Quaternary history of the Lake Magadi Basin, southern Kenya Rift: Tectonic and climatic controls


Keywords:
- Rift basins
- Geochemistry
- Sedimentology
- Palaeolakes
- Diatoms
- Mineralogy

Abstract

Sediments from the Magadi Basin (south Kenya Rift) preserve a one-million-year palaeoenvironmental record that reflects interactions between climatic, volcanic and tectonic controls. Climate changes that impacted sedimentation include wet-dry cycles on variable timescales and an overall progressive trend towards greater aridity. Volcanic influences involved inputs of tephra to the basin, significant inflow of geothermal fluids, and the effects of weathering, erosion and transportation of clastics from trachyte and basalt terrains. Tectonic controls, which were often step-like, reflect the influence of faults that provided pathways for fluids and which controlled accommodation space and drainage directions.

Intensified aridity and evaporative concentration resulted in salinity and pH increasing with time, which led to a change from calcite deposition in mildly saline lakes before 380 ka to the later formation of zeolites from reactions of volcaniclastic debris with highly alkaline lake and pore water. After 105 ka, hyperalkaline conditions led to trona accumulation and increasingly variable rare earth elements (REEs). The presence of mixed saline and freshwater diatom taxa between 545 and 16 ka indicates climate variability and episodic inputs of fresh water to saline lakes. Calcrete formed in lake marginal settings during semi-arid periods.

Tectonic controls operated independently of climate, but they interacted together to determine environmental conditions. Aquatic deposition was maintained during periods of increasing aridity because fault-controlled ambient and geothermal springs continued to flow lakewards. This recharge, in turn, limited pedogenesis: palaeosols are common in other rift floor sequences. Trona formed when aridity and evapoconcentration increased, but its precipitation also reflects increased magmatic CO2 that ascended along faults. Basin fragmentation and north-south fractures caused loss of cross-rift (east-west) drainage from rift-marginal basalts, resulting in reduced transition metals after 545 ka. The Magadi Basin demonstrates how a careful reconstruction of these complex tectono-climatic interactions is essential for accurate palaeoenvironmental reconstruction in continental rifts and in other tectonic settings.

https://doi.org/10.1016/j.palaeo.2019.01.017

Received 15 October 2018; Received in revised form 19 December 2018; Accepted 8 January 2019
Available online 14 January 2019

© 2019 Elsevier B.V. All rights reserved.
1. Introduction

Much of the early palaeoclimate research in East Africa was devoted to understanding Plio-Pleistocene hominin and archaeological sites (see Cohen et al., 2016 and Campisano et al., 2017 for reviews). More recently, efforts have focused on climate variability from various Pliocene to Recent time-slices (Kingston et al., 2007; Owen et al., 2008; Tierney et al., 2010; Junginger and Trauth, 2013; Magill et al., 2013). A range of record types have proven useful in reconstructing palaeoclimates in the region, including Late Quaternary core-records (e.g., Verschuren and Chapman, 2008; De Cort et al., 2013, 2018), outcrop data (e.g., Levin, 2015), and marine cores (e.g., deMenocal, 1995, 2004). More recently, drill cores from extant rift lakes have proven exceptionally useful in documenting palaeoenvironmental histories, including a 1.3-million-year Lake Malawi record (Ivory et al., 2016). In addition, a one-million-year pollen-diatom-mineralogy study of a core at Lake Magadi synthesised the regional climatic history for the southern Kenya Rift as a basis for exploring hominin evolution and mammalian change (Owen et al., 2018a). The latter paper complements this study, which is based on outcrops and two lake cores, and which focuses on aquatic sedimentation and palaeoenvironments using sedimentological, mineralogical and geochemical data, supplemented by diatom analyses.

Lake Magadi lies in the axial trough of the southern Kenya Rift ~605 m above sea level (masl), and is a seasonally-flooded, saline
alkaline pan, currently floored by trona (Fig. 1). Discontinuous Quaternary sedimentary outcrops around the lake include fluvial sediments (channel deposits, alluvium), calcrete, lacustrine limestone (including microbialites), zeolitic mudstone and siltstone, sodium silicate minerals, and chert of diverse origins (Baker, 1958, 1963; Eugster, 1967, 1969, 1980; Hay, 1968; Herrick, 1972; Surdam and Eugster, 1976; Behr, 2002; Brenna, 2016; Felske, 2016; Leet et al., 2016). These sediments accumulated in a N-S axial rift sump, in which the northern depocentre remained a lake or wetland for most of the last million years because of spring recharge during drier periods. Lake Magadi probably united with Lake Natron in northern Tanzania as a single, relatively dilute lake for different periods during the Pleistocene and early Holocene (Eugster, 1986; Casanova and Hillaire-Marcel, 1987; Williamson et al., 1993).

Lake Magadi was cored in June 2014 by the Hominin Sites and Paleolakes Drilling Project (HSPDP), which aims to develop basin-to-regional scale palaeoenvironmental histories that can be compared with local hominin remains and artefacts to infer possible environmental influences on hominin evolution (Cohen et al., 2016). We present here results of detailed analyses of the sedimentology, major- and trace-element geochemistry and mineralogy from two Lake Magadi cores and nearby sediment outcrops (Fig. 1A), supported by diatom records, which collectively provide a history of changing aquatic environments during the past million years. Both drill cores reached the volcanic...
This new evidence gives an opportunity to reconstruct the Pleistocene history of the Magadi basin in much greater detail than previously possible. Specifically, we aim to: 1) reconstruct the environmental history of the Magadi palaeolakes; and 2) relate that sedimentary record to evolving tectonic, volcanic and climatic controls.

2. Previous outcrop and borehole studies in the Magadi Basin

Early descriptions of sediments in the Magadi basin were provided by Parkinson (1914), Gregory (1921), Walter (1922) and Coates (as Anonymous, 1923). The oldest exposed deposits are the fluvial, spring and lacustrine ‘Oloronga Beds’, which rest upon the Magadi Trachytes (~1.4–0.8 Ma; Fig. 1B; Baker, 1958, 1963; Crossley, 1979, Herrick, 1972; Eugster, 1980; Behr, 2002). A radiometric age of 0.78 ± 0.04 Ma for an obsidian flow overlying basal Oloronga sediments (Fairhead et al., 1972; Eugster, 1980, his Fig. 15.21) is consistent with 36Cl/Cl brine estimates of ~0.76 Ma for salt accumulation in the basin (Kaufman et al., 1990). A U/Th date of 0.3 Ma indicates a minimum age for the ‘upper’ Oloronga Beds (Fig. 1B; Röhricht, 1998; Behr and Röhricht, 2000), which are overlain by a calcrete (Eugster, 1980).

Fig. 3. Oloronga Beds, sections OB1 and OB2 with locations also shown on Fig. 1. Inset maps, A and B, show locations of stratigraphic logs at the southwestern corner of Magadi Basin. Key to all logs in Figs. 3, 5 and 6, to the right. The temporal trend is from fluvial (oldest) to lacustrine and then calcrites.
Lacustrine and alluvial sediments are also preserved several km northwest of Lake Magadi, including at Lainyamok (Fig. 1A) (Shipman et al., 1983; Potts et al., 1988). Despite a lack of exposure of geological contacts, Temperley (unpublished, cited by Baker, 1958) reported a ‘Chert Series’ that lay above the Oloronga Beds. Baker (1958) described stratigraphic sections. Eugster (1967, 1969, 1980) and Surdam and Eugster (1976) proposed that the Chert Series formed part of the Late Pleistocene to Holocene ‘High Magadi Beds’ (HMB), despite their non-conformable contact. U/Th dating of chert by Goetz and Hillaire-Marcel (1992), however, showed that some of the chert deposits were much older (98.5 ± 20 ka and 40.0 ± 6.5 ka). Röhrricht (1998), Behr and Röhrricht (2000) and Behr (2002) reassigned sediments with bedded chert above the Oloronga Beds to the newly defined ‘Green Beds’ (Fig. 1B).

