A 18,000 yr record of tropical land temperature, convective activity and rainfall seasonality from the maritime continent

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Abstract

The maritime continent exports an enormous amount of (latent) heat and moisture to the rest of the globe via deep atmospheric convection. How this export has changed through time under evolving boundary conditions, including the inundation of former Sundaland, is critical for the understanding of global climate dynamics. Given its size, relatively few high-resolution and continuous records exist of past hydroclimate, while terrestrial paleotemperature records are still completely absent from the region. In this study we present a 18,000-year multi-proxy record obtained from a lake sediment at the NW corner of former Sundaland. We found that rainfall seasonality was very important over the entire deglacial period, evidenced by biomass burning and C4 vegetation, despite rising atmospheric CO_2 levels and increasing humidity that normally promotes C3 rainforests. The strong seasonality was reduced only upon ongoing inundation of Sundaland, with a clear inflection point around the Older Dryas event (13.8 ka BP), indicating a distinct system change. Land temperatures during the last stadial periods were 5°C colder than today's 27°C. Temperatures rose gradually during the early Holocene to reach 29°C between 7-2 ka BP, accompanied by increasing convection, both driven by insolation power during the wet season. Convection decreased with lowering wet-season (autumn) insolation during the Meghalayan period, concurrent with the known increase of ENSO variability and Northern Hemisphere climate cooling and drying. Our results provide further insight in the role of Sundaland - turned maritime continent for the global climate system in response to sea level rise and orbital forcing.

Significance Statement

We generated a continuous, 18,000-year land-based paleoclimate record from the northwest corner of former Sundaland in present-day southern Thailand. We found evidence for a strongly seasonal climate for most of the deglacial period, causing biomass burning and suppression of rainforest growth. Terrestrial temperatures were *ca.* 5°C cooler than today during the last cold stadial periods, and *ca.* 2°C warmer between 7000-2000 yr ago, a larger range compared to tropical ocean reconstructions. Comparison of our new and other water isotope records from the region with wettest-season insolation strength, suggests a strong control of the latter on the amount of heat and moisture that gets exported away via deep atmospheric convection, thereby influencing global climate dynamics.

Main Text

Introduction

The maritime continent (MC) forms the central part of the Indo-Pacific Warm Pool (IPWP), defined as the equatorial region with sea surface temperatures (SST) above 28ºC. This region is also called the 'steam engine of the world', as it constitutes a critical component of the global climate system by providing large amounts of latent heat to the higher latitudes via deep atmospheric convection, particularly via the monsoon systems (1). The IPWP also forms a key node in the tropical Walker circulation above the Indian and Pacific Ocean, which is modulated by the El Niño-Southern Oscillation (ENSO) (2) and Indian Ocean Dipole (IOD) (3). Changes in rainfall in the MC have large consequences for both society and ecosystems, where drought-induced biomass burning and peat oxidation can induce rapid release of large amounts of carbon to the atmosphere e.g.(4). A major change in the MC over the last glacial-interglacial (G-I) transition was the inundation of formerly exposed Sundaland and the Sahul shelf north of Australia. It has long been recognized that the submergence of this vast tropical landmass must have had substantial consequences for global-scale climate dynamics (3)(5)(6)(7)(8). Palynological data from the former North Sunda and Molengraaff rivers and their deltaic deposits indicate that the region was covered with lowland rainforest that included sedges, reeds, bamboo, palms and ferns, suggesting fairly humid conditions throughout the last glacial maximum (LGM) around Borneo (9, 10). In contrast, other proxy records from the region indicate drier conditions during the LGM (6)(11). Highest ('driest') $\delta^{18}O$ values are recorded in a Borneo speleothem record at that time (12), and evidence exists for forest contraction and generally drier conditions in both peninsular Malaysia and Palawan during the LGM suggesting the existence of a savannah corridor (13)(14)(15). The somewhat conflicting proxy evidence can be explained by the expanse of former Sundaland which is about the size of the Amazon basin. Another explanation may lie in rainfall seasonality, which can strongly impact proxy records of both vegetation and the recorded water isotope signal. A main problem is, however, that the spatial coverage of high-resolution paleoclimate records from the region remains scant, which is partially explained by the fact that much of Sundaland has disappeared under the waves. Insight into the spatial patterns and mechanisms of the, sometimes rapid, climatic changes that occurred during the G-I transition, as well as over the Holocene, is therefore still limited for this climatically important region. Here, we present a high-resolution and continuous 18,000 year-long multi-proxy record from lake Nong Thale Prong (NTP, 8˚17´N, 99˚37´E) located in southern Thailand at the northwestern border of former Sundaland (Fig. 1). Lake NTP is a shallow (<7 m water depth), small (~210 m^2) karst lake at ~60 m above sea level (16). More details of the lake setting can be found in an earlier publication (17) that focused on the ecological evolution over the last 150 years using ancient DNA and lipid biomarkers. We used the stable carbon isotopic composition of leaf wax-derived long-chain *n*-alkanes ($\delta^{13}C_{wax}$) as a proxy for the relative abundance of C3 vs. C4 vegetation, which is influenced by pCO_2, temperature and seasonality (11)(18). This data set was combined with the stable hydrogen isotope composition of the same leaf wax alkanes (δD_{wax}) and charcoal to gain further information about hydroclimate (19) and seasonality. We also present the first high-resolution land-based temperature record of the IPWP, based on bacterial-derived branched glycerol dialkyl glycerol tetraethers (brGDGTs) (20)(21)(22). The combined proxy records reveal how the aerial exposure and subsequent inundation of Sundaland interacted with orbital variation and other climate forcings to impact the hydroclimate of SE Asia.

Results

Proxy validation

We performed a present-day proxy evaluation and calibration by comparing proxy data analyzed from the surface sediments with instrumental data for the last century. δD_{wax} (17) closely follows the annual precipitation amount (Fig. 3a), where a 10‰ decrease in δD_{wax} corresponds to a 25% reduction in rainfall. This confirms earlier work relating convective activity with both greater rainfall and isotopic fractionation (the 'amount effect', e.g. (23) assuming that δD_{wax} predominantly reflects δD_{precip} after biosynthetic fractionation. Previous research has shown that the hydrogen isotopic composition of both terrestrial and aquatic biomarkers generally reflects that of their source water, although with an offset primarily due to biosynthetic fractionation effects (19, 24, 25). In the humid tropics, the fractionation ($\varepsilon_{wax/water}$) was found to be fairly constant at 130‰ (26). Using this fractionation factor, back-calculated δD values for precipitation of the last century ranges between -40‰ and -60‰, reflecting actual measurements for the region (27).

Reconstructed mean annual air temperatures ($MAAT_{RC}$) (Fig 2b) obtained using the relative abundance of microbial-derived branched GDGTs (22)(21), were recalibrated using instrumental temperature data, thus generating a local calibration for this molecular proxy (See Supplementary Information). $MAAT_{RC}$ and δD_{wax} correlate strongly with each other (Fig 2c), and have the same relation to each other as observed during the seasonal cycle: clear skies during the drier seasons and years with less convective rainfall allow for higher surface temperatures, whereas high clouds associated with deep convection result in a cooling of the surface due to reflection and atmospheric absorption of shortwave radiation (28), an effect that is particularly strong in monsoonal Asia (29).

Sedimentology and limnology

The 18,000 year-long lake NTP sequence consist of organic rich gyttja with TOC contents ranging between 10-40% (Fig. S2). TOC contents vary stepwise between 10 and 40% during the deglacial part of the core, high TOC contents between 9.5-4.2 ka BP, turning to somewhat lower and more variable contents over the last few millennia. Besides some variation caused by changes in minerogenic input, we interpret the TOC changes as mainly caused by alternations between meromictic conditions with permanent bottom water anoxia - leading to preservation of organic matter, and monomictic conditions - resulting in greater organic matter oxidation within the sediments. Stratification in tropical lakes is sensitive to small changes in the lake water level between wet and dry seasons, heat budgets and climate (e.g., wind stress), and other limnological or even ecological feedbacks (30). Given this multitude of factors, we do not attempt to interpret the TOC content. Notably, there is no correlation between the variable TOC content and the lipid biomarker proxies presented further below. This indicates that lake stratification and preservation of organic matter did not influence the primary climatic signal of our proxy records.

The continuous occurrence of seeds of the aquatic plant taxon *Najas* (Fig. 3f; SI Table 4) and a robust age model (Fig. S1) indicates that the shallow lake never dried out. *Cyperaceae* spp. remains, mostly seeds, also occur continuously throughout the sequence, except for the last few millennia when they are nearly absent.

The lower part of the sequence, deposited during Heinrich Stadial 1 (HS1, 18-14.7 ka BP), contains unidentified terrestrial plant remains including woody material, often co-occurring with charred plant remains and macroscopic charcoal; this was also the case for the Younger Dryas period (YD 12.8-11.5 ka BP). Charcoal was most abundant during HS1, then declined towards the end of the deglacial period, with irregular occurrences until the early Holocene around 9 ka BP. Ostracods shells are abundant throughout the HS1, leading to high carbonate contents, and this declines during the Bølling (Bø 14.7-14.0 ka).

Temperature reconstruction

$MAAT_{RC}$ (Fig. 3b) stays around 23-24°C during HS1, a 5°C cooling compared to present. This is lower than the most recent estimate for the tropical ocean during the LGM (−4.2 to −3.7°C; (31) but is in line with estimates based on tropical glacier snow line elevations (32). Temperatures rose during the Bø to reach a maximum of 26°C soon after the Older Dryas event (OD 14.0-13.8 ka BP), but declined during the Allerød (Al, 13.8-12.8 ka BP) and again reached stadial values at the end of the YD. With the start of the Holocene temperatures rose steadily to reach 28-29°C between 7-2 ka BP. The last two millennia are characterized by a cooling trend to a present-day $MAAT_{RC}$ of around 27°C.

$\delta^{13}C_{wax}$ as combined proxy for pCO_2, temperature and rainfall seasonality

Stable carbon isotope ($\delta^{13}C$) values of both of the long-chain *n*-alkanes ($\delta^{13}C_{wax}$) (Fig. 3c) and the bulk (SI Fig. 2) reflects a change from a landscape dominated by C4 grasses and sedges at the beginning of the record, to a humid tropical ecosystem dominated by ^{13}C-depleted C3 vegetation – likely forest - during the Holocene (*cf.* 33). The $\delta^{13}C$ record broadly follows the evolution of atmospheric CO_2 (Fig. 3d). This lends support to the hypothesis that low CO_2 concentration favored C4 vegetation during the LGM (34)(35)(18). Our observation compares well to tropical African records (36)(37)(38). Increasing fractionation against ^{13}C at higher pCO_2 levels and greater humidity (39)(40) – regardless of plant type, can explain part of the trend, especially during the Holocene. An exception to the general trend of increasingly more negative $\delta^{13}C$ from the LGM through the Holocene is a large excursion that starts at 16.0 ka BP, reaching the lowest $\delta^{13}C$ values at 13.8 ka BP.

