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Sedimentation controls on the preservation and time resolution of climate-proxy records from shallow fluctuating lakes

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Abstract

Lithological stratigraphies of ^{210}Pb-dated sediment cores from four hydrologically interconnected fluctuating lake basins in Kenya are used to investigate how differences in basin morphometry and physical limnology influence the preservation and time resolution of sedimentary climate-proxy records. The potential of lakes to accumulate an undisturbed sediment record is primarily determined by their relative depth, the ratio between maximum depth and effective wind fetch. Chemically stratified crater lakes accumulate high-quality climate-proxy records because they complement great relative depth with topographical wind shelter, resistance of density stratification to propagation of wind-induced turbulence, and absence of bioturbation in anoxic bottom waters. The changes in water-column circulation and bottom dynamics that accompany lake-level fluctuations affect the time resolution of accumulating climate-proxy signals and thus the apparent magnitude and frequency of inferred climatic events. An episode of low lake level can be both over- or underrepresented in the climate-proxy record depending on the severity of the drawdown relative to the lake's critical depth of sediment accumulation. Decade-scale hiatuses due to non-deposition or erosion at low lake level may be difficult to recognize because mixing of unconsolidated muds deposited before and after the lowstand can obliterate evidence that the record has been truncated. Regional correlation of climate-proxy records must consider both hydrology-related differences in the climatic sensitivity of lakes and sedimentation-related differences in the integrity of their climate-history archive.

1. Introduction

In arid and semi-arid regions worldwide, sediment records from shallow closed-basin lakes are the most prominent source of information on past climate change (Fritz 1996). Closed-basin lakes respond to variations in the regional balance of precipitation and evaporation with changes in lake level and water chemistry that are expressed as characteristic lithological, geochemical, and biological signatures in the sediment record. The selection of a lake for recovery of high-resolution paleoclimate records is complicated by the basic incompatibility between climatic sensitivity and longevity. Lakes that display appropriate hydrological sensitivity to decade-scale climatic change are also most likely to completely dry out during episodes of extreme aridity, at which time deflation and other surface processes may remove or destroy part of the accumulated climate-proxy archive. Conversely, lakes that have demonstrably persisted through episodes of extreme aridity are often located in hydrologic settings that preclude sensitivity to the (relatively) modest climatic variation occurring on historically relevant time scales. Consequently, study sites for paleo-climate research inevitably must strike a compromise between the time frame of the planned climate reconstruction and its desired time resolution.

Comparative paleolimnology of climate-sensitive lake basins has so far mostly focused on the effects of hydrology on the mode and timing of lake response to climate change (e.g., Digerfeldt *et al*., 1992; Harrison and Digerfeldt, 1993; Locke, 1995). But lakes with comparable hydrological sensitivity can differ considerably in the time resolution of their sediment records, because basin morphometry and the regime of water-column circulation influence areal patterns of sedimentation and post-depositional sediment mixing. Large shallow lakes are generally ill-suited for high-resolution paleoclimate study because wind-driven sediment erosion and redistribution is likely to have corrupted the sedimentary archive of past climate change. In contrast, lakes in small volcanic

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crater basins are generally considered choice sites for recovery of high-resolution paleoclimate records (Gasse, 1995; Colman, 1996; Creer and Thouveny, 1996) not only because their simple hydrogeological setting simplifies the relationship between lake history and climate history but also because their great relative depth and additional topographic wind shelter promote undisturbed sedimentation.

Larsen and MacDonald (1993) used physical models and empirical data to describe general relationships between basin morphometry and the quality of offshore depositional environments in small north-temperate lakes as a guide to the selection of study sites for paleoecological studies. This study examines how the changes in basin morphometry and mixing regime over time that accompany the lake-level fluctuations of climate-sensitive lakes affect the integrity and time resolution of their sedimentary climate-proxy records. It does so by comparing lithological stratigraphies of ^{210}Pb-dated sediment cores comprising the last 120 years of sedimentation in four morphometrically and limnologically distinct lake basins in Kenya that share an identical history of documented lake-level change. The hydrological connectedness of these basins is critical, because partial or complete independence of lake-groundwater interactions would have resulted in a different timing and magnitude of each lake's response to climate change and thus created different sediment records *a priori*. Specifically, this study addresses the following questions: (1) Which aspects of basin morphometry and water-column circulation determine the potential of fluctuating lakes to preserve an undisturbed record of past lake-level change? (2) What is the relative importance of bioturbation and wind-driven turbulence on the long-term integrity of these climate-proxy records? and (3) How do changes in bottom dynamics over time affect the time resolution of the sediment record as an archive of climate history?

2. Methods

2.1. Hydrology and limnology of the study lakes

Lake Naivasha and its three satellite basins Lake Oloidien, Crescent Island Crater, and Lake Sonachi are located at about 1885 m a.s.l. in the central valley of the Eastern Rift in Kenya (Fig. 1). Highlands flanking the Rift Valley intercept most of the monsoonal rainfall in the region, resulting in a strong moisture deficit near the lakes with annual rainfall and evaporation averaging 608 and 1865 mm (Åse *et al.*, 1986). Southwest-to-southeast winds channeled through the Rift Valley are strongest in August and October, occasionally exceeding 21 knots (10.8 m s^{-1}). The main basin of Lake Naivasha is a fairly large (about 135 km^2 in 1993), shallow (6 m), and

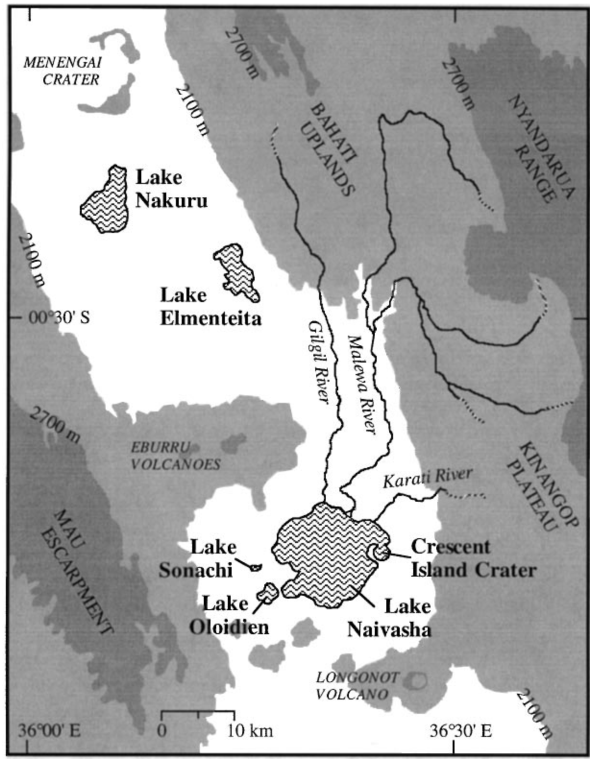

Fig. 1. The Rift valley of central Kenya, with the four studied lake basins and nearby lakes Nakuru and Elmenteita.

wind-stressed freshwater lake maintained by river input primarily from the Malewa River, which drains the wet mountain ranges that form the eastern flank of the Rift Valley. Its drainage basin is topographically closed but hydrologically open, with groundwater flowing towards the lake from the north and exiting in the south and southeast (Thompson and Dodson, 1963; Gaudet and Melack, 1981). The elevation of Lake Naivasha fluctuates in response to short-term rainfall variability (Vincent *et al.*, 1979) essentially as in a hydrologically closed system (Fig. 2), because its water budget is dominated by river input and evaporation from the lake surface (Gaudet and Melack, 1981; Darling *et al.*, 1990). Over the past 110 yr (1883–1993) lake depth has ranged between 4 and 19 m and open-water surface area between <100 and >200 km^2, but lakewater conductivity (K_{25}) has remained fairly constant at 233–499 µS cm^{-1} (Verschuren, 1996).

Lake Oloidien (5.1 km^2) is similarly shallow but hydrologically closed; water loss from this basin is entirely due to evaporation (Gaudet and Melack, 1981). When lake surface elevation stands above 1885.5 m a.s.l., Lake Oloidien is confluent with Lake Naivasha and fresh. Below this elevation Lake Oloidien separates from Lake Naivasha and becomes saline due to evaporative concentration, while being maintained at the level of Lake Naivasha by seepage through the permeable sill between them. The depth of Lake Oloidien has historically ranged

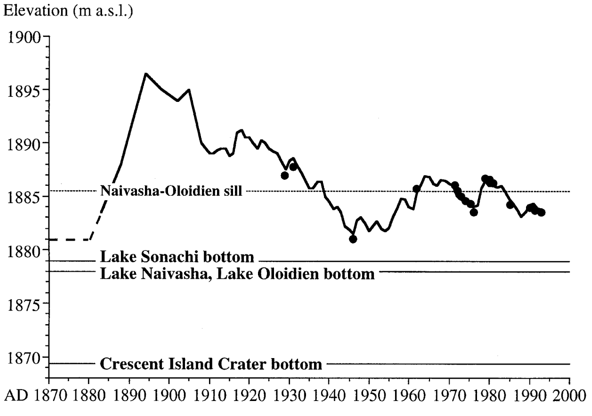

Fig. 2. Historical lake-level record of Lake Naivasha (heavy line) and depth soundings in Lake Sonachi (black dots) in relation to lake-bottom elevation in the four studied basins and the sill elevation between Lake Naivasha and Lake Oloidien. Modified from Verschuren (1996).

between 4 and 19 m and its surface area between 4.0 and 7.5 km^2. Available conductivity data range from 472 to 1040 $\mu S\,cm^{-1}$ (Verschuren, 1996), but evidence from fossil diatoms and chironomids suggests values of 4000–6500 $\mu S\,cm^{-1}$ near the end of the 1940s–1950s lowstand and as high as 12,000–14,000 $\mu S\,cm^{-1}$ before 1890 (Verschuren *et al.*, in press).

