Late Holocene Environmental Reconstructions from Lake Solai, Kenya

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Late Holocene Environmental Reconstructions from Lake Solai, Kenya

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Playa lake systems tend to be overlooked archives of paleoenvironmental change due to the likelihood of a short and intermittent record of deposition. Groundwater-fed wetlands associated with these climate-sensitive playas, however, preserve changes in hydrologic budget and are thus valuable archives for semiarid regions. This study examines the paleoecological record of a groundwater-fed wetland from Lake Solai, Kenya. Biological proxies are used to reconstruct paleoenvironmental change and climate impacts over the past millennium. Dry conditions persisted between CE 1115 and 1490, followed by wetter conditions during the Little Ice Age. Near surface sediments indicate increasing anthropogenic impact through pastoralism. Key Words: biological proxies, East African Rift, groundwater-fed wetlands, land degradation, paleoenvironmental change.

Terminal, or endorheic, basins contain valuable sedimentary archives of climate and environmental change through time (e.g., Stine 1990; Ricketts and Johnson 1996; Magee et al. 2004). In the East African Rift System (EARS), lakes such as Baringo, Naviasha, and Turkana have been the focus of efforts to reconstruct Holocene climate changes (e.g., Verschuren, Laird, and Cumming 2000; Kiage and Liu 2009b; De Cort et al. 2013). These systems, however, have large catchments and surface areas (e.g., Lake Baringo’s catchment is ~8,655 km²), so biologic and sedimentary records deposited within these basins integrate broad-scale climate and environmental change, effectively “drowning out” local changes.
(Jacobson and Bradshaw 1981; Mohammed, Bonnefille, and Johnson 1995). In contrast, smaller playa systems and their associated groundwater-fed wetlands have great potential in the development of local and regional records of climatic variability and human impacts, although the paleoenvironmental archival potential of these small sites is rarely explored.

This article seeks to address this gap in reconstructing local and extralocal environmental change in the EARS. Here we present a late Holocene environmental reconstruction from Lake Solai, Kenya, with a catchment size of only 175 km², making it an ideal site to examine local environmental change (Figure 1). The goals of the study are threefold: (1) Use a suite of complementary biological proxies archived in wetland sediments to reconstruct changes in the paleoenvironment, (2) assess climatic and anthropogenic impacts on the system, and (3) compare the record to regional environmental and climatic records in the EARS.

**Environmental Setting**

Lake Solai is a small playa lake (9 km²) located in the EARS approximately 20 km southeast of Lake Bogoria (Figures 1 and 2). The playa lies at an elevation of 1,500 m and is located in a small graben bounded to the east by an imposing ~500-m-high fault block escarpment and to the west by a low relief fault block that rises ~100 m above the valley floor (Figure 2B). The geology is complex, with Cenozoic volcanics overlying basement rocks. Four major rivers feed the lake (Figure 2A). The Kasururei and Maji Tamu are perennial and arise in the eastern elevated fault block. The Maji Tamu is the principal water source for the...
Figure 2  (A) Satellite image of the Lake Solai region, showing major springs and core locations. White dashed line indicates approximate location of a normal fault, down thrown side is on the east. Image date January 2004. Source: Google, NASA, Digital Globe. (B) Photo of the Maji Tamu deltaic wetland and approximate location of LSA-1. Eastern border fault block, the major source of surface runoff and groundwater to the Solai basin, is in the distance. Photo credit: Gail M. Ashley. (Color figure available online.)
lake and flows into the southern end where it forms a small delta (Figure 2A). The Kamolo and Chemuka rivers emanate from the flanks of the escarpment and are spring sourced, but much of their flow infiltrates and is stored as groundwater (Nyambok, Bhogal, and Bwire-Odhiambo 1993; De Bock 2004). There is no surface outflow from Lake Solai.

The regional climate is controlled by the seasonal migration of the Intertropical Convergence Zone (ITCZ), which results in two rainy seasons (Nicholson 1996). The watershed receives about 700 to 900 mm of precipitation annually (De Bock et al. 2009). Potential evapotranspiration is high (~1,800 mm year\(^{-1}\)), however, resulting in a negative hydrologic budget (De Bock et al. 2009). The playa lake therefore experiences distinct changes in response to the seasonal and interannual effects of evaporation and precipitation, as well as groundwater seepage (De Bock et al. 2009).

