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The late Pleistocene-Holocene African Humid Period as Evident in Lakes

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Summary

From the end of the last glacial stage until the mid-Holocene, large areas of arid and semi-arid North Africa were much wetter than present, during the interval that is known as the African Humid Period (AHP). During this time, large areas were characterized by a marked increase in precipitation, an expansion of lakes, river systems and wetlands, and the spread of grassland, shrub land and woodland vegetation into areas that are currently much drier. Simulations with climate models indicate that the AHP was the result of orbitally-forced increase in northern hemisphere summer insolation, which caused the intensification and northward expansion of the boreal summer monsoon. However, feedbacks from ocean circulation, land-surface cover and greenhouse gases were probably also important.

Lake basins and their sediment archives have provided important information about climate during the AHP, including the overall increases in precipitation and rates, trajectories and spatial variations in change at the beginning and the end of the interval. The general pattern is one of apparently synchronous onset of the AHP at the start of the Bølling-Allerød interstadial around 14,700 years ago, although wet conditions were interrupted by aridity during the Younger Dryas stadial. Wetter conditions returned at the start of the Holocene around 11,700 years ago covering much of North

Africa and extended into parts of the southern hemisphere (SH), including southeastern (SE) Equatorial Africa. During this time, the expansion of lakes and of grassland/shrub land vegetation over the area that is now the Sahara Desert, was especially marked. Increasing aridity through the mid Holocene, associated with a reduction in northern hemisphere (NH) summer insolation, brought about the end of the AHP by around 5000-4000 years before present. The degree to which this end was abrupt or gradual and geographically synchronous or time transgressive, remains open to debate. Taken as a whole, the lake sediment records do not support rapid and synchronous declines in precipitation and vegetation across the whole of North Africa, as some model experiments and other palaeoclimate archives have suggested. Lake sediments from basins that desiccated during the mid-Holocene may have been deflated, thus providing a misleading picture of rapid change. Moreover, different proxies of climate or environment may respond in contrasting ways to the same changes in climate. Despite this, there is evidence of rapid (within a few hundred years) termination to the AHP in some regions, with clear signs of a time-transgressive response both north to south and east to west, pointing to complex controls over the mid-Holocene drying of North Africa.

Keywords Lake sediments; Palaeoclimate; African/Indian Monsoon; Lake-level variations; Late Pleistocene; Holocene

Introduction

The nature of the African Humid Period

Since the last glacial maximum (LGM), around 21,000 years ago[1], the climate and environment of Africa have undergone dramatic change, mainly related to shifts in effective moisture (precipitation minus evaporation, or P-E). Abrupt changes in effective moisture during the transition from the last

[1] Dates quoted in this chapter are in calendar years before present unless otherwise stated

glacial into the Holocene broadly mirrored the temperature changes that took place in the mid to high latitudes of the northern hemisphere, whereas the hydrological shifts that have characterized the Holocene have generally been more dramatic, especially when compared to the oxygen-isotope records from central and northern Greenland ice cores. The period of increased effective moisture that started at the end of the last glacial and continued, albeit with interruptions, until the middle Holocene, is especially noteworthy: this interval is known as the 'African Humid Period' (AHP) (deMenocal et al., 2000). The AHP is best noted for its association with wetter conditions and increased vegetation cover across the area of northern Africa that now comprises the Sahara and Sahel, leading to a so-called 'Green Sahara', a term attributed to the Hungarian explorer László Almásy (Almásy, 1934) following his travels in the Sahara desert and discovery of rock art depicting a very different environment in the early Holocene. The existence of past humid conditions across arid northern Africa had been known since the mid-19th century with the descriptions of rock art by the German explorer Heinrich Barth (Barth, 1890). Subsequent geological and archaeological investigations, coupled with numerical modeling experiments (Claussen et al., this volume), have established an understanding of the nature and timing of climatic changes associated with the AHP, as well as the response of ecosystems and human populations to those changes (Drake et al., 2011; Manning and Timpson, 2014). Geological investigations have included studies of lake and wetland sediments (Street-Perrott et al., 1989; Gasse, 2000), dune fields (Sarnthein, 1978) and marine sediments from the oceans adjacent to the African continent (McGee and deMenocal, this volume). The last of these sources provides information about changing vegetation (Lézine et al., 2005; Hooghiemstra et al., 2011) and land surface conditions (deMenocal et al., 2000a) on the African continent and can also provide information about potential links between ocean circulation and African climate (deMenocal et al., 2000b; liu et al., 2007; Chang et al., 2008) and so is important for our understanding of the causes of change. As a result of these multidisciplinary research efforts, much is now known about the AHP. However, significant gaps in knowledge still remain. Unresolved questions include the rapidity and geographical synchroneity of the onset, peak and termination of the AHP, and the role of atmospheric, oceanic, land-surface and vegetation feedbacks in modulating the response of African climate to gradual, low-amplitude orbital forcing, the underlying driver of changes (Claussen et al., this volume).

In this chapter, we review the response of lakes to the African Humid Period. Following an outline of the modern hydrological and limnological setting, we provide an overview of the sources of lake evidence used to assess the nature and timing of the AHP. We then examine the evidence for the onset, nature and intensity and eventual termination of the AHP before reviewing areas of controversy and uncertainty relating, in particular, to the spatial heterogeneity of the AHP, links between northern and southern equatorial Africa, causes of change including forcing factors and feedbacks, atmospheric circulation during the AHP and the nature and timing of AHP termination. Finally, we conclude with a summary of current understanding, highlight those key questions that remain to be addressed, and assess the ways in which work on lakes may help us answer these questions. We focus primarily on the area that has been affected by the African monsoon for all of part of the past ~15,000 years (i.e. ~8 – 30˚N) and which has been under the influence of changes in northern hemisphere insolation. We extend coverage to southern equatorial Africa, which appears to have been broadly in phase with northern hemisphere changes, thus revealing important information about climatic linkages from the northern hemisphere.

An understanding of the AHP is important for four reasons. First, it provides an opportunity to investigate climate feedback mechanisms (Foley et al., 2003; Braconnot et al., 2007; Lenton et al., 2008; Claussen, 2009). Second, it provides insight into climate teleconnections, especially between tropical and subtropical latitudes and the middle and high latitudes of the northern hemisphere and between northern Africa and southern equatorial Africa (Otto-Bliesner et al., 2014). Thirdly, it has biogeographical and cultural implications, having influenced the distribution of plants and animals, and of human populations (Drake et al., 2011; Manning and Timpson, 2014). Finally, an understanding of past changes in effective moisture may provide insights into likely future changes as a result of anthropogenic increases in greenhouse gas concentrations (Tierney et al., 2015; Dong and Sutton, 2015), despite fundamental differences in forcing compared with the AHP, which was fundamentall drive by insolation.

Modern climate system

The climate of our study area is governed by large-scale atmospheric circulation that is modified regionally by topography and lake effects in some areas. Here, a brief account of the modern climate system is provided: the reader is referred to Claussen et al. (this volume) for further information. The dominant role of large-scale circulation gives rise to a broadly zonal pattern of precipitation amount and seasonal regime (Fig. 1) (Nicholson, 1996; Gasse et al., 2008). **[insert Holmes-Fig1 here]** Along the northern margins of the continent, southward displacement of the mid-latitude winter westerlies gives a winter rainfall maximum. South of this winter rainfall belt, lying to the south of the Atlas Mountains in NW Africa but extending to the Mediterranean coast further to the east, is the zone of low rainfall coinciding with the Sahara desert. Here, occasional rainfall events are associated with westerly depressions or with the northern margins of the Intertropical Convergence Zone (ITCZ: also known as the Intertropical Front – ITF: Nicholson, 2009). The ITCZ marks the equator-ward boundaries of the northern and southern Hadley Cells, which migrate seasonally northwards and southwards in response to seasonal variations in insolation. However, only limited rainfall is associated with the ITCZ itself: instead, most rainfall is associated with the African rainbelt, which lies further to the south between the African Easterly Jet (AEJ) and the Tropical Easterly Jet (TEJ) (Nicholson, 2009). During NH summer, the African rainbelt brings rain to the subhumid to semi-arid Sudano-Sahelian zone that lies to the south of the Sahara desert, associated with the Atlantic and Indian Ocean summer monsoons. Significant westward transport of moisture across the Sudano-Sahelian zone is associated with the easterly jet streams (Joseph et al., 1992). Stretching from SW Africa broadly northeasterly into East Africa, the Congo Air Boundary (CAB) is another important atmospheric feature, marking the zone of convergence between surface airflow from the Atlantic and Indian Ocean, which leads to a marked zonal pattern of precipitation variation in addition to the strong meridional gradient. Atmospheric circulation over East Africa is especially complex, with rain being supplied from three main sources: the Indian Ocean, the Atlantic, and from moisture recycled over the equatorial zone of Africa, especially the Congo Basin. As will become apparent below, understanding the modern climate patterns is important for interpretation of the lake-based palaeoclimate records.

Long-term (orbital) changes in precipitation over northern Africa are paced by changes in insolation, principally on the precessional timescale (Pokras and Mix, 1987; Kutzbach and Liu, 1997). However, these changes are over-printed by variations in high-latitude processes, especially changes in global ice volume (Ruddiman, 2014), ocean circulation (deMenocal et al., 2000a) and patterns of sea-surface temperature (Street-Perrott and Perrott, 1990). Also important are feedback effects from greenhouse gases at the hemispheric scale (Otto-Bliesner et al., 2014), to land-surface conditions, vegetation and soil moisture at a more regional level (Charney and Stone 1975; Kutzbach et al., 1996).

Geographical distribution of key lake regions in Africa

Despite relative aridity over large parts of the continent, Africa hosts a large number of lakes (Fig. 2). **[insert Holmes-Fig2 here]** Moreover, there is widespread geological evidence for the existence of numerous palaeolakes during the more humid past. Some of these were megalakes of a size to rival any on the planet today, whereas others were small. Collectively they show that there was a dramatic expansion in the number and area of lakes and wetlands during the AHP in regions such as the Sahara and Sahel, which are currently arid to sub-humid.

