

Societal response to monsoonal fluctuations in NE Thailand during the demise of Angkor Civilisation

The Holocene
2017, Vol. 27(10) 1455–1464
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DOI: 10.1177/0959683617693900
journals.sagepub.com/home/hol
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Abstract

This paper investigates the possible social responses to changes in the strength of the southwest monsoon in northeastern Thailand during the currency of the Angkor civilisation. These assessments are based on hydrogen and carbon isotope records of leaf waxes (δD_{wax} and $\delta^{13}C_{wax}$) from a 2000-year-long wetland sequence of Pa Kho in northeastern Thailand, a region that formed the northern boundary of the Angkor Kingdom. Our data indicate anthropogenic flooding of the Pa Kho wetland through the control of water through dam construction from c. AD 1300 in response to the fluctuating strength of monsoon rains. δD_{wax} , a proxy for regional hydroclimate variability, corroborates pre-existing evidence that increased summer monsoon rains, which supported the expansion of the agrarian economy, aided the rise of the Angkorian Empire whereas extreme drought contributed to its demise. Interestingly, our δD_{wax} record shows already a gradual decreasing monsoon intensity from c. AD 1000 onwards, although Angkor's prosperity reached its peak at c. AD 1200. We suggest that the complex hydrological system established under royal patronage at Angkor provided a resilient buffer against short-term monsoon fluctuations. The long-term decline in monsoon rains over a ~300-year period, combined with ongoing urbanisation, may have stretched the hydrological systems to their limit. We suggest that this was a major factor that contributed to the demise of Angkor in the mid-15th century.

Keywords

$\delta^{13}C_{wax}$, δD_{wax} , Angkor Civilisation, Lake Pa Kho, northeast Thailand, Southeast Asia, wetland

Received 30 September 2016; revised manuscript accepted 9 January 2017

Introduction

Changes in the Earth's climate have been shown to contribute to human migrations and social changes (Haug et al., 2003; Weiss and Bradley, 2001; Wohlfarth et al., 2016). The history of the civilisation of Angkor is a prominent example of how a technically developed society confronted dramatic shifts in climate conditions around c. AD 1400 (Buckley et al., 2010; Cook et al., 2010; Penny et al., 2007). Encompassing Cambodia and parts of southern Vietnam, Laos and Thailand, the Angkor Civilisation, from the 9th to the 14th centuries AD, was the greatest kingdom in Southeast Asia (Evans et al., 2007). Angkor's economic basis (Hall, 1985) was rice agriculture (Castillo, 2011; Higham, 2015), which was supported by a complex hydrologic system for irrigation during dry seasons (Kummu, 2009).

The cultivation of domestic rice was introduced into the area that was later to come under the political control of Angkor in the first half of the 2nd millennium BC (Higham et al., 2015). Rice was a component of a mixed economy that included the maintenance of domestic cattle and pigs, hunting gathering and fishing, until about AD 200. As far as can be identified from the surviving plant remains, rice was grown as a dryland crop, reliant on rainfall alone for maturation. It was probably not a major component of the human diet (Higham, 2002). From AD 200, however, there was a marked decline in the intensity of the monsoon rains

(Wohlfarth et al., 2016). This took place at the same time as a series of cultural changes that have been described as an agricultural revolution (Higham, 2014).

Iron Age sites proliferated with a growing population, and were now ringed by multiple broad moats and banks to store the water diverted from the nearby watercourses (Scott and O'Reilly, 2015). Indeed, remote sensing has identified possible Iron Age fixed field boundaries surrounding settlements in Northeast Cambodia, on the doorstep of Angkor (Evans et al., 2013; Hawken, 2011). Weeds adapted to wet field systems have been identified at

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Non Ban Jak (Castillo, 2011), and finally, rice itself proliferated in archaeological contexts. It was used in mortuary rituals to fill graves, and rice straw was mixed with clay to construct houses (Higham and Kijngam, 2012).

Wohlfarth et al. (2016) have proposed that the deterioration in the vital monsoon rains stimulated a positive adaptive response involving the establishment of irrigated agriculture. This was a seminal change in the social history of Southeast Asia because irrigated demarcated fields represent an improvement of land that was under personal or family ownership, thereby opening the door to wealth creation and social ranking. Again, evidence of social organisation is precisely what occurred in the archaeological record, the construction of the moats coincided with the interment of the dead with an unprecedented level of wealth expressed as exotic valuables: gold, silver, bronze, carnelian, glass and agate (Higham, 2011). Subsequently, there was a rapid transition from these late Iron Age societies into early states, known under the name Chenla in the area of the later Angkorian kingdom. The textual record between c. AD 550 and c. AD 800 that survives as Sanskrit and Old Khmer inscriptions records elite individuals, known as *pon*, rice field workers, smiths and construction workers all of whom recall developments from the social milieu of the late Iron Age (Jacob, 1979; Vickery, 1998). Hawken (2011) has identified that rice field boundaries matched the epigraphic evidence for the basis of Chenla agriculture.

In about AD 802, a ruler named Jayavarman II defeated the leaders of rival Chenla states, and established the kingdom of Angkor on the northern shore of the Tonle Sap, the Great Lake of Northwest Cambodia. From the foundation of the Kingdom, the diversion of water into reservoirs (*barays*) and thence into rice fields was the basis of state agriculture. This is seen, for example, in the construction of the Indratataka Baray at Hariharalaya under King Indravarman I (reigned c. AD 877–889), soon to be followed by the giant Eastern Baray under Yasovarman I (reigned c. AD 889–910) at Angkor itself. Recent investigations have identified the complex inlet and outlet systems of this reservoir, as well as the pattern of associated rice fields (Pottier, 2001). The Western Baray was the biggest of all and was constructed during the reign of Suryavarman I (reigned c. AD 1006–1050). Pottier (2000) has identified a complex series of outlets that fed water into a canal system and ultimately into the rice fields.

This agricultural system involved exploiting the perennial streams that flowed south from the Kulen upland to the north of Angkor. One was diverted and canalised to feed the larger *barays*, as well as the moats that ringed the temple mausoleum of Angkor Wat, and the canals and moats associated with the city of Angkor Thom (Fletcher et al., 2008). The entire water reticulation system of Angkor was probably built to develop resilience to any fluctuations in monsoon rains. When it faltered, over long periods of time, there would have been stress on the water supply to the populace and the rice fields. Exacerbated by deforestation in the hinterland as the population grew and spread, unusually wet conditions would have caused major problems with sedimentation and the choking of the canal system (Day et al., 2012; Kummu et al., 2008).

In this study, we investigate the effect of southwest monsoon dynamics on the inhabitants of the Khorat Plateau in NE Thailand, a region that formed part of the Angkorian Empire between c. AD 1000 and c. AD 1300 (Welch, 1998). For this, we use carbon and hydrogen isotopes of leaf waxes (δD_{wax} and $\delta^{13}C_{wax}$) derived from a 2000-year-long record of Lake Pa Kho (LPK), a wetland in NE Thailand. $\delta^{13}C_{wax}$ has been used as a proxy for both past vegetation and moisture availability on LPK and δD_{wax} for past regional hydroclimatic conditions in SE Asia, respectively (Yamoah et al., 2016a, 2016b). Together, these provide a unique context to (1) discuss human influence on the environment and (2) assess the impact of climate on Angkor during the late Iron Age through to the ensuing millennium.