Crescent Island Crater is a small and relatively deep crater basin partially submerged in the Naivasha basin, and hydrologically open due to confluence with Lake Naivasha and groundwater throughflow. The depth of Crescent Island Crater has historically ranged between 13 and 28 m, but its surface area has changed little (1.7–2.2 km^2), and it remained fresh with a measured conductivity range of 291–491 $\mu S\,cm^{-1}$ (Verschuren, 1996).

Lake Naivasha, Lake Oloidien, and Crescent Island Crater are continuously warm polymictic (Lewis, 1983), a mixing regime in which a thermal stratification of 3–4°C develops during calm morning hours and is then destroyed by strong afternoon winds and nighttime convective circulation. At the current lake level, Lake Naivasha and Lake Oloidien probably circulate to the bottom almost every night. In Crescent Island Crater, convective circulation extends down to about 10–12 m (Brierley *et al.*, 1987) and reaches the bottom frequently enough to keep it oxygenated (Melack, 1979).

Lake Sonachi is a small (0.14 km^2) alkaline-saline lake located in a volcanic crater basin 3 km west of Lake Naivasha (Fig. 1). It is sheltered from wind by a crater rim towering 50–115 m above the lake surface, and chemically stratified when lake depth exceeds 5 m (MacIntyre and Melack, 1982; Njuguna, 1988). The mixolimnion undergoes a daily cycle of stratification and circulation similar to the other three lakes, but density stratification prevents convective circulation below the chemocline. Lake Sonachi is maintained by groundwater flow from Lake Naivasha (MacIntyre and Melack, 1982; Darling *et al.*, 1990) through the porous volcanic deposits that underlie much of the central Rift Valley (Thompson and Dodson, 1963). Strong correlation between 17 depth soundings in Lake Sonachi over the past 60 years and the continuous lake-level record of Lake Naivasha suggests that the latter can be fully extrapolated to Lake Sonachi (Fig. 2). The solute budget of Lake Sonachi is governed by evaporative concentration at the water surface and dissolution of sedimentary evaporites by groundwater inflow (MacIntyre and Melack, 1982); surface-water conductivity has historically ranged between 3000 and 11,550 $\mu S\,cm^{-1}$ (Verschuren, 1996).

2.2. Field and laboratory techniques

Nine short sediment cores (34–127 cm) with undisturbed mud-water interface were collected with a rod-operated single-drive piston corer (Wright, 1980) at offshore stations in all four lakes (Fig. 3). Selection of core sites aimed to balance maximum distance between sites with avoidance of peripheral areas where sediment accumulation may not have been continuous during recent lowstands. The cores were extruded in the field in 1-cm increments with a fixed-interval sectioning device (Verschuren, 1993), and transferred to Whirl-Pak™ bags for transport. One long sediment core (NC93.2-L: 824 cm) was recovered in 1-m segments with a square-rod piston corer (Wright, 1967) near the deepest point of Crescent Island Crater, from the exact locality of short core NC93.1-S (Fig. 3). Core segments were extruded in the field, wrapped in plastic and aluminium foil, and shipped in PVC tubes. One core from Lake Sonachi (NS93.2-F: 37 cm) was collected by in situ freezing onto a wedge-shaped aluminium corer filled with dry ice and ethanol (Renberg, 1981). This core was returned to Minneapolis intact and cleaned and processed in a cold-room. The frozen sediment profile was divided in 50 increments of variable thickness (3–17 mm) along visible boundaries between stratigraphic horizons (Verschuren, in review). The freeze-coring operation caused slight resuspension of the flocculent sediment-water interface, consequently the top of the sediment profile (0.0 cm core depth) was set at the top of the uppermost undisturbed horizon.

The sedimentary proxy indicators for lake level examined in this study are all measures of bulk-sediment composition commonly used in core logging and core correlation (Dearing, 1986; Colman, 1996): lithology, porosity, organic-matter (OM) and carbonate content, and sediment texture. Water content (% H_2O by weight), porosity (% H_2O by volume), and sediment composition were determined by drying overnight at 105°C, burning at 550°C, and ashing at 1000°C (Bengtsson and Enell, 1986). Total carbon (TC) and inorganic carbon (TIC) of selected samples were measured on a IUC carbon

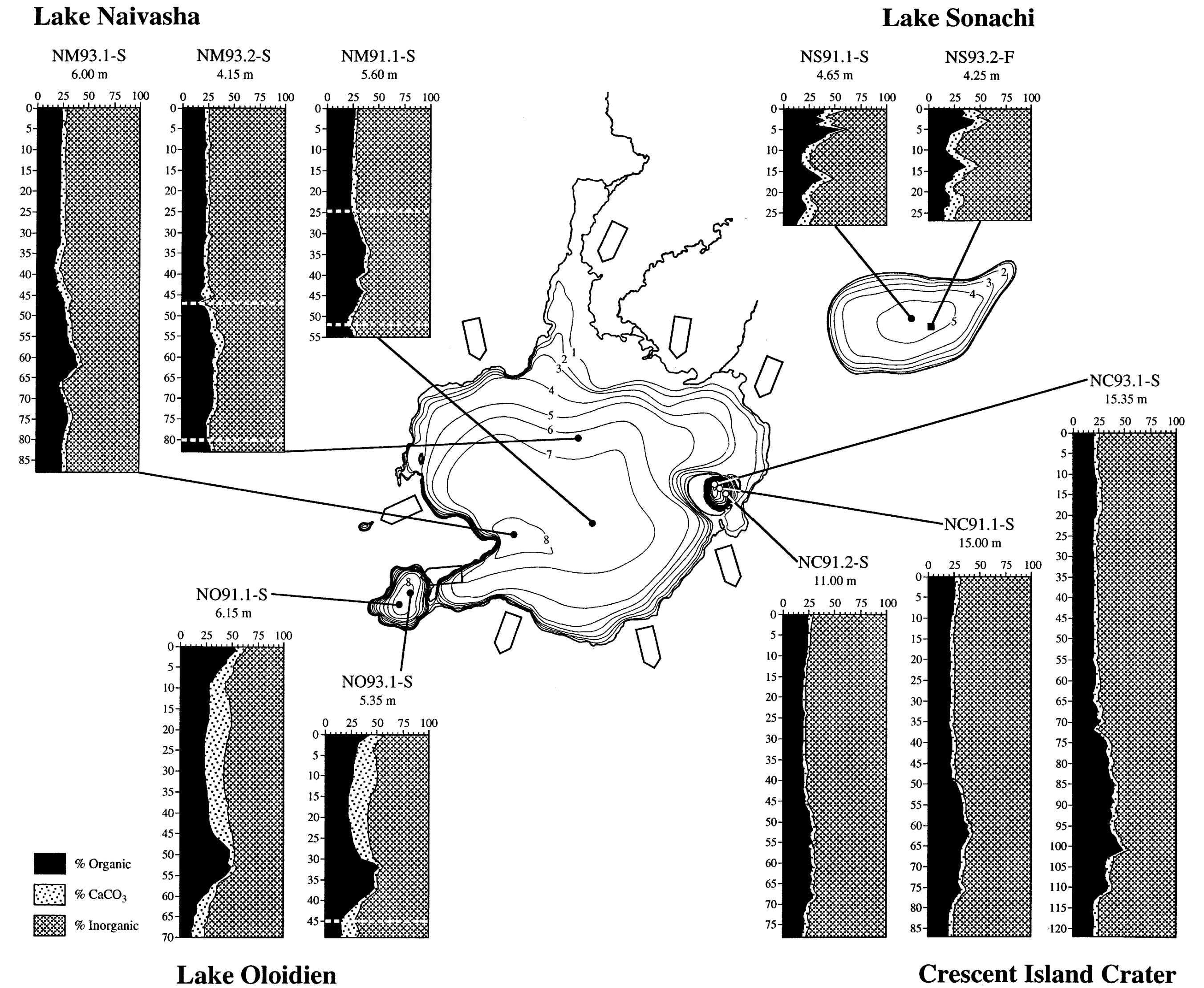

Fig. 3. Sediment composition of cores representing the past 120 yr of sedimentation in (clockwise from upper left) Lake Naivasha, Lake Sonachi, Crescent Island Crater, and Lake Oloidien. Dashed lines across some of the core profiles show positions of sedimentary unconformities, cf. text. Bathymetry of Lake Naivasha (in 1983, at 1886 m a.s.l.) and Lake Sonachi (in 1990, at 1884 m a.s.l.) from Åse *et al.* (1986) and Damnati *et al.* (1991). White arrows indicate groundwater flow.

analyzer; total organic carbon (TOC) was obtained by subtracting TIC from TC. The fractions (% dry weight) of coarse OM derived from aquatic macrophytes (and terrestrial vegetation) and sand-sized mineral particles were quantified by dispersing the sediment matrix in 0.25% Calgon™ followed by rinsing through a 250 μm sieve. The retained residue was transferred to pre-weighed ashless filter paper and analysed by burning and ashing, as above (Digerfeldt, 1986).

