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Late Pleistocene temperature history of Southeast Africa: A TEX_{86} temperature record from Lake Malawi

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A B S T R A C T

We present a TEX_{86}-derived surface water temperature record for Lake Malawi that provides the first continuous continental record of temperature variability in the continental tropics spanning the past ~74 kyr with millennial-scale resolution. Average temperature during Marine Isotope Stage (MIS) 5A was 26.5 °C, with a range from 25.7 to 27.3 °C, comparable to Holocene temperatures. MIS 4 was a relatively cold period with temperatures generally decreasing from 25.5 °C at 68 ka to a minimum of 20 °C at ~60 ka, 1.5–2 °C colder than the Last Glacial Maximum (LGM). Termination of MIS 4 is characterized by a rapid increase of 3–4 °C in only ~0.5 kyr. Temperatures were relatively stable throughout MIS 3 at the resolution of this study, with an average of 23.8 °C and a range from 25.1 to 22.9 °C. The lack of millennial-scale temperature variability during MIS 3 suggests that Lake Malawi's documented response to the bipolar seesaw (Brown et al., 2007) is not reflected in its thermal history. Our temperature estimates for the LGM and Holocene are consistent with a previously published TEX_{86} record from Lake Malawi with a temperature of ~22.6 °C for the LGM, ~25–26 °C in the mid Holocene and ~25–28 for the late Holocene. In general the present extended TEX_{86} record indicates that temperature variability in tropical East Africa during late MIS 5 and MIS 4 was as great as that associated with the deglaciation and Holocene. A decrease in Southern Hemisphere insolation between 70 and 60 ka may have played an important role in forcing temperatures during MIS 4, but after 60 ka other factors, such as the extent of the polar ice sheets, or atmospheric CO_2 may have forced temperature in tropical Africa to a greater extent than local summer insolation.

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1. Introduction

Great progress has been made in recent decades in deciphering the history of thermal change in the surface waters of the ocean. Patterns of past sea surface temperatures (SST's) have elucidated the scope and magnitude of climate change associated with tectonic reorganization of ocean basins, the gradual cooling of the Earth through the Cenozoic, and the waxing and waning of polar ice caps during the past few million years. Moreover, the recognition of temporal shifts in SST patterns has led to an understanding of certain aspects of the history of global climate dynamics for example: the persistence of El Niño or La Niña conditions in the Pacific (Cobb et al., 2003; Lea et al., 2006), the tie between SST patterns in the tropical Pacific and Indian Oceans (Charles et al., 2003) and the meridional gradient of Atlantic SST's and the African monsoon (Schefuss et al., 2003).

Until recently there have been only a few comparable records of past temperature change from the continents that are as quantitative, of comparable temporal resolution and duration, and are as continuous as marine sediment core records. A few records of continental paleotemperature have been produced from thermal gradients measured in drill holes (Chisholm and Chapman, 1992), noble gas concentrations in ancient groundwater (Kulongoski et al., 2004), and pollen and other fossil records from lakes and bogs (Coetzee, 1967; Chalie, 1995), but these typically have been limited in temporal resolution and/or duration.

There are a number of open questions about continental climate dynamics that remain unanswered due to a lack of quantitative continental temperature records. What is the terrestrial thermal history in the tropics through glacial/interglacial cycles? To what extent are tropical temperatures associated with millennial-scale abrupt climate change in polar regions? Have tropical temperatures been as constant through the Holocene as recorded in the ice core records of Greenland?

Most methods that might be used to reconstruct past temperatures from lacustrine systems are equivocal because of complicating

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effects that are often difficult to quantify. Potential temperature proxies such as $\delta^{18}O$ and Mg/Ca ratios of carbonates that have been utilized in marine sediments are not readily applied to lake sediments. Many lake records lack carbonate, but even in lakes that contain calcareous sediment, paleotemperature reconstructions are hampered by uncertainties in the isotopic composition of the lake water, various biological effects, and changes in the regional hydrological cycle (von Grafenstein et al., 1999; Huang et al., 2002; Hu and Shemesh, 2003). Mg/Ca ratios in ostracode shells are affected by dramatic shifts in the Mg/Ca ratio of lake water as a function of evaporation, and by biological effects that vary among species (Holmes, 1996). Long chain alkenones that are used to construct past sea surface temperature records in the ocean (based on the $U^{K'}_{37}$ proxy) are found in certain lakes, but their occurrence is scattered and the manner in which the alkenone based $U^{K'}_{37}$ proxy relates to temperature is different among species (Zink et al., 2001; D'Andrea and Huang, 2005). Lake water temperature reconstructions have been generated by analysis of fossil chironomid assemblages, but such reconstructions are often limited to relatively shallow, oxic and short lived systems, and they can be affected by other environmental and biological effects (Walker, 2001).

A relatively new geochemical proxy for past water temperature, the TEX_{86}, appears to hold promise for certain lacustrine systems. The TEX_{86} index was first established in marine sediments (Schouten et al., 2002) and is based on the relative distribution of glycerol dibiphytanyl glycerol tetraethers (GDGTs), which are membrane lipids produced by aquatic *Crenarchaeota*. These organisms are ubiquitous and abundant in seawater (Hoefs et al., 1997; Karner et al., 2001), and large lakes (Keough et al., 2003; Powers et al., 2004). Schouten et al. (2002) identified a strong correlation between mean annual sea surface temperature (SST) and the relative abundance of cyclopentane rings in GDGTs in the marine water column and surface sediments (Schouten et al., 2002). Enrichment studies of non-thermophilic *Crenarchaeota* also show a strong relationship between relative distributions of cyclopentane rings in archaeal GDGT membrane lipids and growth temperature (Uda et al., 2001), and analysis of particulate organic matter from water column filters and sediment traps also shows a strong correlation of TEX_{86} with temperature (Wuchter et al., 2005, 2006). Thus, by measuring the relative amounts of different cyclopentanes containing GDGTs in sediments, the temperature of the growth environment of the *Crenarchaeota* (i.e., surface waters) can be reconstructed using the TEX_{86} proxy proposed by Schouten et al. (2002). Powers et al. (2004) adapted this method for use in lacustrine systems and subsequently calibrated the TEX_{86} for use in lakes (Powers, 2005) showing a similar strong correlation between mean lake surface temperature and the relative abundance of cyclopentane rings in surface sediments. Powers et al. (2005) applied the TEX_{86} to a sediment core from Lake Malawi and produced one of the first high resolution and continuous lake temperature records since the Last Glacial Maximum (LGM) on the African continent. This temperature record shows that Lake Malawi experienced considerable temperature variability since the LGM and indicates a substantial thermal response of southeastern tropical Africa to deglaciation, as well as significant temperature variability during the Holocene (Powers et al., 2005). More recently (Weijers et al., 2007) produced a mean air temperature record from the Congo basin that extends back to the LGM by using the newly introduces MBT proxy, that is closely related to TEX_{86} (Weijers et al., 2007), but which is based on fluvially transported lipids from soil bacteria to reconstruct the thermal history.