Larger and more permanent features of the present-day landscape include the lakes of the East African Rift Valley such as Lake Malawi and Lake Tanganyika. Lake Victoria, which is one of the East African Great lakes, is located outside the main rift but between the eastern and western branches of the valley. The East African lakes of Victoria, Albert and Edward drain into the White Nile, Lakes Tanganyika and Kivu drain into the Congo River and Lake Malawi drains into the Zambezi. The Blue Nile is fed by Lake Tana, which lies in the Ethiopian highlands. Many other East African lakes are endorheic. West Africa hosts fewer lakes and these are generally smaller and mostly endorheic. Lake Chad, despite being shallow, is the exception as it is a large lake, although also endorheic. Other key lakes that have yielded palaeoclimate records include those of the Biu Plateau in Nigeria (Salzmann et al., 2002) and Lake Bosumtwi in Ghana (Talbot and Delibrias, 1977). There is strong evidence for the existence of lakes in the early Holocene that were greatly enlarged compared to present day, such as Megachad, as well as lakes that occupied basins that are currently dry or ephemeral, such as Lake Fezzan in Libya (Armitage et al., 2007), Lake Ahnet-Mouydir in Central Algeria, and the Chotts palaeolake in Tunisia – Algeria (Drake and Bristow, 2006) (Fig. 2) The megalakes were fed by large catchments that commonly covered climatic zones outside those of the lakes themselves. A large number of smaller palaeolakes also existed in the present-day dryland areas of the Sahara-Sahel region.

The Sahara-Sahel lakes have water balances that are dominated by groundwater, which in some cases may be fossil, having been recharged under a different climatic regime (Gasse, 2000; Lézine et al., 2011). African lakes vary tremendously in chemical composition as well as in size, with water chemistry depending mainly on the presence or absence of outlets, which in turn determines the degree of evaporative enrichment and hence water composition. Over the long term, water composition responds to changes in effective moisture.

The development of our understanding of the AHP from lake records

Types of palaeolimnological evidence

Lake evidence for the AHP has been derived from a number of sources, including geomorphological evidence of lake-level change and investigations of lake sediments. A wealth of climate proxies has been studied (Table 1). **[insert Holmes-Table1 here]** While we do not discuss each of these proxies in detail, it is worth noting that the African lake records suffer from a number of complications to a varying degree, as outlined by Gasse (2000): these complications need to be taken into account when interpreting the records. First, many records suffer from being incomplete; they may have hiatuses in deposition during dry phases and have been truncated as a result of surface deflation following permanent drying at the end of the AHP. The latter, in particular, makes it very difficult to assess the timing and rate of aridification that has occurred with the end of the AHP, since the records will only preserve a maximum age for these events. Second, many lake records have chronological uncertainties – this problem besets sediment sequences as well as dated shoreline features – as a result of various phenomena such as reworking and reservoir effects on radiocarbon dates owing to long lake residence times and/or impacts of groundwater. Third, lake records to some extent always reflect individual conditions within the specific catchment, and therefore may not be more widely representative or suitable for regional palaeoclimate reconstruction. Fourth, a range of palaeolimnological indicators has been investigated and these are proxies for different climate variables including rainfall amount, effective rainfall (precipitation minus evaporation, or P-E), vegetation, land-surface conditions and atmospheric circulation: they should not all be interpreted in terms of rainfall amount. Different proxies may also have different sensitivity to changes in rainfall amount, for example, and some may show threshold responses to a change in moisture. In short, it is unreasonable to expect each climate proxy and each site to deliver the same response to any given change in climate. Finally, many lakes are groundwater fed, and so may show a lagged response to an increase in rainfall at the start of a wet phase as aquifers fill after preceding drought (Lézine et al., 2011).

Our overview of geographical and temporal patterns is primarily based on (1) lake-status time series from published records, which constitute a new update to the records from Hoelzmann et al. (2004): we also make reference to similar compilations in the Oxford lake-level Databank (Street-Perrott et al., 1989), the hydrological status records of Lézine et al. (2011) and the palaeohydrological synthesis of Shanahan et al. (2015); (2) the detailed and relatively well dated lake-level records from selected lakes, mainly located in East Africa but also including Lakes Chad and Bosumtwi in West-Central and West Africa; (3) Gasse's (2002) compilation of oxygen isotope and diatom hydrochemistry data from lake sediment records from Sahara and the Sahel; (4) detailed and well-dated hydrogen-isotope records from East African lakes and from Lake Bosumtwi; (5) Selected continuous and well-dated records from West and North-Central Africa that cover all or part of the Holocene.

Geographical and temporal patterns

The onset of the AHP

The AHP seems to have started at broadly the same time as the northern hemisphere late glacial (Bølling-Allerød) interstadial. Immediately before, the African-Asian monsoon belt experienced extreme drought coincident with Heinrich Stadial 1 (HS1), shown very clearly in numerous marine and terrestrial archives (Stager et al., 2011). Evidence from African lakes supports the occurrence of extreme drought in HS1. Significant lowstands of East African lakes including Lakes Turkana (Bloszies et al., 2015) and Victoria (Stager et al., 2011) occurred at this time (Fig. 3) **[insert Holmes-Fig3 here]** and there is sedimentological evidence for reduced effective moisture from other lakes in the region. In West Africa, both Lakes Chad and Bosumtwi experienced lowstands (Armitage et al., 2015; Shanahan et al., 2015). Diatom-silica oxygen-isotope ($\delta^{18}O_{diatom}$) values from Lake Challa (Barker et al., 2011) and leaf wax hydrogen isotope signatures from Lakes Tanganyika and Bosumtwi (Tierney et al., 2008; Shanahan et al., 2015) (Fig. 4) are consistent with decreased precipitation amount. The Lake Malawi δD_{wax} record (Fig. 4), taken at face value, could represent an *increase* in precipitation and thus imply antiphasing between the climate of northern and southeast equatorial Africa during HS1. However, Konecky et al. (2014) reject this suggestion and propose, instead, that the δD composition of precipitation, and hence that of terrestrial plant leaf wax, is determined primarily by air mass and precipitation source in this climatically complex region, with precipitation amount playing only a secondary control. They therefore conclude that changes in atmospheric circulation leading to shifts in dominant moisture occurred during HS1, and note that other proxies for effective moisture in Lake Malawi suggest drier conditions in the region at the time (Stager et al., 2011). **[insert Holmes-Fig4 here]**

Effective moisture seems to have increased abruptly and near-synchronously at the end of HS1, extending from northern Africa as far south as the latitude of Lake Malawi (9-14˚S) (Otto-Bliesner et al., 2014): further south, variations in moisture were apparently out of phase (Burrough and Thomas, 2013). Large and seemingly abrupt rises in lake level occurred across the region (Fig. 3). During an interval that probably coincides with the Bølling-Allerød interstadial, the Ziway-Shala palaeolake system in east Africa rose to over 70 m above the 1972 level of Lake Shala (Gillespie et al., 1983), one of its present-day constituent waterbodies. Similarly, Lake Turkana rose to more than 70 m above present lake level between about 14.5 and 13 ka, although the timing is not very well constrained (Bloszies et al., 2015) and Lake Nakuru rose over 200 m (Richardson and Dussinger, 1986). Sedimentological and palaeoecological evidence from these and other lakes supports the occurrence of a widespread lake highstand at this time (Gasse et al., 2008; Chalié and Gasse, 2002; Marhsall et al., 2011; Foerster et al., 2012). In West Africa, both Lake Chad (Armitage et al., 2015) and Lake Bosumtwi (Shanahan et al., 2015) underwent transgression. Lake Chad rose to the 289 m shoreline, with an area of 141,000 ± 700 km^2, significantly greater than that of the recent lake (area = 24,000 km^2 in the 1950s, for example) (Armitage et al., 2015). Lake Bosumtwi rose to around 50 m higher than present level at 14 ka (Shanahan et al., 2015). Although many records do not extend pre-Holocene, evidence for the onset of the AHP by about 14.5 ka is quite widespread.

The first systematic attempt to synthesize lake-status records, which resulted in the Oxford Lake-level Data Bank (Street-Perrott et al., 1989), provided evidence for rising lake status after the LGM but before the onset of the Holocene. Although the time-step of the data compilation is quite large (3000 years), it is likely that this rise coincided with the late glacial interstadial. Updated histograms depicting lake status time series presented here (Fig. 5) provide corroboration. Although the number of records with lake status records for the region between 8°-30°N/20°W-50°E is relatively low for the time between 14.5 to 13 ka, rising lake levels are observed for sites from East Africa and for those North African sites where the catchment includes regions with higher precipitation rates (e.g. Lake Chad) and/or for sites in mountainous regions (e.g. Chemchane) with a more positive water balance (Fig. 5). Lézine et al (2011) also note a steady rise in the number of sites classified as 'lacustrine' after 14.5 ka. **[insert Holmes-Fig5 here]**

Hydrogen isotope records of the AHP onset show spatially-variable patterns of temporal change (Fig. 4): those from East Africa, in particular, show inter-site variability. Lakes Tana, Tanganyika, and Albert show a trend to more negative δD values, the decline at Lake Challa is less clear (although more strongly expressed in the $\delta^{18}O_{diatom}$ values: Barker et al., 2011) and at Lakes Victoria and Malawi there is no change. As during HS1, it seems likely that the pattern of δD variation during the early AHP reflects complex changes in air-mass source, which are related in turn to atmospheric circulation (Konecky et al., 2014). In contrast, the clear decrease in δD at Lake Bosumtwi, in West Africa, which also coincides with a rise in lake level, may provide a less ambiguous indication of increasing rainfall amount at this time because in contrast to East Africa, the rain in this region is derived from a single source.

Younger Dryas

The humid conditions of the early AHP were interrupted by a marked drought that coincided with the NH Younger Dryas Stadial, although the evidence suggests that the drought at this time was not as intense as that during HS1. In East Africa, large and abrupt falls in lake level occurred at Ziway-Shala (~ 60 m, Gillespie et al., 1983) and Lake Abhé (~50 m, Gasse, 2000, Fig. 9), Lake Turkana (at least 25 m – Bloszies et al., 2015) and elsewhere (see summary figure in Bloszies et al., 2015) (Fig. 3). In West Africa, the Younger Dryas interval is not resolved in Lake Chad (Armitage et al., 2015) but Lake Bosumtwi fell by about 20 m (Shanahan et al., 2015) and on the Biu Plateau of NE Nigera, lower lake levels occurred at Lake Tilla (Salzmann et al., 2002). In the Manga Grasslands of NE Nigeria, the inter-dune basins were dry and characterized by aeolian sand deposition (Holmes et al., 1999; Waller et al., 2007; Wang et al., 2008). The Younger Dryas is not clearly registered in the lake status histogram (Fig. 5) as the number of sites is still low (14 – 17 sites). However a decline, represented by an increasing number of lakes showing lower lake status, separates the first pulse or early AHP from the early Holocene at c. 13.5 to 12.5 ka BP. Throughout the Younger Dryas just before the Holocene begins, the number of records is still low for the Sahara-Sahel. Changes in leaf wax δD values consistent with a reduction in rainfall occurred in Lakes Bosumtwi, Tana and Challa (also seen in the $\delta^{18}O_{diatom}$ record for this lake; Barker et al., 2011) and less clearly at Tanganyika: there is no clear Younger Dryas signal evident at Lakes Tana, Victoria or Malawi. Taken together, these observed changes suggest that shifts in atmospheric circulation that were inferred for the preceding interval continued into the Younger Dryas.

