The precession phase of the boreal summer monsoon as viewed from the eastern Mediterranean (ODP Site 968)

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Article info
Article history:
Received 31 July 2009
Received in revised form 16 March 2010
Accepted 17 March 2010

Abstract
The astronomical timescale of the Eastern Mediterranean Plio–Pleistocene builds on tuning of sapropel layers to Northern Hemisphere summer insolation maxima. A 3000-year precession lag has become instrumental in the tuning procedure as radiocarbon dating revealed that the midpoint of the youngest sapropel, S1, in the early Holocene occurred approximately 3000 years after the insolation maximum. The origin of the time lag remains elusive, however, because sapropels are generally linked to maximum African monsoon intensities and transient climate modeling results indicate an in-phase behavior of the African monsoon relative to precession forcing. Here we present new high-resolution records of bulk sediment geochemistry and benthic foraminiferal oxygen isotopes from ODP Site 968 in the Eastern Mediterranean. We show that the 3000-year precession time lag of the sapropel midpoints is consistent with (1) the global marine isotope chronology, (2) maximum (monsoonal) precipitation conditions in the Mediterranean region and China derived from radiometrically dated speleothem records, and (3) maximum atmospheric methane concentrations in Antarctica ice cores. We show that the time lag relates to the occurrence of precession-paced North Atlantic cold events, which systematically delayed the onset of strong boreal summer monsoon intensity. Our findings may also explain a non-stationary behavior of the African monsoon over the past 3 million years due to more frequent and intensive cold events in the Late Pleistocene.

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1. Introduction

The largest amplitude swings in global climate during the Quaternary are controlled by cyclic variations in the Earth's orbital parameters, which govern changes in the intensity and distribution of the insolation on the Earth's surface. It is well established that these cycles are the pacemaker of the ice ages (Hays et al., 1976; Imbrie and Imbrie, 1980), with 41kyr variations in the tilt of the Earth's axis (obliquity) playing a dominant role (Huybers and Wunsch, 2005). On the other hand, the intensity of low latitude monsoonal systems responds primarily to the ~23-kyr precession cycle of the Earth's rotational axis (Kutzbach and Guetter, 1986). In theory, the Northern Hemisphere or boreal monsoon should be intensified during precession minima, when Northern Hemisphere summer occurs in perihelion. The induced maximum in peak summer insolation amplifies the atmospheric pressure difference between land and sea intensifying the monsoonal circulation. However, the phase relationship between maximum monsoon intensity and precession minima is still a matter of debate, because data (Clemens and Prell, 2003; Y. Wang et al., 2008) and climate modeling results (Tuenter et al., 2005; Kutzbach et al., 2008) arrived at different estimates for the precession phase of the monsoon. Here, we study the precession phase of the boreal summer monsoon using an example which is considered to be one of the most characteristic sedimentary expressions of changes in African monsoon intensity: the cyclic occurrence of dark, organic-rich sapropel layers in the Eastern Mediterranean Sea (EMS) (Rossignol-Strick, 1983).

According to theory, sapropel layers are formed as consequence of reduced deepwater ventilation, due to diminished excess evaporation from the Mediterranean basin (Rohling, 1994; Emeis et al., 2003; Rohling et al., 2009). The preservation of organic material during sapropel formation is potentially further enhanced by increased primary productivity, the formation of a deep chlorophyll maximum induced by pycnocline shoaling and increased carbon flux due to the formation of diatom mats (Rohling, 1989; Sancetta, 1994; De Lange et al., 2008). Principally, thermohaline circulation and deepwater of the EMS are driven by the formation of saline Levantine-Intermediate-Water during (late) summer and deep-water formation in northern sectors of the basin during winter,
linked to orographically channeled cold continental air outbursts (Casford et al., 2003; Rohling et al., 2009). The African summer monsoon has a remote control on the EMS deepwater ventilation through the discharge of the river Nile. Although the construction of the Aswan High Dam diminishes its influence today, variability in the monsoon-fed discharge is thought to have had a profound effect on the salt-driven part of the deepwater ventilation in the past. Periods of strong African summer monsoon intensity and associated increased river input runoff from the North African continent result in density stratification and breakdown of deepwater formation in the EMS and consequently in enhanced preservation of marine organic matter in oxygen depleted conditions (Rohling, 1994; Rohling et al., 2009). Evidence for such an enhanced freshwater influx into the Eastern Mediterranean is derived from oxygen isotope anomalies in surface dwelling planktic foraminifera (Ganssen and Troelstra, 1987; Rohling and De Rijk, 1999; Rohling et al., 2004).

