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Hydrological changes in the African tropics since the Last Glacial Maximum

Françoise Gasse

CEREGE, B.P. 80, 13545, Aix-en-Provence Cedex 04, France

Abstract

Paleohydrological data from the African tropics and subtropics, including lake, groundwater and speleothem records, are reviewed to show how environments and climates from both hemispheres are inter-related. Although orbitally induced changes in the monsoon strength account for a large part of long-term climatic changes in tropical Africa, the Late Pleistocene–Holocene hydrological fluctuations rather appear to have been a series of abrupt events that reflect complex interactions between orbital forcing, atmosphere, ocean and land surface conditions. During the Last Glacial Maximum (23–18 ka BP), most records indicate that generally dry conditions have prevailed in both hemispheres, associated with lower tropical land- and sea-surface temperatures. This agrees with simulations using coupled ocean–atmosphere models, which predict cooling and reduced summer precipitation in tropical Africa; the global hydrological cycle was weaker than today when the extent of large polar ice-sheets and sea-ice was a prominent forcing factor of the Earth's climate. Glacial-interglacial climatic changes started early: a first wetting/warming phase at ca. 17–16 ka BP took place during a period of rapid temperature increase in Antarctica. Next, two drastic arid-humid transitions in equatorial and northern Africa occurred around 15–14.5 ka BP and 11.5–11 ka BP. Both are thought to match the major Greenland warming events, in concert with the switching of the oceanic thermohaline circulation to modern mode. However, part of the climatic signal after 15 ka BP also seems related to the Antarctica climate. During the Holocene, Africa has also experienced rapid hydrological fluctuations of dramatic magnitude compared to the climatic changes at high latitudes. In particular, major dry spells occurred around 8.4–8 ka and 4.2–4 ka BP in the northern monsoon domain. Comparison with other parts of the world indicates that these events have a worldwide distribution but different regional expressions. In the absence of large polar ice sheets, changes in the continental hydrological cycles in the tropics may have a significant impact on the global climate system. Climate information gathered here allows to identify geographical and methodological gaps, and raise some scientific questions that remain to be solved to better understand how the tropics respond to changes in major climate-forcing factors, and how they influence climate globally.

1. Introduction

The hydrological fluctuations that took place in Africa during the Late Quaternary represent a impressive manifestation of continental climate change. In recent times, exceptional rainfall events over East Africa, associated with El Niño-Southern Oscillation (ENSO) years (Nicholson, 1996), have generated catastrophic floods. During the Sahel drought in the 1970s and 1980s, mean annual rainfall declined by 30% (Hulme, 1992); cattle were decimated and people migrated southward. However, these recent events do not rival those of past millenia. From about 10,000 to 4000 years ago, Neolithic civilizations flourished in a wet and green Sahara, while in northern East Africa closed lakes extended tens or hundreds of metres above their present level. Conversely, from about 20,000 to 15,000 years ago, the largest lake in Africa, Lake Victoria, was desiccated (Talbot and Livingstone, 1989; Johnson et al., 1996), as were many others. The distribution and density of the human population in Africa is regulated primarily by the availability of water to drive biological and human activities. It is thus of relevance to study past hydrological records to understand how regional environmental changes are related to changes in the Earth's climate system.

This paper focuses on the African tropics and subtropics within PAGES-PEP III Time Stream II (the last 250,000 yr at time resolution of 10^2–10^3 yr; PAGES, 1997). It aims to use paleohydrological proxy data to document how environments and climates from both hemispheres are inter-related. Regional paleohydrological records, derived from lake, groundwater, and speleothem archives, are compared. Some pollen data

E-mail address: gasse@cerege.fr (F. Gasse)

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associated with lake studies are also considered. Most records do not provide the resolution and continuity that would be required for detailed correlations with high latitude time-series, especially with the polar ice core records. Nevertheless, the paleohydrological data assembled here can be useful for model-data comparisons, yielding hypotheses on the linkages between high and low-latitude climates. The paper is structured around the following questions:

- How did African climates in both hemispheres respond to orbital forcing versus other glacial boundary conditions during the last glacial maximum?
- How were the successive steps of the last glacial–interglacial transition in Africa linked with deglacial events at high latitudes?
- How did abrupt Holocene changes correlate with climate changes elsewhere in the world, and how did they affect the availability of water resources?

2. Major climatic and hydrological features, and paleohydrological archives

African climates depend on low altitude pressure and winds over the continent, which are the surface expression of the upper air circulation. Climates exhibit a broadly zonal pattern with varying seasonal distribution of precipitation (Figs. 1 and 2). The northern and southern ends of the continent are affected by the equatorward displacement of the mid-latitude westerlies during winter. These temperate regions experience Mediterranean summer-dry climates, receiving most precipitation during winter from westerly cyclonic disturbances. They are both flanked by subtropical deserts, the Sahara north of the Equator, and the Namib coastal desert in southwestern Africa, which are dominated by subtropical anticyclones throughout the year. These arid zones are separated by a wide belt of tropical climates. The tropical climate is governed by the seasonal migration of the Intertropical Convergence Zone (ITCZ; the meteorological equator) in response to changes in the location of maximum solar heating. This results in northern and southern belts of monsoonal climates with summer rains and winter drought, bracketing a humid equatorial zone characterized by a double rainfall maximum. The zonal temperature and moisture patterns are altered by highlands which act as water towers for the surrounding lowlands. The zonal ("Walker") circulation along the equator, caused by East–West differences in surface and tropospheric temperatures, also greatly influences tropical climates. Rainfall fluctuations in many areas of the continent are statistically linked to the ENSO, but these may be more directly a response to sea-surface temperature (SST) fluctuations in the Indian and Atlantic Oceans which occur in the context of ENSO (Nicholson, 1996). The East–West climate asymmetry is linked to topographical features (e.g., the eastern escarpment of southern Africa), and to sea surface conditions (e.g., the cold oceanic Benguela, and the warm Algulhas currents flowing along the western and eastern coasts of southern Africa, respectively).

The major hydrological features of Africa result from the general climate, topography and geological patterns. They are (Fig. 3): (i) large internal drainage basins, e.g., the Lake Chad basin, and enormous groundwater reservoirs, e.g., the geological formations of the "Continental Intercalaire" in the Sahara; (ii) large exoreic rivers, the Congo, the Nile and the Niger Rivers and associated wetlands, which have their headwaters in the humid equatorial belt; (iii) great East African Rift lakes which play a significant role in regional climate through water recycling; (iv) other permanent waterbodies concentrated in, or depending on the equatorial zone (e.g., Lake Chad through the Chari River), or on highlands (e.g., the Ethiopian lakes Ziway-Shala and Abhé).

African lake records are mainly based on geomorphology, sedimentology, stable isotope contents of primary lacustrine carbonates and organic matter, and biological remains (e.g., diatoms, ostracods). Several authors give site locations and syntheses at a continental scale (e.g., Street-Perrott and Perrott, 1993; Jolly et al., 1998). Readers may refer to the compilation on "lake status" (high, intermediate or low water level) by Street-Perrott et al. (1989), and the Global-Lake Level Database available at the World Climate Data Center-A (WDC-A, Boulder, CO), but the original papers used for their construction should also be consulted, since the "lake status" is an interpretation which sometimes deviates from that in the original works.

Palaeoclimatic interpretation of African lake records often remains limited for several reasons. First, few works discuss thoroughly the reliability of their radiocarbon chronologies, although ^{14}C age distortions are numerous in African lakes (Gasse and Fontes, 1992). Second, high time-resolution lake records are rare. Finally, a full interpretation of the lake water, salt or isotope budgets at any given time should be based on the analysis of the response timing and processes of the hydrological system (lake and catchment area) to change in the precipitation minus evaporation balance (P–E). Estimates of past P and E from lake records, based on biological or sedimentological calibration functions and hydrological modelling, is possible as exemplified below. As yet, few approaches of this type have been conducted in Africa. Reliable palaeoclimatic interpretation is particularly difficult for groundwater-fed lakes such as those which prevailed in the Sahara and the Sahel. Groundwater-fed lakes can be sensitive indicators, provided that the recharge area is restricted, the flow patterns are known, and the response time to precipitation change is short (Gasse et al., 1990).

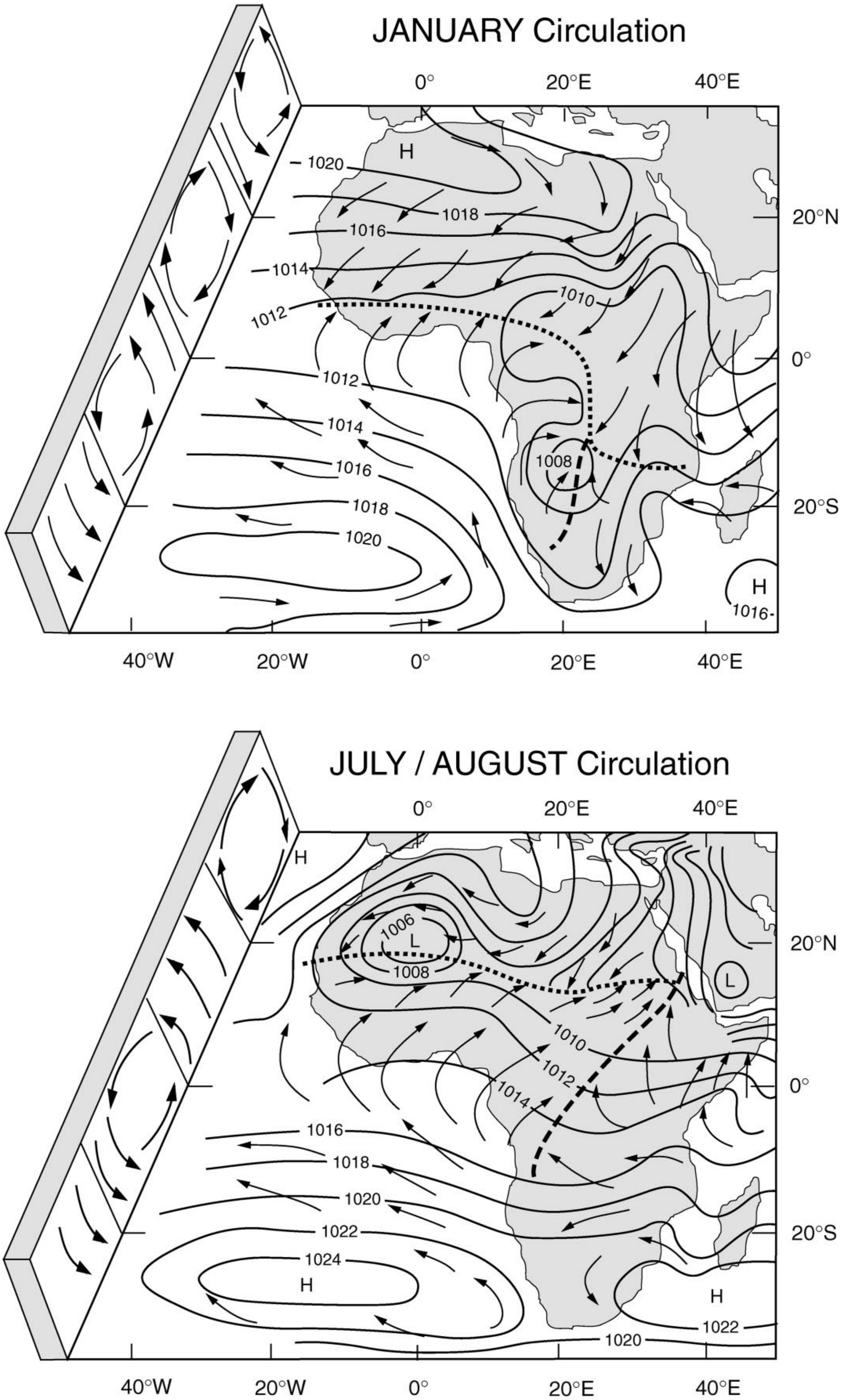

Fig. 1. Schematic of the general patterns of winds and pressure over Africa. Dotted lines indicate the Intertropical Convergence Zone (ITCZ), dashed lines indicate the Congo Air Boundary. From Nicholson (1996).