Stromatolitic limestones encrusting bedrock up to 60 m above modern Lake Magadi (~660 masl), were used to infer former higher lake-phases at ~130 ka and ~12–10 ka, with a maximum palaeoshoreline at ~656 m (Hillaire-Marcel et al., 1986; Casanova, 1987; Hillaire-Marcel and Casanova, 1987). An earlier stromatolite generation, up to 80 m above Lake Natron, was dated at >200 ka. Those higher palaeolakes were inferred to have formed a single water body in the combined Natron-Magadi basins.

The Late Pleistocene to Holocene HMB (Fig. 1B) were deposited in the modern axial graben in a fresh to moderately saline, alkaline lake (White, 1953; Baker, 1958, 1963; Eugster, 1969, 1980; Herrick, 1972; Surdam and Eugster, 1976). White (1953), Baker (1958), Butzer et al. (1972), and Behr (2002) noted a palaeoshoreline ~40 ft (12.2 m) above the modern lake. Partly laminated diatomaceous core sediments dated at 17.71 ± 0.22 to 10.8 ± 0.12 ka from the Northwest Lagoon of Lake Magadi imply freshwater inflow to a former saline lake (Barker et al., 1990; Taieb et al., 1991; Damnati et al., 1992, 2007; Roberts et al., 1993; Damnati and Taieb, 1995).

Baker (1958), Hay (1968), Surdam and Eugster (1976) and Eugster (1980) described the mineralogy of Lake Magadi boreholes drilled in 1953, but gave few stratigraphic details. They inferred low lake-levels after deposition of the HMB, with accumulation of > 40 m of bedded trona and black mud of the Evaporite Series (Fig. 1B). The hydrology, hydrochemistry and brine evolution at Lake Magadi have been discussed by Gregory (1921), Baker (1958), Jones et al. (1967, 1977), Eugster (1969, 1970, 1980, 1986), Eugster and Hardie (1978), Eugster and Jones (1968, 1979), Hillaire-Marcel and Casanova (1987), Allen et al. (1989), Darling (2001) and others. Owen et al. (2018a) described the stratigraphy of the Magadi HSPDP cores, upon which this study is partly based, from a palaeoclimate perspective, using pollen, diatom and mineralogical records.

Fig. 4. Oloronga Beds facies in outcrop. Hammer for scale is 28 cm long. A: Basal fluvial sandstone overlain by lacustrine siltstone and chert, with pisolithic and massive calcrite at top, SW Magadi Basin site OB2. B: Detail of trough cross-bedded sandstone in Fig. 4A. C: Thin undulating chert beds in zeolitic siltstone near outcrop in Fig. 4A. D: Uncemented calcite pisoids from depression in capping calcrite, Site OB1. E: Calcrite with chert clasts capping sequence in Fig. 4A; some clasts show reticulation characteristic of Magadi-type chert. Coin diameter: 23 mm. F: Small abandoned palaeochannel filled with greenish black claystone near Site OB2. HMB: tuffaceous siltstone of upper High Magadi Beds; MT: Magadi Trachyte escarpment. G: Fossil branching root-system in greenish black claystone (F) filled by younger fine silts. H: Small cylindrical tufa chimney in zeolitic siltstone and chert near section OB2. Calcite lip around top implies that the small spring vent was subaerially exposed at times. I: Tufa tower, ~6 m high, located < 80 m from the western edge of a NNE-SSW horst, ~7 km west of Lake Magadi.
Calcrete with claystone and chert clasts; passes laterally (E) into wavy bedded chert

Pale yellow-green massive silty clay

Buff mottled mud and clay with brown plant debris

Grey-green chert with wavy banding

Dark green massive mud with chert granules

Banded grey-green waxy clays with white laminas

Burrowed siltstone with gastropod? shells

Pale green muddy siltstone with rootlets

Laminated red siltstone (zeolitic tuff?)

Dark green laminated mud with simple pellet-filled vertical burrows (1 cm diameter, <5 cm long)

Nodular grey chert in greenish clay matrix

Laminated green mud (fractured, filled by orange and brown granules, massive base)

Dark olive green laminated mud with simple pellet-filled vertical burrows (1 cm diameter, <5 cm long) at upper contact

Orange oxidised mudstone

Massive olive-green mud with fractures and subvertical cm-scale burrows (full relief), some branching and filled by orange and brown granules
Fig. 5. Section GB1 of Green Beds, south Lake Magadi. See Fig. 3 for the key. Inset maps (a, b) show location of the section, excavated on the northeastern edge of the track across a prominent N-S ridge (horst) armoured by bedded Green Beds chert. A: Exposed section showing context of facies photographs in images B to F. B: Lowest part of the section showing burrowed and mottled pale green and orange mud with prominent straight thinly lined vertical burrow preserved in full relief and partly filled by orange and brown granules. C: Horizontal burrow with short blunt side-branches on bedding plane in lower part of section. Preserved as negative epirelief. Burrow fill is finer than host sediments. Similar burrows, horizontal and vertical, are common also in Oloronga Beds outcrops. D: Weakly laminated green and orange muds, locally mottled, with lenses of pale grey mud. E: Grey-green and white ‘laminated’ muds. The discontinuous white laminae, 1–2 mm thick, comprise patches of compacted plant debris and (or) trails mineralised by opal-A? (see inset, same scale). Although they appear to be white laminae in section they are discontinuous. F: Thin chest bed with low-amplitude undulations overlying dark green muds. G: Outcrop of horizontally bedded chert showing crystal pseudomorphs (moulds) of carbonate minerals and fluid-escape structures (FE) that confirm a soft precursor. Eugster (1969) termed these linear moulds “pearl chains of calcite rhombs and their casts”, suggesting they were originally calcite or gaylussite. H: Chert dyke, up to 80 cm high, cutting older, platy bedded chert (left). I: ‘Pillow chert’ of Behr (2002). These common cherts form coalescent, mainly reddish mounds several metres long (typically oriented N-S) with brecciated chert interiors covered by fractured, laminar chert layers (inset: scale bar: 3 cm) producing a biohermal morphology. Arrow points to hammer ~ 30 cm long. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

3. Methods

Sediment outcrops logged between 2006 and 2018 were combined with sampling of modern sediments and waters across the catchment. Lake Magadi was drilled in June 2014 to depths of ~133 mbs (metres below surface; Site 1) and 194 mbs (Site 2) (Fig. 1A; Cohen et al., 2016). Geochemical analyses were carried out on outcrop samples (n = 76) and core sediments, with MAG14-2A sampled at 32 cm intervals, and where distinctive lithologies would have been omitted (n = 344; see Electronic Supplementary Material). Samples were analysed by Activation Laboratories Ltd., Ancaster ON, Canada (4E-exploration package; for methods, see http://www.actlabs.com). Major and trace elements were determined by coupled-plasma and inductively coupled plasma-mass spectrometry. Detection limits are given in the Electronic Supplementary Material. Rare earth element (REE) data were normalised against C1 chondrite composition (Sun and McDonough, 1989). Total organic carbon (TOC) was determined by loss-on-ignition (LOI) at 550 °C without prior removal of carbonates. X-ray diffraction samples were taken at 16 cm-intervals and analysed using a Panalytical Xpert Pro MPD diffractometer using CuKα radiation at 45 kV and 40 mA.

Diatoms are rare or absent in outcrops. In MAG14-2A, diatom samples were collected every ~32 cm (as for geochemical samples) and where facies changed. Synthetic silica microspheres (8 μm diameter) were added to assist quantitative counts. Organic matter and carbonates were removed using H2O2 and HCl. Diatoms were mounted in Naphrax. At least 400 frustules were counted for each microscope slide except where diatoms were rare (see Electronic Supplementary Material).

Dating methods are described by Owen et al. (2018a). Chert samples from core (n = 18) and outcrop (n = 6) samples were dated using U-series methods (Fig. 2A). The uppermost parts of the cores were dated using 14C techniques (n = 11). Those analyses produced inconsistent ages, likely because of recycled old carbon; six ages are compatible with the lithostratigraphy. Nine replicate 40Ar/39Ar single-crystal dates were determined from K-feldspar crystals in a little-altered tephra in MAG14-2A at 96–101 mbs with a single crystal date obtained at 151 mbs. The Brunhes-Matuyama boundary was identified at 174.36 mbs. Bayesian methods were applied to define the age-model and its uncertainties (Fig. 2B; Bacon v. 2.2; Blaauw and Christen, 2011).

