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Fossil diatoms and the mid to late Holocene paleolimnology of Lake Turkana, Kenya: a reconnaissance study

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Abstract

A 12 m sediment core recovered from the south basin of Lake Turkana, northwestern Kenya, reveals four major diatom assemblages that span approximately 5450 to 1070 years BP based on AMS radiocarbon analyses. The oldest assemblage, Zone D (5450 to 4850 yr BP), is dominated by *Melosira nyassensis* and *Stephanodiscus* spp. and is interpreted to reflect higher lake levels, fresher water and more variable seasonal mixing of the water column than the modern lake. *Melosira* dominates the assemblage in Zone C (4850 to 3900 yr BP) with some *Surirella engleri* and *Stephanodiscus*. This assemblage indicates a continuation of relatively high lake levels and seasonal mixing of a stratified lake. The brief peak of *Surirella*, interpreted as benthic, suggests an episode of slightly lower lake level. *Thalassiosira rudolfi* and *Surirella* predominate since the beginning of Zone B (3900 to 1900 yr BP), reflecting a decrease in lake level and increase in water column salinity. Increasing dominance of *Surirella* in Zone A (1900 to 1070 yr BP) may suggest that the lake continued to decrease in depth. Salinity probably rose to levels comparable with the modern lake. These results are consistent with paleoclimatic interpretations based on carbonate abundance, lamination thickness, oxygen isotope and bulk geochemistry profiles from this core and cores recovered from the north basin. It extends the known paleolimnology beyond 4000 yr BP of the earlier research to 5450 yr BP and into the early to mid Holocene pluvial phase in northern intertropical east Africa.

Introduction

Lake Turkana is the largest closed-basin lake in the East African Rift System (Fig. 1). It is approximately 250 km long with a mean width of 30 km. A narrow and shallow bathymetric sill, adjacent to the Turkwel and Kerio Deltas, divides the lake into two basins. Water depths across the sill are less than 40 m whereas maximum depths in the north and south basins are approximately 80 and 100 m, respectively. Moderate salinity (2.5 ppt) and alkalinity (20 meq l^{-1}, pH 9.2) characterize the present lake (Yuretich & Cerling, 1983). The entire water column is well mixed throughout the year by strong, diurnal winds (Ferguson & Harbott, 1982). The sediments are predominantly terrigenous silty clays derived primarily from the Omo River that drains the Ethiopian Highlands to the north (Yuretich, 1979; 1986). Eolian material may also be important, especially in the southern portions of the basin (Yuretich, 1979; Finney *et al.*, submitted). Measured sedimentation rates are between 1 and 5 mm yr^{-1} (Ferguson & Harbott, 1982; Yuretich, 1979; Barton & Torgersen, 1988; Halfman Johnson, 1988; Halfman & Hearty, 1990).

The basin, located in the desolate Northern Frontier District of Kenya, has been the site of active paleoclimatic research in recent years (e.g., Owen *et al.*, 1982; Johnson *et al.*, 1987; Cerling *et al.*, 1988; Halfman & Johnson, 1988). Previous paleoclimatic reconstructions focused on radiocarbon dating of exposed lacustrine deposits within the basin that are up to 80 m above the present lake (Butzer *et al.*, 1972; Owen *et al.*, 1982). At about +80 m, the lake overflows into the Nile system. An early to mid Holocene high stand (+70 to 80 m) from 10000 to 4000 yr BP is consisten with records of lake level throughout northern intertropical east Africa and is predicted by computer simulations of monsoon climates modulated by millennial scale Milankovitch perturbations (Street & Grove, 1976; Hamilton, 1982; Street-Perrott & Harrison, 1983; Kutzbach & Street-Perrott, 1985; COHMAP, 1988; Pachur & Hoelzmann, 1991). Another, slightly lower, highstand (+60 to 80 m) is inferred between about 3800 and 3000 yr BP (Butzer *et al.*, 1972; Owen *et al.*, 1982) that is less consistent with other proxy records from east Africa (Halfman & & Johnson, 1988; Johnson *et al.*, 1991). These elevations are relative to the 1976 level of the lake which has since declined by a few meters (Halfman, personal observations).

Recent paleoclimatic research has focused on several 12 m piston cores (Kullenberg) recovered from the north basin of the lake. Down-core changes in lamination thickness, and abundance and isotopic composition of fine-grained calcite have been related to variability in the Omo River discharge and are thus interpreted to reflect the hydrologic balance within the Omo Basin (Halfman & Johnson, 1988; Halfman *et al.*, 1989;