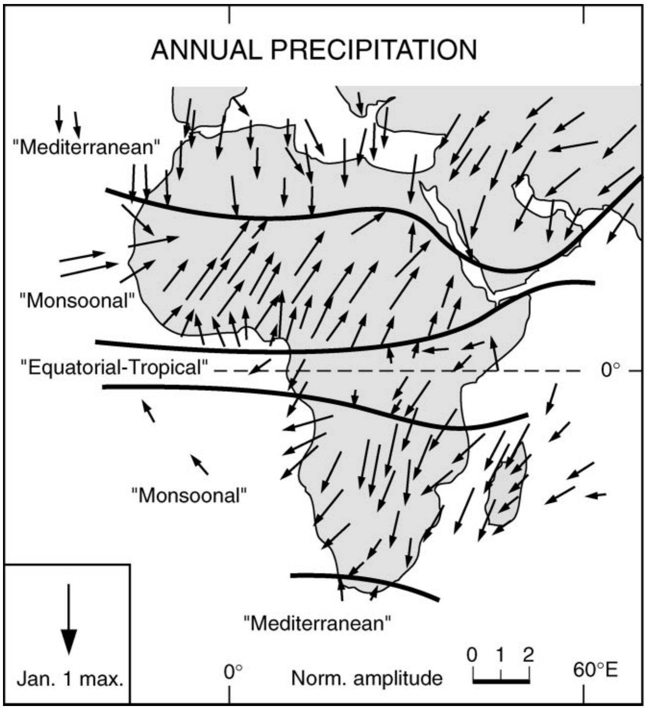

Fig. 2. African precipitation regimes. Annual harmonics of precipitation, from Hsü and Wallace (1976), adapted by deMenocal and Rind (1996). Rainfall time series from individual climatological stations were averaged into monthly means and then subjected to harmonic analysis which defined the phase and amplitude of the annual precipitation cycle. The vector length indicates normalized amplitude. Vector direction indicates month of maximum precipitation. Southward-pointing vectors indicate January 1 rainfall maximum; westward vectors indicate April 1 maxima (Hsü and Wallace, 1976).

Groundwater records may help interpret groundwater-fed palaeolakes. In addition, dated groundwater may provide information about palaeoclimatic conditions when and where the aquifers recharged, through their major and trace element content, stable isotope composition and noble gas content (Stute et al., in PAGES, 1997). Indeed, in confined (closed) aquifer systems, much of the water is fossil water infiltrated during past humid periods. Groundwater archives act as a low-pass filter and thus only provide low-resolution time-series; nevertheless, they lend themselves to quantitative reconstructions of temperature, and yield information on moisture transport patterns (Stute and Talma, 1998). Speleothems provide information about past $P-E$ balance, vegetation cover in the catchment, palaeotemperatures and isotopic composition of precipitation, through their texture, stable isotope content of calcite crystals, and their fluid inclusions (Lauritzen, in PAGES, 1997).

The chronological framework is mainly based on radiocarbon ages (^{14}C yr BP), and some ^{230}Th/^{234}U ages. Radiocarbon ages were converted into calendar estimates (cal. yr BP) using the CALIB 3.0 program (Stuiver and Reimer, 1993) back to 18,000 ^{14}C yr BP, and the equation proposed by Bard et al. (1997) for older ages. In the text below, ages are expressed in ka, which means 10^3 cal. yr BP. Sites have been numbered, S1, S2, … , and are shown in Fig. 3.

3. The last glacial period

The palaeohydrological data in this section were gathered to identify factors, other than orbital forcing, that have controlled the flux of energy and water vapour to the African continent during the last glacial maximum (LGM; ca. 23–18 ka). These data complement pollen and lake organic matter studies which have shown that decreased CO_2 atmospheric concentration induced a lowering of mountain forests and influenced the carbon cycle in mountain lakes in the tropics during glacial times (e.g., Street-Perrott et al., 1997; Jolly and Haxeltine, 1997).

3.1. Evidence for cooler and drier conditions during the LGM

A long (90 m) and continuous (200 kyrs) evaporite-rich depositional sequence was obtained from site S1, a meteoric crater lake, the Pretoria Saltpan (Partridge et al., 1997). Lake sediment texture and chemical composition were quantitatively related to precipitation in the catchment area. The resulting south monsoon precipitation time series is characterized by periodic (23,000 yr) variations, attributed to the orbital precession cycle (Fig. 4). Nevertheless, the 23,000-yr cyclicity is blurred after ca. 50 ka. The precipitation time series shows a negative shift starting at about 30 ka, and ending at the LGM when estimated precipitation values are about 15–20% lower than today.

The stable isotope record of a stalagmite from Botswana (S2) covers the interval 51–21 ka (Holmgren et al., 1995). The interval 51–43 ka was warm and wet. Dry periods at 46, 43, 26, 24 and 22 ka have been associated (Holmgren et al., 1995) with the Heinrich cooling events H5 to H2 (Bond et al., 1992). After 27 ka, gradually drier conditions finally halted speleothem growth at ca. 21 ka. Glacial cooling is estimated at about 2–3°C.

A noble gas temperature and isotope (δ^{18}O) record from palaeogroundwaters at Stampriet (S3, Fig. 5) covers the past 40 ka (Stute and Talma, 1998). At the LGM, it shows a glacial cooling as much as 5.3°C lower than today, similar to that observed at the Uitenhage aquifer (Fig. 3) in the southern Mediterranean climatic zone. At Stampriet, the present-day rainfall originating from the Indian Ocean exhibits a marked isotopic depletion due to the continental effect. The δ^{18}O values higher than today during glacial times, associated with low temperatures, suggest a shift in moisture transport from the

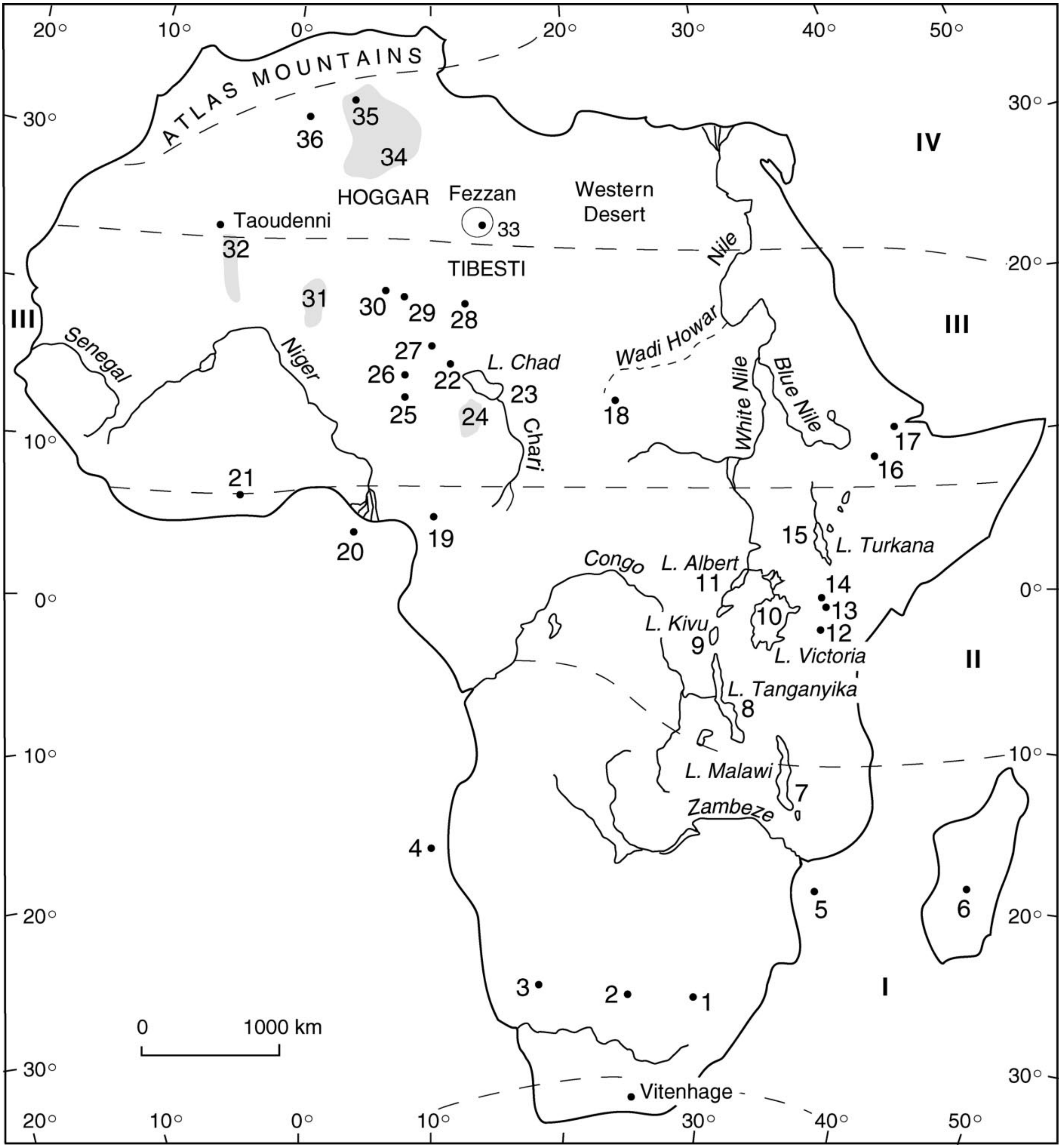

Fig. 3. Major lake and river systems in Africa, and location map of sites (S) cited in text. Sites I are from the southern monsoonal domain, II from equatorial-subequatorial Africa, III from the Sahelian belt and the southern Sahara, and IV from the winter rain domain of the northern Sahara (Fig. 2). 1: Pretoria Saltpan. 2: Lobatse. 3: Stampriet. 4: Marine core GeoBio1023. 5: Marine core MD79257. 6: L. Tritrivakely. 7: L. Malawi. 8: L. Tanganyika. 9: L. Kivu. 10: L. Victoria. 11: L. Albert. 12: L. Magadi. 13: L. Naivasha. 14: L. Nakuru-Elmenteita. 15: L. Turkana. 16: Ziway-Shala lake system. 17: L. Abhé. 18: Jebel Marra. 19. L. Barombi Mbo. 20: Marine core CH22KW31. 21: Lake Bosumtwi. 22: L. Chad. 23: Bahr-el-Ghazal.. 24: Nigeria, recharge zone of the Middle aquifer of the Chad Formation.. 25: L. Bal, Kajemarum Oasis. 26: Bougdouma. 27: Termit. 28: Bilma. 29: Tin Ouaffadene. 30: Adrar Bous. 31: Recharge zone of Northern Niger, and 32: Northern Mali. 33: Western Fezzan. 34: Recharge zone of the "Continental Intercalaire" aquifer. 35: Sebkha Mellala. 36: Hassi el Mejnah.