The correlation of Mediterranean sapropels to summer insolation changes forms the basis of the astronomically tuned timescale for the late Neogene (Rossignol-Strick, 1983; Hilgen et al., 1993; Lourens et al., 1996; Lourens, 2004). Accordingly, maxima in the 21st June insolation curve at 65° N (precession minima) were used as age calibration points for the midpoints of sapropels. A time lag of 3000 years was applied in the tuning of the sapropel record to the maxima insolation. The application of the 3000-year precession lag forms an essential but not well-constrained assumption in the tuning procedure of sapropels. The time lag is derived from radiocarbon ages of the midpoint of the youngest sapropel S1, which occurred at ~8.5 ka, whereas the insolation maximum occurred at 11.5 ka (Lourens et al., 1996). A basin-wide synchronous formation of sapropel S1 has been recently confirmed by detailed radiocarbon measurements on sediment cores throughout the Eastern Mediterranean (De Lange et al., 2008). This study gives the age range of S1 from 10.8 to 6.1 ka cal. BP, thereby confirming the 3000-year time lag of the midpoint of S1. In contrast, transient climate modeling experiments find an in-phase response of the African monsoon with precession minima (Tuenter et al., 2005).

Here we present new high-resolution bulk sediment elemental composition and benthic oxygen isotope records from Eastern Mediterranean ODP Site 968. We compare our proxy records with radiometrically dated speleothem records (Bar-Matthews et al., 2003; Y. Wang et al., 2008), the record of atmospheric methane in Antarctic ice cores (Spahni et al., 2005) and the modeled Late Pleistocene river runoff from North Africa in a transient climate modeling experiment (Tuenter et al., 2005; Ziegler et al., in review). Moreover, we compare the benthic isotope record with the global benthic isotope stack LR04 (Lisiecki and Raymo, 2005) and Antarctic temperatures (Kawamura et al., 2007). Eventually, we introduce a conceptual model, which explains a systematic precession lag of the monsoon response due to the occurrence of precession-paced North Atlantic cold events.

2. Material and methods

2.1. Material

During ODP Leg 160 (April 1995), three holes were drilled at ODP Site 968 to the South of Cyprus at a water depth of 1960 m (Fig. 1). The studied sediments are fine-grained, nanofossil-rich, and contain distinct sapropels, tephra and silt layers. A corrected depth composite record for Hole A and Hole B has been constructed following the approach of Lourens (2004), using color records of Hole A and B for the construction of the composite. Intervals, which are being used in the composite record, are corrected by a stretching-factor to account for differences in sedimentation rates and coring operations between holes, which would otherwise lead to artificial stretching and squeezing in the composite.

2.2. Analytical methods

Bulk concentrations of major elements have been determined by X-Ray Fluorescence (XRF). A few (3–5) grams from each sample were ground in a ceramic pestle and mortar, which was cleaned with alcohol between samples. The ground samples were placed in a Leco TGA (Thermo-Gravimetric Analysis) to remove all residual moisture and oxidize organic matter and carbonates. This removal was achieved by heating the samples to five temperatures (105, 450, 550, 800 and 1000 °C respectively) until the weight of the sample stabilized. Of the residue, 600 mg was mixed with 6 g flux (a mixture consisting of 66% lithium tetraborate, Li₂B₄O₇ and 34% lithium metaborate, LiBO₂) and placed in platinum gold crucibles to which 0.500 ml of a 30% lithium iodide solution was added to prevent sticking of the molten sample to the crucibles. Subsequently the crucibles were placed in a Herzog Hag S automated
fusion machine and fused to pearls. The pearls were measured in an ARL9400 X-ray fluorescence spectrometer twice, once for the major constituents and a second time on a more sensitive scale for minor constituents. Analytical precision, as checked by parallel analysis of international reference material and in-house standards, is better than 1% for Si, Ti, Al, Fe, and better than 5% for Ba and Zr.