Sediment chronology and accumulation rates of selected cores were determined by measuring ^{210}Pb-activity through its granddaughter product ^{210}Po, with ^{208}Po added as an internal yield tracer. Sample preparation followed Eakins and Morrison (1978), and activity was measured in an alpha-spectroscopy system. Supported ^{210}Pb-activity was estimated as the asymptote of total activity at depth, and unsupported activity was determined by subtracting average supported activity from total activity measured at each level. Dates and sedimentation rates were determined with the constant rate of supply (c.r.s.) model (Appleby and Oldfield, 1978; Binford, 1990). The stratigraphic position of lowermost occurrence of unsupported ^{210}Pb in core NS93.2-F from Lake Sonachi was confirmed by additional gamma-spectroscopic analyses of supported ^{210}Pb (Verschuren, in review). Stratigraphic marker horizons near the sediment surface of NS93.2-F are situated 2.5 cm higher than in NS91.1-S (Fig. 3) due to assignment of the 0.0 cm level in NS93.2-F to the top of the uppermost undisturbed horizon. Consistent depth–age relationships were obtained by setting the collecting date of NS93.2-F to 1988, which is the age at 2.5 cm depth in NS91.1-S. The inventory of unsupported ^{210}Pb in core NM93.1-S from Lake Naivasha may be incomplete due to the probable occurrence of sedimentary discontinuities in the recent past; hence, its sediment chronology must be considered less accurate than that of Lake Sonachi or Lake Oloidien. ^{210}Pb-activity data obtained for core NC93.1-S from Crescent Island Crater failed to produce a satisfactory sediment chronology. Strong variation in sediment focusing into the crater depending on its degree of confluence with Lake Naivasha may have violated the assumption of constant ^{210}Pb flux required by the c.r.s. dating model. Scaling of the ^{210}Pb-activity data to simulate the expected increase in ^{210}Pb flux during maximum confluence with Lake Naivasha improved the inferred chronology, but no simple algorithm yielded results consistent with those of the other basins. The chronology for Crescent Island Crater presented here is based on stratigraphic correlation of three marker horizons dated to about 1890 (bottom of Unit II; cf. below), 1940 (top of Unit II), and 1958 (a peak in fossil *Daphnia* ephippia; D. Verschuren, unpublished data) that NC93.1-S has in common with core NM93.1-S from Lake Naivasha.

Five AMS radiocarbon dates (Table 1) were obtained on wood, seeds, or charcoal recovered from near the bottom of selected cores. All five radiocarbon ages correspond to two or more calendar-age windows (Stuiver and Pearson, 1993). The age determinations are reported as uncalibrated ^{14}C yr BP since they are used only to help identify pre-20th-century events of discontinuous sedimentation.

2.3. Basin morphometry, wave theory, and the sedimentary environment

From the viewpoint of a sediment profile being the accumulated archive of lake and climate history, bottom dynamics and sedimentary environments can best be described in terms of the processes of sediment erosion, transport, and accumulation (Dearing, 1997). In the zone of erosion, wind-driven turbulence propagated to the bottom prevents deposition of fine-grained sediments. In the zone of transport, accumulation is discontinuous and frequently or periodically interrupted by episodes of non-deposition or of resuspension and transport of previously deposited sediments. The zone of accumulation is unaffected by wind-driven turbulence, allowing continuous deposition of fine-grained sediments (Håkanson, 1977). Various theoretical and empirical models have been developed to predict from basin morphometry, wave theory and wind-speed data the occurrence and areal distribution of these three zones in individual lakes (Norrman, 1964; U.S. Army Coastal Engineering Research Center, 1977; Håkanson, 1977, 1982; Johnson, 1980; Rowan *et al*., 1992; Blais and Kalff, 1995), and to determine the dominant physical mechanisms by which sediments are redistributed after initial deposition (Hilton, 1985; Hilton *et al*., 1986). Important concepts in these analyses are the effective fetch (L_f) of the open water surface over which wind can generate waves, and the critical water depth (Z_{a-t}) or wave base (WB) separating the zones of accumulation and transport. Table 2 presents morphometric data of the four study lakes, and corresponding values of the critical depth in each basin calculated or estimated using the models of Håkanson (1977, 1982) and Johnson (1980). The depth limit between zones of erosion and transport (Z_{t-e}) is rather diffuse but typically about 40–45% of Z_{a-t} over the range of L_f values relevant to this study (Håkanson and Jansson, 1983). The variation in effective wind fetch associated with historical lake-level change has not significantly affected the absolute value of Z_{a-t}; the areal extent of the zone of continuous sediment accumulation within each basin is mainly affected by changes in lake depth, Z_m.

3. Results and discussion: Contrasts between basins

3.1. Spatial homogeneity of offshore sedimentation

Sediments deposited over the past 120 yr of documented lake history (ca 1871–1993) range in thickness from 27 cm in Lake Sonachi to 122 cm at the deepest point of Crescent Island Crater (Fig. 3). Strong stratigraphical similarities within sets of cores collected in Lake Sonachi, Lake Oloidien, and Crescent Island Crater indicate that the offshore depositional environment in these basins is spatially homogeneous, so that any particular core selected for detailed analysis can be

Table 1
AMS radiocarbon dates obtained on pre-20th-century sediments from the four study lakes in the Eastern Rift in Kenya

Basin	Core	Depth (cm)	Material	^{14}C Age (BP)	Lab No.
Sonachi	NS93.2-F	33.2-39.2	Grass charcoal	520 ± 60	CAMS 31604
Naivasha	NM93.1-S	100-102	Grass charcoal	290 ± 50	CAMS 22003
Oloidien	NO91.1-S	82-83	Wood	120 ± 50	CAMS 22002
Oloidien	NO91.1-S	89-91	*Potamogeton* seed	120 ± 60	CAMS 31605
Crescent Island Crater	NC93.2-L	419-420	Wood	580 ± 40	CAMS 18702

Table 2
Morphometric data for the four studied lake basins, based on the bathymetry shown in Fig. 1

Basin			Sonachi	Naivasha	Oloidien	Crescent I Crater
Level			1884 m	1886 m	1886 m	1886 m
Year			1990	1983	1983	1983
Lake depth	Z_m	m	5	8	8	17
Average depth	Z	m	3.7	4.8	6.1	9.0 (5.5)[a]
Area	A	km^2	0.14	171	5.74	1.95 (3.53)
Volume	V	km^3	0.00053	0.824	0.035	0.018 (0.019)
Volume development	D_V	—	2.10	1.81	2.04	1.68 (1.02)
Basin-averaged slope[b]	S	%	3.1	0.16	1.08	2.18 (1.64)
Range of slope with depth[b]	S	%	0.87–9.05	0.04–0.43	0.35–3.43	0.91–3.65
Effective fetch at coring site	L_f	km	0.14	4.2	1.2	1.2 (1.3)
Area of accumulation[c]	A_a	%	—[d]	26	83	83 (–)[e]
Area of erosion + transport[c]	A_{e+t}	%	—[d]	74	17	17 (–)[e]
Site-specific critical depth[f]	Z_{a-t}	m	—[d]	7.5	2.5	2.6
Wave base[g]	WB	m	< 1.0	6.0	3.3	3.4

[a] Values in parentheses include shelf areas of Lake Naivasha sloping towards Crescent Island Crater.
[b] Averaged along perimeter of depth contours, following Håkanson (1981).
[c] Energy-Topography formula of Håkanson (1982).
[d] Formula does not apply to lakes as small as Lake Sonachi (Håkanson and Jansson, 1983).
[e] Formula does not properly take into account the convex topography of shelf areas.
[f] Erosion–transport–accumulation equation of Håkanson (1977).
[g] Graphically derived from Johnson (1980) for winds of 25 knots (12.8 m s^{-1}), i.e. comparable to the strongest winds at Lake Naivasha.

considered representative for the basin. In Lake Sonachi, time-normalized loss-on-ignition data of cores NS91.1-S and NS93.2-F from stations located about 100 m apart are highly correlated ($R^2 = 0.712$; Verschuren, in review). In Lake Oloidien, cores from stations 800 m apart also display a strong similarity (Fig. 3), except for evidence of sediment focusing (Davis and Ford, 1982) in the form of greater linear sediment accumulation at the site of NO91.1-S than at the slightly shallower site of NO93.1-S. In both Lake Sonachi and Lake Oloidien, spatial homogeneity of offshore sedimentation is promoted by a strongly ellipsoid bathymetry (Lehman, 1975) with steep bottom gradients (slope, S) nearshore and an essentially flat lake floor (volume development (D_V) = 2.10 and 2.04; Table 2). Crescent Island Crater is a paraboloid depression with steep inner crater slopes and a flat profundal lake bottom ($D_V = 1.68$). Averaged along the perimeter of each depth contour (Fig. 3), bottom gradients remain below the 4% limit above which gravity-driven sliding and slumping may start to affect sedimentation (Håkanson, 1977), but gradients as high as 11% occur locally along the steep-walled western side of the crater. Temporal patterns of sedimentation over the past 120 yr appear to have been uniform across the profundal lake bottom, except for a 40–50% higher rate of infilling at the deepest point of the basin than in more peripheral areas (Fig. 3).

The observed homogeneity of offshore sedimentation in Lake Oloidien and Crescent Island Crater is consistent with the limited influence of wind-driven turbulence on bottom dynamics of small lakes predicted by wave theory (Håkanson and Jansson, 1983). In both basins the critical water depth (Z_{a-t}: 2.5 and 2.6 m) separating the zone of accumulation from the zones of erosion and transport is significantly less than the average depth (Z: 6.1 and 9.0 m), at least at the present lake level. Consequently, most of the lake bottom remains unaffected by wave turbulence ($A_a > 75\%$). Lake Sonachi is too small for calculation of Z_{a-t} from the empirical equations of Håkanson (1977, 1982), but the relationship between wind fetch and wave base (Johnson, 1980) indicates that resuspension of fine-grained sediments in Lake Sonachi should be limited to water depths of less than 1 m (Table 2) even without the benefit of wind shelter.

Favorable bottom dynamics in the three smaller basins contrast strongly with the situation in Lake Naivasha. In this large and shallow basin, cores from sites about 4 km apart show substantial differences in sediment stratigraphy and accumulation (Fig. 3). Displacement of depth contours towards the middle of the basin reveal the strong influence on sedimentation by the Malewa River, the dominant source of allochthonous sediment input to Lake Naivasha. The calculated critical water depth (Z_{a-t}: 7.5 m) exceeds the average water depth of the basin (Z: 4.8 m), implying that bottom dynamics of Lake Naivasha are dominated by wind-driven resuspension and random redistribution of previously deposited sediments (Hilton *et al.*, 1986). The small area of lake floor below Z_{a-t} at the 1983 lake level (A_a: 26%) indicates that even only slightly lower levels, such as occurred in the mid-1970s, the late 1980s, and today, may bring about wind-driven sediment disturbance throughout the basin.

