Holocene environmental changes in northeast Thailand as reconstructed from a tropical wetland

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1. Introduction

Excess precipitation and drought associated with the varying intensities of the Asian monsoon impact a region where more than half of the world’s population live. A better knowledge of how tropical ecosystems responded to past shifts in monsoon intensity may help further our understanding of their resilience to changes in effective moisture (precipitation minus evapotranspiration, P-ET) and may provide guidelines for their sustainable management.

The Asian monsoon is generally subdivided into the tropical Indian Ocean and the tropical/subtropical Southeast Asian monsoon, which are roughly delineated at 105° E (Wang et al., 2005a, 2005b). Paleoclimate archives show that both monsoon systems responded to decreasing insolation patterns during the Holocene (Kutzbach, 1981; Wang et al., 2005a), but whether this response was gradual or abrupt (Morill et al., 2003; Dykoski et al., 2005), synchronous (e.g. Zhou et al., 2005, 2007; Zhang et al., 2011) or asynchronous (An et al., 2000; Herzschuh, 2006; Cai et al., 2010; Wang et al., 2010) over larger regions is still debated. An asynchronous pattern would suggest decoupling of the Indian Ocean and Southeast Asian summer monsoon during much of the Holocene (Herzschuh, 2006; Wang et al., 2010). High-resolution precipitation reconstructions for...
the last millennium, however, show that severe drought intervals were registered more or less simultaneously in paleoarchives across the Asian monsoon region (Cook et al., 2010; Sinha et al., 2011), which suggests a synchronous response. The underlying causes of sub-millennial scale intensified and weak monsoon phases are still not resolved, but climatic links to the North Atlantic region (Sinha et al., 2007; Berkelhammer et al., 2010) and to the El Niño Southern Oscillation (Buckley et al., 2007) have been suggested, as well as multiple global forcing mechanisms (Cook et al., 2010; Sinha et al., 2011).

Paleoclimatic archives have greatly advanced our understanding of past shifts in Asian summer monsoon intensity. However, the comparably low number of investigated sites and their often weak chronological framework, the application and correlation of different types of proxies for tracing past hydroclimatic variability, and the varying responses of different ecosystems (and proxies) to shifts in precipitation suggest that a much denser regional network of well-dated, multi-proxy records is necessary to faithfully trace sub-millennial scale shifts in effective moisture throughout the Asian monsoon region.

Thailand is one of the regions that is heavily under-sampled with respect to paleoclimatic and paleoenvironmental records (Penny, 1998; White et al., 2004; Buckley et al., 2007; Boyd, 2008; Marwick and Gagan, 2011), despite having several large lake-wetland systems and extensive cave deposits. Moreover, the country’s tropical climate is influenced by both the Indian Ocean and the Southeast Asian monsoon sub-systems. Summer rainfall intensity shows clear links to the El Niño Southern Oscillation (Singhrattna et al., 2005), but the Indian Ocean and western North Pacific summer monsoon modulate annual rainfall variability over Thailand (Limsakul et al., 2010). Paleoclimate records from Thailand have the potential to allow tracing of the influence of the two monsoon sub-systems and to contribute to the open questions regarding the extent to which they are synchronous and are coupled or decoupled during the Holocene.

One of the few late Pleistocene and Holocene lake sediment records that has been investigated is from Nong (Lake) Han Kumphawapi in northeast Thailand (Kealhofer, 1996; Penny et al., 1996; Penny, 1998; 1999). Based on sediment lithology and vegetation reconstructions of the longest core KUM.3 (Fig. 1B), Kealhofer and Penny (1998) and Penny (1998) concluded that arid conditions interrupted by intervals with higher rainfall prevailed between <14,000 and 10,000 cal years BP, while the early Holocene (10,000–9000 cal years BP) may have experienced higher effective moisture (Kealhofer and Penny, 1998) (Table 1). Marked changes in the local flora between 9000 and 5000 cal years BP suggested higher effective moisture, while an increase in charcoal and the reduction of dry-land taxa after 5000 cal years BP, as well as the re-appearance of dry-land deciduous forests at around 3000 cal years BP were attributed to human activities (Penny, 1998) (Table 1).

Here we combine sediment geochemistry and diatom analysis of a new sediment sequence from Nong Han Kumphawapi, to generate a reconstruction of environmental history at higher temporal resolution than prior studies; we also revaluate whether observed changes in the lake environment are linked to shifts in Asian monsoon intensity or are an expression of human activities in the lake’s catchment.

2. Regional setting

Nong Han Kumphawapi (17° 11′ N, 103° 02′ E) is situated at 166 m above sea level (asl) on the Khorat Plateau of northeast-Thailand (Fig. 1A, B) and is one of Thailand’s few natural freshwater lakes. The Khorat Plateau reaches elevations of 150–500 masl and covers an area of about 170,000 km². It is bordered by 600–1000 m
high escarpments in the west and south, but slopes gently toward the Mekong River into Laos. The NW–SE extending Phu Phan mountain range divides the plateau into two basins: the smaller Sakhon Nakhon basin in the north and the Khorat basin to the south (El Tabakh et al., 2003). Sedimentary rock sequences on the Khorat Plateau include late Triassic to Neogene sandstones, conglomerates, mudstones, siltstones and shales, which accumulated in lacustrine, fluvial and alluvial environments (El Tabakh et al., 2003; Wannakomol, 2005). Of special importance is the Upper Cretaceous Maha Sarakham Formation, an extensive evaporite succession and the main source of soil and groundwater salinization and sinkhole formation (El Tabakh et al., 2003; Wongpokhom et al., 2008). Quaternary sediments are mainly fluvial gravel, sand, silt and clay. Laterites occur frequently, and their formation seems to date to about 0.6–0.7 ma based on tektites contained in the laterites (Tamura, 1992).

The region has a continental and tropical monsoon climate. The dry, cool continental northeast monsoon operates from November to February, and the wet, warm southwest monsoon supplies precipitation between May and October. Mean monthly temperatures for the years 1951–2008 are c. 22–25 °C between November and February, and 27–30 °C between March and October (Klubseang, 2011). Almost no rain falls between November and February, but mean monthly precipitation increases slightly in March and April to ~50–100 mm, and rises to 200–300 mm between May and October (Klubseang, 2011).

