Asian monsoon over mainland Southeast Asia in the past 25 000 years

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This thesis consists of a summary, and three published articles (Paper I-III) and one submitted manuscript (Paper IV);

**Paper I**

**Paper II**

**Paper III**

**Paper IV**
To my parents and my wife
Mainland Southeast Asia is located close to the boundary between the Indian Ocean and the East Asian monsoon sub-systems. Strong convective cells develop over this region during the initial stage of the summer monsoon onset; the start of the summer monsoon season is thus earlier than in neighboring areas. The main objective of this research study is to contribute, analyse, and interpret high-resolution palaeo-proxy data sets from this region to further understanding of Asian summer monsoon variability in the past. This was done by compiling, evaluating, and synthesizing published palaeo-records from the Asian monsoon region; through data – model simulation comparisons; and by analysing new lake sedimentary records from northeast Thailand.

Palaeo-records and climate model output indicate a strengthened summer monsoon over Mainland Southeast Asia during the Last Glacial Maximum at 23 000-19 000 years before present (BP), as compared to dry conditions in other parts of the Asian monsoon region. This can be explained by the Last Glacial Maximum sea level low stand, which exposed the continental shelf of Sundaland and created a large land-sea thermal contrast. The rapid sea level rise ~19 600 years BP led to the reorganization of the atmospheric circulation in the Pacific Ocean and possibly played an important role in the weakening of the summer monsoon between 20 000 and 19 000 years BP.

Reconstructions of semi-quantitative precipitation and temperature for Mainland Southeast Asia show that annual rainfall amounted to more than 1100 mm and that annual temperatures were above 18°C between 15 000 and 5000 years BP. The hydroclimatic records for Mainland Southeast Asia, together with those established for the East Asian monsoon sub-region, moreover point to an earlier Holocene onset of a strengthened summer monsoon, as compared to the Indian Ocean monsoon sub-region. The asynchronous evolution of the summer monsoon and the time lag of about 1500 years between the East Asian and the Indian Ocean monsoon sub-regions can be explained by the palaeogeography of Mainland Southeast Asia, which acted as a land bridge for the movement of the Intertropical Convergence Zone.

The new palaeo-proxy records from northeast Thailand compare well to the semi-quantitative data sets and suggest a strengthened summer monsoon between 10 000 and 7000 years BP and a weakening of the summer monsoon thereafter. Warmer temperatures in the Northern Hemisphere during the early Holocene led to a northward shift in the mean position of the Intertropical Convergence Zone, while the mean position of the Intertropical Convergence Zone moved southward during the mid Holocene following the decrease in insolation.

The multi-proxy, high-resolution lake/peat sequence from Lake Pa Kho provides a picture of summer monsoon variability during the last 2000 years. A strengthened summer monsoon prevailed between BC 170 and AD 370, between AD 800 and 960 and since AD 1450, while it was weaker about AD 370-800 and AD 1300-1450. Although these shifts in summer monsoon intensity can generally also be explained by the movement of the mean position of the Intertropical Convergence Zone, the weakening of the summer monsoon between AD 960 and 1450 was possibly affected by changes in the Walker circulation.
Sammanfattning


Resultat från paleo-data och klimatmodeller indikerar en förstärkt sommarmonsun över Sydostasien under senaste istidsmaximum mellan 23 000 och 19 000 år före vår tideräkning (BP) (innan AD 1950), jämfört med torrare förhållanden i andra delar av den asiatiska monsunregionen. Denna skillnad kan ha sin förklaring i havsnivån, som var betydligt lägre under senaste istidsmaximum, vilket blottade kontinentalsockeln Sundaland och skapade en termalt kontrast mellan land och hav. Den snabba stigningen av havsnivån ~19 600 år BP ledde till en omorganisering av den atmosfäriska circulationen i Stilla havet, vilket kan ha bidragit till försvagningen av sommarmonsunen mellan 20 000 och 19000 år BP.

Semikvantitativa rekonstruktioner av variationer i nederbörd och temperatur för Sydostasien visar att den årliga nederbörden var mer än 1100 mm och att temperaturen var över 18°C mellan 15 000 och 5000 år BP. I övrigt indikerar hydroklimatiska proxys för Sydostasien tillsammans med data från den östasiatiska monsunregionen att den förstärkta sommarmonsunen under Holocene påbörjades tidigare än monsunen ovanför den Indiska Oceanen monsunregionen. Den asynkrona utveckligen av sommarmonsunen och tidsskillnaden på 1500 år mellan den Östasien och Indiska Oceanen monsunregionen kan förklaras av palaeogeografien av sydostasien, som fungerade som en landbrygga för den intertropiska konvergenszonen.

De nya paleo-proxy data från nordöstra Thailand är välkorrelerade med de tidigare nämnda semikvantitativa datansen och föreslår en förstärkt sommarmonsun mellan 10 000 och 7000 år BP samt en försvagning av den efterföljande sommarmonsunen. Varmare temperaturer i den norra hemisfären under tidig Holocene ledde till en nordlig skiftning i medelpositionen för den intertropiska konvergenszonen. Den minskade solinstrålningen under mitten av Holocene gjorde att medelpositionen för den intertropiska konvergenszonen flyttade söderut under denna tid.

Multiproxystudien från sjö/torvsekvensen i Pa Kho sjön ger en indikation av variabiliteten av sommarmonsunen under de senaste 2000 åren. En förstärkt sommarmonsun pågick från BC 170 och AD 370, mellan AD 800 och 960 och efter AD 1450, medans den var svagare mellan AD 370-800 och AD 1300-1450. Trots att dessa intensitetsförändringar i sommarmonsunen generellt sett kan förklaras genom förflyttningen av medelpositionen av den intertropiska konvergenszonen så kan försvagningen av sommarmonsunen mellan 960 och 1450 möjlichen bero på förändringar i Walkercirkulationen.
บทคัดย่อ

เอกชนทั้งมวลในเมืองได้ภาคพื้นที่เป็นพื้นที่ที่มีความสำคัญต่อการศึกษาระบบการเปลี่ยนแปลงของสมรภูมิ
เอกชนเป็นอย่างยิ่ง เนื่องจากถูกมองว่าจะได้รับการสนับสนุนทั้งด้านการศึกษาและด้านการพัฒนาอย่างเต็มที่
ได้รับการพัฒนาจากทั่วถึงระบบ นอกจากนี้ยังเป็นพื้นที่ที่มีการพัฒนาของสมรภูมิที่หลากหลาย
บริภัณฑ์ในออกไปโดยเจ้าของในวงการพัฒนาการอุตสาหกรรม โดยการวิเคราะห์สภาพภูมิอากาศในอดีตจาก
งานวิจัยที่รวบรวมได้มาและนำมาใช้เป็นข้อมูลกับแบบจำลองสภาพภูมิอากาศ ปรับปรุงใช้ในการวิเคราะห์พื้นที่
สภาพภูมิอากาศในอดีตจากหลักการพื้นฐานการผลิต และนำมาใช้ในข้อมูลใหม่จากการวิเคราะห์เครื่องมือของ
สถาบันเหล่านี้ทำให้ผู้ร่วมเป็นอดีตการเปลี่ยนแปลงของสภาพแวดล้อมในอดีต

ผลการวิเคราะห์สภาพภูมิอากาศในอดีตจากหลักการพื้นฐานการผลิต ได้แสดงให้เห็นว่า พื้นที่ที่มีการเปลี่ยนแปลงของสมรภูมิไม่เป็นอย่างชัดเจน
ตลอดสองกำลังพลที่ได้มาจากพื้นที่ของสภาพภูมิอากาศ ซึ่งแสดงให้เห็นว่า การเปลี่ยนแปลงของสภาพแวดล้อมในภาคพื้นที่
มีความชุ่มชื้นกว่าบริเวณเมื่อในที่พื้นที่ที่คูคุกคุกกว่าบริเวณที่คูคุกคุกด้วยประมาณ 2300 ถึง 19000 ปีก่อนปัจจุบัน
(ป.ค. 1950) โดยถ้าจะเป็นแผนกจากการที่ระดับน้ำทะเลลดลงถึง 125 เมตรมีถึงบริเวณปัจจุบัน ทำให้พื้นที่
ทวีปยุโรปด้วยพื้นที่ที่อยู่ภายในบริเวณน้ำทะเลที่มาบริเวณพื้นที่ที่คูคุกคุกขึ้นแสดงให้เห็นสมรภูมิที่มี
กำลังพล นอกจากนี้หลักฐานทางการธรรมชาติยังระบุว่ามีการเปลี่ยนแปลงของสภาพแวดล้อมครั้งใหญ่เมื่อประมาณ
2000 ถึง 19000 ปีก่อนปัจจุบันที่ที่คูคุกคุกขึ้นแสดงให้เห็นสมรภูมิที่มีกำลังพลโดยคำว่าเกิด
จากการที่ระดับน้ำทะเลลดขึ้นอย่างรวดเร็วเมื่อประมาณ 19600 ปีก่อนปัจจุบัน ทำให้เกิดการเปลี่ยนแปลงของการ
ไหลของน้ำโดยการเปลี่ยนแปลงของโมเลกุลของน้ำที่มีซึ่งกันและกันกับ

จากการจัดการสภาพภูมิอากาศในอดีตโดยใช้หลักการพื้นฐานของการผลิตและการผลิต แสดงให้เห็นว่าพื้นที่ที่บริเวณนี้โดยทั่วไปมี
อุณหภูมิรายปีสูงกว่า 18 องศาเซลเซียสและมีปริมาณน้ำในการปัจจุบัน 1100 มิลลิเมตรในช่วง 15000 ถึง 5000 ปี
ก่อนปัจจุบัน การที่ระดับน้ำทะเลเพิ่มขึ้นอย่างดังกล่าวที่อยู่ในคุณค่าของโลกร้อน เมื่อถึงบริเวณน้ำทะเลที่คูคุกคุกขึ้น
อย่างน้อยจะเห็นได้จากแผนที่ที่มีช่วงบริเวณที่มีการเปลี่ยนแปลงของสภาพแวดล้อมสูงประมาณ 1500 ปี และสามารถส่งผลต่อสถานที่นั้น
กว่าการเกิดการผลิตสมรภูมิภูมิที่มีน้ำทะเลอักเสบได้เครื่องไม่พร้อมกัน ซึ่งอาจจะเป็นสาเหตุได้สำหรับการที่ลักษณะของ
เอกชนจะเกิดขึ้นไปว่าภาคพื้นที่ที่มีการเปลี่ยนแปลงพื้นที่ที่สูงและน้ำไหลอย่างคงที่ลดลง

ข้อมูลจากเครื่องมือที่ใช้ในการจัดการสภาพภูมิอากาศในอดีตที่มีผลต่อการเปลี่ยนแปลงของสมรภูมิภูมิที่มีช่วง 2000
ปีที่ผ่านมา สมรภูมิที่มีการเปลี่ยนแปลง ระหว่าง 170 ก่อนคริสตศักราชกับ ค.ศ. 370 ระหว่าง ค.ศ. 800-960 และ
หลังจาก ค.ศ. 1450 สมรภูมิที่มีการเปลี่ยนแปลงประมาณ ค.ศ. 370-800 และ ค.ศ. 1300-1450 ซึ่งมากกว่าอื่นๆ
โดยการเปลี่ยนแปลงของร่องกับภูมิอากาศต่ำدرجةสูงที่ต่ำกว่าภูมิอากาศในช่วง
10000 ถึง 7000 ปีก่อนปัจจุบัน และน้ำที่ไหลที่ลงมาที่ครั้งที่สูงจาก 7000 ปีก่อนปัจจุบัน

ข้อมูลที่ได้มีการถูกอนุมานจากประโยชน์ที่มีในการศึกษาการเปลี่ยนแปลงของสมรภูมิภูมิที่มีช่วงในช่วง 2000
ปีที่ผ่านมา สมรภูมิที่มีการเปลี่ยนแปลง ระหว่าง 170 ก่อนคริสตศักราชกับ ค.ศ. 370 ระหว่าง ค.ศ. 800-960 และ
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10000 ถึง 7000 ปีก่อนปัจจุบัน และน้ำที่ไหลที่ลงมาที่ครั้งที่สูงจาก 7000 ปีก่อนปัจจุบัน
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Appendix I i
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1. Introduction

1.1 The Asian monsoon – a short overview

The monsoon is considered a primary component of the global climate system as it transfers heat energy and humidity from the equator to higher latitudes (e.g. Webster et al., 1998; An et al., 2000; Trenberth et al., 2000; Zhan, 2003; Cook and Jones, 2012; Jo et al., 2014). It is generally described as a seasonal reversing wind associated with distinct changes in precipitation, which is caused by land-sea thermal contrasts (e.g. Ramage, 1971; Wang et al., 2003; Wang et al., 2005; Wang, 2009; Clift and Plumb, 2008). Monsoon variability is fundamentally controlled by changes in insolation (external forcing) and by land, ocean and atmosphere interactions (internal forcing) (Clemens et al., 1991). In summer, the increase in insolation leads to the development of strong convective cells over the continent, which allows high humidity air masses to flow from the ocean to the land. The incursion of humid air masses enhances the amount of precipitation over the monsoon region. A decrease in insolation and relatively warm sea surface temperatures (SST) in winter cause a reversal of the winds from the continent to the oceans and leads to cold and dry climate conditions over the monsoon region.

