**Verrucate:** Sculpture element that is radially isodiametric or non-constricted, the height of which is nearly equal to the diameter.

**Vesiculate:** Possessing sacci.

**Important sources of information for the pollen identification and notes:**


**Other sources of data for pollen identification used in this study:**

1. African pollen collection by Dr. Terry Brncic and colleagues at the Oxford Long-term Ecology Laboratory, Oxford University Centre for the Environment, Dyson Perrins Building, South Parks Road, Oxford, OX1 3QY [Tel lab: +44 (0) 1865 275 883].

2. African Pollen Data Base [http://www.geo.arizona.edu/palynology/apd.html](http://www.geo.arizona.edu/palynology/apd.html) A network serving all scientists involved in African palynology [Unfortunately pollen pictures in this website have been inaccessible for the past three years due to maintenance problems].

3. Pollen reference collection specimens at the Louisiana State University’s Global Change and Coastal Ecology Laboratory (formerly known as Biogeography and Quaternary Paleoecology Laboratory).
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Lawrence M. Kiage, PhD.
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--------------------------------Original Message--------------------------------------

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Editor-in-Chief, International Journal of Remote Sensing

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The manuscript was recently accepted for publication in the IJRS and I am currently finalizing the proofs. This is to request permission from the International Journal of Remote Sensing so that I may include the manuscript in my dissertation.

Thanks,

Lawrence M. Kiage
VITA

Lawrence Morara Kiage was born on 14 November, 1971, in Kisii, Kenya, the second son of George Kiage Ogeto and Hellen Kwamboka Kiage. He spent his childhood in Kapenguria, northwestern Kenya, where he acquired primary and high school education. Upon finishing high school at Chewoyet he joined Kenyatta University in Nairobi, Kenya, for his undergraduate degree in education and graduated with honors in 1995. He briefly taught at Nyatieko high school in Kisii after his undergraduate program before returning to Kenyatta University on a scholarship for a Master of Arts degree in geography. After his second graduation from Kenyatta University he was employed by the Department of Geography, Maseno University, Kenya, as Tutorial Fellow. In the spring of 2002 he was admitted for a doctoral program at the Department of Geography and Anthropology, Louisiana State University. He worked as a Research/Teaching Assistant at the Earth Scan Laboratory to support his doctoral program, and was able to fund his dissertation research through a series of internal and external grants. He will receive the degree of Doctor of Philosophy in geography in May, 2007.