NEW HIGH-RESOLUTION ALKENONE RECORD OF LAST GLACIAL TO HOLOCENE SEA-SURFACE TEMPERATURE CHANGE IN THE EAST-EQUATORIAL SOUTH ATLANTIC OCEAN

BY

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ABSTRACT. A new high-resolution, alkenone-derived record of sea-surface temperature (SST) change covering the last 26000 years was obtained from the east-equatorial Atlantic off the Congo River. Temperature fluctuations correspond to climate change recorded in other marine and terrestrial archives of the region. The maximum temperature difference between 26000 years BP ago and the Holocene climate optimum around 6000 years BP was 4.3°C, corroborating other SST estimates from the same area. The coldest conditions were followed by a warming that began at 24000 calendar years BP, a time when Northern Hemisphere ice sheets were still at their maximal position. This comparatively early warming is in agreement with previous findings from the east-equatorial South Atlantic. After a relatively stable period between 21500 and 14500 calendar years BP, a second warming began at 14500 calendar years BP which coincides with the onset of the African Humid Period. A cool period at 11500 calendar years BP halted this warming but the trend of increasing temperatures began again at 10000 calendar years BP, this time through large-scale oscillations. The warmest time, around 6000 calendar years BP, was followed by a modest cooling that coincides with the end of the African Humid Period and the onset of Neoglaciation on the African continent. Following this the record shows two distinct warming-cooling cycles during the late Holocene.

Introduction

Climate dynamics in the equatorial Atlantic Ocean are important not only for regional climate but also for global conditions as heat is exported to and from other regions via cross-equatorial surface and deep-water flow (Hastenrath 1980). Low-latitude insolation variations in the 23000-year precessional cycle affect the intensity of the African monsoon (e.g. Prell and Kutzbach 1987; de Menocal *et al*. 1993) and moderate the strength and direction of the trade winds that drive the equatorial currents, the cross-equatorial surface flow, and the northwestward movement of cold Benguela Current water (e.g. Imbrie *et al*. 1989; McIntyre *et al*. 1989; Schneider *et al*. 1995, 1996; Jansen *et al*. 1996). The cross-equatorial export of surface water is compensated by southward deep-water flow. A reduction in **North Atlantic Deep Water** (**NADW**) formation may slow the cross-equatorial heat export, thereby elevating the temperature of the equatorial South Atlantic (Mix *et al*. 1986; Crowley 1992).

Modelling of climate dynamics to reconstruct **Last Glacial Maximum** (**LGM**) and Holocene conditions in high- and mid-latitude regions requires climate reconstructions from low-latitude regions. Temperature change at low latitudes during the LGM has been debated and recent studies point to lower glacial temperatures than originally suggested (CLIMAP 1981). Several studies of vegetation, snowlines, lake levels, and high-altitude ice

suggest a tropical temperature change of as much as 5–6°C to 8–12°C (e.g. Rind and Peteet 1985 and references therein; Stute *et al*. 1995; Thompson *et al*. 1995; Farrera *et al*. 1999) and modelling results confirm the possibility of substantial glacial cooling at low latitudes (Webb *et al*. 1997). The various estimates for surface-water cooling in the tropical and subtropical Atlantic Ocean now range from <2° to 10°C (e.g. Sikes and Keigwin 1994; Guilderson *et al*. 1994, 2001; Kirst *et al*. 1999; Arz *et al*. 1999; Nürnberg *et al*. 2000; Niebler *et al*. 2003). It has become clear that the complex oceanic circulation and upwelling systems in this region affect the magnitude of glacial-to-interglacial **sea-surface temperature** (**SST**) change. Different methods for assessing past SST and limitations inherent to the methods may also cause some of the discrepancies among glacial–interglacial SST difference estimates (e.g. Sikes and Keigwin 1994; Nürnberg *et al*. 2000; Bard 2001).

Existing marine records covering the last termination in the east-equatorial South Atlantic have been correlated with the terrestrial climate record (e.g. Hughen *et al*. 1996; Lin *et al*. 1997; Arz *et al*. 1999; Rühlemann *et al*. 1999; de Menocal *et al*. 2000; Shi *et al*. 2000; Guilderson *et al*. 2001). Rapid and distinct climate changes significantly affected the environment in continental Africa during the Holocene (e.g. de Menocal *et al*. 1993; Pastouret *et al*. 1978; Street and Grove 1979; Street-Perrott and Harrison 1984; Gasse *et al*. 1990; Street-Perrott *et al*. 1990; Taylor 1990; Roberts *et al*. 1994; Johnson *et al*. 1996). As these millennial-scale changes likely affected the development in the ocean, comparisons with high-resolution records from, for example, the east-equatorial Atlantic, will further the understanding of the spatial and temporal pattern of climate change in the Atlantic/African region.

This paper presents new SST estimates from the east-equatorial South Atlantic Ocean and compares them with existing data for this area and the nearby African continent. **Ocean Drilling Program** (**ODP**) core 1077A-1H was examined with alkenone analysis to obtain high-resolution SST records for the LGM, Termination I, and the Holocene. Oxygen isotope analysis of multiple species of planktonic foraminifera was performed for hydrographic reconstruction. The core site is located near Africa in the east-equatorial South Atlantic where incursion of open-ocean water, upwelling, and discharge from the nearby Congo River affect the hydrography, affording a tie between ocean and land climate (Wefer *et al*. 1998).

Core site and material

ODP site 1077A was cored on the Congo Fan (5°11'S 10°26'E) at a water depth of 2380 m during ODP Leg 175 in August 1997 (Fig. 1). The core site is part of a transect across the narrow continental margin, and was chosen by ODP for studies of the varying influences of the Congo River, upwelling, and open-ocean advection, and for comparison of marine and terrestrial climate change.