Interannually the playa responds to regional climate drivers such as El Niño that cause lake expansion (e.g., CE 1997–1998), whereas during exceptionally dry La Niña years the lake might completely dry out (e.g., CE 2011). Most recently, the waters flowing into the lake have been the subject of regional disputes with the Kalenjin community on the escarpment diverting water flow for sisal (Agave sisalana) irrigation. The local Tugen community are agropastoralists who recently settled permanently in the area. They primarily use the land around the lake for cattle grazing, particularly during the dry season, and maize is grown using irrigation along several of the stream systems (De Bock 2004; Kervyn de Meerendr 2004).

The vegetation immediately surrounding the lake was mapped in detail (De Bock 2004; De Bock et al. 2009). Acacia woodland is present to the west and south, although the trees are under stress from charcoal production. Cyperus laevigatus dominates the herbaceous zone beneath the woodlands and also borders the lake from north to south on the eastern shore. Saline swamps are present to the east and south, which are dominated by Sporobolus robustus closest to the lake edge and by Cyperus lacovigatus in regions that experience prolonged drying phases. An open water wetland occurs on the delta of the Maji Tamu. Here vegetation is characterized by rooted plants (Typha domingensis, Cyperus dives, and Leersia hexandra) and a floating aquatic flora (Pistia stratiotes and Ludwigia stolonifera; De Bock 2004; De Bock et al. 2009).

Method

Two cores were collected from Lake Solai along with two additional cores recovered from its wetland complex using a Livingstone corer in 2006 (Figure 2). The cores were stored in butyrate tubing and shipped to the United States for analysis. The lithostratigraphy was described and the cores subsampled at approximately 2-cm increments. In this study, the biostratigraphy of the longest and most organic rich core, wetland core LSA-1, is discussed (Figure 2). Total organic carbon (TOC) following standard protocols was also undertaken (Maharjan 2010).

Thirty-nine samples were prepared for pollen analysis. Pollen analysis followed standard protocols (Faegri and Iversen 1989) but deviated slightly in the processes used to remove siliceous material, which was separated by a combination of sieving (200- and 10-μm sieves) and through treatment with sodium pyrophosphate to eliminate fine clays (Munsterman and Kerstholt 1996). Samples were treated with sodium polytungstate as a gravity separator to remove the pollen from the remaining clays (Munsterman and Kerstholt 1996). Samples were stained with safranin and stored in silicon oil. An exotic tablet of Lycopodium was added at the beginning of the preparation as a control (Stockmarr 1971). Pollen and fern spores were identified with counts of 300 or more (not including the control). In sections of the core below 0.40 m, pollen was extremely rare, making counting unfeasible. Non-pollen palynomorphs (NPPs; Van Geel 2001) were identified and counted but are not included in the pollen and fern spore counts. Fungal spores identified in the NPP counts are presented as concentrations of total fungal spores. Macrofossil remains were picked and identified from the material trapped on the 200-μm sieve.

Fossil diatoms were extracted and processed using standard procedures to remove carbonate and organic materials (Batterbee et al. 2001). Diatoms were mounted in naphrax and wherever possible 400 frustules were counted and identified using a light microscope with phase contrast (Ashley et al. 2004). If diatoms were rare, the whole slide was counted. Preservation indicates the percentage of diatom frustules that are intact as a percentage of all diatom material (broken and whole). Corrosion was minimal. Standard identification references were used (Hustedt 1927; Patrick and Reimer 1975; Schoeman and Archibald 1976; Gasse 1986; Krammer and Lange-Bertalot 1986). Transfer functions are based on the East African data set in the European Diatom Database (see http://cricutica.ncl.ac.uk/Eddi/jsp/datasets.jsp).

Four samples were submitted to Beta Analytical for accelerator mass spectrometry (AMS) radiocarbon dating. Dates were calibrated using Intcal13 (Reimer et al. 2013) and depositional models developed using Oxcal 4.2 (Bronk Ramsey 2009).

