A 18,000-year Record of Tropical Land Temperature, Convective Activity and Rainfall Seasonality from The Maritime Continent

Rienk H Smittenberg 1*, Kweku A Yamoah 1†, Frederik Schenk 1,2, Akkaneewut Chabangborn 1‡, Sakonvan Chawchai 1‡, Minna Väliranta 3, Barbara Wohlfarth 1

1 Department of Geological Sciences and Bolin Centre for Climate Research, Stockholm University, Stockholm, Sweden.
2 Rossby Centre, Swedish Meteorological and Hydrological Institute, 601 76 Norrköping, Sweden.
3 Environmental Change Research Unit, Department of Environmental Sciences, University of Helsinki, Finland.
† now at School of Geography, University of Birmingham, Birmingham, UK.
‡ now at Department of Geology, Chulalongkorn University, Bangkok 10330, Thailand.

* Corresponding author: rienk.smittenberg@geo.su.se

Highlights
- First continuous, high resolution record of land temperature from the maritime continent
- Land temperatures of the maritime continent were ca. 5°C colder at the last glacial termination compared to today and ca. 2°C warmer during the mid-Holocene climate optimum
- Strong seasonality at the end of the last glacial until the early Holocene caused a savannah vegetation with regular biomass burning
- The strong seasonality greatly influenced dictating vegetation proxies like $\delta^{13}$C<sub>Wax</sub>, by influencing vegetation (C4 Savanna C4 vs C3 rainforest) and the mean annual water isotopic signal recorded as $\delta$D<sub>Wax</sub> and $\delta^{18}$O
- Orbitally forced 'Wet Season Insolation' is positively correlated with both mean annual tropical temperature and the intensity of deep atmospheric convection.
- Indication of megadroughts at the onset of the Meghalayan period

Keywords:
Maritime continent, leaf wax hydrogen isotopes, carbon isotopes, biomass burning, seasonality, paleotemperature, paleohydrology, Late Glacial, Holocene
Abstract

The maritime continent exports an enormous amount of heat and moisture to the rest of the globe via deep atmospheric convection. How this export has changed through time during the Late Glacial period and through the Holocene is hardly known, yet critical for the understanding of global climate dynamics. In this study, we present a very well dated, continuous paleoclimatic and -environmental record from southern Thailand covering the last 18,000 years, including the first land-based temperature reconstruction of the maritime continent. Confirming a recent climate modelling study, we found evidence for a strongly seasonal climate for most of the late glacial period, causing biomass burning and suppression of rainforest growth, despite rising CO$_2$ levels and increasing mean humidity. Temperatures were $ca.$ 5°C cooler than today during the last cold stadial periods, and $ca.$ 2°C warmer between 7000-2000 yr ago. We also found that tropical wet-season insolation (WSI) is a primary driver of the strength of deep atmospheric convection, exerting a strong influence on both the Monsoon systems and the Walker circulation, and hence on global climate dynamics.

1 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 28°C. This region is also called the ‘steam engine of the world’. 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 (De Deckker, 2016). The MC 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) (Timmermann et al., 2018) and Indian Ocean Dipole (IOD) (Mohtadi et al., 2017). 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. Randerson et al., 2005). 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 – approximately the size of the Amazon basin - must have had substantial consequences for global-scale climate dynamics (Koutavas and Joanides, 2012; DiNezio and Tierney, 2013; Di Nezio et al., 2016; Mohtadi et al., 2017; Yamoah et al., 2021). 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 (Sun Xiang Jun and Sun Xiang Jun, 2002; Wang et al., 2009). In contrast, other proxy records from the region indicate drier conditions during the LGM (DiNezio and Tierney, 2013; Dubois et al., 2014). Highest ('driest') δ18O values are recorded in a Borneo speleothem record at that time (Partin et al., 2007), 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 (Heaney, 1991; Wurster et al., 2010; Wurster et al., 2019). A viable explanation lies in rainfall seasonality, which can strongly impact proxy records of both vegetation and the recorded water isotope signal. A recent climate modelling study (Hällberg et al., 2022) found that the tropical SE Asian climate was strongly seasonal during the Late glacial, with very arid conditions during the northern Hemisphere winter period. The seasonal aridity was driven by orbital forcing and stronger East Asian winter monsoon caused by a much larger latitudinal temperature gradient than today. A breakdown of deep convection during NH winters caused a reorganized Walker Circulation and a mean state resembling El Niño conditions. Spatial coverage of high-resolution paleoclimate records from this vast region remains scant, however, 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. In this study we present a high-resolution, very well dated (Fig. S1) 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 m²) karst lake at ~60 m above sea level (Snansieng et al., 1976). More details of the lake setting can be found in an earlier publication (Yamoah et al., 2016) 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 (δ13C_wax) as a proxy for the relative abundance of C3 vs. C4 vegetation, which is influenced by pCO₂, temperature and seasonality (Dubois et al., 2014; Pinto et al., 2014). 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 (Sachse et al., 2012) and seasonality. We also present the first high-resolution land-based temperature record of the MC, based on bacterial-derived branched glycerol dialkyl glycerol tetraethers (brGDGTs) (Sun et al., 2011; Schouten et al., 2013; Russell et al., 2018). The combined proxy records reveal how the aerial exposure and subsequent inundation of Sundaland interacted with orbital and other climate forcings to impact the hydroclimate of SE Asia.
Fig. 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.

