Paleolimnological evidence for the onset and termination of glacial aridity from Lake Tanganyika, Tropical East Africa

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Abstract

Geochemical (elemental concentrations and Sr isotopes) and sedimentological data (grain size, TOC, MS and BSi%) in a continuous 60,000-year sediment core record from the Kalya horst region of central Lake Tanganyika provide a detailed history of paleoclimate-mediated weathering and overflow events from upstream Lake Kivu. Univariate (elemental profiles), bivariate (elemental ratios) and multivariate analyses of chemical trends show variations between the dry Late Pleistocene (32-18 ka cal yr BP) and the wetter conditions that both preceded and post-date that interval. This record places important new constraints on the timing of LGM aridity in East Africa, based on significant decreases in magnetic susceptibility and soluble cation concentrations, coinciding with increased grain size and biogenic silica. The elemental indicators in the early portion of the sedimentary record (60-50 ka cal yr BP) characterize this interval as a comparatively wet period, similar to modern conditions. Our record demonstrates that the transition toward arid conditions in tropical Africa during high latitude glaciation was a two staged event with conditions similar to modern levels of P/E prior to ~50 ka cal yr BP, intermediate levels of aridity occurring from 50-32 ka cal yr BP, and intense aridity from 32-18 ka cal yr BP.

The initiation of inflow from upstream Lake Kivu into Lake Tanganyika is evidenced at 10.6 ka cal yr BP through its influence on both elemental profiles (Mg, Ca) and through its directional effect on ⁸⁷Sr/⁸⁶Sr ratios. Increases in elemental (Mg, Ca, Sr) concentrations coincide with the timing of the Lake Kivu overflow. Metal geochemistry suggests the overflow from Lake Kivu into Lake Tanganyika may have ceased between 8-6 ka cal yr BP, coinciding with a period of Middle Holocene aridity in East Africa.

1. Introduction

Elemental geochemistry from sediment cores provides a powerful tool for reconstructing the paleoenvironmental and paleoclimate history of lakes and their watersheds (Mackereth, 1966, Ng and King, 2004, Roy et al., in press). For example, studies have found that during periods of increased precipitation elemental concentrations of soluble elements such as K, Be, Mg, Ca, Ba and Sr, often increase in lake sediments, and conversely during dry periods these same elements decline (Mackereth, 1966; Engstrom and Wright, 1984). Similarly, concentrations of beryllium increase with increases in weathering in the surrounding catchment (McHargue et al., 2000; McHargue et al., 2005). These changes result from the fact that under conditions of rapid erosion, exposure and weathering of bedrock within a watershed of high concentrations of original silicate minerals enriched in soluble cations are eroded and transported to downstream lakes.

Elemental profiles in lakes cores can be used to investigate other processes as well. For example, the solubility and/or precipitation of transition metals such as Ti, Co, Cu, Cr, Zn, V, Mn, Fe, Ni are sensitive to redox conditions (Mackereth, 1966; Davison, 1992; Brown et al., 2000). Variations in lake level, windiness or lake temperature can all affect the relative position of the oxicleine with respect to the lake floor, resulting in significant changes in redox sensitive elements. Concentrations of Co, Cr, Fe, Mn, Ni, Pb and Zn increase over an oxic-suboxic boundary (Balistieri et al., 1994), whereas there are noticeable decreases in concentrations of Cu and Cd under oxic conditions (Brown et al., 2000). Fe and Mn in particular are more soluble in anoxic conditions, and low concentrations indicate that the sediments were not oxic at the time of deposition

(Haberyan and Hecky, 1987; Davison, 1992). However, redox-sensitive elements must be used with caution to reconstruct anoxia, as their abundances can be affected by changing redox conditions shortly after burial.

When lake water conductivity is high relatively insoluble metals are more likely to accumulate as precipitates in the sedimentary record, whereas during periods of dilution, often accompanying high lake levels, these same metals are more likely to remain in solution (Davison, 1992). Also, both elemental and isotopic profiles can be used to identify the timing of new hydrologic inflows (e.g. rivers, springs) into a lake when those inflows carry unique or unambiguous chemical signals relative to the receiving basin (Hecky and Degens, 1973; Haberyan and Hecky, 1987).

In contrast to temperate lakes and their watersheds, from which much of our knowledge of elemental geochemistry is derived, elemental geochemistry in tropical lacustrine sediment cores has been much less frequently studied to infer weathering and environmental change (see Brown et al., 2000, and Cardinal et al., 2001 for important exceptions). In this study we present a long record of sedimentary geochemical and sedimentological data from a well-dated core from Lake Tanganyika, which demonstrates the value of elemental geochemical records for inferring paleoenvironmental variability within a tropical lake and its watershed. Our elemental data, in combination with other isotopic, geochemical and stratigraphic information from the core, provides a history of climate and paleohydrologic changes for central Africa over the past 60 ka.

1.2 Lake Tanganyika: Geography, limnology and regional climate

Lake Tanganyika is the second largest freshwater lake in the world by volume, located between 9°S and 3°S. It occupies a series of half graben basins in the western branch of the East African Rift Valley (**Figure 1**) (Tiercelin and Mondeguer, 1991). The bedrock geology of the Lake Tanganyika basin is primarily Proterozoic metasedimentary rocks, with basaltic volcanic rocks only occurring in significant quantities in the upstream basin of Lake Kivu (Tiercelin and Mondeguer, 1991; Cohen et al., 2006). The lake is extremely deep (at >1400m the second deepest in the world) and permanently stratified (meromictic). The lake is currently hydrologically open but loses most of its moisture through evaporation rather than outflow, and consequently is moderately saline (conductivity= 670 $\mu\text{mho/cm}$) (Cohen et al., 1997). Because the lake hovers near the hydrologically open/closed threshold, its lake level is extremely sensitive to watershed precipitation/evaporation ratios.

Lake level in Tanganyika has also been responsive to overflow events in upstream Lake Kivu, which lies in the western rift valley north of Tanganyika. Lake Kivu has overflowed intermittently through the Ruzizi River into Tanganyika starting at 10.6 ka BP, if not earlier (Haberyan and Hecky, 1987; Stoffers and Hecky, 1978). The chemistry of Lake Kivu is dominated by hydrothermal inputs on the lake bottom and is more saline and enriched in numerous metals relative to Lake Tanganyika water (Degens and Stoffers, 1976; Haberyan and Hecky, 1987; Barrat et al., 2000). As a result, its composition (and that of its Ruzizi River outlet) is chemically quite distinct from, and more saline than Lake Tanganyika.

