An uncertainty-focused database approach to extract spatiotemporal trends from qualitative and discontinuous lake-status histories

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Content

Lake Chilwa

Lake Chilwa is a shallow (<3 m depth), enclosed saline lake occupying a shallow tectonic depression of post-Cretaceous age, 100 km SE of Lake Malawi. The basin itself is some 160 km from north to south and 100 km east to west, with an area of c7500 km2. Based on 1950–1976 data, Lancaster (1979a) gives a mean lake level at 622 m asl, with inter- annual fluctuations of up to 1.0 m and up to 2–3 m on a decadal timescale. When at a level of 622 m, the lake has an area of approximately 600 km2, although this changes rapidly in response to budgetary fluctuations. During the historical record the lake has been a closed hydrological system, although Lancaster (1981) suggests that up until the mid-Pleistocene, Lake Chilwa was an open lake, linked to the Indian Ocean via the Lugenda River and Lake Chiuta to the north.

The lake status history is derived from 36 distinct OSL ages. All these ages represent the same level (higher than present, 622m asl). The ages where clustered in 4 different groups following the protocol designed by the GLSDB consortium. These results are coherent with the data presented by Thomas et al. (2009).

Lake Tritrivakely

Tritrivakely is an 80-m-deep maar of the Ankaratra volcanic chain in central Madagascar. The slopes (30–40°) consist of basaltic tuff, occupied today by agriculture. The lake, about 1 km wide, is mainly supplied by rain falling in the crater, has no surface outlet, and is very sensitive to seasonal and interannual changes in P–E. During dry seasons or years (e.g., November, 1992), the water body is an ombrotrophic Cyperaceae marsh with permanent water hollows. Water characteristics are close to those of rain water (electric conductivity, C: 20.6–37 $mScm^{-1}$; pH 5.2–6.3; $SiO2$ content: 1 mg l^{-1}), with maximum values in water hollows. After rains of exceptionally wet years (e.g., May 1994), Tritrivakely becomes a shallow lake, 2–3 m deep. Runoff into the lake tends to increase C (100 $mScm^{-1}$), pH (6–6.7), and $SiO2$ content (4.6 mg l^{-1}). However, no significant change water turbidity takes place because the particle load from the catchment is filtered by vegetation (Gasse and van Campo, 1998).

The lake level history of Tritrivakely is primarily derived from the diatom records, supported by the pollen of aquatic vegetation. The DCA analyses presented by Gasse and van Campo (1998) described four different types of diatom communities that we associated with four different lake levels. An additional 'dry' level was added. The history is then derived from the four components, with the lake level at any time being associated with the dominant component. The samples characterised by an absence of diatoms are considered as dry lake stands.

Lake Lungué

Lake Lungué is located at the eastern margin of the Limpopo River floodplain ca. 30 km north of Xai Xai City and 53 km west of Malehice Locality. The distance to the present Indian Ocean shoreline is ca. 40 km. Today, the lake connects with to the Limpopo River floodplain only during flooding events, e.g. during the 2000 CE flood (cf. Karlsson and Arnberg, 2011). However, the present lake surface is at an altitude of ca. 8 m a.m.s.l. (above mean sea-level; Google Earth information), and the lake is thus not influenced by the high tide, which reaches ca. 3 m a.m.s.l. at the outlet of the river during spring tide (Massinga and Hatton, 1996; Sete et al., 2002).

The diatom record used to reconstruct the lake level history of Lake Lungué indicates that the lake was influenced by the Indian Ocean during the oldest half of the record (zone DAZ 1 to DAZ4). The Lake became fully disconnected from marine influence around 600 cal BP. Two diatom zones characterise the past 600 based on the higher concentration of freshwater taxa observed in zone DAZ5.

Lake Cheshi

Lake Cheshi is a perennial lake of 100km2 lying in a closed basin between Lakes Mweru and Tanganyika in northern Zambia (9”05’S, 29”45’E). It is adjacent to and sometimes confluent with larger Lake Mweru Wantipa, the area of which fluctuates greatly in response to climatic conditions (Lee, 1972). Mweru Wantipa has an outlet to the Kalungwishi River, a tributary of the Congo River. The altitude of Lake Cheshi’s surface in 1971 was 928 m. A

The chronological controls of the diatom record of Cheshi are limited, and most of the glacial times exhibit relatively low accumulation rates of sediments. We followed the original lake level history based on the diatom zonations. We identified four levels from the 9 different zones. Zone A1 was excluded because of its low concentration in diatoms. Zone E and F were merged at they do not document significant differences in the diatoms composition.

Ishiba Ngandu

Ishiba Ngandu occupies a basin formed by gentle down-warping. The lake has a length of about 6km, an area of about 7km2, and a depth of about 7m. The catchment of the lake is about 670km2 in area. The two main rivers that feed the lake flow year-round and are fed in part by at least eight springs. The outflow from the lake also flows year-round and eventually enters Lakes Bangweulu and Mweru before joining the Congo River and draining into the Atlantic Ocean.

The record of Ishiba Ngandu is poorly dated and all the ages after 20,000 cal BP are extrapolated. The start date of the first lake status is therefore highly uncertain. Lake level history was derived from one single diatom record, and we followed the zonation of Haberyan (2018) to define the lake history. Four zones and three different levels were estimated based on the diatom composition of the samples.

Lake Nhauhaque

At the time of sediment coring (July 2008), Lake Nhauhache was about 150 m in diameter and 2.2 m deep. The lake is one of a series of lakes in the same drainage area and is situated at the highest elevation. The lake was circular in shape and located 6 m a.s.l. The surrounding slopes possess an amphitheatre- like shape, except in the southeast, where the topography is flatter. This lower threshold, situated 1-1.5 m above the coring site, allows overland drainage of the lake during high stands, via the Inhamalene River to the Indian Ocean.

The record is ~2300 years long and the diatom data used to code the lake status history suggest a generally drying trend over most of the period, except for the past ~300 years when the lake supported the highest stand recorded. Four different phases have been identified, primarily based on the relative dominance of planktonic and benthic taxa. The classification was facilitated by a CONISS analysis.

Lake Mapimbi

Lake Mapimbi is a small pool, approximately 40 m 9 150 m situated in the Limpopo river valley, in the Pafuri area of the Kruger National Park. The lake was formed as part of the Limpopo river system, as an oxbow lake. Throughout the period under study, it has been entirely cut off from the Limpopo River and is fed by groundwater. The record is short and covers the last 6 centuries.

The lake status history is derived from diatom data and we follow the zonation defined by Gillson and Ekblom (2009). The classification is based on the relative abundance of planktonic and benthic taxa. Three different levels (1. Dominated by benthic taxa, 2. Mix of benthic and planktonic taxa and 3. Dominatied by planktonic taxa) and 4 phases have been identified.

Lake Sibaya

Lake Sibaya, South Africa’s largest freshwater lake (65 km2), is located in Maputaland, northern KwaZulu-Natal and is separated from the Indian Ocean by a dune barrier up to 2 km wide and 130 m high (Stager et al. 2013). Its surface lies 20 m above sea level and the catchment covers 540 km2. Mesozoic volcanics and sediments underlie sands, clays, and calcarenites that were deposited in a barrier-lagoon complex during the Last Interglacial marine highstand (Miller, 2001). The lake is oligotrophic, circumneutral (pH 7-8: Allanson, 1979; Jury and Govender, 2000; measurements by the authors), and dilute (60-70 mS/cm; Allanson, 1979).

Lake Sibaya diatom record coupled with conductivity analysis suggest 3 main phases during the past 2000 years. The original study proposed a much more detailed lake-status history, with levels changing for every diatom sample. Such variability is not evident from the diatom diagram and we prefer to keep a higher order description, and we do not discard the value of the high precision lake level estimate.

Princess Vlei

Princessvlei is a permanent, alkaline, eutrophic, freshwater coastal lake (Harding, 1992) that lies in an inter-dunal depression encroached on by high-density residential, industrial and agricultural land uses. The lake is not, and possibly has never been, directly affected by marine intrusions during the Holocene being slightly inland and at a moderate elevation from the coast. Unconsolidated sand dominates most of the area, originating from river and wind erosion and deposition, with increased transport during periods of summer aridity and anomalously high southerly wind velocities (Harris et al., 1999; Roberts et al., 2009). Freshwater sources into Princessvlei include direct precipitation, river runoff and ground- water via the Cape Flats aquifer (Harding, 1992; Parsons and Harding, 2002). Most of the lakes in the region overlie this aquifer; in winter, the groundwater is just a few centimetres below the surface, but during the drier summer months the water table can drop to several metres below the surface (Harris et al., 1999; Neumann et al., 2011).

The record spans the last ~2,900 years and the lake-status history is derived from diatoms (Kirsten and Meadows, 2016). The CONISS analysis describes three clear phases in the record, and these are well supported by variations in the sediment accumulation rate. However, the diatom assemblage compositions are also strongly influenced by water quality.

Lake Chilau

Lake Chilau is situated in the upper drainage area of River Inhaliave in Mozambique. Lake-status is solely based on the diatom study of Norström et al. (2018). The record is short and covers the last 650 years. Our age-depth model differs slightly from the published model. In ours, the trend after the last radiocarbon age does not change, while it increases in the published model for no apparent reasons. This does not influence the ages of the diatom samples used to reconstruct lake-levels. We also added the age of the top sample as modern.

The three coded levels are derived from the benthic / planktonic ratio of diatom species, with zonation being derived from a standard CONISS analysis.

Lake Nhaucati

Lake Nhaucati is situated about 2km from the coast at an elevation of ~20m above sea level. The lake is one of many smaller lakes in the Vilanculos region. It has no surficial inlets or outlets, and lake levels are instead regulated by direct rainfall and recharge of ground water through shallow aquifers (Coetsee and Hartley, 2001), probably modulated by shifts in evaporation (Norström et al. 2018). The record spans about the past 1500 years. The age-depth model was derived from 14 radiocarbon dates.

The lake-status history is derived from the diatom study of Ekblom and Stabell (2008). The resolution of the record does not allow for an in-depth reconstruction of past lake levels. Four phases were identified from the diatom diagram (2 high and 2 low phases). The classification follows the original interpretations. However, of the 7 zones that were originally defined, we retained only 4 and grouped some of them together because it was not always clear how to position the different zones relative to each other.

Verlorenvlei

The Verlorenvlei catchment occupies 1890 km^2, comprising the Verlorenvlei River, its tributaries, extensive marshes, and a coastal lake, and is distinctive both topographically and geologically (Kirsten et al., 2020). Shifting sands and dune building during the late Holocene led to a southward migration of the mouth and the near isolation of the valley from the sea, forming a narrow, 2.5 km long estuarine channel. Verlorenvlei (18° 26' E; 32° 21' S) is a mesotrophic, shallow lake ~1 m AMSL with a surface area of ~10 km2. The lake has an average depth of 2.5 m, reaching a maximum of 5m, with pH ranging between 6.7 and 9.6. The water is relatively low in salinity, although more saline conditions are experienced in the lower reaches. Water temperature ranges from 15.2 to 20.4 °C and transparency from 17 to 114 cm. Winds maintain turnover of surface waters, creating a thoroughly mixed and homogeneous water column, resulting in neither thermal nor salinity stratification. Occasional breaching of the mouth occurs during storm events and spring tides. The lake is fed by the ephemeral Verlorenvlei River and several springs. Evaporative losses are high, and seasonal fluctuations are such that in the dry, windy summers the channel feeding the lake may be reduced to a series of isolated hypersaline pools.

Information on Verlorenvlei's lake-status history is sourced from Kirsten et al. (2020). The record covers the last ~10,000 years but a strong proportion of marine taxa is visible from the diatom abundance diagram, which suggest full marine conditions at Verlorenvlei during most of the Holocene. The period of high marine influence (10,000 – 2000 cal BP) cannot be used to infer past lake levels and the coding focuses on the past 2,000 years, which corresponds to phase 3 as described in the original publication. Two different subphases can be deduced from the diatom abundance, with first brackish conditions until ~900 years ago, followed by a phase of freshwater until present day. The division in subphases is based on the CONISS analysis.

Lake Naivasha

Lake Naivasha is located on the floor of the Central Kenya Rift Valley, its basin bounded by the rift escarpments to the west and east, by Mount Eburru volcano to the north and by Mount Longonot volcano and the Olkaria volcanic complex to the south. These volcanoes have played a major part in shaping the Naivasha basin over the last 300-400 kyr (Bergner and Trauth, 2004). Lake Naivasha is, compared to its neighbouring lakes in the central Kenya Rift Valley, relatively dilute (conductivity 250-500 µS/cm) and neutral (pH ca. 8) owing to groundwater seepage out of the lake. The two main inflowing rivers, the Malewa and the Gilgil, drain a catchment that includes moist mountain ranges of the eastern and western escarpments, despite the moisture balance on the floor of the rift valley being highly negative. The largest part of Lake Naivasha's surface area is formed by its main basin, which is shallow (z_{max} 6 m in 1993) and wind-stressed because of its high lake-surface area to lake depth ratio (Verschuren, 2001). Close to the lake's eastern shore there is a small (ca. 2 km²) and relatively deep (16 m in 1993) submerged crater basin know as Crescent Island Crater. Although this crater has been connected to the main basin throughout historically documented times, it forms a separate body of water if lake levels drop below 20th-century minima.

The lake status of Lake Naivasha before ca. 11 kyr BP was coded based on a ca. 16-m sediment core from the lake's main basin, collected in 1969-1970. Diatom-derived lake-level changes were reported by Richardson and Dussinger (1986). The major lake-status transitions were radiocarbon-dated by the original authors, and these radiocarbon dates were here recalibrated with the IntCal13NH calibration curve (Reimer et al., 2013). For the period from 11 kyr to the present, the lake-status history of Lake Naivasha was coded based on two sediment cores from the deepest point of Crescent Island Crater, a submerged crater basin within Lake Naivasha. Richardson and Richardson (1972) collected a 27.57-m unnamed sediment core in 1960-1961, Verschuren et al. (2000) collected 8.22-m core NC93.2-L. The top of a corresponding paleosol present in both cores was radiocarbon-dated to 3000 ± 60 ^{14}C yr BP on bulk organic matter by Richardson and Richardson (1972), and to 1800 ± 50 ^{14}C yr on grass charcoal by Verschuren (2001). This allows the calculation of a reservoir effect of 1200 ^{14}C years, which we applied to the other three radiocarbon dates of Richardson and Richardson (1972) to recalculate the calibrated ages of major lake-status transitions as described by the original authors.

Core NC93.2-L's original age model, published by Verschuren et al. (2000), was revised in a later publication (Van der Meeren et al., submitted). The selection of 11 retained ^{14}C dates remained unchanged, along with the additional age-constraining intervals representing three known outbreaks of the exotic water fern *Salvinia*, three stratigraphic marker horizons for which the dates could be transferred from a ^{210}Pb-dated core, and two constraints anchoring known historical lake-level positions to observed sedimentology. The authors recalibrated all radiocarbon dates using the IntCal13NH calibration curve (Reimer et al., 2013). Bacon in R was used, increasing flexibility for sedimentological units to have different sedimentation rates by introducing boundaries, inspired by the specific sedimentological context of Lake Naivasha's Crescent Island Crater. The revised age-depth model by Van der Meeren et al. (submitted) was retained here. Lake-status reconstruction is mainly based on lithological changes throughout the core, assisted by sedimentological composition and diatom- and chironomid-based salinity reconstruction (Verschuren et al. 2000; Van der Meeren et al. submitted).