Results

Lithostratigraphy

Core LSA-1 (1.87 m long) was collected from a wetland on the delta of the Maji Tamu (Figures 2 and 3). The basal material is made up of clayey silts up to 1.42 m. Several lenses of carbonate concretions occur within this zone. Above this, the sediments are clays to 0.87 m depth, where grain size shifts to silty clay. Numerous small (2–4 mm) carbonate concretions are also present. Concretion size increases between 0.88 and 0.67 m depth. An
Figure 3  (A) Calibrated age distributions for radiocarbon dates obtained from Solai wetland core (Table 1). Dates are calibrated with IntCal13 (Reimer et al. 2013) and the age model generated with Oxcal (Bronk Ramsey 2009). Also shown are the modeled age for wetland inception at 0.67 m and the modeled age for the significant increase in organics at 0.32 m. (B) Lithostratigraphy, Munsell coloration, percentage total organic carbon (Maharjan 2010), and location of radiocarbon dates of LSA-1. (Note change in scale of y-axis.) TOC = total organic carbon.
abrupt transition occurs at 0.67 m to organic rich clays. Small rootlets are present within this zone, which, in turn, abruptly changes to a peaty root mat near the surface (0–0.04 m). TOC (Figure 3B) averages 0.4 ± 0.1 percent from the base of the core to 0.7 m, and it rises to 1.4 ± 0.3 percent between 0.7 and 0.4 m. TOC levels increase markedly between 0.4 and 0.2 m (7.4 ± 0.5 percent) but then decline to about half these levels in the top 0.2 m (4.5 ± 0.3 percent; Maharjan 2010).

Chronostratigraphy

Suitable material for dating was sparse due to insufficient sample weight or the unsuitability of the material related to plant physiology (Marty and Myrbo 2014). No obvious terrestrial plant material was found. Nevertheless, three samples of bulk organic sediment and one sample of seeds of *Pistia* were selected for AMS radiocarbon dating. *Pistia* is a floating, nonrooted aquatic plant that photosynthesizes aerially (Spencer and Bowes 1990). The AMS samples were all collected from organic-rich clay (Table 1 and Figure 3). All ages are in stratigraphic order with the oldest returning a median age of CE 1155 at a depth of 0.63 to 0.62 m (Table 1). The *Pistia* seeds at 0.20 to 0.21 m had a postbomb age of CE 1980. Bayesian analysis run in Oxcal 4.2 identified this age as an outlier and it is not used in the age model (Figure 3A; Bronk Ramsey 2009). The lowermost section of the core remains undated due to the absence of organic material suitable for dating. We modeled the age of wetland inception at 0.67 m using a deposition model in Oxcal 4.2 (Bronk Ramsey 2008, 2009).

Table 1  Radiocarbon data for LSA-1 core

<table>
<thead>
<tr>
<th>Lab No.</th>
<th>Material dated</th>
<th>Depth (cm)</th>
<th>C(^{14})</th>
<th>Error ±</th>
<th>Calibrated age range (cal. CE)</th>
<th>Median cal. CE</th>
<th>Rates of sedimentation (cm/year)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Beta 279974</td>
<td><em>Pistia</em> seeds<em>sup a</em></td>
<td>20–21</td>
<td>1.273</td>
<td>0.005</td>
<td>1965–1985</td>
<td>1980</td>
<td>—</td>
</tr>
<tr>
<td>Beta 370470</td>
<td>Organic sediment</td>
<td>38–40</td>
<td>390</td>
<td>30</td>
<td>1440–1630</td>
<td>1490</td>
<td>0.075581</td>
</tr>
<tr>
<td>Beta 370471</td>
<td>Organic sediment</td>
<td>48–50</td>
<td>540</td>
<td>30</td>
<td>1315–1435</td>
<td>1400</td>
<td>0.111111</td>
</tr>
<tr>
<td>Beta 248635</td>
<td>Organic sediment</td>
<td>62–63</td>
<td>880</td>
<td>40</td>
<td>1035–1245</td>
<td>1155</td>
<td>0.055102</td>
</tr>
</tbody>
</table>

Note: Beta = Beta Analytical. Ages calibrated using IntCal13 (Reimer et al. 2013).

*sup a*Identified as outlier by Oxcal 4.2 (Bronk Ramsey 2009).

*sup b*Beta 279974 is not used in this calculation because it was identified as an outlier. Instead, year of collection (2006) is used.

Biological Components

Diatom abundance and preservation were excellent below 0.67 m but with relatively sparse and fragmented diatoms occurring in the more organic-rich sediments. More than forty diatom types were identified that reflect variations in water depth, macrophytes, and water chemistry. Interestingly, analysis of the diatoms from the lake cores (not shown here) contained a sparse diatom flora dominated by *Aulacoseira agassizi* and *A. granulata-agassizi*. These diatoms are often fragmented and scarce in the sediments, which together with common phytoliths suggests both flooding and subaerial exposure of the lake floor. The difference in records between the lake and wetland cores suggests that the wetland record has had continuous deposition at the site at least for the upper 0.67 m.