Lake Tanganyika's equatorial location lies within the migratory path of the Intertropical Convergence Zone (ITCZ) (Nicholson, 2000) (**Figure 2**). The basin

currently receives an average rainfall of 1200 mm/year with a rainy season occurring between Sept.-May (Coulter and Spigel, 1991). The Mahale Mountains of the central Lake Tanganyika basin, adjacent to the core site discussed in this paper, have a significantly higher average rainfall (1800 mm/yr) than the lake basin average (Cohen et al., 2006). Throughout the lake basin, a dry season between May and August coincides with strong winds from the south, whereas during the wet season winds are weaker (Plisnier et al., 1999).

Lake Tanganyika is permanently stratified, with anoxic bottom waters below ~100-130m in our study area (Coulter and Spigel, 1991). Vertical mixing and partial ventilation of deep waters varies seasonally, as the thermocline tilts downwards towards the northern end of the lake during the dry windy season between May and September. This results in upwelling in the southern basin and the subsequent propagation of internal waves as the winds subside (Coulter and Spigel 1991; Plisnier et al., 1999; Naithani et al., 2003).

Long sedimentary records from Lake Tanganyika indicate that, as in much of East Africa, aridity and lowered lake levels occurred during the Last Glacial Maximum (LGM) of higher latitudes (Gasse et al., 1989; Scholz et al., 2003; Talbot et al., 2006). Gasse et al. (1989) for example, suggested a drop in the level of Lake Tanganyika on the order of 400m between 26.0 to 15 ka. Scholz et al. (2003) examined various indicators of climatic changes from a sediment core from the Kavala Island Ridge (total organic carbon, carbon and nitrogen isotopes and diatoms). They argued, based on carbon isotopic data (and the presence of a zone of woody debris in an offshore core) for a shift towards more arid conditions around 55 ka ^{14}C (57 ka) and a dramatic decrease in lake

level. The diatom record from this core shows a similar timing of onset and termination of low lake conditions during the LGM as the Gasse et al., (1989) record. Sediment cores from the southern part of Lake Tanganyika indicate that arid conditions in that region ended around 15 ka, signaled by significant changes in organic matter deposited during a Late Pleistocene lake level transgression (Talbot et al., 2006).

Long paleoclimate records from East Africa are of importance for understanding climatic processes such as the role of solar variability in regulating tropical climates at Milankovitch time scales, and the relationship between abrupt climate changes, changes in ice extent, migration of Intertropical Convergence Zone, and regional climate variability (Nicholson, 2000). Records of pre-Late Pleistocene climate variability from tropical African lakes (>25ka) are still quite rare. Long records from Lake Tanganyika are of particular interest given the lake's antiquity and its demonstrated potential for producing high-resolution (frequently annually laminated) sedimentary records (Cohen et al., 1993). Here we present a new record from the central basin of Lake Tanganyika that provides insight into the timing of critical climate events in East Africa during the Late Quaternary.

2. Methods

2.1 Coring location

In 2004 the Nyanza Project (an NSF-Research Experience for Undergraduates [REU] research training program on tropical lakes) collected a suite of Kullenberg piston

cores from the Kalya horst block and platform, located at the north end of the southern basin of Lake Tanganyika (6°42.827'S, 29°49.957'E) (**Figure 1**). The specific coring site was chosen because ongoing seismic stratigraphic studies of the area of the horst block indicated that this core site was in an area of continuous but relatively slow sedimentation for at least the past 100 ka BP (**Figure 3**). Seismic data shows that our entire core record falls within a single stratigraphic sequence: a profound sequence boundary, indicative of considerably lower lake stands than any inferred from our record, underlies the base of our core by approximately 6.0 m (McGlue et al., 2006). Our most complete and temporally longest sediment core (NP04-KH3) was collected on the Kalya horst in 640 m water depth. This 7.75 m core was shipped to the Limnological Research Center (U. Minn.) core lab, where it was split, digitally photographed and logged for magnetic susceptibility and gamma ray attenuation porosity (GRAPE) density using a GEOTEK core scanner.

2.2 *Elemental analyses*

Geochemical element samples were taken from the split, working half of the core using trace clean spatulas and trace clean containers, for transport to the University of Arizona. The protocol for trace element digestion followed the procedure detailed by Hollocher et al. (1995). Samples for trace element analyses were collected every 8cm throughout the core, corresponding to a nominal sampling interval of 350 years between 0.8 and 15.2 ka BP and 760 years between 15.2 and 59 ka BP. All acids used during the digestion process were distilled. Approximately 0.08 g of dried sample was digested

with nitric acid (HNO₃) and hydrofluoric acid (HF). Then the samples were treated with perchloric acid (HClO₄) and hydrochloric acid (HCl). The samples were diluted in 2M HNO₃ and spiked with 25 ppm of Re and In prior to analysis on an Perkin Elmer DRC II ICP-MS in the Department of Soil, Water, and Environmental Sciences at the University of Arizona. Samples were digested with ultrapure acid reagents and Millipore water to limit contamination. In order to quantify error associated with the ICP-MS a lake sediment standard (LKSD-3, from Natural Resources of Canada) was analyzed along with a sample blank for every 10 Lake Tanganyika samples. Twenty-two elements were analyzed, but three were below the detection limit of the ICP-MS. All elemental concentrations reported had an error less than 10% based on the standard analyzed, sample blank, and instrument detection limits. Error was quantified based on the amount of the lake sediment recovered and elemental spike (Re and In) concentrations.

2.3 *Grain Size*

Grain size analysis was conducted at the University of Arizona on a Malvern Mastersizer 2000 laser diffraction particle size analyzer with a Hydro 200S sample dispersion accessory following a modified version of the methods described by D. Rodbell (<http://www1.union.edu/~rodbelld/grainsizeprep.htm>). Granulometry samples were collected at an 8 cm sampling interval throughout the core (same sampling resolution as trace elements). Granulometry samples were treated with hydrochloric acid, hydrogen peroxide, and sodium hydroxide in order to remove carbonate, organic matter, and diatom frustules in the sample. Each sample was analyzed for its grain size three

times, and each aliquot injected into the instrument was analyzed three times. The grain size value was computed as an average of the nine values.

2.4 *Biogenic Silica and Total Organic Carbon*

Biogenic silica (BSi) and total organic carbon (TOC) measurements were made at the Limnological Research Center, University of Minnesota, Minneapolis.

Approximately 20 mg subsamples were analyzed for %BSi every 8 cm. The samples were analyzed for %BSi using multiple extractions of hot alkaline digestions at 85°C in 0.5M NaOH following the protocol of DeMaster (1979). BSi data were used to determine the Si from detrital material. Total Organic Carbon (TOC) was measured using a UIC Inc. total carbon coulometer, correcting for carbonate concentrations using a UIC Inc. carbonate coulometer where carbonate was noted in smear slides. This technique measures the amount of carbon dioxide during combustion of the sample with a reproducibility of +/- 0.2%.