2. Materials and Methods

2.1 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 description (Table S1) and 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 bulk isotopes (Table S4), 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 (Yamoah et al., 2016).

2.2 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 (Chawchai et al., 2015). Based on these, an age-model (Fig. S1) was constructed using Bacon, a Bayesian statistics-based routine (Blaauw and Christen, 2011) that estimates the accumulation rate for sediment segments based on the radiocarbon dates calibrated using the intCal13 NH calibration curve (Reimer et al., 2013). Radiocarbon dates are given in Table S2.

2.3 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‰. Results are listed in Table S3.

2.4 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 and used to optimize the concentrations for compound-specific isotope analysis.

2.5 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 0.25 μm). 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. Tierney and deMenocal, 2013) as follows: δD$_{wax-c}$ = (δD$_{wax}$ + 1000) / (δO$_{18}$w * 8* 0.001 + 1) - 1000, with interpolated ocean water δO$_{18}$w values (Waelbroeck et al., 2002).

δ$^{13}$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. δ$^{13}$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‰. Results are presented in tables S5 (leaf wax δD) and S6 (leaf wax δ$^{13}$C).

2.6 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 (Rattray and Smittenberg, 2020). 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 (Rattray and Smittenberg, 2020). MBT and CBT proxies were calculated following Weijers et al. (2007). GDGT results and reconstructed MAAT are presented in table S7.

A basic prerequisite for the valid use of brGDGTs is a relatively high branched-over-isoprenoid tetraether (BIT) index, which was 1.0 throughout the core. Reconstructed pH values, based on the CBT index (Weijers et al., 2007) were 8.0±0.2 over the entire core, with lowest values during the
Younger Dryas 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, and that confounding factors like changes in setting (e.g., between peat/wetland and lake) and pH (De Jonge et al., 2021) are likely minimal. At the time of measurement, we had not adopted the new HILIC-based method which separates between 5-methyl and 6-methyl branched GDGTs (Hopmans et al., 2016) but used our own method based on reverse phase chromatography (Rattray and Smittenberg, 2020), similar to the one used by (Zhu et al., 2013), 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 (De Jonge et al., 2014), peats (Naafs et al., 2017), or East African lakes (Russell et al., 2018). 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 compared to colder environments.

3 Results

3.1 Proxy validation

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 (Weijers et al., 2007), we chose to apply the global lake calibration of Sun et al. (2011) for lakes with pH<8.5, which also included data from nearby lake Towuti:

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

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. The MBT/CBT-based MAAT (°C) reconstruction using the global regression model of Sun et al. (2011) 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) (Harris et al., 2014) 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 $MAAT_{\text{Sun-cal}}$ values obtained using the global regression ($\mu_{\text{proxy-global}}$) and
standard deviation ($\sigma_{proxy-global}$) 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 ($\mu_{obs-local}$) and standard deviation ($\sigma_{obs-local}$) of the local observations for the same time period (Eq. 2).

\[
MAAT_{i,local} = \left( x_{i,proxy-global} - \mu_{proxy-global} \right) \frac{\sigma_{obs-local}}{\sigma_{proxy-global}} + \mu_{obs-local}
\]  

(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 (Weijers et al., 2007), 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 (Russell et al., 2018) is approximately 2.5°C.