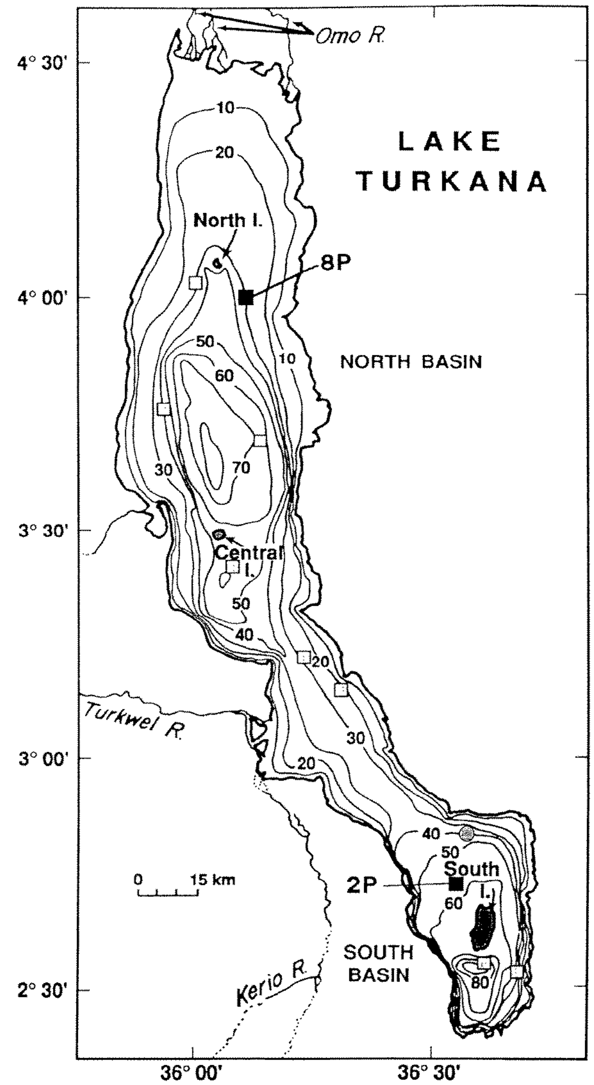

Fig. 1. Bathymetric and core location map of Lake Turkana. Contour interval is 10 m. Solid squares are core sites discussed in this paper. Open squares are other sites for cores recovered in 1984. Open circle is the approximate location of Barton's Mackereth core (Barton & Torgersen, 1988).

Johnson *et al.*, 1991). These records did not recover sediments older than 4000 yr BP and reveal no evidence of a major highstand between roughly 3000 and 3 800 yr BP.

This paper presents diatom abundances from a piston core, LT84-2P, recovered from the south basin of the lake. These data are used to reconstruct the lake's paleolimnology since about

5450 yr BP. Diatoms, organic matter, ostracodes and finegrained carbonate are more abundant in the south basin (Yuretich, 1979; 1986; Halfman & Hearty, 1990). Our goal is to compare the paleolimnological interpretations based on the diatom data from the south basin core with other lake-level proxies from this core and with previous reconstructions derived primarily from the north basin.

Methods

Core LT84-2P was recovered in 1984 from approximately 60 m of water (Fig. 1). The core was lined with PVC tubing with a diameter slightly less than 6 cm. The tubing was sectioned in 1 m lengths, sealed and flown back to the laboratory for storage at 4 °C and subsequent analysis. Twelve new accelerator radiocarbon (AMS) dates were obtained commercially from this core to supplement the geochronology provided by four conventional radiocarbon dates (Halfman & Hearty, 1990). Carbonate was selected for dating because: (1) the organic carbon content is low (< 1.5 wt.%: Halfman, 1987); (2) SEM photography and trace, pore water and stable isotope geochemistry suggest that the carbonate is primarily in two distinct phases: euhedral, fine-grained, low-Mg calcite that precipitates near the surface of the water column, and ostracode carapaces; (3) the lake's high alkalinity and well mixed nature promotes rapid equilibrium of carbon dioxide between the atmosphere and the lake; and, (4) previous results suggest reliable data can be obtained from analysis of the bulk carbonate fraction (Halfman & Johnson, 1988; Halfman & Hearty, 1990).

At four selected depths (Table 1), separate fine-grained calcite and ostracode carapace samples were prepared from 3-cm intervals of core. The fine-grained calcite was isolated from the bulk mud by wet sieving at 63 μm and retaining the fine fraction. Ostracode carapaces (*Sclerocypris clavata*) were hand picked from the coarse fraction and gently washed with distilled water. At 176 cm, the sediment lacked sufficient ostracodes for analysis at a 3-cm interval, so an 8-cm interval was used. The fine-grained calcite fraction was isolated from the 3 and 8-cm intervals and analyzed separately. In addition, fine-grained calcite was isolated and analyzed at three more depths. The reported dates were adjusted for ^{13}C content and used a half-life of 5568 years. For all AMS samples, $\delta^{13}C$ and $\delta^{14}C$ values were measured directly by the accelerator. The $\delta^{13}C$ values were not measured for the previous conventional radiocarbon samples but were assumed to be

Table 1. Radiocarbon AMS ages from core LT84-2P.

Depth (cm)	[a] Radiocarbon age (year BP)	[b] Sample material	ID number
41–44	1930 ± 55	Fine-grained calcite	Beta 40727, ETH 7174
175–178	1910 ± 65	Fine-grained calcite	Beta 36347, ETH 6367
173–181	1860 ± 60	Fine-grained calcite	Beta 37513, ETH 6582
173–181	1670 ± 75	Ostracodes	Beta 36348, ETH 6463
455–458	3170 ± 70	Fine-grained calcite	Beta 36349, ETH 6468
455–458	2735 ± 70	Ostracodes	Beta 36350, ETH 6369
605–608	4085 ± 75	Fine-grained calcite	Beta 36351, ETH 6368
605–608	3385 ± 80	Ostracodes	Beta 36352, ETH 6389
911–914	4600 ± 60	Fine-grained calcite	Beta 40728, ETH 7175
1023–1026	4865 ± 70	Fine-grained calcite	Beta 36353, ETH 6390
1023–1026	4760 ± 85	Ostracodes	Beta 36354, ETH 6354
1157–1160	5600 ± 65	Fine-grained calcite	Beta 40729, ETH 7176

[a] Reported age is adjusted for total isotopic effects.
[b] Fine-grained calcite is all carbonate < 63 μm & Ostracodes are hand-picked *Sclerocypris clavata* larger than 63 μm.

zero. The quoted errors represent one standard deviation of the counting statistics.