Here we apply the TEX_{86} paleotemperature proxy to a drill core from the north basin of Lake Malawi to produce the first continuous, millennial-scale continental temperature record from Africa spanning the past 74 kyr. The time period between 74 ka and the LGM allows us to assess the thermal behavior of southern East Africa during the last glacial period and investigate the thermal response of Lake Malawi to millennial-scale climate variability, which was especially prevalent in MIS 3.

2. Background

2.1. Study location

Lake Malawi (9°S to 14°S, Fig. 1), the most southern of the East African Rift lakes, is situated in the countries of Malawi, Mozambique, and Tanzania. The lake is 560 km long, up to 75 km wide and has a maximum depth over 700 m, currently; the lake is permanently stratified, anoxic below ~200 m (Eccles, 1974) and has relatively high sedimentation rates of 0.5–1.5 mm/yr (Finney et al., 1996). Lake Malawi is thought to be at least 5 million years old (Finney et al., 1996) and has accumulated over 4 km of sediment (Specht and Rosendahl, 1989). Water loss at Lake Malawi takes place primarily through evaporation in the lake rather than outflow, making the lake sensitive to subtle changes in aridity (Spigel and Coulter, 1996). During time of severe droughts, when other African lakes completely desiccated (Gasse, 2000), Lake Malawi continued to contain water and

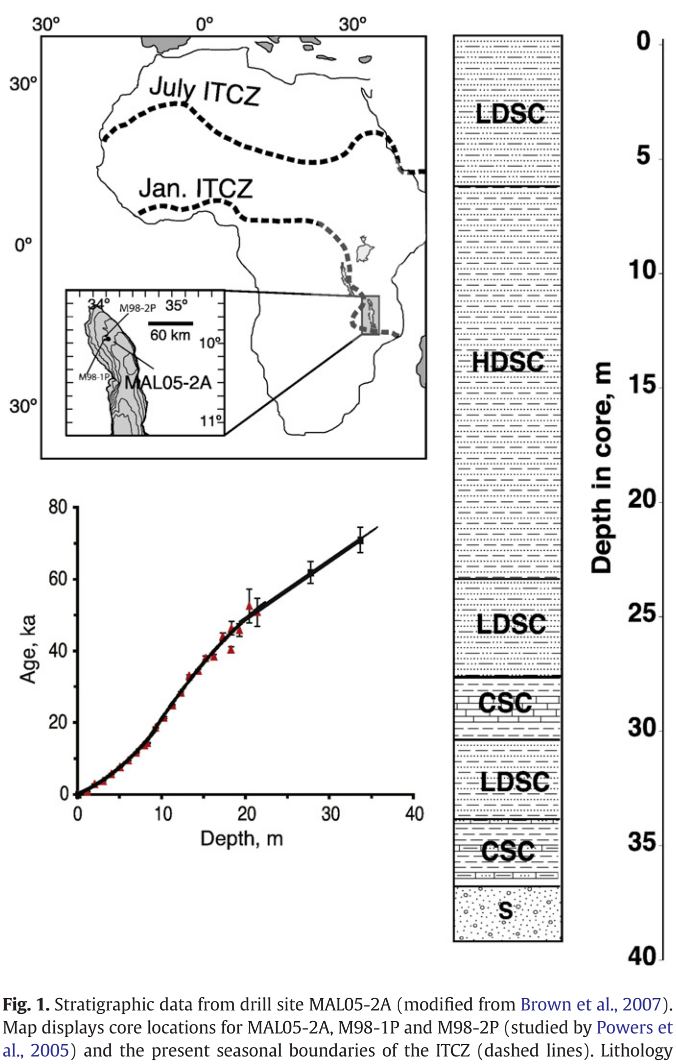

Fig. 1. Stratigraphic data from drill site MAL05-2A (modified from Brown et al., 2007). Map displays core locations for MAL05-2A, M98-1P and M98-2P (studied by Powers et al., 2005) and the present seasonal boundaries of the ITCZ (dashed lines). Lithology abbreviations: LDSC, laminated diatomaceous silty clay; HDSC, homogeneous diatomaceous silty clay; CSC, calcareous silty clay; S, sand. The age model is based on 24 radiocarbon dates (red triangles). The two oldest dates (black triangles) are two carbonate intervals that are stratigraphically correlated with core MAL05-1C, which was dated by OSL, Be-10 and paleomagnetic intensity correlations to an Indian Ocean core (Brown et al., 2007).

accumulated a high-resolution sedimentary record of climate change in its deeper basins (Scholz and Rosendahl, 1988; Johnson, 1996).

Lake Malawi is an important, albeit complex, site for paleoenvironmental reconstructions as it is situated in a densely populated and climatically sensitive location. Lake Malawi is located at the southern limit of yearly migration of the Intertropical Convergence Zone ITCZ (~13°S) and experiences a single rainy season each year. The warm and wet season occurs during austral summer (November–March) when the ITCZ is located just to the south of Lake Malawi, and a cold/dry season characterizes austral winter (April–September) when the ITCZ is located over North Africa. Surface water temperatures at the north basin of Lake Malawi range from a maximum of ~29 °C during the austral summer to a minimum of ~26 °C during austral winter. (Eccles, 1974). Hypolimnetic temperature is presently 23.7 °C (Vollmer et al., 2005).

3. Methods

3.1. Sediment core

We analyzed 60 sediment samples at an average depth interval of about 0.6 m from core MAL05-2A that was drilled in the northern basin of Lake Malawi at 10°01.1′S and 34°11′ E in 345 m water depth during the Lake Malawi Drilling Project in May 2005 (Fig. 1). Sediments in MAL05-2A provide a continuous record of deposition from the lake floor to a prominent seismic reflector at 38 m below the lake floor (mblf). Here near-shore sands overlie an erosional unconformity, marking the end of a series of major droughts that occurred prior to 75 ka (Scholz et al., 2007). The sediments of MAL05-2A consist mainly of homogeneous and laminated diatomaceous silty clays and two intervals of calcareous silty clays at 28–31 and 33–35 mblf overlying near-shore sands at 37 mblf (Fig. 1).