An additional line of evidence of past lake-level change comes from paleoshoreline deposits above the present-day lake surface. Lake-surface elevation, lake-size and lake-volume estimates corresponding to the early- to mid-Holocene high stand were taken from Washbourn-Kamau (1975) and Bergner et al. (2003).

Lake Ashenge

Lake Ashenge's lake-status history was coded based on the study of Marshall et al. (2009), which presents sedimentological, diatom and stable-isotope (on authigenic carbonate) data from a ca. 8-m sediment core (03AL3/2) collected from the northern underwater slope of the lake. The stable-isotope data was not taken into account for relative lake-status coding, but corresponds well with the overall patterns in lake level as deduced from the other lines of evidence and as such probably at least partly also reflects lake level. The exceptionally low d13C and d18O values of authigenic calcite between ca. 7600 and 5500 cal yr BP, and their abrupt increase at the end of this period, did contribute to the hypothesis that the lake was overflowing during this period.

Most of the early-Holocene sediments at the core site of core 03AL3/2 have been removed after deposition, leaving only the heaviest shells and coarse sand as lag deposits and resulting in a ca. 4000-year sediment hiatus at 650 cm core depth. The cause of the sediment removal is surmised to be a (near-)complete desiccation of the core site, and was thus interpreted in terms of lake level. The timing of the onset of this low stand cannot be determined given the available data, and neither can the lake status corresponding to the removed sediment. The best-estimate duration of the low-stand event was conservatively set at 50 years, though it could have been much longer.

Radiocarbon dates reported in the original study were recalibrated using the IntCal13NH calibration curve (Reimer et al., 2013). All dates were determined on bulk organic matter, the possibility of an old-carbon effect was not investigated. As in the original publication, all published ages were retained. The age-depth model was reconstructed using Bacon in R, defining a hiatus at 650 cm as described by Marshall et al. (2009). To cope with variable sedimentation rates in the core section below 650 cm, t.a and t.b of the dates in that section were changed from their default values to 33 and 34, respectively, to approximate a normal distribution for the error distributions of those dates.

Lake Nabugabo

Lake Nabugabo is a small, shallow, swamp-rimmed lake at the northwestern edge of Lake Victoria, separated from it by a ca. 3 m high sandy barrier. The lake receives surface inflow from local springs and the Juma River which enters from the west, while draining eastward through Lwamunda Swamp from whence water seeps through the sand barrier into Lake Victoria.

During the early Holocene, Lake Nabugabo is thought to have been a protected bay of Lake Victoria, only forming an isolated lake as the level of Lake Victoria dropped and a longshore bar closed it off. The mid- to late-Holocene lake-status history is derived from a study by Stager et al. (2005), who employ sedimentological and diatom-assemblage composition of 2.7-m long sediment core NAB-1 from the center of the lake. The age model of NAB-1 was reconstructed using Bacon in R (Blaauw and Christen, 2011), using the original set of retained dates of Stager et al. (2005) and calibrating radiocarbon dates using the SHCal13NH calibration curve (Hogg et al., 2013). All radiocarbon dates were acquired on bulk organic matter without investigating the possibility of an old-carbon effect, which requires that Lake Nabugabo's chronology should be regarded with caution.

The lake-status history of Lake Nabugabo is based on organic and sand content of the sediment and follows the interpretation put forward in the original publication. Phases of high organic content and low sand content are interpreted as lowstands that likely considerably shrunk the area of the lake, and vice versa. The monotonous upper 1 m of the core is interpreted by Stager et al. (2005) as to have underwent disturbance and homogenization by wind-driven currents. Any information on short-term lake changes is therefore probably lost from this section, but the absence of sand layers or paleosols in this section is interpreted as evidence that the lake did not dry out during its corresponding period. Fossil diatom assemblages reported by Stager et al. (2005) suffer from variable preservation and are likely driven by the dynamics of silica and other nutrients in the lake, of which the relation to lake level or water depth is unclear.

Lake Chibwera

Lake Chibwera is a small freshwater maar crater lake in the Bunyaruguru lake district of southwestern Uganda, on the floor of the Edward-George branch of the Albertine Rift just south of Lake George (Bessems et al., 2008). It's water column has a maximum depth of 12.6 m and is seasonally stratified with near-bottom anoxia. The small catchment consists solely of the low crater rim. Hydrological balance is maintained against the negative local precipitation/evaporation balance by the shallow groundwater table around Lake George, and a few ephemeral streams.

Lake Chibwera's lake-status history was determined based on sediment core CHIB-02-1P (Bessems et al., 2008). The core consists of a dry clay layer at the base which was interpreted as a desiccation deposit, overlain by peaty sediments overlain by organic lake mud. The chronology of the core was predominantly determined by 210Pb- and 137Cs-dating, of which the results were only partly documented in the original paper. The time spanned by the core was deemed too short to pursue meaningful Bayesian age-depth modelling. Therefore, status-class period starts and endings were determined by ^{210}Pb dates bracketing them. The two reported radiocarbon dates flanking the desiccation horizon (sampled at 96-98 cm and 114-116 cm core depth in CHIB-02-1P) were recalibrated using the SHCal13 calibration curve. The upper radiocarbon date (170±30 ^{14}C years BP) was found to be compatible with the ^{210}Pb-determined end of the desiccation phase and re-establishment of sedimentation (147±42 cal yr BP). The lower radiocarbon date (185±30 ^{14}C years BP) was taken here to form the start of the lake-status record.

Sediments at the base of the core are characterized by dry low-organic clay, low magnetic susceptibility, fine sand and coarse plant detritus. These are interpreted as complete desiccation with possibility of seasonal flooding. Above this desiccation horizon, the core transitions into moist organic peaty clay of low magnetic susceptibility, with presence of fine sand and coarse plant detritus. This sediment type is interpreted as a shallow-water phase with abundant swamp vegetation and abundant macrophytes. The overlying soft organic clays without plant detritus are interpreted as deeper-water deposits with a dominant role of algae over macrophytes.

Lake Emakat

Lake Emakat occupies the floor of Empakaai Crater, a caldera which is part of the Crater Highlands in northern Tanzania (Ryner et al., 2008). The crater is ca. 6 km in diameter and is bounded by walls rising to 400-900 m above the caldera floor. Lake Emakat is ca. 80 m deep and hydrologically closed, resulting in a saline alkaline water column. The lake is fed solely by rainfall and small streams within the caldera.

The lake-status history of Lake Emakat was coded based on near-shore sediment cores E4 (ca. 15-9.5 ka BP; Ryner et al., 2007) and Emp 1 and Emp2 (ca. 1 ka BP to present; Ryner et al., 2008), recovered in 2000 AD at 10-20 m water depth. Muzuka et al. (2004) reported data on the δ13C and δ15N composition of bulk organic matter in two cores from Lake Emakat, but these data were not used in the assessment of lake status as no clear relation to lake level could be demonstrated. The age model of core E4 was recalculated using Bacon in R (Blaauw and Christen, 2011), retaining all reported 14C dates and calibrating them using the IntCal13SH calibration curve (Reimer et al., 2013). The period of ca. 14480-12700, although represented by part of the core, was not coded since the obtained data could not be interpreted unambiguously, according to the authors of the original study.

The dominant source of information available for cores Emp 1 and Emp2 is pollen (including Cyperaceae). The variations observed in the aquatic pollen record were deemed insufficient to code the period represented by cores Emp 1 and Emp 2 into separate status classes. Similarly, the radiocarbon dates collected from a presently inundated tree, suggesting tree growth and thus a lake low stand ca. 550-250 cal yr BP, were also insufficient to code the period covered by cores Emp 1 and Emp 2 as it does not exclude similar or even more severe low stands before or after tree growth. However, the great difference in Cyperaceae pollen between Emp 1 (100-500 % of pollen sum) and E4 (5-30 % of pollen sum), with both cores having been taken at the same location. This was deemed sufficient evidence to code lake level during the period spanned by core Emp 1 as consistently lower as during the period E4, with the Cyperaceae-dominated littoral zone closer to the core site. The start of the lake-phase represented by core Emp 1 was determined by calibrating the lowermost ^{14}C date of this core (920 ± 85 ^{14}C yr BP) using the IntCal13SH curve.

Lake Victoria

Lake Victoria has the largest surface area of all African lakes. It is situated in a shallow continental sag between the Albertine and Gregory branches of the East African Rift System and has a maximum depth of 68 m. Lake Victoria is an exorheic freshwater lake, overflowing towards the north along the Victoria Nile which further downstream flows through Lake Kyoga and into Lake Albert.

The Late-Pleistocene desiccation of Lake Victoria has been described by seismic profiling and radiocarbon dating of paleosol formations observed in numerous sediment cores taken at different location in the lake (e.g. Kendall, 1969; Johnson et al., 1996). A comprehensive review of available data on this event is presented by Stager et al. (2002). Radiocarbon dates from cores Ibis-1 and Ibis-3 (Damba Channel) and V95-2P and V96-7P (offshore), all used by Stager et al. (2002) to determine the timing of onset and termination of the late-Pleistocene desiccation event, were recalibrated using the IntCal13NH calibration curve (Reimer et al., 2013) and the 330-year reservoir effect (in case of bulk sediment samples) used in the original publication. Although Stager et al. (1986) reports on the period of ca. 25,000-18,000 yr based on core IB5 (offshore, close to Isamba Island), not enough information is available to unambiguously determine a status class for this period.

Determination of post-desiccation lake-status classes was based on the outcome of correspondence analysis of the diatom community assemblage from sediment core Ibis-1 (Damba Channel), presented by Stager et al. (1997; 2002; 2003), where the second axis CA2 was interpreted as reflecting lake level. The age model of this core was reconstructed using Bacon in R (Blaauw and Christen, 2011), using the age-depth data points reported in Stager et al. (1997 and 2002) and recalibrating radiocarbon dates using the IntCal13NH calibration curve (Reimer et al., 2013). These studies were supplemented by conclusions on the open or closed nature of the lake stated by Johnson et al. (2000) and Beuning et al. (2002) based on biogenic silica content and cellulose $\delta^{18}O$ in sediment cores V95-2P and V96-7P, of which age models were recalculated using Bacon in R (Blaauw and Christen, 2011), using the same age-depth data points as the original authors while recalibrating radiocarbon dates using the IntCal13NH calibration curve (Reimer et al., 2013).

In the studies mentioned above, the last millennium is regarded as the driest period since the onset of the Holocene (e.g. Stager et al., 2002). The decade- to century-scale lake-status changes of the last millennium are described in further detail in Stager et al. (2005b) using variability in the abundance of shallow-water diatoms throughout sediment core P2K-1 (Pilkington Bay). The age-depth model of this core was reconstructed using Bacon in R (Blaauw and Christen, 2011), retaining all reported data points and recalibrating radiocarbon dates using the IntCal13NH calibration curve (Reimer et al., 2013). Data of ^{210}Pb dating was not reported in the original publication and was therefore extracted from figures.

One date on a palaeoshoreline, 4 m above the present lake level, has been reported at 3720±120 ^{14}C years (Stuiver et al. 1960). Using the recalibration (IntCal13NH, Reimer et al. 2013), the elevation of this shoreline was connected to the corresponding lake-status period based on the sediment-core data.

Suguta Valley

The Suguta Valley forms the terminal basin of the Northern Kenya Rift Valley, located at the northern edge of the East African Plateau (Junginger et al., 2014). To the north and south, the basin is delimited by the Barrier and Korosi volcanic complexes, respectively, while its eastern and western boundaries are formed by an asymmetric graben structure. The valley is currently almost completely dry, only containing the small ephemeral Lake Logipi at the northern end. In the past, however, the valley has contained the large and by times overflowing palaeo-lake Suguta.

The lake-status history of Lake Suguta was coded mainly based on Junginger et al. (2014), supported by earlier studies by Truckle (1976) and Garcin et al. (2009). The main line of evidence is a series of radiocarbon-dated shorelines at different elevations throughout the catchment. Additionally, outcrops of lacustrine sediments were found that covered the period of ca. 8500-11500 cal yr BP. Stratigraphical evidence from these outcrops could be linked to individual shorelines, and allowed continuous coding during the period covered by the lacustrine sediments they contained. For the part of the record without corresponding outcropping lacustrine deposits, interpolation between dated palaeoshorelines followed the original publication as much as possible.

Garcin et al. (2009) showed that significant tectonic displacement has taken place in the Suguta Valley over the course of the Holocene. These authors observed an across-rift tilting pattern with an offset between the uplifted western flank and the subsiding eastern escarpment by as much as 25 m. Therefore, the shoreline elevations should be considered as merely indicative of lake status, and using these elevations for exact estimations of lake surface area and lake volume is complex. Lake size at times when the lake overflowed into the Kerio River and into Lake Turkana drainage basin has been estimated by using the average elevation of the maximum high stand shoreline (567 m asl; Garcin et al. 2009). An additional source of semi-quantitative lake-size information was the occurrence of *Melanoides tuberculata* fossils in littoral deposits represented in the outcrops. These aquatic snails are known to occur down to 15 m below water depth, which provides an upper and lower limit of water level at their recovery site.

Junginger et al. (2014) applied a reservoir effect on mollusc carbonates of 1900 years, which had not been recognized in earlier studies. Recalibration of the reported dates with the IntCal13NH calibration curve yielded identical results to those reported in the original publication. Not enough information on the compositional data was provided to use reconstructed age-depth models for the outcrops covering ca. 8500-11500 cal yr BP, so the original age-depth models (linear interpolation between dates) were retained. However, as the authors of the original study did not report uncertainty ranges for the outcrop age-depth models, we estimated the mean uncertainty-envelope width using Bacon in R (Blaauw and Christen, 2011) and applied this envelope to the age-depth model reported by Junginger et al. (2014). Mean 2-sigma uncertainty ranges were 320 years for section BG08, 270 years for section EL08 and 100 years for section LN08.

Lake Challa

Lake Challa is a relatively large (4.5 km²) and deep (94 m) crater lake on the border between Kenya and Tanzania. Its catchment is very small (5.6 km² including the lake surface itself), comprising only the steep-walled crater rim. Despite being located in an environment with highly negative precipitation/evaporation balance, Lake Challa's water balance is maintained by inputs from a shallow groundwater aquifer draining the flanks of Mount Kilimanjaro (of which the summit is located some 50 km to the northwest; Payne 1970). Lake water is alkaline and dilute, owing to underground outflow of water and dissolved substances. The water column is meromictic, as testified by finely laminated profundal sediments (Wolff et al., 2011).

The basis for the lake-status coding of Lake Challa covering the last 25,000 years was the seismic-stratigraphic lake-level record of Moernaut et al. (2010). Seismic-reflection data revealed variable patterns of basin infilling depending on lake level, with low lake levels resulting in focusing of sedimentation towards the central bottom plain versus high lake levels leading to uniform draping of sediment across the basin plain and the basin-plain periphery. Mass-wasting events were also interpreted as the result of significantly reduced lake level. Seismic findings were supported by stratigraphic and sedimentological patterns of a 22-m long sediment core spanning the last 25,000 years of Lake Challa's history. As the authors of the original study point out, the exact relative magnitude of the associated lake-level changes is uncertain, because draped sedimentation can occur over a wide range of lake depths.