2.5 *Sr Isotopes*

Strontium was separated from 20 to 100 mg aliquots of sample utilizing Sr Specresin (Eichrome Industries). Strontium samples were loaded on tantalum filaments with Ta gel to enhance ionization and analyzed in a VG sector 54 multi-collector thermalionization mass spectrometer. Analyses of NBS-987 performed during the study yielded a reproducibility of 0.710247 ± 0.000013 (n=5). Lead isotopes data reported here

are from the RdP xenoliths and from samples that major and trace element and Re-Os and Sr data have been previously reported in Chesley et al. (2002). Lead was separated on Sr Spec resin and analysis was conducted on a Micromass Isoprobe multi-collector ICP-MS at the University of Arizona. Samples were introduced into the instrument by free aspiration with a low flow concentric nebulizer into a water-cooled chamber. Prior to analysis, all samples were spiked with a solution of Tl to achieve a Pb/Tl of approximately 10 (Rehkamper and Mezger, 2000). Throughout the analyses, the standard NBS-981 was run to monitor the stability of the instrument. All results were Hg corrected and empirically normalized to Tl using the exponential law correction after Rehkamper and Mezger (2000). To correct for machine and inter-laboratory bias, all results were normalized to values reported by Galer and Abouchami (1998) for the NBS-981 standard ($^{206}\text{Pb}/^{204}\text{Pb} = 16.9405$ $^{207}\text{Pb}/^{204}\text{Pb} = 15.4963$ $^{208}\text{Pb}/^{204}\text{Pb} = 36.7219$). Internal error reflects the reproducibility of the measurements on individual samples, whereas external errors are derived from long-term reproducibility of NIST 981 Pb standard and resulting in part from the mass bias effects within the instrument. In most cases, external error exceeds the internal errors and is reported here. External errors (2σ) associated with each Pb isotopic ratio are as follows: $^{207}\text{Pb}/^{206}\text{Pb} = 0.021\%$, $^{206}\text{Pb}/^{204}\text{Pb} = 0.017\%$, $^{207}\text{Pb}/^{204}\text{Pb} = 0.019\%$, $^{208}\text{Pb}/^{204}\text{Pb} = 0.014\%$.

2.6 *Geochronology*

Samples were submitted to the University of Arizona's Accelerator Mass Spectrometry Laboratory and the Woods Hole Oceanographic Institute National Ocean

Sciences Accelerator Mass Spectrometry Facility for AMS radiocarbon dating. The chronology of the sediment core was determined from seventeen AMS measurements (**Table 1**) (**Figure 4**). The top of the core does not represent the most modern sediments due to overpenetration of ~0.1m from the Kullenberg piston corer. The two oldest samples (give sample numbers here) are beyond the limit of standard ^{14}C extraction systems (i.e., >40 ka). These samples were prepared for AMS analysis in a new low background vacuum extraction system that is designed to provide reliable ^{14}C ages in the 40-60 ka range (Pigati et al. (submitted)).

Prior to calendar year calibration the basal radiocarbon age estimates were corrected for the reservoir effect of old carbon residing in the DIC pool of Lake Tanganyika (**Figure 5**) (**Table 2**). The reservoir effect was calculated from the offset in age estimates between paired bulk organic matter (sediment) and terrestrial plant material samples collected from the same stratigraphic horizons. The results of these paired radiocarbon age offsets appear to show a great increase in the old carbon reservoir in Lake Tanganyika during the late Holocene relative to the previous interval, probably resulting from much longer residence times for lake water in the past ~2000 years. However, regression-defined offset declines to zero for sediments older than 14 ka ^{14}C , and no corrections were applied to age dates older than this. Because only the older three paired sample age offsets (which define a steeper slope on Figure 5) are from the immediate area of this study, it is possible that there have been different residence time histories in different parts of the lake. However, the effects of variations in this offset history are minimal for our record. This is because a) there is no apparent offset in portion of the record older than 14 ka ^{14}C , which makes up the bulk of our history, and b)

only a few data points are young enough at the top of the core to have significant age offsets. Radiocarbon age estimates younger than 25 ka ^{14}C were calibrated with Calib 5.0.1 (Stuiver and Reimer, 1993; Reimer et al., 2004), and the older radiocarbon age estimates were calibrated following Hughen et al. (2004). Two linear regressions were used to accommodate varying sedimentation rates throughout the sedimentary record (8.5 cm yr⁻¹ from 7.75 m to 3.30 m vs. 22.4 cm yr⁻¹ from 3.30 m to 0 m). Age estimates for the bottom sections of the record (below the lowest ^{14}C date) were estimated by extrapolation of the lower core section's age model. All age estimates discussed in this paper are expressed in calendar years BP unless otherwise indicated.

2.7 *Statistical Analyses*

All time series data was plotted in Excel ®. A principal components analysis was conducted to understand the relationship of elemental chemistry and physical properties data within and between samples in the sedimentary record. Principal components analyses on elemental ratio data for all elements above detection limits were performed with JMP IN 5.1 for Macintosh.

3. **Results**

3.1 *Lithostratigraphy, Physical Properties, Biogenic silica and TOC*

NP04-KH3 is characterized by primarily massive, silty light to dark gray clay with occasional diatomaceous beds from its base at 7.75 m, (~60.0 ka) to 3.38 m (16.0 ka) (**Figure 4**). From 4.96m (32 ka) to 3.57 m (18 ka) the core is characterized by weakly bioturbated sediments. The core consists of laminated diatomaceous ooze alternating with dark organic rich horizons between 3.38m (16.6 ka) to 2.25m (10.7 ka). From 2.25m to 0.80 (10.7 ka to 4.3 ka) the core consists of massive clay. The uppermost portion of the core (0.80-0.00m) is silty clay.