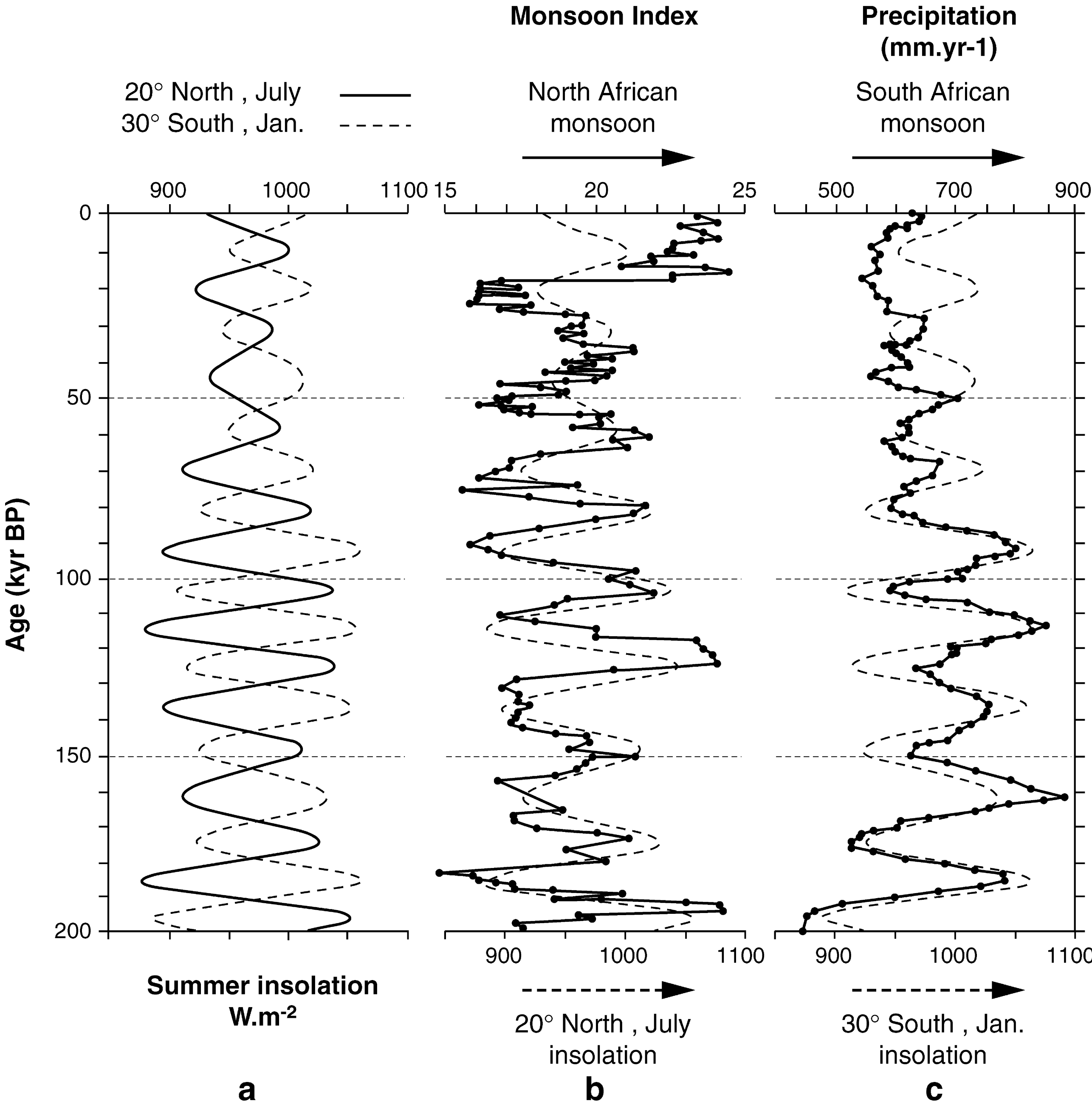

Fig. 4. Orbital precession as a dominant factor regulating late Pleistocene monsoonal precipitation in long-term records. Redrawn from Partridge et al. (1997). Comparison of: (c) variations in South monsoon precipitation reconstructed from a 200 kyr sedimentary record from the Pretoria Saltpan, South Africa (Partridge et al., 1997), with: (b) a monsoonal precipitation index at 20°N based on fossil faunal assemblage variations in deep-sea sediment core RC24-07 (20°N; McIntyre et al., 1989); and (a) changes in summer solar radiation in the northern and southern subtropics (after Berger, 1978).

Indian Ocean (as it is today), to the Atlantic Ocean: the shorter western path for rainwater from the Atlantic Ocean reduced the continental effect (Stute and Talma, 1998). This scenario implies that the westerly winds on the western seaboard extended northward, at least to 25°S during the Glacial period, in agreement with the climatic model of Cockcroft et al. (1987).

No lake record is available for southwest Africa, but a marine pollen record from core GeoBio1023 (site S4) registers climate changes in the continent between 21° and 13°S, west of 24°W (northern Namib desert, Angola-northern Namibian highlands, northwestern Kalahari), since 21 cal. ka BP (Ning Shi et al., 1998). The record shows cold and dry LGM conditions. Alkenone

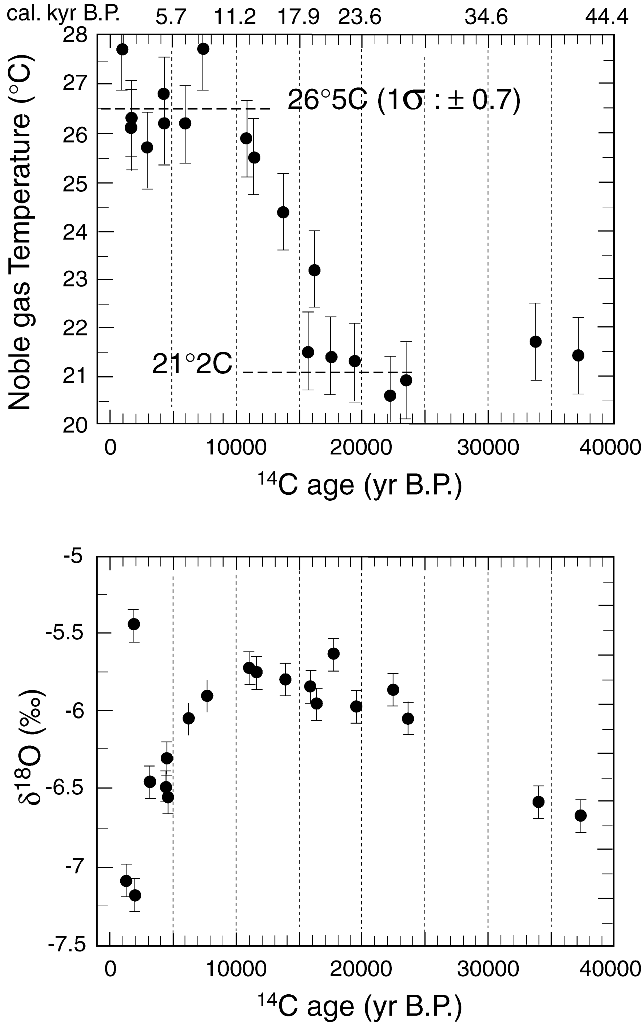

Fig. 5. Groundwater archives as paleoclimatic indicators in Africa. Noble gas temperature and $\delta^{18}O$ versus radiocarbon age for the Stampriet aquifer (Namibia). After Stute and Talma (1998).

paleothermometry also indicates sea-surface temperature (SST) 3–4°C lower than today at neighbouring sites (Schneider et al., 1995). At a similar latitude on the Indian Ocean side, SST alkenone-temperature from core MD79257 (S5) shows that the glacial cooling (about -2.5°C) culminated around 20–19.5 ka (Bard et al., 1997; Fig. 6). In the Madagascar highlands, a small crater lake, Tritrivakely (S6), has provided a multi-proxy record spanning the last 40 ka (Sifeddine et al., 1995; Williamson et al., 1998; Gasse and Van Campo, 1998; Fig. 6). Although the glacial period has been punctuated by several short-term warm-dry intervals, the terrestrial pollen record shows generally low temperature conditions until ca. 17 ka. In the lake, diatom and sedimentary records show a positive P–E balance from about 38 to 32 ka, followed by a step-wise desiccation trend. The LGM was a period of mean annual water deficit, although occasional (seasonal?) rains may have occurred. The lake was at least seasonnally dry from about 22 to 17.5 ka.

A highstand at Lake Malawi (S7) from about 32.5 to 11.5 ka suggested by sedimentological studies (Finney et al., 1996; Fig. 7) appears as an exception within the southern tropics.

In equatorial East Africa, Lake Tanganyika (S8) was at least 300 m lower than today at ca. 21 ka (Fig. 7; Gasse et al., 1989). A water and energy balance model, established for the modern Lake Tanganyika system, was used to estimate past mean annual precipitation and evaporation at the LGM (Bergonzini et al., 1997). Input values used to simulate the LGM situation are insolation, past lake and catchment areas, pollen-inferred temperature (-4.2 ± 2°C; Chalié, 1995), and albedo as estimated from vegetation changes in the region (Vincens, 1991; Bonnefille et al., 1992). In percent of modern means, the simulation yields decreased evaporation from the lake [-5% (-13% to $+3$%)] and land [-8% ($-19/+5$)] bodies, as well as precipitation [-14% ($-24/-2$)]. Decreases in P, E and P–E are substantially amplified when including empirical changes in atmospheric transmission coefficient (Tyson et al., 1997). Sensitivity runs suggest that even large changes in cloud cover and air humidity should not modify these trends. Lake Victoria (S10) was desiccated or nearly so at 20.5 ka (Talbot and Livingstone, 1989). The level of Lake Albert (S11) declined markedly after ca. 22 ka, and was at least 46 m lower than today between 20.2 and 18 ka (Beuning et al., 1997; Fig. 7). The seasonal discharge of the Blue Nile River was considerably reduced (Said, 1993).

In subequatorial West Africa, generally dry LGM conditions are indicated by pollen records from Lake Barombi Mbo (S19; Maley and Brenac, 1998) and Lake Bosumtwi (S21; Maley, 1991), but the interval 23–18 ka is complex. At Lake Bosumtwi, a $\delta^{18}O$ and $\delta^{15}N$ record in lacustrine organic matter spanning the last 32,000 ka shows that the setting of a drier climate, starting at ca. 29 ka, was not linear: dry excursions centred at 25.9, 21.9, and 17.5 ka were followed by relatively humid phases, notably that from 21 to 18.5 ka (Talbot and Johannessen, 1992; Fig. 8).

In the Sahel, the recharge of the confined "Middle aquifer of the Chad Formation" in northern Nigeria (S24), occurred between ca. 28 and 23.2 ka, and then stopped (Edmunds et al., 1999). Noble gas studies suggest that the recharge temperature was at least 6°C lower than at present. The end of this cool and wet recharge period was attributed to the onset of LGM arid conditions, as documented by lake records in the region (e.g., Servant and Servant-Vildary, 1980). The extension of Lake Chad (S22) at the LGM was estimated at 7% of its modern area (Adams and Tetzlaff, 1985). Eastward, lakes depending on the Ethiopian highlands (S16–S17) were low during the LGM (Gasse, 1977; Gasse and Street, 1978; Gillespie et al., 1983). Lake Abhé (S17; Fig. 7) experienced