3.1.1. Age model

The age model for MAL05-2A is based on 24 radiocarbon dates on bulk organic matter in the upper 22 m using the Fairbanks calibration curve (Fairbanks et al., 2005), and stratigraphic correlation of 2 carbonate-rich horizons in MAL05-2A with equivalent dated carbonate intervals from hole MAL05-1C from the central basin of the Lake Malawi (Brown et al., 2007; Scholz et al., 2007). In MAL05-1C these intervals were dated by paleomagnetic, Be^{10} and OSL methods (Scholz et al., 2007). The sedimentation rate in MAL05-2A is assumed to have been steady between the 2 carbonate-rich horizons. This correlation assigns an age for the near-shore sands of 75 ka and 62 ka for the upper calcareous horizon. The uncertainty in the age of these intervals was assumed to be ±5%, yielding an uncertainty of ±3–4 ka for the lower part of the sediment core (Brown et al., 2007). Based on this age model sedimentation rates have ranged between 0.3 and −0.6 m/kyr, with faster sedimentation rates in the Holocene and before 50 ka when lake level was rising after prolonged drought (Brown et al. 2007) (Fig. 1).

3.2. Total organic carbon analysis

Concentrations of total carbon (TC) and total inorganic carbon (TIC) were determined using a UIC Carbon Coulometer. Total organic carbon (TOC) was determined by the difference between TC and TIC.

3.3. GDGT analysis

Between 0.3 and 2 g of freeze-dried, homogenized sediment were extracted with dichloromethane (DCM)/methanol (2:1) using Soxhlet extraction. An aliquot of the total lipid extract was then separated into apolar and polar fractions using an activated Al_2O_3 column and eluting hexane/DCM 9:1 and DCM/methanol 1:1. The polar fraction (DCM/methanol) was dried, dissolved in hexane/2-propanol and filtered over a 0.45 μm PTFE filter prior to analysis. Measurement of the glycerol dibiphytanyl glycerol tetraethers (GDGTs) was performed on an HP 1100 series LC-MS, using an Alltech Prevail Cynano column (2.1 × 150 mm, 3 μm) with isocratic elution of hexane and 2-propanol. GDGTs were detected using atmospheric pressure chemical ionization mass spectrometry (ACPI-MS) in single ion mode (SIM) according to Schouten et al. (2007b).

GDGT peak areas were determined by integration of the area under the peak using a linear baseline. In between our samples we analyzed an external standard containing a C_{46} tetraether lipid external standard (described in (Huguet et al., 2006)) with a known concentration and used its response plus a correction for the difference in ionization efficiency between this lipid and crenarchaeol (isoprenoid GDGT lipid that contains 3 cyclopentane rings and 1 cyclohexane ring), to quantify isoprenoid and branched GDGT lipid concentrations.

TEX_{86} values were calculated as follows: $TEX_{86} = (II + III + IV')/(I + II + III + IV')$ Fig. 2 following Schouten et al. (2002) and converted to temperatures using the (Powers, 2005) lake transfer function of $TEX_{86} = 0.017T + 0.25$. The standard error of the published regression is 0.034 resulting in a calibration error of ±2.0 °C. In addition, the method according to Schouten et al. (2007b) for GDGT determination yields analytical error of ±0.3 °C, observed analytical error in duplicate measurements of the Lake Malawi sediments was 0.13 °C.

The BIT (branched vs. isoprenoid tetraether) index, which is a proxy for determining the ratio of soil vs. aquatic derived GDGT's (Hopmans et al., 2004), was determined by the ratio of the concentration of branched GDGTs/(crenarchaeol + branched GDGTs) according to (Hopmans et al., 2004).

3.3.1. Quality control

Samples of MAL05-2A were measured in duplicate on different days. To examine the analytical reproducibility between our record and the Powers et al. (2005) record, we analyzed 2 samples from Powers et al. (2005) record among each sequence and compared our integrations and TEX_{86} temperatures for these with the previously published values. These were within 0.4 °C of the average temperatures published by (Powers et al., 2005).

Three of the 60 samples analyzed were excluded from our temperature reconstruction. The two deepest samples that just overlie the near-shore sand had GDGT concentrations that were too low to allow determination of TEX_{86}, while the 3rd oldest sample was dominated by terrestrial organic material (as determined by the BIT index). Previous research has shown that the TEX_{86} can be biased by relatively high soil derived GDGT inputs to aquatic sediments (Weijers et al., 2006). A BIT value close to 1 represents pure soil derived GDGT's and a BIT value close to 0 represents pure aquatic derived GDGT's (Hopmans et al., 2004). The third deepest sample yielded a BIT value of 0.75, suggesting that it was likely biased by the influx of land derived GDGTs (Weijers et al., 2006). All other samples from MAL05-2A yielded BIT values in the range of 0.06–0.32, values thought to indicate that terrestrial inputs are relatively low, and therefore are unlikely to bias the TEX_{86} significantly (Weijers et al., 2006).

4. Results

Eighteen samples of a total of sixty from our MAL05-2A core overlap in age with the previously published TEX_{86} record of (Powers et al., 2005). Both records come from the north basin of Lake Malawi, but from different locations (Fig. 1). The overall trend in both records is the same (Fig. 3), although some of the MAL05-2A samples appear to yield slightly lower temperatures It is possible that the apparent deviations between the two records reflect true temperature variation in Lake Malawi, because the samples analyzed from MAL05-2A were not taken from precisely the same time intervals as those of the Powers et al. (2005) record.