The construction of the age-depth model covering the last ca. 25,000 years of Lake Challa's sedimentation history was described in detail by Blaauw et al. (2011, QSR). Radiocarbon dates from 168 bulk-organic samples, corrected for a variable old-carbon age offset which was calculated using wiggle matching or date pairing, were supplemented by a ^{210}Pb chronology for recent sediments. For this coding exercise, we recalibrated all originally used radiocarbon dates with the IntCal13SH calibration curve (Reimer et al. (2013) while retaining the estimations of the variable old-carbon offsets with depth. A new age model, incorporating recalibrated radiocarbon ages and ^{210}Pb-ages, was constructed using Bacon in R (Blaauw and Christen, 2011) with default settings except for a section thickness of 20 cm).

Seismic data is available for the sediments older than ca. 25,000 cal yr BP, but chronological information is lacking. Therefore, this period was not coded.

Nakuru-Elmenteita

Lake Nakuru and Lake Elmenteita today are two separate shallow hypersaline lakes at 1770 and 1786 m asl on the floor of the Nakuru-Elmenteita basin in the central Kenya Rift Valley. Since the lakes were joined into a single large body of water for a significant part of their history, they are regarded here as a single site.

Deriving of the lake-status history was mainly based on two cores collected from Lake Nakuru and Lake Elementeita in 1969-1970. The fossil diatom record from these cores, published by Richardson and Dussinger (1986), formed the main line of evidence, supported by the fossil ostracod record from the Lake Nakuru core of Cohen et al. (1983) which was in very good agreement with the diatom results. The age models of these cores were reconstructed with Bacon in R (Blaauw and Christen 2011), recalibrating radiocarbon ages with the IntCal13NH calibration curve (Reimer et al. 2013). Both core sites suffered from severe disruption or removal of sediments deposited after ca. 6000 cal yr BP, and therefore did not provide information on the period 6000-present.

Additionally, dated paleoshorelines and outcrops of lacustrine sediments described by Washbourn-Kamau (1971) and Dühnforth et al. (2006) provided quantitative lake-size information on the early-Holocene high stand of paleolake Nakuru-Elmenteita, to which their timing corresponds well.

Lake-status information is available for the period AD 1930-1980 from Lake Nakuru's water level gauge (Vareschi 1982), and additionally from two short ^{210}Pb-dated cores from Lake Nakuru (NAK01-1H) and Lake Elmenteita (ELE01-1H) going back from AD 2001 to ca. 1880 (De Cort et al., 2013). Both cores contain a layer of dry, low-organic mud at their base that was interpreted as a desiccation horizon of a severity that has not occurred again since. The originally published ^{210}Pb chronology of NAK01-1H was used to determine the timing of the end of this dry stand at 64 ± 8 (2σ) cal yr BP. The start of the dry phase responsible for the drying of these muds cannot be determined exactly from the available evidence. A duration of five years was conservatively assigned to this dry phase. Gauge data suggest that Lake Nakuru (nearly) fell dry again for much of the 1930's, although this episode did not result in a desiccation layer similar to the one observed for the late 19th century.

Relative correspondence between the statuses of the late Pleistocene-early Holocene and the 19th-20th century was determined through tentative integration of the 20th-century gauge data with the ecosystem reconstructions and synthesis figures published by Richardson and Dussinger (1986).

Lake Kitagata

Lake Kitagata is a small (<1 km²) maar crater lake part of the Katwe-Kikorongo cluster of phreatomagmatic explosion craters in the western arm of the East African Rift System in southwestern Uganda (Russell et al., 2007). The lake is about 9 m deep and permanently stratified. The catchment is small, consisting only of the steep-walled crater rim. While permanent surface inflow is lacking, subsurface inflow from the shallow groundwater table between Lakes Edward and George maintains the lake against the strongly negative local precipitation/evaporation balance. The lake water is hypersaline, dominated by Na^+, SO_4^{2-} and Cl^-.

The lake-status history of Lake Kitagata was determined primarily based on sediment-core stratigraphy described by Russell et al. (2007). The base of a ca. 180-cm long sediment core from the middle of the lake consisted of a crumbly, peaty clay topped by a layer of the sodium silicate mineral magadiite, interpreted as a precipitate during a phase of lake desiccation. Sediments above this desiccation horizon mostly consist of finely laminated clay, suggesting undisturbed deposition under a stratified water column. Two deposits of thenardite-rich clays or sands interrupt the fine laminations and were interpreted as periods of decreased water level.

The age model of the sediment core was recalculated using Bacon in R (Blaauw and Christen, 2011), with default settings except for a prior of acc.mean = 10 yr/cm. The ^{210}Pb-based chronology and radiocarbon ages of the original publication were used, rejecting the same single 14C age at 25.5 cm depth as the authors of the original study. Pb-210 ages and their uncertainties were not explicitly reported in the original paper and thus had to be extracted from figures. Radiocarbon ages were calibrated using the SHCal13 calibration curve (Hogg et al., 2013).

Lake Kibengo

Freshwater Lake Kibengo is currently hydrologically a small embayment of the much bigger Lake George, which itself flows into Lake Edward through the Kazinga Channel. Water depth at the sill separating Lake Kibengo from Lake George is currently 3 m.

Lake Kibengo's status history was coded primarily based on the stratigraphy of a 855-cm long sediment core described by Russell et al., 2007. The lowermost meter of this sequence consists of peat beds separated by muddy silts, interpreted as a severe low-water phase with sporadic desiccation of the basin. Higher up, the sediments mainly consist of massive clayey sapropels with high organic matter content, interpreted as deposited during phases of relatively high water level. These organic-rich clays are at certain intervals replaced by silty, calcite-rich sediments, signifying lower water levels and a more evaporative environment.

The age model of the sediment core was recalculated using Bacon in R (Blaauw and Christen, 2011). The 210Pb- and 137Cs-chronology and radiocarbon ages of the original publication were used, rejecting the single 14C age at 211 cm depth that was also omitted in the original publication. A 200-year reservoir effect was applied for bulk-organic samples as calculated by the original authors, and radiocarbon ages were calibrated using the SHCal13 calibration curve (Hogg et al., 2013). Pb-210 ages and their uncertainties were not reported in the original publication and had to be extracted from figures as depicted in the original publication.

Lake Hayq

Lake Hayq occupies a graben on the eastern margin of the north-central highlands in South Wollo, Ethiopia (Lamb et al., 2007). The lake is relatively deep (max. depth 89 m) and steeply shelving. Despite being topographically closed, lake Hayq is relatively dilute (conductivity 920 µS/cm; Lamb et al. 2007) which suggests groundwater outflow is important in the lake's hydrological budget. The lake is alkaline, with a pH of 9. The principal inflow into the lake comes via the Ankwarka River, which enters the lake from the southeast. Adjacent Lake Hardibo overflowed into Lake Hayq at some point in the past, as indicated by a now permanently dry paleochannel. A newly constructed irrigation channel allows occasional water flow between the two lakes.

The lake-level history of Lake Hayq during the last ca. 3500 years has been studied by Ghinassi et al. (2012; 2015), based on two stratigraphic sections (from the Ankarka River area to the southeast of the lake, and from the Uarababo region at the northern edge of the lake). This study used sequences of dated stromatolites (corrected for a carbon-reservoir effect based on the offset between charcoal and lacustrine gastropod shells from the same deposit), near-shore swamp deposits or fan-delta deposits at paleoshorelines alternated by (undated) erosion surfaces to reconstruct the history of lake high and low stands.

Where age limits of successive phases could not be reliably determined given the available evidence, best-estimate ages were used as in the tentative lake-level curve presented in Ghinassi et al. (2012). The onset of the present-day lake level was set at AD 1969 (i.e. -19 cal yr BP), based on historical documentation by Lamb et al. (2007), stating that by this time the island home to the Istifanos monastery, near the western margin of the lake, became connected to the mainland due to a lake-level fall. Besides climatic phenomena, lake level and other limnological aspects today are probably also influenced by anthropogenic activities in the catchment.

Another available study describes the isotopic signature of authigenic carbonate throughout a sediment core from Lake Hayq spanning the last ca. 2000 years (Lamb et al. 2007). This study was not used here for lake-status coding, since the isotopic signature is not solely influenced by lake level, but also by other factors, such as vegetation development in the catchment or variable input from neighbouring Lake Hardibo (Lamb et al. 2007).

Lake Kasenda

Lake Kasenda is a small crater lake in the Kasenda cluster of phreatomagmatic explosion craters in western Uganda. Its lake-status history of the last ca. 2000 years was coded based on descriptions of stratigraphy, sedimentology and isotopic composition of authigenic carbonate of a sediment core taken from the deepest, central part of the lake by Bessems (2007). In general, the argument of the original author is followed in that laminated sediments high in carbonate signify a deep, permanently stratified water column and that the lake receives significant amounts of inseeping water rich in dissolved bicarbonate because of these high water levels. During lower water levels, the lake becomes holomictic and loses its influx of dissolved bicarbonate, therefore the formation of laminations and authigenic carbonate is impeded. At the base of the sediment sequence, a desiccation horizon overlain by a thick peat layer provides evidence for a severe lowstand.

All radiocarbon ages reported by Bessems (2007) were used to reconstruct the age-depth model of the sediment core, including the single date that was omitted in the original study but of which the in- or exclusion was found to have no significant impact on the outcome of the model. While the author of the original study also performed ^{210}Pb and ^{137}Cs dating, these results were not fully reported. Therefore, we limited the incorporation of the ^{210}Pb and ^{137}Cs dating to the data points that were mentioned in the publication. An age-depth model was constructed using the Bacon software in R with all default settings, calibrating radiocarbon dates using the IntCal13 calibration curve (Reimer et al., 2013).

Additionally, pollen data and diatom-derived conductivity reconstructions spanning the last ca. 1200 years are available from another sediment core taken from this lake. The pollen data (Ssemmanda et al., 2005) do not hold any information on lake-level changes, so were not used in lake-status reconstruction. However, Bessems (2007) transferred two radiocarbon dates from this core to her own sequence, improving the chronological control. We also decided not to use the diatom-derived conductivity reconstructions (Ryves et al. 2011) in the lake-status determination, as the interpretations brought forward in this publication do not match the stratigraphical inferences of lake-level change. The diatom changes in this lake may have been driven by environmental factors other than conductivity, and a follow-up study (Mills and Ryves, 2012) demonstrated that the outcome of the conductivity reconstruction at Lake Kasenda is strongly influenced by which particular training set is used. Therefore, we deemed the diatom studies for Lake Kasenda as too ambiguous regarding the relationship with lake status.

Lake Nyamogusingiri

Lake Nyamogusingiri is part of the Bunyaruguru lake cluster in southwestern Uganda (Mills et al., 2014). The lake is located within the Maramagambo Central Forest Reserve. The lake appears to be an amalgamation of several craters, the majority of which form a broad, flat basin with a depth of 3.9 m. At its western edge however, and connected to the main basin via a sill 1.2 m below the water surface, lies a smaller but deeper crater within which Lake Nyamogusingiri attains its maximum depth (12.5 m). The lake is topographically closed with a conductivity is 554 µS/cm. Emergent dead trees offshore suggest relatively recent episodes of lower-than-present water level, during which the main basin possibly separated from the smaller but deeper western basin.

The lake-status history of Lake Nyamogusingiri was coded based on stratigraphical and diatom-community dated presented by Mills et al. (2014). Sediments that were described as peat-like were interpreted as signifying the lower lake level than at times when organic clay/mud was deposited. Further classification was based on the relative abundance of the diatoms *Nitzschia lancetulla* (interpreted as a deep-water planktonic species), *Cyclotella meneghiniana* (facultatively planktonic, interpreted as lower water levels) and *Amphora coffeaeformis* (a salt-tolerant taxon). As also stated in the original publication, lake-level reconstructions post 150 cal yr BP should be treated with great caution, as there is a clear anthropogenic overprint and influence on the ecology, diatom flora and possibly water level of Lake Nyamogusingiri during this time window.

The age-depth model of the sediment core was reconstructed using the ^{210}Pb and ^{137}Cs data reported in the original publication. Mills et al. (2014) also reported five radiocarbon dates, of which two (at 121.5 and 126.5 cm core depth) were rejected by the authors themselves. In addition to these two, we rejected a third (at 61.5 cm core depth) radiocarbon date given that it produced patterns in the age-depth relationship that were deemed unrealistic. Radiocarbon dates were calibrated using the SHCal13 calibration curve (Hogg et al., 2013). The new age-depth model was constructed using Bacon in R, with default settings except for a prior of acc.mean = 10 yr/cm.

Lake Kyasanduka

Lake Kyasanduka is a small maar crater lake part of the Bunyaruguru lake cluster in southwestern Uganda (Mills et al., 2014). The lake is located within the Maramagambo Central Forest Reserve. Its shallow (max. depth 2 m) water column is dilute (204 µS/cm).

The lake-status history of Lake Kyasanduka was coded based on stratigraphical and diatom-community dated presented by Mills et al. (2014). Sediments that were described as peat-like were interpreted as signifying the lower lake level than at times when organic clay/mud was deposited. Further classification was based on the relative abundance of the aerophilous diatom species *Lutica muticola*, the benthic diatom species *Nitzschia amphibia* and the planktonic diatom species *Aulacoseira ambigua* and *Aulacoseira granulata* (i.e. the 4 most abundant species in the coded part of the record).

Lake status was not coded for the last ca. 230 years (corresponding to the top 133 cm of the sediment core), as it is impossible to distinguish lake-level from anthropogenic impacts as drivers of changes in the diatom community in this time window. This recent anthropogenic overprint was also mentioned by the authors of the original publication.

The age-depth model of the sediment core was reconstructed using the ^{210}Pb and ^{137}Cs data reported in the original publication. Mills et al. (2014) also reported 11 radiocarbon dates, of which 3 were rejected by the authors themselves. We used the same 8 radiocarbon dates retained by the authors of the original study, also excising a slump deposit at 120-125 cm core depth. The new age-depth model was constructed using Bacon in R, with default settings except for a prior of acc.mean = 5 cm/yr. Radiocarbon dates were calibrated using the SHCal13 calibration curve (Hogg et al., 2013).

Lake Tana

Lake Tana is a relatively large (lake-surface area ca. 3150 km²) but shallow (z_{max} 14 m) oligo-mesotrophic freshwater lake on the basaltic plateau of northwest Ethiopia (Marshall et al., 2011). It is fed mainly by four permanent rivers, while its surface outflow is the start of the Blue Nile.

The lake-status history of Lake Tana was coded based on studies by Lamb et al. (2007) and Marshall et al. (2011), which combine high-resolution seismic surveying with data from sediment cores 03TL2, 03TL3 and 03TL4, obtained from the southern edge, deepest central part and northern edge of the lake, respectively. A biomarker study on core 03TL3, focusing on δD of plant leaf waxes (Costa et al., 2014), was not used for coding as this indicator is thought to reflect rainfall source and amount but is not directly related to lake depth or water level.