Nong Han Kumphawapi is located in the southeast part of the Salak Nakhon basin, where it occupies a broad alluvial plain. The lake is surrounded by low-relief hills rising to over 200 m, while higher peaks to the west and east rise to between 500 and 622 m. The Maha Sarakham Formation, which is approximately 130–170 m thick, underlies the Kumphawapi basin and the area immediately to the north and south (Fig. 2). It is made up of alternating clay-, silt- and sandstones and evaporites (rock salt inter-bedded with gypsum, potash and anhydrite). The salt faces impact the surface morphology due to dissolution of underlying salt sequences and diapiric salt domes. The island of Ban Don Kaeo, which rises 10–15 m above the surrounding herbaceous swamp in the southern part of Kumphawapi, constitutes such a salt mound (Figs. 1B, 2). Earlier studies (Rau and Supajanya, 1985) and more recent seismic investigations (Sataruaga et al., 2004) indicate the presence of a salt dome below Kumphawapi and rock salt in variable thickness adjacent to the lake. The formation of the lake basin is therefore likely due to a collapse of sub-surface rock salt cavities.

Kumphawapi today is a shallow circumneutral lake (pH 6.8), with a water depth of < 4 m in its deepest part (Fig. 1B), and is c. 7 km long and 4 km wide. Floating plant communities create extensive herbaceous mats that cover large parts of the lake and form mosaics with sheltered open water areas. The aerial extent of the wetland is estimated to around 56 km², and the open water surface is about 20 km² (Fig. 1B), but this may vary greatly between dry and wet seasons. Numerous perennial and seasonal streams feed the lake from the surrounding hills. Of these, Huai Phai Chan Yai, which rises on the southern slopes of the Phu Phan Range, provides the largest fluvial input. The Lam Pao River drains the lake to the south (Fig. 1B). Groundwater flow, on the other hand, seems to be toward the northwest. Large irrigation work to expand the lake’s water storage capacity during the dry season led to the construction of a wide dam around the lake (completed in 1994). Pump stations now pump water from the lake into irrigation canals, which irrigate 36 km² of agricultural area.

### 3. Materials and methods

Sediment cores from Nong Han Kumphawapi were obtained in January 2009 from a coring platform using a modified Russian corer (7.5 cm diameter, 1 m length). To achieve a continuous sequence, sediment cores were taken with an overlap of 50 cm at each coring site. The water depth at coring sites CP1 and CP2 was 1.50 m and 1.70 m at sites CP3, CP3A and CP3B. The sediment cores were wrapped in plastic and placed in PVC tubes for transport to the Department of Geological Sciences at Stockholm University. Laboratory work included detailed lithostratigraphic descriptions of all sediment sequences and correlation between overlapping core segments. CP3A was further analyzed for long-core magnetic susceptibility, geochemistry, diatoms and 14C dating.

Whole core magnetic susceptibility was measured along the split core at 5-mm resolution with a Bartington MS2EI point sensor core logger (at 0.565 kHz, with a low field intensity of 80 A/m) and expressed as volume specific susceptibility (χv) (×10⁻⁵ SI).

For loss-on-ignition (LOI) analyses, consecutive 1-cm samples were dried at 105 °C and homogenized. Samples were then combusted for 2 h at 550 °C to determine organic matter content, and for 4 h at 950 °C to estimate the carbonate content of the sediments. LOI is expressed as percentage loss of the original dry weight. Selected 1-cm sub-samples were freeze-dried and homogenized prior to analyses of total organic carbon (TOC), total nitrogen (TN), total sulfur (TS), and bulk δ¹³Corg, δ¹⁰Corg, δ¹⁰Corg, δ¹⁰Corg, δ¹⁰Corg, and δ¹⁰Corg were measured on a Carlo Erba NC2500 elemental analyzer, which is coupled to a Finnigan MAT Delta mass spectrometer. Δ¹⁰Corg is expressed as δ (%), relative to the Vienna PeeDee Belemnite (VPDB) standard, and measurement reproducibility is better than 0.15%. The Corg/Norg ratio, which is commonly used to discriminate the origin of lacustrine organic material (aquatic/terrestrial) (Meyers and Lallier-Vergès, 1999; Meyers and Teranes, 2001), was calculated as Corg/(%)/Norg (%) and multiplied by 1.167 to yield atomic mass ratios. Sulfur in lake sediments comprises inorganic and organic sulfur species, which can become incorporated into iron-sulfide compounds, or into organic matter. Although the incorporation of sulfur into lake sediments depends on a variety of different factors, organic matter concentrations and availability of soluble iron have been cited as two of the main factors (Mitchell et al., 1990). Here we mainly use δ²S in conjunction with other proxies, to estimate changes in water column oxygenation.
Subsamples for diatom analysis were treated with 10% HCl to remove any carbonates, heated in H2O2 to oxidize organic matter, and then rinsed multiple times with distilled water to remove oxidation by-products. Afterwards, an aliquot of each treated sample was dried onto a coverslip, and the coverslip was mounted onto a glass slide using a permanent mounting medium (Zrax or Naphrax). At least 300 diatom valves from each depth interval were identified and counted in transects using a 100× oil immersion objective on a Zeiss Axioscop 2 plus microscope (University of Nebraska—Lincoln) or Olympus BH 2 microscope (Stockholm University). In each sample, the abundance of each identified diatom taxon is expressed as a percentage of the total number of valves counted in that sample. Diatom abundance and preservation was variable, and some samples contained very low concentrations of diatoms. In these samples, diatoms were enumerated on up to 5 slides in an attempt to achieve a count of 300 diatom valves. Samples with total diatom counts of <300 valves are not included in the diagram.

Sediment samples selected for 14C dating were sieved under running tap water. The sieve remains were stored in deionized and slightly acidified water, examined under a binocular microscope and identifiable plant remains (seeds, charcoal, leaves, small twigs and wood fragments) were picked out, placed in pre-cleaned glass vials and dried at 105 °C overnight. Charcoal and wood samples were pretreated following the acid-base–acid method (de Vries and Barendsen, 1952) where HCl is used to remove carbonates and fulvic acids and NaOH to remove humic acids. More fragile plant macrofossils were treated with HCl only and are marked in Table 2. The samples were then rinsed in deionized water and dried at 50 °C overnight, then weighed into pre-combusted quartz tubes with silver and CuO and combusted at 850 °C overnight to produce CO2. Samples with less than 0.8 mg of carbon were graphitized in the presence of hydrogen on an iron catalyst at 560 °C for a maximum of 4 h according to the Bosch-Manning Hydrogen Reduction Method (Vogel et al., 1984). The CO2 from the larger samples was converted to graphite on an iron catalyst using the zinc reduction method (Slota et al., 1987). The 14C/12C ratio and 13C/12C were measured on a 0.5MV National Electrostatics Corporation accelerator mass spectrometer (AMS) at the 14CHRONO Centre, Queen’s University Belfast. The radiocarbon age and one standard deviation were calculated following the conventions of Stuiver and Polach (1977) using the Libby half-life of 5568 years and a fractionation correction based on δ13C measured on the AMS which accounts for both natural and machine fractionation. The fourteen 14C dates were calibrated with the Calib 6.0 online program using the northern hemisphere terrestrial calibration curve (Reimer et al., 2009) (Table 2).