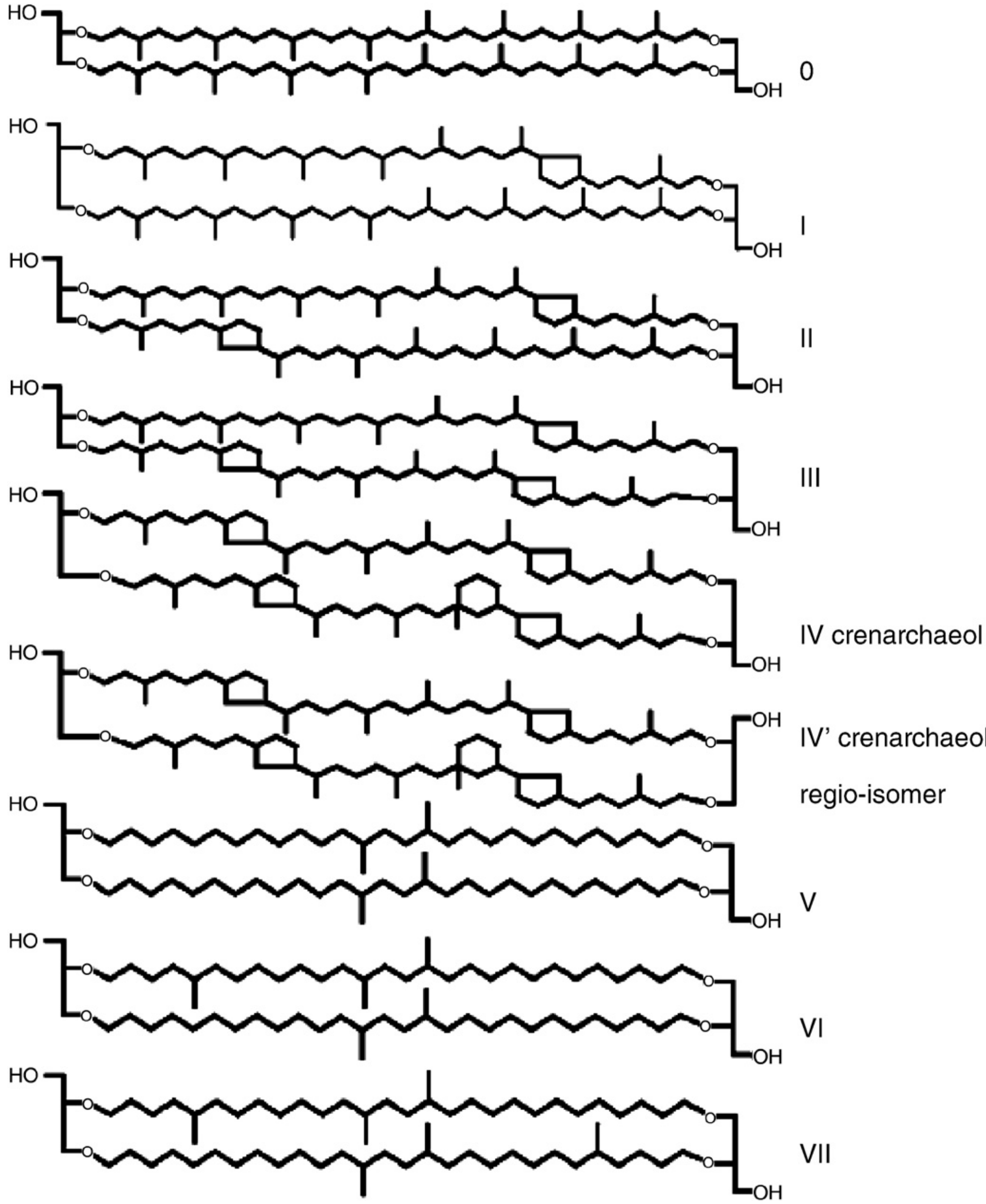

Fig. 2. Chemical structures of the membrane lipids (GDGTs) that were analyzed in this study. Lipids 0–I–II–III–IV and IV′ are isoprenoid GDGTs, which are thought to be mainly produced by archaea. To determine the TEX_{86}, lipids I–II–III and IV's are quantified and then TEX_{86} was calculated: $TEX_{86} = (II + III + IV')/(I + II + III + IV')$. Lipids V–VI–VII are branched GDGTs, that are thought to be mainly produced by a still to be identified soil bacterium. These lipids are quantified and together with GDGT IV are used to determine the BIT index: $BIT = (V–VI–VII)/(V–VI–VII + IV)$.

The MAL05-2A TEX_{86} temperature record shows that Lake Malawi experienced substantial temperature variation between the LGM and 74 ka, exceeding the range observed since the LGM (Fig. 4). Temperatures varied between ~20 and 27 °C from 74 ka to the LGM. The record covers a small part of Marine Isotope Stage (MIS) 5A with an average temperature of 26.5 °C and a range of 25.7–27.3 °C, comparable to temperatures observed during the Holocene. Substantial cooling occurred between 68 and 60 ka, when temperature decreased systematically by more than 5 °C. This cooling period roughly coincides with MIS 4. Average temperature during this period was ~23 °C with a range from 25.5 °C at ~68 ka to 20 °C at ~60 ka. This temperature minimum at ~60 ka was followed by a period of abrupt warming at the beginning of MIS 3, when temperature increased 3–4 °C in less than 1 ky. Temperatures during MIS 3 were relatively stable with mean temperature of ~24 °C, with reconstructed temperatures between 22.5 and 25 °C from 59 to 49 ka and between 23 and 24.5 °C from ~49 to 22 ka. Temperature variance in MIS 3 appears to have been significantly smaller compared to observed variance in the Holocene.

5. Discussion

5.1. Interpretation of TEX_{86} as surface water temperatures

We interpret the TEX_{86}-derived temperatures to reflect lake surface water (or upper water column) temperatures (LST) that are thought to be closely related to mean annual air temperatures (MAT) as has been observed in the lakes of the TEX_{86} calibration dataset (Powers et al., 2005). However, upwelling of cold hypolimnetic waters, a change in the Crenarchaeotal species composition or a change in the ecology of the GDGT producing *Crenarchaeota* could potentially disturb this relationship and could complicate the interpretation of our temperature record.

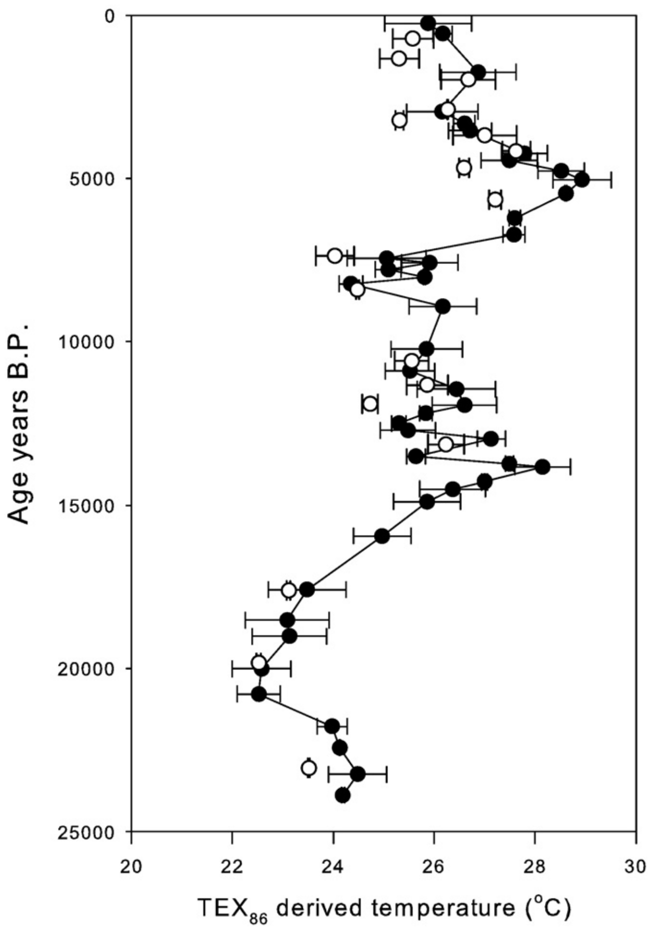

Fig. 3. Comparison of temperatures from 18 samples of MAL05-2A (open circles), with the Powers et al. (2005) reconstruction (black circles) coming from M98-1P and M98-2P shows a similar trend of temperatures between the LGM and the present in cores from different locations in the north basin of Lake Malawi.