More than fifty pollen and spore types were identified, many of which were rare. The most common pollen types were from wetland taxa such as *Typha* and Cyperaceae, as well as herbaceous taxa such as Amaranthaceae, Asteraceae, and Poaceae. These pollen represent the vegetation within the local environment. Overall, terrestrial tree pollen counts are low, which is not uncommon in semi-arid regions where trees are often insect pollinated. Afromontane taxa, such as *Podocarpus* and *Juniperus*, represent the extraregional pollen rain from highland areas (Kiage and Liu 2009b). Pollen preservation and abundance was poor in the lower sections of the core but significantly improved in the upper, organic-rich sections. Pollen concentration is low below 0.40 m, suggesting that pollen preservation was severely affected by environmental conditions at the time of deposition.

Thirty-eight NPP types were identified but their presence was negligible below 0.40 m. Here we present the taxa relevant to determining environmental change, which includes Nymphaeaceae basal cells, rotifer resting eggs (*Brachionus* and *Anuraeoopsis*; Van Geel 2001) and a variety of fungal spore taxa (Van Geel 2001; Kiage and Liu 2009b; Gelorini et al. 2011). The most common types include *Glomus*, UG Type 1109 (Gelorini et al. 2011), and a variety of dung fungal spores.

Seeds of wetland plants were found in the top 0.32 m of the core. *Typha* and Cyperaceae were the most common, but seeds of *Najas* and *Pistia* were also present. *Daphnia ephippia*, primarily from the *Ctenodaphnia* subgenus, and statoblasts of *Plumatella* spp. were found (Birks et al. 2001).

Biostratigraphy

The biostratigraphic record is zoned into three principal sections, based on lithostratigraphy and major changes in various proxy abundances.

Zone 3 (1.87–0.67 m)

This zone is subdivided into seven subzones based on the diatom flora (Figure 4). Pollen and fungal
Figure 4  Percentage diatom stratigraphy for LSA-1. Note that the near-surface samples had only rare diatoms present, which included planktonic and benthonic taxa. For lithostratigraphic key see Figure 3B.
Figure 5 Percentage pollen record for LSA-1. Afrotropical, bush, and herbaceous percentages shown as a function of terrestrial pollen and fern spores. Wetland and aquatic pollen percentages calculated using the sum of total pollen and fern spores. Pollen in levels below 0.40 m were rare; presence is shown by the + sign. For lithostratigraphic key, see Figure 3B.
Fungal spore data are shown as concentration (cm$^3$). For lithostratigraphic key, see Figure 3B.
Figure 7  Distribution of seeds, Nymphaeaceae basal cells, and faunal remains. Data shown as raw counts. For lithostratigraphic key, see Figure 3B.
Table 2  Comparison of biostratigraphic record from Lake Solai with records from select lakes in the East African Rift System

<table>
<thead>
<tr>
<th>Year (CE)</th>
<th>Solai</th>
<th>Challa&lt;sup&gt;a,b&lt;/sup&gt;</th>
<th>Naivasha&lt;sup&gt;c&lt;/sup&gt;</th>
<th>Bogoria&lt;sup&gt;d&lt;/sup&gt;</th>
<th>Lobol&lt;sup&gt;e&lt;/sup&gt;</th>
<th>Baringo&lt;sup&gt;f&lt;/sup&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td>Present</td>
<td>Dry(CE 1750–present)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1800</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Dry conditions (CE 1750–1850)</td>
</tr>
<tr>
<td>1700</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Inception (CE 1650)</td>
</tr>
<tr>
<td>1600</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>No sediment recovery</td>
</tr>
<tr>
<td>1500</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1400</td>
<td>Medieval Climate Anomaly (CE 1100–1400)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1300</td>
<td></td>
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<td></td>
<td></td>
<td></td>
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<tr>
<td>1200</td>
<td></td>
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<td>1100</td>
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<td>1000</td>
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</tbody>
</table>

