An abrupt change in the African monsoon at the end of the Younger Dryas

Michael R. Talbot, Maria Letizia Filippi, Niels Bo Jensen, Jean-Jacques Tiercelin

Geochemistry Geophysics Geosystems G3
AN ELECTRONIC JOURNAL OF THE EARTH SCIENCES
Published by AGU and the Geochemical Society

Article
Volume 8, Number 3
13 March 2007
Q03005, doi:10.1029/2006GC001465
ISSN: 1525-2027

Click Here for Full Article

An abrupt change in the African monsoon at the end of the Younger Dryas

Michael R. Talbot
Department of Earth Sciences, University of Bergen, Allégt. 41, 5007 Bergen, Norway (michael.talbot@geo.uib.no)

Maria Letizia Filippi
Museo Tridentino Scienze Naturali, Via Calepina 14, I-38100 Trento, Italy

Niels Bo Jensen
IRIS Research, Thormølens Gt. 55, 5008 Bergen, Norway

Jean-Jacques Tiercelin
UMR CNRS 6538 "Domaines Océaniques", Institute Universitaire Européen de la Mer, Place Nicolas Copernic, F-29280 Plouzané, France

[1] High-resolution studies of variations in the elemental and stable carbon- and nitrogen-isotope composition of organic matter in cores from Lakes Malawi, Tanganyika, and Bosumtwi (tropical Africa) indicate an abrupt change in the wind-driven circulation of these lakes that, within the limits of available chronologies, was contemporaneous with the end of the Younger Dryas in the northern hemisphere. The change was also coincident with shifts in surface winds recorded in cores from off the west and northeast coasts of Africa. A range of other proxies indicate that these changes in wind regime were accompanied by a marked increase in precipitation in the northern tropics. Africa south of $\sim$5°–10°S, on the other hand, initially suffered drought conditions. Together, the evidence suggests an abrupt northward translation of the African monsoon system at circa 11.5 ± 0.25 ka B.P. The data assembled here contribute to a growing body of work showing that the Younger Dryas was a major climatic excursion in tropical Africa. Furthermore, they add substance to recent suggestions that climatic events in the southern hemisphere may have played a significant role in the abrupt demise of the Younger Dryas.

Components: 10,210 words, 7 figures.

Keywords: paleoclimate; Africa; Younger Dryas; limnogeology; organic matter; stable isotopes.

Index Terms: 4914 Paleoceanography: Continental climate records; 4942 Paleoceanography: Limnology (0458, 1845, 4239).

Received 29 August 2006; **Revised** 1 December 2006; **Accepted** 14 December 2006; **Published** 13 March 2007.

Talbot, M. R., M. L. Filippi, N. B. Jensen, and J.-J. Tiercelin (2007), An abrupt change in the African monsoon at the end of the Younger Dryas, *Geochem. Geophys. Geosyst.*, *8*, Q03005, doi:10.1029/2006GC001465.

1. Introduction

[2] Although it has become clear that orbitally driven variations in insolation provide the overarching control on tropical African climate change [*Kutzbach and Street-Perrott*, 1985; *deMenocal and Rind*, 1993; *Kutzbach et al.*, 1996; *Kutzbach and Liu*, 1997], there is a growing body of evidence that a number of abrupt climatic events have affected tropical Africa for which there is no simple astronomical explanation [*Gasse and Van Campo*, 1994; *Overpeck et al.*, 1996; *Sirocko*, 1996; *deMenocal et al.*, 2000; *Johnson et al.*, 2004; *Barker et al.*, 2004]. Climatic thresholds due to non-linear interactions between surface vegetation cover and the ocean-atmosphere system, particularly in the Sahel-Sahara region, have been invoked as an explanation for some of these [*Claussen et al.*, 1999; *deMenocal et al.*, 2000], but it remains unclear whether similar thresholds exist in regions where ground-cover changes are less dramatic than those that have affected Africa's driest areas. More recently, it has been suggested that some of the rapid, millennial-scale events apparent in northern hemisphere climatic records may have been tied to changes in the tropical monsoons [*Rohling et al.*, 2003]. In seeking explanations for these abrupt events, it is also possible that while some are of intracontinental origin, others may be the local expression of events that were hemispherical or global in extent. In terms of northern hemisphere climatic change, one of the most notable post-last glacial maximum (LGM) abrupt events occurred at the end of the Younger Dryas (YD), circa 11.5 cal. ka B.P., when there was widespread warming, increase in snow accumulation rate, and reduction in surface wind intensity. The main period of transition may have occupied as little as 50 years [*Alley et al.*, 1993; *Björck et al.*, 1996; *Severinghaus et al.*, 1998; *Alley*, 2000; *Lotter et al.*, 2000]. It is now becoming clear that the YD event also had a significant impact on tropical Africa [e.g., *Roberts et al.*, 1993; *Beuning et al.*, 1998; *Stager et al.*, 2002; *Barker et al.*, 2004; *Johnson et al.*, 2004; *Schefuss et al.*, 2005; *Weldeab et al.*, 2005; *Garcin et al.*, 2006]. Here, using high-resolution (typically <100- to 300-year sample spacing) paleoenvironmental data from cores from African lakes, we show that within the limits of available dating methods, there was an apparently contemporaneous, abrupt change in the climate of tropical Africa at the end of the YD, a change that seems to have involved reorganization of monsoonal circulation across the continent and adjacent oceans.

2. Modern Climatic Setting

[3] Rainfall distribution and surface wind patterns over tropical Africa are primarily controlled by the Intertropical Convergence Zone (ITCZ) [*Nicholson*, 1996] which has a northerly position during northern hemisphere summer and a southerly position during winter (Figure 1). In West Africa the convergence is a relatively simple, east-west oriented feature which separates the dry, northeasterly trade winds from a moist, southwesterly airflow. In equatorial and southern Africa the situation is more complex, with three air streams interacting, the northeast and east to southeast trade winds (both dry) and the Congo (humid) with a southwesterly to westerly flow direction. Climatic conditions in this region are thus determined by the positions of two zones of convergence, the ITCZ and the Congo Air Boundary which separates the easterly and westerly airflows (Figure 1) [*Nicholson*, 1996].

3. Note on Chronology

[4] All ages are given in calibrated ^{14}C years [*Stuiver et al.*, 1998]. As will become apparent, not all records show an exact peak-to-peak correspondence, neither within and around Africa, nor with other records (e.g., ice cores). The lack of perfect correlation is to be expected. In addition to local effects, and the possibility that there are real differences in the timing of the events discussed here, there are significant uncertainties associated with dating several of the archives. In large, deep lakes there may be a measurable but poorly constrained reservoir age, and reworked organic matter (OM) can affect age determinations from sediments in all types of water bodies [*Cohen et al.*, 1997; *Barry et al.*, 2002; *Russell et al.*, 2003]. In marine cores the principal uncertainties are reservoir age, particularly from areas of upwelling [e.g., *Kim et al.*, 2002; *Farmer et al.*, 2005], and differences in the ^{14}C age yielded by different proxies from the same stratigraphic level [*Mollenhauer et al.*, 2003; *Farmer et al.*, 2005]. Furthermore, each proxy has a different response time to climatic forcing, an effect that may be particularly significant in discussions of rapid climate change [e.g., *Russell et al.*, 2003; *Beuning et al.*, 2003]. Thus we would not expect exact synchroneity in proxy records of the response of, for example, lake mixing, terres-

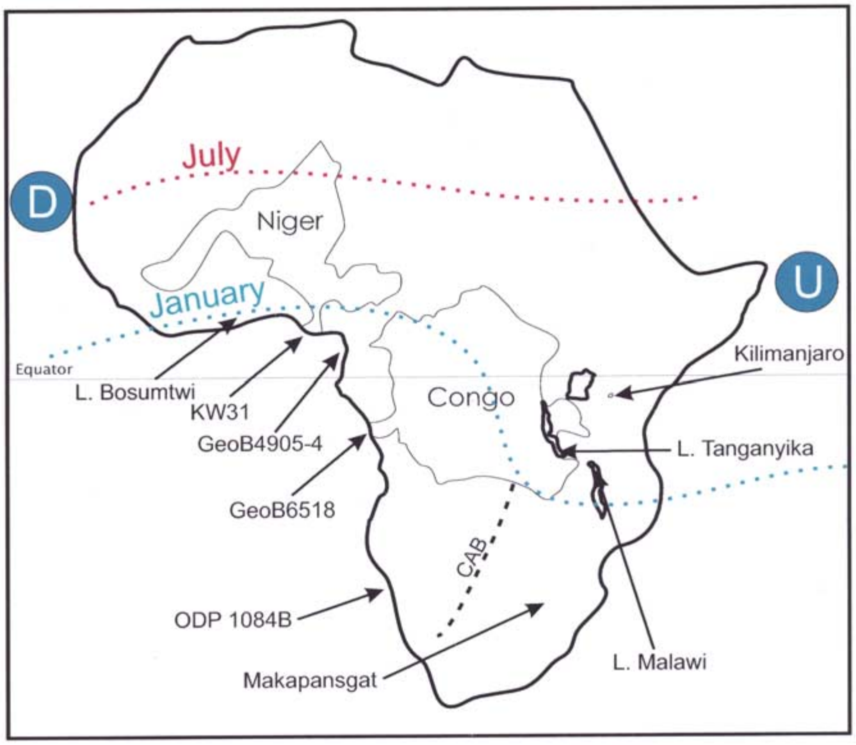

Figure 1. Principal sites mentioned in text and Figures 2–6 with outlines of Congo and Niger River catchments. "D" indicates area off the west coast of Africa where aeolian dust records provide proxy climatic information on the Sahel-Sahara region [*deMenocal et al.*, 2000]. "U" is the area of upwelling in the Arabian Sea where cores provide data on the southwest monsoon [*Sirocko*, 1996]. Also shown are the modern mean positions of the ITCZ in January and July, and the January position of the Congo Air Boundary [after *Nicholson*, 1996].

trial vegetation or sea-surface temperature to regional climate change. Nevertheless, the changes in the archives discussed here are of such a magnitude that we are confident they record responses to the same regional forcing factors.

