LONG-TERM DYNAMICS OF TROPICAL RAINFORESTS, CLIMATE, FIRE, HUMAN IMPACT AND LAND-USE IN INDONESIA

A FOCUS ON THE MONTANE RAINFORESTS IN CENTRAL SULAWESI AND PEAT-SWAMP RAINFORESTS IN SUMATRA

Dissertation

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To the people of Indonesia,

custodians of the Maritime Continent nature' treasures

Lake Lindu, Lore Lindu National Park, Central Sulawesi, Indonesia

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Abstract

The research conducted and compiled in this thesis contributes to the knowledge of the long-term vegetation, climate and fire dynamics as well as human impact on montane and peat-swamp rainforests of Indonesia. We applied multi-proxy palaeoecological methods to three sediment/peat cores taken from Lake Kalimpaa and Lake Lindu in Central Sulawesi and from Air Hitam peatland in Jambi Province, Sumatra.

The Lake Kalimpaa record is the first archive from Central Sulawesi providing information on palaeovegetation dynamics under the background of reconstructed palaeoenvironmental and palaeorainfall conditions throughout the past ca. 1500 years. The palaeovegetation as reconstructed from the palynological analysis reveals that around Lake Kalimpaa the Fagaceae family dominates the entire recorded period, as it still does today. Two disturbance events (ca. AD 1090-1190 and ca. AD 1450-1620) occurred in the catchment area of Lake Kalimpaa as reconstructed from sediment grain size and geochemical analyses. Comparison with fire frequency derived from macro-charcoal analysis indicates that these events were caused by forest fires. The regional correlation of these events with periods of drought registered elsewhere in Sulawesi and in Java suggests that centennial-scale increases in fire frequencies at Lake Kalimpaa were consequences of the vegetation being more prone to fire, probably due to more frequent and/or longer El Niño events. Despite that, Fagaceae did not decrease, indicating resilience towards droughts of at least one species of the family. However, palynological diversity values indicate that within-landscape diversity (Whittaker's gamma diversity) decreased when fires increased. Palynological rate of change and compositional turnover indicate that vegetation communities were more resilient to fire disturbance during periods of average high rainfall.

Palynological, charcoal and diatom reconstructions of the Lake Lindu core reveal that humans have modified the landscape at the Lindu plain for at least 1000 years. Evidence of frequent burning and possible shifting cultivation from an earlier phase from ca. AD 1000 to 1200 might be related to the metal age population which erected the megaliths in the province of Central Sulawesi. From ca. AD 1200-1700 decreases of macro-charcoal concentrations and pioneer vegetation indicators show that the use of the landscape of the Lindu plain had become more permanent. Due to the little research conducted so far on the megalithic culture of Central Sulawesi, it remains uncertain whether the architects of such a cultural change were the megalith people or a different ethnic group. A phase of forest recovery from ca. AD 1730 to 1910 correlated with a decrease in human activities in the valley, which historical reports describe as mostly limited to fishing and cattle grazing. These results suggest, that when human pressure on the landscape decreases as a consequence of different strategies of subsistence, the montane ecosystems possess a great capacity of recovery and fagaceous forest communities can expand within a relatively rapid time scale.

The multi-proxy palaeoenvironmental and palynological results from the Jaw SPT core in the Air Hitam inland peat in central Sumatra show a picture of a highly dynamic system. Since the beginning of peat accumulation around 7800 years ago net balance accumulation exceeded degradation, and the system remained a carbon sink. At first the peatland in Air Hitam was fed by fluvial run-off and the vegetation gradually changed from mixed Dipterocarpaceae swamp to marshy swamp communities dominated by *Durio* trees around 5300 years ago. A marked change in the vegetation community's composition occurred at the beginning of the late Holocene about 4500 years ago. A pole forest established, with *Pandanus* thickets colonizing the area as the dome developed into a rainfall-fed ombrogenous system. At the same time, macro-charcoal peak detection analysis reveals that fire frequency increased, possibly a consequence of the ENSO-onset.

Peat accumulation rates and hence carbon storage in the dome of the Air Hitam was considerable in the past. Higher rates of peat/carbon accumulation were found to correlate with *Pandanus* expansion under ombrotrophic settings. The testate amoebae assemblage indicates that the phases of ombrogenous *Pandanus*-pole forest were characterized by lower water table fluctuations. Despite the lower biomass input the peat accumulation rate was in average higher than during the mixed-swamp community phases. The correlation between high peat accumulation and relatively drier phases linked to ENSO makes the area of Air Hitam potentially one of the most effective carbon sequestering ombrotrophic peatlands in the view of future climate scenarios.

The results of charcoal analysis in montane (Lake Kalimpaa) and peat-swamp (Air Hitam) ecosystems show that although rare, wildfires did occur in the past in Central Sulawesi and Sumatra. The regime of fire, in particular the frequency, was found to correlate with regional scale drought episodes in Central Sulawesi. Both, fire frequency and magnitude of events, increased at Air Hitam in correlation with the increase in the number of El Niño events in the late Holocene. These results indicate that fire regime of montane ecosystems in Central Sulawesi and peat-swamp forests in Jambi can change in the long-term following changes in rainfall patterns and ENSO variability. However, our results indicate that the effects of fire and droughts on the carbon storage functions of the two systems under study might not have been marked. In particular, Fagaceae representation around Lake Kalimpaa was not affected by increases of fires caused by drought and carbon accumulation capacity of the ombrotrophic-*Pandanus* dominated ecosystems in Air Hitam did not change (but rather increased) when frequency and magnitude of fires increased. These results suggest that under natural conditions, fires occur in these ecosystems, but the effects on their carbon storage functions in time can be relatively low.

A strong acceleration of disturbance driven by human activities is underlined by the results in both the regions under study for the past decades. The overall results of this research highlight that the pressure exerted over millennia on montane and peat-swamp rainforests in Indonesia has been of a different magnitude than the modern anthropogenically driven changes.

The results from the research compiled in this thesis highlight the value of using a multi-proxy approach in order to better understand dynamics and functions of tropical ecosystems and to identify the most important drivers of long-term changes.

Acknowledgements

List of Figures, Tables and Appendices

Figures

axes are shown (cumulative explained variation 29%). Scatterplots represent the sample scores. Results are centered by taxa. Group of adjacent samples are marked differently to highlight the different groups corresponding to different prehistoric and historic phases. Palynological sub-zones are given in light grey; a) first and second axes scatterplots; b) second and third axes scatterplots

Tables

Appendices

Acronyms

AD: Anno Domini

AF: stepwise Alternating Field demagnetization

AGB: Above Ground Biomass

AMS: Accelerator Mass Spectrometry

BC: Before Christ

BP: Before Present (1950)

cal: calibrated

CHAR: Charcoal Accumulation Rate

ChRM: Characteristic Remanent Magnetization

Cl: Clay

CSa: Coarse sand

CSi: Coarse silt

DCCA: Detrended Canonical Correspondence Analysis

DI: Diversity Index (calculated with rarefaction analysis)

EFForTS: "Ecological and Socioeconomic Functions of Tropical Lowland Rainforest Transformation Systems" project

ELUC: "Environmental and Land-Use Change in Sulawesi, Indonesia" project

ENSO: El Niño-Southern Oscillation

FRI: Fire Return Interval

FSa: Fine sand

FSi: Fine silt

GAD: Geocentric Axial Dipole model

IOD: Indian Ocean Dipole

IPWP: Indo-Pacific Warm Pool

ITCZ: Intertropical Convergence Zone

ITF: Indonesian throughflow

LGM: Last Glacial Maximum

LIA: Little Ice Age

LLNP: Lore Lindu National Park

LORCA: Long-term apparent carbon accumulation rate

MAB: UNESCO Man and Biosphere Programme

LOWESS: Locally weighted scatter plot smoothings

MSa: Medium sand

MAD: Maximum Angular Deviation

MWP: Medieval Warm Period

MSi: Medium Silt

NRM: Natural Remanent Magnetization

NPP: Non-Pollen Palynomorph

PCA: Principal Component analysis

OP: Oil Palm

PSV: palaeomagnetic Secular Variations

PCC: Palynological Compositional Change

SD: Standard Deviation

ROC: Rate Of Change

Sk: skewness

SE: Southeast (Asia)

STORMA: "Stability of Rainforest Margins in Indonesia" project

SST: Sea-Surface Temperatures

TN: Total Nitrogen

TIC: Total Inorganic Carbon

WPWP: West Pacific Warm Pool

TOC: Total Organic Carbon

Preface

"Everything is the way it is because it got that way"

On Growth and Form, D'Arcy Wentworth Thompson (1917)

In the book, *On Growth and Form*, Thompson coniated this famous aphorism for a scientific view of the universe. In its apparent simplicity, it emphasizes the importance of forces and processes, in other words, of history, to understand why everything is the way it is. The basic concept is that one cannot truly understand modern forms/systems while simply looking (observing) their current state, which is a static picture of something dynamic in an ever-changing state. To understand it is necessary to uncover the mechanisms that generated this modern state.

Ecological research on natural (and anthropogenic) ecosystems is often (if not always) focused on present-day states. Investigating such a relatively short time interval is often insufficient to fully understand ecosystem dynamics such as succession, stability, tipping points and biodiversity patterns. Hence, the history (here palaeoecology) of an ecosystem, how it got to be this way, is equally important. How montane tropical rainforests in Central Sulawesi and peat swamp rainforests in the Jambi Province in Sumatra have responded to climate changes and human activities in the past and how they developed to be what they are today, those are the key questions this thesis aims to investigate.

Thesis aims

This thesis and the research behind it are devoted to investigate long-term vegetation dynamics, climate, fire regime and human disturbance in hot-spot montane ecosystems of the Lore Lindu National Park in Central Sulawesi and an inland peatland of Jambi, Sumatra (Indonesia). The main aims are:

o   Reconstructing past vegetation and climate dynamics in the two study areas

o   Investigating the history of human impact on the landscape

o   Assessing the impact and role of droughts (e.g. El Niño) and fires

o   Adding a historical perspective to the evaluation of current and future changes

Thesis outline

The first 4 chapters are dedicated to the introduction of the two regions under study and a general overview of the methodologies applied. An overview of the importance of tropical rainforests in Indonesia and the current and future drivers of change is given in Chapter 1. Chapter 2 includes a description of the multi-proxy palaeoecological methods applied and studied sites. Sulawesi and the montane ecosystems of the Lore Lindu National Park are described in Chapter 3. In Chapter 4, an

overview of the state of knowledge of peatlands of Southeast (SE) Asia is given with particular attention to the peat-swamp rainforests in Sumatra.

The following chapters are organized according to the manuscripts which are currently published (Chapters 5 to 8) or prepared for submission (Chapter 9) and include the results of the research:

Chapter 5

ENSO and monsoon variability during the past 1500 years as reflected in sediments from Lake Kalimpaa, Central Sulawesi (Indonesia)

Reconstructions of local climate variability as reflected in sediment are indispensable when the final aim is to investigate climate sensibility-resilience of natural tropical rainforests. The multi-proxy investigation included palaeorainfall proxies and fire/disturbance indicators which are used to highlight general trends in climate and periods of disturbance in the catchment area of Lake Kalimpaa for the last ca. 1500 years.

Chapter 6

Assessing resilience/sensitivity of tropical mountain rainforests towards climate variability of the last 1500 years: The long-term perspective at Lake Kalimpaa (Sulawesi, Indonesia)

Following the climate reconstruction from Chapter 5, the vegetation dynamics are reconstructed using palynological analysis and they are used to investigate the long-term effects of climate variability and disturbance on the tropical montane rainforests.

Chapter 7

Unravelling the past 1000 years of history of human-climate-landscape interactions at the Lindu plain (Sulawesi, Indonesia)

Human-landscape interactions in the past are reconstructed in the Lindu plain, an important cultural lake valley and known fish reserve in Central Sulawesi. The effects of pre-colonial and post-colonial cultural changes are discussed in the light of climate variability for the last ca. 1000 years.

Chapter 8

8000 years of vegetation dynamics and environmental changes of a unique inland peat ecosystem of the Jambi Province in Central Sumatra, Indonesia

Despite their importance as global carbon sink, peatlands of SE Asia have rarely been studied and our current knowledge of dynamics and ecology of these ecosystems remains incomplete. Palaeoecological and palynological multi-proxy analyses were carried out on a core taken from the thick Air Hitam inland peat in the Jambi Province of Central Sumatra. The long-term development of the peatland is discussed, highlighting dynamics of vegetation community phases in relation to local hydrology and climate over time.

Chapter 9

Effects of long-term climate and environmental variability on the development of inland tropical peat swamp ecosystems in Jambi Province, Sumatra, Indonesia

The long-term development of the peat-swamp ecosystems are compared to regional climatic and environmental records to investigate the relative importance of each factor in shaping todays system.

Chapter 10

The main findings of the research are summarized and synthesized to give comprehensive conclusions. Further, possible uncertainties are mentioned and recommendations for future research aspects to be covered are given.

CHAPTER 1 - Introduction

Tropical rainforests of Indonesia

At the beginning of the second half of the 20th century the tropical rainforests of Indonesia covered about 160 million hectares. Despite their limited geographic extension compared to the African and South American tropics, these ecosystems are widely acknowledged as biodiversity hotspots (Mittermeier et al. 1999; Myers et al. 2000; Figure 1.1) and are regarded as particularly important due to the numerous ecological and socio-economic services that they provide at the local, regional and global scale. Through photosynthesis and respiration, tropical rainforests absorb carbon dioxide and store the carbon (Figure 1.2) significantly affecting the global carbon flux with consequent impact on climate (Malhi and Grace 2000). Most of the carbon is stored as living biomass (aboveground and belowground) and dead organic matter (i.e. dead trees, leaf litter and soil organic matter).

Additionally, to their global function as carbon sink, tropical rainforests are highly diverse ecosystems and provide essential services for the local human populations. For instance, rainforest ecosystems provide ecological services such as erosion and flood prevention, water filtration, pollination-functions and have resources like timber and medicinal plants.
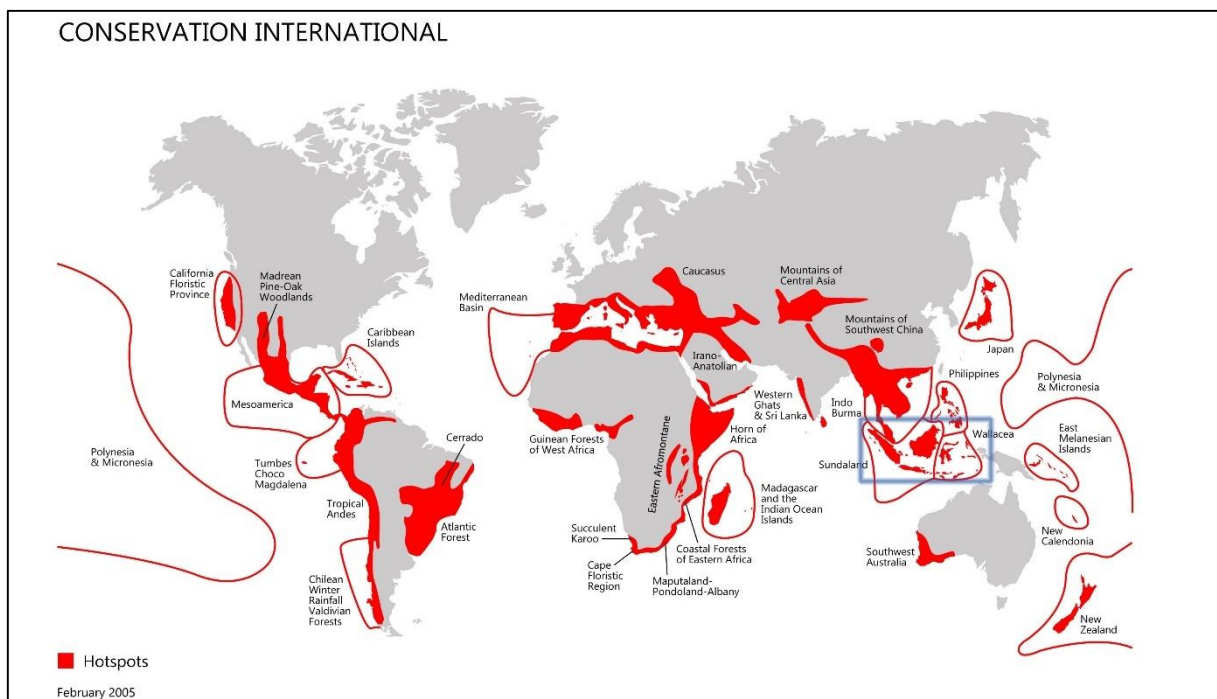
Figure 1.1 Conservation International map showing biodiversity hotspots of the world. These are biogeographical units of quite unequal size featuring exceptional concentrations of endemic plant and vertebrate species and experiencing exceptional loss of habitat. In the blue square approximate position of Indonesia (http://www.conservation.org/How/Pages/Hotspots.aspx)

Phytogeography of Indonesia
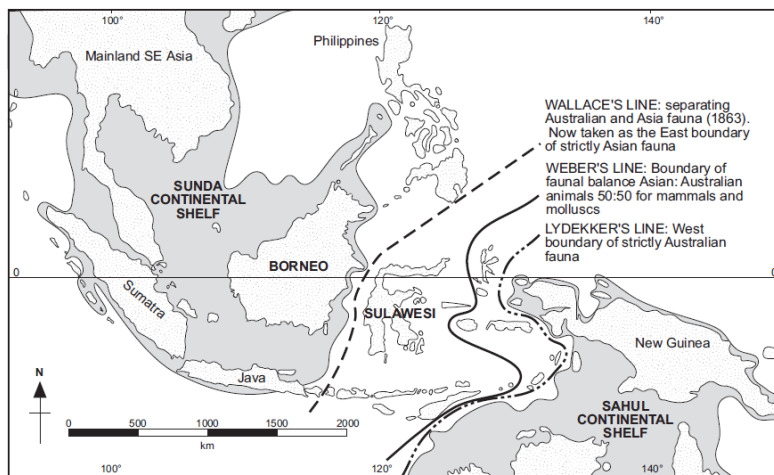
Figure 1.3 Map of Southeast Asia showing the main biogeographic lines. The grey areas represent Sunda and Sahul shelves (modified from Moss and Wilson 1998)

With ca. 41,000 species of vascular plants, the Malesian archipelago, where Indonesia is located, is one of the most species-rich areas of the world (Roos 1993). Alfred Russel Wallace already recognized in 1863 that this area presented islands which were biogeographically markedly different. Following his pioneer work, more "lines" or biogeographic borders were drawn ever since (Figure 1.3). The area cutted centrally by these lines is refered to as Wallacea (Philippines, Sulawesi, the Lesser Sunda Islands and the Moluccas; from Dickerson 1928). Located on the western side of Wallacea is the Sunda shelf or Sundaland (Malay Peninsula, Sumatra, Java and Borneo) and to the eastern side (New Guinea) is the Sahul shelf (Figure 1.3; van Welzen et al. 2011). Despite the fact that these boundaries were based on faunal data, Van Welzen et al. (2011) found in their assessment of phytogeographic regions that, with the exception of Java, the separations apply to plants as well. Their analysis, so far the only study including a complete data set of vegetation distribution, shows that phytogeographically, the Malesian area can be divided in three sections corresponding to the Sunda Shelf, the Wallacea (including parts of Java), and the Sahul Shelf. Therefore, plant species dispersion is also affected by the boundaries marked by the faunal lines. Traditiolally, these distribution patterns have been explained as caused by present day climatic regional differences and past sea-level changes (van Welzen et al. 2011). For instance, when sea-levels were up to 120 m lower during the Last Glacial Maximum (LGM: ca. 21,000 cal yr BP; Figure

Figure 1.4 Reconstruction of sea-level history since the LGM (21,000 cal yr BP) for the Sunda shelf (from Hanebuth et al. 2011)

1.4), large areas currently submerged were exposed (Emmel and Curray 1982; Hanebuth et al. 2000), resulting in dry Sunda and Sahul Shelves (Voris 2000; Bird et al. 2005; Sathiamurthy and Voris 2006). The removal of the marine barriers, would have enhanced terrestrial migration and rainforests probably expanded during glacial periods (Cannon et al. 2009). Opposite to that, dispersal barriers were reinforced durint periods of high sea-levels such it is the current interglacial (Tjia 1996; Woodroffe 2000; Figure 1.4).

Modern climate of Indonesia

The position of Indonesia between two continents (Asia and Australia) and two oceans (Indian and Pacific) makes it climatically one of the most interesting parts in the world (Gunawan 2006). The heat source of Indonesia and the Pacific is a major driver of the global circulation (McBride and Frank 1999). Particularly important is the role of the Indonesian region in providing energy for the operation of the north-south tropical Hadley cell and the east-west Walker circulation (Yan et al. 1992). The on-land main variable of Indonesia's climate is rainfall (Aldrian et al. 2004). In general, monthly rainfall is between 170 and 330 mm, as convective activity is very high in the archipelago (Aldrian and Susanto 2003).

At the intra-annual scale rainfall pattern are mostly driven by the changes in the seasonal insolation which affects the monsoon circulation and the position of the Intertropical Convergence Zone (ITCZ; Dommain et al. 2014). Roughly every six months, the onset of the Asian monsoon brings heavy precipitation from the northwest followed by the dry south easterlies of the Australian monsoon (Figure 1.5). Therefore, a wetter season in most part of Indonesia occurs from October to March during the Asian monsoon and a drier season coincides with the Australian monsoon from April to September (Gunawan 2006). Inter-annual variations in rainfall are influenced by the strengths of the monsoons, the phase of El Niño-Southern Oscillation (ENSO) and, particularly in Sumatra, by the Indian Ocean Dipole (IOD; Saji et al. 1999; Dommain et al. 2014). ENSO is a hemispheric or global phenomenon currently occurring every 2 to 7 years and characterized by two different phases (Torrence and Compo 1998). During El Niño (La Niña) warm (cold) phases, Indonesia experiences lower (higher) rainfall than in other years (Philander 1990; Cane 2005). The IOD is an

surface precipitable water (kg/m²) and wind directions (average 1981-2010)
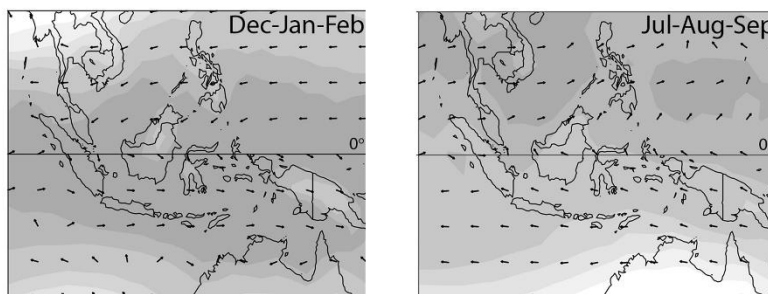
Figure 1.5 Contour maps of NCEP/NCAR reanalysis datasets showing the composite means of surface precipitable water in kg/m² with arrows denoting wind directions for the period 1981–2010. Left: wet season (Dec–Jan–Feb, left) corresponding to the Asian summer monsoon. Right: dry season (Jul–Aug–Sep, right) corresponding to the Australian monsoon (Biagioni et al. 2015a)

aperiodic oscillation of the Indian Ocean SST's. A positive (negative) IOD phase corresponds to the cooling (warming) of waters in the eastern Indian Ocean, which is accompanied by droughts (higher precipitations) in Sumatra (Webster et al. 1999).

Recent drivers of Indonesian rainforest changes

Starting in the 1950s, the commercial exploitation of rainforests and their conversion to plantations in Indonesia have grown rapidly and in less than 50 years more than 50% of the rainforests were converted (Forest Watch Indonesia/Global Forest Watch 2002). Miettinen et al. (2011) have shown that in Indonesia deforestation did not stop with the start of the new millennium, but is continuing at a high rate. The consequences of such a transformation are not restricted to the tropical regions directly affected, but expand beyond to inderectely affect the whole world (Miettinen et al. 2011). One major global concern regards the amount of carbon released into the atmosphere through deforestation of tropical rainforests and its potential contribution to global anthropogenic climate change (Houghton et al. 2000; Defries et al. 2002; Achard et al. 2004; Defries et al. 2007; Skutsch et al. 2007; van der Werf et al. 2009). Beside the indirect negative climatic feedback, the ongoing massive deforestation and forest degradation is causing habitat loss and forest fragmentation, which have a vast impact on biodiversity loss and species extinction in Indonesia.

Besides direct human pressure an additional potential driver of change for tropical rainforests is represented by anthropogenic climate change. The IPCC Fifth Assessment Report (Stocker 2014) provides climate change predictions for the SE Asian region and predicts a warming in the range of 0.8°C to 3.2°C by the end of 21st century. Although mean precipitation predictions sees a moderate increase in rainfall for the whole SE Asia, the scale varies at the regional level and there is an explicit pattern of increasing rainfall during the wet season and decreasing during the dry season (Figure 1.6). Adding to this pattern of increased seasonality, ENSO-related precipitation variability at the inter-annual scale will likely be intensified due to the increase in moisture availability (Li et al. 2007; Stocker 2014).

Figure 1.6 Maps of precipitation changes for Indonesia in 2081-2100 with respect to 1986-2005. Top: dry season from June to September. Bottom: wet season from December to March. Model scenario: RCP4.5 scenario with CMIP5 model. The grey lines represent areas where the signal is smaller than one standard deviation of natural variability (generated from KNMI Climate Change Atlas: http://climexp.knmi.nl/)

How the tropical ecosystems will respond to regional climate changes and how vegetation changes and climate will interact in the tropical

regions of SE Asia is still a matter of speculation. The testing of how forests respond to increasing carbon dioxide, temperature and rainfall variability needs large and difficult manipulation experiments which ecologists have only just started to conduct in the tropics. The complexity of the matter is further underlined by the often contradicting results of vegetation models and field experiments which show remarkable variability in forest sensitivity to climate change (e.g. Allen et al. 2010; Schuldt et al. 2011; Schippers et al. 2015; Van der Sleen et al. 2015).

Undoubtedly, we need to improve our understanding of how tropical forests respond to climate variability and disturbance on the long-term.

Scientific investigations in Indonesia: the projects and regions under study

In order to investigate the ecological and socio-economic consequences of transformation of tropical rainforests in Indonesia, the Georg-August-University of Göttingen has started to work in Indonesia in collaboration with Indonesian Universities in Sulawesi, Java and Sumatra already in 2000.

The project "Environmental and land-use change in Sulawesi, Indonesia (ELUC)" has been conducted between October 2010 and December 2012 in the Lore Lindu National Park (LLNP), Central Sulawesi (https://www.uni-goettingen.de/de/189495.html). The Collaborative Research Centre CRC 990 "Ecological and Socioeconomic Functions of Tropical Lowland Rainforest Transformation Systems (EFForTS)" started in the Jambi Province, Sumatra in 2012 (http://www.uni-goettingen.de/en/310995.html). Within those two interdisciplinary research projects, palaeoecological and palaeoenvironmental investigations were carried out in the studied regions in order to provide a valuable contribution by showing how the natural vegetation has responded to climate variability and human activities in the long-term.

CHAPTER 2 - Palaeoecology

"I shall collect plants and fossils, and with the best of instruments make astronomic observations. Yet this is not the main purpose of my journey. I shall endeavor to find out how nature's forces act upon one another, and in what manner the geographic environment exerts its influence on animals and plants.
In short, I must find out about the harmony in nature"

Alexander von Humboldt, Letter to Karl Freiesleben (June 1799)

The contribution of palaeoecological investigations

Modern ecosystems are the result of the combined action of ecological and historical factors. The relative importance of each of these agents in shaping the present-day world can be evaluated using both ecological and palaeoecological investigations (Rull 2010). Within the tropical ecology there are still many knowledge gaps concerning the dynamic processes that govern the functioning of Indonesian rainforests. One of the main uncertainties regards the long-term impact of climate variability and disturbance on these ecosystems (Cole et al. 2012).

The investigations and reconstructions of long-term vegetation dynamics unravel the ecosystem responses to climate change and disturbance in the long-term and add to the understanding of the state and functioning of today's ecosystems (Schüler 2012). Indirectly, this knowledge increases our understanding on how these ecosystems might respond to current and future change scenarios (Willis et al. 2010). Palaeoecological investigations on natural sediment archives represent a powerful tool which can be used to quantitatively unravel past vegetation dynamics and their response to long-term perturbance. These "palaeo" approaches have extensively been used in temperate ecosystems since the pioneer works of Von Post (1916) and Erdtman (1921) revealed the potential of microscopic fossils to provide insights into past vegetation and climate change. However, the employments of palaeoecological analysis in tropical regions have only recently gained attention. As a consequence, despite their importance, the tropical regions of Indonesia are one of the least studied regions in the world (Figure 2.1).

Montane rainforests and peat-swamp rainforests are amongst the least known and most threatened ecosystems in Indonesia in the view of current and future scenarios of change (Chapter 3 and Chapter 4). With this research we aimed at filling this gap of knowledge for these important ecosystems in the regions under study.
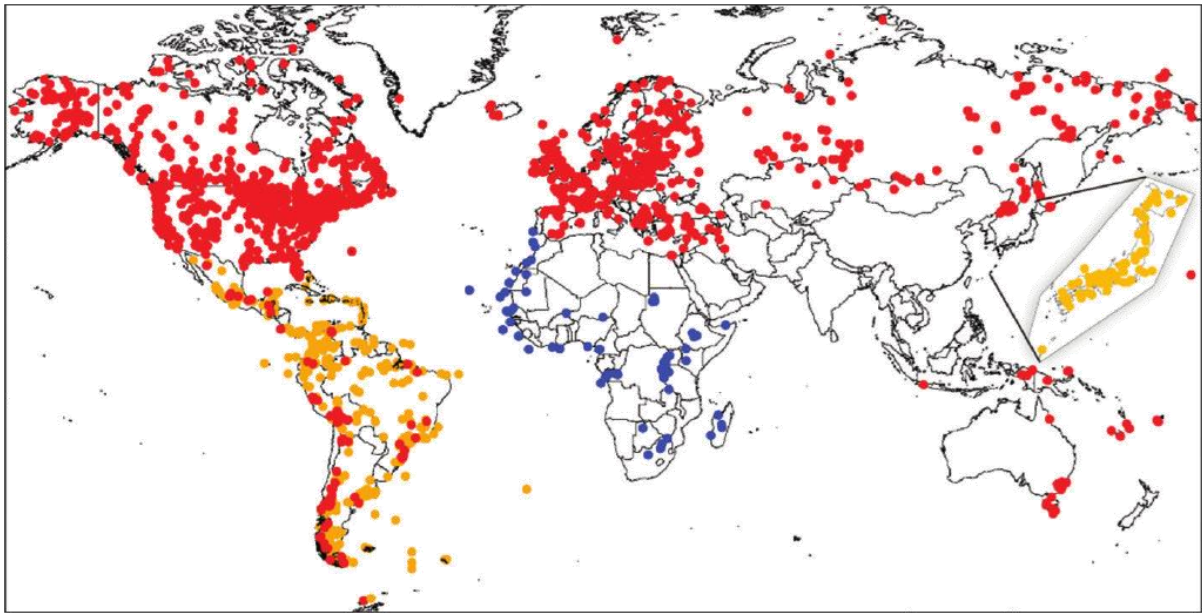
Figure 2.1 NEOTOMA and Latin American Pollen Database pollen records. Red dots: NEOTOMA database; blue dots: African Pollen Database; yellow dots: inventoried sites for inclusion from Latin America and Japan. (Figure from Grimm et al. 2013)
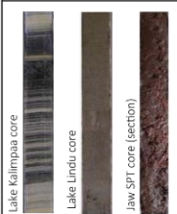
The multi-proxy approach in palaeoecology

In order to extend modern ecological observations on the dynamics of tropical ecosystems, we can look back into the past using sedimentary records with preserved fossil organisms or palaeoecological proxies (Birks and Birks 2006). Palaeoecological proxies include a wide variety of biological and non-biological remains. The basic idea is that, as statistical proxy, a palaeoecological proxy preserves physical characteristics of the past for direct measurements and enables scientists to reconstruct the environmental conditions that prevailed in the past. In other words, they are indirect evidences of certain climatic and environmental conditions in the past, which are no longer present for direct measurement. When more than one independent proxy is included in a palaeoecological study, such an investigation is called multi-proxy (Birks and Birks 2006). The main advantage of applying a multi-proxy approach is that misinterpretations of single proxy-data are reduced when several others are combined in the reconstruction. Additionally, the use of different proxies allows hypothesis on causalities of changes to be tested (Birks and Birks 2006). Coring of natural sediment archives was done in Sulawesi (Chapter 3) at Lake Kalimpaa and Lake Lindu and in an inland peat dome in Sumatra (Chapter 4; Figure 2.2 and Table 2.1). The investigations follow a multi-proxy approach and different palaeoecological and palaeoenvironmental proxies are used in collaboration with other researchers. The main body of the palaeoecological research, however, is made of pollen, spores and charcoal analyses. As a consequence, further details will be given in the following paragraphs on the applications and methods of such proxies.

Vegetation history: methods of palynology

The word palynology derives from the Greek (palunō, "strew, sprinkle" and -logy) and literally means "study of dust or particles that are strewn". A palynological analysis involves the quantitative study of palynomorphs which are both plants and animal remains that are microscopic in size and resistant to decay. Despite wide varieties of objects falling into the category of palynomorphs, the two most widely used are pollen grains and spores. A pollen grain forms in the anthers and is the microgametophyte of seed plants developed from the microspore. Its function is to be transferred in the female gamete for

Table 2.1 Detailed information on the sites and sediment cores studied

	Lake Kalimpaa	Lake Lindu	Jaw SPT
Island	Sulawesi	Sulawesi	Sumatra
province	Central Sulawesi	Central Sulawesi	Jambi
location (latitude, longitude)	1°19'34.8''S, 120°18'31.9''E	1°19'16''S, 120°04'36''E	2°02'52''S, 102°39'55''E
altitude (m asl)	1700	1000	57
ecosystem	mid-montane rainforest	lower montane rainforest	peat-swamp
current land-use	logging from 1960	rice plantations, gardens	oil palm plantations
materials	lake core	lake core	sediment-peat core
approximate age (cal yr BP)	1500	1000	8000
proxies	pollen, spore, charcoal (micro- and macro-), geochemistry, sediment analysis	pollen, spore, charcoal (micro- and macro-), diatoms, sediment analysis	pollen, spore, macro-charcoal, testate amoebae, sediment analysis, C and N isotopes
Chapter	5 and 6	7	8 and 9
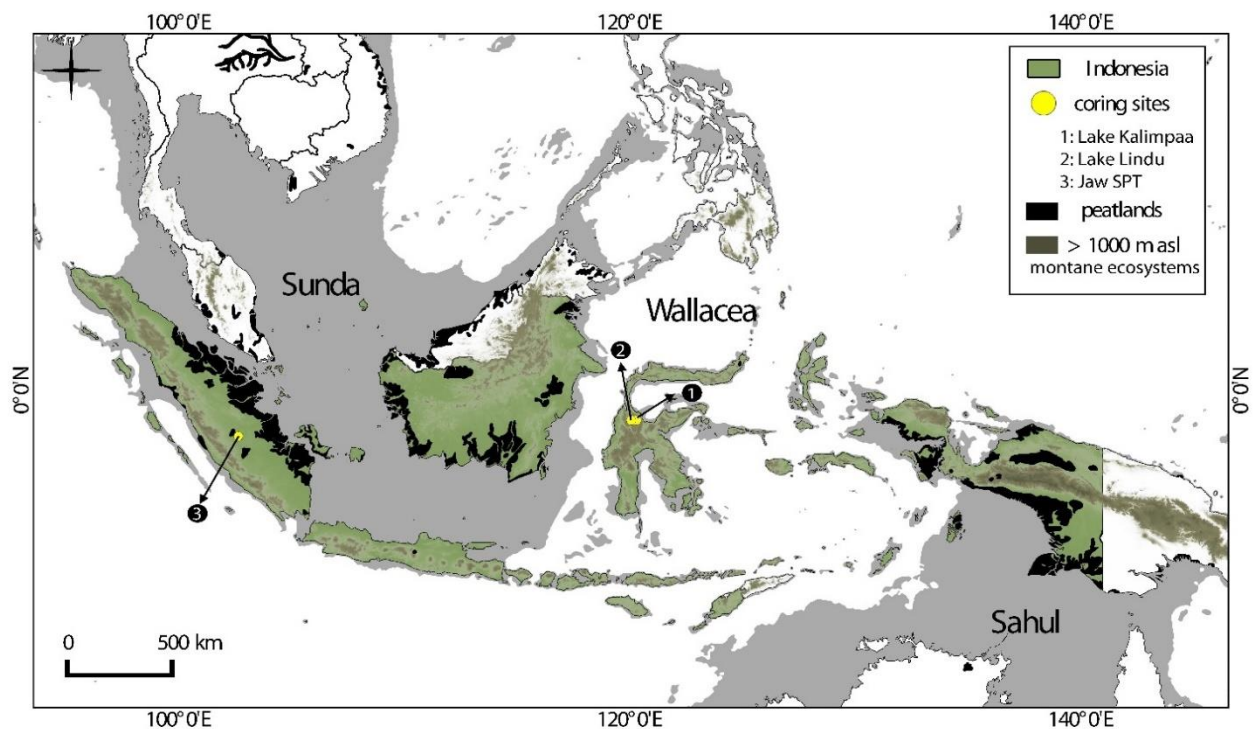
Figure 2.2 Location of the studied sites in Indonesia (yellow circles). Black areas represent peatlands, the brown areas montane ecosystems (at altitude >1000 m asl based on DEM SRTM 90 m). Data source for peat: Land and Water Development Division, FAO, Rome, Harmonized Soil Map of the World (Histosols and Gleysols layers only)

fertilization. Spore is the general term used for microscopically, unicellular asexual or asexual reproductive units of cryptogams (such as algae, bryophytes, ferns and fungi).

A series of circumstances have contributed to make the use of pollen and spores so popular, which Birks and Birks (1980) summarized in the following principles:

- o "Pollen grains and spores are produced in large quantities during the natural reproductive cycles of plants. Vegetation composition can therefore be inferred from the pollen grains released into the environment (pollen rain), as these are assumed to be a function of the number of parent plants"

- o "The majority of pollen grains and spores produced by plants never fulfil their reproductive function, and are deposited within sediments. They preserved as fossils being made of compounds that are highly resistant to most forms of decay other than oxidation"

- o "Fossil pollen grains and spores can be extracted from sediments deposited in anoxic conditions and thanks to their morphological taxonomic features they can be identified down to family/genus/species level under the microscope"

- o "The stratigraphic level at which they are extracted corresponds with particular periods in the past and the timeline can be reconstructed using methods such as radiocarbon dating"

These principles allow palynologists to reconstruct the vegetation dynamics of the past by counting fossil pollen grains and spores preserved under anoxic conditions in sediment deposits.

Laboratory methods for sediment pollen and spores analyses have been defined and standardized by Faegri et al. (1989). The methods employed the use of chemical reactors and sieving in order to remove the large quantities of organic and inorganic materials contained in sediments and "clean" the sample to facilitate the analysis under the light microscope (Figure 2.3).

Figure 2.3 flow diagram:

addition of markers — for concentration estimates, tablet of *Lycopodium clavatum* (Maher 1981)

hydrochloric acid (HCl 10%) — removal of calcareous fraction

potassium hydroxide (KOH 10%) — to break down organic components and remove humic acids (if needed)

hydrofluoric acid (HF 45-75%) — removal of silica and silicates (if needed)

Erdtman's acetolysis

acetic acid (CH$_3$COOH) — for dehydrogenation

mixture of acetic anhydride ((CH$_3$CO)$_2$O 9 parts) and sulphuric acid (H$_2$SO$_4$ 1 part) — removal of polysaccharides (cellulose) (Erdtman 1934)

sieving (150 μm) — removal of large fraction (if needed)

Figure 2.3 Workflow of palynological method applied in the research (modified from Faegri et al. 1989)

Fire regime history: methods of micro and macro-charcoal analyses

Fires occur when there is sufficient combustible biomass (fuel), ignition occurs and the environmental conditions (e.g. weather) allow for the combustion to take place or continue unimpeded (Herawati and Santoso 2011). The source of ignition includes human activities. For instance, in tropical forests, agriculture, logging and hunting are often linked to fires. Environmental conditions such as a dry climate increase the probability of fire as can cause biomass to become more combustible. Changing conditions lead to different fire pattern, intensity and frequency. Typically a fire regime can be described using these three elements (Figure 2.4). It follows that different climatic regions and/or ecosystems are charachterized by different fire regimes.

If oxygen is abundant and combustion is unimpeded ash is the end product of fire. However, incomplete combustion of organic material leads to carbonization or pyrolysis and the product is inorganic carbon compound or charcoal (Clark 1984; Patterson et al. 1987; Braadbaart and Poole 2008). Charcoal is resistant to oxidation and microbial activity and

Figure 2.4 diagram labels: Space, (100 km)^2, km^2, m^2, Time, seconds, days, decades, Climate, Ignitions, Fire regime, Vegetation, Weather, Topography, Wildfire, Fuel, Oxygen, Heat, Flame, Fuel

Figure 2.4 Controls on fire and the make-up of fire regime at different space and time scales (from Moritz et al. 2005)

thus during and following a fire event, charred particles are deposited and preserved in natural palaeo-archives (lakes, swamps, peats). Along the stratigrafic of a sediment core, layers with abundant charcoal particles or charcoal peaks are then used to reconstruct fire history around the site of deposition. Palaeoecologists have used charcoal as a proxy for palaeofire and the applications are many, including: analysis of climate-fire-vegetation linkages (e.g. Clark and Royall 1995; Gavin et al. 2006), investigation on the prehistoric use of fire (e.g. Pitkänen and Huttunen 1999; Wick and Möhl 2006) and the analysis of regional and global biomass burning in connection with carbon cycle and the effects on the global atmospheric CO_2 concentration (e.g. Carcaillet et al. 2002).

Generally, charcoal derived from plant material is black, opaque, brittle and angular, with an elongate-prismatic appearance possibly displaying some cellular structure (Swain 1973; Clark 1982; Griffin and Goldberg 1983; Patterson et al. 1987; Enache and Cumming 2006). Despite these generalizations, charcoal morphology is often surprisingly variable and irregular. This is due to the fact that different kind of fires (e.g. temperature) and source of combustion can produce fragments with different structure and size. Morphologically the charcoal remains differ depending on temperature of formation and the fuel (bark, trunk, branches and leaves). Additionally, the initial morphology is often modified through taphonomic processes (Enache and Cumming 2006), as charcoal fragments can be transported for kilometers before to burial (Garstang et al. 1997; Clark et al. 1998; Pisaric 2002). The source of charcoal in a natural archive can be regional (distant) or local (close or within the deposit area; Whitlock et al. 2001). Mathematical models and empirical studies following modern fires indicate that charcoal abundance of particles >100-125 μm in diameter decreases when the distance from the deposition environment and the fire increases (e.g. Whitlock and Millspaugh 1996; Clark and Patterson 1997; Gardner and Whitlock 2001). These results confirm that charcoal particles larger than >100-125 μm are not transported far from the fire margin. Following that, size classes can be defined based on the distance from fire source and the signal they bare as:

- o   microscopic (micro) charcoal, <125 μm -> regional or long distance-transported signal of fire
- o   macroscopic (macro) charcoal >125 μm -> local signal of fire

Micro-charcoals have mostly been examined on the same slides used for pollen analysis. The sample preparation procedure was therefore standard pollen laboratory preparations (Sadori and Giardini 2007). While it is possible to distinguish and separate larger size of charcoal particles with the pollen-slide method, the standard chemical procedure results in fragmentation of the particles and loss of the local fire signal.

A more appropriate method to quantify local signal of fire, is by quantifying macro-charcoal particles using with the sieving-method. By gently washing and sieving continuously spaced subsamples, only the larger particles are retained (>125 μm), the fragmentation is greatly reduced and the macro-charcoal so quantified embodies the local signal of fire. Since 2004, a diagnostic and analytical tool called

CharAnalysis is available for sediment macro-charcoal analysis (https://sites.google.com/site/charanalysis/). *CharAnalysis* is a program for analyzing sediment-charcoal records via removal of noise and a signal cleaning of the fire peaks to reconstruct local fire regime history. Not all macro-charcoal records are suitable for such a decomposition analysis. A certain number of assumptions must be met as pointed out by Higuera et al. (2007): (I) contiguous sampling at a fine sampling interval is needed as discontinuous sampling misses charcoal peaks; (II) adequate age-depth control and time resolution. They recommend a sampling resolution of less than about 0.12 times the mean fire-return interval (e.g. 12 yr for a 100 yr mean FRI).

Comparison of the pollen-slide and sieving methods at the same site (Carcaillet et al. 2001) reveals that the two charcoal series do not represent exactly the same signal, as only half of the pollen-slide charcoal signal contains the sieving charcoal signal. The pollen-slide method is complex and strongly influenced by long-distance transported micro-charcoal particles. It is therefore best used to infer regional biomass burning in connection to for example climatic reconstructions. The sieving method, on the other hand, shows distinct peaks, and it is more suitable when the goal is to reconstruct the local fire history (Carcaillet et al. 2001).

As a consequence, depending on the final aim of the investigation, either one or both methods should be used.

CHAPTER 3 - Sulawesi

The island of Sulawesi is considered a globally important conservation hotspot due to the unicity of its biogeographic history as a large island between Asia and Australasia and complex geology (Wallace 1869; Hamilton 1979; Holloway and Hall 1998; Wilson and Moss 1999; Cannon et al. 2007). Opposite to the other large islands in South East Asia, Sulawesi remained isolated during period of high sea-level such it was during the Last Glacial Maximum and for the past 8 million years (Audley-Charles 1983; Cannon 2005; Figure 3.1).

Roos et al. (2004) observed that despite this long history of isolation, plant species diversity patterns are not dissimilar from those of the other large islands in SE Asia and level of endemism is relatively low (<15%). However, as pointed out by Cannon (2005), the lack of high endemism in Sulawesi might be at least partly due to the little collection effort being done so far (estimated 23 specimens per 100 km^2; Whitten et al. 2002). An example of this lack of knowledge comes from the recent discovery of new, likely endemic, species of *Magnolia* (*M. sulawesiana*) and *Elaeocarpus* (*E. firdausii*) during tree inventories in montane areas of Central Sulawesi (Brambach et al. 2013; Brambach et al. 2016).

Figure 3.1 Map showing sea level changes of Sunda, Sahul and Wallacea biogeographic regions (at-60 m and-120 m and modern). In green the island of Sulawesi (modified from O'Connell et al. 2010)

Addionally to its status as a biodiversity hotspot, Sulawesi has also been highlighted as an important model or case study for testing the effect of climate change on plant species adaptation and population dynamics (Cannon 2005). Due to the complex geology and isolated biogeographic location, plant species occuping the landscape in Sulawesi have limited migration opportunities and might be particularly sensitive, if not vulnerable, to changes in rainfall patterns in the future (Cannon 2005).

The Lore Lindu National Park

With an area of about 2,300 km², the Lore Lindu National Park (LLNP) is one of the few remaining contiguous areas of montane rainforest in Sulawesi (Culmsee et al. 2010; Figure 3.2). Most parts of the protected area are covered by upland and montane rainforests ranging from 800 m to 2700 m in elevation, on intermediate soils. The forests are classified to be in good to old growth conditions (Cannon et al. 2007). The majority of Sulawesi's unique fauna (over 50% of the mammals and 68% of the birds) are present within the Park making the LLNP one of the most biologically important areas in the Indonesia.

The current area covered by the LLNP resulted from the integration of two natural sanctuaries and a natural preservation area, which were formerly separated (Adiwibowo 2005). The two natural reserves were the Lore Kalamanta Wildlife Reserve (131000 ha) located in Poso District and the Lore Lindu Wildlife Reserve (67000 ha) located in the Donggala and Poso Districts. The nature preserve areas included The Danau Lindu Tourism/Protected Forest (31000 ha) located in Donggala District (Helmi 2001; The Nature Conservancy 2001). In 1977, UNESCO declared Lore Kalamanta and Danau

Figure 3.2 Map of Sulawesi showing Central Sulawesi (grey), the boundaries of the Lore Lindu National Park (LLNP, red) and primary rainforests (green, data source: Land Cover 2011, the Ministry of Forestry, the Republic of Indonesia)

Lindu forests as Biosphere Reserved areas and in the following 20 years the area was slowly converted into the LLNP. During the Third World National Park Congress held in Bali on 1982, the Minister of Agriculture declared the establishment of the LLNP, including 231,000 ha of conservation area (Decree of the Minister of Agriculture No. 736/Menteri/X/1982). Eleven years later in 1993, eventually the government altered the function of the Danau Lindu Tourism/Protected Forest (31,000 ha), the Lore Kalamanta Wildlife Reserve (131,000 ha), and the Lore Lindu Wildlife Reserve (67,000 ha) to that of a national park (Decree of the Minister of Forestry No. 593/Kpts-II/1993). However, the decree stated that the boundary was temporary. The permanent boundary (229,000 ha) was determined after surveys in 1999 (Minister of Forestry and Estate Crops Decree No. 464/Kpts-II/1999). The Lindu and Besoa valleys were moved outside the National Park and the park had 11,000 ha less than previously stated (Adiwibowo 2005).

Geology of the Central Sulawesi

Sulawesi is geologically very complex, being located at the crossroad of three different plates, the Asian, the Australian and the Pacific plates (Figure 3.3). The main orogenic belt of the island formed during the Oligo-Miocene. The central part of Sulawesi contains two metamorphic units. The western unit belongs to the substratum of the western arm and the eastern unit to an Oligo-Miocene suture (Villeneuve et al. 2002). The mountains of the LLNP are part of the western unit at the boarder to the eastern suture unit. They mostly consist of intrusive rocks such as crystalline and metamorphic granites, granodiorites, schists, and phyllitic rocks.
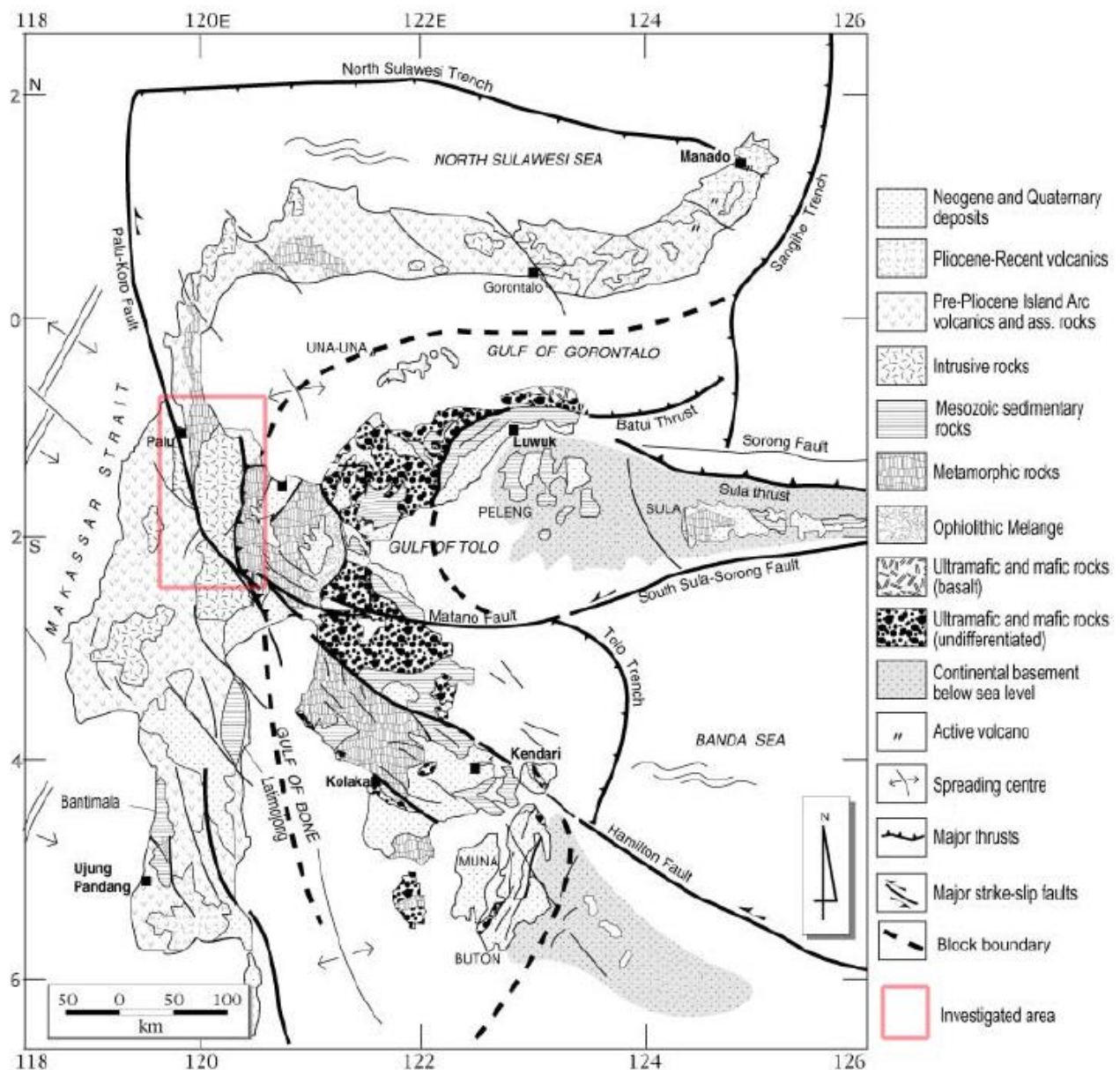
Figure 3.3 Simplified geological map of Sulawesi with approximated location of the LLNP in the red square (modified from Villeneuve et al. 2002)

Vegetation

The LLNP is among the parts of Sulawesi for which the largest plant collections are available (Cannon et al. 2007) and the forests are species-rich tropical montane forests (236 tree species in 62 families Culmsee et al. 2010). A series of investigations conducted in the LLNP along an altitudinal gradient described the main changes in composition and structure (Culmsee and Pitopang 2009; Culmsee et al. 2010; Culmsee et al. 2011), which in tropical mountain rainforests have been often used to understand biogeographic regional affinities (Keßler et al. 2002; Ashton 2003). Based on tree species composition and forest structure changes, the communities of the LLNP can be roughly divided in sub-montane (700-1300 m asl), lower montane (1300-1700 m asl), mid-montane (1700-2000 m asl) and upper montane (above 2000 m asl; F. Brambach personal communication). According to the studies, the lower altitude sub-montane forests are the most diverse in terms of plant tree family. A trend of decreasing diversity with increasing elevation was also observed. The composition and dominance patterns also changed along the elevational gradient, with the exception of the Fagaceae and Myrtaceae, which remain always important components in the tree communities. Diversity was at the lowest level in the lower and mid-montane elevations, where Fagaceae dominate the aboveground (AGB) biomass. Interestingly, the estimated AGB did not decrease from submontane to mid-montane despite the lower diversity. Markedly different were the upper montane communities where the tropical conifers Podocarpaceae became the most important family (Figure 3.4).

Dipterocarpaceae are the dominant trees in the rainforests of Borneo, Sumatra and Malaysia but they are almost completely absent in Sulawesi. Only six species are reported located mostly in the south Sulawesi and at lower elevation (Keßler et al. 2002). This has been explained as the result of timing of evolution and radiation and the geographic history of Sulawesi. The seeds of Dipterocarpaceae cannot be dispersed are not easily dispersed via sea or air and therefore the species need land bridges to expand. The Dipterocarpaceae radiated about 25 Ma (Morley 2000) when Sulawesi was already

Figure 3.4 left: view of the mid-montane rainforest dominated by Fagaceae; right: view of the mossy upper montane rainforest dominated by Podocarpaceae (photos of P. Lembcke)

separated from Sundaland and was an orogenetically active and unstable area. The isolation of Sulawesi earlier in the geological – time scale (Eocene, 45 Ma) might have limited the dispersion of the dipterocarps which today remain in low number in the island (Suzuki et al. 2006). Culmsee et al. (2010) study revealed that the niches filled by the Dipterocarpaceae in western Malesia are, in Sulawesi, occupied by other tropical Gondwanan elements in particular many magnoliids (Culmsee et al. 2010).

Biogeographical patterns are significantly different with elevation (Culmsee and Leuschner 2013). Based on the tree species and family compositions, the lower and mid-montane forests seem comparable to the Bornean forest at 1700 m altitude (Western Wallacean distribution) in having their two most important families in common, the Fagaceae and Myrtaceae. In contrast to that, the geographical isolated mountaintops have a more distinct assemblage with unique phylogenetic diversity despite the low diversity in terms of species number. Western Malesian taxa are found in these communities as well. However, in addiction to that, also elements from Papuasian/eastern Malesian subregions and the Philippines such as Trimeniaceae and Paracryphiaceae are found (Culmsee et al. 2011).

The importance of Fagaceae

Studies of tree families along altitudinal transects have highlighted the importance of Fagaceae mainly with the genera *Lithocarpus* and *Castanopsis* (Culmsee et al. 2010; Culmsee and Leuschner 2013). They were abundant at all elevations, particularly in the lower and mid-montane forests where they contribute for the most part to the AGB.
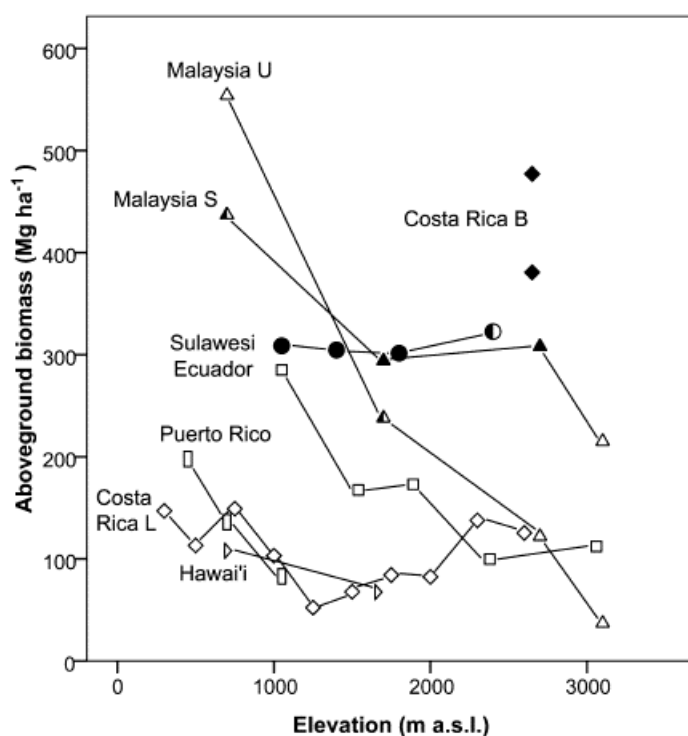
Figure 3.5 Comparison of estimated aboveground tree biomass (Mg ha^{-1}) in relation to elevation (m asl.) in Southeast Asian and Neotropical forests (from Culmsee et al. 2010)

The genera *Castanopsis* and *Lithocarpus* evolved at least 40 Ma and they are today found everywhere in Malesia (Soepadmo 1971; Manos and Stanford 2001; Cannon and Manos 2003; Culmsee and Leuschner 2013). While in western Malesia, the Dipterocarpaceae became dominant in everwet forests about 25 Ma, the Fagaceae became important in the uplands of Malesia (Morley 2000; Dam et al. 2001).

Culmsee et al. (2010) in their assessment of aboveground biomass (AGB) differences between different montane communities, found that, opposite to what was experienced in other tropical montane rainforests, in Sulawesi, AGB does not decrease steadily with elevation (Figure 3.5). Further, they observed that almost all the montane forests where no elevation AGB decrease was shown were dominated by Fagaceae at mid-altitudinal levels. They conclude that the Fagaceae play an important role as carbon stores in the

montane forests of Sulawesi and conservation efforts aiming at preserving ecosystem functions should focus on this family.

Ecologically, the relatively lower temperatures found in the uplands, do not seem to be disadvantageous for tropical Fagaceae, which are today most diverse in the montane forests of SE Asia (Manos and Stanford 2001; Manos et al. 2008). A more important limiting role might be played by rainfall, as the current geographical distribution of *Lithocarpus* and *Castanopsis* suggests an intolerance to seasonal climates (Soepadmo 1971; Cannon and Manos 2003).

Climate

In the LLNP, mean annual temperature decreases with elevation from 21˚C at 1000 m to 14˚C at 2400 m (Culmsee et al. 2010). Precipitation patterns are not easily described as they change markedly due to the complex topography. However, mean annual precipitation is estimated to be between 1900 and 2100 mm (Hijmans et al. 2005; WorldClim 2006). Precipitation patterns in the mountainous regions of the LLNP are controlled by the seasonal migration of the ITCZ across the equator and inter-annual changes in ENSO. As the ITCZ migrates southward during the austral summer, the northwest monsoon delivers humid air and heavy rainfall, whereas during austral winter the southeast monsoon brings relatively cool, dry conditions while the ITCZ is positioned over mainland Asia (Gunawan 2006). In the montane areas of Central Sulawesi the rainfall is strongly determined by the local topography. The air masses reaching the area from the northwest and southeast are lifted orographically, leading to a formation of clouds and rainfall throughout the year. However, the monthly amount of rainfall formed during the southeast monsoon is slightly less than that of the northwest monsoon, as the former brings humid air masses, while the latter brings dry air from the Australian continent and the rainfall is therefore purely orographic. The modern intra-annual climate of the montane areas of the LLNP can be described as perhumid with at most two months of slightly lower precipitation, corresponding to the southeast monsoon peak in August (Gunawan 2006).