The mouth of the Congo River is located southeast of the core site (Fig. 1). Fresh water from the Congo River reaches *c*. 800 km from the river mouth and fluvial sediments have formed the large deep-sea Congo Fan (van Bennekom and Berger 1984; Jansen *et al*. 1984). A submarine canyon incised into the Congo Fan captures most of the coarse-grained material from the river, leaving fine-grained material to be deposited on the continental shelf and shelf edge (Jansen *et al*. 1984; Wefer *et al*. 1998). Oceanic inflow of water from the **South Equatorial Countercurrent** (**SECC**) and the **Equatorial Undercurrent** (**EUC**) affects the core site. This water feeds into the southeastward flowing **Angola Current** (**AC**) that meets the **Benguela Coastal Current** (**BCC**) at about 15–17°S in the **Angola–Benguela Front** (**ABF**), well south of the core site. A permanent cyclonic gyre exists between the equatorial return flow, the Angola Current, and the **Benguela Oceanic Current** (**BOC**) (e.g. Peterson and Stramma 1991; Wacongne and Piton 1992; Stramma and Schott 1999).

There are several causes of upwelling in this area. The constant westward push of surface waters by trade winds leads to compensating upward movement of subsurface water causing the thermocline to penetrate the euphotic zone (e.g. Voituriez 1981; Ravelo and Fairbanks 1992; Wolff *et al*. 1999). Seasonally increased upwelling occurs in August–September along the equatorial divergence zone due to changes in position of the Inter-Tropical Convergence Zone (Voituriez 1981). Doming of subsurface waters centred on 10°S 9°E occurs at 20–150 m depth during January–March (van Bennekom and Berger 1984; Peterson and Stramma 1991; Stramma and Schott 1999) and coastal upwelling and estuarine circulation occur seasonally in the study area (e.g. van Bennekom and Berger 1984; Wacongne and Piton 1992).

Sea-surface temperatures in the Congo Fan area vary between 22.5°C in July–September (during trade-wind-induced upwelling) and 28°C in
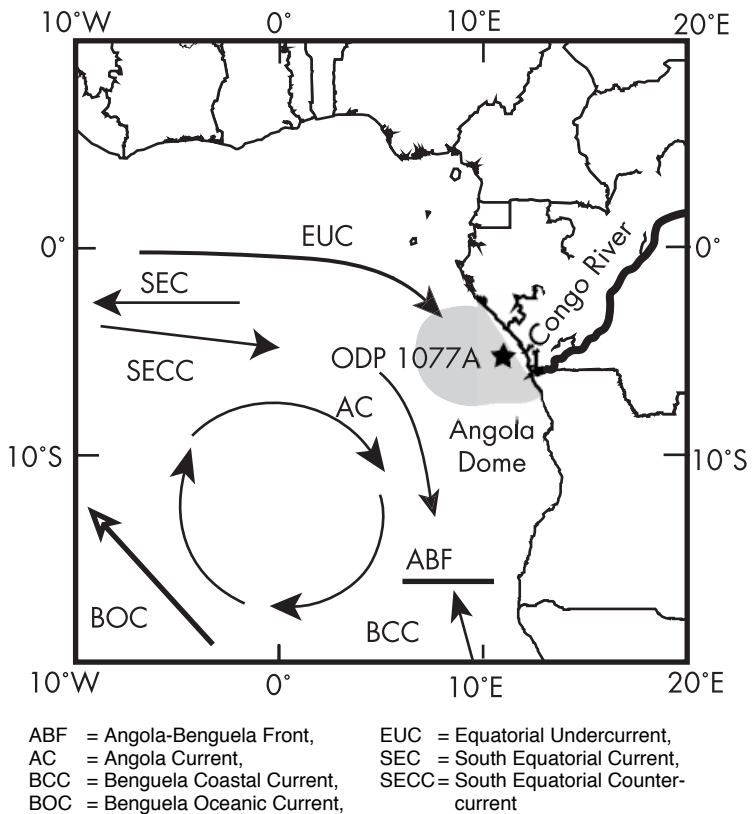
Fig. 1. The eastern tropical and subtropical South Atlantic, after van Bennekom and Berger (1984), Jansen *et al*. (1984), and Schneider *et al*. (1995). Arrows indicate currents; the grey area shows the approximate extent of the Congo River freshwater plume, and a star marks ODP site 1077A.

ABF = Angola-Benguela Front,
AC = Angola Current,
BCC = Benguela Coastal Current,
BOC = Benguela Oceanic Current,

EUC = Equatorial Undercurrent,
SEC = South Equatorial Current,
SECC = South Equatorial Counter-current

January–March (Fig. 2a) (Antonov *et al*. 1998). At 20 m depth, temperatures range between 19° and 24.5°C for the coldest and warmest seasons, respectively.

Despite the generally high surface-water salinity in the equatorial Atlantic, the salinity in the upper 20 m of the water column near the Congo River is clearly influenced by freshwater (Fig. 2b). The river discharge peaks between November and January (Fig. 2c) (van Bennekom and Berger 1984; Vörösmarty *et al*. 1996 1998), the effect reaching farthest horizontally and vertically during January–March (van Bennekom and Berger 1984; Boyer *et al*. 1998).

For the present study, ODP hole 1077A was examined (1077A-1H, 0–5.0 **metres below sea floor (mbsf)**). The material consists of a mixture of fine-grained terrestrial clay and biogenic pelagic material in one lithostratigraphic unit (Wefer *et al*. 1998). High input of terrestrial fine material has enhanced the sedimentation rate and provided conditions that limit breakdown of organic matter. The shipboard examination revealed some signs of bi-

oturbation and a disturbance from the coring process was reported for the uppermost 30 cm, but there was no evidence of turbidite redeposition. Shipboard analysis of total organic carbon (TOC) yielded values between 4.70 and 1.29 **weight per cent (wt%)** (average 2.3 wt%) and measurements of carbonate carbon concentrations using a Coulometrics 5011 carbonate carbon analyser yielded values of 0.8 to 13.2 wt%. Planktonic foraminifera were few and often fragmented, preventing faunal assemblages analysis. Pyrite was commonly found in the foraminiferal tests.