The early Holocene

The onset of the Holocene is marked by major changes in a wide range of records across northern Africa that are consistent with increasing rainfall and effective moisture. Those lakes for which well-resolved lake-level curves exist showed large and probably abrupt transgressions at this time. These include Lakes Turkana (Bloszies et al., 2015), Abhé (Gasse, 2000), Nakuru (Richardson and Dussinger, 1986) and palaeolake Ziway-Shala (Gillespie et al., 1983) in East Africa (Fig. 3). In

West Africa, the Chad Basin was at this time occupied by Lake Megachad, which was up to about 100 m deep and had an area of ~361,000 km^2 (Armitage et al., 2015) and Lake Bosumtwi rose by ~80 m (Shanahan et al., 2015). On a smaller scale, the inter-dune basins of the Manga Grasslands in NE Nigeria, which had been dry during the Younger Dryas, supported lakes at the start of the Holocene (Salzmann and Waller, 1998; Waller et al., 2007; Wang et al., 2008) and provide further evidence for increased effective moisture within the present-day Sudano-Sahel zone. Patterns of lake status for 9000 ^{14}C years BP (~10,000 calendar years) indicate widespread distribution of lakes with mainly high or intermediate status across a large part of northern Africa and extending as far as ~9˚S (Street-Perrott et al., 1989) (Fig. 6). **[insert Holmes-Fig6 here]** Similar patterns can be seen in the lake-status histograms (Fig. 5). The palaeohydrological compilation of Lézine et al. (2011) shows a broadly similar pattern with an abrupt increase in lakes in the Sahara-Sahel domain although with some evidence of time-transgressive behaviour, with the increase occurring earlier in the south and later in the north (10.5 – 7.5 ka from 16-22˚N, versus 9.5 – 6.5 ka north of 23˚N). According to Gasse (2002), there were two major lacustrine phases in the Sahara, from around 10.5 to 8 ka and 7.5 to 4.5 ka, whereas in the Sahel the optimum lacustrine phases lasted from 11.5 or 10.5 until about 7 ka. The maximum lacustrine phase in Lake Chad, compiled from a composite records from the lake margin (Gasse, 2002), occurred between 7 and 5 ka, which is broadly in agreement with more recent estimates (Armitage et al., 2015). The later optimum lacustrine phase in this lake compared with elsewhere in the Sahara-Sahel, reflects a largely Sudano-Guinean source for much of the inflow (Gasse, 2002). Shanahan et al. (2015) also suggest that there was an earlier peak in Holocene wet conditions in more northerly latitudes and a later peak further south, although not all of the evidence is from lakes.

Quantitative estimates of early Holocene precipitation from pollen and lake-level data in W. Africa at 9k (radiocarbon) over Mali north of 20˚N were 150-320 mma^{-1}, accompanied by a change in albedo in the region between 16 and 24˚15'N of -0.10 to -0.14 (Street-Perrott et al., 1990): this change in precipitation accorded well with previous estimates of early Holocene precipitation increase across East and West Africa of 280 ± 138 mma^{-1} (Table 2) (Street-Perrott et al., 1990).

[insert Holmes-Table2 here]

The δD records from East African lakes generally show a clearer and more coherent pattern of change for the early Holocene than for earlier times, with all except that from Lake Malawi displaying a marked decline in δD. Whereas this change might be interpreted as a simple response to increasing rainfall amount, Konecky et al. (2014) note that the magnitude of δD variation differs markedly between sites, suggesting that there was varying sensitivity to hydrological change coupled with changes in moisture source, as for earlier intervals. A marked decrease in $\delta^{18}O_{diatom}$ values occurred at Lake Challa in the early Holocene, consistent with increased effective moisture (Barker et al., 2011).

The compilation of diatom-inferred salinity reconstruction and carbonate oxygen-isotope records from the present-day Sahara-Sahel zone from 13-32˚N (Gasse, 2002) provides further insight into the early Holocene in this region. Although some of the sites discussed suffer from uncertain chronology, there is evidence that significant flooding of the basins lagged the end of the Younger Dryas drought by up to 2000 years, a finding supported by the compilation of Lézine et al (2011), although such a lag is either shorted or even absent in the Manga Grasslands of NE Nigeria (Holmes et al., 1997, 1999; Salzmann and Waller, 1998). Combined carbonate and groundwater $\delta^{18}O$ data from the Sahara suggest that the oxygen-isotope composition of precipitation was very low during the early Holocene and that the pattern of increasing oxygen-isotope values in precipitation ($\delta^{18}O_{precip}$) along the path of the monsoon, which is evident at present day, may have been reversed (Gasse, 2002). Data from the Sahel zone are more limited, but Gasse (2002) suggested that $\delta^{18}O_{precip}$ changes were smaller or even non-existent, although this pattern is not consistent with the few data points from NE Nigera (Holmes et al., 1999), which suggest a larger change. Multiple controls on the $\delta^{18}O_{carbonate}$ values from lakes, however, can lead to large uncertainties in estimates of $\delta^{18}O_{precip}$ (Leng and Marshall, 2004).

Lake-sediment-derived pollen data provide information about vegetation during the early Holocene AHP. There have been several compilations of past vegetation patterns from pollen records (e.g. Hoelzmann et al., 2004; Lézine et al., 2011) and these show broadly similar results. There was significant expansion of Sudanian and Guinean plants from more humid regions that lie to the south, to around 16˚N in the western Sahara, coupled with a possible slight southward migration of the northern margin of the Sahara. The southern margins of the desert lay at 20-22˚N in the western Sahara and 20.5-22.5˚N in the east, whereas south of around 22˚N Sudano-Sahelian savanna prevailed (Hoelzmann et al., 2004). Despite this general pattern, multiple pollen profiles in key areas suggest that a model of northward migration of phytogeographical zones is too simple. For example, for the Manga Grasslands in NE Nigeria, Salzmann and Waller (1998) showed that the increases in Guinean plant pollen during the AHP actually represented the expansion of swamp forest in the inter-dune depressions, and that the surrounding dune fields still supported open sudano-sahelian savanna and grassland.

The dust record from marine sediments from ODP Site 658C, off the coast of Mauritania, has provided an iconic picture of the onset, duration and end of the AHP (deMenocal et al., 2000). Dust records from the lakes of the Manga Grasslands of NE Nigeria show broadly similar patterns, with low dust fluxes during the early- to mid-Holocene. However, the two localities have contrasting dust sources and transport pathways. Dust transported over Northeast Nigeria is transported by the northeasterly Harmattan winds and sourced from the Greater Lake Chad basin, which lies to the northeast, as well as from river floodplains lying to the south and east of the Manga Grasslands (Cockerton et al., 2015). The equatorial Atlantic, on the other hand, is sourced by dust that is mainly derived from the Sahara desert and transported at high level by the Saharan Air Layer (SAL).

Low early Holocene dust fluxes at Site 658C therefore reflect the existence of the vegetated surface over large parts of the western Sahara, whereas low fluxes to the Manga Grasslands lakes were the result of the Chad basin being occupied by the early Holocene megalake.

Abrupt events in the Holocene AHP

There is abundant evidence for abrupt drought events having occurred across northern Africa during the early to mid Holocene AHP (Fig. 3). Abrupt regressions of the Ziway-Shala palaeolake occurred around 8 ka and possibly around 6.5 ka. The first of these was especially marked and involved an abrupt lake-level fall of more than 100 m (Gillespie et al., 1983). In West Africa, Lake Bosumtwi fell by around 80 m at ~8ka and there is evidence of a regression of up to about 20 m of Lake Chad from 10.4 – 9.4 and 8.1 – 6.6 ka (Armitage et al., 2015). The detailed lake-level record from Lake Turkana shows a regression of at least 40m from the preceding highstand, at around 8 ka and a 90 m regression at about 6 ka, along with a series of more minor, but still quite substantial, regressions at other times in the early Holocene (Bloszias et al., 2015). If similar, detailed and high resolution lake-level records were available for other lakes it is possible that this pattern of multiple, abrupt changes in lake level in the early Holocene might be seen elsewhere, although the general patterns described for Lake Turkana accord well with lake-level evidence from across northern Africa (Fig. 3) (Street-Perrott and Perrott, 1990; Gasse, 2000). The continuous, well-dated records from the Manga Grasslands in NE Nigeria also provide evidence for multiple low-stands during the early to middle part of the Holocene, with desiccation surfaces and substantial layers of aeolian sand within the sediment records attesting to low-stands of the lakes at various times (~11200, 9000, 5500 and 4100 aBP) (Waller et al., 2007; Wang et al., 2008). Variability within early Holocene carbonate oxygen-isotope records from the Sahara-Sahel (Gasse, 2002), as well as the presence of desiccation surfaces in some of these records, also points to abrupt drought events in the early Holocene. Finally, the δD record at Lake Bosumtwi shows early Holocene variability (Shanahan et al., 2015) (Fig. 4) although other δD records do not, possibly because they are too low in resolution (e.g. Lakes Albert and Victoria) or maybe because no significant changes in air mass source occurred during abrupt events.