3.2. Lithostratigraphy and chronostratigraphy

Recent sediments in all four study lakes consist of unconsolidated organic muds. Sediment porosity exceeds 95% in the upper portions of all cores and is as high as 98–99% near the sediment surface in Lake Sonachi, Lake Oloidien, and Crescent Island Crater (Figs. 4 and 6–8). In Lake Naivasha and Crescent Island Crater, the two basins that recharge shallow aquifers, porosity is high throughout the cores, with little evidence of sediment compaction downcore. In Lake Sonachi and Lake Oloidien, the two hydrologically closed basins, porosity decreases to 73–82% in stiff low-organic clays at the bottom of the cores.

The carbonate content of recent sediments similarly separates the hydrologically open from the hydrologically closed basins. Measurements of TIC in selected samples show that the carbonate content of Lake Naivasha and Crescent Island Crater sediments is insignificant ($<0.1\%$); the values of percent carbonate obtained as weight loss between 550°C and 1000°C (averaging 4.9 and 4.6%; Figs. 6 and 8) overestimate the true carbonate content due to release of clay-bound water (Bengtsson and Enell, 1986). The TIC content of recent sediments in Lake Sonachi and Lake Oloidien is variable but generally high, with a range of 1.6–3.1% in Lake Sonachi and 0.5–4.5% in Lake Oloidien; the corresponding ranges of carbonate content determined by loss-on-ignition are 7–18% and 4–23% (Figs. 5 and 7).

The stratigraphy of OM content defines four major stratigraphic units common to all four lake basins (Figs. 4 and 6–8). The boundaries of these units reflect changes in sedimentation regime at the core sites that can be linked to the major trends of historical lake-level change. Unit I was deposited at the low to intermediate lake levels prevailing during much of the 19th century, including the 1870s–1880s when early explorers' reports imply that Lake Naivasha stood lower than at any time during the 20th century (Åse *et al.*, 1986). The OM content of Unit I sediments is low (12–20%) in Lake Sonachi and Lake Oloidien and intermediate (20–25%) in Lake Naivasha and Crescent Island Crater. Unit II corresponds to the long period of high lake level between about 1890 and 1940, defined here as the period when water stood above the sill between Lake Naivasha and Lake Oloidien (Fig. 2). The Unit I–Unit II transition is characterised by a significant increase of OM content to a subsurface maximum of 30–50% that is common to all basins but with considerable variability in timing and magnitude. In the upper part of Unit II, OM content decreases again to reach intermediate values of 20–30% at the Unit II–III boundary. Unit III was deposited at the low and intermediate lake levels prevailing since 1940 (Fig. 2). In Lake Naivasha and Crescent Island Crater, OM content remains fairly constant throughout Unit III up to the sediment surface, whereas cores from Lake Sonachi and Lake Oloidien display a strong increase near the top of Unit III to values exceeding those in Unit II.

The section of stiff, low-organic silty clays at the bottom of cores from Lake Sonachi and Lake Oloidien is labeled Unit 0 and interpreted as a desiccation horizon formed by oxidation and compaction of lake sediments during a prolonged period of subaerial exposure when these basins stood dry. In Lake Sonachi, renewed onset of subaquatic sedimentation at the bottom of Unit I is

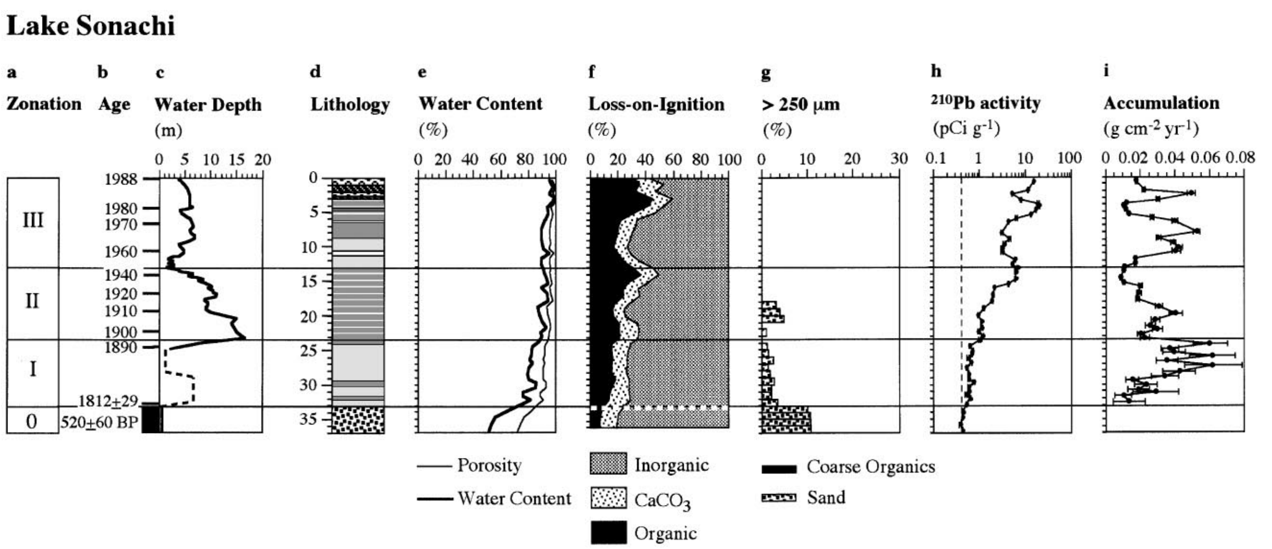

Fig. 4. Lake Sonachi freeze-core NS93.2-F: (a) stratigraphic zonation, (b) sediment age, (c) water depth at ^{210}Pb-inferred time of deposition, (d) lithology, (e) water content and porosity, (f) bulk-sediment composition, (g) percent abundance of sand-sized (>250 μm) organic and mineral sediment components, (h) ^{210}Pb-activity, and (i) rate of sediment accumulation versus core depth. See Figs. 5 and 8 for a key to the lithology.

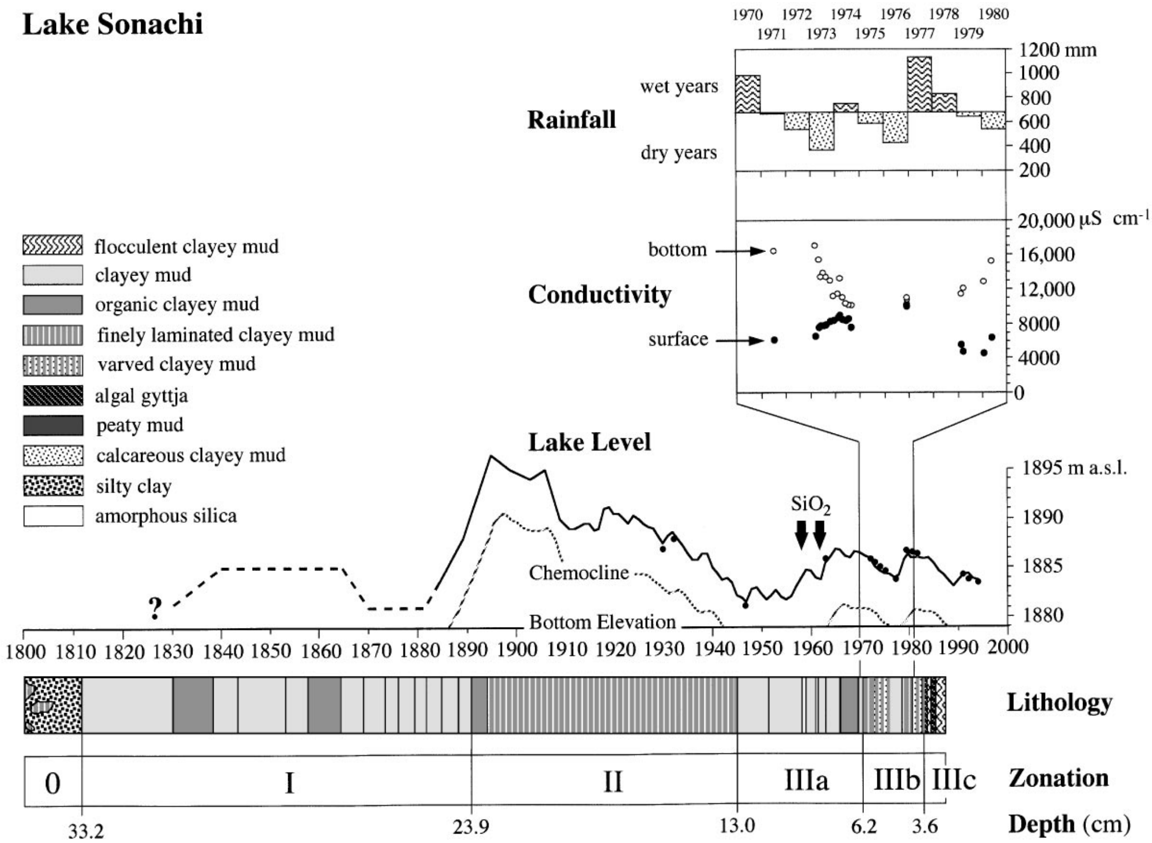

Fig. 5. Lake Sonachi freeze-core NS93.2-F: sediment chronology, lithology, and stratigraphic zonation in relation to historical fluctuations in water depth (full line), chemocline depth (stippled line), annual rainfall, and conductivity of surface- and bottom waters. Core depth is adjusted to fit a linear time axis. Modified after Verschuren (in review), with rainfall and conductivity data from MacIntyre and Melack (1982) and Njuguna (1988).