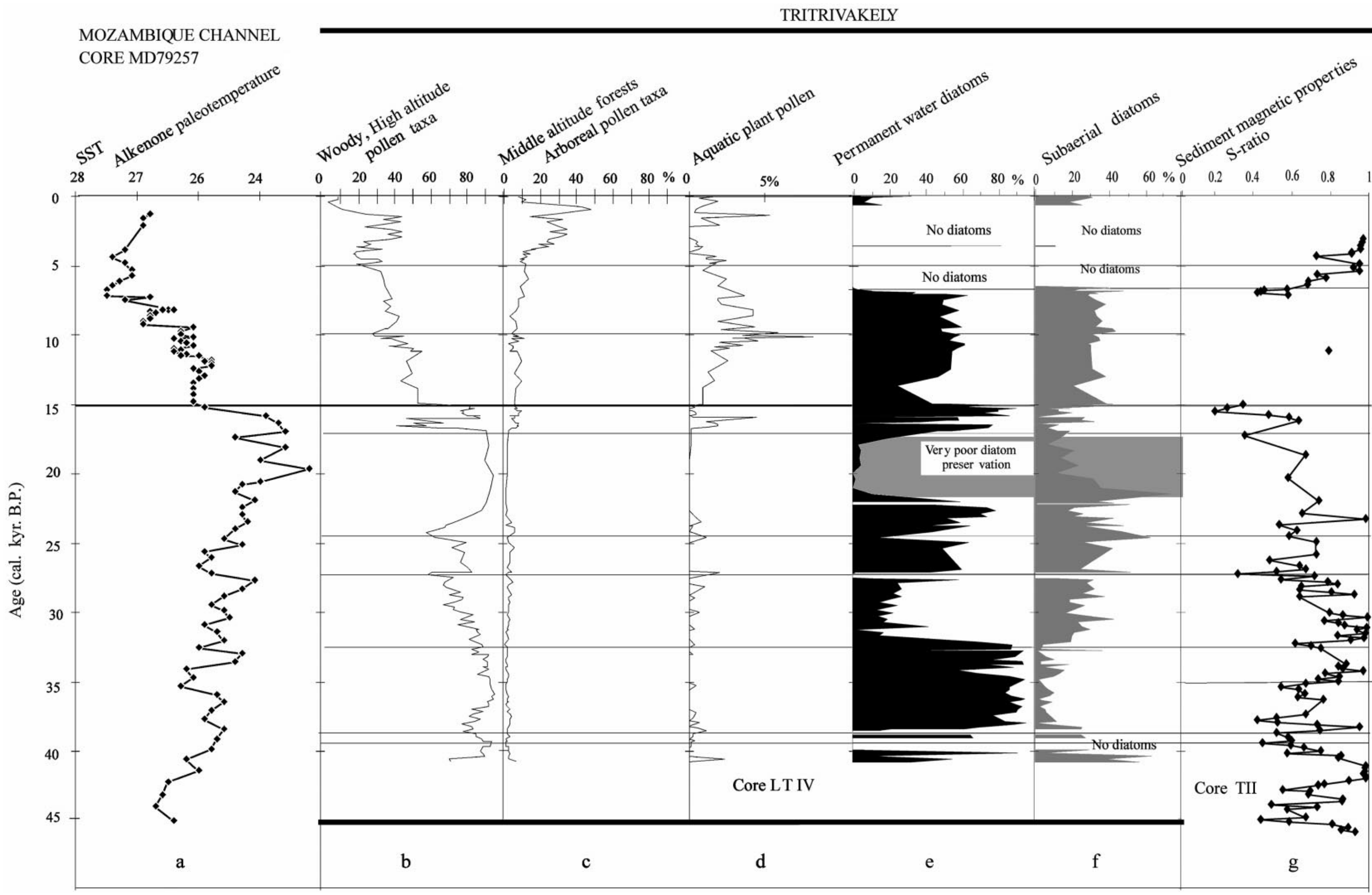

Fig. 6. Summary results from the Lake Tritrivakely core record (Madagascar highlands). Comparison with reconstructed SST in core MD79257 in the Mozambique Channel. (a) Alkenone paleotemperature in core MD79257 (after Bard et al., 1997). (b) Mountain forest taxa in the terrestrial pollen spectrum, reflecting changes in both atmospheric CO_2 concentration and temperature. Abrupt fluctuations are attributed to temperature changes. (c): arboreal pollen taxa from middle-altitude forests. (d) Aquatic plant pollen. Percentage is calculated on the terrestrial pollen sum. (e): Diatom diagram of taxa from permanent, slightly alkaline, mesotrophic water. (f) Diatom diagram of aerophilous, acidobiontic taxa reflecting subaerial, ombrotrophic conditions. Poor preservation and the absence of diatom valves indicate ephemeral conditions or seasonal/pluriannual desiccation phases. (g) *S*-ratio, a sediment magnetic parameter reflecting changes from high run-off and reducing bottom conditions (high *S*-ratio values) to low run-off and oxic environments (low *S*-ratio values). (b) to (f): after Gasse and Van Campo (1998); (g) after Williamson et al. (1998).

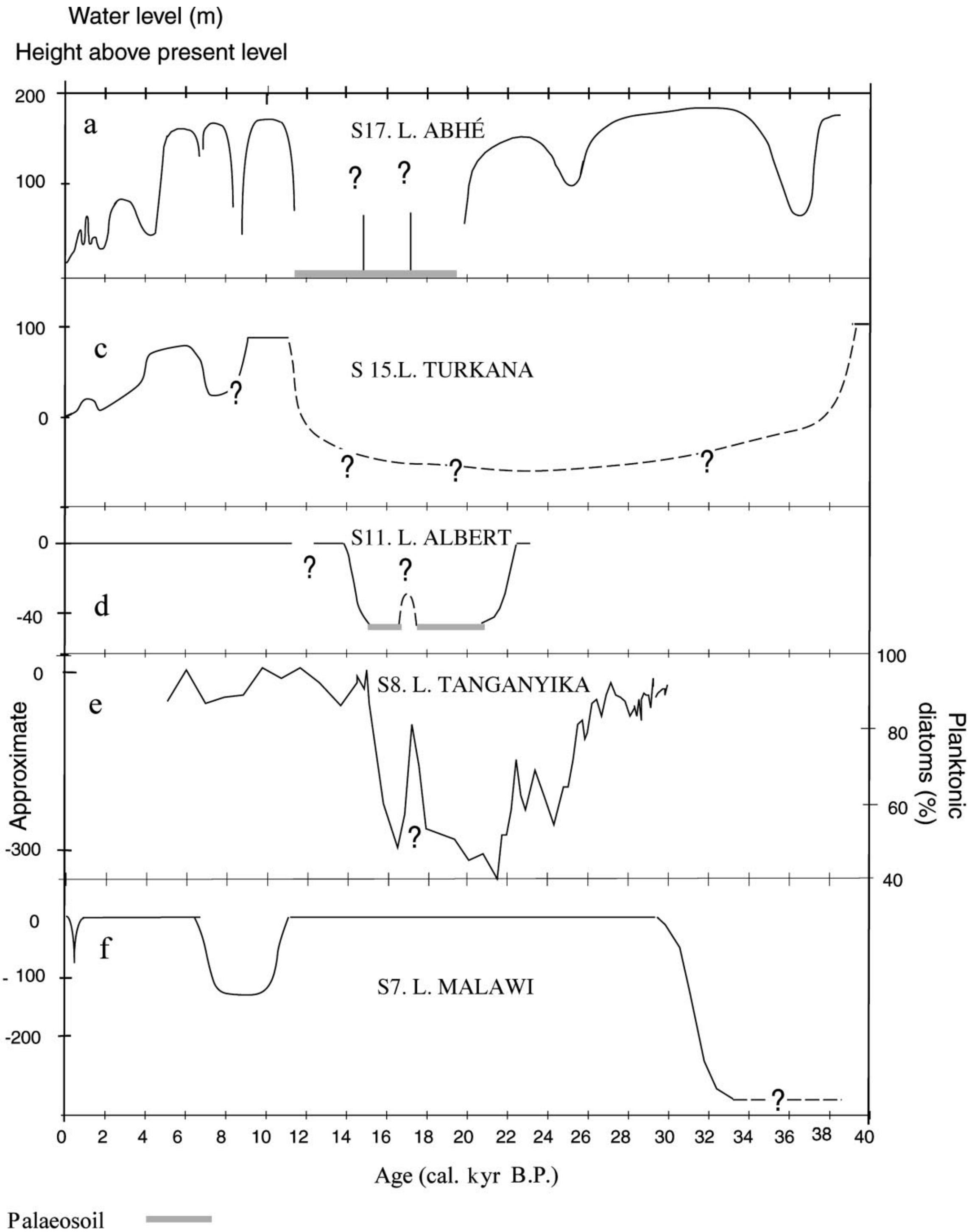

Fig. 7. The amplitude of water-level fluctuations in rift lakes from East Africa. Examples of: (a) Lake Malawi (after Finney et al., 1996 and Johnson, 1996). (b) Lake Tanganyika, Planktonic diatom percentage in core MPUXII (after Gasse et al., 1989). (c) Lake Albert. Lake level curve based on data from Talbot and Livingstone (1989) and Beuning et al. (1997). (d) Lake Turkana (after Johnson, 1996). (e) Lake Abhé (after Gasse and Street, 1978).

a step-wise shrinking of more than 160 m between about 27 and 20 ka when the lake dried up. The White Nile River channel was partly blocked by sand dunes (Said, 1993). In the southern Sahara, the Jebel Marra crater lake (S18) experienced its minimum level at ca. 20.2 ka (Williams et al., 1980).

In the northern Sahara, the chronology of the few records sometimes assigned to the LGM on the basis of