The end of the AHP, and after

Multiple lines of evidence indicate that the Green Sahara environment came to an end sometime in the mid Holocene. Lake status records from the Oxford lake-status databank show a change from high status of lakes over much of northern and southern-equatorial Africa at around 9,000 ^{14}C years BP (~10,000 calendar years) to predominantly low or intermediate status by levels 3000 ^{14}C years (~3200 calendar years) and with some basins showing a fall by 6000 ^{14}C years (Street-Perrott et al., 1989). The histograms (Fig. 5) show the same peak at 8.5 ka BP and a second peak at 7.75-7.0 ka BP, which is separated by a decline from 8.5 to 8.0 ka BP which probably represents the 8.2 ka BP event. Moreover, the palaeohydrological compilation of Lézine et al. (2011) shows a steady reduction in the number of lacustrine sites from a peak 8.5 ka until 3.5 ka, with a sharp fall around 7.5 ka. However, beyond this general agreement on the pattern of progressive drying over the course of the mid to late Holocene, the details of the changes, including regional synchroneity and rates of change, have been debated. Since the publication of the terrigenous flux records from marine ODP site 658C, which shows a doubling of aeolian dust flux around 5.5 ka (deMenocal et al., 2000), the end of the AHP has been assumed by some authors to be an abrupt and regionally-synchronous event (e.g. Foley et al., 2003). However, some of the lake evidence for the end of the AHP and the period after points to a more complex picture. Moreover, lake-status records from basins that dried out after the end of the AHP may provide a misleading picture of the timing and rate or aridification, because of deflation of exposed lake sediments: this needs to be borne in mind when interpreting the lake-status records. Therefore, continuous sedimentary records like Lake Yoa, located in the core region of the central Sahara around 20˚31'E, become increasingly important (Kröpelin et al. 2008; Francus et al. 2013). First signs of regional aridity at Lake Yoa are seen 5.6 ka BP and continue to increase gradually until 2.7 ka BP when the modern-day desert landscape conditions were established. Shanahan et al. (2015) note an east-west trend in the timing of the AHP end: it occurred around 4 -5 kaBP east of 20˚E, earlier than in the west. This trend is also evident in the lake-status histograms (Fig. 5), where the total number of records decreases earlier in the east (~7 kaBP) than the west (~5.5 kaBP).

Some individual lake-level records do show abrupt and large regressions around the middle Holocene, although the timing of the events is variable. At Lake Ziway-Shala, there was a c. 100 fall in level from the mid-Holocene highstand between about 5.5 and 4 ka. Earlier significant regressions are seen in Lakes Suguta, Nakuru and Abhé beginning a little earlier, around 6ka (Fig. 3). The detailed lake-level curve for Lake Turkana reveals a more complex pattern, with multiple, rapid variations in water level between about 8 and 4-5 ka, with a major regression commencing around 5.2 ka. In West Africa, Lake Chad underwent a major regression around 5.2 ka and at Lake Bosumtwi, a major regression began after about 5-6 ka. Also, other lake-sediment data allude to significant changes in the middle Holocene. For example, at Lake Abiyata in the Ziway-Shala system, diatom-inferred conductivity rose sharply at about 5.5 ka (Gasse, 2001) and at Lake Ashenge, in the northern Ethiopian highlands, there was a shift towards drier conditions indicated in diatom and stable isotope data at about the same time (Marshall et al., 2009). In contrast Lake Tana, also in northern Ethiopia, showed a more gradual change, with evidence for aridification commencing around 8.5 ka and culminating in intense drought at 4.2 ka (Marshall et al., 2011). Interestingly, an abrupt drought event around 4.2 ka is seen across the Near East and South and East Asia (e.g. Cullen et al., 2000; Dixit et al., 2014).

The hydrogen-isotope records show variable patterns, as for earlier intervals (Fig. 4). At lakes Tanganyika and Challa, the sharp increase in δD starting around 5 ka is consistent with a decrease in precipitation at this time: the $\delta^{18}O_{diatom}$ values show a steadier increase, however (Barker et al., 2011). No sharp increase in δD is seen at Lakes Malawi and Tana and Lake Victoria shows a steady, rather than abrupt increase in δD over the course of the mid Holocene. This complex pattern is therefore interpreted in terms of changing atmospheric circulation (Costa et al., 2014). δD variations at Lake Bosumtwi are also complex, with a peak in values around 5ka followed by a decline.

Continuous pollen records across the terminal AHP show complex changes but no abrupt shifts as shown by the changes in tropical Sudanian and Guinean pollen at sites in northern Nigeria and Chad (Fig.7). The compilation of Shanahan et al. (2015), which is based partly on lake records, reveals a latitudinal trend in the timing of the AHP end, indicating that it commenced earlier in regions that lie north of the present-day monsoon limit. Latitudinally-grouped records from this synthesis show a time-transgressive end to the AHP, around 6 ka at 28˚N and near to 3 ka at about 2˚S (deMenocal, 2015). A return to wetter conditions is seen in some records after the end of the main AHP, e.g, in lake-levels (Ziway-Shala, Lake Abhé, Lake Chad, possibly Lake Tanganyika: Fig. 3), in the Bosumtwi δD record (Fig. 4) and possibly the NE Nigerian pollen data (Fig. 7).

[insert Holmes-Fig7 here]

Questions and controversies

Rate and synchroneity of onset and termination

Although the evidence for the onset of the AHP is less abundant than that for events during the Holocene, much of it indicates that the start of humid conditions after HS1 was regionally synchronous and abrupt. Similarly, other major dry 'interruptions' to the AHP, such as the Younger Dryas and the 8200-yr BPdroughts, also appear to have been regionally coherent. However, the timing and nature of the AHP termination, in the mid Holocene, are much less clear. As discussed in the previous section on 'Geographical and temporal pattens', some of the detailed lake-level records do show substantial and abrupt regressions during the middle Holocene. However, many of the lakes in question are hydrologically closed 'amplifier lakes' (Street, 1980), which are highly responsive to changes in effective moisture and which may, therefore, show a larger and more rapid response than that seen in other proxies. The lake-level compilations, on the other hand, combine a range of lake types, some of which would be expected to be more responsive to changes in effective moisture than others although the actual magnitude of change is effectively smoothed by the lake-status coding method. In short, the way in which the AHP termination is recorded in lake-sediment archives may be strongly determined by lake type. In a similar way, lake characteristics might explain the response of lake salinity and lake water isotope composition, both of which are variables that have been used to attempt to reflect climate change during the AHP. In contrast, plant leaf wax δD has the potential to capture a signal of the isotopic composition of rainfall, since the leaf waxes are derived from terrestrial plants, despite the fact that the biomolecules are preserved in aquatic sediment. That said, the hydrogen-isotope composition of precipitation (δD_{precip}) is a complex variable that is controlled by air mass source and trajectory as well as rainfall amount: hence presence or absence of abrupt changes in leaf wax plant leaf wax δD may not necessarily reflect abrupt or gradual changes in rainfall amount. This is especially the case for East Africa, which is climatologically complex. Changes in vegetation type, while related to precipitation amount, may not respond to rainfall in a linear way. In summary, it is important to consider that different proxies may react differently to a change in rainfall amount.

The balance of available evidence suggests that the termination of the AHP was ‘locally abrupt’ (deMenocal, 2015), but showed time transgressive behavior. In West Africa particularly, this seems to have occurred earlier at more northerly latitudes (Shanahan et al, 2015) and also earlier in the eastern Sahara than in the west. However, whether the ‘locally abrupt’ end to the AHP represents rapid reductions in rainfall or is in part an artifact of thresholds within the palaeoclimate proxies remains open to debate.

Forcing of the AHP

Orbital forcing

The link between orbital forcing and monsoon strength is well established (e.g. Kutzbach and Street-Perrott, 1985) and the general trends seen in African lake records are broadly consistent with insolation variations over the period from HS1 through to the late Holocene (Fig. 8). However, several uncertainties about the role of orbital forcing remain, namely (1) the timing of AHP onset, when precessionally-forced insolation was not at its peak, although rising rapidly (2) the lag between the early Holocene peak in the AHP and the peak in insolation and (3) the in-phase behaviour of lake records in southeast equatorial Africa with those in northern Africa, when insolation in the two hemispheres was in antiphase. Abrupt events in the AHP also require alternative or complementary mechanisms to explain them adequately. **[insert Holmes-Fig8 here]**

The onset of the AHP around 14.5 ka coincided with an interval when summer insolation was significantly lower than the maximum value reached in the early Holocene. Moreover, rapid onset of the AHP has to be reconciled with slow orbital forcing, although the rate of increase in insolation was at a maximum at the start of the AHP, which may be significant. Otto-Bliesner et al. (2014) argue that rapid onset was either a non-linear response to orbital forcing, or a response to the rapid re-establishment of Atlantic Meridional Overturning Circulation (AMOC) after HS1: there is some evidence for the latter, given that the Pa/Th AMOC proxy shows a very sharp increase at this time (McManus et al., 1994) (Fig. 8).

The lag between the peak of the AHP and peak insolation in the early Holocene may relate to complex relationships between the latitudinal gradients and seasonal distribution of insolation in relation to the peak monsoon season, which occurs in later summer. Alternatively, Ruddiman (2014) suggested that the delayed peak in the Holocene AHP was the result of atmospheric cooling from remnant northern hemisphere ice sheets. A further consideration relates to the groundwater-fed lakes of the Sahara and Sahel, the filling of which in the early Holocene may have been delayed as the aquifer was recharged following the dry glacial period (Lézine et al., 2011).

The latitudinal variation in the timing of AHP termination fits well with orbital forcing (deMenocal, 2015), whereas the broadly synchronous behavior of monsoon rainfall at the start of the AHP in SE Equatorial Africa with that in northern Africa cannot be explained directly by orbital forcing because SH insolation changes are in antiphase with those in the north. Otto-Bliesner et al. (2014) argue that the reactivation of the monsoon after the LGM was strongly forced by greenhouse gas concentrations; the fact that this effect extends south of the equator explains the synchronous behavior in the two regions. South of latitude ~9˚S, this effect is diminished, however, and here there is antiphase behavior with that further to the north.

Ocean circulation

Links between abrupt change in the African monsoon and AMOC are well established. Street-Perrott and Perrott (1990) noted that reductions in Sahelian rainfall seen in instrumental records are associated with cold-warm sea-surface temperature (SST) anomalies in the North-South Atlantic. They further noted that such patterns of SST anomalies were associated with inputs of freshwater into the North Atlantic in the late glacial and Holocene, which they reasoned would have reduced AMOC strength. Subsequently, deMenocal et al. (2000b) showed that Holocene cooling events with a periodicity of 1500 ± 500 years in the subtropical North Atlantic were synchronous with cooling in the northern North Atlantic. Shanahan et al. (2009) linked multi-centennial drought in subtropical West Africa to changes in SST patterns and to AMOC strength. The occurrence of major drought episodes seen in the lake records from across Africa is linked to major cooling events in the North Atlantic, especially the Younger Dryas, the 8200-yr BP event and the 4200-yr BP event. Reduced ocean – land temperature contrasts during these events led to a diminished West African monsoon, confirming that AMOC has been a major driver of abrupt change in this system, mainly through its impact on the position of the ITCZ and the African rainbelt (Chang et al., 2008). Indian Ocean

SSTs have a major impact on convection and hence rainfall over East Africa (Tierney and deMenocal, 2013): ENSO variations may also have an impact on monsoon rainfall across Africa through their impact on SSTs in both the equatorial Atlantic and Indian Oceans.