Methods

ODP core 1077A-1H was sampled at 5 cm intervals between 0 and 480 cm and stored under refrigerated conditions. Samples were weighed wet and split into subsamples. Wet material was sent to respective laboratories for alkenone analysis and radiocarbon dating. Oxygen isotope samples were dried, weighed, and soaked in deionized water and sodium-metaphosphate to disaggregate before wet-
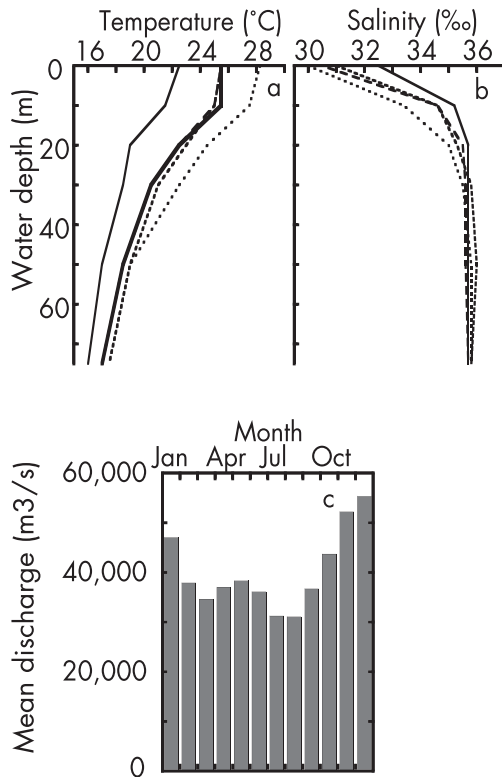
Fig. 2. Vertical profiles of modern (a) seasonal and annual mean temperature (in °C) and (b) seasonal mean salinity (in per mil) for the core location based on data from Antonov *et al*. (1998) and Boyer *et al*. (1998). Annual mean temperature (thick solid line), temperature and salinity for January–March (sparsely dotted line), April–June (densely dotted line), July–September (thin solid line), and October–December (dashed line). (c) Modern monthly mean discharge from the Congo River at Brazzaville (in m³/s) (Vörösmarty *et al*. 1996, 1998).

sieving at 63 μm. After drying overnight (maximum temperature 50°C), the isotope samples were dry-sieved at 150 μm.

Age control

Sediment samples from eleven depth levels in ODP core 1077A-1H were dated by accelerator mass spectrometry (AMS) ^{14}C dating at the Ångström Laboratory, Division of Ion Physics ^{14}C-laboratory at Uppsala University, Sweden. Sediment samples were treated with HCl and NaOH to remove calcium carbonate and to separate soluble and insoluble components. The soluble organic fractions were dated for this study. In addition to radiocarbon

dates, the carbon isotopic signal ($\delta^{13}C_{org}$) was measured and reported with respect to the **Pee Dee Belemnite (PDB)** standard.

The $\delta^{13}C_{org}$ measurement of a sample can indicate the predominance of either terrestrial or marine sources for the organic carbon in the sample. However, since $\delta^{13}C_{org}$ signals from marine and terrestrial material overlap to a certain extent, a clear distinction is difficult. We used the reservoir correction of 400 years (Bard 1988) for the raw ^{14}C AMS dates, assuming a dominantly marine surface-ocean source of carbon because of the high marine productivity in the area. The $\delta^{13}C_{org}$ measurements for the dated samples do not conflict with this assumption. The reservoir-corrected AMS dates were calibrated to calendar years using the CALIB program (Stuiver and Reimer 1993; Stuiver *et al*. 1998) for dates younger than 20,760 ^{14}C years BP and the Bard *et al*. (1993, 1997) equation for the oldest date. An age model was produced by linear interpolation between the calibrated dates (rounded to the nearest ten years) and extrapolation below and after the oldest and youngest dates, respectively. All ages are in calendar years before present (cal ka BP) unless otherwise stated.

Alkenone analysis

Analysis of alkenones (long-chained unsaturated methyl and ethyl ketones $C_{37:2}$, $C_{37:3}$, and $C_{37:4}$) produced by microalgae from the class *Prymnesiophyceae* has become a useful tool for assessing oceanic palaeotemperatures, because of the direct relationship between the degree of alkenone unsaturation and the water temperature at time of algal growth (e.g. Volkman *et al*. 1980, 1995; Brassell *et al*. 1986; Prahl and Wakeham 1987; Prahl *et al*. 1988; Müller *et al*. 1997). Several calibrations to SST have been proposed from studies of sediment traps, core tops and culture experiments, and though the overall relationship is similar, the degree of deviation varies (e.g. Prahl and Wakeham 1987; Prahl *et al*. 1988; Sikes *et al*. 1991; Sonzogni *et al*. 1997; Müller *et al*. 1998). The most commonly used calibration of $U^{k'}_{37}$ to SST is from Prahl and Wakeham (1987) and Prahl *et al*. (1988) ($U^{k'}_{37} = 0.034T + 0.039$) for the temperature range 8–25°C. In the eastern tropical South Atlantic, Müller *et al*. (1998) showed that the $U^{k'}_{37}$/SST relationship is most robust for the upper 10 m of the water column, suggesting that the coccoliths grow primarily in the mixed layer. Variations in salinity, degradation in the sediment, storage methods, carbonate dissolu-

tion, or nutrient stress do not appear to have a discernible effect on the $U^{k'}_{37}$/SST relationship and no bias of the temperature calibrations towards any particular season is evident (e.g. Prahl *et al*. 1989; Sikes *et al*. 1991; Madureira *et al*. 1995; Sonzogni *et al*. 1997; Müller *et al*. 1998).