Seasonal upwelling can occur in the north basin of Lake Malawi. Comparison of our TEX_{86} temperature record with a record of upwelling intensity based on Si/Ti ratios in MAL05-2A (Brown et al., 2007) does not show a consistent relationship, suggesting that the MAL05-2A temperature record is not significantly influenced by changes in upwelling intensity. This observation is consistent with the conclusions of the Powers et al. (2005) TEX_{86} study in the north basin of Lake Malawi which compared TEX_{86}-derived temperatures to the mass accumulation rate of biogenic silica, a reflection of upwelling intensity, in the same core (Johnson et al., 2002). Further evidence that upwelling did not significantly affect the TEX_{86} temperature record in the north basin in Lake Malawi comes from a 17 ky record from the central basin, where upwelling is expected to be less intense than at either end of the rift lake; that record strongly resembles the Powers et al. (2005) record from the north basin over the same interval (Berke et al., 2007).

We do not know if a Crenarchaeotal species composition change would alter the relationship between the TEX_{86} and LST. Batch culture experiments on marine *Crenarchaeota* observed genetically different species of group 1 *Crenarchaeota* dominating during different incubation experiments, with lower and higher temperature culture experiments yielding almost identical correlation between growth temperature and TEX_{86} (Wuchter et al., 2004; Schouten et al., 2007a). This result suggests that the different species of *Crenarchaeota* that grew in the experiments altered their membrane composition in a similar fashion in response to temperature (Schouten et al., 2007a). Likewise the correlation between temperature and TEX_{86} is very similar between the marine and lacustrine calibration datasets, implying that adjustment of cell membrane composition to temperature is essentially the same between lacustrine and marine *Crenarchaeota* species (Powers 2005; Schouten et al., 2007a; Kim et al., 2008).

A change in the ecology of the *Crenarchaeota* that produce the lipids used in the TEX_{86} could potentially alter the prevailing water depth at which temperature is reflected by the TEX_{86}. If most of the GDGTs settling to the lake floor were formed by Crenarchaeota at substantial depth below the lake surface, colder temperatures than LST's would be recorded. A change in temporal ecology to a certain season would create a seasonal bias in the temperature signal. At present day Lake Malawi's thermocline lies at 80–100 m depth (Vollmer et al., 2005). It is likely that the GDGTs that are settling toward the sediment are produced mainly in the epilmnion, but the vertical distribution and bloom period of *Crenarchaeota* has not yet been studied in Lake Malawi. At present we have no basis for speculating on past ecological changes that may have potentially affected our temperature record. Thus with recognizing the above caveats, we interpret the TEX_{86} temperature record as representing a history of LST and recognize that this interpretation may have to be modified after ecological studies of Crenarchaeota are carried out on Lake Malawi and the other African tropical lakes.

5.2. MIS 4

The most striking feature of the MAL05-2A TEX_{86} record is the cold temperature observed during MIS 4, especially around 60–62.5 ka, when the lake temperature was even colder than during the LGM. Interestingly, during this period TOC and isoprenoid GDGT concentrations are the highest of the entire record (Fig. 4). We do not see a consistent relationship between the TEX_{86} temperature record and the Si/Ti ratio, a measure of the abundance of diatom frustules and an independent record of upwelling intensity (Fig. 3) (Brown et al., 2007). It is therefore unlikely that the temperature signal at this time is solely the result of changes in upwelling intensity in the north basin.

We compare the Lake Malawi TEX_{86} temperature record to two ice core records, the EDML (Barbante et al., 2006) and NGRIP (NGRIP-members, 2004) oxygen isotopic records, which can be used to infer temperature changes in the polar regions of Antarctica and Greenland, respectively, and to the closest marine record that covers MIS 4, a Mg/Ca based sea surface temperature (SST) record from the Gulf of Guinea in the equatorial Atlantic ocean (MD03-2707 (Weldeab et al., 2007), Fig. 6). Taking into account the uncertainty in the MAL05-2A age model for this time section (±3–4 kyr), there is a fairly good correlation between the timing and duration of cool temperatures during MIS 4 among all records. The Lake Malawi record shows a unique trend compared to the other records, in that it displays minimum temperatures toward the end of MIS 4 at ~60 ka, whereas the other records display relatively cool temperatures early in MIS 4, followed by gradual warming into MIS 3 in the EDML and Gulf of Guinea records. Temperatures in Lake Malawi during the last ~3000 years of MIS 4 were lower than those of the LGM, in contrast to the EDML record that displays colder temperatures at the LGM, while the NGRIP and Gulf of Guinea records show similar temperatures for both periods. The rate of warming at the termination of MIS 4 in Malawi was significantly faster than that observed in the EDML and Gulf of Guinea records, and was as abrupt as the rate of warming in the NGRIP record. This correlation might suggest the possibility that warming in Malawi and Greenland were forced by a similar mechanism, though no solid conclusion can be drawn based upon the current data.

MIS 4 stands out from the rest of the MAL05-2A record; it has the largest temperature variation, coldest temperatures, highest TOC content and largest relative isoprenoid GDGT flux. A high TOC content and high relative isoprenoid GDGT flux may indicate higher production and/or better preservation of organic matter during MIS 4 than at other