The different monsoon regions have been generally classified according to seasonal differences in surface wind components and are confined to South and East Asia, North Africa, and Australia (Ramage, 1971; Black, 2002; Wang et al., 2005; Wang, 2009; Wang et al., 2014) (Fig. 1A). Recent studies, which are based on the development in meteorological observations and satellite technology, however introduce a new concept, the ‘global monsoon’ (Trenberth et al., 2000; Wang, 2009; Wang et al., 2014); following these studies, monsoon regions are defined by seasonal changes in precipitation associated with the migration of the rain belt following the monsoon trough and the Intertropical Convergence Zone (ITCZ). Since the ITCZ is a global-scale, low-latitude feature, the ‘global monsoon’ concept (Wang, 2009) includes six monsoon regions: Asia, Australia-Indonesia, North and South Africa, and North and South America (Figs. 1B, 1C). Among these six monsoon regions, the Asian and African monsoons are regarded as the most prominent and interesting (Wang, 2009).

Here I follow Wang et al. (2003) and Wang et al. (2005) to constrain the Asian monsoon region; it extends from the Arabian Sea to the South China Sea and from northern Australia to northern China (Fig. 2). According to this definition, the Asian monsoon controls the annual hydroclimate variability over an area, which is a home to more than half of the world's population (An et al., 2000; Wang et al., 2003; Wang et al., 2005; Clift and Plumb, 2008; Buckley et al., 2014). Most of the people living in this part of the world work in agriculture and their produce is strongly dependent on the intensity of annual rainfall (Webster et al., 1998; An et al., 2000; Clift and Plumb, 2008; Buckley et al., 2014). Moreover, the Asian monsoon is the largest of the different monsoon systems; its significant northward incursion onto the continent (~40°N) makes it a potentially important player linking low and high latitudes (Fig. 1A). Rainfall variability over a region with a high population density and the importance of the monsoon as a global-scale climate system makes the monsoon highly sensitive to future climate change; negative or positive changes will have large regional and global scale effects on environment, society and economy (IPCC 2012).

The Asian monsoon system is generally divided into the Indian Ocean (IOM) and East Asian monsoon (EAM) sub-systems, according to summer wind circulation patterns (Wang et al., 2003; Wang et al., 2005) (Fig. 2). The IOM is described by a distinct clockwise circulation across the equatorial Indian Ocean and associated with the Somali Jet and southeasterly winds generated by the Mascarene High. The EAM is sometimes divided into the EAM and the Western North Pacific Monsoon (WNPM), because of the significant relationship between the western Pacific Warm Pool and the El Niño Southern Oscillation (ENSO) (Wang et al., 2003). Since changes in ENSO do not occur on an annual basis and since it originates in the eastern Pacific and outside of the Asian Monsoon region, I here consider this factor as a remote forcing. Consequently, the EAM sub-system can be regarded as a convergence of the Pacific trade wind, the
subtropical front near the east coast of China and the southwest monsoon from the Indian Ocean. The boundary of the IOM and EAM sub-systems extends along longitude 105°E, i.e. from the eastern edge of the Tibetan Plateau through mainland Southeast Asia and into the Indonesian Archipelago (Fig. 2).

Modern observations show that the Asian monsoon is a much more complicated system than a simple large-scale land-sea breeze, as it is modulated by various factors, e.g. ENSO and/or the Western Pacific Warm Pool (Goswami and Xavier, 2005); the Indian Ocean Dipole (IOD) (Weller and Cai, 2014); the Atlantic Meridional Overturning Circulation (AMOC) (Yu et al., 2009); the movement of the ITCZ (Chao and Chen, 2001); change in SST (Li et al., 2001) and atmospheric constituents (Ramanathan et al., 2005).

Climate models simulate an increase in land-sea thermal contrast and a subsequent strengthening of the summer monsoon under future global warming scenarios (IPCC 2012). However, IPCC (2012) stressed that there is low confidence regarding future changes in the Asian monsoon because the different climate models provide highly divergent output. Meteorological observations and data analysis (1950-2000) however provide a more differentiated
picture. Annual precipitation has increased over the western part of China and the southeast coast of China, over the northeastern Tibetan Plateau and the Philippines, and decreased over Southeast Asia; the patterns over India are uncertain (Ramahathan et al., 2005; Cruz et al., 2007; Zhang et al., 2011).

Fig. 2: Association of mean precipitable water content (in colour) and wind circulation patterns (arrows) in winter (January and February) and in summer (June, July and August) based on the reanalysis of the National Centers for Environmental Prediction and the National Center for Atmospheric Research (NCEP/NCAR) from 1981 to 2010. The Indian and East Asian monsoon sub-regions are separated by the white dashed line (105°E).

These uncertainties and discrepancies necessitate a better understanding of the long-term hydroclimate history in the region, which allows placing modern climate change into a geological and historical perspective. Extreme climate states in the past (e.g., the Last Glacial Maximum; Heinrich events), which are recorded in geological archives, moreover provide a possibility to test the performance of climate model and the influence of external and internal forcing.
1.2 The Asian monsoon during the late Quaternary and outstanding questions

Palaeo-studies are mainly restricted to the late Quaternary because this time interval offers more reliable geochronologies and has also been relatively stable from a tectonic perspective. Reconstructed cool and dry conditions between 25 and 17 thousand years (ka) before present (BP) have conventionally been interpreted as a weakening of the Asian summer monsoon (e.g. An et al., 1991; Chen et al., 1997; Huang et al., 1997). Some palaeo-records moreover suggest that cool and dry conditions prevailed over the Asian monsoon region until 12 ka BP (e.g. Bhattacharyya et al., 1989; van der Kaars, 1997; Huang et al., 1999; Anshari et al., 2001; van der Kaars et al., 2001; Mingram et al., 2004). Other studies however point to a gradual increase in precipitation and a stronger summer monsoon between 17 and 12 ka BP (e.g. Vasathy et al., 1988; An et al., 1991; Yang et al., 1995; Chen et al., 1997; Rajagopalan et al., 1997; Williams et al., 2008). Most palaeo-records agree regarding an abrupt increase in precipitation during the early Holocene due to a strengthening of the summer monsoon (e.g. Morinaga et al., 1993; Chen et al., 1999; Hodell et al., 1999; Huang et al., 1999; Dam, 2001; van der Kaars et al., 2001; Wang et al., 2002; Mingram et al., 2004; Zhang et al., 2009). After 7 ka BP, the Asian summer monsoon intensity started to weaken due to the overall decrease in insolation (Yang et al., 1995; Bernasconi et al., 1997; Yan et al., 1999; Mingram et al., 2004; Wang et al., 2010a; Stevenson et al., 2011).

Recent studies with better chronologies however show that the general trend of a strengthening/weakening of the Asian monsoon was punctuated by several shorter intervals of abrupt climate shifts. Short intervals of wetter conditions are for example reported for 25 to 17 ka BP (e.g. Rashid et al., 2007; Saher et al., 2007; Govil and Naidu, 2011; Mahesh et al., 2011) and 14 to 13 ka BP, and drier intervals between 12.5 and 11.5 ka BP and ~8 ka BP (Dykoski et al., 2005; Sinha et al., 2005; Shakun et al., 2007). Palaeo-records from China moreover discuss an asynchronicity between the IOM and EAM sub-systems in respect to the onset of the summer monsoon during the early Holocene (An et al., 2000; Herzschuh, 2006; Maher, 2008; Wang et al., 2010b). These new studies highlight the fact that the temporal variability of the Asian monsoon is still an open question.

Several of these palaeo-records have been assembled and synthesized on a regional scale, e.g. China (Herzschuh, 2006; Wang et al., 2010b), India (Tiwari et al., 2003 and 2011), Southeast Asia (Cook and Jones, 2012), Indonesia, and the northern part of Australia (Reeves et al., 2013). Since the Asian monsoon affects a vast area and since each of these compilation studies used different techniques, proxies and assumptions, valid spatial comparisons across the entire Asian monsoon region are limited.

Although summer monsoon intensity is thought to follow changes in insolation, most of the palaeo-studies agree that the peak in summer monsoon intensity lags maximum insolation by approximately 3 ka (Overpeck et al., 1996). The reason for this lag between maximum summer monsoon intensity and insolation has been debated and has been attributed to a number of factors. ENSO, atmosphere-ocean interactions (Wang et al., 2005), a southward shift of the ITCZ (Broccoli et al., 2006; Zhang and Delworth, 2005; Braconnot et al., 2007), the Pacific Warm Pool (De Deckker et al., 2002; Partin et al., 2007), and North Atlantic circulation (Overpeck et al., 1996; Tiwari et al., 2009; Pausata et al., 2011; Stager et al., 2011) may have played an important role in modulating Asian monsoon variability through time.

Several of these outstanding questions are likely to be related to a lack of high-resolution palaeo-proxy studies from mainland Southeast Asia, which is considered as a key location for the study of the Asian monsoon (e.g. Cook and Jones, 2012; Buckley et al., 2014).

1.3 Environmental setting in MSEA

Southeast Asia is conveniently divided into mainland (continental) and maritime Southeast Asia based on its geographical setting. Mainland Southeast Asia (MSEA) is characterised by
more seasonal climate variability than over the maritime continent of Southeast Asia, which experiences more uniform humid conditions (Chuan, 2005). MSEA covers the area of present day Myanmar, Thailand, Laos, Cambodia, Vietnam, the Malay Peninsula and Singapore.

Orogeny has played a major role in shaping the topography of MSEA. The series of collisions between the Indian and Eurasia plates since the Eocene, accompanied by climate change during the Quaternary, are the main factors behind the geographical development of MSEA (Gupta, 2005). The northern and northwestern part of MSEA is made up of high mountain areas, plateaus and hills, which form part of the Himalayas and which are intersected by deep valleys (Gupta 2005). The central mountain ranges extend into Myanmar and Thailand, along the Malay Peninsula, and into the Indonesian archipelago, while the eastern range continues from northern Laos to southern Vietnam. The major rivers, Irrawaddy, Chao Phraya, Mekong and Sông Hồng (Red River), created wide alluvial plains between the mountain ranges. An exception is the region between the Chao Phraya Plain and the Mekong Lowland, which is separated by the gently sloping Khorat Plateau. The major MSEA rivers originate on the eastern Tibetan plateau, except for the Chao Phraya River, which has its source in the MSEA region.