4. Paleowind Records From African Lakes

[5] Although direct or proxy evidence of climatically driven changes in lake level have provided the most widespread evidence of past climate change in tropical Africa [*Street-Perrott and Perrott*, 1993; *Gasse*, 2000; *Barker and Gasse*, 2003; *Barker et al.*, 2004; *Hoelzmann et al.*, 2004; *Gasse and Roberts*, 2005], most lakes in the region, particularly large, deep water bodies, are also sensitive to variations in wind regime. Primary production in many of the larger lakes is dependent upon nutrient regeneration in the epilimnion, and subsequent return by vertical mixing of the regenerated nutrients into the photic zone. In the absence of marked seasonal contrasts in surface temperature, vertical mixing is mainly controlled by the wind; any change in wind direction or intensity can thus impact the rate of primary production and, depending upon nutrient availability, the composition of the phytoplankton assemblage [*Beadle*, 1981; *Hecky and Kling*, 1987; *Hecky et al.*, 1991; *Patterson and Kachinjika*, 1995; *Plisnier et al.*, 1999]. Variations in either of these may in turn lead to changes in the flux and preservation of autochthonous biogenic material at the lake floor. Fossil diatom assemblages, biogenic silica (BSi) accumulation rates, and the amount and composition of sedimentary OM have previously been used to infer past changes in the wind-driven mixing of lakes in both West and East Africa [*Haberyan and Hecky*, 1987; *Talbot and Johannessen*, 1992; *Talbot and Lærdal*, 2000; *Talbot*, 2001; *Gasse et al.*, 2002; *Johnson et al.*, 2002, 2004; *Filippi and Talbot*, 2005]. Data from Lakes Malawi and Tanganyika, East Africa, and Lake Bosumtwi, West Africa, document a major change in the mixing regime of these water bodies at the end of the Younger Dryas.

4.1. Lakes Malawi and Tanganyika

[6] Lakes Tanganyika and Malawi lie south of the equator in the western arm of the East African rift system; they are respectively the largest and next-largest of the East African rift lakes. Both lakes are strongly influenced by the seasonal migration of the Intertropical Convergence Zone (Figure 1). Vertical mixing is most intense during the austral winter when the ITCZ is located north of Lake Tanganyika and dry, strong southeasterly trade winds cause upwelling, stimulating high rates of primary production at the south end of each lake [*Coulter and Spigel*, 1991; *Hecky et al.*, 1991; *Patterson and Kachinjika*, 1995; *Nicholson*, 1996]. Although the ITCZ migrates south to the Lake Malawi region during the boreal winter, the modern northeasterly trades which characterize the north side of the ITCZ, are relatively slack in this region and have limited impact upon vertical mixing in the two lakes. At present both water bodies are meromictic with a substantial anoxic hypolimnion; complete mixing has never been observed. High rates of primary production combined with meromixis provide ideal conditions for the accumulation and preservation of phytoplankton OM in deep-water sediments. Previous studies of the elemental (C, N), stable isotopic (C, N) and Rock-Eval pyrolysis composition of this OM have provided insights into past changes in nutrient cycling which can be related to variations in the wind-driven mixing of the water column [*Talbot*, 2001; *Filippi and Talbot*, 2005; *Talbot et al.*, 2006].

4.1.1. Lake Malawi

[7] Three cores, M98-1P, -2P and -3P, have been studied in detail [*Filippi and Talbot*, 2005]. Core chronology is based upon 6 accelerator mass spectrometric dates from 1P, 8 from 2P, and 3 from 3P; additional correlation between the cores is provided by tephras and other distinctive marker beds [*Barry et al.*, 2002]. To reduce high-frequency noise, data from the three cores has been stacked [*Paillard et al.*, 1996]. In the individual cores and the stacked data, one of the most robust records is provided by the Rock-Eval Hydrogen Index (HI) (Figure 2) which provides a measure of the composition and state of preservation of the bulk organic matter. Especially striking is the abrupt fall in HI, centered at circa 11.7 cal. ka, from values in excess of 400, to $\sim$200, the lowest value for the last 25 kyr [*Filippi and Talbot*, 2005]. Coincident with the HI minimum are spikes in C/N and δ^{15}N (Figure 2). In sediments deposited prior to this excursion, stacked HI values are generally moderate to high (>350); afterward, they are generally lower (<350). TOC, C/N and δ^{15}N also tend to be higher in pre-excursion sediments than they are in those that accumulated post circa 11 ka (Figure 2) (a full discussion of the interpretation of these geochemical data is given by *Filippi and Talbot* [2005]). A lithological change accompanies the geochemical changes; sediments deposited before circa 11.5 ka are homogeneous or only vaguely laminated with some traces of bioturbation, thereafter they are finely laminated and, in some intervals, varved [*Barry et al.*, 2002]. Paleotemperature estimates based upon the TEX_{86} proxy suggest that Lake Malawi's surface waters cooled by $\sim$2°C during the YD [*Powers et al.*, 2005].

[8] High HI values (>400) indicate the presence of labile OM derived mainly from phytoplankton; low values (<250) are typical of sediments containing degraded autochthonous OM or refractory, hydrogen-poor OM derived from terrestrial plants [*Talbot and Livingstone*, 1989; *Talbot and Lærdal*, 2000]. Visual inspection of the OM assemblage indicates that although a slight increase in terrestrial plant fragments coincides with the fall in HI, these do not dominate the OM. The HI excursion is thus likely to have been caused by an abrupt change to conditions that were less favorable to the preservation of unstable OM, leaving the sediment with a mixed assemblage of degraded phytoplankton and terrestrial plant debris. Thin turbidites and an increase in periphytic diatoms also characterize this section [*Barry et al.*, 2002; *Gasse et al.*, 2002]. Previous studies have shown that Lake Malawi was low during the early Holocene [*Finney and Johnson*, 1991; *Brown et al.*, 2000], Together, the evidence suggests a large and abrupt fall in lake level from the generally highstand conditions that characterized much of the period between 18 and 12 ka [*Gasse et al.*, 2002; *Filippi and Talbot*, 2005]. This fall brought the core sites into the zone of permanent or frequent mixing where oxic conditions led to the destruction of the more unstable, hydrogen-rich OM components. The turbidites and associated spikes in C/N and δ^{15}N support the suggestion that there was an influx of terrestrial and reworked, degraded OM, probably from sediments exposed as lake level declined. Generally lowstand conditions persisted until circa 10 ka B.P. [*Filippi and Talbot*, 2005].

[9] Consistently high to very high HI and TOC values in the pre-excursion deposits are thought to reflect high rates of primary production in the north basin that were probably maintained by persistent

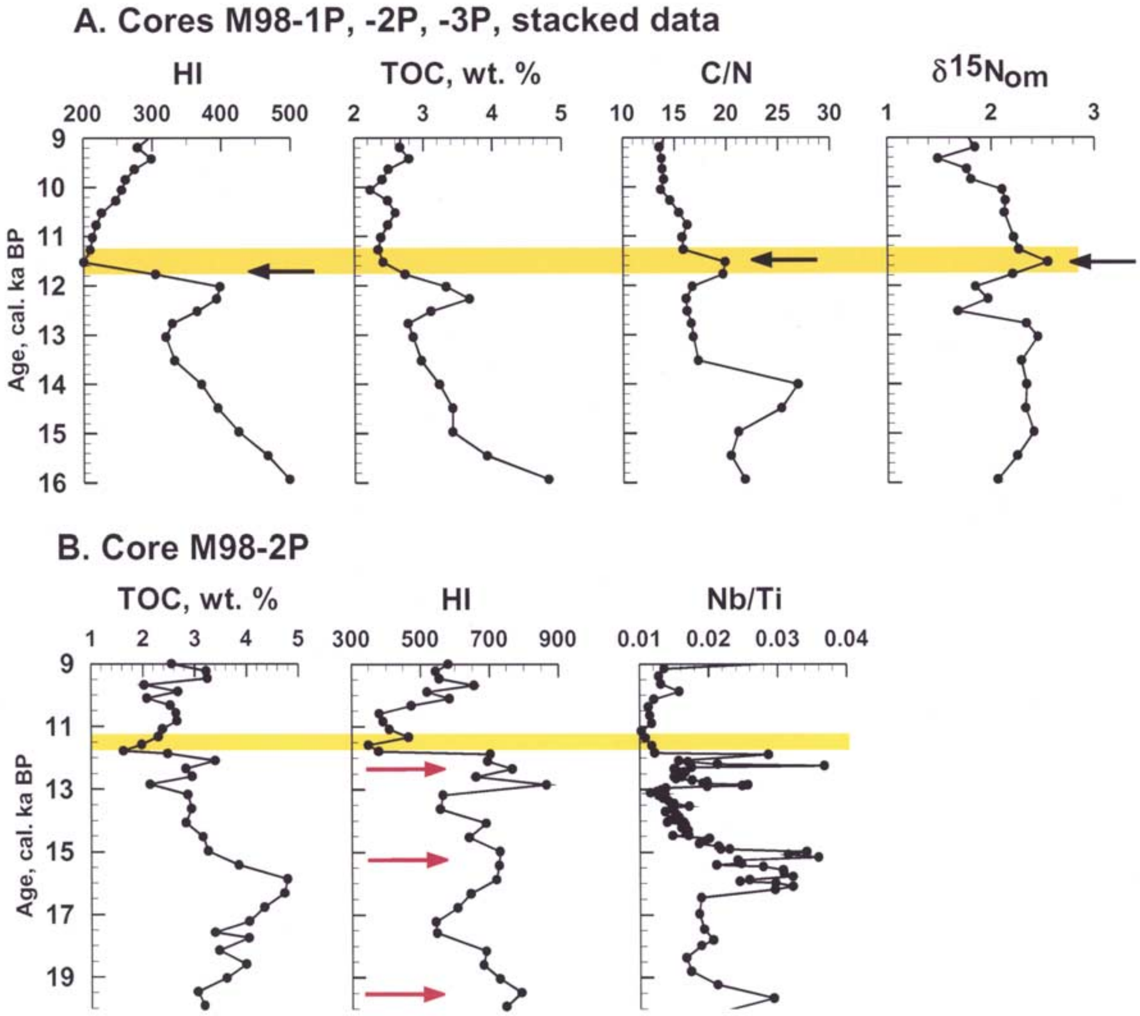

Figure 2. (a) Stacked data from organic matter in Lake Malawi cores M98-1P, -2P, and -3P [*Barry et al.*, 2002; *Filippi and Talbot*, 2005]. Arrow on Hydrogen Index (HI) plot indicates abrupt fall in HI to a minimum for the ~25 kyr covered by these cores. Arrows on C/N and δ^{15}N plots indicate peaks that were probably due to an influx of degraded OM during a lake-level lowstand. (b) Total organic carbon (TOC) and HI (corrected for dead carbon content [*Filippi and Talbot*, 2005]) compared with Nb/Ti values for core M98-2P. Nb/Ti provides a measure of the content of volcanic debris, presumed to be derived from the Rungwe volcanic center to the north of L. Malawi [*Johnson et al.*, 2002]. HI values have been corrected for the presence of dead carbon [*Filippi and Talbot*, 2005]. Arrows indicate correlation between peaks in HI and Nb/Ti, suggesting enhanced primary productivity due to upwelling during periods of strong northerly winds. Shaded zone: end of the Younger Dryas (circa 11.5 ka ± 0.25 ka).