Feedbacks

Notwithstanding debates about rapid versus gradual end to the AHP, there is clear evidence of non-linear responses to gradual forcing within the African monsoon system during the period of the late glacial and Holocene. Understanding the feedbacks involved in the non-linear responses is key to our understanding of long-term monsoon behaviour. Different feedback processes operate at contrasting spatial and temporal scales: such feedbacks include ocean circulation and SSTs (Braconnot et al., 1999); vegetation and landsurface conditions, both of which affect albedo (Charney and Stone, 1975); soil moisture (Kutzbach et al., 1996); large waterbodies (Krinner et al., 2012); aquifers (Lézine et al., 2011). Greenhouse gases are also important (Otto-Bliesner et al., 2014), as discussed above in the section on orbital forcing.

Charney and Stone (1975) suggested that vegetation may be an important feedback in controlling recent Sahel drought, reasoning that anthropogenic removal of vegetation increases albedo, leading to surface cooling, a decrease in convection and hence monsoon weakening. Vegetation feedbacks have also frequently been invoked to explain an abrupt end to the AHP during the mid Holocene (Claussen et al., 1999; this volume). In this scenario, reductions in rainfall as a result of gradually-weakening precessionally-forced insolation lead to a threshold collapse in vegetation, which then has a negative feedback on monsoon strength. Feedbacks from soil moisture, which impact on surface heating (e.g. Shanahan et al., 2015), are also important; the evidence for a decrease in albedo during the AHP was discussed above in the section on 'The early Holocene'. However, given the debates about the rapidity or otherwise about the end of the AHP, as well as for other responses, the role of vegetation feedbacks in the monsoon system remains a matter for discussion.

For example, Kröpelin et al (2008) maintained that evidence for a gradual end to the AHP meant that vegetation forcing across northern Africa was weak or absent, a point disputed by Brovkin and Claussen (2008). Shanahan et al. (2015) further argued that the termination of the AHP in West Africa could be explained by the progressive southward migration of the African rain belt due to weakening insolation forcing during the mid Holocene, which would lead to locally abrupt reduction in rainfall at any given locality without the need to invoke vegetation feedback. This would be especially the case close to the northernmost limit of the monsoon, where small southward movements of the rainbelt would lead to abrupt decline in precipitation and hence vegetation cover. Tierney and deMenocal (2013) suggested that the nature of feedbacks might differ geographically, arguing that vegetation feedback would be most pronounced in the moderately vegetated Sahel and less marked in the heavily vegetated equatorial regions and more sparsely vegetated Eastern Sahara. They suggested that weakened vegetation feedbacks in the equatorial region might explain the muted response to the end of the AHP in that region.

The presence of large lakes in northern Africa during the early Holocene can also have a number of feedback effects. For lake Megachad and the other North African megalakes (Drake and Bristow, 2006) the presence of large expanses of open water during the early Holocene may have increased precipitation over the western and central Sahara by as much as 50% (Krinner et al., 2012) although this has been challenged in subsequent modelling experiments (Contoux et al., 2013). Given that Lake Chad is mainly sourced by precipitation from subtropical regions to the south of the lake itself, this may represent a teleconnection between the desert region and the more humid south and a mechanism by which rainfall in the present-day desert region can be impacted by rainfall further to the south, although the magnitude of the impact clearly needs further investigation. The large lake basins also represent major dust 'hotspots' (e.g. Washington et al., 2009). Here, changes in water level can expose dust source areas, such as the Bodélé Despression, which can lead to abrupt increase in dust export. Such changes partly explain the late Holocene increases in dust flux over the lake basins of NE Nigeria, which lie downwind of the Megachad Basin and falls under the pathway of the dust-exporting Harmattan wind system (Cockerton et al., 2014).

Tropical lakes and wetlands, including those in Africa, play an important role in modulating concentrations of greenhouse gases, especially methane (Street-Perrott, 1993; Loulergue et al., 2008). The pattern of change in atmospheric methane during the late glacial to early Holocene strongly mirrors changes in hydrology across the topics as a whole (Fig. 8). The high concentration of methane in the early Holocene is consistent with the large expansion of lakes and wetlands at this time, although the occurrence of the Holocene methane peak at the very start of the Holocene is hard to reconcile with suggestions of a lag in the AHP onset after the Younger Dryas (Gasse, 2002; Lézine et al., 2011) if Africa were indeed a major methane source at this time.

Clearly feedbacks are an important component of the African climate-environment system that may operate differently in different regions and on different timescales. However, the exact details of how they operate have yet to be fully understood.

Atmospheric circulation during the AHP

The AHP is classically linked to increased northward penetration of the ITCZ during the northern hemisphere summer monsoon season as a result of orbitally-forced summer heating. Although there is strong evidence that the ITCZ did migrate northwards in boreal summer during the AHP, the palaeoclimate evidence suggests that this is only a partial explanation of the reconstructed changes in climate. Moreover, much of the region's rainfall is not in fact directly associated with convergence along the ITCZ, but with the African rainbelt that lies to the south (Nicholson, 2009), as discussed in the section on 'the modern climate system' above.

Shanahan et al. (2015) have pointed out that long-term variations in rainfall depend on monsoon intensity at any given location, as well as the latitudinal position of the African rainbelt. The former should have a broadly even effect spatially, whereas changes in the latter will have a time-transgressive effect. They argue that changes in both explain the patterns seen in palaeoclimate records from West Africa, including the termination of the AHP earlier in the north than the south and the secondary precipitation maximum seen in the late Holocene as the rainbelt migrates south, as discussed in the previous section on ‘feedbacks’. Collins et al. (2011) suggest that oscillations in the seasonal range of the African rainbelt explain coherent changes in precipitation in SW equatorial Africa and northern Africa; whether this explanation holds for East Africa is uncertain.

Stable isotopes provide an excellent means of investigating temporal changes in moisture source and hence atmospheric circulation. Over East Africa there are three main moisture sources, each with distinctive isotopic signatures. Costa et al. (2014) argue that changes in δD_{wax} values in East African lakes are strongly controlled by changes in Indian Ocean (high δD_{precip}) values versus recycled, and hence more D-deplete, Congo Basin moisture sources. The latter are ultimately derived from the Atlantic (high δD_{precip}), reflecting in turn changes in the position of the CAB (Tierney et al., 2011). Pachur and Hoelzmann (2000) as well as Gasse (2002) point to very low inferred in $\delta^{18}O_{precip}$ values over the Sahara during early Holocene, and suggested that these were the result of intensification of squall-line showers along the ITCZ in the Southern Sahara at this time (Fontes et al., 1993): squall-line precipitation provides about 90% of the present-day precipitation of the Sahel region (Peters and Tetzlaff 1988). Evidence for this control operating in the later Holocene is also seen in the NE Nigerian Sahel. Here, a 3 ‰ increase in the $\delta^{18}O$ of lake water at a time when the lake was freshening is best explained by a rise in the $\delta^{18}O_{precip}$, possibly as a result of the weakening of convection along the ITCZ after the end of the main AHP (Fig. 9) (Street-Perrott et al., 2000). **[insert Holmes-Fig9 here]**

Conclusions and unresolved questions

Lake evidence has contributed significantly to our understanding of the AHP. It shows, in broad agreement with evidence from other palaeoclimate archives (McGee and deMoenocal, this volume) and with modeling (Claussen et al., this volume), that the AHP began abruptly after HS1, as insolation was rising. As a result of insolation forcing, the AHP was accompanied by large increase in rainfall across large parts of northern Africa and it extended into southeast equatorial Africa, most likely as a result of feedbacks from increased greenhouse gase concentrations. Following widespread tropical drought during the Younger Dryas, the AHP resumed at the start of the Holocene, although the timing of the peak in the AHP and also its end, seem to have varied geographically occurring earlier at more northerly latitudes. There is good evidence from lake sediment records and other archives, especially from marine sediments (McGee and deMenocal, this volume), that the end of the AHP occurred abruptly at some individual sites although whether this is a result of an abrupt decline in rainfall amount or an artifact of threshold responses of the proxy records, is somewhat unclear. Although feedbacks are clearly an integral component of the African monsoon system, their role in abrupt regime shifts remains open to debate. There is gathering evidence, especially from stable isotopes, that significant changes in atmospheric circulation occurred during the AHP and after: the compilation of further isotope records from lake systems has potential to shed further light on atmospheric circulation changes, and this will enhance our understanding of the past behavior of the African monsoon.

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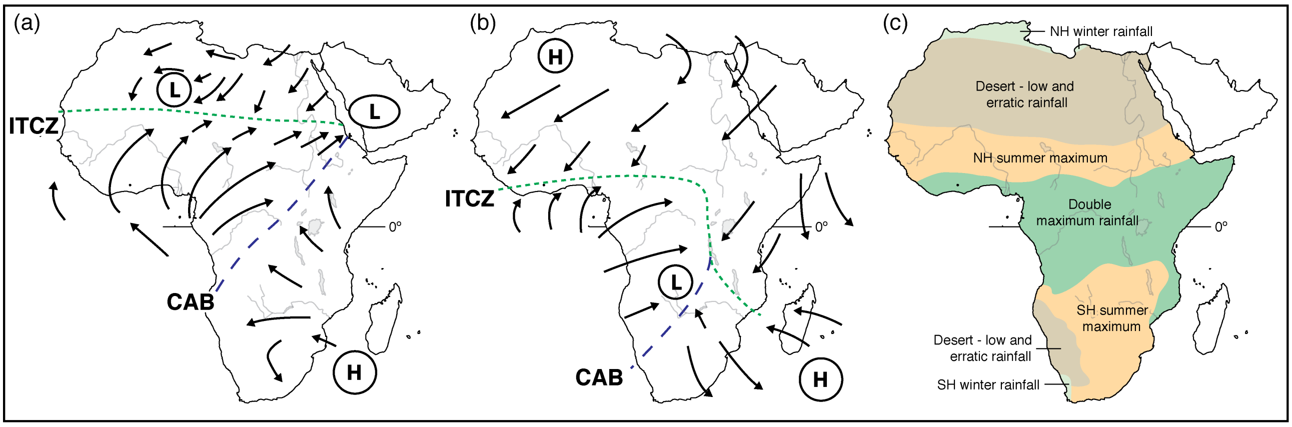

Fig. 1. Modern climatic setting. Atmospheric circulation patterns over Africa during (a) northern hemisphere summer and (b) northern hemisphere winter. ITCZ = inter-tropical convergence zone; CAB = Congo Air Boundary. L and H are major centres of surface low and surface high pressure, respectively. Arrows show surface winds (adapted from Nicholson, 1996). (c) Distribution of rainfall regimes over Africa (redrawn from Verschuren et al., 2009).