The inter-annual variability of rainfall is influenced by the coupled ocean-atmosphere phenomenon El Niño-Southern Oscillation (ENSO). In particular, during El Niño warm phases, Central Sulawesi experiences lower rainfall than in other years.

Prehistory and history of Central Sulawesi

Central Sulawesi is rich in Metal Age megaliths, large worked stones in the shape of cylindrical vats, statues, urns and mortars (Kaudern 1938; Sukendar 1976; Bellwood 1979; Sukendar 1980a; Sukendar 1980b; Kirleis et al. 2012; Figure 3.6). The majority are located in the valleys of Napu, Besoa and Bada. The absolute age of the megaliths has not been determined with precision (archaeological estimates range from 3000 BC to 1300 AD). At the Pokekea site in the Besoa valley Kirleis et al. (2011 and 2012) established a *terminus ante quem* for the erection of the large stone vats called "kalambas" of ca. AD 830. Palynological analysis of sediment cores show that the opening of the forest in the valley via

Figure 3.6 Cup-marked stones named *Kalambas* and one decorated disc-shaped lid close to Kalambas from Pokekea archaeological site in Besoa valley

continuous burning started from ca. 2000 years ago and the authors suggest a link between this landscape transformation and the early construction phase of the megalith monuments. The relationship between the megalithic culture and the indigenous people that live around the Park today remains open to speculation. It seems likely that Proto-Malay and Palaeo-Mongoloid people migrated into the area, but much more research is needed in order to answer this fundamental question related to the megaliths and their creators.

The first Europeans who visited Central Sulawesi came from Portugal and Spain via the Philippines in the middle of the 16th century. Their arrival marked the introduction of new plant and animal species in particular corn, tomato, chili peppers and horses (Davis 1976). A more important impact was the arrival of the Dutch, who opened up the areas in the lowlands starting in the 17th century. Before the Dutch arrival, there was little wetland rice growing, and agriculture activities were mostly focused on upland dryland rice, corn, and tubers grown under a shifting system. Population density in Central Sulawesi was very low. The small isolated communities lived in the mountains and were ruled by kings and it seems the settlements were not permanent at the time (Kreisel et al. 2004). Contrary to the inhabitants of the

Poso region, methods of wet-rice cultivation were already well known to inhabitants in western Central Sulawesi before the arrival of the Dutch, as reported by Valentyn (1724).

Today, the Central Sulawesi region is ethnically and culturally heterogeneous and comprises 15 indigenous groups, speaking 24 distinct languages. However, most of the people living around the National Park today are of recent arrivals, or descendants of recent arrivals. The majority moved into the area as participants in the government's transmigration programs (at a peak in the 1970-80s) and as a result of conflicts elsewhere (Kreisel et al. 2004).

CHAPTER 4 - Sumatra

Peatlands of Southeast Asia: an overview

The peatlands of SE Asia are the largest peatland areas in the tropics (Dommain et al. 2011). Estimates indicate they cover an area of about 236,000 km², which corresponds to about 60% of tropical peat area. The formation of such an extended area of peatland in SE Asia was favoured by a series of factors including low topography, high rainfall throught the year and substrated which impeded water permeation (Page et al. 2009). These conditions are found mostly in delta costastal areas and along riparian depressions (Page et al. 2011; Figure 4.1). The vast majority of peatlands in SE Asia are found in Indonesia (about 210,000 km²) and Malaysia (23,000 km²). These peatlands are major terrestrial carbon reservoirs. Current estimates indicate they store about 65 Gt of carbon, corresponding to 9% of the global peatland area and 11-14% of the global peat carbon stocks (Joosten 2009; Page et al. 2011).

Besides their function as carbon sinks, tropical peatlands are covered by forested ecosystems (peat swamp rainforests) that are unique (Schrier-Uijl et al. 2013) and provide a large number of ecosystem services. These ecosystems function as important climatic mitigating areas, they supply water and prevent erosion (Yule 2010) during extreme climatic events such as the occurrence of El Niño and La Niña episodes. Additionally, they are valuable as a form of biological heritage and genetic resources. Furthermore, as shown by Kuniyasu (2002) in the lowland plain of the Kampar River in Sumatra, local villages and the majority of capital-less people, greatly benefit from the collection of forest products and fishing in peat-swamp areas.

Figure 4.1 Distribution of peatlands in SE Asia (modified from Wüst et al. 2007)

Peatland conversion and fire in Sumatra

Despite the fact that indonesian legislation stipulates peatlands with deep peat deposits should not be developed (Presidential Decree No. 32/1990), peat swamp forests are one of the ecosystems most threatened by deforestation in Indonesia, particularly in Sumatra (Miettinen et al. 2011). According to Giesen (1993; 1994) at the end of the last century, peat swamp and freshwater swamp forests in Sumatra covered an area of about 93,000 km². However, since the 1970es, the majority of Sumatran peatlands have been converted, logged and drained to such an extent that only a small portion (about 9.3%) remain under pristine conditions (Miettinen et al. 2011). Based on the analysis of land cover maps of Miettinen et al. (2011) the Riau and the Jambi Provinces are the areas in Sumatra where deforestation and degradation of peatlands are concentrated. In 10 years, from 2000 to 2010, these two Provinces alone lost 40% of their peat swamp forest cover and it is estimated that only 35% of the peatlands in Riau and Jambi provinces remained forested in 2010.

Conversion and degradation are causing loss of peat swamp forest species and elevated carbon emissions due to the decomposition of organic carbon-rich matter stored belowground in the peat (Couwenberg et al. 2010; Hooijer et al. 2010; Miettinen and Liew 2010). Addionally, drainage and exposure due to deforestation have increased the vulnerability of peat soil to fires (Page et al. 2002; Heil et al. 2007; Figure 4.2).

Under natural conditions, tropical peat and swamp rainforests are protected from fires due to their permanently wet settings (Rieley et al. 1996). However, conversion to agricultural use requires both removal of the primary vegetation and soil drainage, resulting in peat shrinkage through a combination of water loss, enhanced aerobic decomposition and compaction (Figure 4.2). Drained peatland areas
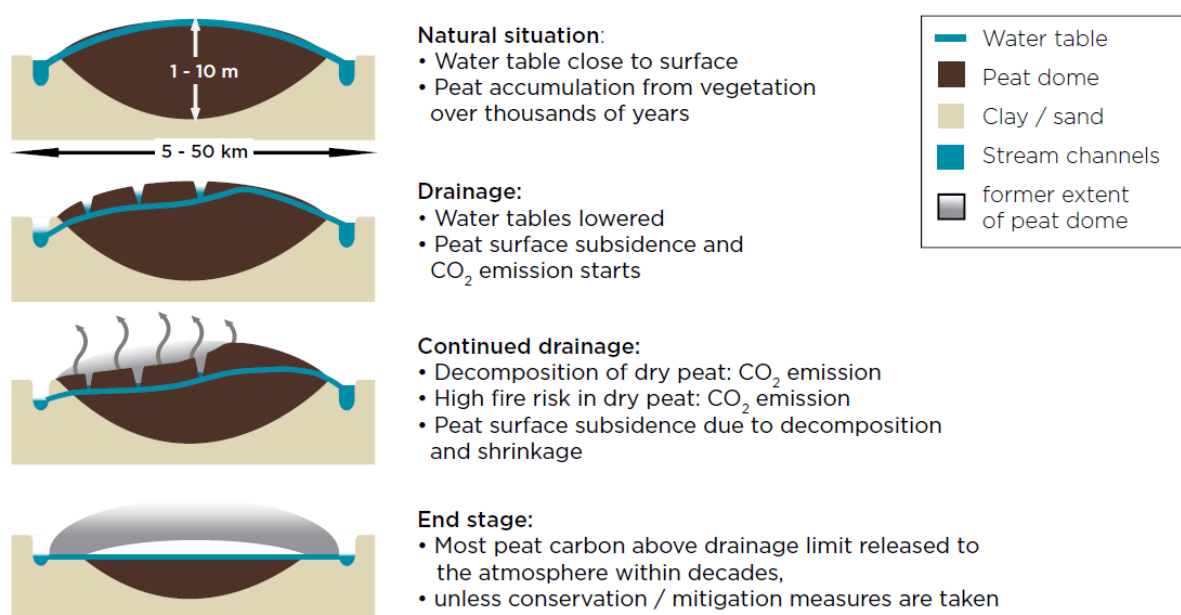
Figure 4.2 Schematic illustration of drainage effects on peatland dome (from Page et al. 2011)

with degraded opened canopy vegetation become extremely vulnerable to annual fires that further degrade these ecosystems (Hoscilo et al. 2011).

Occasional but catastrophic fires on peatland can release immense quantities of carbon into the atmosphere from peat combustion (Page et al. 2002; Heil et al. 2007). For instance, widespread fires during the strong El Niño year of 1997-98 burned a total of 308,000 hectares of peat swamp and freshwater swamp forest only in Sumatra (Tacconi 2003). According to the Asian Development Bank study (Tacconi 2003) about 156 million (about 75%) of the 206.6 million tons of carbon emission estimated from the fires occurring in 1997-1998 were produced by burning peat.

This suggests, as pointed out by Dommain et al. (2011), that in the future carbon emission from burning and oxidizing SE Asian peatlands might increase, if dry-season rainfall will decrease and extreme precipitation events linked to ENSO will increase, as predicted by climate models (Li et al. 2007; Stocker 2014).

Oil palm plantations

Currently the majority of the plantations on peatlands are oil palm (OP; *Elaeis guineensis*) and pulpwood (*Acacia*) plantations (Miettinen et al. 2012; Figure 4.3). These plantations are for the most part managed by large-scale industries, while small-holder farmers mostly manage other kind of crops on peatlands such as coconut, pineapple, sago palm and rubber (Miettinen et al. 2012). OP plantation for palm oil production has expanded rapidly in the the past three decades and today Indonesia and Malaysia are the main global exporters providing 85% of the global palm oil demand (Fargione et al. 2008; Danielsen et al. 2009). This increase in demand has contributed to deforestation across the SE Asian region on peat soils (Figure 4.3). Between 1990 and 2010, more than 1 Mha of peatlands were converted to OP plantations (more than 15% of the total area) and only in the Jambi Province more than 75,000 ha (Miettinen et al. 2012). Miettinen et al. (2012) projections indicate that the extension of OP plantations on peatland may double by 2020.

■ Oil palm plantations	
■ Acacia plantations	Figure 4.3 Extent of oil palm and *Acacia* plantations on peatland in 2010 (modified from Miettinen et al. 2012)
■ Other \ unknown plantations	
■ Other mapped peatland	
■ Unmapped peatland	
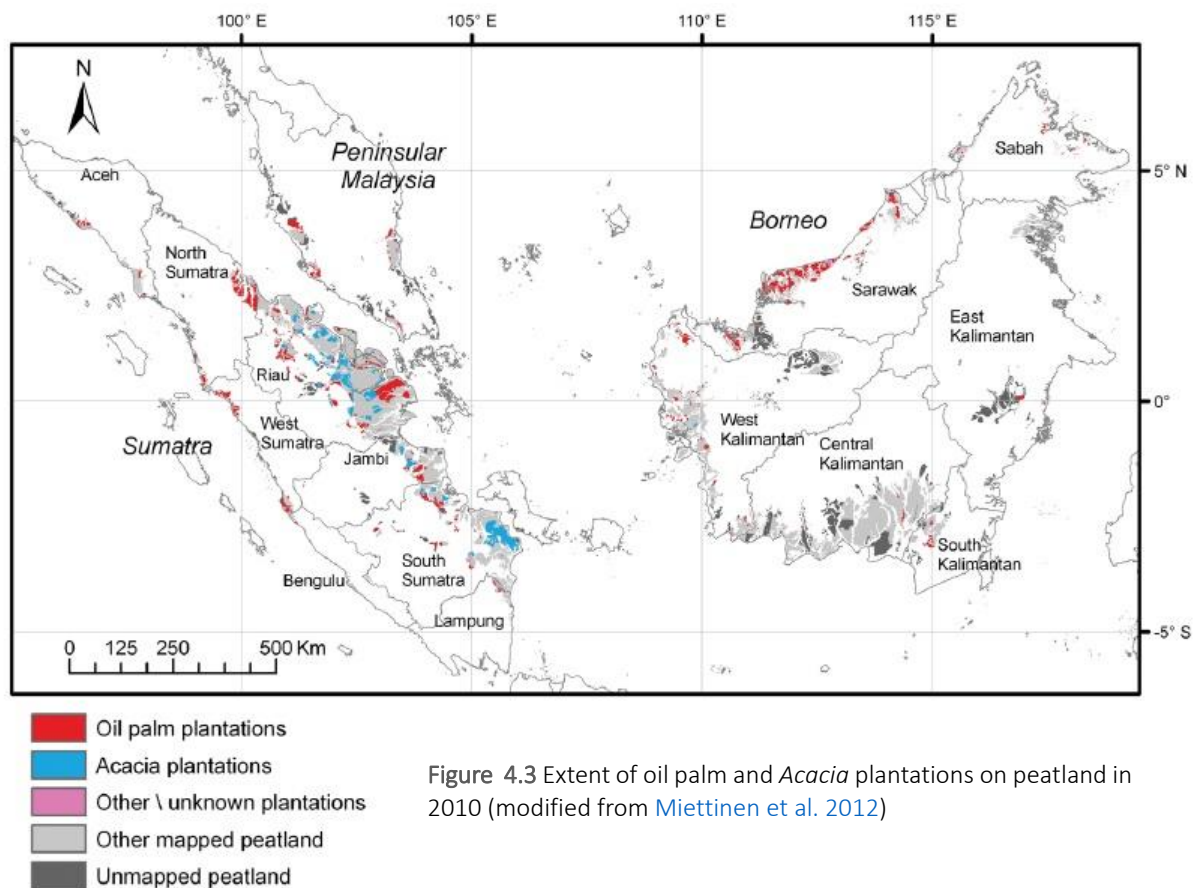
Peat formation

Peat deposits can be regarded as both orgainic soils and sedimentary deposits due to their mode of formation throught accumulation of organic material (Wüst et al. 2007). The main feauture of peat deposits beside having high organic content (50-60% of carbon by dry weight), is to have a very low bulk density with approximately 90% water (Neuzil 1997; Page et al. 1999; Hooijer et al. 2010; Page et al. 2011).

Peat accumulation occurs when net organic production exceeds net organic decomposition by bacteria and fungi and therefore largely depends on environmental conditions and settings. Despite the fact that the tropical climate (high temperature and annual precipitation) greatly favors decomposer organisms, the same conditions promote high primary production, with the result that often, production exceeds degradation, leading to net organic matter accumulation (Wüst et al. 2007). In the lowland tropics, input of organic matter derives mainly from the belowground root mass and partly from the aboveground biomass of the peat swamp forest (Cameron et al. 1989; Brady 1997; Wüst and Bustin 2004).

On the long-term, precipitation, radiation, temperature and most importantly local hydrology and geomorphology influence the development and differentiation of peatlands (Moore and Bellamy 1974; Barber 1981; Cecil et al. 1993; Esterle and Ferm 1994; Winston 1994). Similar to temperate peatlands,

the development of peatland can lead to deposits accumulating above river influence. Once this level is reached, peatlands may be more affected by factors such as climate variability, peat composition and hydrology (Clymo 1983).

Rates of peat accumulation are recorded from tropical ombrogenous deposits in SE Asia (Dommain et al. 2011). They are calculated based on radiocarbon dating with the assumption that the decay of peat below the water table is not significant. Holocene average rates within deposits ranged from 0.54 mm/yr in Central Kalimantan to 1.83 mm/yr for deposits in Sumatra. The accumulation in SE Asia is much higher than the average temperate and boreal peat accumulation rates (0.2-0.8 mm/yr; Gorham 1991; Page et al. 2004; Page et al. 2010).

Type of peatland in Southeast Asia
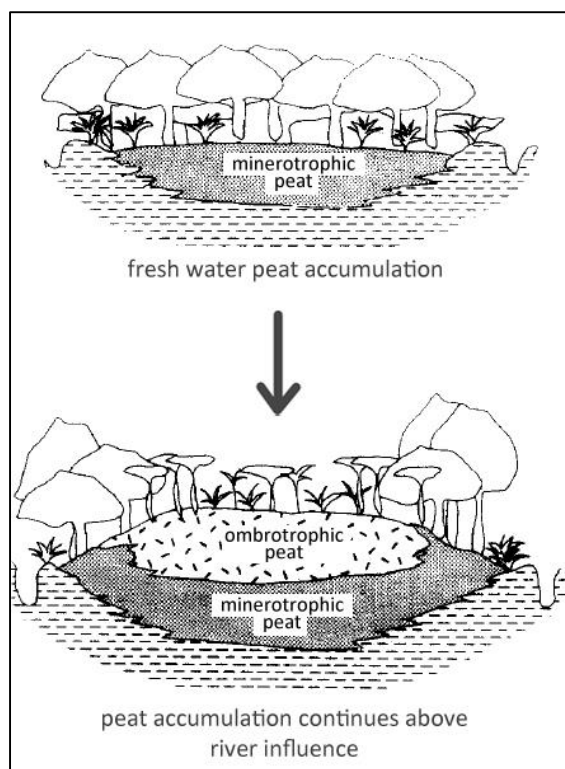
Figure 4.4 Diagrammatic cross sections of the stages of peat development (modified from Cameron et al. 1989)

Both ombrotrophic (rain-fed) and minerotrophic (receiving surface runoff and/or groundwater) peatlands can be found in SE Asia. Most of the ombrotrophic deposits evolved from a "primitive" minerotrophic peat stage with the influence of mineral rich groundwater (minerotrophic settings; Figure 4.4). Since ombrotrophic peat deposits receive all their water and nutrients from precipitation, they are generally acidic and nutrient-poor (Anderson and Muller 1975; Sieffermann et al. 1988; Rieley 1992; Neuzil 1997; Page et al. 2004). Tropical peatlands have been classified in several ways mostly as a tool for land-use planning and peatland development using attributes such as topographic position (e.g. Sieffermann et al. 1988; Morley 2013), vegetation communities (e.g. Anderson 1983) and fibre and peat composition (e.g. Esterle and Ferm 1994). According to Morley (2013), in SE Asia two main types of ombrotrophic peatland forests can be found: kerapah or watershed peats (Morley 2000), which mainly form inland, and basinal peats forming behind mangrove swamps. Kerapah peats are less common than basinal peats and they have been described so far only in Sarawak and Kalimantan (see Morley 2013 for an overview). The fact that the two types form on different locations make them different in regards to origin, substrate and long-term rates of peat accumulation (Anshari et al. 2001; Dommain et al. 2011). While kerapah peat swamp rainforests have developed in areas of podzolic soils (Brünig 1974; Brünig

1990; Morley 2013; Figure 4.5), basinal peats are found in Sarawak, Sumatra and Kalimantan both in inland and coastal locations. Close to the coastline, basinal peat swamp forests developed on prograding deltas behind mangrove swamps (e.g. Caline and Huang 1992). More in the inland, they can form along lowland river depressions in areas such as South Kalimantan (Sieffermann et al. 1992) and Central and South Sumatra (Sabiham 1990). Coastal peatlands developed more recently compared to the kerapah peatlands, usually following the stabilization of sea levels in the mid-Holocene and the uncovering of coastal areas following the lowering level of the sea after 5000 years ago (Dommain et al.
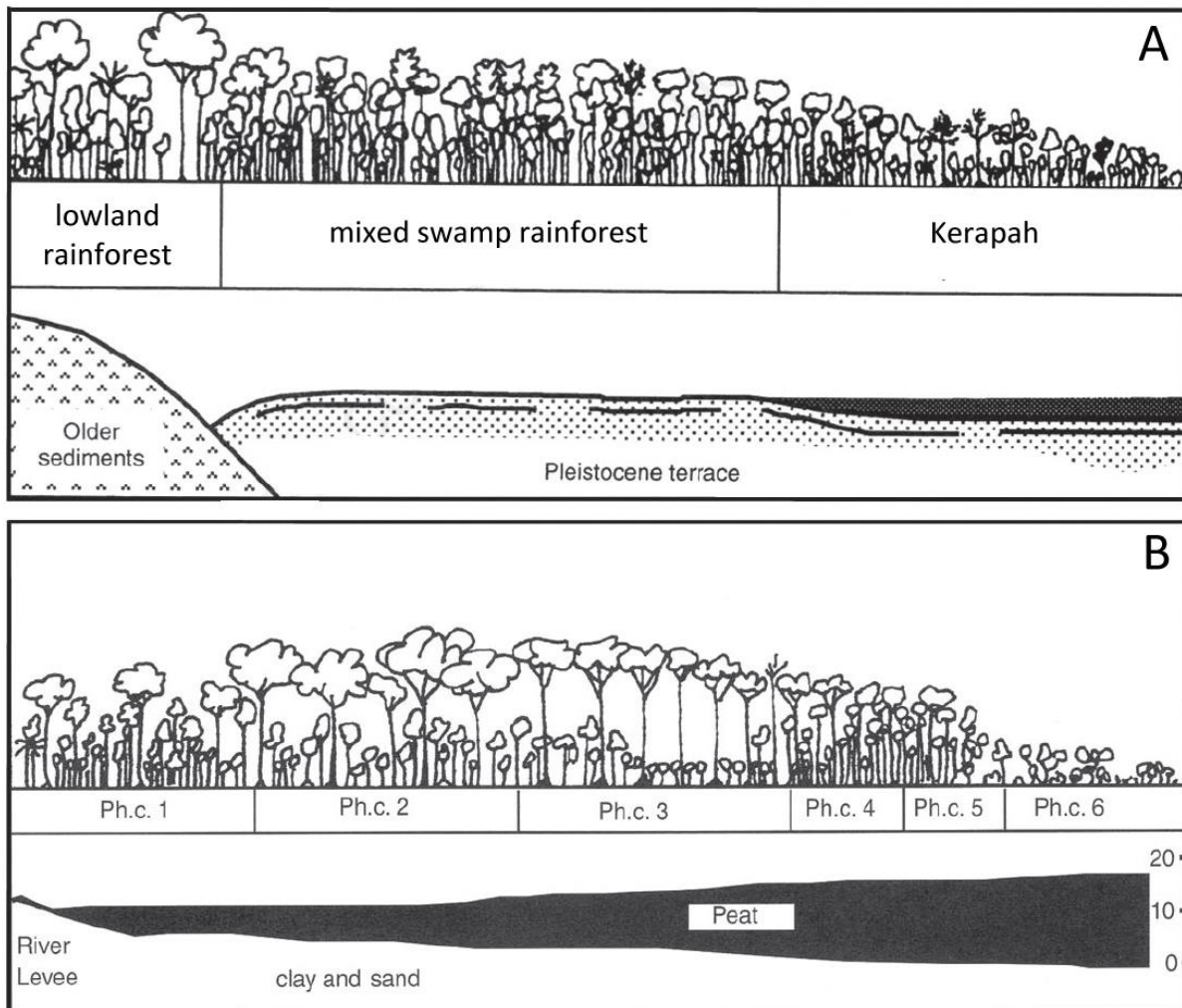
Figure 4.5 Kerapah (A) and basinal (B) peat catena compared. Ph.c. stands for Phasic communities (modified from Morley 2013)

2011).

The importance of vegetation communities

Although in general the water table closely follows the peat doomed surface, when Brady (1997) monitored the water table levels on peat with different thicknesses in Sumatra, he observed declining water table levels with increasing peat depths. The smallest amplitude of water level movement also occurred in thicker deposits. It follows that surface flooding seems to be less common in deeper peat compared to shallow peat. Brady (1997) additionally noted that surface flooding was found to be uncommon in the pole forests over deep peat in Kalimantan (Kostermans 1958) and Sarawak (Anderson 1961). Based on his results in Sumatra, Brady (1997) suggested that on thick peat deposits, peat accumulates more because of the accumulation of root biomass and declining decomposition, rather than from stagnant water logged conditions. The study of Grady et al. (1993) confirmed his hypothesis. While analyzing the fungal degradation of the maceral content of peat from Siaksriindrapura, Riau, they found that the root-dominated fibric peat found in deep deposits was more aerobic than that of the wood-dominated sapric peat found in thinner deposits. They suggested that peat in thicker deposits may be associated more with drier surface conditions compare to shallower deposits, where the influence of river flooding plays an important role. Similar results were found in Sumatra (Esterle and Ferm 1994) and Kalimantan (Moore and Hilbert 1992; Dehmer 1993). The accumulation of peat under these conditions, is therefore more strongly associated with the material deposited and therefore, plant species composition.

These studies highlight how crucially important the vegetation cover is for the internal development and peat accumulation in space and time. The lowland peatlands of SE Asia are covered with peat swamp forest that is arranged into concentric forest communities associated generally with increasing peat depth on ombrotrophic peat domes (Anderson 1983; Brünig 1990). Anderson (1961) used the term "catenary stages" or "catena" to refer to this sequential pattern of forest types in Sarawak. Botanically, tropical peatlands are more diverse than temperate peatlands. Different tropical domed peat swamp rainforests are similar in respect to the vegetation structure but they often markedly differ in their forest species compositions. So far, one of the best classifications of peat forests composition and structure is the one of Anderson (1961) based on peat swamp forests of Sarawak and Brunei. He distinguished six forest types or phasic communities along a gradient of peat depth. Forests of thin layers of peat at the edges of peat deposits showed high species richness and biomass. With increasing peat depth, tree heights, diameters, and species numbers declined. In the deepest peat in the central peat deposits, the forest types were characterized by a low canopy of pole-sized trees. These forest types have been referred to as "Padang or pole forest" (Figure 4.5).

Similarly to what was found by Anderson in Sarawak and Brunei, an investigation of the peat swamp forests in Sumatra revealed that vegetation communities on shallow and deep peat are markedly different in their composition and structure (Brady 1997; Rieley and Ahmad-Shah 1996). The vegetation

changes with increasing peat depth include a reduction in tree species, a decrease in tree height, a decrease in basal area, an increase in tree stem density, and an increase in surface root biomass from mixed forest (on shallow peat) to pole forest (on thicker peat deposits; Brady 1997).

The factors that drive these changes are linked to the decline in nutrient and moisture as the peat becomes deeper (Anderson 1961; Brünig 1971; Whitmore 1984). Nutrients enter the thick peatlands mainly via precipitation as river waters do not reach these peats which are located at higher elevations. As a result, only specialized plant species that have or can develop roots that can reach down to the underlying soil layers can trive under such conditions (Driessen 1977; Cameron et al. 1989).

Temporal and spatial development of Southeast Asian peatlands: state of knowledge

The onset and development of modern peat deposits in SE Asia range from the Late Pleistocene to the Holocene (from ca. 40 cal ka BP; Page et al. 2004). However, according to the review of Dommain et al. (2014), peat formed and accumulated for the most part after the LGM. Sea-level change has been identified by the authors as the primary driver for both peatland formation and expansion. This is related to the effect of sea-level change on both the regional atmospheric moisture availability and the hydrological gradient on the islands of the Sunda shelf (Dommain et al. 2014; Figure 4.6).
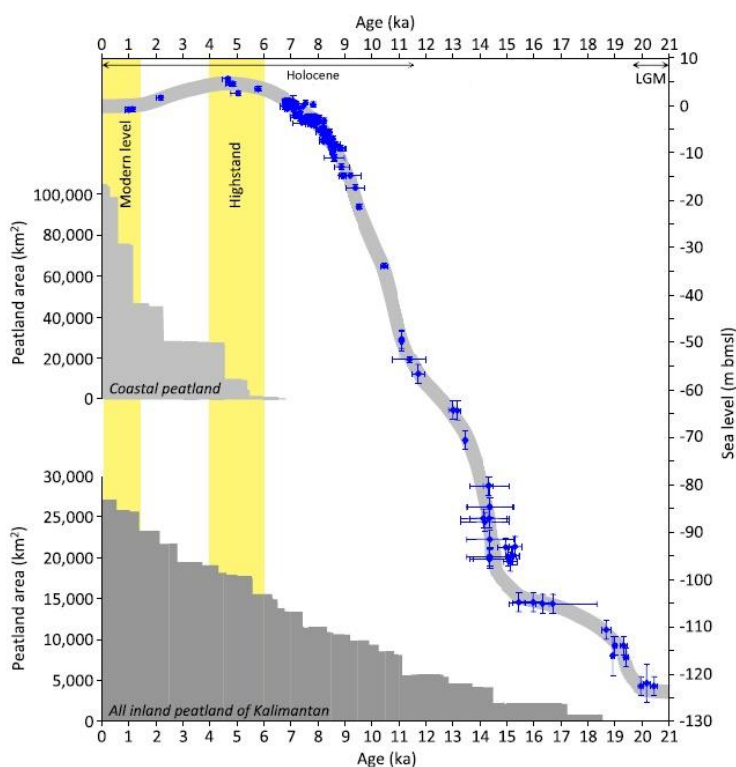
Figure 4.6 Areas of coastal (top) and inland (bottom) peatlands in time with sea-level curve of Sunda shelf (modified from Dommain et al. 2014)

The post LGM sea-level history of the Sunda Shelf has been divided in three main events: 1) rapid sea-level rise during deglaciation from ca. 19,000 to 7000 cal yr BP, 2) a slow rise to the Holocene high stand at about 5000 cal yr BP and 3) the subsequent lowering of sea level by about 5 m (Hanebuth et al. 2000; Hanebuth et al. 2011; Dommain et al. 2014). During the deglatiation, sea-level rose rapidly to flood a large portion of the Sunda Shelf and causing a rise in the water table and paludification of the depression and low lying areas in land. However, the rapid transgression prevented the expansion of peatlands until about 7000 cal yr BP, when the sea level rise

stabilization allowed for the first coastal peatlands in Sumatra to form (Sabiham 1988; Dommain et al. 2011). Dommain et al. (2011) noted a connection between higher seasonality and El Niño droughts during the late Holocene and reduced accumulation and peat truncation in Borneo. However, in Sumatra carbon accumulation was uninterrupted, possibly as a result of the large increase in land availability in the coastland regions due to sea-level regression (Dommain et al. 2011). In contrast to northern temperate peatlands, limited carbon-related field research has been conducted in tropical peat-swamp rainforests, thus the influence of climate and land-use on the ecosystems and on the carbon storage are not well understood (Farmer et al. 2011).

Palaeoecological studies represent a powerful tool, as they can be used to assess the impact of climate and environmental variability in the past as comparison to modern transformation led by human activities. However, little has been done in this direction so far in the peatlands of SE Asia, as only few studies have included palaeoenvironmental investigations (see Chapter 8 for details).

CHAPTER 5 - paper 1

ENSO and monsoon variability during the past 1500 years as reflected in sediments from Lake Kalimpaa, Central Sulawesi (Indonesia)

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Abstract

The climate of Sulawesi is driven by the monsoon system as well as the El Niño-Southern Oscillation (ENSO). Until now, mechanisms and long-term variations of these complex interacting climate processes have been poorly understood. This paper uses a sediment record from Lake Kalimpaa to investigate long-term rainfall trends of the past ca. 1500 years. Granulometric and geochemical data provide indications for an increasingly wetter climate (higher rainfall intensities and/or mean rainfall) on centennial to millennial time scales from approximately AD 560 to the 20th century. Highest rainfall intensities probably occurred at the end of the Little Ice Age (LIA). The trend towards wetter conditions during this time could also be detected in other palaeoclimatic studies from the region. A plausible explanation for these observations is the southward displacement of the Intertropical Convergence Zone (ITCZ) associated with changes in monsoon dynamics. However, comparison of the results with other proxy and model data indicates that the long-term rainfall variability in Central Sulawesi is also determined by variations in ENSO. During the 20th century the climate signal in the Kalimpaa record is superimposed by human impact. Moreover, the data suggest that two periods of disturbance occurred within the lake catchment between about AD 1090-1190 and AD 1450-1620. Comparison with fire frequency derived from macro-charcoal analysis indicates that these events were caused by forest fires which likely took place during periods of drought. Broadly simultaneous drought periods have been detected in records from East Java suggesting a regional drought occurrence affecting at least East Java and Sulawesi.

Introduction

Sulawesi belongs to the Greater Sunda Islands located in the Indo-Pacific warm pool (IPWP), which is the largest storage of warm surface ocean water globally (Gagan et al. 2004; Oppo et al. 2009). Consequently, this area is a substantial source of latent heat and has a great importance for understanding modern climate dynamics (Hope 2001).

Several rainfall regimes occur in Indonesia in response to, for example, variations in the monsoon system and the El Niño-Southern Oscillation (ENSO; Aldrian and Susanto 2003). The seasonal rainfall in eastern Indonesia is mainly driven by the intensity of the Australian-Indonesian summer monsoon (Griffiths et al. 2010), whereas the ENSO cycle (Diaz and Kiladis 1992) is responsible for the complex inter-annual climatic variability in this region. While ENSO cold events (La Niña) enhance the rainfall in Sulawesi, ENSO warm events (El Niño) result in dry periods, which occasionally have led to extreme droughts in the past (Salafsky 1994; D'Arrigo et al. 2006; Keil et al. 2008). These drought events facilitated the occurrence of forest fires in tropical rainforests, as recorded in Borneo and Sulawesi during the strong El Niño of 1997/98 (Rowell and Moore 2000; Siegert and Hoffmann 2000; Sastry 2002).
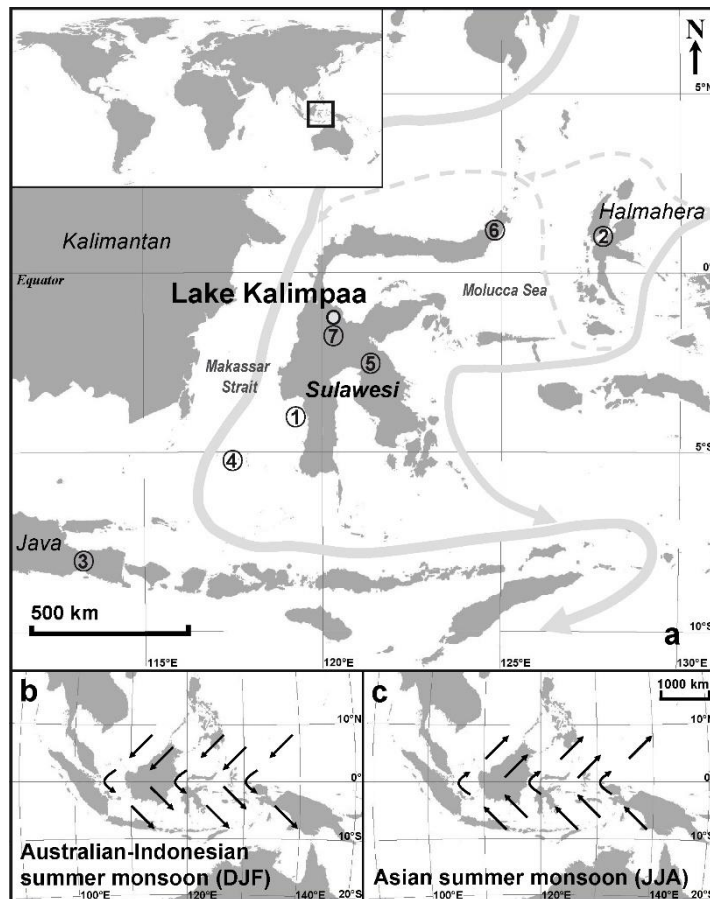
Figure 5.1 a) Map of the research area showing the location of Lake Kalimpaa on the island of Sulawesi. Also depicted are the locations of other study sites mentioned in the text: (1) Makassar Strait (Oppo et al. 2009; Tierney et al. 2010); (2) Halmahera: Kau Bay (Langton et al. 2008); (3) East Java: Lake Lamongan (Crausbay et al. 2006), Lake Logung (Rodysill et al. 2012), Lake Lading (Konecky et al. 2013); (4) Makassar Strait (Newton et al. 2006); (5) South Sulawesi: Wanda site near Lake Matano (Hope 2001); (6) North Sulawesi: Lake Tondano (Dam et al. 2001); (7) Besoa Valley, Central Sulawesi (Kirleis et al. 2011). The grey solid arrows represent the sea currents of the Indonesian throughflow (ITF) referring to Godfrey (1996) and Morey et al. (1999), whereas the dashed arrows depict further possible pathways. b) The generalized main wind directions over Sulawesi during the Australian–Indonesian summer monsoon (DJF: December-January-February) and c) during the Asian summer monsoon (JJA: June-July-August). The basic map was created using MapCreator 2.0

The annual influence of the monsoon is dominant in most parts of Sulawesi, where the wet season coincides with the Australian-Indonesian summer monsoon from October to March and the dry season with the Asian summer monsoon from April to September (Aldrian and Susanto 2003; Gunawan 2006; Figure 5.1b-c). An 'anti-monsoonal' rainfall pattern has been observed in the northeast of Sulawesi. This area representing only a vast minority of the island is characterized by inverse occurrences of wet seasons from April to August and dry seasons from September to March as well as a more pronounced impact of ENSO (Aldrian and Susanto 2003; Aldrian et al. 2004; Gunawan 2006). The mechanism underlying this pattern is the strong influence of the ocean and its sea surface temperatures (SST) in this region. Due to the Indonesian Through flow (ITF), marine currents from a Pacific warm pool northeast of New Guinea can enter the Indian Ocean (Godfrey 1996; Morey et al. 1999). These currents mainly pass the Makassar Strait, but also influence the SST in the Molucca Sea northeast of Sulawesi (Figure 5.1a). During the boreal winter, the ITF brings cooler water from the warm pool to the Molucca Sea inhibiting the formation of a convective zone in the northeastern part of Sulawesi (wet season in the monsoonal regions of Indonesia). In contrast, during the boreal summer, warmer water supplied from the warm pool promotes the formation of a convective zone in northeastern Sulawesi (dry season in most parts of Sulawesi; Godfrey 1996; Gordon et al. 1999; Morey et al. 1999; Aldrian and Susanto

2003; Gunawan 2006). According to Gunawan (2006), Central Sulawesi is characterized by a mixture of both monsoonal and 'anti-monsoonal' type of rainfall, and by the strong influence of ENSO.

Topography also influences rainfall and causes substantial variations over short distances. In the mountainous regions of the LLNP rainfall is mainly generated orographically (Gunawan 2006). During boreal winters, the Australian-Indonesian summer monsoon brings moist air masses to Central Sulawesi reaching the island from the northwest. In the mountains around Palu Valley the ascending air leads to orographic rainfall with a strong correlation between altitude and precipitation totals. In contrast, dry air masses reach Sulawesi during austral winters, when precipitation is lower and solely generated orographically (Gunawan 2006).

Thus far, neither the mechanisms nor the long-term variations of these complex interacting climate processes have been completely understood. Palaeoenvironmental data are needed to achieve more detailed knowledge of palaeoclimatic changes in Sulawesi; there are only a few palaeoclimatic studies from this region covering the past millennia. Marine records are available from Makassar Strait (Newton et al. 2006; Oppo et al. 2009; Tierney et al. 2010; Figure 5.1a) and Kau Bay (Halmahera; Langton et al. 2008); lacustrine sediments from East Java (Lake Lamongan: Crausbay et al. 2006; Lake Lading: Konecky et al. 2013; Lake Logung: Rodysill et al. 2012) provide terrigenous archives with a suitable temporal resolution. Palynological evidence are available from two sites in the Besoa Valley, Central Sulawesi, and the pollen data of one of these exhibits a cooling trend during the Little Ice Age (LIA; Kirleis et al. 2011). However, palaeoclimatic studies from Sulawesi are rare and cover longer time frames with lower temporal resolutions (Lake Tondano: Dam et al. 2001; Wanda site near Lake Matano: Hope 2001; Figure 5.1a). Therefore, it is necessary to carry out further investigations on palaeoenvironmental archives from Sulawesi to enhance the spatial and temporal resolution of proxy data.

This study aims to contribute to a better understanding of palaeoclimatic changes in Sulawesi during the past 1500 years. The main objective is to investigate various sedimentological characteristics of a core from Lake Kalimpaa (Central Sulawesi; Figure 5.1a) in order to provide indications for long-term rainfall trends, palaeoenvironmental changes and fire history in the lake catchment.

Site description

Lake Kalimpaa (1°19'34.8''S, 120°18'31.9''E; Figure 5.1a), sometimes also referred to as Danau Tambing, is located at 1660 m asl in the LLNP in Central Sulawesi, Indonesia. It occupies an area of 6.5 ha and has a maximum water depth of 6.6 m. The lake is surrounded by small reed belts and a swamp area to the north (Haberzettl et al. 2013); it has an inflow from the northeast and an outflow to the southwest.

Lake Kalimpaa is located in a mountain pass at the end of a side valley of the Palu Valley. Surrounding mountaintops in the northeast and southwest exceed altitudes of 2000 m asl. The lake's catchment area of approximately 1.8 km² extends mainly to the north. The geological setting is characterized by the

Pliocene Kambuno Granite (Simandjuntak et al. 1991; Priadi et al. 1994; Leemhuis 2005) and its weathering products. Regional vegetation is characterized by montane rainforest which in the LLNP, is presently characterized by a 90 % intact canopy cover (Kirleis et al. 2011) and is dominated by Fagaceae, mostly represented by the two genera *Lithocarpus* and *Castanopsis* (Culmsee et al. 2010).

There is no representative meteorological station in the vicinity of Lake Kalimpaa itself but there are records for the mountains of western Central Sulawesi and these can be used as representative of the catchment. Because of the close proximity to the equator temperatures are almost constant throughout the year. Daytime temperatures at higher altitudes range from 16 to 22 °C; annual precipitation is around 2000 to 3000 mm (Weber 2006).

Materials and methods

A 211 cm long composite sediment record consisting of three overlapping sections (KAL 1-1, KAL 1-2, KAL 1-3) was recovered from Lake Kalimpaa using a Livingstone piston corer (Livingstone 1955). Palaeomagnetic analyses showed that only KAL 1-1 and KAL 1-2 (153 cm composite length) were suited for palaeoenvironmental reconstruction since no reliable chronology could be established for the oldest part of the core (Haberzettl et al. 2013). Each section was split, photographed and described lithologically. The composite profile was compiled based on macroscopic marker layers. The chronology was adapted from Haberzettl et al. (2013; **Figure 5.2a**). Several sets of bulk samples which had been sent for radiocarbon dating revealed inconsistencies, i.e., ages were not in stratigraphic order. Since hard-water effects can be excluded for Lake Kalimpaa another kind of reservoir effect is assumed. Probably
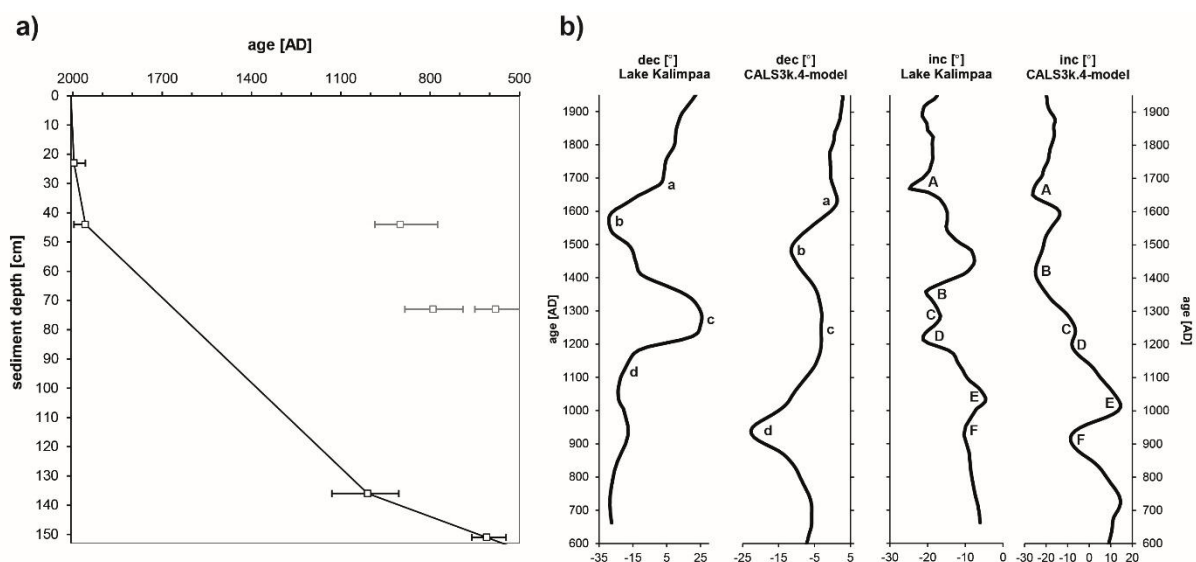
Figure 5.2 a) The age-depth model for the Lake Kalimpaa record based on calibrated radiocarbon ages which are presented as medians referring to the 2σ ranges (Haberzettl et al. 2013). The grey data points are regarded as outliers and have been excluded from the final age depth model (Haberzettl et al. 2013). b) Comparison of declination and inclination of the Lake Kalimpaa record (Haberzettl et al. 2013) with the CALS3k.4 model output for the location of Lake Kalimpaa (Korte and Constable 2011)

the dated material of the bulk samples comprised both autochthonous and allochthonous organic matter. Therefore, only the youngest ages were used for the age-depth model that is based on linear interpolation (Haberzettl et al. 2013; Figure 5.2a). Despite these dating uncertainties the created age-depth model seems to be a first order approximation which is corroborated by magnetostratigraphy. The comparison of palaeosecular variation data of the Kalimpaa record with the output of the CALS3k.4 model (Korte and Constable 2011) shows many isochronic similarities for inclination as well as for declination (Figure 5.2b). Therefore, the age-depth model can be used as basis for further multi-proxy approaches on the Lake Kalimpaa record. A marked increase in sedimentation rate is obvious for the most recent sediments. While the mean sedimentation rate between 151 and 44 cm is about 0.8 mm/yr, for the youngest sediments between 44 cm and the top of the core the age-depth model provides a sedimentation rate of ca. 9.2 mm/yr on average.

After palaeomagnetic measurements u-channels of KAL 1-1 and KAL 1-2 were subsampled for subsequent geochemical and granulometric analyses. Grain size distribution was measured at an interval of 1 cm (N=153) with a Laser Diffraction Particle Size Analyser (Beckman Coulter LS 13320) utilizing the Aqueous Liquid Module and 10 s ultrasonic for dispersion. The Fraunhofer diffraction theory was used for optical modelling of light scattering (De Boer et al. 1987). Organic matter and carbonates were removed with H_2O_2 (30 %) and HCl (10 %), respectively and sodium pyrophosphate solution ($Na_4P_2O_7 \cdot 10\,H_2O$; 0.1 M) was used as a dispersion medium. The measurements were carried out in several runs until a reproducible signal was obtained. The first reproducible run was taken for further statistical treatment. The grain size fractions were calculated according to Ad-hoc AG Boden (2005). The skewness (ϕ-scale) was determined by the logarithmic method of moments by means of a modified version of Gradistat 4.2 (Blott and Pye 2001).

Element concentrations of Al, Ca, Fe, K, Mg, Mn, P, S and Ti were measured at intervals of 2 cm (N=77) using ICP-OES (Varian 725-ES). For this purpose, oven-dried (50°C) aliquots were pestled (< 40 µm) and homogenized. The samples were digested with a microwave based procedure using modified aqua regia, which consists of 1 ml deionized water, 2 ml HCl (30 %) and 4 ml HNO_3 (30 %).

With the use of syringes, sediment samples were obtained at ca. 4 cm intervals (N=41) from the core and freeze-dried. Aliquots were ground and analysed for total organic carbon (TOC) and total nitrogen (TN) contents with a CNS element analyser (Euro EA 3000). TOC was determined after the destruction of carbonates with 2M H_3PO_4. Subsequently the molar TOC/TN ratio was calculated.

Additionally, mineralogical investigations on the pestled samples (N=10) were carried out using an X-ray diffractometer (D8-Discover, Bruker AXS) equipped with a CuKα X-ray tube and a gas proportional counter (HI-STAR area detector, Bruker AXS). The evaluation of the data was performed by means of the software Match! 2.0.9 and MacDiff 4.2.6.

Macro-charcoal particles (>150 µm) were counted for samples evenly spaced at 1 cm intervals along the upper part of the sediment core (first 145 cm). The samples (1 cm^3 each) were prepared following the method of Stevenson and Haberle (2005), which is a modification of a method developed by Rhodes (1998). Weak hydrogen peroxide (6% H_2O_2) was used to partially digest and bleach organic material in the sediment samples when counted under a binocular dissecting microscope. The sample preparation procedure aims to ensure that little particle fragmentation occurs during preparation. Results are expressed as number of charred particles per cm³.

Charcoal raw data were interpolated to constant 5 years, corresponding approximately to the median temporal resolution. Interpolated charcoal concentrations (number of particles cm^{-3}) were multiplied by estimated sedimentation rate (cm² yr^{-1}) to obtain the charcoal accumulation rate (CHAR, particles cm^{-2} per 5 yr) of each sample. Low-frequency variations in a charcoal record ($C_{background}$) represent changes in charcoal production, sedimentation, mixing, and sampling. $C_{background}$ was estimated with a locally weighted regression using a 100 yr window in order to maximize the signal-to-noise index and the goodness-of-fit between the empirical and the modeled C_{noise} distributions (Higuera et al. 2009). $C_{background}$ was subtracted to obtain a residual series, C_{peak}. It is assumed that C_{peak} is composed of two subpopulations (Higuera et al. 2008; Higuera et al. 2009): C_{noise}, representing variability in sediment mixing, sampling, and analytical and naturally occurring noise, and C_{fire}, representing charcoal input from local fires. A Gaussian mixture model was used for each sample to identify the C_{noise} distribution. The 99th percentiles of the C_{noise} distributions were considered as thresholds separating samples into 'fire' and 'non-fire' events. The peaks which passed the threshold criterion were subjected to a "Poisson minimum-count" screening in order to eliminate the peaks that result from statistically insignificant variations in charcoal counts. Peak fire episodes refer to one or more fires occurring within the time span of the charcoal peak. Past fire regime characteristics were inferred based on the temporal pattern of identified charcoal peaks via calculation of fire frequencies smoothed with a 200-year window. All statistical treatments were done using the program CharAnalysis (Higuera et al. 2009).

Proxy determination for palaeoenvironmental processes

Palaeorainfall proxies

Grain size can be used as a proxy for variations in transport energy or lake levels and hence, climate variability (Conroy et al. 2008). In detail, high rainfall intensities and/or amounts possibly result in enhanced erosion in the catchment as well as an increased transport capacity and competence of the tributary which might lead to the deposition of coarser clastic material in the lake (Håkanson and Jansson 2002). According to Nichols (2009) grain size can yield information on the flow velocity and hence the runoff during the time of sediment deposition.

In the Lake Kalimpaa catchment sediment erosion and transport is likely driven by runoff. Channel erosion and denudation of soil material are probably the major causes for the transport of sediments, although dense vegetation cover accompanied by high interception protects the soil surface.

It is assumed that the deposition of coarse grain sizes reflects periods or events characterized by high runoff and hence, higher rainfall intensities. In contrast, the deposition of fine grain sizes is linked to periods of lower average rainfall and thus, lower transport energy of the inflow. In such a small system probably lower runoff facilitates the deposition of finer particles due to the absence of turbulences in the lake water which are caused by the inflow during periods of greater rainfall and runoff.

The skewness (Sk) of grain size distributions of deposited sediments is a result of the composition of the source material as well as the energy level of the transport process. If transport processes exhibit a high-energy level, the deposited sediments become coarser and their grain size distributions are more positively skewed (on φ-scale). If the energy level is low, the deposited sediments become finer and more negatively skewed (McLaren and Bowles 1985).

It is supposed that the main source of Al, Ca, K, Mg and Ti are the minerals of the Kambuno Granite and its weathering products. K is mainly associated with K-feldspars, biotite and muscovite as well as, to a lower extent, with illite. Mg is related to biotite and clay minerals, such as montmorillonite; Ca occurs for instance in plagioclase and illite. Ti is chemically immobile and occurs mainly in heavy accessory minerals like rutile and ilmenite, which are extremely resistant against weathering (Goldich 1938; Shotyk et al. 2001; Li et al. 2003). Therefore, Ti was used in many other studies on lake sediments as an indicator for the input of clastic, terrigenous material (e.g. Haberzettl et al. 2005; Whitlock et al. 2008; Kasper et al. 2012) a process that is often driven by precipitation and runoff in the lake catchment.

During the chemical weathering of feldspars and micas to clay minerals, K, Ca, and Mg get dissolved. In contrast, Al which is part of feldspars and micas as well as clay minerals, is nearly insoluble and less mobile than the alkali and alkaline earth elements (Nesbitt et al. 1980; Middelburg et al. 1988). Kaolinite and gibbsite, which are common weathering products of granites under tropical conditions (West and Dumbleton 1970), contain Al but no K, Ca and Mg. Therefore, Al is used as denominator in various element/Al ratios in the following to compensate variable depletion effects which are primarily caused by concentrations of the redox sensitive element Fe.

In many other studies element/Al ratios like K/Al and Ti/Al are related to weathering intensities in the source area, input pathways or the strength of transport processes (Boyle 1983; Lückge et al. 2001; Muller et al. 2001; Zabel et al. 2001). Engstrom and Wright (1984) and Mackereth (1966) found that alkali and alkaline earth elements (e.g. Ca, K and Mg) accumulate in lake sediments during periods of intense erosion, when mineral matter is transported into the lake. In contrast, low values of alkali and alkaline earth elements in lake sediments occur when erosion is low and leaching of the catchment soils is dominant. Granite is usually deeply weathered in tropical climates and thus, unweathered minerals

like feldspars are more prominent in areas with higher slope angles (Berry and Ruxton 1959). Therefore, the source area of feldspars and micas is possibly the steeper slopes around the lake and their transport into the sediment occurs via terrigenous runoff during periods of high erosion. Accordingly, K/Al, Ca/Al and Mg/Al reflect the variability of the proportion of chemically less weathered feldspars and micas to clay minerals, especially kaolinite. The Ti/Al ratio mainly reflects changes in grain size (Boyle 1983; Zabel et al. 2001). Ti is associated with coarser material while Al rather represents the fine-grained fraction so that the Ti/Al ratio is linked to the strength of fluvial transport which reflects hydrological variability. Therefore, Ti/Al, K/Al, Mg/Al and Ca/Al ratios may be used as proxies for palaeorainfall similar to other studies on tropical lake sediments (Felton et al. 2007; Warrier and Shankar 2009). High ratios are interpreted as periods and/or events of high erosion and, accordingly, high rainfall intensities in the lake catchment.

Proxies for catchment disturbances and changes in redox conditions

Fire frequency data derived from macro-charcoal analysis is used to reconstruct the fire history around Lake Kalimpaa. Such forest fires may cause substantial disturbances within the lake catchment regarding changes in vegetation and sedimentological processes. Enhanced erosion as well as an increased supply of organic matter entering the lake can be consequences of these disturbances.

Fe, Mn, P, S, TOC and TN typically are part of a common reaction and transport cycle in sediments (Van Cappellen and Wang 1996). The significantly higher presence of allogenic Fe and Mn is characteristic of tropical lakes (Crowe et al. 2008). In soil samples from the upper horizons of nine sites in the catchment area of Lake Kalimpaa, average contents of ca. 2.5 % of Fe and ca. 340 ppm of Mn were measured (Markussen 2000). As it will be demonstrated later, Fe and Mn concentrations are significantly higher in certain sections of the sediment core than those in the topsoil. This cannot be explained solely by an increased input but rather by the accumulation of these elements due to another process(es), such as redox reactions in the sediment and/or the water column. Since this is accompanied by relatively high TOC/TN ratios which can indicate a shift to a more terrestrial origin of the organic matter (Mayr et al. 2005; Meyers 1994; Haberzettl et al. 2008), it is hypothesized that sharp increases in TOC, TN, S, Fe and Mn concentrations result from the supply of organic material from the catchment. This promotes reducing conditions in the sediment due to microbial decomposition. In contrast, low molar TOC/TN ratios likely arise from a comparatively low input of terrestrial organic matter or a higher occurrence of phytoplankton, which is typically characterized by TOC/TN ratios ranging from 4 to 10 (Meyers 1994; Kasper et al. 2013).

The microbiological decomposition of organic material in lake water and sediment is an oxygen-consuming process that produces anoxic environments (Davison 1993). In lakes, Fe and Mn occur in various oxidation states (Fe^{II}, Fe^{III}; Mn^{II}, Mn^{III}, Mn^{IV}) dependent on the given redox conditions. Under

oxidizing conditions both elements exhibit low solubility, where Mn is soluble at higher redox potentials than Fe (Sigg and Stumm 1991). Fe that enters lakes via rivers, occurs mainly in particulate form (> 99 %; Salomons and Förstner 1984) as ferric (oxyhydr)oxide or bound in the lattice of micas and clay minerals. Mn is also predominantly supplied by solids, such as MnO_2, and to lower amounts as dissolved Mn^{2+} (Engstrom and Wright 1984; Davison 1993). There seems to be a succession of redox processes in the sediment record from Lake Kalimpaa, during which particulate Fe(III)-(oxyhydr)oxides are dissolved to Fe^{2+} which precipitates as amorphous Fe(II)-sulphide or for instance mackinawite, or it becomes mobile and reprecipitates as Fe(III)-(oxyhydr)oxide at the oxic/anoxic boundary. The identification of Fe(II)-sulphides is inferred from the black sediment colour and its association with high Fe and S bulk concentrations (Emerson 1976; Engstrom and Wright 1984). The affinity of P to be adsorbed on the surfaces of Fe-(oxyhydr)oxides is known from the literature (e.g. Lijklema 1980; Buffle et al. 1989). According to López et al. (2006), high Fe and P values are indicative for the precipitation and accumulation of authigenic Fe. Moreover, P is also associated with organic matter.