4. Results

4.1. Exposed Quaternary sediments

The Early to Middle Pleistocene Oloronga Beds overlie Magadi Trachyte in scattered outcrops north, south and west of Lake Magadi (Baker, 1958; Fig. 1A). Sections OB1 and OB2, southwest of Magadi, contain trough cross-bedded fluvial sandstone (Figs. 1A, 3, 4A, B) overlain by bedded, wavy lacustrine chert and grey zeolitic siltstone (Fig. 4C) covered locally by thin (~5 cm) dark grey limestone. A laterally extensive calcrite up to 40 cm thick, with massive, pisolithic and laminated facies caps the Oloronga Beds inconformably across much of the basin (Fig. 4A, D; Eugster, 1980; Felske, 2016), including at Section OB2 where the massive facies contains subangular chert clasts (Fig. 4E). This calcite is one of several of different ages (Felske, 2016). Locally, SW-NE paleochannels that incise the fluvial sandstone are filled with blackish brown mudstone with lighter cm-scale vertical and horizontal branching trace fossils (roots?) (Fig. 4F, G). Small cylindrical pale grey-brown tufa ‘chimneys’ (<70 cm high, <40 cm diameter, ~1 cm-thick walls) with a flat upper rim are interstratified with grey silts and laminar cherts at sites ~150 m from the modern SW Magadi hot springs (Fig. 4H). The calcite tufa cements angular chert and siltstone fragments. Most chimneys align along N-S to NNE-SSW trends.

About 10 km west of northern Lake Magadi (Fig. 1A), several tall N-S-aligned tufa towers up to ~6 m high (Fig. 4I), and linear mounds of spring carbonate (up to ~2.5 m high with variable orientation), lie upon Magadi Trachyte (Baker, 1958). Internally, most towers are composed of clusters of cm-scale vertical pipes with subhorizontal plates and cement. About 150 m south of the tallest mound (Fig. 4I), poorly-preserved rimstone dams are present. These spring deposits lie ~120 m east of a fault-scarp up to 20 m high that marks the edge of the N-S horst upon which the towers are rooted.

Section GB1 (Fig. 5), located at the southern margin of Lake Magadi (Fig. 1A), exposes a 2.6 m sequence of Middle to Late Pleistocene Green Beds (Fig. 5A). The section includes basal massive, burrowed green and orange mud (Fig. 5B, C) that gives way upwards to laminated green and orange mud with thin white flakes (silicified plant fragments?) on lamination planes (Fig. 5D, E), thinly shelled spired gastropods, and three bedded, undulating light and dark grey quartz-chert horizons < 3 cm thick (Fig. 5F). The latter can be traced northwards for > 700 m and, on the land surface, show open crystal moulds of calcite, trona, gaylussite and other salts (Fig. 5G), fluid-escape structures, shrinkage cracks, tepee structures, trace fossils and possible raindrop-impact prints (Eugster, 1969; Behr, 2002; Scott, 2010). Northeast of the lake, low (~30 cm) terraces of pale-green chert fragments might reflect palaeoshorelines of unknown age. Many chert precursors (opaline species, gels or sodium silicates?) were remobilised while partly soft to form dykes, decimetres thick and at least 50 m long (Fig. 5H). Other intrusive chert outcrops consist of reddish domal mounds of quartz with concordant fractured and brecciated laminae, some with couplets and triplets of repeating laminae typical of microbialites (Fig. 5I).

Palaeoshorelines that formed when Lake Magadi was a deeper expanded lake are present in many locations up to ~12 m above the modern lake (Figs. 6, 7A–B). Exposures in the ‘Dry Lagoon’ southeast of Lake Magadi (Fig. 1A) show sections through the Late Pleistocene to Early Holocene HMB (Fig. 6). The Lower HMB include a basal imbricated trachyte breccia (subangular conglomerate) over lain by interbedded gravel lenses and pale brown tuff with thin lensoid to bedded magadite horizons and isolated magadite patches. The top of the lower HMB in many outcrops is a dark brown, laminated clay with well-preserved Tilapia fish fossils (Fig. 7D). In contrast, the Upper HMB are a crudely to well-bedded, zeolitic (mainly erionite) tuffaceous silt.
Volcaniclastic trough-cross-bedded sandstone overlain by Upper HMB near Karamai (SE Magadi) provides evidence of fluvial-deltaic inflow to the palaeolake (Figs. 6D, 7C).

Trona, which covers Lake Magadi during most dry seasons (Fig. 7E, F), is the contemporary surface of the Evaporite Series (Baker, 1958), which comprises bedded and massive trona and nahcolite, interbedded with anoxic black zeolitic mud. The evaporites and mud extend > 65 m below the lake floor.

4.2. Core data

4.2.1. Sedimentology and mineralogy

Cores MAG14-1A and MAG14-1C form a composite sequence that reached trachyte at 133 mbs. MAG14-2A contains 194 m of sediment, including gaps (Figs. 1A, 2A). Core sediments consist of detrital siliciclastic grains, primary (airfall) and reworked ash and pumice, detrital and authigenic clay minerals and zeolites, carbonate (micrite, gastropods, ostracods), siliceous microfossils (diatoms, sponge spicules, phytoliths) and organic matter (pollen, spores, charcoal, diffuse organic matter, microbial matter). Evaporite deposits, common in the uppermost 65 m of MAG14-2A, include nahcolite and abundant trona. Calcite, Mg-calcite and dolomite are present below ~100 mbs. Many minerals, both deposited and precipitated, have undergone early diagenetic alteration. Zeolites, including erionite, phillipsite, mordenite, natrolite, clinoptilolite, chabazite and analcime, are common, especially between ~100 and 10 mbs. K-feldspar, clay minerals, biotite, pyrite and hematite are also common. Chert (as cryptocrystalline quartz and chaledony) and magadinite are present at some levels.

Fifteen lithofacies are distinguished (Table 1; Fig. 8). Gastropods and ostracod-bearing carbonate-grainstone are present in the basal parts of both cores (Facies 1–3, Table 1). Overlying sediments are zeolitic, massive, thinly bedded or laminated, green, brown or black silt, clay, mud and ash (F4–F8, Table 1, Fig. 8). Nodular and bedded chert (F13) and silicified mudstone (F12) are also present. Massive (F11) and bedded (F10) trona, and massive black, zeolitic mud with trona (F9) dominate the uppermost 65 m of core MAG14-2A.

Coarse-grained siliciclastic sediments include matrix-supported paraconglomerate and rare sand-and-gravel (orthoconglomerate) intervals (F15), which are most common in MAG14-1A. Palaeosols are absent, but minor root development was observed (F14).

4.2.2. Geochemical zonation, MAG14-2A

Correlations between outcrops and geochemical zones are shown in Fig. 9. Six geochemical zones (G1–G6) can be distinguished using changes in total organic carbon (TOC) and Ca/Na and (K+Na)/Al ratios (Fig. 9A). TOC is low and Ca/Na ratios are highest in Zone G1 (194.3–186 mbs, ~1056–930 ka; Figs. 2, 9A). Ca/Na ratios and (K+Na)/Al decrease in G2 (186–169 mbs, ~930–740 ka), but P2O5 concentrations increase. Calcite is present throughout both zones with analcime confined to G1, and Mg-calcite present in lower G1. Zone G3 (169–132 mbs, ~740–545 ka) includes variable Ca/Na and slightly elevated (K+Na)/Al ratios. Calcite is present throughout, with Mg-calcite common, and dolomite recorded at some levels. Analcime is present through most of G3 but other zeolites are absent. Zone G4 (132–102 mbs, ~545–380 ka) has variable Ca/Na ratios but with a moderate increase in (K+Na)/Al. Calcite and Mg-calcite are present, but less common than in Zone G3. Zeolites other than analcime appear for the first time in a few horizons in G4. Zone G5 (102–60 mbs,
~380–105 ka) is characterised by low Ca/Na and (K+Na)/Al ratios and LOI values that intermittently increase in upper G5. Calcite and Mg-calcite are absent, but a more varied assemblage of zeolites is present. Zone G6 (60–0 mbs, ~105–0 ka) shows an increase in TOC, low Ca/Na ratios, and very high (K+Na)/Al ratios, reflecting the dominance of trona and nahcolite. The dominant zeolites vary in G6. Using trace elements, nineteen subzones are recognised (Fig. 9B; Table 2).