Smear slides for diatom enumeration and assemblage analysis were prepared using standard techniques at approximately 10-cm intervals down core (92 samples). Qualitatively, the same amount of sediment (i.e., approximately 1 mg of sediment from a well-mixed, cubic-cm, subsample of core) was used for each slide. Diatoms were counted at 400 × and 1000 × on a light microscope until 400 frustules were identified. Only recognizable fragments representing more than one-half of a valve were counted. Because diatoms are less concentrated in the upper 3 m of core (counts averaged less than 400 frustules per slide), we counted all recognizable valves on the slide. Identifications were based on Hustedt (1930, 1949), Van Meel (1954), Gasse (1980) and Stoermer & Håkanson (1983), which are summarized in Gasse (1986). *Stephanodiscus astraea* has been previously identified in the literature as either *S. hantzschii*, *S. minutula*, or *Cyclostephanos* forms, the taxonomy is in a state of flux. We refer to our specimens as *S. astraea*. Percent frequency of occurrence was calculated for common species (> 10% of all frustules). Less common species were lumped into an 'Other Species' category.

Geochronology

The relationship between radiocarbon age and depth is generally linear in core LT84-2P with some deviations and a non modern core-top age (Fig. 2;. Table 1). Two previously published conventional radiocarbon dates from 730 and 1175 cm down core are about 550 and 900 years younger than suggested by a best-fit, linear regression of all of the data. The range of independent $\delta^{13}C$ measurements on finegrained and ostracode carbonate fractions in this core (−1.2 and 1.3‰, PDB: Halfman *et al.*, 1989) is too small to cause the discrepancy because it could only alter the age by at most 50 years. The discrepancy may reflect differences between conventional and AMS techniques, and different origins of the carbonate fractions analyzed.

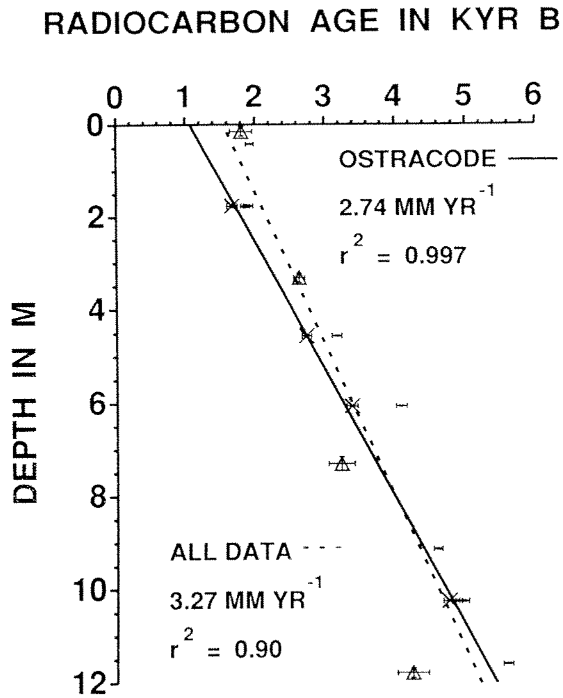

Fig. 2. Radiocarbon age versus depth down core LT84-2P. The AMS ages are adjusted for total isotopic effects in nature and during physical and chemical laboratory procedures. Triangles are conventional radiocarbon analyses of bulk carbonate samples (Halfman & Hearty, 1990), crosses are AMS radiocarbon analyses of hand-picked *Sclerocypris clavata* ostracode carapace samples, and the remainder are AMS radiocarbon analyses of fine-grained calcite samples. The solid and dashed lines are the best-fit, linear regressions through the ostracode and all the samples, respectively. The error bars represent the depth interval and 1 standard deviation of the counting statistics. The ages for the bulk carbonate samples are adjusted for total isotopic effects assuming a $\delta^{13}C$ value of 0‰.

Four pairs of fine-grained calcite and ostracode samples, each pair isolated from the same depth interval, reveal statistically different ages. The ostracode material is always younger than the fine-grained calcite. The differences average 350 yr, but are not consistent down core. All of these dates are corrected for isotopic fractionation effects, thus exclude natural (e.g., photosynthetic) and laboratory induced fractionation from consideration. In addition, core-top samples from two cores recovered from the north basin of the lake reveal 'modern' ages (Halfman & Johnson, 1988; Halfman, unpublished data) and two fine-grained calcite samples from approximately 175 cm down core LT84-2P reveal statistically identical ages. This supports an earlier hypothesis that the carbonate is precipitating in equilibrium with atmospheric carbon dioxide. Another possibility is that a variable but relatively small amount of old detrital carbonate, that is smaller than 63 μm, is accumulating in the sediment. Changes in the supply of old material through time could account for the variable difference in ages. Possible mechanisms to supply older carbonate include eolian transport, which is important for the input of silicate material to the lake (Yuretich, 1976; Finney *et al.*, submitted), and reworking of exposed lacustrine deposits, which probably occurs when lake-level lowers.

Ostracode carapaces sieved from the sediment contained abundant quantities of juvenile forms, were not fragmented and lacked signs of abrasion. This suggests that they are probably less susceptible to eolian transport, and/or reworking, and subsequent transport over long distances than the fine-grained calcite. In addition, SEM observations, trace element and stable isotope geochemistry data lacked evidence for calcite recrystallization on ostracode carapaces (Halfman *et al.*, 1989). For this paper, we will use the time scale based on the ostracode samples rather than using all the data.

Linear regression of the ostracode ages versus depth ($r^2 = 0.997$) and all the radiocarbon ages versus depth ($r^2 = 0.90$) are generally similar in this core (Fig. 2). The largest exception is at the top of the core where the difference for the interpolated core-top ages of the ostracode samples and all the samples is approximately 500 years. This may reflect an increased supply of detrital fine-grained calcite as lake level lowers within the upper portion of the core. The ages of the diatom assemblage boundaries (see next section), calculated by the two linear regressions, are within approximately 25 (statistically indistinguishable) to 300 years. The best agreement corresponds to the major shift in diatom assemblage at about 3900 yr BP. We interpret these differences to reflect the accuracy of the radiocarbon chronology.