δD_{wax} as proxy for precipitation

To further investigate past precipitation changes, we analyzed δD_{wax} with higher resolution between 17-10 ka BP to discern trends during deglaciation (Fig. 3a). δD_{wax} was corrected for the effect of global ice volume (41). A confounding factor in the interpretation of δD_{wax} is the potential effect of changing vegetation and associated change in fractionation (42). For instance, C3 and C4 plant types tend to fractionate differently against deuterium and may moreover respond differently to drought in order to minimize water loss while still allowing gas exchange through the stomata (33)(43). The generally stronger biosynthetic fractionation against deuterium of C3 plants compared to C4 would however lead to an opposite behavior of δD_{wax} as observed: the increase in C4 during the Bølling period is associated with more negative δD_{wax}, not more positive. The same argument can be made from a possible transition from a grassy to more woody vegetation during the G-I transition, which would be expected to lead to less negative δD_{wax} values (42), but again the opposite is observed. From the perspective of vegetation change, our δD_{wax} record might thus even underestimate the original variations in source water δD.

Discussion

Deglacial climate evolution

The unusual $\delta^{13}C$ excursion that starts at 16.0 ka BP suggests a renewed contribution of C4 vegetation to the carbon pool in this interval, even though the excursion is coincident with continued warming and its onset correlates with a change in the rate of increase in atmospheric pCO_2 (Fig. 3). The behavior of the $\delta^{13}C$ record indicates that the tropical lowland ecosystem of Sundaland represented an ecotone inhabiting the C3/C4 crossover line during the deglacial period. The ecosystem was thus sensitive to the antagonistic effects of rising pCO_2 and rising temperature on C3 versus C4 plants, where higher temperatures but lower pCO_2 favor C4 plants. However, a third important climatic factor also favors non-perennial C4 vegetation: rainfall seasonality (11). Seasonal dryness was likely promoted by the presence of Sundaland, which only became fully inundated around 11 ka BP during Meltwater Pulse 1b (44). This large landmass prevented the dry northern winds of the Asian winter monsoon from picking up moisture over the Sunda Sea as they do today. This effect was probably promoted by orbital forcing: insolation during NH winter declined while summer insolation increased over the deglacial period, favoring the strength of both the winter and summer monsoon. Strong seasonality promotes biomass production during the wet season, which then serves as fuel for biomass burning during a longer dry season (45). This severely limits the establishment of perennial C3 forests that would otherwise outcompete non-perennial C4 vegetation as atmospheric CO_2 levels rose. The charcoal record (Fig. 3f, SI Table 4) provides evidence that fires were a persistent feature during the entire deglacial period, especially during HS1. We therefore conclude that the return towards a larger contribution of C4 vegetation after 16 ka BP arose from a combination of both rising temperatures and greater seasonality in rainfall patterns, temporarily offsetting the C3-promoting effect of increasing pCO_2. The return of C3 vegetation after 13.8 ka BP, despite the stagnation of CO_2 rise until 12 ka BP, indicates that the region started to have year-round precipitation, reducing seasonal drying and fire. The general trend in $\delta^{13}C$ observed at NTP also is evident in the lower resolution IPWP record from Lake Towuti on Sulawesi (46)(Fig. S3), supporting the interpretation of the combined influence of pCO_2 and rainfall seasonality over the entire IPWP over G-I timescales.

Starting at 18 ka BP, the δD_{wax} record increases to reach highest (least negative) values around 16 ka BP (Fig. 3), indicating that the driest conditions with the weakest convection and greatest evapotranspiration (47) occurred during HS1. This is followed by a rapid decrease during the Bø, and, similar to the $MAAT_{RC}$ and $\delta^{13}C$ records, a sharp reversal at the start of the Al. δD_{wax}, $MAAT_{RC}$

and $\delta^{13}C$ track each other until the YD, with lower (higher) δD_{wax} and higher (lower) $MAAT_{RC}$ - suggesting warmer and wetter (colder and drier) conditions – consistent with inferences from the $\delta^{13}C$ record of patterns of change in C4 vegetation. The combined records suggest that the period of high rainfall seasonality also had wetter wet seasons. Yet despite increased humidity and warmer conditions during the wet season, C4 grasses and sedges were not completely replaced by C3 plants.
We suggest that the steep change in δD_{wax} between 16 and 14 ka BP was driven by an increase in the convective strength over Sundaland with rising temperatures during the Bø. Large-scale convective activity and rainfall amount are the dominant factors that influence water isotope values in tropical SE Asia, in addition to changes in moisture source region (27). Today, during NH summer (JJA), most moisture in southern Thailand is derived from the Indian Ocean, but during the wettest autumn season (SON) there is also a contribution from the South China Sea. In the past, however, moisture derived from evapotranspiration over Sundaland likely also contributed to the isotopic signature. Moreover, longer air mass trajectories over land would have caused a larger rainout effect, leading to lower water isotope values similar to those of present-day mainland SE Asia (27). The lower values might have been exacerbated by the seasonality of rainfall, because the final isotopic signal of water available for plant growth is biased towards that of the wet season (with lowest δD_{precip}) because of its larger contribution to the weighted annual mean (see supplementary information).

The rapid sea level rise during MWP1a changed the hydrologic gradient and reduced the flow of Sundaland river systems. Together with monsoon intensification this most likely transformed the entire Sundaland region into a vast expanse of tropical wetlands (1) with abundant moisture and isotope recycling comparable to the present-day Amazon basin. The parallel reversal of δD_{wax}, $\delta^{13}C$ and $MAAT_{RC}$ around 13.8 ka BP, coincident with the OD event (Fig. 3), indicates a system change towards decreasing rainfall seasonality and a more marine climate. Higher year-round moisture availability would result in a greater contribution of less-depleted δD_{prec} during the cooler winter monsoon months, thereby raising annual mean δD. Lowering of MAAT can occur because of an increase in latent heat production and hence evaporative cooling throughout the year, at the expense of sensible heat. It is also possible that the cold winter monsoon had already started to strengthen during the AI period in response to a southward movement of the mean position of the intertropical convergence zone (ITCZ) caused by NH cooling, something that continued until the end of the YD (~11.5 ka BP). The hypothesis of a southward ITCZ is supported by the coherent patterns in the variability of δD_{wax} during the YD and the Greenland ice core $\delta^{18}O$ record, with shifts in the mean position of the ITCZ in response to latitudinal temperature gradients (48). After the YD, however, δD_{wax} continues to increase until 11 ka BP, in opposition to the rapid change in the Greenland record, but interestingly enough also opposite to the local $MAAT_{RC}$. We attribute this to the development of an even more equable hydroclimate throughout the year, with an increased relative influence of 'dry'-season rainfall with higher δD values, sourced from the Gulf of Thailand and the South China Sea.

Orbital forcing of Holocene and deglacial climate

After 11 kyr BP, δD_{wax} and $MAAT_{RC}$ vary again in tandem. Both show a generally asymptotic trend towards the warmest and wettest conditions peaking at ~4.5 ka BP. This indicates that the 'steam engine of the world', the IPWP, was at full power during the mid-Holocene thermal maximum, exporting greatest amounts of latent heat, i.e. moisture, to the Northern Hemisphere during this time. This long-term coupling between δD and $MAAT_{RC}$ at orbital to millennial scales is opposite to that of higher frequency relationships at annual to decadal scales (Fig. 2), where the total insolation is distributed between latent and sensible heat. Orbital-scale changes in the seasonal distribution of insolation apparently steer $MAAT_{RC}$ and convective strength in the same direction. The precessional cycle has indeed long been identified as the dominant component of orbital forcing influencing tropical and monsoonal climate (e.g. 29). NH summer insolation (JJA) is most commonly used to explain the waxing and waning of monsoon strength, even though leads and lags between proxy records exist. In the tropics, however, the season of most intense rainfall does not occur during JJA. Thus, we compare our records with 'wet season' insolation (WSI), i.e. the mean monthly insolation during the wettest part of the annual cycle at 8ºN. Indeed, δD_{wax} follows the insolation curve for the wettest months, September-November (Fig. 4) (50), although with a notable excursion during the Bø/Al-YD periods, which were caused by the influence of Sundaland.

The 7% variation of WSI over the last 18,000 years (418 - 446 kW/m^2) (50) thus appears to be a main driver of both surface (temperature) and atmospheric (latent, convective) heat flux. This observation is consistent with a Borneo (4ºN) speleothem record (51), where $\delta^{18}O$ is correlated with the wettest months at that latitude (Fig. 4). The δD_{wax} record from Lake Towuti (Sulawesi) has been interpreted as being driven primarily by changes in moisture source and air trajectories (52), but it also shows a strong correspondence with WSI at 2ºS (Dec&Jan, during the passing of the ITCZ) (Fig. 4). Both δD_{wax} records (NTP and Towuti) show a sensitivity to WSI of -1.4‰ per W/m^2, as does the Borneo record when scaled by a factor of eight for $\delta^{18}O$ according to the global meteoric water line. Combined, these records provide further evidence for the influence of the precessional cycle on the isotopic composition of regional precipitation, via the combined mechanisms of regional convective activity and associated amount of precipitation. This is exacerbated by secondary effects of seasonality, which also affects the distribution between latent and sensible heat. In the tropics there is a clear correlation between insolation and rainfall amount, with at present lowest values in June and July (68). Over the course of a precessional cycle, the shift in seasonal distribution of solar energy can be as much as 15%, which must be causing a large effect on rainfall seasonality. The lowest δD_{wax} values in the mid Holocene indicate strongest convective activity during the autumn, i.e. at highest WSI, but precipitation was likely lower in spring at low insolation level, causing a stronger bias in the annual mean towards the autumn. At the same time, the highest MAATs observed at NTP in the mid Holocene are likely caused by a shift to mean dryer conditions with clearer skies during winter and spring, even with relatively lower insolation levels. At the same time, higher mid Holocene MAATs result from a combination of drier and sunnier spring months, compensating for relatively low insolation levels (more sensible heat, less latent heat), and cloudy wet months that however receive highest solar inputs. Our data are thus consistent with the theory that the precessional cycle caused greater seasonality in the mid Holocene, compared to the low-seasonality period we currently experience. The seasonal bias on the mean annual isotopic composition outlined here may explain many water isotope records from the tropics, without a large need to infer changes in rainfall amount, or changes in moisture source. We discuss the interaction between precession and the annual cycle and its influence on precipitation seasonality and the mean annual isotope signal, together with MAAT, in further detail in the supplementary information (SI).