4.2.3. Diatom stages

Fig. 10 shows selected diatom taxa in core MAG14-2A plotted against age, highlighting periods when diatoms were preserved between ~545 and ~16 ka (132–38 mbs). Correlations with outcrops are also shown in this figure. Nine stages are distinguished. Stages D1 (194.3–132 mbs, ~1056–545 ka) and D9 (37–0 mbs; ~16–0 ka) lack diatoms. Both saline (e.g., *Thalassiosira faurii*, *T. rudolfi*, *Cyclotella meneghiniana*) and freshwater (*Aulacoseira granulata* and varieties, *A. agassizii*) taxa are present in D2–D8 sediments. Stage D2 (132–108 mbs, ~545–415 ka) contains rare diatoms dominated by planktonic *Aulacoseira* spp., but with *C. meneghiniana* and *T. faurii* recorded in a few horizons. Overall diatom diversity is low. Diatoms vary in abundance in D3–D8 and include well-preserved diverse floras with episodic increases in benthic taxa. Stages D3 and D5 are distinguished by abundant *C. meneghiniana*. Planktonic *A. granulata* and *A. agassizii* are common throughout D3–D8, with the former dominant in most samples from D4, D6 and D8 and the latter dominant in D3 and D7.

5. Discussion

5.1. Palaeoenvironmental interpretation

5.1.1. Geochemical stratigraphy

Lacustrine deposition began soon after volcanic eruptions in the Magadi Basin ceased, given the lack of weathering of the underlying trachytes. Zone G1 (~1056–930 ka) is characterised by ostracod-rich grainstone without diatoms. In contrast, ostracod- and gastropod-rich grainstone in the basal part of MAG14-1A and MAG14-1C (Fig. 2A) contains freshwater benthic and epiphytic diatoms (*Epithemia, Rhopalodia, Encyonema*), implying a marsh setting. Grainstone is absent above the G1-G2 boundary, but calcite in silts throughout G2 (~930–740 ka) indicates relatively low palaeolake salinity. Calcite might have formed during photosynthetic microbial blooms as happens in modern rift lakes (cf. Ng’ang’a et al., 1998; Stone et al., 2011).

Principal component analyses confirm a close relationship among REE, Al₂O₃, Fe₂O₃ and TiO₂ and feldspars, implying similar detrital sources (Fig. 11A). REE patterns for G1 and G2 resemble those for the Magadi Trachyte Formation (Fig. 11B), which are characterised by a
Table 1
Representative major facies. See Fig. 8 for typical examples in core. Based on HSPDP Initial Core Description (ICD) logs. Letters refer to facies codes.

<table>
<thead>
<tr>
<th>Primary lithology</th>
<th>Facies code</th>
<th>Lithology, structure and diatom content</th>
<th>Associated features</th>
<th>Environmental interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>F1. Gastropod limestone</td>
<td>Lg</td>
<td>Limestone (L) with abundant in situ and reworked gastropod (g) shells in micritic and sparitic matrix.</td>
<td>Shallow-water diatoms, ostracods.</td>
<td>Shallow to moderately deep, fresh lakes.</td>
</tr>
<tr>
<td>F2. Carbonate grainstone</td>
<td>Lgr</td>
<td>Bedded fine to coarse carbonate sand grainstone (shells, trachyte) (r).</td>
<td>Shallow water diatoms, ostracods, gastropods.</td>
<td>Shallow to moderately deep, fresh lakes.</td>
</tr>
<tr>
<td>F3. Carbonate mud</td>
<td>Lu</td>
<td>Poorly bedded, light brown carbonate mud (u).</td>
<td>Black pyrite grains (1–1.5 cm) with bluish rims. Light coloured grains of chert. Chert with open vugs.</td>
<td>Moderately deep or deep, fresh to mildly saline lake.</td>
</tr>
<tr>
<td>F4. Laminated silt and clay</td>
<td>Czl</td>
<td>Greenish grey, light brown or brown zeolitic or calcitic silt (z) clays (c). Lamine (l) 1–3 mm thick.</td>
<td>Convolute laminae, microfractures. Pollen. Pyrite, patchy chert, unidentified white crystals.</td>
<td>Lacustrine.</td>
</tr>
<tr>
<td>F5. Bedded silty clay</td>
<td>Czb</td>
<td>Alternating beds (1–6 cm thick) (b) of light to medium and darker green zeolitic silt (C) or clayey silt. Darker bands are siltier.</td>
<td>Scattered pyrite. Mottling. Unidentified white crystals.</td>
<td>Lacustrine.</td>
</tr>
<tr>
<td>F6. Massive clay</td>
<td>Cmb</td>
<td>Massive (m) to weakly bedded light green zeolitic clay. Wispy banding (1–4 cm).</td>
<td>Scattered white crystals and pyrite. Also present in subhorizontal wisps and laminae (1–2 mm).</td>
<td>Lacustrine.</td>
</tr>
<tr>
<td>F7. Massive silty ash</td>
<td>Zm</td>
<td>Massive (m) dark grey zeolitic silt (Z). Subhorizontal fractures hint at layering.</td>
<td>Local chert nodules. Pyrite in some levels.</td>
<td>Lacustrine.</td>
</tr>
<tr>
<td>F9. Massive mud with trona</td>
<td>Mt</td>
<td>Black mud (M) with mm-scale trona (t) crystals. Interbedded with massive or bedded trona.</td>
<td>Some larger trona crystals.</td>
<td>Saline alkaline lake.</td>
</tr>
<tr>
<td>F10. Bedded trona</td>
<td>Tb</td>
<td>Bedded trona (T) (1–3 cm layers) with clusters of radiating crystals.</td>
<td>(Sub)horizontal dissolution surfaces.</td>
<td>Very shallow (&lt; 1 m?), saline, alkaline lake.</td>
</tr>
<tr>
<td>F11. Massive trona</td>
<td>Tm</td>
<td>Massive recrystallized trona. Crystals &lt; 1 to 3 cm with mostly random orientations.</td>
<td>Interstitial greenish black mud is common.</td>
<td>Saline alkaline lake.</td>
</tr>
<tr>
<td>F12. Silicified mudstone</td>
<td>Ui</td>
<td>Bedded green mud (U) with variable degrees of silicification (i). Very fine clay with some silt.</td>
<td>Pyrite present at some levels.</td>
<td>Lacustrine.</td>
</tr>
<tr>
<td>F13. Chert</td>
<td>Hb; He; Hn</td>
<td>Bedded (b), lensoid (e) to nodular (n) cherts (H).</td>
<td>Compaction around chert indicating early lithification. Some chert has serrated or reticulate surface textures.</td>
<td>Shallow saline lake. Replacement of magadiite, trona, mudstone or silica gel in saline lake sediments.</td>
</tr>
<tr>
<td>F14. Massive mud with roots</td>
<td>Umr</td>
<td>Massive green mud with plant remains, roots (r) and local oxidation.</td>
<td>Pyrite. Locally forms paraconglomerates.</td>
<td>Palaeosol, but with reducing conditions dominant? Alluvial sandflat with local debris flows?</td>
</tr>
</tbody>
</table>
negative Eu anomaly that is absent in basalts of the region (Le Roex et al., 2001; Owen et al., 2011, 2014). Furthermore, low concentrations of Co, Cr, Cu, Ni and V (Fig. 9B) support a trachytic source lithology because these transition metals are less abundant in lavas of the Magadi Trachyte Formation (2.6–3.9, 1.6–4.9, 6.3–12, < 8.4, < 3 ppm, respectively) than in rift basalt (41–51, 53–254, 110–184, 21–124 and 232–307 ppm, respectively) (Le Roex et al., 2001).