Fig. 3: (A) Mean position of the Intertropical Convergence Zone (ITCZ) and path of tropical cyclones (dashed orange line) over MSEA for each month. The red and blue arrows mark the southwest and northeast monsoon, respectively. The dashed black line (longitude 105°E) marks the boundary between the Indian Ocean (IOM) and East Asian (EAM) monsoon sub-systems. (B) Monthly rainfall data for 1981 to 2010 shows a distinct difference in seasonal precipitation over Thailand (data from Thai Meteorological Department) (B).
Based on the Köppen-Geiger climate classification, MSEA is assigned to a tropical climate system that is characterised by high annual precipitation, and annual temperatures exceeding 18°C (Peel et al., 2007). Seasonal temperature difference is very small in most of the MSEA region but increases with latitude (Chuan 2005). The National Center for Atmospheric Research (NCAR) Community Climate Model (CCM1) simulates mean annual temperatures of ~25°C, precipitation of 3170 mm a⁻¹ and effective moisture (precipitation minus evaporation) of 1825 mm a⁻¹ over MSEA and the maritime continent (Zhang et al., 1996). Annual precipitation over MSEA is controlled by the Asian monsoon, the seasonal position of the ITCZ and receives contributions from tropical cyclones during the summer/winter and winter/summer transition (Takahashi, 2011; Reid et al., 2013). The establishment of strong convection and reversing winds over the Bay of Bengal, MSEA and the South China Sea occurs during the Asian summer monsoon onset in May (Matsumoto, 1997; Zhang et al., 2005; Buckley et al., 2014). These allow a northward shift of the ITCZ, winds blowing from the surrounding oceans to the continent and transport of humidity (Fig. 3A). The ITCZ reaches the northern part of MSEA in September, and moves southward about a month later. This leads to high precipitation over MSEA between May and October (Fig. 3B). The precipitation amount induced by the Asian summer monsoon, however, also receives a strong influence from tropical cyclones that develop over the west Pacific Ocean, the South China Sea and the north Indian Ocean during the transition from summer to winter and vice versa (Takahashi, 2011). A southward shift of the ITCZ allows the penetration of cold and dry air masses from Siberia, which leads to relatively drier conditions in MSEA between November and February (Figs. 3A, 3B).

1.4 Land-sea configuration during the late Quaternary

Dramatic sea level changes during the late Quaternary strongly altered the land-sea configuration in MSEA (Fig. 4). Since the land-sea distribution is a primary driver of monsoon variability, significant changes in MSEA's palaeogeography may have had an important impact on land-ocean-atmosphere interactions and past Asian monsoon dynamics.

During the last glacial maximum (LGM) (23-19 ka BP) the sea level attained a low stand of ~125 m below present levels (BPL), but increased on average by about 10 m/ka between 19 and 8 ka BP (Voris, 2000; De Deckker et al., 2002; Bird et al., 2005; Clack et al., 2009; Hanebuth et al., 2011) (Fig. 4A). However, this averaged sea level rise was modified by many short intervals of rapid increase in sea level, e.g. at 19.6 ka BP (15 m); 14.5 ka BP (20 m); 8.5 ka BP (5 m); and 7.5 ka BP (5 m) (Hanebuth et al., 2011) (Fig. 4A). At ~4.5 ka BP sea level reached its highest stand of ~5 m above present levels and subsequently decreased toward the present level prior to 2 ka BP (Hanebuth et al., 2011) (Fig. 4A).

Advances in topographic and bathymetric observations allow generating approximate shoreline configurations for the late Quaternary (Figs. 4B-4G). The sea level low stand of ~125 m BPL at ~23 ka BP exposed the Sunda shelf and connected MSEA to Borneo, Sumatra, and Java (Voris, 2000; De Deckker et al., 2002; Bird et al., 2005; Hanebuth et al., 2011) (Fig. 4B). This emerged area was potentially characterised by a low gradient and three major river systems: (i) the Malacca Straits river between the Malay Peninsula and Sumatra running in northwest-southeast direction to the Andaman Sea; (ii) the Siem river extending from the Chao Phaya lowland across the Gulf of Thailand in northwest-southeast direction; and (iii) the North Sunda river flowing from Sumatra to the Gulf of Thailand in northeast-southwest direction (Voris, 2000; Sathiamurthy and Voris, 2006) (Fig. 4B).
Fig. 4: (A) Sea level curve for MSEA for the past 21 ka BP (Hanebuth et al., 2011); (B) palaeo-shoreline at 20 ka BP for a sea level low stand of 125 m below present level (BPL), and at (C) 14.3, (D) 12.7, (E) 10.8 ka BP, (F) 9.5, and (G) 4.5 ka BP for sea level of 80, 40, 20 m BPL, and 5 m above present level, respectively (modified from Voris, 2000 and Hanebuth et al., 2011).
The palaeo-shoreline at ~14.3 ka BP was at 80 m BPL and was fairly similar to the sea level low stand at 23 ka BP. However, a significant transgression occurred near the Siam and North Sunda estuaries (Fig. 4C). This transgression migrated further inland near the Siam, North Sunda and Malacca Straits estuaries ~12.7 ka BP (Fig. 4D). A sediment core obtained off the east coast of the Malay Peninsula indicated that the freshwater lakes or swamps possibly existed in the Gulf of Thailand (Emery and Nino, 1963; Biswas, 1973; Voris, 2000). A significant change occurred at 10.8 ka BP when sea level was about 40 m BPL (Fig. 4E). Although this transgression significantly incised into the Gulf of Thailand, MSEA was still connected to Sumatra and Java. The connection between Borneo and Sumatra became narrower, or a long narrow channel possibly separated them (Voris, 2000). MSEA became likely separated from Sumatra, Java and Borneo at 9.5 ka BP when sea level reached about 20 m BPL (Fig. 4F). The palaeo-shoreline is slightly different from present day during the sea level high stand (5 m above present levels) at ~4.5 ka BP (Fig. 4G). This transgression was only dominant near the main river deltas, i.e. the Irrawaddy, Chao Phraya, and Mekhong Rivers.

1.5 Palaeoclimate studies in mainland Southeast Asia

Mainland Southeast Asia (MSEA) is an important area for Asian monsoon studies because it is located close to the boundary between the IOM and EAM sub-systems and is the only continental area that develops strong convective cells during the initial state of the summer monsoon onset (Zhang et al., 2004; Buckley et al., 2014). However, only a few low-resolution palaeo-records are available for MSEA (see for example Kealhofer and Penny, 1998; Penny 2001; White et al., 2004; Penny and Kealhofer, 2005; Cook et al., 2010; Wohlfarth et al., 2012 for literature overviews). In addition, these studies focused mainly on pollen, coastal change or archaeology, and were published in local sources only (e.g. Pramojanee and Hastings, 1983; Chaimanee, 1989; Udomchoke, 1989; Maloney and McAlister, 1990).

The palaeo-vegetation studies from Thailand by Kealhofer and Penny (1998), Maloney (1999), and Penny (2001), from Cambodia by Maxwell (2001) and Penny (2006), and from Vietnam by Proske et al. (2011) provide an overview of past environmental change based on the succession and distribution of vegetation. Cook and Jones (2012) compiled these studies and assessed them in terms of semi-quantitative climate change during the last 30 ka. Their reconstruction suggests cool/dry and moderate cool/dry conditions from 30 to 13 ka BP, moderate warm/wet and warm/wet conditions from 13 to 6 ka BP, and moderate warm/wet and moderate cool/dry conditions from 6 to 0 ka BP (Cook and Jones, 2012). These palaeoenvironmental reconstructions however assume continuous deposition based on the few poor chronologies and thus dismiss breaks in deposition, which are frequently found in recent studies (e.g. Smittenberg et al., 2011; Wohlfarth et al., 2012; Chawchai et al., 2013). Moreover, the general assumption of cold/dry or warm/wet conditions is possibly an overestimation since palaeo-proxies may respond to either temperature or precipitation.

1.6 Thesis objectives

This thesis research is part of the Asian monsoon project (Table 1). The specific focus of my thesis is:

- To contribute new well-dated lake/peat sequences and high-resolution proxies in MSEA
- To better understand hydroclimate patterns reflected in Late Quaternary Asian summer monsoon variability.

My contribution to the Asian monsoon project is detailed in Table 1.
Table 1: Summary of my contribution to the Asian monsoon project. The location of the different sites is shown in Figure 5. {\textsuperscript{1}}Included in this thesis; \textsuperscript{2}ongoing project; \textsuperscript{3}no sediments present in lake basin; \textsuperscript{4}future studies.

<table>
<thead>
<tr>
<th>Site of studies</th>
<th>Fieldwork and/or coring</th>
<th>Laboratory work</th>
<th>Data interpretation</th>
<th>Figures and Tables</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Lake Kumphawapi\textsuperscript{1,2}</td>
<td>x</td>
<td>x</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>2. Lake Pa Kho\textsuperscript{1,2}</td>
<td>x</td>
<td>x</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>3. Lake Han Sakon Nakhon\textsuperscript{3}</td>
<td>x</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4. Lake Laung\textsuperscript{3}</td>
<td>x</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5. Lake Kway\textsuperscript{3}</td>
<td>x</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6. Lake Phayao\textsuperscript{3}</td>
<td>x</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>7. Lake Leng Sai\textsuperscript{2}</td>
<td>x</td>
<td>x</td>
<td></td>
<td></td>
</tr>
<tr>
<td>8. Lake Sifai\textsuperscript{3}</td>
<td>x</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>9. Lake Boraphet\textsuperscript{3}</td>
<td>x</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>10. Sam Roy Yod\textsuperscript{2}</td>
<td>x</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>11. Lake Thale Pron\textsuperscript{2}</td>
<td>x</td>
<td>x</td>
<td></td>
<td></td>
</tr>
<tr>
<td>12. Lake Thale Song Hong\textsuperscript{3}</td>
<td>x</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>13. Lake Thale Bun\textsuperscript{3}</td>
<td>x</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14. Cave exploration for speleothems\textsuperscript{2}</td>
<td>x</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>15. Myanmar project\textsuperscript{4}</td>
<td>x</td>
<td></td>
<td>x</td>
<td></td>
</tr>
</tbody>
</table>

Fig. 5: Location of the sites investigated within the Asian monsoon project in MSEA. The red squares represent the palaeo-records included in this research thesis and the yellow squares mark ongoing projects. The blue circle shows the location of a future project in northwest Myanmar. The lake sites with no sediment present are marked with white circles. See Table 1 for details.
2. Methods of study

The thesis consists of three published articles and one submitted manuscript. Papers I and II are compilations of palaeo-records from the entire Asian monsoon region during the Last Glacial Maximum (LGM) (23-19 ka BP) and from MSEA between 10.5 and 5 ka BP. Papers III and IV focus on a reconstruction of hydroclimate conditions using multi-proxy lake sediments from Thailand, which is located almost in the middle of MSEA.

2.1 Palaeo-records compilation

Published palaeo-records were assembled based on the criteria detailed in Table 2. All published \(^{14}\)C dates were calibrated with the Calib 6.0 online program ([http://calib.qub.ac.uk/calib/calib.html](http://calib.qub.ac.uk/calib/calib.html)) (Reimer et al., 2009). Age-depth curves were further created for each of the sites listed in Paper I. However, for Paper II, the poor chronology of several of the MSEA palaeo-records did not allow creating valid age-depth curves. We therefore only used pollen-stratigraphic intervals that are covered by a \(^{14}\)C date.