The age–depth curve for CP3A was constructed using the Bacon age-modeling software (Blaauw and Christen, 2011), which explicitly models past accumulation rates and their variability. The routine assumed an average accumulation rate of 25 years/cm, with other rates possible though less likely (using a gamma prior distribution with shape 2). The CP3A sequence was divided into 5 cm thick sections, and the accumulation rate of each subsequent cm was allowed to vary to a certain degree (average 50%, a beta distribution with strength 10 and mean 0.5) from its underlying cm, therefore modeling an evolving accumulation rate. In order to accommodate for outlying dates, we assume that the 14C dates

**Fig. 2.** Geological map of Kumphawapi, redrawn from the Geological Map of Udon Thani Province (2005) (Department of Mineral Resources, Thailand; http://www.dmr.go.th/ewt_news.php?nid=8879). The coordinate system is based on the UTM Grid System (Indian 1975 zone 47). The Upper Jurassic Sao Khua Formation to the west of the Kumphawapi Basin includes silt- and sandstones with calcrite and silcrete horizons, and the Lower Cretaceous Phu Phan Formation consists of gravelly sandstone and siltstones. The middle Cretaceous Khok Kruat Formation with siltstones, calcareous sandstones and conglomerates with calcite horizons is found to the west and east of the basin. The Upper Cretaceous Maha Sarakham Formation, which is approximately 130–170 m thick, underlies the Kumphawapi basin and the area immediately to the north and south. It is made up of alternating clay-, silt- and sandstones and evaporites (rock salt inter-bedded with gypsum, potash and anhydrite). Dissolution of the rock salt, salt domes and salt anticlines are common features. River terrace and alluvial deposits are made up of gravel, sand, silt, clay and laterites.
The sedimentary units were re-drawn to make them comparable to our new sediment succession. Charcoal, seeds, leaves, wood fragments, small twigs, and other calibrated materials were used as proxies to complement the dated sequence. The location of the coring points is shown in Fig. 1B. Sedimentary units 1–3 are well-represented in the new sediments, whereas the presence of a hiatus or a very low sedimentation rate is suggested for the large age difference between these close-by dates (UBA-14170, UBA-12662, UBA-14168, UBA-12661), and one sample (UBA-14166) resulted in older ages than expected. Possible explanations for the large age difference between these close-by dates are the presence of a hiatus, or a very low sedimentation rate. The gradual transition between layers 14 and 13 (Table 3) and the geochemical proxies (Fig. 4) do not provide support for a hiatus at 1.60 m. However, the break in peat growth and a hiatus. Samples UBA-16763, UBA-16766, UBA-16768, UBA-16769, and UBA-16770 were dated in the lower and upper parts of the sequence plot sequentially according to depth, but five 14C dates between 1.18 and 1.60 m do not fit this general pattern (UBA-14170, UBA-12662, UBA-14168, UBA-12661, and UBA-14166), and resulted in older ages than expected. Four of these samples consisted of wood fragments (UBA-14170, UBA-12662, UBA-14168, UBA-12661), and one sample was composed of charcoal and seeds (UBA-14166) (Fig. 3B). Older ages for wood samples may result from reworking, since wood is more resistant to decomposition than seeds and leaves. Reworking of undetermined fragile plant remains (Table 2; Fig. 3A–C). The 14C dates in the lower and upper parts of the sequence plot sequentially according to depth, but five 14C dates between 1.18 and 1.60 m do not fit this general pattern (UBA-14170, UBA-12662, UBA-14168, UBA-12661, and UBA-14166), and resulted in older ages than expected. 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The gyttja clays of unit 1 (c. 9400–c. 6700 cal yr BP) have bulk mineral magnetic susceptibility (MS) values of around 5 × 10⁻⁵ SI, suggesting some contribution of magnetic minerals. LOI (at 550 °C) is ~6% and TOC around 2–3%, which indicates that the organic matter content of the sediments is low. Carbonates are present in the sediments at between 1 and 5% (Fig. 4). The C/N ratio of 15–20 indicates a mix of terrestrial and lacustrine organic matter sources (Meyers and Teranes, 2001), which is also suggested by δ¹³Corg values of −21 to −24% (Figs. 4, 5A). Sediment composition and geochemical parameters thus provide the picture of a low productivity lake, which received most of its input through catchment run-off. The comparably low organic matter content of the sediments is, however, somewhat unexpected. The catchment supported open woodlands and dry/mixed deciduous forests (Kealhofer and Penny, 1998; Penny, 1998) (Table 1), whose plant remains should have been transported into the lake by seasonal streams. The low organic matter content of the sediments could therefore be a product of extensive decomposition and oxidation of the organic material. Low %TS and slightly higher values of δ¹³Corg (Fig. 5A, B) favor the assumption of an oxic environment and higher rates of decomposition.

MS values decline in the clay gyttja (layers 19–18) of unit 2a (c. 6700–5900 cal yr BP), coincident with a gradual increase in organic carbon content (Fig. 4). C/N ratios of 14–18 and δ¹³Corg values of around −21 to −22% are similar to unit 1, indicating a mix of terrestrial and aquatic organic matter. The increase in %TS in the upper part of unit 2a precedes the rise in %TOC and suggests reducing bottom water conditions. The sediments in unit 2a can therefore be interpreted as reflecting less catchment run off and increased lacustrine productivity, and/or enhanced preservation of the organic material under less oxic conditions.