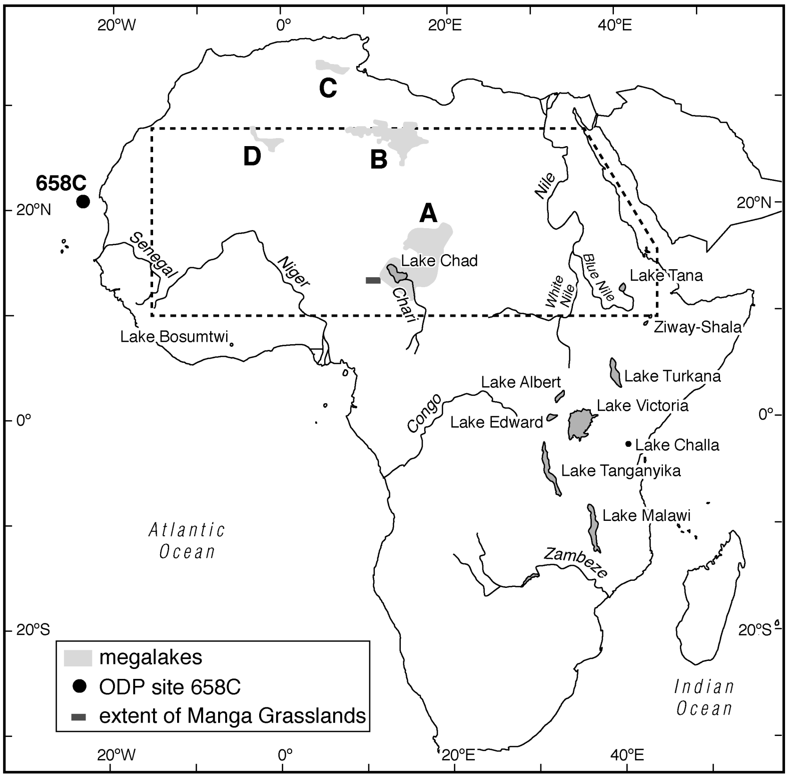

Fig. 2. Map of the study area showing major lakes and river systems and deep ocean-core from ODP site 658C. Shading shows approximate extent of megalakes during the AHP: A: Megachad, B: Lake Fezzan, C: Chotts palaeolake, D: Lake Ahnet-Mouydir. The location of the Manga Grasslands is also shown. Dashed line delimits the area within which the basins represented in Fig. 5 are located.

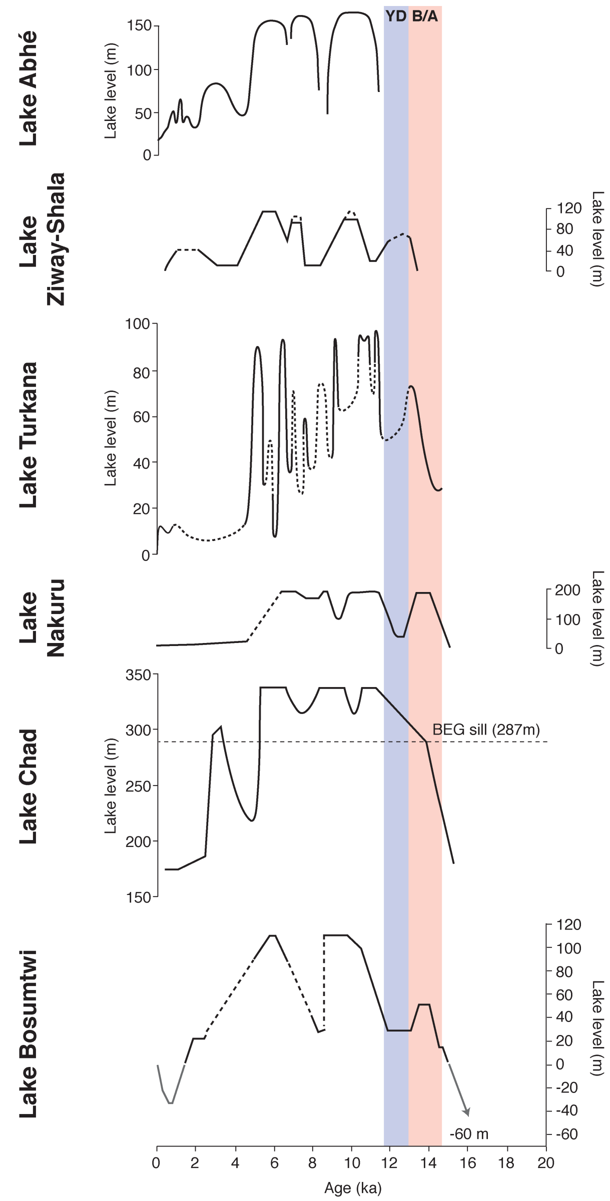

Fig. 3. Compilation of lake-level curves for African lakes. Data from (top to bottom) Lake Abhé (original curve from Gasse et al., 2000, replotted versus calendar age in Verschuren et al., 2009); Lake Ziway-Shala (original curve from Gillespie et al., 1983, replotted versus calendar age in Bloszies et al., 2015); Lake Turkana (Bloszies et al., 2015); Lake Nakuru (original curve from Richardson and Russinger, 1986, replotted versus calendar age in Bloszies et al., 2015); Lake Chad (Armitage et al., 2015); Lake Bosumtwi (Shanahan et al., 2015). Shaded intervals labeled YD and B/O refer to the Younger Dryas and Bølling/Allerød, respectively. BEG refers to the Bahr el Ghazal Channel, which links the southern basin of Lake Chad to the Bodélé Depression during times of low lake level.

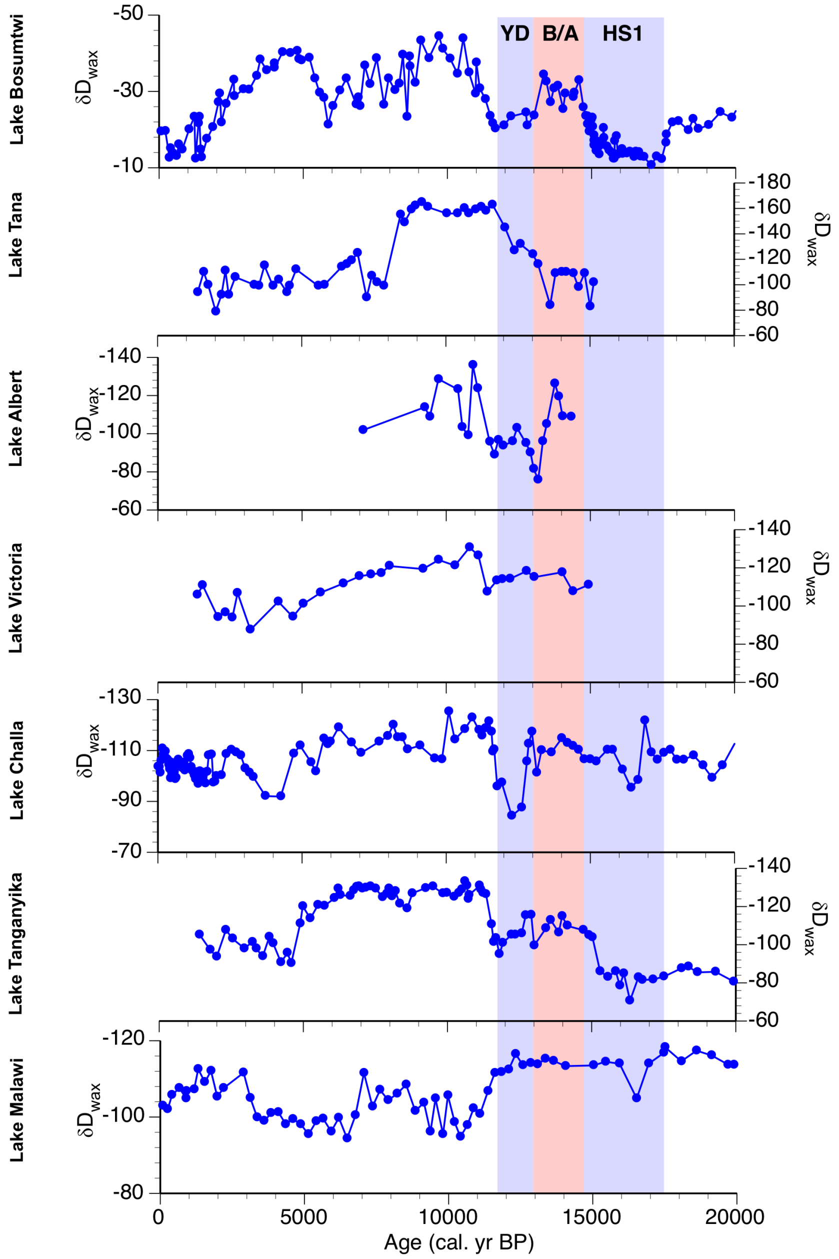

Fig. 4. Compilation of hydrogen-isotope records from African lakes. Data from (top to bottom) Lake Bosumtwi (Shanahan et al., 2015); Lake Tana (Costa et al., 2014); Lake Albert (Berke et al., 2014); Lake Victoria (Berke et al., 2012); Lake Challa (Tierney et al., 2010); Lake Tanganyika (Tierney et al., 2008); Lake Malawi (Konecky et al., 2011). Shaded intervals labeled YD, B/O and HS1 refer to the Younger Dryas, Bølling/Alllerød and Heinrich Stadial 1, respectively.