<table>
<thead>
<tr>
<th>Table 2: Selection criteria for the palaeo-data set compilation of Papers I and II.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Paper I</td>
</tr>
<tr>
<td>1. Published marine and terrestrial proxy records from the Asian monsoon region (15°S-40°N, and 40–160°E) (see Fig. 6) covering the Last Glacial Maximum were selected (see details in Appendix I).</td>
</tr>
<tr>
<td>2. Each record should contain at least one (^{14})C, U/Th, and/or TL date between 23 and 19 ka BP.</td>
</tr>
<tr>
<td>3. Records with age errors of &gt;1000 years and (^{14})C dates on carbonate bulk sediment were excluded.</td>
</tr>
</tbody>
</table>

2.1.1 Palaeo-proxy assessment in Paper I

The various palaeo-proxies (pollen, faunal remains, \(\delta^{18}\)O of speleothems and planktonic foraminifera, \(\delta^{13}\)C measured on bulk sediment) were assessed in terms of qualitative effective moisture (P-E) and precipitation during the LGM (23-19 ka BP) (Clark et al., 2009). The obtained results were subsequently compared for each type of proxy to reconstruct hydroclimate patterns and, in a second step compared to simulation output derived from the Climate Community System Model version 3 (CCSM3).

Reconstructed P-E and precipitation were assessed by the following procedures. Pollen records were biomized following the approach of Prentice et al. (1996), which is based on the relationship between plant functional types (PFTs) and biomes. These relationships have been established for China (Yu et al., 2000a), Japan (Takahara et al., 2000), Southeast Asia, (Pickett et al., 2004) and the IOM region (Kramer et al., 2010). Although affinity scores were not calculated here, qualitative P-E was directly approximated from the biomes based on the relationship between biomes and mean annual precipitation and temperature, as suggested by Mader (2012). Inferred P-E was grouped into three categories, representing low (1), medium (2) and high (3) (see Table 3). For the temporal analysis, the expansion/shrinkage of grassland was assessed using non-arboreal pollen taxa (e.g. *Artemisia*, Compositae, Rosaceae, Chenopodi-
aceae, Poaceae) and higher/lower run-off was reconstructed using changes in other major ecological groups (e.g., Pteridophyta spores). Changes in grassland and run off were assumed to reflect local and regional hydroclimate changes.

Fig. 6: Location of marine and terrestrial records included in the compilation of Paper I. See Appendix I for details on each record.

Fig. 7: Location of published pollen-stratigraphic records included in Paper II. See details for each record in Appendix II.

For the speleothem records, we assumed that the speleothems grew in a closed system ($\Delta T \sim 0^\circ C$) and that variations in $\delta^{18}O$ reflect the precipitation amount over the site (Wang et al., 2001; Yuan et al., 2004). The mean LGM precipitation was reconstructed based on average $\delta^{18}O$ values between 23 and 19 ka BP for each site following the categorization in Table 3. The
speleothem $\delta^{18}$O values were then smoothed using a five-point running mean to provide insight into Asian summer monsoon variability between 25 and 17 ka BP.

$\delta^{18}$O records of *Globigerinoides ruber* (*G. ruber*) were selected from marine stratigraphies. For this, we assumed that (i) LGM sea level was stable (Clark and Mix, 2002; Clark *et al.*, 2009), (ii) major change in salinity did not occur during the LGM (Schulz *et al.*, 1998; Clark and Mix, 2002), and (iii) the $\delta^{18}$O values of LGM global ocean effect were constant ~1.1‰ higher than present (Adkin *et al.*, 2002; Ravelo and Hillaire-Marcel, 2007). SST differences were calculated from the MARGO (2009) and the World Ocean Atlas (WOA) 1998 (Conkright *et al.*, 1998) data sets, and converted to $\delta^{18}$O values using the relationship of a 0.25‰ increase in $\delta^{18}$O with a SST decrease of 1°C (Erez and Luz, 1983). P-E was calculated by subtracting the $\delta^{18}$O values of the LGM global ocean effect and the SST difference from the $\delta^{18}$O values of *G. ruber*. The obtained P-E values were then smoothed using a five-point running mean between 25 and 17 ka BP.

Table 3: Climate variables used to assess P-E and precipitation in Paper I.

<table>
<thead>
<tr>
<th>Palaeo-proxy assessment</th>
<th>Effective moisture</th>
<th>Qualitative precipitation</th>
<th>Speleothem $\delta^{18}$O values</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Biome</td>
<td>P-E</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Warm-temperate rain forest, tropical deciduous broadleaf forest and woodland, wet sclerophyll forest, and broadleaved evergreen/warm forest</td>
<td>&lt;-1.4</td>
<td>&lt;-7.70</td>
</tr>
<tr>
<td>2</td>
<td>Mix of steppe and warm mixed forest</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Cool mixed forest, mix of steppe and cool mixed forest and steppe</td>
<td>&gt;0.3</td>
<td>&gt;-0.58</td>
</tr>
</tbody>
</table>

Bulk sediment $\delta^{13}$C, hardwood and faunal remains were used as supporting information following the original interpretation of the respective authors.

*Climate model output data*

The Community Climate System Model version 3 (CCSM3) is a fully coupled global atmosphere-ocean-sea-ice-land surface climate model (Collins *et al.*, 2006). CCSM3 output compares well with palaeo-data in other studies about extreme climate conditions (e.g. Kjellström *et al.*, 2009; Brandefelt *et al.*, 2011). The CCSM3 simulation used here is described in detail in Otto-Bliesner *et al.* (2006) and Brandefelt and Otto-Bliesner (2009). CCSM3 simulated precipitation and P-E were compared with the qualitative data set of the compiled palaeo-proxy data. The mean position of the ITCZ was further calculated based on the simulated precipitation according to the method outlined by Braconnot *et al.* (2007) as:

$$ \text{loc}_{-}\text{ITCZ}(\text{lon}) = \frac{\sum_{y=\text{lat}(\text{pr} \text{ max})}^{30^\circ N} pr(y) \text{lat}(y)}{\sum_{y=\text{lat}(\text{pr} \text{ max})}^{30^\circ N} pr(y)} $$

where $\text{loc}_{-}\text{ITCZ}(\text{lon}) = \text{mean position of the ITCZ in each longitude}$
lat(pr max) = latitude of maximum precipitation in each longitude

pr(y) = precipitation over latitude y

2.1.2 Reconstruction of semi-quantitative precipitation and temperature in Paper II

The pollen diagrams were digitized and the biomization followed Prentice et al. (1996). The biomization is based on the relationship between PFTs and the biomes that had been established by Picket et al. (2004). The affinity scores were calculated as:

\[ A_{i,k} = \sum \delta_{i,j} \sqrt{p_{j,k} - \theta_j} \]

where \( A_{i,k} \) represents the affinity score of biome \( i \) in sample \( k \).

\( \delta_{i,j} \) is entry of taxon \( j \) in biome \( i \) based on taxon×biome matrix (if taxon \( j \) is included in biome \( i \), \( \delta_{i,j} = 1 \), otherwise \( \delta_{i,j} = 0 \)).

\( p_{j,k} \) is the percentage of pollen taxon \( j \) in sediment sample \( k \).

\( \theta_j \) denotes the threshold of pollen taxon \( j \) which here is 0.5% in order to eliminate the influence of long-transported pollen (e.g. Prentice et al., 1996; Herzschuh et al., 2006).

Affinity score percentages of each biome were subsequently calculated and analysed together based on the assumption that all pollen taxa contained in the sediment samples had been directly or indirectly influenced by hydroclimatic variations. Semi-quantitative precipitation and temperature were separately assessed using the Southeast Asian and Pacific (SEAPAC) model of biome distribution in climate space (Pickett et al., 2004) (Table 4).

Precipitation — The precipitation range of each time interval \( (P_t) \) was assumed to be a function of precipitation \( (P_i) \) and affinity score percentage \( (x_j) \) of each biome (see Table 4) (Picket et al., 2004):

\[ P_t = \left( \frac{x_1}{100} \right) P_1 + \left( \frac{x_2}{100} \right) P_2 + \left( \frac{x_3}{100} \right) P_3 \quad (P_1 < P_2 < P_3) \]

where \( P_1 \) is the precipitation of temperate or tropical grassland and xerophytic shrubland (STEP), and of xerophytic woods/scrub (XERO) biomes. We here assume that \( P_1 \) is less than 500 mm a\(^{-1}\) (Pickett et al., 2004).

\( P_2 \) represents the precipitation of tropical deciduous broadleaf forest and woodland (TDFO), and of temperate sclerophyll woodland and shrubland (DSFW) biomes and is estimated to be between ~500 and ~1100 mm a\(^{-1}\) (Picket et al., 2004).

\( P_3 \) denotes the precipitation of tropical semi-evergreen broadleaf forest (TSFO), wet sclerophyll forest (WSFW), tropical evergreen broadleaf forest (TRFO), warm-temperate rainforest (WTRF), cool-temperate rainforest (CTRF) and cool evergreen needle-leaf forest (COCO) biomes. \( P_3 \) is here assumed to be higher than ~1100 mm a\(^{-1}\) (Picket et al., 2004).

\( x_1, x_2 \) and \( x_3 \) are the sums of the percentages of the affinity scores of STEP and XERO, of TDFO and DSFW, and of TSFO, WSFW, TRFO, WTRF, CTRF and COCO, respectively;

\[ x_1 = x_{STEP} + x_{XERO} \]
\[ x_2 = x_{TDFO} + x_{DSFW} \]
\[ x_3 = x_{TSFO} + x_{WSFW} + x_{TRFO} + x_{WTRF} + x_{CTRF} + x_{COCO} \]
We further divided $P_3$ into $P_{3,1}$ and $P_{3,2}$ which represent the precipitation of TSFO and WSFW, and TRFO, WTRF, CTRF and COCO biomes, respectively. The moisture index ($\alpha$) between $P_{3,1}$ and $P_{3,2}$ is 0.8 and $P_{3,1}$ is higher than $P_{3,2}$ (Picket et al., 2004). $x_{3,1}$ and $x_{3,2}$ are the affinity score percentages of $P_{3,1}$ and $P_{3,2}$, respectively. The semi-quantitative precipitation was approximated from $x_{3,1}$ and $x_{3,2}$ ratio, which deviate from the moisture index of 0.8.

Table 4: Conceptual model of biome distribution in climate space (modified from Picket et al., 2004).

<table>
<thead>
<tr>
<th>Increasing mean temperature of the coldest month</th>
<th>Increasing moisture index ($\alpha$)</th>
<th>P~500 mm a$^{-1}$</th>
<th>P~1100 mm a$^{-1}$</th>
<th>$\alpha$: 0.8</th>
</tr>
</thead>
<tbody>
<tr>
<td>Xerophytic woods/scrub (XERO)</td>
<td>Tropical deciduous broadleaf forest and woodland (TDO)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Temperate or tropical grassland and xerophytic shrubland (STEP)</td>
<td>Tropical semi-evergreen broadleaf forest (TSFO)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Temperate sclerophyll woodland and shrubland (DSFW)</td>
<td>Tropical evergreen broadleaf forest (TRFO)</td>
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</tr>
<tr>
<td></td>
<td>18°C</td>
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<tr>
<td></td>
<td>Frost limit</td>
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<tr>
<td></td>
<td>5°C</td>
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<tr>
<td></td>
<td>Cool temperate rain-forest (CTRF)</td>
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<tr>
<td></td>
<td>Cool evergreen needle-leaf forest (COCO)</td>
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<td></td>
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<tr>
<td>Tundra (TUND)</td>
<td></td>
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</tbody>
</table>

**Temperature** — Semi-quantitative temperatures at each time interval ($T_i$) were estimated following the same assumption and procedure as for precipitation. STEP and XERO biomes were here excluded because of their distribution in all climate spaces (Picket et al., 2004) (Table 4).

$$T_i = \left(\frac{y_1}{100}\right)T_1 + \left(\frac{y_2}{100}\right)T_2 \quad (T_1 < T_2)$$

where $T_1$ and $T_2$ represent the temperatures of the DSWF, WSWF, WTRF, CTRF and COCO biomes, and of the TDO, TSFO and TRFO biomes, respectively.