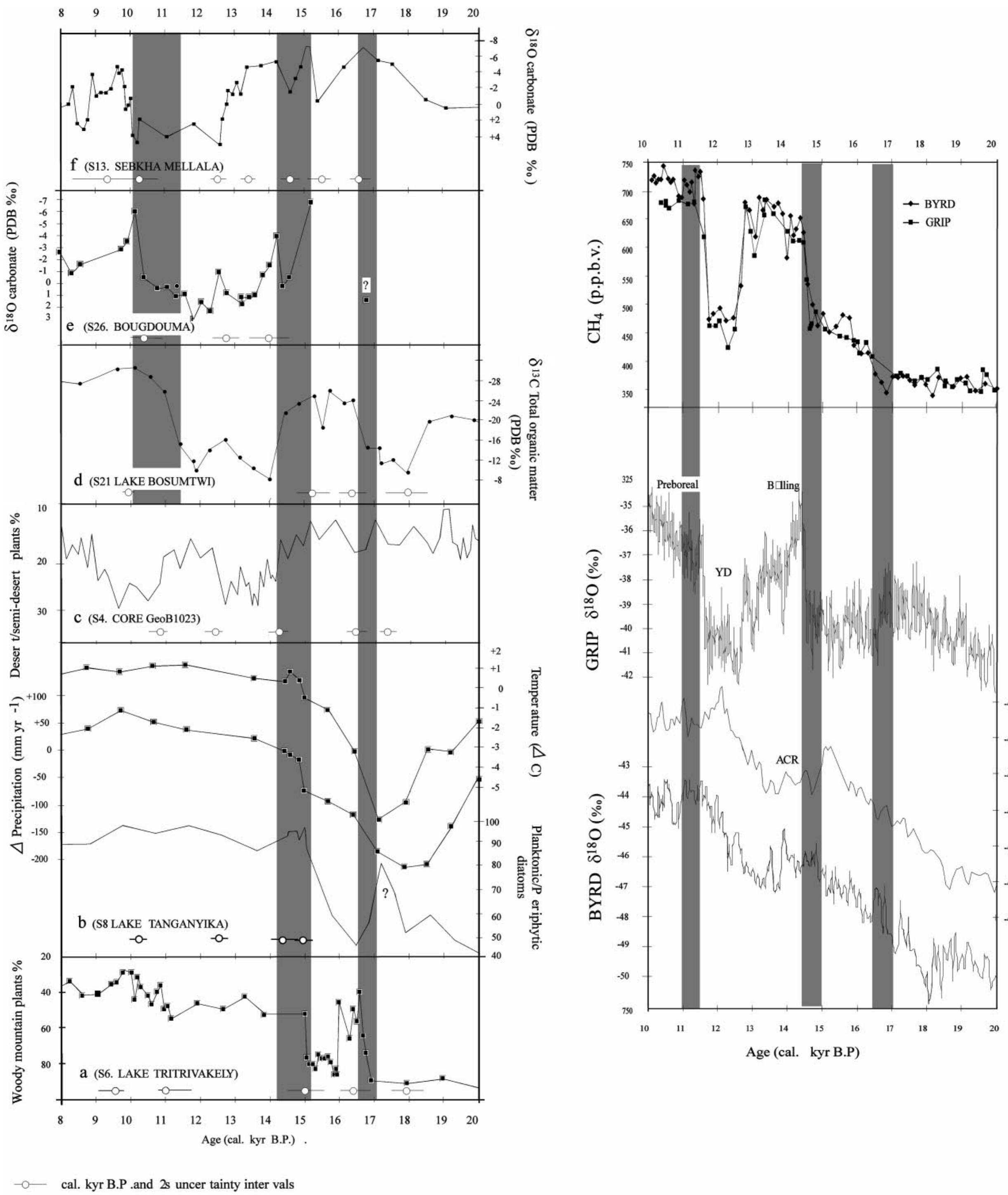

Fig. 8. Left: summary of some African lake and pollen records (20–8 ka). (a) Lake Tritrivakely. Frequency of woody mountain pollen taxa (as for Fig. 6b). (b) Lake Tanganyika, core MPU XII (southern basin). Percentage of planktonic diatom taxa (as for Fig. 7b) and pollen-inferred difference (■) from modern temperature and precipitation (after Chalié, 1995, and F. Chalié, pers. comm., 1999). (c) Marine core GeoB1023. Percentage pollen diagram of desert/semi-desert plants showing arid intervals at 14.4–12.5 and 10.9–9.3 ka (adapted from Ning Shi et al., 1998). (d) Lake Bosumtwi. Changes in $\delta^{13}C$-values of total organic matter (after Talbot and Johannessen, 1992). Positive excursions are interpreted as the result of arid climatic intervals which favoured a dominance of C4 plants in the crater and caused evaporatively driven changes in the chemical and isotopic composition of surface water dissolved inorganic carbon pool. (e and f) Changes in ^{18}O content of authigenic carbonates in waterbodies from the southern (e: Bougdouma) and northern (f: Sebkha Mellala) margins of the Sahara, indicating fluctuations in the ratio of groundwater influx to evaporation. Negative excursions reflect wet pulses. High values reflect arid intervals (after Gasse et al., 1990). Right: GRIP (Greenland), Vostok and Byrd (Antarctica) isotopic records (after Johnsen et al., 1972, 1992; Jouzel et al., 1987, 1992), and CH_4 (GRIP and Byrd) records on a common time scale (Blunier et al., 1998). Vertical dashed bands show the major warming/wetting phases in the African tropics during the last deglaciation. YD: Younger Dryas; ACR: Antarctica Cold Reversal.

^{14}C ages (e.g., Street-Perrott et al., 1989; Jolly et al., 1998) has to be reassessed but U/Th dating (Causse et al., 1988). The aquifer of the "Continental Intercalaire" (S34) experienced a period of intense recharge from ca. 45 to 23.5 ka when noble gas-inferred temperature was 5°C lower than the modern mean annual temperature (Guendouz et al., 1998). Water supplying the aquifer was depleted in heavy isotopes compared with late Holocene palaeowater and modern mean precipitation. From Morocco to the Western Desert of Egypt, an overall west-to-east decrease in heavy isotope content of Late Pleistocene fossil groundwaters is observed, suggesting that the relative influence of the mid-latitude westerlies was reinforced on the Sahara (Sonntag et al., 1978; Sultan et al., 1997) during the glacial period, in symmetry with the Namibian desert.

To sum up the above data, much of the continent experienced cooler conditions than today during glacial times, and a marked decrease in P or in P–E at the LGM.

3.2. *African climates, precession cycles, and glacial boundary conditions.*

The main points arising from these observations are as follows.

(1) Due to the geometry of orbital precession, changes in summer insolation are in antiphase between hemispheres (Berger, 1978: Fig. 4). Increased summer insolation should result in reinforced monsoon circulation by enhancing the ocean-land pressure contrast which draws the monsoon winds inland (Kutzbach et al., 1993; Kutzbach and Street-Perrott, 1985). Orbital forcing predicts dry climates in the northern tropics, and enhanced monsoon rainfall in the southern tropics during the LGM. In the northern hemisphere, long deep-sea records (e.g., McIntyre et al., 1989; Clemens and Prell, 1991) have demonstrated that the Pleistocene monsoonal climates have been paced by Milankovitch cycles (Fig. 4); low lake-levels and low aquifer recharge during the LGM are consistent with orbital forcing. In the southern subtropics, the Pretoria Saltpan time series of summer precipitation (Partridge et al., 1997; Fig. 4) provides a test of the role of orbital forcing on global climate from 200 to 50 ka, although results should be confirmed by other records supported by a better chronology.

(2) Nevertheless, the fact that the Pretoria Saltpan climate became drier after 30 ka does not fit with the Milankovitch theory. Dry LGM conditions observed at sites S1, S2, S4 and S6 in the southern tropics cannot be accounted for by changes in the Earth's orbital parameters. Hydrological records appear to show that the tropical African climate has rather responded to changes in sea-surface conditions and ocean–atmosphere interaction. It is now established that tropical SST fell by several degrees C at the LGM (Guilderson et al., 1994; Schneider et al., 1995; Beck et al., 1997; Bard et al., 1997), contrary to CLIMAP (1981) estimates. Lower tropical SST induced a decrease in evaporation at low latitudes, the major source of water vapour, that further intensified cooling through a decrease in the atmospheric concentration of this greenhouse gas. Recently, a coupled ocean–atmosphere climate model simulates a cooling and a decrease in summer precipitation in most of the tropics, although a slight increase in winter rainfall occurs in some regions of southern Africa (Ganopolski et al., 1998a); the tropical cooling is due to increased oceanic heat transport out of the tropics induced by enhanced meridional temperature gradient and to stronger northern hemisphere trade winds, and to the reduction of atmospheric water vapour and CO_2 concentrations. Bush and Philander (1998) present a LGM simulation which shows that a large part of the LGM temperatures lowering is due to a decrease in water vapour atmospheric concentration; the global hydrological cycle is weaker: both evaporation and precipitation are reduced by about 10%. Model simulations thus provide some explanation for the geological observations summarized here, although, in all models, precipitation simulations are less reliable than temperature predictions (Ganopolski et al., 1998a).

(3) Spatial and temporal variations are observed. The apparent exception of the Lake Malawi basin, believed wet at the LGM may be due to the complexity f the regional topography and rainfall pattern. Some records (e.g., S19, S21) show that the interval 23–18 ka has been complex, with relatively wet periods e.g. around 19 ka.

4. The last glacial–interglacial transition

The question of interhemispheric symmetry or asymmetry of major climate changes during the last deglaciation is currently a matter of debate (e.g., Sowers and Bender, 1995; Lowell et al., 1995; Bard et al., 1997; Ariztegui et al., 1997; Steig et al., 1998; Blunier et al., 1998). Discussions are mainly based on polar ice core records (Fig. 8). While the Greenland isotopic records (Johnsen et al., 1992; Blunier et al., 1998) evidence two major abrupt warmings at the onset of the Bölling and Preboreal periods (14.6 and 11.6 cal. kyr BP, respectively), the temperature increase in Antarctica began as soon as 21 cal. kyr BP at Byrd and 18.5 at Vostok (Johnsen et al., 1972, 1992; Jouzel et al., 1987; Blunier et al., 1998). Paleoclimatic changes in Africa in the time interval 20–8 ka (Figs. 7 and 8) appear compatible with both an early ending of full glacial conditions, as observed at Vostok or Byrd, and further major steps toward full interglacial climates, as observed in Greenland.

4.1. Evidence for the step-wise onset of postglacial climate.

At site S1 (Pretoria Saltpan), a trend toward higher monsoon precipitation is observed from about 18 ka, although precipitation remained low until the mid-Holocene (Partridge et al., 1997; Fig. 4). Several terrestrial records in South Africa indicate that increased wetness is associated with warming which started at 17–16 ka (Partridge, 1997); dry conditions then established during the early Holocene (Partridge, 1997; Scott, 1993). At Stampriet (S3), deglacial warming started at 17.5–17 ka, and was about 60% achieved at ca. 16 ka (Fig. 5; Stute and Talma, 1998). The marine pollen record at site S4 (Ning Shi et al., 1998) shows a marked temperature and humidity increase from ca. 17.5 to 16.5 ka, and very dry periods at 14.4–12.5 ka and at 10.9–9.3 ka (Fig. 8). At Tritrivakely (S6; Figs. 6 and 8), the interval 22–7 ka shows a very low sedimentation rate and some short-term desiccation episodes. However, detailed pollen and diatom studies indicate that warm conditions and a water body favourable to aquatic life re-established in two steps, around 17.5–17 and 15 ka, respectively (Gasse and Van Campo, 1998). The Lake Malawi record (S7) suggests high lake level until ca. 11.5 ka, when the lake level declined rapidly (Finney et al., 1996; Fig. 7).

In the equatorial East African Rift lakes, the major event is the lake basins refilling around 15 cal. kyr BP, but an earlier wetting is apparent, although poorly dated. In Lake Tanganyika (S8; Figs. 7 and 8), a generally positive water budget was restored from ca. 21 ka; a first positive oscillation may have occurred around 17 ka; deep conditions and perennial stratification were established by 15 ka (Gasse et al., 1989). From pollen-based reconstructions in the Southern Tanganyika basin, precipitation and temperature steeply increased from ca. 18–17 to 15–14.5 ka (Chalié, 1995; F. Chalié, pers. com. 1999; Fig. 8). At Lake Victoria (S10), two major late Pleistocene lowstands were registered by palaeosols, and were separated by a minor transgression: the earlier and more intense dry interval was between ca. 20.5 and 17.9 ka, and the second ended at ca. 15.3 ka (Talbot and Livingstone, 1989), when the basin refilled rapidly (Johnson et al., 1996). This record fits well with the hydrological history of Lake Albert (S11; Fig. 7): a brief and limited transgression is also evidenced by lake sediments bracketed between two palaeosols that developed between ca. 20.7 and 15.3 ka; shortly after 15.3 ka, a major transgression led to open-water conditions (Beuning et al., 1997). A dry episode fitting the cool Younger Dryas interval (YD) has been identified in several equatorial lakes, e.g. Lake Kivu (S9; Haberyan and Hecky, 1987) and Lake Magadi (S12; Roberts et al., 1993).

In subequatorial West Africa, a major arid–humid transition starting around 11.5 ka follows several climatic oscillations. At Lake Bosumtwi (S21), the stable isotope record in lacustrine organic matter (Fig. 8) registers relatively wet periods between 21.5–18.5 and 17–15 ka followed by very dry intervals centred around 17.5, 14.0 and 12 ka, and a dramatic wetting event at 11.5–10.5 ka (Talbot and Johannessen, 1992). The latter is coincident with a major lake-level rise (Talbot et al., 1984) when the closed forest settled (Maley, 1991). At Lake Barombi Mbo (S19), changes in the Cyperaceae pollen frequencies suggest that the lake level, intermediate around 19, 16 and 15–14 ka, later dropped drastically; the major rise occurred at 11.5–11 ka (Maley and Brenac, 1998).

North of 5°N, the prominent postglacial hydrological change was the flooding of a multitude of depressions at the onset of the Holocene. In the Sahelian belt, it has long been established that the monsoon reactivation occurred in two steps, at ca. 15–14.5 and 11.5–11 ka, separated by a return to drier conditions coincident with the YD (e.g., Street-Perrott and Perrott, 1990; Gasse et al., 1990; Gasse and Van Campo, 1994; Fig. 8). The former step only induced a water-level rise of moderate amplitude (Fig. 9). Earlier moisture increases are, however, indicated by isotope and sediment studies of a marine core (S20) from the Niger River delta: first runoff increases in the Niger basin occurred prior to 15.8 ka (Pastouret et al., 1978; Durand, 1993). In the southern Sahara, the Jebel Marra crater lake (S18) was 7.5 m higher than the lowest LGM level at 16.5 ka (Williams et al., 1980). In northern Niger (S31) and northern Mali (S32), dated groundwaters suggest that the post-glacial recharge of the aquifers began around 16.5 ka (Fontes et al., 1993); shallow lakes and swamps developed in the depressions around 15 ka (e.g. at S30; Gasse and Fontes, 1992). In Libyan Sahara (S33), U/Th datings of travertines indicate a wet episode in the Tadrart Acacus massif from 15.6 to 9.7 ka (Carrara et al., 1998). At the northern margin of the Sahara, the Sebkha Mellala record (S35; Fig. 8) shows a first increase in precipitation in the M'zab heights or in the Saharan Atlas mountains centred at 16.4 ka; further short-term wet pulses occurred at ca. 15 and 14.0 ka, before an arid period (12.3–10.3 ka) that preceded the onset of Holocene permanent lacustrine conditions (Gasse et al., 1990).