Influence of IPWP hydroclimate on the Asian monsoon and ENSO

The trends in the NTP δD_{wax} record are similar to those in the Asian speleothem $\delta^{18}O$ records (53)(54) (Fig. 5), including the OD event and the 'peak isotope' feature at the beginning of the Al. NTP δD_{wax} also tracks the Greenland ice core record (Fig. 5), reflecting the impact of high-latitude NH forcing on tropical climate. NTP receives most of its moisture from the Indian Ocean, in contrast to the East Asian speleothems, which also receive significant summer monsoon moisture from the East (27)(54). The shared patterns of variation are consistent with modeling studies (55)(56), which have shown that East Asian speleothem $\delta^{18}O$ records reflect the isotopic composition of the advected moisture, as much or more so than rainfall amount, and that large-scale convection patterns are the main drivers of the isotopic composition of precipitation (27). Our results, which are similar to those from a recent study in northern Thailand (8), demonstrate that the exposure and inundation of Sundaland played a decisive role in affecting the water isotopic composition not across mainland East Asia, but also in Thailand. The same factors that lowered δD_{wax} at our site (more rainout and more land-derived moisture from Sundaland, and greater seasonality), must also have applied further inland. Remote processes upstream of the SE Asian Monsoon, such as the presence / inundation of Sundaland, as well as precession-forced changes in WSI in the lower tropics, need to be considered when interpretating SE Asian water isotope records in sediments and speleothems. Experiments with isotope-enabled general circulation models are needed to gain further insight.

Another notable feature of the δD_{wax} record are the positive ('dry') excursions between 4 and 3 ka BP, which is coincident with the onset of the Meghalayan age (Fig. 3), characterized by megadroughts observed in multiple regions (57). The dry events occur on top of a general decline in convective activity, which follows the decrease in WSI after 5 ka BP. Our results of a wettest and warmest mid Holocene extend the finding (58) of a warmer mid Holocene thermocline in the IPWP east of 115°E, caused by greater September insolation. The warmer and deeper thermocline causes a stronger zonal thermal difference across the equatorial Pacific, which further promotes deep atmospheric convection and rainfall over western equatorial Pacific in a positive feedback mechanism, inducing a stronger Walker circulation and suppression of ENSO activity. The interaction of the precessional and seasonal cycles that act upon the IPWP, being the 'steam engine of the world', thus appears to play a decisive role in global climate dynamics by regulating the amount of latent heat exported to the higher latitudes. In this respect, we even speculate that the inundation and warming of Sundaland and may have provided a key positive feedback mechanism during the last G-I transition, and possibly also earlier ones.

Materials and Methods

Sampling and sample processing

Two parallel sediment cores were retrieved in one-meter sections using a rod-operated Russian corer from a small raft at the deepest part of the lake. After recovery, the sections were wrapped in foil and secured and transported in PVC tubes to Stockholm University, where they were stored at 4°C until further analysis. Sub-samples were taken in contiguous 1-cm increments and split to accommodate subsequent analyses. One half of the samples was utilized for macrofossil and charcoal analysis and radiocarbon dating. The other half of the samples was freeze-dried and analysed for loss-on-ignition (LOI), bulk total organic carbon (TOC), nitrogen (TN) and their isotopes, lipid biomarkers and compound-specific hydrogen and carbon isotopes. For LOI, samples were dried overnight at 105°C, ground and then combusted at 550 °C for 3h. LOI was calculated as a percentage of the dry sample weight to obtain an estimate of the organic matter and carbonate content. In parallel, a sediment-water interface surface core covering the last 150 years was retrieved and sampled on site in one cm slices (17).

Macrofossil analysis and radiocarbon dating

Approximately 380 samples were sieved under running water (mesh sizes 0.5 and 0.25 mm) to recover plant macrofossils for radiocarbon dating. Plant remains were picked with tweezers under a binocular microscope, described, and rinsed multiple times in deionized water, placed in pre-cleaned glass vials and dried overnight at 105 °C. 59 samples were dated at the 14Chrono Centre, Queen's University Belfast, where pre-treatment and measurement followed the methodology described in (59). Based on these, an age-model (SI Fig. 1) was constructed using Bacon, a Bayesian statistics-based routine (60) that estimates the accumulation rate for sediment segments based on the radiocarbon dates calibrated using the intCal13 NH calibration curve (61).

Bulk geochemistry

%TOC, %TN and bulk $\delta^{13}C_{org}$ and bulk $\delta^{15}N_{bulk}$ were measured on a Carlo Erba NC2500 elemental analyser, coupled to a Finnigan MAT Delta^{+} mass spectrometer. To remove carbonates, samples were fumigated with HCl within a dessicator prior to analysis. $\delta^{13}C_{bulk}$ is expressed in ‰ against the Vienna PeeDee Belemnite (VPDB) standard, and had an analytical error of less than ±0.15‰. $\delta^{15}N_{bulk}$ are reported in ‰ relative to air (N), with an analytical error of ±0.15‰.

Lipid biomarkers

Lipid extraction was performed on freeze-dried samples by sonication with a mixture of dichloromethane and methanol (DCM-MeOH 9:1 v/v) for 20 minutes and subsequent centrifugation. The process was repeated three times and supernatants were combined. Aliphatic hydrocarbon fractions were isolated from the total lipid extract using silica gel columns (5% deactivated) that were first eluted with pure hexane (F1) and subsequently with a mixture of DCM-MeOH (1:1 v/v) to obtain a polar fraction (F2). A saturated hydrocarbon fraction was obtained by eluting the F1 fraction through 10% $AgNO_3$ impregnated silica gel using pure hexane as eluent. The saturated hydrocarbon fractions were analyzed by gas chromatography – mass spectrometry for identification and quantification, using a Shimadzu GCMS-QP2010 Ultra. C_{21} to C_{33} *n*-alkanes were identified based on mass spectra from the literature and retention times. The concentrations of individual compounds were determined using a calibration curve made using mixtures of C_{21}-C_{40} alkanes of known concentration.

Leaf wax hydrogen and carbon isotope analysis

The hydrogen isotopic composition of *n*-alkanes (expressed in delta notation in ‰ against VSMOW) was analyzed 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 the HTC reactor of a GC Isolink II and Conflo IV system. Helium was used as a carrier gas at constant flow mode and the compounds separated on a Zebron ZB-5HT Inferno GC column (30 m x 0.25 mm x 0.25µm). A standard set of alkanes with known isotopic composition (obtained from A. Schimmelmann, Indiana University, USA) was used for daily calibration of the reference gas. The average standard deviation of δD values was 5‰. The reported δD_{wax} values are the average of the most abundant long chain *n*-alkanes: C_{27}, C_{29} and C_{31}. To correct for the higher global average of global oceanic δD during lower sea levels, the δD values of the *n*-alkanes were ice volume corrected (*c.f.* (41) as follows: $\delta D_{wax\text{-}c} = (\delta D_{wax} + 1000) / (\delta O^{18}{}_{w} * 8 * 0.001 + 1) - 1000$, with interpolated ocean water $\delta O^{18}{}_{w}$ values (62).

$\delta^{13}C_{wax}$ was measured on the same compounds on the same system and the same isotope standards, except for the use of the combustion reactor. $\delta^{13}C_{wax}$ values are the average of C_{27}, C_{29} and C_{31} alkanes, expressed in delta notation in ‰ against VPDB, with an average standard deviation of 0.5‰.

Glycerol dialkyl glycerol tetraether (GDGT) analysis

Branched glycerol dialkyl glycerol tetraethers (brGDGTs) were measured on the F2 fractions after reconstituting in MeOH:DCM 9:1 and subsequent filtration through 0.45 µm PTFE filters, following published protocols (63). Analysis was done using a Thermo-Dionex HPLC connected to a Thermo Scientific TSQ quantum access triple quadrupole mass spectrometer, using an APCI interface. Chromatographic separation was achieved on a Kinetex C18-XB reverse phase column using a gradient of mobile phase A: MeOH with 0.04% formic acid and mobile phase B: propan-2-ol with 0.04% formic acid. GDGTs were detected in SIM mode at *m/z* 1020 (scan width 7, 0.2s), 1034 (width 7, 0.2s), 1048 (width 7, 0.2s), 1296 (width 17.5, 0.5s). Quantification was performed using Excalibur software, using the (M+) and (M+1) ions of the GDGTs. More details can be found elsewhere (63). MBT and CBT proxies were calculated following (64).

Acknowledgments

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The data presented in this paper is available online at the Bolin Centre of Climate Research Database: (link).

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Figures

Figure 1. Location of Lake Nong Thale Prong (NTP) and other records mentioned in the text: GB: Gunung Buda National Park speleothem, Borneo; TWT: Lake Towuti, Sulawesi; CS: Chinese Speleothems; TAN: Tangga cave, Sumatra. The map shows the extent of the emerged landscapes of former Sundaland during the last glacial maximum sea level low stand.

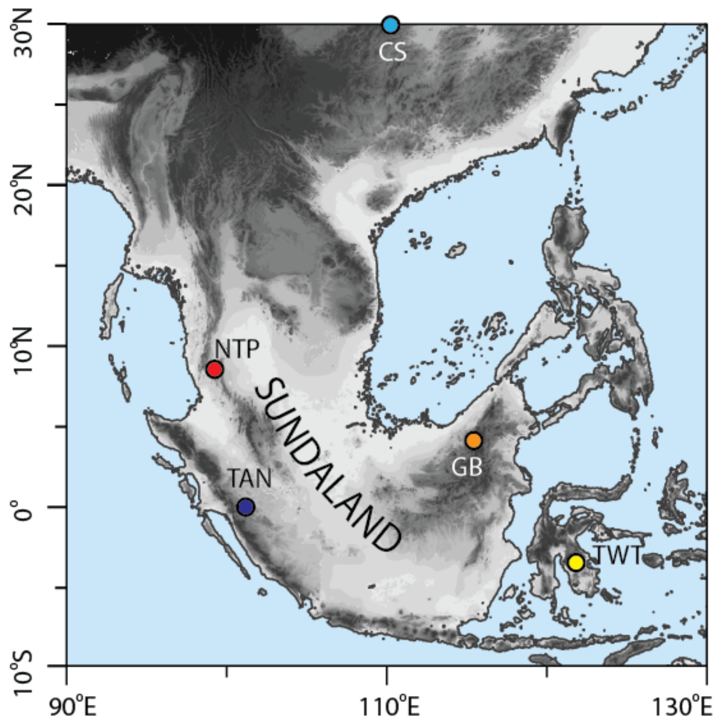


Figure 2. Comparison of instrumental climate data with proxy data measured on the NTP surface sediments. **a**) Mean Annual Air Temperature ($MAAT_{RC\text{-}Sun}$) reconstructed using bacterial-derived branched GDGTs (right axis, grey), compared to observations (left axis, stippled black). To obtain a local calibration the reconstructed MAAT was scaled for amplitude and mean (see SI) to correspond with the instrumental record (black, left axis). **b**) Instrumental rainfall data (right axis, stippled red) compared with δD_{wax} data from the same samples (17) suggests a good correlation between the two. **c**) Scatter plot of δD_{wax} and reconstructed MAAT from the same samples. Instrumental climate data are taken from the CRU TS monthly high-resolution gridded multivariate climate dataset, Version 4 (65).