<sup>a</sup>Tierney et al. (2011).
<sup>b</sup>Wolff et al. (2011).
<sup>d</sup>De Cort et al. (2013).
<sup>e</sup>Ashley et al. (2004) and Ashley et al. (2013).
<sup>f</sup>Kiage and Liu (2009a, 2009b).
spores are sparse throughout this zone, although Ceratophora and Glomus are present (Figure 5 and 6). Basal Subzone 3g is dominated by diatom taxa Anomoeoneis sphaerophora with less common Thalassiosira faurii and T. rudolfii. Encyonema Muelleri, Epithemia adnata, and Cocconeis placentula vary in abundance. Subzone 3f is represented by increases in Aulacoseira granulata var. angustissima and Synedra ulna. In Subzone 3e there is a major increase in Rhopalodia gibberula and Staurastrum leptostraturn with A. granulata var. angustissima remaining common. A. granulata also appear during this period. A. granulata var. angustissima again remain common in Subzone 3d but with A. granulata increasing and A. agassizi and A. granulat-aga-giszi common toward the top of the zone. Ceratophora and Glomus fungal spores are present in this subzone. Subzone 3c is characterized by the disappearance of A. granulata var. angustissima and a major increase in A. agassizi, A. granulata-agassizi and A. agassizi var. malayensis. Subzone 3b continues with abundant planktonic species, but with a change to dominance by A. agassizi and its variety malayensis and a decline in A. granulata, and A. granulata-agassizi. The diatoms in Subzone 3a resemble those of Subzone 3c but with a small increase in epiphytic species (Cocconeis placentula, Gomphonema spp., and Synedra ulna). Glomus fungal spores are present in notable concentrations.

Zone 2 (0.67–0.40 m)

Zone 2, like Zone 3, is marked by the near absence of pollen and macrofossils; one seed from the Amaranthaceae family was identified (Figure 7) and pollen of herbaceous and wetland taxa are present (Figure 5). The NPP record is dominated by the Glomus type (Figure 6). TOC shows a doubling from Zone 3 (Figure 3). This zone is characterized by a sparse diatom flora and common fragmentation (Figure 4). The flora is dominated by rare planktonic species, including A. agassizi var. malayensis, A. granulata, A. agassizi, and A. granulata-agassizi. Nymphaeaceae (water lily family) basal cells are present throughout Zone 2 (Figure 7). Rates of sedimentation are initially low, but a near doubling occurs at the transition with Zone 1 (Table 1).

Zone 1 (0.40–0 m)

Zone 1 is subdivided into two subzones: Subzone 1b (0.40–0.19 m) and Subzone 1a (0.19–0 m). Rates of sedimentation return to levels similar to the early part of Zone 2 (Table 1). Subzone 1b is characterized by herbaceous pollen taxa (e.g., Poaceae, Asteraeaceae, and Amaranthaceae) and tree and shrub taxa such as Podocarpus and to a lesser extent Juniperus, Justicia, and Phyllanthus. Typha dominates the wetland taxa (both pollen and seeds; Figure 5 and 7). Achenes of Typha and undifferentiated Cyperaceae seeds are found, as were seeds of Najas, a rooted submersed aquatic plant, and Pistia (Figure 7). Basal cells from Nymphaeaceae plants are also present (Figure 7). Rotifer resting eggs are present, as are statoblasts of Plumatella (Figure 7). The diatom flora in this subzone is comparable with those found in Zone 2 (Figure 4). Fungal spore types Chaetomium, Ceratophora, and the UG Type 1109 are present in high concentrations. Glomus concentrations exhibit a dramatic decline at the beginning of the subzone but concentration rises toward the boundary with Subzone 1a.

Subzone 1a depicts an interesting, albeit subtle shift from Subzone 1b, with Asteraeaceae pollen slightly more abundant, as Amaranthaceae declines (Figure 5). Podocarpus pollen becomes more abundant, as do other tree and bush pollen such as Juniperus and Justicia. Poaceae pollen shows a slight decline. Pollen of Zea mays is found at 0.12 m. Seeds of Typha and Cyperaceae are less common (Figure 7), although their pollen remains a significant component of the wetland taxa. Pollen, seeds, and microfossils of obligate aquatics, Pistia and Nymphaeaceae, are also present throughout Subzone 1a. Glomus and fungal spore UG Type 1109 concentrations are high throughout this subzone although both show a decline from 7 cm to the surface (Figure 6). Dung fungal spores Chaetomium, Sporomellia, and Ceratophora are common throughout Subzone 1a. Podspora is only found in the near-surface sediments (Figure 6). Rotifer resting eggs drop out of the record and Daphnia ephippia, D. barbata, are present (Figure 7; Mergesey, Verschuren, and De Meester 2005). The numbers of Plumatella decline. The diatom record is characterized by an abundance of epiphytic species, particularly Gomphonema parvulum and G. lanceolatum. The near-surface sample is characterized by rare planktonic (Aulacoseira spp.) and benthonic taxa (Figure 4).

Discussion

Environmental Reconstruction

In this section we interpret the results from Lake Solai and place the reconstruction in the context of climate and environmental change in the broader East African Rift for the past 1,000 years. Table 2 provides a comparative summary with other EARS studies.