The sediments in unit 2b (c. 5900–1600 cal yr BP) change from gyttja (layers 17–15) to peaty gyttja (layer 14) and peat (layer 13), and back to peaty gyttja (layers 12–11) (Table 3). Soil mineral particles in sieve remains are present in low amounts throughout unit 2b, but show a peak at 1.60 m (layer 13) and 1.30–1.20 m (layers 12–11). Charcoal appears at 1.60 m and has a distinct peak at 1.25–1.20 m (Fig. 4). MS values are mainly negative, but start to increase at 1.40 m, in the upper part of layer 13. LOI and TOC mirror this trend with high values of 50–60% and 30–35%, respectively in layers 17–13, and a subsequent decline to 30% and 18%, respectively (Fig. 4). C/N ratios increase and are −20–25 throughout unit 2b, and δ¹³Corg values of −25% (Figs. 4, 5A) are typical of higher plants (Meyers and Teranes, 2001), indicating that most of the organic material is of terrestrial origin. %TS is highest in layers 16–15 with 0.8%, but declines upward, more or less following the pattern of %TOC and %LOI (Figs. 4, 5B). Anoxic conditions could explain the initially high TS values and the gradual decline in %TS could be ascribed to subsequent sulfur limitation. The sediments and the geochemical proxies thus document a distinct change around 5900 cal yr BP to a lake with anoxic bottom water conditions and increasing terrestrial organic matter contributions. About 500 years later, the shallow lake transformed into a wetland, which became increasingly dominated by terrestrial vegetation. The shift from lake to wetland and subsequently to peatland shows that the water level in the basin had decreased substantially. The transition from wetland to peat, which is seen in the stratigraphy at 1.60 m (Fig. 4), coincides with a hiatus as suggested by the ¹⁳C dates and by high amounts of soil mineral particles, while the age model points to very low sedimentation rates (Fig. 3B, C). These two assumptions do not necessarily contradict each other, since very low sedimentation rates can also conceal

### Table 3
Lithostratigraphic description of sediment sequence CP3A and modeled age range of the lower boundary of each layer. MAP = mean age point; gLB = gradual lower boundary; sLB = sharp lower boundary.

<table>
<thead>
<tr>
<th>Depth below surface (m)</th>
<th>Lithostratigraphic description</th>
<th>Layer</th>
<th>Unit</th>
<th>Age range (MAP) (cal yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.00–0.11</td>
<td>Gray in field, now oxidized reddish black clay gyttja, gLB</td>
<td>1</td>
<td>3c</td>
<td>66–386 (307)</td>
</tr>
<tr>
<td>0.11–0.25</td>
<td>Gray in field, now oxidized clay gyttja with reddish spots, light laminations, gLB</td>
<td>2</td>
<td>3c</td>
<td>155–515 (440)</td>
</tr>
<tr>
<td>0.25–0.42</td>
<td>Gray brown in field, now greenish-brown slightly oxidized clay gyttja, gLB</td>
<td>3</td>
<td>3c</td>
<td>288–648 (551)</td>
</tr>
<tr>
<td>0.42–0.51</td>
<td>Dark brown clay gyttja, gLB</td>
<td>4</td>
<td>3c</td>
<td>384–734 (602)</td>
</tr>
<tr>
<td>0.51–0.58</td>
<td>Oxidized greenish brown clay gyttja with reddish-black/brown spot, gLB</td>
<td>5</td>
<td>3c</td>
<td>476–836 (615)</td>
</tr>
<tr>
<td>0.58–0.65</td>
<td>Olive green gyttja, gLB</td>
<td>6</td>
<td>3b</td>
<td>558–903 (653)</td>
</tr>
<tr>
<td>0.65–0.77</td>
<td>Dark brown gyttja, gLB</td>
<td>7</td>
<td>3b</td>
<td>727–1217 (1043)</td>
</tr>
<tr>
<td>0.77–0.91</td>
<td>Dark greenish brown clay gyttja, gLB</td>
<td>8</td>
<td>3a</td>
<td>1213–1608 (1409)</td>
</tr>
<tr>
<td>0.91–1.05</td>
<td>Dark brown clay gyttja, some FeS stains, gLB</td>
<td>9</td>
<td>3a</td>
<td>1469–2089 (1500)</td>
</tr>
<tr>
<td>1.05–1.18</td>
<td>Dark brown/blackish brown clay gyttja, sLB</td>
<td>10</td>
<td>3a</td>
<td>1553–2228 (1562)</td>
</tr>
<tr>
<td>1.18–1.35</td>
<td>Dark brown-black peaty gyttja, possibly some clay, gLB</td>
<td>11</td>
<td>2b</td>
<td>3043–3403 (3168)</td>
</tr>
<tr>
<td>1.35–1.60</td>
<td>Blackish brown peat, gLB</td>
<td>12</td>
<td>3b</td>
<td>4963–5358 (5317)</td>
</tr>
<tr>
<td>1.60–1.67</td>
<td>Blackish brown peaty gyttja, gLB</td>
<td>13</td>
<td>2b</td>
<td>5026–5741 (5554)</td>
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<tr>
<td>1.67–1.72</td>
<td>Dark brown gyttja, gLB</td>
<td>14</td>
<td>2b</td>
<td>5089–5924 (5715)</td>
</tr>
<tr>
<td>1.72–1.75</td>
<td>Transition zone: greenish brown and black gyttja, gLB</td>
<td>15</td>
<td>2b</td>
<td>5128–5998 (5786)</td>
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<tr>
<td>1.75–1.85</td>
<td>Greenish brown gyttja, gLB</td>
<td>16</td>
<td>2b</td>
<td>5285–6295 (5874)</td>
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<tr>
<td>1.85–1.95</td>
<td>Olive brown clay gyttja, gLB</td>
<td>17</td>
<td>2a</td>
<td>5493–6573 (6037)</td>
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<tr>
<td>1.95–2.16</td>
<td>Olive brown clay gyttja, gLB</td>
<td>18</td>
<td>2a</td>
<td>6030–7115 (6659)</td>
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<tr>
<td>2.16–1.85</td>
<td>Clay gyttja</td>
<td>19</td>
<td>2a</td>
<td>9294–9574 (9420)</td>
</tr>
</tbody>
</table>

UBA-14169 and UBA-14170 are also from approximately the same depth, but vary in age by 400–3000 years (Table 2, Fig. 3B). The boundary between layers 11 and 10 is sharp (Table 3), which would suggest a hiatus at 1.18 m; moreover, the second peak in soil mineral particles between 1.20 and 1.30 m depth, just below this transition, points to desiccation, and the high number of charcoal particles at 1.18–1.25 m could indicate surface burning (Fig. 4). Taken together, these observations would support a break in deposition in layers 11–12, and at the transition between layers 11 and 10.