$y_1$ and $y_2$ are the percentages of the affinity score of $T_1$ and $T_2$. 

22
\[ y_1 = y_{DSFW} + y_{WSFW} + y_{WTRF} + y_{CTR} + y_{COCO} \]
\[ y_2 = y_{TTR} + y_{NT} + y_{TRF} \]

\( T_1 \) is less than \( T_2 \) which is higher than 18°C. Semi-quantitative temperature was assessed by \( y_2 \) and \( y_1 \) which deviated from 18°C.

### 2.2 Lake sediment analysis

Sediment cores from Lakes Kumphawapi and Ph Kho in northeast Thailand were collected during fieldwork in January 2010 and 2011 using a modified Russian corer (7.5 cm, diameter) (sites #1 and 2 in Fig. 5 and Fig. 8). To achieve a continuous sequence, we collected each one-meter sequence with an overlap of 50 cm.

The sediment cores were transported to the Department of Geological Sciences, Stockholm University and stored in a cold room prior to sub-sampling and laboratory analyses.

The sediment cores were carefully cleaned prior to sub-sampling and the stratigraphy of each sediment core was described in detail. Specific marker horizons were used to correlate between overlapping core segments to create a composite sequence for each site.

Each 1-m long sediment core was subsequently scanned every 5 mm using an Itrax XRF core scanner with a Mo tube (30 kV and 30 mA for measuring 60 sec/point). All elements obtained from XRF scanning were smoothed using a 3-point running mean to create 1-cm resolution profiles. Some selected elements were normalized by incoherent (Compton) and coherent (Rayleigh) scattering and then used in a statistical approach to assess their relationship and to reconstruct environmental changes.

Consecutive 1-cm samples were dried overnight at 105°C to measure loss-on-ignition (LOI). The dried samples were combusted for 6 hr at 550°C and for 4 hr at 950°C, respectively. The weight loss of the samples combusted at 550°C and 950°C, respectively was calculated in respect to the dried sample weight and is reported as % LOI.

Selected sediment samples were freeze-dried and homogenized prior to analyses of total organic carbon (TOC), nitrogen (TN) and sulphur (TS), and their isotopic composition (\( \delta^{13}C \), \( \delta^{15}N \) and \( \delta^{34}S \)). These proxies were analysed using a Carlo Erba NC2500 elemental analyzer connected to a Finnigan MAT Delta+ mass spectrometer. \( \delta^{13}C_{org} \), \( \delta^{15}N \) and \( \delta^{34}S \) were measured relative to the Vienna PeeDee Belemnite (VPDB), AIR and Canyon Diablo Troilite (CDT) standard, respectively. Analytical errors for \( \delta^{13}C_{org} \) and \( \delta^{15}N \) were ±0.15‰ and for \( \delta^{34}S \) measurements the uncertainty was ±0.2‰. The carbon:nitrogen ratio (C/N) was multiplied by 1.167 to yield atomic ratios.

For biogenic silica (BSi), sediment samples were pre-cleaned with \( H_2O_2 \) and HCl to remove organic matter and carbonate. The cleaned samples were then extracted by 40 mL of 1% Na\(_2\)CO\(_3\) solution for 5 hr and neutralized with 0.21N HCl. The dissolved silica concentration was analysed after the extraction by using ICP-OES (Varian Vista Ax) and corrected by a simultaneous dissolution of minerogenic silica to determine weight percentages (wt %) of BSi.

Sediment samples were sieved under running tap water by using sieves with a mesh size of 0.5 mm. The sieved plant remains were stored in deionized water. The plant remains were carefully identified under a stereomicroscope, cleaned multiple times with deionized water, and dried overnight at 105°C in glass vials. Selected samples were sent to the ¹⁴CHRONO Centre at Queen’s University, Belfast for radiocarbon analysis. The radiocarbon age and one standard deviation were calculated following the conventions of Stuiver and Polach (1977) using the Libby half-life of 5568 years and a fractionation correction based on \( \delta^{13}C \) measured on the AMS. The age-models were constructed using Bacon, a Bayesian statistics-based routine that models accumulation rates by dividing a sequence into many thin segments and estimating the (linear) accumulation rate for each segment based on the (calibrated) \( ^{14}C \) dates, together with
assumptions about accumulation rate and its variability between neighbouring segments (Blaauw and Christen, 2011).

Fig. 8: (A) Location of Lakes Kumphawapi and Pa Kho, and (B) topography around the lakes and coring points. Sediment cores from Lake Kumphawapi (core KMP-CP4) and Lake Pa Kho (core PK-CP3) were used for Papers III and IV.
3. Summary of the papers

Below I outline my contributions to the analysis, data interpretation and writing of the four papers included in this thesis (Table 5).

Table 5: Author contribution to the analysis, data interpretation and writing of the four papers included in this thesis.

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4. Results

The results from the three published papers and one submitted manuscript included in this thesis are summarized below.

4.1 Paper I: Asian monsoon climate during the Last Glacial Maximum: palaeo-data—model comparisons

The Asian monsoon region was separated into three sub-regions: the IOM, EAM, and Sundaland and Sahulland (SSS) (Fig. 9). The IOM and EAM sub-regions were divided along longitude 105°E following Wang et al. (2003) and Wang et al. (2005). The LGM sea level low stand exposed the vast area of the Sunda and Sahul continental shelves, which are included as a third domain (Bird et al., 2005; De Deckker et al., 2002; Hanebuth et al., 2011) (Figs. 9A, 9B).

Palaeo-data and model comparison

Indian Ocean monsoon sub-region – Enrichment in δ¹⁸O values of speleothems from the western Arabian Sea (site #33) (Shakun et al., 2007) implies low precipitation, which agrees well with the CCSM3 simulation (<1000 mm a⁻¹) during the LGM (Table 3) (Fig. 9A). Reconstructed P-E based on planktonic foraminifera obtained from the northern Arabian Sea is low (site #34) (van Rad et al., 1999), while high P-E was reconstructed by the same proxy off south India (site #29) (Tiwari et al., 2006). These wet and dry conditions are consistent with the model output (Fig. 9B).

Simulations of moderately wet conditions (0-800 mm a⁻¹) compare moderately well with reconstructions of high P-E based on δ¹⁸O values of planktonic foraminifera in marine records from the Bay of Bengal (site #32) (Rashid et al., 2007) (Fig. 9B).

Data-model discrepancies are visible in Sri Lanka (site #30) (Premathilake and Risberg, 2003; Premathilake, 2006), southern India (site #31) (Rajagopalan et al., 1997), and in the northern part of the IOM sub-region (site #35) (Sharma and Chatterjee, 2007); (site #36) (Kotlia et al., 2010); (site #37) (Fujii and Sakai 2002; Hayashi et al., 2009); (site #38) (Cook et al., 2011); (site #39) (Tang et al., 1998; Tang et al., 2004); (site #40) (Wattanapituksakul, 2006) (Figs. 9A, 9B). CCSM3 simulation output shows moderate P-E, while pollen records from Sri Lanka (site #30) (Premathilake and Risberg, 2003; Premathilake, 2006) and the northern IOM sub-region (site #36) (Kotlia et al., 2010); (site #37) (Fujii and Sakai 2002; Hayashi et al., 2009); (site #38) (Cook et al., 2011); (site #39) (Tang et al., 1998; Tang et al., 2004), a distinct barren pollen zone from northern India (site #35) (Sharma and Chatterjee, 2007) and a dominance of dry-tolerant C4 plants in sediments from the Nilgiri Hills (site #53) (Rajagopalan et al., 1997) indicate arid conditions during the LGM (Figs. 9A, 9B). Reconstructions of wet conditions based on faunal remains from west Thailand (site #40) (Wattanapituksakul, 2006) also diverge from the CCSM3 simulations of moderately wet conditions.

Sundaland and Sahulland sub-region - Reconstructed wet conditions correspond well with simulated high precipitation (>2000 mm a⁻¹) and P-E (>800 mm a⁻¹) over the SSS sub-region during the LGM (Figs. 9A, 9B). Pollen assemblages were assigned to warm-temperate rain forest and tropical deciduous broadleaf forest and woodland biomes in western (site #7) (Newsome and Flenley, 1988); (site #9) (Maloney and McCormac, 1996); (site #8) (Maloney, 1980), southern (site #2) (Stuijts et al., 1988); (site #3) (van der Kaars and Dam, 1997); (site #4) (van der Kaars et al., 2010), and central (site #6) (Anshari et al., 2001) Sundaland; from the northern Molucca Sea region (site #10) (Barmawidjaja et al., 1993), the Banda Sea (site #5) (van der Kaars et al., 2000) and the eastern part of the SSS sub-region (site #1) (Hope, 2009), which implies high P-E (Table 3) (Fig. 9B). The plant remains of Dipterocarpaceae, which can be assigned to warm-temperate rain forests (Pickett et al., 2004), also indicate high P-E over the Malay Peninsula (site #11) (Wüst and Bustin, 2004) during the LGM. In addition, wet con-
ditions are consistent with low $\delta^{18}O$ values of speleothems from northern Borneo (site #12) (Partin et al., 2007) (Table 3).

Fig. 9: Comparison between qualitative reconstruction and quantitative simulation of (A) precipitation and (B) P-E during the LGM. The CCSM3 precipitation was classified as low (<1000 mm a$^{-1}$), medium (1000-2000 mm a$^{-1}$) and high (>2000 mm a$^{-1}$). Simulated P-E was also categorized as dry (<0 mm a$^{-1}$), moderately wet (0–800 mm a$^{-1}$), and wet (>800 mm a$^{-1}$) conditions (modified from Chabangborn et al., 2013).

**East Asian monsoon sub-region** - Pollen assemblages from Southern China (site #16) (Zheng and Lei, 1999) (site #17) (Wang et al., 2010c) were assigned to a mix of warm mixed forest and steppe biomes and indicate moderate P-E during the LGM (Table 3). While in Japan (site #26) (Hayashi et al., 2010); (site #27) (Takahara and Takeoka, 1992); (site #28) (Yasuda, 1982) the assemblages were assigned to and a mix of steppe and cool mixed forest biomes suggesting low P-E, (Table 3). These hydroclimate conditions compare well with CCSM3 model output of
moderate P-E (0-800 mm a⁻¹) over the southern part of China and low P-E (>800 mm a⁻¹) over Japan (Fig. 9B).

Reconstructions of low P-E based on pollen stratigraphies from Korea (site #25) (Chung et al., 2006) and from the Chinese loess plateau (site #23) (Sun et al., 1997); (site #24) (Wang and Sun, 1994) generally disagree with simulations of moderate P-E (Fig. 9B). The reconstructed palaeo-vegetation from Taiwan (site #18) (Liew et al., 2006) indicates high P-E which markedly diverges from simulations of low P-E. Data-model discrepancies are also found in the reconstructed precipitation based on δ¹⁸O values of speleothems from Jintawan Cave (site #20) (Cosford et al., 2010) and Hulu Cave (site #21) (Wang et al., 2001) (Fig. 9A). The low δ¹⁸O values of the speleothems imply wet conditions, while CCSM3 simulates moderately wet conditions during the LGM (Fig. 9A).