Zones G1 and G2 are also distinguished by relatively high $P_2O_5$.
Fig. 9. Geochemical zones (bulk geochemistry) and authigenic mineralogy versus time for core MAG14-2A. A: Geochemical zones are based on changes in Ca/Na and (K + Na)/Al ratios, P₂O₅, LOI and mineralogy and broadly reflect a transition from fresh and mildly saline waters (G1 to G3) to saline (G4) and hypersaline (G5 and G6) conditions. Major authigenic minerals to right (excluding silica species: quartz, chalcedony, magadiite and other sodium silicates, which are present at several levels). Siliciclastic detrital minerals are shown in Table 2. B: Temporal variability in trace elements, Core MAG14-2A. Subzones are defined using trace elements (Table 2) and reflect changes in salinity of the palaeolake, spring inputs, variations in bottom water conditions and the supply of transition metals from basaltic and trachytic lithologies. All elements in ppm, except for Au (ppb) and S (%). See Table 2 for descriptions of each zone and subzone. Magadi outcrop correlations shown to the right (ES = Evaporite Series; HMB = High Magadi Beds). Green Beds dating range includes younger dates from Behr (2002) and results from this study (191–158 ka).
<table>
<thead>
<tr>
<th>Zone</th>
<th>Sub zone</th>
<th>Age (ka)</th>
<th>Major elements</th>
<th>Trace elements</th>
<th>Major lithologies</th>
<th>Mineralogy/LOI</th>
</tr>
</thead>
<tbody>
<tr>
<td>G6</td>
<td>d</td>
<td>5–0</td>
<td>Low Ca/Na ratios; moderate to very high K + Na/Al; high LOI; low P₂O₅, CaO and MgO; relatively low Al₂O₃, SiO₂</td>
<td>Low trace-element concentrations</td>
<td>Trona and black anoxic mud</td>
<td>Mg-calcite and dolomite near top of core; trona, nahcolite, pyrite, quartz, albite, anorthoclase, K-feldspars, analcime, natrolite, mordenite, erionite, phillipsite, chabazite, diopside, natrolite. High LOI</td>
</tr>
<tr>
<td></td>
<td>c</td>
<td>28–5</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>82–20</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>a</td>
<td>100–82</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>G5</td>
<td>e</td>
<td>130–100</td>
<td>Low Ca/Na ratio; moderate K + Na/Al; low to moderate LOI (with one very high value); low P₂O₅, CaO and MgO</td>
<td>Relatively high As, Be, Br, Cs, Mo, Rh, Sb, Sc and U compared to G5d</td>
<td>Siliciclastic silt, sand and mud; chert, tufts</td>
<td>Pyrite, quartz, albite, anorthoclase, K-feldspars, analcime, erionite, phillipsite, natrolite. Low LOI</td>
</tr>
<tr>
<td></td>
<td>d</td>
<td>181–130</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>c</td>
<td>192–181</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>285–192</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>a</td>
<td>375–285</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>G4</td>
<td>d</td>
<td>395–375</td>
<td>Similar to G3 but with higher K + Na/Al</td>
<td>E levated Sc, V compared to G4c Similar to G4b, but with lower Ba, Sr, Th, and Y variable REE</td>
<td>Siliciclastic silt, sand mud and chert</td>
<td>Calcite, Mg-calcite, dolomite, pyrite, quartz, albite, anorthoclase, K-feldspar, feldspathoids, erionite, analcime, phillipsite, natrolite. Low LOI</td>
</tr>
<tr>
<td></td>
<td>c</td>
<td>505–395</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>520–505</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>a</td>
<td>545–520</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>G3</td>
<td>c</td>
<td>650–545</td>
<td>Low to moderate Ca/Na ratio; moderate K + Na/Al; low P₂O₅; moderate LOI, CaO and MgO</td>
<td>Similar to G3a, but with Be increasing upwards Increased Br, Co, Cu, Mo, Ni and W, with lower REE compared with G3a</td>
<td>Siliciclastic silt, sand and mud and chert</td>
<td>Calcite, Mg-calcite, dolomite, pyrite, quartz, albite, anorthoclase, K-feldspar, feldspathoids. Low LOI</td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>660–650</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>a</td>
<td>730–660</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>G2</td>
<td>b</td>
<td>845–730</td>
<td>Moderate to high Ca/Na ratios; low K₂O + Na₂O/Al₂O₃ and LOI; low to high P₂O₅; low CaO and MgO</td>
<td>Low trace-element concentrations</td>
<td>Mud, chert</td>
<td>Calcite, pyrite, quartz, albite, anorthoclase, K-feldspar, analcime. Very low LOI</td>
</tr>
<tr>
<td></td>
<td>a</td>
<td>930–845</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>G1</td>
<td>1040–930 ka</td>
<td>Very high Ca/Na ratios; low K₂O + Na₂O/Al₂O₃; low to moderate LOI; high P₂O₅; moderate to high CaO but low MgO</td>
<td>Trace elements low; Moderately high Be, Cr, Rh, Sc and REE</td>
<td>Carbonates, siliciclastic mud and chert</td>
<td>Calcite, Mg-calcite, dolomite, pyrite, quartz, albite, anorthoclase, K-feldspar, feldspathoids, analcime. Very low LOI</td>
<td></td>
</tr>
</tbody>
</table>
Phosphate, which is released following weathering becomes available to plants, and is then concentrated in soils following decay of litterfall (Ruttenberg, 2014). Consequently, the elevated P$_2$O$_5$ could reflect P delivery to the palaeolake from soils enriched in P, implying a relatively wet climate. Dericquebourg et al. (2015), for example, suggested that phosphatic sediments in Miocene Lake Lukeino (central Kenya Rift) record high runoff to that palaeolake from organic-rich forest-covered soils. Ostracods and gastropods in the early Magadi palaeolake also indicate that it was then shallow and dilute. The absence of pollen, common in overlying sediments, implies oxic bottom waters, which would have favoured phosphorus deposition (Cosmidis et al., 2014). Chert began to accumulate from about 1 Ma ago, implying that early north-south rift faulting might have provided pathways for silica-rich deep fluids to reach the surface (Owen et al., 2018b), supplementing the silica derived from chemical weathering and runoff. Later, intermittent development of chert implies continued inflow of silica-rich runoff, spring- and groundwaters, with periodic development of evaporative alkaline brines that underwent dilution, cooling or both, perhaps with microbial mediation of some silica precipitation.

Zone G3 (~740–545 ka) correlates with upper Stage D1, which lacks diatoms, perhaps reflecting dissolution, competitive exclusion or high turbidity. K + Na/Al ratios increase and P$_2$O$_5$ decreases in G3, with the sediments containing Mg-calcite and dolomite. Those minerals typically form in alkaline waters with high Mg/Ca ratios (Last et al., 2012), and imply lake- or pore-fluids with a Mg/Ca ratio high enough to induce primary carbonate precipitation or early diagenetic alteration of a carbonate precursor (Murphy et al., 2014).

REE data for G3 (Fig. 11) show similarities to G1–G2 indicating weathering of trachytic bedrock. However, relatively steep REE patterns, reflected in normalised La/Lu ratios, suggest an increase in basaltic bedrock sources. This inference is supported by increased Co, Cr, Cu, Ni and V, which are more abundant in rift basalt. The REE changes and increase in transition metals might have been induced by N-S axial-rift faulting that fragmented the basin and eroded trachyte, locally exposing older rocks. Potential sources include the Ol Tepesi, Singaraini and Kirikiti Basalts, and basalt along the rift margins or beyond (Baker, 1958; Guth and Wood, 2014). In all cases, cross-rift lateral drainage is implied for mafic siliciclastics to have reached the MAG14-2A site. Increased organic carbon and sulfur (Fig. 9), and development of euhedral pyrite crystals (~3 mm) in G3b upwards indicate increased bottom-water anoxia. Redox-sensitive trace-metals (Mo, Cd, U) also increase in G3, as they do today in the deeper, anoxic and sulfidic waters of Lake Tanganyika (Brucker et al., 2011).