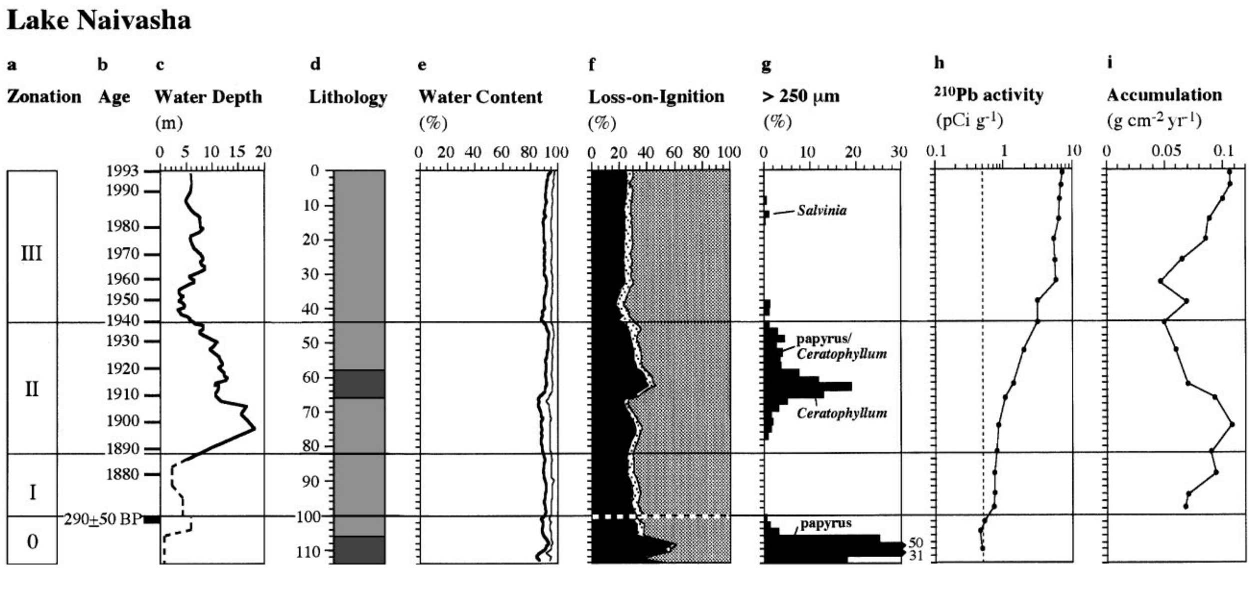

Fig. 6. Lake Naivasha piston-core NM93.1-S: (a–i) as in Fig. 4.

dated to AD 1812 $\pm$ 29 yr (Verschuren, in review; Fig. 4). Refractive terrestrial OM recovered from Unit 0 is dated to 520 $\pm$ 60^{14}C yr BP, indicating that deflation of lacustrine deposits from before the early 19th-century drystand caused a significant temporal hiatus in the sediment record. Radiocarbon dates of 120 $\pm$ 60 yr BP and 120 $\pm$ 50 yr BP for both the top of Unit 0 and the base of Unit I in Lake Oloidien (Fig. 7) are also

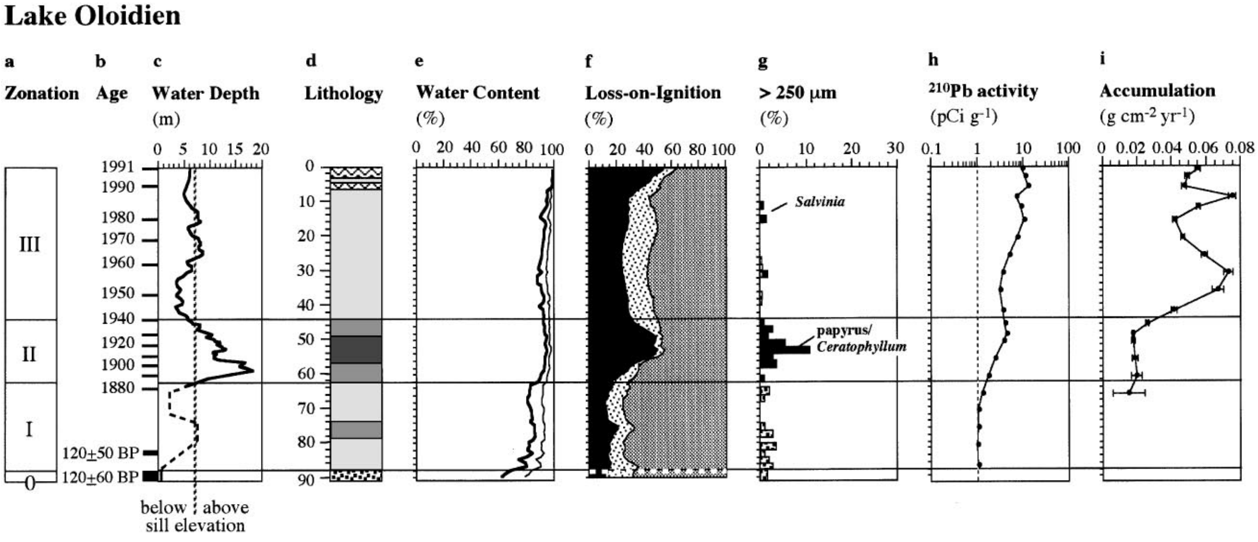

Fig. 7. Lake Oloidien piston-core NO91.1-S: (a–i) as in Fig. 4.

compatible with an early 19th-century start of renewed lake filling. The presence of desiccation horizons underlying 19th- and 20th-century lacustrine deposits in these hydrologically closed basins is consistent with historical proxy data indicating drought in East Africa from the late 1700s to the 1820s–1830s and a return to wetter conditions after that (Nicholson, 1995).

In Lake Naivasha, the bottom of core NM93.1-S consists of peaty mud with abundant seeds of papyrus, *Cyperus papyrus*. This horizon is interpreted as representing an episode when near-complete desiccation transformed Lake Naivasha into a fragmented shallow wetland overgrown by papyrus swamp. A radiocarbon date of 290 ± 50 yr BP for lake mud overlying the peat argues against contemporaneity with the early 19th-century phase of lake filling identified as the Unit 0–Unit I boundary in Lake Sonachi and Lake Oloidien. In NM93.1-S the latter boundary must be located somewhere between 89 and 100 cm depth (Fig. 6), with a high probability of discontinuous sedimentation below it. The sharp drop in ^{210}Pb-activity between 99 and 102 cm depth suggests truncation of the record at that level, i.e. immediately above the dated horizon. Core NC93.1-S from Crescent Island Crater (Fig. 8) contains only the upper part of Unit I. The lower part of Unit I and pre-19th-century deposits equivalent to Unit 0 are represented by core NC93.2-L, which has a radiocarbon date of 580 ± 50 yr BP at 420 cm depth.

4. Results and discussion: Four times a sediment record

4.1. Lake Sonachi

Verschuren (in review) used the historical sediment record of Lake Sonachi to investigate relationships between lake level and mixing regime and the texture and composition of offshore sediments in a fluctuating tropical soda lake. He described 19th- and 20th-century sediments in Lake Sonachi as consisting of alternating sections of dark-brown finely laminated muds and somewhat lighter colored coarsely laminated muds (Fig. 4d). The absence of fine (millimeter-scale) lamination in Unit I sediments (33.2–23.9 cm depth) suggests a holomictic regime during most of the 19th century, but preservation of distinct sub-centimeter-scale lamination (Fig. 5) implies that bioturbation was insignificant. Similar to today, Lake Sonachi at that time may have remained chemically stratified for most of the year, with periodic events of deep circulation failing to inject sufficient oxygen into the lower water column to permit immigration and persistence of zoobenthos. An extreme lowstand of the Lake Naivasha basin during the 1870s and early 1880s (Åse *et al.*, 1986) is reflected in the Lake Sonachi record by strongly fluctuating rates of sediment accumulation at the top of Unit I (28.8–23.9 cm; Fig. 4i), evidence for shifting sedimentation and periodic redistribution of shallow-water sediments offshore (Dearing, 1983).

Unit II (23.9–13.0 cm) consists entirely of finely laminated muds. The ^{210}Pb-inferred date of 1891 ± 3 yr for the Unit I-II boundary correlates well with the prominent transgression of Lake Naivasha between 1883 and 1894 (Fig. 5) that followed a decade of above-average rainfall (Nicholson, 1981). If the depth of deepest seasonal water-column circulation at that time was comparable to the chemocline depth observed during more recent episodes of meromixis (4–5 m; MacIntyre and Melack 1982; Njuguna 1988), it appears natural that a 14-m lake-level rise resulted in permanent chemical stratification and preservation of fine lamination. A 6-m lake-level decline in the period 1906–1910 failed to