Approximately 3–4 g of wet bulk sediment were separated for alkenone analysis. Samples were stored in plastic bags under refrigerated conditions and analyses were performed at Brown University, following procedures described in Herbert *et al*. (1998). For this study, the $U^{k'}_{37}$ index and the most commonly used $U^{k'}_{37}$/SST calibration from Prahl *et al*. (1988) were used. The analytical error was on the order of 0.2°C for most of the samples. We treat the alkenone SST record here as a measurement of average annual SST for the surface mixed layer.

The total abundance of C_{37} in sediment can be used as a rough estimate of marine productivity provided that degradation of the alkenones is relatively minor (Prahl *et al*. 1989; review in Grimalt *et al*. 2000). Weaver *et al*. (1999) reported a broad agreement between C_{37} abundance and coccolith abundance in North Atlantic sediment, suggesting that high productivity increases the C_{37} abundance. However, oxidative degradation and differences in environmental conditions, growth stages, species, and strains of species can affect the alkenone abundance (e.g. Prahl *et al*. 1989) and limit the use of C_{37} abundance for estimates of marine productivity. Because of problems with assessing the cause of abundance changes, data from this study are shown but no interpretation is offered.

Oxygen isotope analysis

Oxygen isotope analysis of planktonic foraminifera was performed on *Globigerinoides ruber* (p = pink variety and w = white variety) and *Neogloboquadrina dutertrei*. *Globigerinoides ruber* is a near-surface-dwelling species (e.g. Bé and Hamlin 1967; Bé and Tolderlund 1971; Fairbanks *et al*. 1982) with highest abundances in the surface mixed layer or at the top of the thermocline (upper *c*. 30 m) in the east-equatorial Atlantic (Ravelo and Fairbanks 1992). The total SST range for *G. ruber* is 14–30°C (Bé and Hamlin 1967) and abundances peak between 21° and 29°C (Bé and Tolderlund 1971). *Globigerinoides ruber* (w) is characteristic of warm, high-salinity Equatorial Undercurrent water, whereas *G. ruber* (p) is abundant in the warm, low-salinity outflow from the Congo River (Ufkes *et al*. 1998). *Neogloboquadrina dutertrei* is

a subsurface water species (e.g. Bé and Tolderlund 1971; Ravelo and Fairbanks 1992) and occurs in highest abundances at the chlorophyll maximum (Fairbanks and Wiebe 1980; Fairbanks *et al*. 1982). In the east-equatorial Atlantic, the chlorophyll maximum coincides with the seasonal thermocline (Ravelo and Fairbanks 1992), which currently is located at 10–50 m depth at ODP site 1077A (Fig. 2a) based on temperature data from Antonov *et al*. (1998). *Neogloboquadrina dutertrei* appears to tolerate relatively low salinities and often reaches peak abundances during upwelling conditions (Guptha *et al*. 1997, and references therein). Previous studies show that *N. dutertrei* calcifies in equilibrium with the predicted $\delta^{18}O$ of ambient water (e.g. Fairbanks *et al*. 1980; Ravelo and Fairbanks 1992). The *G. ruber* (p) $\delta^{18}O$ values are often offset from predicted $\delta^{18}O$ by around –0.5‰, whereas the *G. ruber* (w) $\delta^{18}O$ values were offset in some studies but not in others (e.g. Williams *et al*. 1979; Fairbanks *et al*. 1980; Ravelo and Fairbanks 1992).

Foraminifera were analysed in the size fraction 250–500 μm for *N. dutertrei* and 150–500 μm (mostly 250–500 μm) for *G. ruber*. Between one and three specimens of *N. dutertrei* and three to six specimens of *G. ruber* (pink and white varieties measured separately) were used for each individual measurement. Because of low foraminiferal abundance, measurements could only be replicated at a few depths and some levels had too few foraminifera for isotope analysis. The isotope signal was similar to other studies for these species, geographic area, and time frame (Ravelo and Fairbanks 1992; Schneider *et al*. 1995).

Analyses were performed at the Stable Isotope Laboratory, University of Maine, in a VG PRISM II mass spectrometer with an automated carousel. Samples were reacted in phosphoric acid at 90°C for ten minutes to produce CO_2, which was purified cryogenically. Laboratory precision was <0.1‰ for $\delta^{18}O$ and <0.06‰ for $\delta^{13}C$. Values are reported with respect to PDB. These values were corrected for the daily average deviation of six NBS-20 standards analysed during each run from the published values of NBS-20, –4.14‰ for $\delta^{18}O$ and –1.06‰ for $\delta^{13}C$, respectively (Coplen *et al*. 1983).

Results
Age control and sedimentation

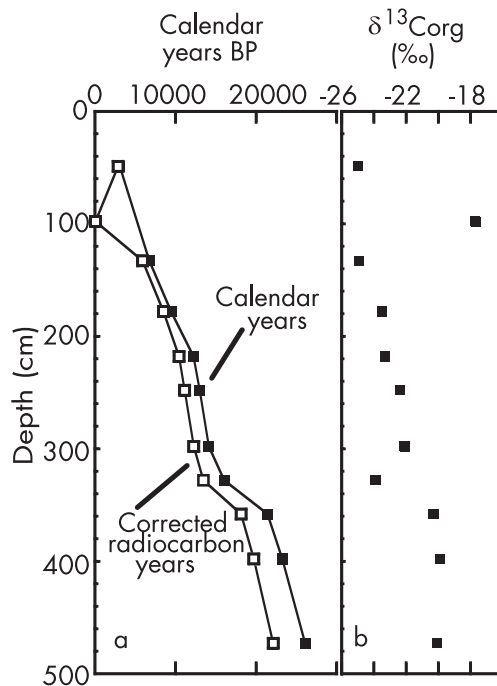All radiocarbon dates except the date at 98 cm (which was discarded) are in chronological order

Fig. 3. (a) Age versus depth in ODP 1077A-1H: corrected radio-carbon dates (open squares) and calendar years BP (filled squares); (b) $\delta^{13}C_{org}$ of the dated samples (in per mil) (PDB) plotted against depth.