$\delta D_{wax}$ of the surface core (Yamoah et al., 2016)(table S8) closely follows the annual precipitation amount (Fig. 2a), where a 10‰ decrease in $\delta D_{wax}$ 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. (Bony et al., 2008) assuming that $\delta D_{wax}$ predominantly reflects $\delta D_{precip}$ 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 (Sachse et al., 2012, 2004; Sessions et al., 1999). In the humid tropics, the fractionation ($e_{wax/water}$) was found to be fairly constant at 130‰ (Feakins et al., 2016). Using this fractionation factor, back-calculated $\delta D$ values for precipitation of the last century ranges between -40‰ and -60‰, reflecting actual measurements for the region (Wei et al., 2018).

MAAT$_{RC}$ and $\delta D_{wax}$ 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 (albedo) and atmospheric absorption of shortwave radiation (Sobel et al., 2010), an effect that is particularly strong in monsoonal Asia (Jalihal et al., 2019).

Fig. 2. Comparison of instrumental climate data with proxy data measured on the NTP surface sediments. a) Mean Annual Air Temperature (MAAT$_{RC-Sun}$) 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 to correspond with the instrumental record (black, left axis). b) Instrumental rainfall data (right axis, stippled red) compared with $\delta$D$_{wax}$ data from the same samples (Yamoah et al., 2016) suggests a good correlation between the two. c) Scatter plot of $\delta$D$_{wax}$ 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 (Harris et al., 2020).

3.2 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 Late Glacial 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 (Lewis Jr, 1996). 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, Fig. S2) and a robust age model showing no large changes in accumulation rate 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 Late Glacial 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).
Fig. 3. Proxy records of lake NTP of the last 18,000 years. a) $\delta^{13}$D$_{\text{wax}}$, both as measured and corrected (stippled) for global sea level change. b) Reconstructed mean annual air temperature ($\text{MAAT}_{\text{RC}}$). c) $\delta^{13}$C$_{\text{wax}}$. d) Atmospheric $\text{CO}_2$ levels (Monnin, 2006). e) Sea level reconstruction for the Sunda Shelf region (Hanebuth et al., 2011). 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.

3.3 Temperature reconstruction

$\text{MAAT}_{\text{RC}}$ (Fig. 3b) stays around 23-24°C during HS1, a 5°C cooling compared to the present. This is more than the most recent estimate for the tropical ocean during the LGM (−4.2 to −3.7°C; (Tierney et al., 2020) but is in line with estimates based on tropical glacier snow line elevations (Porter, 2000). 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.

3.4 $\delta^{13}$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 (Fig. S2) 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. Garcin et al., 2014). 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 (Ehleringer et al., 1997; Collatz et al., 1998; Pinto et al., 2014). Our observation compares well to tropical African records (Street-Perrott, 1997; Cerling et al., 1998; Sinninghe Damsté et al., 2011). Increasing fractionation against $^{13}$C at higher pCO$_2$ levels and greater humidity (Diefendorf and Freimuth, 2017; Hare et al., 2018) – regardless of plant type, can explain part of the trend. 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.

3.5 $\delta$D$_{wax}$ as proxy for precipitation

To further investigate past precipitation changes, we analyzed $\delta$D$_{wax}$, with higher resolution between 17-10 ka BP, to discern trends during deglaciation (Fig. 3a). $\delta$D$_{wax}$ was corrected for the effect of global ice volume (Tierney and deMenocal, 2013). A confounding factor in the interpretation of $\delta$D$_{wax}$ is the potential effect of changing vegetation and associated change in fractionation (Liu and Yang, 2008). 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 (Wang et al., 2013; Garcin et al., 2014). The generally stronger biosynthetic fractionation against deuterium of C3 plants compared to C4 would however lead to an opposite behavior of $\delta$D$_{wax}$ as observed: the increase in C4 during the Bølling period is associated with more negative $\delta$D$_{wax}$, 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 $\delta$D$_{wax}$ values (Liu and Yang, 2008), but again the opposite is observed. From the perspective of vegetation change, our $\delta$D$_{wax}$ record might thus even underestimate the original variations in source water $\delta$D.
4. Discussion