northerly winds driving the upwelling of nutrient-rich deep water [*Filippi and Talbot*, 2005]. A similar conclusion has been reached by *Johnson et al.* [2002, 2004], based upon the accumulation rates of BSi, used as a proxy for diatom productivity, and volcanic debris presumably derived from the Rungwe volcanic center, north of Lake Malawi. The coincidence between peaks in HI (including the YD) and a high abundance of volcanic debris in the M98-2P core is especially striking (Figure 2b), confirming the coupling between OM production and the supply of volcanogenic debris. Before 12 ka the nutrient cycle in the north basin of Lake Malawi thus seems to have been dominated by upwelling driven by northerly winds. The change to generally lower HI, TOC, C/N and δ^{15}N after 11.5 ka, on the other hand, is compatible with lower rates of primary production at the northern end of the lake, suggesting that upwelling and the locus of maximum productivity had now shifted to the south basin. Our OM record indicates that the switch was abrupt and that it was accompanied by the onset of a period of aridity. The most likely reason for such a change was an overall northward displacement of the ITCZ from a position generally south of the north basin, to one significantly farther north. Indeed, for perhaps as much as 1500 years following the shift, the ITCZ may have been more or less permanently located north of the lake, leaving Malawi largely under the influence of the dry southeasterly trade wind system and resulting in low lake levels [*Finney and Johnson*, 1991; *Filippi and Talbot*, 2005].

4.1.2. Lake Tanganyika

[10] Any major displacement of the ITCZ north of Lake Malawi is also likely to have impacted at least the southern end of Lake Tanganyika. Figure 3 shows geochemical data from core MPU-10 collected from a depth of 422 m. from the southern basin of this lake as part of the Georift project [*Tiercelin and Mondeguer*, 1991]. Additional data come from cores MPU-3 [*Talbot et al.*, 2006], MPU-12 [*Gasse et al.*, 1989; *Hillaire-Marcel et al.*, 1989] and T-2 [*Livingstone*, 1965; *Haberyan and Hecky*, 1987]. Correlation of the terminal Pleistocene section in these cores is aided by a prominent tephra that occurs throughout the southern basin (Figure 3). The ash has a calibrated age of circa 12.8 ka B.P. Additional chronology is provided by 19 AMS ^{14}C age determinations from MPU-3, -10 and -12, and 5 conventional ^{14}C dates from T-2.

[11] In Lake Tanganyika, $\delta^{13}C_{OM}$ provides the most informative record of environmental change. After a gradual decline from 16 to 12 ka B.P., δ^{13}C fell abruptly to reach a minimum value at circa 11.25 ka. In the high-resolution record from the deep-water MPU-10 core the change exceeds 3‰. A similar δ^{13}C change is recorded in MPU-3 and -12 [*Hillaire-Marcel et al.*, 1989; *Talbot et al.*, 2006]. The excursion is also marked by a HI maximum, and in all the cores by a TOC maximum (Figure 3). Following these maxima (and minima), the geochemical trends all reverse direction, δ^{13}C rises, TOC and HI fall. Other changes accompany the geochemical shifts. In the diatom flora, the previously dominant *Cyclostephanos* was abruptly replaced by *Aulacoseira* (*Melosira*) [*Haberyan and Hecky*, 1987; *Gasse et al.*, 1989]. Smear slides show that in MPU-3 and -10 this change occurs over a sediment thickness of just a few cm, within the interval where δ^{13}C falls to a minimum.

[12] In large African lakes, *Aulacoseira* is commonly regarded as typical for turbulent conditions with relatively deep mixing and *Haberyan and Hecky* [1987] suggest its abrupt appearance here marks the onset of upwelling at the southern end of Lake Tanganyika. Hydrogen Index values in excess of 500 (Figure 3) and smear-slide analysis indicates that the OM is predominantly of phytoplankton origin, so the accompanying TOC, HI and δ^{13}C excursions probably reflect the sudden availability of abundant nutrients and isotopically light, dissolved inorganic carbon (DIC), both products of the remineralization of sinking OM, brought into the photic zone by the upwelling. The nutrients would have stimulated massive phytoplankton blooms (hence the high TOC and HI values), while photosynthetic fixation of the isotopically light DIC would have produced new cell material with low δ^{13}C [*Talbot et al.*, 2006]. As we have noted, this event also marked a turning point in the general compositional trends for OM at the southern end of Lake Tanganyika. The preceding falling trend in δ^{13}C was largely a consequence of flooding of the basin as the lake recovered from the LGM lowstand [*Talbot et al.*, 2006], but it had almost certainly reached overflow level well before the switch in compositional trend. The reversals in TOC, HI and δ^{13}C thus reflect readjustment of the lake and its nutrient cycle to the change from a southward- to northward-directed surface flow with upwelling at the southern end of the basin. Persistence in the dominance of *Aulacoseira* after the change indicates that this limnological regime continued into the early Holocene.

[13] The change in water column mixing in Lake Tanganyika was contemporaneous with the switch in circulation that affected the northern end of Lake Malawi. In common with the latter basin, the most likely explanation for the transition was a general northward migration in the position of the ITCZ, bringing the southern end of L. Tanganyika under the influence of the SE trade winds.

4.2. Lake Bosumtwi

[14] In West Africa, one of the best late Pleistocene-Holocene records of the monsoon system comes from Lake Bosumtwi, Ghana. The lake, currently closed, occupies a 1.06 Ma meteorite impact crater in the forest zone of southern Ghana where the climate is dominated by the seasonal N-S migration of the ITCZ (Figure 1). The basin experiences two rainy seasons as the ITCZ crosses the region on its way north and south, and is subject to notably dry conditions during the boreal winter when the ITCZ may be located well south of

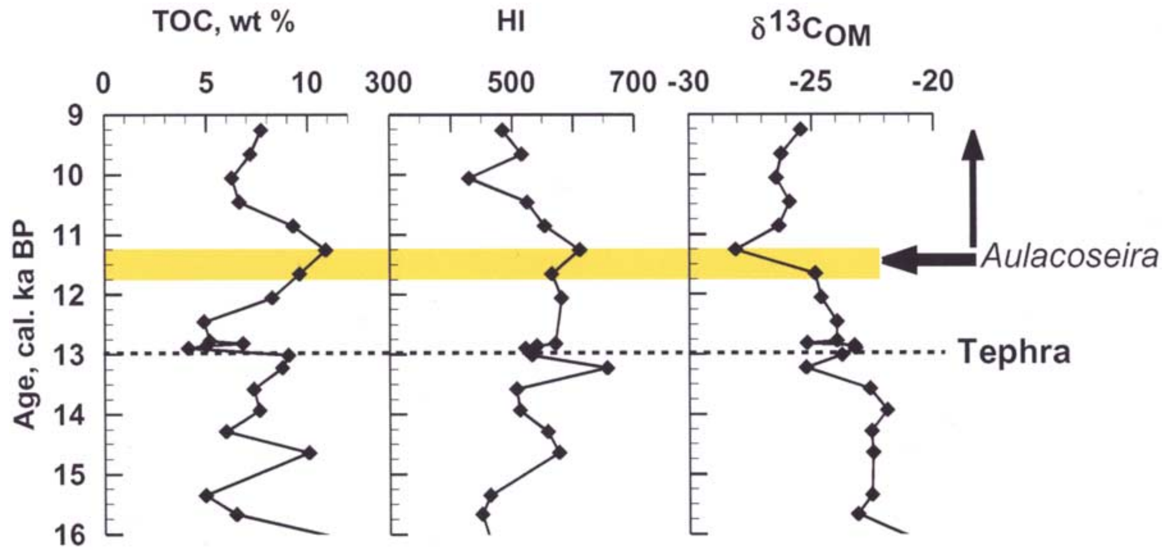

Figure 3. Data from organic matter in core MPU-10 (depth 422 m) from the south basin of Lake Tanganyika. The abrupt appearance of *Aulacoseira*, thought to indicate the onset of upwelling at the south end of the lake, is followed by peaks in TOC and HI, and then a reversal in the compositional trends for these and $\delta^{13}C_{OM}$ as the lake's nutrient cycle adjusted to a new mixing regime. Continuing dominance of *Aulacoseira* indicates the persistence of upwelling in the south basin after the end of the Younger Dryas (shaded zone: circa 11.5 ± 0.25 ka).

the lake [*Nicholson*, 1986]. Partial mixing of the water column typically occurs in August, during the cool, windy season; the hypolimnion is permanently anoxic. Cores from the lake have yielded detailed insights into the environmental history of the Bosumtwi region over the last ~30 kyr years [*Talbot et al.*, 1984; *Talbot and Johannessen*, 1992; *Maley*, 1997; *Russell et al.*, 2003; *Beuning et al.*, 2003; *Peck et al.*, 2004]. In particular, it has been demonstrated that variations in the stable carbon and nitrogen isotopic composition of the OM in these cores preserve a sensitive record of changes in basin hydrology that can be related to regional climate [*Talbot and Johannessen*, 1992; *Russell et al.*, 2003; *Beuning et al.*, 2003]. One of the most striking isotopic events is centered on circa 11.8 ka B.P. when both $\delta^{13}C_{OM}$ and $\delta^{15}N_{OM}$ fell by several per mil within the space of a few hundred years (Figure 4). This geochemical excursion accompanies a marked change in sediment lithology, from finely laminated deposits, some of which may be varved, to a massive or vaguely laminated sapropel rich in cyanophyte remains. The transition has been interpreted [*Talbot and Johannessen*, 1992] to reflect a rise in lake level and the rapid change from a seasonally influenced limnological regime (laminated deposits) to one with much less marked seasonality and little tendency to vertical mixing, which allowed N-fixing cyanophytes with characteristically low $\delta^{15}N$ values to dominate the lake's phytoplankton (sapropel accumulation). New field data supported by numerous AMS age determinations confirm that the change coincided with a major lacustrine transgression following a low-stand contemporaneous with the YD [*Shanahan et al.*, 2006]. High lake levels after the YD [*Talbot and Delibrias*, 1980; *Shanahan et al.*, 2006], and the development of a diverse forest vegetation [*Maley*, 1991] at the time of sapropel accumulation, indicate that the climate was humid, but the lack of well-developed lamination suggests little seasonality in the runoff that supplied clastic sediment to the lake. Furthermore, the apparent absence of significant vertical mixing indicates that surface winds were too slack and seasonal temperature contrasts too small to cause significant overturn of the water column at that time [*Talbot et al.*, 1984; *Talbot and Johannessen*, 1992]. The evidence of minimal seasonality and low wind intensity at Bosumtwi suggests that the ITCZ may have been permanently located north of southern Ghana during this period, putting the region under the influence of the equatorial doldrums (the equatorial zone of calm or light winds).