Study site

A peat sequence was obtained in January 2010 from Nong (Lake) Pa Kho (LPK; 17°06'70" N 102°56'17" E) located on the Khorat Plateau at approximately 186 m a.s.l. in the northeastern Thailand, a region that formed the northern frontiers of the Angkorian Empire (Figure 1). A small and shallow lake of approximately 1 km² and <2 m water depth, respectively, LPK was formerly a wetland before dredging and damming in the 1990s (Chawchai et al., 2015; Penny, 2001). The vegetation in and around the wetland, prior to dredging, was dominated by grasses, sedges and ferns with floating communities, including *Typha angustifolia*, *Nelumbo nucifera*, *Persicaria attenuate*, *Alocasia macrorrhiza*, *Ipomoea aquatica*, *Ludwigia adscendens*, *Sagittaria sagittifolia*, *Jussiaea linifolia*, *Nepenthes thorei*, *Alternanthera sessilis*, *Eupatorium odoratum*, *Hewittia sublobata*, *Nymphoides indicum* and *Physalis angulata* (Penny, 2001). The current vegetation around LPK includes sugar cane, paddy fields and *Eucalyptus* plantations (Klubseang, 2011).

Thailand has a monsoonal hydroclimate, which consists of the southwest and northeast monsoonal trade winds. The southwest monsoonal winds arrive from the Indian Ocean, reaching Thailand between May and October, bringing most of the annual precipitation. The northeast monsoonal wind emanates from the high atmospheric pressure areas in Mongolia and China and brings dry air. This reaches Thailand during the winter period from November to April. The movement of the Intertropical Convergence Zone (ITCZ) (Thai Meteorological Department, 2011), El Niño–Southern Oscillation (ENSO) variability (Singhrattana et al., 2012; Yamoah et al., 2016a) and tropical cyclones (Thai Meteorological Department, 2011) contribute substantially to the variation of annual precipitation in Thailand. Mean annual precipitation in the study region is around 1455 mm, whereas air temperatures vary from ~22°C to 30°C (Chawchai et al., 2015).

Methods

Sampling and lithostratigraphy

The sequence from LPK was obtained in January 2010 using a modified Russian corer (1 m length, 10 cm diameter). The cores were described in the field, wrapped in plastic foil and placed in PVC (polyvinyl chloride) tubes and were transported to the Department of Geological Sciences at Stockholm University, where they were kept in a cold room at 4°C. The lithostratigraphy (Figure 2a) was described in more detail in the laboratory based on physical properties. Specific marker horizons were used to create a composite sequence based on overlapping core segments. The 20 AMS ¹⁴C dates measured on charcoal, seeds, leaves, insects and small wood fragments from 3 to 5 cm depth intervals (Chawchai et al., 2015) formed the basis for the Bayesian age models (Figure 2b and c). A detailed description of the chronology and the age modelling is given by Chawchai et al. (2015).

Biomarker analysis

Guided by the chrono- and lithostratigraphy (Figure 2a–c), which was established by Chawchai et al. (2015), samples for biomarker analyses were taken from 2.01 to 2.51 m (AD 1500–2000) at 5 cm intervals; from 2.51 to 2.81 m (AD 550–1500), continuous 1 cm samples were taken; and from 2.81 to 3.41 m (c. 150 BC to c. AD 550), samples were again taken at 5 cm intervals. A total of 52 samples were analysed. Details about the sample preparation and biomarker analyses are given in Yamoah et al. (2016a, 2016b). Briefly, the samples were ground to a fine powder prior to freeze drying, after which squalane was added acting as an internal hydrocarbon standard. The total lipid extract was fractionated by a solid phase chromatography procedure where the hydrocarbon

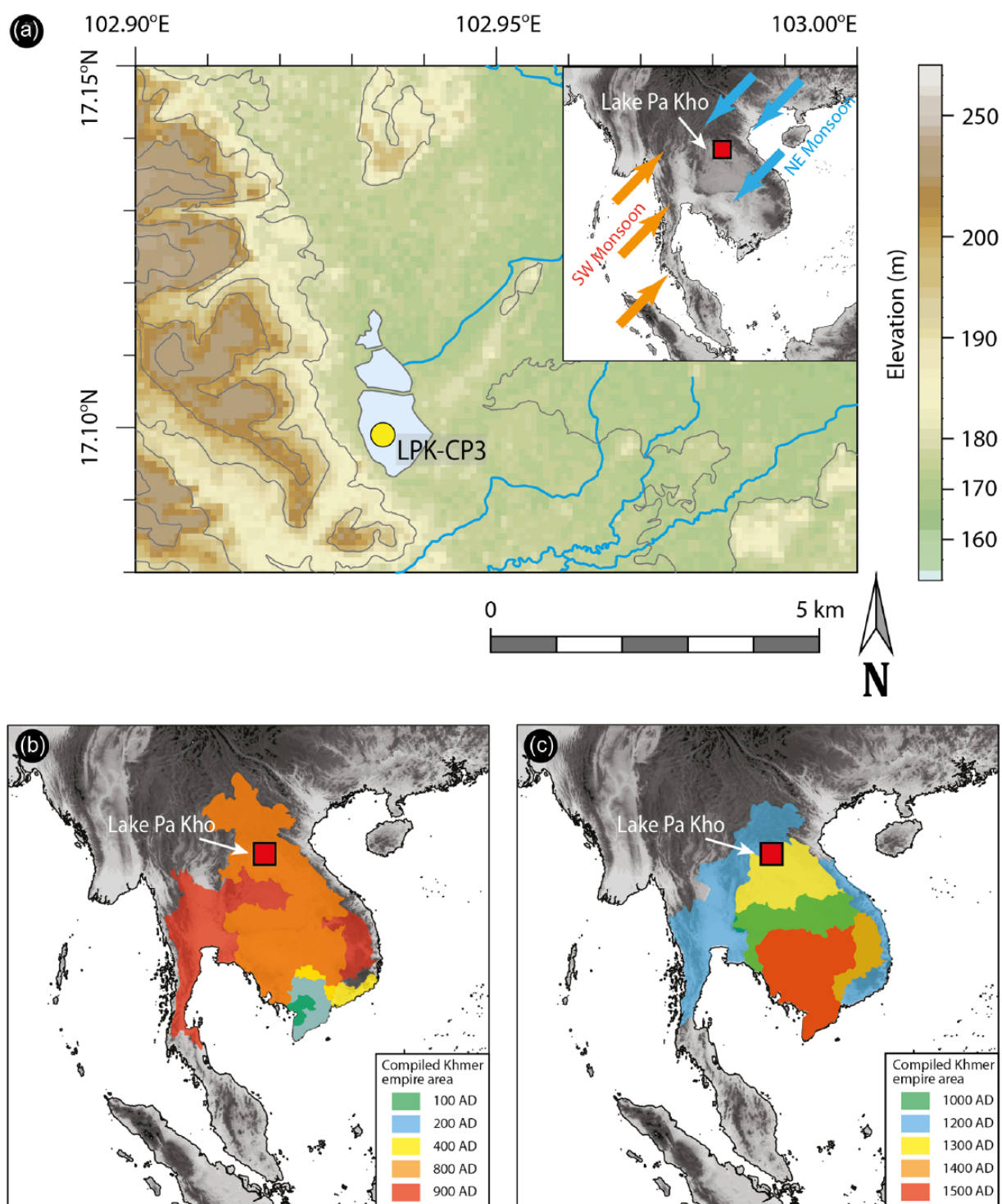


Figure 1. Map of the study site showing (a) the topography and the location of the retrieved sequence (yellow dot) on the Khorat Plateau and an insert of the map of Thailand, (b) the extent of Angkorian Civilisation between c. AD 100 and c. AD 900 and (c) also between the period from c. AD 1000 to c. AD 1500. (The map of the Angkorian Empire was modified from the TimeMap Project, University of Sydney, Australia. <http://www.timemap.net>). Readers are referred to the web version for the colour interpretation.

fraction was eluted with pure hexane over silica gel columns, while more polar fractions were subsequently recovered with more polar solvents. A saturated hydrocarbon fraction was obtained by eluting the hydrocarbon fraction over a pipette column filled with 10% AgNO₃-coated silica gel. Long-chain *n*-alkanes were quantified by gas chromatography–mass spectrometry (GC–MS) on a Shimadzu GCMS-QP2010 Ultra, equipped with an AOC-20i auto sampler and a split–splitless injector operated in splitless mode. A Zebtron ZB-5HT Inferno GC column (30 m × 0.25 mm × 0.25 μm) was used for separation.