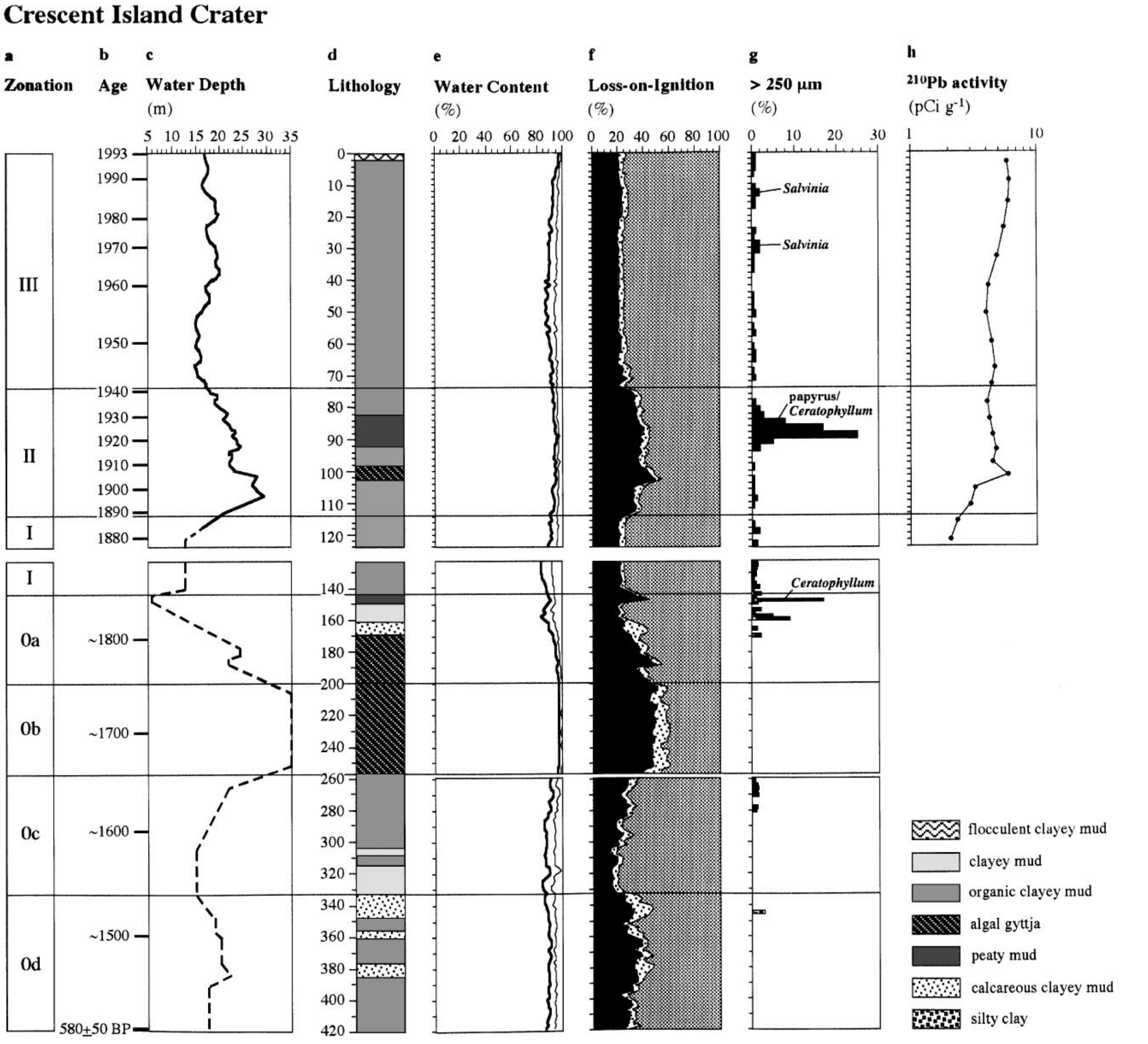

Fig. 8. Crescent Island Crater piston-cores NC93.1-S and NC93.2-L: (a–h) as in Fig. 4. Depth scale for core section 124–420 cm is half that of section 0–124 cm.

disrupt meromixis but did cause a temporary reversal in the trend toward reduced rates of sediment accumulation (Fig. 4i). Lake level settled around 1890 m a.s.l. for two decades, after which it started a progressive decline that culminated in the historic lowstand of 1946 and a water depth of only 2.5 m (Fig. 5). The ^{210}Pb-derived age for the Unit II–III boundary suggests that loss of fine lamination and inferred disruption of meromixis happened shortly before the lowstand was reached.

The varied composition of Unit III (13.0–0.0 cm) requires definition of three subunits. Subunit IIIa covers both the historical lowstand of 1945–1956 and the ensuing lake-level rise, and consists of coarsely laminated ochreous and dark-brown muds interrupted by two pale horizons of colloidal amorphous silica (Verschuren, in review). This silica precipitated directly out of the water column when downmixing of a large freshwater input caused a sudden drop in pH (Jones *et al.*, 1967; Stumm and Morgan, 1970). In agreement with the ^{210}Pb-chronology, silica precipitation most likely occurred during two episodes of above-average rainfall in 1956–1958 and 1961–1963 (Flohn, 1987; Nicholson *et al.*, 1988) that led to the two-step lake-level rise from a depth of 3 m in 1956 to 7 m by 1964 (Fig. 5).

Although the depth of 7 m reached in 1964 was sufficient for formation of a present-day chemocline, meromixis does not appear to have been re-established until about 1970, at the Subunit IIIa–IIIb boundary (Fig. 5). Failure of chemical stratification to persist year-round before that time suggests that the density stratification was weaker than normal. Local Rift Valley rainfall was below average during 1964–1967 (Nicholson *et al*., 1988), consequently the high level of Lake Sonachi during the mid-1960s must have been sustained by groundwater-driven adjustment to the level of Lake Naivasha rather than local rainfall. Continuous inputs of salt-charged groundwater (MacIntyre and Melack, 1982) combined with below-average freshwater input and strong evaporation at the lake surface may have delayed meromixis by preventing development of a sufficiently stable density stratification (Verschuren, in review).

The stratigraphy of Subunit IIIb can be compared directly with field data collected by MacIntyre and Melack (1982) and Njuguna (1988) between 1970 and 1981 (Fig. 5). Lake Sonachi started this period with stable meromixis and preservation of millimeter-scale laminae. As a result of below-average rainfall in 1972–1973 and again in 1975–1976, lake level declined 2.5 m to a water depth of about 5 m. This period of lake-level decline is incorporated in the sediment record as three clear varve couplets of white calcite laminae alternating with organic clayey mud (Fig. 5). Surface waters were gradually down-mixed into the monimolimnion, so that by November 1976 surface and bottom conductivities were nearly identical and chemical stability was very low (MacIntyre and Melack, 1982). Although field data for the period 1975–1978 are scarce, meromixis was lost shortly before or after November 1976 and not re-established until 1978 or early 1979. This episode of holomixis is represented by a 4-mm thick unlaminated horizon dated to the years 1976 through 1978 (Fig. 5). Following two years of heavy rainfall in 1977–1978, lake level rose 2.5 m and meromixis was restored (MacIntyre and Melack, 1982), as reflected in a new sequence of finely laminated sediments first without and then with calcite laminae forming part of the varve couplets. Subsequently meromixis was lost again by the mid-1980s after lake depth dropped to less than 5 m. Subunit IIIc was deposited under the present regime of holomixis at low lake level and consists of layers of olive-green gyttja alternating with brown flocculent muds containing aggregations of decomposing algae (Fig. 5).

The richness of detail displayed by the sediment record of Lake Sonachi is unmatched by the other three basins. Excellent agreement between the ^{210}Pb-inferred ages of stratigraphic marker horizons and the documented or probable age of the changes in bottom dynamics that created them, justifies the use of Lake Sonachi's sediment record as a reliable and accurate archive of 19th- and 20th-century lake history with a time resolution of individual years (Verschuren, in review). The formation of this climate-proxy record resulted from frequent changes in sedimentation patterns associated with the crossing of limnological thresholds. Its excellent preservation resulted from a unique combination of basin-morphometric and limnological factors that protected the sediment-water interface against both physical and biological disturbance.

Foremost among the factors that contributes to undisturbed sedimentation in Lake Sonachi is the superior wind shelter provided by its crater rim. Expressed as the ratio between the minimum height of the rim above the water surface and lake width in the predominant wind direction, Lake Sonachi ranks among the most sheltered of East African crater lakes (Melack, 1978, 1981). Equally important is its status as a hydrologically closed basin. Evaporative concentration of dissolved salts created elevated salinities that led to development of strong density stratification and increased the resistance of the lower water column to propagation of wind-induced turbulence. Density stratification also prevented convective circulation to adequately replenish hypolimnetic oxygen supplies; together with high oxygen needs for bacterial decomposition of sedimented algal blooms (Wood *et al*., 1984), this has led to persistent bottom anoxia both during meromixis and holomixis and prevented bioturbation in lake depths as shallow as 2 m. The rather low rate of average linear sediment accumulation in Lake Sonachi (Fig. 3) implies that if bioturbation had occurred, the resulting homogenization of surface deposits would have instantly reduced the resolution of sedimentary climate-proxy signals to a timescale of decades.

4.2. Lake Naivasha

Recent sediments in Lake Naivasha are dark-brown organic muds with no visible lamination (Fig. 6d). Near-constancy of ^{210}Pb-activity down to 15 cm core depth (Fig. 6h), substantially deeper than the typical depth of bioturbation (Robbins *et al*., 1977), points to wind-driven resuspension and settling as the dominant cause of sediment mixing (Krishnaswami and Lal, 1978). This radionuclide evidence for wind stress affecting the lake floor at 6 m water depth is consistent with the critical depth of 7.5 m derived from wave theory (Table 2). Notwithstanding basin-wide physical sediment disturbance at today's lake level, the sediment record of Lake Naivasha does contain three well-defined stratigraphic units (Fig. 6f) equivalent to those in Lake Sonachi. As in Lake Sonachi, Unit II sediments in Lake Naivasha display an initial increase in OM content coincident with the lake-level rise of 1883–1894, followed by a decrease during the lake-level decline of 1906–1910. Reduction of lake depth from 17 to 11 m at that time resulted in a substantial increase in bottom areas subject to sediment erosion and transport, causing enhanced focusing of low-OM shallow-water sediments to the core site.