4.3. Other Wind-Related Records

4.3.1. South of the Equator

[15] On Mt. Kilimanjaro the most recent phase of ice accumulation seems to have commenced at circa 11.8 ka [*Thompson et al.*, 2002]. The precipitation required to construct these ice fields is

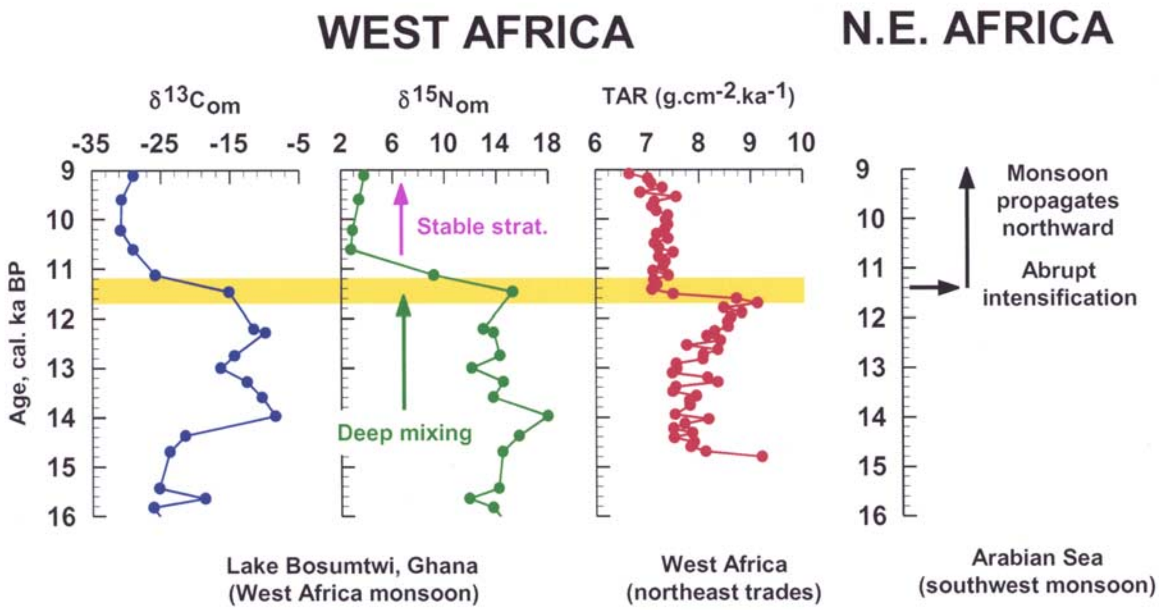

Figure 4. Organic-matter isotopic data from cores B-6/B-7, Lake Bosumtwi, Ghana [*Talbot and Johannessen*, 1992], compared with the terrigenous accumulation rate (TAR) from Site 658 off the west coast of Senegal [*deMenocal et al.*, 2000]. The abrupt decline in TAR coincides with very large excursions in $\delta^{13}C_{OM}$ and $\delta^{15}N_{OM}$ that reflect changes in the mixing regime of L. Bosumtwi. Events in West Africa seem to have been contemporaneous with an abrupt intensification of the southwest monsoon in the Arabian Sea, off the northeast coast of Africa [*Sirocko*, 1996]. Shaded zone: end of Younger Dryas (circa 11.5 ± 0.25 ka).

brought largely by the SE trades [*Hastenrath*, 1984; *Kaser and Osmaston*, 2002], suggesting that the change to a positive mass balance may have been the result of a shift in circulation over the mountain. If so, it coincided with onset of persistent SE winds over Lake Malawi.

[16] Cores from the Benguela upwelling region off the southwest coast of Africa record a change from relatively cool to warmer sea-surface temperatures at the end of the YD [*Kim et al.*, 2002; *Farmer et al.*, 2005] (Figure 5). The transition was particularly rapid close to the coast of Namibia where, on the basis of changes in Mg/Ca molar ratios, surface waters are inferred to have warmed by approximately 2°–3°C, a consequence, it is suggested, of shifts in both wind strength and direction [*Farmer et al.*, 2005]. Similarly, a sea-surface temperature (SST) record from the upwelling zone off Angola shows an abrupt fall in SST within the YD, thought to reflect intensification of the SE trade winds, followed by a gradual decline in wind intensity [*Kim and Schneider*, 2003].

4.3.2. North of the Equator

[17] Here, additional evidence for changes in wind regime at the end of the YD comes from both terrestrial and marine records. Offshore West Africa, *deMenocal et al.* [2000] document major variations in the supply of terrigenous, windborne dust to the eastern tropical Atlantic during the terminal Pleistocene. Of particular note is the very abrupt decline in terrigenous flux at 11.6 ka (Figure 4). Although some of this decrease must have been due to increasing vegetation cover on the adjacent continent [*deMenocal et al.*, 2000; *Hoelzmann et al.*, 2004; *Gasse and Roberts*, 2005], it may also reflect an abrupt decline in the transport capacity of the winds blowing off the continent. This change may in part have been due to generally lower wind velocities, but could also be a consequence of greater northward penetration of the ITCZ reducing the seasonal influence of the NE Trades. Such a scenario is compatible with the general decline in aeolian activity across western and northern Africa at the end of the YD [*Swezey*, 2001; *Lancaster et al.*, 2002].

[18] A similar pattern is seen in proxy records of monsoon wind direction and intensity from the Arabian Sea. Here, *Sirocko* [1996] (Figures 1 and 4) documents an abrupt intensification in the southwesterly monsoon between circa 11.6 and 11.45 ka B.P., following several millennia when this wind system was of limited importance. Change at this time is also coincident with one of the periods of rapid

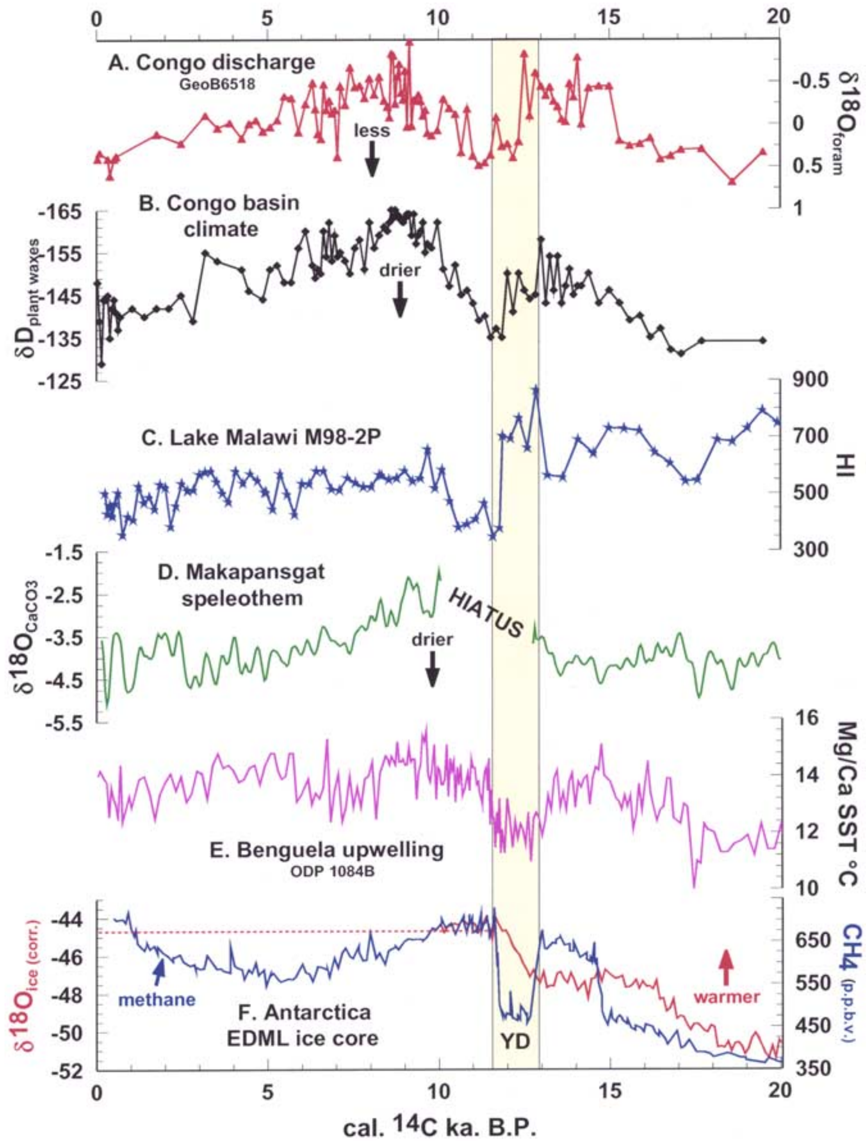

Figure 5. (a and b) Proxy records from core GeoB6518 from the mouth of the Congo (replotted from *Schefuss et al.* [2005]) and southern Africa ((c) *Filippi and Talbot* [2005]; (d) *Holmgren et al.* [2003]; (e) ODP core 1084B [*Farmer et al.*, 2005]). Note how in comparison with the Gulf of Guinea (Figure 6), the records shown in Figures 5a–5c and 5e suggest a more gradual transition from the cool, dry, windy conditions of the YD. All five records indicate that the fully humid conditions of the early to mid-Holocene were not achieved south of the equator until circa 10 ka B.P. Figure 5f is from the Dronning Maud Land (EDML) ice core [*EPICA Community Members*, 2006] and shows how the YD, here clearly defined by the ice core methane record, was coincident with a period of particularly rapid warming in the area of Antarctica adjacent to the southern Atlantic Ocean. Dashed line shows the mean $\delta^{18}O_{ice}$ value for the Holocene segment of the curve, which is linked to the LGM by a transitional period of rising $\delta^{18}O$. The end of the YD coincided with the end of this post-LGM warming in Antarctica.

intensification of the SW Indian monsoon identified by *Overpeck et al.* [1996].