3a). Sedimentation rates vary between 6 cm/1000 years and 38–45 cm/1000 years. The lowest sedimentation rate occurs between 358 and 328 cm (*c.* 21.5 to 16 cal ka BP) and the highest between 298 and 218 cm (*c.* 14 to 12 cal ka BP).

The measurements of $\delta^{13}C_{org}$ in the sediment yielded maximum values of around –20‰ between 473 and 358 cm, followed by a gradual decrease to more negative values (–25‰) in the upper part of the core (Fig. 3b). The $\delta^{13}C_{org}$ values may reflect changes in sources of organic carbon. Grossly generalized, organic carbon from marine phytoplankton (–12‰ to –23‰) and terrestrial and aquatic C_4 and CAM plants that reflect dry conditions (–6‰ to –19‰) should yield a relatively high $\delta^{13}C_{org}$ value. Organic carbon from terrestrial C_3 plants reflecting relatively humid conditions afford relatively low $\delta^{13}C_{org}$ values (–24‰ to –34‰) (overview in Faure 1986). Our $\delta^{13}C_{org}$ values of –19.9 to –25.0‰ are tentative support of a predominantly marine source for the dated organic carbon.

Alkenones

The total difference in glacial-to-interglacial mean annual temperature recorded by alkenones in ODP 1077A-1H is *c.* 4.3°C. The lowest temperatures vary between 19.5° and 20.5°C at 470 to 415 cm followed by a gradual but fluctuating increase to maximum values between *c.* 23.5° and 24°C, recorded from 185 to 110 cm. Above 110 cm, temperatures oscillate between *c.* 22.5° and 24°C at lower frequencies than previously, forming longer periods of relatively cool or warm temperatures

(Table 1). This suggests that the material has not been severely disturbed by bioturbation or reworking, other than the upper 30 cm that was disturbed during coring. The analytical errors from the AMS dating range between 70 and 380 ^{14}C years (Fig.

Table 1. Radiocarbon dates, δ^{13}C values, and calendar years.

Ångström Lab. No.	Depth (cm)	Raw ^{14}C age (years BP)	Corrected ^{14}C age (years BP)	Calendar age[a] (years BP)	Calendar year error range[b] (1 sigma)	^{13}C (‰ *vs.* PDB)
Ua-15979	49	3265 ± 95	2865	3090	2944–3227	–25.0
Ua-14234	98	475 ± 70	–	–	–	–17.7
Ua-15980	133	6325 ± 100	5925	6760	6666–6894	–24.9
Ua-15168	178	8895 ± 110	8495	9460	9099–9773	–23.5
Ua-15981	218	10835 ± 125	10435	12230	11741–12621	–23.3
Ua-15169	248	11530 ± 250	11130	13010	12873–13180	–22.4
Ua-14235	298	12675 ± 300	12275	14130	13839 – 15172	–22.1
Ua-15170	328	13850 ± 380	13450	16040	15557–16536	–23.9
Ua-15171	358	18535 ± 195	18135	21430	21040–21837	–20.3
Ua-14739	398	20120 ± 235	19720	23260	22808 – 23765	–19.9
Ua-14740	473	22520 ± 315	22120	26080	–	–20.1

[a] Radiocarbon dates were calibrated to calendar years using the CALIB program (Stuiver and Reimer 1993; Stuiver *et al.* 1998) and the U/Th calibration (Bard *et al.* 1993, 1997) after correction for the reservoir effect by 400 years (Bard 1988). Calendar years are rounded to nearest 10 years.

[b] Error range (1-sigma) for calendar years estimated by using the CALIB program (Stuiver and Reimer 1993; Stuiver *et al.* 1998).
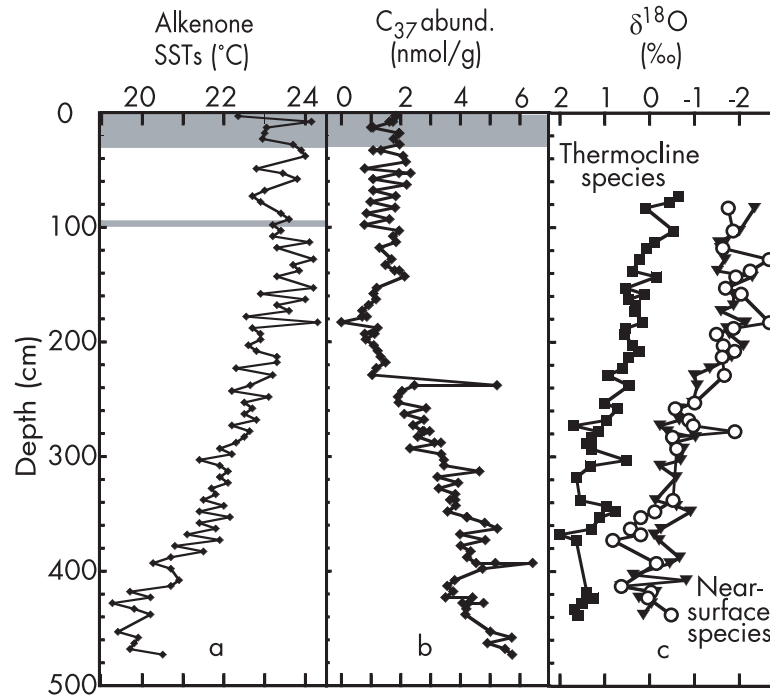
Fig. 4. (a) Alkenone-derived SSTs (in °C), (b) abundance of C_{37} (in nmol/g), and (c) $\delta^{18}O$ records from *N. dutertrei* (squares), *G. ruber* (p) (triangles), and *G. ruber* (w) (open circles) (in per mil) (PDB) against depth in core. Shaded area reflects the disturbed upper 30 cm of the core and the level of the reversed date.