The linear regression of the ostracode data reveals a core-top age of 1070 yr BP. The non-modern age probably reflects over penetration of the core head beyond the sediment/water interface, a conclusion supported by: (1) offsets in pore water geochemistry between the modern lake and the trigger core, and between the trigger core and the piston core; (2) the lack of excess ^{210}Pb in the trigger core; and, (3) very fluid nature of the sediments (>95% water contents, Halfman & Hearty, 1990). The best-fit, linear sedimentation rate of 2.74 mm yr^{-1} is consistent with other estimates at other sites based on ^{210}Pb, paleomagnetic and radiocarbon chronologies (Barton & Torgersen, 1988; Halfman & Johnson, 1988).

Diatom stratigraphy and paleolimnology

The diatom profiles were divided into four major diatom zones, with Zone A being the most recent and D the oldest (Fig. 3). This is the first enumeration of diatom variability within the upper 12 m of sediments from Lake Turkana. The core-top diatom assemblage is similar to those found in the modern sediments, so the recovery of more recent sediments probably will not define a more recent zone, and we hope that future coring will penetrate deeper into the section. Ages were assigned to sample depths based on a linear sedimentation rate of 2.74 mm yr^{-1} and a core-top age of 1070 yr BP (Fig. 2).

Zone D (1200 to 1040 cm, 5450 to 4850 yr BP) is dominated by relatively abundant but rapidly fluctuating *Melosira* and *Stephanodiscus* populations. The dominant *Melosira* species is *M. nyassensis* var. *victoriae*, but also includes rare amounts of *M. italica* var. *bacilligera* and *M. granulata*. The major *Stephanodiscus* species are *S. damasii* and *S. astraea*, with the former representing typically 50 to 70% of the *Stephanodiscus* spp. Limnologically, *M. nyassensis* requires turbulent waters that have relatively low alkalinity, conductivity, moderate pH, and an elevated ratio of dissolved silica to phosphorus (Lund, 1955; Hecky & Kling, 1987; Haberyan & Hecky, 1987). *M. nyassensis* is presently most common in the southern ends of

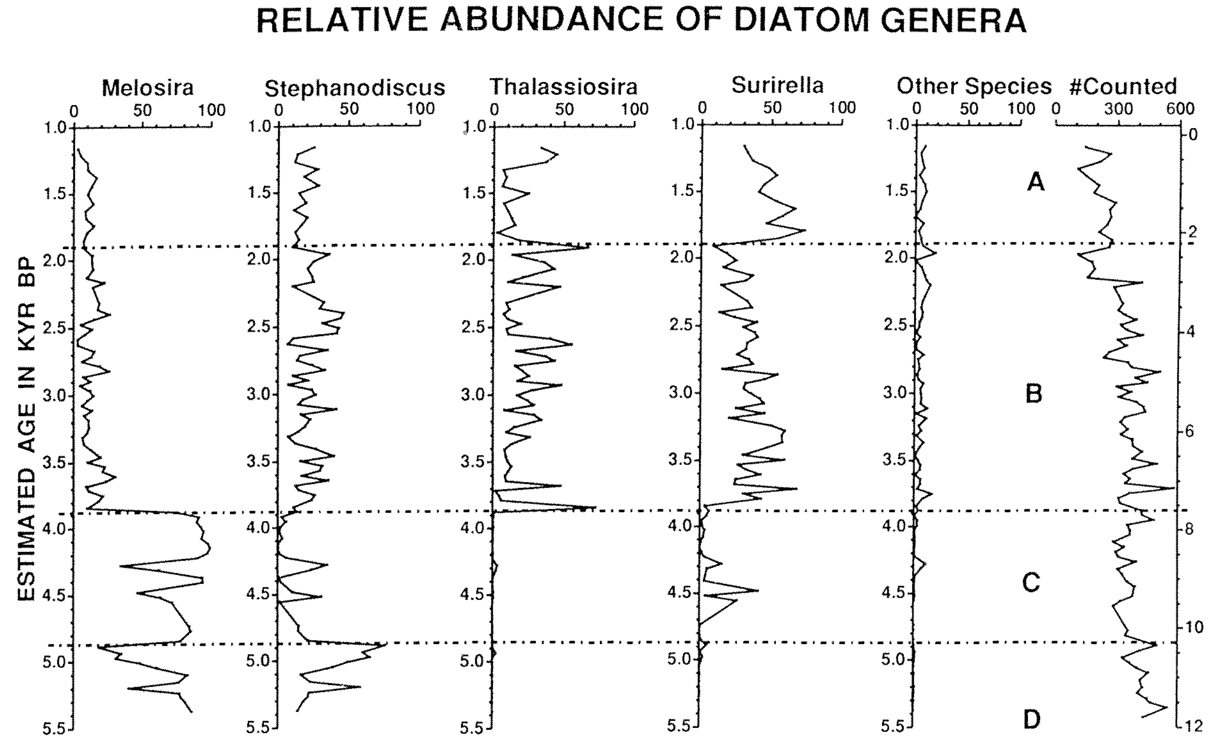

Fig. 3. Relative abundance of common diatom taxa versus depth (right) and estimated age (left) from core LT84-2P. Other species are all forms representing less than 10% of the total count at any depth. Labels A-D are diatom assemblage zones discussed in the text. Note that the age scale starts at 1000 yr BP.