In order to generate a benthic foraminiferal oxygen isotope record, we had to rely on different species, as no single species is present throughout the whole record. Specimens were hand-picked from size fraction >212 μm and analyzed with an automated carbonate reaction device (Kiel III) coupled to a Thermo-Finnigan MAT253. Each sample reacted with 103% phosphoric acid (H₃PO₄) for 7 min at 70 °C. Calibration to the international carbonate standard NBS-19 and in-house standard NAXOS revealed an analytical precision better than 0.03‰ and 0.1‰ for δ¹³C and δ¹⁸O, respectively. Following species have been used: Gyroidina altiformis, Gyroidina neosoldanii, Cibicides kullenbergi, Hoeglundina elegans, Cibicides ungerianus, Melonis pompiliana. The oxygen isotope data of H. elegans have been corrected for its enrichment (0.78‰) relative to equilibrium calcite (Grossman, 1984).

The color reflectance records were scanned with a hand-held Minolta CM 2002 spectrophotometer at 2 cm resolution on board the JOIDES Resolution just after the cores were opened (Emeis et al., 1996).

### 3. Results

#### 3.1. Bulk sediment geochemistry

Sapropels S1 to S10 are clearly identifiable in the color record (i.e., 540 nm reflectance %). Using the sapropel chronology of Lourens (2004), this implies that the record covers approximately the last 350,000 years (Fig. 2). Weak insolation minima can prevent the formation of true sapropelic layers, but moreover post-depositional oxidation can remove previously formed sapropels (Thomson et al., 1995; van Santvoort et al., 1997; Langereis et al., 1997). To overcome diageneric effects, bulk elemental geochemistry has been successfully applied in order to identify periods of increased African Monsoon intensity and runoff from North Africa (Calvert and Fontugne, 2001). Fig. 2 shows different bulk sediment elemental ratios, which show in general a good correlation with the color record. In Calvert and Fontugne (2001) a detailed discussion can be found on the origin of variations in the geochemical composition, which characterize the cyclic changes from marls to sapropels. To summarize the results from Site 968: Fe/Al and Ba/Al are higher, whereas Si/Al, Ti/Al, K/Al and Zr/Al are lower in the sapropels. High Ba/Al ratios are very characteristic for sapropel layers and are related to increased primary productivity in the surface layers (Thomson et al., 1995; van Santvoort et al., 1997; Calvert and Fontugne, 2001; De Lange et al., 2008). Enrichments in Fe/Al
ratios within sapropels are due to the formation of iron sulphide minerals, mostly pyrite, under sulphate-reducing conditions and the degradation of organic matter (Thomson et al., 1995; Wehausen and Brumsack, 2000; Calvert and Fontugne, 2001). The low Ti/Al ratios are related to an increased fluvial supply of Al-rich elements and a decreased input of Ti-rich eolian material (Wehausen and Brumsack, 2000; Lourens et al., 2001). Low Zr/Al and Si/Al have been interpreted as indicators of decreased wind speeds and eolian transport (Calvert and Fontugne, 2001). The individual elemental ratios are to a varying degree influenced by additional factors, e.g. glacial–interglacial variability, clay mineral composition.

In particular the Ti/Al ratio has been successfully applied as African monsoon proxy, as it appeared to show a more linear relationship with the insolation forcing when compared to other proxies (Lourens et al., 2001). Variations in organic carbon content, color, Ba/Al and redox-sensitive elements, are sensitive to oceanic processes and therefore may reflect a higher degree of non-linear response to the monsoonal forcing. One aim of this study was to use the Ti/Al to investigate the precession phase of the monsoon, because of its more direct relation with atmospheric and fluvial processes. The more gradual nature of the Ti/Al variations is clearly visible during marine isotope stage 5, which encompasses sapropels S3, S4 and S5. However, the Ti/Al of Site 968 appears to be also sensitive to secondary influences within the studied interval. Firstly, volcanic ash layers are enriched in Al and are represented by extreme and sharp minima in the Ti/Al (Fig. 2). Moreover, during the Holocene, marine isotope stage (MIS) 4 and early MIS 6, the record shows deviations from the regular sapropel pattern. Sapropel S6 for instance is well developed in the color and Ba/Al records, but not in the Ti/Al record. Within S6 silty layers (up to 1.5 cm in thickness) are preserved, which are interpreted as thin turbidites. Mediterranean turbidites appear enriched in titanium (Wehausen and Brumsack, 2000; Lourens et al., 2001), and it is therefore likely that they may have caused the high Ti/Al values within S6.