Pre-wetland environment (Zone 3): Older than CE 1115

The environmental and climatic reconstruction of Zone 3 sediments is challenging. The diatom flora show an apparent transition from shallow alkaline lake conditions (Subzones 3g–3d) to “deeper” and fresher water conditions in Subzones 3c through 3a. The diatom flora in Subzone 3g indicate the highest reconstructed salinity conditions for the core with estimated conductivities of 1,000 to 10,000 µSm cm⁻¹ and a pH of about 8.7 to 9.8. Further, epiphytic and benthic diatoms (e.g., Encyonema Muelleri) vary in abundance and
indicate a period of shallow water. In Subzone 3f, taxa indicate shallow mildly saline to fresh waters with transfer function data for the overall flora suggesting gradually freshening conditions (conductivities ranging from about 500 to 3,000 £Sm cm\(^{-1}\)). Epiphytic and benthic diatoms reach maxima at the base and top of this zone, indicating more expansive macrophytes, although no organic material was recovered. In Subzone 3e, planktonic taxa suggest a deepening and freshening of lake water with pH falling to 7 to 8. The rarity of epiphytes in Subzone 3d further indicates open shallow water and estimated conductivities of 500 to 2,000 £Sm cm\(^{-1}\) and a pH of 7.5 to 8. The planktonic flora of Subzone 3c suggest fresh (conductivities = 100–200 £Sm cm\(^{-1}\); pH ~ 8), moderately deep waters with macrophytes absent from the core location. Conditions in Subzone 3b were probably similar to those of Subzone 3c. The diatom assemblage of Subzone 3a is indicative of Subzone 3c, although the presence of epiphytic taxa toward the top indicate the growth of larger plants in the area. This is somewhat supported by the finding of the only plant macrofossils within Zone 3, as one Typha and one Cyperaceae seed were found within Subzone 3a. Overall diatom preservation and abundance declined upward through this zone, perhaps reflecting reworking of diatoms within a shallowing lake or as a result of periodic desiccation.

The scarcity of plant macro- and micromaterial combined with the preservation of siliceous diatoms with distinct stratigraphic distributions suggests that conditions might have been sufficiently oxic, alkaline, or both to preclude plant material preservation. This is supported by the low TOC values (Figure 3) and the presence of carbonate nodules (Figure 3), which likely indicate pedogenic formation. This further suggests periods when the lake was shrinking or completely dry that permitted soil formation.

Elsewhere in the EARS, at Crescent Island Crater, Lake Naivasha, diatom and lake sediment data indicate that between CE 200 and 1000 this crater lake experienced low to intermediate lake depths (Figure 1 and Table 2; Verschuren, Laird, and Cumming 2000; Verschuren 2001). At Lake Challa, a deep freshwater crater lake located on the eastern flank of Mount Kilimanjaro (Figure 1), sediments are varved (Wolff et al. 2011). Varve thickness is controlled primarily by the dry season diatom layer and thickness is correlated with wind conditions. Thick varves correlate with La Niña and thin ones with El Niño years. The 1,000-year time period prior to CE 900 exhibits generally thin varve layers indicative of overall wetter conditions. The prominent variability in varve thickness, however, particularly between CE 400 and 750, suggests that conditions were also characterized by extreme drought for extended time periods. Similarly, to the north at Lake Bogoria (Figure 1), a thick deposit of lath-shaped trona crystals was deposited between CE 450 and 700 and indicates a shallow hypersaline lake phase (De Cort et al. 2013). At Hausberg Tarn, Mount Kenya, the organic carbon content from sediments in a proglacial lake indicate glacial advances between CE 650 and 850. Advances appear to be correlated with low temperatures (Karlén et al. 1999; see Figure 1).

Unfortunately, the absence of a chronology for the lower section of the Lake Solai record makes direct comparisons to other watersheds difficult. Deposition models place the end of Zone 3 near CE 1115. The time period represented by Zone 3 at Lake Solai might therefore correspond with the generally wetter conditions to the south at Lake Challa, as reflected by the presence of diatom assemblages indicating fluctuating salinity and pH. The record is complicated, though, by the absence of preserved organic material. These findings suggest that although water was present in the lake, conditions were oxic, possibly reflecting a well-mixed shallow Lake Solai and overall windier conditions. The presence of discrete carbonate concretions is also suggestive of periodic lake desiccation and pedogenesis.