4.1 Late Glacial climate evolution

4.1.1 Inferences from $\delta^{13}$C

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 Late Glacial period. This ecosystem was sensitive to the antagonistic effects of rising pCO$_2$ and rising temperature on C3 versus C4 plants, where higher temperatures and/or lower pCO$_2$ favor C4 plants. However, a third important climatic factor also favors non-perennial C4 vegetation: rainfall seasonality (Dubois et al., 2014). Seasonal dryness was likely promoted by the presence of Sundaland, which only became fully inundated around 11 ka BP during Meltwater Pulse 1b (Hanebuth et al., 2011). 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 between the LGM and the end early Holocene, 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 (Murphy and Bowman, 2007). 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 Late Glacial 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 rising temperatures with the continued rainfall seasonality, thereby offsetting the C3-promoting effect of increasing pCO$_2$. Our interpretation of strong seasonality concurs with CESM1 climate model simulation results that focused on the Late glacial period (Hällberg et al., 2022), which indicate that a large part of island SE Asia, apart from Borneo, experienced not only much reduced total rainfall compared to today, but especially a dry season lasting several months during NH winters. The trend toward increasing C4 (savannah) vegetation reversed after 13.8 ka BP, coinciding with highest MAAT$_{RC}$ and one of the fastest increases of pCO$_2$. Hällberg et al. (2022) found a strong seasonality for both 13 ka BP (i.e., Al) and 12 ka BP (i.e., YD) so gradual return to an increasing contribution of C3 (forest) vegetation was likely driven by the cooling trend that started already during the Al, supported by a further rise in pCO$_2$ during the YD. It is also possible that the local area started to experience a lesser rainfall seasonality, and/or
higher general humidity, due to the ongoing inundation of Sundaland resulting in an ever more maritime climate despite the continuation of strong and dry winter monsoons until the start of the Holocene. The general trend in δ¹³C observed at NTP also is evident in the lower resolution IPWP record from Lake Towuti on Sulawesi (Russell et al., 2014)(Fig. S3), supporting the interpretation of the combined influence of pCO₂ and rainfall seasonality over the entire IPWP over glacial-interglacial timescales.

4.1.2 Inferences from δD_wax

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 (Douglas et al., 2012) occurred during HS1, culminating at Heinrich Event 1. This is followed by a rapid decrease during the Bø, and, similar to the MAAT_RC and δ¹³C records, and subsequent a sharp reversal at the start of the Al. δD_wax, MAAT_RC and δ¹³C track each other until the YD, with lower δD_wax accompanying higher MAAT_RC, suggesting warmer and wetter conditions, and vice versa. This is consistent with inferences from the δ¹³C record of patterns of change in C4 vegetation. The combined records suggest that the period of high rainfall seasonality may have had wetter wet seasons. However, climate model simulations (Hällberg et al., 2022) indicate generally dryer conditions even during the NH summer, besides strong NH droughts.

To explain the steep change in δD_wax between 16 and 14 ka BP, despite NH winter dryness, the convective strength over Sundaland must have increased, caused by 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 (Wei et al., 2018). 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. At that time, 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 (Wei et al., 2018). Lastly, 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.
4.1.3 Inferences from the combined $\delta D_{\text{wax}}$, $\delta^{13}C$, and MAAT$_{\text{RC}}$ records

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 (De Deckker, 2016) with abundant moisture and isotope recycling comparable to the present-day Amazon basin. The parallel reversal of $\delta D_{\text{wax}}$, $\delta^{13}C$ and MAAT$_{\text{RC}}$ 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 $\delta D_{\text{prec}}$ during the cooler winter monsoon months, thereby raising annual mean $\delta D$. 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 Al 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 $\delta D_{\text{wax}}$ 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 (Yuan et al., 2018). After the YD, however, $\delta D_{\text{wax}}$ 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$_{\text{RC}}$. We attribute this to the development of a more equable hydroclimate throughout the year, with an increased relative contribution of ‘dry’-season rainfall with higher $\delta D$ values, sourced from the Gulf of Thailand and the South China Sea.
Fig. 4. **Comparison of isotope records.** 
a) Greenland ice core $\delta^{18}$O as a reference for NH temperature (Andersen et al., 2004). 
b) Combined Chinese speleothem $d^{18}$O (Cheng et al., 2016). 
c) NTP $\delta D_{wax}$ corrected for sea level effect (this study). 
d) Borneo speleothem $\delta^{18}$O record (dark, Partin et al., 2007) and light (Chen et al., 2016), orange. 
e) Sumatra speleothem $\delta^{18}$O record (Wurtzel et al., 2018). 
f) Sulawesi $\delta D_{wax}$ record (Konecky et al., 2016). b-f are all plotted on the same scale where one unit in $d^{18}$O corresponds to 8 units in $d D$ 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 (Laskar et al., 2004). No clear wettest period could be defined for Sumatra. Time periods are shown as in in Figure 3.