Lakes Tanganyika and Malawi when silica-rich bottom waters are upwelled to the photic zone during the windy season (Haberyan & Mhone, 1991). *S. damasii* is common in Lakes Albert and Edward, and thus seems to prefer medium conductivity and low alkalinities (500–1000 μS cm^{-1}: Gasse, 1986). *S. astraea* is widespread in many lakes of varying salinities and alkalinities. *S. astraea* is found in the same lakes as *M. nyassensis*, but flourishes when dissolved silica concentrations are relatively depleted (Kilham, 1971; Richardson, 1968). In Lake Malawi, *Stephanodiscus* is prevalent during the initiation of and later discontinuance of upwelling events, i.e., it brackets the dominance of *Melosira* species (Haberyan & Mhone, 1991).

We interpret Zone D to represent a lake that is much fresher and deeper than present. Both *Melosira* and *Stephanodiscus* species are reported from the Galana Boi Formation that was deposited during the early Holocene pluvial phase. Specifically, these localities are approximately 70 to 80 m above the present level of Lake Turkana (Owen *et al.*, 1982). Seasonal stratification probably occurred; epibenthic ostracodes in this core suggest that the bottom waters were at least periodically oxygenated (D. Van Nieuwenhuise, personal communication, 1991). This is in contrast to the continually well-mixed water column by diurnal winds at the lake today. Subsections of Zone D reflect alternating dominance by *Melosira* or *Stephanodiscus* and suggest prolonged periods (> 100 yr) of altered water column mixing, or fortuitous sampling of upwelling and non-upwelling seasons. If the former is true, then such changes may be induced by variations in the duration or velocity of the prevailing winds, or the intensity of seasonal cooling and thermally-induced bottom water mixing.

Zone C (1040 to 770 cm, 4850 to 3900 yr BP) is very much like Zone D except that *Melosira* species dominate the entire interval. This may imply continual upwelling of nutrient rich bottom waters that maintains an elevated ratio of dissolved silica to phosphorus in the photic zone on par with the *Melosira* dominated intervals in the previous zone. Perhaps the change is due to increased wind velocity or duration of the windy season, or seasonal cooling. *Surirella engleri* appears in this zone as well. *Surirella* species are generally regarded as benthic forms (Richardson, 1964; Round, 1961; Haberyan & Mhone, 1991), even though they are found in the plankton of the shallow, wind-swept Lake Victoria (Talling, 1966; 1969). In Lake Malawi, *Surirella* is nearly exclusively benthic and littoral (Haberyan & Mhone, 1991). A down-core increase in the abundance of *Surirella* in Lake Tanganyika led Haberyan & Hecky (1987) to suggest that the core site was relatively closer to shore, a source of resuspended littoral material, due to a lower lake level. The bathymetry of the south basin at Lake Turkana is steep; variations in lake level will not dramatically change the proximity of a core site to the shoreline. However, lowered lake levels increases the availability of recently deposited lacustrine material that is easily resuspended, or subaerially eroded by water or wind. The occurrence of *Surirella* throughout Zone C may imply lake levels slightly lower than those during earlier times, and the peak of *Surirella*, centered at about 4500 yr BP, represents an abrupt lowering to the lowest lake levels during Zones C and D. A lack of *Thalassiosira* suggests this episode was brief enough to preclude a major increase in water salinity, alkalinity or colonization by *Thalassiosira*. Alternatively, the occurrence of *Surirella* may represent a period of greater wind strength and resuspension of shallow water sediments. This interpretation is more consistent with the dominance of *Melosira*.

Zone B (770 to 220 cm, 3900–1900 yr BP) is dominated by *Surirella* and *Thalassiosira rudolfi* with minor concentrations of *Melosira* and *Stephanodiscus* species. The transition from Zone C to B is abrupt, within one sample interval. The boundary marks the first sustained appearance of *Thalassiosira*. Only one of 32 depths below the transition revealed more than 5 valves per slide and most depths did not reveal any valves of *Thalassiosira*. *Thalassiosira* prefers medium to high conductivity and alkalinity, with a pH over 8.5 (Richardson *et al.*, 1978; Gasse, 1986; Haberyan, 1987). Because the sharp rise in *Thalassiosira* is not an artifact of the reduction in *Melosira*, the transition to Zone B probably reflects an decrease in the precipitation:evaporation ratio in the hydrology of the lake. Within Zone B, there are smaller-scale fluctuations in the relative abundance of the dominant forms. A decrease in the relative abundance of *Thalassiosira*, along with elevated levels of *Surirella* between approximately 3700 and 3200 yr BP, may represent the influx of easily erodible lacustrine sediments abandoned by a recent lowering of the lake. Another decline in *Thalassiosira* abundance between 2600 to 2300 yr BP along with lowered levels of *Surirella* and an increase in *Stephanodiscus* may suggest an interval of slightly fresher conditions.

Zone A (220 to 0 cm, 1900–1070 yr BP) is dominated by *Surirella engleri* and a *Surirella* species that resemble species 1 described by Gasse (1986) and identified from the littoral mud of modern Lake Turkana. We suspect that this form may be another variety of *S. engleri*. *Stephanodiscus* and *Melosira* are also present throughout this zone. *Thalassiosira* is present throughout this interval and is the most abundant species after about 1250 yr BP. The lower portion of this zone is similar to the diatom assemblage found in the recent sediments of the lake (Gasse *et al.*, 1983). This zone is also characterized by a marked decline in the abundance of diatoms that actually initiated at about 2100 yr BP. This decline may reflect: (1) an increase in the pH of the lake to levels greater than 9 thus enhancing frustule dissolution; (2) a decrease in diatom productivity, possibly replaced by unpreserved algae; or, (3) an increase in the dilution of the plankton by detrital material eroded from the exposed lacustrine sediments. The pores of *Stephanodiscus* reveal evidence for dissolution, e.g., pore enlargement, that is perhaps more evident near the top of the core but is not noticed in the other taxa. The dissolution did not hamper identification of the taxa presented here. *Microcystis* species dominate the lake's plankton today, with other blue-green and green algae (Harbott, 1982). The rise in *Thalassiosira* at the top of the core is to levels typical of Zone B and may suggest a brief rise in lake level. We interpret Zone A as a continuation of low lake levels approaching conditions common at Lake Turkana today with perhaps a brief reversal at about 1250 years.