The base of core 03TL4 contains a desiccation horizon consisting of compacted sediments that were deposited during an earlier phase. This horizon is also visible on the seismics as a strong, high-amplitude reflector. The start time of the desiccation event was estimated based on a single radiocarbon date from the base of core 03TL4. All other status-period demarcations are based on the reworked age-depth model of sediment core 03TL3, which does not contain the desiccation horizon itself but does contain the sediment unit immediately above it.

All radiocarbon ages were recalibrated using the IntCal13NH calibration curve (Reimer et al. 2013). The age-depth model of core 03TL3 was reconstructed using 16 of the 17 radiocarbon dates reported by Marshall et al. (2011), supplemented by the core top which represents the time of core collection. The radiocarbon date at 640.5 cm core depth was omitted because its inclusion created unrealistic changes in the age-depth relationship at an unexpected core interval.

At ca. 15,270 cal yr BP (935 cm core depth in 03TL3), Lake Tana switches from a closed to an overflowing lake, as evidenced by the rapid disappearance of halophilic diatoms and a decrease to low Ca and Sr levels. Two small and short-lived increases in Ca and Sr at ca. 13280-12970 and 12600-12200 cal yr BP, corresponding to seismic reflectors, are interpreted as lowstands that potentially temporarily lowered the lake below its overflow level leading to evaporative concentration. Holocene seismic reflectors are interpreted as less severe lowstands during which the lake remained open, as no coincident increase in Ca and Sr (and thus evaporative concentration of the lake water) is observed. Apart from lowstands evidenced by seismic reflectors, no other lake-status changes could be defined for the Holocene. Despite XRF and magnetic data showing variability throughout the Holocene, these signals are thought to be driven by precipitation instead of lake level. Bulk sedimentology and stratigraphy suggest that Lake Tana's outflow was never significantly interrupted during the Holocene. Diatom preservation in the Holocene part of the core is reported as poor (Lamb et al. (2007), and no data is available to allow the identification of Holocene events or patterns other than those recognised in the seismic data.

Lake Abhé

Lake-status coding of Lake Abhé was based on the data presented in Gasse (1977) and Gasse and Street (1978), two articles synthesizing the results of the unpublished PhD thesis of F. Gasse (1975). The main line of evidence was a large collection of ^{14}C-dated littoral deposits sampled throughout the Lake Abhé basin. These littoral deposits are described as “excellently preserved”. They have been recognized in the form of thick calcitic, stromatolitic crusts on rocky escarpments; beaches existing of shell accumulations associated with flattened cobbles; or other geomorphic features. Unfortunately, details on the individual littoral deposits used for constraining lake level could not be deduced from the publications used for lake-status coding. Radiocarbon ages and their associated errors had to be extracted from Fig. 9 and 10 in Gasse and Street (1978).

The late-Pleistocene part of Lake Abhé’s history is further documented by a 50-m long sediment core collected a few kilometres southeast of the present lake, 5 m above the present water level. Stratigraphical, mineralogical, sedimentological and diatom data from this core was used in the lake-level coding of late-Pleistocene Lake Abhé. The Holocene section of this core was deemed unusable as a suitably representative archive of the Holocene, since i) the upper 6 m were lost during core collection, and ii) a hiatus in sedimentation between 7,250 and 16,400 ^{14}C yr BP, related to a drought that ended at the timing of re-establishment of sedimentation above this hiatus.

The chronology of late-Pleistocene lake-status change was based on a new age-depth model of the Lake Abhé core starting from 13.77 m core depth downward. The soil stratum between 13.77 and 8.90 m core depth was not included in the age model, since a very different sedimentation rate can be expected for this deposit as compared to underlying strata. Instead, the age of the upper limit of this stratum was estimated using the radiocarbon date reported for its upper boundary (16,400 ± 400 ^{14}C yr BP). For the underlying section of the core, where chronology was based on the new age-depth model, ages reported for corresponding littoral deposits were checked for compatibility with the age-depth model. In all cases, the littoral dates were compatible with the core-derived age model. The chronology of Holocene lake-status change was based solely on radiocarbon dates obtained from littoral deposits, as reported in Fig. 10 in Gasse and Street (1978).

While sediment-core data predating 30,990 ± 1575 ^{14}C yr BP is available, we have not incorporated it here since any form of chronological control older than this date is lacking.

Lake Abiyata (Ziway-Shala)

Lake Abiyata belongs to a chain of four residual lakes of decreasing altitude and increasing salinity that occupy the endorheic Ziway-Shala lake system: Ziway (1636 m a.s.l.), Langano (1585 m a.s.l.), Abiyata (1578 m a.s.l.), Shala (1558 m a.s.l.).

The main basis for coding the lake-status history of Lake Abiyata, and the entire Ziway-Shala system, was a 12.6-m long sediment core taken from Lake Abiyata under 7 m water in 1995 (core AB95II; Chalié and Gasse, 2002). Diatom community assemblage was the main data reported for this core. The core was dated using ^{14}C dating on bulk sediment, and while the authors report no old-carbon effect for these samples, a reservoir effect has been reported for more recent sediments in this lake (Legesse et al., 2002). The age-depth model of core AB95II was reconstructed using all dates reported in the original study (without applying an old-carbon effect), calibrating them with the IntCal13NH calibration curve (Reimer et al., 2013), and using Bacon with default settings and a section thickness of 20 cm.

Earlier lake-level reconstructions in the Ziway-Shala basin were based on radiocarbon-dated palaeoshorelines or other geomorphic features spread throughout the basin (Gasse and Street, 1978; Gillespie et al., 1983). These demonstrated high water levels in the early- to mid-Holocene, although exact lake-level elevations estimated from these were mostly tentative. Radiocarbon dates from these features reported by Gillespie et al. (1983) were recalibrated using the IntCal13NH calibration curve (Reimer et al., 2013) before assessing their compatibility with the lake-status history as inferred from sediment core AB95II. We found that all dates were compatible. The dates on the highest shorelines allowed to assign an absolute elevation to the highest status class.

High-resolution sedimentological and diatom data from a short core taken in 1998 (E98AB05-CV; 116 cm) provided information on lake-level changes in isolated Lake Abiyata during recent centuries (Legesse et al., 2002). Dating of this short core is problematic, as the ^{210}Pb dating is not supported by ^{137}Cs dating and likely underestimates true age uncertainty. Both reported bulk-OM ^{14}C dates are rejected by the authors of the original article, based on the suspicion of contribution of reworked material and of an old-carbon effect on lake water in the period covered by the core. It should be noted that the latter argument at least partly contradicts Chalié and Gasse (2002), who argue against an old-carbon effect affecting bulk-OM ages in Lake Abiyata. The lake-status history represented by core E98AB05-CV was divided into three main periods, following the original authors. Without clear correlation to the lake-status periods of the longer time scales of the late-Pleistocene and Holocene (see above), we assumed that the variability during the last centuries was of a lesser magnitude than the lower-resolution variability deduced from core AB95II. An age-depth model was made using the ^{210}Pb chronology, with the dates extracted from figures in the original article. Default settings in Bacon were adjusted to acc.shape = 0.5, acc.mean = 2 and mem.mean = 0.3. The ^{210}Pb chronology had to be extrapolated downward to cover the bottom 36 cm of the core.

Lake Duluti

Lake Duluti is a small freshwater crater lake situated south of Mount Meru at an elevation of 1260 m a.s.l. The catchment is very small, consisting only of the crater walls. The lake has no surface inlet or outlet, but freshwater conditions are maintained probably by subsurface seepage. Two known water-column measurements, taken in January and September 2008, suggest a permanently stratified water column with anoxic bottom waters.

In 2007, three short cores (LD1, 36 cm, taken at 4 m water depth; LD2, 36 cm, taken at 5 m water depth; LD3, 59 cm, taken at 9 m water depth close to the deepest point of the lake; Öberg et al. 2012) and one longer core (LD6, 191, taken at the same location as LD3; Öberg et al. 2013) were collected from Lake Duluti. The top 44 cm was missing from LD6, and no overlap could be established between LD3 and LD6.

The principal source of information on lake-status history from these cores is the fossil diatom community. Interpretation of species abundances is aided by four modern samples collected from plankton, nearshore sediment, stone scrapes and floating macrophytes (Öberg et al. 2012). The plankton sample was dominated by *Aulacoseira ambigua*, while *Nitzschia amphibia* was the most abundant species in the shallow-water and stone-scrape samples. The macrophyte samples dominated by *Gomphonema*, *Eunotia* and *Pinnularia*. Limited stratigraphical information is described for short cores LD1-3 but is not available for LD6, which is why stratigraphy was not taken into account for lake-status coding.

Chronological control of core LD6 consists of 19 AMS ^{14}C radiocarbon dates, all of which were retained (including two that were rejected in the original publication) using the IntCal13NH calibration curve and Bacon in R for the reconstruction of the age-depth model. Dating of cores LD1, LD2 and LD3 was for an important part achieved through ^{210}Pb dating. Unfortunately, uncertainties of ^{210}Pb dates were not reported in the original publication, which means ^{210}Pb dates could not be incorporated here. A Bayesian age-depth model (using Bacon in R) was reconstructed for core LD3, using the core top, the clear ^{137}Cs marker interpreted as AD 1963, and three ^{14}C dates (one of which was dated as modern, calibrating the other two using the IntCal13NH calibration curve). Diatom community composition of this core was used to code the recent lake-status history of Lake Duluti, since the core was taken at the same location as long core LD6 which was used to code pre-20th century lake-status history. While their diatom data shows similar patterns to LD3, short cores LD1 and LD2 were not considered as a primary source of information during lake-status coding, as they were collected close to the shore which makes their diatom record less easily comparable to the observations for long core LD6.

Lake Tilo

Lake Tilo is a small crater lake in the Ethiopian Rift Valley. Its lake-status history during the last ca. 10,000 cal yrs was coded based on the 23.30-m composite core combining cores Tilo95-1 and Tilo97-1, collected in 1995 and 1997.

The age model of this core was reconstructed using all originally reported radiocarbon dates (Telford and Lamb, 1999), calibrating the dates with the IntCal13NH calibration curve. Bacon in R was used, with default settings except for a section thickness of 25. Tephra deposits were removed before running the age model, but their positions had to be estimated from Fig. 4 in Telford and Lamb (1999) as their exact core depths were not provided by the authors of the original study.

The multiple sources of evidence available for this sediment core are often unsuitable for direct translation to lake level, but suggest that the lake has known a complex history with an important role for geothermal activity. During the early- to mid-Holocene, groundwater flux into the lake is thought to have been very high, contributing to high lake levels. The authors propose that the groundwater flux into Lake Tilo diminished abruptly at the time corresponding to ca. 800 cm core depth (ca. 6,000 cal yr BP according to our reworked age model), as indicated by an abrupt decrease in the accumulation rates of calcite and silica. The authors do not invoke climate-related causes for this event.

Isotope information from different sources (authigenic calcite, organic matter, diatom frustules) is available (Lamb et al. 2000; 2004; 2005) but was mostly not used to reconstruct lake level. However, $d^{13}C$ and $d^{18}O$ of authigenic calcite was invoked to differentiate between the earliest part of the record (weak correlation; interpreted as short water-residence times in an open basin) and the later part of the early Holocene (strong correlation; interpreted as reflecting evaporation in a closed basin).

Lake Sonachi

Lake Sonachi is a small and shallow saline alkaline crater lake in the semi-arid central Kenya Rift Valley. It is located at ca. 3 km west of Lake Naivasha, to which it is hydrologically connected through substantial subsurface inflow from Naivasha into Sonachi.

The Holocene history of Lake Sonachi is covered by 15.5-m long sediment core SI collected from the center of the lake (Damnati and Taieb 1996), which contains stratigraphical and sedimentological evidence for lake-level change. Three radiocarbon dates were obtained on bulk sediment from this core, of which the depths were not reported in Damnati and Taieb (1996) and had to be estimated from a figure. The possibility of an old-carbon effect was not investigated and can't be ruled out. These dates were recalibrated using the SHCal13 calibration curve (Hogg et al., 2013) and incorporated into a Bayesian age-depth model, including also the core top assumed to contain the intact sediment-water interface. This assumption is debatable, as core SI seems to be missing the features described for a short freeze core retrieved from a nearby location in the lake (see below; Verschuren, 1999). Bacon (Blaauw and Christen, 2011) in R was used for Bayesian age-depth modelling, with default settings.

The short sediment freeze core NS93.2-F (37 cm) presented by Verschuren (1999) provides high-resolution evidence of lake-level change during the 19th and 20th centuries. This core consists of organic unlaminated and finely laminated clays, with at the base a deposit of stiff low-organic sandy clay that is interpreted as a desiccation horizon. The start of the desiccation event represented by this base unit can't be determined exactly, so a minimum duration for this event was estimated conservatively at 10 years. Chronological information on NS93.2-F consists of the intact sediment surface corresponding to the date of recovery, combined with ^{210}Pb dating. Age-depth modelling using these chronological data points was undertaken with Bacon in R, adjusting the default settings of acc.shape to 0.5, acc.mean to 5, mem.mean to 0.1, mem.strength to 6 and employing a normal error distribution for the provided dates.

The likely relative lake-status relationships between facies of core NS93.2-F and older facies in core SI were estimated based on available sedimentological characteristics in both cores.

Lake Albert

Lake Albert is the northernmost lake of the Albertine branch of the East African Rift System and occupies a half-graben rift basin that originated in the Miocene (Hecky and Degens, 1973). The lake lies along the western edge of the rift margin with steeply sloping topography to the west of the lake, while to the east the landscape gently slopes upward towards Lake Victoria, which overflows into the Lake Albert basin. Lake Albert also receives significant inflow from the Rwenzori Mountains and Lake Edward to the southwest via the Semliki River. The outflow of Lake Albert is located at the northern end of the lake, where it forms the start of the White Nile River.

The lake-status history of Lake Albert incorporated here is based on data from core F, a 9.2-m sediment core collected in 1971 at a location where water depth was 46 m (Beuning et al. 1971). Overpenetration resulted in loss of the sediment-water interface. Radiocarbon dates from a second core (core G, 10.6 m long) from the same site were reported versus depth in core F, based on core-correlation using lithological features.

Age-depth modelling for core F was revised, excluding the part of the core above 84 cm core depth. Unlike in the original publication, all dates in the interval 84-907 cm core depth were incorporated, with the exception of date Gif-6697 which returned a date of > 25,000 ^{14}C yr BP. Material used for radiocarbon dating was of mixed origin, including bulk sediment for which an old-carbon effect is not discussed in the original publication and could therefore not be excluded. All radiocarbon dates were recalibrated using the IntCal13 calibration curve (Reimer et al., 2013). Due to marked changes in sedimentation rate in the lower part of the core, separate age-depth models were constructed for the sections 905-840 cm and 840-84 cm core depth. The former was constructed with Bacon's defaults, whereas for the latter section a 'boundary' was introduced at 660 cm. Values for acc.mean and mem.mean were set to 10 and 30, 0.5 and 0.5 above and below this boundary, respectively.