Results

Core lithology and mineralogy

 The core sections KAL 1-1 and 1-2 consist mainly of finely laminated silts with a few homogeneous sections and distinct sand layers. Based on obvious changes in sediment structure, colour and macroscopic grain size, six lithological units can be distinguished (Figure 5.3). Unit I (153-128 cm; ca. AD 560-1090) exhibits finely laminated blackish grey and light grey silts. Unit II (128-118 cm; ca. AD 1090-1190) is characterized by homogeneous black sediments composed of fine silts to fine sandy coarse silts. Finely laminated, light greyish to grey layers consisting of medium to fine sandy coarse silts occur in unit III (118-93 cm; ca. AD 1190-1450; Figure 5.3). Mica particles, probably muscovite, are conspicuous between ca. 110 and 40 cm (ca. AD 1275-1965). Partially laminated, dark brown to black fine sandy coarse silts occur in unit IV (93-77 cm; ca. AD 1450-1620). Unit V (77-48 cm; ca. AD 1620-1910) exhibits a greyish colour and contains partially laminated, fine sandy coarse silts with an intercalated sand layer. The uppermost unit VI (48-0 cm; ca. AD 1910-2006; Figure 5.3) consists of brownish medium to fine sandy silts. It is assumed that human impact in the catchment area (e.g. by road construction, camp site) superimposes any effects of climatically induced changes during the 20th century (Haberzettl et al. 2013). Nevertheless, it cannot be excluded that also climate could have caused the marked increase in sedimentation rate for the most recent sediments. Since the distinction between climatic and human-driven impacts seems not possible with certainty, no palaeoclimatic interpretation has been done for unit VI in the following. X-ray diffractograms of all measured samples show comparable spectra. The most common sediment minerals are quartz, alkali feldspars, plagioclase,

muscovite, biotite (chemically unweathered minerals from the Kambuno Granite), kaolinite and to a lesser extent illite (weathering products of the Kambuno Granite). In some samples goethite and gibbsite were detected, which are typical for weathered granites in tropical regions as well as clay minerals (West and Dumbleton 1970). Furthermore, X-ray diffraction spectra indicate that carbonates are of minor importance in this record which was confirmed by multiple negative tests with hydrochloric acid.
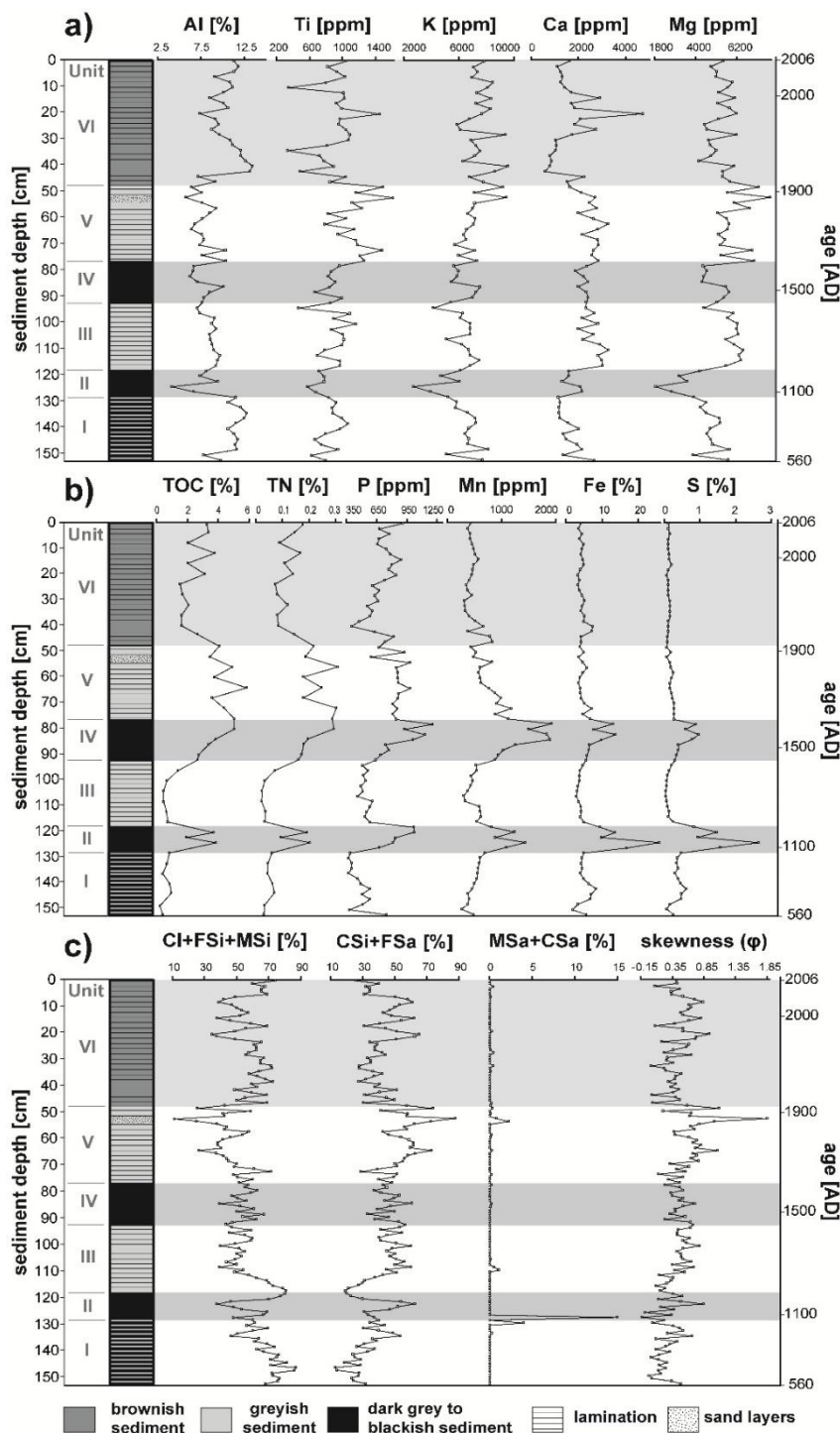
Figure 5.3 a) and b) Geochemical and c) granulometric (Cl: Clay, FSi: fine silt, MSi: medium silt, CSi: coarse silt, FSa: fine sand, MSa: medium sand, CSa: coarse sand) parameters of the Kalimpaa sediment core plotted versus sediment depth (age scale is also depicted on the right). A schematic illustration of the lithology is given on each plot on the left. The grey and white bars represent the lithological units

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Grain size analysis

The core sections KAL 1-1 and 1-2 are dominated by medium to coarse silt and accessorily contain some fine sand layers. With a mean of ca. 78.3 % in all samples, silt is the most common fraction followed by sand (ca. 13.5 %) and clay (ca. 8.2 %). A correlation matrix of the calculated grain size fractions (Table 5.1) exhibits positive correlations between clay (Cl), fine silt (FSi) and medium silt (MSi). Coarse silt (CSi) and fine sand (FSa) are also positively correlated but negatively correlated to the aforementioned. Moreover, there is a positive correlation between medium sand (MSa) and coarse sand (CSa) which are only sporadically present in the record (Table 5.1). On this basis, the grain size fractions can be grouped into classes, viz. one of Cl+FSi+MSi and one of CSi+FSa. They show an opposing trend (Figure 5.3c) but depend on each other and thus, probably represent the same palaeoenvironmental signal. In contrast, the third class consisting of MSa+CSa is interpreted as an additional signal.

Units I and II (153-118 cm; ca. AD 560-1190) are characterized by low but slightly increasing CSi+FSa and Sk values as well as high but slightly decreasing Cl+FSi+MSi values (Figure 5.3c). The most prominent peak in MSa+CSa occurs at the base of unit II (at 128 cm; ca. AD 1090). The lowermost section of unit III (118-112 cm; ca. AD 1190-1250) exhibits low values for CSi+FSa and Sk as well as high values for Cl+FSi+MSi again. For the remaining part of unit III, IV and V (112-48 cm; ca. AD 1250-1910), CSi+FSa and Sk values first show a decreasing and then an increasing trend reaching their maxima within unit V (at 53 cm; ca. AD 1860; Cl+FSi+MSi vice versa). Over the entire record (units I to V), an increasing trend of CSi+FSa and Sk values as well as a decreasing trend of Cl+FSi+MSi values can be identified.

Table 5.1 Correlation matrix including the Pearson product–moment correlation coefficient and the related p-value of the grain size fractions from all measured samples of the Lake Kalimpaa record. The limits of grain size fractions are according to Ad-hoc AG Boden (2005) as follows: clay (Cl), fine silt (FSi), medium silt (MSi), coarse silt (CSi), fine sand (FSa), medium sand (MSa) and coarse sand (CSa). Highly significant correlations ($p < 0.001$) are highlighted (bold: highly positively correlated, bold italic: highly negatively correlated)

	Cl	FSi	MSi	CSi	FSa	MSa	CSa
	Pearson's r						
	<2 μm	2–6.3μm	6.3–20 μm	20–63 μm	63–200 μm	200–630μm	630–2000μm
	p						
Cl	–	0.92	0.74	−0.81	−0.79	−0.10	−0.07
FSi	<0.001	–	0.77	−0.88	−0.80	0.01	0.03
MSi	<0.001	<0.001	–	−0.57	−0.94	−0.19	−0.11
CSi	<0.001	<0.001	<0.001	–	0.50	−0.18	−0.14
FSa	<0.001	<0.001	<0.001	<0.001	–	0.14	0.04
MSa	0.23	0.93	0.02	0.02	0.09	–	0.95
CSa	0.37	0.76	0.18	0.10	0.65	<0.001	–

Geochemistry

With an average concentration of 9.3 % Al is the most abundant of the measured elements. It shows a decreasing trend from unit I to unit V (153-48 cm; ca. AD 560-1910; Figure 5.3a). Concentrations of K, Mg and Ti show an increasing and compared to Al opposite trend. Mg and Ti reach their maximum at 53 cm (ca. AD 1860) together with maximum values in CSi+FSa and Sk. Ca values are low in unit I and II (153-118 cm; ca. AD 560-1190) and on a higher level in the units III-V (118-48 cm; ca. AD 1190-1910; Figure 5.3a).

All graphs of the element/Al ratios exhibit a similar pattern (Figure 5.4). They are characterized by low values in unit I (153-128 cm; ca. AD 560-1090) and exhibit an increasing trend starting from unit II to V
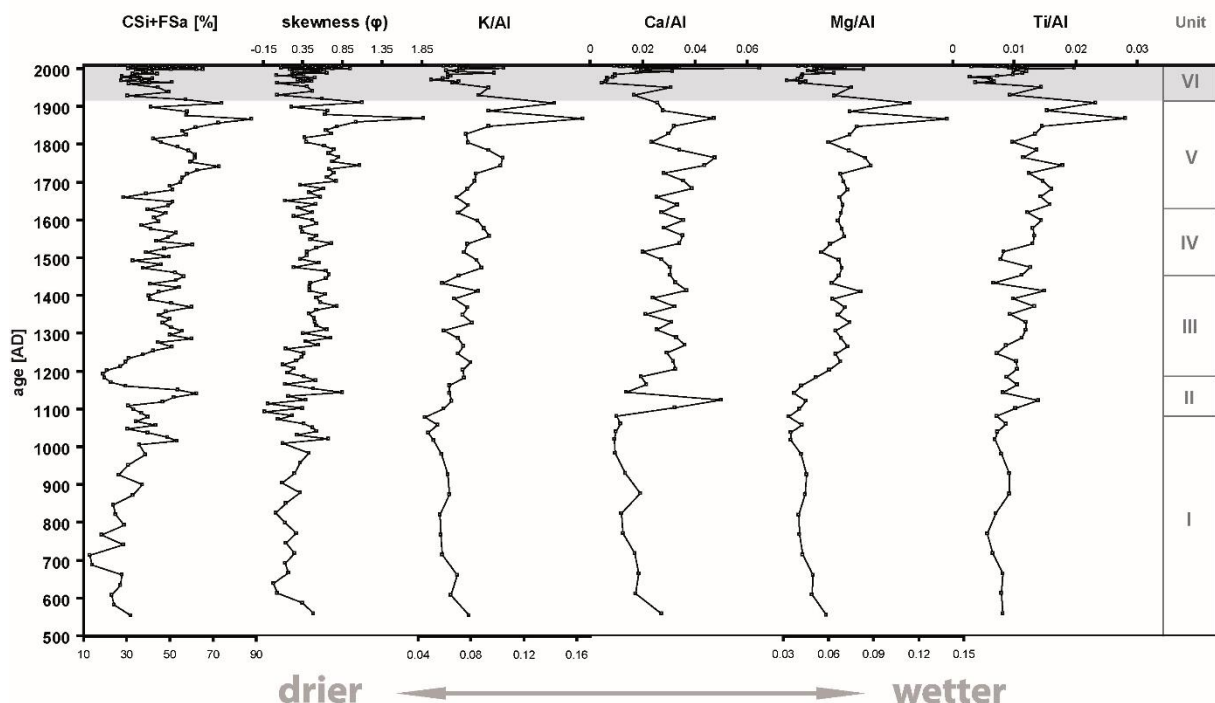
Figure 5.4 Granulometric data (CSi + FSa: coarse silt + fine sand; skewness) as well as selected element/Al ratios of the Kalimpaa record plotted versus age. All graphs are interpreted to exhibit a rising trend to wetter conditions (higher rainfall intensities and/or mean rainfall) on centennial to millennial time scale from AD 560 to 1910. Note that the uppermost unit VI (grey bar) is not included in the palaeoclimatic interpretation

(128-48 cm; ca. AD 1090-1910) with maximum values at 53 cm (ca. AD 1860). Increased element/Al ratios coincide for large parts of the record with the macroscopic conspicuously presence of mica particles. A similar pattern in Ti/Al, K/Al, Mg/Al, and grain size data is especially notable for the sections comprising the highest peaks within unit V (at 53 cm, ca. AD 1860 and 49 cm, ca. AD 1910).

TOC and TN are highly correlated (r = 0.97; p < 0.001). TOC values range from 0.24 % up to 5.85 %, TN from 0.02 % up to 0.31 % (Figure 5.3b). Both are low in unit I (153-128 cm; ca. AD 560-1090) and increase markedly in unit II (128-118 cm; ca. AD 1090-1190), with lower levels in unit III (118-93 cm; ca.

AD 1190-1450) and subsequently higher again in unit IV (93-77 cm; ca. AD 1450-1620). TOC and TN values remain on high levels in unit V (77-48 cm; ca. AD 1620-1910), where they reach their maximum of 5.85 % for TOC and 0.31 % for TN. The molar TOC/TN ratio shows two levels (Figure 5.5). The lower level in unit I (153-128 cm; ca. AD 560-1090) exhibits values of around 14.3 and the higher level within units II-V (128-48 cm; ca. AD 1090-1910) is characterized by mean values around 22.0. Fe concentrations average 5.7 % in the Kalimpaa record (Figure 5.3b) and reach a maximum of 25.7 % within unit II (128-118 cm, ca. AD 1090-1190). Fe and S concentrations are strongly correlated (r = 0.96; p < 0.001) and exhibit distinctive peaks in the units II and IV (128-118 cm, ca. AD 1090-1190 and 93-76 cm, ca. AD 1450-1620; Figure 5.3b). Mn and P values also show peaks in these two sections, but are characterized by higher values in the sediments above and below the peaks of Fe and S. Especially in unit V (77-48 cm; ca. AD 1620-1910) Mn and P values are above average and show a similar pattern to TOC and TN (Figure 5.3b).

Macro-charcoal and fire frequency

A total of 11 fire episodes occurred locally during the past 1300 years, 7 additional fire episodes failed to pass the screen test (Figure 5.5). Fire frequencies were higher in unit II (up to 2 fires per 200 years at
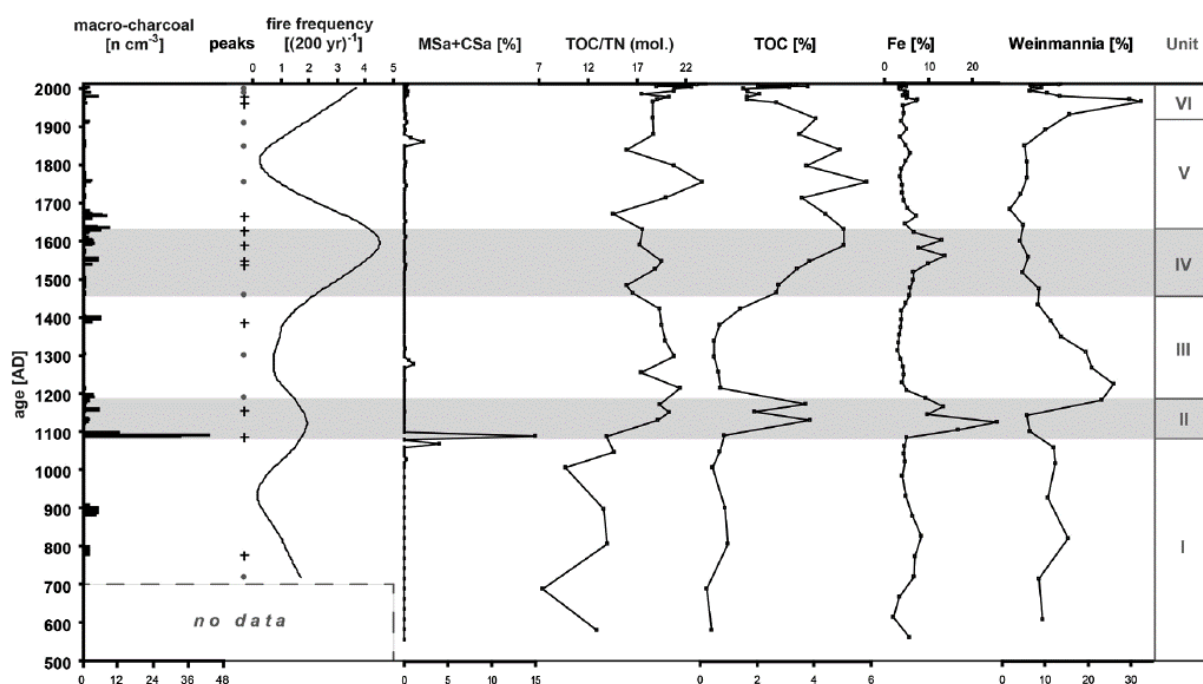
Figure 5.5 Macro-charcoal concentrations (number of particles per cm³), fire frequency, Fe, TOC and TOC/TN (molar) as well as the medium and coarse sand content (MSa + CSa) of the Kalimpaa record plotted versus age. Fire frequency calculated on the fire peaks detected with the software CharAnalysis, smoothed with a 200-year window. Fire peaks which passed the 'Poisson minimum-count' screening are indicated with black crosses, and the insignificant peaks are indicated as grey circles. Additionally shown are the *Weinmannia* pollen data (Biagioni et al. 2015a) which probably act as secondary forest species in the Lake Kalimpaa catchment. The grey bars represent the lithological units II and IV (AD 1090–1190 and AD 1450–1620) which are interpreted as periods of disturbance events in the Kalimpaa catchment

128-118 cm; ca. AD 1090-1190) and unit IV (up to 4 fires per 200 years at 93-76 cm; ca. AD 1450-1620). In unit VI (48-0 cm; ca. AD 1910-2006) fire frequencies gradually increased until today.

Discussion

Palaeorainfall in Central Sulawesi

Based on variations in the established palaeorainfall proxies (K/Al, Ca/Al, Mg/Al, Ti/Al, CSi+FSa, Sk) the Lake Kalimpaa record likely reveals a centennial to millennial-scaled trend towards wetter conditions characterized by higher rainfall intensities and possibly higher mean rainfall from ca. AD 560 to the 20th century. The period from ca. AD 560-1090 was characterized by drier conditions, while an increasingly wetter climate can be inferred from ca. AD 1090 to 1910 (Figure 5.4).

A long-term trend towards wetter conditions is also observed in sediments from two other lakes in East Java (Rodysill et al. 2012; Konecky et al. 2013; Figure 5.6e). Rodysill et al. (2012) explain the trend as a consequence of migration of the ITCZ with increased precipitation during its southward displacement. In comparison, Konecky et al. (2013) conclude that the migration of the ITCZ influences the climate variability on multidecadal to centennial time scales while it is the strengthening of the Walker circulation and its associated changes in ENSO variability that produces the increasingly wetter climate during the last millennium. Yan et al. (2011) arrive at similar conclusions, namely, precipitation changes in response to the combined influence of the migration of the ITCZ and the position and strength of the Pacific Walker circulation in the western tropical Pacific.

The palaeorainfall proxies obtained from the Lake Kalimpaa record show long-term similarities with ENSO variability from the eastern Pacific region as well as the South American continent (Figure 5.6a-c). An ENSO record derived from the sediments of Laguna Pallcacocha, southern Ecuador, reveals that the number of El Niño events per 100 years decreased since ca. AD 1150 until today (Moy et al. 2002) consistent with increasing terrigenous input to Lake Kalimpaa (Figure 5.6a-c). Another lacustrine archive from Galápagos exhibits a decreased ENSO frequency between AD 1300 and 1850 (Conroy et al. 2008). From Kau Bay (Halmahera; Figure 5.1a), a region where rainfall is strongly dependent on ENSO, it is known that the El Niño activity decreased steadily from ca. AD 1300 (Langton et al. 2008). A long-term trend starting from ca. AD 560 onwards is visible in both the established palaeorainfall proxies for the Kalimpaa record and the output of the coupled ocean atmosphere model from Clement et al. (2000; Figure 5.6a-b) which reflects the number of El Niño events occurring over a 500 year interval. These similarities of the ENSO and CSi+FSa data (Figure 5.6a-b, c) indicate that palaeorainfall in Central Sulawesi is related to the variability of ENSO on centennial to millennial time scale and thus, changes in the Pacific Walker circulation.

Two studies carried out at Makassar Strait (Newton et al. 2006; Oppo et al. 2009; Figures 5.1, 5.5d) revealed that the SST in this region was ca. 0.5 to 1.5 °C lower during the LIA (ca. AD 1550-1850) compared with modern SST and those during the Medieval Warm Period (MWP; ca. AD 900-1300; Oppo et al. 2009). The authors infer that a cooling of the North Pacific surface water, which is transported by ocean currents through the Makassar Strait, is responsible for lower SST during this period. These cooler SST phases are associated with the southward displacement of the Intertropical Convergence Zone (ITCZ) during the LIA, which led to wetter conditions at Makassar Strait (Newton et al. 2006), similar to the region around Lake Kalimpaa in Central Sulawesi (Figure 5.6a-d). Sachs et al. (2009) note that the ITCZ reached its southernmost position between AD 1400 and 1850 with the result that drier conditions occurred north of the equator and wetter conditions in the southern tropics (Newton et al. 2006). Tierney et al. (2010), who examined marine sediments off the coast of Southwest Sulawesi (Figure 5.1a), inferred that the hydrological variability and the monsoon strength in the Indo-Pacific warm pool (IPWP)
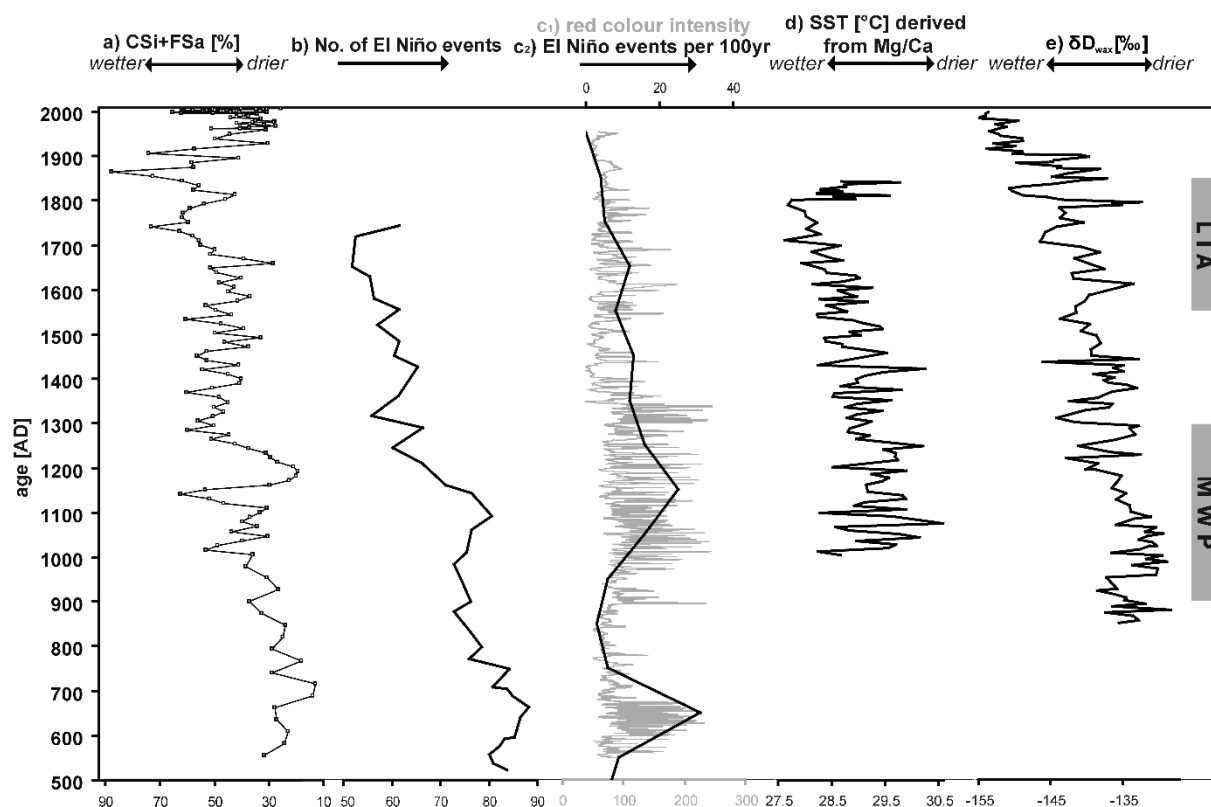
Figure 5.6 Comparison of a) coarse silt and fine sand data (CSi + FSa) representing terrigenous input to Lake Kalimpaa with other studies; b) modelled number of El Niño events using input data from the eastern Pacific region. Shown is the number of events in 500-year overlapping windows (overlapping every 10 years; digitized from Clement et al. 2000); c1) red colour intensity of sediments from Laguna Pallcacocha, Ecuador, representing the layers deposited during El Niño events (grey) and (c2) the resulting modelled number of El Niño events per 100 years (black; Moy et al. 2002); d) reconstructed sea surface temperatures of Makassar Strait derived from Mg/Ca ratios (Newton et al. 2006); and e) leaf wax δD from Lake Lading (East Java) as palaeorainfall proxy (Konecky et al. 2013). The periods of the 'Little Ice Age' (LIA; AD 1550 to 1850) and the 'Medieval Warm Period' (MWP; AD 900–1300) according to Oppo et al. (2009) are shown on the right

are dependent on migrations of the ITCZ. These authors suggest teleconnections between ENSO and the monsoon with a weak Indian monsoon and a more El Niño-like mean state during the MWP and a strong Indian monsoon and a more La Niña-like mean state during the LIA (Tierney et al. 2010). It seems therefore that the southward displacement of the ITCZ may have caused higher rainfall intensities and possibly higher mean rainfall around Lake Kalimpaa during the LIA.

Both the monsoon and ENSO climate systems are interacting through teleconnections (Ju and Slingo 1995; Soman and Slingo 1997; Torrence and Webster 1999). Variations in the Pacific Walker circulation and/or SST anomalies during El Niño events could influence the global scale divergence, which can result in a shift of the ITCZ and thus, changes of the monsoon dynamics (Ju and Slingo 1995; Soman and Slingo 1997; Torrence and Webster 1999). The position of the ITCZ as well as the mean state of ENSO/Pacific Walker circulation, therefore, may be considered interactive and co-responsible for the long-term rainfall variability in Central Sulawesi on centennial to millennial time scale. The interpretation of the Kalimpaa record suggests a general weakening of the Pacific Walker circulation and a more El Niño-like mean state from ca. AD 560 to 1090. In contrast, the strengthened Pacific Walker circulation and the southward displacement of the ITCZ probably are the causes for the wetter climate in Central Sulawesi during the LIA, which is consistent with the interpretation of Yan et al. (2011).

Palaeoenvironmental disturbance events in the catchment of Lake Kalimpaa

In the Lake Kalimpaa record, peaks in Fe, Mn, P, S, TOC and TN occur almost simultaneously with increased fire frequency in unit II (128-118 cm; ca. AD 1090-1190) and unit IV (93-76 cm; ca. AD 1450-1620; Figure 5.5). The most prominent peaks of these elements coincide in unit II with macro-charcoal and MSa+CSa (Figure 5.5). Increased fire frequency data indicate periods during which forest fires occurred more frequently in the drainage basin of Lake Kalimpaa. Hence, it is assumed that these disturbance events are a potential cause for the increased supply of organic material into the lake, which in turn may lead to enhanced TOC, TN and TOC/TN values. A similar pattern showing increased TOC and charcoal values was found in the sediments from Lago dell'Accesa (Tuscany, Italy; Vannière et al. 2008). As a result of microbial decomposition of the organic matter, anoxic conditions prevailed at Lake Kalimpaa and thus, enhanced the formation of black sediment layers which contain high Fe, Mn, P and S amounts.

As noted, the deposition of MSa+CSa seems to reflect a different process compared to the other grain size fractions (Table 5.1). Due to the simultaneous occurrence of the most prominent peaks in MSa+CSa and macro-charcoal (Figure 5.5), the deposition of these coarse materials is interpreted as an input signal after catchment disturbance (Cerdà and Lasanta 2005) which is likely associated with a forest fire around ca. AD 1090 (Figure 5.5). Thus, the fire likely facilitated the erosion and deposition of MSa+CSa.

Following the first period of inferred increasing fire frequency (ca. AD 1090-1210) changes in the vegetational composition occurred within the catchment. These alterations are reflected within the units II and III (123-95 cm; ca. AD 1140-1430) and were characterized by an expansion of *Weinmannia* which probably acts as secondary forest species in the Kalimpaa drainage basin (Biagioni et al. 2015a). In contrast, *Weinmannia* pollen concentrations remained low after the second period of disturbance (ca. AD 1460-1620). However, pollen data of fast-growing pioneer species like *Macaranga* and *Mallotus* increased during this period indicating disturbance in the forest (Biagioni et al. 2015a). *Weinmannia* increased again in unit VI (47 to 35 cm; ca. AD 1930-1975), when the vegetation cover in the catchment area was likely disturbed due to the road construction (Figure 5.5).

Considering the macro-charcoal and sedimentological findings from KAL 1-1 and 1-2, at least two disturbance events occurred in the catchment area that are reflected in units II and IV (ca. AD 1090-1190; ca. AD 1450-1620). The changes in fire frequencies between ca. AD 1090-1190 and ca. AD 1450-1620 have affected the geochemical composition of the lacustrine sediments. The alterations within the catchment during the 20th century are possibly caused by human impact (Haberzettl et al. 2013).

From archaeological and palynological investigations in Central Sulawesi it has been concluded that anthropogenic impact in the Besoa Valley, ca. 25 km south of Lake Kalimpaa, started ca. 2000 years ago (Kirleis et al. 2011; Kirleis et al. 2012) when the montane rainforest was replaced by grassland. However, considerable human modifications of the landscape in the catchment of Lake Kalimpaa are not assumed before the 20th century.

Natural fires occurring during drought periods seem to be the most likely triggers of the assumed disturbances in the Kalimpaa drainage basin. Probably drought stress accompanied by increased plant mortality (McDowell et al. 2008; Zach et al. 2010) fostered the forest fires as it is known from eastern Borneo during the strong El Niño event from 1997/98 (Siegert and Hoffmann 2000; Van Nieuwstadt and Sheil 2005), when fires also occurred on Sulawesi (Rowell and Moore 2000; Sastry 2002). Droughts in Indonesia result from the failure of the monsoon, which often coincides with ENSO warm events (D'Arrigo et al. 2006). According to Quinn et al. (1978), over 90 % of droughts in Indonesia during the period from AD 1861 to 1976 are associated with a warm phase of ENSO.

The ca. 800 year-record obtained from Lake Lamongan (Crausbay et al. 2006; Figure 5.1a) reveals two periods of multidecadal drought from ca. AD 1275-1325 and ca. AD 1450-1650 (Figure 5.7), probably as a result of ENSO variations. More recent investigations on the timing of droughts in East Java have been carried out by Rodysill et al. (2013) who use U-series dating and suggest that the onset of the latter drought at Lake Lamongan was more than 300 years later around ca. AD 1790.

In sediments from Lake Logung (also East Java; Figure 5.1a) spanning the past ca. 1400 years, the long-term trend towards wetter conditions was superimposed by four decadal to centennial-scale droughts between ca. AD 930-1130, ca. AD 1460-1640, ca. AD 1790-1860 and ca. AD 1985-2008 (Rodysill et al. 2012; Figure 5.7). The authors discuss these drought occurrences in relation to both migration of the ITCZ and variability in ENSO. Two of the four droughts (ca. AD 1460-1640 and ca. AD 1790-1860) took place when the ITCZ was displaced to the south, a period that is actually characterized by a wetter mean climate on centennial to millennial time scale. These droughts hence represent unusual events on an inter-annual to multidecadal time scale.

Comparisons of the data from Lake Kalimpaa with the drought occurrences observed by Crausbay et al. (2006) and Rodysill et al. (2012, 2013) show that disturbance events at Lake Kalimpaa partly coincide with drought periods in East Java. The older two of four drought periods observed in the Lake Logung record, match temporally with disturbance events at Lake Kalimpaa (Figure 5.7; Rodysill et al. 2012). Therefore, it is suggested that they are regional in spatial extent since age differences are within the range of dating uncertainties. The second drought period (ca. AD 1450 to 1620) at Lake Kalimpaa coincides well with findings from Lake Logung (ca. AD 1460-1640) and the radiocarbon dated drought
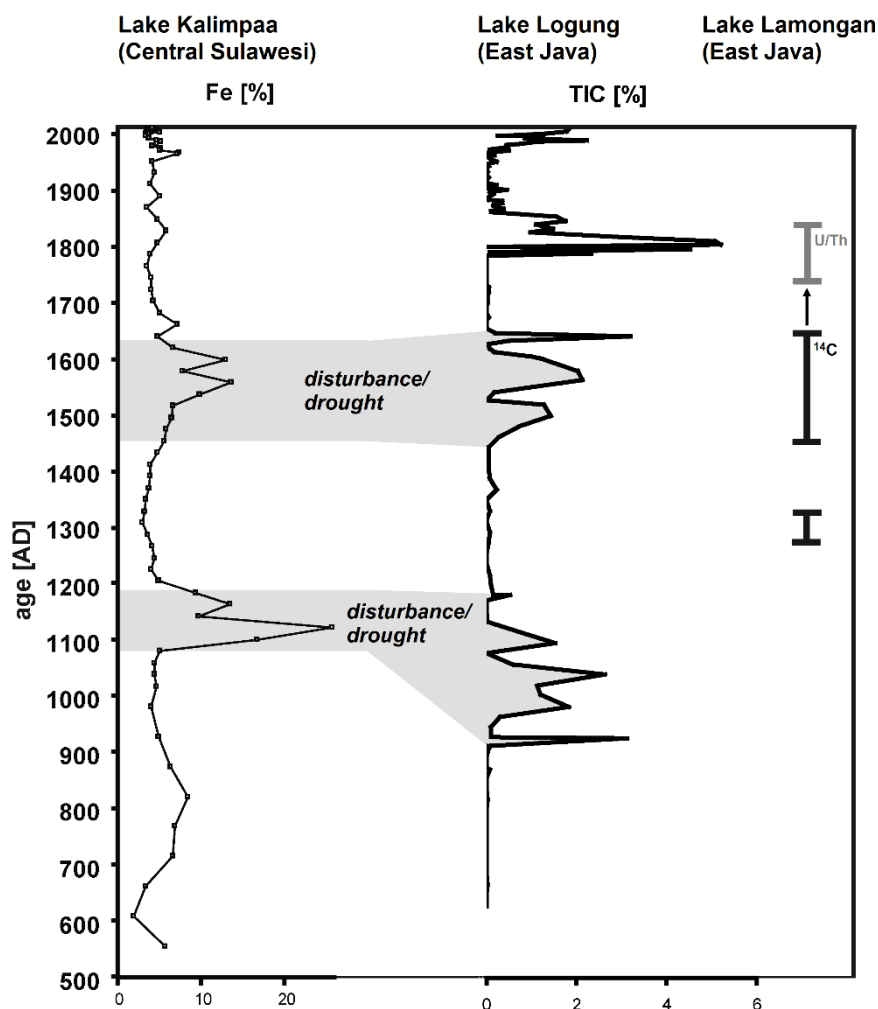
Figure 5.7 Comparison of the Fe data from Lake Kalimpaa indicative for disturbance events with total inorganic carbon (TIC) data representing periods of drought at Lake Logung, East Java (Rodysill et al. 2012). The black bars on the right represent periods of drought (based on radiocarbon dating) observed from Lake Lamongan, also East Java (Crausbay et al. 2006). The grey bar on the right represents the time period of the more recent drought at Lake Lamongan on the basis of U-series dating (Rodysill et al. 2013). The greyish bars (representing the core units II and IV of the Kalimpaa record which were deposited when disturbance events occurred in the catchment) are linked to periods of drought obtained from the Lake Logung record

at Lake Lamongan (ca. AD 1450-1650; Crausbay et al. 2006), but not with the same drought period applying U-series dating (Rodysill et al. 2013; Figure 5.7).

These similarities indicate the occurrence of two drought periods that found their expression at least in Central Sulawesi and East Java during this time. Thus, it seems likely that the long-term trend towards higher rainfall intensities and possibly higher mean rainfall was superimposed by individual, inter-annual to multidecadal-scaled drought periods which were likely associated to intense ENSO warm events. The drought indicated from ca. AD 1450 to 1620 occurred during a period when the ITCZ was displaced to the south (Sachs et al. 2009) and wetter conditions were prevailing at Lake Kalimpaa on the centennial to millennial time scale. This may indicate that the movement of the ITCZ is not the main trigger for inter-annual to multidecadal drought occurrences in Central Sulawesi, but rather the variability of the Pacific Walker circulation and hence, ENSO.

Conclusions

The Lake Kalimpaa record is one of only a few terrestrial archives from Sulawesi providing information on palaeoenvironmental and palaeorainfall changes as well as fire history throughout the past ca. 1500 years. The two main conclusions of this study are: a) in Central Sulawesi a long-term trend towards wetter conditions (higher rainfall intensities and/or mean rainfall) probably occurred on the centennial to millennial time scale starting from ca. AD 560 to the 20th century with highest rainfall intensities during the LIA; b) two disturbance events (ca. AD 1090-1190 and ca. AD 1450-1620) caused by forest fires occurred in the catchment area of Lake Kalimpaa. A comparison with other records exhibits that the long-term trend towards wetter conditions is associated with migrations of the ITCZ and the millennial-scale variability of ENSO/Pacific Walker circulation. The disturbance events around Lake Kalimpaa are probably related to regional droughts affecting at least East Java and Sulawesi. The occurrence of droughts in Indonesia which can be accompanied by forest fires, are mostly caused by the failure of the monsoon during ENSO warm events (El Niño years). While a major human impact in the Kalimpaa catchment cannot be excluded completely before the 20th century, this seems rather unlikely considering the regional correlation of drought periods at these times.

Acknowledgements

CHAPTER 6 - paper 2

Assessing resilience/sensitivity of tropical mountain rainforests towards climate variability of the last 1500 years: the long-term perspective at Lake Kalimpaa (Sulawesi, Indonesia)

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Abstract

The tropical montane rainforests of the LLNP in Sulawesi, Indonesia provide many ecosystem services for the population inhabiting the area and harbour unique biodiversity in a key area for phytogeography. The mountain regions of Central Sulawesi experience perhumid climate conditions with few seasonal changes in precipitation, making the vegetation a possible sensitive target for future changes of precipitation patterns. The ecological consequences are hard to predict due to the lack of knowledge of the dynamical processes that govern these tropical forests. This research aims to shed light on the long-term response of the montane vegetation of LLNP to stress caused by climate variability and human activities in the past. Palynological data are used to reconstruct forest vegetation dynamics and are compared to centennial time scale data of fire frequencies, palaeorainfall proxies and regional climate reconstructions to assess the drivers of these changes. Results reveal that the Fagaceae family dominates the entire recorded period, as they still do today. Fire episodes occurred locally only ten times in the last 1500 years but two periods were characterized by higher frequencies: between ca. AD 1070 and 1200 and between ca. AD 1450 and 1660. The regional correlation of these events with periods of drought registered in Java suggests that centennial-scale increases in fire frequencies at Lake Kalimpaa were consequences of the vegetation being more prone to fire, probably due to more frequent or more intense El Niño events. In both cases Fagaceae did not decrease, indicating resilience towards droughts and fires of at least one species of that family. Following the first period of increased fire frequencies, the vegetation went through a long secondary forest phase lasting about two and a half centuries (ca. AD 1200-1450). *Weinmannia* was co-dominant together with *Lithocarpus/Castanopsis*. The second period of increased fire frequencies corresponds to a phase when records across the tropics show that the Intertropical Convergence Zone (ITCZ) was displaced to the south. High effective rainfall enhanced the growing of swamp taxa like *Pandanus* around the lake. Human-landscape interactions are evident only starting from the 20th century (from ca. 1950 to present) with *Weinmannia* rising probably due to the logging of emergent *Agathis* trees and/or landslides caused by the construction of the road which today passes near the lake. In general, palynological diversity values indicate that within-landscape diversity (Whittaker's gamma diversity) decreased when fires increased. Palynological rate of change and compositional turnover indicate the vegetation communities were more resilient to fire disturbance during period of high rainfall. A different trend is apparent starting from the second half of the 20th century, suggesting a change in the dynamical response of the vegetation communities to forest fires, possibly as a consequence of increasing human activities around the lake. The emergent tree *Agathis,* while being more responsive to long-term rainfall variability in the past, did not reestablish itself following the years of intensive selective logging in the second part of the last century. These findings improve our knowledge of the long-term ecology of Central Sulawesi, one of the world's hotspots of biodiversity.

Introduction

At the dawn of the 19th century Indonesia was one of the richest tropical countries in the world in terms of forest cover and biodiversity (Baas et al. 1990). However, forest cover decreased from estimated 170 million ha around 1900 to 98 million ha by the end of the 20th century. The deforestation process is still accelerating today, due to extensive logging and conversion of forest to agricultural land (Miettinen et al. 2011). Aware of the consequences of these rapid changes, national and international efforts to develop forestry conservation plans started from the 1980s (Wardojo and Masripatin 2002; e.g. UNESCO Man and Biosphere Programme and Biodiversity Action Plan for Indonesia). These efforts include strategies aimed at protecting biodiversity and restoring ecosystem services by monitoring natural forest ecosystems in Indonesia. The Lore Lindu Biosphere Reserve and National Park (LLNP) in Central Sulawesi is part of the UNESCO Man and Biosphere Programme (MAB) and represents one example of these national and international management measures aimed to protect both natural ecosystems and cultural heritage. The protected area of 217982 ha comprises one of the largest continuous montane rainforests of Sulawesi and shows unique biodiversity (Cannon et al. 2007) representative of Sulawesi's key position for biogeographical questions in SE Asia and Malesia (e.g. van Balgooy 1987; van Welzen et al. 2011). Studies of tree families along altitudinal transects have highlighted the importance of Fagaceae mainly with the genera *Lithocarpus* and *Castanopsis* (Culmsee et al. 2010). However, developing long-term forest ecosystem management plans is not an easy task. Apart from encroachment on the forest by population pressure and largely uncontrolled extraction of forest resources, an additional potential stress factor to consider is climate change.

The main variable of Indonesia's climate is not temperature, which is rather constant throughout the year, but rainfall (Aldrian et al. 2004). The inter-annual variability of rainfall is influenced by the coupled ocean-atmosphere phenomenon El Niño-Southern Oscillation (ENSO): during El Niño (La Niña) warm (cold) phases, Indonesia experiences lower (higher) rainfall than in other years (Philander 1990; Cane 2005). Predictions of changes in ENSO variability are uncertain, as it is not clear how this phenomenon is related to climate forcing factors and to what extent the increasing anthropogenic greenhouse gas emission will influence ENSO variability. Climate models predict a wide range of responses of ENSO from weaker to stronger, from more El Niño-like to more La Niña-like average conditions (e.g. Collins 2005; van Oldenborgh et al. 2005; Guilyardi 2006; Merryfield 2006). The tropical regions of SE Asia could experience a shift in precipitation patterns leading to more frequent and/or more severe droughts in the future (Christensen et al. 2007; Sheffield and Wood 2008). The ecological consequences of these scenarios for forests of the LLNP are hard to predict. Palaeoecological and palaeoenvironmental studies in Sulawesi are still rare, but they are important as they show vegetation, fire and climate history and add to the understanding of vegetation response to climate variability and human disturbance. In conservation science, palaeoecological data are invaluable for making well-founded predictions on how

the biological component of ecosystems may respond to future perturbations such as climatic changes (Willis et al. 2010; Cole 2012). The last 2000 years are of particular interest as they include marked regional to global scale climate variations that can be reconstructed at decennial to centennial resolution. However, the few available studies in Sulawesi (Gremmen 1990; Dam et al. 2001; Hope 2001) discuss long-lasting variations in the climate regime and have a relatively low temporal resolution for the last few thousand years. The aim of this study is to evaluate the response of the montane rainforests of LLNP to past stress caused by climate variability and human activities at the decennial to centennial time scale for the last 1500 years. We present results of palynological analysis of a sediment core taken from Lake Kalimpaa, located in the center of the Fagaceae dominated montane rainforest of LLNP.

Study area

The study area is located in the northern part of the LLNP in Central Sulawesi (Figure 6.1). Sulawesi is the biggest island of the Wallacea biogeographical region, a relatively young, geologically highly complex island world (Hall 2009). The crossroad position between the Sunda shelf (part of East Asia) to the west and the Sahul shelf (part of Australasia) to the east and the biogeographic complexity makes it a global biodiversity hotspot for plants and animals, with high levels of endemism (Myers et al. 2000).
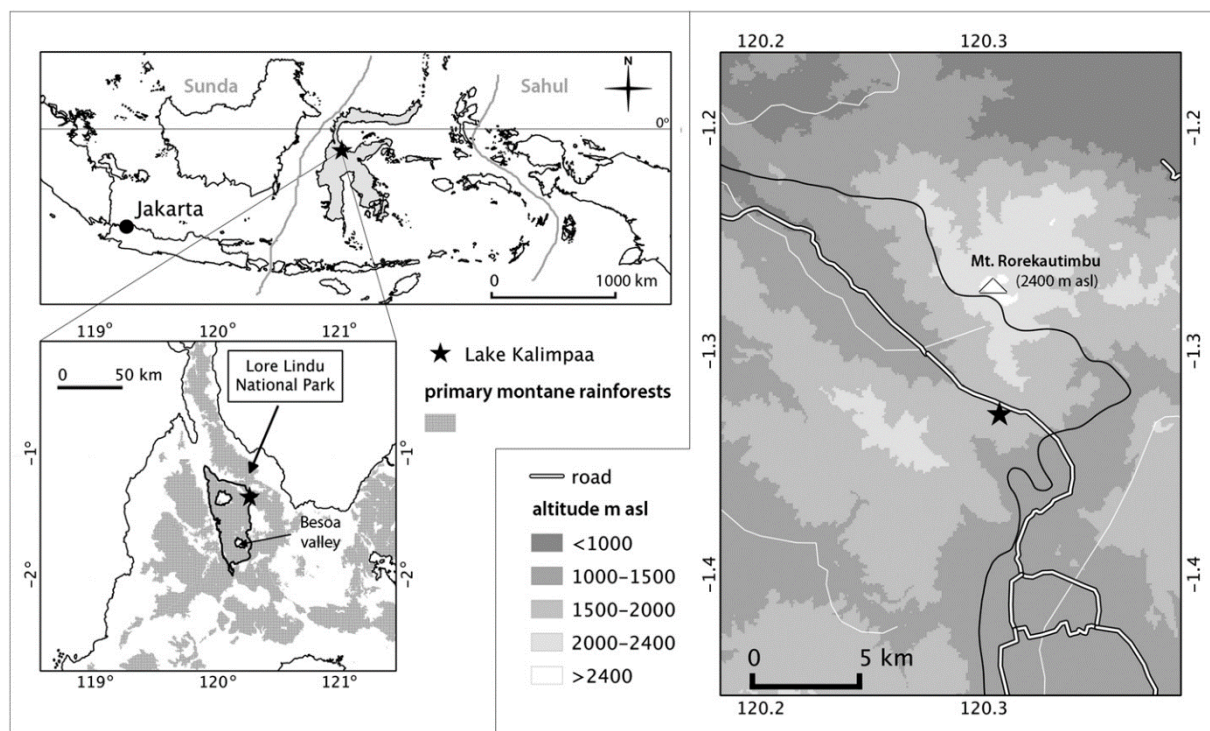
Figure 6.1 Map of the study region with the black star marking the location of Lake Kalimpaa. Upper left: location of the site with grey lines marking the separation of Sunda, Wallacea and Sahul biogeographic regions. Bottom left: location of Lore Lindu National Park in Central Sulawesi and primary upland rainforests cover (data source: Land Cover 2001, the Ministry of Forestry, Republic of Indonesia). Right: detailed altitudinal map of the location of Lake Kalimpaa showing the border of Lore Lindu National Park (black line) and the road which passes near the lake

Lake Kalimpaa (1°19'35''S, 120°18'32''E, also known as Lake Tambing) lies at 1660 m asl and was chosen because of its small catchment and its remote position far from the valleys in the north and south of LLNP which are more strongly impacted by human activities. The lake has an area of ca. 6.5 ha and a maximum water depth of 6.6 m and is therefore likely to record local vegetation changes in the past. About 200 m northeast of the shoreline the main asphalt road through the National Park passes and separates the lake area from the steeper mountainous area leading up to the peak of Mt. Rorekautimbu (ca. 2400 m asl). A small inflow reaches the lake mainly through a small swamp forest area in the northeast, dominated by *Pandanus* and palm species, and a small outflow is located in the southwest.

Modern vegetation

The rainforests of the LLNP are species-rich tropical forests. Most of the forests are still in good or old-growth condition and are situated in mountain areas (Cannon et al. 2007). The vegetation gradient ranges from upper montane rainforest, above 2000 m asl, dominated by conifers and Myrtaceae to lower montane rainforest between 1000 and 2000 m asl. Tree transect studies along an altitudinal gradient in the LLNP revealed that Fagaceae, mainly represented by the genera *Lithocarpus* and *Castanopsis,* are important at all elevations but are dominant in the lower montane rainforest between 1200 and 1800 m asl, followed by Myrtaceae, Theaceae, Symplocaceae, Magnoliaceae, Melastomataceae and Juglandaceae (Culmsee et al. 2011).

Around Lake Kalimpaa, *Castanopsis acuminatissima* (Fagaceae) was found to be the dominant species, followed by *Bischofia javanica* (Phyllanthaceae), *Calophyllum soulattri* (Clusiaceae), *Castanopsis argentea* (Fagaceae), *Prunus arborea* (Rosaceae) and *Ficus* sp. (Moraceae) (Febriliani et al. 2013).

Fagaceae play a key role in these forests in terms of aboveground biomass, which maintains a steady value despite altitudinal changes in forest structure and composition as opposed to the decline of biomass with increasing elevation usual for non-Fagaceous montane forests. Therefore, the Fagaceae dominated forest of LLNP is of particular interest in light of climate mitigation initiatives, which aim to reduce CO_2 emissions via forest preservation and restoration (Culmsee et al. 2010). The modern distribution of *Lithocarpus* and *Castanopsis* (Soepadmo 1971) indicates that these two tropical genera avoid seasonal climates. Their actual presence in Central Sulawesi seems to represent no exception.

Climate

The modern climate of Indonesia is controlled by the seasonal migration of the ITCZ across the equator and inter-annual changes in ENSO (Figure 6.2). As the ITCZ migrates southward during the austral summer, the northwest monsoon delivers humid air and heavy rainfall to Indonesia, whereas during austral winter the southeast monsoon brings relatively cool, dry conditions while the ITCZ is positioned over mainland Asia (Gunawan 2006). In the montane areas of Central Sulawesi the rainfall is strongly

determined by the local topography. The air masses reaching the area from the northwest and southeast are lifted orographically, leading to the formation of clouds and rainfall throughout the year. However, the monthly amount of rainfall formed during the southeast monsoon is slightly less than that of the northwest monsoon, as the former brings humid air masses, while the latter brings dry air from the Australian continent and the rainfall is therefore purely orographic. The modern intra-annual climate of the montane areas of the LLNP can be described as perhumid with at most two months of slightly lower precipitation, corresponding to the southeast monsoon peak in August (Gunawan 2006). Mean annual precipitation ranges between 1800 and 2100 mm, whereas mean annual temperature decreases with elevation from 21°C at 1000 m asl to 14°C at 2400 m asl (Hijmans et al. 2005; WorldClim 2006; Culmsee et al. 2010).
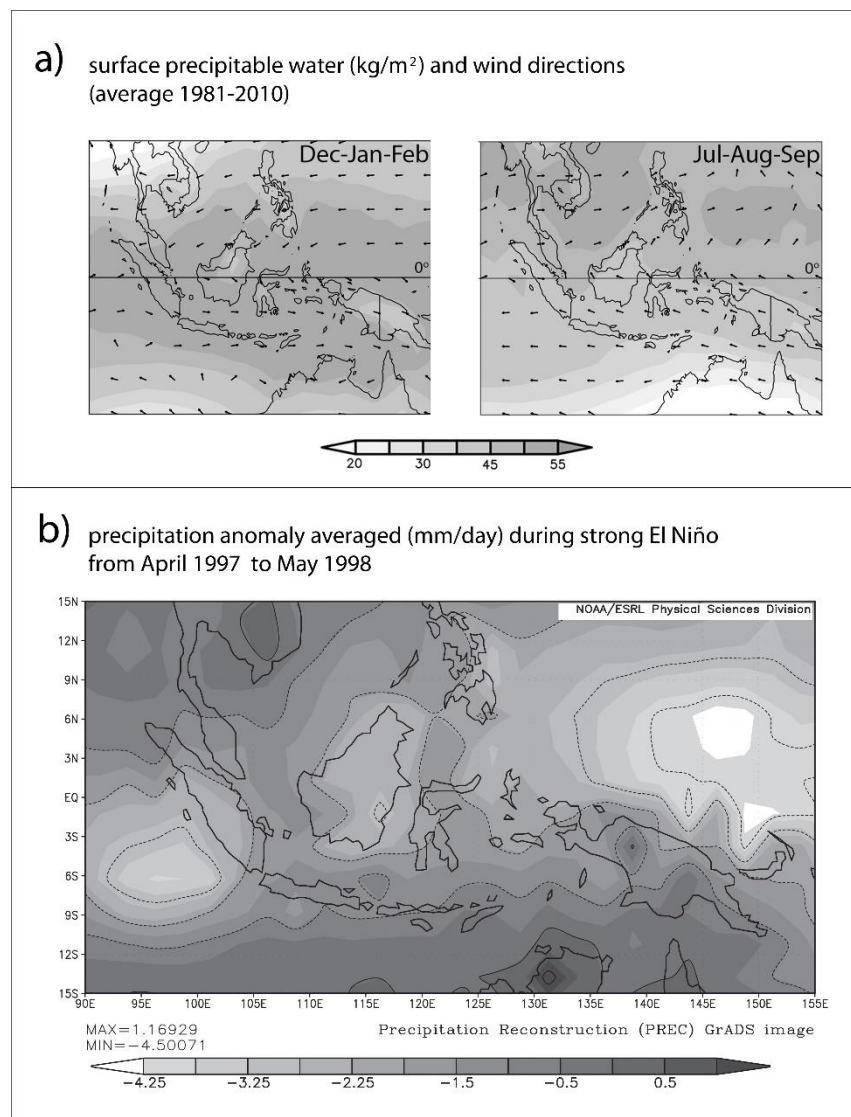
Figure 6.2 Contour maps of NCEP/NCAR reanalysis datasets. a) Composite means of surface precipitable water in kg/m^2 with arrows denoting wind directions for the period 1981–2010 for wet season (Dec–Jan–Feb, left) and dry season (Jul–Aug–Sep, right); b) composite means of precipitation anomalies in mm/day during the strong El Niño of 1997–1998

Human-landscape interactions

Few human activities were attested in the area of Lake Kalimpaa before the nearby road was constructed from 1970 to 1980. The following decades have seen the intensification of human disturbance of the forest, which was subjected to logging of *Agathis* by the company Kebun Sari PT (Adiwibowo 2005 p.113 and references therein).

Prehistoric evidence of human-landscape interactions in Central Sulawesi are provided by combined palynological and archeological investigations in the Besoa valley around 40 km from the site of Lake Kalimpaa. The oldest indication of human impact might derive from the pollen record from the northeastern part of Besoa valley, which shows that an abrupt change of the vegetation composition occurred around 2000 years ago. Fagaceae-dominated montane rainforests were replaced by grassland vegetation and fires increased (Kirleis et al. 2011; Kirleis et al. 2012).

Material and methods

A 211 cm long sediment core was recovered from Lake Kalimpaa in 2006 using a Livingstone piston corer (Livingstone 1955). At the Department of Palynology and Climate Dynamics, University of Göttingen, Germany the core was split lengthwise, photographed and described lithologically and then stored in darkness at 4°C. Due to the complexity of the age-depth modeling of Lake Kalimpaa sediment core, the chronology here is adapted from Haberzettl et al. (2013, Figure 5.2a) where further details can be found. Several sets of bulk samples which had been sent for radiocarbon dating were not in stratigraphic order. A hard water-effect was excluded for Lake Kalimpaa due to the negligible sedimentary calcite contents (Wündsch et al. 2014). A mixture of allochthonous and autochthonous organic matter was considered as possible cause for the dating uncertainties. As a consequence, the older ages were considered as too old and only the youngest ages were used for the age-depth model, which is based on linear interpolation. The age-depth model was constructed for the first 155 cm, the lowermost reliable date being at 151 cm sediment depth (Haberzettl et al. 2013). This age-depth model was then corroborated by magnetostratigraphy. The palaeosecular variation of the Kalimpaa core was compared with the output of the CALS3k.4 model (Korte and Constable 2011). Both inclination and declination show isochronic similarities (Figure 5.2b), validating the age-depth model. Therefore, the upper 155 cm of the core was used for multi-proxy palynological and palaeoenvironmental analyses. The reconstruction of the palaeoclimatic conditions based on granulometric and geochemical analyses and the local fire history using macro-charcoal analysis were done on the same sediment core by Wündsch et al. (2014), see Chapter 5 for details.

Palynological analysis

In total 40 sediment subsamples (0.5 cm³) were taken at 4 cm intervals along the core for analyzing pollen, spores and micro-charcoal particles. The samples were prepared using standard methods (Faegri et al. 1989) including 70% HF treatment. Before sample processing, the marker *Lycopodium clavatum* was added to the samples for the calculation of the concentration. Pollen and spore identification is based on the reference collection of tropical pollen and spores at the Department of Palynology and Climate Dynamics at the University of Göttingen, pollen keys and atlases for SE Asia (Flenley 1967; Powell 1970; Huang 1972; Garrett-Jones 1979; Stevenson 2000) as well as the online Australasian Pollen and Spore Atlas (APSA) hosted at Palaeoworks, Australian National University, Canberra (http://apsa.anu.edu.au). Identified pollen grains were counted to a sum of 300, excluding aquatic pollen. Percentages were calculated relative to the pollen sum. *Castanopsis* and *Lithocarpus* were counted together because pollen grains from these two genera are not distinguishable morphologically. The concentrations of the Pteridophyta spores were calculated as well in order to isolate the local climatic signal of this group. Concentrations are expressed in the diagrams as counts per cm^3 of sediment. Pollen taxa are grouped in lower montane, upper montane and herbaceous taxa according to their altitudinal distribution and ecology (Keßler et al. 2002; Culmsee et al. 2010; Flora Malesiana collection: http://floramalesiana.org; Prosea collection: http://prosea.nl). For plotting and calculations the software C2 was used (Juggins 2007).

Micro-charcoal analysis

Micro-charcoal particles were counted on the same 40 slides analyzed for pollen and spores. Only micro-charcoal particles that were found under a normal light microscope as black and completely opaque with sharp edges have been considered (10-150 μm). Following Finsinger and Tinner (2005), 200 items (total of micro-charcoal particles and *Lycopodium clavatum* spores) were counted and the concentration per unit of volume was calculated. The 95% confidence intervals of the micro-charcoal concentrations were calculated using Mosimann's equation (Maher 1981) to evaluate the errors derived from the standard deviation of the number of markers in *Lycopodium* tablets and volumetric sampling, and thus isolate samples with significant variation in the amount of micro-charcoal.

Numerical analysis

Clustering

Local pollen assemblage zones are defined numerically according to the dissimilarity matrix of Euclidian distances of squared root transformed percentage data, via constrained cluster analysis using the software CONISS (Grimm 1987; Grimm 1993). All pollen and spore taxa were included in the analysis.

Palynological diversity index (PDI)

From the palynological results palynological diversity indices were estimated via 'rarefaction analysis' (Siegel 1986) which is the calculation of the expected number of taxa $E_{(Tn)}$, in a random sample of n individuals from a smaller population of N individuals containing T taxa (Birks and Line 1992). The small size of Lake Kalimpaa indicates that it collects a large portion of pollen from the near surroundings and the PDI is probably close to Whittaker's gamma diversity/within-landscape diversity or at least within pollen-source diversity (Whittaker 1977; Odgaard 2007). Rarefaction was calculated with the software PAST (Hammer et al. 2001).

Palynological compositional change (PCC) or compositional turnover

Detrended canonical correspondence analysis (DCCA) was used to determine the amount of palynological compositional change or compositional turnover using age as the external constraint (Birks 2007; Birks and Birks 2008). Results were scaled in standard deviation (SD) units of compositional turnover (= beta diversity or diversity between habitats or communities within a landscape: Hill and Gauch 1980). Changes in palynological composition were estimated by looking at the range of sample scores on the first, time-constrained DCCA axis. Larger variation in the sample scores within the sequence implies greater compositional turnover between time-constrained samples. All pollen and spores were included in the analysis. Percentages were transformed to square roots to stabilize variances, rare taxa were not down-weighted, detrending was by segments and non-linear scaling was applied. The DCCA was implemented using CANOCO 5 (ter Braak and Smilauer 2002).

Rate of Change (ROC) or velocity

A linear interpolation of pollen and spore percentages was done to provide a time series of equal time intervals of 10 years. Time steps were then computed by dividing the chord distance of adjacent subsamples by the time interval in calibrated years (AD). The analysis used the distances computed from square root transformed pollen and spore percentages data. A LOWESS smoother (span=0.3) has been fitted to the plot to highlight the major trends in the estimates. The analysis was performed by MULTIV software (http://ecoqua.ecologia.ufrgs.br/ecoqua/MULTIV.html).

Results

Stratigraphy of the Lake Kalimpaa core

The Lake Kalimpaa profile consists of laminated silt-clay sediments with organic matter. The laminae are irregularly distributed and they are characterized by alternating darker and lighter bands. Between 153 cm and 128 cm (ca. AD 560–1090) the sediments are finely laminated blackish grey and light grey silts. From 128 cm to 118 cm (ca. AD 1090–1190) sediments are homogeneous black composed of fine silts

to fine sandy coarse silts. From 118 cm to 93 cm (ca. AD 1190–1450) the sediments are finely laminated, with light greyish to grey layers consisting of medium to fine sandy coarse silts. Partially laminated, dark brown to black fine sandy coarse silts occur from 93 cm to 77 cm (ca. AD 1450-1620). The sediments from 77 cm to 48 cm (ca. AD 1620–1910) exhibit a greyish color and contain partially laminated, fine sandy coarse silts with an intercalated sand layer. The top of the core consists of partially laminated brownish medium to fine sandy silts (Wündsch et al. 2014).

Chronology

According to the age-depth model (Haberzettl et al. 2013; Figure 5.2a), the first 155 cm of the Lake Kalimpaa core represents the past 1500 years. A marked increase in sedimentation rate characterizes the most recent sediments. While the mean sedimentation rate between 151 and 44 cm is about 0.8 mm year^{-1}, for the youngest sediments from 44 cm to the top of the core, the age-depth model provides a sedimentation rate of ca. 9.2 mm year^{-1} on average. A single pollen grain of *Pinus* was identified at 30 cm depth. *Pinus* was only recently introduced in Central Sulawesi during a restoring campaign which started in 1960-70 in three different locations near the study site (Tamandue, east Napu valley; Bolapapu-Boya and Nano, Kulawi valley). Therefore, the presence of *Pinus* confirms that the first 30 cm of the Lake Kalimpaa core records the past 30-40 years.