Medieval Climatic Anomaly (Zone 2): CE 1115–1490

The Zone 3–2 contact is abrupt, indicating a significant change in the factors affecting deposition at the core site. Starting around CE 1100 there was a shift from the prevailing wetter conditions to an overall drier or ephemeral water situation. The presence of Nymphaeaceae basal cells and the diatom record indicates standing water but with poor preservation, perhaps indicative of water only persisting for part of the time (Figures 4 and 7). The increasing concentration of Glomus, a mycorrhizal fungus found on a variety of plants, is associated with erosional conditions (Figure 6; Van Geel 2001). Gelorini et al. (2012) associated a rise in Glomus with lake bed desiccation at Lake Chibwera, Uganda. The absence of dung fungus is further suggestive that the environment was not hospitable to wild herbivores (Figure 6; Kiage and Liu 2009b).

The results from Solai support the evidence for overall dry conditions during the Medieval Climatic Anomaly (MCA) within the EARS. Although lake records show a shift to generally drier climatic conditions, the transition and duration appears to vary throughout the EARS. At Lake Challa, varves and an index of soil bacterial versus aquatic archaeal membrane lipids (BIT) indicate severe drought conditions between CE ~1100 and 1300 (Figure 1, Table 2; Verschuren et al. 2009; Tierney et al. 2011; Wolff et al. 2011). At Lake Naivasha, diatom, sediment composition, and chironomid data indicate a pronounced low stand between CE 1000 and 1270 (Verschuren, Laird, and Cumming 2000; Verschuren 2001). At Lake Bogoria, water chemistry and lake level reconstructions indicate that relatively dry conditions lasted until about CE 1100, followed by increasing lake levels until CE 1550 (De Cort et al. 2013). Just to the north of Lake Bogoria, at the wetland site of Loboi, paleoenvironmental data from several sediment cores indicate that prior to ~CE 1250 this region was characterized by fine-grained, highly weathered sediments deposited by
the Loboi River, as well as wind-blown sediment (Ashley et al. 2004; Ashley et al. 2013).

Little Ice Age to Present (Zone 1): CE 1490–present

The Solai wetland record indicates that a shallow lake persisted throughout Subzone 1b as rooted and free-floating aquatics are present (e.g., *Najas, Pistia*, and *Nymphaeaceae*) as well as planktonic diatoms (e.g., *Aulacoseira* spp.). Interestingly, no taxa from the *Nymphaeaceae* family are present at the lake today, although they are found in other EARS lakes (Lamb, Darbyshire, and Verschuren 2003; De Bock 2004). *Plumatella* statoblasts are also common during this phase, suggesting clear still or gentle flowing waters (Birks et al. 2001; Francis 2001). These indicators all suggest higher, persistent, and also flowing water at the core site. The local vegetation assemblage is dominated by grasses and other herbaceous taxa and the bush taxa *Pbyallanthus*. Afromontane taxa (e.g., *Podocarpus, Juniperus*, and *Olea*) are present, indicating long-distance transport into the system. The marked decline in *Glogen* in Subzone 1b and presence of *Coniochaeta* is further evidence for overall wetter conditions as soils were anchored by the vegetation (Gelorini et al. 2012). The fungal spore UG Type 1109 is abundant in Subzone 1b and is associated with drier conditions (Gelorini et al. 2011). Its presence is possibly reflective of seasonal expansion and contraction of the lake at this time. Fungal spores associated with dung become increasingly important, reflecting the movement and possible expansion of animal herds about the lake as wetter conditions persisted.

The findings at Lake Solai for the Little Ice Age (LIA) are comparable to evidence for generally wetter conditions throughout the EARS at this time, although there is some variation within the region, with wetter conditions generally prevailing in the south, but with periods of marked aridity also occurring during this time frame. At Lake Naivasha, fairly wet conditions existed, although this was interrupted by two pronounced severe phases of aridity (CE 1560–1620 and CE 1760–1840; Table 2; Verschuren et al. 2000). The Lake Challa record indicates wetter conditions, with below-average varve thickness occurring (Wolff et al. 2011). At Lake Bogoria, the presence of magadiite in the core sediments between CE 1400 and 1550 indicates relatively dilute conditions and the highest water levels (De Cort et al. 2013). At Loboi Swamp, wetland sedimentation began about CE 1630 (Ashley et al. 2004; Driese et al. 2004), which is consistent with the initiation of sedimentation reflecting wetter conditions (CE 1650) at Lake Baringo (Kiage and Liu 2009a, 2009b). The wetter conditions within the EARS were likely forced by minimum values of the Indo-Pacific warm pool sea surface temperatures (Tierney et al. 2013). At Lake Solai a subtle shift in the record occurs at about CE 1750. TOC drops to about half its Subzone 1b levels (Figure 3) and there is a general decline in the number of preserved seeds (Figure 7). There is a paucity of diatoms in Subzone 1a (Figure 4), which is surprising but has also been observed in parts of modern *Typha* and *Papyrus* swamps at Loboi near Lake Bogoria (Owen et al. 2004) and might relate to the dense networks of reeds that inhibit light penetration. These findings coupled with dramatic increases in fungal UG Type 1109 suggest a lowering and shrinking in lake size. Further, the overall increase in the transport of Afromontane taxa to the site, particularly *Podocarpus*, implies an onset of drier conditions in the region that persist today (Ashley et al. 2004; Kiage and Liu 2009b).