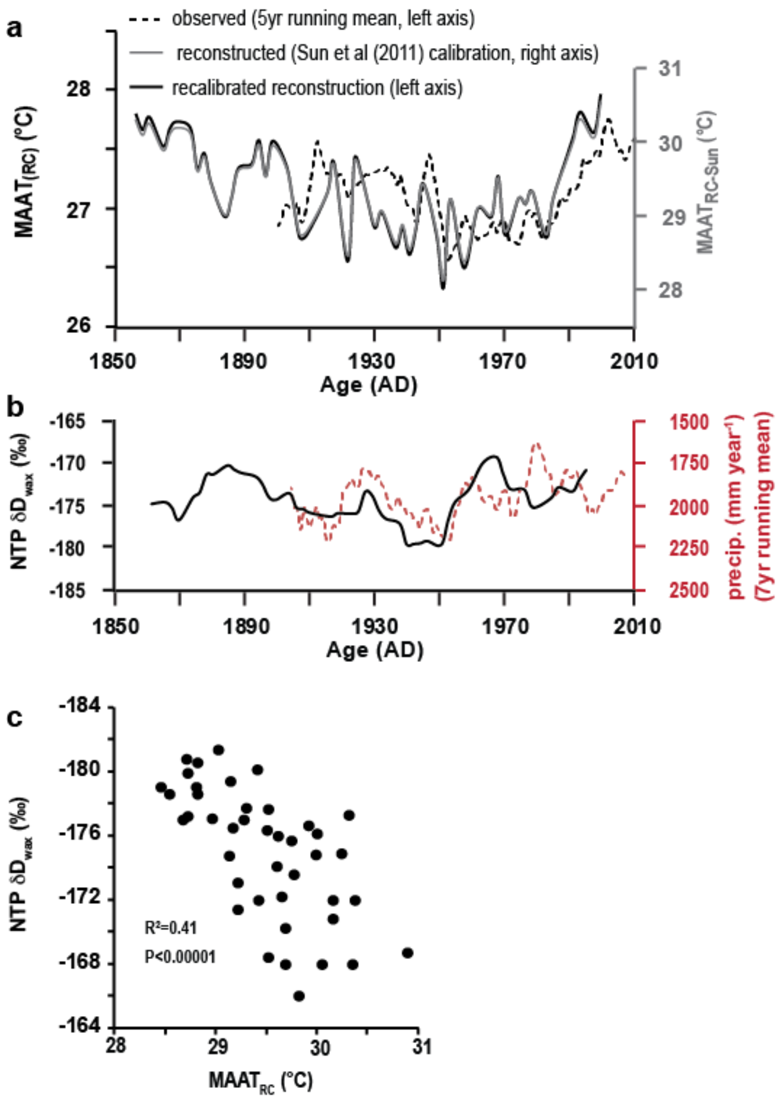


Figure 3. Proxy records of lake NTP of the last 18,000 years. **a**) δD_{wax}, both as measured and corrected (stippled) for global sea level change. **b**) Reconstructed mean annual air temperature ($MAAT_{RC}$). **c**) $\delta^{13}C_{wax}$. **d**) Atmospheric CO_2 levels (66). **e**) Sea level reconstruction for the Sunda Shelf region (44) **f**) Macrofossil and charcoal results. Thick line: very abundant; Thin line: present. Meg: Meghalayan period; YD: Younger Dryas; Al: Allerød; OD: Older Dryas; Bø:Bølling; HS1: Heinrich Stadial 1. MWP: Meltwater pulses.

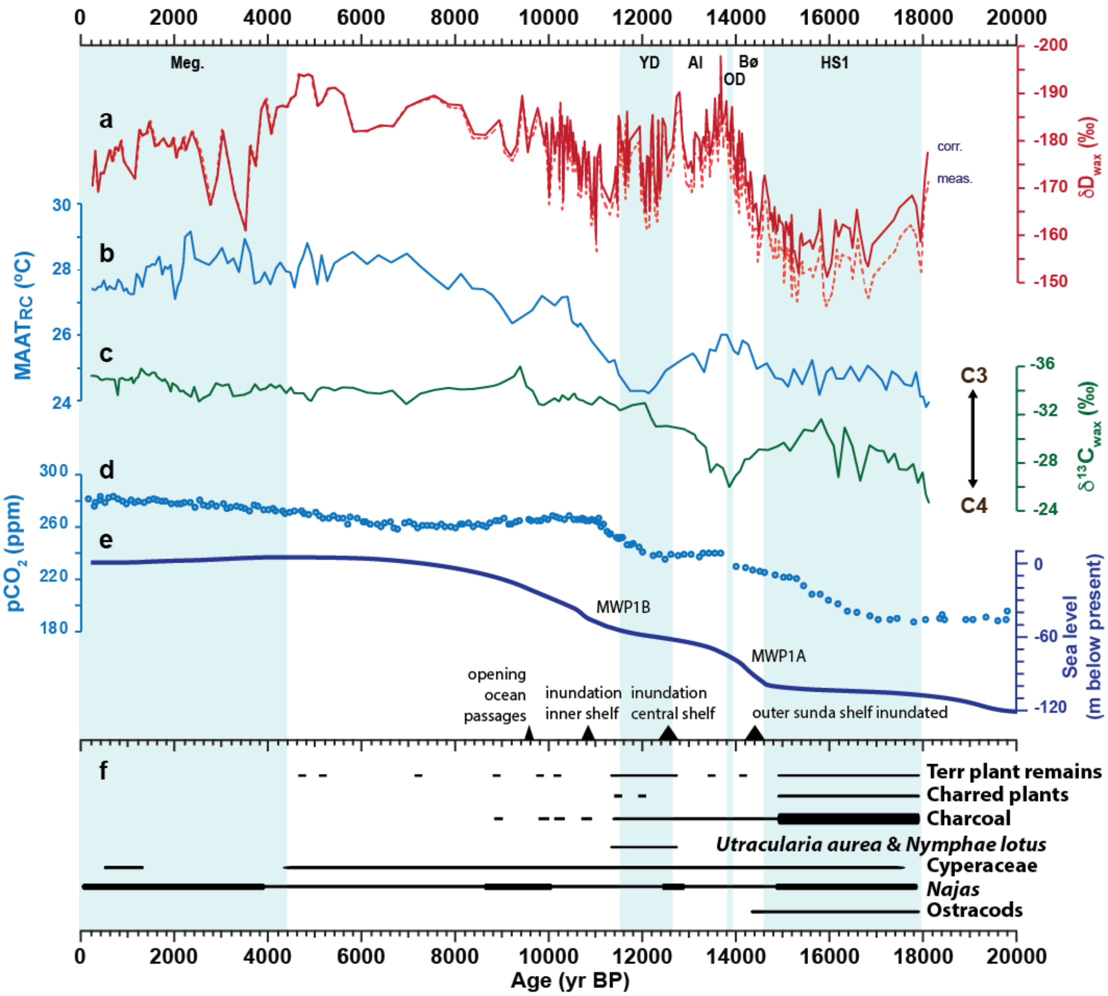


Figure 4. Comparison of isotope records. **a**) Greenland ice core $\delta^{18}O$ as a reference for NH temperature (67). **b**) Combined Chinese speleothem $\delta^{18}O$ (53). **c**) NTP δD_{wax} corrected for sea level effect (this study). **d**) Borneo speleothem $\delta^{18}O$ record (dark(12) and light (51) orange). **e**) Sumatra speleothem $\delta^{18}O$ record (68). **f**) Sulawesi δD_{wax} record (52). b-f are all plotted on the same scale where one unit in $\delta^{18}O$ corresponds to 8 units in δD space, according to the global meteoric water line. Grey dotted lines over b-d and f show the solar irradiation averaged for the 2 or 3 wettest months (WSI:Wet Season Insolation) for the latitudes of the respective records (50). No clear wettest period could be defined for Sumatra (see SI). Time periods are shown as in in Figure 3.