Although no lithological changes are reported in the upper 84 cm of the core, dates at 84 and 35-39 cm core depth are separated by ca. 5000 14C yr and therefore suggest the presence of a stratigraphic hiatus, most probably caused by prevention of sedimentation or erosion of previously deposited sediments. The cause of this event could be low water level, although there is no evidence of exposure or of a distinct lag deposit or mineralized surface. There is, however, a relative concentration of coarser siliciclastic grains and fish debris around 70-80 cm core depth. Increased wind strength, resulting in a lowering of the wave base, could be an alternative explanation for this stratigraphical unconformity. The time window during which this non-deposition or erosion took place, caused by low lake level or another mechanism, could not be accurately dated and is therefore not incorporated in the consensus lake-status history of Lake Albert. Timing of the upper 39 cm of the core, for which lake status is incorporated, is derived from direct calibration of date Gif-6694 at 35-39 cm core depth which provides a minimum age for the resumption of sedimentation.

Lake Turkana

Lake Turkana is a large tectonic endorheic lake in arid northern Kenya and southern Ethiopia. The largest inflow of water comes from the Omo River, which drains the Ethiopian Highlands and is complemented by the Kerio and Turkwell rivers draining the Kenyan Highlands. Presently, the only significant loss of water from Lake Turkana is through evaporation. Multiple studies have described and dated geomorphic shoreline features attesting to higher-than-present lake levels during the late Pleistocene and Holocene. These were synthesized most recently in Bloszies et al. (2015). Additionally, continuous sediment-core data is available from three sediment cores collected in the southern basin of the lake, complemented by high-resolution seismic survey data (Morrissey and Scholz, 2014).

For the coding of Lake Turkana's lake-status history, stratigraphic and sedimentological (TOC) data from the sediment cores together with seismics were used to define lake-status periods. These were then assigned relative and quantitative statuses based on the elevations of corresponding dated shorelines presented in Bloszies et al. (2015). This means the shoreline dates were not directly used to construct the age model but were visually matched to corresponding episodes recognized in the sediment record. Elevations were binned into statuses that represent 10-m increments in lake-level elevation. Minimum and maximum estimates of status were based on minimum/maximum shoreline elevation estimates in Bloszies et al. (2015). Lake-level data for the late-19th and 20th century was taken from Fig. 10 in Bloszies & Forman (2015) and is based on sporadic historical observations and instrumental measurements. During this period, statuses represent 2.5- to 5-m increments of lake-level elevation.

The age-depth models of cores 4P, 14P and 46P were reconstructed using Bacon with default settings, recalibrating radiocarbon ages using the IntCal13NH calibration curve (Reimer et al. 2013). The old-carbon effect in Lake Turkana is considered as small to negligible in most studies (e.g., Morrissey and Scholtz, 2014) and here was set at 0 ± 50 years.

Magadi-Natron

Lakes Magadi (600 m a.s.l.) and Natron (608 m a.s.l.) are at present two separated shallow, seasonally dry salt pans on the floor of the endorheic Magadi-Natron basin at the southern end of the Kenya Rift Valley. They are treated here as one since the available evidence suggests that, for significant time windows in the past, they were united into one body of water.

Evidence for past changes in lake level is present in the form of a near-continuous, well-developed band of stromatolites above the present lake level. Elevations reported for these stromatolites range from 645 to 656 m above sea level (Hillaire-Marcel et al., 1986). While rare occurrences at higher elevation are ascribed to post-depositional uplift in one particular part of the basin, tectonic displacement is thought to have been of negligible influence in the majority of the basin. Dating of the stromatolites revealed deposition as three distinct generations, separated by long periods without stromatolite formation. The first and second generations were dated with U/Th, with the former being out of reach of the dating method. For the second generation of stromatolite growth, only two of the obtained OSL dates were deemed reliable and were used to bracket the implied high-water episode. The third (youngest) generation was dated by radiocarbon dating. Using isotopic characteristics of the stromatolite carbonate, the authors claim an equilibrium of lake-water TIC with the atmosphere and thus an absence of a large carbon-reservoir effect. All radiocarbon ages obtained from stromatolites were recalibrated using the SHCal13 calibration curve (Hogg et al., 2013). Additional dates were obtained from travertine pipes thought to signify increased groundwater flow near the paleoshoreline during times of high lake level (Hillaire-Marcel et al., 1986). However, these were described as suffering from an unquantified old-carbon effects from geothermal CO_2 and were therefore not used to reconstruct lake-level history.

Additional lake-status evidence is available from sediment core NF, retrieved from the ‘Flamingo Nursery’ bay in the north-western part of Lake Magadi (Taieb et al., 1991). The core spanned a total length of 8.7 m, and consisted of three main stratigraphical units. There is evidence for at least three episodes of discontinuous deposition, possibly with erosion of previously deposited sediments, separating the three main units and at the core top. Several radiocarbon dates were obtained from the uppermost unit of core NF. However, radiocarbon dates from the lower two units were found to be unreliable, and chronological evidence is only present as sparse U/Th dates with very large uncertainties (1 dated interval in each of the lower two units). This sparseness, in combination with the depositional hiatuses demarcating the three units, precludes a single age-depth model for core NF. Lake-status information has only been incorporated for the uppermost unit, for which enough chronological data points are available to allow the meaningful reconstruction of an age-depth model. Reported AMS dates were used for age-model reconstruction, after calibration using the SHCal13 calibration curve (Hogg et al., 2013) and using Bacon with a prior accumulation rate of 10 yr/cm (Blaauw and Christen, 2011).

Lake-status coding of the upper unit of core NF was based on the interpretation and lake-level reconstruction of Roberts et al. (1993), which is based on a combination of fossil diatoms, organic-matter content and magnetic parameters (the latter of which are interpreted to act as an ‘on-off’ switch to record the connection between Lakes Magadi and Natron by inundation of the sill presently at 35 m above lake level). The stromatolite ages, thought to represent maximum lake-level elevation, where then compared to the core-based lake-level history. These were found to be compatible up to the youngest point represented by the sediment core. Furthermore, the stromatolite dates were able to extend the known duration of the high-water phase beyond the hiatus at the top of the sediment core.

The present-day status of the lake, i.e. a seasonally flooded salt pan, was assigned the minimum status 1 and has been present at the lake since at least 50 cal yr BP (when commercial trona mining at Lake Magadi began; Hughes, 2008) but very likely much earlier.

Lake Edward

Lake Edward is a large rift lake on the border between Uganda and the Democratic Republic of the Congo, just south of the equator in the western arm of the East-African Rift System. Much of the lake's paleoenvironmental history is known from a set of four overlapping sediment cores collected in 1996 by the International Decade for East African Lakes (IDEAL) team (Laerdal et al., 2002; Russell et al., 2003). Of these four, cores E96-2P and E96-5M are used here to code lake-level history since these are the most favourably placed in the basin to record lake-level changes, represent the longest time span, and have the best chronological control. All available dates for these cores are radiocarbon-derived, and separate age-depth models for these two cores were reconstructed using Bacon in R (Blaauw and Christen, 2011) with default settings, calibrating the dates with the SHCal13 calibration curve (Hogg et al., 2013). As in the original publication, hiatuses were inferred at 135 cm (E96-2P) and at 193 cm (E96-5M). Also as in Russell et al. (2003), two dates that were considered outliers in the original publication (20,600 ± 200 ^{14}C yr in E96-2P and 1905 ± 40 ^{14}C yr in E96-5M) were omitted.

Coding is mainly based on stratigraphy and follows the interpretation of Russell et al. (2003). For interpretation of the severity and timing of the 4-ka lowstand, the conclusions of Laerdal et al. (2002) were rejected in favour of the interpretation of Russell et al. (2003), the latter of which was an update and revision of previous findings by the same research team that produced the former publication. Additional evidence comes from sedimentological and mineralogical parameters described in Russell et al. (2003), such as organic carbon content, carbonate content, biogenic silica content and % Mg in calcite. The latter is interpreted by the authors as the most reliable representative of P-E, with % Mg increasing as conditions become drier. We have chosen not to directly translate %Mg in calcite into lake status, although an argument could be made for this approach. In a similar way, calcite content is also interpreted to reflect P-E through the promotion of calcite precipitation by evaporative concentration of the lake water. Biogenic silica content is thought to be driven mainly by the dissolved Si reservoir of the water column, which in turn depends on river input of dissolved Si and Si dissolution and burial rate (Russell and Johnson, 2005).

The early Holocene high-water phase, inferred from the sediment-core stratigraphy, might correspond to paleoshoreline features ca. 14.5 m above the present lake-water surface along the northern and eastern shores of Lake Edward and the Kazinga Channel, as well as alluvial terraces roughly 14 m above the present outflow of the Semliki River (Musisi, 1991; Brooks and Smith, 1987; Bishop, 1969; de Heinzelin and Verniers, 1996). However, these geomorphological features have not been dated and thus could not be formally correlated to the sediment-core reconstruction.

<u>Lake Hora</u>

Lake Hora is a small lake on the shoulder of the South Eastern Plateau of the Ethiopian Rift, between the cities of Harar and Dire Dawa.

Lake Hora's lake-status history was translated into the data base based on a single radiocarbon date from a paleoshoreline 1 m above 'present' lake level, published by Williams et al. (1977). The date was obtained from shells, published uncalibrated, and recalibrated here using the NHCal13 calibration curve (Reimer et al., 2013). The existence of an old-carbon reservoir effect has not been investigated and therefore cannot be ruled out. The duration of this high-water event during which the shoreline was formed, was estimated conservatively at 50 years.

Lake Besaka

Lake Besaka is a graben lake situated at the junction of the Afar Rift and the Ethiopian Rift. Lake-status coding is based on Williams et al. (1977) and Williams et al. (1981), who describe a significant influence of groundwater inflow on the levels of Lake Besaka based on 20th-century observations. Also, several volcanic eruptions and lava flows have taken place in the catchment of Lake Besaka during the period under study here, some of which very close to the lake. It is very likely that these have played a role in reshaping the basin currently occupied by the lake. The last major such lava flow occurred in AD 1820 (Williams et al., 1981).

Lake Besaka's original lake-level reconstruction is based on two stratigraphic transects between the western shoreline and the eastern fault-scarp. Further stratigraphic interpretation of the original study was published by Williams et al. (1981). Radiocarbon dates, mostly obtained on shells from *Melanoides tuberculata* and originally reported uncalibrated, were calibrated using the NHCal13 calibration curve (Reimer et al., 2013). The existence of an old-carbon reservoir effect was not investigated but cannot be ruled out. Two dates labelled SUA-459, dated to 4490 ± 115 and 5010 ± 110 ^{14}C yr BP, were not used in coding as they were obtained on calcium carbonate nodules that are thought to have precipitated post-depositionally in older sediments (Williams et al., 1977; Williams et al., 1981).

Williams et al. (1981) also discuss a possible pre-Holocene history of Lake Besaka's water level, with freshwater conditions preceding a terminal-Pleistocene regression. However, this period could not be incorporated in the data base as no suitable dates have been reported for it.

Loboi Swamp

Loboi Swamp is a wetland at the southern end of the Loboi Plain in the Baringo-Bogoria half-graben of the central Kenya Rift Valley. Loboi Swamp lies in a narrow, westward tilted, graben-like structure adjacent to one of the fault blocks of the Bogoria Plateau, a densely grid-faulted lava field (< 1 Ma) directly south of the Loboi Plain. Multiple springs feed the swamp, while the Loboi River drains it flowing northward into Lake Baringo. Loboi Swamp is characterized by dense stands of cattail (*Typha domingensis*) which form a wide belt encircling a zone of papyrus (*Cyperus papyrus*).

The history of this wetland was studies by Driese et al. (2004) and Ashley et al. (2004) using a soil pit dug at the edge of the swamp, in combination with two sediment cores (core 1 and core 7) from two locations deeper in. The soil pit and sediment core 1 demonstrate a clear, abrupt transition from floodplain deposits to sediments deposited in a wetland environment, with the associated appearance of diatoms and a shift in organic-matter stable-isotope composition suggesting the instalment of emergent wetland vegetation. Two radiocarbon dates are reported, of which one (obtained from sediment core 7, at a core depth of 309 cm) resulted in a modern age and was not used in the original study. To estimate the age of the wetland establishment, at 110 cm core depth in sediment core 1, an age-depth model for core 1 was made based on two ages; the core top and the ^{14}C age obtained for the 93-95 cm depth interval. The date was calibrated using the NHCal13 calibration curve (Reimer et al., 2013), and Bacon (Blaauw and Christen, 2011) was run using a prior accumulation rate of 1 mm/yr. The age-depth model was not extrapolated into the underlying floodplain sediments, as significant but undocumented changes in accumulation rate are expected there. We do not have a realistic idea of the start time of the period represented by these floodplain deposits. Therefore, the starting time window of Loboi Swamp, which has been allocated a status-class of 1, has been estimated at 100 years preceding the transition to wetland conditions, implying the floodplain-phase of (at least) one century.

Driese et al. (2004) and Ashey et al. (2004) argue that the cause of the abrupt water-level rise (i.e. wetland establishment) might not be exclusively climate-related. Tectonic subsidence might also have played a role, diverting water flow and providing space for water accumulation.

Lake Rukwa

Lake Rukwa is a large, shallow (ca. 13.5 m in 1996), closed-basin tectonic lake southeast of Lake Tanganyika and northwest of Lake Malawi.

The water-level history of Lake Rukwa, as synthesised here, is based on several publications. A first, 23.1-m long core was taken from the southern basin of the lake in 1960, and described in Talbot and Livingstone (1989). An age-depth model was made for this core, recalibrating reported radiocarbon dates using the SHCal13 calibration curve (Reimer et al. 2013). This was done to get age estimates of three episodes of lake desiccation, which are represented as dry, crumbly layers interrupting the otherwise dark, wet muds. These exposure surfaces are interpreted here as desiccation horizons that may have eroded an unknown amount of previously deposited sediments. Therefore, it is impossible to accurately assign a timing to the start of these lowstand episodes. They have been arbitrarily assigned a 'best-estimate' length of 100 years, loosely based on the fact that shorter episodes of lake desiccation, known from historical records from the last 200 years, have not resulted in similar stratigraphical features. The diatom composition of this core has been studied by Haberyan (1987).

With exception of these three desiccation events, the pre-historic part of Lake Rukwa's lake-level history has been synthesised based on a 12.8-m sediment core collected in 1996 from the deepest part of the southern basin of Lake Rukwa. The sedimentological aspects of this core have been described in Thevenon et al. (2002), while diatom analyses are reported in Barker et al. (2002). While the environmental reconstructions presented in these two papers are very similar, the diatom-based reconstruction of lake level presented by Barker et al. (2002) provides slightly more detail and is well founded. Their lake-level interpretation of diatom composition is followed here. The age-depth model of this sediment core was reconstructed, calibrating radiocarbon dates using the SHCal13 calibration curve (Reimer et al. 2013) and using Bacon with a section thickness of 15 cm and a memory strength of 0.3. Additionally, the two tephra layers at 60-80 and 792-904 cm depth were considered as event deposits, and boundaries were introduced at 510 and 990 cm to allow for differences in sediment accumulation rate between the main lithological units. Core depths were not consistently reported in the original paper and thus had to be approximated using figures in Thevenon et al. (2002) and Barker et al. (2002).