Different age models were run to test how the outlying dates would influence the modeled age–depth curve and whether the break in sedimentation, as suggested around 1.60 m and 1.30–1.20 m, could be reproduced. Exclusion of the outliers (UBA-14170, UBA-12662, UBA-14168) provided a fairly smooth age–depth curve through most of the accepted data points, with higher sedimentation rates below 1.60 m and above 1.20 m, and low sedimentation rates around 1.60–1.50 m and 1.30–1.18 m depth (Fig. 3C), but failed to produce clear breaks in sedimentation. Modeled minimum and maximum errors between 2.65 and 1.60 m are 180 and 1200 years, respectively (mean 660 years), 260 and 1400 years (mean 390 years) between 1.60 and 1.20 m, and 240 to 760 years (mean 210 years) for the upper 1.20 m. In the following we use the modeled mean age points for assigning age estimates to the different layer and unit boundaries (Table 3), but also assume breaks in peat growth and sedimentation, respectively between 1.60 and 1.18 m depth.

4.2. Lithostratigraphy and geochemistry
a hiatus. Peat growth seems to have been interrupted by periodic desiccation, as suggested from the presence of soil mineral particles and the frequent appearance of charcoal, although this is not corroborated by the geochemical proxies (Fig. 4). The lithological change from peat to peaty gyttja at 1.35 m and the geochemistry indicate a rise in water level around 3200 cal yr BP and the subsequent establishment of a wetland/shallow productive lake. The occurrence of soil mineral particles in the peaty gyttja of layers 12–11 and high amounts of charcoal at 1.20 m depth, just below the transition to unit 3a, however suggests that this shallow lake also experienced periodic dryness (Fig. 4). Moreover, the sharp boundary between layers 11 and 10 at 1.18 m depth (Table 3) and very low sedimentation rates may indicate another hiatus (Fig. 3).

Unit 3a (c. 1600–1000 cal yr BP) is made up of clay gyttja with varying colors and FeS stains (layers 10–8) (Table 3). The distinct lithological transition observed between layers 11 and 10 is not reflected in the geochemical parameters, which indicate a gradual decline in %LOI, %TOC, %TS and in the C/N ratio, and δ¹³Corg values of −25 to −24‰ (Fig. 4). The change in sediment composition between units 2b and 3a and the geochemical parameters thus document a return to a higher water level, and the slightly higher proportion of mineral material could signify higher catchment run-off. LOI and TOC values are around 20% and 10%, respectively, and carbonate percentages increase slightly in unit 3a. C/N ratios of 11–17 and more depleted δ¹³Corg values in layers 9 and 8 suggest a higher proportion of algal organic material as opposed to terrestrial derived organic matter (Figs. 4, 5A).
4.3. Diatom assemblages and lake status changes

Diatom abundance and diversity in the sediments are highly variable. The lowermost sediments in diatom zone A (3.70–1.95 m; 9200–6000 cal yr BP) are dominated by Aulacoseira granulata (>75%), with low relative abundances (<5%) of both Aulacoseira ambigua and Cyclotella meneghiniana (Fig. 6). Aulacoseira granulata is a cosmopolitan planktonic species that is abundant in conditions of low light and high turbulence (Kilham and Kilham, 1975; Anderson, 2000). This includes both monomictic and dimictic lakes, where the average daily irradiance necessary for diatom growth is low because of deep mixing of surface waters, as well as shallow polymeric lakes where turbidity is high. Both A. granulata and C. meneghiniana (Tuchman et al., 1984) are common in relatively shallow nutrient-rich tropical lakes (Gasse et al., 1983).

Diatom concentrations are very low in diatom zone B between 1.95 and 1.765 m (c. 6000–5800 cal yr BP), which suggests either very low diatom production or dissolution of diatom silica. The sediments in this zone are gyttja, which suggests algal production may have been moderately high. Thus, one explanation for the absence of diatoms is that they were excluded as a result of resource competition with other algal groups. None of the geochemical data provide a clear picture of a driver of such a shift. It also is possible that the diatoms were dissolved after their deposition in the sediments. Many very different conditions can dissolve diatom valves, including high pH, high salinity, low dissolved silica, and sediment drying and oxidation. Given the inferred hiatuses subsequent to the deposition of this unit, one logical scenario is that the periodic desiccation of the sediments enhanced the breakage and subsequent dissolution of diatoms in the underlying sediments. However, presently insufficient information is available to confidently identify the likely cause of the absence of diatoms in this zone of the core.

The highest diversity of diatoms is in diatom zone C between 1.765 and 0.895 m (Fig. 6). In the lower part of this zone (1.765–1.345 m, subzone C1; c. 5800–3200 cal yr BP) Aulacoseira granulata declines in relative abundance to <35% and is replaced by Aulacoseira ambigua and a diverse array of benthic diatoms. The higher abundance of A. ambigua (5–20%) and benthic diatoms (55–75%) suggests increased water clarity. The assemblage of benthic diatoms consists of species characteristic of low-alkalinity waters, such as Brachysira vitrea, Frustulia rhomboides, multiple Eunotia species, and several Pinnularia species (Charles, 1985). The benthic assemblage is typical of a low-alkalinity shallow lake with abundant aquatic plants or a wetland, as suggested by the lithostratigraphy. In the middle part of zone C (1.345–1.045 m, zone C2; c. 3200–c. 1500 cal yr BP) benthic diatom diversity and abundance (<30%) declines and A. granulata increases (>70%) (Fig. 6), suggesting somewhat reduced water clarity and increased water depth. In the top of zone C (1.045–0.895; c. 1500–c. 1400 cal yr BP), diatom concentrations are low and planktonic diatoms decrease relative to benthic diatoms. The low diatom concentration may reflect increased clastic inputs in runoff, consistent with the higher mineral content in the sediments.

In the uppermost sediments (<0.895 m, diatom zone D; <1400 cal yr BP) diatom abundance is very low (Fig. 6). In contrast, in the contemporary lake, living diatoms are abundant on both plants and mud (Inthongkaew, unpublished data), which suggests that the absence of diatoms in the uppermost sediments is a result of dissolution of biogenic silica. This hypothesis is corroborated by the absence of phytoliths in the correlative unit of cores taken by prior researchers (Kealhofer and Penny, 1998). The processes producing silica dissolution again are unclear.