4.2 Orbital forcing of Holocene and Late Glacial climate, and seasonality effects

4.2.1 Wet Season Insolation

After 11 kyr BP, $\delta D_{wax}$ 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 $\delta D$ 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 (Clement et al., 1999; Jalihal et al., 2019). 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
$8^\circ$N. Indeed, $\delta D_{wax}$ follows the insolation curve for the wettest months, September-November (Fig.
4) (Laskar et al., 2004), although with a notable excursion during the Bø/AI-YD periods, which we
attribute to the influence of increasing seasonality combined with a much stronger (cold) Asian
Winter monsoon under Late glacial conditions when the higher latitudes were still significantly
colder than today, as discussed above.

The 7% variation of WSI over the last 18,000 years (418 - 446 kW/m$^2$) (Laskar et al., 2004) thus
appears to be a main driver of both surface (temperature) and atmospheric (latent, convective) heat
flux. This observation is consistent with a Borneo ($4^\circ$N) speleothem record (Chen et al., 2016),
where $\delta^{18}$O is correlated with the wettest months at that latitude (Fig. 4). The $\delta D_{wax}$ record from
Lake Towuti (Sulawesi) has been interpreted as being driven primarily by changes in moisture
source and air trajectories (Konecky et al., 2016), but it also shows a strong correspondence with
WSI at $2^\circ$S (Dec&Jan, during the passing of the ITCZ; Fig. 4). Both $\delta D_{wax}$ 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 (Fig. S7), with at present lowest values in June and July (Fig. 5)
(Wurtzel et al., 2018). 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. 5),
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 (Fig. 4).

Figure 5. Monthly insolation curves for 0° (equator). Present-day (0 BP) June and July insolation are at their
precessional low, and these months have correspondingly lowest rainfall amounts (compare with Wurtzel et al.,
2018), while the months of December and January, with the same angle of the sun, have stronger insolation and
greater rainfall (See Fig S7). Assuming a dominant influence of insolation on convective activity, the annual
precipitation patterns likely change over the course of the precessional cycle.

Fig. 6. Annual insolation curves at 8°N over selected periods from the last 18,000 years (Laskar et al., 2004) clearly
showing the two maximums in April and August/September. Months are in numbers.

4.2.2. Relation between WSI and mean annual temperature water isotopes
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. 6), when the sun's altitude is 90° at noon.
The annual movement of the ITCZ and the Monsoon system behaves in an attenuated fashion (Fig.
S5). From January onwards, temperatures rise (Fig. S6) 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. S6). The ITCZ passes over quickly going northwards during May and June, to merge with the Asian Summer Monsoon system during the NH summer (Fig. S5). 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. S5 and S6). 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. S6).

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. 6). This stronger 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. All this explains that lowest δD_{wax} values are observed in the mid Holocene. On top of that, precipitation was likely lower in spring with low insolation levels (Figs. 7 and Fig S5), causing a stronger bias of autumn rainfall towards the annual mean.

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. 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 below.
Figure 7. Mean monthly insolation (W/m²) over the last 20,000 years for 8°N (Laskar et al., 2004), 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. S6). 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.

Looking further back at 20 kyr BP, the seasonal pattern of insolation is similar as today (Fig. 7), but over the ensuing Late Glacial period (i.e., towards 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'; Fig. 7), 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 dD of precipitation. Results for nearby Phuket (Wei
et al., 2018) indicate only a relatively small range in $\delta^{18}$O through the year, from -2‰ (i.e. $\delta D = -7‰$) in April to -8‰ ($\delta D = -55‰$) in November, with an annual weighted average of -5.5‰ ($\delta D = -35‰$). 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 isotopically relatively heavy spring precipitation would have contributed less to the annual mean, while the stronger convective activity during the wet autumn season would have caused more depleted wet season precipitation. Together, this causes a bias towards lower mean annual $\delta D$ 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 is likely a factor that has influenced other isotope records from the tropics as well, instead of being solely caused by changes in rainfall amount or moisture source.