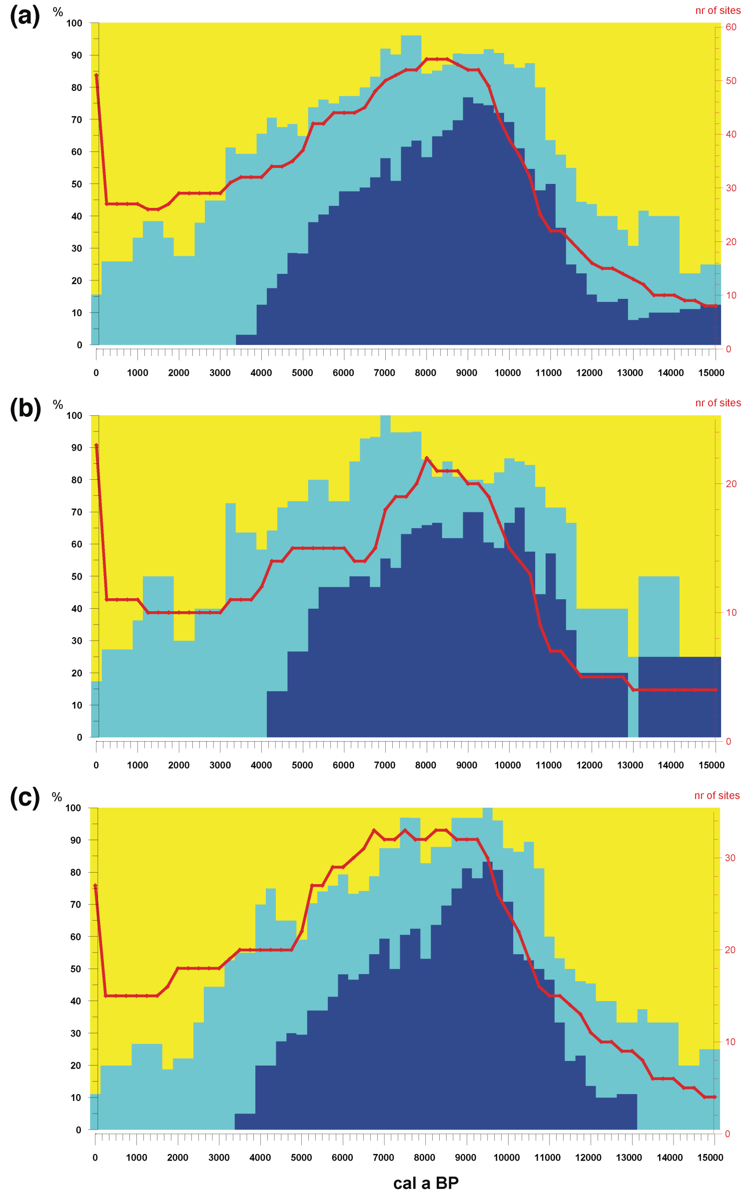

Fig. 5. Histograms of lake status for (a) all sites between 10°N to 27°N; (b) “east”: all sites between 10°N to 27°N and between 15°E to 43°E; (c) “west” all sites between 10°N to 27°N and between 12°W to 15°E. All histograms show collapsed codings (yellow = low status; light blue = intermediate status; dark blue = high status) for every 250 calendar year time-step between 0 and 15 cal ka BP. The red line indicates the number of sites represented in each time slice. The boundaries of status classes were set so that for each lake record the class “high” corresponds to the upper quartile and “low” to the lower quartile of that lake’s variation in level during the entire period. Methodology for lake-status coding follows Yu et al. (2001).

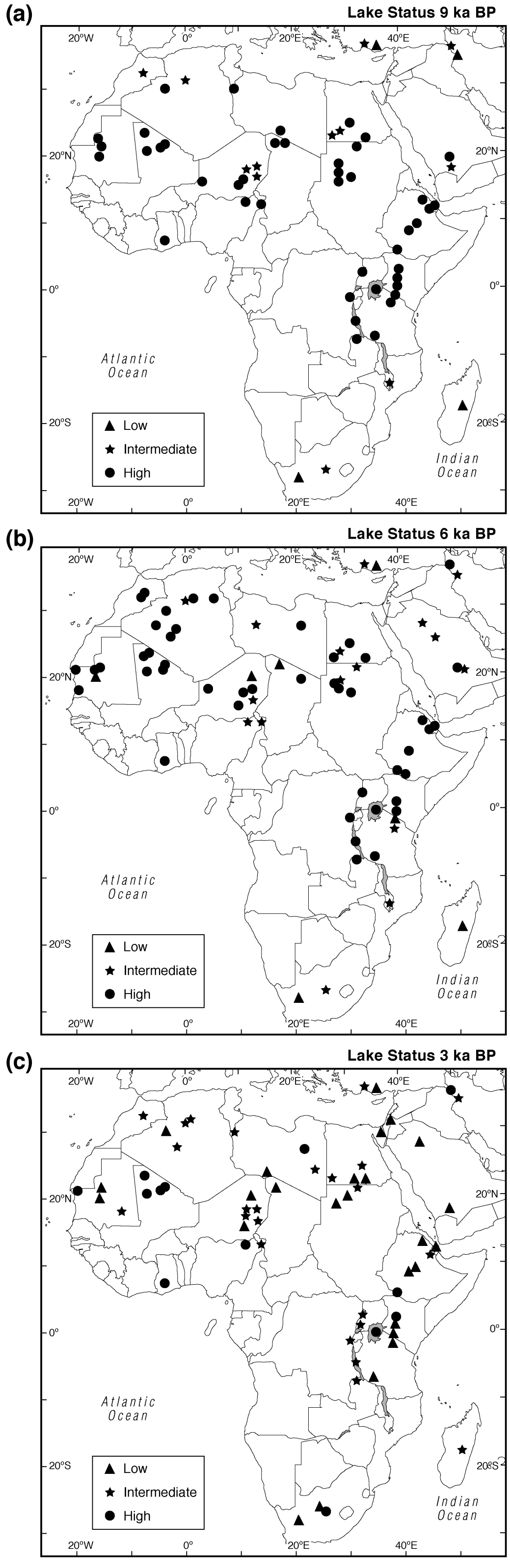

Fig. 6. Lake status at (a) 9000, (b) 6000 and (c) 3000 ^{14}C yrBP: data from the Oxford lake-level databank (redrawn from Street-Perrott et al., 1989).

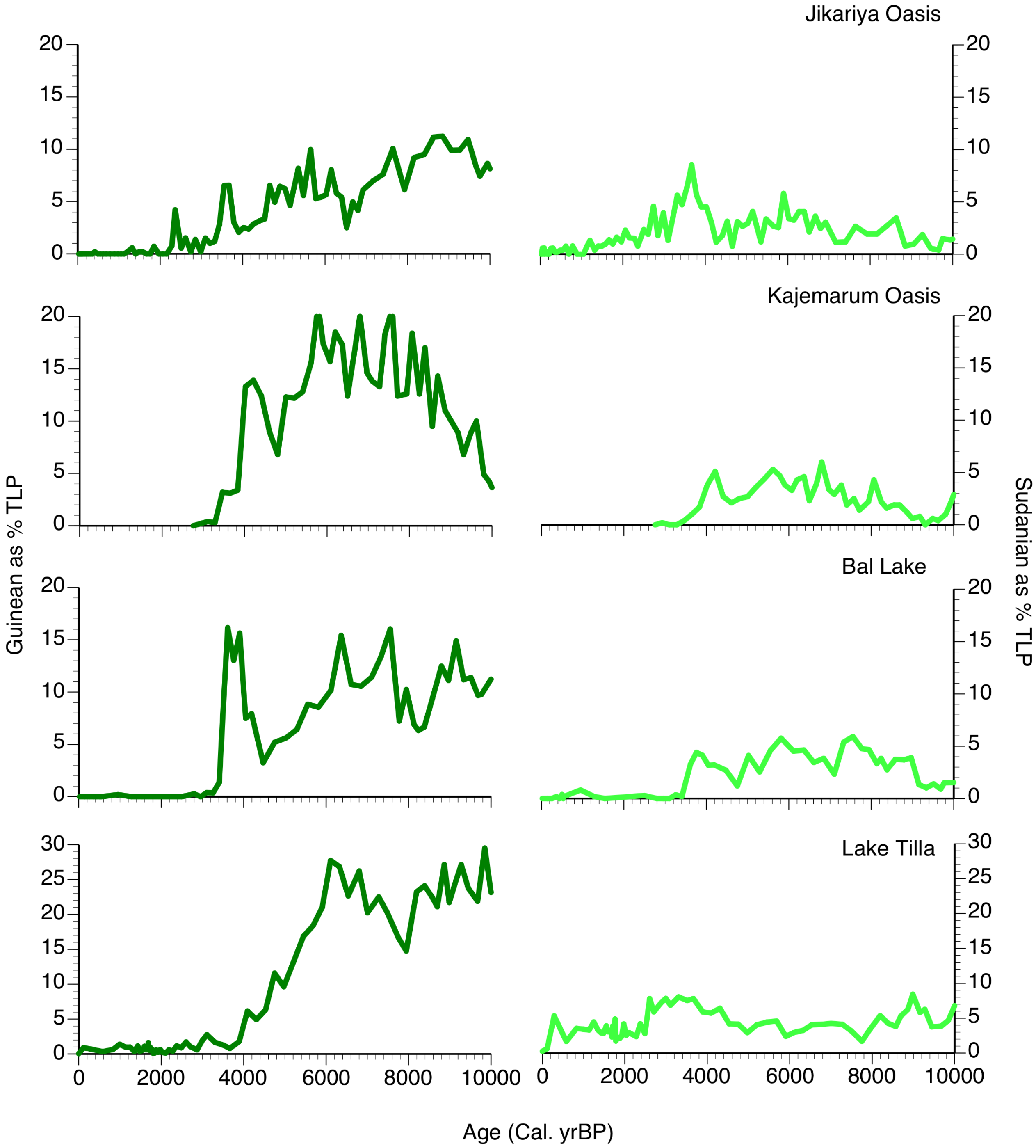

Fig. 7. Guinean and Sudanian pollen taxa as a percentage of total land pollen (TLP) in selected sites from northern Africa (data from Salzmann and Waller, 1998; Salzmann et al., 2002).