Aquatic sedimentation during G4 times (~545–380 ka) appears to have been continuous at the MAG14-2A site despite outcrop distributions that indicate that the palaeolake had shrunk in size. This zone overlaps with stages D2, D3 and lower D4, in which diatoms are rare. Floras are dominated by planktonic species that include freshwater
Aulacoseira mixed with saline Cyclotella meneghiniana and Thalassiosira faurii (Fig. 10), with mean transfer-functions implying pH of ~7.4–8.5 and conductivities of ~300–3000 μS cm⁻¹. Co decreases sharply at the base of G4, together with declining Cr, Cu, Ni and V (Fig. 9B), reflecting reduced basaltic inputs, perhaps due to (1) diminished weathered basaltic terrains, (2) increased aridity and (or) reduced fluvial flow from basaltic terrains, or (3) rising horst-blocks that deflected drainage axially, in common with regional N-S tilting (Owen et al., 2014).

Ca/Na and (K + Na)/Al ratios increase slightly in G4, with analcime common and rare natrolite, erionite and phillipsite, indicating higher salinity, consistent with increasing regional aridity (Fig. 9A). P₂O₅ remained low during G4 deposition with well-preserved organic matter and lamination implying an anoxic lake floor. Variable chondrite-normalised La/Lu ratios and Eu anomalies, first observed in G3, continue. G4 REE patterns show strongly negative Eu anomalies, declining LREE and flattening to declining HREE, with variable total REE (Fig. 11B). The REE patterns resemble those for rocks of the Magadi Trachyte Formation (Owen et al., 2011), but with a distinctive depletion in Lu similar to
samples from Late Pleistocene spring tufa at Olorgesailie (Lee et al., 2013).

Zone G5 (~380–105 ka) coincides with upper D4 to lower D8, which contain rare to common, mixed saline and freshwater, planktonic diatoms dominated by Aulacoseira, Thalassiosira and Cyclotella. Episodic increases in freshwater benthic species (Fig. 10) may indicate dilute inflow during floods (Barker et al., 1990). One explanation for mixing of saline and freshwater floras could be that a saline alkaline lake was periodically, or seasonally, flooded by fresh, lower-density nutrient-rich waters and became meromictic, similar to modern Lake Bogoria (Fig. 12D; De Cort et al., 2018). Potential water sources include axial rivers and dilute overflow from the Koora Graben to the east. Diatoms could have lived in the fresh surface waters (mixolimnion) if essential nutrients remained available and evaporative concentration did not exceed their salinity tolerances.

Zone G5 is characterised by very low Ca/Na and low (K + Na)/Al ratios (Fig. 9A). Models of brine evolution reflect critical geochemical divides and pathways that help to explain the Magadi hydrochemistry (Hardie and Eugster, 1970; Deocampo and Jones, 2014). Today, Lake Magadi brines form following the evaporation of dilute inflow with HCO$_3^-$ ≪ Ca + Mg. Early precipitation of calcite and Mg-calcite (plus minor primary or replacement dolomite?) in subsurface flow paths, at or near spring discharges, at the lake margin, or in the water column leads to a Ca- and Mg-free Na-CO$_3$-SO$_4$-Cl brine (Deocampo and Renaut, 2016). Microbial sulfate reduction then produces Na-CO$_3$-Cl fluids from which trona and halite precipitate (Hardie and Eugster, 1970; Jones et al., 1977; Eugster, 1986). Lower Ca content in G5 likely reflects these processes and (or) proportionally more hydrothermal inflow with low Ca and high Na, as is typical for most of the modern hot springs (Jones et al, 1977; Allen et al., 1989).

Zone G5 sediments contain many zeolites (natrolite, chabazite, erionite, phillipsite, clinoptilolite) but lack calcite or Mg-calcite, reflecting increasingly saline and alkaline waters (Rabideaux, 2018). Many zeolites formed during early diagenesis (cf. Hay, 1966, 1970; Manega and Bieda, 1987). Erionite, the most common zeolite, forms where trachytic glass reacts with water. Later zeolite phases formed where erionite or other precursors reacted with more concentrated brines (Herrick, 1972; Surdam and Eugster, 1976). Erionite + Na$^+$...
produces analcime and is favoured by high Na\(^+\)/H\(^+\) ratios and low silica activity. In contrast, phillipsite forms by reaction of trachytic glass with Ca\(^+\) and K\(^+\)-bearing fluids associated with high silica activity. Herrick (1972) noted that analcime might form from aluminosilicate gels, which are often present in modern sediments at Nasikie Engida (Fig. 1A), a small hyperalkaline lake northwest of Lake Magadi (Eugster and Jones, 1968). Hay (1964) inferred that dominance of erionite over phillipsite reflects high Na/K ratios in brines, with phillipsite associated with ratios of 28–34. In contrast, chabazite requires more Ca to form (Singer and Stoffers, 1980), and clinoptilolite needs high silica activity (Herrick, 1972).

Zone G5 REE patterns vary more than those in G1–G4. Some show negative Eu anomalies, but others do not. Tb is lower in some G5 samples. Nd and Sm also change with HREE showing falling, flat and rising trends. Kerrich et al. (2002) reported REE variability in outcrops of the Green Beds and High Magadi Beds, which correlate with parts of G5 and G6, respectively (Fig. 9A). Although REE are commonly used in provenance studies because of their stability (Taylor and McLennan, 1985), they vary in response to changes in pH, redox conditions, adsorption/desorption and cation exchange in highly saline alkaline lakes. Lee and Byrne (1993), for example, noted that carbonate-rich brines have a significant impact on REE speciation and in complexing REEs. Lee and Byrne (1993), for example, noted that carbonate-rich brines have a significant impact on REE speciation and in complexing REEs. For example, carbonate-rich brines may complex REEs, preventing their mobilization and deposition in lacustrine sediments (Lee and Byrne, 1993).

Zone G6 (~105–0 ka) sediments, characterised by very low Ca/Na, high (K + Na)/Al ratios and high LOI\(_{550}\) percentages, include many zeolites, abundant trona and minor nahcolite (Fig. 9A). Stage D8 (120–16 ka, Fig. 10) overlapped with G6a–G6c, with diatoms declining in abundance upwards. Taxa include mixed freshwater Aulacoseira spp. and saline Thalassiosira and Cyclotella taxa. The mixed flora, similar to that in G5, implies periodic flooding and the formation of meromictic lakes of variable duration, with sparse diatoms after about 0.08 ka consistent with development of a trona saline pan.

REE data vary more than in lower zones (Fig. 11B). Detailed G6 REE patterns show that trona-bearing mud has a negative Eu anomaly and that REEs decline from HREE to LREE (Fig. 11) except for Lu, which shows both negative and positive anomalies. In contrast, trona shows positive anomalies for Nd and Tb with low La, Sm, Eu and Yb and a small positive anomaly for Lu, probably reflecting the influence of strongly alkaline carbonate-rich brines.

Relatively high Au concentrations in G6c, which partially overlaps with the African Humid Period (AHP) and HMB deposition, contrast with Au concentrations below detection limits in G6a–G6b and G6d at times when the basin was a trona pan. Zone G6a has less As, Br, Sb and REE than G6b (Fig. 9B) with G6c distinguished from G6b by higher concentrations of Br, Ag, Pb, Th, U and Zn, together with less As, Mo and Sb. Several of these elements (Br, Pb, Zn, Sb, Ag) have been related to springs feeding carbonate lakes at Sassykul in Tajikistan (Volкова, 1998). Other elements were considered tracers of evaporative concentration (U, Mo, As) at Sassykul. Variations in these elements might reflect contrasting spring sources and brine evolution in palaeolake Magadi. Br shows very-low positive or negative correlations with other elements, with Br concentrations increasing and Ca/Na ratios decreasing upwards (Fig. 9B), which may indicate a link to increasing salinity and possible hydrothermal inflow.

5.1.2. Outcrop to core correlation and palaeogeography

Outcrop to core correlations are shown in Figs. 9, 10 and 13. The Magadi outcrop and core sediments lie upon the Magadi Trachyte Formation (~1.4–0.8 Ma), which consists of flood trachyte up to 120 m thick that filled the pre-existing horst-and-graben topography. Rivers and lakes then deposited fluvial and lacustrine sediments upon the volcanic substrate (Baker, 1958; Crossley and Knight, 1981; Guth and Wood, 2014).