$\delta^{13}\text{C}$ and δD analysis

The $\delta^{13}\text{C}$ and δD composition of the long-chain *n*-alkanes (C₂₇–C₃₅) was analysed by gas chromatography–isotope ratio monitoring–mass spectrometry (GC–IRMS) using a Thermo Finnigan Delta V mass spectrometer interfaced with a Thermo Trace GC 2000 using a GC Isolink II and Conflo IV system. Helium was used as a carrier gas at constant flow mode and the compounds were separated on a Zebtron ZB-5HT Inferno GC column (30 m × 0.25 mm × 0.25 μm). More details about the GC–IRMS method, including GC oven

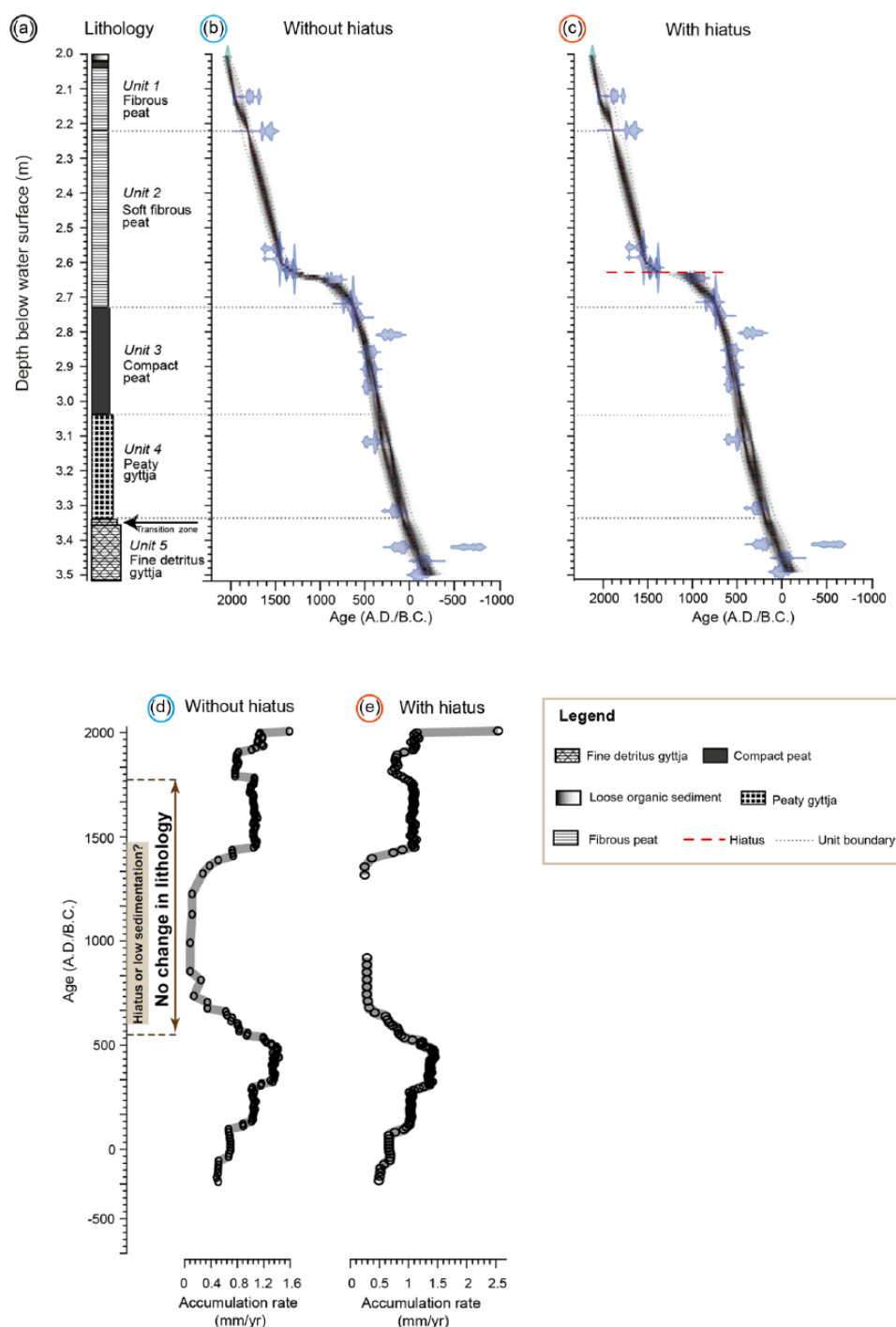


Figure 2. Chrono- and lithostratigraphy of the LPK sequence based on 20 AMS ^{14}C dates (Chawchai et al., 2015): (a) Lithostratigraphy, (b) age model (without hiatus), (c) age model (with hiatus), (d) accumulation rate (without hiatus) and (e) accumulation rate (with hiatus). The blue shapes show the calibrated ^{14}C dates with two standard deviations, the grey shading indicates the likely age model and the dotted lines show the 95% confidence ranges of the age model. Details of the stratigraphy and chronology are given in Chawchai et al., (2015).

temperature program, instrument performance and reference gases used are given in Yamoah et al. (2016a, 2016b). The average standard deviations for δD and $\delta^{13}\text{C}$ values of the n -alkanes were around 5‰ and 0.5‰, respectively.

Results and discussion

Age model and factors affecting sediment accumulation

This study relies on one of the two previous age models published by Chawchai et al. (2015) (Figure 2b and c), one with, and one without a hiatus. Chawchai et al. (2015) argued in preference

for the hiatus model between 2.68 m and 2.63 m depth based on enriched $\delta^{13}\text{C}_{\text{bulk}}$ values and a slowdown in accumulation rate despite plant remains suggesting moisture availability on the wetland. However, the introduction of a hiatus does not explain the slowdown in accumulation rate (Figure 2d and e) and the enriched $\delta^{13}\text{C}_{\text{bulk}}$ values between 2.68 m and 2.63 m depth. Neither is there any lithological evidence to support the presence of a hiatus. Indeed, a hiatus would imply a parched period with zero preservation of organic matter (OM). However, the hiatus was introduced between the wettest years, inferred from δD (Yamoah et al., 2016a) and plant macrofossils (Chawchai et al., 2015), and there are no arguments to suggest extreme dry periods or unconformity.