3.2. Chronology

The sapropel chronology of Lourens (2004) provides sapropel midpoint ages, which we used to establish an initial age model for our ODP Site 968 record. In the following we compared the sapropel (color) record of Site 968, which is linked to the intensity of the African summer monsoon, with the Chinese speleothem oxygen isotope ($\delta^{18}$O) record of the Sanbao and Hulu caves (Y. Wang et al., 2008) (Fig. 3). The speleothem record reflects changes in the strength of the East Asian summer monsoon (EAM). A correlation of the two records is in-line with the idea of a global monsoon, which considers monsoon as a manifestation of the seasonal migration of the intertropical convergence zone (ITCZ) and, hence, a climate system of the global scale (P. Wang, 2009). An in-phase response of the Asian and African monsoon system to external forcing is moreover consistent with transient climate modeling (Tuenter et al., 2005; Ziegler et al., in review). The speleothem $\delta^{18}$O record is dated by 186 U/Th dates back to 224,000 years ago with an average dating error of less than 1% (Y. Wang et al., 2001; Y. Wang et al., 2008). Currently, it is the best-dated, high-resolution paleoclimate record covering the last two glacial cycles. Virtually identical fluctuations in the $\delta^{18}$O record of independently dated individual stalagmites and an excellent agreement with records from Hulu (Y. J. Wang et al., 2001) and Dongge (Y. Wang et al., 2005) caves indicate the high-precision of the dating. During glacial periods the speleothem record is characterized by Dansgaard–Oeschger type variability, with rapid transitions between stadial and interstadial periods (Y. Wang et al., 2008).

We find that the color record of Site 968 matches very well with the variability in the Chinese speleothem oxygen isotope ($\delta^{18}$O) record of the Sanbao and Hulu caves, which reflects changes in the strength of the East Asian summer monsoon (EAM) (Y. Wang et al., 2008) (Fig. 3). Several sapropel layers of Site 968 display interruptions of the organic-rich layers by homogenous marly intervals. Sapropel interruptions are commonly observed, suggesting that sapropel formation is not only forced by changes in insolation (Rohling et al., 1997; De Rijk et al., 1999; Meyers and Arnaboldi, 2005). The most obvious interruption in our record occurs during sapropel S4, which consists of two equally dark layers. This was also found in other records (Calvert and Fontugne, 2001). In addition, sapropels S1, S3 and S4 are preceded by minor peaks, which are also clearly visible in the Sanbao—Hulu $\delta^{18}$O record. Sapropel S5 is one massive dark layer and similarly to the Sanbao—Hulu $\delta^{18}$O record shows no distinct interruption or precarious event.

We used the consistent sub-Milankovitch features of the speleothem $\delta^{18}$O record and our sapropel record to refine our age model. For this purpose, we assigned 18 age calibration points, which are tied to the sharp transitions before and after the major wet phases in the speleothem record to the boundaries of the sapropels as defined by the color record (Fig. 4, Table 1). For

![Fig. 3. $\delta^{18}$O record of the Sanbao—Hulu caves (Y. Wang et al., 2008) and ODP 968 color record on the initial age model, applying sapropel midpoint ages from Lourens (2004) as age control points. Black dots mark age control points.](image-url)
sapropels S9, S' and S10 we used the midpoint ages of Lourens (2004). A comparison between the radiometrically constrained ages and the astronomically derived ages of the midpoints of sapropels S1 to S8 reveals differences of less than a few hundred years, which is within the uncertainty limits of the two age models (mean age difference: 0.38 ± 0.65 ka) (Table 2). We note that for S3 and S4 midpoint ages would become substantially younger if the small precursory events are taken into account. Similarly, the midpoint of S5 shifts to an older age when the Ti/Al record is used with the precursory increase included.