From the base of the core (7.75 m) to 6.77 m (50 ka) the sediments are distinguished by high levels of magnetic susceptibility (21.5 ± 8.8), low to moderate levels of organic carbon ($4.5 \pm 0.8\%$), low levels of biogenic silica ($2.25 \pm 1.6\%$), and low average grain size ($8.73 \pm 3.0 \mu\text{m}$) (Figure 5). The core sediments from 6.77 to 4.98m (50 to 32 ka) are characterized by intermediate levels of magnetic susceptibility (12.8 ± 3.5), intermediate levels of organic carbon ($4.0 \pm 1.2\%$), low to intermediate levels of biogenic silica ($8.8 \pm 8.5\%$), and relatively fine grain size ($10.8 \pm 3.0 \mu\text{m}$), (**Figure 6**). There is a major increase in %BSi ($38.5 \pm 12.9\%$), a profound decrease in magnetic susceptibility (4.0 ± 1.8), a minor decrease in mean TOC ($4.0 \pm 0.7\%$), and an increase in mean grain size ($27.9 \pm 13.1 \mu\text{m}$) at 4.98m (32 ka) which persists up to 3.59m (18 ka). From 3.59 to 3.02m (18 to 14 ka) there are increasing values of TOC ($9.6 \pm 1.9\%$), moderate to high levels of biogenic silica ($17.9 \pm 6.0\%$), low levels of magnetic susceptibility (3.17 ± 1.3), and a fine mean grain size ($12.5 \pm 4.1 \mu\text{m}$). The period between 3.02 and 1.67m (14 to 8 ka) is characterized by moderate levels of biogenic silica ($13.9 \pm 11.5\%$), low levels of magnetic susceptibility (9.0 ± 5.7), a continued increase values of TOC ($8.4 \pm 3.4\%$), and fine mean grain size ($13.4 \pm 7.4 \mu\text{m}$). The

uppermost section of the sedimentary record (1.67 to 0m; 8 to 0.87 ka) has increasing magnetic susceptibility (19.6 ± 7.9), fine grain size ($8.9 \pm 2.1 \mu\text{m}$), and low biogenic silica ($2.0 \pm 1.3\%$) except around 4 ka where the BSi% increases to 15%. TOC is low between 1.67 and 0.62m (8 to 3.5 ka) ($4.9 \pm 0.9\%$), but increases dramatically at the top of the sedimentary record ($11.5 \pm 2.4\%$).

3.2 *Elemental Analyses*

Trends in Al, Fe, K and Ti (as well as many other elements) show a strong inverse correlation with biogenic silica because the latter strongly influences the bulk composition of the core (Figure 7). Therefore, we will present all subsequent discussion of our elemental data in terms of ratios against Al, the most insoluble (under both oxic and anoxic conditions) and common, terrestrially-derived fraction (Brown et al., 2000) (**Figures 8 and 9**).

Between 60 ka and 32 ka (7.75 to 4.98m) values of Fe, Be, Na, Cu (and possibly Ti) ratios against Al are higher than in the following interval. Ratios of these elements, as well as Zn, Mn, K) are also more variable from 60 to 32 ka. From 32 ka to 18 ka BP (4.98 to 3.59 m) Zn, Na, Be, Fe, Cu, and possibly Ti to Al ratios decline. Other elements aside from Co show no significant trend in this interval relative to the earlier one. Co/Al begins to increase at ~32 ka BP and rises dramatically after 22 ka until 10 ka BP (**Figure 9**). At around 18 ka Fe and V begin to increase, whereas Mn, Ti, Cr, Cu, Pb, K, Ni and Be ratios begin to increase later, between 14.3 ka BP (e.g. Mn) to around 12 ka BP, but with the greatest change in all these elements after 12 ka BP (**Figures 8 and 9**). Mg, Ba,

Sr, Co and Zn increase dramatically at 8 ka BP (**Figure 8**). The ratios of Na/Al and Ca/Al increase significantly in the last 0.8 m of the sedimentary record at 4 ka BP.

Several elemental ratios (Fe, Mn, Ti, V, Cr, Cu, and Pb) increase in two distinct periods over the Holocene (**Figure 8**). These elemental ratios are elevated between 12 and 8 ka BP (2.60 to 1.88 m) and again between 6 ka BP (1.18 m) and the top of the sedimentary record.

3.3 *Principal Component Analyses*

The Principal Components Analysis of elemental concentration/Al concentration ratio using correlations data indicates two major components of variance in the data (**Figure 10**) (**Table 3**). The first component accounts for 35.55% percent of the variance (eigenvalue = 6.4), and shows a strong increase at the Pleistocene/ Holocene transition. Elemental ratios (to Al) with strong loadings on the first PCA axis include Mg, Ni, Pb, Cu, Ba, Mn, Cr, K, and V. The first PCA axis appears to combine elements with relatively low variability in the Pleistocene with strongly rising concentrations in the Holocene.

The second PCA axis for the elemental ratio data accounts for 13.3% of the total variance (eigenvalue= 2.4), and corresponds to the dramatic decrease in total non-Si elemental concentrations between ~30 ka BP and ~17 ka BP. Elemental ratios (to Al) with strong positive loadings on the second PCA axis are Sr, Ca and K, whereas there are strong negative loadings for Cu, Be and Fe. The contrast in loadings might be expected if the axis is driven to higher values in correlation with an overall decrease in precipitation,

which would lead to less Be weathering and increased Ca and Sr precipitation. However, the lack of TIC in this section of the sedimentary record leads to evidence of the influence of Lake Kivu input. The strong influence of Lake Kivu tends to mask the more subtle record of the metals earlier in the sedimentary record. The fact that Mg does not show this correspondence with PC2 may indicate that pre-Lake Kivu input into Lake Tanganyika was a system that trended towards low Mg calcite/aragonite precipitation rather than the current Mg calcite from Kivu influence.

3.4 *Sr isotopes*

The preliminary data on $^{87}\text{Sr}/^{86}\text{Sr}$ indicate values of 0.738 and 0.739 at 23.8 ka BP (4.16 m) and 14.6 ka BP (3.21m) respectively. The $^{87}\text{Sr}/^{86}\text{Sr}$ values in the Holocene are 0.731 and 0.725 at 1.8 ka BP (0.23 m) and 1.1 ka BP (0.08 m) respectively (**Figure 11**).

4. Discussion

The geochemical record from core NP04-KH3 shows considerable variability over the past ~60 ka BP which can be interpreted in terms of variation in weathering rates, variable redox conditions in bottom waters, and the influence of upstream contributions from Lake Kivu (**Figure 12**).

4.1 *60 to 50 ka BP- Unit 5*

This interval is marked by the deposition of primarily fine grained, massive silty clays, displaying relatively high but variable magnetic susceptibilities, high Fe/Al, and moderately high Be/Al ratios, all of which point towards relatively strong terrestrial weathering intensity and probably high precipitation/evaporation (P/E) ratios within the adjacent Mahale Mountains watershed. Similarly, low relative concentrations of the divalent cations (Ca, Mg and Sr) point towards little or no carbonate precipitation, also consistent with low salinities and relatively humid conditions during this time. There is evidence of strong weathering of feldspars characterized by increased values of K/Al throughout this interval.