Between 58 and 66 cm depth, Unit II sediments contain significant amounts of coarse OM of non-algal origin (8–19% of total OM; Fig. 6g). This organic debris primarily consists of partial roots and shoots of the submerged macrophyte *Ceratophyllum demersum*, and its presence in Unit II sediments creates a texture similar to that of surface muds in the vicinity of submerged weed-beds (D. Verschuren, personal observation). This horizon is interpreted to represent growth of submerged vegetation during the period of stable lake level between about 1910 and the early 1920s (Fig. 6c), possibly down to about 8 m water depth (cf. Beadle, 1932). Lesser amounts of coarse OM at 44–58 cm depth contain both *Ceratophyllum* and remains of papyrus. They cover the period of lake-level decline during the late 1920s and 1930s, when lake-fringing papyrus swamp grew from seeds on freshly exposed mud flats and then died after being stranded (Gaudet, 1977).

The historical lowstand of 1945–1956 is represented in core NM93.1-S by a subsurface minimum in OM content at 34–40 cm depth, immediately above the Unit II–Unit III boundary (Fig. 6f). With the lake's open-water surface area reduced to about 100 km^2, periodical flushing of low-OM material from exposed peripheral mud flats must have diluted autochthonous OM and contributed to increased accumulation in the deepest part of the basin (Fig. 6i). The lack of an equivalent horizon in lower Unit III deposits of cores NM91.1-S and NM93.2-S (Fig. 3) reflects local sediment erosion or non-deposition during the lowstand, translated in an unconformity at the Unit II–III boundary. Both of these sites are shallower than the site of NM93.1-S; more important however is their greater exposure to the predominantly southerly winds (L_f: 5.4 and 8.6 km) creating considerably greater critical depths (Z_{a-t}: 9.2 and 13.1 m versus 7.5 m; Table 2).

The uniform color and OM content of upper Unit III sediments (Fig. 6d and f) appears to suggest that Lake Naivasha's offshore bottom dynamics remained virtually unchanged over the past three decades, notwithstanding lake-level fluctuations that caused water depth at NM93.1-S to vary between 5 and 9 m (Fig. 5c). This is unexpected, considering the substantial lateral shifts in the sedimentation boundary that accompany lake-level changes in this large and shallow basin. Since lake depth continuously balanced around the critical depth of 7.5 m, with drawdowns well below Z_{a-t} during much of the 1970s and from 1985 until today, wind stress must regularly have caused deep sediment mixing and obliterated any stratigraphic evidence of short-term changes in bottom dynamics. Although continuous polymixis and adequate oxygenation allowed persistence of an offshore bottom fauna throughout this period, bioturbation has evidently played only a secondary role in sediment mixing. The rather low macrobenthos densities in Lake Naivasha (870 oligochaetes and 390 chironomids m^{-2} in 1982–1984; Clark *et al.*, 1989) are consistent with field data from other wind-stressed lakes that document a general negative correlation between macrobenthos density and the depth of wind-driven sediment disturbance (Darlington, 1977; Wiederholm, 1978).

4.3. Lake Oloidien

Recent sediments in Lake Oloidien (Fig. 7) display a more striking stratigraphy than those in Lake Naivasha, primarily because the high carbonate content of Unit I and III sediments gives them a contrasting light color (Fig. 7d). Sedimentary evidence for carbonate precipitation in Lake Oloidien is consistent with field data indicating high pH (9.0–9.2), supersaturation with regard to calcite, and a low dissolved Ca/Na ratio in comparison with Lake Naivasha (Gaudet and Melack, 1981; Brierley *et al.*, 1987). The low-carbonate, high-OM sediments of Unit II were deposited during the 1890–1940 period of high lake level when Lake Oloidien was broadly confluent with Lake Naivasha and fresh. The substantial variation in OM content of recent Lake Oloidien sediments (Fig. 7f) can be explained in large part by dilution or concentration of algal OM during times of enhanced or reduced influx of allochthonous mineral sediment components (Fig. 7i). Significant amounts of papyrus remains together with *Ceratophyllum* in a dark plant-debris horizon at 49–57 cm depth (Fig. 7d and g) reflects local growth of sizable papyrus reefs during confluence with Lake Naivasha between about 1900 and the 1930s (Worthington, 1932; Lady S. Cole and M. D. Carnelley, *personal communication*, 1993). The separation of Lake Oloidien from Lake Naivasha in about 1938 and further lake-level decline to a lake depth of 4 m by the mid-1940s started a period of evaporative concentration and rising salinity that eventually eliminated *Ceratophyllum* and other freshwater plants; papyrus reefs became stranded and died off or were cleared for horticulture (Gaudet, 1977, Harper *et al.*, 1990). Salinity decreased again following lake-level rise in the 1960s, but papyrus failed to become re-established (Verschuren *et al.*, in press).

According to the ^{210}Pb-chronology, the uppermost 9 cm of unconsolidated surface deposits in Lake Oloidien represent just 3 years of accumulation. The sharp increase in OM content towards the surface reflects incomplete diagenesis of algal OM before permanent burial, suggesting that processes of OM degradation are less effective in Lake Oloidien than in Lake Naivasha. The occurrence near the sediment surface of distinct horizons containing variable amounts of flocculent algal remains (Fig. 7d) implies that sediment mixing, from either bioturbation or wind-induced turbulence, is insignificant. Wave theory predicts that offshore sedimentation in Lake Oloidien at today's lake level is unaffected by wind stress (Table 2). Macrobenthos densities in Lake Oloidien (680 oligochaetes and 270 chironomids m^{-2}; Clark *et al.*, 1989) are an order of magnitude lower than those reported capable of complete homogenization of surface sediments (Robbins, 1982). Possible factors contributing to these low densities include the occasional occurrence of bottom anoxia (Melack, 1979), the poor support provided by a substrate of flocculent surface muds (porosity 98%; Fig. 7e), and inpalatability of algal detritus derived from cyanobacteria (Johnson *et al*., 1989).

Rates of offshore sediment accumulation in Lake Oloidien are inversely proportional to lake level (Fig. 7c and i). Low accumulation rates ($<0.02\ \mathrm{g\,cm^{-2}\,yr^{-1}}$) during the highstand of 1890–1940 can be attributed in part to the role of shorefringing papyrus swamp as a buffer against soil run-off. Disappearance of this swamp buffer during lake-level decline in the 1930s–1940s thus set the stage for enhanced sediment influx. However, although Lake Oloidien has lacked papyrus swamp or other littoral vegetation since the 1940s, sediment accumulation retained its strong inverse correlation with lake level throughout the recent sediment record (Fig. 7c and h). This indicates that variation in the rate of offshore accumulation may be determined more by changes in the amount of shallow-water sediments being redeposited offshore than changes in the yield of terrestrial run-off. Peak accumulation rates of $0.07–0.08\ \mathrm{g\,cm^{-2}\,yr^{-1}}$ were reached during the lowstands of 1945–1956 and 1986–1990, when drawdown to lake depths of 5 m or less caused the boundary between zones of accumulation and transport (i.e. where water depth equals Z_{a-t}) to move far offshore, but not during the less severe lowstand of 1975–1977 when this boundary stayed closer nearshore (Fig. 7c and h; compare with the depth contours in Fig. 3).

4.4. Crescent Island Crater

Unit II sediments in Crescent Island Crater consist mostly of dark-brown organic muds (Fig. 8d). Peak OM concentrations occur at 99–102 cm depth in a horizon of somewhat gelatinous algal gyttja (Fig. 8d). Since this material reflects a regime of hemipelagic sedimentation with little influence of shallow-water environments, it appears that for some time during the 1890–1940 highstand the northeast sector of Lake Naivasha stopped contributing materials to profundal sedimentation in the crater. This most likely occurred between 1892 and 1908, when water depth above these shelf areas exceeded the critical depth (Z_{a-t} of Lake Naivasha re-evaluated for the local L_f) and shallow-water sediments were retained on the shelves. The probable presence of extensive macrophyte beds at that time (Beadle, 1932; Milbrink, 1977) may also have helped stabilising the lake floor (Anderson, 1990). Sediment transport into the crater resumed shortly thereafter, however, possibly initiated by the lake-level decline of 1906–1910.

As in Lake Naivasha and Lake Oloidien, lake-level decline during the 1920s and 1930s is incorporated as a horizon of peaty mud with organic debris derived from stranded papyrus reefs and uprooted macrophyte beds (Fig. 8g), followed by a sharp drop in OM content at the Unit II–Unit III boundary (Fig. 8f). In contrast with Lake Naivasha, evidence for the lowstand of 1945–1956 is limited to a few irregularities in the OM profile. With lake elevation at 1882 m a.s.l., Lake Naivasha's northeast sector had fallen dry (cf. the 4 m depth contour in Fig. 3), so that the crater rim was closed and Crescent Island became sedimentologically separated from Lake Naivasha. Groundwater flow maintained Crescent Island Crater at the same level as Lake Naivasha, however, and there is no evidence that evaporative concentration significantly affected salinity (C. Cocquyt, *unpublished fossil-diatom data*). Crescent Island Crater was reunited with Lake Naivasha after renewed lake-level rise in the late 1950s.