Palynological and micro-charcoal results

The pollen diagram shows the most important pollen and spore taxa (Figure 6.3) out of the 121 different types which were identified 122 rare types are still unknown. Based on the constrained cluster analysis the diagram is divided into 7 pollen and spore assemblage zones from Kal-I to Kal-VII. Zone characteristics are summarized in Table 6.1. Throughout the whole record the pollen assemblage reflects the lower montane rainforest tree taxa, represented mainly by *Lithocarpus/Castanopsis* (21-43%)*, Weinmannia* (1-32%), Euphorbiaceae (1-14%), *Clethra* (up to 11%), Elaeocarpaceae (up to 4%) and Myrtaceae (up to 4%). Swamp taxa are also well represented by *Pandanus* (up to 14%). Fern spores are highly diverse (66 morphological types) and abundant (up to 18%). Only few taxa are restricted to the upper montane vegetation: *Phyllocladus*, Ericaceae, Chlorantaceae and Myricaceae (Culmsee et al. 2011). Pollen concentration ranges between ca. 14x10^3 and 273x10^3 grains cm^{-3} (mean 97x10^3 grains cm^{-3}) with the exception of an unusual high value at 43 cm sediment depth (680x10^3 grains cm^{-3}). Micro-charcoal particles show constantly high values along the studied core but only one peak at 143 cm (ca. AD 820) is statistically significant.

Table 6.1 Zone characteristics, pollen and spores percentages, PDI, PCC, ROC, charcoal and fire frequencies of the Lake Kalimpaa core (* macro-charcoal data from Wündsch et al. 2014)

Zone	Core depth (cm)	Age (AD)	Sedimentation rate (mm yr^{-1})	Pollen zone characteristics	Palynological Diversity Index (PDI), Palynological Compositional Change (PCC) and Rate of Change (ROC)	Charcoal results and fire frequencies *
Kal-I	155-128	ca. 550-1070	from 0.4 - 1	Relatively high proportion of herbaceous taxa especially *Gunnera* (up to 2%) and Poaceae (up to 2%); low values of *Agathis* (up to 1%); total concentrations of Pteridophyta the lowest of the entire record (<5614 spores cm^{-3}), in particular tree ferns (<4%)	High values of PDI and increasing trend from 62 to 75 pollen and spore types at the end of the zone; maximum PDI value at 139 cm depth (ca. AD 930); PCC constant; increasing trend of ROC corresponding to PDI increases	One significant peak in the curve of micro-charcoal at 143 cm (ca. AD 820); macro-charcoal particles absent except at 144 (ca. AD 794) and 140 (ca. AD 900) cm; increase of fire frequency at the end of the zone (from 0 to 1.93 fires 200 yr^{-1}); low values of peak magnitude (<3 pieces cm^{-2} per peak); one local fire episode at 140 cm depth (ca. AD 880)
Kal-II	128-121	ca. 1070-1170	1	Increasing proportion of *Agathis* (up to 2%); decrease of *Gunnera* and Poaceae; high values of *Clethra* characterize the upper boundary of the zone (up to 11%); increase of Pteridophyta spore concentration and rise of tree fern (up to 3%) and *Blechnum* spores (up to 2%)	Decreasing trend of PDI; minimum value at 123 cm (ca. AD 1140, 59 pollen and spore types)	Rise of fire frequencies up to 125 cm (ca. AD 1120, 2.93 fires 200 yr^{-1}). High values of peak magnitude, maximum at 128 cm (ca. AD 1090, 57 pieces cm^{-2}); three fire episodes
Kal-III	121-93	ca. 1170-1450	1	Rise of *Weinmannia* curve (up to 23%) and mirrored decrease of *Lithocarpus/Castanopsis; Agathis* less represented at the beginning of the zone	Decreasing trend of PDI continues with minimum value at 107 cm (ca. AD 1306, 52 pollen and spore types) and then increasing trend towards the end of the zone; PCC and ROC changes occur at 115 cm (ca. AD 1200)	Decrease of fire frequencies with minimum at 107 cm (ca. AD 1306, 0.78 fires 200 yr^{-1}) and then increasing trend towards the end of the zone. One single local fire event detected at 100 cm (ca. AD 1386)
Kal-IV	93-73	ca. 1450-1660	1	Increase of *Pandanus* pollen (up to 15%); frequent occurrence of pioneer, light-demanding Euphorbiaceae: *Macaranga/Mallotus* (up to 8%) and *Baccaurea/Bischoffia* (up to 3%); tree ferns decrease	High, oscillating values of PDI (between 66 and 74 pollen and spore types); minimum at 83 cm (ca. AD 1554, 56 pollen and spore types)	Increasing trend of fire frequencies continues, with maximum at 83 cm (ca. AD 1554, 3.56 fires 200 yr^{-1}); four fire episodes
Kal-V	73-45	ca. 1660-1950	1	Decreasing trend of *Pandanus;* decreased proportions of *Macaranga/Mallotus* and *Baccaurea/Bischoffia;* frequent occurrence of Lycopodiaceae (up to 9%), *Huperzia* (up to 2%), *Asplenium* (up to 1%) and tree ferns (up to 9%)	Beginning of a decreasing trend of PDI	Decrease of fire frequencies, with minimum at 38 cm (ca. AD 1826, 0.14 fires 200 yr^{-1}) and then rising trend towards the end of the zone; no fire peak passed the minimum-count screen test
Kal-VI	45-36	ca. 1950-1970	from 1 to 5.5	Rise of *Weinmannia* curve (up to 32%) and Fabaceae (up to 18%); decreased proportion of *Agathis;* spores of Pteridophyta almost absent (total sum less than 1% at 43 cm, ca. AD 1960); pollen concentrations jump to 677,247 grains cm^{-3}	Lowest PDI value at 43 cm (ca. AD 1960, 44 pollen and spore types); ROC increases	Rising trend of fire frequency continues; no fire peak passed the minimum-count screen test
Kal-VII	36-0	ca. 1970-2006	from 5.5 to 23	Decrease of *Weinmannia;* low values of *Agathis* (less than 1%); increase of *Pandanus* (up to 12%) and Moraceae/Urticaceae (up to 3%); first appearance of Lamiaceae pollen type and *Isoetes* spores	Rapid increase of PDI (between 60 and 70 pollen and spore types); greatest change in PCC and ROC	Rising trend of fire frequency continues to present; one local fire event detected at 34 cm (ca. AD 1976)

Palynological diversity, amount of compositional turnover and rate of change

Palynological diversity values change markedly along the record. The lowest PDIs are recorded at 123 cm (ca. AD 1140, 59 pollen and spore types), at 107 cm (ca. AD 1310, 52 pollen and spore types), at 83 cm (ca. AD 1550, 56 pollen and spore types) and at 43 cm (ca. AD 1960, 44 pollen and spore types). In general, periods of higher (lower) palynological diversity correspond to periods of low (high) fire frequency (Wündsch et al. 2014; Figure 6.3b). Low PDIs however characterize zones Kal-III and Kal-VI due to the dominance of pollen from *Weinmannia* and *Lithocarpus/Castanopsis* which together account for up to 54% of the total sum of pollen and spores in Kal-III and up to 58% of the total sum of pollen and spores in Kal-VI. The zone representing approximately the last 40 years, Kal-VII, is characterized by a different pattern. Here, increasing fire frequencies correspond to higher PDIs.

The length of the first time-constrained DCCA axis provides an estimate of the amount of palynological compositional change (PCC) or compositional turnover within the considered time period (Birks 2007). The total turnover estimate for the entire pollen record of Lake Kalimpaa is 0.924 SD. This gradient length indicates that samples of the sequence differ in their palynological composition by ca. 25% of the total assemblage (Hill and Gauch 1980). This suggests that important changes occurred in the community composition of the lower montane rainforests. These dynamics are not clearly visible when looking at the percentage curves of the most important taxa represented in the pollen assemblage of Lake Kalimpaa. The range of compositional turnover changes markedly at the beginning of Kal-III corresponding to the rise in *Weinmannia* proportion. Afterwards, when *Weinmannia* values decrease again, the range of compositional turnover changes slowly and linearly until Kal-VII where the greatest compositional change occurred. Since the PCC tends to emphasize changes in taxonomic composition (Anderson et al. 2006), we calculated also the rate of compositional changes or velocity (ROC) that estimates both changes in assemblage percentage values as well as changes in taxonomic composition, revealing also the changes in relative abundance (Anderson et al. 2006). As for PCC, ROC increases in zones Kal-III, and Kal-VII, suggesting that both taxonomic composition and relative abundance of taxa changed. Opposite to PCC, ROC values are high also in Kal-VI.
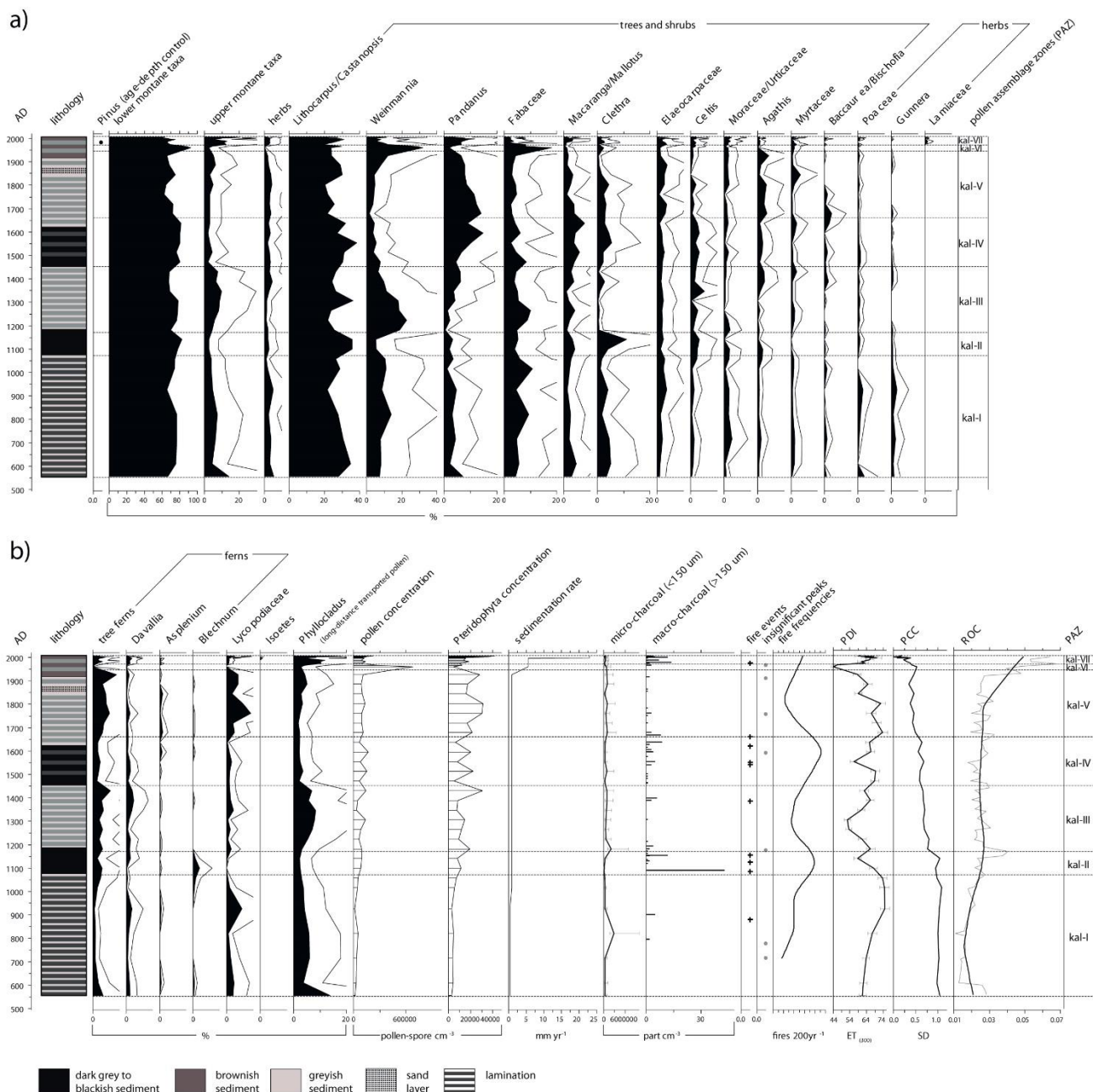
Figure 6.3 Summary diagram from Lake Kalimpaa showing the lithological table and: a) more common and significant pollen taxa (expressed in percentages of total pollen sum); b) more common spore taxa (percentages of total sum), total pollen and Pteridophyta concentrations (pollen-spore cm⁻³), micro charcoal concentrations. Macro-charcoal concentrations (particles cm⁻³), fire peaks which passed the "Poisson minimum-count" screening represented with + symbols and the grey circles representing the peaks which did not pass the screening and fire frequencies (per 200 yr) (from: Wündsch et al. 2014), palynological diversity index (PDI based on sum of 300), palynological compositional change or turnover (PCC in standard deviation units) and rate of change per 10 years (ROC, in black: 0.3 lowess smooth, in grey: raw data). Grey bars on micro-charcoal and PDI values represent 0.95 confidence intervals. Outline curves showing an exaggeration of 3× to highlight variations of the taxa

Interpretation and discussion

For the interpretation of possible drivers, the change in pollen assemblage zones is compared with palaeoclimatic and fire history reconstructions based on macro-charcoal, grain size and geochemical investigations on sediment samples of the Lake Kalimpaa core as interpreted by Wündsch et al. (2014) as well as other regional records of palaeoclimate (**Figure 6.3** and **Figure 6.4** and Chapter 5).

Palaeoclimatic evidence from continental Asia, Africa and the Americas (e.g. Haug et al. 2001; Tierney and Russell 2007; Tierney et al. 2010) suggests that the ITCZ mean position has changed during the past millennium, leading to a substantial rearrangement of rainfall patterns across the tropical and subtropical regions of the world. Superimposed onto the latitudinal movements of the ITCZ is the decade- to century-scale rainfall variability caused by changes in amplitude and frequency of the ENSO phenomenon. El Niño phases are associated with cooler sea surface temperatures (SSTs) in the West Pacific Warm Pool (WPWP) region and warmer SSTs in the central and eastern Pacific, which drive anomalous easterly winds across the equatorial Pacific and Indonesia and weaken the Walker circulation and vertical convection over the WPWP. The anomalous easterlies prevail into austral summer, delaying the arrival of the northwest monsoon causing droughts in SE Asia (Cane 2005).

The analysis of palaeorainfall and fire proxies on the Lake Kalimpaa core, revealed a trend from ca. AD 560 to the 20th century of increasing rainfall intensities associated with a southern migration of the ITCZ. Additionally, two disturbance events (ca. AD 1090-1190 and ca. AD 1450-1620) caused by forest fires occurred in the catchment area of Lake Kalimpaa. The authors interpreted these events associated with increased fire frequency around Lake Kalimpaa as probably related to regional droughts affecting at least East Java and Sulawesi (Wündsch et al. 2014; **Figure 6.4**).

Vegetation composition and fires

Pollen and spore analyses of the sediment core from Lake Kalimpaa illustrate the dynamics of the lower montane rainforest during the past 1500 years. If *Phyllocladus* pollen grains are excluded, the sum of taxa restricted to the upper montane rainforest never reaches 3% (mean value 1%). Therefore, the Lake Kalimpaa pollen record reflects local vegetation of the lower montane rainforest, with little influence from long-distance transported pollen from upper montane vegetation. The pollen assemblage reveals that Fagaceae (*Lithocarpus*/*Castanopsis*) dominated the vegetation surrounding the lake for the entire period recorded, together with Cunoniaceae, Euphorbiaceae, Clethraceae, Elaeocarpaceae, Myrtaceae, Pandanaceae and Pteridophyta. In the background of the Fagaceae-dominated pollen spectrum, changes in certain local taxa indicate a dynamic response of the forest to climate variability. The lack of correlation between micro-charcoal concentrations and local fire frequencies confirms that the former were transported from long distances, probably from lower altitudes, as macro-charcoal analysis of a

sediment core taken at Mt. Rorekautimbu (ca. 2400 m asl) indicates that the upper montane mossy forests were free of fires for at least the last 9000 years (Biagioni data unpublished).

Long-term palaeoecology and drivers of change

From ca. AD 550 to 1070 (zone Kal-I) values of tree ferns, represented by species of Cyatheaceae and Dicksoniaceae were low. In the lower montane rainforest of the LLNP most tree fern species are slow growth forest species occurring only in rather shady forest and do not tolerate strong seasonality (de Winter and Amoroso 2003). Additionally, drier conditions might be inferred by the low presence of *Agathis*, a very large emergent tree, which is confined to regions with rainfall well distributed over the year (de Laubenfels 1986). Small openings in the forest might be indicated by the frequent occurrence of herbaceous species, in particular Poaceae and *Gunnera*. Pollen-based vegetation reconstruction at Lake Kalimpaa suggests that the climate was drier from ca. AD 550 to 1070. This is confirmed by run off proxies which are related to palaeorainfall such as grain size and geochemical element ratios of the sediment of the Kalimpaa core (Wündsch et al. 2014, Figure 6.4).

The rising trend in palynological diversity index (PDI) observed from ca. AD 900 to 1070 corresponds to increasing mean average rainfall as indicated by run off proxies (Wündsch et al. 2014, Figure 6.4). This suggests that rainfall was a limiting growth factor for certain taxa (for example *Agathis* and tree ferns), reducing within-landscape diversity (gamma diversity). A gradual increase in the rate of change curve (ROC) is also observed for this period, while the range of the palynological compositional change or turnover (PCC) remained constant confirming that changes occurred within the vegetation communities and no marked changes in taxonomic composition occurred.

Natural fires started by lightning or spontaneous combustion occur in SE Asian rainforests but usually do not completely destroy the forest and merely cause the death of a few trees (Mueller-Dombois and Goldammer 1990; Flenley and Butler 2001). However, if lightning occurs during a period of frequent or longer than normal droughts the rainforests become more susceptible to fire and the frequency of fire episodes increases. The low frequency of local fire episodes from AD 550 to 1070 might indicate that the vegetation composition around Lake Kalimpaa was less prone to fire during this period and thus long periods of drought were not common despite permanent drier than average conditions. An exception is recorded at ca. AD 800-900 when regional and local fires were more common, as indicated by both micro- and macro-charcoals.

From ca. AD 1070 to 1170 (zone Kal-II) a change in the frequency of fire episodes occurred towards more frequent fires per 200 yr. Long-term disturbance can be inferred from the increase of pollen of *Clethra*, a genus of light-demanding trees which tend to remain sterile in dense forest and only flower at forest borders and in more open vegetation. The maximum of *Blechnum* might be the consequence of frequent fires as some species of *Blechnum* are colonizers after fire disturbances and can re-sprout from

trunks after fire. Increasing fire frequencies in Lake Kalimpaa suggest that longer and or more frequent droughts occurred during this century. Furthermore, grain size data of the Lake Kalimpaa sediment core show high variability of sand and clay at 136-118 cm, ca. AD 1010 to 1190, suggesting that the rainfall regime was characterized by alternating intensive rain and dry periods (Wündsch et al. 2014). The geochemical composition of the lacustrine sediments also shows variation as a consequence of disturbance around the lake. Fe and S markedly increased as a result of enhanced supply of organic
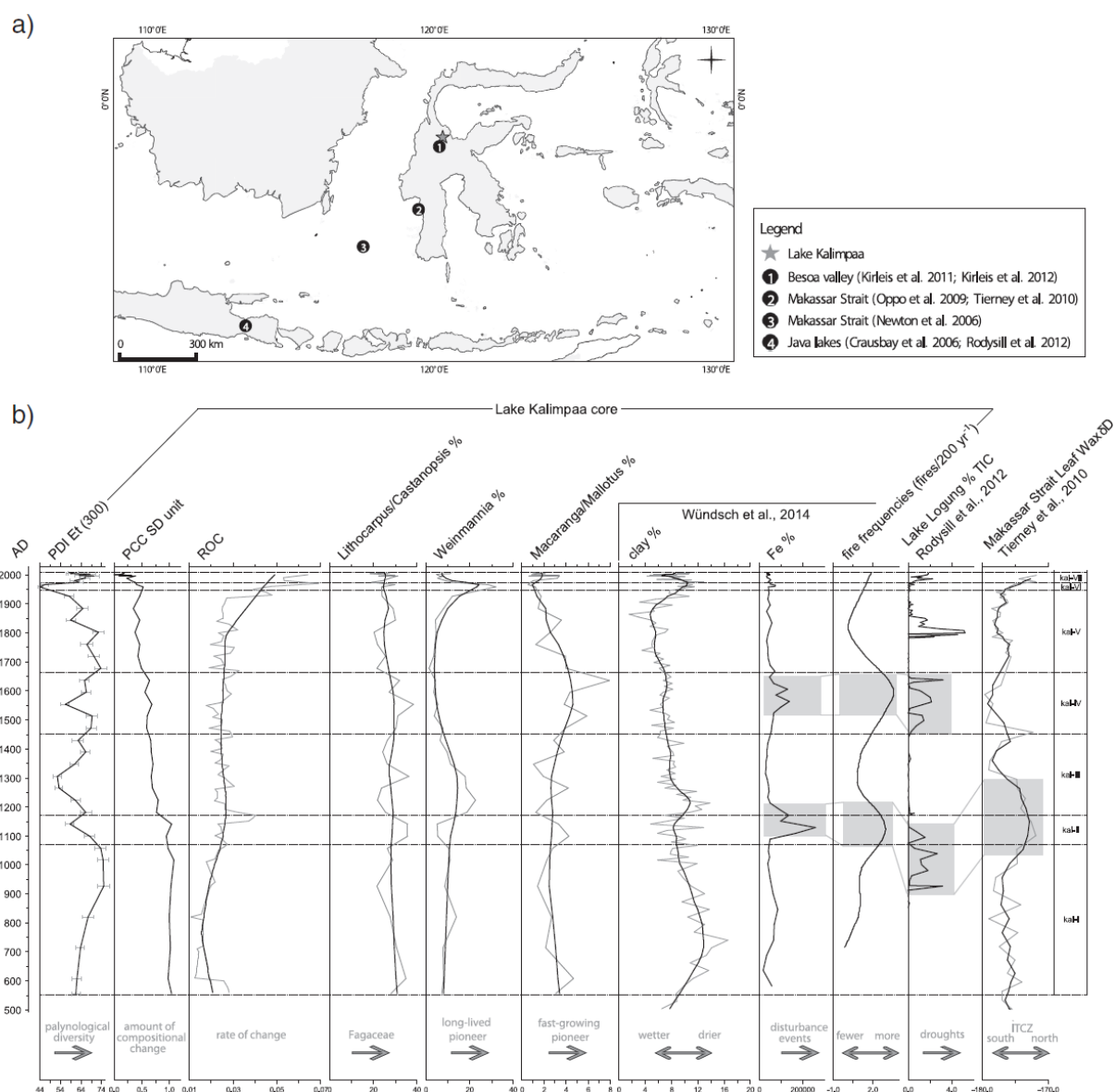
Figure 6.4 a) Locations of the sites mentioned in the text; b) comparison of palynological diversity index (PDI), palynological compositional change or turnover (PCC), rate of change (ROC), selected % pollen taxa (Fagaceae, *Weinmannia* and *Macaranga/Mallotus*)with % of clay grain size, % of Fe, number of fire episodes per 200 years of Lake Kalimpaa sediment core (Wündsch et al. 2014) and total inorganic carbon (%, TIC) from Lake Logung, Java, as drought indicator (Rodysill et al. 2012), δD value of leaf waxes from the Makassar Strait cores 31MC and 34GGC (3 point moving average of raw data) interpreted to represent rainfall amount changes as a consequence of the ITCZ mean position changes (Tierney et al. 2010). In black: 0.3 lowess smooth, in grey: raw data. Grey areas underline the correlation of disturbance events and regional droughts as discussed in the text and interpreted by Wündsch et al. (2014)

material into the lake which promotes reducing conditions in the sediment due to its microbial decomposition (Wündsch et al. 2014).

Results from a sediment core from the megalithic Pokekea complex in the Besoa Valley (Kirleis et al. 2011; Kirleis et al. 2012), about 40 km south of Lake Kalimpaa, show that the sediment composition changed from lacustrine to peaty between ca. AD 710 and 1280 and the pollen concentrations decrease due to hydrological variations. This indicates an aridification of the floodplain area and may have been caused by either human impact or drought (Kirleis et al. 2011). Similar to Lake Kalimpaa, leaf wax δD data from Makassar Strait (Tierney et al. 2010) and geochemical data from East Java (Rodysill et al. 2012), record droughts during this period. The regional correlation suggests that the increased number of fire episodes at Lake Kalimpaa were not caused by human activities but more likely were consequences of natural fires resulting from more frequent and/or longer droughts. Droughts in Indonesia result from the failure of the monsoon, which often coincides with El Niño warm events (Cane 2005).

The change in fire regime was accompanied by decreasing PDIs. However, *Lithocarpus/Castanopsis* did not decrease during this period, indicating that at least one species of Fagaceae was not affected by the increase in fire episodes.

During the following three centuries, from ca. AD 1170 to 1450 (Kal-III), the greatest change in composition occurred as indicated by PCC. At the beginning of the zone (from ca. AD 1170 to 1300) fire frequencies were very low. However, drier climate conditions can be inferred from the relatively low abundance of *Agathis* and high (low) values of clay (sand) (Wündsch et al. 2014). This might be related to a more northerly position of the ITCZ and the corresponding weakening of the northwest monsoon over Indonesia (Tierney et al. 2010). This is consistent with records from SE Asia and mainland Asia which suggest that the ITCZ reached its northernmost extent between AD 900 and 1100, shifted southward and reached its southernmost extent between AD 1420 and 1640, and then began to shift northward again between ca. AD 1700 and 1800 (Wang et al. 2005; Zhang et al. 2008; Sachs et al. 2009). A more northerly position of the ITCZ would have shifted the tropical rain belt away from Indonesia on a mean annual basis and could have resulted in a longer dry season in the montane areas of Central Sulawesi compared to today's annual conditions. In contrast, a southward migration of the ITCZ would have strengthened the northwest monsoon, causing wetter conditions with rainfall more evenly distributed over the year.

At the same time, PDIs decreased and ROC increased as a consequence of the rise of the *Weinmannia* proportion. The co-dominance of *Weinmannia* trees in the lower montane rainforests surrounding Lake Kalimpaa might represent a secondary forest phase following the previous hundred years of more intense fire disturbance and the onset of drier/more seasonal conditions. Pioneer species like *Macaranga/Mallotus* are little represented in this period, possibly due to their faster replacement in

the forest. *Weinmannia* species can act as pioneers but also grow into tall canopy trees (unlike *Macaranga* and *Mallotus*). Therefore, they are more adapted in terms of competition for space with other canopy trees and can be considered late succession secondary forest trees or 'long-lived pioneer' trees (Lusk 1999). Little information is available about this genus in the montane vegetation of the LLNP. However, Febriliani et al. (2013) reported *Weinmannia descombesiana* as one of the common species found as seedling in the understory around Lake Kalimpaa and *Weinmannia descombesiana* has been observed growing along roads and disturbed patches of forest (Brambach, personal communication). Starting from ca. AD 1300 *Agathis* increased, pointing to increasing average rainfall consistent with the southward migration of the ITCZ, while *Weinmannia* slowly decreased.

From ca. AD 1450 to 1660 (Kal-IV) a major change in fire regime towards more frequent fire episodes occurred. Increasing fire frequencies suggest longer or more frequent periods of drought. Similarly, multidecadal- to centennial-scale droughts are recorded in Lake Logung (Rodysill et al. 2012, Figure 6.4) and Lake Lamongan (Crausbay et al. 2006) in Java. In contrast, the increasing presence of *Pandanus*, a small tree nowadays locally abundant in swamp depressions around the lake, indicates that, on average and in normal conditions, rainfall was high during this period. The ITCZ reached its southernmost position during this period, when archives across Indonesia record the wettest conditions for the last 2000 years. A confirmation that higher rainfall characterized the central Sulawesi area during this period comes also from the sediment of the Pokeka site core where the growth of peat indicates a hydrological change towards wetter conditions (Kirleis et al. 2011). Cooler SSTs are inferred from foraminifera from Makassar Strait marine cores (Newton et al. 2006; Oppo et al. 2009). Modeling experiments show that if there was a cooling over the entire tropics, the Pacific would change more strongly in the west than in the east, where strong upwelling holds the temperature closer to pre-existing values. Hence, the east-west temperature gradient would weaken, leading to a more El Niño-like state (Cane 2005). The persistence of a high rainfall regime on the centennial to millennial time scale, superimposed onto decadal scale periods of drought, might explain the apparently contradictory abundance of swamps and more frequent fire episodes. Disturbance of the area around the lake is also underlined again by increased Fe and S in the Lake Kalimpaa sediment core (Wündsch et al. 2014). Fast-growing, light-demanding pioneer trees like *Macaranga/Mallotus* and *Baccaurea/Bischoffia* increased while *Lithocarpus/Castanopsis* show no variation. Both PCC and ROC remained on the ranges of the previous zone, suggesting that vegetation communities did not change markedly. Diversity was high, probably due to the presence of pioneer and fire tolerant taxa. However, an exceptionally low PDI value characterized the 16th century while, at the same time, fire frequencies were supposedly very high.

The following centuries from ca. AD 1660 to 1950 (Kal-V) were characterized by an unclear mixture of climate and disturbance indicators. Initially, fire frequencies decreased and fast-growing pioneer species like *Macaranga/Mallotus* decreased. The increased presence of Lycopodiaceae and *Asplenium* might

indicate either wetter conditions or fast colonization after disturbance. However, from the second half of the 19th century a decreasing trend in PDI is apparent while at the same time fire frequencies increased and *Pandanus* decreased.

Two decades of intense disturbance around Lake Kalimpaa are clearly recorded during the last century (ca. AD 1950-1970, Kal-VI). The pollen concentration increased notably, while spores of Pteridophyta almost disappeared during that period. Spores of ferns are relatively heavy and are usually transported over short distances by water. The absence of fern spores point towards an interruption of water inflow from rivers and/or small streams into the lake, while the increase of long-distance transported pollen grains such as *Phyllocladus* indicates that the deposition of allochthonous materials increased. From the 1960s to the beginning of the 1980s the logging company Kebun Sari was active in the surrounding area. The road, which today enables access to the mountain areas surrounding Lake Kalimpaa, was constructed in this period. It is possible that the construction works changed the course of an influent stream, temporarily blocking the inflow into the lake while sediment deposition increased as a consequence of higher erosion caused by the activities of the company in the area. Kebun Sari's main target was the valuable wood of *Agathis.* The logging of these trees is reflected in the pollen diagram by the decrease of *Agathis* percentages. A second phase of high frequencies of *Weinmannia* in this period might confirm the role of at least one *Weinmannia* species as a secondary forest tree, as indicated by the relative synchronicity between the disturbance event and the rise of *Weinmannia* during this period. After the disturbance around the lake ended, the forest quickly recovered its former state, as indicated by the decrease of *Weinmannia*. The exceptionally low PDI value and high ROC might be partly artifacts due to the dominance of *Weinmannia* and the absence of most of the Pteridophyta spores from the pollen assemblage.

During the last decades, from ca. 1970 to present (Kal-VII), the record indicates that a highly diverse lower montane rainforest surrounded Lake Kalimpaa, as shown by the high values of PDI. The greatest change in PCC and ROC occurred in this period. For the first time in 1500 years, increasing fire frequencies correspond to higher diversity. This suggests a change in the dynamical response of the vegetation communities after disturbance occurred, possibly as a consequence of increasing human activities around the lake. However, changes in PDI, PCC and ROC values might be a consequence of the high sediment accumulation rate that characterized the uppermost 30 cm of the Lake Kalimpaa core (Haberzettl et al. 2013). The percentage of *Agathis* pollen did not return to the values shown before the Kebun Sari logging activities started in the second half of the 20th century.

Conclusions

Pollen, spore, and charcoal records from Lake Kalimpaa reveal the long-term response of the Fagaceae-dominated montane rainforest towards climate variability and human disturbance for the past 1500

years. The pollen results show the resilience of the Fagaceae towards the decennial to centennial scale climate variability of the past 1500 years. However, the vegetation seems to be more prone to fires during long periods of drought despite the annual ever-wet conditions and high rainfall regime. The combination of decreasing average rainfall, long-term fire and human disturbances increased the chance for at least one species of *Weinmannia* to become common in open patches of the lower montane forest of the LLNP. However, the system was more resilient to fire disturbance when average rainfall was higher.

Important human manipulation of the landscape is only visible starting from the 20th century with changes in sedimentary processes as well as vegetation composition.

While the long-term results suggest that these ecosystems have shown resilience to natural climate disturbances in the past, the more recent anthropogenic disturbances have changed the dynamical response of the vegetation communities to climatic disturbance events.

Acknowledgments

CHAPTER 7 - paper 3

Unravelling the past 1000 years of history of human–climate–landscape interactions at the Lindu plain, Sulawesi, Indonesia

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Abstract

The Lindu plain, located in the northern mountainous region of the LLNP in Sulawesi, Indonesia, provides many ecosystem services for the population inhabiting the area and harbours a unique biodiversity. Palynological, charcoal and diatom analyses of a lake sediment core from Lake Lindu (Danau Lindu) reveal that during the last 1000 years the Lindu plain has been modified by human activities. Evidence of frequent burning and possible shifting cultivation from an earlier phase from ca. AD 1000 to 1200 might be related to the metal age population which erected the megaliths in the province of Central Sulawesi. From ca. AD 1200–1700 there followed 500 years of wetter climate conditions, corresponding to the southward movement of the Intertropical Convergence Zone. At the same time, decreases of macro-charcoal concentrations and pioneer vegetation indicators show that the use of the landscape of Lindu plain had become more permanent. Following a phase of forest recovery from ca. AD 1730 to 1910, the most recent part of the Lake Lindu record shows a trend towards deforestation that started in the late 20th century, lasting until now. The lake level started to fall at the beginning of the 20th century, as shown by the increase of sedimentation rate and supported by low pollen concentration and palaeomagnetic data. Such a change was unprecedented for the last 1000 years covered by the record, and it has no link to the climate variability as reconstructed for the last hundred years. If deforestation increases and a larger amount of water is channeled away from the lake for irrigation purposes, the lake level will continue to fall. This suggests that there is a need for better management of the forests surrounding the plain and of the irrigation systems in the area open for cultivation.

Introduction

Present-day tropical montane rainforests in the area of the LLNP in the province of Central Sulawesi, Indonesia, are rich in biodiversity and an important source of ecosystem services for the local population inhabiting the area. The montane rainforests within the LLNP, a UNESCO Man and Biosphere reserve since 1977, are mostly untouched old-growth forests (Cannon et al. 2007). However, during the last three decades, a growing population and political and economic initiatives have increased the pressure on previously relatively isolated communities, leading to conflicts for land and the opening up of further forested areas for agriculture (Acciaioli 2001). Within the context of sustainable management of conservation areas and agro-landscapes, palaeoecological and palaeoenvironmental studies provide a valuable contribution by showing how the vegetation and the environment have changed as a consequence of long-term climate variability and human-landscape interactions. In order to understand

the present and future landscape dynamics, it is important to have a historical perspective when analyzing the effects of environmental changes caused by human activities. However, little is known about the prehistory and history of the LLNP.

Megaliths found in the area of the park indicate that well-organized human societies have been present there for at least the last 2000 years. Closer investigations have only been conducted in the Besoa valley in the southern part of the park (Figure 7.1), where pollen analysis has revealed that the valley has been deforested for 2000 years (Kirleis et al. 2011; Kirleis et al. 2012).
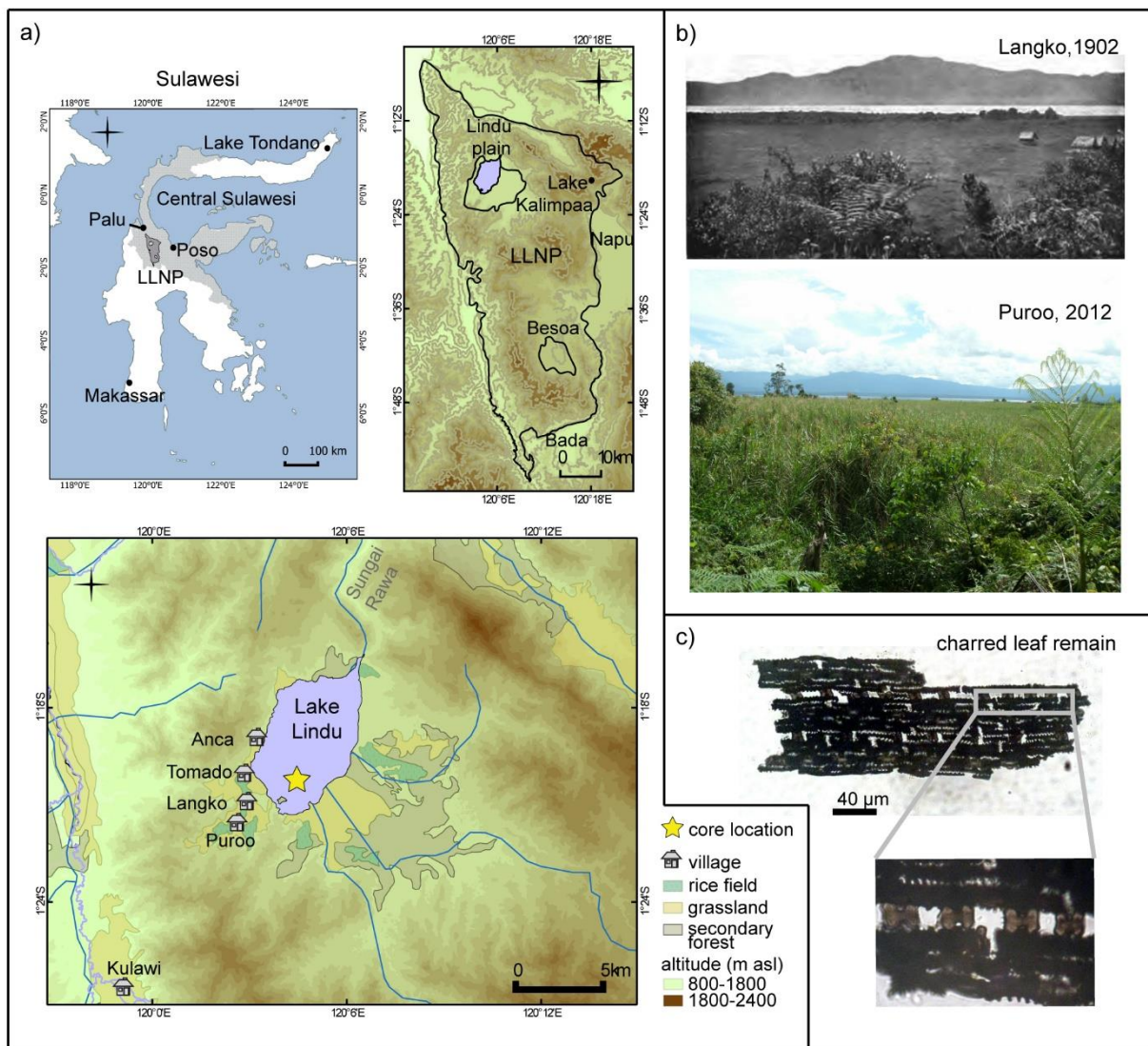
Figure 7.1 a) Map of the study region showing the location of the Lake Lindu sediment core (star). Upper left location of the Lore Lindu National Park (LLNP, dark grey) in Central Sulawesi (light grey); upper right location of the Lindu plain, Bada, Besoa and Napu valleys, borders of the LLNP and other places mentioned in the text; bottom locations of the villages of the Lindu plain and the coring site. Data source, Land Cover 2011, the Ministry of Forestry, the Republic of Indonesia (http://appgis.dephut.go.id/appgis/download.aspx); b) upper photo view of the Langko village and grassland on the southwestern shore of the Lindu lake in 1902 (from Sarasin and Sarasin 1905). Bottom photo; view of the alang-alang (*Imperata cylindrica*) grassland from the village of Puroo in 2012, photo by S. Biagioni; c) example of charred leaf remain from grass found at 120 cm (ca. AD 1100) and details of dumb-bell silica body

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Isolated from the valley to the north and south, the present-day economy of the four villages on the shore of Lake Lindu, Anca, Tomado, Langko and Puroo, is based on the trading of rice cultivated in the plains surrounding the lake and fish caught in the lake itself. Little is known of the prehistory of the Lindu plain where megaliths are also found.

We present the results of a multi-proxy palaeoenvironmental study of a 123 cm long sediment core taken from Lake Lindu (1°19'16''S, 120°04'36''E at 960 m asl.), spanning the past ca. 1000 years. Palynological, charcoal and diatom data are used to reconstruct the vegetation and fire history of the Lindu plain, as well as the history of eutrophication of the lake. The aim is to characterize the timing and intensity of human activities during prehistoric and historic times. Results from the Lindu sediment core will shed light on the history of human-landscape interactions on the Lindu plain, a well-known fish reserve, and contribute to the understanding of the complex but still poorly known history of Central Sulawesi.

Study area

The Lindu valley is located in the highlands of Central Sulawesi and has an area of ca. 1000 km^2 in the Takoekadju mountain range. In the northwest portion of the valley lies Lake Lindu, an ancient permanent freshwater lake, which is 10 km in length and 5-6 km in width (Figure 7.1a). Small streams originating from the surrounding mountains drain into the lake. At the north-eastern corner, the river Sungai Rawa is the only outlet and it flows northwards to join a tributary to the Palu valley (Sudomo et al. 1990). High mountain peaks and steep topography within a short distance characterize the area, as the result of the uplift that started in the Pliocene, following the juxtaposition of the east and north arms of Sulawesi (Moss and Wilson 1998). At an altitude of ca. 960 m asl, the lake basin covers an area of 32 km^2. It is the eighth largest lake in Sulawesi and the largest water body within the area of the LLNP. The Lindu plain is the only large wetland habitat in the National Park.

Climate and vegetation

Central Sulawesi is characterized by high humidity and temperatures. Mean annual precipitation ranges between 1800 and 2100 mm, and mean annual temperature decreases with elevation from 21 °C at 1000 m to 14 °C at 2400 m (Hijmans et al. 2005; WorldClim 2006; Culmsee et al. 2010). The modern climate of Indonesia is controlled by the seasonal movement of the Intertropical Convergence Zone (ITCZ) across the equator and inter-annual changes in the phase of the El Niño Southern Oscillation (ENSO) (Gunawan 2006). As the ITCZ moves southwards during the austral summer, the northwest monsoon delivers humid air and heavy rainfall to Indonesia, while during the austral winter the southeast monsoon brings relatively cool, dry conditions when the ITCZ is positioned over mainland Asia. According to Gunawan (2006), in the montane areas of Central Sulawesi, the rainfall is strongly

determined by the local topography. The air masses reaching the area from the northwest and southeast are lifted orographically, leading to the formation of clouds and rainfall throughout the year. As a consequence, the monthly amount of rainfall during the dry southeast monsoon is only slightly less than that of the wet northwest monsoon. The modern intra-annual climate of the montane areas of the LLNP can be described as perhumid with at most two months of slightly lower precipitation, corresponding to the southeast monsoon peak in August (Gunawan 2006). The inter-annual variability of rainfall is influenced by the coupled ocean-atmosphere phenomenon ENSO: during El Niño (La Niña) warm (cold) phases, Indonesia experiences lower (higher) rainfall than in other years (Philander 1990; Cane 2005). When El Niño occurs, the Lake Lindu catchment has experienced yearly low water and mean discharge despite the high retention capacity (Leemhuis and Gerold 2006), negatively affecting the fish population and the fishing market established by the local villages (Acciaioli 2000).

The eastern, northern and northwest corners of the lake are covered with low-lying areas of marshy grassland. These open swamps are used by the local population for cultivating rice and grazing cattle. The mountain areas surrounding the lake are dominated by lower montane vegetation, in which the dominant tree families are Lauraceae, Fagaceae, Sapotaceae, Moraceae and Euphorbiaceae (Culmsee and Pitopang 2009). The higher peaks range from 2000 to 2400 m asl are covered with upper montane vegetation, in which Podocarpaceae, Myrtaceae and Fagaceae are the dominant tree families (Culmsee et al. 2011).

Prehistory and history of Central Sulawesi (Table 7.1)

Central Sulawesi is rich in metal age megaliths, large worked stones in the shape of cylindrical vats, statues, urns and mortars (Kaudern 1938; Sukendar 1976; Bellwood 1979; Sukendar 1980a; Sukendar 1980b; Kirleis et al. 2012). The majority are located in the Napu, Besoa and Bada valleys, but megaliths are also found on the Lindu plain. The absolute age of the megaliths has not been determined with precision, but archaeological estimates range from 3000 BC to AD 1300. At the Pokekea site in the Besoa valley Kirleis et al. established a *terminus ante quem* for the erection of the large stone vats called *kalambas* of ca. AD 830 (Figure 7.1a; Kirleis et al. 2011; Kirleis et al. 2012). The authors further suggested a link between the opening of the forest in the valley when continuing burning started from ca. 2000 years ago and the early construction phase of the monuments. The relationship between the megalithic culture and the indigenous people who live around the park today remains open to speculation. It seems likely that Proto-Malay and Palaeo-Mongoloid people migrated into the area, but much more research is needed in order to answer this fundamental question about the megaliths and their creators.

The first Europeans who visited Central Sulawesi were the Portuguese in the middle of the 16th century, followed by the Spanish, who arrived via the Philippines. They never settled in large numbers, but their influence is still visible. They introduced maize, tomato, chili peppers and horses (Davis 1976).

Table 7.1 Overview of prehistory and history of Central Sulawesi and Lindu plain, as discussed in the text

Phases	Time	Central Sulawesi	Lindu plain
Pre-colonial period	5000 to 700 years ago (?)	Metal age population inhabits Central Sulawesi and builds a large number of megaliths	Megaliths also found on the plain
	16th century	Portuguese and Spanish visit Central Sulawesi, introduction of maize, tomato, chili pepper and horses	
Dutch colonial period	1605	Dutch East Indian company (DEI) in Palu	
	1648	Central Sulawesi comes under the rule of the Dutch East India company, but stays as a liege under the sultanate of Ternate (Moluku)	
	1668	The Dutch conquer Makassar, Bugis flee from Makassar to Central Sulawesi	
	17th cent.	Palu becomes an important sea trade harbour, the interior of the island remains isolated	Lindu plain inhabited by an ethnic group named after Lindu
	1710	Central Sulawesi comes under the rule of Makassar (South Sulawesi)	
	1891	Christianization of the Poso region (Central Sulawesi)	Remote mountain populations of western Central Sulawesi including the Lindu plain, remained relatively isolated
	End of 19th cent.		First Europeans in the area, south of Palu
	1905	Start of direct impact on the cultural landscape	Sarasin brothers visit Lindu plain and report of fishing traps, horses and lowering lake level
	1930		Discovery of schistosomiasis, the Dutch colonial administration maintained a policy of isolating the lake
Japanese rule and independence	1942	End of the Dutch rule and start of the Japanese colonial rule	
	1945	Independence of Indonesia, Dutch wars of re-conquest	
	1949	Final admission of independence	
	1950s		Bugis start fish market and clove gardens and open new areas for wet rice plantations
	1970s-80s	Lore Lindu National Park	Declaration of Lore Lindu National Park led to increased migration to the Lindu plain

A more important impact was the arrival of the Dutch, who opened up the areas in the lowlands starting from the 17th century. Before the Dutch arrival, there was little wetland rice growing, and agriculture activities were mostly focused on upland dryland rice, maize, and tubers grown under a shifting system. Population density in Central Sulawesi was very low. The small communities were ruled by kings and were relatively isolated from each other, and they were mainly located in the surrounding mountains, with no permanent settlements existing at that time (Kreisel et al. 2004). In contrast to the inhabitants of the Poso region, methods of wet rice cultivation were already well known to the inhabitants of western Central Sulawesi before the arrival of the Dutch, as reported by Valentyn (1724).

Today, the Central Sulawesi region is ethnically and culturally heterogeneous and comprises 15 indigenous groups, speaking 24 distinct languages. However, most of the people living today around the National Park are recent arrivals, or their descendants. The majority moved into the area as participants

in the government's transmigration programs which were at a peak in the 1970-80s, and as a result of conflicts elsewhere (Kreisel et al. 2004).

History of the Lindu plain

According to Kaudern (1925), the Lindu plain has been inhabited since at least the 17th century by an ethnic group known by the same name. Accounts of the land-use strategies of the indigenous groups which lived in the area are derived mostly from reports by Dutch missionaries, in particular Albert Christian Kruyt and Nicolaus Adriani. They explored the mountainous regions in the first half of the 20th century following the Dutch Christianization mission, which began in 1891. The activities of the missionaries were centered in the eastern part of Central Sulawesi and particularly in the region around the river Poso, while the remote mountain populations of Central Sulawesi, including the Lindu plain, remained relatively isolated. This is probably due to the discovery on the Lindu plain in the 1930s of infestations of the snails that harbour the blood flukes causing schistosomiasis (Clarke et al. 1974). Indeed, according to Acciaoli (1989), the Dutch colonial administration maintained a policy of isolating the region after the initial attempts in the 1910s and 1920s to improve wet rice cultivation. After Kruyt and Adriani, the next Europeans to visit the Lindu plain were the Sarasin brothers, two natural scientists, who mapped the watercourses from Palu to Palopo (Sarasin and Sarasin 1905).

The first people who started the exploitation of the resources of the Lindu plain were of Bugis origin from South Sulawesi, in the late 1960s. They first migrated into the area after the Kahar Muzakkar regional rebellion in the 1950s. They expanded the local economy by selling fresh and salted fish and starting the transport of fish from the lake by horse cart. Their arrival marked the opening up of new areas for wet rice cultivation, and clove gardens were established near the shore line (Weber et al. 2003). Soon after the establishment of the Lore Lindu protected area in the late 1970s, the government granted a special status for the Lindu plain. A buffer zone was established in the surrounding forest to allow the Lindu villagers to maintain their fields and to have access to forest products. As a consequence, migration to the Lindu plain increased in the 1980s when arable land in Central Sulawesi became scarce following the establishment of the LLNP (Kreisel et al. 2004).

Material and methods

A 123 cm long sediment core (LINDU_3) was recovered from Lake Lindu in 2006 using a Kajak corer (Renberg 1991) at a water depth of 46.5 m. The core was split into two halves and transported to the Department of Palynology and Climate Dynamics, University of Göttingen, Germany, where it was photographed and described lithologically and then stored in darkness at 4°C.

Radiocarbon dating and palaeomagnetic analyses

Altogether eight bulk sediment samples have been sent for radiocarbon dating, four to the Leibniz Laboratory for Radiometric Dating and Isotope Research at the University of Kiel, Germany, three to the AMS ^{14}C Laboratory in Erlangen, Germany and one to the Poznań Radiocarbon Laboratory, Poland (Table 7.2). Ages were calibrated using the R script CLAM with the SHCal_13 and postbomb_SH1-2.14C calibration datasets. Since some age reversals occurred, only the youngest ages were used to establish the chronology, and the resulting age-depth-model was tested using magnetostratigraphy.

Table 7.2 Accelerator mass spectrometry radiocarbon dates from Lake Lindu, calibrated age ranges at 95% confidence intervals ([a] Ages used for chronology are in bold; calibration done with R script CLAM, calibration curves SHCal13.14C and postbomb_SH1-2.14C for postbomb dates)

Sediment depth (cm)	^{14}C age BP / negative ^{14}C	Age AD (mean; 2σ range; probability)[a]	Delta ^{13}C	Laboratory	Laboratory code
17	-1372 ± 20	1988 (1985-1990; 57%)	-28.8	Kiel-2012	KIA47353
20	**-730 ± 40**	**2000 (1998-2004; 93%)**	**-31.9**	**Erlangen-2008**	Erl-12489
32	530 ± 30	1430 (1405-1450; 95%)	-30.3	Poznan-2008	Poz-24226
53	350 ± 30	1570 (1490-1645; 95%)	-27.2	Kiel-2012	KIA47354
62	**-482 ± 40**	**1958 (1957.61-1958.37; 5%)**	**-13.1**	**Erlangen-2008**	Erl-12490
77	**105 ± 20**	**1900 (1880-1930; 51%)**	**-28.2**	**Kiel-2012**	KIA47355
86	542 ± 41	1420 (1390-1460; 94%)	-33.5	Erlangen-2008	Erl-12491
100	**455 ± 20**	**1470 (1440-1500; 94%)**	**-27.3**	**Kiel-2012**	KIA47356

The sediment core was sub-sampled with a u-channel and sent to the Sedimentary Paleomagnetism and Marine Geology Laboratory at the Institut des sciences de la mer de Rimouski (ISMER) of the University of Québec at Rimouski, Canada. The natural remanent magnetization (NRM) was acquired at 1 cm intervals on the u-channel using a 2G Enterprises 755 cryogenic magnetometer with stepwise alternating field (AF) demagnetization at peak fields of 0 to 90 mT with 5 mT increments from 0 to 80 mT. Inclination and declination of the characteristic remanent magnetization (ChRM) were calculated using an Excel spreadsheet developed for that purpose (Mazaud 2005) with AF demagnetization steps from 5 to 90 mT (17 steps). This macro also allows calculation of component magnetizations and maximum angular deviation (MAD) values using principal component analysis (Kirschvink 1980). Due to the response function of the magnetometer pick-up coils some smoothing occurs, and the top and bottom ca. 7 cm of the u-channel have to be considered cautiously.

Palynological and micro-charcoal analyses

Sediment samples of 0.5 cm³ were taken at 4 cm intervals (31 samples) along the core for analyzing pollen, spores and non-pollen palynomorphs (NPPs), and at 2 cm intervals for micro-charcoal particles (62 samples). The samples were prepared using standard methods including 70% HF treatment (Faegri et al. 1989). Before sample processing, the spores of the marker *Lycopodium clavatum* were added to the samples for the calculation of the concentrations. Pollen and spore identification is based on the reference collection of tropical pollen and spores at the Department of Palynology and Climate Dynamics at the University of Göttingen, which includes specimens collected from the LLNP in 2011 and 2012, pollen keys and atlases for SE Asia (Flenley 1967; Powell 1970; Huang 1972; Garrett-Jones 1979; Stevenson 2000) and the online Australasian Pollen and Spore Atlas (APSA) hosted at Palaeoworks, Australian National University, Canberra (http://apsa.anu.edu.au). Identified pollen grains were counted to a sum of 300 and percentages were calculated relative to the pollen sums. Moraceae and Urticaceae were counted together because pollen grains from these two families are not distinguishable morphologically. The values of Moraceae-Urticaceae are overrepresented in the pollen assemblage due to the high production of pollen from these families (Jantz et al. 2014). Therefore, they are excluded from the total pollen sum for the calculation of the percentages of the remaining taxa. The opposite situation occurred for the Lauraceae, which is today one of the most important tree families in the area, but due to the thin and fragile exine, it is almost completely absent from the pollen assemblage. Concentrations are expressed in the diagrams as counts per cm^3 of sediment. Pollen taxa are grouped into lower montane rainforest, swamp, pioneer and secondary rainforest, anthropogenic indicator, palm and upland long-distance transported according to their altitudinal and ecological distributions, based on field observation and available literature (Keßler et al. 2002; Culmsee et al. 2010; Flora Malesiana collection: http://floramalesiana.org; Prosea collection: http://prosea.nl). Important NPPs are presented as concentrations per unit of volume (counts/cm^3). Poaceae pollen grains larger than 40 μm were counted separately. Although it is not possible to distinguish the grains of *Oryza sativa* from other Poaceae under the compound light microscope, pollen grains belonging to this group are larger than 35-40 μm (Chaturvedi et al. 1998). This group is therefore used, in combination with other anthropogenic proxies, as a possible indicator of rice cultivation.

Micro-charcoal particles that were seen under a normal light microscope as black and completely opaque with sharp edges have been counted (10-150 μm). Following Finsinger and Tinner (2005), at least 200 items, the total of micro-charcoal particles and *Lycopodium clavatum* spores, were counted and the concentration per unit of volume was calculated. Concentrations and proportions of the taxa are plotted against depth, and the ages of the record are discussed as time-windows in order to minimize the error due to the uncertainty of the age-depth model. For plotting and calculations, the software C2 was used (Juggins 2007).

Macro-charcoal analysis

Macro-charcoal particles (>150 μm) were counted in samples which were evenly spaced at 1 cm intervals along the sediment core (123 samples). The samples, of 2 cm^3 each, were prepared following the methods of Stevenson and Haberle (2005) and Rhodes (1998). Weak hydrogen peroxide (6 % H_2O_2) was used to partially digest and bleach organic material in the sediment samples when counted under a binocular dissecting microscope. The sample preparation procedure aimed to ensure that little particle fragmentation occurred during preparation. Results are expressed as the number of charred particles per cm³.

Diatom analysis

Sediment samples for diatom analysis were prepared according to standard methods at 2 cm intervals (62 samples) (Wang et al. 2013). 0.5 cm^3 sediment samples were treated with 30 % H_2O_2 and mounted using mounting media with a high refractive index (Mountmedia, Wako). At least 300 diatom valves were counted for each sample using an optical microscope with 1000× magnification. The diatom taxa were identified on the basis of reference collections and literature (Krammer and Lange-Bertalot 1986; Wang et al. 2010).

Numerical analysis

Local pollen assemblage zones were defined numerically by constrained cluster analysis using the software CONISS (Grimm 1987; Grimm 1993). The dissimilarity matrix was calculated as Cavalli-Sforza's chord distances of squared root transformed percentage data. All pollen and spore taxa were included in the analysis. Unconstrained multivariate statistical analysis was done to characterize the changes in vegetation composition of the past 1000 years using the software CANOCO 5 (ter Braak and Šmilauer 2002). The length of the compositional taxa gradient had a value of only 1.2 standard deviation (SD) units, so a linear model was chosen and a principal component analysis (PCA) was carried out with all identified pollen and spore percentage data. Data were centered and square root transformed to downscale the weight of a few dominant taxa.

Results

Lithology and chronology

Sediment core LINDU_3 consisted entirely of sediments with laminations that measured in millimeters. From 123-78, 62-36 and 6-0 cm sediment depths dark blackish colors predominated. Intercalated were light brown to whitish colors between 78-62 cm and whitish colors between 36-6 cm.

As the sediment-water interface was intact, the top of the record represents the year of coring, 2006. According to a linear extrapolation of the lowermost accepted ages, the base of the Lake Lindu record has an age of AD 1030. Since the radiocarbon dating results were rather heterogeneous and had a number of age reversals, only the youngest ages were linearly interpolated (Table 7.2; Figure 7.2a). Such results for radiocarbon dating seem to be rather common on the island of Sulawesi, where dating has turned out to be a challenge in many sediment archives (Haberzettl et al. 2013).
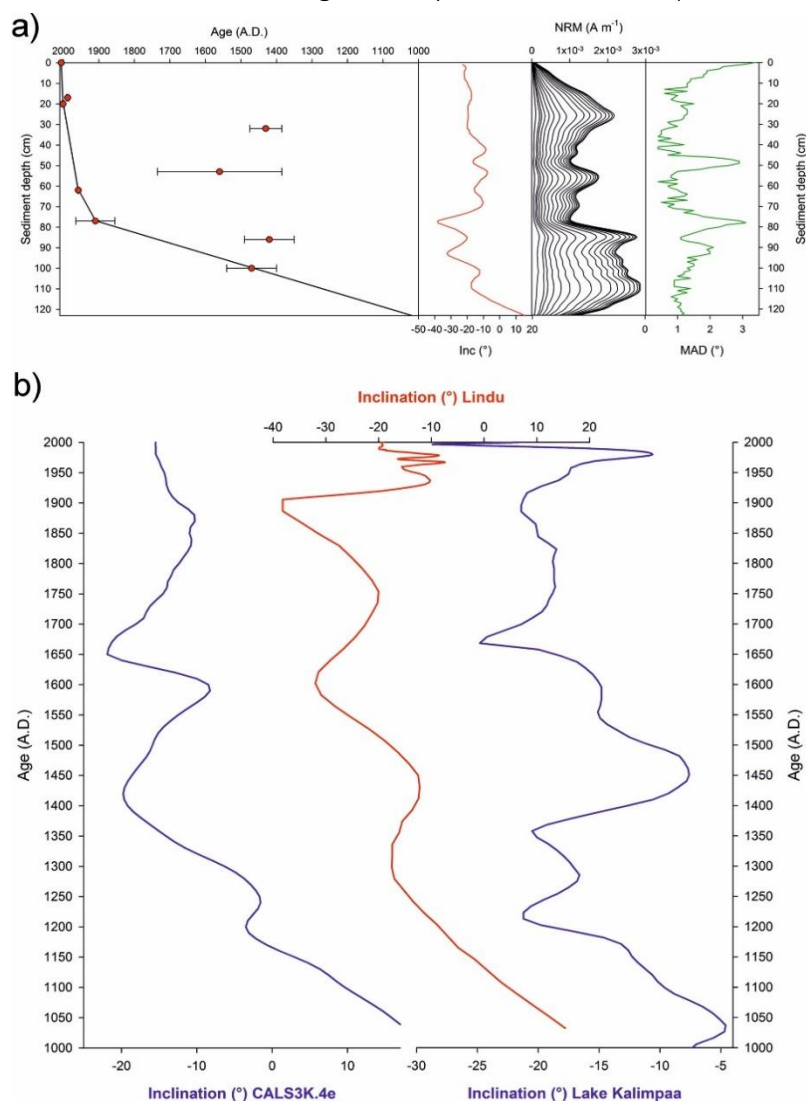
Figure 7.2 a) Chronology of the Lake Lindu sediment record as well as inclination (inc), the natural remanent magnetization (NRM) and maximum angular deviation values (MAD); b) comparison of the Lake Lindu inclination record to that from Lake Kalimpaa (Haberzettl et al. 2013) and the CALS3 K.4e (Korte and Constable 2011) model output calculated for the location of the sediment core from Lake Lindu

Palynological results

In total, 209 different pollen taxa were encountered, of which 77 rare taxa remain unidentified. The most representative lower montane forest taxa in the pollen assemblage belong to the Moraceae and Urticaceae families (average 38 %; min 10 %, max 50 %) (Figure 7.3). The next most representative taxa of lower montane rainforest are Fagaceae, mostly represented by *Lithocarpus-Castanopsis* (average 20%; min 9%, max 36%, percentages based on a total pollen sum excluding Moraceae-Urticaceae), *Acalypha* (average 6%, up to 17%), *Peperomia* (average 6%; min 2%, max 13%), Myrtaceae (average 3%, up to 9%), *Engelhardtia* (average3 %, up to 9%), Elaeocarpaceae (average 2%, up to 5%), *Ilex* (average

2%, up to 5%), *Celtis* (average 2%, up to 6%), *Myrica* (average 1%, up to 5%) and *Freycinetia* (average 1%, up to 4%). All these taxa together account for 53% of the pollen sum on average. The group of swamp plants is represented mostly by *Typha* (up to 4%) and *Callitriche* (average 2%, up to 8%). Pioneer and secondary forest taxa are represented mostly by *Macaranga* (average 4%, up to 11%), *Trema* (average 2%, up to 8%) and *Neonauclea* pollen (average 2%, up to 5%). Anthropogenic indicators are represented mostly by Poaceae (average 13%, up to 20%), Mimosoideae (up to 4%) and *Plantago* (up to 0.8%). Palms are little represented (average 1%, up to 6%). Upper montane pollen taxa are rare and mostly represented by *Phyllocladus* (average 1%, up to 4%). Pteridophyta spores are abundant and diverse (average of the total sum of pollen plus spores 29%, min 12%, max 45%, total number of taxa 81), although most of the taxa remain unidentified.