Maximum lake depth, inferred from sedimentological and diatom-community composition, occurred during the late Pleistocene to early Holocene. The magnitude of this deep-water phase is demonstrated by paleoshorelines ca. 200 m above the present lake, which have been dated to 9615 ± 95 (Delvaux et al. 1998) and 9740 ± 140 ^{14}C yr BP (Clark et al. 1970). The elevation suggests that at that time Lake Rukwa overflowed into Lake Tanganyika via the Ilyandi watershed, to the northwest of the current lake.

Additionally, Nicholson (1999) presents a reconstruction of the water level of Lake Rukwa since the mid-1800's based on historical documentation by early European travellers. Nicholson (1999) also suggests that the lake was completely dry during the late-18th to early-19th century AD. However, this hypothesis is based on indirect evidence from Lake Malawi, and as such is not incorporated in the lake-status synthesis presented here.

Sacred Lake

Sacred Lake is a small (0.51 km² in 1989), shallow (max. 5 m in 1989) lake in an explosion crater ca. 1 km across on the eastern flank of Mount Kenya. The lake is surrounded by a belt of floating swamp. It is topographically closed but is maintained dilute through outward underground seepage.

The lake-level coding included here is based on 1634-cm sediment core SL1, collected in 1989 (Olago et al., 1999; Olago et al., 2000; Olago et al., 2001). Earlier material was used for pollen analysis and vegetation reconstruction but contained no information in lake-level variation (Coetzee, 1964; Coetzee, 1967). Most climate proxies derived from sedimentological and biogeochemical analyses on SL1 (Huang et al., 1999; Olago et al., 1999; Olago et al., 2000; Loomis et al., 2012; Konecky et al., 2014) were not directly indicative of water-level change and thus could not be incorporated here. Lake-status history as derived here is based on stratigraphic features, with three occurrences of root mats interpreted by the authors of the original study (Huang et al., 1999; Olago et al., 2000) as indicative of a (nearly) desiccated lake bed interrupting the deposition of lake mud or waterlily peat. The latter only occurs in the top 3.5 m of the core, and might therefore be the product of gradual lake infilling rather than an actual decrease of water-level.

The age-depth model of core SL1 was reconstructed using the same set of radiocarbon dates as Loomis et al. (2012), i.e. 17 dates originally reported by Olago et al. (2001) supplemented with 11 dates originally reported in Loomis et al. (2012). U/Th dates reported by Olago et al. (2001) are not incorporated, following the approach of Loomis et al. (2012). Radiocarbon dates were recalibrated using the NHCal13 calibration curve (Reimer et al., 2013). The age-depth model was constructed with Bacon (Blaauw and Christen, 2011), using all default settings. The lake-level history as incorporated here does not extend beyond a depth of 1230 cm, which is the upper boundary of a tephra deposit sitting directly above a layer of glacial diamict. The ca. 3 m of the core below this glacial diamict is not radiocarbon-dated and would require unwarranted extrapolation of the age-depth relationship.

Lake Baringo

Lake Baringo is a shallow freshwater lake in the semi-arid volcanic region of the central Kenya Rift Valley. The modern lake is the successor to a series of precursor lakes that have occupied the rift valley at this latitude since the middle Miocene. Lake Baringo lies in the northern part of a rhomb-shaped half-graben basin, ca. 21x13 km. The lake is separated from saline, alkaline Lake Bogoria 16 km to the south by the Pleistocene and Holocene fluviatile and lacustrine sediments of the Loboi Plain.

The lake sits on the dry rift floor but drains wetter volcanic uplands, mainly to the southwest. Most annual recharge comes from the Molo and Perkerra rivers which discharge at the Molo Delta at the southern end of the lake. Freshwater conditions are maintained, despite topographical closure, through underground outflow northward. The lake surface in the 1980's to early 2000's was at ca. 970 m asl, with lake level fluctuating over a vertical range of 4 m during that period (Tiercelin et al., 1987; Renaut et al., 2000).

The only source of information on late-Pleistocene and early-Holocene lake-level history comes from a number of radiocarbon dates on shoreline features of the Kokwob formation west of the lake (Williams and Johnson, 1976). These dates were recalibrated using the NHCal13 calibration curve (Reimer et al., 2013). Since they seem to be centred on three different elevations, three distinct lake-status classes were inferred from them. The dates on the shorelines at 987 and 980 m a.s.l. were considered close enough together to be considered as a continuous sequence of lake-status periods, whereas a gap of unknown status was allowed between the periods represented by the shorelines at 980 and 975 m a.s.l.

Lake-status information concerning the last two to three centuries is derived from mid-lake sediment core Baringo03-2P (Bessems et al., (2008). The base of this core consists of a desiccation horizon, the product of a drought event during which the lake stood dry. The age of refilling, and uncertainty thereof, was derived from the reconstructed age model based on the date of core collection, two reported ^{210}Pb dates, and a single ^{14}C date on terrestrial plant material from the base of the core. Bacon in R was used (Blaauw and Christen, 2011), with default settings except for a prior of a mean accumulation rate of 1 yr/cm. The start of this desiccation event is unknown and cannot be derived from the available data. Here it was estimated conservatively at 10 years before the re-establishment of lacustrine sediment deposition.

Additional lake-level information for the 20th century is derived from Tiercelin et al. (1987), who present a lake-depth curve based on historical observations and, later, gauge measurements, spanning the period of 1940-1980. Additionally, the recent severe flooding event of 2013-2015, when the water level reached the 980 m altitude line (Obando et al., 2016), is also incorporated.

Lake George

Lake George is a shallow (average depth 2.4 m; Greenwood, 1976) eutrophic freshwater lake on the equator in southwest Uganda. It is fed by several rivers draining the eastern slopes of the Rwenzori Mountains and the eastern escarpment of the Albertine Rift. Discharge in these rivers is relatively high year-round, even during the driest times of the year (Laerdal et al., 2003). In turn, Lake George drains via the Kazinga Channel in the west into the larger Lake Edward. There is no sill at the outflow of Lake George. Greenwood (1976) reports a remarkable stability of the water level throughout the year, unlike other shallow East-African lake systems.

Evidence for the lake-level history of Lake George comes from Viner (1977), who reports data on five cores taken between 1967 and 1972 from the flat lake bottom. These cores have similar a stratigraphy, i.e. organic-rich lake muds overlying a desiccation horizon consisting of dry clay and sand. Viner (1977) also reports four radiocarbon dates on one of these cores (core 4), which were determined on bulk sediment and are stratigraphically consistent. These dates were recalibrated here using the NHCal14 calibration curve, and used to build an age-depth model with Bacon in R (Blaauw and Christen, 2011) with all default settings except for a prior accumulation rate of 10 yr/cm. This allowed the estimation of the age of re-onset of sedimentation, i.e. the timing of the termination of the desiccation event. Since it was impossible to estimate the start of this event (due to lack of underlying chronological information and the unknown amount of inflation during the dry stand), its duration was arbitrarily set at 100 years. Although the lake probably has known more subtle changes in water level since then, these could not be deduced from the available data on the soft organic mud.

Lake Tanganyika

Lake Tanganyika is one of East-Africa's largest (32,600 km²) lakes, and the world's second deepest (1,470 m; Gasse et al. 1989). It occupies the Albertine Rift over a length of 650 km and includes parts of 4 countries – Tanzania, the Democratic Republic of the Congo, Burundi and Zambia. The water column is permanently stratified because of its great depth. The lake is currently hydrologically open, with surface outflow by the Lukula River currently forming the most important control on maximum water level.

Lake Tanganyika has been the subject of extensive paleolimnological research. Consensus lake-status history here was compiled based on several studies, archives and proxy types. The earliest information that was used comes from core T97_52V (Scholz et al. 2003). This 10.5-m sediment core was collected from the Kavala Island Ridge (KIR), where depth is more modest (to within 40 m from the lake surface at the ridge crest) and sedimentation rate is an order of magnitude lower than in most other locations in Lake Tanganyika. Geochronology was based on radiocarbon-dating of bulk sediment. Three of these samples were also used for pollen extraction, dating of which suggested no significant reservoir effect. The age-depth model of core T97-52V was reconstructed using Bacon in R, retaining all originally used dates. The greatest core depth for which a date was obtained, was 766.5 cm (44500 ± 2200 ^{14}C yr). The age-depth relationship was linearly extrapolated below this depth down to the base of the core. The base of the core captures a seismic unconformity, which likely corresponds to severe lowstand or desiccation at the core site.

Lake-status history during more recent parts of the late Pleistocene and Holocene was informed by the stratigraphy of core NP04-KH3 (Felton et al. 2007). The age-depth model of this core was reworked using Bacon in R, retaining all originally used dates and recalibrating radiocarbon dates using the IntCal13SH calibration curve (Reimer et al. 2013). The variable old-carbon reservoir effect reported by the original study, restricted to sediments younger than 14,000 ^{14}C years, was applied.

The late-Holocene portion of Lake Tanganyika's lake-status history was additionally informed by ostracod community changes throughout core LT-97-56V (Alin and Cohen 2003), collected in 56 m water depth along the central eastern shore, and by diatom community changes throughout core LT03-05 (Stager et al. 2009), a nearshore sediment core obtained from the north basin. Age-depth models of both cores were reconstructed using Bacon in R. Age control of core LT-97-56V consists of radiocarbon dates obtained on terrestrial plant fragments to avoid the complex reservoir problem. All reported ages were retained and were calibrated using the IntCal13 calibration curve (Reimer et al. 2013). The core top, described as intact and thus corresponding to the time of collection, was also included into the model. The age-depth model of core LT03-05 was reconstructed in a similar fashion. Age control for this core consists of a combination of bulk sediment and terrestrial botanical radiocarbon dates. These were intercompared to determine a variable reservoir effect, which we also applied in our reconstruction. Our reconstruction also followed the original study of Stager et al. (2009) in omitting the same three outlying dates from the model. Radiocarbon dates were calibrated using the IntCal13SH calibration curve (Reimer et al. 2013).

Finally, lake-status information from historical sources contributed to the last two centuries of our consensus history of Lake Tanganyika. This information was obtained from Nicholson 1999, whose study is based on reports of European visitors, settlers and explorers and from oral accounts of peoples inhabiting the area.

Lake Manyara

Lake Manyara is a shallow, hypersaline and alkaline endorheic lake in northern Tanzania. It lies in an asymmetrical half-graben in the Gregory Rift. Maximum depth is reported at 4 m by Casanova and Hillaire-Marcel (1992b), but water level has been variable throughout its known history (Greenway and Vesley-Fitzgerald 1969).

The main line of evidence used in deducing the lake-level history of Lake Manyara is provided by stromatolite assemblages ca. 20 m above the present-day water level (Casanova and Hillaire-Marcel 1992b). These stromatolites are interpreted to have formed in shallow water, but do not constrain the exact elevation of the paleolake surface at the time of their formation. Carbonate from these stromatolites has been radiocarbon- and U/Th-dated. Radiocarbon dates, not calibrated in the original publication, were calibrated using the SHCal13 calibration curve (Hogg et al., 2013). Radiocarbon dates of which samples yielded ^{14}C activities < 2% were rejected by the authors of the original study (Casanova and Hillaire-Marcel 1992b) and were not used here. Except for one sample where the obtained result confirms the radiocarbon results, U/Th dating results proved problematic and are not used here. The retained radiocarbon dates seem to indicate two distinct generations of stromatolite growth, of which the most recent is characterised by a single date. The duration of growth of this recent generation was conservatively estimated at 100 years.

Further paleolimnological evidence from Lake Manyara consists of mainly diatom community assemblages throughout two sediment cores from the lake. Diatom data from a 56-m core and from a 14-m core (MANE-87) have been presented by Holdship (1976), and by Barker (1992) and Barker and Gasse (2003), respectively. However, the dating of these cores is extremely problematic. Additionally, Barker (1992) argues that differential frustule dissolution has most likely played an important role in determining the fossil diatom assemblages throughout Lake Manyara's sediment record, and significantly alters ecological interpretations and the result of hydrochemical transfer functions. These two factors have led us to exclude the diatom information from Lake Manyara's entry into the data base.

Lake Masoko

Lake Masoko occupies a maar crater in the Rungwe volcanic highlands in the western branch of the East African Rift System, 35 km north of Lake Malawi. The crater has steep slopes and a flat bottom, with a high lake/catchment surface area. Maximum depth of the lake is ca. 38 m, with lake level fluctuating seasonally by ca. 1-2 m (Garcin et al., 2007a). Lake Masoko is oligotrophic and stratified, with a thermocline located at ca. 11-16 m water depth (Delalande et al. 2005).

Paleolimnological information from Lake Masoko is available from a large number of sediment cores from various lengths taken between 1996 and 2003. Here, the attention is focused on 52-cm core MKJ-03-II (spanning the last ca. 500 years; Garcin et al. 2007a) and ca. 30-m composite core MM8/M96 (providing the longest continuous record available, spanning the last ca. 40,000 years; Gibert et al. 2002). Chronology of the former is mainly based on 210Pb dating, complemented with tephrostratigraphy and one radiocarbon date. The age-depth model of this core was reconstructed using the Bayesian approach of Bacon (Blaauw and Christen, 2011), calibrating the single radiocarbon date using the SHCal13 calibration curve (Hogg et al., 2013). Unfortunately, exact data of the 210Pb ages was not reported in the original publication. Depths and midpoint ages had to be derived from Fig. 3c in Garcin et al. (2007a), whereas the errors on the 210Pb ages, on which information was lacking altogether, were set to ± 10 years. The age-depth model of sediment sequence MM8/M96 could not be reliably reconstructed as the exact depths of event deposits (tephras and turbidites), which made up a significant share of the total core length, were not reported. The original age-depth model (Gibert et al. 2002; Garcin et al. 2006a), which was established through linear interpolation and did not report chronological uncertainty ranges, was therefore used to derive the best-estimate start and end times for the defined status periods. Since chronological uncertainty ranges were not reported for the originally published age model, a reconstruction of the age-depth relationship was made using Bacon in R, using all originally published radiocarbon dates (calibrated using the IntCal13SH calibration curve) and default software settings except for a section thickness of 20 cm. The width of the 95-% confidence envelope in this Bayesian age-depth model was then transferred to the corresponding midpoint ages in the originally published age model.

The interpretation of several hydroclimate indicators obtained from the sediments of Lake Masoko has led to conflicting hypotheses on the moisture-balance history of the lake. Discussions have mainly focused on the Last Glacial Maximum (LGM) and the Younger Dryas (YD) event. On one hand, there are diatom-based reconstructions of lake depth and salinity, which seem to indicate that water level was at a minimum during the LGM and YD (Barker et al. 2003; Barker and Gasse 2003). On the other hand, there is the low-field magnetic-susceptibility record which, based on modern samples from the lake shore and catchment, interpret increased magnetic susceptibility as a sign of low lake levels and/or increased seasonal variability and wind stress recruiting the titanomagnetite-rich shoreline reservoir (Garcin et al. 2006a). According to these data, the LGM and YD seem to have experienced the highest lake levels of at least the last ca. 40,000 years. Here, the latter, magnetic susceptibility-based interpretation is followed because of two reasons. First, as discussed by Garcin et al. (2006b), the observed changes in the diatom community assemblages might to an important degree be affected by mixing processes and nutrient status of the lake. Secondly, although not a direct indicator for lake status, pollen assemblages indicate an increase in drought-vulnerable species during periods with low magnetic susceptibility levels, which would confirm the interpretation of wet conditions during the LGM and YD (Garcin et al. 2006b). However, given these contradictory lines of evidence in the literature, the lake-status history of Lake Masoko should be regarded with caution.