Crescent Island Crater does not enjoy much wind shelter from its partly submerged crater rim, but at historical lake depths of at least 13 m its profundal bottom has never been affected by wind-driven turbulence. The rather uniform composition and OM content of Unit III sediments deposited since that time can be attributed in part to the activities of a dense macrobenthos community (9000 oligochaetes and 4000 chironomids $\mathrm{m^{-2}}$ in 1971–1973; Milbrink, 1977). But in fact bioturbation may have little impact on the preservation of climate-proxy signatures, because the very high rates of sediment accumulation (averaging $14\ \mathrm{mm\,yr^{-1}}$ over the past 40 yr) may outpace the rate at which the macrobenthos can process and homogenize the sediment (Robbins, 1982; Cohen, 1984). Another issue is whether a 4-m fluctuation of lake depth between 14 and 18 m could have sufficiently affected profundal bottom dynamics in this steepsided crater basin to create sedimentary climate-proxy signatures even in the absence of bioturbation. Although at current lake depths the basin is continuously polymictic and usually oxygenated throughout, deoxygenation of the hypolimnion appears to have been common in the early 1980s when deepwater oxygen supply from seasonal convective circulation was unable to meet increased bacterial oxygen demand for the decomposition of sunken rafts of *Salvinia molesta* (Brierley *et al*., 1987). An outbreak of this exotic water fern in the period 1980–1984 had formed a vast floating mat along Lake Naivasha's north shore, which then broke up and drifted around the lake and into Crescent Island Crater (Harper, 1984; Harper *et al*., 1990). Early-1980s sediments in Crescent Island Crater, Lake Naivasha, and Lake Oloidien all contain fossil *Salvinia* leaves (Figs. 6g, 7g and 8g), and the slightly higher OM content of early-1980s sediments in Crescent Island Crater (21–11 cm: 23% versus 20% above and below) may reflect the enhanced preservation of OM in an anoxic hypolimnion. The Crescent Island Crater record also documents a second *Salvinia* outbreak that occurred during in the late 1960's (Gaudet, 1976). Thus, while profundal bottom dynamics may not have varied enough recently to substantially affect lithology, it does appear that Crescent Island Crater has the potential to preserve signatures of short-term events at a timescale of individual years. The occurrence of hypolimnetic oxygen depletion during the early 1980s suggests that the crater basin is balancing close to its threshold toward discontinuous polymixis (Lewis, 1983). Enhanced stratification at the lake depths of 20–28 m that prevailed between 1890 and 1940 may well have caused seasonally persistent loss of oxygen, which would have eliminated the bottom fauna and further promoted the time resolution of the sediment record.

Pre-20th century sediments in Crescent Island Crater, as recovered in core NC93.2-L (Fig. 8), comprise four stratigraphic units here identified as subunits of Unit 0 (Fig. 8a). Subunits 0d (420–332 cm) and 0c (332–258 cm) encompass lake history from about AD 1400 to 1650 and consist primarily of brown to dark-brown clayey muds with OM contents between 15 and 39%. Using the positive correlation between OM content and lake level observed in the recent sediment record of Crescent Island Crater (Fig. 8) as a guide, these subunits can be interpreted as having been deposited in water depths of between 10 and 25 m, i.e. comparable to the historical range. The gradual increase of OM content in Subunit 0c and subsequent transition from dark-brown clayey mud to gelatinous olive-brown algal gyttja at the Subunit 0c–0b boundary suggests a trend of rising lake level eventually leading to a regime of pure hemipelagic sedimentation. Lake level must have stood well above 1900 m a.s.l. (lake depth >30 m) at that time, high enough to turn the wind-stressed northeast sector of Lake Naivasha into a zone of undisturbed sediment accumulation. Stratigraphic evidence for a pre-19th century highstand of Lake Naivasha is consistent with other records of high lake levels in Africa during the 17th and early 18th century, such as at Lake Chad in the Sahel region of North Africa (Maley, 1976), and Lake Turkana in northern Kenya (Halfman *et al*., 1994). Discussion of climate history is outside the scope of this paper, but it is significant that evidence for this pre-19th-century highstand appears to be all but missing from the sediment records of lakes Sonachi, Naivasha, and Oloidien. In these shallower basins, sediments dated to sometime before the highstand are immediately overlain by 19th-century deposits. Clasts of finely laminated clayey muds incorporated in Unit 0 (Fig. 5) at Lake Sonachi could be reworked remnants of a highstand sequence.

Subunit 0a of core NC93.2-L (198–144 cm) records the long and progressive lake-level decline that ensued. Algal gyttja above the Subunit 0b–0a boundary gradually becomes more clayey than gelatinous, indicating that Crescent Island Crater started to again receive resuspended shallow-water sediments from the increasingly wind-stressed Lake Naivasha bottom. The horizon of silty peat capping Subunit 0a at 144–148 cm testifies to a brief episode in the early 19th century when Crescent Island Crater was so shallow that submerged *Ceratophyllum* weedbeds could grow on the profundal basin floor (Fig. 8g). It represents the culmination of late 18th- and early-19th century aridity (Nicholson, 1981), a time when Lake Naivasha, Lake Oloidien, and Lake Sonachi stood dry. It is overlain by the dark-brown clayey muds of Unit I, deposited during the early- to mid-19th-century phase of renewed lake filling and continuous with Unit I sediments at the bottom of core NC93.1-S.

5. Synthesis and conclusion

The results of this study prompt four conclusions. First, they generally support the predictions of empirical models based on wave theory (Håkanson, 1977; Håkanson and Jansson, 1983; Larsen and MacDonald, 1993) that the potential of a lake to accumulate a high-quality climate-proxy record is primarily determined by its relative depth: undisturbed sediment accumulation requires greater water depths in a large lake than in a small lake. In these sedimentation models, the effective fetch of predominant winds is the significant horizontal dimension. For example, in Lake Naivasha more favorable bottom dynamics at NM93.1-S than at the central and north-central core sites result not so much from the (only slightly) greater local water depth but from a significantly smaller critical depth due to the shorter fetch of southerly winds. However, wave theory may not be applicable to very small lakes (<0.5 km^2), such as Lake Sonachi and many other lakes in phreatomagmatic explosion craters: it predicts that wind-driven sediment disturbance in Lake Sonachi should be restricted to a water depth of less than 1 m (Table 2) even without the benefit of topographic wind shelter, while sedimentation patterns (Fig. 5) reveal at least periodical wind-driven disturbance at depths approaching chemocline depth. The volume of Lake Sonachi is sufficiently small for most sediment focusing to occur during these infrequent episodes of deep circulation, i.e. below the average critical depth of sediment accumulation (Pennington, 1974; Hilton, 1985). In these circumstances, the main advantage of topographic wind shelter is that it reduces the frequency and energy of such deep-circulation events.

Second, a variety of local conditions can promote the integrity and time resolution of a climate-proxy record beyond what is predicted by sedimentation models. The superior time resolution of the Lake Sonachi record resulted from its combination of great relative depth, wind shelter behind a high crater rim, the additional resistance to deep circulation provided by density stratification, and absence of bioturbation due to high bacterial oxygen demands for the decomposition of dense algal blooms. As demonstrated by the preservation of distinct sediment horizons through the 1940–1950s lowstand, these conditions expand by about 5 m the lowest lake depth that Lake Sonachi can tolerate without disturbance of offshore sedimentation, as compared to Lake Naivasha. Nevertheless, in the long run Lake Sonachi is just as likely as Lake Naivasha and Lake Oloidien to dry out. Only the deeper Crescent Island Crater basin survived early-19th century aridity without losing part of its sediment archive.

Third, lake-level change has a major impact on rates of net sediment accumulation in shallow lakes. If lake depth remains greater than the critical depth throughout an episode of low lake level, such as occurred in Lake Sonachi and Lake Oloidien during the 1940–1950s lowstand, accumulation will tend to increase because peripheral wave action or intermittent deep mixing will focus relatively greater amounts of shallow-water sediments into a relatively smaller zone of accumulation. However, if lake depth falls below the critical depth, such as occurred in Lake Naivasha, periodic non-deposition or erosion of previously deposited sediments will decrease net accumulation and may result in significant truncation of the climate-proxy record. Consequently, if linear sediment accumulation is used as the measure of time, an episode of low lake level can be both over- or underrepresented in the climate-proxy record depending on the severity of the drawdown relative to the local critical depth. This phenomenon may significantly complicate attempts to correlate high-resolution climate-proxy records between lakes even when their hydrological sensitivity to climate change is comparable.

Fourth, bioturbation has relatively minor importance to the integrity and time resolution of climate-proxy records from shallow fluctuating lakes. In part this is due to the often high rates of sediment accumulation in these systems. For example, the near-surface sedimentary signals representing recent short-lived events in Lake Oloidien and Crescent Island Crater were preserved because sediment accumulation outpaced the capacity of the macrobenthos community to homogenize them: in Lake Oloidien the uppermost 5 cm of uncompacted surface muds subject to bioturbation (Davis, 1974; Robbins *et al*., 1977; Krezoski *et al*., 1978) represent just 1.5 yr of accumulation (Fig. 7c). Compared to the smoothing of sedimentary signatures caused by invertebrate activity (Davis, 1974; Leavitt and Carpenter, 1989), wind-driven mixing and erosion can inflict much more pervasive damage to record integrity. This is usually well appreciated when it concerns the major, century-scale unconformities that result from deflation during an episode of complete desiccation and that are easily recognized by clear lithological evidence for sediment erosion or compaction. In contrast, unconformities caused by decade-scale episodes of non-deposition in shallow wind-stressed lakes may often be difficult to trace because mixing of unconsolidated muds deposited before and after the lowstand obliterates the evidence of record truncation at that level. In Lake Naivasha, for example, evidence for a lowstand during the 1940s and 1950s is lacking in core NM91.1-S because of local non-deposition or erosion during that time. The presence of the unconformity that resulted is revealed only through comparison with a complete sequence in core NM93.1-S.

The use of a lake-sediment record to reconstruct climatic variability on time scales shorter than the resolution of its supporting radiocarbon chronology requires demonstration that the sedimentary column is a reliable measure of time. The data presented here show that the changes in bottom dynamics that accompany lake-level fluctuation in shallow lakes can significantly affect the rate of net sediment accumulation and so distort this presumed known relationship between core depth and time. They also vividly illustrate how basin-specificity of the sedimentary environment can translate an identical record of past lake-level change into markedly different climate-proxy records. Regional correlation of inferred climate events needs to account both for hydrology-related differences in the climatic sensitivity of lakes and sedimentation-related differences in the integrity and time resolution of the accumulated archives of climate history.

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