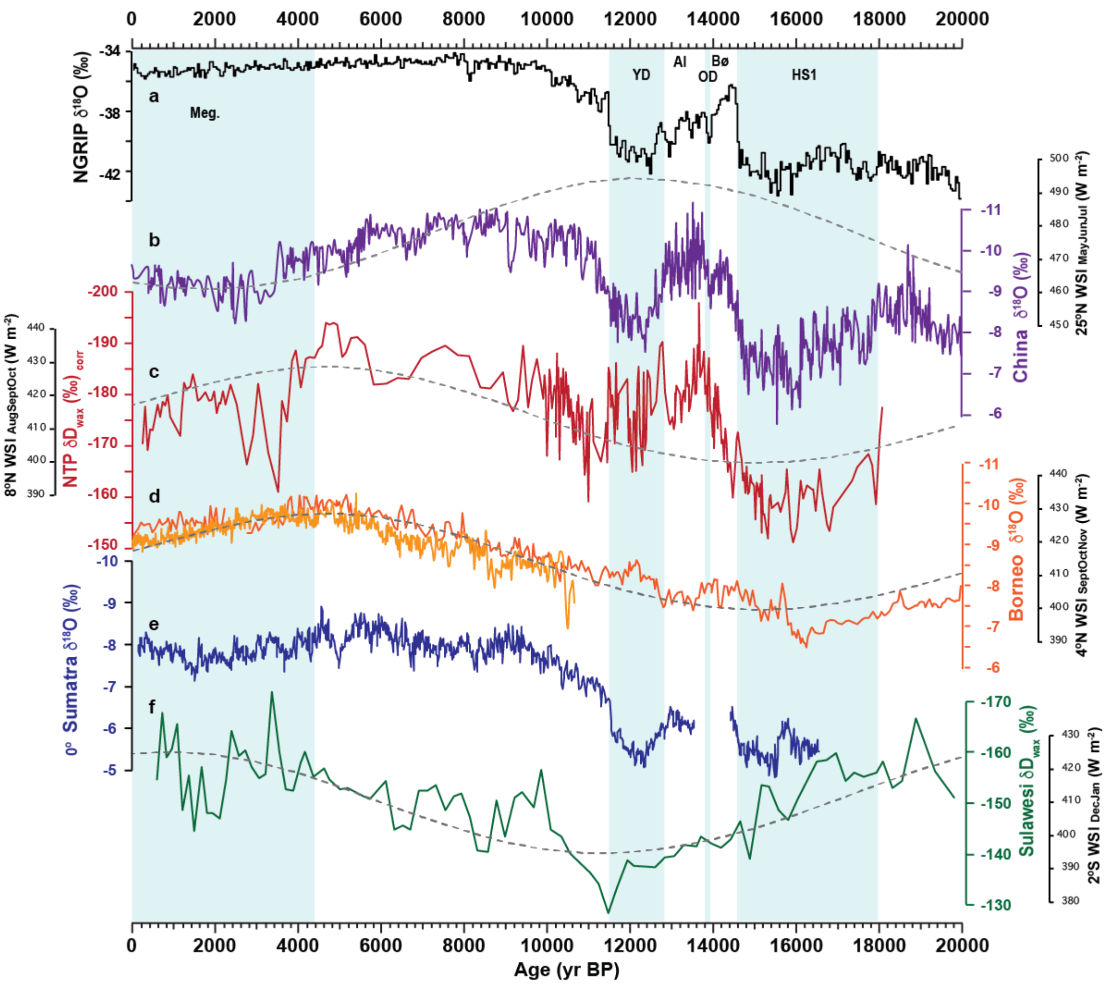
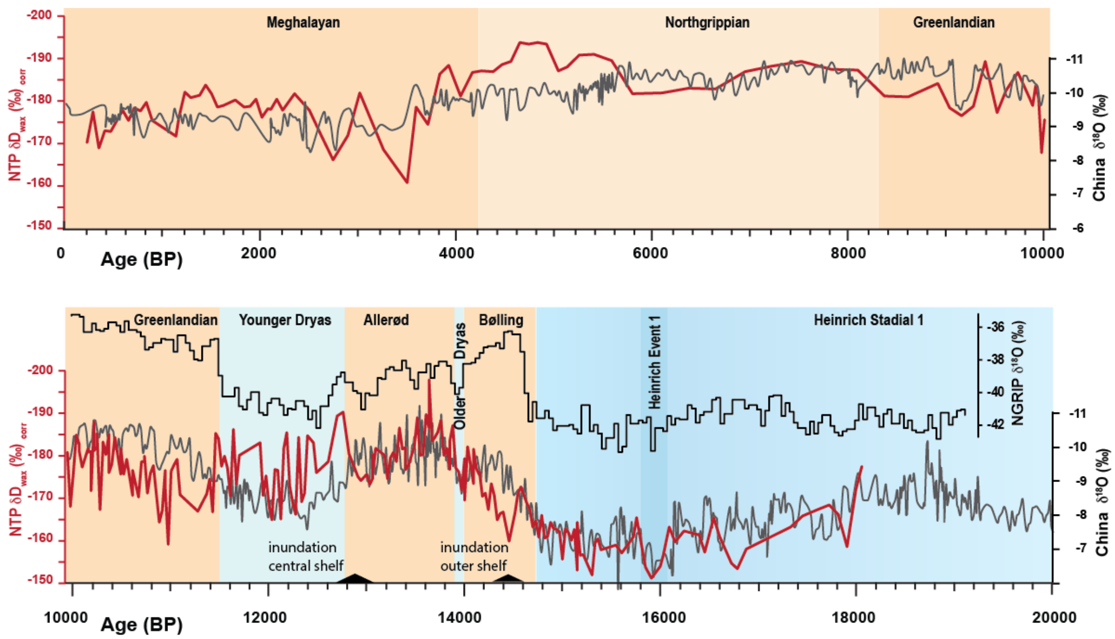


Figure 5. Comparison of the sea-level-corrected NTP δD_{wax} (this study) with the Chinese speleothem $\delta^{18}O$ record over the last 20,000 years (53) and Greenland $\delta^{18}O$ for reference (67). The records are scaled in the same way as in Figure. 4.



Supplementary Information for

A 18,000 yr record of tropical land temperature, convective activity and rainfall seasonality from the maritime continent

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This PDF file includes:

- Supplementary text
- Figures S1 to S10
- SI References

Other supplementary data for this manuscript will be made available at the Bolin Center of Climate research Database: https://bolin.su.se/data/

Supplementary Tables
Table 1. Composite stratigraphy
Table 2. Radiocarbon Data
Table 3. Plant macrofossil and charcoal data
Table 4. Bulk geochemistry: TOC, TN, LOI, $\delta^{13}C_{bulk}$ and $\delta^{15}N_{bulk}$
Table 5. Leaf wax δ D
Table 6. Leaf wax $\delta^{13}C$
Table 7. GDGTs and reconstructed MAAT
Table 8. Surface Core GDGTs, reconstructed MAAT, leaf wax δD and instrumental MAAT

Temperature reconstruction using branched GDGTs

A basic prerequisite for the valid use of brGDGTs, a relatively high branched-over-isoprenoid tetraether (BIT) index, is easily met with values of 1.0 throughout the core. Reconstructed pH values, based on the CBT index (1) were 8.0±0.2 over the entire core, with lowest values during the YD and a downward trend for the last 2000 years (Fig. S4). This means that temperature is the dominant environmental factor exerted on the brGDGT distribution. At the time of measurement, we had not adopted the new HILIC-based method which separates between 5-methyl and 6-methyl branched GDGTs (2) but used our own method based on reverse phase chromatography (3), similar to the one used by Zhu *et al*. (4), and which compared well with the original method using a cyano column. As a consequence, we do not have individual quantifications of 5-methyl and 6-methyl branched GDGT isomers used in the revised MBT'_{5me} temperature proxy for mineral soils (5), peats (6), or East African lakes (7). However, for high temperatures as is the case for our site, the main response to temperature is a shift between tetra- and pentamethylated GDGTs, which makes the differentiation between 5- and 6-methyl GDGTs less relevant than in cold environments. The relative abundance of tetra-, penta- and hexamethylated GDGTs plot in the same region as datasets produced with the HILIC method from east African lakes and from global soils and peats (Fig S4). This strengthens the confidence that the brGDGTs we measured can be used as a temperature proxy. Among the various GDGT-temperature calibrations that have been developed since the original one (1), we chose to apply the global lake calibration of Sun *et al.* (8) for lakes with pH<8.5, which also included data from nearby lake Towuti:

$$MAAT_{Sun\text{-}cal} = 3.949 + 38.213\ MBT - 5.593\ CBT \qquad \text{(Eq. 1)}$$

The MBT/CBT-based MAAT (°C) reconstruction using the global regression model (8) shows a good agreement with temperature observations in the region (closest grid point at 8.25° N; 99.25° E from the University of East Anglia Climate Research Unit dataset CRU TS3.23) (9) for the overlapping period of 1903-2001. There is however an offset and overestimation of variability in the proxy reconstruction using the global lake calibration relative to the local temperature. To adjust the reconstruction to our local conditions, we re-calibrated the global reconstruction by replacing the mean of the $MAAT_{Sun\text{-}cal}$ values obtained using the global regression ($\mu_{proxy\text{-}global}$) and standard deviation ($\sigma_{proxy\text{-}global}$) with those of local conditions from CRU TS3.23. This was done by first normalizing the proxy record for the overlapping period 1903-2001 and then re-normalize it using the mean ($\mu_{obs\text{-}local}$) and standard deviation ($\sigma_{obs\text{-}local}$) of the local observations for the same time period (Eq. 2).

$$MAAT_{i,local} = \left[(x_{i,proxy-global} - \mu_{Proxy-global})\frac{\sigma_{Obs-local}}{\sigma_{Proxy-global}}\right] + \mu_{Obs-local} \qquad \text{(Eq. 2)}$$

resulting in a record of recalibrated $MAAT_{RC}$ values. This re-calibration effectively adjusts the intercept and slope of the original calibration so that the proxy data reflects the mean and annual variability observed over the instrumental record. Generating a new calibration by regression of the GDGT data with the instrumental record is not straightforward, because the samples do not correspond to annual measurements but approximately 3 years, with an error of the age estimate based on ^{210}Pb dating that increases with depth.

We even performed the same exercise using the original calibration (1), and came to the same results (Fig S4a). It is important to note that all data come from one location where the microbial ecology of the brGDGT-producing organisms and the dominant environmental factors vary much less compared to the globally distributed surface sediment datasets used to generate the GDGT calibrations. For reference, the RMSE of the East African lake calibration (7) is approximately 2.5°C.

About the influence of the precessional cycle on seasonal insolation, its effect on tropical hydroclimate and the annual mean water isotopic composition.

Solar radiation, or insolation, is the main source of energy reaching the earth, which gets distributed between reflection (albedo), sensible heat (temperature), and latent heat upon evaporation of water. At our site lake Nong Thale Prong at 8°N, the present-day annual insolation curve exhibits two highs: one in April and one in August/September (Fig. S5), when the sun's altitude is 90°C at noon. The annual movement of the ITCZ and the Monsoon system behaves in an attenuated fashion (Fig. S6). From January onwards, temperatures rise (Fig. S7) but precipitation remains low until May, because the ITCZ remains south. Dry conditions with low cloud cover cause low albedo, resulting in highest surface temperatures in April (Fig. S7). The ITCZ passes over quickly going northwards during May and June, to merge with the Asian Summer Monsoon system during the NH summer (Fig. S6). The Monsoon/ITCZ moves back towards the equator in NH autumn, causing the strongest period of convective precipitation over the northern IPWP from September-November (Figs. S6 and S7). During this time, much of the incoming radiation is reflected by high convective clouds, or is used to generate latent heat, leading to reduced surface temperatures (Fig. S7).