The successive steps in the re-establishment of rainfall in North Africa show good consistencies with the major abrupt events of Indian monsoon intensification recorded in core 74 KL from the Arabian Sea (Sirocko et al., 1996), and dated at 16.0, 14.5 and 11.45 ka.

4.2. Implications for north–south climatic connections

Summing-up the above observations, the last deglacial period in Africa appears to have been a series of abrupt transitions from arid to humid, or humid to arid conditions. Major events occurred around 17–16, 15–14.5, and 11.5–11 ka, but their expression and relative strength varies with latitude (Figs. 7 and 8).

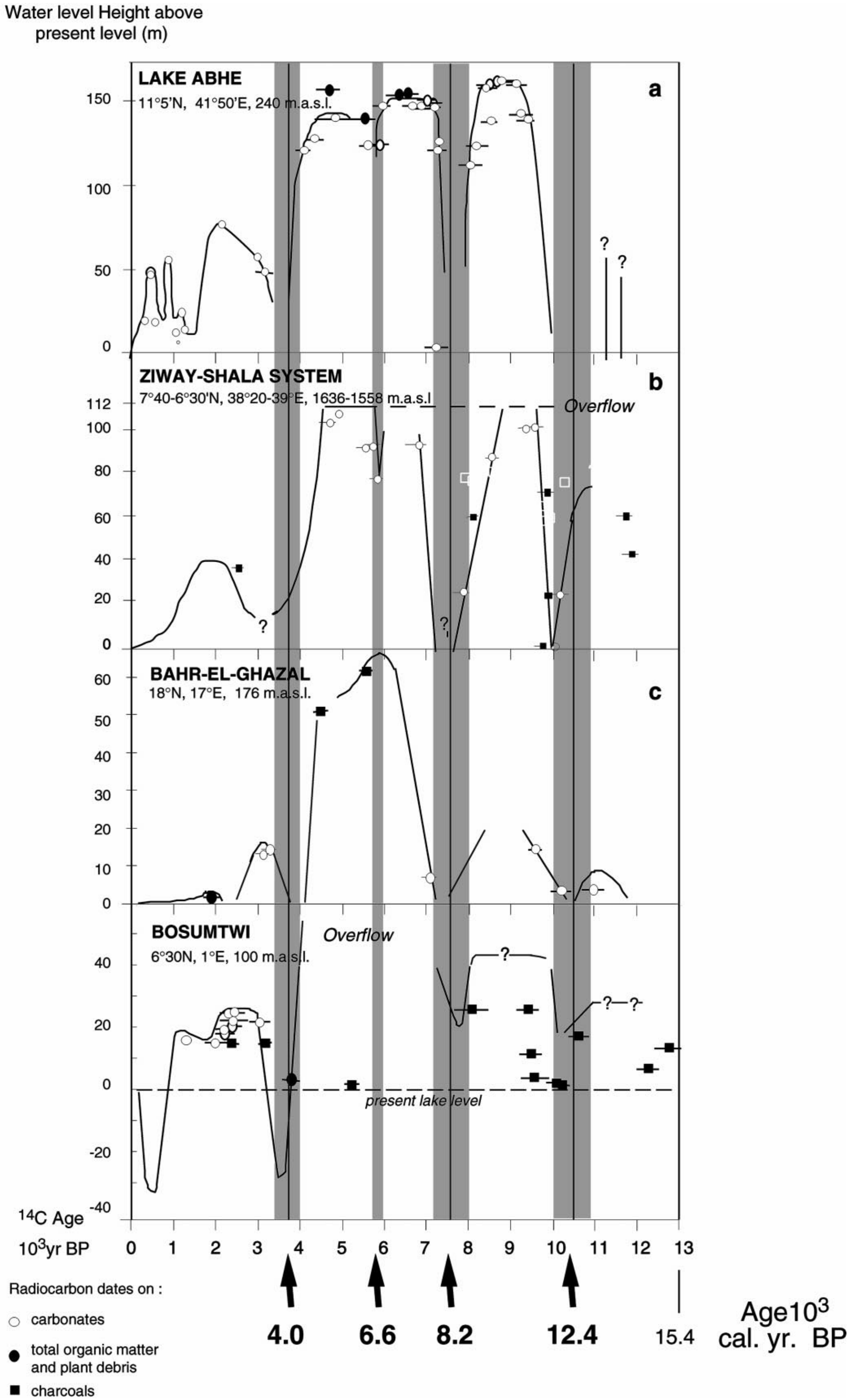

Fig. 9. Changes in lake level over the past 13^{14} ka in the northern monsoon domain of Africa. After Gasse and Van Campo (1994). From East to West: (a) Lake Abhé (from Gasse, 1977); (b) Lake Ziway-Shala (from Gillespie et al., 1983); (c): Bahr-el-Ghazal (from Servant and Servant-Vildary, 1980); (d) Lake Bosumtwi (from Talbot et al., 1984). Dashed lines underlined weak monsoon episodes.

The 17–16 ka event is an arid–humid transition of low amplitude in both hemispheres. Associated warming is documented in southern Africa (S3, S4, S6) and in the southern Lake Tanganyika basin (S8). It appears coincident with the largest part of the temperature increase at Byrd and Vostok in Antarctica, with the starting of the increase in global atmospheric methane concentration (Fig. 8), and with rapid climate changes recorded at many other sites in the southern hemisphere. These include: SST increase in the Southern Ocean (Pichon et al., 1992); rapid warming around 16.5 ka which induced a widespread glacier collapse in South America and New Zealand (Lowell et al., 1995); a significant increase in lake-level in the Bolivian Altiplano prior to the major water rise at ca. 15 ka (Sylvestre et al., 1998); and the initial warming in the Huascaran ice core in Peru (Thompson et al., 1995). Initial warming in the southern hemisphere and tropical regions may have increased the atmosphere evaporative power, when the North Atlantic deep-water (NADW) production and the oceanic thermohaline circulation were still in glacial mode (Charles and Fairbanks, 1992). Thus, water vapour may have been redistributed over the tropics through the atmospheric circulation. The 17–16 ka event leads the major deglacial events of the northern high latitudes by about 2000 yr. This delay could be due to a stationary polar front located at 45°N in the North Atlantic until about 15 ka; it may have insulated areas to the north from the global warming (Sowers and Bender, 1995).

A return to dry conditions, poorly dated but between ca. 16 and 15 ka, is identifiable at Madagascar (S6), Lakes Victoria (S10) and Albert (S11). This reversal event has no clear equivalent in polar ice sheets (Fig. 8).

The warming/wetting events around 15 and 11.5 ka dominate north of 10°S, but the former is weak in the northern tropics and equatorial West Africa. Both roughly match the abrupt temperature rises observed in Greenland ice cores and separated by the YD cold spell, and large and rapid increases in atmospheric methane concentration (Fig. 8). The African events were already associated with the switch to modern mode of NADW production (Street-Perrott and Perrott, 1990; Gasse et al., 1990; Gasse and Van Campo, 1994; Broecker et al., 1998). The reactivation of the oceanic thermohaline conveyor belt at ca. 15 ka, and thus of the Algulhas current, may have acted on climates in southeastern Africa and Madagascar.

On the Atlantic side of Africa, the dry interval recorded at 14.4–12.5 ka at site S4, also apparent at S19 and S21, rather appears coincident with the Antarctic Cold Reversal (ACR; Jouzel et al., 1995; Fig. 8) than with climate events in the North Atlantic region. It possibly originated from an increased influence of the cold Benguela current along the Atlantic coast. But clearly, the southern S4-record is in antiphase with the S21-record after 12.5 ka.

Extremely wet conditions established in equatorial and northern Africa at 11.5–11 ka. At the same time, African climates south of 10°S turned drier when the warming trend stopped in Antarctica (Fig. 8). Only after glacial boundary conditions decayed, did Southern Africa become sensitive to orbital forcing, in agreement with climate model sensitivity experiments at 10 ka (e.g., deMenocal and Rind, 1996).

Four conclusions arise from these observations.

(1) In tropical Africa, full glacial conditions ended early. A first warming/wetting phase starting around 17 ka led the abrupt changes in Greenland by several millenia, but appears coincident with the steepest temperature increase recognized at Byrd and Vostok in Antarctica.

(2) Drastic arid–humid transitions around 15 and 11.5 ka north of 10°S are thought to match the major Greenland deglacial warming events, in concert with the switching of the oceanic thermohaline circulation to modern mode.

(3) However, part of the signal after 15 ka in southern and Atlantic equatorial Africa rather seems connected with climate changes at high southern latitudes, as observed in continental Antarctica.

(4) A synchrony is observed between major wet and dry pulses north of 10°S (the largest part of the African continent), and rises and falls of the global atmospheric methane concentration. The amplitude of the hydrological steps even follows that of the CH_4 shifts. Bearing in mind that most of the modern methane is produced in tropical wetlands, the step-wise transition toward wetter conditions supports the hypothesis of positive feedbacks from the African tropics to the global deglacial warming through greenhouse gas production (Chappellaz et al., 1993; Blunier et al., 1998).

5. The Holocene period

Hydrological changes of extremely large amplitude occurred during the Holocene in the African tropics and subtropics. Although their general envelope is roughly consistent with orbitally induced variations of the monsoon strength, changes in efficient moisture did not simply respond to the smooth sinewaves of orbital forcing.

5.1. Evidence for major Holocene climatic changes

In southwest Africa, the Stampriet isotope record (S3) suggests a transition from a dry to a wet climate around 7 ka (Fig. 5; Stute and Talma, 1998). The S4 record shows: a very dry interval from 10.9 to 9.3 ka and the warmest and wettest Holocene period between 6.3 and 4.8 ka (Ning Shi et al., 1998) in phase with increased river discharge and weathering as indicated by clay mineral

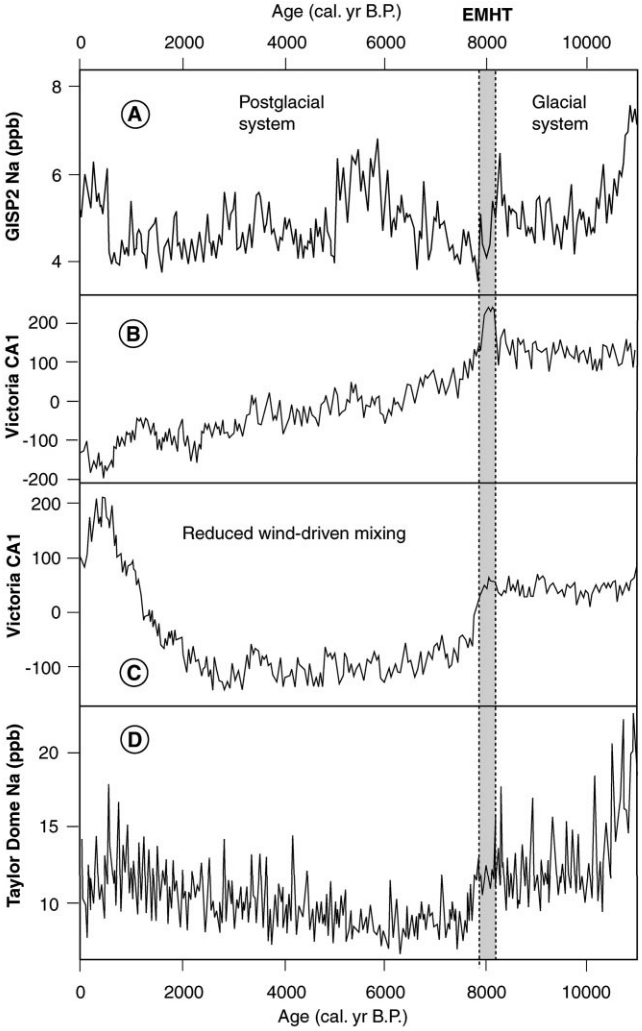

Fig. 10. Changes in limnological conditions at Lake Victoria over the past 11 ka as inferred from diatom assemblages (b and c), compared with windblown dust concentration in the GISP ice core in Greenland (a), and in the Taylor Dome ice core in Antarctica (d). From Stager and Mayewski (1997). Dust concentration after O'Brien et al. (1995), and Mayewski et al. (1995).

data (Gingele, 1996). In South Africa, dry/cool conditions prevailed during the early Holocene (Scott, 1989,1993; Tyson, 1991; Partridge, 1997), and "optimal" Holocene temperatures from latitudes 23–28°S occurred at 8–7 ka (Scott, 1993). At Lake Malawi (S7), an early Holocene low lake-level at -100 m, -150 m (Finney et al., 1996; Fig. 7) may correspond to a decrease in precipitation of ca. 29–25%, according to simulations based on the modern hydrological budget of this lake (Bergonzini, 1997).