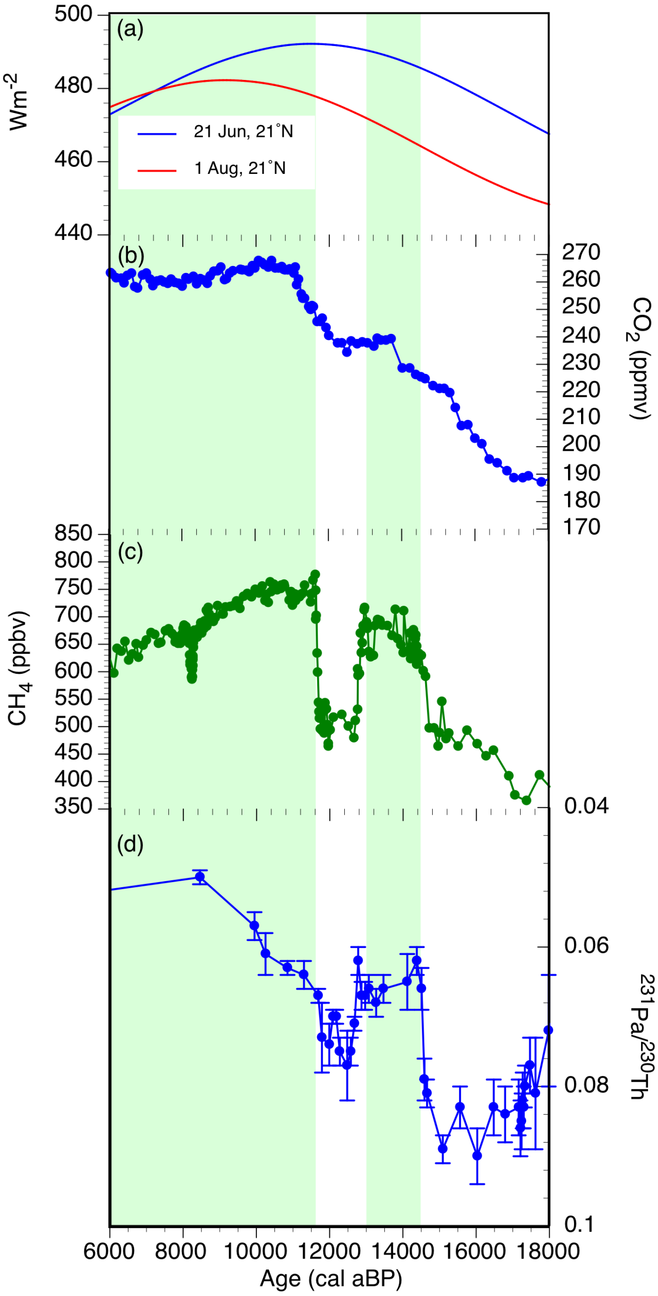

Fig. 8. Late glacial through mid-Holocene forcing of the AHP. (a) Insolation values, computed using AnalySeries 2.0.4.2 (Paillard et al., 1996) (b) carbon dioxide concentration (Monnin et al., 2001. Data from Monnin, E., et al., 2001, Dome C Last Glacial Termination Atmospheric CO2 Data, IGBP PAGES/World Data Center A for Paleoclimatology Data Contribution Series #2001-004. NOAA/NGDC Paleoclimatology Program, Boulder CO, USA) (c) methane concentration (data from Greenland GISP2 ice core, on GICC05 time scale.) (d) Pa/Th AMOC strength proxy (McManus et al., 1994). Green shading shows main intervals of AHP.

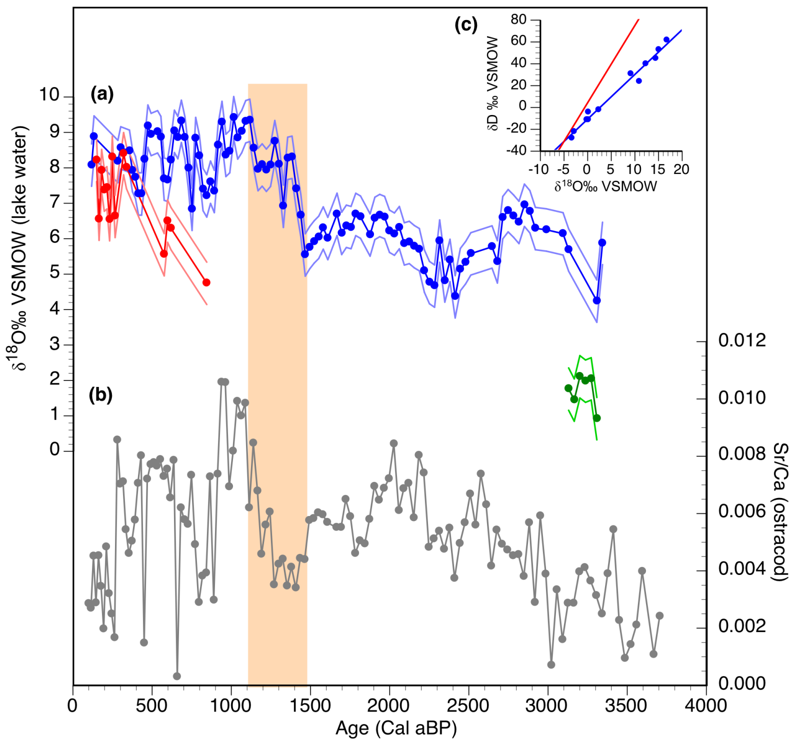

Fig 9. Late Holocene variations in inferred salinity and lakewater δ^{18}O ($\delta^{18}O_{lake}$) at Kajemarum Oasis, Manga Grasslands, NE Nigeria. (a) Reconstructed $\delta^{18}O_{lake}$ values from vital-offset-corrected oxygen-isotope determinations of ostracod shells, assuming water temperatures similar to those of late 20th century (Beyerle et al., 2003). Calculations using water temperature of 27˚C (heavy lines) and the equation of Kim and O'Neil (1997); feint lines show reconstructions for 24˚C (lower line) and 30˚C (upper line). Blue = calculations from δ^{18}O determinations of shells of *Limnocythere inopinata* (vital offset = 0.73 ‰; von Grafenstein et al., 1999), red = *Heterocypris giesbrechtii* (using vital offset of 1 ‰ for *H. punctata*; Pérez et al., 2013), green = *Candonopsis* sp. (using vital offset of 2.3 ‰ for Candoninae; von Grafenstein et al., 1999). (b) Sr/Ca in shells of *Limnocythere inopinata*, a proxy for water salinity. Original data from Holmes et al. (1997), but plotted against revised age model of Cockerton et al. (2014). Shading shows interval of significant decoupling of the two time series, as discussed in the text (c) Isotope composition of modern lakes (blue) in the Manga Grasslands, NE Nigeria (data from Edmunds et al., 1999) and precipitation at Kano (red) (data from IAEA/WMO (2016). Global Network of Isotopes in Precipitation. The GNIP Database. Accessible at: http://www.iaea.org/water).

Table 1. Palaeoe nvironmental proxies from African lakes.

Variable	Proxy for	Key references
Shoreline features	Water level, water volume	Schuster et al., 2005; Ghoneim and El-Baz, 2007; Bloszias et al., 2015; Armitage et al., 2015
Lake sediments		
Physical sedimentology		
Grain size	Catchment sediment influx, catchment hydrology	Lamb et al., 2007
Dust content	Dust flux; local and regional aeolian activity	Cockerton et al., 2014
Palaeoecology		
Pollen	Vegetation	Lézine et al., 20011; Salzmann and Waller, 1998
Diatoms	Water salinity and hydrochemistry; lake level	Gasse et al., 1995; Gasse, 2002
Ostracods	Water salinity and hydrochemistry; lake level	Holmes et al., 1998
Geochemistry		
Mineral magnetic composition	Sediment inwash, dust	Marshall et al., 2011; Kröpelin et al., 2008; Wang et al., 2008
Bulk sediment geochemistry	Catchment runoff	Marshall et al., 2011
	Water salinity	Hoelzmann et al., 2010
TEX86	Water temperature	Konecky et al., 2011
Mg/Ca and $Sr/Ca_{ostracod}$	Water salinity	Gasse et al., 1987; Holmes et al., 1997
BIT index	water balance	Verschuren et al., 2009
Stable isotopes		
$\delta^{18} O_{carbonate}$	P-E, basin hydrology, $\delta^{18} O_{precip}$	Holmes et al., 1997; Gasse, 2002
δD_{wax}	$\delta D_{precip,}$ rainfall amount	Tierney et al., 2011
$\delta^{18} O_{diatom}$	P-E	Barker et al., 2011

Table 2. Palaeo-precipitation estimates for the early Holocene AHP (updated from Street-Perrott et al., 1990). Mean value is based on estimates from pollen, combined water-balance and energy balance and water-balance methods (for last of these, only those estimates with 0-2˚C cooling are included in the calculations).

Site	Latitude	Precipitation increase (mma^{-1})	Method	Assumed temeprature change (˚C)	Source
Chemchane, Mauritania	20˚56'N	440	Pollen		Lézine et al., 1990
Oyo, Sudan	19˚16'N	395	Pollen		Ritchie et al., 1985
Chad, Chad/Nigeria/Niger	13˚00'N	>300	Water & energy balance		Kutzbach, 1980
		300	Pollen		Amaral et al., 2013
Ziway-Shala, Ethiopia	7˚45'N	>450	Water balance	0	Street, 1979
		>268	Water balance	-2	Street, 1979
Bosumtwi, Ghana	6˚30'N	0	Water balance	0	Talbot & Delibrais, 1977
Turkana, Kenya	5˚00'N	>110	Water & energy balance		Hastenrath & Kutzbach, 1983
Nakuru/Elmenteita, Kenya	0˚25'S	>625	Water & energy balance	+2 to +3	Butzer et al., 1972
		>505	Water balance	-2	Butzer et al., 1972
		>280	Water & energy balance		Hastenrath & Kutzbach, 1983
Naivasha, Kenya	0˚41'S	>>225	Water balance	0	Butzer et al., 1972
		>>130	Water & energy balance		Hastenrath & Kutzbach, 1983
Victoria, Tanzania/Uganda/Kenya	1˚00'S	220	Pollen		Laseski, 1983
Manyara, Tanzania	3˚37'S	237	Water balance	0	Holdship, 1976
Northern Mali	20-24˚15'N	150-320	Vegetation		Street-Perrott et al., 1990
West Nubian palaeolake	18˚30'N	470	Water balance		Hoelzmann et al., 2000
	Mean	280 ± 138			