It is important to note here, that during sapropel formation, not only the intensity of the African monsoon and associated North African runoff was intensified, but that the precipitation in the Mediterranean region increased as well. The radiometrically dated speleothem records from the Israeli Soreq cave provide important evidence that regional climate variability varied synchronously with the changes in the monsoon intensity (Bar-Matthews et al., 1999, 2000, 2003). The oxygen isotope composition of the speleothem calcite has been interpreted in terms of changes in local, annual precipitation changes (Bar-Matthews et al., 1996) Accordingly, marked decreases in the speleothem oxygen isotope composition have been interpreted as periods of enhanced rainfall. As alternative, it has been suggested that the speleothem data reflect an overall δ18O change in the source for the eastern Mediterranean vapor that eventually precipitated over the Levant (Marino et al., 2009). Another speleothem record of regional importance was derived from the Italian Argentarola cave (Bard et al., 2002). This oxygen isotope record shows a period of enhanced rainfall during the penultimate glacial which coincides with the timing of sapropel S6. Overall these regional archives suggest that during sapropel events dilution of ocean surface waters was not restricted to the Nile discharge but was rather widespread over the entire Mediterranean Sea due to increased rainfall (Bard et al., 2002).

3.3. Benthic oxygen isotope data of Site 968

We evaluate the refined sapropel based age model by comparison of our new Eastern Mediterranean benthic isotope record of Site 968 with the global benthic isotope stack LR04 (Lisiecki and Raymo, 2005). In order to produce a benthic oxygen isotope record we had to combine measurements from different species (Fig. 5). The benthic foraminiferal fauna revealed strong fluctuations in diversity and number as well as in the species composition throughout the record, probably triggered by the extreme environmental changes in terms of oxygen and nutrient availability. The laminated sapropel intervals represent gaps in the isotope record due to complete absence of benthic foraminifera. Nevertheless, the record of Site 968 shows the major features of the global ocean benthic isotope stack LR04 (Lisiecki and Raymo, 2005), although absolute values and amplitudes differ considerably. During glacial periods, values in Site 968 are roughly 0.5–1‰ lighter compared to the global ocean stack, whereas interglacial values are roughly 1.5–1.8‰ lighter. We attribute the general lighter values of Site 968 to the relative warm deepwaters in the Eastern Mediterranean. The higher amplitude between glacial and interglacial conditions demonstrates moreover that temperature and/or salinity effects are amplified over glacial terminations.

Several intervals show unexpected light values and a large degree of scatter. They typically follow on or occur at the top of the...
that the precipitation maxima of the Asian summer monsoon lag age differences with chronology of Lourens (2004). Table 2 (Schmiedl et al., 1998). On the other hand, these values have to be compared with sapropel formation as found for planktonic foraminiferal species and precipitation conditions in the Eastern Mediterranean during sapropel layers. Possibly, these values are related to higher runoff and precipitation conditions in the Northern Hemisphere are sources of methane when organic material decays under reducing conditions. The methane record from EPICA Dome C over the last 800,000 years is dominated by the 100 kyr–interglacial cycles and the ~23 kyr precession component (Spahni et al., 2005; Loulergue et al., 2008). Widespread wetland areas during periods of increased summer monsoon precipitation in the Northern Hemisphere are sources of methane when organic material decays under reducing conditions. The methane record from EPICA Dome C over the last 800,000 years is dominated by the 100 kyr–interglacial cycles and the ~23 kyr precession component (Spahni et al., 2005; Loulergue et al., 2008). Widespread wetland areas during periods of increased summer monsoon precipitation in the Northern Hemisphere are sources of methane when organic material decays under reducing conditions.