During the late Holocene, cooler/drier conditions in the northern Kalahari occurred between 4.8 and 3.3 ka (Ning Shi et al., 1998). A temperature decrease in the Mediterranean belt of South Africa was also inferred between 4.5 and 2 ka from stable isotope data from Elands bay (Cohen et al., 1992), and around 4.5 ka from speleothems of Cango Cave (Talma and Vogel, 1992). A maximum in aridity is suggested around 4.2 ka at site S6 (Gasse and Van Campo, 1998).

North of 10°S, the general picture is a high P–E balance during the early-mid Holocene. In equatorial East Africa, maximum P–E balance may have occcurred during the early Holocene. A diatom record from Lake Victoria (S10; Stager and Mayewski, 1997; Fig. 10) shows a high P/E ratio (or lake level) and good wind-driven mixing from ca. 11 to 7.8 ka, with a maximum between 8.2 and 7.8 ka. An abrupt decrease in wind activity and in lake-level then occurred, in response to enhanced rainfall seasonality and/or a tendency toward aridity. Wind-driven mixing remained low until ca. 3–2 ka; the P/E ratio progressively dropped until today. Hastenrath and Kutzbach (1983) estimated from a water and energy balance model that, at around 9 ka, mean annual rainfall in the basins of lakes Victoria (S10), Naivasha (S13), and Nakuru-Elmenteita (S14) was respectively 20, 15, and 35% higher than today. Lakes Tanganyika (S8), Victoria (S10), Albert (S11) overflowed, and supplied the Congo and the White Nile rivers. Lake Kivu (S9) rose rapidly by 11.5 ka, and overflowed toward Lake Tanganyika at ca. 10.5 ka. This lake remained high up to ca. 4.5 ka, and then dropped. It was closed again from ca 3.5 to 1.3 ka (Haberyan and Hecky, 1987). A lowering of water level took place at Lake Turkana (S15) from ca. 5 ka (Fig. 7) and a closed-basin lake status was achieved permanently at ca. 4.2 ka (Johnson, 1996).

In equatorial West Africa, maximum wetness occcurred during the mid-Holocene, from about 8–7.5 to 4 ka; drier conditions then established and culminated around 2.2 ka, before a return to wetter conditions around 1.5–1 ka (Maley, 1991,1997; Maley and Brenac, 1998). Lake Bosumtwi (S21) experienced marked drops in lake level centred around 8.3–8 ka, 4.2–4, and 0.5 ka (Talbot et al., 1984; Fig. 9).

In the Sahelian belt, the Bahr-el-Ghazal depression (S23), which received the outflow of Lake Chad, and the Ethiopian lakes (S16–S17) experienced an evolution very similar to that of Lake Bosumtwi (Fig. 9), especially in the timing of short-term dry events around 8.3–8 ka and 4.2–4 ka. In response to dry conditions in both Ethiopia and equatorial Africa, the Main Nile flood was considerably reduced around 4.2 ka (Said, 1993). A minor dry spell is also recorded at 7–6.5 ka in Ethiopia (Fig. 9) and in several Sahelian sites (e.g., S26–S27; Gasse and Van Campo, 1994). High resolution records of Bal Lake and Kajemarum Oasis (S25; Holmes et al., 1999) show that the early mid-Holocene organic silt accumulation, which reflects wet conditions, was interrupted by eolian sand layers at 8.1 and 4.1 ka; a marked climate deterioration started at ca. 4.1 ka with an extremely severe drought at 1.2–1.0 ka.

An early-mid-Holocene "green" Sahara is evidenced by numerous palaeolake, palaeodrainage, vegetation, and archeological records (see, e.g., Hillaire-Marcel et al.,

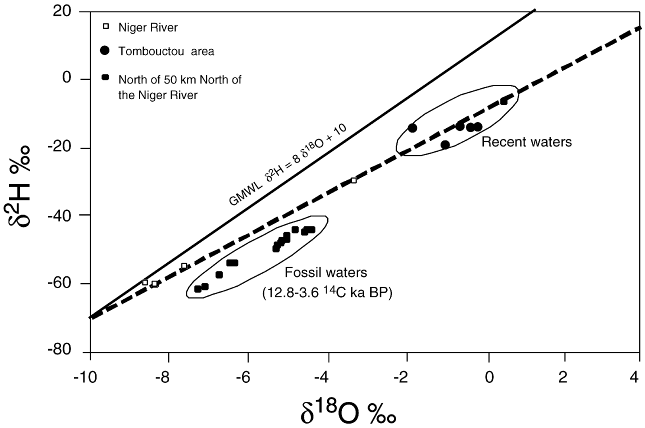

Fig. 11. Groundwater stable isotope composition from the semi-confined aquifer complex of the "Route du Sel" in northern Mali, as an example of Holocene paleowaters in the Sahara. Comparison with modern water from the Niger River, and recently infiltrated waters close to Tombouctou. GMWL: Global Meteoric Water Line. After Fontes et al. (1993).

1983, Baumhauer, 1991; Fontes and Gasse, 1991; Petit-Maire et al., 1993; Pachur and Altmann, 1997; Hoelzmann et al., 1999; Cremaschi and Di Lernia, 1998, for regional syntheses). Between 17 and 18°N, a savanna ecosystem was maintained without interruption during the early and mid-Holocene, when parts of the large, now arid, endorheic region of Western Nubia drained into the Nile via the Lower Wadi Howar (Fig. 3; Kroepelin, 1993). In western Fezzan (S33), a wet savannah landscape inside the mountains and shallow lakes in the ergs developed from ca. 10 to 6 ka, connecting the Sahelian area to the northern Sahara; a dry period at 8.5–8 ka interrupted this wet interval (Cremaschi and di Lernia, 1998). In the western Sahara, the northern fringe of the palaeo-Sahel has reached about 22°N (Petit-Maire and Riser, 1993). Some depressions (e.g., S28) experienced two lacustrine episodes between 10 and 4.5 ka separated by a long dry phase centred around 8 ka; others remained dry after 8 ka (e.g., S29). Permanent lakes also occurred along the northern margin of the Sahara during the early-mid-Holocene: the Sebkha Mellala (S35) experienced two freshwater lacustrine episodes between 10.3 and 5.7 ka, separated by a desiccation phase at ca. 8.5-8.0 ka (Gasse et al., 1990); shallow waterbodies have occurred at site S36 from ca. 11–10 ka to 3.3 ka (Gasse et al., 1987), while high lake-level persisted until today in the Atlas Mountains of Morocco (Lamb et al., 1995). Dry conditions prevailed after 6–5.5 ka in the eastern Sahara, and 4.5–4 ka in the whole Sahara, although a further late Holocene wet episode of minor amplitude is recorded at several places (Maley, 1997). General aridity took place at ca. 2 ka.

The wet Holocene period has also been a period of intensive recharge of the Saharan deep water reservoirs. Early-mid-Holocene waters of northern Mali and northern Niger (e.g. Fontes et al., 1993; Dodo and Zuppi, 1997), Sudan (Darling et al., 1987) and Libya (Edmunds and Wright, 1979) differ from late Holocene and modern rainfalls, especially in their very low deuterium excess (Fig. 11) for which no satisfactory explanation has yet been proposed. Their very low $\delta^{18}O$-values ($\delta^{18}O$: $-10‰$ to $-12‰$) were attributed to convective showers generated by squall lines that develop along the ITCZ (Fontes et al., 1993). This hypothesis implies an ITCZ northwards movement of at least 500 km, in agreement with palaeolake data. Such a migration of Sahelian environments is confirmed by the high dissolved nitrogen content in Holocene groundwater (up to 45 mg kg^{-1} of nitrate in northern Mali), reflecting biologically active soils (Fontes et al., 1993).

5.2. *The greening of the Sahara*

Climate and vegetation simulations have been used to examine the influence on the African monsoon, of Earth orbital changes, SST changes, and feedbacks associated with changes in soil moisture at 6 ka, but most models fail to simulate conditions much wetter than today in the western Sahara: in simulations by AGCMs (Kutzbach et al., 1993; de Noblet et al., 1996; Jolly et al., 1998), a coupled ocean-atmosphere model (Kutzbach and Liu, 1997), and a coupled atmosphere-biome model (Texier et al., 1997), the amplitude of the climate change appears underestimated. Recently, a synchronously coupled atmosphere–ocean–vegetation model showed that changes in vegetation cover during the mid-Holocene modify and amplify the climate system's response to an enhanced seasonal cycle of insolation in the northern hemisphere (Ganopolski et al., 1998b). This model results show strong synergetic effect of changes in vegetation cover, ocean temperature, and sea-ice at boreal latitudes, but the atmosphere–vegetation feedback is most important in the subtropics. The model simulates increased precipitation over the whole northern Africa, and thus provides a better level of agreement with geological data from northwest Africa. It should be noted, however, that

(1) The 6 ($\pm$ 0.5) ka snapshot which was selected for simulations makes the model-data comparison difficult in Africa, where this time slice falls during a period of climate instability.

(2) Many of the early-mid-Holocene waterbodies of the Sahara-Sahel were the surface expression of water-table rises in the aquifers. Waterbodies and soil moisture in the lowlands were generated by local precipitation, but may have also resulted from groundwater flows from remote highlands. Furthermore, a delay may have occurred between the onset of arid conditions and the final desiccation of lowlands. For instance, in the Nubian aquifer system, the very low rate of drawdown fits the

idea of continuously decreasing groundwater levels since the recharge period, about 7 ka. The hydraulic head decreased to a regional mean of 60 m below the ground surface, but groundwater is still fairly close to ground surface in the depression areas (Christman and Sonntag, 1987).

(3) Today, the Sahara north of 20–24°N belongs to the "winter rainfall pattern", while precipitation in the southern Sahara is associated with the summer monsoon (Fig. 2). Hydrological changes at sites at 30–32°N presently supplied by winter precipitation on the Atlas and the northern Sahara heights can hardly be attributed to a monsoonal climate. It is not possible yet to determine the seasonal distribution of past precipitation, and thus to identify the sources of water vapour responsible for "greening" in the northern Sahara.

5.3. Holocene hydrological and climatic instability

Attention is paid here to some major events apparent at numerous sites around 8.5–7.8, 7–6.6 and 4.5–3.5 ka, and which have affected regions well beyond the tropics and subtropics of Africa. These events have different regional expressions. They are either a step-over shift to a new millenial-duration background climate, or a short-lived fluctuation rapidly followed by a return to initial conditions.