4.2 50 to 32 ka BP- Unit 4

After 50 ka BP there is a significant decrease in magnetic susceptibility and a shift towards lower and less variable elemental ratios for a number of elements, notably in Co, and Fe, accompanied by a shift in lithology to massive silty clays in the older portion of the interval (50-43 ka BP) and diffuse laminations later (43-32 ka BP). The lithologic contrast also corresponds to some elemental differences, with a period of slightly elevated Ti/Al, V/Al and Cu/Al between 50-43 ka BP and a return to lower levels of all of these from 43-32 ka BP. Cumulatively, the evidence points towards a somewhat lower P/E ratio in the surrounding watershed during the 50-32 ka BP interval relative to the preceding period, with possible fluctuations in the depth or stability of the oxicline and mixing zone accounting for the difference in metals and lithology accounting for the differences observed between 50-43 ka BP (deeper oxicline, more vigorous mixing) and

the 43-32 ka BP period. The reduced values of magnetic susceptibility are caused by a reduction in watershed weathering rates and delivery of suspended terrigenous sediment to this topographic high.

4.3 32 to 14 ka BP- Unit 3

This section of the sedimentary record is characterized by occasional intervals of bioturbation, indicative of significantly lowered lake levels or deeper mixing induced by stronger winds and lower temperatures. High grain size, low TOC and low magnetic susceptibility shifts are all consistent with relatively drier conditions during this interval. The coarser terrigenous silt fraction is likely to be of eolian origin, given the core site's location on a topographic ridge, isolated from turbidity flows or other coastal inputs. Proximal dust accumulation in lakes is commonly marked by a marked coarsening of grain size (in the range of 10 to 1000 μm) relative to hemipelagic fines of terrigenous origin (1 to 10 μm) (Fan, 2005). Studies of Saharan dust indicate that eolian grain size ranges from 8 to 50 μm from sedimentary records off the west coast of Africa (Stuut et al., 2005), which correlates with out increased average grain size during the LGM. Grain size in the Kalya core peaked at ~23 ka BP, suggesting that may have been the period of maximum aridity. There are low levels of elemental concentrations (Fe, Be) proportional to Al that are diagnostic of decreased weathering intensity, consistent with decreased precipitation in the surrounding watershed. Co/Al values increase during this interval, dramatically so after 23 ka BP indicating a significant enrichment and reduction in weathering of feldspars and soluble mafic minerals.

4.4 14 to 10 ka BP- Unit 2

Much finer grain size and increases in Be/Al, Ni/Al, Cr/Al, Fe/Al indicate a transition from the previous drier period to a wetter interval. The beginning of this transition is difficult to pinpoint. Some indicators such as grain size pointing towards an earlier start, with grain size falling after about 23 ka BP. Elemental indicators of increased watershed weathering are apparent anywhere from 18-15 ka BP. The transition from diffuse to finely laminated sediments and shifts in TOC and BSi also coincide with this transition. The second PCA axis (**Figure 10**) shows increases in elemental concentrations throughout this section that are also consistent with rising lake level conditions.

Between 12-10 ka BP divalent cations ratios are low, redox sensitive elements (Cr, Cu, Co, Mn, and V) are high, and Fe remains variable, suggesting a deepening in the oxycline. Co/Al values typically follow the trends of Fe and Mn, so during this interval there is increased variation in the Co/Al values, while other metals do not show similar trends (Brown et al., 2002).

4.5 10 ka to 0.8 ka BP- Unit 1

Metal/Al ratios increase at about 10 ka BP, and this trend accelerates in the early Holocene with dramatic rises in K, Mg, Ba, Sr, Be, Cr, Cu, Pb and Ni. Ca and Na ratios increase during the late Holocene around 4 ka BP. These trends indicate a new source of

metals to Lake Tanganyika during the Holocene relative to the Pleistocene portion of our record. Almost certainly this records the first influxes of relatively hard and saline Lake Kivu waters (Haberyan and Hecky, 1987), which is accentuated in the Late Holocene, consistent with earlier investigations of carbonate accumulation and saturation in Lake Tanganyika (Cohen et al., 1997; Alin and Cohen, 2003). Apparently even during the very arid conditions of the Late Pleistocene the solute load derived from relatively insoluble bedrock sources surrounding Lake Tanganyika was insufficient to raise the lake's chemical concentration to the point of carbonate saturation, whereas this could occur under much less arid conditions during the Holocene with the addition of Kivu's saline hydrothermal waters. There is no carbonate in the sediments that predates ~3 ka BP. This would also explain why similar weathering intensities between the 60-50 ka BP period and the late Holocene (reflected in grain size and magnetic susceptibility trends) are marked by such different metal ratios. Lake Kivu did not overflow into Lake Tanganyika until between 16.7 to 13.0 ka BP (14 to 11 ^{14}C ka BP) when the Virunga volcanics to the north of Lake Kivu blocked its northern outflow (Coulter, 1991).

The increase in soluble cation ratios also correlates with the dramatic shift in $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic composition of lake sediments in the early Holocene, interpretable as a signal of the influx of Tertiary basalt-derived Sr into the lake, which previously had only more radiogenic Precambrian and Karoo, Late Paleozoic-aged sources of Sr (**Figure 11**). Some hydrothermal activity has been recorded in the northern portion of Lake Tanganyika (Tiercelin and Mondeguer, 1991), and it might be argued that the dramatic shift in Sr isotopes is a record of hydrothermal activity changes within Lake Tanganyika independent of Kivu sources. However Barrat et al. (2000) analyzed the strontium

isotopes in gastropod shells and lake water from northern Lake Tanganyika near hydrothermal vents. Their results showed isotopic ratios of $^{87}\text{Sr}/^{86}\text{Sr}$ from 0.7152 to 0.7165 for lake waters and 0.72183 to 0.72495 for the hydrothermal aragonite chimneys. Thus the primary source of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in hydrothermal inputs within Lake Tanganyika proper is also a reflection of older, more radiogenic sources and not Tertiary magmatic sources. Our $^{87}\text{Sr}/^{86}\text{Sr}$ ratio differences prior to the Holocene (0.738 and 0.739) vs. during the Holocene (0.731 and 0.725) therefore provide an unambiguous indication of either continuous or episodic overflow of Lake Kivu into Lake Tanganyika.