Other species of diatoms are ubiquitous in low concentrations through Zones A and B, and are rarely observed in Zones C and D. *Rhopalodia gibba, R. gibberula, R. hirundiniformis, Cymbella muelleri, Anomoeoneis sphaerophora,* and *Navicula, Nitzschia* and *Gomphonema* spp. represent the bulk of the 'other' genera. These species, except *Nitzschia* spp., are primarily benthic.

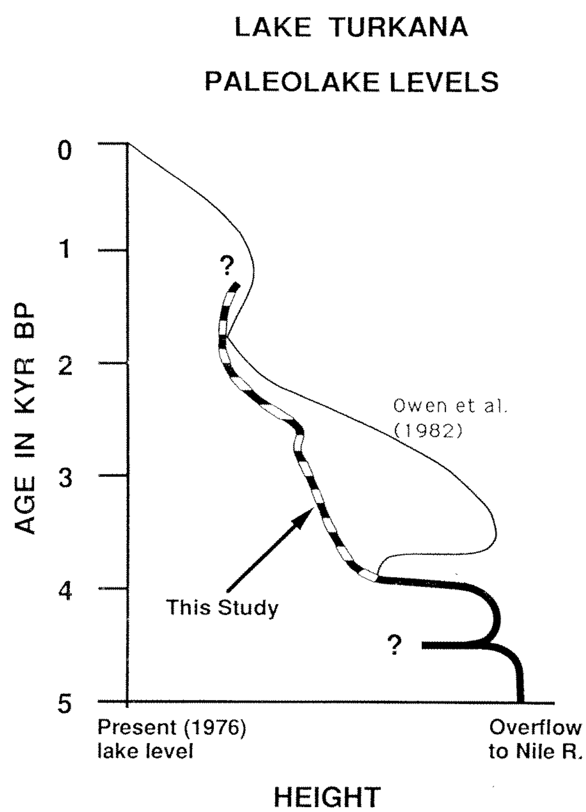

Fig. 4. Inferred lake level curve from this study (thick line) and earlier work (thin line: modified from Owen *et al.*, 1982). The dashed portion represents lake levels that are lower than those prior to 3900 yr BP but the extent of the lowering is unconstrained by the diatom data.

Discussion

The general paleolimnological sequence represented by the diatom assemblages consists of declining lake levels, increasing water conductivity, and decreasing water column stratification from roughly 5450 years ago to the present with the greatest change at 3900 yr BP. Similar patterns in diatom assemblages are observed at another site (Barton's Core in Fig. 1: C. Stager, personal communication, 1991). Climatically, the long-term trend suggests a decrease in the precipitation-evaporation budget for the basin (Fig. 4). An early pluvial phase and later desiccation through the Holocene is consistent with paleoclimatic reconstructions for northern intertropical east Africa and simulated by computer models that are forced by Milankovitch related changes in Northern Hemisphere insolation (Street & Grove, 1976; Hamilton, 1982; Street-Perrott & Harrison, 1983; Kutzbach & Street-Perrott, 1985; COHMAP, 1988; Pachur & Hoelzmann, 1991).

Superimposed on this general trend are some abrupt and high frequency changes. A brief excursion to lower lake levels may have occurred at about 4500 yr BP (Fig. 4). The rapid transition between Zones C and B suggests an abrupt change in lake conditions at about 3900 yr BP. Perhaps the decrease in *Melosira* represents the time when seasonal stratification was permanently broken down. The modern lake is continually mixed by strong diurnal winds and *Melosira* is in low abundances in the plankton (Gasse *et al.*, 1983). We suggest that lake level lowered to initiate continual mixing of the entire water column at about 3900 yr BP. The sudden appearance of *Thalassiosira* indicates more saline and alkaline conditions, and supports this hypothesis. A pause in lowering, or possibly an episode of higher lake levels, though not as high as the pre-3900 yr levels, may be suggested between about 2600 to 2300 yr BP and beginning at 1250 yr BP to at least the top of the record 1070 yr BP.

The magnitude of lake-level change is partially constrained by a published lake level curve for Lake Turkana based on radiocarbon dating of shoreline deposits of known height above the modern lake (Owen *et al.*, 1982). This curve is consistent with our results in that high lake levels (+ 70 to 80 m elevations relative to 1976 lake level) are indicated prior to Zone B, i.e., 3900 yr BP (Fig. 4). It also reveals two lowstand dates (+ 45 and + 55 m elevation) at roughly 4500 yr BP that may correspond to the peak in abundance of *Surirella* within Zone C. Another group of lowstand dates at + 43 to + 47 m cluster at about 4000 yr BP. We correlate this lowering with the diatom transition from Zone C to B.