Pollen concentration values are high at the bottom of the sediment core between 123 and 115 cm (average 120×10^3 grains/cm^3), but they decrease from 115 to 75 cm (average 60×10^3 grains/cm^3) and have even lower values in the top part of the core from 75 to 0 cm (average 40×10^3 grains/cm^3).

The record is divided into two large clusters by the CONISS analysis. The first zone is lin-1 from 123 to 76 cm, and the second is lin-2 from 76 to the top of the core. Based on important changes and composition of the palynological, charcoal and NPP data, the two zones are additionally divided into six sub-zones: lin-1a (123-116 cm), lin-1b (116-84 cm), lin-1c (84-76 cm), lin-2a (76-64 cm), lin-2b (64-52 cm) and lin-2c (52-0 cm).

a)

b)

lower montane rainforest taxa — swamp taxa

Acalypha · *Freycinetia* · *Lithocarpus-Castanopsis* · Myrtaceae · *Celtis* · *Ficus* type · *Peperomia* · Elaeocarpaceae · *Typha* · *Callitriche*

% of tot pollen sum (Moraceae-Urticaceae excluded)

pioneer and secondary forest taxa — anthropogenic indicators — Pteridophyta

Macaranga · *Trema* · *Glochidion* · *Baccaurea* · *Neonauclea* · Poaceae < 40 um · Poaceae > 40 um · *Plantago* · Mimosoideae · Apiaceae · Cyatheaceae · *Nephrolepis*

% of tot pollen sum (Moraceae-Urticaceae excluded) — % of tot pollen and spores

Zones / Sub-zones

lin-2 post-colonial

- lin-2c (late 20th cent. to present)
- lin-2b (mid 20th cent.)
- lin-2a (early 20th cent.)

lin-1 pre-colonial

- lin-1c (ca. AD 1730-1910)
- lin-1b (ca. AD 1200-1730)
- lin-1a (ca. AD 1030-1200)

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Description of the palynological sub-zones

Lin-1a (123-116 cm; ca. AD 1030-1200)

The basal sub-zone of the record is characterized by low values of *Lithocarpus-Castanopsis* pollen (average 11 %), while other montane rainforest genera like *Freycinetia* and *Acalypha* have the highest values of the record (averages 2% and 16%). Poaceae pollen grains are well represented and they increase towards the top of the sub-zone (from 13% to 16%). Within this group, grains larger than 40 μm are found. Swamp pollen taxa have low values (average 2%) mirrored by high micro-charcoal particles (average 120×10^3 part/cm^3). Pioneer pollen taxa have high values, represented by *Macaranga* (average 7%) and *Trema* (average 5%).

Lin-1b (116-84 cm; ca. AD 1200-1730)

Lithocarpus-Castanopsis pollen values slightly increase (average 18%) while *Acalypha* decreases. Swamp pollen taxa values increase notably, represented mostly by *Typha* (up to 4%). Micro-charcoal concentrations decrease while *Glomus* spores increase. *Tilletia* and *Plantago* occur for the first time in this sub-zone. On average, the pollen concentrations decrease.

Lin-1c (84-76 cm; ca. AD 1730-1910)

In this sub-zone *Lithocarpus-Castanopsis* pollen values continue to increase (average 25%). Poaceae values decrease slightly (average 10%) and grains larger than 40 μm are no longer found. Swamp pollen taxa decrease (average 1%), mirrored again by high micro-charcoal values (average 80×10^3 particles/cm^3) and low *Glomus* concentrations (average 470 spores/cm^3).

Lin-2a (76-64 cm; early 20th century)

In this sub-zone pollen preservation is poor, and grains show corrosion and are often folded. The pollen grains of Moraceae-Urticaceae are particularly badly preserved.

Primary lower montane pollen taxa continue to rise, especially *Lithocarpus-Castanopsis* (average 29%). However, *Freycinetia* and *Acalypha* values markedly decrease (averages 0.5 and 3%), while *Celtis* increase (up to 6%). Pioneer and secondary pollen taxa are mostly represented by *Macaranga* (average

Figure 7.3 (previous 2 pages) Summary diagram from the Lake Lindu sediment core divided into temporal zones and sub-zones. The black lines are the locally weighted scatter plot smoothings (LOWESS) fitted to the sample values (light grey bars) to highlight trends. X-axes are rescaled for a better visualization of the least abundant taxa. **a)** Upper diagram Moraceae-Urticaceae (expressed in percentages of the total pollen sum); sum of lower montane rainforest, swamp, pioneer, secondary forest, anthropogenic, palms and long-distance transported pollen taxa (expressed as percentages of total pollen sum excluding Moraceae-Urticaceae); total Pteridophyta spores (expressed as percentages of sum of pollen and Pteridophyta spores); pollen, Pteridophyta and diatom concentrations (counts/cm^3). Lower diagram: macro- and micro-charcoal concentrations (counts/cm^3); swamp pollen taxa sum (percentages of the total pollen sum excluding Moraceae-Urticaceae); selected non-pollen palynomorph (NPP) concentrations (counts/cm^3); selected diatoms (counts/cm^3); CONISS dendrogram of the square root transformed proportions of all taxa (dissimilarity coefficient Edwards and Cavalli-Sforza's chord distance); **b)** most significant pollen taxa within the groups (percentages of the total pollen sum excluding Moraceae-Urticaceae); selected Pteridophyta taxa (expressed as percentages of sum of pollen and Pteridophyta spores)

5%) and *Neonauclea* (average 5%). As in the previous sub-zone, swamp pollen taxa continue to decrease, mirrored by high micro-charcoal and low *Glomus* concentrations. *Tilletia* values decrease markedly (average 47 counts/cm^3).

Lin-2b (64-52 cm; mid-20th century)

The difference between this and the previous sub-zone is in the increase of swamp pollen taxa (average 4%), while micro-charcoal concentrations decrease and *Glomus* increases. Peaks in *Plantago* pollen and *Botryococcus* colonies are recorded at the beginning of the sub-zone (1% and 7500 colonies/cm^3).

Lin-2c (52-0 cm; late 20th and beginning of 21st century)

This sub-zone marks the start of a decreasing trend for primary lower montane and pioneer pollen taxa, which continues until the top of the record. In contrast, Poaceae pollen values start to increase and grains >40 µm are found again. *Baccaurea* values increase from 44 to 28 cm, followed by the increase of *Neonauclea* and anthropogenic pollen indicators like Poaceae > 40 µm, Mimosoideae and Apiaceae. *Botryococcus* values increase, starting from 14 cm and continuing to the top. Several peaks of *Tilletia* are recorded, starting from 38 cm.

Macro-charcoal results

Macro-charcoal particles are found in all the samples along the core. There are exceptionally high concentrations in Sub-zone lin-1a at the bottom of the record, from 123 to 113 cm. At the same depths, charred particles were larger than in the rest of the record and remains of grass leaves were observed (Figure 7.1c).

Diatom results

Overall, 42 diatom taxa were identified, of which planktonic diatom, *Aulacoseira granulata* is the most important species with an average of 97% of the total diatom sum throughout the core. The values of the benthic diatoms, *Nitzschia frustulum*, *N. perminuta* and *Staurosira construens* peak shortly in Sub-zones lin-1a, lin-1b, and lin-2a. In Sub-zones lin-2b and lin-2c, the benthic diatom *Staurosira construens* is no longer present. The diatom concentration is stable with a lower average (3.25x10^8 valves/cm^3) in Zone lin-1, while the mean concentration doubles (6.39x10^8 valves/cm^3) with marked variations in Zone lin-2.

Multivariate statistical analysis

Principal component analysis (PCA) of all identified pollen and spore taxa was used to compare and characterized the patterns of palynological composition variation across the different prehistoric and historic phases in the Lindu results from ca. 1000 years ago (Figure 7.4). The first and second axes of the ordination diagram explain 22% of the variance. The samples in the middle Sub-zones lin-1c and lin-2a score mostly positively on the first axis, while the remaining samples score negatively (Figure 7.4a). The

scores of the second axis separate the assemblages into two groups (Figure 7.4b). Samples in Zone lin-1 score negatively (pre-colonial period), while those in Zone lin-2 (post-colonial period) score positively.

Discussion and interpretation

According to the age-depth model, a clear increase in the sedimentation rate from 0.52 to 40 mm/yr can be observed from the base to the top of the Lake Lindu record (Figure 7.2). This is an even more distinct change than what was observed at the much smaller lacustrine system of Lake Kalimpaa, where the sedimentation rate increased from 0.8 to 9.2 mm/yr due to human disturbances in its surroundings (Haberzettl et al. 2013; Wündsch et al. 2014; Biagioni et al. 2015a). At Lake Tondano, a modern age was obtained at 100-90 cm (Dam et al. 2001). This date was explained as the probable result of the admixture of recent soil organic matter into the sediment, since its properties did not suggest a radical change in depositional activity (Dam et al. 2001). However, if this age is assumed to be correct, a similar change in sediment accumulation occurred as at Lake Lindu. This is in accordance with the observation that Lake Tondano is seriously threatened by increasing silting up, which according to references dating back to 1979 has been reaching values of 200 mm/yr (Lehmusluoto 1997; Dam et al. 2001). These comparisons make such a drastic change in sediment accumulation conceivable. However, since such a change without the occurrence of mass wasting events is unusually high, the chronology has been tested using magnetostratigraphy and comparison to geomagnetic field model output, since palaeomagnetic secular variations (PSV) can be used as a significant tool to correlate Holocene regional records (Yang et al. 2009; Barletta et al. 2010; St-Onge and Stoner 2011; Ólafsdóttir et al. 2013).

A strong and stable ChRM (characteristic remanent magnetization) was isolated between 5 and 90 mT (Figure 7.2a). A viscous remanent magnetization was hardly observed and, when present, was easily removed at 5 mT. Maximum Angular Deviation (MAD) values of the ChRM of the Lake Lindu palaeomagnetic record are entirely below 3.5°, indicating a very well preserved magnetization (Stoner and St-Onge 2007). Unfortunately, declination seems to suffer from core twisting, which has often been observed with soft sediments (Ali et al. 1999; Haberzettl et al. 2013) and hence is not plotted. The inclination shows a trend from 15.4° to -38.2° from 123 to 78 cm, intersected by two high amplitude maxima in between. Values close to the ones expected, based on a geocentric axial dipole model (GAD = 2.64°S for Lake Lindu) for the site latitude, are only reached at the base of the core. After the decreasing trend, a change to -10.4° at 69 cm is observed. Thereafter, the amplitude of the variations is much lower, although distinct differences ranging between -7.4 and -16.4° (=9° difference) can be observed. From 32 cm to the top of the record, the amplitude further decreases to values between -17 and -23.7° (=6.7° difference) (Figure 7.2a). This change in amplitude is consistent with the observed increase in the sedimentation rate. The lower the sedimentation rate, the higher the amplitude in inclination variations, because a longer period of time is recorded during intervals of lower

sedimentation rate. During this longer period, larger variations in the inclination can occur. In contrast, high sedimentation rates only record short time intervals with minor variations in the inclination. Therefore the synchronous occurrence of higher amplitudes during phases with lower sedimentation rates, and lower amplitudes when there are higher sedimentation rates, is in agreement with the chronology. Further support for the chronology comes from the comparison of the inclination record from Lake Lindu with the data obtained from Lake Kalimpaa (Haberzettl et al. 2013) and the CALS3K.4e spherical harmonic geomagnetic model output for the coring site (Figure 7.2b; Korte and Constable 2011). While individual swings in the inclination curve of Lake Lindu are also found in the Lake Kalimpaa record, a similar general trend is found in the CALS3K.4e model. If one takes into account the error of the radiocarbon dating method itself of ±70 years for the accepted ages (Table 7.2) and the error in the chronology determined by age-modelling artefacts owing to linear interpolation, as well as the uncertainties contained in the CALS3K.4e model, an even better fit might be conceivable.

Although the radiocarbon-based chronology is conservative, the palaeomagnetic analyses support this approach and indicate that the age-depth model is a good first order approximation as a basis for palaeoenvironmental reconstruction. In addition, the inclination data extend palaeomagnetic knowledge into an area where such information is very scarce.

According to the age-depth model, the palaeoecological analyses of the Lindu core illustrate the vegetation, climate and fire history of the Lindu plain for the past ca. 1000 years. The following discussion divides the record into a pre-colonial period corresponding to Zone lin-1 starting from ca. AD 1030 and a post-colonial period, lin-2 after ca. AD 1910, which also includes the post-independence period from 1949 and more recent history.

The change in the sedimentation rate from lin-1 to lin-2 matches the change in pollen concentrations. The first zone is characterized by a low sedimentation rate and fits the higher than average pollen concentrations. The opposite is observed in Zone lin-2, where a high sedimentation rate corresponds to an average lower concentration of pollen. The increase in sedimentation rate can be linked to the fall in lake level at the beginning of the 20th century as observed by the Sarasin brothers in 1902 during their visit to the Lindu plain (Sarasin and Sarasin 1905).

The stable high percentage values of *Aulacoseira granulata*, a widespread planktonic diatom, common in carbonate-rich, eutrophic lakes (van Dam et al. 1994; Gómez et al. 1995), indicates that Lake Lindu continued to have a high nutrient content. However, an increase in human activities in Zone lin-2 can be inferred from the unstable and higher diatom concentrations, corresponding to cultural eutrophication and increasing amounts of nutrients being washed into the lake (Horner et al. 1990; Kirilova et al. 2010).

Pre-colonial period (lin-1, 123-76 cm, ca. AD 1030-1910)

At the beginning of the record (lin-1a, ca. AD 1030-1200), a well-developed montane rainforest surrounded the lake, as suggested by high values of *Freycinetia*, a climber found in old-growth montane rainforests. The grassland plain which today surrounds the lake shore was already developed, as indicated by high values of Poaceae. Also, indicators of human activities suggest a long history of landscape exploitation on the Lindu plain. Large Poaceae grains (>40 µm) which might include pollen of *Oryza sativa* type were already encountered. Local fires were very frequent or intense in this period, as shown by the high concentration of macro-charcoal. Large charred particles of grass were found, suggesting that fires burned the grassland on the lake shore. If humans caused the fires, it is possible that the opening of the forest was following a shifting system of cultivation, as a more permanent occupation would not have allowed pioneer fast growing taxa like *Macaranga* and *Trema* to proliferate around the lake. High values of micro-charcoal indicate high regional biomass burning, suggesting that the climate then was drier and/or periods of droughts were frequent. This is further confirmed by the low values of swamp taxa, indicating that the river discharge was low. The same period of drought and disturbance of the vegetation was recorded at Lake Kalimpaa, 15 km southeast (Wündsch et al. 2014; Biagioni et al. 2015a) and in the Besoa valley, 30 km south of the lake (Kirleis et al. 2011; Kirleis et al. 2012).

The increase in fungal spores of *Glomus*, starting from ca. AD 1200 to AD 1730 (lin-1b), is a good indicator of soil erosion in the lake catchment (Scott Anderson et al. 1984). Increasing soil erosion and the development of larger swamp areas around the lake point toward an increase in run-off and precipitation in this period and/or less frequent periods of drought, as also indicated by low micro-charcoal values. The reconstruction at Lake Lindu matches with the reconstructions of the average position of the ITCZ. Palaeorecords from the Southern Hemisphere, anticorrelated with palaeorecords from the Northern Hemisphere (Haug et al. 2001; Tierney and Russell 2007; Tierney et al. 2010), show that the ITCZ moved southwards, reaching its southernmost position of the past 2,000 years during the period commonly known as the Little Ice Age (LIA), ca. 15th-17th century. Such a changed position of the ITCZ to being more centered on the LLNP area would have caused seasonality to decrease and average annual precipitation to increase. Similar results were also found at Kalimpaa and Besoa (Kirleis et al. 2011; Kirleis et al. 2012; Wündsch et al. 2014; Biagioni et al. 2015a). Characteristic swamp taxa were *Typha* and *Callitriche*, indicating that stagnant swampy depressions spread around the lake.

Tilletia is a genus of smut fungi in the Tilletiaceae family, species of which are plant pathogens that affect various grasses including rice (Duran and Fischer 1961; Carris et al. 2006). Their occurrence for the first time in Sub-zone lin-1b and the decrease of pioneer taxa and macro-charcoal particles might indicate that a change occurred towards a more permanent and organized use of the plain, starting from around AD 1200. It is possible that increased precipitation had allowed wet rice cultivation to start on the shores

around the lake. The high values of *Freycinetia* indicate well-developed montane rainforests surrounding the lake, however Euphorbiaceae decreased, especially *Acalypha*, suggesting that disturbance of the forest also occurred.

Starting from ca. AD 1730 (lin-1c), average precipitation decreased and/or periods of drought increased, as indicated by low values of swamp taxa and *Glomus* spores and high micro-charcoal values. *Lithocarpus-Castanopsis* increased, Poaceae decreased with no more evidence of large grains from this family, suggesting abandonment or decrease of human activities around the lake. When the Sarasin brothers visited the Lindu plain in 1902, they found that a small group of people were living on the shore line of the lake (Sarasin and Sarasin 1905). The grassland plain which is visible today was already present, but no wet rice cultivation was encountered, the plain being grazed by horses (Figure 7.1b). The inhabitants lived off garden products and fishing. In fact, Lake Lindu has been well known for a long time for the abundance of its fish. When Adriani and Kruyt first visited the lake in 1897, they reported that the indigenous people used fish traps to provide for local consumption in the village of Langko near the southwestern shore (Adriani and Kruyt 1898). The cultural difference observed from ca. AD 1730 might have been caused by decreased precipitation and/or occurrences of long periods of drought. The people living around the lake might have been forced to limit their activities to fishing and cattle grazing in consequence of the no-longer favorable climatic conditions. However, additional causes of cultural changes might have played a role. For instance, conflicts and/or spread of diseases can severely affect both human populations and cultivated plants, but such events are not detectable with pollen and palaeoecological analyses.

Post-colonial period (lin-2, 76-0 cm, from ca. AD 1910)

Taxa from montane rainforest continued to increase at the beginning of the 20th century (lin-2a), in particular *Lithocarpus-Castanopsis*. Moraceae-Urticaceae decreased, but this was possibly an artefact of the bad preservation of the pollen grains in this sub-zone. At the same time, secondary forest taxa increased, indicating recovery of the forest, while macro-charcoal values remained low and anthropogenic indicators like *Tilletia* concentrations decreased. Dutch missionaries arrived in this area following the submission of the Kulawi *raja* in 1905. According to Acciaioli (1989), the area around Lake Lindu remained relatively isolated, despite the substantial modifications to the surrounding lowlands, introduced by the Dutch.

A clear increase in the sedimentation rate started at the beginning of the 20th century as indicated by the age-depth model, the change in amplitude of palaeomagnetic inclination and decreased pollen concentrations. Such a change in the sedimentation rate can be linked to the falling lake level observed by the Sarasin brothers in 1902 during their visit to the Lindu plain (Sarasin and Sarasin 1905).

Low values of *Glomus* and swamp taxa as well as high micro-charcoal values were also recorded in the previous sub-zone, lin-1c, suggesting that rainfall decreased and/or droughts increased before the increase in sedimentation rate observed from the early 20th century onwards. However, a link between human activities and the initial lowering in lake level could not be established, as this sub-zone records the recovery of the forest.

The *Botryococcus* colonies in lake sediments are related to periods with a high delivery of nutrients, and their presence is often used as a palaeoenvironmental proxy (Guy-Ohlson 1992). The arrival of Bugis communities in the area during the second part of the 20th century marked the beginning of new landscape opening around the shore of wet rice fields (Acciaioli 2001).The peak in colonies in Sub-zone lin-2b in the mid-20th century might be a consequence of increasing nutrients washed into the lake, as human activities increased around it. At the same time, *Glomus* and swamp taxa increased and micro-charcoal decreased, suggesting increased rainfall. However, the sedimentation rate remained high and the silting of the lake continued despite the changes in the rainfall regime. In this period, there is a clear lack of correlation between the change in the lake level and rainfall variability as reconstructed from the palynological assemblage. It seems likely that the trend towards a falling lake level continued from the mid-20th century until now, as a consequence of the diversion of larger amounts of water away from the lake for irrigation purposes (Acciaioli 2001) and erosion/sedimentation increased due to intensification of land-use practices.

Clear palynological evidence for increasing human activities are recorded, starting from the late 20th century (lin-2c) with the re-occurrence of Poaceae pollen >40 μm and the gradual decrease in primary lower montane rainforest taxa up to the present time. Secondary forest taxa like *Neonauclea* and anthropogenic indicators like Mimosoideae, Apiaceae, *Tilletia* and *Botryococcus* increase markedly, starting from very recent time. The palynological evidence for an increase in human activities matches the increase in people migrating to the Lindu plain in search of available land to cultivate after the 1970s, following the establishment of the LLNP (Acciaioli 2001; Weber et al. 2003).

Phases of prehistoric and historic human-landscape interactions on Lindu plain and the link to climate variability in Central Sulawesi

Various phases of cultural use of the Lindu plain are evident from the Lake Lindu record (Figure 7.3; Figure 7.4). In particular, changes occurred from ca. AD 1200 to ca. AD 1730 showing a more permanent use of the landscape. At the same time, the reconstructed rainfall regime indicates wetter conditions. Such a cultural change can be explained in two different ways. The first hypothesis is that the builders of the megaliths changed their strategies on the Lindu plain by establishing more permanent settlements, taking advantage of the wetter climatic conditions for wet rice cultivation. Accordingly, the disappearance of the metal age culture that produced the megaliths would be recent, ending on the

Lindu plain only in the 16ᵗʰ-17ᵗʰ century. These populations might have persisted in isolation from South Sulawesi and Palu, with which the contacts with other populations from South Sulawesi, Europe and mainland Asia were established long before. Alternatively, the changes recorded at ca. AD 1200 might represent the end of the activities of the megalith builders on the Lindu plain. A different ethnic group with knowledge of wet rice cultivation techniques might have replaced the people of the megalith culture.
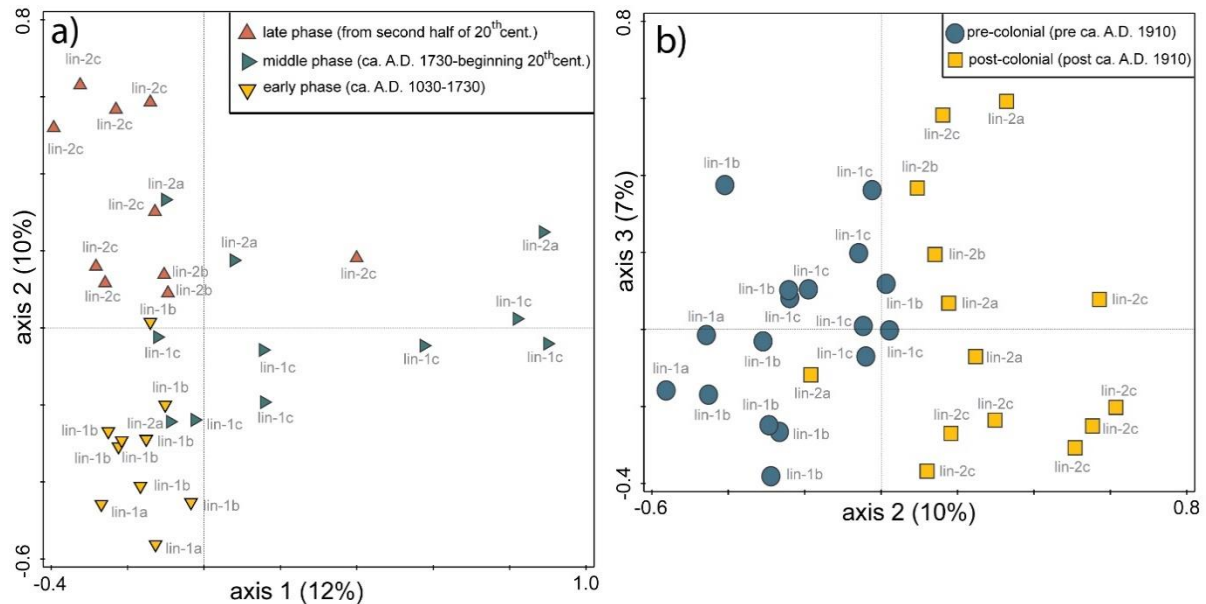
Figure 7.4 Principal component analysis (PCA) of all percentage data of identified pollen and spore taxa. Percentages calculated on the total sum of all taxa square root transformed. First, second and third axes are shown (cumulative explained variation 29%). Scatterplots represent the sample scores. Results are centered by taxa. Group of adjacent samples are marked differently to highlight the different groups corresponding to different prehistoric and historic phases. Palynological sub-zones are given in light grey; **a)** first and second axes scatterplots; **b)** second and third axes scatterplots

There is not yet a clear chronology for the megalithic culture in Central Sulawesi. At the megalithic Pokekea site, the dating of the bottom sediments of one of the large stone vats called *kalambas* established that the site was at least 900 years old (Kirleis et al. 2011; Kirleis et al. 2012). However, in contrast to the Lake Lindu record where there was a phase of forest recovery, the pollen analysis of the Besoa valley shows that open grassland persisted uninterrupted from 2000 years ago when deforestation started. It is not known when the megalithic culture in Central Sulawesi ended, therefore in order to confirm the validity of these hypotheses, more research is needed on these megalith sites. The forest recovery phase recorded from ca. AD 1730 to the early 20ᵗʰ century matches with historical reports of a different use of the plain by the so-called Lindu people, whose activities there were mostly limited to fishing and cattle grazing (Sarasin and Sarasin 1905). Although different causes might explain such a radical cultural change, climate might have been one important factor. Indeed, drier conditions

characterized this period as indicated by the low values of swamp taxa and high micro-charcoal concentrations.

Starting from the mid-20th century, the arrival of Bugis communities and the more recent population encroachment onto the Lindu plain following the establishment of the LLNP caused major changes (Acciaioli 2000; Kreisel et al. 2004), as also shown by the change in palynological composition of the Lake Lindu record (Figure 7.4). The lake level fall began already in the early 20th century, possibly a consequence of long-term drier conditions. However, such a trend has continued until now, although the increase in swamp taxa and decrease in micro-charcoal indicate wetter conditions characterizing the mid-20th century. This suggests that the silting of the lake is not being caused by natural climatic conditions, but rather by increasing human activities. Indeed, local inhabitants of the Lindu plain recently declared that migrants to the area, who cut down forests for coffee and cocoa plantations, were responsible for the increasing shallowness of the lake and the streams feeding into it (Acciaioli 2001). It is possible that, as deforestation increases, and larger amounts of water are used to irrigate new gardens and fields, river discharge will further decrease, damaging the cultivation and fishing activities on the Lindu plain and aggravating the effect of droughts caused by the occurrence of El Niño. A better management of the forested areas around the plain and the water resources of the basin would decrease the erosion and the sedimentation that have been decreasing the depth of Lake Lindu, as shown by its receding shoreline.

Conclusions

Palaeoecological analyses of a sediment core from Lake Lindu reveal a long history of human-landscape interaction on the Lindu plain, which has been modified by human activities during the past 1000 years. Although further investigations are needed, evidence of intense burning and possible shifting cultivation from an earlier phase from ca. AD 1030 to 1200 might be related to the metal age population which built megaliths in Central Sulawesi. From ca. AD 1200 to 1730, the climate became wetter as a consequence of the southward movement of the ITCZ. At the same time, fires decreased, and a more permanent effect on the landscape began. It remains open to speculation whether the architects of such a cultural change were the megalith people, or a different ethnic group. A phase of abandonment or less intense activities characterized the period from ca. AD 1730 to 1910. Following this phase of forest advancement, the more recent part of the Lake Lindu record shows a trend towards deforestation which started in the late 20th century until the present. The increase in sedimentation rate and lowering of lake level started at the beginning of the 20th century and these have continued until now, despite changes in the rainfall regime which occurred in the last hundred years, as reconstructed from the palynological record.

In conclusion, the Lindu record represents one further step in the increase of the knowledge of human and landscape history in Central Sulawesi and it highlights the potential for further palaeoecological and archaeological investigations in the area.

Acknowledgments

CHAPTER 8 - paper 4

8000 years of vegetation dynamics and environmental changes of a unique inland peat ecosystem of the Jambi Province in Central Sumatra, Indonesia

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Abstract

Despite their importance as global carbon sinks, peatlands of Southeast Asia have been rarely studied and our current knowledge of the dynamics and ecology of these ecosystems remains incomplete. Paleoecological and palynological multi-proxy analyses including analysis of pollen, spores, charcoal, testate amoebae and sediments were carried out on a 733 cm-long core taken from the Air Hitam peatland in the Jambi Province, Central Sumatra, Indonesia. The radiocarbon chronology reveals that peat formation started ca. 7800 years ago. The site was covered by dipterocarp-swamp mixed rainforest during the first 2000 years, after which freshwater swamp taxa became more important, in particular *Durio* trees. This lasted until ca. 4500 years ago, when the swamp vegetation shifted to a pole forest with abundant *Pandanus* thickets in response to a system change from minerotrophic to ombrotrophic conditions. For this period, macro-charcoal analysis reveals that fire frequency increased, possibly as a consequence of climate change linked to the onset of the modern El Niño Southern Oscillation (ENSO) in the late Holocene. The ombrogenous *Pandanus*-pole forest phases were characterized by lower water table fluctuations and higher peat accumulation rates. The area of Air Hitam has been for thousands of years a highly effective carbon sequestering peatland. Natural climate variability in the past did not affect the carbon storage function as suggested by the general increase in peat accumulation during relatively drier phases and increasing ENSO variability in the late Holocene. However, the recent changes caused by selective logging, drainage and conversion to oil palm plantations, have caused a decline in the peat swamp forest communities changing the ecological functions of the peatland. It is likely that the accumulated carbon will be released in the atmosphere due to exposure to aerobic conditions and its function will be lost unless a better management of the watershed is applied.

Introduction

The largest peatland areas in the tropics are currently found in Southeast (SE) Asia, where thick deposits of peat have accumulated as a consequence of conditions such as low topographic relief, high biomass production and high rainfall (Page et al. 2009; Page et al. 2010; Dommain et al. 2011). The majority of these peatlands formed in delta areas and along river valleys in low-altitude, watershed positions (Page et al. 2011). Estimates indicate that they cover an area of about 240,000 km^2, of which the vast majority is in Indonesia (82%), Papua New Guinea (8.8%) and Malaysia (8.3%; Page et al. 2011). Their global importance as carbon reservoirs is well acknowledged, as they store an estimated 17–19% (65 Gt) of the global peat carbon pool (Page et al. 2011). In addition to their carbon storage function, the peat swamp ecosystems growing on tropical peatlands are regarded as unique (Schrier-Uijl et al. 2013) and provide a large number of eco-system services. For instance, tropical peat swamp rainforests are important climatic mitigating areas as they supply water and prevent erosion (Yule 2010) during extreme climatic events such as the occurrence of El Niño and La Niña episodes. Furthermore, as shown by

Kuniyasu (2002) in the lowland plain of the Kampar River in Sumatra, local villages and the majority of capital-less people, greatly benefit from the collection of forest products and fishing in peat swamp areas.

For thousands of years, these extensive peat deposits have functioned as a large storage for carbon (Page et al. 2004; Page et al. 2010). However, since the 1970s, rapid land-use changes have turned the majority of the SE Asian peatlands into degraded ecosystems, which are now emitting the formerly stored carbon (Hooijer et al. 2010; Jauhiainen et al. 2010; Page et al. 2011; Kurnianto et al. 2015). Within Indonesia, the island of Sumatra has the largest area of peatland (72,000 km^2; Wahyunto et al. 2004). According to Giesen (1993, 1994) at the end of the last century, peat swamp and freshwater swamp forests in Sumatra covered an area of 92,865 km^2. However, less than 9.3% of the large areas of peatland in Sumatra remains in pristine conditions (Giesen 1994). The establishment and rapid increase of palm oil plantations and production in SE Asia have contributed to deforestation across the SE Asian region. Around one third of the peatlands of peninsular Malaysia, Borneo and Sumatra were converted between 1990 and 2007 (5.1 Mha of the total 15.5 Mha) and most of the remaining peatland forests were intensively logged (Page et al. 2011). The conversion of peatlands to agricultural use requires both removal of the primary vegetation and soil drainage, resulting in peat subsidence through a combination of water loss, enhanced aerobic decomposition and compaction (Couwenberg et al. 2010; Miettinen et al. 2012). Such a change of the peatlands, leads to increasing sensitivity of these systems towards fires, with devastating results. For instance, widespread fires during the strong El Niño year of 1997–1998 burned a total of 308,000 ha of peat swamp and freshwater swamp forests in Sumatra alone (Tacconi 2003). It is estimated that the conversion of the coastal peatlands in SE Asia have caused the burning and emission of 1400 Mg per ha of carbon, which correspond to 2900 years of carbon accumulation in 1 ha of coastal peat (Kurnianto et al. 2015). Peatland fires and climate are closely linked, as increase in fires were found to be coupled with periodical droughts caused by the occurrence of El Niño events (van der Werf et al. 2008). This suggests, as pointed out by Dommain et al. (2011), that in the future carbon emission from burning and oxidizing SE Asian peatlands might increase, if dry-season rainfall will decrease and extreme precipitation events linked to ENSO will increase, as predicted by climate models (Li et al. 2007; Stocker 2014).

Because of the complex network of interactions of peatland dynamics, a better understanding of the main components and processes forming a peat system is needed when aiming at managing and/or restoring peatlands. In order to evaluate and predict peat processes that control peat accumulation, these components should be investigated (Brady 1997). These include age of the deposits, vegetation, hydrology and soil, as peat swamp ecosystems are the result of combined action of allogenic (climate, hydrology) and autogenic (vegetation) factors which vary in space and time depending on the thickness of the peat deposit and environmental/hydrological settings (Brady 1997). The relative importance of

each of these factors in shaping the system functions on the long-term can be evaluated using paleoenvironmental and paleo¬ecological investigations. This is due to the fact that in general the horizontal pattern of vegetation types found on the ground can also be found in a temporal succession in the center of the raised bogs (Anderson 1963), as palynological analysis carried out on peat deposits indicate (e.g. Anderson and Muller 1975). Thus peat deposits are important natural archives for dynamics and developments of peatlands. However, despite their global and local importance, in SE Asia only a few studies have included paleoecological and paleoenvironmental analyses (Figure 8.1).

The goal of this multi-proxy study is to reconstruct the paleoecology and paleoenvironment of an Indonesian peatland, and to investigate the ecosystem development and dynamics since the peat began to accumulate. We present results of a 733 cm-long core taken from the Air Hitam peatland in the Sarolangun district, in the inland of the Jambi Province in Central Sumatra, Indonesia. We use pollen and spore data to infer the vegetation history and community phases, testate amoebae to infer paleohydrology and water table fluctuations and peat characteristics to infer the rate of decomposition of peat in relation to vegetation phases and peat accumulation rate. We apply the decomposition approach on macro-charcoal data, as developed by Higuera et al. (2009), in order to isolate peaks which are considered a signal of local fires and thus infer fire regime characteristics through time. We compare the reconstructed history of fire to vegetation and hydrology, to unravel the effects of changing fire regime on the peatland ecosystems for the past ca. 7800 years. The combined results of Air Hitam represent a further step forward to the gaining of a more comprehensive understanding of modern SE Asian peat swamp ecosystem dynamics and their functions in the long-term.

Environmental settings of the study site

Central Sumatra lies within the influence of the Intertropical Convergence Zone (ITCZ) and experiences a wet tropical climate. The mean annual rainfall for Air Hitam area is 2900 mm and annual temperatures average by 26.8 °C (Hijmans et al. 2005; Figure 8.2). Seasonality of rainfall is not usually marked, but a long wet season of 9–10 months is alternated with a short drier season of two or three months from May to September corresponding to the onset of the southeast monsoon (Aldrian and Susanto 2003). At the inter-annual scale variability of rainfall is influenced by changes in the phase of the El Niño Southern Oscillation (ENSO) and the Indian Ocean Dipole (IOD). ENSO is a coupled ocean–atmosphere phenomenon consisting of two phases. During the El Niño warm phase, Indonesia experiences lower rainfall than in other years, while higher rainfall characterizes the cold phase La Niña (Philander 1990; Cane 2005). The IOD is an aperiodic oscillation of the Indian Ocean sea-surface temperatures. A positive IOD phase is accompanied by cooling of waters in the eastern Indian Ocean and droughts in Sumatra. The negative phase of the IOD brings the opposite conditions (Webster et al. 1999).
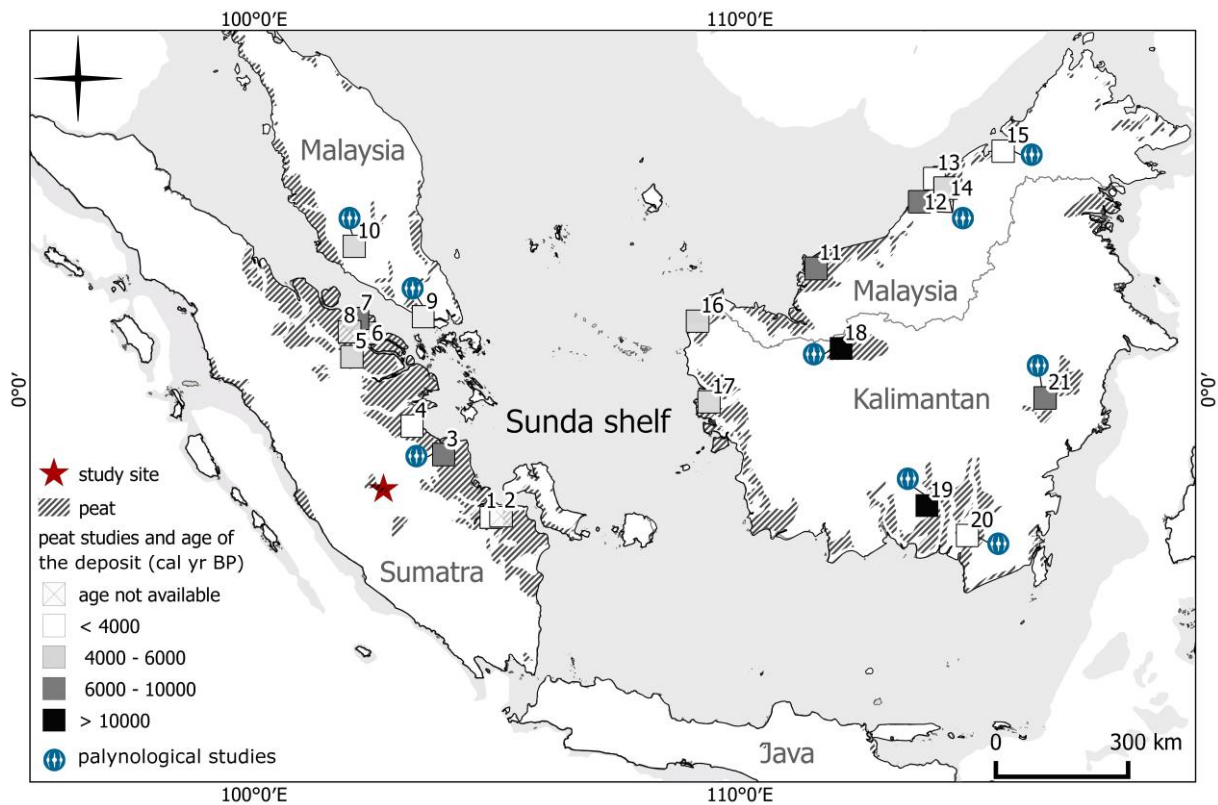
Figure 8.1 Map of the Sunda shelf with the red star representing the location of the Air Hitam inland peat. The light shaded area delineates the Sunda Shelf as it was exposed at the time of the Last Glacial Maximum (based on Sathuamurthy and Voris 2006). Line markers show the extension of peat deposits (data source: FAO Harmonized Soil Map of the World, Histosols and Gleysols layers only). Square symbols correspond to the approximate position of the available studies on peatlands in the area modified from Dommain et al. (2011): **1)** Muara Telang (Furukawa 1994); **2)** Sugihan (Brady 1997); **3)** Batang Hari Delta (Cameron et al. 1989; Esterle and Ferm 1994; Sabiham 1988; Sabiham and Furukawa 1986; Silvius et al. 1984); **4)** Pulau Kijang (Furukawa 1994; Shibata et al. 1997); **5)** Siak Kanan (Diemont and Supardi 1987a; Neuzil 1997; Ruppert et al. 1993; Supardi et al. 1993); **6)** Palau Padang (Brady 1997); **7)** Pulau Bengkalis (Neuzil 1997; Supardi et al. 1993); **8)** Siak Kecil (Gunawan et al. 2012); **9)** Pekan Nanas (Haseldonckx 1977); **10)** Tasek Bera (Morley 1982; Phillips and Bustin 1998; Wüst et al. 2002; Wüst and Bustin 2004); **11)** Rajang River Delta (Staub and Esterle 1993; Staub and Esterle 1994; Staub and Gastaldo 2003); **12)** Batu Niah (Cole 2012; Cole et al. 2015); **13)** Sungai Dua Forest Reserve and Senadin (Cole 2012; Cole et al. 2015); **14)** Baram River area (Anderson 1964; Anderson and Muller 1975; Brünig 1974; Dommain et al. 2015; Esterle and Ferm 1994; Morley 2013; Muller 1963; Sabiham 1990; Tie and Esterle 1992; Wilford 1960; Woodroffe 2000); **15)** Lawas River area (Anderson 1963; Morley et al. 2011; Muller 1963; Wilford 1960); **16)** Teluk Keramat (Neuzil 1997); **17)** Rasau Jaya (Anshari et al. 2010; Diemont and Supardi 1987b; Notohadiprawiro 1981); **18)** Pemerak (Anshari et al. 2001; Anshari et al. 2004); **19)** Sebangau (Kershaw et al. 2000; Morley 1981; Morley 2013; Neuzil 1997; Page et al. 1999; Page et al. 2004; Page et al. 2006; Rieley and Page 1997; Sieffermann et al. 1988; Sieffermann et al. 1992; Sieffermann et al. 1996; Yulianto and Hirakawa 2006; Yulianto et al. 2004; Wüst 2009; Wüst et al. 2008); **20)** Pulau Petak Delta (Notohadiprawiro 1981; Sabiham 1988; Sumawinata 1998); **21)** Kutai lake area (Hope et al. 2005). Different colors of the squares denote different ages for the initiation of the peat and the pollen symbol (blue) indicates the sites where palynological analysis was conducted

The eastern side of Sumatra is dominated by alluvial soils while the swampy areas of Riau and Jambi Province as well as South Sumatra are dominated by organosols (Whitten 2000).

The Air Hitam area in the Sarolangun Regency is a topographic depression constricted to the north, south and west by higher elevation areas (>70 m asl, **Figure 8.3**). The watershed is characterized by three different watercourses as part of the Batang Asai and Air Hitam rivers which join to the east in the proximity of Pauh village. Behind the interfluvial terraces of the watercourses that crossed the area, two
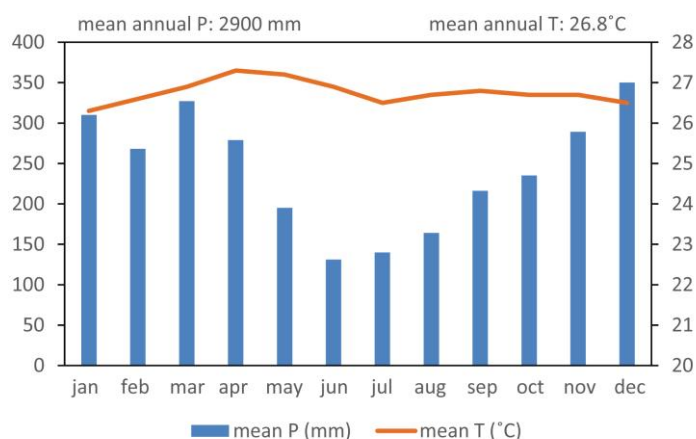
Figure 8.2 Monthly mean precipitation, temperature values and annual means for the Air Hitam area (2.047685°S 102.665362°E) based on the interpolations of observed data representative of the period 1950–2000. Data source: WorldClim — Global Climate Data, http://www.worldclim.org/

peat domes are found and cut by the central course to a characteristic bilobate shape covering an area of about 400 km² (Figure 8.3). Wetlands International mapped the area of peatland in Sumatra in 2003. According to the report, the Air Hitam domes are characterized by thick peat deposit with average depths of more than 4 m above the mineral substrate (Wahyunto et al. 2003). Inland peatlands, such as the Air Hitam peat domes, form along river valleys and their initiation seems to have been the result of rising ground water levels and paludification due to changes in sea level (Dommain et al. 2011). The mineral substrate restricted water discharge leading to permanent waterlogged conditions and, under high rainfall conditions, accumulation of peat. The Air Hitam peat domes represent the largest contiguous area of thick peat outside the coastal environment in the Jambi Province. Cross-sections of the peatland reveal a series of hummocks and hollows rather than a perfect dome due to the complex geomorphology of the mineral and alluvial substrate.

Most of the forest cover on the northern side of the peat dome was burned between 1990 and 2000 (WWF Indonesia 2010) when the palm oil companies PT Era Mitra Agro Lestari (PT Emal) and PT Jambi Agro Wijaya (Jaw SPT) started to drain the Air Hitam peat domes to convert the area to palm oil plantations. Canals and ditches were built to facilitate the removal of water, necessary for starting palm oil plantations. There is no information on the vegetation communities covering the peat domes before the conversion started and currently the land is covered with palm oil plantations, degraded bare lands and a small forest remnant on the southeastern part of the northern dome (Figure 8.3).

Materials and methods

In 2012 three sediment/peat cores were taken on the northern section of the Air Hitam peat dome using a Russian Corer. The cores were named according to the two oil companies active in the area; PT Era Mitra Agro Lestari (PT Emal) and PT Jambi Agro Wijaya (PT Jaw). The first two cores B21 (2.053377°S 102.694207°E; 53 m asl; 750 cm-long) and D21 (2.053321°S 102.721165°E; 47 m asl; 330 cm-long) were taken on young oil palm plantations (4–6 years) at the time under concession of PT Emal. The third core Jaw SPT was taken in the forest remnant under PT Jaw concession (2.047685°S 102.665362°E; 57 m asl; 733 cm-long; Figure 8.3). The three cores were photographed and described at the University of Jambi

following the classification systems for SE Asian tropical peat (Esterle and Ferm 1994; Wüst et al. 2003) using peat attributes like color and plant part composition. Afterwards the cores were moved to the Department of Palynology and Climate Dynamics, University of Göttingen, Germany. Fourteen samples of small organic bulk samples, plant remains or charcoal (see Table 8.1) were collected from the three cores and sent for Accelerator Mass Spectrometry (AMS) radiocarbon dating o the Laboratories at the University of Erlangen in Germany and at the National Taiwan University to establish an absolute chronology for the deposits. The core Jaw SPT presents a complete peat section and the peat surface is the least disturbed of the three. It was therefore selected as master core and used for multi-proxy palaeoenvironmental analyses. The age-depth model of the Jaw SPT core was constructed with CLAM 2.2 (Blaauw 2010) script in R (R Core Team 2012), using the Southern Hemisphere calibration curve SHCal13.14C.
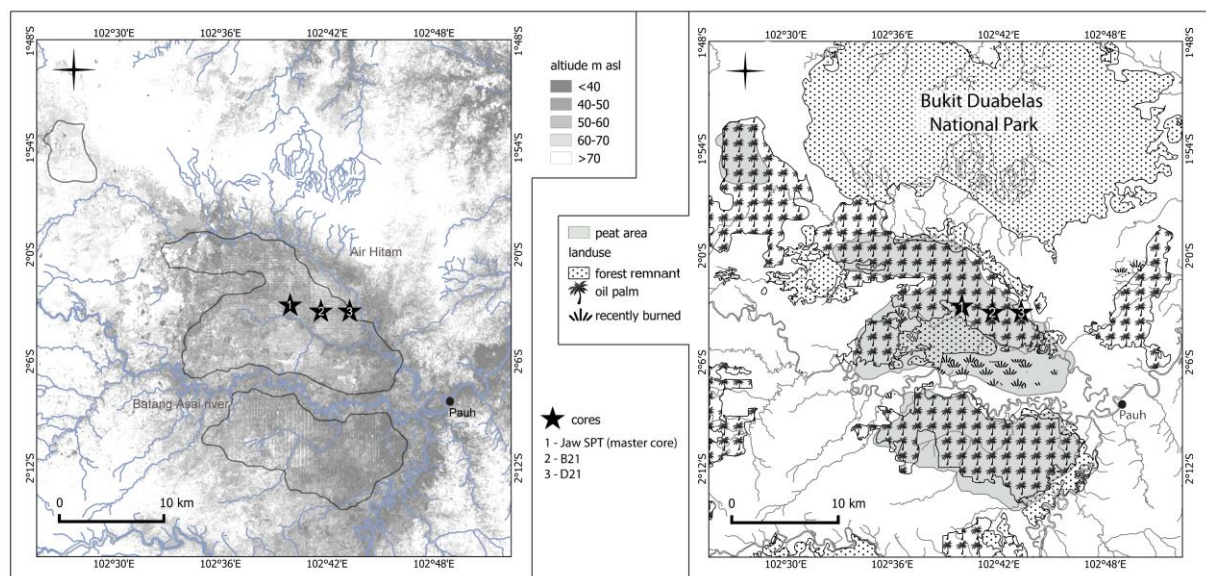
Figure 8.3 Maps of the study areawith black star symbolsmarking the locations of the coring. Left: altitudinal map of the Air Hitam watershed (Digital Elevation Model: ASTER GDEM Version 2, METI and NASA). Right: current land-use of the study area. Data source: Land Cover 2011: the Ministry of Forestry, the Republic of Indonesia (http://appgis.dephut.go.id/appgis/download.aspx); river data: digitalized from Bakosurtanal, Peta Rupa Bumi Indonesia scale 1:50.000, 2013

Palynological analysis

In total 37 sediment subsamples (0.5 cm³) were taken in 20 cm intervals along the core for analyzing pollen and spores. The samples were prepared using standard methods (Faegri et al. 1989) and before sample processing, the marker *Lycopodium clavatum* was added to the samples for the calculation of the concentrations. Pollen and spore identification is based on the reference collection of Indonesian tropical pollen and spores at the Department of Palynology and Climate Dynamics at the University of Göttingen, peat swamp pollen and spore taxa at the Oxford Long-Term Ecology Laboratory at the Department of Zoology at the Oxford University and available literature and databases on SE Asian

tropical pollen and spore taxa (e.g. Flenley 1967; Powell 1970; Huang 1972; Anderson and Muller 1975; Garrett-Jones 1979; Sabiham and Furukawa 1986; Tissot et al. 1994; Stevenson 2000; Premathilake and Nilsson 2001; Kodela 2006; Cole et al. 2015; Jones and Pearce 2015; The Australasian Pollen and Spore Atlas http://apsa.anu.edu.au). Identified pollen and spore grains were counted until a sum of more than 300 pollen grains was reached. Pollen and spore percentages were calculated relative to the pollen and spore total sum. Pollen taxa are grouped in dipterocarp-swamp mixed rainforest, freshwater swamp, peat swamp and open/pioneer taxa according to their distribution and ecology (Anderson 1963; Anderson and Muller 1975; Esterle and Ferm 1994; Brady 1997; Haseldonckx 1977; Cameron et al. 1989; Phillips and Bustin 1998; Page et al. 1999; Wüst and Bustin 2004; Hope et al. 2005; Yulianto and Hirakawa 2006; Gunawan et al. 2012; Morley 2013; Cole et al. 2015; Flora Malesiana collection: http://floramalesiana.org; Prosea collection: http://prosea.nl). A complete list of pollen and spore taxa included in the groups can be found in the Table 8.2. For calculations the software C2 was used (Juggins 2007).

Testate amoebae analysis

Peat samples for testate amoebae analysis were taken from the same depth as those for pollen analysis. For extraction of testate amoebae 4 cm^3 of sediments and two *Lycopodium clavatum* tablets were boiled in distilled water for 10 min and sieved over a filter of 500-µm mesh, with the filtrate then back-sieved over 10-µm mesh. Microscope slides were prepared and the tests (shells) of testate amoebae were identified and counted at 200x and 400x magnification. Due to the variable concentrations of tests, a count of 100 was not always achievable. However, it has been demonstrated elsewhere that a meaningful paleoenvironmental signal may still predominate over random noise and major changes for counts of less than 100 but more than 50 (Payne and Mitchell 2009). To downscale the effect of different counts, the percentages were calculated based on the initial calculation of concentrations per unit of volume. Determination of testate amoebae was based on Charman et al. (2000), Mazei and Tsyganov (2006), for details on taxonomic references see also Krashevska et al. (2007). Testate amoebae with similar habitat preferences were classified into four environmental groups: "water" group, includes testate amoebae predominantly occurring in water bodies and/or in the regularly flooded areas; "peat moss" group, includes testate amoebae predominantly occurring in submerged and/or dry peat bryophytes; "green moss" group, includes testate amoebae predominantly occurring in non-peat bryophytes, a habitat with high water table fluctuations; "soil" group, includes testate amoebae predominantly occurring in different soil horizons, a habitat with higher water table fluctuations. Additionally, based on the established literature dry indicators were distinguished and separately grouped (e.g. Chardez 1965; Bobrov et al. 1999; Charman et al. 2000; Bobrov et al. 2004; Mazei and Tsyganov 2006). The test size variation between and within the species can be indirectly related to

hydrology and or temperature (Bobrov et al. 1999; Mitchell et al. 2008). In order to highlight the relationship between microhabitat changes and testate amoebae size, the tests were also clustered to the three different size classes based on the average longer axes measurements (small: <75 μm, medium: 75–105 μm and large: >105 μm).

Macro-charcoal analysis

Macro-charcoal particles (>150 μm) were counted for samples evenly spaced at 1 cm intervals on the first 661 cm (peat-organic layers). The samples (1 cm^3 each) were prepared following the method of Stevenson and Haberle (2005) adapted from Rhodes (1998), which is a procedure that greatly limit particle fragmentation. Weak hydrogen peroxide (6% H_2O_2) was used to partially digest and bleach organic material in the sediment and samples were wet-sieved using a 125 μm filter. All charcoal particles were counted under a binocular dissecting microscope. Concentrations are expressed as number of charred particles per cm^3.

Numerical analysis

Clustering

Palynological zones are defined numerically according to the dissimilarity matrix of Euclidian distances of squared root transformed percentage data, via constrained cluster analysis using the software CONISS (Grimm 1987; Grimm 1993). All pollen and spore taxa were included in the analysis.

Diversity index and palynological compositional change

From the palynological and testate amoebae data, diversity indices (DIs) were estimated via 'rarefaction analysis' (Siegel 1986) which is the calculation of the expected number of taxa $E_{(Tn)}$, of a smaller population (n) as compared to the original (N) in a random sample of n individuals from a smaller population of N individuals (Birks and Line 1992).

Empirical data suggests that the use of a pollen type diversity index based on the rarefaction analysis calculated on a small sample of pollen taxa (i.e. 10), well correlate to the landscape diversity around the deposit (Matthias et al. 2015). Additionally, tropical peat deposits are covered by forest, and flowers or inflorescences drop and become part of the peat. As a consequence, the pollen signature in the peat is dominated by this local surface vegetation, with little influence from the long-distance wind transported pollen component (Anderson and Muller 1975). For tropical peat swamp deposits, palynological DI is therefore probably close to Whittaker's gamma diversity/within-landscape (Whittaker 1977; Odgaard 2007) which correspond to plant diversity of the peatland community over the site. Rarefaction was calculated with the software PAST (Hammer et al. 2001).

The amount of palynological compositional change or turnover (PCC) was estimated by detrended canonical correspondence analysis (DCCA) using as external constraint the sample age following the method described by Birks (2007). The gradient length, measured in standard deviation units, of the first time constrained axis is used as indicator of compositional turnover (Birks 2007). First, all pollen and spore percentages were transformed to square-roots to stabilize variances and rare taxa were not down-weighted. Detrending was done by segments with non-linear scaling. With the application of detrending and scaling options, the sample scores are scaled in standard deviation (SD) units of compositional change or turnover with time (Hill and Gauch 1980). The DCCA was implemented using CANOCO 5 (ter Braak and Smilauer 2002).

Principal component analysis

The relationships between vegetation communities in the different palynological zones were analyzed using principal component analysis (PCA) as implemented in CANOCO 5 (ter Braak and Smilauer 2002). PCA was carried out with all identified pollen and spore percentage data. Environmental groups and size classes of testate amoebae were included as supplementary environmental variables. Data were centered and square root transformed to downscale the weight of a few dominant taxa.

From macro-charcoal to local fire history

Charcoal raw data were treated using the peak detection analysis developed in the software Charanalysis (Higuera et al. 2009). The data were first interpolated to the median temporal resolution (11 years). The interpolated charcoal concentrations were multiplied by the estimated sedimentation rate (cm/yr) to obtain the charcoal accumulation rate (CHAR, particles/cm^2 per yr) of each sample. The low-frequency variations in the charcoal record ($C_{background}$) are assumed to represent changes in secondary charcoal caused for instance by sediment remixing. To remove $C_{background}$ a locally weighted regression was applied. A 400-year window which maximized the signal-to-noise index and the goodness-of-fit between the empirical and the modeled Cnoise distributions was used (Higuera et al. 2009). The subtracted residual series, C_{peak} is assumed to be composed of two subpopulations (Higuera et al. 2008; Higuera et al. 2009): C_{noise}, representing variability in sampling and analytical and naturally occurring noise, and C_{fire}, representing charcoal input from local fires. A Gaussian mixture model was used to identify the C_{noise} distribution. To separate these two populations, the 99th percentiles of the C_{noise} distributions were considered as thresholds. An additional test, 'Poisson minimum-count' was done to eliminate the peaks from statistically insignificant counts. Finally, we inferred aspects of the local past fire regime based on the frequency of peaks and their magnitude. Peak magnitude, the number of charcoal pieces from all samples defining a given peak (i.e. all samples above the threshold value; number of pieces per cm^2 per peak), is an estimation of total charcoal deposition per fire event (Higuera et al. 2009). Therefore, changes in peak magnitude at millennial time scales were used as a

proxy for fire size and/or fuel consumption for a given area burned. We used the distribution of fire frequencies (number of fire episodes per 800 years) to characterize the temporal characteristics of fire regimes at Air Hitam for the past ca. 7800 cal yr BP.