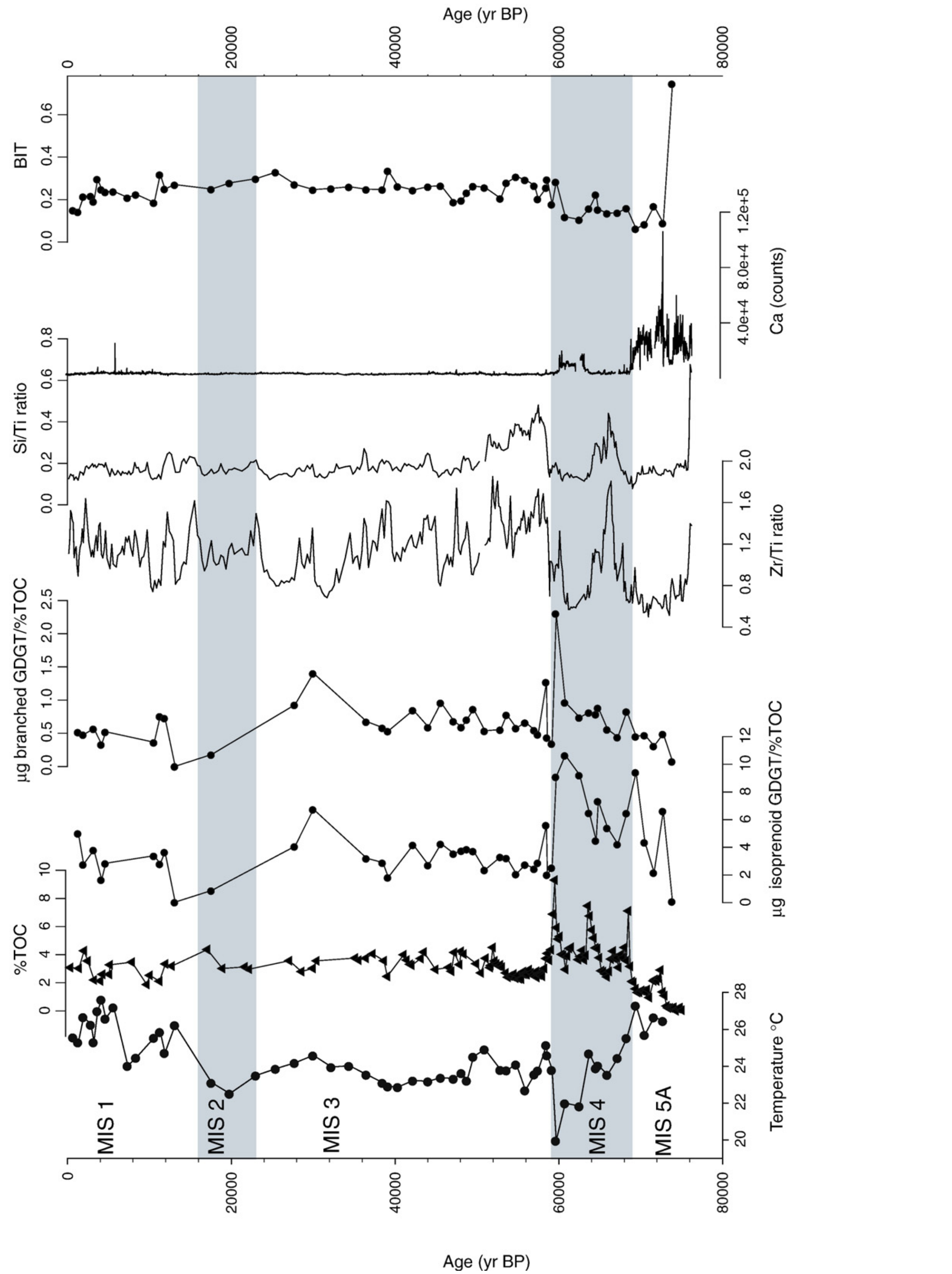

Fig. 4. Different parameters analyzed in MAL05-2A showing TEX_{86}-derived temperature, % TOC, isoprenoid GDGT/%TOC, µg isoprenoid GDGTs/%TOC, µg branched GDGTs/%TOC, Zr/Ti, Si/Ti ratios, calcium concentration (Brown et al., 2007) and BIT index. Glacial MIS are shaded. Isotopic stages after (Shackleton and Opdyke, 1973).

times, especially between ~62.5 and 60 ka, the period when temperatures were lower than the LGM. The MAL05-2A age model (Fig. 1) for the period of MIS 4 is not sufficiently constrained to determine whether the increased concentrations of TOC and GDGT's are the result of greater burial rates of these components or of less dilution by other sediment components.

The lake level history is reflected somewhat by the Ca abundance curve (Fig. 4), which is a measure of endogenic calcite abundance in the sediments preserved during low stands of the lake (Brown et al., 2007; Scholz et al., 2007). Lake level apparently was relatively low between 65 and 62.5 ka, after which it started to rise again (Scholz et al., 2007). Si/Ti ratios in MAL05-2A indicate high diatom abundance, suggesting that upwelling in the north basin was relatively intense around ~68.5–64 ka (Brown et al., 2007). This could potentially have resulted in lower surface water temperatures and shifted our TEX_{86} temperatures for this time period. Kim et al. (2008) observed in an upwelling area that TEX_{86} temperature is be lower compared to the annual mean surface water temperature, which may be the result of higher GDGT fluxes occurring during the upwelling seasons, which are characterized by colder nutrient rich waters. However, lowest TEX_{86} temperatures within MIS 4 do not occur at this time, but between ~63 and 60 ka, a period without signs of intense upwelling, but of some carbonate preservation (Fig. 4).

The non-thermophilic *Crenarchaeota* that produce the GDGT lipids utilized in the TEX_{86} paleothermometer are thought to be aerobic ammonium oxidizers, though other metabolic functions may also be present within this understudied Kingdom level group of microbes (Venter et al., 2004; Francis et al., 2005; Konneke et al., 2005). Thus, high relative fluxes of isoprenoid GDGTs produced by *Crenarchaeota* during MIS 4 could indicate a period of high ammonium availability in the water column. High TOC values during MIS 4 could indicate periods of high overall production in the lake which would require sufficient nutrient availability, but the Si/Ti ratio indicates that diatom productivity did not increase during this time. Further research on the ecology of the *Crenarchaeota* that produce the lipids used in the TEX_{86} paleothermometer could provide more information about the conditions may have stimulated Crenarchaeotal production in Lake Malawi.

5.3. MIS 3

MIS 3 is characterized by millennial-scale temperature variability associated with D–O events in both ice cores and, to a lesser extent, in 2 nearby marine sediment core SST records that cover at least a portion of MIS 3 from the Gulf of Guinea (Weldeab et al., 2007) and the Mozambique channel (MD792575) (Bard et al., 1997). The MAL05-2A record shows relatively stable temperatures without major cooling or warming trends throughout MIS 3. This record differs from a recent TEX_{86} record from Lake Tanganyika that reflected some substantial temperature fluctuations with a general cooling trend from 60 to 40 ka (Tierney et al., 2008) (Fig. 6). MAL05-2A displays a strong signal of D–O-type oscillations in the wind field and hydrological cycle (attributed to ITCZ dynamics) (Brown et al., 2007), but the TEX_{86} temperature record does not show millennial-scale variability during MIS 3. Sample resolution might not have been high enough to detect these events in the TEX_{86} record; however, 6 samples of the Lake Malawi TEX_{86} temperature reconstruction were deliberately selected to coincide with exact depths where Brown et al. (2007) observed D–O-type peaks in Zr/Ti ratios, while 4 samples in the TEX_{86} record were obtained from depths with low Zr/Ti ratios from the same cores. (Fig. 5). No significant temperature difference was observed between Zr/Ti peak and non-peak samples, so we conclude that it is likely that there is no significant temperature response of Lake Malawi to D–O type events.

The absence of a Malawi temperature response to millennial-scale events in MIS 3 is consistent with a similar record from nearby Lake Tanganyika, in which significant millennial-scale variability in precipitation (δD of plant leaf waxes) is not matched by one in temperature (TEX_{86}) (Tierney et al., 2008). Although the Lake Malawi and Tanganyika temperature records show a similar response to deglaciation and no response to millennial-scale events in MIS 3, the LST trends within MIS 3 differ between these two lakes. Between 60 and 40 ka the Tanganyika record shows relatively large temperature variability with an overall cooling trend of ~3.5 °C while the MAL05-2 record in this interval exhibits less variability without a substantial warming or cooling trend (Fig. 6). The lower resolution of the MAL05-2A record might contribute to the lack of temperature variability that is seen in Tanganyika. Tierney et al. (2008) report the Tanganyika record tracks Northern Hemisphere summer insolation on an orbital time scale (Laskar et al., 2004) and, over the same time interval (60 ka to present), the Malawi temperature record follows much the same trend (Fig. 7). There are minor, regional differences, such as between 35 and 30 ka, when the temperature trends between the two lakes are opposite (Fig. 6), but for the most part there appears to be a Northern Hemisphere influence on the temperature of the Tanganyika and Malawi basins over the past 60 ka. This relationship does not hold up in the Malawi basin prior to this time (Fig. 7).