The increase in concentration of soluble cations during the Holocene appears to have occurred in two phases (**Figure 9**), interrupted by an interval of lower metal/Al ratios that may indicate a brief, more arid period in the middle Holocene. At that time the outlet of Lake Kivu may have ceased flowing, causing soluble metal ratios to fall as solute sources reverted to the older bedrock of the immediate Lake Tanganyika watershed. The increase in all the elements in the late Holocene (~4 ka BP to top of core) is consistent with a renewal of Lake Kivu overflow into the Tanganyika basin in the Late Holocene. By about 2.5 ka BP prior studies have shown that this saline inflow had raised solute concentrations in Tanganyika to the point of allowing carbonate precipitation (Alin and Cohen, 2003).

4.6 Correlation of NP04-KH3 with other regional and global paleoclimate records

Several paleoclimate records are available for East Africa that cover some or the entire interval recorded in the NP04-KH3, allowing us to put our record into a regional perspective. Some interpretations from our record are consistent with these earlier

studies, for example the inference of extreme aridity in the late Pleistocene followed by evidence of dramatic lake level rise and wetter climates during the latest Pleistocene (Haberyan and Hecky, 1987; Gasse et al., 1989; Talbot et al., 2006). However, the NP04-KH3 record provides a number of important new interpretations of regional paleoclimate, in some cases clarifying the timing of important climate transitions, and in others apparently contradicting earlier inferences.

Our record shows a clear signal of significantly relatively moist conditions between 60-50 ka BP, roughly comparable to modern conditions, then a transitional period of moderately humid conditions from about 50-32 ka BP and a sharp onset of aridity thereafter. The only other regional record, which extends back to the 60-50 ka period, is that of Scholz et al. (2003) from the Kavala Ridge of central Lake Tanganyika. Based on TOC, MS, diatom and stable isotope data those authors argued for a significant arid period starting around 55 ka ^{14}C BP (57 ka BP) and continuing into the latest Pleistocene, with brief pulses of stronger aridity at 42, 29 and 23 ka ^{14}C BP (44, 35, 26 ka, respectively). Their interpretation of the early (57 ka BP) onset of aridity was based primarily on a shift toward increased $\delta^{13}\text{C}$ values in organic matter, which is sensitive to a variety of terrestrial and limnological processes (Talbot et al., 2006). Talbot et al (2006) indicate that the $\delta^{13}\text{C}$ signal in Lake Tanganyika is primarily lacustrine and may not indicate a terrestrial weathering signal. A thin bed of woody material in the Kavala sediment core, which they note could have been deposited from floating vegetation, may mark a lowstand at about 57 ka BP (55 ^{14}C ka BP), but our record shows no evidence of a significant lowstand, and indicates relatively wet conditions until 32 ka BP. The

transition at 50 ka BP in the Kalya record may be related to the change at 57 ka (55 ^{14}C ka BP) in the Kavala site, because neither transition is directly dated.

A major result of our study is a clear delineation at a basinal scale of the onset of increased aridity associated with the onset of higher latitude glaciations (MIS 2) at 32 ka BP. This conclusion is in general agreement with pollen stratigraphic evidence in the equatorial highlands of East Africa who conclude that arid conditions prevailed generally in East Africa from 35 to 18.3 ka BP (30 to 15 ka ^{14}C BP) (Bonnefille and Chalié, 2000) (**Figure 12**).

It differs however from the interpretation of Gasse et al. (1989), who argued for an onset of arid conditions at 26 ka BP based on an abrupt increase in benthic/planktonic diatom ratios at that time. We suspect the explanation for the evidence of an earlier onset of aridity in terrestrial (pollen and weathering indicators) vs diatom ratios lies in a taphonomic artifact of benthic fossil accumulation (including benthic diatoms) in Lake Tanganyika. Because this rift lake has extremely steep lake floor slopes in its upper few hundred meters and much gentler slopes in deeper water, the proportion of littoral habitat (where benthic diatoms can be produced) to deepwater environments is relatively insensitive to lake level declines in Tanganyika of up to several hundred meters. Such proportional differences in lake floor morphometry are known to have significant effects on the accumulation of benthic diatoms (Stone and Fritz, 2004). Furthermore, once falling lake level has declined sufficiently to expose the more gently sloping, previously deep water environments, erosion of the now more-expansive littoral deposits would yield an even higher proportion of benthic diatoms to offshore depozones. This combination of effects would produce a benthic/planktonic diatom ratio record that is

both insensitive to the early effects of lake level fall, and then magnifies the ratio signal once the lake falls below the morphometric threshold of decreasing bottom slope. Significantly, the turnaround to higher lake levels would be recorded accurately by benthic/planktonic fossil ratios and it is significant in this regard that all records, whether derived from terrestrial or aquatic indicators, are in agreement on that timing.

Our record suggests a gradual termination of aridity starting perhaps as early as 22 ka BP and continuing until 16 ka BP. Gasse et al. (1989) found evidence for the lowest lake levels around 21.2 ka BP, the latter timing being consistent with our study and the record from Kavala Island Ridge (Scholz et al., 2003). Based on the diatom record from Lake Massoko, a small volcanic crater lake located between Malawi and Tanganyika, Barker and Gasse (2003) proposed that arid conditions existed between 22 and 17.5 ka BP. Stager et al. (2002) analyzed diatoms from Lake Victoria and indicate that the lake level dropped dramatically between 18 and 17 ka BP and also between 15.9 and 14.2 ka BP. Beuning et al. (1997) showed that the Lake Albert region of the western rift valley, north of our study area, was already relatively arid by 35 ka BP (30 ka C^{14} BP) and that this aridity persisted until 14.6 ka BP.

Dates for the termination of East African aridity show that the event was quite uniform throughout the region. Our record, based on both sedimentology and elemental ratio data, indicates a gradual transition to more humid conditions during the late Pleistocene, with some indicators of aridity ending as early as 17 ka BP (Grain size) and others, such as increases in (metals/Al) not apparent until 15-12 ka BP. $\delta^{13}C$ evidence from the Kavala Island Ridge (Scholz et al., 2003), further north in Lake Tanganyika, may indicate an end to arid conditions at around 15 ka BP. Diatom records from the

southern Lake Tanganyika basin cores demonstrated that major lake level increases occurred after the 21.2 ka BP extreme low stand (Gasse et al., 1989). Talbot et al. (2006), in their study of cores from the same region suggested that that major transgressions in the Mpulungu basin of Tanganyika began between 20 and 18 ka BP.

Elsewhere in the region, Fillipi and Talbot (2005) inferred from Lake Malawi records that the post-LGM transgression began at 17.9 until 16.5 ka BP. In contrast, Johnson et al (2002), in their analysis of the transition of periphytic to planktonic diatom dominance in the same lake, suggested that lake levels began to rise at 15.7 ka BP. Street-Perrott et al. (2004) examined a sediment core from Sacred Lake on Mt Kenya indicates the onset of wetter conditions at 14.3 ka BP.