4.4. Other Records

4.4.1. Southern Africa

[19] There appear to be no high-resolution records showing an abrupt change in atmospheric circulation at the end of the YD, but records of other climatic proxies show some striking changes over the same time interval. Of particular note is the speleothem $\delta^{18}O$ record from Makapansgat, South Africa (Figure 1), which is interrupted by a ~2.5 kyr hiatus spanning the period from circa 12.7 to 10.2 ka [*Holmgren et al.*, 2003] (Figure 5). The gap is thought to have been caused by drought

conditions and corresponds very closely in time to the inferred lowstand conditions in Lake Malawi (see above). Additional evidence for aridity at this time comes from the Tswaing (Pretoria) Saltpan [*Partridge et al.*, 1997] and Wonderkrater [*Holmgren et al.*, 2003], South Africa, and from Botswana and Namibia [*Shi et al.*, 1998; *Thomas and Shaw*, 2002].

4.4.2. Equatorial Regions

[20] A multiproxy study of a core from a high-accumulation-rate site off the mouth of the Congo River provides an outstanding record of changes in river discharge and basin-scale climatic conditions over a large area of equatorial Africa [*Schefuss et al.*, 2005] (Figure 5). The δD of terrestrial plant waxes, a proxy for water balance in the Congo basin, and surface water δ^{18}O, which at the core site is strongly influenced by the Congo River, both indicate the abrupt onset of drier conditions with reduced Congo River discharge during the YD, followed over the next ~1500 years by a gradual transition to wetter conditions (Figure 5). Pollen data from Lake Albert indicate an arid YD, and diatom evidence from Lake Victoria suggests lowered water levels over the same interval, but in both these cases there seems to have been a relatively rapid return to more humid conditions at the end of the YD [*Beuning et al.*, 1998; *Stager et al.*, 2002].

4.4.3. Gulf of Guinea

[21] Oxygen isotopic records of discharge from the Niger delta and the Sanaga and Nyong Rivers (Cameroon) indicate low runoff during the early to middle part of the YD, followed by an abrupt rise in freshwater discharge during the later YD [*Pastouret et al.*, 1978; *Lézine et al.*, 2005; *Weldeab et al.*, 2005] (Figure 6). High sediment accumulation rates and elemental analyses of sediment deposited over the same period show an abrupt rise in river-supplied sediment contemporaneous with the increased discharge [*Pastouret et al.*, 1978; *Adegbie et al.*, 2003].

4.4.4. Northern Tropical Africa

[22] Across the Sahel and Sahara, and in Ethiopia, a major phase of lake filling commenced between circa 11.5 and 11.2 ka [*Gasse*, 2000; *Chalié and Gasse*, 2002; *Gasse and Roberts*, 2005; *Hoelzmann et al.*, 2004]. Prior to this, many basins were desiccated or contained saline water bodies. The main phase of basin filling occurred from 11.5 to 10.8 ka and has been identified by *Hoelzmann et al.* [2004] as a major climatic event in the post-LGM evolution of northern Africa. The arrival of a moister climate in these areas north of the equator strongly suggests a northward shift in the influence of the ITCZ at the end of the YD.

5. Discussion

[23] The data we have assembled extend over about 40° of latitude and confirm that in Africa the Younger Dryas was a continent-wide event marked by generally dry and windy conditions. The inferred climatic conditions are in agreement with several other studies of the same period in tropical Africa (see above), and are consistent with a postulated southerly position for the ITCZ over Africa, the tropical Atlantic, the Caribbean and South America (Figure 7) (see *Peterson and Haug* [2006] and *Seager and Battisti* [2007] for recent reviews). Our interpretation of the geochemical records from Lakes Malawi, Tanganyika and Bosumtwi suggests that this situation came to an end in the later part of the YD, which was marked by an abrupt change in atmospheric circulation with a switch in dominant wind direction from NE to SE in southern Africa and to doldrum-like conditions in equatorial West Africa. This change coincided with the penetration of monsoonal conditions into arid northern Africa. Between circa 11.8 and 11.2 ka B.P. the African monsoon system apparently shifted abruptly northward by ~200–300 km (Figure 7).

[24] Although there are strong regional similarities in the continent's climatic evolution, there are also some clear differences, especially in the immediate post-YD period. In southern Africa, including at least parts of the Congo basin, dry conditions persisted beyond the end of the YD, the climate only achieving typical Holocene-like levels of humidity at circa 10 ka (Figure 5). North of the equator, on the other hand, the transition was much more rapid with notably humid conditions reaching arid northern Africa within a few hundred years of the end of the YD (Figure 6) [*Gasse and Roberts*, 2005]. Eventually, this change pushed the northern limit of the ITCZ several 100 km into the heart of northern Africa; *Gasse and Roberts* [2005] estimate that at its maximum early Holocene extent the ITCZ lay around 3°–5° north of its present mean position. At the southern margin of the monsoon belt, on the other hand, a corresponding northward translation of the southern limit of the ITCZ initially put Lake Malawi mainly beyond the limit of monsoon rains, leaving it under the influence of

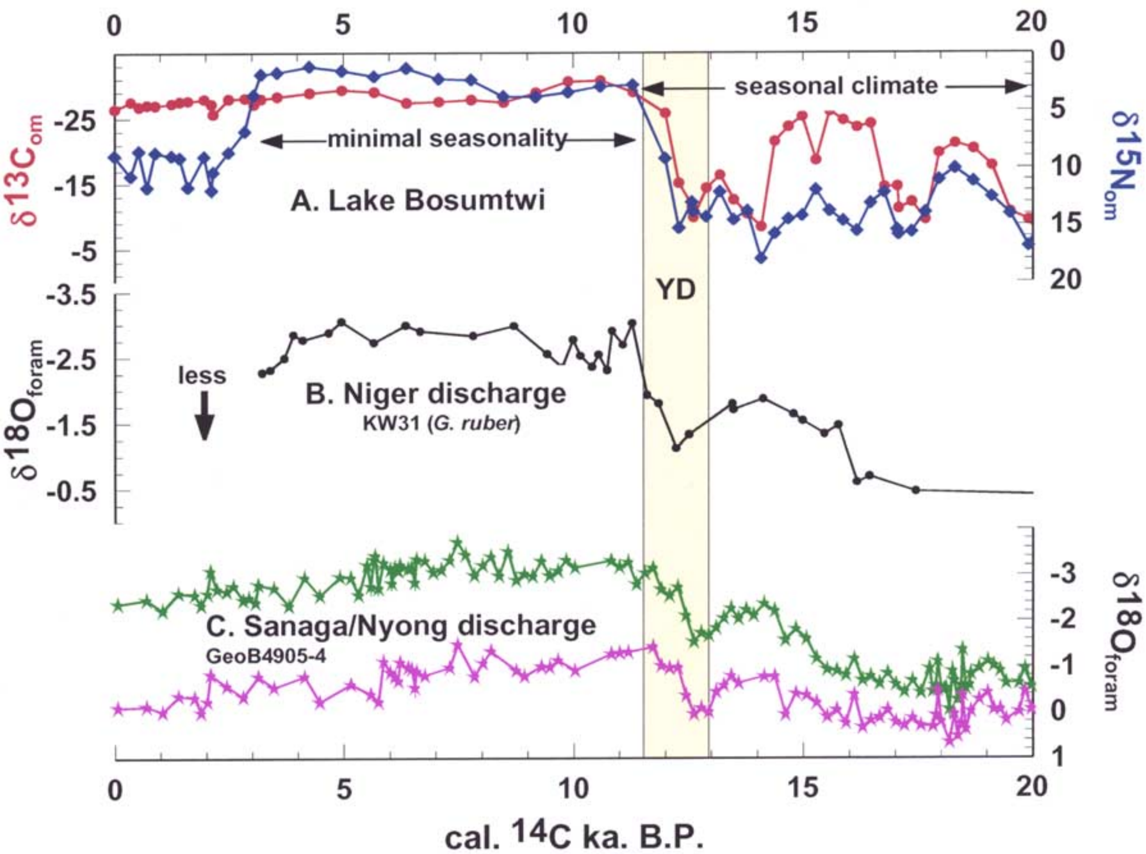

Figure 6. Proxy records from the Gulf of Guinea. (a) *Talbot and Johannessen* [1992]; updated timescale based in part on *Peck et al.* [2004]. (b) Isotopic data from core KW31 [*Pastouret et al.*, 1978]; calibrated timescale from *Lézine et al.* [2005]. (c) Core GeoB4905-B [*Weldeab et al.*, 2005]. Note the relatively abrupt onset of humid conditions following the arid, low-discharge conditions of the Younger Dryas (compare to Figure 4).

the dry SE Trades, possibly for much of the year. Notably dry conditions seem to have persisted over most of southern Africa at this time and by extending into at least the southern part of the Congo basin, were probably responsible for the significantly reduced discharge of the Congo River (Figure 5). Thus the immediate effect of the end-YD climatic event seems to have been a wholesale northward displacement of the African monsoon system. Subsequently, over a period of about 1500 years, the southern limit of monsoon influence gradually extended south, bringing humid conditions to the whole of the Congo basin, to Malawi, and eventually southern Africa. Isotopically enriched rainfall and a warm climate that favored the formation of aragonite speleothems with relatively high δ^{18}O values in the Makapansgat cave (Figure 5), indicate that after circa 10 ka B.P. monsoonal rainfall reached at least as far south as northern South Africa [*Holmgren et al.*, 2003]. With the ITCZ also extending far into the Sahara at this time [*Hoelzmann et al.*, 2004; *Gasse and Roberts*, 2005], it seems that the early Holocene saw a significant expansion in the monsoon belt to both the north and south.