(Fig. 4a). Between 370 and 250 cm, the magnitude of the oscillations is smaller than the rest of the record. The uppermost 30 cm (six depth levels) are not interpreted here as this section was disturbed during coring (shaded area). On the whole, the abundance of C_{37} in ODP 1077A-1H decreases gradually from around 6 nmol/g at the bottom of the core to *c.* 0–1 nmol/g between 230 and 150 cm (Fig. 4b). Values increase slightly to *c.* 1–2 nmol/g above 150 cm.

Oxygen isotopes

Both near-surface and thermocline oxygen isotope values change gradually from high $\delta^{18}O$ values at the core bottom to low values at the top (Fig. 4c). As expected, the $\delta^{18}O$ values for the near-surface-dwelling species *G. ruber* (p) and *G. ruber* (w) are lower than the thermocline-dwelling *N. dutertrei* values due to high temperatures and low salinity of surface waters in this area. The $\delta^{18}O$ values range between 0.9‰ and –2.8‰ for *G. ruber* and between 2‰ and –0.9‰ for *N. dutertrei*, consistent with previous studies from the equatorial Atlantic (Ravelo and Fairbanks 1992; Schneider *et al.* 1995). *Globigerinoides ruber* (p) and *G. ruber* (w) generally recorded similar $\delta^{18}O$ values

with a few exceptions, e.g. between 380 and 350 cm, and there is no consistent offset between the records.

Discussion

Our high-resolution east-equatorial SST record features a three-stepped glacial–interglacial warming of *c.* 4.3°C (Fig. 5a and b). High-frequency fluctuations superimposed on the general trends occur throughout the alkenone record. The lowest SSTs (*c.* 19.5–20.5°C) were recorded between 26 and 24 cal ka BP and were replaced by the first warming of around 2°C beginning at 24 cal ka BP. Relatively stable temperatures of 21.5–22°C followed between 21.5 and 14.5 cal ka BP, although lower temporal resolution due to low sedimentation rates at this time may artificially smooth the record (Fig. 5c). It is important to note that the warming beginning at 24 cal ka BP occurred during the LGM (24–23 to 19–18 cal ka BP, as defined in Mix *et al.* 2001) when Northern Hemisphere ice sheets were still at their maximum position and size (Peltier 1994).

The second warming began abruptly around 14.5 cal ka BP and ended with the onset of a comparatively cool period between *c.* 11.5 and 10 cal ka
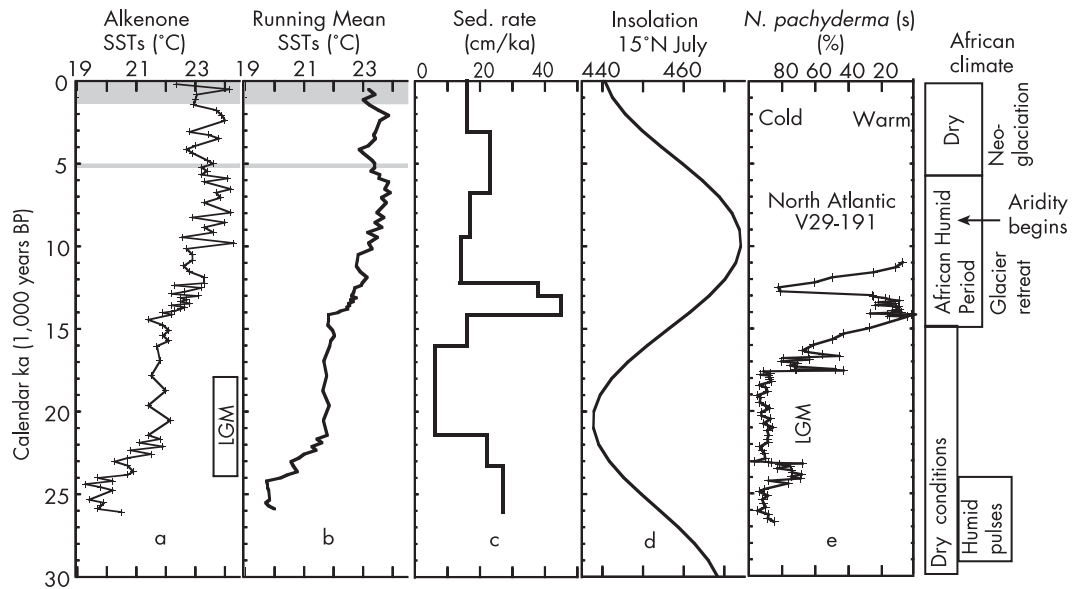
Fig. 5. (a) Alkenone-derived SSTs (in °C) against thousands of calendar years BP. The time frame for the Last Glacial Maximum is marked by LGM. (b) Three-point running mean of alkenone-derived SSTs (in °C) against thousands of calendar years BP. (c) sedimentation rate. (d) July insolation at 15°N (from Berger 1992), against calendar age. (e) Relative abundance of *N. pachyderma* (s) (in %) in V29-191 from the North Atlantic (Lagerklint and Wright *1*999; Lagerklint 2001), against calendar age. Conditions on the African continent are marked to the right in the figure.

BP. The third warming trend followed between 10 and 6 cal ka BP in the form of large-magnitude temperature fluctuations with four SSTs peaks around 24°C, indicating highly variable ocean conditions. The running mean SST record defines the time for the maximum warming as 7 to 6 cal ka BP (Fig. 5b). From 6 to 4 cal ka BP, temperatures decrease to *c.* 22.5°C and the Late Holocene was subsequently characterized by two very distinct warming-cooling sequences with temperatures fluctuating between 22.5° and 24°C.