Fig. 10: (A) Qualitative precipitation change inferred from biomes and δ¹⁸O of speleothems in the Asian monsoon region between 25 and 17 ka BP. Thick orange and blue lines represent the calibrated age points and the thin orange and blue lines are interpolations between these. The blue and orange colours represent relatively wet and dry climatic condition, respectively. (B) The map shows the location of the sites included in the temporal variability analysis (modified from Chabangborn et al., 2013). See Appendix I for details on the sites.
Temporal variability

Temporal changes in hydroclimate patterns were assessed using palaeo-records with at least two dating estimates between 25 and 17 ka BP. In the IOM sub-system, most palaeo-records indicate that hydroclimate conditions shifted from dry to wet between 20 and 19 ka BP (site #30) (Prematilake and Risberg, 2003; Prematilake, 2006); (site #33) (Shakun et al., 2007); (site #35) (Kotlia et al., 2010); (site #36) (Fujii and Sakai, 2002; Hayashi et al., 2009); (site #37) (Cook et al., 2011) (Fig. 10).

A pollen stratigraphy from western Sumatra (site #7) (Newsome and Flenley, 1988) and δ¹⁸O values of speleothems from northern Borneo (site #12) (Partin et al., 2007) indicate wetter conditions ~20 ka BP and drier conditions thereafter (Fig. 10). In contrast, a decrease in P-E was reconstructed by pollen records from the Sumatra plateau for approximately the same time interval (site #9) (Maloney, 1980); (site #7) (Maloney and McCormac, 1996) (Fig. 10).

Pollen records from South China (site #17) (Wang et al., 2010c) and Taiwan (site #18) (Liew et al., 2006) depict drier conditions around ~20-19 ka BP (Fig. 10). On the other hand, the decrease in δ¹⁸O values of speleothems from sites around 30°N (site #34) (Cosford et al., 2010); (site #35) (Wang et al., 2001); (site #36) (Zhou et al., 2008) imply an increase in precipitation between ~20 and ~19 ka BP. These wetter conditions seem consistent with a reconstructed increase in P-E based on pollen records from Japan (site #42) (Yasuda, 1982). Other pollen records from Japan (site #26) (Hayashi et al., 2010) however do not show any hydroclimate change throughout the LGM.

4.2 Paper II: Climate over mainland Southeast Asia 10.5-5 ka BP

The number of data points available for this reconstruction varied greatly (Fig. 11A). No data were available for the interval between 12.5-11.5 ka BP and the few data points between 15-12.5 ka BP were excluded from the published paper. However, there are 12-17 data points for 8.5-7 ka BP, and 3-5 data points for 10.5-8 ka BP and for 7-5 ka BP.

Pollen assemblages were mainly assigned to warm-temperate rainforest (WTRF), tropical deciduous broadleaf forest and woodland (TDFO), tropical semi-evergreen broadleaf forest (TSFO), and tropical evergreen broadleaf forest (TRFO) biomes (Fig. 11B). These biomes indicate that precipitation and temperature generally exceeded 1100 mm a⁻¹ and 18°C over MSEA between 15 and 5 ka BP (Picket et al., 2004).

Semi-quantitative precipitation and temperature variability

The dominance of the WTRF biome (x_{WTRF} >50%) implies cool and wet conditions from 15 to 13.5 ka BP (Figs. 11B, 11C). x_{WTRF} abruptly decreases at ~13 ka BP when x_{TDFO} and x_{TSFO} rapidly increase. These changes in pollen assemblages suggest a climate shift from cool/wet to warm/dry climate conditions over MSEA ~13 ka BP (Fig. 11C).

x_{TDFO} is almost constant ~40%, while x_{WTRF} varies by 23-44% from 11.5 to 5 ka BP (Fig. 11B). x_{TSFO} increases from 13 to 22% ~10 ka BP and varies about 20% thereafter. x_{TRFO} increases from 5 to 10% between 11.5 and 10 ka BP and gradually decreases from 10 to 5 ka BP. The biome assignments thus indicate that precipitation decreased between 11.5 and 10.5 ka BP. The amount of precipitation seems to have been similar at 10.5 and at 7 ka BP and generally declined after 7 ka BP (Fig. 11C). However, this precipitation pattern was punctuated by dry intervals between 10-9 ka BP and 6.5 ka BP. Despite insignificant changes in temperature, warmer intervals coincide with drought periods (Fig. 11C).
Spatial analysis

Precipitation and temperature over MSEA during the early (10.5-8.0 ka BP) and mid Holocene (8.0-5.0 ka BP) were reconstructed based on the semi-quantitative data set (Fig. 11). The inferred biomes suggest wet to moderately wet conditions over MSEA during the early Holocene (Fig. 12A). Reconstructed dry conditions in site #3 (Kealhofer and Penny, 1998) and #5...
(Li et al., 2012) however are possibly caused by the influence of local conditions (e.g. catchments, soil type and/or human disturbances), or due to the low diversity pollen spectra. High to moderate precipitation during the mid Holocene was reconstructed for sites located south of 12°N (Fig. 12B). The reconstruction of wet conditions over site #4 (Penny, 2001) may come from an underestimation of the non-arboreal pollen assemblages as discussed by Burke (2014). Pollen records suggest warm to moderately warm conditions over MSEA between the early and mid Holocene (Figs. 12C, 12D). Reconstructed cool conditions at site #13 (Rugmai, 2006) may also be due to the low diversity pollen spectra.

Fig. 12: (A, B) Precipitation and (C, D) temperature over MSEA during the early (10.5-8 ka BP) and mid Holocene (8-5 ka BP). The grey land areas show the palaeo-shoreline after the sea level rise at 10.5 ka BP and for the sea level high stand at ~6 ka BP (Hanebuth et al., 2011; Chabangborn and Wohlfarth, 2014).
4.3 Paper III: Lake Kumphawapi – an archive of Holocene palaeoenvironmental and palaeoclimatic changes in northeast Thailand

Several lake sediment cores and a variety of proxies were analysed from Lake Kumphawapi, Thailand (Figs. 5, 8) to reconstruct hydroclimatic changes during the Holocene. Sediment sequence KMP-CP4, which is the focus of this paper, was collected from the southern part of the lake. The age-depth model suggests continuous sediment and peat deposition until 6 ka BP, a hiatus of several thousand years, and sediment deposition since 1.6 ka BP.

The lowermost sediments (units 1a, b) between 7.00 and 3.80 m date to 9.8-7.6 ka BP and are made up of silty gyttja clay and gyttja clay with low %LOI and %TOC (units 1a, b). This suggests low lake organic productivity (Fig. 13). The increase in Zr/Rb ~6.00-5.50 m, which indicates coarser grain sizes (Dypvik and Harris, 2001), coincides with an increase in the C/N ratio (Figs. 13, 14C). This could point to higher run off to the lake between 9.80 and 9.60 ka BP and in-wash of terrestrial organic matter. Si, Ti, Rb, and Ca values are well correlated with each other between 5.40-3.80 m, which would suggest transport of clay- and silt-sized particles to the lake and relatively stable lake conditions.

The organic matter content increases slightly in the gyttja clay of unit 2a (Fig. 13). The geochemical proxies suggest a stable lake with increasing aquatic productivity and less run off between 7.6 and 7 ka BP.

Fig. 13: Lithostratigraphy, %LOI, %TOC, C/N ratios, %BSi and $\delta^{13}$C values of core KMP-CP4 (Chawchai et al., 2013).
The transition between the gyttja clay in unit 2a and the peaty gyttja and peat in unit 2b together with changes in the geochemical proxies suggest that the lake developed into a wetland and eventually a peatland. Peaks of Si, Zr, Ti and Zr/Rb in the peat of unit 2b compare well with low %LOI and %TOC between 2.91 and 2.93 m (Figs. 13, 14B) and might be attributed to wind transport under dry conditions. The lithological and geochemical proxies thus indicate a gradual decrease in P-E between 7 to 6.5 ka BP. The hiatus, which is present between 6 and 1.6 ka BP may also relate to dry conditions and to no or low sedimentation/peat growth.

Fig.14 : XRF core scanning profiles of major elements (Si, Zr, K, Rb, and Ca), and Zr/Rb ratio of (A) units 3a-3b, (B) unit 2b and (C) unit 1a-2a (modified from Chawchai et al., 2013).

The deposition of clayey gyttja and algae gyttja in units 3a and 3b suggests the reestablishment of a shallow lake and possibly an increase in P-E after 1.6 ka BP (Fig. 13). The increase in %LOI and %TOC and the decrease in the C/N ratio suggest an increase in the lake
organic productivity and a predominance of aquatic sourced organic matter ~0.6 ka BP (Fig. 13). All major elements (Si, Zr, K, Ti, Rb, and Ca) decrease at 2.50 m, which would indicate lower run off into the lake (Fig. 14A). However, the sediments and the geochemical proxies do not show signals of higher run off. This discrepancy may be explained by the influence of human activities (higher nutrient transport due to agriculture) after 0.6 ka BP, since archaeological evidence suggests a settlement at Ban Don Kaew ~1.1 ka BP (Fig. 8).

4.4 Paper IV: Hydroclimatic shifts in northeast Thailand during the last two millennia – the record of Lake Pa Kho

The sediment sequence used in Paper IV is derived from Lake Pa Kho, which is located ~15 km to the southwest of Lake Kumphawapi (Fig. 8). Paper IV focused on lake status changes and hydroclimatic conditions based on multi-proxy analyses of the sediment sequence between 2 to 3.5 m, corresponding to BC 170 - AD 2001.

Lithological units 5 (3.50-3.36 m) and 4 (3.36-3.34 m) consist of fine detritus gyttja and peaty gyttja, respectively. These sedimentary units are characterised by a slight change in C/N ratios (27-24) and by a gradual decline in $\delta^{13}$C values (-24 to -21‰), which imply a mix of terrestrial and aquatic sources of the organic matter (Fig. 15). This is consistent with the presence of both phytolith and diatoms, and of terrestrial, telmatic and aquatic species in the plant macrofossil record. The elevated values of Si, Ti and Ca suggest that run off was important, which is also supported by the terrestrial plant input. These evidences indicate that Lake Pa Kho was a shallow water lake, which received run off from the surrounding catchment between BC 170 and AD 370.

The lithological transition from peaty gyttja (unit 4) to compact peat (unit 3: 3.34-2.73 m) at ~AD 370 is marked by a decrease in $\delta^{13}$C values, and the presence of phytoliths and of terrestrial plant remains (Fig. 15). $\delta^{13}$Corg values increase again between 2.98 and 2.93 m (AD 410-450) and fluctuate between -30 and -27‰ until 2.68 m (AD 800). Despite relatively constant %TOC, the gradual decline in C/N ratios from 2.93 to 2.68 m can be potentially explained by the consumption of organic carbon during peat decomposition (Chimner and Ewel, 2005; Ise et al., 2008). High amounts of charcoal can be observed between 2.98 and 2.73 m (AD 410-650). Plant remains change from terrestrial species at the bottom to terrestrial/telmatic species at the top of unit 3. The multi-proxies thus indicate low P-E from AD 370 to 800, which was however punctuated by a short interval of higher P-E between AD 410 and 450.

The re-establishment of a wetland and higher P-E may be inferred from the abrupt increase in $\delta^{13}$Corg values and the shift from terrestrial/telmatic plant species to telmatic/aquatic plant species between AD 800 and 970 (Fig. 15). Although the age model suggests a 330-year long hiatus at 2.63 m, the stratigraphy does not give any indications for a break in peat growth. Therefore we present two alternative hypotheses: one including a hiatus, and one inferring slow peat growth.