[25] The cause of the dramatic rebound from the cooler, dry, windy conditions of the Younger Dryas remains an enigma, as, indeed, does the profound impact of the YD itself on African environments. The YD in its "type," circum-North Atlantic area is generally thought to have been caused by a major reduction in the thermohaline circulation (THC) due to freshwater influx [*Broecker et al.*, 1988; *Bond et al.*, 1997; *Broecker*, 2006]. Modeling simulations, while confirming that the North Atlantic thermohaline circulation does influence the climate of tropical Africa, suggest that the impact of a THC reduction should have been significantly less than most paleoenvironmental proxies indicate was in fact the case [*Dahl et al.*, 2005]. A dominant role for a THC shutdown in

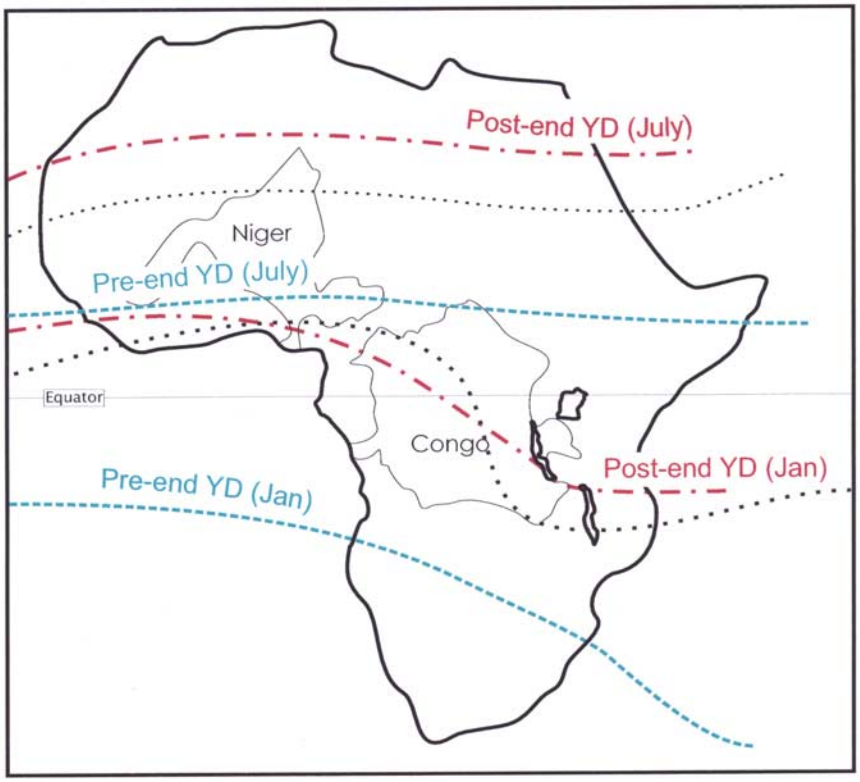

Figure 7. Positions (speculative!) of the ITCZ just before the YD and immediately following the end of the YD; the mean modern January and July positions (dotted lines) are shown for comparison. In the former, the ITCZ is displaced south of its modern position, in keeping with proxy evidence from tropical Africa, the eastern tropical Atlantic, Caribbean, and South America (see text). The end-YD changes in circulation saw the ITCZ shifted significantly farther north, bringing dry conditions to South Africa, Lake Malawi, and the southern part of the Congo basin (see also text and Figure 4), which were now situated south of ITCZ. On the other hand, northward translation of the monsoon system now brought moist conditions to the Sahel and Sahara, putting the whole of the Niger basin under the influence of the ITCZ (July position of the post-end YD ITCZ based in part on *Gasse and Roberts* [2005]).

forcing the YD climatic excursion in the tropics has also recently been questioned on theoretical grounds by *Seager and Battisti* [2007]. Weakened thermohaline circulation may thus not have been the only climatic forcing function in the region. In recent years there has been considerable focus on the possible role of SSTs, both in the adjacent oceans and farther afield in the South Atlantic, in forcing tropical African climate [e.g., *Camberlin et al.*, 2001; *Kim et al.*, 2002, 2003; *Barker and Gasse*, 2003; *Weldeab et al.*, 2005]. However, *Dupont et al.* [2004] have specifically discounted a relationship between Atlantic SST and the YD in southwest Africa. Attention has also been directed toward possible changes in Southern Hemisphere atmospheric circulation particularly with respect to wet and dry phases in southern Africa [e.g., *Tyson*, 1986; *Partridge*, 2002; *Holmgren et al.*, 2003; *Dupont et al.*, 2004; *Stuut et al.*, 2004]. Interest has, in particular, been focused upon the role of the southern hemisphere polar vortex in forcing rainfall and wind patterns over southern African today and in the past [*Tyson*, 1986; *Stuut et al.*, 2004; *Partridge et al.*, 2004]. During warmer climatic intervals in Antarctica and the southern oceans, when sea-ice cover is reduced, the polar vortex is in a southerly position and dry conditions prevail over southern Africa. The opposite situation occurs during colder periods when sea-ice cover increases. Under these conditions, the polar vortex expands northward, pushing moisture-bearing air masses into southern Africa. It is striking therefore that the end-YD shift in circulation over Africa coincided with the end of several millennia of warming in Antarctica, marking the transition [*Röthlisberger et al.*, 2002] from a glacial to interglacial climate mode (Figure 5). This warming is particularly clearly seen in the new EPICA Dronning Maud

Land (EDML) ice core which shows that temperature rise in the area of Antarctica adjacent to the South Atlantic was at its most rapid during the YD and ended, with no apparent transitional period, when the YD ended [*EPICA Community Members*, 2006] (Figure 5). A cold northern and warm southern hemisphere would have enhanced the interhemispheric thermal gradient and caused the ITCZ to be located at a southerly position [cf. *Camberlin et al.*, 2001], thus explaining the dominance of strong, northeasterly trade winds over southern Africa during the YD. Once Antarctic warming ceased and the northern hemisphere warmed, the polar vortex moved northward, eventually bringing moister conditions into Africa south of the equator. Our explanation for the abrupt end of the YD in tropical Africa clearly involves a dramatic reorganization of atmospheric circulation over the continent that may have been associated with major changes in Antarctica and its adjacent oceans. Reorganization of Southern Hemisphere atmospheric circulation at this time may thus have played a central role in the northward displacement of the African monsoon system, from where it was transmitted to the Northern Hemisphere, contributing to the end of the YD there. The enigma remains the apparent rapidity of these changes. They may have been related to the existence of the so-called "bipolar seesaw" [*Stocker and Johnsen*, 2003; *EPICA Community Members*, 2006], but are also consistent with the mode switch in tropical circulation recently postulated by *Seager and Battisti* [2007] as a potential explanation for the rapid transition from stadial- (e.g., YD) to interstadial-like (e.g., Holocene) conditions in the tropics.

6. Conclusions

[26] Across tropical Africa, from ~23°S to at least 20°N, the end of the Younger Dryas was marked by an abrupt change in climate which terminated the cool, dry windy conditions of the earlier part of the YD. In southern Africa, dry conditions seem to have persisted and the change was chiefly marked by a switch in dominant wind direction from NE to SE. In equatorial and northern tropical Africa, on the other hand, there appears to have been a sudden and significant increase in precipitation, raising lake levels and swelling river discharge. Wind speeds seem to have decreased and northern equatorial regions may have become subject to more or less permanent doldrum-like conditions with minimal seasonality. The changes are consistent with an abrupt, 200–300 km northward translation of the African monsoon system at the end of the YD. Although this dramatic reorganization in tropical atmospheric circulation initially caused drought in southern Africa south of 5°–10°S, wetter conditions subsequently spread gradually southward between 11.5 and 10 ka B.P., raising the level of Lake Malawi and bringing monsoonal rains to northern South Africa.

Acknowledgments

[27] The authors acknowledge support for their work on the African lakes from the Norwegian Research Council (NFR-MRT, M.L.F. and N.B.J.), the EU's Marie Curie Post-Doctoral Programme (M.L.F.), and CNRS (J.-J.T.). We have all benefited greatly from interaction with members of the IDEAL program, in particular Tom Johnson (Large Lakes Observatory, University of Minnesota, Duluth). Anne-Marie Lézine, Karin Holmgren, Christa Farmer, and Peter deMenocal are thanked for providing data files of their results, and Françoise Gasse is thanked for unpublished AMS age details from Lake Tanganyika. Two anonymous reviewers helped improve an earlier version of the paper. Jim Kennett (UC, Santa Barbara) is thanked for comments on an early draft of the paper.

References

Adegbie, A. T., R. R. Schneider, U. Röhl, and G. Wefer (2003), Glacial millennial-scale fluctuations in central African precipitation recorded in terrigenous sediment supply and freshwater signals offshore Cameroon, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *197*, 323–333.

Alley, R. B. (2000), The Younger Dryas cold interval as viewed from central Greenland, *Quat. Sci. Rev.*, *19*, 213–226.

Alley, R. B., et al. (1993), Abrupt increase in snow accumulation at the end of the Younger Dryas, *Nature*, *362*, 527–529.

Barker, P., and F. Gasse (2003), New evidence for a reduced water balance in East Africa during the Last Glacial Maximum: Implication for model-data comparison, *Quat. Sci. Rev.*, *22*, 823–837.

Barker, P. A., M. R. Talbot, F. A. Street-Perrott, F. Marret, J. Scourse, and E. O. Odada (2004), Late Quaternary climatic variability in intertropical Africa, in *Past Climate Variability Through Europe and Africa*, edited by R. W. Battarbee, F. Gasse, and C. E. Stickley, pp. 117–138, Springer, New York.

Barry, S. L., M. L. Filippi, M. R. Talbot, and T. C. Johnson (2002s), Sedimentology and geochronology of late Pleistocene and Holocene sediments from northern Lake Malawi, in *The East African Great Lakes: Limnology, Palaeolimnology and Biodiversity*, edited by E. O. Odada and D. O. Olago, pp. 369–391, Springer, New York.

Beadle, L. C. (1981), *The Inland Waters of Tropical Africa*, 433 pp., Longman, New York.

Beuning, K. R. M., K. Kelts, and J. C. Stager (1998), Abrupt climatic changes associated with the arid Younger Dryas interval in Africa, in *Environmental Change and Response in East African Lakes*, edited by J. T. Lehman, pp. 147–156, Elsevier, New York.

Beuning, K. R. M., M. R. Talbot, D. A. Livingstone, and G. Schmukler (2003), Sensitivity of carbon isotopic proxies to paleoclimatic forcing: A case study from Lake Bosumtwi, Ghana, over the last 32,000 years, *Global Biogeochem. Cycles*, *17*(4), 1121, doi:10.1029/2003GB002072.

Björck, S., et al. (1996), Synchronized terrestrial-atmospheric deglacial records around the North Atlantic, *Science*, *274*, 1155–1160.