Between 6-4 ka BP, perihelium (the moment the earth is closest to the sun during its elliptical orbit) occurred in September-October, causing 5% greater insolation in September compared to today (Fig. S5). At the same time, insolation during early spring was lower (Figs. S6 and S8). At the equator, a clear correlation exists between insolation and rainfall amount, with at present lowest values in June and July (10) (Figs. S9 and S10). Over the course of a precessional cycle, the shift in seasonal distribution of solar energy can be as much as 15%, which must be causing a large effect on seasonality. At and near the equator, the 'dry' season may even have shifted from NH summer to SH summer (Fig. S9), and the wettest season more towards or away from the March and September annual maximums, depending on the orbital phase. Because of this we did not assign a wet season insolation curve to the Tangga Cave record at Sumatra (main Fig. 4). We reconstructed highest mean annual temperatures at our NTP site during the mid-Holocene. Given the fact that total annual isolation hardly changes over the precessional cycle, this must be caused by a shift to generally dryer conditions with clearer skies, most likely during winter and spring. At the same time, our δD_{wax} record exhibits lowest values during the mid-Holocene, indicating stronger convective activity during the autumn and/or a greater relative contribution of isotopically more depleted precipitation in the (wettest) months. Our data are thus consistent with the theory that the precessional cycle caused greater seasonality in the mid Holocene, compared the low-seasonality period we currently experience. At lake NTP, the water isotopic composition of precipitation is primarily dependent on the convective activity as well as the isotopic composition of water evaporated from the Indian Ocean ('source effect') (11). In the mid Holocene, stronger insolation in NH summer and autumn, i.e. the wettest season insolation (WSI) for the SE Asian Monsoon and the northern IPWP, will have caused warmer ocean surfaces and subsequently greater evaporation and convective activity both in the northern Indian Ocean (specifically the Bay of Bengal), as well as the South China sea.

At 20 kyr BP, the seasonal pattern of insolation is similar as today (Fig. S7), but over the ensuing deglacial period (e.g., 14 ka BP) perihelion shifts towards NH spring. Being a mirror case of the situation at 6 ka BP, this would cause higher convective activity in the northern IPWP during spring with moisture sourced from the Pacific side. In NH autumn, the lower insolation would have caused a weakened ITCZ convection. Different to today, however, was the presence of Sundaland. Air masses coming from the northeast would not have been able to pick up as much moisture as they can today over South China Sea. Consequently, the greater NH spring insolation only could lead to more rainfall when Sundaland became a large wetland, allowing more land surface evapotranspiration. Until then, the annual sum of precipitation would have derived almost exclusively from the autumn. After 14 ka BP, the perihelion moves towards NH summer, and insolation remains high from spring through summer and into the autumn. After 12 ka BP, insolation becomes ever more focused on the autumn (all autumn months go 'up', see Fig. S8), until 6 ka BP, thus aligning ever more with the annual movement of the ITCZ and the period of strongest convection. Over the last millenniums, perihelium has shifted from NH winter towards spring.

The relative strength of insolation and related convective activity distributed over the year will have had its effect on the annual weighted mean of δD of precipitation. Results for nearby Phuket (11) indicate only a relatively small range in $\delta^{18}O$ through the year, from -2‰ (i.e. δD = -7‰) in April to -8‰ (δD = -55‰) in November, with an annual weighted average of -5.5‰ (δD = -35‰). The moment a shorter season is responsible for the majority of the annual sum, i.e., when the perihelion aligns with the wettest months in autumn, then the weighted mean annual isotope value will shift towards that season. This is the likely situation in the mid Holocene around 6 ka BP where relatively heavy (isotopically) spring precipitation will have contributed less, however the stronger convective activity during the wet autumn season will likely have caused more depleted wet season precipitation as well. Together this causes a bias towards lower mean annual δD values at times of strong seasonality. This seasonal bias also means that there does not need to be any close relation between total annual rainfall and the mean isotopic composition. The seasonal bias on the mean annual isotopic composition can explain many isotope records from the tropics, without a large need to infer changes in rainfall amount, or changes in moisture source.

Lastly, we comment on the effect of the precessional cycle on advected moisture. In the early Holocene, perihelion (i.e., highest insolation) occurred during the start of the Asian summer monsoon, when the advected moisture in mainland SE Asia does not yet reach very depleted values ($\delta^{18}O$ = -18‰ (11)) In the mid-late Holocene, however, perihelion has shifted to the autumn, at a time when the moisture reaching mainland SE Asia is already much more depleted. The expected shift of the Monsoon strength towards the autumn will thus also cause a shift in the mean isotope composition - even if at the local scale insolation and therefore total monsoon strength has already decreased. The end result is an attenuation of 'peak isotope', because of the source effect, and not because of the amount effect. This effect can explain the temporal shift of 'peak isotope' away from the time perihelium occurred during classical NH summer (JJA, between 10-8 ka BP) towards later (8-6 ka BP) as observed in the Chinese speleothem records.

Supplementary figures

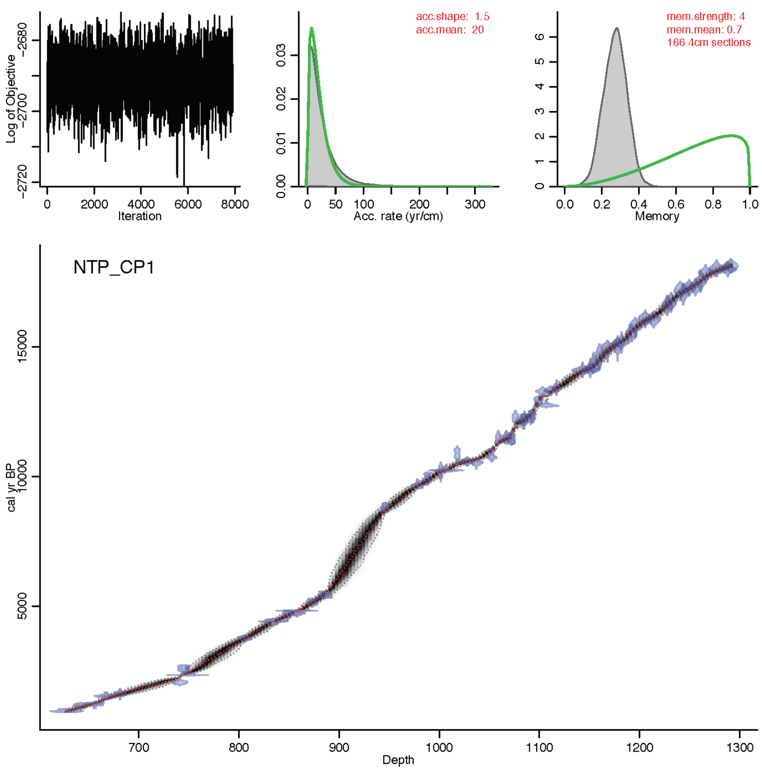

Figure S1. Age model of Lake Nong Thale Prong. Depth is expressed in meter below lake level.

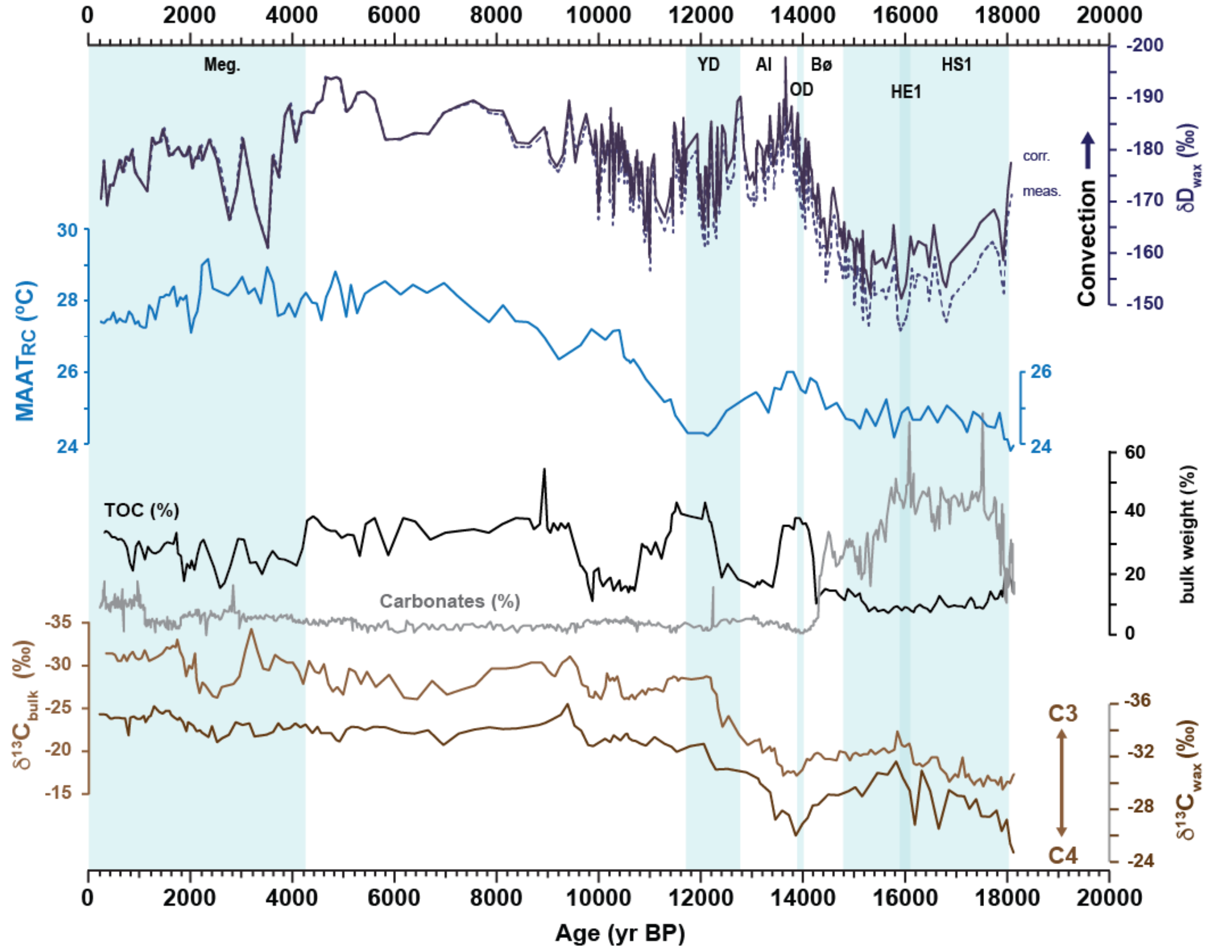

Figure S2. Proxy records of lake Nong Thale Prong, with elements of Fig. 3 in the main paper, extended with TOC content, bulk δ^{13}C, and carbonate content based on loss-on-ignition.