The $\delta^{13}$Corg values of -28‰ indicate deposition of predominantly terrestrial plant remains between 2.63-2.62 m, which would suggest lower P-E between AD 1300-1340. This interval is followed by a gradual increase in $\delta^{13}$Corg values, by the presence of diatoms and a shift from telmatic to aquatic plant species, which all signify wetter conditions after AD 1450. The step-wise increase in Si, K, and Ti at 2.17 m depth (AD 1850) and at 2.05 m depth (AD 1970) could be related to higher run-off due to land-use changes.
Fig. 15: Lithostratigraphy, and profiles of geochemistry, charcoal, plant macro remains and selected elements for sediment sequence PK-CP3 (Chawchai et al., submitted).
5. Discussion
5.1 The last glacial maximum (23-19 ka BP): Paper I

Reconstructed P-E and precipitation agree well with CCSM3 output, which implies wet conditions in the SSS domain as well as off southern India, moderately wet conditions in south China, and dry conditions in the west Arabian Sea (Figs. 9A, 9B). Reconstructions of wet conditions over the Bay of Bengal compares fairly well with CCSM3 simulations of moderately wet conditions. Data-model discrepancies are observed in Sri Lanka, southern India and north of latitude 30°N, where palaeo-proxies suggest dry conditions, but CCSM3 suggests wet conditions (Fig. 9B). Differences between reconstructions and simulations are also found in northwest Thailand and central China, where palaeo-proxies provide wetter conditions than those simulated by CCSM3 (Fig. 9A).

The data–model discrepancies can be caused by the palaeo-proxy assessment and the CCSM3 simulation. The qualitative data assessed by palaeo-proxies are difficult to categorize with a proper method. The reconstructed conditions consequently may not have been as dry or wet as here assumed. Moreover, the proxies are possibly showing more response to other environmental and climatic parameters other than precipitation or P-E. The δ¹⁸O values of speleothem records from China, for example, are generally interpreted as precipitation amount (e.g. Wang et al., 2001; Cosford et al., 2010), while recent studies challenge this and argue that they possibly reflect changes in atmospheric circulation (Clemens et al., 2010; Pausata et al., 2011; Maher and Thomson, 2012). On the other hand, the low-resolution of CCSM3 simulation and the selected LGM boundary conditions may not well capture ocean-atmosphere interactions and other control parameters, e.g. LGM dust (Otto-Bliesner et al., 2006; Marchant et al., 2007).

A weakening of the summer monsoon over the IOM and the northern EAM sub-regions was inferred from reconstructed dry conditions, which could be explained by a decrease in land-sea thermal contrast caused by a slight change in palaeo-shorelines and a decrease in sea surface temperatures (Yokoyama et al., 2000; MARGO, 2009; Hanebuth et al., 2009, 2011). Aridity may have been caused by reduced transport of humid air masses to the western part of the IOM sub-region, and a strengthening of the Siberian high-pressure system and westerly winds in the northern part of EAM sub-region (An et al., 2012). On the other hand, reconstructed wet conditions in the southern part of the EAM domain and in the SSS sub-area, where a vast area was exposed by the LGM sea level low stand, indicate a strengthening of the summer monsoon. The stronger summer monsoon intensity in the SSS sub-region and in south China can be explained by an increase in land-sea thermal contrast cause by decreased SSTs and increased land areas (De Deckker et al., 2002; Bird et al., 2005). All these together indicate that the LGM mean position of the ITCZ was located off southern India and in the southern part of China (Fig. 16). Reconstructions of the mean position of the ITCZ correspond well with CCSM3 output in the west Arabian Sea, and compare moderately well over MSEA. Palaeo-data – model discrepancies are observed in southern India and in the Bay of Bengal, where the CCSM3 simulated ITCZ shifts northward, while the palaeo-records suggest dry conditions.

A distinct hydroclimate reversal, potentially related to the northward shift of the mean position of the ITCZ, is reconstructed from wetter conditions in the northern part of the EAM sub-region and in the IOM domain (Figs. 10, 16, 17). A strengthening summer monsoon and a northward shift in the mean position of the ITCZ was possibly caused by the rapid sea level rise, which led to the reorganization of the atmospheric circulation at ~19 ka BP (Hanebuth et al., 2011) (Fig. 4).
Fig 16: Reconstruction of the mean position of the ITCZ (dashed red line) in the Asian monsoon region during the last glacial maximum (LGM) (23-19 ka BP) and 19.5 ka BP (modified from Chabangborn et al., 2013). The green lines represent the seasonal shift of the present-day ITCZ according to Wang (2009). The grey land areas mark the palaeo-shorelines during the LGM sea level low stand (~120 m) and after the rapid sea level rise ~19.5 ka BP following Hanebuth et al. (2011).

5.2 The early to mid Holocene (10.5-5 ka BP): Paper II and III

The semi-quantitative data set indicates that precipitation and temperature generally exceed 1100 mm a\(^{-1}\) and 18°C, respectively, over MSEA between 15 and 5 ka BP. In addition, the data set suggests that cool/wet and warm/dry conditions alternated. This assumption diverges from the general assumption of cool/dry and warm/wet conditions (e.g. Mingram et al., 2004; Herzschuh et al., 2009; Zhang and Mischke, 2009). The association of cool/wet and warm/dry conditions can be explained by the importance of cloud cover that plays an important role in Earth's energy budget in this area (Trenberth and Shea, 2005). A decrease in the amount of cloud cover due to lower humidity can enhance solar irradiance and thus causes warmer surface temperature, and vice versa.

The reconstruction of high precipitation over MSEA (unpublished data sets) compares well with speleothem records from the IOM and EAM sub-systems (Sinha et al., 2005; Dykoski et al., 2006; Partin et al., 2007; Shakun et al., 2007; Yang et al., 2010), which indicate a strengthening of the summer monsoon over the entire Asian monsoon region between 15 and 13 ka BP (Figs,
The interval of intense summer monsoon coincides with warmer conditions in the Northern Hemisphere, and concomitant cooler conditions in the Southern Hemisphere, as seen from Greenland and Antarctic ice core records (NGRIP Members, 2004; EPICA Community Members, 2006). The strengthening of the summer monsoon can be explained by a northward shift of the mean position of the ITCZ, which was caused by the Northern and Southern Hemisphere temperature gradient (Linzen and Nigam, 1987; Lindzen and Hou, 1988; Chiang and Bitz, 2005; Kang et al., 2009; Cvijanovic, 2013).

Fig. 17: Summary of hydroclimate variability over mainland Southeast Asia from 25 ka BP to present day, based on the results of Papers I, II, III and IV.

Drier conditions associated with warmer temperature over MSEA occurred between 13 and 12.5 ka BP. This is slightly later than the decrease in precipitation reconstructed for Oman (Shakun et al., 2007) and Borneo (Partin et al., 2007), but earlier than that observed for India (Sinha et al., 2005) and China (Dykoski et al., 2006; Yang et al., 2010) (Figs. 11, 17). The interval of a weaker Asian summer monsoon corresponds with the Younger Dryas cooling in northern hemisphere records (e.g. Rasmussen et al., 2006; Broecker, 2010; Fiedel, 2011; Björck, 2013; Carlson, 2013). Cooler temperatures in the Northern Hemisphere led to a decrease in the north-south temperature gradient and subsequently to a southward shift of the mean position of the ITCZ (e.g. Stouffer et al., 2006; Yancheva et al., 2007; Gasse et al., 2008; Calson, 2013).

Distinctly higher precipitation over MSEA is observed already ~11.5 ka BP (unpublished data set), which is earlier than the increase in precipitation over the EAM and IOM sub-systems (Dykoski et al., 2005; Fleitmann et al., 2007; Berkelhammer et al., 2012) (Figs. 11, 17). Wet conditions over MSEA possibly relate to the transport of humidity from Borneo (Partin et al., 2007) over the exposed land areas in the Sundaland and correspond in time with the predominance of La Niña-like conditions at the west coast of Mexico (Marchitto et al., 2010).

The semi-quantitative precipitation reconstructed for MSEA compares moderately well with the qualitative P-E inferred from the new studies at Lake Kumphawapi. The latter studies indicate wet conditions between 10 and 7 ka BP and dry conditions after 7 ka BP (Fig. 17).
Fig. 18: Reconstruction of the mean position of the ITCZ (dashed red line) in the Asian monsoon region at 10.8, 9.5 and 6.0 ka BP. The red arrows represent the direction of the ITCZ movement with respect to the reconstructed mean position of the ITCZ. The green lines show the seasonal shift of the present-day ITCZ according to Wang (2009). The grey land areas are the palaeo-shorelines at 10.8, 9.5 and 6.0 ka BP, following Hanebuth et al. (2011).
Reconstructed wet conditions over MSEA and the EAM between 10 and 7 ka BP differ from the precipitation pattern over the IOM sub-region, where $\delta^{18}O$ values of speleothems from Mawmluh and Quft Cave indicate an insignificant change in precipitation between 10.5 to 9.0 ka BP, but show an increase after 9 ka BP (Fleitman et al., 2005; Berkelhammer et al., 2012). A strengthening of the summer monsoon seems thus to have occurred earlier in the EAM sub-system than in the IOM sub-system. These results support the hypothesis of an asynchronous onset of the Holocene optimum (An, 2000; Herzschuh, 2006; Wang et al., 2010b) (Fig. 17). One possible explanation is the location of MSEA, which acts as a land bridge for the ITCZ.

The dominance of the summer monsoon over the entire Asian monsoon region between 9 and 7 ka BP can be explained by a northward shift in the mean position of the ITCZ, which was triggered by warmer Northern Hemisphere and cooler Southern Hemisphere temperatures as seen in Greenland and Antarctic ice cores, respectively (Linzen and Nigam, 1987; Lindzen and Hou, 1988; NGRIP Members, 2004; Chiang and Bitz, 2005; EPICA Community Members, 2006; Kang et al., 2009; Cvijanovic, 2013) (Fig. 18). Summer monsoon intensity gradually declined after 7 ka BP following the decrease in insolation and the southward shift of the mean position of the ITCZ (e.g. Yang et al., 1995; Bernasconi et al., 1997; Yan et al., 1999; Mingram et al., 2004; Wang et al., 2010a; Stevenson et al., 2011) (Fig. 18).

5.3 The late Holocene (the last 2 ka): Paper IV

Lake Pa Kho was a shallow lake from BC 170 to AD 370, which compares in time with the reappearance of a shallow lake in Lake Kumphawapi about AD 150-350 (Fig. 17). The formation of a peatland suggests that P-E started to decrease ~AD 370. Dry conditions prevailed until AD 800, but were interrupted by wetter conditions between AD 420 and 450. The development of a wetland ~AD 800 to 970 points to an increase in P-E. A further dry interval may be recognized between AD 970 and 1300, corresponding to the hiatus or to very low peat accumulation. Higher peat accumulation and lower $\delta^{13}C$ values indicate an increase in P-E about AD 1300-1450, although overall dry conditions seem to have prevailed until AD 1450. The increase in aquatic plant remains, the reappearance of diatoms and the isotope proxies suggest the reestablishment of a wetland and higher P-E after AD 1450.