Rapid climate changes occurring in the interval 8.5–7.8 ka probably correlate with the most prominent Holocene climatic event observed in Greenland ice-cores, and recorded there by the only distinct Holocene ^{18}O shift and a significant decrease in methane at ca. 8.4–8.0 ka (Chappellaz et al., 1993; Alley et al., 1997). The Greenland event fits with one of the Holocene cool, ice-bearing water episodes (nr 5) in the North Atlantic (Bond et al., 1997). It is coincident with a peak in aerosol concentrations in polar ice cores from Greenland and Antarctica. This peak was followed by an abrupt shift at 7.8 ka (Fig. 10). After 7.8 ka, windblown chemical indicators recovered quickly in Greenland ice, while they remained low in Antarctica.

At Lake Victoria (Fig. 10), the 8.5–7.8 ka event led to a long period of decreasing P–E balance that maintained until today. Stager and Mayewski (1997) concluded that Holocene climatic fluctuations in equatorial East Africa are linked to changes in Antarctica weather and circumpolar ocean currents. In subequatorial Atlantic Africa, e.g. at Lake Bosumtwi (Fig. 9), a 8.5–7.8 ka dry event preceeded the episode of maximum $P - E$ that lasted over about 4000 yr. It thus corresponds with the onset of specific mid-Holocene climate conditions, as at many places in the world: generally cooler, wetter conditions in the western Mediterranean basin (Kallel et al., 1998); very low effective moisture in the Amazon basin and in the Chilian and Bolivian Altiplano (Valero-Garcès et al., 1996); reduced precipitation and increased windiness throughout much of the eastern two-thirds of the United States (Bradbury and Dean, 1993). In the African northern tropics and in the northern Sahara, a dry spell centred at 8.5–8 ka interrupted a generally wet climate episode. This short-life dry event, apparently in phase with a weak monsoon episode in the Indian monsoon domain of West Asia (Gasse and Van Campo, 1994), is followed by a return to "initial" conditions, as the cool spell that affected high latitudes of the northern hemisphere.

A dry event in the northern tropics around 7–6.5 ka was recorded in Africa (e.g., S16, S16, S26, S27) and in West Asia (Gasse and Van Campo, 1994). In western Tibet, $\delta^{18}O$-records of lacustrine carbonates suggest that this event marks the start of the monsoon retreat and a change in the origin of air moisture, from monsoonal to locally re-evaporated moisture under drier conditions (Wei and Gasse, 1999).

As with the ca. 8.2 ka event, climate changes at the mid to late Holocene transition, ca. 4.5–4 ka, either corresponded to a short-term dry/cool episode, or inaugurated new atmospheric, climatic conditions. At many sites (S9, S15–17, S21, S23, S27, S28), the 4.5–4 ka event is a dry spell followed by a further wet phase of low amplitude. It coincides with an unusual dust event identified at 4.2 ka in the eastern Mediterranean and across Mesopotamia, from Lake Van to the Gulf of Oman (Dalfes et al., 1997), and with the episode of maximum Holocene aridity in Western Tibet (Gasse and Van Campo, 1994). A trend toward aridity, culminating around 2.2 ka, started at ca. 4.5–4 ka in equatorial West Africa. Cooler/drier climates took place in Southwest Africa (S4). The changes were opposite to those of the early-mid-Holocene transition, the 4.5–4 ka event coinciding with the re-establishment of drier and warmer conditions in the West Mediterranean basin (Kallel et al., 1998), the onset of generally cooler climates and increased moisture in the north-central United States (Bradbury and Dean, 1993), increased wetness in Amazonia and Central Andes leading to drastic lake-level rises of Lake Titicaca (Martin et al., 1993; Mourguiart et al., 1998). In South America, this transition toward wetter conditions was interpreted as a change from warm environments without El Niño episodes to the present cooler climate interrupted by El Niño every few years (Markgraf et al., 1991).

The causes of these abrupt Holocene events are not known. Alley et al. (1997) hypothesized a decrease in North Atlantic thermohaline circulation around 8.2 ka caused by an increase in freshwater flux to the North Atlantic. Gasse and Van Campo (1994) noted that the weak monsoon episodes around 8 and 4 ka are roughly coincident with decreases in North Atlantic SST and surface salinity. The antiphase between rainfall changes in Equatorial West Africa and in the Amazon may reflect changes either in the Walker circulation and ENSO phenomena, or in the Hadley circulation, e.g. a strengthening of the northern trade winds at ca. 4 ka, bringing dry, harmattan wind flows in Africa, and water vapour from the Atlantic in South America. Whatever their cause, the worldwide impacts of these events imply that rapid changes in the water cycle in Africa are the expression of global climatic reorganizations.

6. Conclusions

The palaeohydrological evidence collected here yields some insights on the sensitivity of the African climate to changes in global climate forcing factors. It also raises a series of questions that remain to be solved. The main points arising from the palaeohydrological evidence are as follows.

(1) As a whole, climate information in Africa suffers large geographical and methodological gaps. First, the climate evolution in the southern tropics is poorly documented and the large Congo basin is practically unknown. Second, whatever the region, most records are undermined by low time resolution and large dating uncertainties. Third, African palaeohydrological data are mainly qualitative while model-data comparisons require quantitative estimates of past hydroclimatic variables. Further studies are needed to estimate palaeo-temperature, palaeo-precipitation and palaeo-evaporation from noble gas groundwater records, speleothem isotope records, and modelling of individual lake systems combining water, salt, isotope, and energy balance equations. Fourth, multi-proxy and multi-archive studies need to be developed further. Recent works (e.g. Cremaschi and Di Lernia, 1998; Holmes et al., 1999) have shown that the combination of surface water (lakes, rivers), groundwater (aquifers, speleothems) and sand dune studies leads to better understand the system evolution. Such approaches should be made whenever possible, especially when different dating techniques can be applied, allowing a cross-control of the chronological framework. Fifth, correlations between inland and marine records, poorly developed so far, will be required to understand ocean–continent interactions. Marine cores close to the major African rivers mouth may help establish such correlations.

(2) At long-time scale, orbitally induced changes in summer solar radiation certainly account for a large part of hydrological changes in the tropics. The orbital precession appears to be a satisfactory explanation for the cyclicity in summer precipitation changes in the Pretoria Saltpan record (South Africa) from 200 to 50 ka, but the signal is blurred during the LGM. Very few long records are available. No comparison is possible between the penultimate and the last glaciations, or between the Holocene and the last interglacial periods. Other palaeoclimatic sequences spanning at least one climatic cycle will be required from both hemispheres to estimate the relative roles of the Milankovitch cycles and of ocean–land interactions on African climates under glacial and interglacial conditions. Potential sites to obtain long continuous records are the large Rift lakes of East Africa and numerous crater lakes from East and equatorial West Africa, and Madagascar. Archives for remote periods are rare in arid regions because of the wind deflation, but groundwaters and speleothems are promising.

In fact, over the period for which radiocarbon dating is possible, the hydrological history of Africa appears to have been a complex alternation of wet and dry episodes with abrupt transitions, which cannot be directly accounted for by orbital forcing. As already highlighted for the last deglaciation (Broecker et al., 1998), the Earth's climate jumped from one mode of operation to another rather than proceeding in smooth sinewaves. These abrupt changes reflect reorganizations in the transport of energy and water vapour at the global scale. The few records covering the past 40,000 yr (e.g., S6-record) show that the glacial period has been punctuated by short-term warm/dry pulses. Further work is needed to understand if theses pulses were related to the interstadial warm events in the GRIP ice cores and/or with SST variations in the surrounding tropical oceans.

(3) During the LGM, most records indicate generally dry conditions in both hemispheres, while orbital forcing predicts enhanced summer rainfall in the southern tropics. Geological evidence presented here roughly agrees with recent coupled ocean–atmosphere climate models which simulate cooler conditions and decreased summer rainfall in the tropics. Lower tropical sea-surface conditions have played a role on tropical climates by reducing the flux of water vapour inland.

We need, however, to obtain a better picture of regional LGM climates to fully understand the impact of changes in atmospheric circulation and oceanic currents along the eastern and western coasts of Africa. Because of the possible complexity of the 23–18 ka interval, model-data comparison should be based on a chronology more accurate than that available in many records. More improved paleoclimatic data, and more simulations with detailed models will be required to arrive at a robust quantitative understanding of hydrological changes in the tropics associated with a global hydrological cycle weaker than today.

(4) The first signs of the deglaciation in the African tropics and subtropics started around 17 ka, much earlier than the major deglacial events observed in the North Atlantic regions. Initial warming in the southern hemisphere may have affected the water cycle in the entire African continent at ca. 17–16 ka through changes in the atmospheric circulation. North of 10°S, sudden and very large increases in the water budget at ca. 15 and 11.5 ka were coincident with the onset of the Bölling-Allerod and

Pre-Boreal warm intervals identified in northern high latitudes. It was the reactivation of the oceanic thermohaline circulation that was apparently instrumental in bringing the interglacial shift to the continents.

Linkages between high southern latitudes and tropical Africa climates apparently maintained after 15 ka. The dry interval from 14.5–14 to 12.5 ka in southern and equatorial West Africa may be related to the ACR. Holocene climatic fluctuations in equatorial East Africa, e.g. at Lake Victoria, also appear to have been linked to the Antarctica climate. More detailed marine and continental records in the tropics are needed to reconcile low and high latitude climate records of both hemispheres.

(5) Once the glacial boundary conditions decayed, generally wet climate established in the northern tropics while the southern tropics became drier during the early Holocene, in response to orbital forcing. However, changes in summer insolation alone cannot account for some regional climate patterns, e.g., conditions much wetter than today in the northwestern Sahara around 6 ka, when the western Mediterranean basin was colder and wetter than today and when the eastern Sahara was becoming drier. Climate models showed that such a wetting induced complex interactions and feedback processes between the atmosphere, the oceans, and vegetation cover. However, the origin of water vapour over the Sahara is not identified. Stable isotope studies of Holocene groundwater along North–South and East–West transects could bring important insights on the moisture transport patterns. In this region where many waterbodies were supplied by aquifers, the water cycle could be better understood by combining quantified surface and groundwater data, and by modelling the entire hydrological systems. Annually laminated sediments from lakes or speleothems may also provide information about the rainfall season.

(6) Until recently, the Holocene climate was commonly thought to have been fairly stable. This idea was mainly based on polar ice and North Atlantic deep-sea core records where Holocene fluctuations are indeed of very low amplitude compared to those of the glacial-deglacial transition. On the contrary, dramatic hydrological changes have long been apparent in Africa, and appear as large in magnitude as their glacial counterparts. Dramatic changes in water resources have enormous consequences on human populations, generating famines, migrations, civilization foundations and collapses. The advanced urban civilizations in Egypt Hassan, 1997, Mesopotamia and India mysteriously collapsed around 4000 yr ago (Dalfes et al., 1997). This collapse of the Old Word societies is likely to be related to the 4.2–4 ka dry event that affected the monsoon domain and eastern Mediterranean. There is no doubt that Holocene drought and flood episodes have been more dramatic and more persistent than those of the instrumental record of the last century, which does not represent the full range of natural climatic variability of the current interglacial period. There is no guarantee that similar shifts with serious social impact cannot happen again. We need to estimate their amplitude, to better constrain their timings, to obtain high resolution Holocene records from the surrounding oceans, before concluding on their relations with short-term events known elsewhere (e.g. in the North Atlantic region) and to analyze potential mechanisms. An understanding of natural fluctuations in water resources that occur at the time scale of human lives should be a first priority in future research.

(7) Palaeohydrological data summarized in this paper show that Africa has experienced rapid changes in the continental water cycle of considerable magnitude, regardless of the presence or absence of large polar ice sheets. Because water vapour is the most powerful greenhouse gas, large hydrological changes in the tropics may have significant feedback effects on global climate. Further paleohydrological research should aim at obtaining solid chronologies, but also at analysing the mechanisms of water storage and losses in surface and groundwater reservoirs, obtaining quantitative reconstructions of hydrological cycles, and identifying moisture transport patterns at regional scales. Together with the development of climate-system models with finer spatial and temporal scales, this will allow a better understanding of how hydrological changes in the tropics respond to changes in major-forcing factors, and how the tropics influence climate globally.

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