The interpretation of the two records diverge after this time. The curve suggested by Owen *et al.* (1982) reveals a return to higher lake levels at about 3800 yr BP lasting about 500 years followed by a gradual decline to modern levels. This high phase is poorly constrained by 3 dates on either bone apatite or mixed shell material. These samples are collected from elevations above a + 45 m datum, but are restricted to the Omo Basin (compiled in Street-Perrott *et al.*, 1989). There is no evidence for lake levels above + 50 m after 4000 yr BP in the Galana Boi Beds at Koobi Fora (Owen *et al.*, 1982). The first appearance of *Thalassiosira* in the exposed Holocene deposits is found at + 35 to + 40 and is stratigraphically above the cluster of 4000 yr BP dates (Owen *et al.*, 1982). We suggest that this first appearance may correspond to the beginning of our Zone B. Our diatom data suggest that lake level generally declined after about 3900 yr BP. The resulting exposure of older lacustrine deposits at 3900 yr BP may have contributed to the deviation between the ostracode and fine-grained calcite radiocarbon results. The largest deviations occur stratigraphically above 3900 yr BP (770 cm), at 605 and 455 cm down core. Thus, a combination of our data with a reinterpretation of the exposed deposits suggests that lake levels lowered from a highstand at roughly + 70 to + 80 m to roughly + 43 to + 47 m beginning at 3900 yr BP. The lake did not regain a sustained high-phase following this time.

Other lake-level proxies

Piston cores from the north basin have been analyzed for carbonate abundance (Halfman & Johnson, 1988; Halfman & Hearty, 1990), lamination thickness (Halfman & Johnson, 1988; Halfman & Hearty, 1990), stable isotope variability of the finegrained calcite (Halfman *et al.*, 1989; Johnson *et al.*, 1991), and bulk sediment geochemistry of the aluminosilicate fraction of the sediment (Finney *et al.*, submitted). The long-term decrease in carbonate abundance and increase in lamination thickness from a north basin core, LT84-8P, has been interpreted to be an indicator of Omo River discharge of detrital silicates. A greater influx of Omo material to a core site results in greater dilution of carbonate and thicker laminae, and was interpreted to reflect higher lake levels. The $\delta^{18}O$ profile, which reflects the isotopic composition of the water, has been interpreted in terms of relatively low lake levels for most of the past 4000 yr BP except for brief highstands during the past century, at 2000 yr BP, and possibly at the base of the core. Two results are critical to this discussion: (1) core LT84-8P did not recover sediments older than 4000 yr BP, and thus lacks the pluvial phase represented by Zones C and D in core LT84-2P; and, (2) previous interpretations of carbonate, lamination, and isotope proxies from LT84-8P suggest that lake level at 3900 yr BP was at elevations near or below the present lake level.

The chemical composition of the silicate fraction of the sediments can be partitioned using linear programming. Material derived from the Omo River can be identified due to its enrichment in Al, Fe and other transition elements (Yuretich, 1979; 1986; Finney *et al.*, submitted). Factors that may influence Omo mass accumulation rates include delta migration, lake circulation, and sediment discharge. The mass accumulation rate of the Omo material exponentially decreases with distance from the Omo Delta. This observation, combined with the low gradient of the Omo basin, suggest that distance from the Omo delta to any core site may influence the long term record of these earlier lake-level proxies. For example, lake levels at $+80$ m result in a delta migration of about 100 km farther away from a core site. This migration may result in a reduced detrital load to the site even though Halfman & Johnson (1988) argued that higher lake levels might imply higher precipitation, greater runoff and higher sediment discharge. Vegetation changes in the drainage basin may provide additional complications. For the purposes of this study, the carbonate abundance, lamination thickness and oxygen isotope proxies may indeed be influenced by varying lake levels, but they may also reflect the relative position of the Omo delta, especially in the north basin. For example, low lake levels move the Omo delta closer to a core site and may result in greater Omo flux, greater carbonate dilution, thicker laminae and depleted $\delta^{18}O$ values that are more directly influenced by the oxygen isotopic composition of the Omo River than the entire lake. With this in mind, we suggest that the north basin data do not indicate lowered lake levels to or slightly below the present level but lake levels at an intermediate depth around $+45$ m (Fig. 4).

Whereas the north basin cores should experience a significant impact, the south basin cores may experience a more limited change in Omo flux due to delta migration, and thus should be more influenced by constricted circulation across the bathymetric high between the north and south basins, especially at low lake levels. The highest flux of Omo material at core LT84-2P, in the south basin, occurred during Zones C and D, and peaked during Zone C (Fig. 5). The percentage of Omo material shown in Fig. 5 probably parallels the flux as sedimentation rates and bulk densities are relatively constant down core. The occurrence of Omo material 200 km from the present delta suggests a vigorous, lake-wide circulation to distribute the sediment to the south basin of the lake. This interpretation is consistent with the elevated abundance of *Melosira*. The decrease in Omo material roughly 5050 to 4850 BP in Zone D parallels a decline in *Melosira* and rise in *Stephanodiscus* suggesting a period of decreased lake circulation. An abrupt decrease in Omo material at about 4100 yr BP suggests a rapid decrease in circulation or Omo discharge which may be related to lake level. However, complete exposure of this sill is not necessary to restrict circulation and isolate the south basin from the Omo sediments. The timing is approximately 200 years before the corresponding decrease in *Melosira*, which may represent the time required to break down seasonal stratification or an increased constriction between the north and south basins that would reduce the influx of Omo material to the south basin as lake level lowers. The flux of Omo material generally decreases from 4100 to 3500 yr BP and is found in low but varying abundances since that time.

Oxygen isotope ratios of ostracode carapaces reveal increasing values from 2.5‰ (PDB) near the base of the core to 3.8‰ at the top of the core, and are interpreted to reflect changes in the isotopic composition of the water (Fig. 5: Halfman, 1987; Halfman *et al.*, 1989; Johnson *et al.*, 1991). The long-term trend indicates lowering lake levels from the bottom to the top of the core (5450 to 1070 yr BP). The coarse sampling precludes a detailed comparison to the other data presented here and to published oxygen isotope profiles from the north basin.