Records from the northern subequatorial region of the Indian Ocean (Gulf of Aden, Arabian Sea) recording conditions in the Sahara and Arabian Peninsula suggests an onset of wetter conditions at 14.8 ka BP, based on dust accumulation (deMenocal et al., 2000), or 15 ka BP based on trace element geochemistry (Sirocko et al., 2000), both of which postdate the timing of evidence from the Kalya Horst region for increased weathering and rising lake levels.

The dramatic increase of the ratio of soluble cations to Al in the Holocene correlates with the overflow of Lake Kivu to the north. Lake Kivu's earlier northerly outflow was blocked at some time during the Late Pleistocene by the Virunga volcanic field (Stoffers and Hecky, 1978; Haberyan and Hecky, 1987). However, outflow through the lower elevation spillway into the Lake Tanganyika basin was delayed during the Late Pleistocene by the prevailing arid regional climate, which kept Lake Kivu well below its threshold. Based on diatom studies in cores taken from both lakes Haberyan and Hecky

(1987) proposed that Lake Kivu began overflowing via the Ruzizi into Lake Tanganyika by 10.6 ka. They further proposed that this flow continued until 3.8 ka BP, when volcanic activity occurred at the south end of the Kivu Basin, cutting off the overflow. Increased weathering and the overflow of relatively saline Lake Kivu waters to the north were occurring simultaneously, resulting in increasing metal/Al ratios for a wide range of elements during the Holocene. Overflow from Lake Kivu continued at 1.3 ka under moister conditions (Haberyan and Hecky, 1987).

Our record suggests a slightly different history. We observe two distinct intervals of high metal concentrations (Figure 8), from ~13 to ~8 ka BP and again from ~6.5 ka BP to the present, which may correspond to different overflow intervals from Lake Kivu separated by a regionally more arid period between ~8 and ~6.5 ka BP when outflow from Lake Kivu ceased. Diatom analyses from a sediment core in the Mpulungu basin (southern Lake Tanganyika) indicate that lake level decreased during this interval (8.4 to 8 ka BP), but not as drastically as the LGM (Gasse et al., 1989; Gasse, 2000). Cohen and Nielsen showed that Lake Elmenteita (Kenya) became smaller and shrank in size starting around 8.8 ka BP (8.0 ka ¹⁴C BP) and persisting for at least several ka.

The elemental record and magnetic susceptibility signal in NP04-KH3 shows a striking correlation with some parts of the global Marine Oxygen Isotope chronology for Marine Isotope Stage (MIS) 3 through 1 (**Figure 12**). Increased elemental concentrations and magnetic susceptibility from 60 to 32 ka BP correspond to MIS 3 and the transition to cooler conditions and MIS 2 at 32 ka BP. However the transgression of Lake Tanganyika at the Kalya horst occurs around 18 ka BP (with some indications of a shift away from maximal aridity such as grain size occurring even earlier) and precedes the

global transition from MIS 3 to MIS 2 at 13 ka. Lake level in Tanganyika begins to increase before the global record indicates warming. This record coincides with evidence of early southern tropical deglaciation in the Andes from Lacustrine sedimentary records (Seltzer et al., 2002).

The Kalya horst record corresponds to rainfall records derived from pollen assemblages in the Burundian highlands and southern Lake Tanganyika (Mpulungu) (Bonnefille and Chalié, 2000). The Marine Isotope record corresponds to increases in elemental ratios in the late Holocene and the fluctuations in the magnetic susceptibility (Martinson et al., 1987). Rainfall comes from the Atlantic Ocean and the Indian Ocean. Increases in rainfall are related to increases in sea-surface temperatures in the western Indian Ocean (Barker and Gasse, 2003). The insolation record at 4°S does not correspond strongly with the records from the Kalya horst indicating that the forces influences the climate changes in the region are more complex than solely astronomical forcings.

5. Conclusions

We have used geochemical and sedimentological records from a core collected in central Lake Tanganyika to infer climatic variability for East Africa over the last 60 ka BP. Geochemical and sedimentological analyses have proven useful at Lake Tanganyika for examining dramatic changes in a lakes' chemistry resulting from both weathering variability related to climate and variability of upstream watershed inputs from the volcanic/hydrothermal sources of Lake Kivu.

We have found evidence for relatively intense weathering and humid conditions (comparable to modern P/E) for the interval 60-50 ka BP. After 50 ka BP and until 32 ka BP weathering records suggest that climate conditions became somewhat drier, although still moderately humid. This was followed by intense aridity with possible evidence for eolian dust deposition between 32-14 ka BP, during the same period of the Last Glacial Maximum. Our findings suggest a rapid transition to the highly arid conditions of the LGM at 32 ka BP, earlier than some previously published estimates for the region, but a more gradual transition out of the arid conditions of the LGM.

The overflow of Lake Kivu into Lake Tanganyika is recorded in the sedimentary record in the Kalya region by increases in elemental concentrations and decrease in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. The addition of Lake Kivu saline/hydrothermal waters starting about 12 ka BP completely reset the baseline of metal concentrations in Lake Tanganyika when contrasting humid vs. arid intervals. The overflow of saline/metal rich water from Lake Kivu into Tanganyika appears to have been punctuated by an arid period from 8 to 6.8 ka BP, when Lake Kivu would have been a hydrologically closed basin. We propose that there were two distinct periods of overflow from Lake Kivu indicated by increased in elemental concentrations.

The paleoclimate record of the Lake Tanganyika basin recorded at the Kalya horst site over the past 60 ka BP shows striking similarities to the marine isotope record of glacial ice volume, and much less correspondence to precessionally driven insolation forcing over the same time interval. The onset of maximal aridity associated with the LGM appears to have been abrupt in the Kalya region, whereas the termination of aridity was more gradual.