Results

Core descriptions and stratigraphy

In general, the three cores mainly consist of dark brown hemic peat layers. Several layers with charred particles are observed in the core (Figure 8.4).
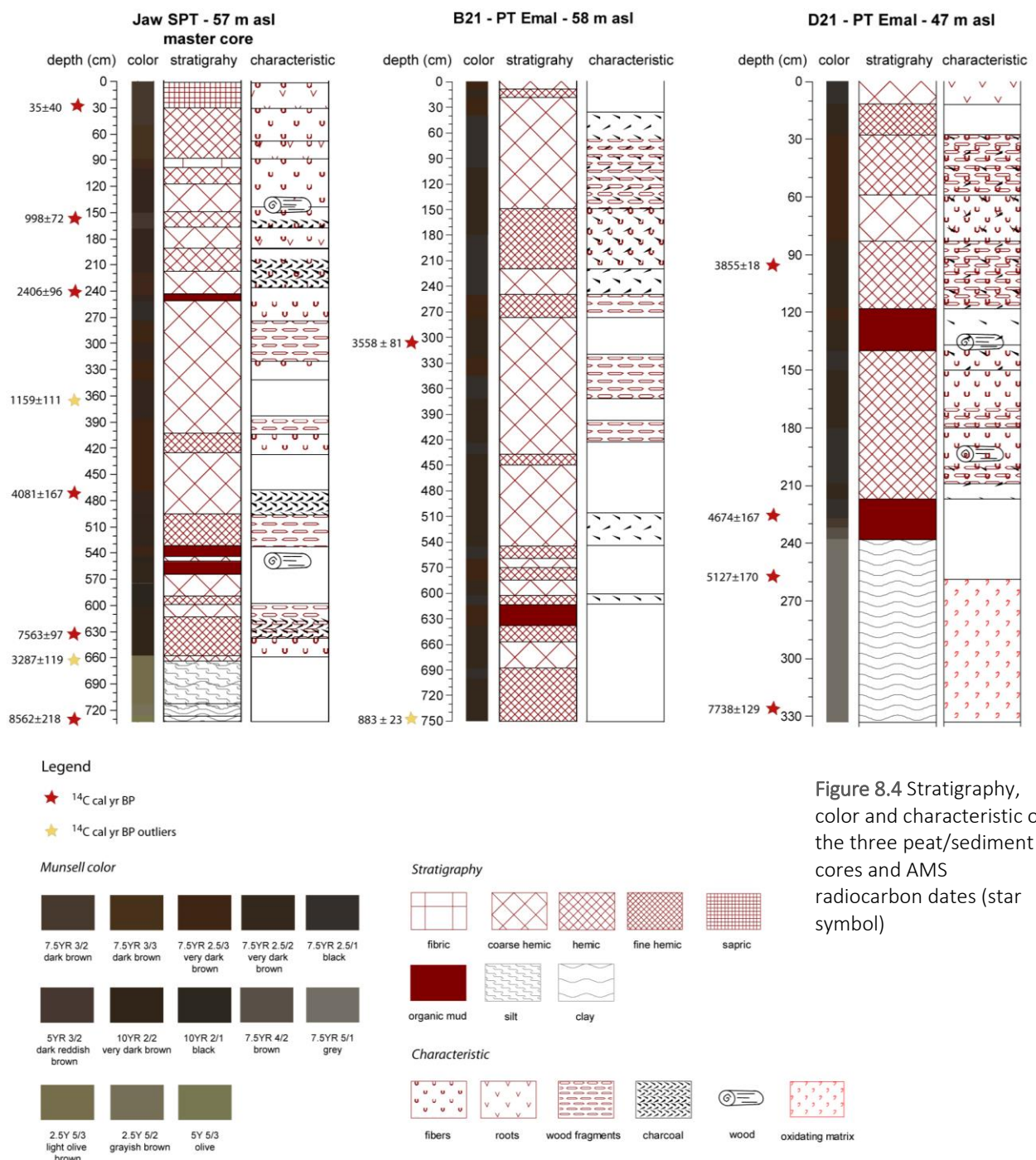

Figure 8.4 Stratigraphy, color and characteristic of the three peat/sediment cores and AMS radiocarbon dates (star symbol)

The bottom part of the eastern, more marginal, core D21 from 333 to 237 cm depth represents the mineral substrate consisting of gray clay with red mottles of oxidizing material upon exposure to air, suggesting permanent anaerobic conditions at the time of deposition (Figure 8.4). Between 237 to 227 cm core depth a layer of organic mud is found. Hemic peat with fibers and large wood remains characterizes the depth interval from 227 to 82 cm. A second organic mud layer is deposited from 140 to 118 cm depth. The top 82 cm of the core consists of fine and coarse hemic peat with fibers, wood fragments and roots.

The mineral substrate of the central core B21 could not be reached after 750 cm coring and deeper deposits could not be cored due to the difficulties in the field. The core mainly consists of coarse hemic peat and several layers of fine hemic peat. Macro-remains of wood fragments and fibers are observed from 424 to 68 cm depth. A layer of organic mud characterizes the core from 637 to 614 cm.

The bottom mineral substrate of the master core Jaw SPT consists of olive clay from 733 to 727 cm. From 727 to 713 the sediments change to grayish-brown clay with silt and from 713 to 664 cm the deposits consist of olive-brown silty clay. A transitional peaty-clay zone is found from 664 to 658 cm and marks the beginning of peat accumulation. Fine to coarse hemic peat with wood fragments and fibers characterizes the depths from 658 to 496 cm. From 496 to 219 cm depth the deposits consist mainly of coarse hemic peat with fibers and wood fragments. From 219 up to 30 cm depth the sediment are mainly fine hemic to coarse hemic peat with fibers and large wood remains. The top 30 cm of the master core Jaw SPT consists of sapric (highly decomposed) peat with roots. Layers of organic mud are found at 252–244, 545–532 and 565–550 cm core depth.

Chronology and peat accumulation rate of the master core Jaw SPT

Of fourteen samples dated with the AMS radiocarbon method, the three derived from woody material are stratigraphically inconsistent (Table 8.1). The most likely explanation is that the wood samples were in fact root material. As a consequence, two of the dates from the core Jaw SPT are treated as outliers in the age-depth model (Figure 8.5). The best age-depth model fitted to the dates is a smoothing spline, and reveals that the Jaw SPT record spans the last ca. 8600 cal yr BP. The peat accumulation starts at 664 cm core depth, ca. 7800 years ago. Peat accumulation rates per sample (mm/yr) are calculated based on the sample ages estimated from the age-depth model. Based on this estimate, the average peat accumulation rate from 664 cm to the top of the core is 0.9 mm/yr (min 0.5; max 1.6 mm/yr).

Table 8.1 Accelerator mass spectrometry radiocarbon dates from Air Hitam cores, calibrated age ranges at 95% confidence intervals. Calibration done with R script CLAM 2.2, calibration curve SHCal13.14C. Outliers are in light gray

| Core name	Depth (cm)	Material dated	^{14}C age BP	^{14}C age cal yr BP 2σ range	Laboratory code
Jaw SPT	29	organic bulk sediment	78 ± 38	35 ± 40	Erl-19242
Jaw SPT	157	charcoal	1140 ± 43	998 ± 72	Erl-18296
Jaw SPT	242	charcoal	2404 ± 42	2406 ± 96	Erl-18297
Jaw SPT	367	wood	1265 ± 55	1159 ± 111	Erl-18298
Jaw SPT	474	leaf and partially charred wood	3780 ± 52	4081 ± 167	Erl-18299
Jaw SPT	633	organic bulk sediment	6732 ± 54	7563 ± 97	Erl-19243
Jaw SPT	662	wood	3138 ± 50	3287 ± 119	Erl-18300
Jaw SPT	730	bulk of sediment	7770 ± 101	8562 ± 218	Erl-18301
B21	307	wood	3394 ± 19	3558 ± 81	NTUAMS-447
B21	746	wood	979 ± 5	883 ± 23	NTUAMS-326
D21	96	leaf and wood	3601 ± 18	3855 ± 18	NTUAMS-327
D21	226	wood	4187 ± 80	4674 ± 167	NTUAMS-328
D21	257	organic bulk sediment	4516 ± 54	5127± 170	NTUAMS-451
D21	326	organic bulk sediment	6955 ± 70	7738 ± 129	NTUAMS-330
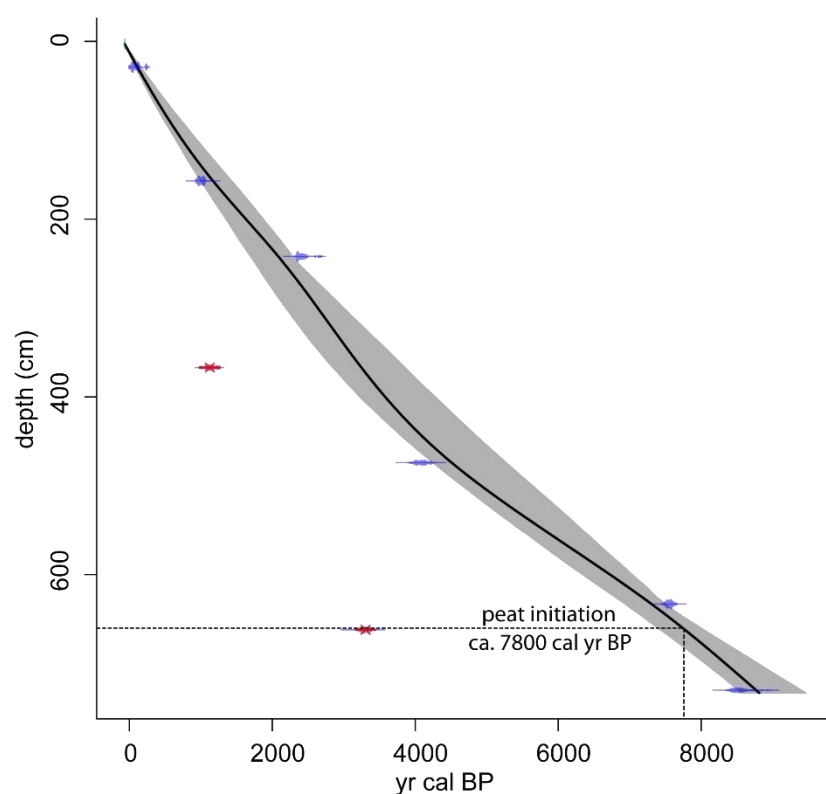
Figure 8.5 Age-depth profile of Jaw SPT core. Smoothing spline proved the best fitting model, with extrapolated basal points and surface (3 cm) age set at –62 years (AD 2012). Red symbols depict outliers. Dotted line indicates peat initiation depth (664 cm) and estimated age

peat initiation
ca. 7800 cal yr BP

Palynological results

The pollen diagram shows the paleovegetation groups and the most important pollen and spore taxa (Figure 8.6a and b). A total of 242 different pollen and spore types were identified including 87 rare still unknown types. The complete list of taxa groups can be found in the Table 8.2. Based on the constrained cluster analysis, 7 pollen and spore assemblage zones, Jaw-I to Jaw-VII are separated (Table 8.3). The most abundant group is represented by peat swamp taxa which oscillate between percentages of 3 and 86%. Within this group, particularly from zone Jaw-III to Jaw-VI, pollen grains of *Pandanus* dominate (average for these zones 35%). After *Pandanus*, the most abundant pollen taxa of peat swamp are *Parastemon* (average 7%), *Campnosperma* (average 5%), *Ilex* (average 4%), *Garcinia cuneifolia* type (average 3%) and *Austrobuxus* (average 2%). The group of dipterocarp-swamp mixed rainforest shows mostly high values in zones Jaw-I and Jaw-II (average for the zones 21%; up to 43%). The Dipterocarpaceae are represented in order of importance by *Hopea*, *Shorea*, *Dryobalanops* and *Dipterocarpus*. Beside dipterocarps, important taxa of the group are *Stemonurus* type (average 2%), *Cephalomappa* (average 2%) and *Stenochlaena palustris* (up to 2%). *Durio* (average 6%; up to 44%), *Blumeodendron* (average 2%; up to 23%) and *Calamus* (average 1%; up to 5%) are the most common of the freshwater swamp group, particularly in zone Jaw-II. Open and pioneer taxa are mostly represented by *Nephrolepis biserrata* (average 4%), *Ficus* type (average 3%), Elaeocarpaceae (average 3%), and *Macaranga/Mallotus* (average 2%; Table 8.3).

Testate amoebae results

A total of 78 testate amoebae taxa were recorded. The most common taxa are *Hyalosphenia subflava* (average 56% of the total sum; 21 size-based morphotypes), *Hyalosphenia minuta* (6.4%), *Cyclopyxis eurystoma parvula* (4.9%) and *Pyxidicula invisitata* (2.9%). For details on most important environmental groups and taxa see Figure 8.6c, d and Table 8.3.

Diversity indices and palynological compositional change or turnover

The palynological diversity indices (DIs) remain rather uniform along the core (average 6 $ET_{(10)}$). However, lower values are found in zone Jaw-IV when *Pandanus* pollen dominates the palynological assemblage (average 4 $ET_{(10)}$; Figure 8.6a). The testate amoeba DI varied more (Figure 8.6c) with $ET_{(28)}$ averaging by 6.

The first time-constrained DCCA axis shows that total turnover or PCC at Air Hitam for the past ca. 7800 cal yr BP is 2.1 SD (Figure 8.6a). Such a value is indicative of large differences between samples in the dataset (Hill and Gauch 1980), thus it indicates that important changes occurred in the community composition.

The range of compositional turnover changes strongly from zone Jaw-II to III and Jaw-VI to VII. A lesser-marked change is recorded at the passage from Jaw-V to VI.

Ordination results (PCA)

The PCA clearly separates the vegetation compositions of the zones Jaw-I and II from the remaining zones Jaw-IV to VI (axis 1; Figure 8.7), while vegetation compositions of the last anthropogenic zone Jaw-VII markedly separates from all other zones (Figure 8.7). The testate amoebae environmental groups and size classes, used as supplementary variables, explain 32% of the total variation. Water group and the larger size tests score negative on the second axis and are driven towards the first two zones Jaw-I and II. The opposite trend is observed for the smaller tests, green moss and soil groups, which score positive. The peat moss group and the medium size tests score negative on both axes.

Macro-charcoal and fire history

The macro-charcoal analysis of the Jaw SPT record, characterize the fire history at the site for the past 7800 years (Figure 8.6a). The local signal-to-noise index always exceeded 3, indicating a good separation between peak and non-peak values. The most recent peak-fire event is detected at ca. AD 2001 (−51 yr cal BP; 5 cm depth). The approximated synchronicity between the period of forest conversion (WWF Indonesia 2010) and the recent charcoal peak support the assumption that identified charcoal peaks detectmajor local fire episodes. The mean fire return interval (95% confident interval) for the whole record is 133 yr (106–160 yr) indicating that large fire episodes have been rare on the peat dome for the past 7800 years. However, changes in fire regime are visible from the fire frequencies at the centennial time scale (Figure 8.6a). A total of 60 fire episodes occurred locally during the past 7800 years, 1 fire episode failed to pass the 'Poisson minimum-count' screening test. Fire frequencies are low in the zones Jaw-I, II, V and VI. Higher fire frequencies are recorded in the zones Jaw-III, IV and VII. High magnitude peaks are found in the zones Jaw-II, V and VI (Table 8.3; Figure 8.6a).

Interpretation and discussion

Opposite to most temperate and boreal peatlands, the vegetation of natural lowland tropical freshwater swamp and peat swamp forests is dominated by trees (Page et al. 1999). The majority of the tree families of lowland dipterocarp forests of SE Asia can live and proliferate in freshwater swamp and peat swamp forests. However, increasing peat thickness leads to markedly different nutrient and oxygen availability and as a consequence vegetation composition (Anderson 1983). One important parameter is represented by flooding and riverine influence on the system. In general, the margin of the domes, where flooding occurs are richer in nutrients, while the more central thick peat deposits are poorer in nutrients, are acidic and have low dissolved oxygen (Anderson 1961; Esterle and Ferm 1994).

Table 8.2: Ecological groupings of plant taxa identified (pollen and spores)

dipterocarp-swamp mixed rainforest: taxa restricted or common in dry lowland rainforests and mixed swamp rainforests on shallow peat

Plant family	Pollen/spore taxon	Plant family	Pollen/spore taxon
Araceae	Araceae	Loranthaceae	Loranthaceae
Blechnaceae	*Stenochlaena palustris*	Malvaceae	Malvaceae
Cannabaceae	*Gironniera*	Malvaceae	subfam. Tilioideae
Celastraceae	*Lophopetalum*	Meliaceae	*Aglaia rubiginosa* type
Dipterocarpaceae	*Dipterocarpus* type	Pandanaceae	*Freycinetia*
Dipterocarpaceae	*Dryobalanops* type	Phyllanthaceae	*Bischofia*
Dipterocarpaceae	*Hopea* type	Primulaceae	*Ardisia*
Dipterocarpaceae	*Shorea* type	Proteaceae	Proteaceae
Ericaceae	Ericaceae	Rosaceae	Rosaceae
Euphorbiaceae	*Cephalomappa*	Sapindaceae	*Pometia*
Euphorbiaceae	*Homalanthus*	Stemonuraceae	*Stemonurus* type
Juglandaceae	*Engelhardia*	Vitaceae	cf. *Cissus*

freshwater swamp: taxa more abundant on peat deposit where river flooding occur (minerotrophic settings)

Plant family	Pollen/spore taxon	Plant family	Pollen/spore taxon
Amaryllidaceae	*Narcissus*	Malvaceae	*Durio*
Arecaceae	*Calamus*	Malvaceae	subfam. Sterculioideae
Arecaceae	*Oncosperma*	Menispermaceae	*Stephania*
Cannabaceae	*Celtis*	Phyllanthaceae	*Antidesma*
Euphorbiaceae	*Blumeodendron*	Polypodiaceae	*Microsorum heterocarpum* type
		Rhizophoraceae	Rhizophoraceae

peat swamp: taxa more abundant in ombrotrophic peat-swamp rainforests on thick peat deposits

Plant family	Pollen/spore taxon	Plant family	Pollen/spore taxon
Anacardiaceae	*Campnosperma*	Clusiaceae	*Garcinia cuneifolia* type
Anisophylleaceae	*Combretocarpus rotundatus*	Ebenaceae	*Diospyros*
Aquifoliaceae	*Ilex*	Pandanaceae	*Pandanus*
Araliaceae	Araliaceae	Picrodendraceae	*Austrobuxus nitidus*
Blechnaceae	*Stenochlaena areolaris*	Primulaceae	*Myrsine* type
Burseraceae	*Canarium*	Rubiaceae	*Randia*
Chrysobalanaceae	*Parastemon*	Sapotaceae	Sapotaceae

open-pioneer: if taxon increases in abundance, indicates early successional plant community and/or disturbed-open vegetation

Plant family	Pollen/spore taxon	Plant family	Pollen/spore taxon
Apiaceae	Apiaceae	Lamiaceae	Lamiaceae
Aspleniaceae	*Asplenium*	Lycopodiaceae	*Lycopodium cernuum*
Blechnaceae	*Blechnum*	Moraceae	*Ficus* type
Cannabaceae	*Trema*	Myrtaceae	Myrtaceae
Caryophyllaceae	Caryophyllaceae	Nephrolepidaceae	*Nephrolepis biserrata*
Convolvulaceae	*Ipomoea* type	Plantaginaceae	*Plantago*
Cyperaceae	Cyperaceae	Poaceae	Poaceae
Cyperaceae	*Thoracostachyum*	Polypodiaceae	*Microsorum unctatum* type
Elaeocarpaceae	Elaeocarpaceae	Sapotaceae	*Palaquium*
Euphorbiaceae	*Macaranga/Mallotus*	Selaginellaceae	*Selaginella*
Flagellariaceae	*Flagellaria*		

other: pollen-spore taxa with no habitat preference in peat-swamp rainforests

| Plant family	Pollen/spore taxon	Plant family	Pollen/spore taxon
Actinidiaceae	Actinidiaceae	Meliaceae	*Aglaia*
Adiantaceae	*Adiantum*	Meliaceae	Meliaceae
Anacardiaceae	Anacardiaceae	Menispermaceae	Menispermaceae
Apocynaceae	*Alstonia*	Moraceae/Urticaceae	Moraceae/Urticaceae
Apocynaceae	*Anodendron*	Myristicaceae	*Knema*
Apocynaceae	Apocynaceae	Myristicaceae	Myristicaceae
Arecaceae	*Areca*	Oleaceae	*Ligustrum*
Arecaceae	Arecaceae	Phyllanthaceae	*Aporosa*
Aristolochiaceae	Aristolochiaceae	Phyllanthaceae	*Phyllanthus*
Davalliaceae	*Davallia*	Phyllanthaceae	*Phyllanthus urinaria* type
Dennstaedtiaceae	Dennstaedtiaceae	Piperaceae	*Peperomia*
Dryopteridaceae	*Elaphoglossum*	Piperaceae	*Piper*
Euphorbiaceae	*Acalypha*	Polypodiaceae	Polypodiaceae
Hymenophyllaceae	Hymenophyllaceae	Polypodiaceae	*Selliguea*
Hypodematiaceae	*Leucostegia*	Pteridaceae	*Pteris*
Leguminosae	Leguminosae	Rubiaceae	*Hedyotis* type
Leguminosae	*Koompassia*	Rubiaceae	*Lasianthus*
Leguminosae	subfam. Mimosoideae	Rubiaceae	*Mussaenda*
Lindsaeaceae	Lindsaeaceae	Rubiaceae	*Nauclea*
Lycopodiaceae	*Huperzia*	Rubiaceae	Rubiaceae
Lycopodiaceae	Lycopodiaceae	Rutaceae	Rutaceae
Melastomataceae	Melastomataceae	Thelypteridaceae	Thelypteridaceae
		Thymelaeaceae	Thymelaeaceae

Table 8.3 Zone characteristics and results of the Jaw SPT core analyses

Palynological zone and core depth (cm)	Age (cal yr BP); average peat accumulation rate (mm/yr); peat thickness from zone bottom (cm)	Sediment and peat characteristics	Pollen zone characteristics	Testate amoeba environmental and size groups	Macro-charcoal and fire regime
Jaw-I; 700-570	- ca. 8340 to 6230 - 0.6 - peat starts from 664 cm	olive-brown silty-clay from 700 to 664 cm; peaty-clay from 664 to 658 cm; fine hemic to coarse hemic with wood fragments, fibers and charcoal	Relatively high proportion dipterocarp-swamp mixed rainforest taxa (average 22%), in particular Dipterocarpaceae (average 7%) and peat-swamp taxa (average 26%); pioneer-secondary taxa peak at 640 cm (58%; ca. 7400 cal yr BP); average low values of *Pandanus* (15%) with a peak at 580 cm (56%; ca. 6400 cal yr BP). Characteristic taxa of the zone are *Shorea* type, *Hopea* type, *Blumeodendron*, *Combretocarpus rotundatus*, *Stemonurus* type and *Stenochlaena palustris*	Peat moss group is dominant (average 66%). From 660 to 640 cm (ca. 7740–7420 cal yr BP) soil group has high values (up to 18%). Water group is present with relatively high values (average 2.6%). Medium to large tests (>75 µm) have high values (tot. average 50%)	Macro-charcoal concentration values and fire frequency low
Jaw-II; 570-470	- ca. 6230 to 4500 - 0.6 - 94	organic mud with wood to fine hemic with wood fragments to coarse hemic with charcoal	Decrease of Dipterocarpaceae and peat swamp taxa (average 4% and 16%) and increase of freshwater swamp taxa (average 34%), in particular with *Durio* (average 20%) and *Calamus* (average 4%). Characteristic taxa of the zone are *Durio, Calamus, Cephalomappa, Blumeodendron, Stemonurus* type	From 580 to 540 cm (ca. 6350–5620 yr BP) few or no tests are preserved. From 520 cm (ca. 5260 cal yr BP) peat moss group increases to the top of the zone (average 46%) and water group values are high (average 8%). Medium to large tests (>75 µm) dominate in this zone (tot. average 53%)	Fire frequency low. High peak in magnitude at 529 cm (9000 particles per peak/cm^2; ca. 5430 cal yr BP).
Jaw-III; 470-290	- ca. 4500 to 2500 - 1 - 194	coarse hemic with wood fragments and fibers; one layer of fine hemic with fibers from 404 to 427 cm	Dipterocarp-swamp mixed taxa and freshwater swamp taxa decrease (average 4% and 6%) while peat swamp taxa increase (average 55%). *Pandanus* proportions highly increase to the top of the zone (average 22%). Open/pioneer taxa increase (average 18%). Characteristic taxa of the zone are *Pandanus, Ilex, Garcinia cuneifolia* type, *Campnosperma, Parastemon, Austrobuxus nitidus, Stenochlaena areolaris* and *Nephrolepis biserrata*	Green moss group is found more frequently and dry indicators appear in this zone. Water group decreases (average 1%). Small tests (<75 µm) increase from 400 cm (ca. 3580 yr BP) to the end of the zone (average 51.2%)	Fire frequency high. No high magnitude peaks recorded
Jaw-IV; 290-230	- ca. 2500 to 1900 - 1 - 374	coarse hemic with fibers and wood fragments; one layer of organic mud from 244 to 252 cm	Freshwater swamp taxa decrease markedly. Pioneer-secondary forest taxa and *Pandanus* increase continue (average 21 and 65%). Characteristic taxa of the zone are *Pandanus, Macaranga* and *Ficus* type	Soil group increases markedly. Small tests (<75 µm) continue to be dominant (54%)	Fire frequency slightly lower compares to Jaw-III. No high magnitude peaks recorded
Jaw-V; 230-150	- ca. 1900 to 1100 - 0.9 - 434	hemic to coarse hemic with fibers, roots and charcoal	Pioneer-secondary forest taxa and *Pandanus* decrease (average 11% and 15%) while peat-swamp and freshwater swamp taxa increase (average 53% and 4%). Characteristic taxa of the zone are *Garcinia cuneifolia* type, *Durio, Parastemon, Stenochlaena palustris* and *Nephrolepis biserrata*	Green moss group increases (average 1.4%) while soil group disappears. Water group increases to the end of the zone (up to 11%). Larger tests, in particular the medium size class (75–105 µm) gradually increase (average 52%)	Fire frequency decreases to the top of the zone. Two high magnitude peaks recorded at 204 cm (9700 particles per peak/cm^2; ca. 1680 cal yr BP) and 165 cm (5000 particles per peak/cm^2; ca. 1250 cal yr BP)
Jaw-VI; 150-30	- ca. 1100 to 120 - 1.2 - 514	hemic to coarse hemic with fibers and roots	Peat-swamp taxa increase (average 70%) with *Pandanus* proportions increasing markedly (average 54%). Freshwater taxa decrease (average 1%). Characteristic taxa of the zone are *Pandanus, Koompassia, Campnosperma* and Myrtaceae	Water group decreases (average 1.5%). Starting from 100 cm (ca. 650 cal yr BP) soil group and green moss group increase (average 6% and 2%). Small tests (<75 µm) gradually increase (average 54%)	Fire frequency slowly increase to the top of the zone. Two high magnitude peaks recorded at 99 cm (4300 particles per peak/cm^2; ca. 650 cal yr BP) and 45 cm (1190 particles per peak/cm^2; ca. 240 cal yr BP)
Jaw-VII; 30-3	- modern - 1.5 - 634	sapric with roots	Pioneer-secondary forest taxa increase (average 20%) while *Pandanus* decrease (average 2%). Characteristic taxa of the zone are *Parastemon* and *Nephrolepis biserrata*	Green moss and soil group have high values (average 8% and 2%). Peat moss group decrease (average 26%). Small tests (<75 µm) dominate (average 84%)	Fire frequency high. No high magnitude peaks recorded

Vegetation composition changes depending on the adaptation of different taxa to these boundary conditions. Marginal and more central-high communities were classified accordingly to Anderson into species associations or "Phasic Communities" (Anderson 1961; Anderson 1964; Anderson 1983). Available studies on peat domes in SE Asia (Figure 8.1) reveal that highly diverse swamp forests, where trees are large (>45 m height, 2–4 m girth) and understory plants are dense, are found on the margins of the deposits, where peat is shallow and flooding by rivers common. In response to flooding and standing water, trees develop buttressed roots and pneumatophores. These marginal forest communities have been described by Esterle and Ferm (1994) as freshwater swamp or well flooded swamps. In the central more topographically elevated areas, river flooding does not occur and most of the nutrients available to plants come from rainfall (Esterle and Ferm 1994). As a consequence, vegetation is less diverse and the trees that can cope with the limited nutrients are reduced in their size. To these forests is given the descriptive name of 'pole forest' (Brady 1997) or Phasic Community IV after Anderson (1963). Understory and subcanopy plants are sparse and adapted to poor in nutrient-acidic conditions such as *Pandanus*. Anderson described a 'climax' phase in the Baram River area in Borneo which was named 'savanna like' due to the characteristic open appearance. This forest was dominated by *Combretocarpus rotundatus* and small shrubs with the ground covered with moss and pitcher plants (Phasic Community VI after Anderson 1963).

The following interpretation of the paleoenvironment and vegetation phases at the Air Hitam peat dome take into account the vegetation phasic community descriptions currently available from SE Asian peatlands (Figure 8.1).

Studies on pollen dispersal in peat and swamp rainforests are rare (e.g. Anderson and Muller 1975; Morley 1981) and cannot be applied outside the specific site studied due to the uniqueness of each deposition environment and vegetation structure. However, in general, the pollen signal is strongly localized due to the closed canopy of tropical peat swamp rainforests, which limits regional deposition by wind and the strategy of tropical rainforests plants which are mostly animal pollinated as noted by Anderson and Muller (1975). However, on the more marginal or shallower peat deposits, where river flooding occurs, pollen can be transported by water from a longer distance and incorporated into the peat deposit (Morley 1981). Additionally, gap-phases within the peat swamp rainforest caused by wind disturbance or even fires could lead to a misinterpretation of a disturbance/secondary vegetation phase as true vegetation succession phase. Although these effects on the pollen deposition cannot be ruled out for the Jaw SPT palynological reconstruction, the high presence of peat swamp taxa through the record, suggests the main component of the palynological assemblage was made of the plants growing on the Air Hitam peatland. In order to downscale the effects of short-lived gap phases and local overrepresentation, the results are interpreted according to average trends via locally weighted scatterplot smoothing (LOWESS) fitted to the sample values (black lines on Figure 8.6a and b).
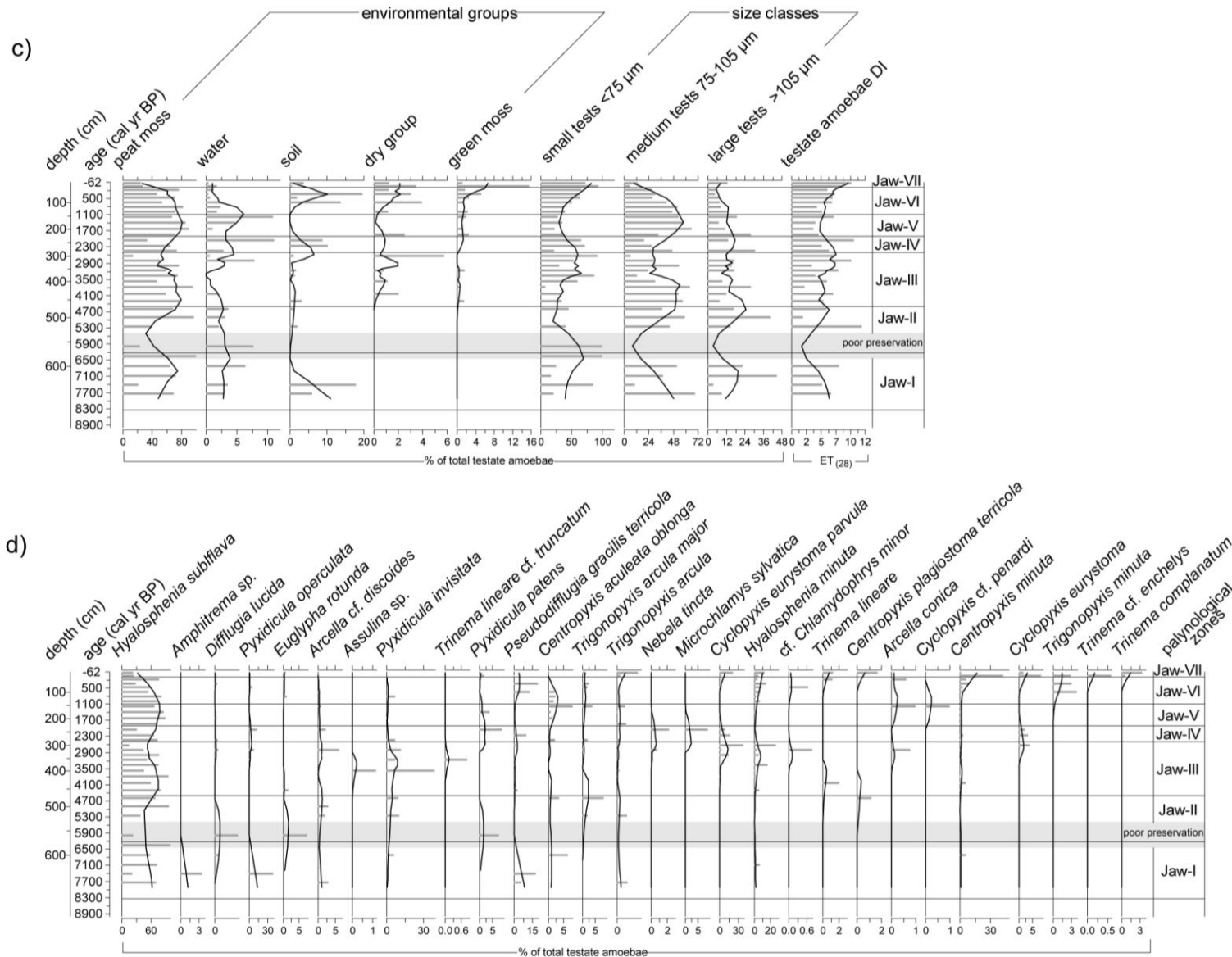
a)

b)
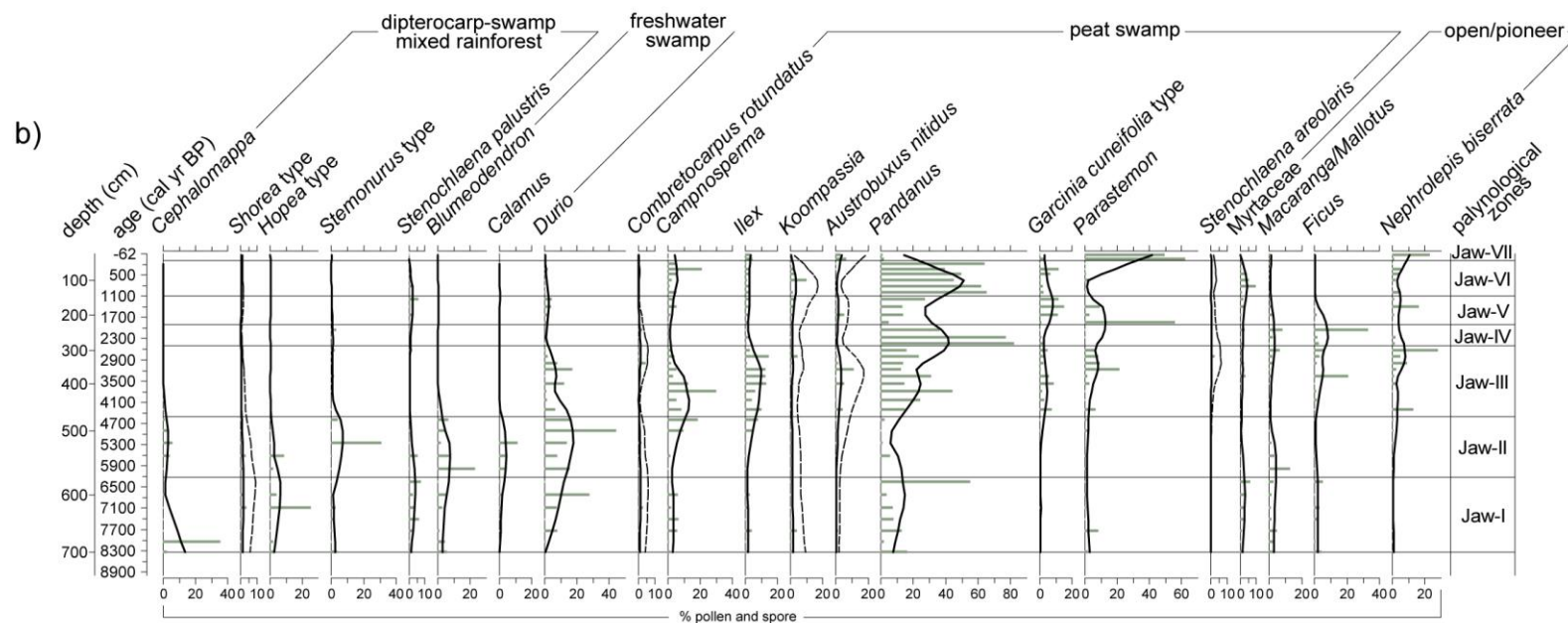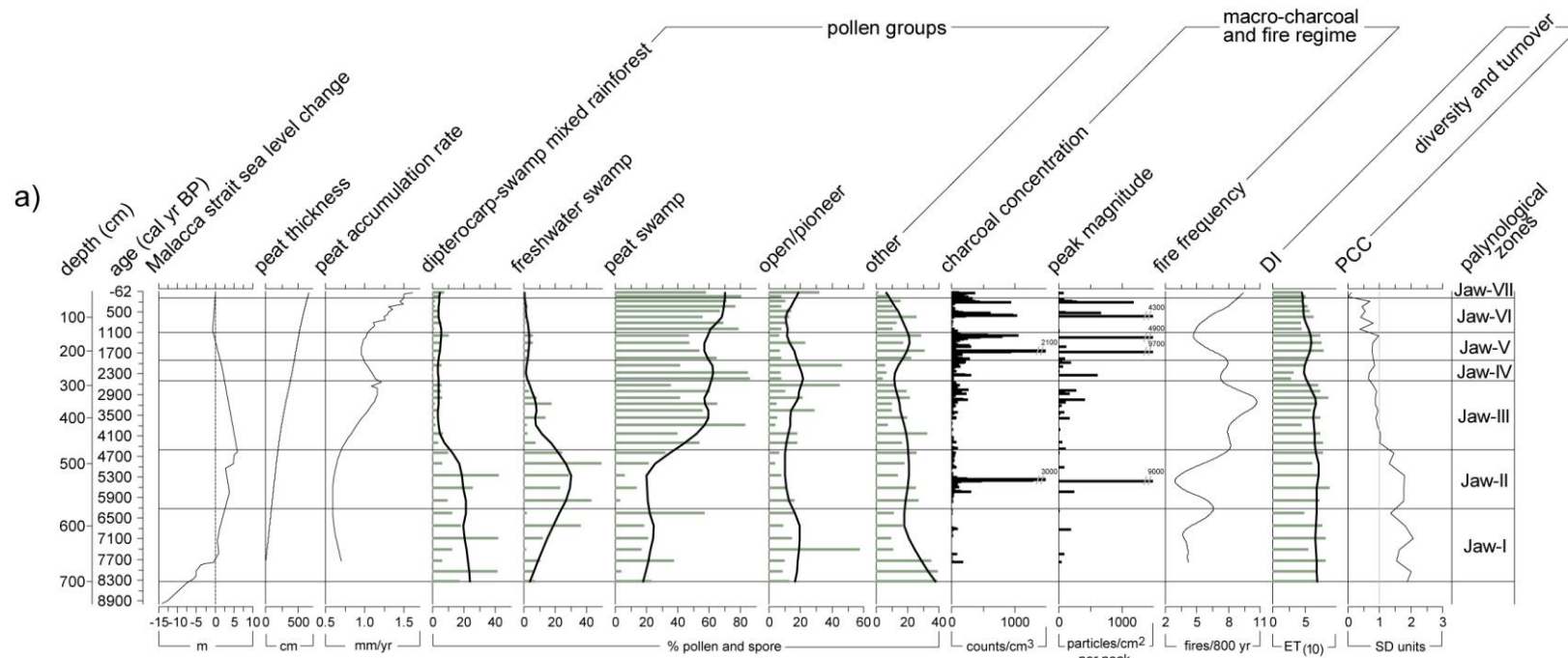
Figure 8.6 (also previous page) Summary diagram of the Jaw SPT core analyses with zonation based on constrained cluster analysis and palynological data. The black lines are the locally weighted scatterplot smoothing (LOWESS) fitted to the sample values (light green and gray lines) to highlight trends. a) Sea-level relative changes of the Malacca Strait (Geyh et al. 1979); peat thickness (cm); peat accumulation rate (mm/yr); pollen groups (expressed as percentage of total pollen and spores); macro-charcoal concentrations (counts/cm^3), peak magnitude (particles/cm^2 per peak) and fire frequency (number of fires per 800 years). The gray symbol "//" indicates a scale break used on large values on the x-axis to highlight smaller peaks; palynological diversity index (DI, ET$_{(10)}$) and palynological compositional change or turnover (PCC, standard deviation units). b) Most important and common pollen and spore taxa expressed as percentage of total pollen and spore sum. Dashed black lines are the exaggerated LOWESS lines (5x) for less abundant taxa. c) Testate amoeba environmental groups and size classes (expressed as percentage of the total sum of testate amoebae); testate amoebae diversity index (DI, ET$_{(28)}$). d) Most important and common testate amoebae taxa (expressed as percentage of the total sum of testate amoebae)

Environmental changes and palaeovegetation community phases

Peat forming vegetation phase Jaw-I (ca. 8340-6230 cal yr BP, 700-570 cm)

According to the basal radiocarbon age of the Jaw SPT core, peat started to accumulate on the northern section since at least 7800 years ago (640 cm core depth; Figure 8.5).

Testate amoebae were not preserved in the basal mineral deposits. As the peat started to accumulate, medium to large tests dominated the assemblages suggesting that thewater table was high. However, indicators of wet (for e.g. *Amphitrema* sp. and *Pyxidicula operculata*) and drier (for e.g. *Hyalosphenia minuta*) conditions were alternatively abundant, suggesting that large fluctuations also occurred in this phase. *Stenochlaena palustris* is a scrambling high-climbing epiphyte common in periodically inundated areas, where the lower parts of the rhizome are frequently submerged. The high frequency of spores of *Stenochlaena palustris* in this phase confirms that the site was periodically subject to flooding.

In this initial phase fine hemic peat indicates that the decomposition of the organic matter and humification were relatively high, while the presence of abundant wood fragments confirms that dense arborescent vegetation covered the area, possibly maintained by flooding, which can promote high biomass production but also high degree of decay. The vegetation as reconstructed from palynological analysis indicates a well-developed lowland rainforest covered the area with the high canopy constituted by Dipterocarpaceae and *Koompassia* trees. The sub-canopy was a mixed-swamp rainforest, with species such as *Blumeodendron*, *Combretocarpus rotundatus* and *Campnosperma* trees.

Freshwater swamp rainforest phase Jaw-II (ca. 6230-4500 cal yr BP, 570-470 cm)

During this period the palynological analysis reveals that the lowland Dipterocarpaceae-mixed rainforest taxa were slowly replaced by freshwater swamp taxa. The rattan *Calamus* is common in freshwater swamp forest and it has also been found in peat swamp forest and on alluvial flats (Anderson and Muller 1975). Currently 8 species (1 endemic) of *Durio* are found in Sumatra (Lemmens et al. 1995). These are small to large trees (up to 50–60 m) with buttresses and pneumatophores (knee roots) present in trees growing in marshy places. *Durio carinatus* was found to be the dominant tree in the mixed swamp rainforest in Riau in Sumatra (Anderson 1976). The increasing trend of, first *Calamus* (from 560 to 520 cm; from 6000 to 5300 cal yr BP) and then *Durio* pollen (from 520 to 480 cm; from 5300 to 4600 cal yr BP) suggests that a gradual change from freshwater to waterlogged-marshy conditions occurred in this phase. The poor preservation of testate amoebae in the lower part of this zone until ca. 5620 cal yr BP (540 cm) corresponds to a phase in which organic mud was deposited. Therefore, the reduced presence of testate amoebae might be a consequence of increasing river flooding and anoxic conditions, as also indicated by the peak of very small cosmopolitan less-sensitive (eurybiotic) and water demanding (hydrobiotic) testate amoebae taxa like *Eughlypha rotunda*, *Difflugia lucida* and *Pyxidicula patens*.

Following this initial phase, the testate amoebae assemblage is dominated by medium-large tests confirming that the water table was high, suggesting waterlogged conditions (Figure 8.6c). Parallel to the establishment of a swampy *Durio*-dominated vegetation, the reappearance of testate amoebae such as *Trigonopyxis arcula* and the peaks in *Trigonopyxis arcula major* and *Centropyxis plagiostoma terricola* indicate a gradual change from a system where most of water is coming from water flowing through mineral soil (minerotrophic settings) and waterlogged anoxic conditions to relatively drier rainfall-fed ombrotrophic acidic conditions.

Pole forest phase Jaw-III (ca. 4500-2500 cal yr BP, 470-290 cm)

A strong change in vegetation composition occurred around 4500 years ago as indicated by the marked shift in PCC. Freshwater swamp pollen taxa decreased abruptly indicating that the effects of river flooding became negligible in this phase. The strong increase of *Pandanus* pollen indicates thickets formed in the area as also confirmed by the occurrences of spores of *Stenochlaena areolaris*, a common epiphyte on *Pandanus* (Anderson and Muller 1975), so far reported only in the Philippines and Papua New Guinea (Chambers 2013) and found extinct in Borneo (Anderson and Muller 1975). The water group of testate amoebae decreased in this zone and the larger tests indicative of free standing water were gradually replaced by smaller sized taxa such as *Centropyxis minuta*, *Pseudodifflugia gracilis terricola* and *Hyalosphenia minuta* toward the top of the zone, confirming decreased water availability. Testate amoebae indicating drier conditions such as *Assulina* sp. and *Trinema lineare* cf. *truncatum* also occurred in this zone for the first time, suggesting that the water table was more regularly below the ground surface. A wet phase characterized the period from ca. 2900 to 2700 cal yr BP (330–310 cm depth) with the increase of testate amoebae taxa such as *Arcella* cf. *discoides* and *Arcella conica*, *Pyxidicula operculata*, *P. invisitata* and cf. *Chlamydophrys minor*.

The increase of *Campnosperma* pollen started at the end of the previous phase and the presence remained strong in this period. Of the two species of *Campnosperma* found in Sumatra, *Campnosperma coriaceum* possess pneumatophorous roots and occurs gregariously in peat swamp with oligotrophic water while *Campnosperma auriculatum* can be codominant in freshwater swamps in places regularly inundated by euthropic water (Endert 1920; Wyatt-Smith 1959). It is not possible to distinguish palynologically species of the genus *Campnosperma*. However, the decrease of freshwater swamp taxa in this phase suggests that the major contributor of the *Campnosperma* pollen was *Campnosperma coriaceum*. Dense thickets of the terrestrial fern *Nephrolepis biserrata* are often formed in completely open secondary forest especially after fires (Anderson 1963). The fire frequency increased markedly in this zone as a greater number of fire episodes are detected from the macro-charcoal peak detection analysis. The increase of spores from *Nephrolepis biserrata* also indicate that the structure of the forest changed to a more open as the spread of *Nephrolepis biserrata* is enhanced by open light conditions. The more open structure of the vegetation is further highlighted by the increasing presence of

Autrobuxus and *Ilex* which are common trees in more open stands of peat swamp rainforests (Phasic community IV after Anderson).

To the genus *Parastemon* belong tall trees which are common in peat swamp forests. *Parastemon urophyllus* is the only species reported in Sumatra and it is common in open scrub forest and in the Padang forest from Riau (notophyllous pole forest or Phasic community IV after Anderson). The increase of *Parastemon* pollen in this phase suggests that trees belonging to this genus were adapted to more open disturbed conditions. The relative high abundance of pollen of the giant trees of the genus *Koompassia* indicates that high canopy trees grew around the site.

The environmental conditions reconstructed in this phase resemble the "Padang forest" after Anderson (Phasic communities IV–VI, Anderson 1983) or pole forest of Brady (1997). However, dominant species composition differs from that found elsewhere with the most abundant taxa being *Pandanus*, *Ilex*, *Campnosperma*, *Garcinia cuneifolia* type, *Parastemon* and *Austrobuxus*.

In this phase the peat accumulation rate increased from an average 0.6 mm/yr to 1 mm/yr suggesting that the decomposition decreased and/or the accumulation of biomass increased (Figure 8.6a and Table 8.3). A decrease in decomposition rate is suggested from the characteristics of the peat which changes to less decomposed-coarser hemic material. The increase in peat accumulation rate closely followed *Pandanus* expansion while testate amoeba results suggest that the water table was on average lower and fluctuations smaller compared to the previous dipterocarp-mixed and freshwater swamp phases. *Pandanus* thickets have massive interlocking root system which are quite resistant to decay (Phillips and Bustin 1998) thus their expansion might have act as an important driver for increasing peat accumulation rate at Air Hitam. Additionally, since the river influence decreased as indicated by the reduced importance of freshwater swamp taxa, all water and nutrients were received from precipitation leading to a nutrient-poor environment (ombrotrophic settings). Such conditions might have reduced the efficiency of decomposer organisms leading to decrease in humification as also indicated by peat composition.

"Savanna" phase Jaw-IV (ca. 2500-1900 cal yr BP, 290-230 cm)

In this phase *Pandanus* thickets dominated and the lower palynological diversity (DI) suggests fewer species were present at the site. High values of *Ficus* type and pioneer taxa such as *Macaranga*, suggest a strongly opened vegetation. Opposite to the model of peat development in the Batang Hari coastal peat dome (Esterle and Ferm 1994), this phase of *Pandanus* thickets are not accompanied by ponds on the surface as indicated by the dominance of small tests. Fire episodes continue to be common as indicated by high fire frequency. The relatively dry and open connotation of this phase resembles the "savanna" phase observed in the Baram area, on thick peat in central areas and disturbed patches (Anderson 1963). However, the vegetation composition reconstructed from Air Hitam is markedly different in respect to species composition. While in Baram, *Combretocarpus rotundatus* dominates the

communities, palynological results from this phase, suggest that *Pandanus*, *Ficus* and *Macaranga* were the most important taxa at Air Hitam. The increase of green moss group and dry indicator taxa like *Nebela tincta* in the testate amoeba assemblage further confirm this reconstruction. *Cyclopyxis eurystoma* and *C. eurystoma parvula* started to increase in the previous zone and are characteristic of the assemblage in this phase. They are small eurybiotic taxa and their presence in this zone suggests conditions were less suitable for other taxa more sensitive to environmental disturbance.

Peat-swamp rainforest phase Jaw-V (ca. 1900-1100 cal yr BP, 230-150 cm)

From ca. 1900 cal yr BP, *Pandanus* and pioneers started to decrease while freshwater swamp and peat swamp taxa increased, suggesting the recovery of peat swamp vegetation at the study site. The peat accumulation slightly decreased, likely as a consequence of increasing decomposition as suggested by the occurrence of finer hemic peat layers in this phase. The increase in the medium to large-size tests and hydrobiotic taxa (e.g. *Centropyxis aculeata oblonga*, *Arcella conica*, *Cyclopyxis* cf. *penardi*) indicate increase in water availability. It is possible that the water channel, which today passes close to the Jaw SPT, formed in this phase. The flooding of the area occurred as is also indicated by the increased frequency of *Stenochlaena palustris*. The fire frequency decreased markedly. However, several fire peaks of high magnitude were recorded in this phase, suggesting that fire episodes were accompanied by a local increase in biomass burning. This could either indicate that the vegetation which characterized this phase was more sensitive to fire and/or fires affected larger areas. However, the increase in fire magnitude might also be linked to paleoclimatic conditions. Numerous paleorecords recording ENSO amplitude and frequency in the Holocene, indicate that the period between 2000 and 1200 cal yr BP might have been characterized by particularly severe El Niño episodes (Dommain et al. 2011 and reference therein). These strong paleo-El Niños and drought associated, would have increased fire susceptibility of the moisture sensitive peat swamp taxa, thus enhancing and magnifying the effects on the vegetation when fires occurred.

Pole forest phase Jaw-VI (ca. 1100-120 cal yr BP, 150-30 cm)

A marked increase in *Pandanus* characterizes this phase. Small-size tests and indicators of drier condition such as *Hyalosphenia minuta* and *Trigonopyxis minuta*, increased to the top of the zone. The reconstruction suggests that water table fluctuations were low and the ecological conditions were similar to those reconstructed from the zones Jaw-III and IV. As in these zones, again peat accumulation rates increase from average 0.9 mm/yr to 1.2 mm/yr, parallel to the expansion of *Pandanus*. However, despite that *Pandanus* dominated this phase, diversity remained relatively high (average 5 $Et_{(10)}$), suggesting that a well-developed peat swamp rainforest surrounded the site with *Koompassia*, *Shorea* and *Hopea* as emergent trees and *Campnosperma*, *Ilex* and *Garcinia cuneifolia* as understory trees. Fire frequencies increased to the top of the zone and several peaks of high magnitude were recorded.

Modern anthropogenic disturbance Jaw-VII (modern, 30-3 cm)

The top 30 cm of the Jaw SPT core are markedly different from the remaining deposit (Figs. 6 and 7). Highly decomposed sapric peat with root remains characterize this layer and the peat accumulation rate is high. However, as pointed out by Dommain et al. (2011) the rates of peat accumulation on the surface peat layer (acrotelm) are biased by the greater presence of decomposing organic material and living biomass and might not represent a realistic estimate of the recent rates of peat accumulation. The presence of this top layer of sapric peat might conform to the model of Esterle and Ferm (1994) in which they assume intense degradation of surface material and preservation of root biomass beneath the surface of the peat. As peat accumulate, the stratigraphic sequence is a function of vegetation (particularly roots) and humification at the time of the burial of peat underneath the water table (Esterle and Ferm 1994).

The greatest vegetation turnover as reconstructed from the palynological analysis occurred in this phase. *Pandanus* and emergent trees such as *Hopea*, *Shorea* and *Koompassia* disappear and *Parastemon* and thickets of *Nephrolepis biserrata* dominates at the site, also observed in the field during the coring. The strong human impact in this modern phase is further highlighted by the increase in fire frequency. It remains uncertain why *Parastemon* was not subjected to logging while virtually all other large trees were removed. It might be related to the fact that the wood from this species is hard to use because of the silica content. The testate amoebae assemblage shows a marked change from peat moss to green moss and soil dominating communities and small-size tests dominate in this phase. The high presence of eurybiotic species such as *Trinema* spp., *Cyclopyxis eurystoma* and *Cyclopyxis eurystoma parvula*, indicates that environmental conditions were less favorable for testate amoeba species compare to the previous phases. While most of the Air Hitam peat dome was converted to oil palm plantations following intense logging in the 1990s, this small portion of land was not converted. However, the logging and the creation of a net of drainage channels on the dome to reduce water table height strongly changed the structure and composition of the vegetation and therefore conditions of the peat at the site.

Similarities/dissimilarities of the vegetation phases

With the exception of the last modern phase, the results from the PCA including all taxa, generally confirm a clear separation between the vegetation composition in the first two phases when minerotrophic conditions persisted and the following ombrotrophic phases (Figure 8.7). As indicated by the large shift in PCC from Jaw-II to Jaw-III, species composition changed overall. General observations on tropical SE Asian peatlands, indicate that freshwater swamp and peat swamp forests on shallow peat are generally more diverse, compare to the most advanced peat swamp communities on thick peat deposits. Opposite to that, the DI index estimated from the Jaw SPT core, suggests diversity did not change from minerotrophic to the ombrotrophic settings. The PCA results show that zones Jaw-III/pole

forest and Jaw-V/peat swamp phases were highly variable or dynamic and were more closely connected with the peat forming phase Jaw-I. The greatest distance, suggesting more marked differences in vegetation composition, is observed between the Jaw-VI/pole forest and Jaw-IV/"savanna" phases where *Pandanus* dominated with the freshwater swamp composition which characterized Jaw-II.
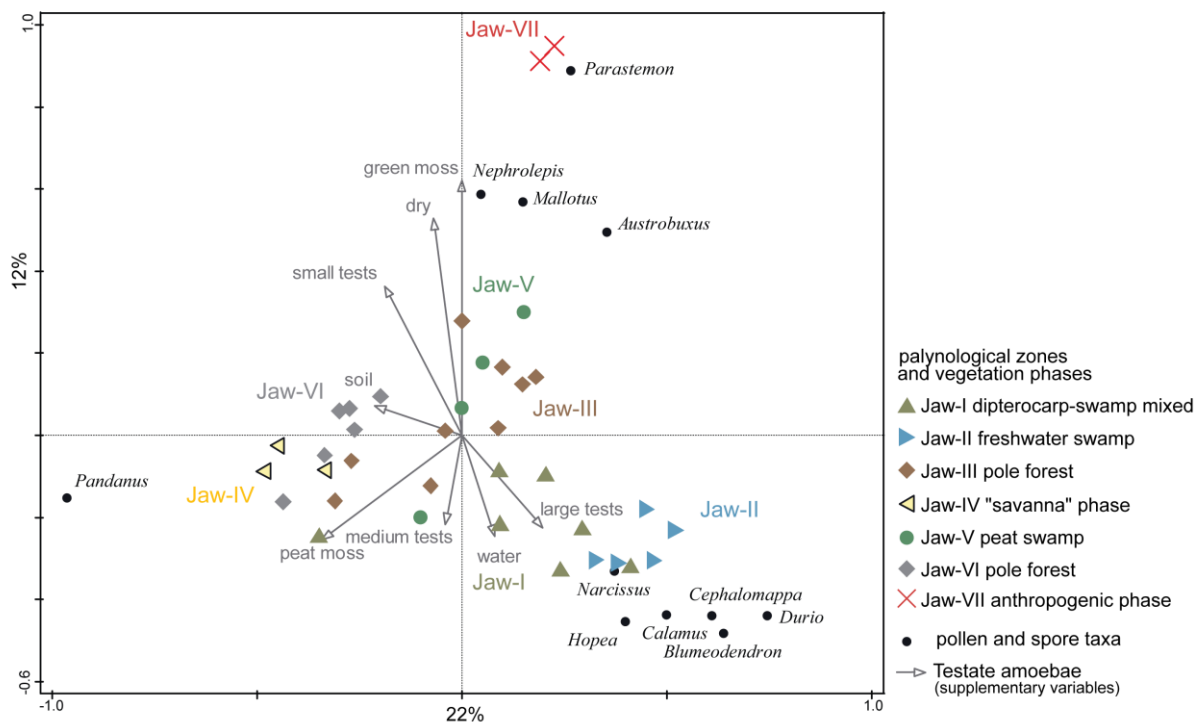

Figure 8.7 Samples scatterplot of the principal component analysis (PCA) of all percentage data of identified pollen and spore taxa. Percentages calculated on the total sum (*Pandanus* excluded) of all taxa square root transformed. First and second axes cumulative explained 34% of the variation in pollen and spore taxa data. Differently marked palynological zones correspond to different vegetation phases. Gray arrows correspond to the variation of testate amoeba environmental groups (water, peat moss, green moss, soil and indicators of drier conditions) and size classes (large, medium and small) used as supplementary variables. Only most important taxa are shown

Water table and peat accumulation rate changes

Initially, from ca. 7800 to 4500 cal yr BP, minerotrophic conditions characterized the watershed in the Air Hitam area. Water table fluctuations were high and flooding of the area occurred frequently as indicated by the dominance of medium to large size tests and the alternative domination of wet and drier taxa. A well developed Dipterocarpaceae mixed-swamp rainforest covered the area and biomass input must have been very high. However, this was outbalanced by a high decomposition rate and peat accumulation remained low for the first 4500 years of the record (average 0.6 mm/yr). The change to ombrogenous conditions became apparent from ca. 4500 cal yr BP onwards, when a marked change in vegetation occurred and *Pandanus* thickets colonized the site, leading to an increase in peat accumulation rate (average 1 mm/yr) likely as a consequence of increasing root biomass associated with the formation of *Pandanus* thickets and decrease in decomposition under oligotrophic conditions.

Testate amoebae taxa tolerant of drier conditions appeared and the small size tests slowly replaced larger tests confirming decreased permanent water availability. The peat accumulation rate remained high to the top of the core although it slightly decreased from ca. 1900 to 1100 yr cal BP (Jaw-V), when *Pandanus* decreased sharply and a mixed-swamp rainforest, resembling the initial minerotrophic phases became established. As observed in Jaw-I and II, the increase in water table fluctuations and possibly river flooding, as indicated by the increase in the testate amoebae water group and freshwater swamp pollen taxa, were accompanied by higher decomposition and hence lower peat accumulation rate.

Although in general the water table closely follows the peat dome surface, Brady (1997) found that the water table levels monitored on different peat thickness in Sumatra exhibited patterns of declining average level and fluctuations with increasing peat depth, with the lowest average water table depth occurring in the thickest peat site. The study of Brady (1997) suggests that on thick peat deposits, peat accumulates more because of the larger amount of fresh roots and declining rates of litter decay rather than from rising water levels and permanent stagnant conditions. This hypothesis was supported by the study of Grady et al. (1993). They analyzed the maceral content of peat from different layers of a deep deposit near Siak Sri Indrapura, Riau (Sumatra), and used the increase in fungal degradation of plant cells in peat as an evidence of higher oxygen levels in peat during degradation. The fungal degradation results suggest that the peat found in deep deposits is more aerobic than that found in thinner deposits. Thus it is thought peat accumulation in the thicker peat deposits of Riau are associated with drier surface conditions and changes in species composition rather than with a higher water table. In this model, the declining influence of river flooding along the gradient plays an important role. Vertical sequences of peat analyzed from deposits in Sumatra (Esterle and Ferm 1994) and in Kalimantan (Dehmer 1993; Moore and Hilbert 1992) suggest that increasing peat accumulation with thickness of the deposits was associated with reduced flooding in addition to a lower mean water table level, declining resource quality of plant inputs, and an increase in the root biomass.