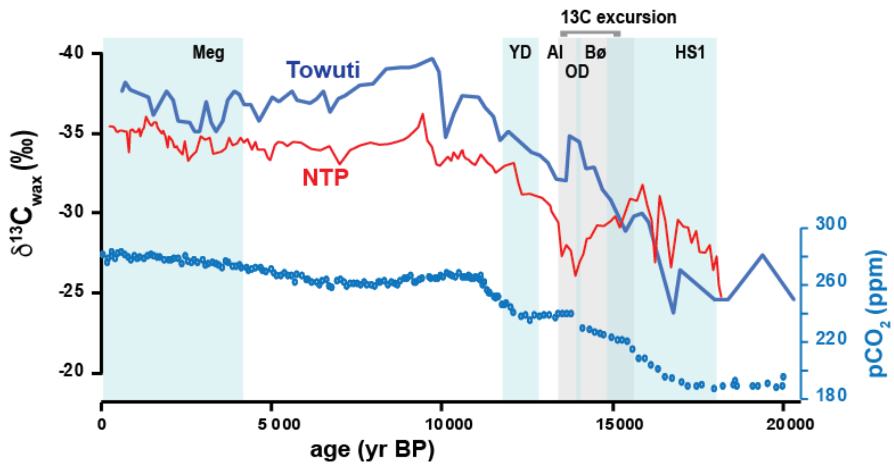

Figure S3. Comparison of the $\delta^{13}C_{wax}$ records of lake NTP (this study) and lake Towuti (12) and atmospheric CO_2 levels (13).

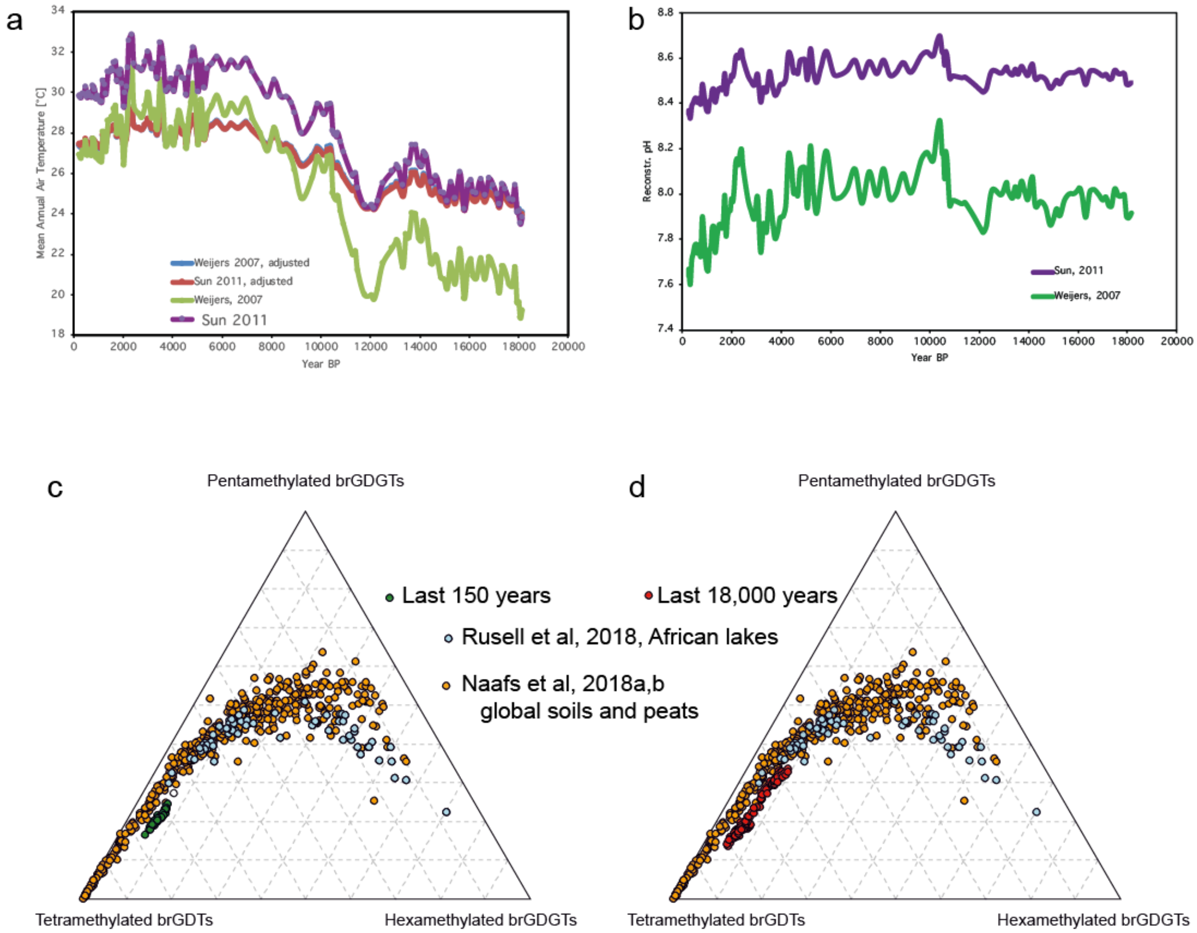

Figure S4. a) Reconstructed MAAT using the MBT/CBT ratios according to two calibrations (1)(8), and after local recalibration (adjusted) as described in the text. b) reconstructed pH using the CBT ratios (1)(8), c) Triplot of the relative abundance of tetra, penta- and hexamethylated GDGTs in the surface core (green); a pooled soil and peat (6)(14) (orange) and an African lake dataset (7)(light blue) are plotted for reference. d) the same as c, but for the long core NTP data (in red). The reference data set includes both the 5- and 6-methyl GDGTs, while the NTP dataset includes all isomers of the same *m/z*. Cindy de Jonge is acknowledged for making the triplots.

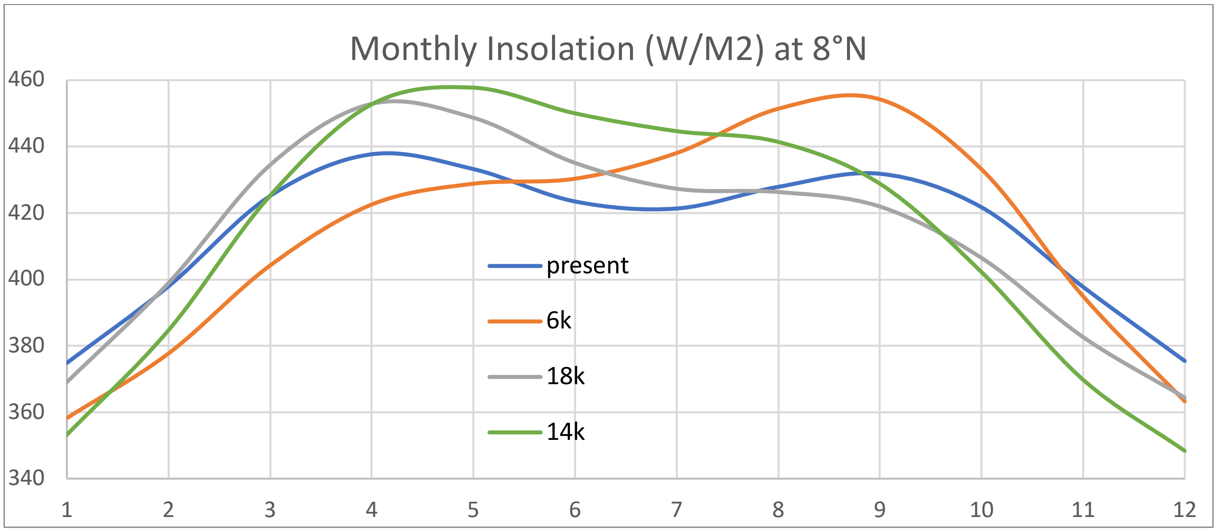

Figure S5. Annual insolation curves at 8°N over selected periods from the last 18,000 years (15) clearly showing the two maximums in April and August/September. Months are in numbers.

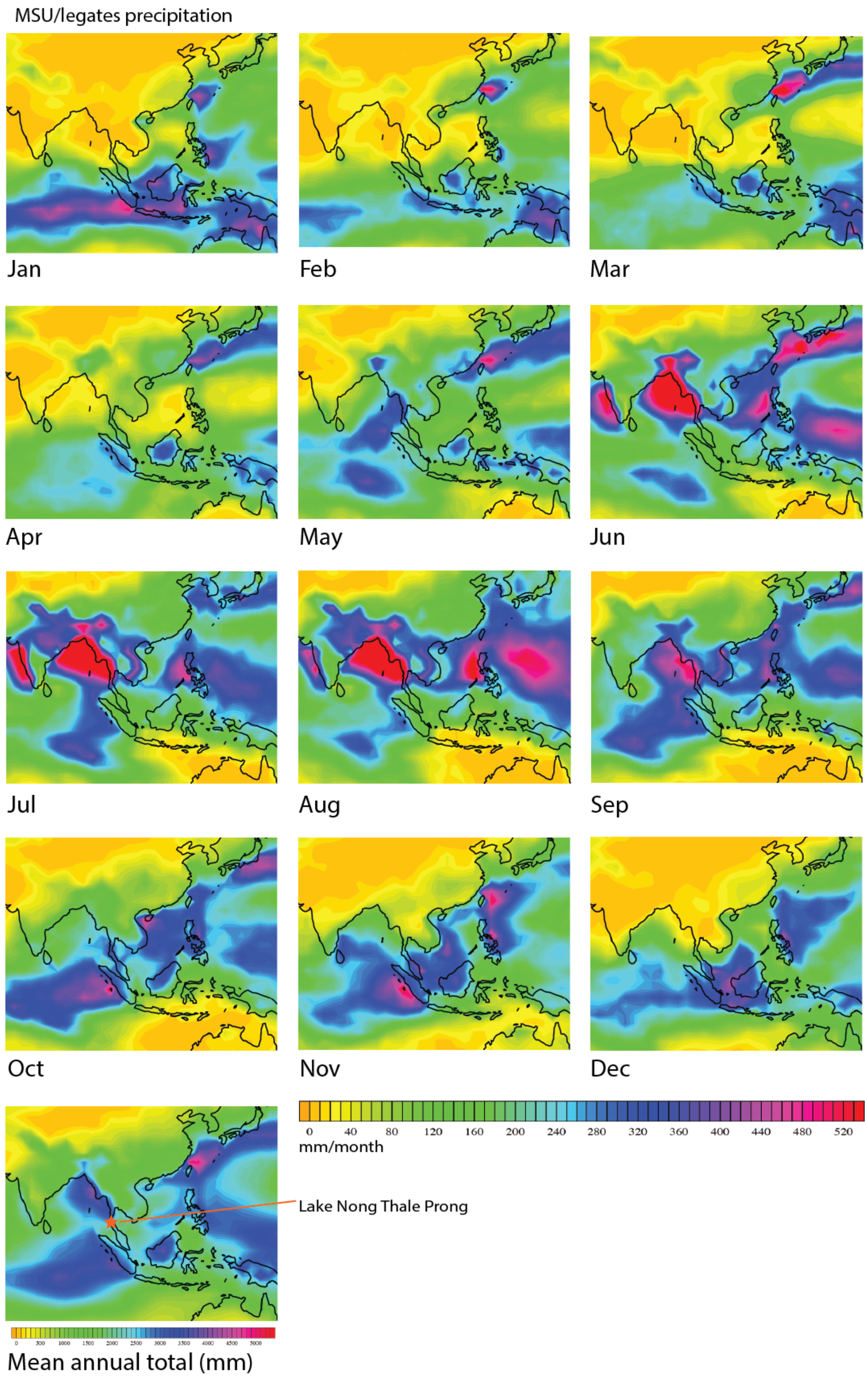
MSU/legates precipitation
Jan
Feb
Mar
Apr
May
Jun
Jul
Aug
Sep
Oct
Nov
Dec
0 40 80 120 160 200 240 280 320 360 400 440 480 520
mm/month
Lake Nong Thale Prong
Mean annual total (mm)