A profile of carbonate abundance for core LT84-2P reveals increasing values to the top of the core with a sharp increase between 4600 and 4350 yr BP (Fig. 5: Halfman & Hearty, 1990). The lack of correlation between the carbonate and Omo flux profiles in this core indicates that carbonate abundance in the south basin is not strongly controlled by Omo River discharge as suggested by Halfman & Hearty (1990). A possible scenario to explain this sharp increase at roughly 4600 yr BP is that lake level first lowered just enough to close the basin. The timing of the initial closure is speculative at this time, but may correspond to the peak in appearance of *Surirella* at about 4500 yr BP. This may cause an increase in carbonate precipitation, as incoming dissolved Ca would have then been trapped within the basin. However, the abrupt increase in carbonate abundance precedes the decrease in Omo material and *Melosira*. The decrease in Omo flux may have lagged this event if continued lowering of lake level was required to restrict circulation. As lake levels continued to drop, increased salinity,

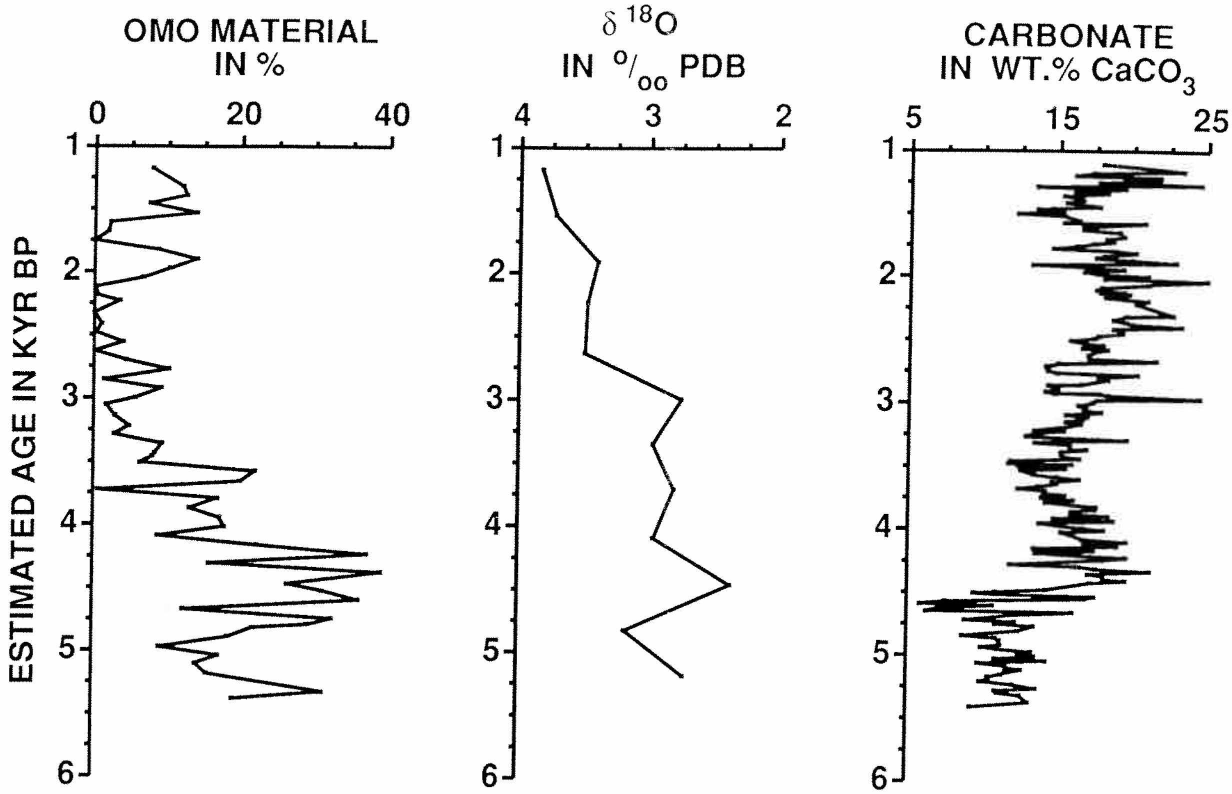

Fig. 5. Additional paleolimnological proxies versus age down core LT84-2P. Left: Abundance of Omo derived material (Finney *et al.*, submitted). Center: $\delta^{18}O$ results (‰ PDB) of hand-picked *Sclerocypris clavata* ostracode carapaces (Halfman *et al.*, 1989). Right: Bulk carbonate abundance (weight % $CaCO_3$ of the dry sediment: Halfman & Hearty, 1990). Note that the age scale starts at 1000 yr BP.

and a break-down in water-column stratification may have resulted in the decline in *Melosira* and increase in *Thalassiosira*. Variations in the carbonate profile since 4300 yr BP may reflect dilution by detrital silicates or changes in calcite precipitation rates.

In conclusion, diatom assemblages from a south basin core recovered from Lake Turkana reveal a general trend of decreasing lake levels from 5450 yr BP to the present with the most rapid lowering at about 3900 yr BP. Lake level is fixed at 70 to 80 m above the present lake prior to this time and is within 45 m of the present lake level after 3900 yr BP. The long term trend is consistent with proxies of lake level variability from this core, other cores within the basin, and other lakes in Africa. A previously proposed late Holocene highstand (+ 70 m at roughly 3500 yr BP) based on radiocarbon dating of exposed shoreline deposits is not supported by these data.

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