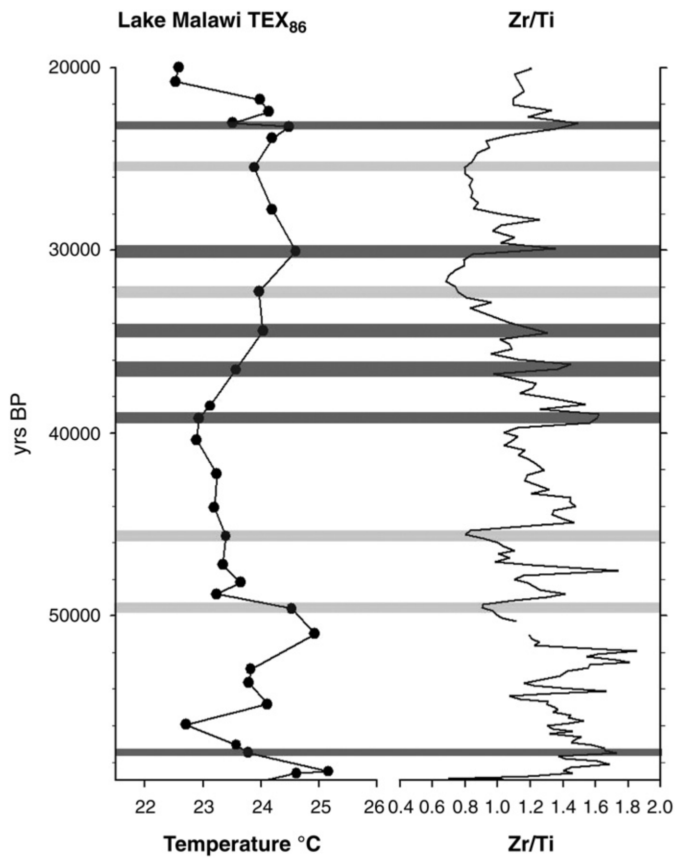

Fig. 5. Comparison of MAL05-2A TEX_{86} temperature record with MAL05-2A Zr/Ti record (Brown et al., 2007) during MIS 3. Light shading indicates events that line up with Greenland interstadials and dark shading indicates events that line up with Greenland stadials. Six of the TEX_{86} samples line up with Greenland interstadials and four with stadials.

5.4. Correlation of GISP-2 methane and temperatures in Lake Malawi

Temperature change in the Malawi basin has been observed to correlate positively with changes in precipitation (Castaneda et al., 2007). Brook et al. (1999) and Dallenbach et al. (2000) postulated that trends of atmospheric methane concentration observed in high latitude ice cores primarily reflect hydrological change in the tropics. We do not observe a high degree of synchroneity between the MAL05-2A temperature reconstruction and the GISP-2 methane record over the 74 kyr time span of the present study (Fig. 7). However, there appears to be some correlation between the warming observed at ~59 ka and the increase in atmospheric methane concentrations in the GISP-2 record at ~58 ka. Another temperature increase that appears to be synchronous with atmospheric methane can be seen at ~51–49.5 ka in the Malawi record compared to 52.5–49 ka in GISP-2. These two intervals may potentially reflect wetter conditions in the