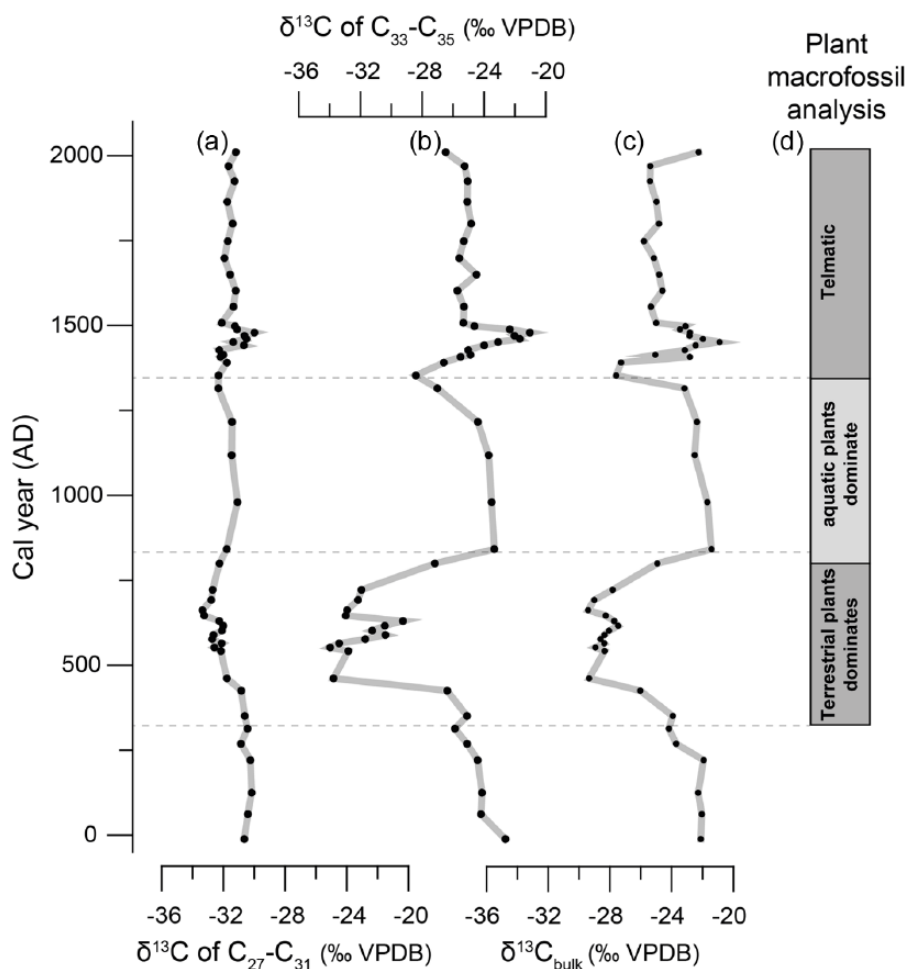


Figure 3. Age profiles of (a) $\delta^{13}\text{C}$ of $\text{C}_{27}\text{--C}_{31}$ n -alkanes, (b) $\delta^{13}\text{C}$ of $\text{C}_{33}\text{--C}_{35}$ n -alkanes, (c) $\delta^{13}\text{C}$ of bulk and (d) plant macrofossils. The average standard error of measurement is $\pm 0.5\text{‰}$ for the n -alkane homologues. The figure was modified from Chawchai et al., 2015 and Yamoah et al., 2016b.

Subsequently, Yamoah et al. (2016b) showed that the local wetland vegetation dominated the bulk of the total OM and that the variability of $\delta^{13}\text{C}_{\text{bulk}}$ appeared to be driven mainly by the input of wetland plants that highly varied in their carbon isotopic composition depending on the wetland wetness, where high (less negative) $\delta^{13}\text{C}$ values indicate wet conditions and dominance of aquatic plants over terrestrial vascular plants, as described further below. Extensive degradation of OM (Wüst, 2001), especially soft-tissued aquatic plants dominate between 2.68 and 2.63 m depth (as inferred from plant macrofossils) also explains the low accumulation rates and why there were no noticeable changes in the lithology (Figure 2a). Based on more in-depth understanding of the factors affecting accumulation rate, this study adopts the non-hiatus age model, as presented in Yamoah et al. (2016a, 2016b).

$\delta^{13}\text{C}$ values as a proxy for wetland wetness

According to an earlier study, higher plants form the primary source of OM input to the peaty deposits of LPK (Yamoah et al., 2016b). Based on biomarker and carbon isotope analyses, this source could be divided into two: terrestrial plants that generate predominantly $\text{C}_{27}\text{--C}_{31}$ n -alkanes (Figure 3a) and aquatic plants (macrophytes) producing mainly $\text{C}_{33}\text{--C}_{35}$ n -alkanes (Figure 3b). The $\delta^{13}\text{C}$ values of $\text{C}_{33}\text{--C}_{35}$ n -alkanes ($\delta^{13}\text{C}_{33-35}$) had a wide variability ($\sim -34\text{‰}$ to -22‰) mimicking the isotopic variability of bulk OM ($\delta^{13}\text{C}_{\text{bulk}}$) (Figure 3c). Yamoah et al., (2016b) argued against the typical explanation for $\delta^{13}\text{C}$ variability – shifts between C3 and C4 plant input – since the (expected) distribution of leaf wax derived n -alkanes does not show any changes, nor is there any evidence for nearly complete transitions between C3 and C4 plant input in the macrofossil record.

Indeed, nearly all the identified plant species by Penny (1998) and Chawchai et al. (2015) were mainly C3 plant types (e.g. *Eupatorium odoratum*, *Alocasia macrorrhiza*, *Ludwigia adscendens*, *Nymphaea lotus*, *Nelumbo nucifera*, *Typha angustifolia* and *Potamogeton* spp.).

Various lines of evidence based on n -alkane distributions, plant macrofossils and $\delta^{13}\text{C}$ values of leaf waxes led Yamoah et al. (2016b) to the interpretation that the variability observed both in $\delta^{13}\text{C}_{\text{bulk}}$ and $\delta^{13}\text{C}_{33-35}$ values was related to changes in water level and primary productivity, with wetter conditions leading to high $\delta^{13}\text{C}$ values. They argued that the presence of large amounts of aquatic plant remains at depths between 2.22 and 2.73 m (Chawchai et al., 2015) indicated a wet period that stimulated a high aquatic productivity. Increase in the aquatic productivity led to a drawdown of dissolved CO_2 available for (partially) submerged macrophytes, especially at higher water levels and low rates of air–water gas exchange due to dense vegetation. This drawdown leads to increased pH values, shifting the carbonate system away from dissolved CO_2 in favour of bicarbonate. The isotopic effect is double: the limitation of dissolved CO_2 not only leads to smaller isotopic fractionation (causing higher leaf wax $\delta^{13}\text{C}$ values), but certain aquatic macrophytes (e.g. *Potamogeton* spp.) may also start utilising bicarbonate as their carbon source, which has much higher $\delta^{13}\text{C}$ values than dissolved CO_2 (Yamoah et al., 2016b).

$\delta\text{D}_{\text{wax}}$ values as a proxy for SE Asian summer monsoon intensity

The hydrogen isotope value of leaf waxes ($\delta\text{D}_{\text{wax}}$) reflects the isotopic composition of the source water available to the plant, and the isotopic fractionation involved in biosynthetic processes of