Although the availability of multiple sediment sequences in the deeper part of the lake and along its eastern shore (Fig. 1B) would...
offer an excellent opportunity for detailed lithostratigraphic correlations, the sediment descriptions provided in Kealhofer and Penny (1998) and Penny (1998) only allow for general comparisons to our new cores. We illustrate a possible correlation between Kealhofer and Penny’s (1998) and Penny’s (1998) sediment sequences and CP3A along a southwest–northeast transect in Fig. 7. The bottommost fine sand and sandy clay sediments, which were reached in sediment cores KUM.3, KUM.2, KUM.1, and KUM.9 (Kealhofer and Penny, 1998; Penny, 1998) and dated to ~ >10,500 cal yr BP, could not be obtained in CP3A. The transect, however, shows that the surface of these sediments is very irregular, indicating large variations in lake bottom topography (Fig. 7). The gray gyttja clay of sedimentary unit 1 in CP3A correlates with the loam, clay and silt in KUM.3, with the clay to silty clay in KUM.2 (Penny, 1998) and the clay/silt in KUM.1 and KUM.9. This correlation is supported by 14C ages, which estimate the deposition of these sediments to between ~10,000 and 7000 cal yr BP (Kealhofer and Penny, 1998; Penny, 1998). The transect suggests that these layers drape the underlying sediments and fill the deeper parts of the basin (Fig. 7). The succession of organic-rich layers between 2.16 and 1.18 m in CP3A, which were assigned to sedimentary units 2a and 2b (6700–1600 cal yr BP), compares well to the peat layers described for KUM.3, KUM.2 and KUM.1 (Kealhofer and Penny, 1998;
Penny, 1998) and is also likely correlative to the black organic clay and silt in KUM.9 (Penny, 1998). The upper boundary of these organic sediments coincides with a hiatus of approximately 1500 years in CP3A, but is dated to c. 2100 cal yr BP in KUM.2 and to c. 3000 cal yr BP in KUM.3. Given that all 14C dates for KUM.2 and also the uppermost 14C date for KUM.3 were made on pollen concentrates, this age difference would suggest that peat formation ended earlier in KUM.3 and that the southern part of the basin is slightly deeper (Figs. 1B, 7). The hiatus observed in CP3A, at or just below the boundary of unit 3a, could on the other hand suggest erosion of the upper part of the peat in KUM.3. The peat layers in the three sequences KUM.3, KUM.2 and CP3A are overlain by clay gyttja and gyttja sediments, which indicate a rise in lake level. In contrast, peat formation continued at KUM.1 and KUM.9, sites situated closer to the shore of the present lake (Fig. 7).

4.5. Paleoclimatic and paleoenvironmental interpretation

Kealhofer and Penny (1998) and Penny (1998) demonstrated that sand and clay layers accumulated in the Kumphawapi basin before ~10,500 cal yr BP and suggested deposition in a floodplain or back swamp environment (Table 1). The inference of arid climatic conditions was based on the reconstructed species-poor and strongly seasonal vegetation that surrounded the site. By c. 10,000–9400 cal yr BP a lake had formed in the basin, as documented by the proxies analyzed in CP3A. The shallow freshwater lake, where terrestrial and aquatic organic material accumulated together with clastic sediments derived from the catchment, was characterized by low light conditions, high turbulence in the water column, and oxygenated bottom water. This suggests high run-off and consequently higher moisture availability (Fig. 8). Pollen and phytolith assemblages provide evidence for the establishment of a mosaic of open dry-land vegetation, open woodlands, and dry/mixed deciduous forests between ~10,500 and ~9000 cal yr BP (Table 1), which has been interpreted as a change toward more humid climatic conditions (Kealhofer and Penny, 1998; Penny, 1998). This assumption compares well to the proxy record from CP3A.

Gradually increasing lake organic productivity between 6700 and 5900 cal yr BP and less oxic bottom water conditions suggest decreased catchment run-off (Fig. 8). This development could have been a response to denser vegetation around the lake, but could also have been initiated by a decrease in effective moisture. The subsequent rapidly increasing organic content of the sediments, reduced bottom water conditions, and a decrease/increase in planktonic/benthic diatoms, indicate a distinct lowering of the water level in the already shallow lake. The fairly abrupt change in lake status at c. 5900 cal yr BP indicated by multiple proxies suggests that this shift was not a simple response to basin infilling, but was caused by a change to lower effective moisture. By 5400 cal yr BP an extensive wetland with predominantly terrestrial vegetation had become established. According to the chronology of Kealhofer and Penny (1998) and Penny (1998), lowering of the lake level commenced already ~6800 cal yr BP, coincident with a significant reduction in dry-land taxa (Table 1). This is approximately 1000 years earlier than the water level lowering evidenced in our proxy records. Given that most of Kealhofer and Penny’s (1998) and Penny’s (1998; 1999) 14C dates for KUM.3 were made on bulk sediment (Table 2), this older age could be explained by incorporation of reworked organic material.

The transition from wetland to peatland coincides with a hiatus, or alternatively with a change to very low peat accumulation starting around 5200 cal yr BP, as suggested by the age model (Figs. 3B, 8). Both a hiatus and/or slow peat growth would indicate no accumulation or very low accumulation rates, which in turn could be interpreted in terms of low effective moisture availability. The peatland persisted until about 3200 cal yr BP and seems to have been subject to periodic desiccation. The change from peatland to wetland around 3200 cal yr BP, indicated by the lithological shift from peat to peaty gyttja and by a decrease in organic matter content and an increase in planktonic diatoms, would suggest a rise in water level and flooding of the peat surface. According to the age model, deposition of the peaty gyttja would have occurred between 3200 and 1600 cal yr BP (Fig. 3, Table 3), which implies very slow sediment deposition, or a hiatus. The high number of soil mineral particles between c. 2700 and 2500 cal yr BP, large amounts of charcoal between 1900 and 1600 cal yr BP, and the sharp lithological transition between units 2b and 3a would argue for hiatuses, possibly due to surface dryness. This in turn would indicate at least two intervals of reduced effective moisture availability, one around 2700–2500 cal yr BP, and one around 1900–1600 cal yr BP (Fig. 8). The latter coincides with the transition to the clay gyttjas of unit 3a, which document the establishment of a shallow lake. Most of the northern and northeast parts of the basin (KUM.1, KUM.6 and