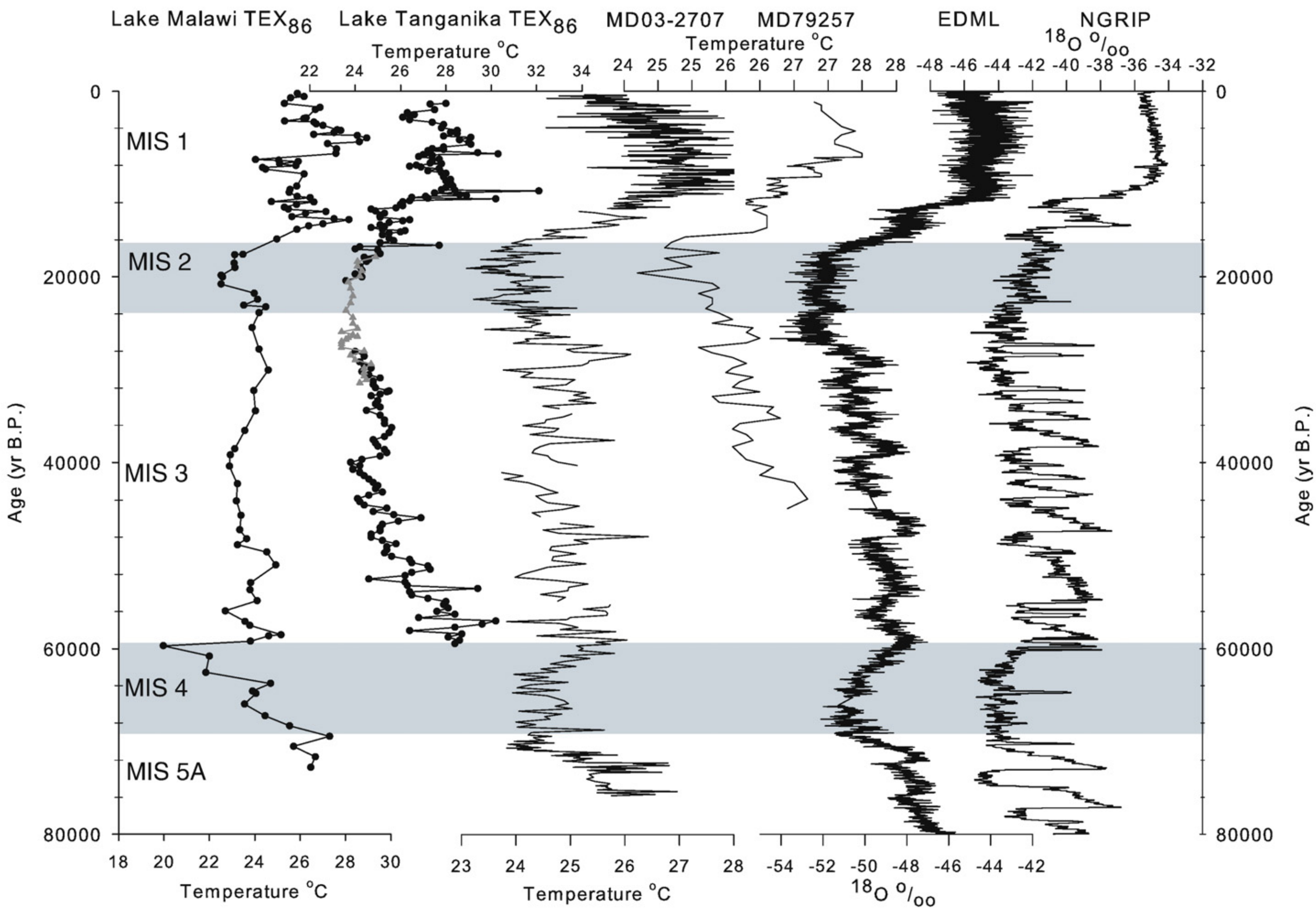

Fig. 6. Malawi TEX_{86} temperature reconstruction (combination of MAL05-2A and (Powers et al., 2005) TEX_{86} records) compared to the Lake Tanganyika TEX_{86} temperature record (Tierney et al., 2008), oxygen isotopic records from the Antarctic EDML (Barbante et al., 2006) and Greenland NGRIP (NGRIP-members, 2004) ice cores and two marine sediment SST reconstructions; MD03-2707 from the equatorial Atlantic ocean in the Gulf of Guinea (Weldeab et al., 2007) and MD79257 from the Indian Ocean in the Mozambique Channel (Bard et al., 1997). Glacial MIS are shaded. Isotopic stages after (Shackleton and Opdyke, 1973).

Malawi basin, contributing to the apparently enhanced methane production in the tropics during these times. After these two periods there appears to be little correlation between the Malawi temperature and GISP-2 methane records, especially with regard to millennial-scale variability during MIS 3. Sediment samples from MAL05-2A are currently being analyzed for proxies of past hydrological conditions in the Lake Malawi watershed to assess possible ties to global methane production during the last glacial period.

6. Conclusions

The MAL05-2A record is the first record of continental temperature from the east African tropics that extends to 74 kybp. This record shows that Lake Malawi experienced large temperature variability during MIS 4, but relatively low variability in MIS 3, at least at the resolution of our sampling interval. At the end of MIS 4 temperatures in Lake Malawi were significantly colder than during the LGM, a feature different from both ice core and nearby marine records. Temperatures during MIS 3 show no sign of millennial-scale temperature response to the bipolar seesaw, which is in agreement with a TEX_{86} temperature record from Lake Tanganyika.

Covariance between insolation and the MAL05-2A TEX_{86} record during ~70–60 ka suggests that insolation may have played an important role in forcing lake temperatures in Lake Malawi. However after 60 ka these two parameters are anti-phased just as in Lake Tanganyika, suggesting that summer insolation in the Northern Hemisphere is somehow having significant influence over temperature in the southern tropics of East Africa.

More quantitative, high-resolution climate records from the African continent are needed to assess whether observed trends in Lake Malawi and Tanganyika temperature are unique to their basins or are representative of the African tropics in general. Understanding the dynamics of the abrupt temperature shifts observed during the termination of MIS 4 and the apparent relative temperature stability of MIS 3 in the Malawi basin will improve our ability to predict future climate in the region, and offer new insight into the influence of climate on the dispersal of African flora and fauna, including our human ancestors during their migration within and out of Africa.

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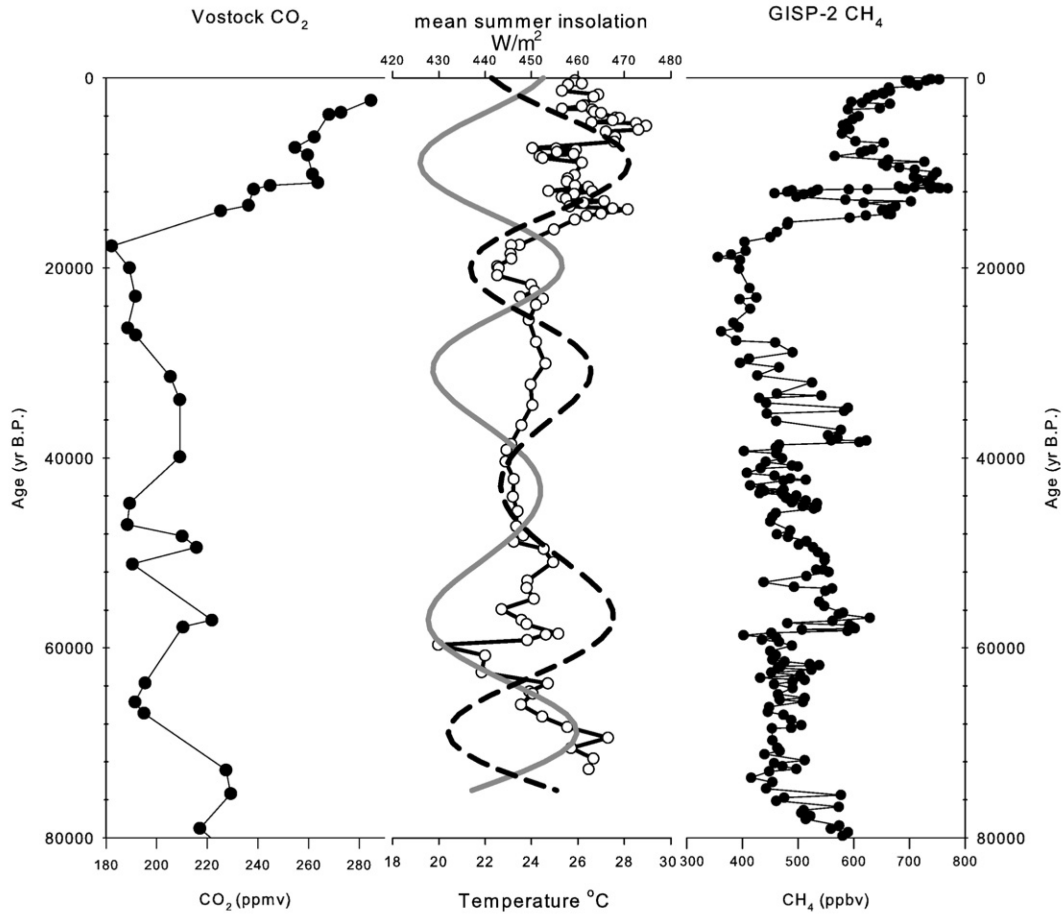

Fig. 7. Vostok CO_2 record (Petit, 2005) plotted together with mean summer (3 month) insolation at 10° S (solid grey line) and 30°N (Laskar et al., 2004), MAL05-2A TEX_{86} temperature record (white circles), and the GISP-2 methane record. ppmv = parts per million by volume, ppbv = parts per billion by volume.

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.palaeo.2010.02.013.

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