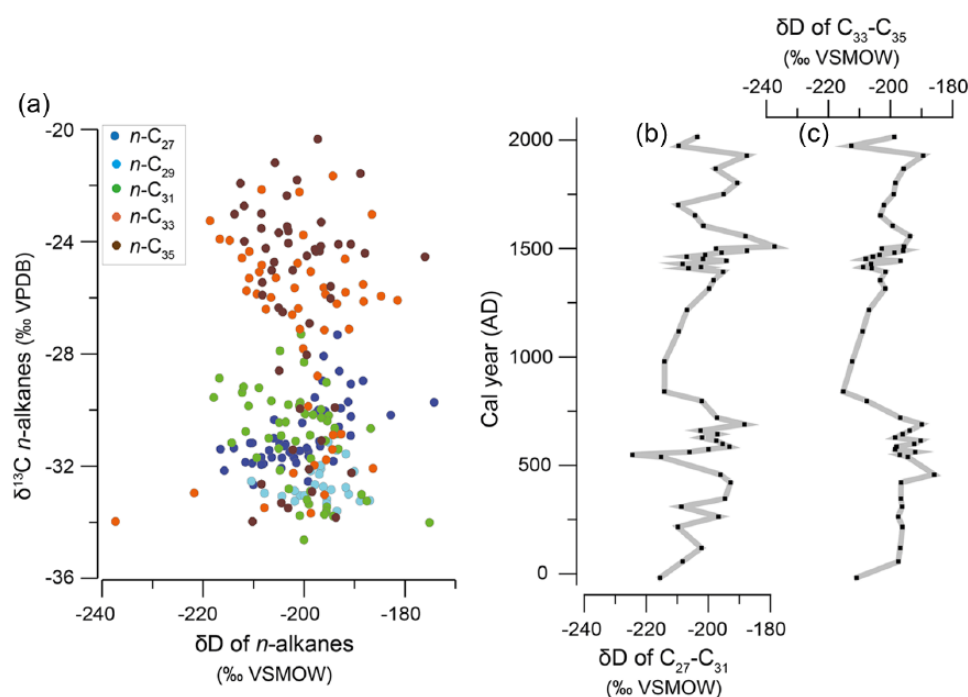


Figure 4. (a) Correlations between δD and $\delta^{13}\text{C}$ of $\text{C}_{27}\text{--C}_{35}$ *n*-alkanes, (b) age profile of weighted average of δD of $\text{C}_{27}\text{--C}_{31}$ and (c) age profile of weighted average of δD of $\text{C}_{33}\text{--C}_{35}$ *n*-alkanes. Error bars are omitted for clarity. The average standard error of measurement for δD is ± 5 ‰ (modified from Yamoah et al. (2016a)).

the plant (Sachse et al., 2004; Sauer et al., 2001; Sessions et al., 1999). Therefore, different plant types, due to variations in their biosynthetic fractionation, may influence $\delta\text{D}_{\text{wax}}$ interpretations (Sachse et al., 2012). As earlier argued by Yamoah et al. (2016a), $\delta^{13}\text{C}$ and δD values of $\text{C}_{27}\text{--C}_{35}$ *n*-alkanes from LPK show very weak internal covariance (Figure 4a), which suggests limited influence of changes in plant functional types to the recorded δD signal. Additionally, there are no significant differences in trends between the δD values of the various *n*-alkane homologues ($\text{C}_{27}\text{--C}_{35}$) (Figure 4b and c), indicating that both terrestrial and aquatic plants have been using the same source water and have undergone similar environmental and climatic stress. Thus, the source water, mainly derived from precipitation, is a major driver influencing alkane δD values. Factors influencing the isotopic ratio of precipitation in Thailand include the amount of precipitation, ENSO and the extent of evapotranspiration (Yamoah et al., 2016a). All these processes lead to more positive $\delta\text{D}_{\text{wax}}$ values during drier conditions and more negative values during wetter conditions.

Reconstructed hydroclimatic conditions for Thailand, based on $\delta\text{D}_{\text{wax}}$ values, suggest a decrease in summer monsoon rains between *c.* 50 BC and *c.* AD 450. From *c.* AD 500 to *c.* AD 700, the hydroclimate conditions display greater variability between wet and dry periods, but an overall trend towards wetter summers. This time interval was followed by a rapid increase in summer monsoon rains from *c.* AD 700 to *c.* AD 1000. $\delta\text{D}_{\text{wax}}$ values suggest that summer monsoon intensity started to decrease steadily until *c.* AD 1300 and was succeeded by greater variability between *c.* AD 1300 and *c.* AD 1500. The trend towards decreasing summer monsoon intensity reversed for around a century between *c.* AD 1550 and *c.* AD 1650 and then continued to decline until the start of the 20th century.

In order to understand the mechanisms underlying the reconstructed hydroclimatic changes in Thailand, Yamoah et al. (2016a) compared the $\delta\text{D}_{\text{wax}}$ record from LPK with other paleoclimate records from the broader tropical Pacific region (Figure 5). A coherent pattern of monsoonal changes was inferred from records of $\delta\text{D}_{\text{wax}}$ (Figure 5a) and $\delta\text{D}_{\text{botryococcene}}$ (Zhang et al., 2014) (Figure 5b) from Thailand and the Galapagos Islands, respectively. Over

the last millennium, a $\delta^{18}\text{O}_{\text{ostracode}}$ record (Figure 5c) from Dongdao Island, South China Sea (Yan et al., 2011), shows a general intensification of the summer monsoon rains similar to trends observed in the $\delta\text{D}_{\text{wax}}$ record (Figure 5d) retrieved from Lake Lading in East Java (Konecky et al., 2013). These records show opposite long-term trends compared with both the δD records from Thailand and the Galapagos Islands. Overall, the proxy records revealed a low-frequency centennial scale tripole pattern between LPK, the maritime West Pacific and the central-eastern Pacific where most of mainland Southeast Asia became wet during El Niño events and relatively dry during La Niña's (Yamoah et al., 2016a). This proxy record pattern mimicked composite climate data covering the strongest El Niño/La Niña events of the last 50 years (Yamoah et al., 2016a). Since (variations in) the $\delta\text{D}_{\text{wax}}$ record of LPK compares well with other hydroclimate proxy records from the Asia-Pacific region, it can be assumed to capture the long-term regional hydroclimate dynamics in Southeast Asia.

Evidence of anthropogenic activity

The $\delta^{13}\text{C}$ signal, modulated by moisture availability on the wetland, can be driven by either increasing monsoon intensity or anthropogenic activity such as damming (Yamoah et al., 2016b). The $\delta^{13}\text{C}_{\text{wax}}$ values from the lower part of the record, up until *c.* AD 1300, correspond generally with the $\delta\text{D}_{\text{wax}}$ record (Figure 6a and b). Low carbon isotope values correspond to dryer conditions and more vascular terrestrial plants input, whereas an increase in moisture availability corresponds to higher $\delta^{13}\text{C}$ values of the bulk and $\text{C}_{33}\text{--C}_{35}$ *n*-alkanes, and more aquatic plants (Figure 6a and b). This general correspondence between $\delta\text{D}_{\text{wax}}$ and $\delta^{13}\text{C}_{\text{wax}}$ values, however, breaks up in the upper part of the core from *c.* AD 1300 to AD 1500. We observe a switch of the natural isotope response of the wetland – where a decrease in rainfall results in lower $\delta^{13}\text{C}_{33\text{--}35}$ values to one where a decline in rainfall is suddenly accompanied by higher $\delta^{13}\text{C}_{33\text{--}35}$ values, suggesting wetter wetland conditions (Figure 4a and b). Similarly, relatively high C/N ratios, increased $\delta^{13}\text{C}_{\text{bulk}}$ values and the presence of