![Fig. 8. Reconstructed lake level, run-off and moisture history for Kumphawapi. The different settlement periods at Ban Chiang are according to Pietrusewsky and Douglas (2002) and White (2008).](image-url)
Kumphawapi's present outlet toward the south runs through a several kilometer-wide, low-relief area (Clubseng, 2011). Damming of the outlet at its narrowest point would have involved construction of a 0.5 to 1 km long dam by the former settlers. Geoarchaeological investigations in the Upper Mun River valley, some 250 km to the southeast of Kumphawapi, have shown that intensified landscape management and construction of drainage channels became common during the Iron Age (c. 2500–1500 cal yr BP) due to limited water availability (Boyd, 2008). On the other hand, the well-known Iron Age settlement of Ban Chiang had been abandoned around 1800 cal yr BP (Pietrusewsky and Douglas, 2002; White, 2008), which predates the higher lake level in Kumphawapi. The construction of a large dam to raise water levels in Kumphawapi seems therefore less probable. Uplift of the salt dome, which forms the small island of Ban Don Kao to the south of the lake (Figs. 1B, 2), could have played a role in blocking the outlet. Active salt dome uplift due to regional tectonic compression or buoyancy has been reported at rates varying between an extreme of 82 mm yr$^{-1}$ and 2 mm yr$^{-1}$ (Davison, 2009). Given these uplift rates, Ban Don Kao could theoretically have been uplifted by between 33 and 0.8 m during the c. 400 years, which cover the upper hiatus. However, on-going uplift for Ban Don Kao has not been described, although many topographic features on the Khorat Plateau are explained by the formation of salt domes and by the corrosion/dissolution of salt beds of the Maha Sarakham Formation (Hisao and Wichaidit, 1989; Malilla, 2001). Moreover, a gradual uplift of the salt dome would likely not have caused a hiatus, but would show up as a fairly gradual transition from peaty gyttja to clay gyttja. Subsidence of the Kumphawapi basin, due to the dissolution of underlying salt beds, could be another explanation for the rise in water level. However, none of the analyzed proxies in the sediments of CP3A indicates a change in lake water salinity, which would be expected as a result of salt bed dissolution. Rising groundwater levels and higher effective moisture is another explanation for the gradual increase in lake level. Given that anthropogenic impact and salt-related processes seem less probable, we hypothesize that climate-induced changes led to higher water levels in Kumphawapi.

Kumphawapi's lake phase continued up to the present, but was interrupted by a change in lake status between c. 1000 and 600 cal yr BP (Fig. 8). The distinct increase in sediment organic matter content could have been caused by intensified agriculture in the catchment (i.e. deforestation, construction of irrigation channels), which led to increased nutrient flux to the lake. Boundary stones from the island of Ban Don Kao (Fig. 1) date the presence of settlements to 800 AD (1200 cal yr BP) (Penny, 1998), which compares approximately to the interval of higher organic matter content. Another explanation could be expansion of the shore vegetation due to a lowering of the water level and lower effective moisture availability, which could have increased the organic matter content in the shallow lake. In addition, it could be speculated that a stronger human impact on the lake's catchment could have been a result of a change in climatic conditions.

4.6. Is Lake Kumphawapi recording local or regional past climatic changes?

Given the scarcity of paleo-precipitation records from tropical lakes in Southeast Asia, it is important to examine whether Lake Kumphawapi can be added as an archive of tropical climate change or whether the environmental signals stored in its sediments are records of local catchment processes and/or anthropogenic impact. We therefore compare our environmental reconstruction to terrestrial records from the Asian monsoon region (Fig. 9).

Geoarchaeological investigations in the Upper Mun River valley to the southeast of Kumphawapi suggest that drier climate conditions starting around 3500 cal yr BP led to decreased water availability, to gradually intensified landscape management, and to the final abandonment of Iron Age settlements after 1500 cal yr BP (Boyd, 2008). This observation is in line with the general view of a mid-Holocene decline in Asian monsoon strength (Morill et al., 2003; Wang et al., 2005a), but diverges from the record of Lake Kara in northern Cambodia (Maxwell, 2001). The data from this site suggest higher effective moisture c. 9500–6200 cal yr BP and during the past 2700 cal yr BP, and lower effective moisture c. 6200–2700 cal yr BP (Maxwell, 2001). These trends compare fairly well to the lake-level history of Kumphawapi (Fig. 9). Variable lake levels are also reported for Lake Tonle Sap since the mid-Holocene, but these are explained by decreased rainfall and increased seasonality (Penny, 2006).

Paleoclimate reconstructions based on the record from Lake Huguangyan in southern China (Wang et al., 2007) indicate high effective moisture between 9500 and 8000 cal yr BP, lower effective moisture between 7800 and 4200 cal yr BP, and distinctly drier conditions between 4200 and 350 cal yr BP (Fig. 9). The Dongge cave stalagmite $^6$H record suggests that the East Asian summer monsoon was stronger between 9000 and 5600 cal yr BP, and then declined in a step-wise sequence with marked shifts at 5600 and 3500 cal yr.
Superimposed on a number of short-term events when the monsoon was distinctly weaker; these were centered at around 8300, 7200, 6300, 5500, 4400, 2700, 1600, and 500 cal yr BP (Wang et al., 2005b). Comparisons between speleothem δ¹⁸O records from southern and central China moreover suggest that East Asian summer monsoon precipitation decreased asynchronously from south to north, i.e., starting around 7000 cal yr BP in Dongge cave and around 4500 cal yr BP in Jixuan cave (Cai et al., 2010). These findings contrast with those of An et al. (2000), who had suggested that the Holocene precipitation maximum reached southern China as late as around 3000 cal yr BP, and also diverge from the findings of Zhou et al. (2005; 2007) and Zhang et al. (2011), who show that Holocene climate was broadly synchronous across the monsoon region. Moreover, syntheses of paleorecords from central Asia suggest that the Indian Ocean and East Asian summer monsoon behaved in an asynchronous way during much of the Holocene (Herzschuh, 2006; Wang et al., 2010). The wettest interval occurred between 10,900 and 7000 cal yr BP, with continued moderately wet conditions until 4400 cal yr BP for sites influenced by the Indian Ocean monsoon. In contrast, sites influenced by the East Asian summer monsoon show generally wet conditions between 11,500 and 1700 cal yr BP, with the wettest period between 8300 and 5500 cal yr BP. A stronger early Holocene summer monsoon is in line with the moisture history reconstructed for Kumphawapi, although the decline in monsoon strength seems to have been registered earlier in Lake Huguangyan in southern China than in northeast Thailand and in Cambodia (Fig. 9). The early Holocene short-term events of a weaker summer monsoon evidenced in the Dongge record (Wang et al., 2005b) are not seen in Kumphawapi, but the timing of late Holocene weak monsoon events at 2700, 1600 and 500 cal yr BP corresponds approximately in time to the hiatuses in the Kumphawapi record, and to the assumed low lake level phases (Figs. 8, 9). The timing of lower effective moisture availability reconstructed for Lake Huguangyan between 8000 and 4200 cal yr BP and the subsequent low moisture availability only partly compare to the environmental history of Kumphawapi.