The poorly dated Oloronga Beds began to accumulate before 0.78 ± 0.04 Ma based on a K-Ar date from obsidian within basal Oloronga lake beds (Fairhead et al., 1972; Eugster, 1980), his Fig. 15.21). A U-Th dated hippo tooth gave a minimum age of 300 ka for the upper Oloronga Beds (Röhricht, 1998; Behr and Röhricht, 2000), although the actual ages of these beds could be much older. The lack of pedogenesis upon the trachyte substrate at MAG14-2A implies that deposition started soon after eruptions ceased (~1078.3 ± 3.6 ka). Together, the evidence from cores and the distribution of the Oloronga Beds, show that the earliest palaeolakes (Zones G1–G2, ~1056–740 ka) were fresh to mildly saline and more extensive than today (Fig. 12A).

Tufa-travertine towers on a horst west of Lake Magadi have been radiocarbon dated at 16.7 ± 0.4 to 25.5 ± 0.7 ka (Hillaire-Marcel et al., 1986) and were interpreted to have formed while partly submersed near the HMB palaeolake shoreline. However, they are possibly much older, given that the base of the largest tower (~712 m asl) is > 50 m higher than the maximum inferred Late Pleistocene-Holocene HMB lake level (~660 m asl) (Hillaire-Marcel et al., 1986). It is unclear if the tufa deposits are subaerial or sublacustrine or both, but the towers lack external drapery that might be expected with subaerial outflow (cf. Bargá, 1978). Possible rimstone dams south of the towers imply subaerial formation and the dense travertine fabrics of those dams are compatible with high-temperature fluids (Jones and Renaut, 2010). In contrast, the towers, with internal vertical tubes, are similar to those described from lake floors (e.g., Lake Abhé: Dekov et al., 2014). A sublacustrine origin (Casanova and Hillaire-Marcel, 1987, their Fig. 9) would imply a Holocene lake at ~715–720 m or higher, for which there is no recorded evidence. Either the horst block has been uplifted > 50 m during the terminal Pleistocene and Holocene, or the tufa towers may be contemporary with the small Oloronga tufa chimneys at Section OB1, having formed before axial-rift faulting, and contemporary with Zone G1 (1080–930 ka) when carbonate grainstone was deposited (Fig. 12A). Baker (1958) found no evidence for tectonic disturbance of the HMB and recorded the ‘40ft’ (+12.2 m: ~617–620 m: Fig. 7A) shoreline at many locations across the basin. Behr (2002, p. 260) similarly found no sedimentary or geomorphological evidence for a higher precursor lake at that time. The age of the tufa towers remains uncertain.

The distribution of sediment outcrops that are contemporary with Zone G3 (~740–545 ka) suggests a similar areal extent for the Magadi palaeolake to that inferred for G1–G2 times (Fig. 12B). Faulting increased the topographic expression of north-south trending grabens, some of which might have hosted isolated or periodically interconnected lakes.

In outcrop, the Oloronga Beds are commonly capped by pisolitic to massive calcite (Fig. 4A, D and E) that implies prolonged semi-aridity and relatively low lake level. In cores, the boundary between Zones G4 and G5 (Fig. 9A) is characterised by reduced Ca/Na ratios and a change from calcite and Mg-calcite to zeolites, which indicates an increase in palaesoilinity consistent with increased aridity at ~380 ka. The widespread calcite overlies laterally extensive and eroded Oloronga Beds (Eugster, 1980; Felske, 2016) with younger outcrop sediments restricted to the modern axial lake basin, which suggests that the G4 (~545–380 ka) palaeolake had permanently shrunk (Fig. 12C), likely due to faulting and basin fragmentation that developed an axial horst- and-graben topography. Increased tectonic activity at Magadi during this interval is consistent with evidence from the Olorgesailie Basin, ~20 km to the northeast, where faulting disrupted that basin after ~500 ka (Behrensmeier et al., 2018; Potts et al., 2018).

During G5 times (~380–105 ka), a saline, alkaline lake developed that might have been periodically flooded by axial rivers, or perhaps overflow from a contemporary lake in the neighbouring Koora Basin to the east (Fig. 12D–E) (Baker, 1958, 1986; Marsden, 1979; Muiruri, 2018). Although the timing of any overflow episodes is uncertain, evidence for such events are preserved in a deep NW-SE fault-controlled channel incised into Magadi Trachyte and by palaeowaterfalls to the southeast of modern Lake Magadi, near Karamai. Composite cores MAG14-1A and MAG14-1C (Fig. 12D), between the north and central Lake Magadi sub-basins, contain orthoconglomerate that correlates with G5. These gravels accumulated on a shallow E-W tectonic sill.
contrast, MAG14-2A lies in the deeper north Magadi sub-basin and remained wet even during times of increased aridity, perhaps due to spring inflow.

The Green Beds, south and northeast of the modern lake, are contemporary with upper Zone G5 and possibly lower G6. They were previously dated at 98.5 ± 20 ka and 40.0 ± 6.5 ka using U/Th techniques (Goetz and Hillaire-Marcel, 1992). New U/Th dates from bedded cherts obtained for this study indicate ages of 180.6 ± 19.5, 176.6 ± 23.7, and 158.4 ± 17.4 ka. Chert dykes that intrude the Green Beds have ages of 191.8 ± 3.6, 166.9 ± 3.3 and 163.0 ± 3.3 ka. The latter dates might reflect initial chert formation rather than the age of dyke injection but all are older (~191–158 ka) than those proposed by Goetz and Hillaire-Marcel (1992) and Behr (2002). The revised chert ages correlate with zones G5c–d (Fig. 9B) and diatom stage D7 (Fig. 10). Outcrops of the Green Beds imply deposition on a gently sloping playa margin during the later stages of Zone G5 with upward-shallowing cycles, strong evaporation, and with microbialites and bedded chert forming periodically in shallow saline, alkaline waters.

Zone G6 is poorly represented in outcrops as it represents a time when Lake Magadi was generally low and confined to the axial graben as a small highly saline lake or trona pan (Fig. 12F). However, there were wetter intervals when the lake expanded as it did during the African Humid Period (Fig. 12G). Butzer et al. (1972), for example, reported a 14C date of 9120 ± 120 yr BP for a fish-bearing clay marker-bed in the lower HMB (Figs. 6, 7D). They also reported “corrected” 14C trona dates from a mid-basin core of 4600 (11 m depth), 5750 (21.4 m) and 10,010 (47.5 m) yr BP (errors unreported) in the Evaporite Series but questioned their reliability. Despite the age uncertainty, these deposits represent Holocene periods when the palaeolake resembled the modern evaporative lake (Fig. 12H). Williamson et al. (1993) also obtained U/Th dates from a core in the ‘Northwest Arm’ of Lake Magadi of
40,000 ± 6500 and 23,700 ± 6000 yr BP with HMB high lake deposits dated at 12,090 ± 120 and 10,800 ± 120 yr BP. The HMB and Evaporite Series would therefore correlate with Zones G6c and d in MAG14-2A, and parts of Stage D9 (Figs. 9 and 10).

5.2. Controls on sedimentation

Sedimentation in the Magadi Basin was influenced by climate, tectonics and volcanism. The importance of climate is confirmed by pollen and diatom data, which record a drying trend after ~575 ka and many links to global climate trends (Owen et al., 2018a). For example, diagonal shading in Fig. 13 shows correlations between interglacials (Antarctica data: Jouzel et al., 2007) and influxes of benthic diatoms (high PCA values) that imply flooding of the palaeolake. Periods when conditions were wetter in the Magadi Basin (750–525 ka) were also recorded in the Olorgesailie Basin, 20 km to the northeast, as shown by correlations between high pollen PCA values (Fig. 13) and inferred high lake-levels (Potts et al., 2018). These reversible changes are unlikely to have been driven by uplift or subsidence, which tends to be directional in rift settings, especially on the timescales involved.

In contrast, the geochemical data (Figs. 9 and 13), discussed below, include episodes when abrupt step-like changes occurred that might reflect tectonic controls. For example, a major transition at ~740 ka led to an increase in anoxic-euxinic elements (Mo, Cd, U, S) in the sediments. LOI percentages also increased, pollen started to be well-preserved, La/Lu ratios became more variable and transition metals appeared in the geochemical record. A second step-like change occurred at ~105 ka ago with increased Na and the first appearance of trona in cores.