Bond, G., W. Showers, M. Cheseby, R. Lotti, P. Almasi, P. DeMenocal, P. Priore, H. Cullen, I. Hajdas, and G. Bonani (1997), A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates, *Science*, *278*, 1257–1266.

Broecker, W. S. (2006), Was the Younger Dryas triggered by a flood?, *Science*, *312*, 1146–1148.

Broecker, W. S., M. Andree, W. Wolli, W. Oeschger, G. Bonani, J. Kennett, and D. A. Peteet (1988), A case in support of a meltwater diversion as the trigger for the onset of the Younger Dryas, *Paleoceanography*, *3*, 1–9.

Brown, E. T., L. Le Callonnec, and C. R. German (2000), Geochemical cycling of redox-sensitive metals in sediments from Lake Malawi: A diagnostic paleotracer for episodic changes in mixing depth, *Geochim. Cosmochim. Acta*, *64*, 3515–3523.

Camberlin, P., S. Janicot, and I. Poccard (2001), Seasonality and atmospheric dynamics of the teleconnection between African rainfall and tropical sea-surface temperature: Atlantic vs. ENSO, *Int. J. Climatol.*, *21*, 973–1005.

Chalié, F., and F. Gasse (2002), Late Glacial–Holocene diatom record of water chemistry and lake level change from the tropical East African rift Lake Abiyata (Ethiopia), *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *187*, 259–283.

Claussen, M., C. Kubatzki, V. Brovkin, A. Ganopolski, P. Hoelzmann, and H.-J. Pachur (1999), Simulation of an abrupt change in Saharan vegetation in the mid-Holocene, *Geophys. Res. Lett.*, *26*, 2037–2040.

Cohen, A. S., M. R. Talbot, S. M. Awramik, D. L. Dettman, and P. Abell (1997), Lake level and paleoenvironmental history of Lake Tanganyika, Africa, as inferred from late Holocene and modern stromatolites, *Geol. Soc. Am. Bull.*, *109*, 444–460.

Coulter, G. W., and R. H. Spigel (1991), Hydrodynamics, in *Lake Tanganyika and Its Life*, edited by G. W. Coulter, pp. 49–75, Oxford Univ. Press, New York.

Dahl, K. A., A. J. Broccoli, and R. J. Stouffer (2005), Assessing the role of North Atlantic freshwater forcing in millennial scale climate variability: A tropical Atlantic perspective, *Clim. Dyn.*, *24*, 325–346.

deMenocal, P. B., and D. Rind (1993), Sensitivity of Asian and African climate to variations in seasonal insolation, glacial ice cover, sea surface temperature, and Asian orography, *J. Geophys. Res.*, *98*, 7265–7287.

deMenocal, P., J. Ortiz, T. Guilderson, J. Adkins, M. Sarnthein, L. Baker, and M. Yarusinsky (2000), Abrupt onset and termination of the African Humid Period: Rapid climate responses to gradual insolation forcing, *Quat. Sci. Rev.*, *19*, 347–361.

Dupont, L. M., J.-H. Kim, R. R. Schneider, and N. Shi (2004), Southwest African climate independent of Atlantic sea surface temperatures during the Younger Dryas, *Quat. Res.*, *61*, 318–324.

EPICA Community Members (2006), One-to-one coupling of glacial climate variability in Greenland and Antarctica, *Nature*, *444*, 195–198.

Farmer, E. C., P. B. deMenocal, and T. M. Marchitto (2005), Holocene and deglacial ocean temperature variability in the Benguela upwelling region: Implications for low-latitude atmospheric circulation, *Paleoceanography*, *20*, PA2018, doi:10.1029/2004PA001049.

Filippi, M. L., and M. R. Talbot (2005), The palaeolimnology of northern Lake Malawi over the last 25 ka based upon the elemental and stable isotopic composition of sedimentary organic matter, *Quat. Sci. Rev.*, *24*, 1303–1328.

Finney, B. P., and T. C. Johnson (1991), Sedimentation in Lake Malawi (East Africa) during the past 10,000 years: A continuous paleoclimatic record from the southern tropics, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *85*, 351–366.

Garcin, Y., D. Williamson, M. Taieb, A. Vincens, P.-E. Mathé, and A. Majule (2006), Centennial to millennial changes in maar-lake deposition during the last 45,000 years in tropical Southern Africa (Lake Masoko, Tanzania), *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *239*, 334–354.

Gasse, F. (2000), Hydrological changes in the African tropics since the Last Glacial Maximum, *Quat. Sci. Rev.*, *19*, 189–211.

Gasse, F., and C. N. Roberts (2005), Late Quaternary hydrologic changes in the arid and semiarid belt of northern Africa, in *The Hadley Circulation: Present, Past and Future*, edited by H. F. Diaz and R. S. Bradley, pp. 313–345, Elsevier, New York.

Gasse, F., and E. Van Campo (1994), Abrupt post-glacial climate events in West Asia and North Africa monsoon domains, *Earth Planet. Sci. Lett.*, *126*, 435–456.

Gasse, F., V. Lédée, M. Massault, and J. C. Fontes (1989), Water-level fluctuations of Lake Tanganyika in phase with oceanic changes during the last glaciation and deglaciation, *Nature*, *342*, 57–59.

Gasse, F., P. Barker, and T. C. Johnson (2002), A 24,000 yr diatom record from the northern basin of Lake Malawi, in *The East African Great Lakes: Limnology, Palaeolimnology and Biodiversity*, edited by E. O. Odada and D. O. Olago, pp. 393–414, Springer, New York.

Haberyan, K. A., and R. E. Hecky (1987), The Late Pleistocene and Holocene stratigraphy and paleolimnology of Lakes Kivu and Tanganyika, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *61*, 169–197.

Hastenrath, S. (1984), *The Glaciers of Equatorial Africa*, 353 pp., Springer, New York.

Hecky, R. E., and H. J. Kling (1987), Phytoplankton ecology of the great lakes in the rift valleys of central Africa, *Ergeb. Limnol.*, *25*, 197–228.

Hecky, R. E., R. H. Spigel, and G. W. Coulter (1991), The nutrient regime, in *Lake Tanganyika and Its Life*, edited by G. W. Coulter, pp. 76–89, Oxford Univ. Press, New York.

Hillaire-Marcel, C., A. M. Aucour, R. Bonnefille, G. Riollet, A. Vincens, and D. Williamson (1989), ^{13}C/palynological evidence of differential residence times of organic carbon prior to its sedimentation in East African rift lakes and peat bogs, *Quat. Sci. Rev.*, *8*, 207–212.

Hoelzmann, P., F. Gasse, L. Dupont, U. Salzmann, M. Staubwasser, D. C. Leuscher, and F. Sirocko (2004), Palaeoenvironmental changes in the arid and sub-arid belt (Sahara-Sahel_Arabian Peninsula) from 150 kyr to present, in *Past Climatic Variability Through Europe and Africa*, edited by R. W. Battarbee, F. Gasse, and C. E. Stickley, pp. 219–256, Springer, New York.

Holmgren, K., J. A. Lee-Thorp, G. R. J. Cooper, K. Lundblad, T. C. Partridge, L. Scott, R. Sithaldeen, A. S. Talma, and P. D. Tyson (2003), Persistent millennial-scale climatic variability over the past 25,000 years in southern Africa, *Quat. Sci. Rev.*, *22*, 2311–2326.

Johnson, T. C., E. T. Brown, J. McManus, S. Barry, P. Barker, and F. Gasse (2002), A high-resolution paleoclimate record spanning the past 25,000 years in southern East Africa, *Science*, *296*, 113–132.

Johnson, T. C., E. T. Brown, and J. McManus (2004), Diatom productivity in northern Lake Malawi during the past 25,000 years: Implications for the position of the intertropical convergence zone at millennial and shorter time scales, in *Past Climatic Variability Through Europe and Africa*, edited by R. W. Battarbee, F. Gasse, and C. E. Stickley, pp. 93–116, Springer, New York.

Kaser, G., and H. Osmaston (2002), *Tropical Glaciers*, 207 pp., Cambridge Univ. Press, New York.

Kim, J.-H., and R. R. Schneider (2003), Low-latitude control of interhemispheric sea-surface temperature contrast in the tropical Atlantic over the past 21 kyears: The possible role of SE trade winds, *Clim. Dyn.*, *21*, 337–347.

Kim, J.-H., R. R. Schneider, P. J. Müller, and G. Wefer (2002), Interhemispheric comparison of deglacial sea-surface temperature patterns in the Atlantic eastern boundary currents, *Earth Planet. Sci. Lett.*, *194*, 383–393.

Kim, J.-H., R. R. Schneider, S. Mulitza, and P. J. Müller (2003), Reconstruction of SE trade-wind intensity based on sea-surface temperature gradients in the Southeast Atlantic over the last 25 kyr, *Geophys. Res. Lett.*, *30*(22), 2144, doi:10.1029/2003GL017557.

Kutzbach, J. E., and Z. Liu (1997), Response of the African monsoon to orbital forcing and ocean feedbacks in the middle Holocene, *Science*, *278*, 440–444.

Kutzbach, J. E., and F. A. Street-Perrott (1985), Milankovitch forcing of fluctuations in the level of tropical lakes from 18 to 0 kyr BP, *Nature*, *317*, 130–134.

Kutzbach, J., G. Bonan, J. Foley, and S. P. Harrison (1996), Vegetation and soil feedbacks on the response of the African monsoon to orbital forcing in the early to middle Holocene, *Nature*, *384*, 623–626.

Lancaster, N., G. Kocurek, A. Singhvi, V. Pandey, M. Deynoux, J.-F. Ghienne, and K. Lô (2002), Late Pleistocene and Holocene dune activity and wind regimes in the western Sahara desert of Mauritania, *Geology*, *30*, 991–994.

Lézine, A.-M., J.-C. Duplessy, and J.-P. Cazet (2005), West African monsoon variability during the last deglaciation and the Holocene: Evidence from fresh water algae, pollen and isotope data from core KW31, Gulf of Guinea, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *219*, 225–237.

Livingstone, D. A. (1965), Sedimentation and history of water level changes in Lake Tanganyika, *Limnol. Oceanogr.*, *10*, 607–609.

Lotter, A. F., H. J. B. Birks, U. Eicher, W. Hofmann, J. Schwander, and L. Wick (2000), Younger Dryas and Allerød summer temperatures at Gerzensee (Switzerland) inferred from fossil pollen and cladoceran assemblages, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *159*, 349–361.