4.3 Relation of IPWP climate with the Asian monsoon system and ENSO

4.3.1 Export and attenuation of 'peak isotope' in SE Asia

Lastly, we discuss 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‰ (Wei et al., 2018)). 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.
Fig. 8. Comparison of the NTP δD\textsubscript{wax} (this study) with the Chinese speleothem δ\textsuperscript{18}O records (Cheng et al., 2016). Both records resemble each other very well, including a number of short-term events like Heinrich Event 1. For reference, the Late Glacial Greenland δ\textsuperscript{18}O (Andersen et al., 2004) record is also plotted. The records are scaled in the same way as in Fig. 4.

4.3.2 Influence of IPWP hydroclimate on the Asian monsoon and ENSO

The trends in the NTP δD\textsubscript{wax} record are similar to those in the Asian speleothem δ\textsuperscript{18}O records (Cheng et al., 2016; Zhang et al., 2019) (Fig. 8), including the OD event and the 'peak isotope' feature at the beginning of the Al. NTP δD\textsubscript{wax} also tracks the Greenland ice core record (Fig. 8), 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 (Wei et al., 2018; Zhang et al., 2019). The shared patterns of variation are consistent with modeling studies (Yang et al., 2014; Pausata et al., 2011), which have shown that East Asian speleothem δ\textsuperscript{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 (Wei et al., 2018). Our results, which are similar to those from a recent study in northern Thailand (Yamoah et al., 2021), demonstrate that the exposure and inundation of Sundaland played a critical role in affecting the water isotopic composition not only across mainland East Asia, but also in Thailand. Hällberg et al. (2022) found a complete breakdown of convection over former Sundaland during NH winter for Lateglacial conditions, resulting in an El-Niño-like mean state with extended dry seasons. They
attribute this mainly to orbital forcing, combined with a still much colder NH hemisphere causing a much large temperature gradient between the tropics and the higher latitudes. The same factors that lowered \( \delta D_{\text{wax}} \) 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, precession-forced changes in WSI in the lower tropics, and seasonality, need to be considered when interpreting SE Asian water isotope records in sediments and speleothems. Experiments with isotope-enabled general circulation models are needed to gain further insight.

4.4. Late Holocene droughts

Another last notable feature of the \( \delta D_{\text{wax}} \) 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 (Kathayat et al., 2018). 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 recent finding (Dang et al., 2020) 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. This mechanism weakened when WSI in the northern IPWP lessened, which may have allowed the crossing of a climate tipping point allowing ever greater 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 and dictating the existence and strength of ENSO. 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.

5. Conclusions

A main conclusion that can be drawn from our multi-proxy record is that seasonality is a major factor that needs to be taken into consideration when interpreting climate and vegetation proxies like \( \delta^{13}C_{\text{wax}} \) and \( \delta D_{\text{wax}} \). Our data show that over the Late Glacial period the aerially exposed Sundaland experienced a more continental and especially more seasonal climate than today with
biomass burning during dry winters, favoring C4 (Savannah) vegetation. This feature is most
apparent during the Bølling period that saw a rapid warming and strong increase in seasonal
precipitation conditions. A key turning point in a tug-of-war between pCO$_2$, temperature and
seasonality as the three driving factors determining the ratio of C3 and C4 vegetation was the Older
Dryas event at 13.8 ka BP, after which climate evolved towards that of the year-round humid
climate known from the present day. Our Holocene record shows a clear mid-Holocene optimum
of deep convection in the northern IPWP, indicating that the ‘steam engine of the world’ was at full
power exporting greater amounts of (latent) heat, i.e. moisture, to the Northern Hemisphere during
this time. This declined over the last 5000 years, with dramatic effects starting in the Meghalayan
age at 4.2 ka BP where we find some evidence of severe droughts. Inferred from our own and from
other records, we argue that ‘wet season’ insolation (WSI), following the precessional cycle,
predicts convective strength, rainfall amount and intensity, distilling into an isotope effect of -1.4‰
for $\delta^D$ (and -0.175‰ for $\delta^{18}O$) per W/m$^2$ for the WSI. It is this weighted mean isotope signal that
gets predominantly recorded in water isotope-based proxies. Moreover, we observe a long-lasting
coupling between the hydrological cycle and MAAT, where temperatures are driven by the
cumulative effects of rainfall (cloud) seasonality imposed by the precessional cycle, even though
mean annual insolation hardly changes. Our first, continuous record of mean annual terrestrial
temperature from tropical SE Asia confirms earlier compiled evidence that tropical temperatures in
the LGM were 4-5°C lower than today. The close resemblance of our record with other Asian
speleothem $\delta^{18}O$ records indicates that the tropical SE Asian climate is dictated by the combined
effects of the precessional cycle, seasonality, and the changing continentality of the IPWP region
over glacial cycles due to sea level change. Our results highlight the importance of the IPWP as the
'steam engine of the world' to global climate, and how it responds to orbital forcing and sea level
change.