The glacial–interglacial transition is also clear in the oxygen isotope records, both for thermocline and for near-surface waters (Fig. 4), although the δ¹⁸O and alkenone records do not show the same outline. However, Mollenhauer *et al.* (2003) showed that radiocarbon dating of alkenones and foraminifera from the same core depth may give substantial differences in measured ages, potentially caused by the alkenones being resuspended by currents in tidal and upwelling areas. Such potential age differences could explain some of the discrepancy between our alkenone and δ¹⁸O records. Due to the sparseness of foraminifera in our samples, we could not separately determine the radiocarbon age of the foraminiferal carbonate. There-

fore, we have chosen not to plot the isotope records in the existing age model based on radiocarbon dates of organic carbon.

Magnitude of Glacial-Holocene temperature difference

The magnitude of glacial–interglacial SST change estimated in different studies is highly dependent on local oceanography at each study site, the methods used, and whether the Holocene comparison is made to the LGM or the time of lowest temperatures during the glacial (which in some records precedes the actual LGM). Glacial–interglacial SST difference estimates from tropical and subtropical Atlantic records range from 1.8 to 10°C (Sikes and Keigwin 1994; Guilderson *et al.* 1994, 2001; Schneider *et al.* 1995; Rühlemann *et al.* 1999; Kirst *et al.* 1999; Nürnberg *et al.* 2000; Niebler *et al.* 2003). It is clear that upwelling and current dynamics create varying results, and study sites in areas with strong coastal upwelling or in the path of a cold current will experience other changes than mere atmospheric cooling or warming of surface waters. Although our study site off the Congo is in an area of general upwelling, it is not located in the imme-

diate paths of the Benguela Coastal Current or the Benguela Oceanic Current. The temperature difference of *c*. 4.3°C in the ODP 1077A-1H alkenone record is consistent with that measured in another core from the Congo Fan, GeoB 1008-3 (Schneider *et al.* 1995).

Another important factor for determining the magnitude of glacial–interglacial change is whether to use the defined time period of LGM or the coldest conditions recorded during the whole glacial period for comparison with the Holocene temperature optimum. Minimum temperatures before the LGM time period have been noted in records from the tropical South Atlantic and near-shore core sites off Namibia (Schneider *et al.* 1995, 1996; Kirst *et al.* 1999). These records feature the coldest temperatures between 50 and 40 cal ka BP. Although our data do not range as far back in time, it is still clear that the coldest conditions in our alkenone record also occur before the LGM. Our glacial–interglacial SST difference of *c*. 4.3°C is based on comparing minimum and maximum temperatures of the record. If instead the comparison is made using LGM temperatures, the estimate would be roughly 1.5°C lower since our alkenone record shows a warming during this time.

A comparison with our near-surface $\delta^{18}O$ records reveals discrepancies between the two proxies with respect to magnitude of change (Fig. 4). After correcting our isotope records for global ice-volume change, the total differences between glacial–interglacial $\delta^{18}O$ values from *G. ruber* (p) and *G. ruber* (w) range between 1.68 and 2.69‰, depending on the correction value: 1.3‰ (Fairbanks 1989) or 0.8‰ (Schrag *et al.* 1996). These differences in $\delta^{18}O$ correspond to temperature changes of *c*. 7.5° to 12°C, using the relationship of *c*. 0.22‰ per 1°C change (Epstein *et al.* 1953). The discrepancy between the alkenone temperature difference of *c*. 4.3°C and that calculated from foraminiferal $\delta^{18}O$ may have been caused by changes in physical parameters that affect foraminiferal shell growth but not alkenone production. This is consistent with salinity variations, which would change the $\delta^{18}O$ of the foraminiferal shell but have not been shown to affect alkenone values (Sonzogni *et al.* 1997). Variations in freshwater outflow from the Congo River are likely the primary reasons for the discrepancies between our alkenone and isotope records. As our isotope data are based on analyses of few foraminifera per sample and few replicating measurements, we refrain from further interpretation of the records. However, it should be
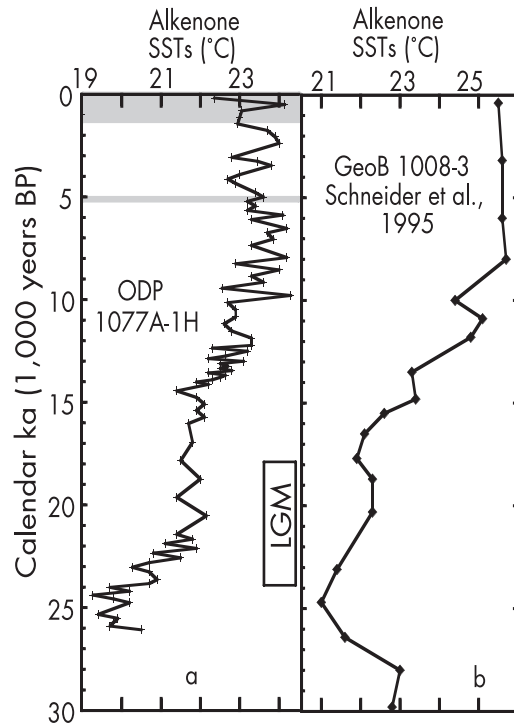

Fig. 6. (a) Alkenone-derived SSTs (in °C) as in Fig. 5 for ODP 1077A-1H and (b) alkenone-derived SSTs (in °C) from the nearby core GeoB 1008-3 (from Schneider *et al.* 1995).

noted that our analysis of species with different depth habitats clearly revealed the expected gradient between thermocline and near-surface waters, as previously documented in other studies (e.g. Williams and Healy-Williams 1980; Steens *et al.* 1992).

Onset of warming during the LGM

An interesting feature of the ODP 1077A-1H alkenone record is the 2°C warming that began around 24 cal ka BP at the height of Northern Hemisphere glaciation, illustrated by a relative abundance record of *Neogloboquadrina pachyderma* (sinistral) from North Atlantic core V29-191 located on the Feni Drift west of the British Isles (Lagerklint and Wright 1999; Lagerklint 2001) (Fig. 5e). This warming and the subsequent stable temperatures between 21.5 and 14.5 cal ka BP were also recorded in the GeoB 1008-3 record (Schneider *et al.* 1995) (Fig. 6). Records from more southerly or central locations and different oceanographic set-

tings than those of the Congo Fan show evidence of warming at the end of or after the LGM (e.g. Kirst *et al.* 1999; Kim *et al.* 2002).