Table 1

<table>
<thead>
<tr>
<th>Event</th>
<th>Composite depth (cm)</th>
<th>Age (ka)</th>
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</thead>
<tbody>
<tr>
<td>Coretop</td>
<td>0.00</td>
<td>0</td>
</tr>
<tr>
<td>S1 (top)</td>
<td>0.58</td>
<td>6.5</td>
</tr>
<tr>
<td>S1 (bottom)</td>
<td>0.84</td>
<td>10.2</td>
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<tr>
<td>S3 (top)</td>
<td>5.40</td>
<td>77.3</td>
</tr>
<tr>
<td>S3 (bottom)</td>
<td>5.58</td>
<td>83.9</td>
</tr>
<tr>
<td>Precursory S3 (top)</td>
<td>5.64</td>
<td>86.6</td>
</tr>
<tr>
<td>Precursory S3 (bottom)</td>
<td>5.86</td>
<td>91</td>
</tr>
<tr>
<td>S4 (top)</td>
<td>6.47</td>
<td>99.6</td>
</tr>
<tr>
<td>S4 (bottom)</td>
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<td>104</td>
</tr>
<tr>
<td>Precursory S4 (top)</td>
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<td>105.9</td>
</tr>
<tr>
<td>Precursory S4 (bottom)</td>
<td>6.81</td>
<td>109.7</td>
</tr>
<tr>
<td>S5 (top)</td>
<td>7.63</td>
<td>121.4</td>
</tr>
<tr>
<td>S5 (bottom)</td>
<td>7.96</td>
<td>129.5</td>
</tr>
<tr>
<td>S6 (top)</td>
<td>10.36</td>
<td>165.5</td>
</tr>
<tr>
<td>S6 (interruption)</td>
<td>10.69</td>
<td>169.1</td>
</tr>
<tr>
<td>S6 (bottom)</td>
<td>10.95</td>
<td>178.5</td>
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<tr>
<td>S7 (top)</td>
<td>11.77</td>
<td>191.9</td>
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<td>198.5</td>
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<td>SB (bottom)</td>
<td>13.22</td>
<td>224.1</td>
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<td>S9 (midpoint)</td>
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<tr>
<td>S8 (midpoint)</td>
<td>16.18</td>
<td>288</td>
</tr>
<tr>
<td>S10 (midpoint)</td>
<td>17.50</td>
<td>331</td>
</tr>
</tbody>
</table>

4. Discussion

4.1. Phasing of the boreal summer monsoon

The radiometrically dated Chinese speleothem δ18O indicates that the precipitation maxima of the Asian summer monsoon lag precession minima by ~2700 years (Y. Wang et al., 2008). The consistency between this precession phase lag and the lag of ~3000 years adopted in the original sapropel chronology indicates that the intensity of the African and Indian–Asian summer monsoon respond with the same phase lag to the precession forcing. This is also consistent with transient climate modeling (Tuenter et al., 2005; Ziegler et al., in review).

Atmospheric methane records reconstructed from Antarctic ice cores provide additional constraints on the precession phase, because they reflect the strength of tropical monsoons with a secondary input from boreal sources (Ruddiman and Raymo, 2003; Fischer et al., 2008; Loulergue et al., 2008). Widespread wetland areas during periods of increased summer monsoon precipitation in the Northern Hemisphere are sources of methane when organic material decays under reducing conditions. The methane record from EPICA Dome C over the last 800,000 years is dominated by the 100 kyr–interglacial cycles and the ~23 kyr precession component (Spahni et al., 2005; Loulergue et al., 2008) (Fig. 6), with the latter showing a 32° ± 13.5° (2000 ± 850 years) lag with respect to precession (for the interval 0–250 ka on the EDC3 chronology (Fairén, 2007)).

The lag of the Chinese speleothem δ18O record was interpreted as being in-phase with July insolation in June insolation (Y. Wang et al., 2008), building on the fact that modern monsoon precipitation maximum occurs in July. Our and other climate modeling experiments indicate however that the maximum monsoon precipitation in July varies at orbital timescales in-phase with precession minima and thus with June insolation maxima (Tuenter et al., 2005; Kutzbach et al., 2008). These transient climate modeling experiments therefore elegantly demonstrate that seasonal signals cannot simply be extrapolated to orbital timescales as they are related to inertia in the system that produces lags of a few weeks and thus does not have to occur synchronously with the real forcing.

Initially, Hilgen et al. (1993) and Lourens et al. (1996) suggested that the 3000-year precession lag of the African monsoon could be the consequence of a direct thermal response at low latitudes to the insolation forcing as simulated by simple energy balance models (Short and Mengel, 1986), assuming that moisture availability in the monsoon region changes in-phase with the thermal response. Later simulations with a climate model of intermediate complexity (CLIMBER-2) showed, however, that the intensity of the African monsoon changes in-phase with precession (i.e., maximum summer insolation at the northern hemisphere) (Tuenter et al., 2005; Kutzbach et al., 2008). The lack of variable ice sheets may have led, however, to an underestimation of the monsoon response time in the modeling experiments (Kutzbach et al., 2008). To overcome this problem, the transient simulations were run again with prescribed glacial-bound ice fraction and height of the Eurasian and North American ice sheets, because CLIMBER-2 does not include an interactive ice-sheet module (Ziegler et al., in review).