The hydroclimatic conditions reconstructed for Lake Pa Kho compare moderately well with reconstructed precipitation based on $\delta^{18}O$ values of speleothems from Dandak, Jhumar and Wah Shikar Caves in the IOM sub-region, and Wanxiang Cave in the EAM sub-region between BC 170 and AD 800 (Zhang et al., 2008; Berkelhammer et al., 2010) (Fig. 19). The lead and lag of wet or dry intervals can be explained by chronological differences. The reconstructed hydroclimatic conditions for Pa Kho and for the EAM and IOM regions show a divergent pattern to the precipitation reconstructed from biomarker in Southwest Sulawesi, Indonesia (Tierney et al., 2010). While wet conditions and a stronger summer monsoon prevailed over MSEA and the EAM sub-region between BC 170 and AD 370, precipitation was lower over Southwest Sulawesi. This would indicate a northward shift in the mean position of the ITCZ (Tierney et al., 2010). Between AD 400 and 800 summer monsoon intensity weakened over EAM and IOM region, but increased over Indonesia, which indicates a southward shift in the mean position of the ITCZ. The ITCZ possibly moved again northward after AD 800, when wet conditions were reconstructed for MSEA and Indonesia, while the IOM and EAM sub-regions remained under dry conditions. The Pa Kho sequence suggests drier conditions and thus a decrease in summer monsoon intensity between AD 1300-1450. This feature may be related to the Walker Circulation, because this dry interval is recognized in Indian caves, on Dongdao Island in the South China Sea and in Indonesia (Berkelhammer et al., 2010; Yan et al., 2011; Tierney et al., 2010). Relatively wetter conditions reconstructed for Pa Kho after AD 1450 compare well to records from India, the South China Sea and Vietnam, but diverge from those from Wanxiang Cave in China (Zhang et al., 2008; Berkelhammer et al., 2010; Buckley et al., 2010; Yan et al., 2011).
These results suggest that the northward shift in the mean position of the ITCZ may not have reached as far north after AD 1450 as during the former wet intervals.
Fig. 19: Reconstructed hydroclimate shifts for Lake Pa Kho compared with selected high-resolution records from the Asian monsoon region for the past 2000 years: (A) δ¹⁸O values of speleothem from Wanxiang cave (Zhang et al., 2008); (B) composite δ¹⁸O values of speleothems from Dandak, Jhumar and Wah Shikar caves (Berkelhammer et al., 2010); (C) δ¹³C values from Lake Pha Kho (this study); (D) grain size analysis of sediments from Cattle Pond, Dongdao Island (Yan et al., 2011); (E) Palmer Drought Severity Index (PDSI) derived from the Monsoon Asia Drought Atlas (MADA) for the region between 10-20°N and 95-115°E (Buckley et al., 2007, 2010; Sano et al., 2009; Cook et al., 2010; D’Arrigo et al., 2011); (F) δDwax values derived from a marine core from Southwest Sulawesi (Tierney et al., 2010). The dark and light grey bars represent high effective moisture/stronger summer monsoon and low effective moisture/weaker summer monsoon, as interpreted by the original authors. (G) The location of the selected records is shown on the map below (modified from Chawchai et al., submitted).
6. Conclusions

Mainland Southeast Asia (MSEA) is influenced by the Indian Ocean (IOM) and East Asian monsoon (EAM) sub-systems. It is also the only continental area in the Asian Monsoon region that generates strong convective cells during the initial stage of the summer monsoon onset. MSEA is thus considered a key region for studying the evolution of the Asian monsoon over time.

Published palaeo-records from the Asian monsoon region (including MSEA) have been compiled, evaluated, compared to climate model simulations, and complemented by new palaeo-proxy data sets derived from lake/peat sequences in northeast Thailand. Thailand is located almost in the middle of MSEA and occupies therefore an important position for tracing past shifts in summer monsoon intensity and for assessing the migration of the mean position of the Intertropical Convergence Zone through time.

Palaeo-data - model comparisons suggest a strengthened summer monsoon over MSEA during the Last Glacial Maximum (LGM) (23-19 ka BP). This can potentially be explained by an increase in land-sea thermal contrast due to the LGM sea level low stand, which exposed the vast Sunda continental shelf. The weakening of the summer monsoon intensity ~20-19 ka BP can be related to the rapid sea level rise ~19.6 ka BP, which led to a reorganization of the atmospheric circulation over the Pacific Ocean.

Reconstructed semi-quantitative precipitation and temperature indicate that climate conditions over MSEA between 15 and 5 ka BP alternated between cool/dry and warm/wet states. Precipitation and temperature generally exceeded 1100 mm a⁻¹ and 18 °C, respectively, during this time interval. The semi-quantitative data set derived from MSEA compares well with the Holocene hydroclimatic variability assessed from multi-proxy sediment sequences from Lake Kumphawapi in northeast Thailand. The onset of a stronger summer monsoon at 11.5 ka BP, as reconstructed from MSEA, compares well with the moisture history inferred from palaeo-records in the EAM sub-region, but differs slightly from that of the IOM sub-region, where the onset occurred about 1500 years later. These results support the hypothesis of an asynchronous onset of the summer monsoon between the IOM and EAM sub-regions. This asynchrony is possibly related to the land-sea configuration in this part of the Asian monsoon region, where MSEA, which connects the equator and higher latitudes, acted as a land bridge for the northward movement of the ITCZ. All data sets from the Asian monsoon region indicate a strengthened summer monsoon until 7 ka BP, which corresponds to a northward shift in the mean position of the ITCZ. Reconstructed drier conditions in the mid Holocene suggest a weakening of the summer monsoon after 7 ka BP, as a result of the decrease in insolation, and a southward shift in the mean position of the ITCZ.

The high-resolution, multi-proxy study of the sediment/peat sequence from Lake Pa Kho in northeast Thailand provides a detailed picture of hydroclimatic changes during the past 2000 years. Intervals of a strengthened summer monsoon are recognized between BC 170 and AD 370, between AD 800 and 960 and since AD 1450 and intervals of a weaker summer monsoon can be observed between AD 370 and 800 and between AD 1300 and 1450. These shifts were mainly controlled by the movement of the ITCZ, except for the drought interval between AD 1300 and 1450, which was possibly triggered by a change in the Walker circulation.
7. Ongoing works and future perspective

7.1 Asian monsoon variability during the marine isotope stage 3 (MIS 3)

Radiocarbon dating of the 10 m long sediment sequence from Lake Pa Kho shows that the sediments at 6.8 m date to >47 ka BP and that a long hiatus is present between 3.8 and 3.5 m, i.e. between 21.8 ka BP and 2.1 ka BP (BC 170). This means that sediments dating to the LGM to late Holocene are missing in the record. This contrasts with the findings of Penny (2001), who had assumed constant deposition since 40 ka BP.

The fact that sediments dating to between 21.8 and >47 ka BP are present provides an excellent opportunity to study hydroclimatic shifts and Asian monsoon variability during a climatically highly variable time interval, namely Marine Isotope Stage (MIS) 3. This time interval (59-29 ka BP) is characterised in ice cores, in marine and lacustrine sediment sequences and in speleothem records by rapid shifts between cold stadials and warm interstadials (Dansgaard et al. 1993; NGRIP Members 2004; Wang et al., 2001; Voelker et al., 2002). Most notable is the asynchrony between Northern and Southern Hemisphere records and the so-called sea-saw effect (Broecker, 2010; Stocker and Johnsen, 2001).

Preliminary geochemical analyses (between 6.8 and 3.8 m depth) of the peaty gyttja in sediment core PK-CP3 from Pa Kho comprise total organic carbon and nitrogen and their isotopes (Fig. 20). Since the sediment composition does not change, I assume that the lake level was relatively stable and that mineral contributions from catchment run off were minor.

Meyers and Teranes (2001) and Meyers (2003) suggested that C/N ratios of <10 and >20 in lake sediment samples would indicate phytoplankton and vascular plant sources, respectively. δ¹³C values of -30 to -35‰ have been reported for limnic algae and terrestrial C₃ plants, while δ¹³C values of -15 to -10‰ are indicative of C₄ plants and values of -11 to -39‰ represent aquatic macrophytes (Meyers and Teranes, 2001; Farquhar et al., 1989).

The TOC values for PK-CP3 range between 35 and 55% between 6.8 and 3.8 m depth (corresponding to 47.2-22 ka BP) and do not display any significant changes. Therefore I used the C/N ratio to preliminary subdivide the geochemical record into eight units (Fig. 20). In the bottom unit 1 (47 ka BP), C/N ratios are ~25-29 and δ¹³C values about -25‰, which would suggest a dominance of C₃ terrestrial organic matter sources. In unit 2 (47-44 ka BP), C/N ratios decline to 21-25 and δ¹³C values show a slight overall increase to -23 to -19‰, which indicates that more aquatic organic matter became incorporated in the sediments. Unit 3 (44-42 ka BP) displays an increase in C/N ratios (22-28) and a further increase in δ¹³C values to a maximum of -17‰, which could possibly be explained by higher contributions of C₄ plants. The C/N ratio is ~25 in unit 4 (42-39 ka BP) and δ¹³C values decrease to about -21 to -18‰, which suggests terrestrial organic matter sources and possibly higher contributions of C₃ plants, as compared to unit 3. The overlying unit 5 (39-35 ka BP) is characterised by an increase in the C/N ratio to 25-30 and a gradual increase in δ¹³C values from -19 to -17‰. This signifies terrestrial organic matter input, and an increase in the contribution of C₄ plants. The C/N ratio fluctuates ~24 in unit 6 (35-28 ka BP) and δ¹³C values display an overall increase to -16‰, which points to higher supply of aquatic organic material. In unit 7 (28-25.5 ka BP), C/N ratios increase from 23 to 28 and δ¹³C values decrease from -19 to -17‰. This would suggest an increase in the contribution of C₃ plants. The uppermost unit 8 (25.5-22 ka BP) shows a further increase in the C/N ratio and increasing values of δ¹³C from -17 to -16‰, which points to higher input of C₄ plants.

The tentative interpretation of the geochemical record for Pa Kho thus suggests distinct alternations in the supply of organic matter to the lake, i.e., dominantly C₃ plants, a mix of aquatic and terrestrial organic material, and dominantly C₄ plants. The shifts between C₃ and C₄ vegetation and changes in the supply of organic material into the basin potentially indicate hydroclimatic changes. This would mean wetter conditions (47-44 ka BP), drier conditions (44-42 ka BP), wetter conditions (42-39 ka BP), drier conditions (39-35 ka BP), wetter conditions (35-25 ka BP) and drier conditions (25-22 ka BP).
Since the age model for PK-CP3 is still in progress, detailed correlations to the Greenland and Antarctic ice cores are premature. However it is obvious that hydroclimate fluctuations are recorded in the sequence, and that these resemble the temperature evolution over the Northern or Southern Hemisphere.

Fig 20: Results of C/N, $\delta^{13}$C, and TOC measurements in the peaty gyttja of core PK-CP3 from Lake Pa Kho. The sequence between 6.8 and 3.8 m dates to between 47.5 and 21 ka BP. The location of the site is shown in Figure 8.

7.2. Myanmar Project

Sediment cores from two crater lakes in northwest Myanmar were collected in November 2013 using a modified Russian corer (7.5 cm diameter, 1 m length) (yellow square) and a gravity corer (10 cm diameter, 1 m length) (red square) (Figs. 5, 21). To achieve a complete sequence, sediment cores were taken with 50-cm overlap with the Russian corer. The gravity cores were sub-sampled in the field in 1.0 and 0.5 cm segments. The sediment cores and samples are stored in the cold room of the Department of Geological Sciences. Preliminary lithostratigraphic descriptions and loss-on-ignition measurements (LOI; 550°C and 950°C for organic matter and carbonate content) have recently been completed. Further analyses ($^{14}$C chronology, geochemistry) are planned for fall 2014/spring 2015.
Fig. 21: Topography of the area hosting the northeast-southwest chain of crater lakes in northwest Myanmar. Sediment cores were obtained from four lakes. The red squares mark the location of gravity cores and the yellow and red squares the locations where both long and short sequences were obtained.
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## Appendix I

Palaeo-records and palaeo-proxies used for the compilation in Paper I. The age assignments are based on published $^{14}$C, TL and U/Th dates. The number of dates between 25-17 ka BP for each sequence is given in brackets.

<table>
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<th>Site no.</th>
<th>Site Name</th>
<th>Lat (°)</th>
<th>Long (°)</th>
<th>Elevation (m a.s.l.)</th>
<th>Archive</th>
<th>Proxy</th>
<th>Age assignment</th>
<th>References</th>
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<td>1</td>
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<td>Pollen</td>
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<td>van der Kaars and Dam (1997)</td>
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<td>Pollen</td>
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ii
### Appendix II

Palaeo-vegetation records used in Paper II.

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