The findings at Solai Subzone 1a are consistent with reconstructions throughout the EARS. Palynological work at Lake Baringo found an abrupt change to drier conditions starting around CE 1750 (Kiage and Liu 2009b), and at Lake Bogoria mineralogy indicates that lake level dropped around CE 1800 (De Cort et al. 2013). To the south, at Lake Naivasha, aridity returned about CE 1760 (Verschuren, Laird, and Cumming 2000). At Bogoria, Baringo, and Naivasha, aridity ended sometime after CE 1850. At Solai, however, the multiproxy data do not indicate a shift back toward the peak wet conditions experienced during Subzone 1b.

Human Impacts

The first indication of potential human presence in the region occurs in Subzone 1b when dung fungal spores become significant. It is probable that the region was populated by wild herbivores; however, pastoralists with domesticated herds also frequented the site. This is suggested by the importance of *Chaetomium* and *Cercophora* spores but with low levels of *Sporormiella*, which is often associated with domestic herds (Kiage and Liu 2009b). This indicates that pastoralism was likely occurring at Lake Solai prior to European colonization (late CE 1800s).

The impacts of cattle ranching rather than agricultural cultivation are significant in Subzone 1a. Although introduced crops are important on the escarpment today, they are not significant in the immediate lakes' history, as only one grain of maize was found in the near-surface sediments. The overall increase in grass and Asteraceae pollen, however, indicate that there was an increase in herbaceous vegetation that is likely reflective of not only drier conditions but cattle grazing as these taxa are encouraged. Indeed, this is supported by the significant increase in dung fungal spores in Subzone 1a that reflects an increase in the cattle population. *Sporormiella* and *Cercophora* are particularly important components
of the dung fungal record at this time and *Podospora* appears in the near surface sediments. The shift in dung fungal types might reflect a decline in wild herbivores and concomitant increase in domesticates.

It is currently estimated that there are 40,000 head of cattle and more than 60,000 sheep and goats within the region (Kervyn de Meerendré 2004). With the increase in livestock, erosion of soil, as reflected in the high *Glomus* concentrations, has resulted within the catchment, a phenomenon also seen to the north at Baringo (Kiage and Liu 2009b). This has likely affected the quality and clarity of Lake Solai waters. The decline and recent absence of *Plumatella* but increase in *Ctenodaphnia* species indicates an increase in turbidity within the lake waters, which might be caused by the increase in cattle numbers and associated increase in erosion. The presence of *Nitzschia perminuta*, which is a eutrophic diatom tolerant of moderate pollution, as well as the overall absence of benthic diatoms, is a further indication that cattle grazing is having an impact on the lake system, which is highly turbid today (Figure 2). At Lake Baringo, a Landsat analysis demonstrated significant deforestation and associated increases in lake turbidity between 1986 and 2000 (Kiage et al. 2007). The results from Lake Solai suggest that erosion and land degradation have been affecting turbidity in the lake for at least fifty years.

Conclusions

The multiproxy data at Solai indicate that wetland preservation, growth, and persistence are reflective of broader regional climate controls that are likely forced by changes in sea surface temperature in the Indian Ocean (Tierney et al. 2013). Unlike larger lake systems, however, local anthropogenic impacts can be detected in the record, specifically the rise in the importance of cattle to the local economy. The macrofaunal remains indicate that cattle numbers are having a negative effect on water quality in the system.

This analysis indicates that groundwater-fed wetlands associated with small terminal lake basins have the potential to archive paleoclimate and local land use signals, despite the propensity for the lake itself to desiccate on annual to interannual timescales. These wetland deposits thus represent a valuable, but overlooked, environmental repository and can provide critical data on understanding environmental responses to anthropogenic impacts and resource pressures.

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