Table 2

<table>
<thead>
<tr>
<th>Sapropel</th>
<th>New midpoint ages</th>
<th>Midpoint ages (Lourens, 2004)</th>
<th>Age difference (ka)</th>
</tr>
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<tbody>
<tr>
<td>S1</td>
<td>8.35</td>
<td>8.5</td>
<td>0.15</td>
</tr>
<tr>
<td>S3</td>
<td>80.6 (84.15 incl. prec. event)</td>
<td>81</td>
<td>0.4</td>
</tr>
<tr>
<td>S4</td>
<td>101.8 (104.65 incl. prec. event)</td>
<td>101</td>
<td>-0.8</td>
</tr>
<tr>
<td>S5</td>
<td>125.45</td>
<td>124</td>
<td>-1.45</td>
</tr>
<tr>
<td>S6</td>
<td>172</td>
<td>172</td>
<td>0</td>
</tr>
<tr>
<td>S7</td>
<td>195.2</td>
<td>195</td>
<td>-0.2</td>
</tr>
<tr>
<td>S8</td>
<td>216.8</td>
<td>216</td>
<td>-0.8</td>
</tr>
</tbody>
</table>
precipitation, we find that the resulting phase relations with precession minima also differ only by up to 500 years. We derive a maximum lag of 560 years, if we use JIAS precipitation in the orbital plus ice run. Similar results were obtained for the Indian–Asian monsoon (Ziegler et al., in review). Evidently an alternative mechanism must be responsible for the observed 3000-year phase lag.

4.2. The impact of non-linear responses to insolation forcing on the monsoon record

A major difference between the model simulations and the data is that the modeled monsoon intensity appears to respond in a linear way to the insolation forcing. The precession cycles show a maximum lag of 560 years, if we use JJAS precipitation in the orbital plus ice run. Similar results were obtained for the Indian–Asian monsoon (Ziegler et al., in review). Evidently an alternative mechanism must be responsible for the observed 3000-year phase lag.

4.3. The impact of North Atlantic cold events on the precession phase of the boreal monsoon

RAPIDLY OCCURRING MILLENNAL SCALE CHANGES SHOW AN IDENTICAL PATTERN OF ABRUPT CLIMATE CHANGE DURING THE LAST GLACIAL CYCLE, AS IT HAS BEEN DESCRIBED IN MANY RECORDS FROM THE LOW LATITUDES: IN A RECORD OF RIVER DISCHARGE INTO THE GULF OF GUINEA DUE TO CHANGES IN THE WEST AFRICAN MONSOON INTENSITY (Weldeab et al., 2007), IN RECORDS OF INDIAN SUMMER MONSOON RELATED UPEWELLING IN THE ARABIAN SEA (Schulz et al., 1998; Altabet et al., 2002), IN A RECORD OFFSHORE MAURETANIA RECORDING THE CHANGE BETWEEN DRY AND HUMID PERIODS IN NORTHWEST AFRICA (Tjallingii et al., 2008), IN THE CARICO BASIN OFFSHORE VENEZUELA RECORDING THE RIVER DISCHARGE IN THE NORTHERN SOUTH-AMERICA (Peterson et al., 2000; Haug et al., 2001). Evidently, millennial scale variability forms also a persistent feature of the sapropel and the Chinese speleothem δ18O records in the form of interruptions of maximum monsoon intensities and precursory events. The timing of these features coincides with the occurrence of cold spells (i.e., Heinrich events) in the North Atlantic (Y.J. Wang et al., 2001; Y. Wang et al., 2008). A rapid ventiliation of the EMS water column during these periods was most likely also facilitated by the intensified cold northerly winds over the Adriatic and Aegean Seas during winter and spring (Rohling et al., 2002; Casford et al., 2003). The African summer monsoon and related fluvial discharge to the Mediterranean was probably weakened during these periods by warmer sea surface temperatures in the South Atlantic and the reduced thermal contrast between land and sea (Chang et al., 2008). According to the idea of a bipolar see-saw effect, the Southern Hemisphere warms due to the decreased intensity of the North Atlantic overturning circulation. This is well documented in an Antarctic temperature record from the Dome Fuji ice core, which has a robust chronology, building on the linear relation of oxygen/nitrogen ratios in the ice with variations in local summer insolation (Kawamura et al., 2007). The record demonstrates that the North Atlantic cold event related monsoon minima coincide with sharp Antarctic temperature maxima (Fig. 6). The termination of Antarctic warming coincides with rapid onset of warming and hence increase in summer monsoon intensity in the Northern Hemisphere (Blunier and Brook, 2001).