The results from the Jaw SPT core fits the model of Brady (1997) and Grady et al. (1993) as the increase in accumulation rate was likely a consequence of belowground biomass increase (in particular *Pandanus* roots) and decreasing decomposition characteristic of the oligotrophic conditions on ombrotrophic peatlands. This suggests a strong link between vegetation communities and peat accumulation rate changes over time at Air Hitam, as increase in peat accumulation rate closely followed increasing representation of *Pandanus* at the site.

Drivers of change: sea-level, climate, fire and human impact

Peat accumulation started at the Air Hitam watershed after 7800 years ago corresponding to the mid-Holocene maximum in regional rainfall and the lowering and stabilization of the rate of sea-level rise (Geyh et al. 1979; Figure 8.6a). As a consequence of these changing conditions, the regional water table

in the islands of Sunda rose, causing flooding and paludification of the more depressed areas (Dommain et al. 2014). Initially, the peatland was influenced by the river flooding, which caused a high rate of decomposition. As a consequence, opposite to what observed in Central Kalimantan (Brünig 1996), the presence of a highly biomass productive dipterocarp-mixed swamp rainforest did not induce high peat accumulation rate in this initial phase. Starting from ca. 5000 cal yr BP, sea level started to decrease slightly due to hydro-isostatic adjustment of the Sunda shelf (Horton et al. 2005; Steinke et al. 2003; Dommain et al. 2011) causing the decrease of the regional water table in the inland, likely affecting the interfluvial peat domes. Additionally, peat thickness at Air Hitam likely reached a level above the influence of rivers discharge, causing a change towards a purely rainfall fed oligotrophic system. From ca. 4500 year ago, a more open pole forest colonized the area, with *Pandanus* thickets covering the site. Parallel to this, local fire episodes became more frequent, corresponding to the increase in ENSO variability in Indonesia as indicated by several paleoclimatic records (Haberle et al. 2001; Donders et al. 2008). The burned open patched areas were colonized by the terrestrial fern *Nephrolepis bisserrata* and *Parastemon* trees. A stronger disturbance signal was found from ca. 2500 to 1900 cal yr BP. The diversity decreased markedly and pioneer taxa like *Macaranga* codominated the site together with *Pandanus* and *Ficus*. It remained uncertain if this phase represented an advanced phase of the catena as found by Anderson (Phasic Community VI; Anderson 1963) or a long-lived disturbance phase following fire burning of the canopy. Despite those major changes in the vegetation cover, peat accumulation rates remained high and decreased only slightly from ca. 1900 cal yr BP when the paleoenvironmental reconstructions indicate that a renewed mixed-swamp rainforest expanded on the site. According to Dommain et al. (2011), the occurrencesof strong paleo-El Niños starting from ca. 2000 cal yr BP, caused peat accumulation in many Central Kalimantan peat domes to cease. While fire frequency decreased overall in this phase, major fire peaks were recorded possibly in connection with the occurrences of strong El Niño episodes.

Following this temporary mix-swamp rainforest phase dominated by *Garcinia cuneifolia* type and *Durio*, a low pole forest reestablished with abundant *Pandanus* thickets and the peat accumulation rate increased again from ca. 1100 cal yr BP onwards. The general increase in peat accumulation during the relatively drier phases and of increasing ENSO variability indicate that the peatland of Air Hitam has been potentially an effective carbon sequestering ombrotrophic peatland against the millennial scale climatic and environmental variability.

Despite the resolution for the top modern 30 cm of the core being low (2 samples analyzed for paleoenvironmental reconstructions), major transformation of the landscape could be detected. *Pandanus* species and the majority of the trees were removed from the site and *Parastemon* and *Nephrolepis biserrata* dominate the assemblages. The general net greenhouse gases balance of tropical peatlands is affected by the conversion to oil palm plantations as the carbon stored is lost during the

conversion and subsequently for management processes when water table is kept artificially belowground (Verwer et al. 2008; Hooijer et al. 2010; Murdiyarso et al. 2010; Page et al. 2011; Kurnianto et al. 2015). The consequences of clearance and drainage of the Air Hitam peat dome are not predictable without long-term monitoring of the peat deposit. However, it is likely that the accumulated carbon will be transferred to the atmosphere due to exposure to aerobic conditions. The potential value as carbon sink of this area might soon be lost unless a better management of the watershed is applied.

Conclusions

In general, the domed shape of a peatland is derived from the combination of differential rates of biomass production, decomposition and preservation (Tie and Esterle 1992) and the long term function of a peatland as a carbon sink depends largely on the regional and local environmental settings for much of its history. The multi-proxy paleoecological and paleoenvironmental results from the core Jaw SPT of the Air Hitam peatland increase our knowledge on what were the most important factors driving the development of these ecosystems on a time scale of thousands of years, thus contributing to the understanding of how ecosystem dynamics and fire are linked to peat development through time.

As for several other modern peatlands in SE Asia (Figure 8.1), peat initiated at the Air Hitam peat dome during the mid-Holocene rainfall maxima in combination with a slowing of the pace of sea level rise, which likely raised the water table in these inland settings leading to paludification (Griffiths et al. 2009; Partin et al. 2007; Dommain et al. 2011). Between the fluvial levees, in the depression that characterizes the Air Hitam area, peat started to accumulate. The peatland at Air Hitam was initially fed by fluvial run-off and the vegetation gradually changed from mixed Dipterocarpaceae swamp to marshy freshwater swamp communities dominated by *Durio* trees at around 5300 years ago. The change from a minerotrophic to ombrotrophic setting occurred during the late Holocene about 4500 years ago, when peat growth reached a level above river influence leading to a marked change in vegetation community composition. A pole forest established with *Pandanus* thickets colonizing the area and freshwater swamp taxa decreased.

The occurrence of fire in humid SE Asian peat swamp forests is often reported as an anomaly, due to the fact that pristine tropical peat swamps are permanently wet (Rieley and Ahmad-Shah 1996). Both historical records and charcoal reconstructed from paleo archives have demonstrated that fires do occur in the humid tropical forests but they are rare with return intervals of hundreds to thousands of years (Cochrane 2003). The results of the Jaw SPT core confirm for Air Hitam, fire return interval has been in the range of a hundred years. However, the correspondence between the increase in fire magnitude and frequency and ENSO onset in the late Holocene, suggests that the fire regime changed at the centennial time scale following regional climatic variability. Interestingly, the drier ombrotrophic *Pandanus*-dominated phases in the late Holocene were not accompanied by a decrease in peat

accumulation rates but rather increase. *Pandanus* thickets might be one of the most important drivers of peat accumulation at Air Hitam, downscaling the effects of the occurrence of more frequent regional droughts and less-suitable for peat development drier surface conditions. This might suggest that, under natural conditions, long-term climatic variability does not affect peatland development in the inland of Central Sumatra. However, recent changes in vegetation cover and hydrological settings following agricultural conversion might make the Air Hitam peatland more sensitive to drought, and fires might increase as observed elsewhere already in SE Asia peatlands (van der Werf et al. 2008). If extreme climatic events should increase in the future, longer or more severe drought might cause the loss of 7800 years of carbon stored at Air Hitam. A better management of the watershed and, in particular, restoring measurements act at reestablishing the natural vegetation cover, might be fundamental in order for the Air Hitam peatland to maintain its state as carbon sink in time.

In conclusion, the paleoenvironmental and palynological results from the Jaw SPT core further confirm the value of the inland peat system in Air Hitam and highlight the importance of conducting paleoenvironmental and charcoal analyses in order better reconstruct ecosystem-climate interactions in the tropical peatlands of SE Asia.

Acknowledgments

CHAPTER 9 - paper 5

Effects of long-term climate and environmental variability on the development of inland tropical peat swamp ecosystems in Jambi Province, Sumatra, Indonesia

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Abstract

We studied the area of Air Hitam, a tropical peatland in the Jambi province (Sumatra) in order to investigate the long-term effects of environmental and climate variability on the development of the peatland and on its carbon accumulation capacity. We used multi-proxy palaeoenvironmental and palaeoecological analyses on a 733 cm-long and ca. 7800 cal yr BP old peat core. The combined results of pollen, spores, testate amoebae, peat characteristics, ^{15}N natural abundance and macro-charcoal analyses reveal that changes in peat and long-term carbon accumulation rates were mostly a function of vegetation phases and sea-level changes. Higher rates of peat and long-term carbon accumulation were found to correlate with *Pandanus* expansion under ombrotrophic settings. The occurrences of several centennial-scale period of vegetation disturbance were reconstructed from pioneer and open pollen taxa and peat ^{15}N natural abundance signatures. Increase in number of disturbance phases of forest community was paralleled with increase in fire frequency, as reconstructed from the macro-charcoal analysis, which in turn followed the onset of modern ENSO amplitude during the late Holocene. Although the increase in fire occurrences caused disturbance of the vegetation communities for ca. 7800 years, these ecosystems showed resilience to these relatively short-lived disturbance events.

Introduction

Large areas of tropical peatlands are found in the lowlands of SE Asia where they provide important ecosystem functions (Dommain et al. 2011; Page et al. 2011; **Figure 9.1a**). Despite their importance as major carbon sinks, peat swamp forests are one of the ecosystems most threatened by deforestation in Indonesia (Miettinen et al. 2011). Beside land-use change, it has been hypothesized that climate change and particularly, the increase in extremes rainfall anomalies and drought episodes, might disrupt the delicate equilibrium of these ecosystems in the future (Dommain et al. 2011). Palaeoecological multi-proxy analysis of peat cores provides a longer temporal view of the dynamics of these ecosystems and can help understanding what are the most important drivers affecting their develop and functionality in time. In this study we use vegetation phases as reconstructed from pollen and spore data, palaeohydrology and water table fluctuations inferred from testate amoebae and local fire regime history as reconstructed from the macro-charcoal peak detection analysis from the peat core of Jaw SPT (**Figure 9.1b**), in the province of Jambi, Sumatra, Indonesia (Biagioni et al. 2015b; Chapter 8). Additionally, the long-term carbon accumulation rate (LORCA) are reconstructed using isotope analysis from the same core and peat characteristics are used to infer the rate of decomposition of peat in relation to vegetation phases and LORCA. [15]N natural abundance and pollen and spore pioneer relative abundance are used as proxies for ecosystem disturbance. We use these results and we plot them against external potential drivers which include climate and sea-level change with the final aim of

assessing the relationship between peatland development, palaeoclimate and palaeoenvironment at the centennial to millennial time scale.

Figure 9.1 a) Map of Sundaland, showing the study area (yellow square) in the Province of Jambi (red line) in Sumatra. Grey areas represent the Sunda shelf. Black areas represent peatlands (data source for peat FAO Harmonized Soil Map of the World, Histosols and Gleysols layers only); b) location of the Jaw SPT core (yellow circle) in the Air Hitam peat dome (dark grey area)
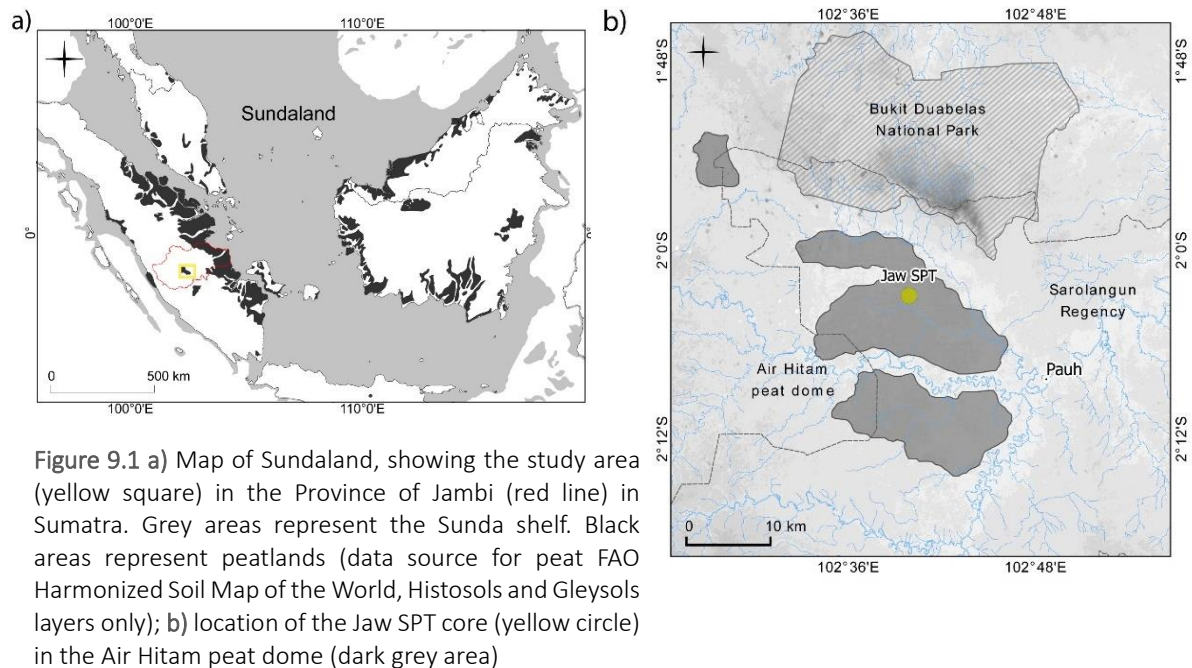
Modern settings and palaeo-environmental dynamics

Climatically, the eastern part of the Jambi province in Sumatra, Indonesia, experiences a wet tropical climate with mean annual precipitation ranging between 2400 and 2900 mm and annual temperatures average of 27˚C (WorldClim - Global Climate Data, http://www.worldclim.org/). Seasonality of rainfall is not marked, although the months from May to September are characteristically drier due to the southeast monsoon (Aldrian and Susanto 2003). The inter-annual scale rainfall patterns change when the El Niño Southern Oscillation (ENSO) and the Indian Ocean Dipole (IOD) occur. The El Niño warm phase of ENSO and the positive IOD cause decreasing rainfall, while the opposite trend with increasing rainfall characterizes the cold phase La Niña of ENSO and the negative IOD phase (Philander 1990; Webster et al. 1999; Cane 2005).

The results from the palaeoecological reconstruction of the Jaw SPT core (2.047685°S 102.665362°E; Biagioni et al. 2015b; Chapter 8) indicates that peat started to accumulate in the Air Hitam depression at least 7800 years ago, and the vegetation reconstruction from pollen and spores suggests that until ca. 4500 years ago the vegetation was dominated by dipterocarp-mixed swamp rainforest taxa. Following, a change occurred towards more specialized vegetation adapted to live under oligotrophic conditions (Biagioni et al. 2015b; Chapter 8).

Methods

In this study we use the palaeoenvironmental and palaeoecological proxies including pollen and spores, testate amoebae and macro-charcoal results from the Jaw SPT core. Methods applied and data analyses for these proxies are described in details by Biagioni et al. (2015b) and in Chapter 8.

Using the classification systems for SE Asian tropical peat (Esterle and Ferm 1994; Wüst et al. 2003), we infer level of decomposition and humification from the peat attributes and plant composition (Cameron et al. 1989). According to the morphology of the botanical components, the peat is classified in fibric, hemic and sapric. The hemic peat is further divided in to coarse and fine (Esterle and Ferm 1994; Wüst et al. 2003). The degree of decomposition or humification increase from fibric to sapric and it is a function of the biomass characteristics of the vegetation communities and the hydrological settings.

Pollen and spore taxa percentage results are grouped into two main groups based on the available accounts for the vegetation communities and the ecological descriptions of taxa currently available from SE Asian and Sumatran peat-swamp rainforests (e.g. Anderson 1963; Sabiham and Furukawa 1986; Cameron et al. 1989; Brady 1997; Page et al. 1999; Cole 2012; Gunawan et al. 2012; Cole et al. 2015). These groups are Dipterocarp-mixed swamp, which includes taxa commonly found in shallow marginal peat deposits and on fresh swamp rainforests in SE Asia and peat swamp taxa which includes taxa more representative of the communities growing on thick peat deposits (Table 8.2).

Similarly, to infer average water table level and fluctuations, testate amoebae with similar habitat preferences were classified into environmental groups. Testate amoebae are unicellular protists, which can be found commonly in wet environments such as lakes, rivers, mosses, soils and estuaries all around the world (Mitchell et al. 2008). Their responses to the major ecological gradients in peatlands have been established (e.g. Chardez 1965; Bobrov et al. 1999; Charman et al. 2000; Bobrov et al. 2004; Mazei and Tsyganov 2006; Swindles et al. 2014), where the response is mostly related to moisture. For instance, seasonal changes in soil moisture in peatlands due to water table level fluctuations can affect testate amoebae to such an extent to cause shifts in abundance and community composition (Quinn 2003; Warner et al. 2007; Mitchell et al. 2008). Two environmental groups are used as particularly significant: the water group, which includes taxa indicating high water availability and standing water, and the dry group, which includes taxa which can tolerate and live under drier conditions, increasing surface light and aerobic conditions (Chardez 1965; Bobrov et al. 1999; Charman et al. 2000; Bobrov et al. 2004; Mazei and Tsyganov 2006; Table 9.1).

For the local fire regime reconstructions, we used the fire frequency and magnitude of macro-charcoal peaks as described in details in Biagioni et al. (2015b) and in Chapter 8.

Additionally, dried and ground samples were analyzed every 10 cm along the core (total of samples 68) for total C and N by a CNS Elemental Analyzer (CN Elementar Analyzer Vario EL, Hanau, Germany) and ^{15}N natural abundance using isotope ratio mass spectrometry (IRMS; Delta Plus, Finigan MAT, Bremen,

Germany). Short-term phases of soil N cycle disruption, following ecosystem disturbance, might be detected in palaeoenvironmental core analyses. We use ^{15}N natural abundance signatures of the peat to identify short-term disturbance events at the centennial time-scale. We compared the ^{15}N signatures to the "open and pioneer" palynological group, which included pollen and spore taxa that indicate openness and disturbance of vegetation. Additionally, to deduce possible causes of change, the ^{15}N signatures are plotted against fire frequencies, as reconstructed from the peak detection macro-charcoal analysis.

Long-term carbon accumulation rates (LORCA; g C /m^2 yr) were calculated from the dry bulk density (g/cm^3), carbon content (C %) and peat accumulation rates (mm/yr; Chambers et al. 2011). Peat accumulation rates (mm/yr) are calculated based on the age depth models as reconstructed from ^{14}C AMS radiocarbon dates (Biagioni et al. 2015b; Chapter 8). All results are plotted using the program C2 (Juggins 2007).

Table 9.1: Environmental grouping of testate amoebae

Water: taxa commonly living in very wet environment	
Amphitrema cf. *lemanense*	*Difflugia lucida*
Amphitrema sp.	*Euglypha* cf. *acanthophora*
Arcella conica	*Euglypha rotunda*
Arcella cf. *discoides*	*Euglypha tuberculata*
Centropyxis aculeata oblonga	*Heleopera petricola*
Centropyxis aerophila sphagnicola	*Microchlamys* cf. *patella*
Centropyxis cf. *hirsuta*	*Phryganella paradoxa*
Chlamydophyrus minor	*Pyxidicula* cf. *cymbalum*
Cryptodifflugia sacculus	*Pyxidicula operculata*
Cyclopyxis ambigua	*Pyxidicula patens*
Cyclopyxis cf. *penardi*	*Trigonopyxis arcula major*
Cyclopyxis cf. *puteus*	
Dry: taxa commonly living in drier conditions and/or in conditions with high aeration, surface light and water fluctuations	
Arcella cf. *arenaria*	*Hyalosphenia minuta*
Assulina sp.	*Playfairina* cf. *valkanovi*
Centropyxis plagiostoma terricola	*Trigonopyxis arcula*
Corythion dubium	*Trigonopyxis minuta*
Geamphorella lucida	*Trinema lineare* cf. *truncatum*
Other taxa	
Centropyxis aerophila	*Heleopera sylvatica*
Centropyxis cf. *constricta*	*Hyalosphenia* cf. *ovalis*
Centropyxis cf. *laevigata*	*Hyalosphenia* cf. *schoutedeni*
Centropyxis elongata	*Hyalosphenia sublflava*
Centropyxis minuta	*Microchlamys sylvatica*
Centropyxis ssp.	*Nebela* cf. *lageniformis*
Cryptodifflugia oviformis fusca	*Nebela tincta*
Cryptodifflugia sp.	cf. *Plagiopyxis* sp.
Cyclopyxis arcelloides	cf. *Pseudodifflugia gracilis terricola*
Cyclopyxis eurystoma	*Pseudodifflugia* sp.
Cyclopyxis eurystoma parvula	*Pyxidicula* cf. *invisitata*
Cyclopyxis kahli	*Pyxidicula* ssp.
Cyclopyxis cf. *machadoi*	cf. *Trachelocorythion pulchellum*
Cyclopyxis ssp.	*Trigonopyxis microstoma*
Euglypha ciliata glabra	*Trinema complanatum*
Euglypha compressa glabra	*Trinema* cf. *enchelys*
Euglypha cristata decora	*Trinema lineare*
Euglypha laevis	

Results and discussion

Vegetation community changes and phases of peat development

The vegetation of natural lowland tropical peat swamp forests is dominated by trees (Page et al. 1999). Despite that most of the tree families of lowland dipterocarp forests of SE Asia are found in lowland peat swamp forests, the tree species composition of peat swamp forest varies considerably in response to peat thickness, acidity, nutrient and oxygen availability (Anderson 1961; Anderson 1964; Anderson 1983). The spatial distribution of the different communities depends largely on the peat thickness and thus it can also be seen as a temporal succession as peat develop and accumulate over time (Anderson and Muller 1975; Morley 2013). The results from the pollen and spore analysis of the Jaw SPT highlight two distinct vegetation phases (Figure 9.2). The period from ca. 8400 to ca. 4500 cal yr BP (700-470 cm) was dominated by taxa commonly growing on shallow marginal peat that is representative of dipterocarp-swamp mixed rainforests. Peat swamp rainforest established from 4500 cal yr BP up to the present (470-3 cm).

Minerotrophic settings (ca. 7800-4500 cal yr BP)

According to the age-depth model, peat started to accumulate at least 7800 years ago (664 cm core depth). Until ca. 4500 cal yr BP, minerotrophic fresh-water setting characterized the area as indicated by the alternation of high and low values of water testate amoebae group, suggesting occurrence of large fluctuations of the water table (Figure 9.2). Additionally, the limited occurrence of dry indicators, suggests that the water table was commonly above ground and stagnant conditions might have persisted. In this initial phase. the peat was fine to coarse hemic, indicating that the decomposition of the organic matter and humification were relatively high. The vegetation, reconstructed by pollen and spore analyses, indicates that well-developed dipterocarp-swamp mixed rainforests covered the area. Characteristic taxa in this phase were Dipterocarpaceae, *Durio, Blumeodendron*, *Pandanus* and *Stenochlaena palustris* (Biagioni et al. 2015b; Chapter 8). The presence of abundant wood fragments in this core section supported the interpretation that dense arborescent vegetation covered the area, maintained by flooding, which promotes high biomass production and high degree of decay (Biagioni et al. 2015b; Chapter 8). As a consequence, LORCA values were relatively low in this initial phase. Fire episodes were not frequent during the whole phase, indicating that fires were not a common occurrence. However, one large magnitude peak (at ca. 5400 cal yr BP) suggests that a large fire episode occurred.

Ombrotrophic settings (ca. 4500 cal yr BP to present)

A change in the environmental conditions occurred at around 4500 cal yr BP. The testate amoebae dry indicators are consistently more present with higher values, indicating that surface wetness decreased

compared to the previous phase (Figure 9.3). However, the degree of humification decreased as indicated by the coarser hemic peat. Together, these proxies suggest that river flooding ceased to influence the network of interactions of the peat system. The decrease in fresh, oxygen- and nutrient-rich, river water caused a change toward a system that was mostly influenced by nutrient poor rainfall, establishing ombrotrophic conditions. Starting from ca. 4500 cal yr BP, *Pandanus* increased and began to dominate together with *Campnosperma*, *Ilex*, *Austrobuxus* and *Nephrolepis biserrata,* to form a peat swamp rainforest (Biagioni et al. 2015b; Chapter 8). Interestingly, this reconstruction suggests that a positive correspondence between *Pandanus* and fire frequency existed on the long-term. Higher occurrence of *Pandanus* corresponds to periods of increased occurrences of fire episodes. Despite that the resolution of the palynological data did not allow us to deduce causalities, it is likely that the increase in fires and the formation of dense *Pandanus* thickets were interlinked (Figure 9.2).
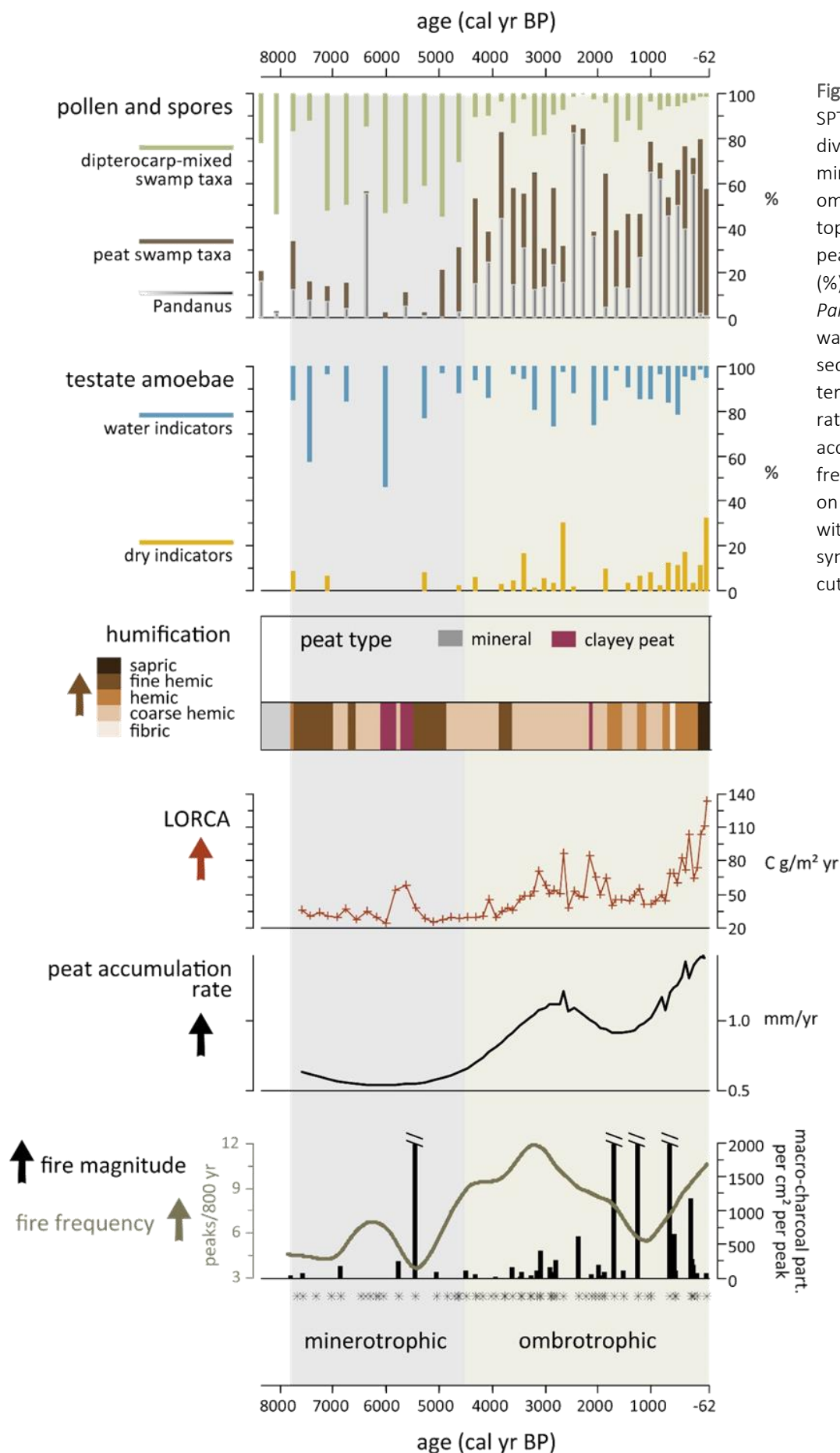
Figure 9.2 Selected results of the Jaw SPT core analyses. The diagrams are dived in the two phases: minerotrophic (light blue) and ombrotrophic (light red). From the top: dipterocarp-mixed swamp and peat swamp pollen and spore taxa (%). In light gray the % values of *Pandanus* pollen. Testate amoebae water and dry indicators (%); sediment/peat characteristics; long-term apparent carbon accumulation rate (LORCA, C g/m^2 yr) and peat accumulation rates (mm/yr); fire frequency and magnitude calculated on the fire peaks (asterisks) detected with the software CharAnalysis. Cut symbol denote fire peaks which were cut to highlight trend in minor peaks

LORCA changes over time

Lacks of evidence for hiatuses in the records indicate that peat accumulated continuously since peat initiation ca. 7800 years ago (Biagioni et al. 2015b; Chapter 8). Average peat accumulation rate calculated based on the age depth model from 664 cm to the top of the core is 0.9 mm/yr (min 0.5; max 1.6 mm/yr), average LORCA is 51 C g/m^2 yr (min 25; max 133).

Conditions that favor peat preservation are found where primary production exceeds degradation by bacteria and fungi. Tropical climate favors decomposition but at the same time also highly productive rainforests and thus often production exceeds degradation leading to net organic matter accumulation (Wüst et al. 2007). Estimates of LORCA are available from few tropical ombrogenous deposits in SE Asia (Dommain et al. 2011; Dommain et al. 2014). When compared to other sites in SE Asia, the Jaw SPT average LORCA is within recorded ranges (Figure 9.3a). The larger difference is found within the record between the minerotrophic and the ombrotrophic phases (Figure 9.3b). Additionally, LORCA seems to increase markedly in correspondence to the expansion of the peat swamp forests and *Pandanus* thickets (Figure 9.4). This suggests that a strong link existed between ecosystem settings and LORCA at Air Hitam, as a peat swamp rainforest became established, LORCA increased. Such increase occurred despite that the establishment of a *Pandanus*-dominated vegetation possibly caused a decrease of aboveground biomass, compared to the woody and more diverse dipterocarp-mixed swamp communities of the minerotrophic phase. However, *Pandanus* thickets develop a massive interlocking root system and have a large belowground biomass quite resistant to degradation (Phillips and Bustin 1998). These results from the Air Hitam core highlight how crucially important are the vegetation communities covering ombrotrophic peat in relation to peat and carbon accumulation in space and time.
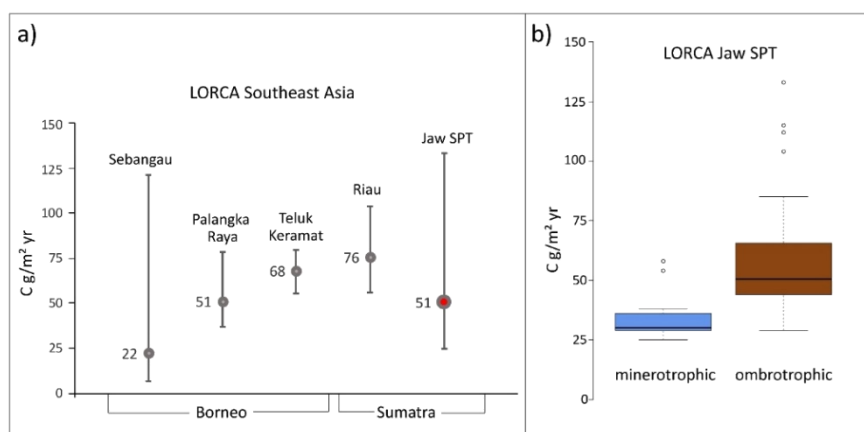
Figure 9.3 a) Comparison of LORCA (expressed as C g/m^2 per yr) during the Holocene of Sebangau, Palangka Raya (Central Kalimantan; Page et al. 2004; Neuzil 1997), Teluk-Kemarat (West Kalimantan; Neuzil 1997) and Siak Kanan (Riau, Sumatra; Diemont and Supardi 1987a) with the Air Hitam peat dome as calculated from the Jaw SPT core analysis. Average (dots) and min, max values (whiskers) are shown. Data source for all other sites other than Air Hitam: Dommain et al. 2011; b) Boxplots of the Jaw SPT LORCA (average, first and third quartiles, min and max values) divided in the minerotrophic (ca. 7800-4500 cal yr BP) and ombrotrophic phase (4500 cal yr BP-present)
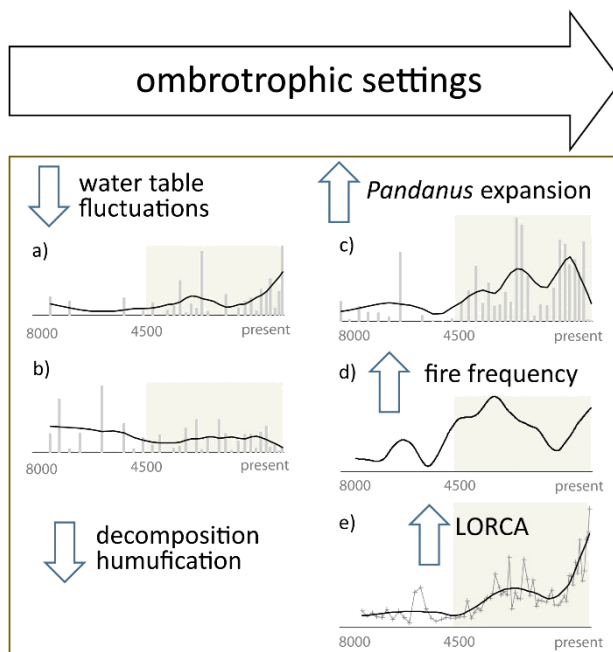
Figure 9.4 Schematic overview of the network of interactions as the system changed from the minerotrophic to ombrotrophic settings. Large blue arrows represent changes (increase/decrease) of the most important parameters reconstructed from the proxy results from minerotrophic to ombrotrophic phase. Graphs show the minerotrophic (white) and ombrotrophic (light green) phases. Black lines where present represent the locally weighted scatter plot smoothings (LOWESS) fitted to the sample values (grey bars) of: a) dry and b) water testate amoeba groups (%), c) *Pandanus* pollen (%), d) macro-charcoal peaks frequency per 800 yr and e) LORCA (C g/m^2 per yr)

Effects of sea-level changes and climate variability on the development of the Air Hitam peat dome

As shown by the palaeoenvironmental data, internal dynamics linked to vegetation and hydrological changes were important factors influencing the development of the Air Hitam peat dome. However, regional changes in climate, in particular precipitation variability, as well as changes in sea-level were found to be equally important for the development and differentiation of peatlands in SE Asia (e.g. Dommain et al. 2011; Dommain et al. 2014).

In order to investigate possible effects of climate and sea-level changes at the Air Hitam peat system, we compared the results from the Jaw SPT core to regional proxies of rainfall and sea-level changes for the last 8000 years (Figure 9.5).

Sea-level changes

According to the review study of Dommain et al. (2014), peatland formation in the lowlands of SE Asia was primarily triggered by sea-level change. As sea-level decreased during the deglaciation, land became submerged enhancing the atmospheric moisture and the hydrological gradient on land. The peatland formation in the inland Air Hitam peat dome occurred since at least 7800 years ago, corresponding to a phase in which the sea-level rise pace lowered (Figure 9.5a). This suggests that the formation of peat in Air Hitam was delayed, possibly as a consequence of the fact that the basin was constantly flooded until the rate of sea-level rise slowly decreased and regional water table stabilized. By 5000 years ago the sea-level had reached a level above its present modern level and it began to

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slowly retrocede. In the inland the marine regression would have caused the regional water table to decrease. This phase of sea-level decrease corresponds to the lowering of the water table fluctuations and the establishment of drier surface conditions at the Air Hitam peat as indicated by the testate amoebae dry indicators increase (Figure 9.5f). This suggests that the hydrological changes reconstructed for the ombrotrophic phase since 5000 years ago followed the sea-level decrease.

Rainfall variability

At the centennial- to millennial-time scale, the rainfall variability in the lowlands of Jambi is mainly a function of strength of the monsoons, amplitude and frequency of ENSO and IOD. The major climatic change from the late glacial to the Holocene in the Indo-Pacific Warm Pool was the increase in moisture due to both rising sea-level and the southward migration of the mean position of the ITCZ (e.g. Haug et al. 2001; Wang et al. 2005). Stalagmite analysis of $\delta^{18}O$ from Borneo (Partin et al. 2007; Figure 9.5b) record a smooth trend of increasing rainfall following the deglaciation up to ca. 5000 years ago. When compared to records north (Haug et al. 2001) and south (Wang et al. 2005) of the equatorial line, this Holocene trend has been explained with a southward migration of the ITCZ in response to precessional forcing. While sea-level changes seem to have greatly influenced the development of the peat accumulation at the Air Hitam peat dome, increasing rainfall (associated with the migration of the ITCZ through the mid- to late Holocene) seems to have not had a major impact (Figure 9.5g). In fact, opposite to what Dommain et al. (2014) found in the coastal peatlands of Kalimantan and Sumatra, the maxima in LORCA at Air Hitam did not correspond with the Holocene rainfall maxima associated with the southernmost position of the ITCZ (Figure 9.5g) at around 5000 cal yr BP. This is likely related to the position of the site in the center of the ranging shift of the ITCZ. The migration north or south of its average position, likely did not have a marked effect on the rainfall patterns in Jambi.

During the late Holocene, ombrotrophic settings, vegetation communities and lowering sea-level/water table fluctuations, corresponded to the increase in LORCA at the Air Hitam peat dome. At the same time, higher seasonality and El Niño droughts led to reduced accumulation and in some cases peat stopped to accumulate in Borneo (Dommain et al. 2011; Dommain et al. 2014). Since in the humid tropical rainforests, recurrent and large wildfires are mostly associated with drought periods, palaeo-fire regime reconstructions in SE Asia have been used to infer ENSO variability at the centennial to millennial time scale. Similar to what were found in several other records across SE Asia (e.g. Haberle et al. 2001; Gagan et al. 2004), fire frequency increased at the Air Hitam site starting from ca. 4500 cal yr BP, likely as a consequence of modern ENSO onset in the late Holocene (Figure 9.5e). Phases of more frequent El Niño events were attested in particular during the last 3000-2000 years (e.g. Moy et al. 2002; Woodroffe et al. 2003; Conroy et al. 2008). Within the age uncertainties due to radiocarbon dating and age-depth modeling, periods of frequent El Niño episodes as reconstructed from the red color intensity at the lake Laguna Pallcacocha in the southern Ecuadorian Andes, correlated to peaks of high fire magnitude of the

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Jaw SPT core (Figure 9.5d-e). This suggests that increasing periods of drought led to increasing chances for the Air Hitam forests to burn. Despite the legacy left by ENSO and related fires, peat accumulation did not cease as in the peatlands of Borneo. This suggests that the effects of water table lowering and enhanced periods of droughts did not reach a threshold beyond which enhanced aerobic decay would have markedly affected the carbon balance of the Air Hitam peat dome. One explanation might be that natural wildfires were most likely affecting the aboveground vegetation, but not the below-water table peat deposits via smoldering fires, as no evidence are found of decrease in LORCA during phases of increased fire frequency and magnitude.

The concomitant occurrences of a strong El Niño event and IOD positive anomaly in 1997/1998 have proven that when these two events occur simultaneously, the effects on negative rainfall anomaly (droughts) over western Indonesia is magnified. In order to test if the centennial scale IOD mean state and ENSO variability increased the chances of fires, we compare our record to the available IOD reconstruction from northwestern Sumatra stable hydrogen compositions of terrestrial plant waxes (Niedermeyer et al. 2014; Figure 9.5c). The pink band indicate phases of increasing δD plant wax interpreted as IOD positive mean state in the Indian ocean from 6500 to 4500 yr BP and from 2500 to 2000 yr BP (Niedermeyer et al. 2014; Figure 9.5c). The first IOD positive mean state phase correspond to a period in which ENSO variability decreased as a consequence of reduced number of Eastern Pacific (EP) flavors events (Karamperidou et al. 2015). The second phase, on the other hand, lack evidences of changes in fire regimes as compared to previous or later periods. Altogether, this suggests that IOD mean state was not an important driver of change for fire regime changes in the Air Hitam peatland. One possible explanation might be related to the geographic location of Air Hitam on the eastern side of the Barisan mountain range. The location behind this topographic barrier might explain the little influence played by the Indian Ocean dipole mode at the site. However, more records recording rainfall anomalies linked to the Indian warm pool state are necessary in order to better constrain these results.
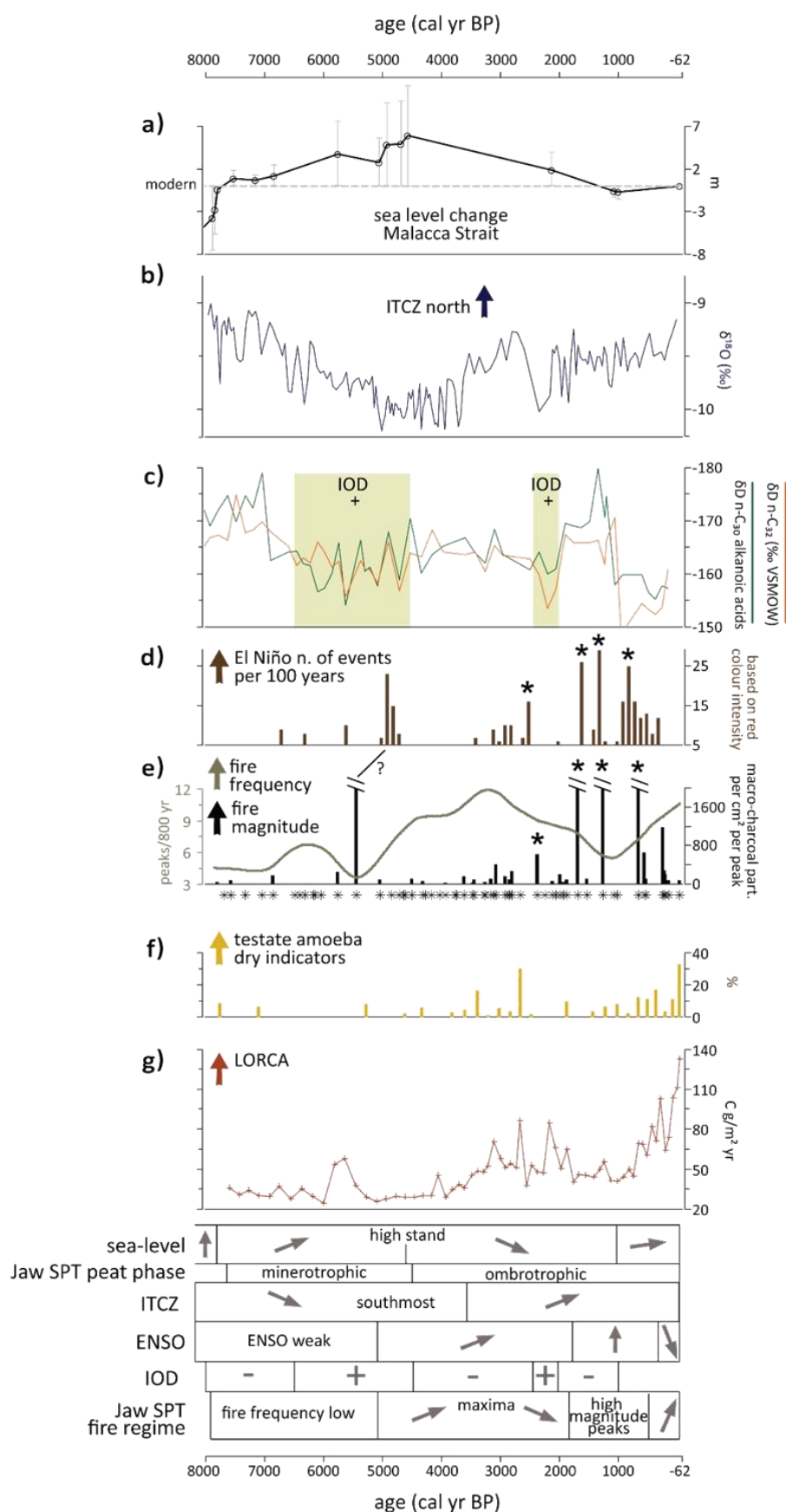
Figure 9.5 Comparison of Jaw SPT core results with other proxies of environmental and climatic variability for the last 8000 years. a) sea-level changes (median, min and max values in m) in the Strait of Malacca (Geyth et al. 1979); b) Borneo stalagmite δ^{18}O record from Gunung Buda National Park (Partin et al. 2007); c) δD records of n-C30 and n-C2 alkanoic acids from marine sediment cores in northwest Sumatra (Niedermeyer et al. 2014); d) modelled number of El Niño events per 100 years based on red colour intensity of sediments from Laguna Pallcacocha, Ecuador (Moy et al. 2002); selected results of the Jaw SPT core analyses (Biagioni et al. 2015b; Chapter 8): e) fire peak magnitude and fire frequency (asterisks indicate correspondence of high in magnitude fire peaks as reconstructed from macro-charcoal analysis of Jaw SPT core and high numbers of El Niño events modelled from Laguna Pallcacocha. Cut symbol denote fire peaks which were cut to highlight trend in minor peaks); f) testate amoeba dry indicators as % of total sum; g) LORCA

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Centennial scale fire disturbance and ecosystems recovery

The ^{15}N natural abundance signatures of peat may reveal environmental changes caused by the disturbance of the forest since ^{15}N signature is essentially a function of N input and soil-N cycling processes over time (Nadelhoffer and Fry 1994). The soil ^{15}N natural abundance signatures have been linked with soil N availability (e.g. gross rates of N mineralization and nitrification) of forests in Brazil, northwestern Ecuador and Panama (Sotta et al. 2008; Arnold et al. 2009; Corre et al. 2010), where high rates of mineral N production reflected high ^{15}N signatures in soils and high soil N2O fluxes (Purbopuspito et al. 2006; Koehler et al. 2009; Wolf et al. 2011). This is linked to the fact that soil N availability controls losses of N from an ecosystem, and N losses are isotopically depleted, owing to fractionation during nitrification (aerobic condition) and denitrification (anaerobic condition). Thus, the higher the soil N availability and consequently the N losses, the more isotopically-enriched is the soil N (Amundson et al. 2003). Disturbance like fire can loss large fraction of biomass N, depleting the pool of actively cycling ecosystem N and provoking a N limitation during the earlier years after fire (e.g. Kauffman et al. 1995; McGrath et al. 2001). Decades to a century after fire, soil N cycle recuperates as biomass N pool buildup, and actively cycling ecosystem N and ^{15}N signature increase. This was shown from a study of forest-age chronosequence (covering 3-200 years old successional forests after abandonment from slash-and-burn agriculture) in the Brazilian Amazon, where soil N availability, gaseous N losses and foliar ^{15}N signatures of trees increase with chronosequence age (Davidson et al. 2007). As a consequence, changes in the organic geochemical properties of peat may reveal environmental changes caused by the disturbance of the forest. Opposite to the C/N ratio, the results of Zaccone et al. (2011) on a peat core in Switzerland, suggest a lack of correlation between ^{13}C and ^{15}N with the humification proxies, suggesting their trends are conservative and preserved along the depth profile. yr BP. Several positive peaks in ^{15}N natural abundance were recorded in the Jaw SPT core for the last ca. 7800 cal yr BP (Figure 9.6). The majority of the peaks (66%) occurred in correspondence of phases in which open and pioneer pollen and spores taxa increased, indicating disturbance of the primary vegetation communities occurred. The ^{15}N natural abundance signatures of the Jaw SPT core suggest forest changes, caused possibly by fire disturbance at the site. The pattern of increase in fire frequency followed by the increase in ^{15}N signatures (Figure 9.6) suggests that the repeated occurrence of fire*s* was the major cause of ecosystems disturbance for the past ca. 7800 years. The majority of the peaks were recorded in the ombrotrophic phase from ca. 4500 cal yr BP to present, while only two phases of ^{15}N enrichment were detected in the initial minerotrophic phase. The vegetation after disturbance seemed to recover relatively quick, and phases of more open or pioneer vegetation lasted less than a hundred years. This suggests that while the peat-swamp communities at the Air Hitam peat

dome were sensitive towards short-term fire disturbance, the rate of recovery remained stable through time despite the large changes in the environmental settings.
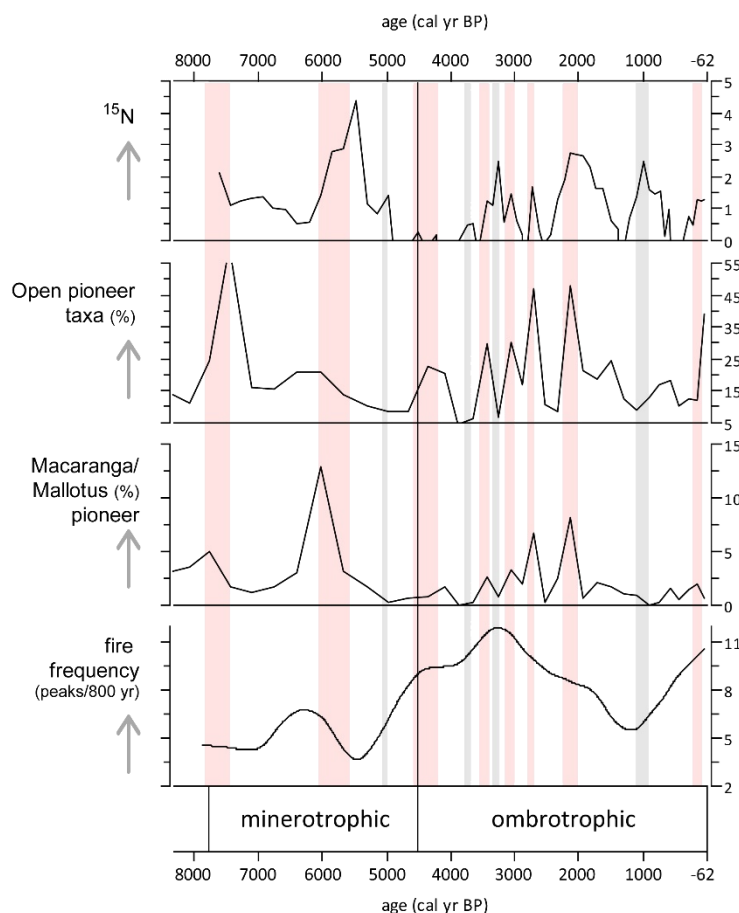
Figure 9.6 Diagram comparing ^{15}N natural abundance, % of total open pioneer pollen and spore group and % of the pioneer pollen taxa *Macaranga/Mallotus* as compared to reconstructed fire frequency. The diagram is dived in minerotrophic and ombrotrophic phases. Pink bars highlight phases in which increase in ^{15}N natural abundance corresponds to increases of the open pioneer group and/or *Macaranga/Mallotus*, while grey bars highlight phases where there was no correspondence

Conclusions

Peat started to form in the Air Hitam area at least 7800 years ago corresponding to a phase of lowering rate of sea-level rise in the Sunda region. As peat accumulated over time, the hydrological and environmental conditions gradually changed. Initially, vegetation communities representative of productive and diverse dipterocarp-mixed swamp rainforests dominated the area. The hydrological gradient changed when sea-level started to decrease following the high stand in sea-level around 5000 years ago. The combined effects of a decreasing regional water table level and the increased in peat depth above river influence caused a decrease water table fluctuations of the peat domed areas in Air Hitam. The decrease of fresh, oxygen- and nutrient-rich river water caused a change toward oligotrophic conditions, a system which was mostly fed by nutrient poor rainfall. The vegetation communities were replaced by peat swamp taxa and *Pandanus* thickets expanded in the area. Interestingly, LORCA increased in this later phase, following the expansion of *Pandanus*. The lack of correlation with regional rainfall maxima as reconstructed from monsoon and ITCZ records suggests that the most important factor influencing the increase in LORCA at Air Hitam was the expansion of *Pandanus* and their extensive roots system. Additionally, the millennial scale changes in ENSO played an important role. The

occurrence of more frequent period of droughts correlated with the increase in fire frequency and fire magnitude in Air Hitam area. The link between phases of forest disturbance and fire suggests fires caused by El Niño occurrences affected the forest communities. However, the decrease of pioneer taxa was relatively rapid, lasting less than a hundred years. Thus the vegetation communities have shown a resilience capacity towards the centennial scale disturbance caused by natural fires.

Acknowledgments

CHAPTER 10 - Synthesis and implications

The research conducted and compiled in this thesis contributes to the knowledge of the long-term vegetation, climate and fire dynamics as well as human impact on montane and peat-swamp rainforests of Indonesia. The use of a multi-proxy palaeoecological approach increased our understanding of how these ecosystems have responded to natural and anthropogenic driven changes in the past. Altogether, these results represent a further step forward to the gaining of a more comprehensive understanding of modern tropical ecosystem dynamics and their functions in the long-term.

Synthesis of the main results on vegetation, climate, fire dynamics and human-landscape interactions

Montane ecosystems of Central Sulawesi

The Lake Kalimpaa record is the first archive from Central Sulawesi providing information on palaeovegetation dynamics under the background of reconstructed palaeoenvironmental and palaeorainfall conditions throughout the past ca. 1500 years. Today's climate of Sulawesi is driven by the monsoon system as well as by ENSO. The mountain regions of Central Sulawesi experience perhumid climate conditions with few seasonal changes in precipitation, making the vegetation a possible sensitive target for future changes of precipitation patterns (Wündsch et al. 2014; Biagioni et al. 2015a). The palaeovegetation, as reconstructed from the palynological analysis, reveals that around Lake Kalimpaa, the Fagaceae family dominates the entire recorded period, as they still do today.

Granulometric and geochemical data from the Lake Kalimpaa record provide indications for an increasingly wetter climate (higher rainfall intensities and/or mean rainfall) on centennial to millennial time scales from approximately AD 560 to the 20th century. Since this trend towards wetter conditions was also detected in other palaeoclimatic studies from the region, a likely explanation for this is that the ITCZ migrated southward in this period of time (Wündsch et al. 2014).

Two disturbance events (ca. AD 1090-1190 and ca. AD 1450-1620) occurred in the catchment area of Lake Kalimpaa. The comparison with the fire frequency record derived from macro-charcoal analysis indicates that these events were caused by forest fires. The regional correlation of these events with periods of drought registered elsewhere in Sulawesi and in Java suggests that centennial-scale increases in fire frequencies at Lake Kalimpaa were consequences of the vegetation being more prone to fire, probably due to more frequent and/or longer El Niño events.

Despite that, Fagaceae did not decrease in their representation, indicating resilience towards droughts of at least one species of the family. However, palynological diversity values indicate that within-landscape diversity (Whittaker's gamma diversity) decreased when fires increased. Palynological rate of

change and compositional turnover indicate that the vegetation communities were more resilient to fire disturbance during periods of average high rainfall.

At Lake Kalimpaa, human-landscape interactions are evident only starting from the 20th century (from ca. 1950 to present), with *Weinmannia* rising probably due to the logging of emergent *Agathis* trees and/or landslides caused by the construction of the road which today passes near the lake (Biagioni et al. 2015a; Chapter 6).

The Lindu valley is located only 30 km west from Lake Kalimpaa but, opposite to what was found at Lake Kalimpaa, a longer history of human-landscape interactions was found from the palynological, charcoal and diatom reconstructions of the Lake Lindu core for the past ca. 1000 years (Biagioni et al. 2016; Chapter 7). Evidence of frequent burning and possible shifting cultivation from an earlier phase from ca. AD 1000 to 1200 might be related to the metal age population which erected the megaliths in the province of Central Sulawesi. From ca. AD 1200-1700, decreases of macro-charcoal concentrations and pioneer vegetation indicators show that the use of the landscape of the Lindu plain had become more permanent. Due to the little research conducted so far on the megalithic culture of Central Sulawesi, it remains uncertain whether the architects of such a cultural change were the megalith people or different ethnic groups. A phase of forest recovery from ca. AD 1730 to 1910 correlated with a decrease in human activities in the valley, which historical reports describe as mostly limited to fishing and cattle grazing (Sarasin and Sarasin 1905). These results indicate that when human pressure on the landscape decreases as a consequence of different strategies of subsistence, the montane ecosystems possess a great capacity of recovery and fagaceous forest communities can expand on a relatively rapid time scale. The most recent part of the Lake Lindu record shows a trend towards deforestation that started in the late 20th century, lasting until now, following the increase of human population encroachment on the margins of the LLNP.

Inland peat-swamp ecosystems in Jambi, Sumatra

The extended tropical peatlands of SE Asia have retained a large amount of carbon, functioning as a large sink for thousands of years (Page et al. 2004; Page et al. 2010). Currently, however, these peatlands are facing large scale degradation and conversion, turning from carbon sink to source (Hooijer et al. 2010; Jauhiainen et al. 2010; Page et al. 2011).

The need to manage peatlands requires a better understanding of the main components and processes of peat-forming systems. The study conducted in the peat dome of Air Hitam demonstrates that indicators used to evaluate and predict peat processes should be based on an understanding of the factors (in addition to age) that control peat accumulation. These include vegetation, hydrology and soil, as peat-swamp ecosystems are the result of combined action of allogenic (climate, hydrology) and autogenic (vegetation) factors which vary in space and time depending on the thickness of the peat

deposit and environmental/hydrological settings. Multi-proxy results depict a picture of a highly dynamic system (Chapter 8). Since the beginning of peat accumulation around 7800 years ago net balance accumulation exceeded degradation, and the system remained a carbon sink for ca. 7800 years. At first the peatland in Air Hitam was fed by fluvial run-off and the vegetation gradually changed from mixed Dipterocarpaceae swamp to marshy swamp communities dominated by *Durio* trees at around 5300 years ago. A marked change in the vegetation community's composition occurred at the beginning of the late Holocene about 4500 years ago. A pole forest established with *Pandanus* thickets colonizing the area as the dome developed into a rainfall-fed ombrogenous system. At the same time, macro-charcoal peak detection analysis reveals that fire frequency increased, possibly as a consequence of the ENSO-onset.

Peat accumulation rates and carbon storage in the dome of Air Hitam have been considerable in the past. Higher rates of peat/carbon accumulation were found to correlate with *Pandanus* expansion under ombrotrophic settings. The testate amoebae assemblage indicates that the ombrogenous *Pandanus*-pole forest phases were characterized by lower water table fluctuations. Despite the lower biomass input peat accumulation rate was in average higher than during the mixed-swamp community phases. The correlation between high peat accumulation and relatively drier phases linked to ENSO makes the area of Air Hitam potentially one of the most effective carbon sequestering ombrotrophic peatlands in the view of future climate scenarios (Chapter 9).

Impact of El Niño (droughts) and fires on the carbon storage function of the natural forest ecosystems

Evidence from both historical records and charcoal particles embedded in stratigraphic records indicate that fires occurred in the everwet tropical forests although rarely (Cochrane 2003). Similarly, the results of charcoal analysis in montane (Lake Kalimpaa, Chapter 5 and Chapter 6) and peat-swamp (Air Hitam, Chapter 8 and Chapter 9) ecosystems show that although rare, wildfires did occur in the past in Central Sulawesi and Sumatra. The regime of fire, in particular the frequency, was found to correlate with regional scale drought episodes in Central Sulawesi. Both fire frequency and magnitude of events increased at Air Hitam in correlation with the increase in El Niño number of events. These results indicate that the fire regime of montane ecosystems in Central Sulawesi and peat-swamp forests in Jambi can change in the long-term following changes in patterns of rainfall and ENSO variability. However, our results indicate that the effects of fire and droughts on the carbon storage functions of the two systems under study might not have been marked. In particular, Fagaceae species representation around Lake Kalimpaa was not affected by increases of fires caused by drought, and the carbon accumulation capacity of the ombrotrophic-*Pandanus* dominated ecosystems in Air Hitam did not change (but rather increase) when frequency and magnitude of fires increased. These results

suggest that under natural conditions, fire occurs in these ecosystems, but the effect on their carbon storage function in time can be relatively low.

A historical perspective of current and future changes

Investigations on the long-term dynamics of vegetation communities using a multi-proxy palaeoecological approach extend our knowledge on how tropical ecosystems have responded to past climate variability and human activities and help putting the current and future scenarios of change into a more comprehensive historical perspective.

The results from the investigation on the fagaceous montane rainforests surrounding Lake Kalimpaa in Central Sulawesi (Biagioni et al. 2015a; Chapter 6) suggest that the main driver of change for the past ca. 1500 years was rainfall variability. In particular, the occurrences of regional centennial scale long periods of drought and changes in the amount of rainfall following the migration of the ITCZ across the tropics were affecting the composition and diversity of the ecosystems. Interestingly, species turnover was less marked when drought episodes occurred in combination with millennial scale enhanced rainfall at the intra-annual scale. This suggests that the natural forest ecosystems are more resilient to the occurrences of episodes of droughts when the overall seasonality is less marked and perhumid conditions are established. Predictions of climate change in the area are still uncertain. However, if seasonality will increase and drought episodes will occur more frequently, the response of vegetation communities in the future might closely resemble the conditions reconstructed between ca. AD 1070 and AD 1450. The occurrences of wildfire will likely increase in frequency, while communities' diversity may decrease and long-living pioneers such as *Weinmannia* could become more common in disturbed patches of the forest. The growth of the valuable *Agathis* tree might be greatly reduced, as results indicate that species of this genus were sensitive towards patterns of decreasing rainfall and rates of recovery following selective logging have been slow. However, the overall function as biomass storage of the Fagaceae forests might persist even under changing rainfall conditions, as our results indicate that species from this family have been resilient towards rainfall variability and droughts. However, if changes in temperature caused by anthropogenic climate change should occur (Stocker 2014), competition and pressure from species currently growing at lower altitudes might increase and the overall composition of the forest could change. As no comparable conditions were found for the past ca. 1500 years, the consequences of temperature change over these ecosystems are currently hard to predict.