Maley, J. (1991), The African rain forest vegetation and palaeoenvironments during Late Quaternary, *Clim. Change*, *19*, 79–98.

Maley, J. (1997), Middle to late Holocene changes in tropical Africa and other continents: Paleomonsoon and sea surface temperature variations, in *Third Millennium BC Climate Change and Old World Collapse*, *NATO ASI Ser., Ser. I*, vol. 149, edited by H. N. Dalfes and G. Kukla, pp. 611–640, Springer, New York.

Mollenhauer, G., T. I. Eglington, N. Ohkuchi, R. R. Schneider, P. J. Muller, P. M. Grootes, and J. Rullkotter (2003), Asynchronous alkenone and foraminifera records from the Benguela upwelling system, *Geochim. Cosmochim. Acta*, *67*, 2157–2171.

Nicholson, S. E. (1986), The spatial coherence of African rainfall anomalies: Interhemispheric teleconnections, *J. Clim. Appl. Meteorol.*, *25*, 1365–1381.

Nicholson, S. E. (1996), A review of climate dynamics and climate variability in eastern Africa, in *The Limnology, Climatology and Paleoclimatology of the East African Lakes*, edited by T. C. Johnson and E. O. Odada, pp. 25–56, Gordon and Breach, New York.

Overpeck, J., D. Anderson, S. Trumbore, and W. Prell (1996), The southwest Indian Monsoon over the last 18000 years, *Clim. Dyn.*, *12*, 213–225.

Paillard, D., L. Labeyrie, and P. Yiou (1996), Macintosh program performs time-series analysis, *Eos Trans. AGU*, *77*, 379.

Partridge, T. C. (2002), Were Heinrich events forced from the southern hemisphere?, *S. Afr. J. Sci.*, *98*, 43–46.

Partridge, T. C., P. B. deMenocal, S. A. Lorentz, M. J. Paiker, and J. C. Vogel (1997), Orbital forcing of climate over South Africa: A 200,000-year rainfall record from the Pretoria saltpan, *Quat. Sci. Rev.*, *16*, 1125–1133.

Partridge, T. C., L. Scott, and R. R. Schneider (2004), Between Agulhas and Benguela: Responses of Southern African climates of the Late Pleistocene to current fluxes, orbital precession and the extent of the Circum-Antarctic vortex, in *Past Climatic Variability Through Europe and Africa*, edited by R. W. Battarbee, F. Gasse, and C. E. Stickley, pp. 45–68, Springer, New York.

Pastouret, L., H. Chamley, G. Delibrias, J. C. Duplessy, and J. Thiede (1978), Late Quaternary climatic changes in western tropical Africa deduced from deep-sea sedimentation off the Niger delta, *Oceanol. Acta*, *1*, 217–232.

Patterson, G., and O. Kachinjika (1995), Limnology and phytoplankton ecology, in *The Fishery Potential and Productivity of the Pelagic Zone of Lake Malawi/Niassa. Scientific Report of the UK/SADC Pelagic Fish Resource Assessment Project*, edited by A. Menz, 67 pp., Nat. Resour. Inst. (ODA), Chatham, UK.

Peck, J. A., R. R. Green, T. Shanahan, J. W. King, J. T. Overpeck, and C. A. Scholz (2004), A magnetic mineral record of Late Quaternary tropical climate variability from Lake Bosumwi, Ghana, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *215*, 37–57.

Peterson, L. C., and G. H. Haug (2006), Variability in the mean latitude of the Atlantic Intertropical Convergence Zone as recorded in the riverine inputs of sediments to the Cariaco Basin (Venezuela), *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *234*, 97–113.

Plisnier, P.-D., D. Chitamwebwa, L. Mwape, K. Tshibangu, V. Langenberg, and E. Coenen (1999), Limnological annual cycle inferred from physical-chemical fluctuations at three stations of Lake Tanganyika, *Hydrobiologia*, *407*, 45–58.

Powers, L. A., T. C. Johnson, J. P. Werne, I. S. Castañeda, E. C. Hopmans, J. S. Sinninghe Damsté, and S. Schouten (2005), Large temperature variability in the southern African tropics since the Last Glacial Maximum, *Geophys. Res. Lett.*, *32*, L08706, doi:10.1029/2004GL022014.

Roberts, N., M. Taieb, P. Barker, B. Damnati, M. Icole, and D. Williamson (1993), Timing of the Younger Dryas event in East Africa from lake-level changes, *Nature*, *366*, 146–148.

Rohling, E. J., P. A. Mayewski, and P. Challenor (2003), On the timing and mechanism of millennial-scale climate variability during the last glacial cycle, *Clim. Dyn.*, *20*, 257–267.

Röthlisberger, R., R. Mulvaney, E. W. Wolff, M. A. Hutterli, M. Bigler, S. Sommer, and J. Jouzel (2002), Dust and sea salt variability in central East Antarctica (Dome C) over the last

45 kyrs and its implications for southern high-latitude climate, *Geophys. Res. Lett.*, *29*(20), 1963, doi:10.1029/2002GL015186.

Russell, J., M. R. Talbot, and B. J. Haskell (2003), Mid-Holocene climate change in Lake Bosumtwi, Ghana, *Quat. Res.*, *60*, 133–141.

Schefuss, E., S. Schouten, and R. R. Schneider (2005), Climatic controls on central African hydrology during the past 20,000 years, *Nature*, *437*, 1003–1006.

Seager, R., and D. S. Battisti (2007), Challenges to our understanding of the general circulation: Abrupt climate change, in *The General Circulation of the Atmosphere*, edited by T. Schneider and A. S. Sobel, Princeton Univ. Press, Princeton, in press.

Severinghaus, J. P., T. Sowers, E. J. Brook, R. B. Alley, and M. L. Bender (1998), Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice, *Nature*, *391*, 141–146.

Shanahan, T., J. T. Overpeck, C. W. Wheeler, J. W. Beck, J. S. Pigati, M. R. Talbot, C. A. Scholz, J. Peck, and J. W. King (2006), Paleoclimatic variations in West Africa from a record of late Pleistocene and Holocene lake level stands of Lake Bosumtwi, Ghana, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *242*, 287–302.

Shi, N., L. M. Dupont, H.-J. Beug, and R. Schneider (1998), Vegetation and climate changes during the last 21000 years in S.W. Africa based on a marine pollen record, *Vegetation Hist. Archaeobot.*, *7*, 127–140.

Sirocko, F. (1996), The evolution of the monsoon climate over the Arabian Sea during the last 24,000 years, *Palaeoecol. Afr.*, *24*, 53–70.

Stager, J. C., P. A. Mayewski, and L. D. Meeker (2002), Cooling cycles, Heinrich event 1, and the desiccation of Lake Victoria, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *183*, 169–178.

Stocker, T. F., and S. J. Johnsen (2003), A minimum thermodynamic model for the bipolar seesaw, *Paleoceanography*, *18*(4), 1087, doi:10.1029/2003PA000920.

Street-Perrott, F., and R. A. Perrott (1993), Holocene vegetation, lake levels, and climate of Africa, in *Global Climates Since the Last Glacial Maximum*, edited by H. E. Wright et al., pp. 318–356, Univ. of Minn. Press, Minneapolis.

Stuiver, M., P. J. Reimer, E. Bard, J. W. Beck, G. S. Burr, K. A. Hughen, B. Kromer, G. McCormac, J. VanderPlicht, and M. Spurk (1998), INTCAL98 radiocarbon age calibration, 24,000-0 cal BP, *Radiocarbon*, *40*, 1041–1083.

Stuut, J.-B. W., X. Crosta, K. Van der Borg, and R. Schneider (2004), Relationship between Antarctic sea ice and southwest African climate during the late Quaternary, *Geology*, *32*, 909–912.

Swezey, C. (2001), Eolian sediment responses to late Quaternary climate changes: Temporal and spatial patterns in the Sahara, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *167*, 119–155.

Talbot, M. R. (2001), Nitrogen isotopes in palaeolimnology, in *Tracking Environmental Change Using Lake Sediments*, vol. 2, *Physical and Geochemical Methods*, edited by W. M. Last and J. P. Smol, pp. 401–439, Springer, New York.

Talbot, M. R., and G. Delibrias (1980), A new late Pleistocene-Holocene water-level curve for Lake Bosumtwi, Ghana, *Earth Planet. Sci. Lett.*, *47*, 336–344.

Talbot, M. R., and T. Johannessen (1992), A high-resolution palaeoclimatic record for the last 27,500 years in tropical West Africa from the carbon and nitrogen isotopic composition of lacustrine organic matter, *Earth Planet. Sci. Lett.*, *110*, 23–37.

Talbot, M. R., and T. Lærdal (2000), The Late Pleistocene–Holocene palaeolimnology of Lake Victoria, East Africa, based upon elemental and isotopic analyses of sedimentary organic matter, *J. Paleolimnol.*, *23*, 141–164.

Talbot, M. R., and D. A. Livingstone (1989), Hydrogen index and carbon isotopes of lacustrine organic matter as lake level indicators, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *70*, 121–137.

Talbot, M. R., D. A. Livingstone, P. G. Palmer, J. Maley, J. M. Melack, G. Delibrias, and S. Gulliksen (1984), Preliminary results from sediment cores from Lake Bosumtwi, Ghana, *Palaeoecol. Afr.*, *16*, 173–192.

Talbot, M. R., N. B. Jensen, T. Lærdal, and M. L. Filippi (2006), Geochemical responses to a major transgression in giant African lakes, *J. Paleolimnol.*, *35*, 467–489.

Thomas, D. S. G., and P. A. Shaw (2002), Late Quaternary environmental change in central southern Africa: New data, synthesis, issues and prospects, *Quat. Sci. Rev.*, *21*, 783–797.

Thompson, L. G., et al. (2002), Kilimanjaro ice core records: Evidence of Holocene climate change in tropical Africa, *Science*, *298*, 589–593.

Tiercelin, J.-J., and A. Mondeguer (1991), The geology of the Tanganyika Trough, in *Lake Tanganyika and Its Life*, edited by G. W. Coulter, pp. 7–48, Oxford Univ. Press, New York.

Tyson, P. D. (1986), *Climatic Change and Variability in Southern Africa*, 220 pp., Oxford Univ. Press, New York.

Weldeab, S., R. R. Schneider, M. Kölling, and G. Wefer (2005), Holocene African droughts relate to eastern equatorial Atlantic cooling, *Geology*, *33*, 981–984.