Accordingly, we propose that the observed precession phase lag of 3000 for the boreal summer monsoon originates from the suppression of monsoon intensity during North Atlantic cold
events. As an example we compared the Chinese speleothem record with the insolation forcing (represented by summer insolation record 65 N, June 21st, Fig. 7). This comparison reveals two major features. In first instance the Chinese speleothem record is non-linearly related to the insolation forcing, because it exhibits a clear threshold value or cutoff point. Secondly, major deviations in the Chinese speleothem record from the insolation curve (grey bars) reflect the superposition of prominent monsoon minima induced by North Atlantic cold events. Because the most prominent North Atlantic cold events almost all precede precession minima, they introduce a systematic shift of the midpoint of the monsoon maximum to younger ages.

To further test this conceptual model, we performed cross spectral analysis between the original summer insolation record (June 21st, 65° N) and one which is clipped above 510 W/m² and includes minimum values at the timing of the cold events. These two simple modifications to the insolation series introduced a lag in the precession band of 32° ± 4° (approx. 2000 years). Obviously, the value of the lag depends on the threshold value and the minimum values assumed for the cold events. Nevertheless, this simple conceptual model demonstrates in an empirical way that North Atlantic cold events act as an important negative feedback mechanism on the strength of the boreal monsoon.

Such an interpretation is unlike other explanations not in conflict with climate modeling experiments on both, the Milankovitch but also the sub-Milankovitch scale. Y. Wang et al. (2008) ascribed deviations of the Sanbao–Hulu speleothem record from the chosen July insolation to sub-Milankovitch stadial as well as interstadial events. In contrast we explain deviations from the insolation forcing entirely by the overprint of cold events (stadials). Modeling results support only the latter explanation. Sub-Milankovitch variability observed in monsoon proxies is commonly associated with changes in the Atlantic meridional overturning in response to freshwater forcing in the North Atlantic and a weakened monsoon (Y.J. Wang et al., 2001; Rahmstorf, 2002; Weldeab et al., 2007). In addition, previous
studies with the CLIMBER-2 model demonstrate that a North Atlantic cold event, induced by a prescribed freshwater pulse, significantly weakens both the African and Indian–Asian summer monsoon (Claussen et al., 2003; Jin et al., 2007; Tjallingii et al., 2008). Because large North Atlantic ice surge events do not occur prior to the Late Pleistocene (Hodell et al., 2008), we expect that the phase relation between Northern Hemisphere summer insolation and Indian and East Asian Monsoon will be more in-phase during the Pliocene and the early Pleistocene. This supports the in-phase tuning approach of African monsoon records for the Pliocene (Lourens et al., 2001).

5. Conclusion

The African and Asian summer monsoons vary synchronously during the Late Pleistocene on (sub-)Milankovitch timescales. Proxy records indicate that both African and Asian monsoon lags precession minima by ~3000 years, while climate modeling experiments simulate an in-phase relationship with precession. Our benthic isotope record demonstrates that the application of such a lag in dating the sapropel record is in accordance with the marine isotope chronology. We conclude that the ~3000-year precession lag is the result of a non-linear response to the insolation forcing in combination with the weakening of the monsoon during North Atlantic cold events.

Acknowledgements

This study is supported by the Research Council for Earth and Life Sciences (ALW) with financial aid from the Netherlands Organization for Scientific Research (NWO) to L.J. Lourens (grants 853.00.032 and 834.04.003). We thank three anonymous reviewers whose comments improved the manuscript. A. van Dijk, G. Ittmann, and Verheul, M. are thanked technical support. Steven Clemens, Frits Hilgen, Gert-Jan Reichart and Nanne Weber are thanked for discussions.

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