The magnetic-susceptibility data, which formed the main line of evidence for deducing lake status, exhibits strong short-term variability which is superimposed on long gradual, seemingly precessionally forced fluctuations (Garcin et al. 2006a). Since the short-term variability of the long record (core MM8/M96) is not discussed by the original authors, it was not fully translated to lake-status variability here. Instead, the major periods of interest in the original literature, i.e. pre-LGM, LGM, deglaciation, YD and Holocene, were delineated and translated to three qualitative lake-status classes. To this end, the magnetic-susceptibility data was low-pass filtered with a Butterworth filter with a cut-off setting of 10 cm which, assuming constant sedimentation rate, corresponds to 160 years. In contrast, multidecadal variability in core MKJ-03-II is discussed in depth by Garcin et al. (2007a), and could therefore also be incorporated here. The magnetic-susceptibility record of this short core allowed the distinction of five distinct lake-status classes, and followed the interpretation of the original authors.

Lake Bogoria

Lake Bogoria is a saline, alkaline tectonic lake in the central Kenya Rift Valley, ca. 25 km north of the equator. The lake is hydrologically closed with Na^+, CO_3^{2-} and HCO_3^- as major ions, and receives substantial amounts of water (an estimated ca. 34 % of its total water budget; Onyando et al., 2005) from geothermal springs along the western and southern shoreline and below the lake surface. The lake consists of three basins along its north-south oriented longitudinal axis. These basins (north, central and south) are separated by two sills, with the south sill being deeper underwater than the north sill. The overflow point of the lake is formed by the Loboi Plain just north of the lake, at 999 m a.s.l. relatively large lake-level fluctuations have been noted during the 20th century, ranging from a maximum depth of ca. 9-16 m. During relatively high-water phases of the 20th century, all three basins hold a stably stratified water column, which breaks down as water levels decrease toward the lower end of the 20th-century range (De Cort et al., 2018).

The main source of evidence for the lake-level history of Lake Bogoria used here is De Cort et al. (2018). This study reconstructed the water-level history over the last ca. 1,300 years, based on sediment cores collected from five locations in the lake, namely the deepest point of the three basins, and the crest of the two sills separating them. The main source of evidence comes from mineralogical and sedimentological changes throughout these cores, which reflect connectivity between the basins depending on whether the sills are submerged, and concentration of solutes triggering mineral precipitation as lake level drops. All radiocarbon dates for these cores were derived from charcoal or terrestrial plant macrofossils. The original age-depth model of this publication was used, as radiocarbon dates were calibrated using the NHCal13 calibration curve (Reimer et al., 2013) and were combined with 12 $^{210}Pb/^{137}Cs$ dates in a Bayesian age-depth model using Bacon (Blaauw and Christen, 2011). Additional information on late-19th and 20th-century lake level fluctuations was included from Tiercelin et al. (1987), who reported a lake-level curve covering the period AD 1890-1980 based on sporadic historical observations and gauge measurements. No gauge data is available after AD 1980, but isolated observations by members of Ghent University's Limnology Unit (Ghent, Belgium) were used to constrain parts of the last few decades. Like many of the other Kenya Rift Valley lakes, Lake Bogoria has known a severe lake-level rise in 2013-2015, attaining levels not seen since the early 20th century.

Additional lake-status information has been reported in older studies. An older set of sediment cores is described in Tiercelin et al. (1987) and Vincens et al. (1986b; pollen). Also, a double band of stromatolites has recorded two episodes during which the lake was at about 999 m asl (at or close to the overflow level) and 995 m asl (close to, possibly slightly below, the 2013-2015 high-stand levels; Vincens et al., 1981; McCall, 2010). These previously described cores and stromatolites have been dated on bulk organic matter (for the cores) and carbonate (for the stromatolites). However, a study by De Cort et al. (2013) demonstrated the existence of a very large old-carbon effect on lake-derived carbon. Indeed, correlation between charcoal-dated cores from De Cort et al. (2018) and the bulk-dated cores from Tiercelin et al. (1987) show that dates in the latter have an offset of multiple millennia. In this geothermal setting, the assumption of a constant reservoir effect can't be made and requires a thorough investigation. Therefore, although these older sources contain compelling evidence for past changes in lake status, their chronological control was deemed too unreliable to be incorporated here.

Chew Bahir

Chew Bahir is an endorheic tectonic basin between the Omo-Turkana basin to the west and the southern sector of the Main Ethiopian Rift to the east. Today, it holds a 30x70 km saline mudflat that episodically fills to a shallow lake during the rainy season. Water and sediment is brought in by the perennial Weyto and Segen rivers feeding into the northern part of the basin, as well as alluvial fans draining the highlands to the east and west of Chew Bahir.

Indication of at least one previous lake high stand is derived from a single radiocarbon date obtained on an oyster shell isolated from limestone encrustations with countless lacustrine shells on a small volcanic island (Grove et al., 1975). These encrustations reached a height of 18-24 m above the basin floor. The authors argue that under such conditions, Chew Bahir was overflowing into the basin of Lake Turkana. Additional evidence on lake status is obtained from historical accounts of early European explorers who visited the area in the late 19th century and from sporadic observation in the 20th century (described in Grove et al., 1975).

Lake Afrera

Lake Afrera is a hypersaline lake occupying the depression between the Erta Ale and the Alaita-Tat Ali ridges of the Ethiopian Danakil Desert in the northern Afar. The area is subject to intense seismicity, hydrothermal activity and volcanism. Lake Afrera presently comprises two basins connected to each other by a narrow passage. Bathymetric surveying suggests that the deepest point is found in the north-eastern basin at ca. 80 m (ca. 180-190 m below sea level), which probably corresponds to the lowest topographic point of the entire Afar (Bonatti et al. 2017).

The lake is for the most part fed by warm saline springs, which are localised around the border of the lake particularly at the western side and most probably also below the water level. These springs are probably driven by infiltrated rain water that falls in the Ethiopian high lands and easily circulates in the strongly fissured basalts of the Afar (Gasse et al. 1974).

Evidence for former lake high stands is present in lacustrine deposits found above the current lake level. Carbonate-rich diatomites 10-30 m above the present lake surface were dated to the early Holocene and interpreted as being produced in a dilute lake with a surface area of ca. 250 km^2 (as opposed to the current ca. 80 km^2; Gasse et al. 1974). Dated calcareous sediments devoid of organisms were interpreted as stages in a progressively decreasing lake-level trend (Bannert et al. 1970; Gasse et al. 1974, Delibrias et al. 1974). Estimates of lake-surface elevation in 1969 and 2015 suggest that the lake level may have dropped significantly over the last ca. 50 years (Bonatti et al., 2017).

Dobi

Dobi is a graben in the Danakil Desert of Djibouti. Among the many grabens in the area, it stands out because of its steep faults and its depth, at some points reaching over 600 m below the surrounding plateau. The lowest parts of the basin floor are occupied by salty wetlands fed by warm, salt-water springs. These springs are probably fed at least in part by water from the Awash River that infiltrates the porous volcanic rock of the lower Danakil before it reaches the terminal basin (Lake Abhé). Part of this water resurfaces in a series of warm springs along the boundary of the Dobi, mainly on the eastern side of the graben, and forms small permanent streams that drain towards the wetlands on the basin floor.

Evidence for past episodes during which the Dobi Graben held a deep, dilute lake are present in the form of diatomitic deposits found above the basin floor. Close to the deepest part of the graben, Gasse et al. (1974) describes an undated 3-m thick finely laminated, carbonate-rich diatomite which testifies of a deep, permanently stratified and dilute lake. Towards the northwest, two diatomitic deposits containing shells (which were dated) provide evidence for at least two distinct lacustrine episodes with lake levels at least 50 m above the lowest part of the graben floor. The oldest bank, dated to 10870 ± 320 14C yr BP, has been severely affected by neotectonism. Due to lack of evidence on the relative depths of this older and more recent episode (the latter dated to 1960 ± 160 14C yr BP), both episodes have been assigned a status of 2, i.e. higher than the 20th- and 21st- century lake status. All dates were originally reported as uncalibrated, and were recalibrated using the IntCalNH13 calibration curve (Reimer et al. 2013).

Lake Malawi

Lake Malawi is a large rift lake at the southern end of the East African Rift System. Extending over 9-14 °S, it is about 650 km long, 40 km wide and up to 700 m deep. The lake and its catchment have a strongly seasonal tropical climate, with a single rainy season that typically lasts from November to April alternating with a dry season, characterized by dry southerly winds, from May to October. Its waters are very dilute (salinity = 0.2 ‰) and anoxic below a depth of 200 m (Wüest et al., 1996; Johnson et al. 2001). Evaporation is the main control on the hydrological budget, as the Shire River, draining the lake towards the south, accounts for only ca. 20 % of the total water loss (Beadle, 1981). Lake Malawi comprises seven half-graben units alternating asymmetrically along a single axis (Specht and Rosendahl, 1989). These form three main bathymetric basins and depositional provinces. The deep north and central basins are each ca. 150 km in length are characterized by steep offshore slopes in many areas and are surrounded by high mountains associated with the uplift of border fault footwall blocks (Specht and Rosendahl, 1989; Scholz et al., 2011). Offshore slopes are less steep in the southern basin, where maximum water depth is only 450 m, the lake floor is generally blanketed by fine-grained hemiplegic sediments, and the surrounding landscape consists of lower, rolling hills (Filippi and Talbot, 2005).

Evidence for past changes in lake level comes from several studies, which have investigated a variety of paleolimnological indicators. Scholz and Rosendahl (1988) and Scholz and Finney (1994) analyzed seismic profiles of the Songwe Sequence, the uppermost of four depositional sequences identifiable on multichannel seismic data from lake Malawi. Unconformities in the seismic data are interpreted as prolonged periods of major lowstands. Based on very long extrapolation of deposition rate of more recent sediments, two periods are estimated to have occurred between 28000 and >40000 yr BP and prior to 78000 yr BP. Because these deposits were not directly dated, these lowstand events are not incorporated into our data base.

Owen et al. (1990) collected echo soundings and geopulse recordings from the southern part of the lake, combined with grab samples and short (1 m on average) sediment cores. Geopulse traces reveal a major reflector, for which sampling revealed a varying nature of coarse sands and gravels with fish bones, faecal pellets and locally developed vivianite crusts (near-shore) and hard, dewatered mud-breccia (off-shore). A mollusc shell isolated from these deposits was dated to 10740 ± 130 ^{14}C yr. Here we assume a minimum duration of 100 years for this episode, which is interpreted by Owen et al. (1990) as a "major emergent event in the history of southern Lake Malawi". Owen et al. (1990) also describes several minor reflectors in the diatomaceous mud overlying the major reflector. The most recent of these is correlated to an erosional surface which, in a series of short gravity cores, typically consists of a cracked sandy layer sharply overlying the lower diatomaceous oozes. This is interpreted as a severe lowstand, during which water level was at least 121 m lower than during the late 20th century. The end of this lowstand, and reestablishment of deep-water sedimentation in the southern basin, was estimated through ^{210}Pb dating at ca. 1860 CE. These ^{210}Pb dates could unfortunately not be used to create an updated age depth model since uncertainties, as well as year of core collection, were not reported in the original publication. However, a multidecadal lowstand preceding the mid-19th century is also supported by a series of radiocarbon dates from archeological artefacts obtained near the present-day shoreline of Lake Magadi. A series of 19 such dates, which were recalibrated here using the SHCal13 calibration curve (Hogg et al., 2013), display remarkable hiatuses between ca. 350 and 100 cal yr BP and between 830 and 590 cal yr BP. Such hiatuses suggest that during these times, the lake receded enough to make fishing villages close to the present-day shoreline non-viable. De dates suggest that an earlier period of prolonged abandonment might have occurred between 1450 and 1050 cal yr BP. As this period was not discussed by Owen et al. (1990), it was inserted into the data base as a period of unknown water level bracketed by episodes of modern-day levels, rather than as a low stand.

Longer sediment cores provide evidence for hydrological variability further back in time. From a series of six 8-9-m long piston cores and sixteen short multi-cores collected during an IDEAL (International Decade of East African Lakes) cruise in 1998, most data has been published on 901-cm long core M98-2P. The lithostratigraphical composition of these cores consists of stiff homogenous brown clays, annually laminated muds (varves, consisting of a diatomaceous component deposited during the dry windy season and a terrigenous component deposited during the wet season), and silty homogenites which typically interrupt and erode the laminites (Barry et al., 2002). No clear link between lithology and water level could be established. Data on diatom community composition throughout sediment core M98-2P shows the highest relative percentages of periphytic diatoms (10-20 %) between ca. 870 and 700 cm, and between ca. 590 and 550 cm (Gasse et al., 2002). These are interpreted as episodes of major lowstands, with water level falling below the level of the more recent lowstands described above. The age-depth model of core M98-2P was reconstructed using Bacon in R. All ten radiocarbon dates reported in Barry et al. (2002) were recalibrated using the SHCal13 calibration curve (Hogg et al., 2013) and used in the construction of the model. As in the original publication, an old-carbon reservoir of 450 years was employed for bulk samples. Although the sediment-water interface of core M98-2P was not recovered, chronology above the uppermost ^{14}C date was anchored using three varve-counted time points tied to short cores from nearby locations (Gasse et al., 2002).

The lake status of Lake Malawi during the early to mid-Holocene has been a subject of debate. Prior to the IDEAL work, a Duke University expedition collected 33 piston cores with an average length of 10 m from Lake Malawi in 1986. Bulk-sedimentology and elemental-composition data of these cores is discussed in, among others, Finney and Johnson (1991) and Ricketts and Johnson (1996), who infer a period of low water level between 10000 and 6000 yr BP based on the deposition of calcium carbonate during this time. However, Gasse et al. (2002) argue that the early to mid-Holocene was a period of intermediate, rather than low water levels. A clearer picture was drawn by Van Bockxlaer et al. (2012), who date periods of deposition at Chipalamawamba (south of Lake Malawi's most southern tip) where several subsequent units of shallow-water sedimentation alternate with erosional contacts, as exposed by downcutting of the upper Shire River. All dates were obtained from radiocarbon dating in-situ extracted mollusc shells. The mollusc reservoir effect of 175 years estimated by Van Bockxlaer et al. (2012) was retained, and all radiocarbon dates were recalibrated using the SHCal13 calibration curve (Hogg et al., 2013). This sequence suggests three multi-century long periods during which Lake Malawi stood at least 5 m higher than at present. Because of subsequent erosion, timing of the ending of these periods can only be estimated conservatively, while lake level during the erosional periods can't be determined. This unambiguous evidence for oscillating lake level contrasts with previous hypotheses of rather continuous high or low phases during the early to mid-Holocene. Interestingly, the first of the three high stands at Chipalamawamba partly overlaps, and thus contrasts, with a period of increased deposition of periphytic diatoms at the site of core M98-2P (core depths 591-551 cm; Gasse et al., 2002). Because of the better chronological control and the more robust indicator of lake level employed in the study of Van Bockxlaer et al. (2012), the short interval of periphytic diatom deposition from Gasse et al. (2002) was not incorporated as a period of low water level in this data base.