Previous studies have reported that intensified and more zonal trade winds deflected and strengthened the flow of the Benguela Oceanic Current to the northwest during the end of the LGM (Jansen *et al.* 1984, 1996; Jansen and van Iperen 1991). The warm conditions recorded during the LGM in the eastern South Atlantic were explained as the result of increased equatorial return flow of warm waters (as countercurrents and undercurrents), which caused temperatures to rise in this region (Schneider *et al.* 1995, 1996). At the same time, reduced NADW formation would have caused a reduction of the cross-equatorial export of warm water, which may also have resulted in a warming of the tropical Atlantic (Crowley 1992).

On the African continent, conditions were generally dry between *c.* 34.5 and 14 cal ka BP (30 and 12 ^{14}C ka BP) (Nicholson and Flohn 1980; Gasse *et al.* 1990; Johnson *et al.* 1996). These dry conditions were interspersed with several humid climatic fluctuations between 29 and 24 cal ka BP (25 and 20 ^{14}C ka BP) (e.g. Pastouret *et al.* 1978; Street and Grove 1979; Gasse *et al.* 1989; Lézine and Vergnaud-Grazzini 1993).

Deglacial and Holocene temperature changes

The increase of SSTs in two steps between 14.5 and 6 cal ka BP in ODP 1077A-1H coincides with maximum low-latitude summer insolation in the Northern Hemisphere (Fig. 5d). Weakened and more meridional trade winds resulted in overall reduced upwelling and high SSTs in the eastern Angola Basin (Schneider *et al.* 1995; McIntyre *et al.* 1989; Molfino and McIntyre 1990). The strong African low pressure caused increased monsoonal activity and rainfall in northern and central Africa (de Menocal *et al.* 2000). Major changes in palaeohydrology took place in Africa during the Pleistocene–Holocene transition. A relatively long humid period, known as the African Humid Period, between 14.8 and 5.5 cal ka BP has been registered in many records from tropical and subtropical Africa (Street and Grove 1979; Street-Perrott and Harrison 1984; Street-Perrott *et al.* 1990; Taylor 1990; Gasse *et al.* 1990; Roberts *et al.* 1994; Johnson *et al.* 1996; de Menocal *et al.* 2000; Gasse 2001). In the Uganda highlands, both temperature and precipitation increased after 17 cal ka BP (14 ^{14}C ka BP) (Taylor 1990). Tropical glaciers commenced their retreat at this time, indicating that temperatures were rising (Rosqvist 1990; Shanahan and Zreda 2000). Tropical Africa (Congo, Ghana, lowland Cameroon), including East Africa, experienced a pronounced rise in precipitation and forest expansion in the early Holocene (Nicholson and Flohn 1980; Talbot and Johannessen 1992; Lézine and Vergnaud-Grazzini 1993; Beuning *et al.* 1997; Vincens *et al.* 1998; Peyron *et al.* 2000; Thompson *et al.* 2002; Stager *et al.* 2003). Barker *et al.* (2001) infer a wet phase between 11.1 and 8.6 cal ka BP in an isotopic record from Mt Kenya in East Africa, which they suggest might be linked to positive SST anomalies over the tropical South Atlantic and Indian Ocean. Increased rainfall and high SSTs were recorded by Lézine and Vergnaud-Grazzini (1993) in the Gulf of Guinea around 14.5 cal ka BP (12.5 ^{14}C ka BP), contemporary with increases in the Niger and Senegal River discharges (Pastouret *et al.* 1978; Sarnthein *et al.* 1982).

The mid-to-late Holocene reduction in temperatures recorded in our alkenone record between *c.* 6 and 4 cal ka BP coincides with changing conditions on the African continent (Fig. 5) and with evidence of cooling in the east-equatorial South Atlantic (Mix *et al.* 1986; Kim *et al.* 2002). Major shifts characterized mid-Holocene climate in subtropical and tropical Africa when aridity and rainfall seasonality increased (e.g. Maley 1997; de Menocal *et al.* 2000). The African Humid Period terminated abruptly at 5.5 cal ka BP when conditions became much drier in Sahara (de Menocal *et al.* 2000). At this time glacier advances mark the onset of Neoglaciation on Mount Kenya, East Africa (Karlén and Rosqvist 1988; Karlén *et al.* 1999). A dry episode occurred between 4 and 3 cal ka BP in lowland African rain forests north of the Gulf of Guinea, associated with low SST and upwelling (Maley 1997).

Conclusions

Our alkenone record indicates a glacial–interglacial SST temperature change of *c.* 4.3°C in the east-equatorial South Atlantic. The alkenone SST record shows that the coldest conditions occurred prior to the LGM, followed by an initial warming starting at 24 cal ka BP while Northern Hemisphere ice sheets were still at their maximum extent. These findings are in agreement with previous results from the area. This first warming occurred when Africa experienced generally dry conditions. The

renewed warming recorded at 14.5 cal ka BP correlates temporally with the onset of the African Humid Period. A temperature decrease beginning at 6 cal ka BP and subsequent periods of alternating warm and cold conditions coincide with the end of the African Humid Period and the onset of Neoglaciation in Africa.

Our data provide new high-resolution SST estimates for the Glacial–Holocene transition in the east-equatorial South Atlantic. The alkenone record confirms previous findings of climate development off the coast of Congo differing from those seen in areas to the south and west during the LGM. The study emphasizes that comparisons between studies must take into account local oceanographic differences, selection of criteria when calculating glacial–interglacial differences (maximum and minimum temperatures or certain defined time periods), and differences in recording potential between analytical methods.

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