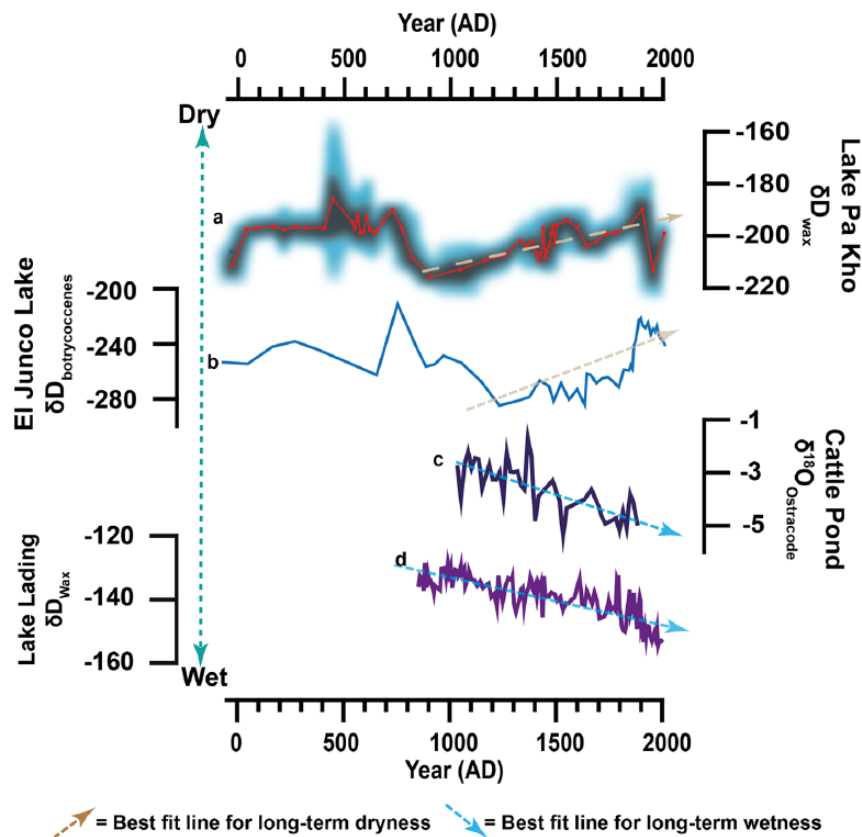


Figure 5. Comparison of hydroclimate proxy records from the tropical Asia-Pacific region. (a) LPK's δD_{wax} record (17°N, 102°E) with shadings indicating empirical 95% uncertainty bounds based on analytical and age model; (b) a botryococcene δD record from the Galapagos (4°N, 160°W) (Zhang et al., 2014); (c) the ostracode $\delta^{18}O$ record from Cattle Pond, Dongdao Island (16°N, 112°E) (Yan et al., 2011); (d) leaf wax δD records derived from Lake Lading, East Java (8°S, 118°E) (Konecky et al., 2013). The figure was modified from Yamoah et al., (2016a).

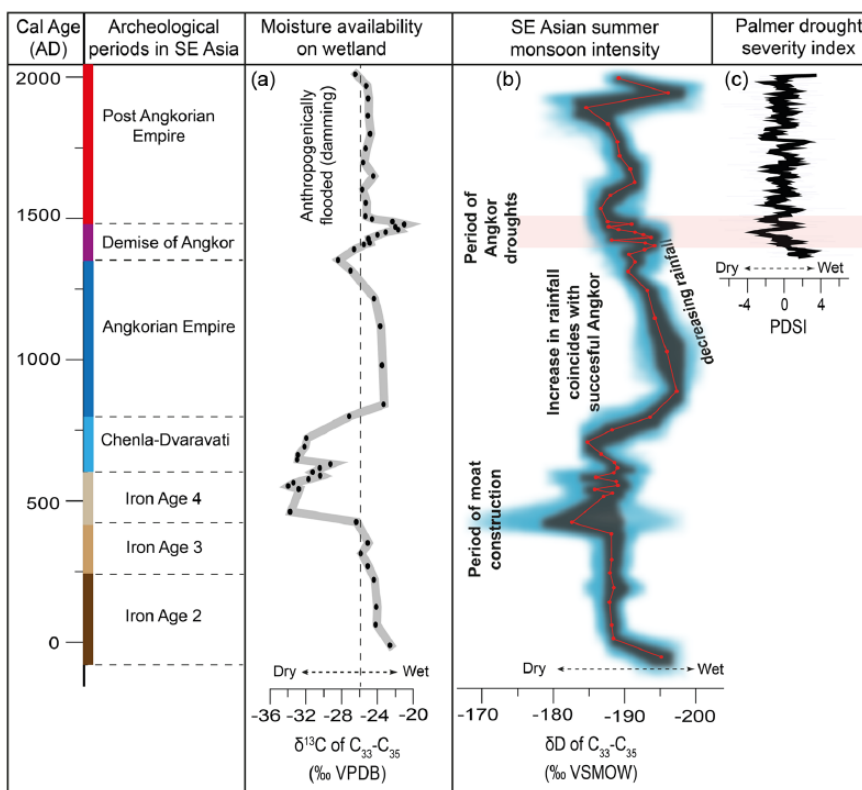


Figure 6. Archaeological periods in Northeast Thailand (Higham, 2014; Higham et al., 2015; Wohlfarth et al., 2016) compared to (a) $\delta^{13}C$ of $C_{33}-C_{35}$ n-alkanes (modified from Chawchai et al., 2015; Yamoah et al., 2016b), (b) δD of $C_{33}-C_{35}$ n-alkanes; shadings indicate the empirical 95% uncertainty bounds based on analytical and age-model errors (modified from Yamoah et al., 2016a), and (c) the Palmer Severity drought index (Buckley et al., 2010).

telmatic–aquatic plant material around this period have been interpreted as wetter wetland conditions in LPK (Chawchai et al., 2015).

From *c.* AD 1500 onwards, both $\delta^{13}\text{C}_{33-35}$ and $\delta^{13}\text{C}_{\text{bulk}}$ values remain relatively high and very stable. The presence of diatom species such as *Eunotia yanomami*, *Eunotia incisa*, *Eunotia intermedia*, *Eunotia monodon*, and *Gomphonema gracile* together with the plant macrofossil composition indicated a wetland environment (Chawchai et al., 2015). Meanwhile, $\delta\text{D}_{\text{wax}}$ continues to show much variability suggesting a decoupling between wetland response and hydroclimate variability in SE Asia. A probable explanation for this change would be anthropogenic flooding (damming) of the wetland. Under that condition, the carbon isotopic signal is mainly dictated by the availability of moisture on the wetland, which tends to drive aquatic productivity and CO_2 draw-down thereby leading to higher $\delta^{13}\text{C}$ values. The effect of damming the wetland during dry periods would result in an evaporative enrichment of the hydrogen isotopes of the source (lake) water. Continuous flooding of the wetland over a period would result in decoupling the $\delta^{13}\text{C}_{\text{wax}}$ values from the hydroclimate variability inferred from $\delta\text{D}_{\text{wax}}$ values. Evidence for anthropogenic activities in and around LPK may be gained from the gradual increase in biogenic silica, indicative of higher nutrient availability, from *c.* AD 1300 onwards due to either changes in land-use or intensification of agriculture around the lake (Chawchai et al., 2015). Human impact likely increased nutrient availability in LPK, which overprinted the climatic signal from *c.* AD 1300 onwards as observed in the decoupling of $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{wax}}$.

The agricultural revolution, driven by irrigated farming in NE Thailand (Higham, 2014) that necessitated large-scale infrastructure and substantial human impact on the landscape, started during late Bronze Age (Boyd and McGrath, 2001; Wohlfarth et al., 2016). However, isotopic evidence of anthropogenic activity in LPK is only seen from *c.* AD 1300 onwards, which coincides with just about the beginning of the demise of Angkor. As already elucidated, the $\delta^{13}\text{C}_{\text{wax}}$ signal is very localised and therefore extrapolating it to interpret the environmental conditions of the broader regional context of the Angkorian Empire might be farfetched. However, $\delta\text{D}_{\text{wax}}$ reflects the hydroclimatic conditions in Southeast Asia. Combined, the $\delta^{13}\text{C}$ and δD records suggest a human response to climatic changes, being the damming of LPK from *c.* AD 1300 onwards, likely necessitated by dwindling summer monsoon rains.