Figure S6. Monthly precipitation of the maritime continent and SE Asia. The wettest months at Lake Nong Thale Prong are associated with the southward passing of the ITCZ from September to November. Maps from http://research.jisao.washington.edu/legates_msu/#analyses (Legates, D. R. and C. J. Willmott, 1990. Mean seasonal and spatial variability in gauge-corrected, global precipitation. Int. J. Climatology, 10, 111-127;   Spencer, R. W., 1993: Global oceanic precipitation from the MSU during 1979-91 and comparisons to other climatologies. J. Climate, 6, 1301-1326)

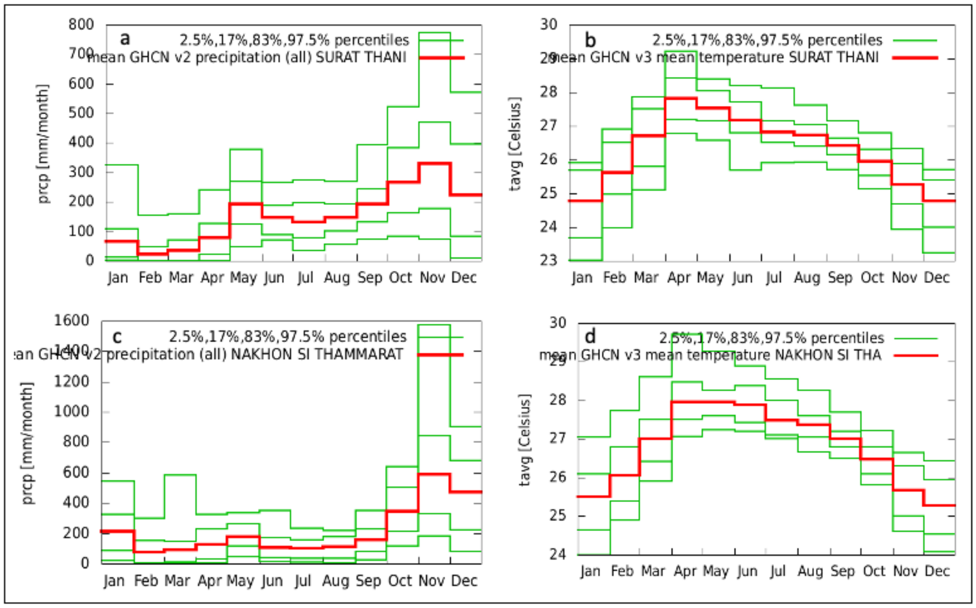

Figure S7. Monthly precipitation and mean temperature data from the two nearest weather stations to lake NTP, Surat Thani (9.12N, 99.35E) (a,b) and Nakhon Si Thammarat (c,d) (8.47N, 99.97E), obtained from the Global Historical Climatology Network (GHCN-Monthly) database Version 2.

The wettest period is September-November, running even into December (left panels); the warmest months are April-May. Reference: Thomas C. Peterson and Russell S. Vose (1997): Global Historical Climatology Network - Monthly (GHCN-M), Version 2. NOAA National Centers for Environmental Information. doi:10.7289/V5X34VDR [accessed 15 October 2020 using http://climexp.knmi.nl]

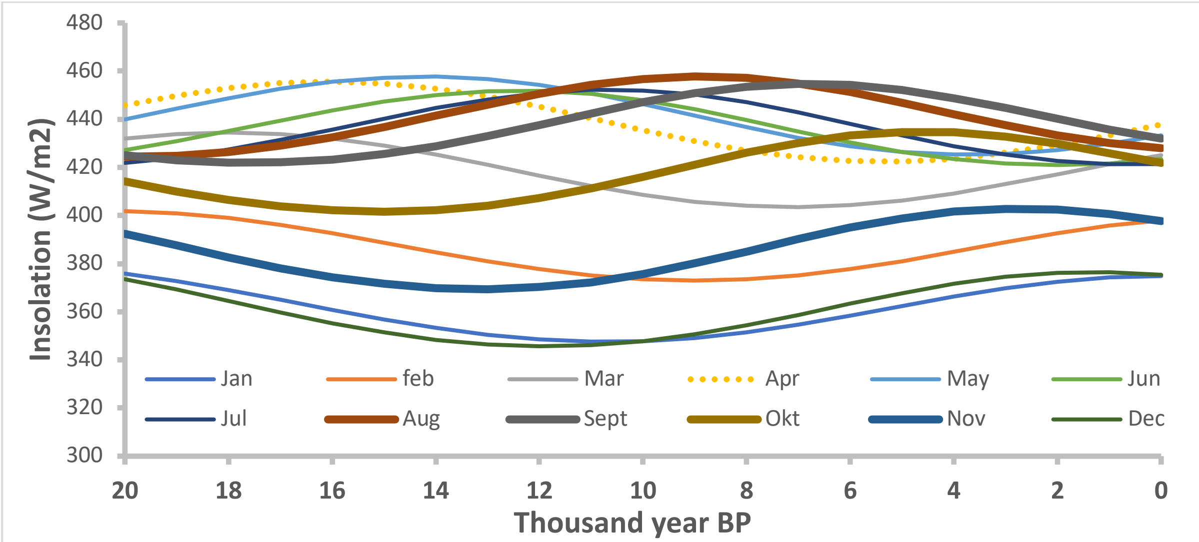

Figure S8. Mean monthly insolation (W/m^2) over the last 20,000 years for 8°N (15), showing the waxing and waning of insolation energy over the precessional cycle for the various months. Insolation maximizes between 6-4 kyr BP for the wettest period SON (See Fig. S7). The insolation curves have the same shape for higher latitudes, but have different absolute values. The mainland SE Asian summer monsoon peaks in JAS, with highest insolation between 10-8 kyr BP and very low insolation at the present. Note that the age axis is reverse compared to proxy records.

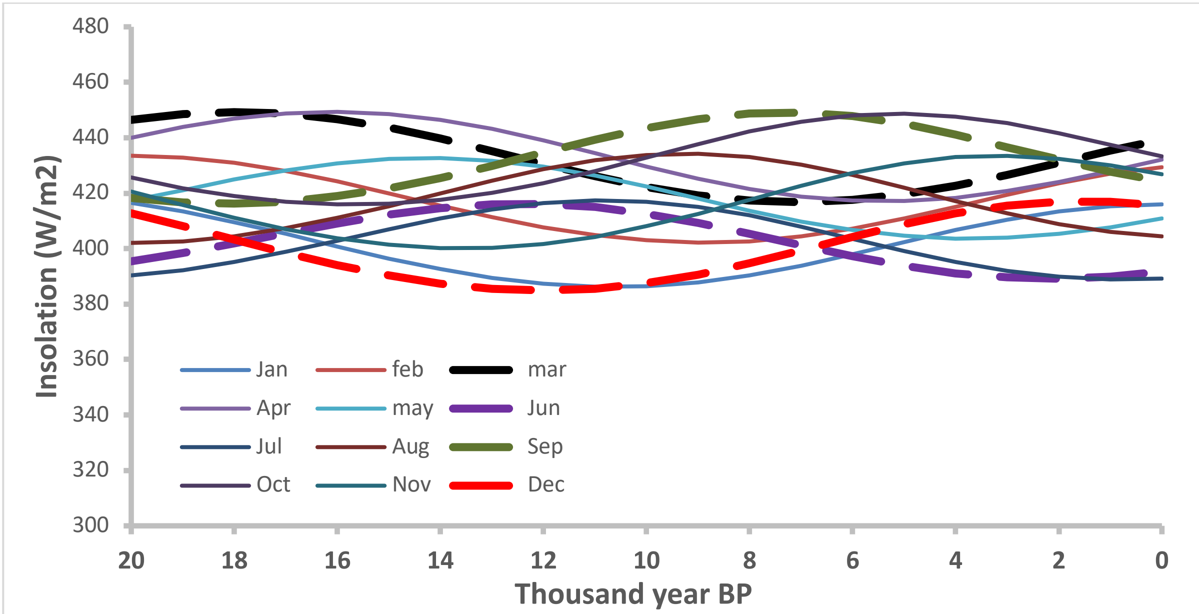

Figure S9. As Figure S8 but for 0° (equator). Present-day June and July insolation are at their precessional low, and these months have correspondingly lowest rainfall amounts (see (10), while the months of December and January, with the same angle of the sun, have stronger insolation and greater rainfall (See Fig S10). Assuming a dominant influence of insolation on convective activity, the annual precipitation patterns likely changes over the course of the precessional cycle.

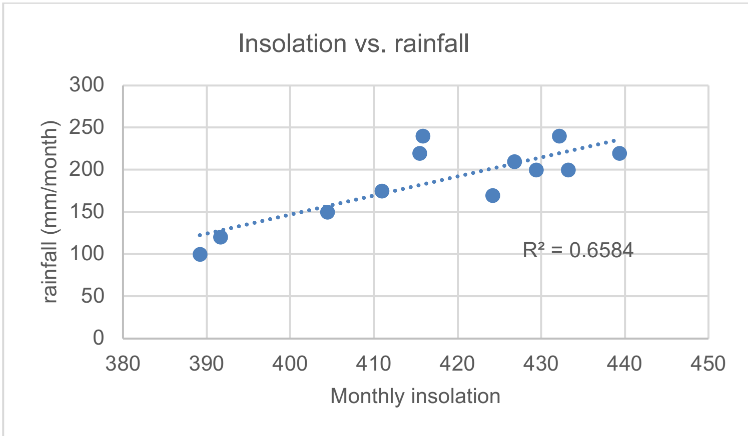

Figure S10. Cross plot of monthly rainfall against monthly insolation for 0° (equator), showing a clear correlation between the two. Rainfall data taken from (10) and insolation for the present day (0 ka BP) of Fig. S9.

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