Information on lake level after 1860 is available from historical observations and, later, from gauge measurements. The water-level curve presented in Johnson and McCave (2008) was used for incorporation in this data base, subsectioning the observed 1867-2000 range (470-477 m asl) into 2-m increments.

In 2005 an international team collected the first long, continuous and high-fidelity cores from a lake in tropical East Africa at Lake Malawi. The data and interpretation of these cores was mainly targeted at multi-millennial climate and lake-level change over the period 145,000-10,000 cal yr BP (cf. Scholz et al. 2011). These reconstructions were not incorporated into the data base as they fall largely outside of the time window targeted here.

Lake Nkuruba

Lake Nkuruba is a small, moderately deep, slightly alkaline (pH = 8) freshwater lake (Saulnier-Talbot et al. 2018). It is sheltered by high forested walls that average 48 m above current water level. The lake has no know surface inflow or outflow and is likely fed solely by rain. It is one of many volcanic explosion crater lakes that dot the landscape in southwestern Uganda.

The lake-level history of Lake Nkuruba was taken from Saulnier-Talbot et al. (2018), who present data on lithology and algal pigments from a 150-cm sediment core collected in January 2009. The sole evidence for water-level variability is based on lithology, which shows a contrast between dark organic muds and coarse organic sandy mud at times rich in snail shells or pebbles. The latter type of sediment is observed between 122 and 74 cm core depth, and is interpreted as reflecting a severe lowstand. To anchor the sediment core chronologically, the study presents five time points (core top, one tie point to a ^{210}Pb-dated short core, and three ^{14}C-dated wood samples). To obtain a continuous age-depth model, the authors constructed a Bayesian age-depth model with the Bacon software in R, in which the sandy unit was regarded as a slump deposit. However, the original publication also argues against this unit being the product of a single slumping event. Therefore, the timing of the lowstand was here determined by directly using the two radiocarbon dates immediately bracketing the sandy unit as its start and end date. These dates were recalibrated using the NHCal13 calibration curve (Reimer et al. 2013). Similarly, the base of the core (and thus the age of oldest available information on lake level) was estimated by using the radiocarbon age extracted from the core base.

Urwi Pan

Under present-day conditions Urwi pan is dry year-round. We use the author's interpretation that stromatolites (formed in permanent wave agitated waters) and found 1-1.5m above the present pan floor level represent a time where there was a standing body of water occupying the pan. We use 14C dates from these stromatolites to indicate a higher lake status than present.

Tsodilo Hills

Under present day conditions at Tsodilo (NW Botswana) there is currently no lake (or any historical record of one) but there is well documented evidence of a flat plain with a calcrete sheet containing gastropods and diatoms. This depression currently has no inflow and would likely have been fed by runoff and seepage from the adjacent Tsodilo Hills.

Brook et al. (1992) dated two carbonate samples from the deepest part of the lake. They also analysed diatoms from these samples and inferred the first (C1) was from ephemeral shallow lake and C2 was more permanent deep lake.

Thomas et al. (2003) built on this work by taking more samples from the lake bed and also samples from the now degraded bounding shoreline. Dated calcretes from this study were not included here as there some speculation that these are post-depositional and may have formed as the lake dried up. As there is no altimetric significance to these calcrete ages it is therefore very difficult to attribute any accurate lake status to them. However, dates associated with hydrological conditions that were indicated by mollusc and diatom assemblages were included.

Additional dates from fish fossils in a nearby cave site (Robbins et al., 2000) also imply the use of nearby water resources. However, there is no consensus on specifically where these came from and so conservatively we do not include them here.

≠Gi

≠Gi is a small pan located in northeast Botswana close to the Namibian border which ephemerally holds water across an area up to 16 km^2 after heavy rainfall.

This site was investigated in the 1970s as important MSA/LSA site within interior southern Africa (Yellen, 1971).

The pan lies in the Dobe valley which is described as containing outcrops of lacustrine limestones overlain by grey and white sands.

Detailed descriptions of section stratigraphy are given by Helgren and Brooks, (1983). Former water levels in the pan are identified via terraces marking lake high-stands 2-4 m above the present pan floor together with sedimentological evidence. Arid periods are identified by inorganic sediments containing groundwater calcretes interpreted as pan-margin alluvia and colluvia.

Significant 'lake' phases are identified by massive indurated limestones. The lake at Gi is linked to standing water in the Dobe valley (not dated) where the authors identify algal matts and subaqueous slumps. Older sections of the sediments are 14C dated but return infinite ages.

Alexandersfontein

Located near the northern cape/orange free state border, this salt pan depression is believed to be formed by periodic deflation but, it is suggested, previously held a lake with a surface area of 44 km^2. The current basin margin is defined at 151m (+32m) in the past forming an overflow outlet to the Modder River. Only the younger lake phases are considered in this paper. The current basin catchment is 328 km^2 (Butzer et al., 1973).

Primary information is taken from Butzer et al. (1973). We are aware new research at this site is forthcoming but has not yet been published.

Beach ridges mark the southern side of the pan with shoreline cliffs locally cut into shale on the northern side. The authors identify 3 levels +17-19, +12m and +6m. Only the highest level is dated via 14C in two sections using inorganic carbonate. However, the authors suggest one of these dates is contaminated by young carbon via secondary calcification. We have taken this interpretation at face value.

They infer high lake level represents “a long interval with appreciable oscillations of the shoreline”. MSA archaeology (with in situ fossils at one site) have been identified here on both the 19m and 12m shoreline. This may cast doubt on the age of the 19m lake but it is unclear to what extent the archaeology had been investigated.

Presently, the pan only holds water intermittently after heavy rains.

Ngami

Lake Ngami is a 3000 km^2 basin at the distal end of the Okavango delta in northern Botswana which, together with the Mababe depression and Makgadikgadi pans, makes up the palaeo-lake Makgadikgadi system. Because it is possible for the level of the lake to be independent from other parts of the basin (up to a certain elevation) we treat the data from this basin independently from the system as a whole.

Dated calcretes from the Ngami basin (Shaw 1985) have been interpreted to imply minimum ages for calcification due to aerial exposure following a decline in groundwater level. These dates therefore represent low lake levels following lake high-stand declines.

Subsequently luminescence dating was applied to the shorelines themselves by Shaw et al (2003) as well as investigations into diatom deposits from the eastern end of the basin.

Huntsman-Mapila et al. (2006) used a section dug into the basin to infer from rare earth elemental profiles and diatom assemblages that high lake levels occurred during periods where Al_2O_3 and organic matter (LOI) were elevated. Her section was dated using TL dating and 14C but it is not clear whether the dates were calibrated or not in the original study. We assume not as no mention of calibration is made.

Burrough et al. (2007) resampled and OSL dated a large number of shoreline sediments from known altitudinal contexts around the basin. Shoreline dates on sand are assumed to represent that shoreline elevation/lake extent even where taken at depth.

Robbins et al. (2008) dated in situ archaeological sites within the Ngami basin (potentially offering an insight into low stands) but excavations lack altimetric control so are considered to have an uncertain context as far lake status is concerned.

Makgadikgadi

Historically, Makgadikgadi has been subject to a large number of studies, with dating efforts on lake levels beginning in the 1980s and continuing to the present day. These include C14 dates on shells and calcrete and OSL dates on depositional shorelines. Some of the C14 dates are derived from opportunistic contexts where the geomorphic significance in relation to lake level is unclear. Where there is no clear altimetric or geomorphological significance to the dates (as far as lake level determination is concerned) we have not included this data in the final analyses.

TL dates have also been reported from a section in the eastern side of the pan (Ringrose et al., 2005). These are interpreted as dating the deposition of cal-silcrete within the sediment. However, it is not possible to date calcrete/silcrete using TL. The TL dates more realistically put an age on the last sediment deposition of grains within this matrix and not the post-depositional cal-silcrete formation within them. We interpret them as such and include them where there is clear and relatively precise altimetric data relating to lake levels.

With respect to OSL dates on shorelines (Burrough et al., 2009), the oldest OSL ages in these series have uncertainties that are so large that the dates are not useful in any climate analysis and we exclude these dates from final analyses.

Where there is conflict between dated shorelines and calcrete ages (within errors) we have used the age with the smallest uncertainty as the bracketing age for a lake phase.

Diatom records (Schmidt et al., 2017 and Shaw et al., 1997) have also been used to interpret lake level and status as these are found in a distinct context where water was ponded back behind a lake along a fluvial valley and subsequently exposed through fluvial incision. The altitudinal level of these deposits is therefore relevant in this context as well as the water body conditions the diatom assemblages in these studies.

Mababe

The Mababe depression is a 50 km by 90 km heart shaped basin in northern Botswana that, together with lake Ngami and the Makgadikgadi pans makes up the palaeo-lake Makgadikgadi system. Because it is possible for the level of the lake in this depression to be independent from other parts of the basin (up to a certain elevation) we treat the data from this basin independently from the system as a whole.

A series of calcrete dates on river terraces from within the basin are available from studies in the late 80s and early 90s (Shaw, 1985; Shaw and Thomas, 1988; Shaw and Cooke, 1986). These were generally interpreted to imply minimum ages for calcification due to aerial exposure following a decline water level (either fluvial or lacustrine). These dates therefore represent times when the lake was lower than or fluctuating at this level. Where the altitudinal context is unclear, dates were omitted from the final analyses.

Subsequently luminescence dating was applied to the basin shorelines by Burrough and Thomas (2008) and used where altimetric control had been previously established. Where there was evidence of mixing within sediments, dates were excluded.

Etosha

Etosha basin is a huge (4760 km^2) salt pan located in the southernmost part of the Owambo basin in northern Namibia and has been subject to several intensive investigations, many of which have inferred its former status as a closed palaeolake. These include TL and OSL ages on shorelines (Buch and Zoller, 1992; Brook et al., 2007; Hipondoka et al., 2014) and ^{14}C ages from stromatolite beds left stranded a few metres about the lake floor. We use both shoreline elevation and the elevation of stromatolites to indicate standing water at this level, in line with the interpretations of these authors.

Where altitudinal data is absent or where the authors indicate possible contamination by older organic material, we exclude dates from the final analyses.

Additional evidence for former wetter conditions than present comes from dated fossil faunal assemblages (Hipondoka, 2005; Hipondoka, 2006) but these provide little information on lake level within the basin so have not been included in this analysis.

Tsondab

The Tsondab area in Namibia has been investigated many times, by many different authors. It falls within the category of 'lake' only because it has been interpreted to hold standing water as result of increased floods along fluvial channels in the Namib. These floods left silt and mud deposits within an otherwise entirely sandy system. Its interpretation as a 'lake' is tenuous but meets the criteria as a standing body of water which has left a long term record in the landscape. We include it here because it provides a data in a region where there is almost none.

Initial attempts at dating these standing water episodes were undertaken using radiocarbon on associated deposits (particularly shells, terrace calcretes and calcified roots). Subsequent work using OSL dating obtained ages on bracketing sands. This provided a much cruder estimate with larger uncertainties but demonstrated that the radiocarbon ages were likely contaminated by both older and younger carbon as had been suspected by previous researchers.

The bracketing ages appear inverted (though not within errors because of the large uncertainties on these dates). We excluded the older OSL dates as there was no stratigraphic consistency to them. We used a summary approach for the younger bracketing ages of all the mud units (as provided by Stone et al., 2010). The uncertainty in doing so is strongly represented in the errors captured by the database particularly within the oldest/youngest age parameters.

Branddam East and Omongwa Pans

The SW Kalahari pans are drainless basins which receive hydrological input by a combination of direct precipitation, surface and sub-surface inflow. Under current conditions water coverage is ephemeral occurring only for some days or weeks during the year.

Of the four pans studied by Schuller et al., (2018), geochemical data was available for only two. These two sets of data were used as hydrological indicators.

In the database we only considered two proxies from the suite that were studied (K/Al and sedimentation rate), all other proxies, reflecting aeolian processes, wind strength or some integration of indicators, were omitted as they lacked the criteria necessary to define hydrological change. In addition, only periods where there was clear agreement between both sedimentation rate and geochemical proxy data (K/Al) were used to infer pan status. Where there was no clear agreement between these two indicators, we did not allocate a status for the pan/basin.

The ratio of K/Al in this study, as determined by XRF measurements is taken by the authors to reflect the amount of water in the pan and the fluvial input from the catchment into the pan. This is based on other studies that have used this criteria. It is not clear whether there is enough information with regard to local processes and catchment specific inputs to make this proxy 100% robust in this region but we have used the interpretation as found.

There is sparse alternative data available for this region.

Kathu Pan

This area contains a series of 11 dolines that are developed within the Tertiary sequence of the Kalahari Group and have undergone infilling during the late Quaternary (Beaumont et al., 1984). There are several specific excavated sites at Kathu Pan, which have been studied predominantly with the goal of recording and analysing the rich stone age archaeological record present here. Some sites have only a single 14C date, others have more. The interpretation of environmental conditions is sometimes based largely on pollen data (e.g. KP2). Other sites are interpreted based on their sedimentology (e.g. KP5). We have used data where there is more than one 14C age within a stratigraphy (in order to construct a more robust age model) and where interpretations are hydrologically rather than ecologically focused e.g. the existence of peats (higher water table) vs sands and calcrete (lower water table) whilst still remaining true to the accepted interpretation by the author. Note that while some cores are not used, their sediments e.g. the existence of peat beds with one or more radiocarbon dates, is occasionally at odds with other cores from the same pan. We have used data from KP5 which provides the greatest number of 14C ages and interpretations that hinge more strongly on hydrological rather than ecological data.

The age model was coded in 2 sections to deal with strong chrono-stratigraphic break between Str2a and Str2b.

The pumping of groundwater for the town of Kathu began to drop the water table in the mid-1970s, resulting in the formation of a number of sinkholes (dolines) subsiding in the pan deposit (Walker et al., 2013). The first research to be conducted here was in 1975. Presently, the water-table rises above the surface in summer, but lies 1-2 m below it in mid-winter (Beaumont et al., 1984).

Lebatse Pan

A study by Holmgren and Shaw (1997) used three cores from now dry/ephemeral pan in SE Botswana to identify four periods of deposition separated by hiatuses. These hiatuses are interpreted as marking dry conditions with possible pan deflation. Potential for unknown loss of sediment means authors are unable to put clear timeframe on hiatuses.

The chronology is sparse for this pan and is dated only by one radiocarbon date and one TL date derived from two separate cores. Uncertainties on the age model are therefore large. Sedimentary structure between cores is very similar which renders the interpretations therein slightly more robust. Sediment interpretations by the authors are summarized below. Where interpretations appear uncertain or in conflict, we have not ascribed a hydrological status to the pan.

Witpan

Witpan is located in a particularly dry area of the Kalahari where most research focus has been centred on establishing detailed chronologies of dune accumulation. Several studies have noted that pans in the region have potential to preserve an independent record of hydrological change.

Under present day conditions at Witpan, there is no accumulation of sediments and there is a net loss of material from the pan due to aeolian deflation during the dry season. We use the author's interpretation that any net accumulation of the silts and clays dated in their study would require a surface significantly wetter than present (Telfer et al., 2009).

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