Paleorecords for the Indian Ocean monsoon region (Fig. 9) show high lake levels and increased precipitation between approximately 7200 and 6000 cal yr BP, and an onset of aridity around 5300 cal yr BP in NW India (Prasad and Enzel, 2005; Singhvi and Kale, 2008). High lake levels and higher moisture availability have been reported for Lake Sambhar between 9600 and 7500 cal yr BP, and between 6800 and 2500 cal yr BP, while lower lake levels and decreased precipitation are reconstructed between 7500 and 6800 cal yr BP and during the past 2500 years (Sinha et al., 2006) (Fig. 9). Pokhara Playa, also in the Thar Desert of Northwest India registered a high rainfall regime between 4000 and 2300 cal yr BP and low rainfall conditions between 2300 and 1100 cal yr BP (Roy et al., 2009). The rainfall history reconstructed from the two Thar Desert lakes (Fig. 9) compares partially to that observed in Siddha Baba Cave in Nepal (Denniston et al., 2000), where arid conditions are observed between 2300 and 1500 cal yr BP. However the increase in summer monsoon precipitation around c. 500 cal yr BP (1550–1640 AD) in Nepal (Fig. 9) and the humid phase seen around 1000 cal yr BP (peaking 500–750 cal yr BP) in northern Indian cave deposits (Yadava and Ramesh, 2005) are not registered in the desert lakes. These latter changes also do not compare in time to the decadal to centennial long intervals of a stronger monsoon between 950 and 1200 AD (c. 1000–800 cal yr BP) and the series of decadal droughts between 1250 and 1450 AD (c. 800–600 cal yr BP) (Buckley et al., 2007; 2010; Sinha et al., 2007, 2011; Cosford et al., 2008; Cook et al., 2010) reconstructed from speleothem and tree-ring records in Southeast Asia. The changes in Kumphawapi’s lake level and inferred effective moisture availability are not comparable to the above-mentioned Indian Ocean monsoon paleoclimate records. The decadal droughts between 1250 and 1450 AD (c. 800–600 cal yr BP) seen in tree-ring and speleothem archives, however, overlap approximately with the assumed lake level lowering in Kumphawapi between 1000 and 600 cal yr BP.

The discrepancy between the inferred lake-level changes in Kumphawapi and moisture records from the Indian and East Asian monsoon regions (Fig. 9) highlights the problem of interpreting environmental proxies in lake sediments in terms of hydroclimate, and the difficulties in correlating moisture histories over large geographical distances. Each lake basin had and has its own specific setting, different lakes had different threshold responses to changing climatic conditions, and different paleoproxies provide a range of possible environmental responses. In contrast to temperature, precipitation is spatially heterogeneous; annual and rainy season precipitation totals from the Asian monsoon region can, for example, only be compared at distances of around 500 km (Dayem et al., 2010), which would mean that a spatially dense network of well-dated sites is needed for southeast Asia to track changes in monsoon intensity in greater detail. Such a network already exists for the past 1000 years based on tree-ring series (Cook et al., 2010). Any conclusions regarding asynchronous/synchronous changes between the different regions remain premature until more well-dated records have become available.

5. Conclusions

Sediment geochemistry combined with diatom assemblage changes provides a detailed record of past environmental changes in northeast Thailand. Nong (Lake) Han Kumphawapi formed around 10,000–9400 cal yr BP as a result of higher run-off and higher moisture availability due to a stronger summer monsoon. The sediments deposited between c. 9400 and c. 6700 cal yr BP indicate a well mixed water column and oxygenated bottom waters. A decrease in run-off and the concomitant increase in lake organic productivity between c. 6700 and 5900 cal yr BP suggests the development of dense shore vegetation or alternatively a decrease in effective moisture. Multiple proxies indicate a marked lowering of the already shallow lake around 5900 cal yr BP and the development of an extensive wetland around 5400 cal yr BP. The fairly abrupt change in lake status suggests that this shift was caused by lower effective moisture. The subsequent transition to a peatland coincides with a hiatus, which indicates very low accumulation rates due to very dry climatic conditions. The re-establishment of a wetland around 3200 cal yr BP is interpreted as a rise in groundwater and lake level, indicative of slightly higher effective moisture. However, the sediments deposited between c. 3200 and 1600 cal yr BP provide evidence for at least two hiatuses at c. 2700–2500 cal yr BP, and at c. 1900–1600 cal yr BP, respectively. These hiatuses suggest surface dryness and consequently intervals of low effective moisture. Around 1600 cal yr BP a new shallow lake became re-established in the basin, as documented by the sediment lithology and various other proxies. Although the underlying causes for this new lake phase remain unclear, we hypothesize that higher effective moisture was the main driving force. This shallow lake phase continued up to the present, but was interrupted by higher nutrient flux to the lake around 1000–600 cal yr BP. Whether this was caused by intensified agriculture in the catchment or, whether this signals a lowering of the lake level and exposure of the shore vegetation due to reduced effective moisture, needs to be corroborated by further studies in the region.

The lake level changes and inferred effective moisture history reconstructed for Kumphawapi compare well to the general observation of a strong Asian summer monsoon during the early Holocene and a weakening of the monsoon during the mid-Holocene. The assumption of higher effective moisture during the later part of the Holocene is in line with reconstructions based on lake sediments from Cambodia and also with precipitation changes inferred from the Dongge Cave speleothems, but contrasts with other reconstructions, especially from the Indian subcontinent. Whether this is due to coupling/decoupling between the Indian Ocean and the East Asian monsoon...
systems, or due to local influences on the archive and/or dependent on the choice of the studied proxies, needs to be evaluated by a much denser network of hydroclimatic reconstructions.

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