Owen et al. (2018b) documented several broad relationships between rifting and sedimentation across the East African Rift System, observing, for example, that axial subsidence can cause rainshadow effects which increase aridity on the rift floor. Fig. 14 shows major controls on Magadi basin sedimentation, emphasising the role of tectonics. The Magadi Rift was initiated at ~7 Ma (Foster et al., 1997; Lee et al., 2017). The oldest deposits are buried below the Magadi Trachytes (1.4–0.8 Ma). Early uplift along the rift margins would have cut off lateral drainage into the rift (Fig. 14A) restricting sediment supply.
reducing clastic input, and changing bedrock source lithologies and elemental compositions (solid and aqueous) delivered to the basin. Changes in subsidence and uplift would also have altered erosion, transportation and deposition rates that, in turn, would have controlled when rocks were exposed to weathering and its intensity.

Prior to the step-like change at ~740 ka, the volume of water inflow and sediment infill were broadly equal to the evolving accommodation (subsidence and compaction) in the rift grabens, producing ‘balanced-filled’ lakes (cf. Carroll and Bohacs, 1999). The earliest lakes were di- lute to moderately saline, locally accumulating carbonate grainstone. After ~740 ka, the lakes became ‘underfilled’ when horsts and grabens fragmented the formerly flatter rift-floor. As axial- rift faulting pro- ceeded and the climate became more arid, potential accommodation exceeded the volume of water and sediment supply. Closed hydro- logical basins then enabled the development of increasingly saline, alkali- naline lakes with anoxic bottom waters.

The presence of chert in the oldest Magadi sediments implies that early rising thermal fluids, and runoff with compositions derived from silicate weathering, contained enough silica for siliceous deposits to form after evaporation, evapotranspiration, dilution at a chemical in- erts, cooling or microbial biodegradation. Crustal thinning above an elevated asthenosphere might have enabled faults to tap deep geo- thermal reservoirs (Fig. 1B) but subaerial sinter deposits, linked to elevated asthenosphere might have enabled faults to tap deep geo- thermal reservoirs (Fig. 1B) but subaerial sinter deposits, linked to boiling water, are rare in the south Kenya Rift. Palaeosinters (fossil geysers) are known only at Eremet ~15 km northeast of Olor- gesailie (Owen et al., 2014). Most modern hydrothermal fluids at Maga- gadi contain Na > Ca and abundant HCO3- and CO2. These frown for- mation of Na-CO3-SO4-Cl brines with a high pH after evaporation (Eugster, 1970, 1980, 1986). Those fluids have exerted a major in- fluence on deposition, especially since 55 times (after ~380 ka) when salinities increased and alkaline carbonate-rich fluids reacted with volcanoclastic particles to form zeolites. However, the contemporary alkaline water did not precipitate extensive trona until the second step- like change in the geochemical profile at about 105 ka ago.

That event might have involved a tectonic control. Earman et al. (2005), for example, emphasised the role of magmatic CO2 in producing extensive deposits of trona in the USA and Mexico. Renaud and Tiercelin (1994) had earlier proposed that geothermal CO2 contributed to trona formation at Lake Bogoria, Kenya. Gaseous CO2 is trapped today below an extensive nafocolite (NaHCO3) crust on the floor of Nasikie Engida, a small lake northwest of Lake Magadi (Fig. 1A). Darling et al. (1995) showed that abundant CO2 of mantle origin is issues along the rift floor. Lee et al. (2016, 2017) reported that ~4 Mt yr−1 of mantle-derived CO2 is released along faults in the Magadi–Natron Basin. It is possible, therefore, that formation of trona at Magadi was enhanced by CO2-enriched gas discharge into the lake along sublacustrine faults, in addition to strong evaporation in an underfilled basin.

The origins of the Magadi chert deposits remain enigmatic (Eugster, 1967, 1969; Hay, 1968; Behr, 2002; Brenna, 2016; Leet et al., 2016). Eugster (1980) attributed the Magadi chert to replacement of magadiite [NaSiO4(OH)2·4H2O], other sodium silicates, or Na-silicate gels in the HMB sediments and at Nasikie Engida. In outcrop, most bedded chert in the Oloronga Beds and Green Beds are concordant within la- custrine sediments, whereas other cherts are intrusive, forming dykes, irregular masses and mounds (some biothermal?). Intrusions of silica- rich fluid, soft and semi-lithified silica gel, and chert into older shallow sediments are probably related to tectonic events (Behr and Röhricht, 2000). Some bedded chert in outcrop provides sedimentological evi- dence that implicates microbial influences in its formation and shows that the quartz might originally have been soft gelatinous silica (Behr, 2002; Brenna, 2016). Evidence from the HSPDP cores implies several origins for the chert, but some chert horizons were clearly diagenetic (Leet et al., 2016). Behr (2002) reported that some exposed chert might have replaced carbonate. Rare examples of carbonate replacement by chert are exposed ~10 km SSE of Magadi townsite. High aqueous silica re- sulting from weathering (silicate hydrolysis of volcanic rocks), hydrothermal inflow, and highly alkaline (high pH) brines makes it difficult to differentiate the origins of the parent fluids and specific factors leading to silica precipitation (e.g., evaporation, fluid mixing, cooling and replacement of carbonate in various combinations). These processes nonetheless have produced what is perhaps the most ex- tensive outcrop of lacustrine chert in the world.

6. Conclusions

The Magadi Basin preserves a one-million-year record of aquatic deposition under dry, tropical conditions. The basin experienced pro- gressive increases in aridity superimposed on wet-dry cycles and step- like changes that resulted from tectonic processes. Major tectonic controls include:

- Axial rift faulting that tapped geothermal fluid reservoirs, introduc- ing silica via springs from early in the basin history.
- Faulting of the rift floor that diverted cross-rift (E-W) rivers that reduced inputs of transition metals derived from rift-paramarginal ba- salts after 540 ka.
- The addition of magmatic CO2 to evaporated soda brines that en- abled thick trona deposits to form after 105 ka.
- The development of a horst-and-graben topography that modified accommodation space and which confined the palaeolake to its present narrow N-S axial setting after deposition of the Oloronga Beds.

Climatically, the basin was characterised by many wet-dry cycles, but with a trend towards increasing aridity after 575 ka. This increased evaporative concentration of spring, stream and lake waters, resulting in:

- Higher pH and salinity. Mildly saline, calcite-precipitating, waters developed before 380 ka. Later high-pH waters reacted with volca- niclastic grains to form zeolites; trona precipitated after ~105 ka.
- Increased REE instability, with these elements complexing with carbonate as waters became more alkaline after 400 ka and espe- cially after 105 ka.
- Saline lakes that received episodic fresh waters, leading to mixed assemblages of saline and freshwater planktonic diatoms between 545 and 16 ka.

The Quaternary sediments of the Magadi Basin record long-term climatic change (increasing aridity) and contemporary tectonics that frequently modified the hydrology and hydrogeology. The sedimentary record confirms the importance of considering all potential environ- mental controls when performing environmental reconstructions, especially in tectonically active settings.

Acknowledgements

Drilling was funded by ICDP and NSF grants (EAR-1123942, BCS- 1241897, and EAR-1425953). Analyses were supported by the Hong Kong Research Grants Council (HKBU-201912 and 12304018). We thank the National Museums of Kenya, the Kenyan National Council for Science and Technology, the Kenyan Ministry of Mines, and the National Environmental Management Authority of Kenya for providing permits. We also thank DOSECC Exploration Services for drilling super- vision, the Operational Support Group of ICDP for downhole logging and the US National Lacustrine Core Facility. Tata Chemicals Magadi Limited and the Magadi Administrative District of Kajiado County provided local support. Research was undertaken with support of the local Maasai community. The late Jean-Jacques Tiercelin, with John Ego (NOCK) and George Muia, helped to log outcrops. Luis Buatois provided help with description of the ichnofossils. This is publication # 18 of the Hominin Sites and Paleolakes Drilling Project.
Appendix A. Supplementary data

Geochemical and diatom data in Tables S1 and S2. Supplementary data to this article can be found online at https://doi.org/10.1016/j.palaeo.2019.01.017.

References


Amsterdam, pp. 177–224.