Effect of SE Asian summer monsoon dynamics on Angkor civilisation

Several climate hypotheses concerning the waxing and waning of the Kingdom of Angkor have been formulated (Buckley et al., 2010; Day et al., 2012; Lieberman and Buckley, 2012; Vickery, 1996). These have shown that an increase in summer monsoon rains (Lieberman and Buckley, 2012) contributed immensely to the expansion of the agrarian economic system of Angkor (Hall, 1985). Similarly, it has been argued that a decrease in summer monsoon rains and multiple extreme droughts contributed to their demise (Buckley et al., 2010, 2014; Cook et al., 2010; Day et al., 2012). Other researchers have rather attributed the demise of Angkor to the over-exploitation of the environment (Nitta, 1992).

Here, we discuss, within the context of hydrological changes, a new way of testing these pre-existing studies during major time events: pre-Angkorian (*c.* AD 500–800), Angkorian (*c.* AD 800–1430) and post-Angkorian (*c.* AD 1430–1900) (Figure 6). During the pre-Angkorian period, we observe a general pattern of continuous decrease in summer monsoon rains. The reduction in monsoon rains is suggested to have necessitated a strong social response to regulate water for irrigated agriculture by

expanding moat reservoirs (Wohlfarth et al., 2016) and in the process also stimulating wealth and social ranking (Higham, 2011). Indeed, this agricultural revolution (i.e. irrigated agriculture) may have laid the foundations for the success of the Angkorian Empire, which was reliant on irrigation to feed the extensive rice fields (Wohlfarth et al., 2016). Reconstructed rainfall intensity remained low until *c.* AD 700, but increased significantly between *c.* AD 700 and *c.* AD 1000. This period, which covered the latter part of the pre-Angkorian period and the early phase of the Khmer Empire, coincided with large-scale agriculture in NE Thailand (Boyd, 2008; Boyd and McGrath, 2001; Wohlfarth et al., 2016).

In about *c.* AD 800, King Jayavarman II united the Chenla states, through conquest, into the Angkor civilisation (Higham, 2014; Welch, 1989, 1998). The summer monsoon rains were still intense during this period, and indeed, it is not surprising that most of the notable barays were constructed within *c.* AD 877 and *c.* AD 1050. Water stored in the barays supported the agricultural base of the Angkor people leading to the massive production of rice all year round and sustaining a population that through a system of corvée labour, constructed the numerous Angkorian temples, roads and bridges (Tully, 2005). Summer monsoon rains declined gradually and steadily from *c.* AD 1000 onwards up to *c.* AD 1350, followed by greater variability until *c.* AD 1500. Interestingly, the period between *c.* AD 1350 and *c.* AD 1500, where we observe greater summer monsoon variability, coincides with the intervals of extreme droughts and wet periods suggested by Buckley et al. (2010) (Figure 6c). Due to the difference in their chronological resolutions, it is difficult to compare specific periods. However, the similarities between these intervals cannot be overlooked.

The summer monsoon rains were most intense around AD 1000 but started to decrease gradually thereafter, preceding the more extreme climatic conditions suggested by Buckley et al. (2010). The gradual decline in summer monsoon rains did not slow the expansion of the Angkorian Empire, which reached an apogee in about AD 1200. Presumably, this was in part due to the constructed barays, which were built to develop resilience to any fluctuations in the monsoon. We therefore suggest that the hydrological systems were not resilient enough to the long-term fluctuations in the summer monsoon rains, which pushed the kingdom closer to the boundary of its ecohydrological carrying capacity amidst ongoing urbanisation and expansion. This long-term weakening of the monsoon may have rendered the Angkorian society too vulnerable for the more extreme climate variability of the late 14th and 15th century (Buckley et al., 2010; Cook et al., 2010; Day et al., 2012).

Conclusion

Anthropogenic flooding and/or damming of LPK from *c.* AD 1300 onwards, amidst decreasing monsoon rains as revealed from the combination of $\delta\text{D}_{\text{wax}}$ and $\delta^{13}\text{C}_{\text{wax}}$ records, show a social response of the populace in adapting to increasing aridity. Our results corroborate pre-existing evidence of contemporaneous hydroclimate dynamics and the growth and collapse of Angkor (Buckley et al., 2010, 2014; Cook et al., 2010; Lieberman and Buckley, 2012). Within this framework, the expansion of the agrarian economy during the rise of the Angkorian Empire was likely aided by liberal amounts of steady summer monsoon rains. Subsequently, we observe from the $\delta\text{D}_{\text{wax}}$ records that summer monsoon rains had already begun to decrease gradually from *c.* AD 1000. The gradual decline in favourable conditions two centuries earlier while Angkor's prosperity reached its peak at *c.* AD 1200 could be due to the hydrological systems established by the Angkor people, which provided a buffer to the faltering summer monsoon rains and enabled their civilisation to be more resilient

against short- to medium-term climate fluctuations. However, the gradual dwindling of the monsoon rains over a 300-year period while urbanisation increased may have reduced the volume of water stored in the barays. The long-term weakening of the monsoon may have rendered the Angkor society too vulnerable for the more extreme climate variability of the late 14th and 15th century (Buckley et al., 2010; Cook et al., 2010; Day et al., 2012). These, possibly in conjunction with some sociopolitical factors such as warfare, may have contributed to the abandonment of Angkor and the establishment of a new capital centre to the east.

Acknowledgements

Special thanks go to Sherilyn Fritz, Ludvig Löwemark, Suda Inthongkaew, Wichuratree Klubseang, Jayne Rattray and Anna Hägglund for help in the field and in the laboratory.

Funding

Support from the Swedish Research Council (VR) research grants 621-2008-2855, 621-2011-4684 and 348-2008-6071 to Barbara Wohlfarth, and 621-521 2011-4916 to Rienk Smittenberg is acknowledged. The Delta Facility, funded by the faculty of Science, Stockholm University, is acknowledged for support with the isotope analyses.