Acknowledgments
We wish to thank Sherilyn Fritz, Wichuratree Klubseang and Sudo Inthonkaew for sampling
assistance and discussion. Jayne Rattray and Anna Hägglund and Camilla Bredberg are thanked for
laboratory assistance. Paula Reimer from Queen's University of Belfast conducted the radiocarbon
analyses.

Funding This work was supported by Swedish Research Council (VR) research grants 621-2008-
2855 (RHS), 348-2008-6071 (BW) and 621-2011-4916 (BW).
Author contributions

- Conceptualization: RHS, BW
- Sampling: BW, KAY, AC, SC
- Analysis: BW, KAY, RHS, MV, SC
- Supervision: RHS, BW
- Writing: RHS
- Commenting: BW, KYA, FS, MV, SC, AC

Competing interests: Authors declare that they have no competing interests.

Data and materials availability: The data presented in this paper is available online as csv files and as excel file at the Bolin Centre of Climate Research Database: Barbara Wohlfarth, Rienk Smittenberg (2022) Temperature and hydrological data for the last 18,000 years from Lake Nong Thale Prong, Southern Thailand. Dataset version 1. Bolin Centre Database.
https://doi.org/10.17043/wohlfarth-2022-nong-thale-prong-1

Supplementary Materials

Figures S1-S7

Tables S1-S8 are available at the Bolin Center for Climate Research database:
https://doi.org/10.17043/wohlfarth-2022-nong-thale-prong-1

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- Table 7. GDGTs and reconstructed MAAT
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Supplementary Figures

A 18,000-year Record of Tropical Land Temperature, Convective Activity and Rainfall Seasonality from The Maritime Continent

Rienk H Smittenberg 1*, Kweku A Yamoah 1†, Frederik Schenk 1,2, Akkaneeewut Chabangborn 1‡, Sakonvan Chawchai 1‡, Minna Välimäki 3, Barbara Wohlfarth

1 Department of Geological Sciences and Bolin Centre for Climate Research, Stockholm University, Stockholm, Sweden.
2 Rossby Centre, Swedish Meteorological and Hydrological Institute, 601 76 Norrköping, Sweden.
3 Environmental Change Research Unit, Department of Environmental Sciences, University of Helsinki, Finland.
† now at School of Geography, University of Birmingham, Birmingham, UK.
‡ now at Department of Geology, Chulalongkorn University, Bangkok 10330, Thailand.
*
* Corresponding author: rienk.smittenberg@geo.su.se
**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 δ¹³C, and carbonate content based on loss-on-ignition.
Figure S3. Comparison of the δ¹³C_wax records of lake NTP (this study) and lake Towuti (Russell et al., 2014) and atmospheric CO₂ levels (Monnin, 2006).
Figure S4. a) Reconstructed MAAT using the MBT/CBT ratios according to two calibrations (Weijers et al., 2007)(Sun et al., 2011), and after local recalibration as described in the text. b) reconstructed pH using the CBT ratios (Weijers et al., 2007)(Sun et al., 2011), c) Triplot of the relative abundance of tetra, penta- and hexamethylated GDGTs in the surface core (green); a the pooled soil and peat (B.D.A. Naafs et al., 2017)(B. D. A. Naafs et al., 2017) (orange) and an African lake dataset (Russell et al., 2018)(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.
**Figure S5.** 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 S6. Monthly meteorological data from the two nearest weather stations to lake NTP, Surat Thani (9.12N, 99.35E) and Nakhon Si Thammarat (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 S7.** Cross plot of monthly rainfall against monthly insolation for 0° (equator), showing a clear correlation between the two. Rainfall data taken from (Wurtzel et al., 2018) and insolation for the present day (0 ka BP) of Fig. 6 (main text).

![Insolation vs. rainfall](image)

\[ R^2 = 0.6584 \]

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