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Table 1- Radiocarbon age estimates

Lab accession number	Depth (cm.)	Material	^{14}C Age	Error (^{14}C yr)	Reservoir Corrected (^{14}C yr)	Calib Age (yr BP)	1_range	2_range
WHOI-NOSAMS-50145	72.5	Bulk OM	4380	40	3970	4410	4410-4520	4300-4530
AA61733	174	Bulk OM	7840	51	7560	8330	8340-8410	8210-8450
AA61732	174	Bulk OM	8140	51	7870	8760	8590-8760	8550-8980
AA64614	210	Bulk OM	9110	60	8880	9970	9910-10160	9740-10190
AA67019	232	Bulk OM	9550	59	9330	10500	10430-10650	10300-10700
AA64615	252	Bulk OM	10270	62	10080	11660	11410-11810	11340-11970
AA67020	282	Bulk OM	10620	58	10440	12380	12170-12590	12110-12640
AA67021	312	Bulk OM	12170	84	12060	13920	13820-14000	13750-14100
AA61729	332	wood	13250	69	NA	15930	15500-15890	15340-16110
AA61728	332	Bulk OM	13320	71	13250	15720	15500-15890	15330-16110
WHOI-NOSAMS-50146	362	Bulk OM	14650	70	NA	17540	17590-17960	17240-18020
WHOI-NOSAMS-50142	413	Bulk OM	24600	170	NA	26700	*	*

WHOI- NOSAMS-50143	473	Bulk OM	28300	290	NA	31900	*	*
WHOI- NOSAMS-50392	563.5	Bulk OM	39400	500	NA	41800	*	*
WHOI- NOSAMS-50144	653	Bulk OM	44500	760	NA	44800	*	*
AA68966	593.5	Bulk OM	40090	418	NA	42000	*	*
AA68967	627.5	Bulk OM	38480	384	NA	41500	*	*

(* denotes ages beyond Calib 5.0.1 range-see text for methods.)

Table 2- Radiocarbon reservoir effect
(*-denotes approximate location)

Location	Longitude	Latitude	Water depth (m.)	Age ¹⁴ C years	offset ¹⁴ C years	1sd	Material
Luiche Platform, TZ LT-03-05	29.61E *	4.95 S *	130	500	980	35	stick
Luiche Platform, TZ LT-03-05	29.61E *	4.95 S *	130	595	1225	35	plant
Luiche Platform, TZ LT-03-08	29.60E *	4.97 S *	336	1310	920	40	wood
Kigoma Bay, TZ LT- 03-02	29.585E *	4.876S *	109	1360	925	35	wood
Kigoma Bay, TZ LT- 03-02	29.585E *	4.876S *	109	1435	515	35	wood
Kigoma Bay, TZ LT- 03-02	29.585E *	4.876S *	109	2100	695	35	wood
Luiche Platform, TZ LT-03-05	29.61E *	4.95 S *	130	2430	630	35	charcoal
Kigoma Bay, TZ LT- 03-02	29.585E *	4.876S *	109	2550	505	35	wood
Kigoma Bay, TZ LT- 03-02	29.585E *	4.876S *	109	2550	575	40	wood
Kalya Platform Slope, TZ (LT-00-02)	29 58.306E	6 33.154S	309	2845	458	48	wood
Kalya Platform Slope, TZ (NP04-KH1)	29 58.480E	6 33.147S	303	7838	297	51	wood
Kalya Platform Slope, TZ (NP04-KH1)	29 58.480E	6 33.147S	303	13250	71	69	wood

Table 3- Principal components analysis (PCA) of Me/Al data from NP04-KH3.
Moderately-strongly positive and negative loading ratios are underlined.

Principal Components: on Correlations		
Eigenvalue	6.393	2.4362
Percent	35.5165	13.5343
Cum Percent	35.5165	49.0509
Eigenvectors		
Fe/Al	0.14918	-0.24528
Mg/Al	<u>0.31966</u>	0.02641
Ca/Al	0.06388	<u>0.45233</u>
K/Al	<u>0.25409</u>	0.29563
Na/Al	0.14941	0.12725

Ti/Al	0.23735	-0.04302
Ba/Al	<u>0.26523</u>	0.18228
Zn/Al	0.22461	-0.13511
Mn/Al	<u>0.2592</u>	0.19789
Sr/Al	0.2178	<u>0.45161</u>
V/Al	<u>0.25124</u>	-0.2097
Ni/Al	<u>0.31619</u>	-0.13395
Cr/Al	<u>0.25544</u>	-0.17364
Cu/Al	<u>0.26941</u>	<u>-0.30953</u>
Co/Al	0.17221	0.20454
Pb/Al	<u>0.28586</u>	0.07268
Be/Al	0.2035	<u>-0.28104</u>
Tl/Al	0.19778	-0.14328

Figure 1: Site map with core location and seismic lines from a survey conducted in 2004. Dashed line indicates general location of horst block.

Figure 2: Schematic of two different climate regimes over Lake Tanganyika (modified from Nicholson, 2000). Lake Tanganyika is denoted by (LT).

Figure 3: Seismic line 6a adjacent to core NP04-KH3 location on the west side of the Kalya horst.

Figure 4: *Stratigraphy of sediment core NP04-KH3, and calendar year age model derived for this study.*

Figure 5: *Variability in old-carbon reservoir effect on bulk organic matter radiocarbon ages in Lake Tanganyika sediments. Offsets are expressed as $^{14}\text{C yr}_{\text{bulk lake sediment organic matter}} - ^{14}\text{C yr}_{\text{terrestrial organic matter from same horizon}}$. Data are from Cohen et al. (1997 and 2006), Stager unpubl. and this study.*

Figure 6: *Physical stratigraphy of NP04-KH3 cast against calendar age for: a) biogenic silica, b) magnetic susceptibility, c) total organic carbon, and d) grain size. The lines indicate boundaries of climatic and/or stratigraphic events discussed in the text.*

Figure 7: Concentrations of Al, Fe, K and Ti in the NP04-KH3 core record. The lines indicate different climatic and/or stratigraphic events.

Figure 8: *Elemental ratios relative to Al for a) K, b) Ba, c) Sr, d) Zn, e) Be, f) Ca, g) K, h) Na cast against age and depth in NP04-KH3. The lines indicate different climatic/stratigraphic events described in the text.*

Figure 9: *Elemental ratios for the NP04-KH3 core ratioed against Al for; a) Fe, b) Mn, c) Ti, d) V, e) Cr, f) Cu, g) Pb, h) Co, and i) Ni. Bars indicate different events described in the text.*

Figure 10: *Principal components axes (PCA) 1 and 2 from metal/Al ratio and (correlation matrix) plotted against time*

Figure 11: *Strontium isotope values prior and after the Lake Kivu overflow*

Figure 12: *Summary figure of climate and lake level fluctuations in Lake Tanganyika over the past 60 ka, b) $\delta^{18}\text{O}$ from marine record Martinson et al. (1987); c) Fe/Al (black) and magnetic susceptibility (gray) for NP04-KH3; d) estimation of rainfall as it deviates from modern day mean annual rainfall from Burundi highlands (black) from (Bonnefille and Chalié, 2000); e) estimation of rainfall as it deviates from modern day mean annual rainfall from Mpulungu basin (southern Tanganyika) (Bonnefille and Chalié, 2000); f) insolation at 4°S .*