References

- Boyd WE (2008) Social change in late-Holocene mainland SE Asia: A response to gradual climate change or a critical climatic event? *Quaternary International* 184: 11–23.
- Boyd WE and McGrath RJ (2001) The geoarchaeology of the prehistoric ditched sites of the upper Mae Nam Mun valley, NE Thailand, III: Late-Holocene vegetation history. *Palaeogeography, Palaeoclimatology, Palaeoecology* 171: 307–328.
- Buckley BM, Fletcher R, Wang S et al. (2014) Monsoon extremes and society over the past millennium on mainland Southeast Asia. *Quaternary Science Reviews* 95: 1–19.
- Buckley BM, Anchukaitis KJ, Penny D et al. (2010) Climate as a contributing factor in the demise of Angkor, Cambodia. *Proceedings of the National Academy of Sciences* 107: 6748–6752.
- Castillo C (2011) Rice in Thailand: The archaeobotanical contribution. *Rice* 4(3–4): 114–120.
- Chawchai S, Chabangborn A, Fritz S et al. (2015) Hydroclimatic shifts in northeast Thailand during the last two millennia – the record of Lake Pa Kho. *Quaternary Science Reviews* 111: 62–71.
- Cook ER, Anchukaitis K, Buckley BM et al. (2010) Asian monsoon failure and megadrought during the last millennium. *Science* 328: 486–489.
- Day MB, Hodell DA, Brenner M et al. (2012) Paleoenvironmental history of the West Baray, Angkor (Cambodia). *Proceedings of the National Academy of Sciences* 109: 1046–1051.
- Evans D, Pottier C, Fletcher R et al. (2007) A comprehensive archaeological map of the world's largest preindustrial settlement complex at Angkor, Cambodia. *Proceedings of the National Academy of Science* 104: 14277–14282.
- Evans DH, Fletcher RJ, Pottier C et al. (2013) Uncovering archaeological landscapes at Angkor using lidar. *Proceedings of the National Academy of Sciences* 110(31): 12595–12600.
- Fletcher R, Penny D, Evans D et al. (2008) The water management network of Angkor, Cambodia. *Antiquity* 82: 658–670.
- Hall KR (1985) *Maritime Trade and State Development in Early Southeast Asia*. Honolulu, HI: University of Hawaii Press, pp. 187–190.
- Haug GH, Gunther D, Peterson LC et al. (2003) Climate and the collapse of Maya Civilization. *Science* 299: 1731.
- Hawken S (2011) *Metropolis of ricefields: A topographic classification of a dispersed urban complex*. PhD Thesis, 2 vols, Department of Archaeology, University of Sydney.
- Higham CFW (2002) Languages and farming dispersals: Austroasiatic languages and rice cultivation. In: Bellwood P and Renfrew AC (eds) *Examining the Farming/Language Dispersal Hypothesis*. Cambridge: McDonald Institute for Archaeological Research, pp. 223–232.
- Higham CFW (2011) The Iron Age of the Mun Valley, Thailand. *The Antiquaries Journal* 91: 1–44.
- Higham CFW (2014) From the Iron Age to Angkor: New light on the origins of a state. *Antiquity* 88: 822–835.
- Higham CFW (2015) Debating a great site: Ban Non Wat and the wider prehistory of Southeast Asia. *Antiquity* 89: 1211–1220.
- Higham CFW and Kijngam A (eds) (2012) *The Origins of the Civilization of Angkor Volume 5: The Excavation Ban Non Wat: The Iron Age, Summary and Conclusions*. Bangkok: The Fine Arts Department of Thailand, 404 pp.
- Higham CFW, Douka K and Higham TF (2015) A new chronology for the bronze age of Northeastern Thailand and its implications for Southeast Asian Prehistory. *PLoS ONE* 10(9): e0137542.
- Jacob JM (1979) Pre-Angkor Cambodia: Evidence from the inscriptions concerning the common people and their environment. In: Smith RB and Watson W (eds) *Early South East Asia*. Oxford: Oxford University Press, pp. 406–26.
- Klubseang W (2011) *Paleogeography and paleoenvironment of Nong Han Kumphawapi, Changwat Udon Thani*. MSc Thesis, Chulalongkorn University, pp. 1–110.
- Konecky BL, Russell JM, Rodysill JR et al. (2013) Intensification of southwestern Indonesian rainfall over the past millennium. *Geophysical Research Letters* 40: 386–391.
- Kummu M (2009) Water management in Angkor: Human impacts on hydrology and sediment transportation. *Journal of Environmental Management* 90(3): 1413–1421.
- Kummu M, Penny D, Sarkkula J et al. (2008) Sediment: Curse or blessing for Tonle Sap Lake? *AMBIO: A Journal of the Human Environment* 37: 158–163.
- Lieberman V and Buckley B (2012) The impact of climate on Southeast Asia, circa 950–1820: New Findings. *Modern Asian Studies* 46: 1049–1096.
- Nitta E (1992) Iron-smelting and salt-making industries in Northeast Thailand. *Journal of Indo-Pacific Archaeology* 16: 153–160.
- Penny D (1998) *Late quaternary palaeoenvironments in the Sakon Na-Khon Basin, North-East Thailand*. PhD Thesis, Monash University, 260 pp.
- Penny D (2001) A 40,000-year palynological record from northeast Thailand; implications for biogeography and palaeo-environmental reconstruction. *Palaeogeography, Palaeoclimatology, Palaeoecology* 171: 97–128.
- Penny D, Kummu M, Barbeti M et al. (2007) Hydrological history of the West Baray, Angkor, revealed through palynological analysis of sediments from the West Mebon. *Bulletin de l'École Française d'Extrême-Orient* 92: 497–521.
- Pottier C (2000) Some evidence of an inter-relationship between hydraulic features and rice field patterns at Angkor during ancient times. *The Journal of Sophia Asian Studies* 18: 91–97.
- Pottier C (2001) Charles Higham, The civilization of Angkor. *Aséanie* 8: 167–176.
- Sachse D, Billault I, Bowen GJ et al. (2012) Molecular paleohydrology: Interpreting the hydrogen-isotopic composition of lipid biomarkers from photosynthesizing organisms. *Annual Review of Earth and Planetary Sciences* 40: 221–249.
- Sachse D, Radke J and Gleixner G (2004) Hydrogen isotope ratios of recent lacustrine sedimentary n-alkanes record modern climate variability. *Geochimica et Cosmochimica Acta* 68: 4877–4889.

- Sauer PE, Eglinton TI, Hayes JM et al. (2001) Compound-specific D/H ratios of lipid biomarkers from sediments as a proxy for environmental and climatic conditions. *Geochimica et Cosmochimica Acta* 65: 213–222.
- Scott G and O'Reilly DJW (2015) Rainfall and circular moated sites in north-east Thailand. *Antiquity* 89: 1125–1138.
- Sessions AL, Burgoyne TW, Schimmelmann A et al. (1999) Fractionation of hydrogen isotopes in lipid biosynthesis. *Organic Geochemistry* 30: 1193–1200.
- Singhrattana N, Babel MS and Perret SR (2012) Hydroclimate variability and long-lead forecasting of rainfall over Thailand by large-scale atmospheric variables. *Hydrological Sciences Journal* 57: 26–41.
- Thai Meteorological Department (2011) Tropical cyclones in Thailand historical data 1951–2010. Available at: www.tmd.go.th/en/archive/thailand_climate.pdf
- Tully JA (2005) *A Short History of Cambodia: From Empire to Survival, Short History of Asia Series*. Crows Nest, NSW: Allen & Unwin, 44 pp.
- Vickery M (1996) What to do about the Khmers, review article on David P. Chandler and Ian Mabbett, *The Khmers*. *Journal of Southeast Asian Studies* 27: 389–404.
- Vickery M (1998) *Society, Economics and Politics in Pre-Angkor Cambodia: The 7th–8th Centuries*. Bunkyo: The Centre for East Asian Cultural Studies for UNESCO.
- Weiss H and Bradley RS (2001) What drives societal collapse. *Science* 291: 609–610.
- Welch DJ (1998) Archaeology of northeast Thailand in relation to the pre-Khmer and Khmer historical records. *International Journal of Historical Archaeology* 2: 205–233.
- Welch DJ (1989) Late prehistoric and early historic exchange patterns in the Phimai Region, Thailand. *Journal of Southeast Asian Studies* 20: 11–26.
- Wohlfarth B, Higham C, Yamoah KA et al. (2016) Human adaptation to mid- to late-Holocene climate change in Northeast Thailand. *The Holocene* 26(11): 1875–1886.
- Wüst RAJ (2001) *Holocene evolution of the intermontane Tasek Bera peat deposit, Peninsular Malaysia: Controls on composition and accumulation of a tropical freshwater peat deposit (DE-6 (2/88))*. MSc Thesis, University of Bern, pp. 105–127.
- Yamoah KKA, Chabangborn A, Chawchai S et al. (2016a) A 2000-year leaf wax-based stable carbon and hydrogen isotope record from Southeast Asia, suggest low frequency ENSO-like teleconnections on a centennial timescale. *Quaternary Science Reviews* 148: 44–53.
- Yamoah KKA, Chabangborn A, Chawchai S et al. (2016b) Large variability in *n*-alkane $\delta^{13}\text{C}$ values in Lake Pa Kho (Thailand) driven by wetland wetness and aquatic productivity. *Organic Geochemistry* 97: 53–60.
- Yan H, Sun L, Oppo DW et al. (2011) South China Sea hydrological changes and Pacific Walker Circulation variations over the last millennium. *Nature Communications* 2: 293.
- Zhang ZH, Ludec G and Sachs JP (2014) El Niño evolution during the Holocene revealed by a biomarker rain gauge in the Galápagos Islands. *Earth and Planetary Science Letters* 404: 420–434.