A rapid acceleration of disturbance driven by human activities is underlined by the results in both regions under study for the past decades. The overall results of this research highlight that the pressure exerted over millennia on montane and peat-swamp rainforests in Indonesia has been of a different magnitude as compared to the modern anthropogenic driven changes. In particular, increased in

siltation, as a consequence of forest conversion, is causing lake level decreases in both small basins, such as Lake Kalimpaa, and larger water basins, such as Lake Lindu in the montane regions of Central Sulawesi. The consequences of these rapid changes are likely to affect the ecosystem functions thought a loss of the ecosystems capacity to prevent flooding and landslides, affecting human welfare of the isolated communities living in the valleys. Natural forest cover on the upper slopes of the valleys has been underlined to be a particular important factor in relation to flood protection, erosion and siltation (Whiteman and Fraser 1997). Given the rapid rate of forest recovery experienced at the Lindu valley in the phase from ca. AD 1730 to 1910, a better management of irrigation channels and restoration of the degraded forest on the slopes of the valley might reverse this observed trend if actions are taken in the immediate future.

The results from the multi-proxy reconstruction of the peat-swamp ecosystems of Air Hitam in Sumatra suggest that the most important factor influencing the increased carbon storage at Air Hitam was the expansion of *Pandanus* and its extensive roots system, while water table change, above ground biomass and rainfall variability overall played a less important role.

However, the recent degradation of the area and the conversion to oil palm plantations greatly changed the vegetation communities on the peatland. Intensive logging activities virtually removed all of the large trees in the area and the constructions of channels and ditches have artificially lowered the water table. If peat accumulation is largely controlled by the quantity of belowground organic matter inputs (i.e. roots of *Pandanus*), as evidence in the study suggests, then the peat forests where root mats have declined or disappeared by natural or artificial causes will remain in equilibrium or subside, but will not expand. Net accumulations of peat will likely occur where high water levels and thick root mats still exist, and possibly in altered areas where these conditions are re-established. Under these conditions, climate-induced changes in precipitation will probably not alter factors to an extent in which peat accumulation ceases in the long-term, even under scenarios of fire regime changes. However, in order to preserve the potential value of this area as a carbon sink, a better management of the water table level should be applied and restoring measurements, act at re-establishing the natural vegetation cover, should be taken.

Concluding remarks, uncertainties and perspectives

The results from the research compiled in this thesis highlight the value of using a palaeo multi-proxy approach in order to better understand dynamics and functions of tropical ecosystems and the identification of the most important drivers of change.

Despite their potential, so far the number of such investigations in Indonesia is still very limited. As a consequence, several uncertainties currently remain. For instance, a large question mark remains on the role of prehistoric communities in shaping present-day tropical ecosystems. The example of Central

Sulawesi is an emblematic one. Virtually nothing is known about the megalithic culture despite few studies (Kirleis et al. 2011; Kirleis et al. 2012), and the research conducted at the Lindu valley (Chapter 7), suggests that this population might have played an important role, and it might have acted as a strong driver of change already thousands of years ago.

Trying to predict the fate of montane tropical rainforests under future climate change scenarios requires an understanding of the long-term effect of climate variability on the different communities occupying different altitudinal niches. So far, no such study exists and one large uncertainty remains on how these ecosystems might respond to increases in temperature.

The need to manage peatlands has recently gained large attention due to the fast rate of degradation and conversion these important ecosystems are experiencing in SE Asia. However, when a management of the systems is required in order to restore and/or preserve the ecosystem functions, a better understanding of the main components and processes of peat-forming systems is essential. Few restoration programs have taken this into consideration and so far the planning and approaches are limited to short-term studies on recovery and rate of survival of pioneer vegetation cover during early stage succession. Due to the fact that peat-swamp ecosystems in SE Asia are spread over large areas (and on different islands), environmental conditions are greatly different from one system to the other and generalizations are difficult to make. There is a need for more integrative investigations in key sites, in order to act in the most efficient way.

Although there is still much to do before we can gain a deep understanding of the stability, resilience and recovery of tropical ecosystems in Indonesia, with our investigations in Central Sulawesi and Sumatra we have found that the use of a palaeoecological integrative approach can be a powerful tool and much can be learnt from the past.

Slowly, we can move forward to our goal to unravel the processes and factors that brought tropical ecosystems to be the way they are today.

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Appendix A

Identified pollen, spores and non-pollen palynomorphs (NPP) of the records – complete list and photos of selected taxa

Appendix A1: Table of identified pollen, spores and NPP taxa

Abbreviations for palynological records:

- kal: Lake Kalimpaa, LLNP, Central Sulawesi (Chapter 6)
- lin: Lake Lindu, LLNP, Central Sulawesi (Chapter 7)
- jaw: Jaw SPT, Air Hitam, Jambi Province, Sumatra (Chapter 8 and Chapter 9)

Family	pollen type	photo n.	record
Actinidiaceae	*Saurauia*	1	kal
	Actinidiaceae		jaw
Adoxaceae	*Viburnum*		kal
Alismataceae	*Sagittaria*		lin
Amaranthaceae	*Alternanthera*	2	lin
	Amaranthaceae	3	lin
Amaryllidaceae	*Narcissus*	4	jaw
Amaryllidaceae/Liliaceae	Amaryllidaceae/Liliaceae		kal
Anacardiaceae	Anacardiaceae		kal, jaw
	Campnosperma	5	jaw
Anisophylleaceae	*Combretocarpus*	6	jaw
Annonaceae	*Cyathocalyx*		lin
	Annonaceae		lin
Apiaceae	Apiaceae	7	lin, jaw
Apocynaceae	*Alstonia*		jaw
	Anodendron	8	jaw
	Apocynaceae		kal, lin, jaw
Aquifoliaceae	*Ilex*	9	kal, lin, jaw
Araceae	Araceae		lin, jaw
	Pothos	10	lin
Araliaceae	Araliaceae	11	lin, jaw
	Hydrocotyle	12	lin
	Polyscias		kal
	Schefflera	13	kal
Araucariaceae	*Agathis*	14	kal, lin
Arecaceae	*Areca*	15	lin, jaw
	Arecaceae		jaw
	Arecaceae cf. Gronophyllum		lin
	Arecaceae echinate type		kal
	Arecaceae indif.		lin
	Arecaceae trichotomocolpate reticulate		kal
	Arenga	16	lin
	Calamus	17	kal, lin, jaw
	Caryota		lin
	Cocus type		lin

Family	pollen type	photo n.	record
Arecaceae	*Oncosperma*	18	jaw
	Pinanga annulocolpate		lin
Aristolochiaceae	Aristolochiaceae		lin, jaw
Asparagaceae	*Dracaena*		lin
Asteraceae	Asteraceae	19, 20	kal, lin
	Senecio type		lin
Begoniaceae	Begoniaceae		kal, lin
Brassicaceae	Brassicaceae		lin, jaw
Burseraceae	*Canarium*	21	lin, jaw
	Burseraceae		kal
Cannabaceae	*Celtis*	22	kal, lin, jaw
	Gironniera		lin, jaw
	Trema	23	kal, lin, jaw
Caryophyllaceae	Caryophyllaceae	24	jaw
Celastraceae	Celastraceae	25	lin
	Euonymus	26	kal
	Lophopetalum	27	jaw
Chlorantaceae	Chlorantaceae	28	kal
Chrysobalanaceae	*Parastemon*	29	jaw
Clethraceae	*Clethra*	30	kal
Clusiaceae	Clusiaceae		kal
	Garcinia	31	jaw
Combretaceae/Melastomataceae	Combretaceae/Melastomataceae	32	kal, lin, jaw
Convolvulaceae	*Ipomoea* type		jaw
	cf. *Erycibe*		lin
Cornaceae	*Alangium*		kal
	Mastixia		lin
Costaceae	*Costus*	33	lin
Cucurbitaceae	Cucurbitaceae		kal, lin
Cunoniaceae	Cunoniaceae dicolporate type		kal
	Weinmannia	34	kal, lin
Cyperaceae	Cyperaceae	35	kal, lin, jaw
	Cyperus	36	lin
	Scirpus		lin
	Thoracostachyum/Mapania		jaw
Daphniphyllaceae	*Daphniphyllum*		kal, lin
Dilleniaceae	Dilleniaceae		lin
Dipterocarpaceae	*Dipterocarpus*	37	jaw
	Dryobalanops	38	jaw
	Hopea	39	jaw
	Shorea	40	jaw
Droseraceae	*Drosera*	41	kal
Ebenaceae	*Diospyros*	42	lin, jaw
Elaeocarpaceae	Elaeocarpaceae	43	kal, lin, jaw
Ericaceae	Ericaceae	44	kal, lin, jaw

Family	pollen type	photo n.	record
Euphorbiaceae	*Acalypha*	45	kal, lin, jaw
	Blumeodendron	46	jaw
	Cephalomappa	47	jaw
	Homalanthus		kal, lin, jaw
	Homonoia		lin
	Macaranga/Mallotus	48	kal, lin, jaw
	Neoscortechinia cf. *nicobarica*		lin
	Euphorbiaceae		kal
Fabaceae	Caesalpinioideae indif.		lin
	Caesalpinoideae - *Cassia* type		lin
	Koompassia	49	jaw
	Mimosoideae		lin, jaw
	Mucuna	50	lin
	Pterocarpus type		kal
	Fabaceae indif.		kal, lin, jaw
Fagaceae	Fagaceae indif.		kal
	Lithocarpus/Castanopsis	51	kal, lin
Flacourtiaceae	Flacourtiaceae		kal, lin
Flagellariaceae	*Flagellaria*		jaw
Gentianaceae	*Fagraea*	52	lin
Gnetaceae	*Gnetum*		lin
Gunneraceae	*Gunnera*	53	kal
Haloragaceae	*Myriophyllum*		lin
Hamamelidaceae	Hamamelidaceae		kal
Hydrangeaceae	*Dichroa febrifuga*		lin
Hydrocharitaceae	Hydrocharitaceae		kal
Icacinaceae	Icacinaceae		lin
	Platea	54	lin
Iridaceae	Iridaceae		kal
Juglandaceae	*Engelhardtia*	55	kal, lin, jaw
	Juglandaceae indif.		kal
Lamiaceae	Lamiaceae		kal, lin, jaw
Lauraceae	Lauraceae		kal
Lecythidaceae	*Barringtonia*	56	kal
Liliaceae	Liliaceae		lin
Loganiaceae	*Geniostoma*	57	lin
Loranthaceae	Loranthaceae	58	kal, lin, jaw
Lythraceae	Lythraceae	59	kal, lin
Malvaceae	*Helicteres*		lin
	Hibiscus		jaw
	Malvaceae		kal
	Sterculioideae		jaw
	Tilioideae		lin
Malvaceae (Brownlowioideae)	*Brownlowia*	60	jaw
Malvaceae (Bombacoideae)	*Durio*	61	jaw

Family	pollen type	photo n.	record
Melastomataceae	*Astronia*		lin
Meliaceae	*Aglaia*	62	kal, lin, jaw
	Aglaia rubiginosa type		jaw
	Dysoxylum	63	kal
	Meliaceae		kal, lin, jaw
Menispermaceae	*Fibraurea*		kal
	Fibraurea tinctoria		jaw
	Menispermiaceae	64	jaw
	Stephania	65	kal, lin, jaw
Menyanthaceae	Menyanthaceae		kal
Moraceae/Urticaceae	Moraceae/Urticaceae	66	kal, lin, jaw
Moraceae	*Ficus*	67	lin, jaw
Myricaceae	*Myrica*	68	kal, lin
Myristicaceae	*Knema*		jaw
	Myristicaceae		kal, lin, jaw
Myrtaceae	Myrtaceae	69	kal, lin, jaw
Oleaceae	*Chionanthus*		lin
	Ligustrum		jaw
	Oleaceae		kal
Pandanaceae	*Freycinetia*	70	kal, lin, jaw
	Pandanus	71	kal, lin, jaw
Paracryphiaceae	*Quintinia*	72	kal, lin
Pentaphylacaceae	*Ternstroemia*		kal, lin
Phyllanthaceae	*Antidesma*	73	jaw
	Aporosa	74	jaw
	Baccaurea/Bischofia	75	kal, lin, jaw
	Breynia	76	lin
	Glochidion	77	lin
	Phyllanthus		lin, jaw
	Phyllanthus urinaria type		jaw
Picrodendraceae	*Austrobuxus*	78	jaw
Pinaceae	*Pinus*		kal
Piperaceae	*Peperomia*	79	kal, lin, jaw
	Piper		lin, jaw
Plantaginaceae	*Callitriche*	80	lin
	Plantago	81	kal, lin, jaw
Poaceae	Poaceae	82, 83	kal, lin, jaw
Podocarpaceae	*Dacrycarpus*	84	kal, lin
	Phyllocladus	85	kal, lin
	Podocarpaceae indiff.		jaw
	Podocarpus	86	kal, lin
	Sundacarpus		kal
Pontederiaceae	Pontederiaceae	87	lin
Potamogetonaceae	*Potamogeton*		kal
Primulaceae	*Ardisia*		jaw

Family	pollen type	photo n.	record
Primulaceae	*Discocalyx*		lin
	Maesa		lin
	Myrsine	88	kal, lin, jaw
Proteaceae	Proteaceae		kal, jaw
Rhizophoraceae	Rhizophoraceae	89	jaw
Rosaceae	Rosaceae		jaw
Rubiaceae	*Gardenia*		lin
	Hedyotis	90	jaw
	Lasianthus		jaw
	Mussaenda		jaw
	Nauclea	91	jaw, lin
	Neonauclea	92	lin
	Randia	93	kal, lin, jaw
	Timonius		lin
	Rubiaceae indif.		kal, jaw
Rutaceae	*Acronychia*		lin
	Rutaceae indif.		kal, lin, jaw
	cf. *Melicope hookeri*		lin
	Zanthoxylum	94	kal
Sapindaceae	*Acer*		kal
	Allophylus	95	lin
	Dodonaea		lin
	Pometia	96	jaw
	Sapindaceae		kal
Sapotaceae	*Palaquium* type	97	jaw
	Sapotaceae	98	kal, lin, jaw
Scrophulariaceae	*Buddleja*		kal
Smilacaceae	*Smilax*		lin
Solanaceae	Solanaceae		kal, lin
Staphyleaceae	*Turpinia*		lin
Stemonuraceae	Stemonuraceae	99	lin, jaw
Styracaceae	*Styrax*	100	kal
Symplocaceae	*Symplocos*	101	kal, lin
Theaceae	*Gordonia*	102	lin
Theaceae	Theaceae indif.		lin
Thymelaeaceae	Thymelaeaceae		jaw
Thymelaeaceae	*Wikstroemia*		kal
Trimeniaceae	*Trimenia*		lin
Typhaceae	*Typha*	103	kal, lin
Violaceae	Violaceae	104	lin
Vitaceae	cf. *Cissus*		jaw
Winteraceae	Winteraceae		Lin

Family	NPP type	photo n.	record
Glomeraceae	*Glomus*	105	lin
Tilletiaceae	*Tilletia*	106	lin
Botryococcaceae	*Botryococcus*	107	Lin
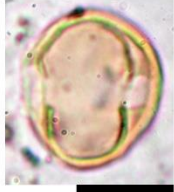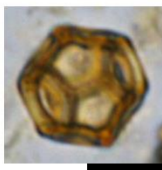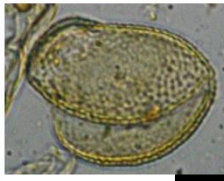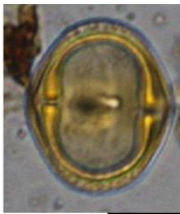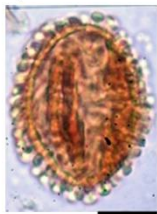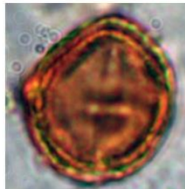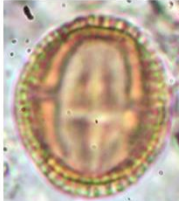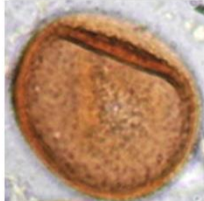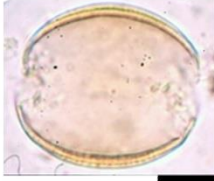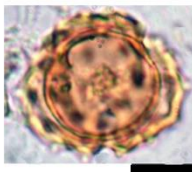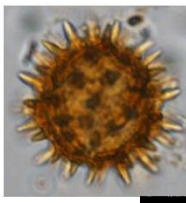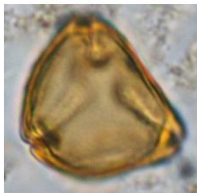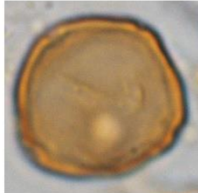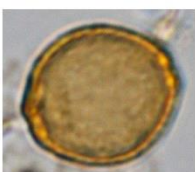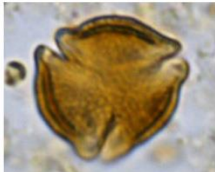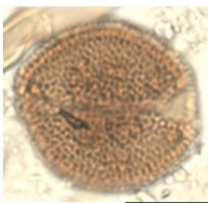
Family	spore type	photo n.	record
Aspleniaceae	*Asplenium*	108	kal, lin, jaw
	Asplenium belangeri	109	lin
	Asplenium macrophyllum	110	lin
Blechnaceae	*Blechnum*	111	kal, lin, jaw
	Stenochlaena		lin
	Stenochlaena areolaris	112	jaw
	Stenochlaena palustris	113	jaw
Cyatheaceae	Cyatheaceae	114	kal, lin
Davalliaceae	*Davallia*	115	kal, lin, jaw
Dennstaedtiaceae	Dennstaedtiaceae		jaw
	Histiopteris		jaw
	Microlepia	116	lin
	Pteridium		lin
Dicksoniaceae	*Dicksonia*	117	kal
Dryopteridaceae	*Arachniodes*	118	lin
	Dryopteris		kal
	Elaphoglossum	119	lin, jaw
	Polystichum	120	lin
Hymenophyllaceae	Hymenophyllaceae	121	kal, lin, jaw
Hypodematiaceae	*Leucostegia*	122	jaw
Isoetaceae	*Isoetes*		kal
Lindsaeaceae	Lindsaeaceae		lin, jaw
Lycopodiaceae	*Huperzia*	123	kal, lin, jaw
	Lycopodiaceae indif.		jaw
	Lycopodiella cernua	124	kal, lin, jaw
	Lycopodium indif.		jaw
	Lycopodium clavatum		kal
	Lycopodium verrucato		kal
Lygodiaceae	*Lygodium*	125	lin
Marattiaceae	*Angiopteris*	126	lin
Nephrolepidaceae	*Nephrolepis*	127	lin, jaw
Ophioglossaceae	*Ophioglossum*	128	kal
Osmundaceae	*Osmunda*		kal
Polypodiaceae	*Grammitids*		kal, lin
	Microsorum heterocarpum	129	jaw
	Microsorum unctatum		jaw
	Polypodiaceae	130	kal, lin, jaw
	Selliguea		jaw
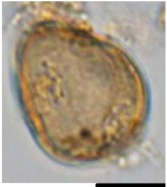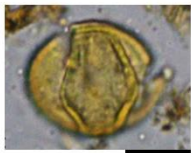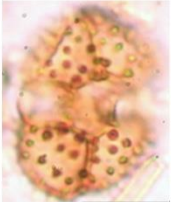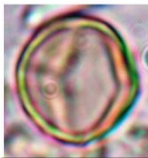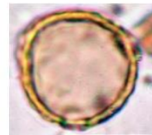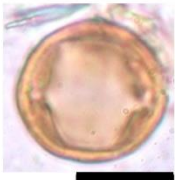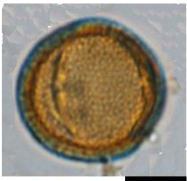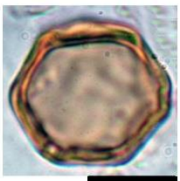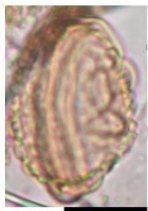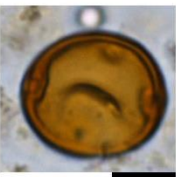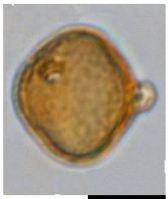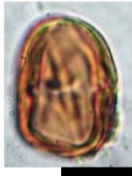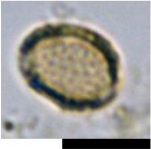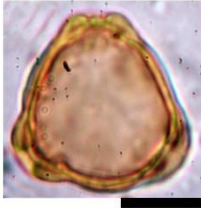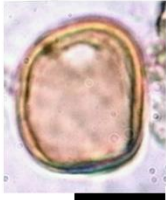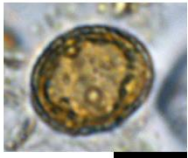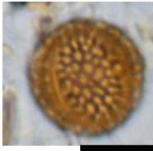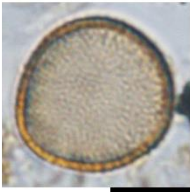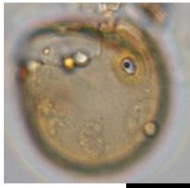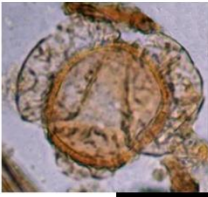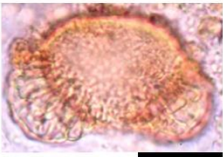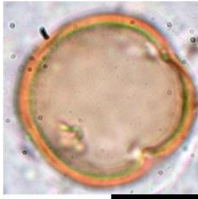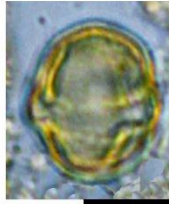
Family	spore type	photo n.	record
Pteridaceae	*Adiantum*	131	kal, lin, jaw
	Anogramma		jaw
	Pteris	132	kal, jaw
Selaginellaceae	*Selaginella*	133	kal, lin, jaw
Thelypteridaceae	*Amauropelta*	134	lin
	Thelypteridaceae		kal, jaw
Woodsiaceae	Woodsiaceae	135	lin
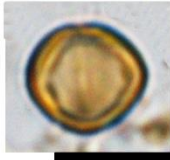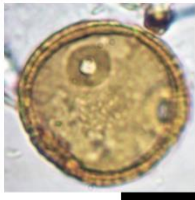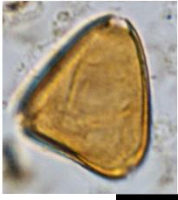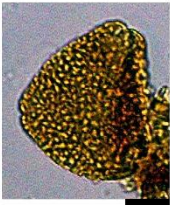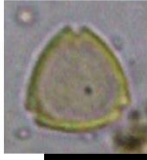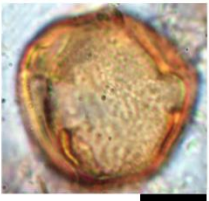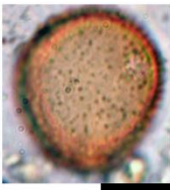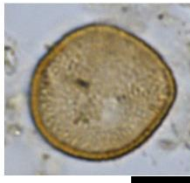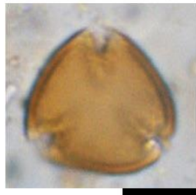
Appendix A2: Plates of selected pollen, spores and NPP taxa
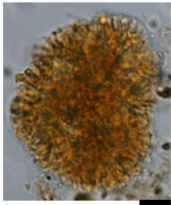
Black line on photos represents the scale bar = 10 µm
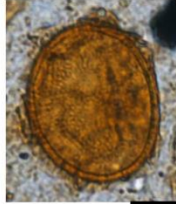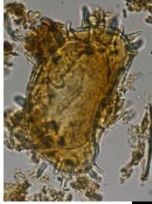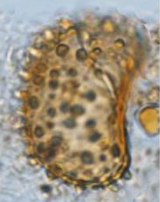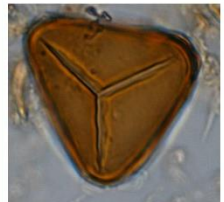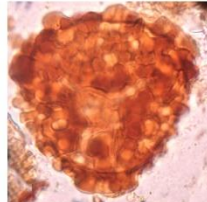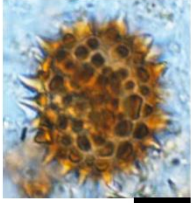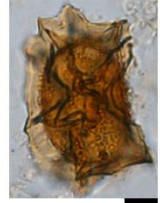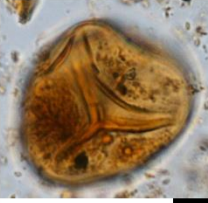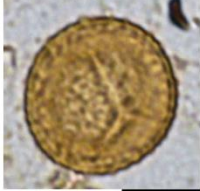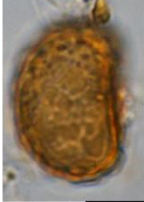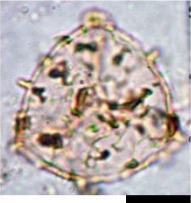
Actinidiaceae	Amaranthaceae		Amaryllidaceae	Anacardiaceae
1) *Saurauia*	2) *Alternanthera*	3) Amaranthaceae	4) *Narcissus*	5) *Campnosperma*
Anisophylleaceae	Apiaceae	Apocynaceae	Aquifoliaceae	Araceae
6) *Combretocarpus*	7) Apiaceae	8) *Anodendron*	9) *Ilex*	10) *Pothos*

Araliaceae

11) Araliaceae	12) *Hydrocotyle*	13) *Schefflera*	14) *Agathis*	15) *Areca*
			Araucariaceae	Arecaceae

Arecaceae

Asteraceae

16) *Arenga*	17) *Calamus*	18) *Oncosperma*	19) Liguliflorae	20) Tubuliflorae

Burseraceae	Cannabaceae		Caryophyllaceae	Celastraceae
21) *Canarium*	22) *Celtis*	23) *Trema*	24) Caryophyllaceae	25) Celastraceae
Celastraceae		Chloranthaceae	Chrysobalanaceae	Clethraceae
26) *Euonymus*	27) *Lophopetalum*	28) Chloranthaceae	29) *Parastemon*	30) *Clethra*

Clusiaceae	Combretaceae/Melastomataceae	Costaceae	Cunoniaceae	Cyperaceae
31) *Garcinia*	32) Combretaceae/Melastomataceae	33) *Costus*	34) *Weinmannia*	35) Cyperaceae

Cyperaceae	Dipterocarpaceae			
36) *Cyperus*	37) *Dipterocarpus*	38) *Dryobalanops*	39) *Hopea*	40) *Shorea*

Droseraceae	Ebenaceae	Elaeocarpaceae	Ericaceae	Euphorbiaceae
41) *Drosera*	42) *Diospyros*	43) Elaeocarpaceae	44) Ericaceae	45) *Acalypha*

Euphorbiaceae			Fabaceae	
46) *Blumeodendron*	47) *Cephalomappa*	48) *Macaranga*	49) *Koompassia*	50) *Mucuna*

Fagaceae	Gentianaceae	Gunneraceae	Icacinaceae	Juglandaceae
51) *Lithocarpus/ Castanopsis*	52) *Fagraea*	53) *Gunnera*	54) *Platea*	55) *Engelhardtia*

Lecythidaceae	Loganiaceae	Loranthaceae	Lythraceae	
56) *Barringtonia*	57) *Geniostoma*	58) Loranthaceae	59) Lythraceae	

Malvaceae		Meliaceae		Menispermaceae
60) *Brownlowia*	61) *Durio*	62) *Aglaia*	63) *Dysoxylum*	64) Menispermaceae

Menispermaceae	Moraceae/Urticaceae	Moraceae	Myricaceae	Myrtaceae
65) *Stephania*	66) Moraceae/Urticaceae	67) *Ficus*	68) *Myrica*	69) Myrtaceae

Pandanaceae		Paracryphiaceae	Phyllanthaceae	
70) *Freycinetia*	71) *Pandanus*	72) *Quintinia*	73) *Antidesma*	74) *Aporosa*

Phyllanthaceae			Picrodendraceae	Piperaceae
75) *Bischofia*	76) *Breynia*	77) *Glochidion*	78) *Austrobuxus*	79) *Peperomia*

Plantaginaceae		Poaceae		Podocarpaceae
80) *Callitriche*	81) *Plantago*	82) grain <40 µm	83) grain >40 µm	84) *Dacrycarpus*

Podocarpaceae		Pontederiaceae	Primulaceae	Rhizophoraceae
85) *Phyllocladus*	86) *Podocarpus*	87) Pontederiaceae	88) *Myrsine*	89) Rhizophoraceae

Rubiaceae

90) *Hedyotis*
91) *Nauclea*
92) *Neonauclea*
93) *Randia*

Rutaceae

94) *Zanthoxylum*

Sapindaceae

95) *Allophylus*
96) *Pometia*

Sapotaceae

97) *Palaquium* type
98) Sapotaceae

Stemonuraceae

99) Stemonuraceae

Styracaceae

100) *Styrax*

Symplocaceae

101) *Symplocos*

Theaceae

102) *Gordonia*

Typhaceae

103) *Typha*

Violaceae

104) Violaceae

Glomeraceae

105) *Glomus*

Tilletiaceae

106) *Tilletia*

Botryococcaceae

107) *Botryococcus*

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Aspleniaceae			Blechnaceae	
108) *Asplenium*	109) *Asplenium belangeri*	110) *Asplenium macrophyllum*	111) *Blechnum*	112) *Stenochlaena areolaris*

Blechnaceae	Cyatheaceae	Davalliaceae	Dennstaedtiaceae	Dicksoniaceae
113) *Stenochlaena palustris*	114) Cyatheaceae	115) *Davallia*	116) *Microlepia*	117) *Dicksonia*

Dryopteridaceae			Hymenophyllaceae	Hypodematiaceae
118) *Arachniodes*	119) *Elaphoglossum*	120) *Polystichum*	121) Hymenophyllaceae	122) *Leucostegia*

Lycopodiaceae		Lygodiaceae	Marattiaceae	Nephrolepidaceae
123) *Huperzia*	124) *Lycopodiella cernua*	125) *Lygodium*	126) *Angiopteris*	127) *Nephrolepis*

Ophioglossaceae	Polypodiaceae		Pteridaceae	
128) *Ophioglossum*	129) *Microsorum heterocarpum*	130) Polypodiaceae	131) *Adiantum*	132) *Pteris*

Selaginellaceae	Thelypteridaceae	
133) *Selaginella*	134) *Amauropelta*	135) Woodsiaceae

Appendix B

Complete pollen, spores, NPP and charcoal records

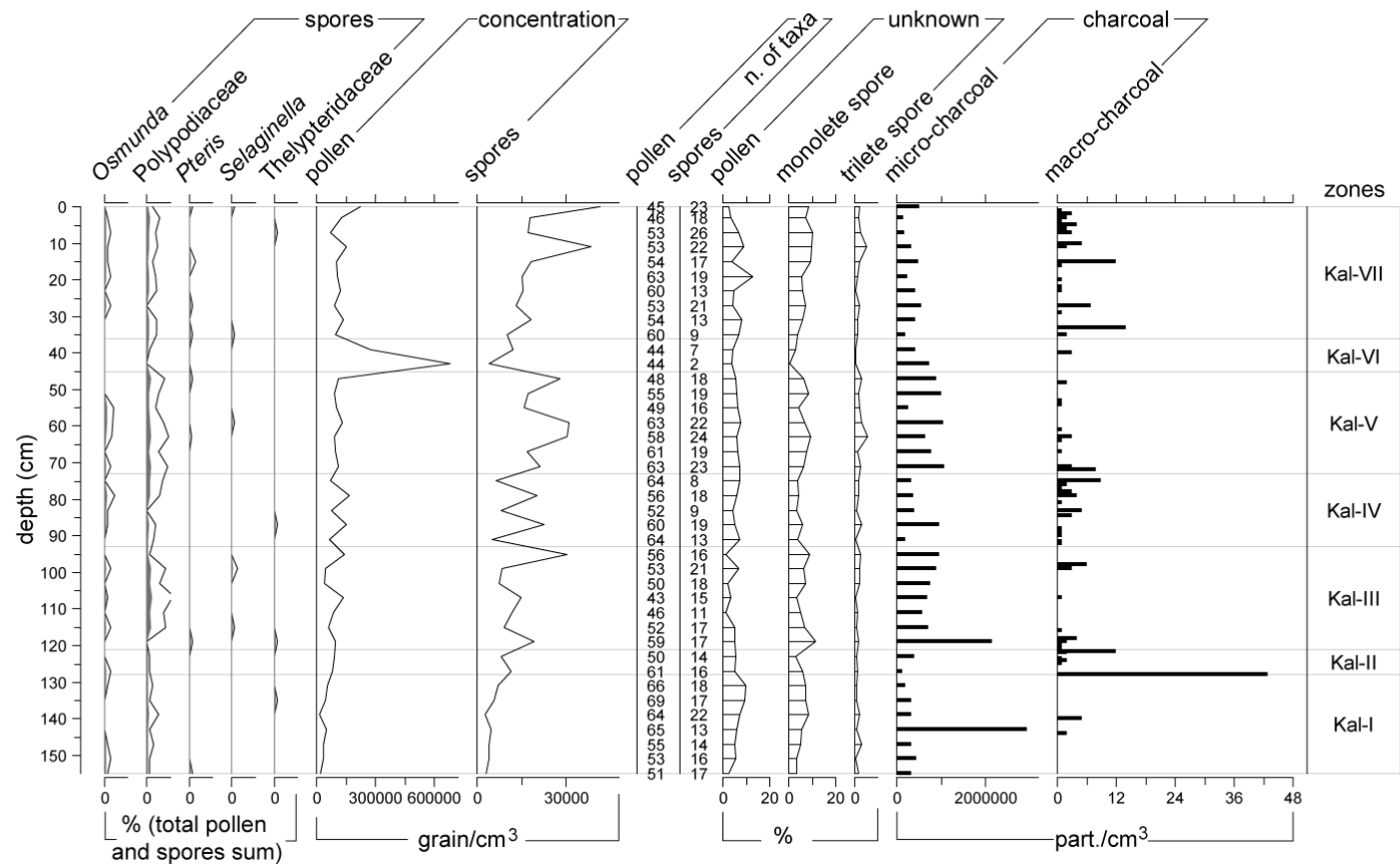Outline curves showing an exaggeration of 5x to highlight variations of the taxa
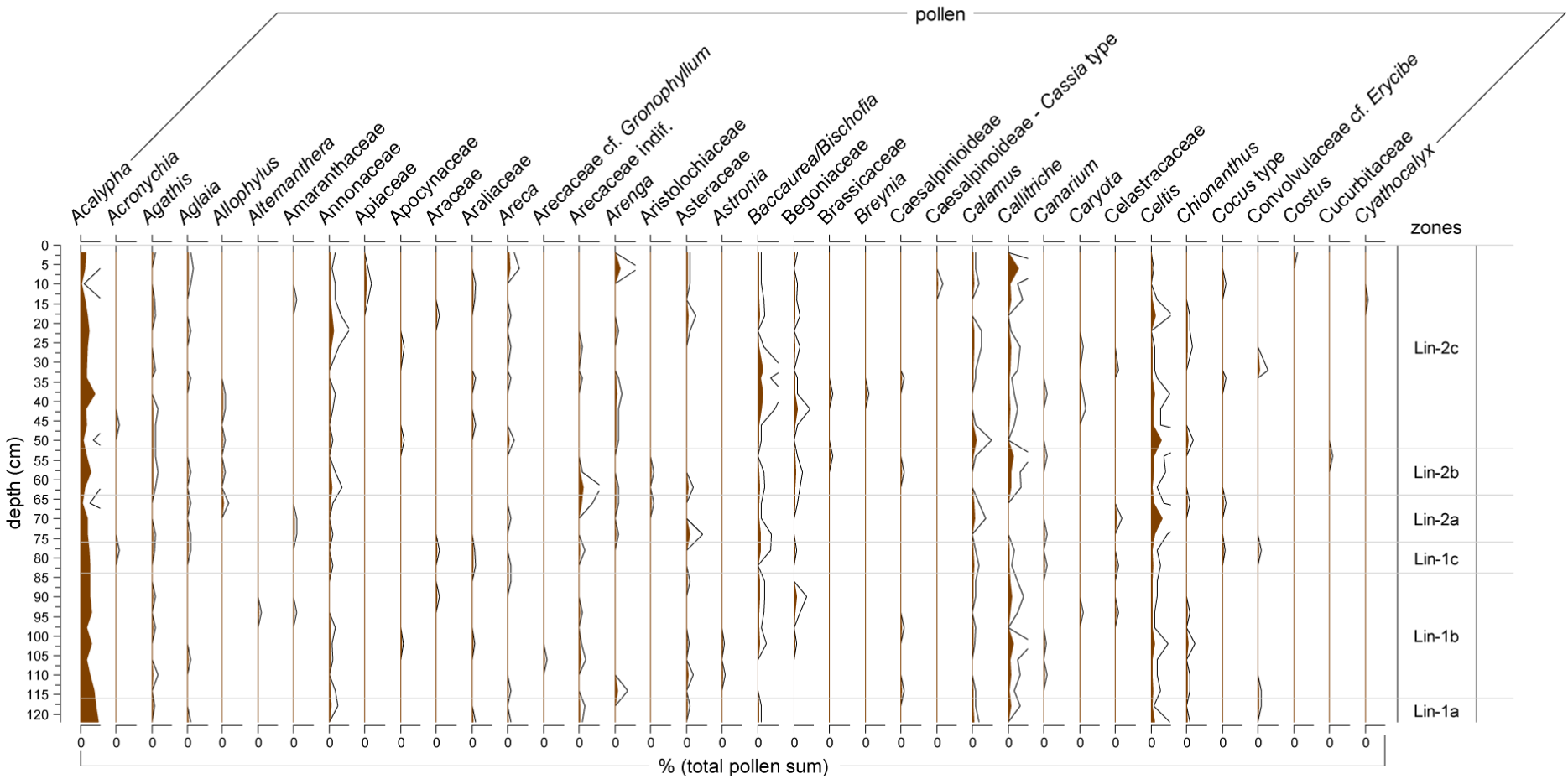
Lake Kalimpaa
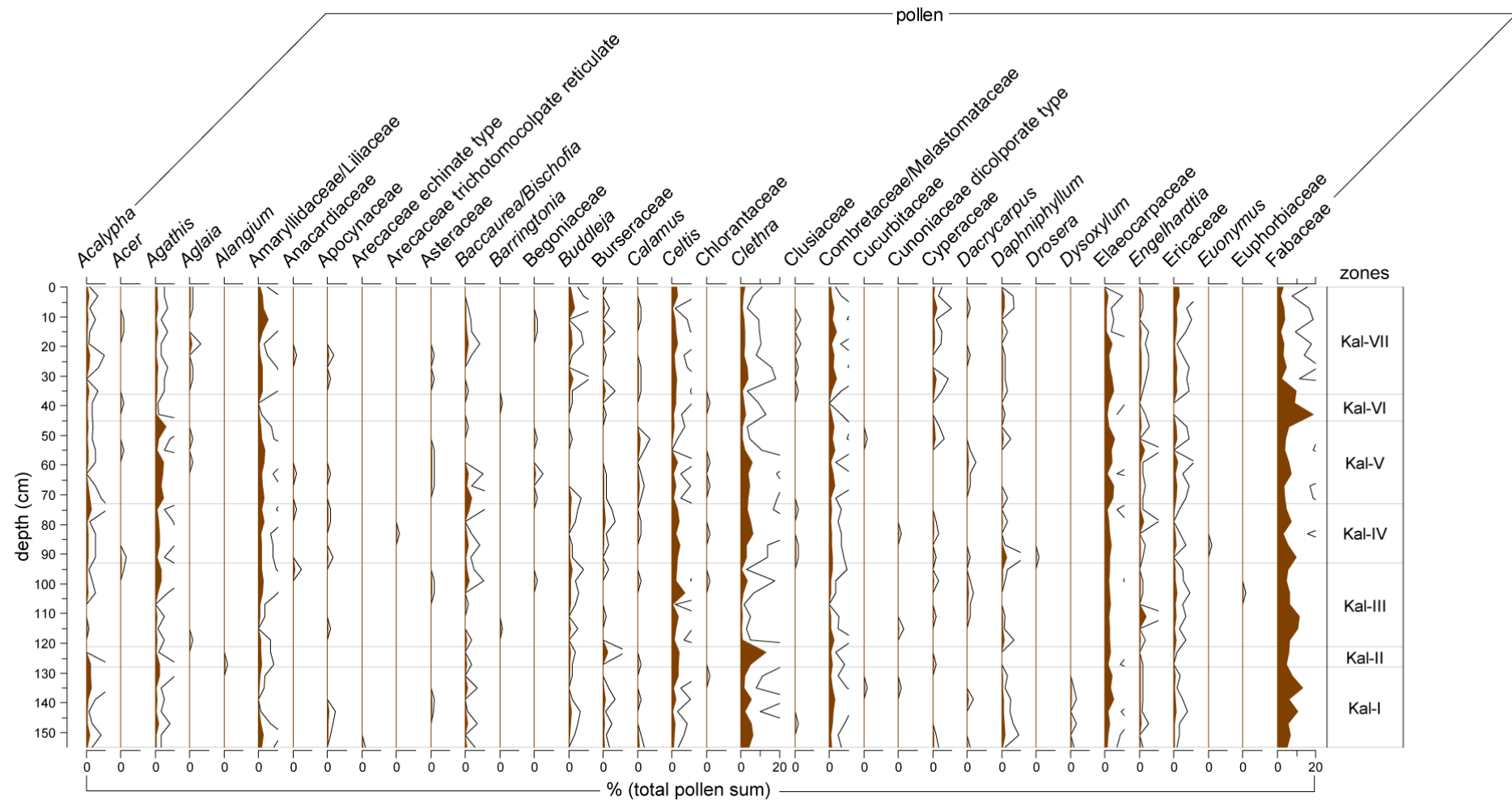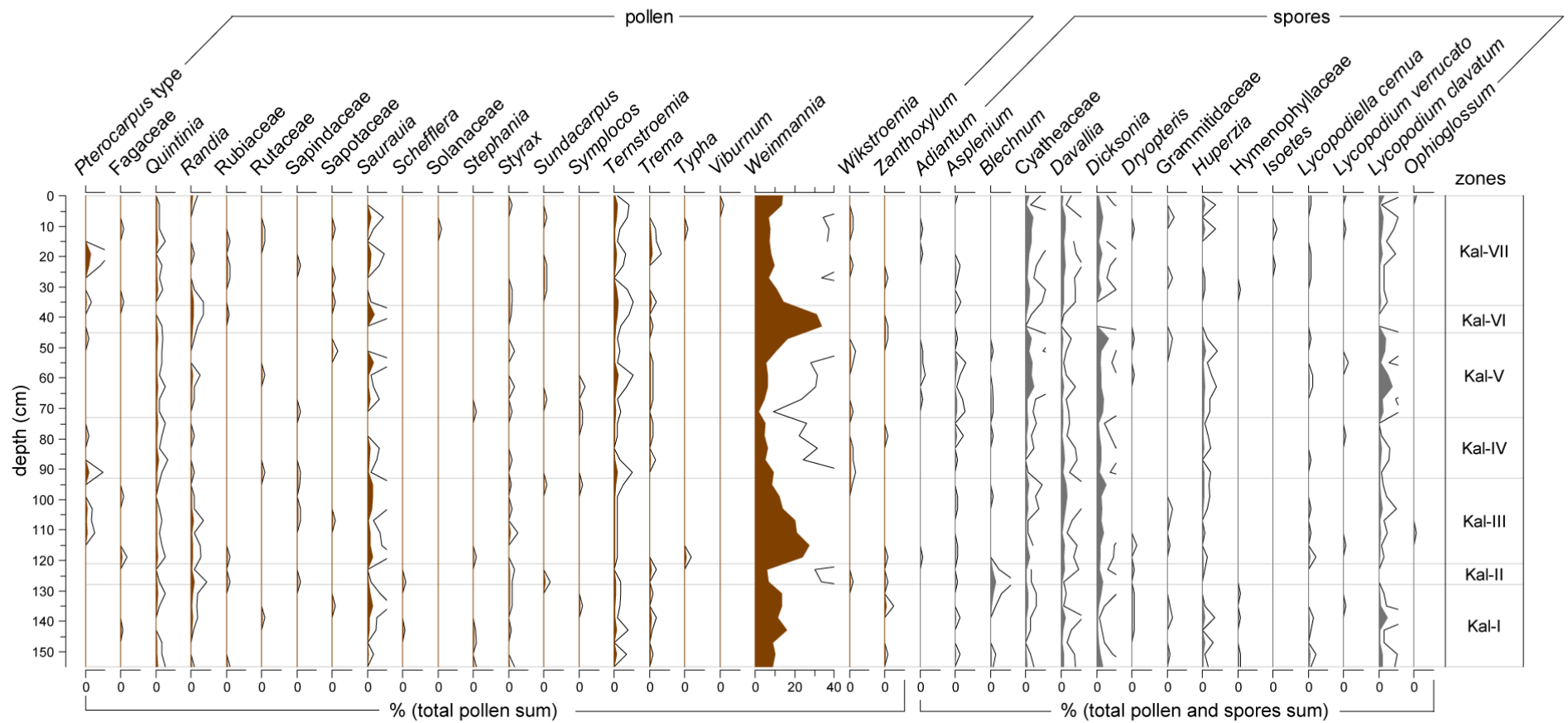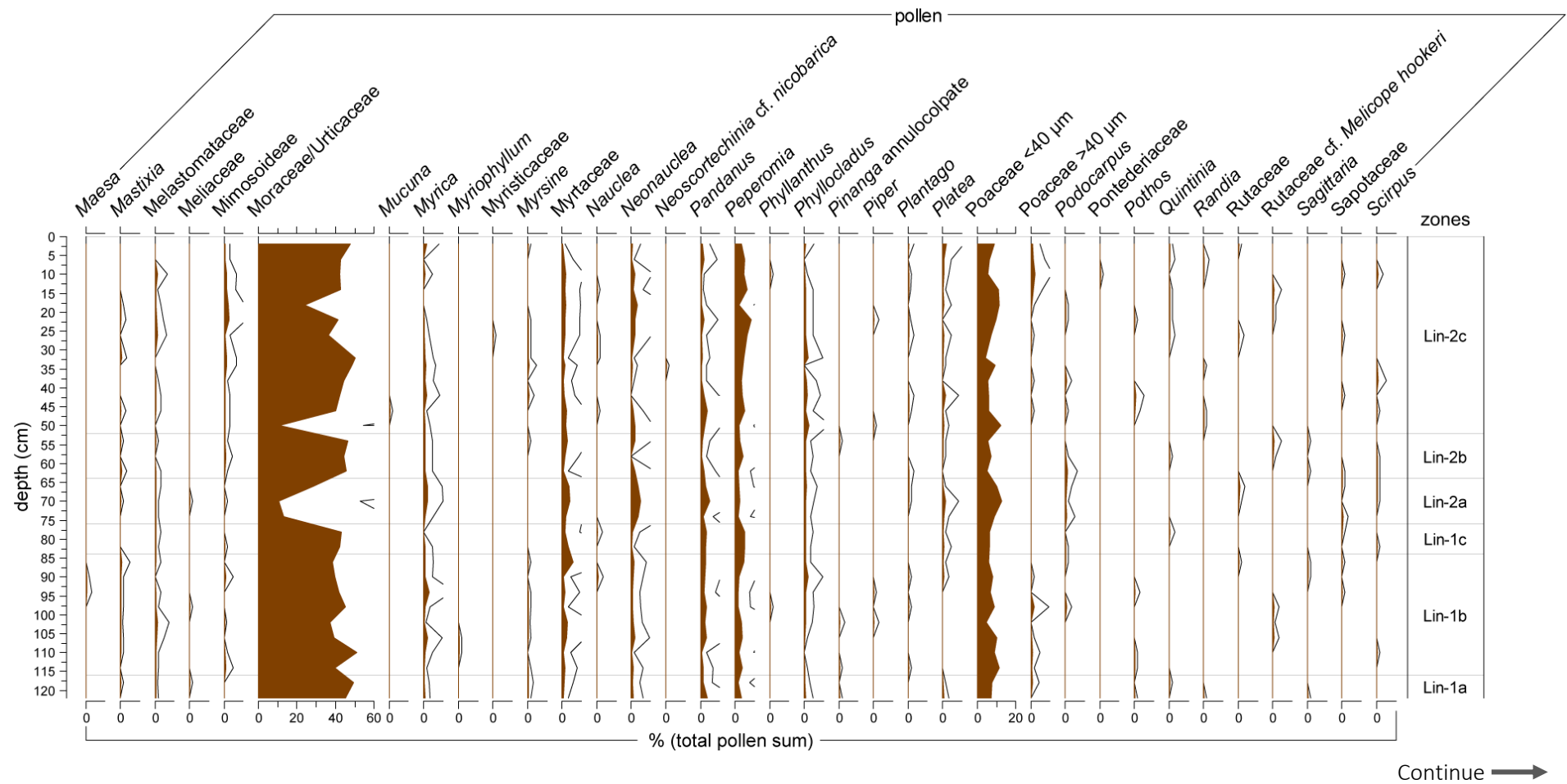
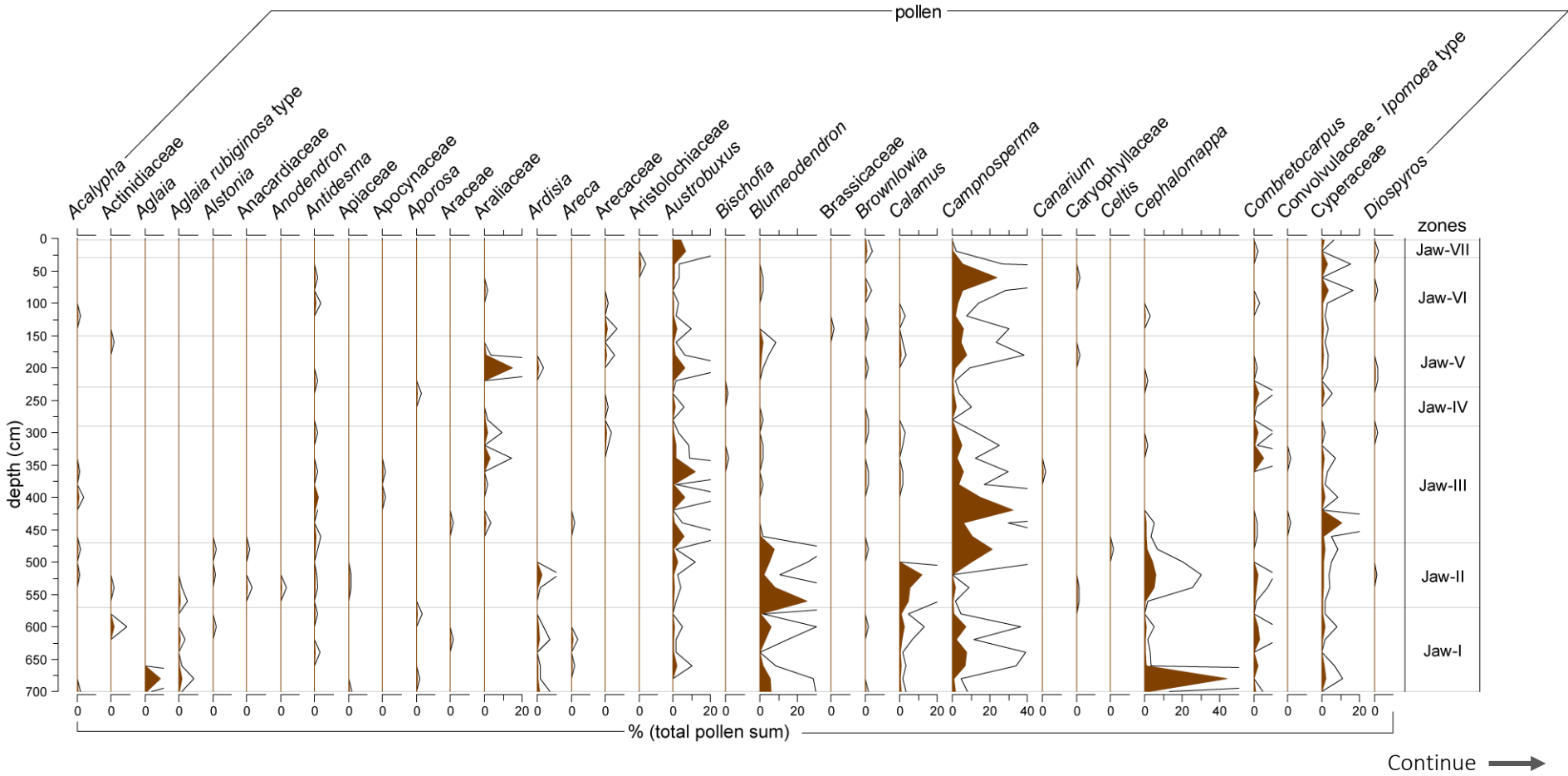LLNP, Central Sulawesi (Chapter 6)

Lake Lindu

LLNP, Central Sulawesi (Chapter 7)

Jaw SPT, Air Hitam

Jambi Province, Sumatra (Chapter 8 and Chapter 9)

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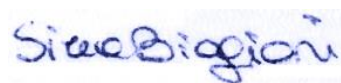pollen

% (total pollen sum)
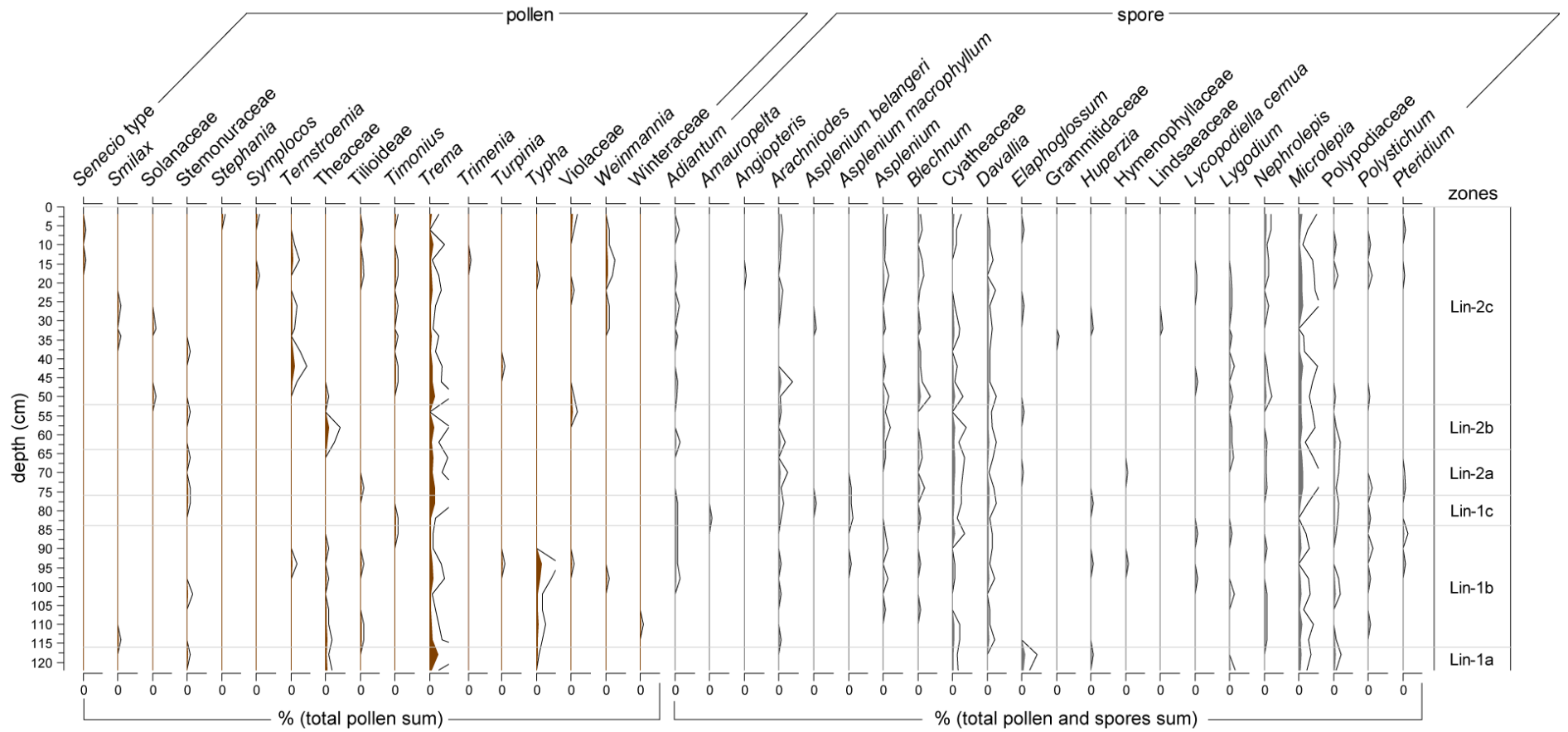
Continue →
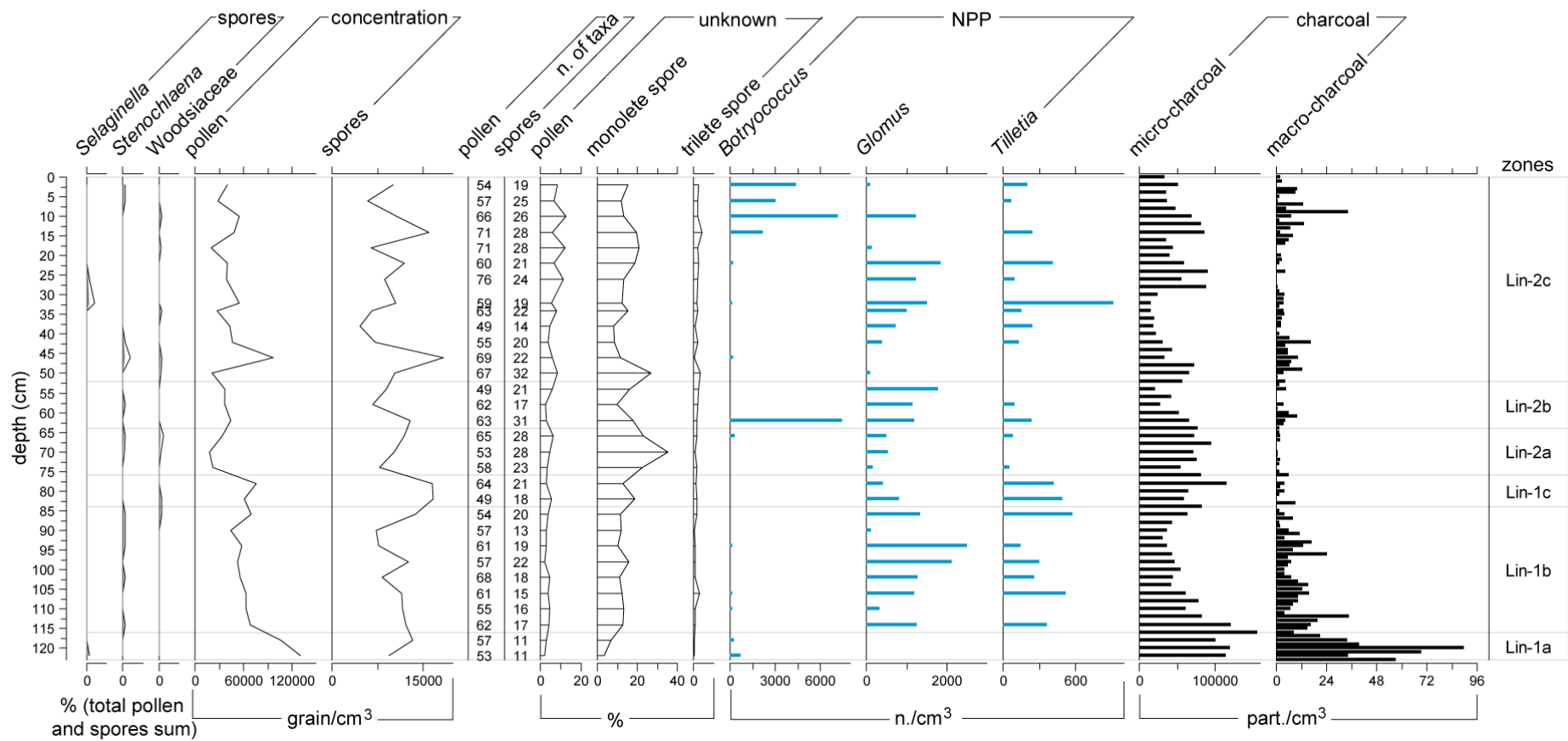
199
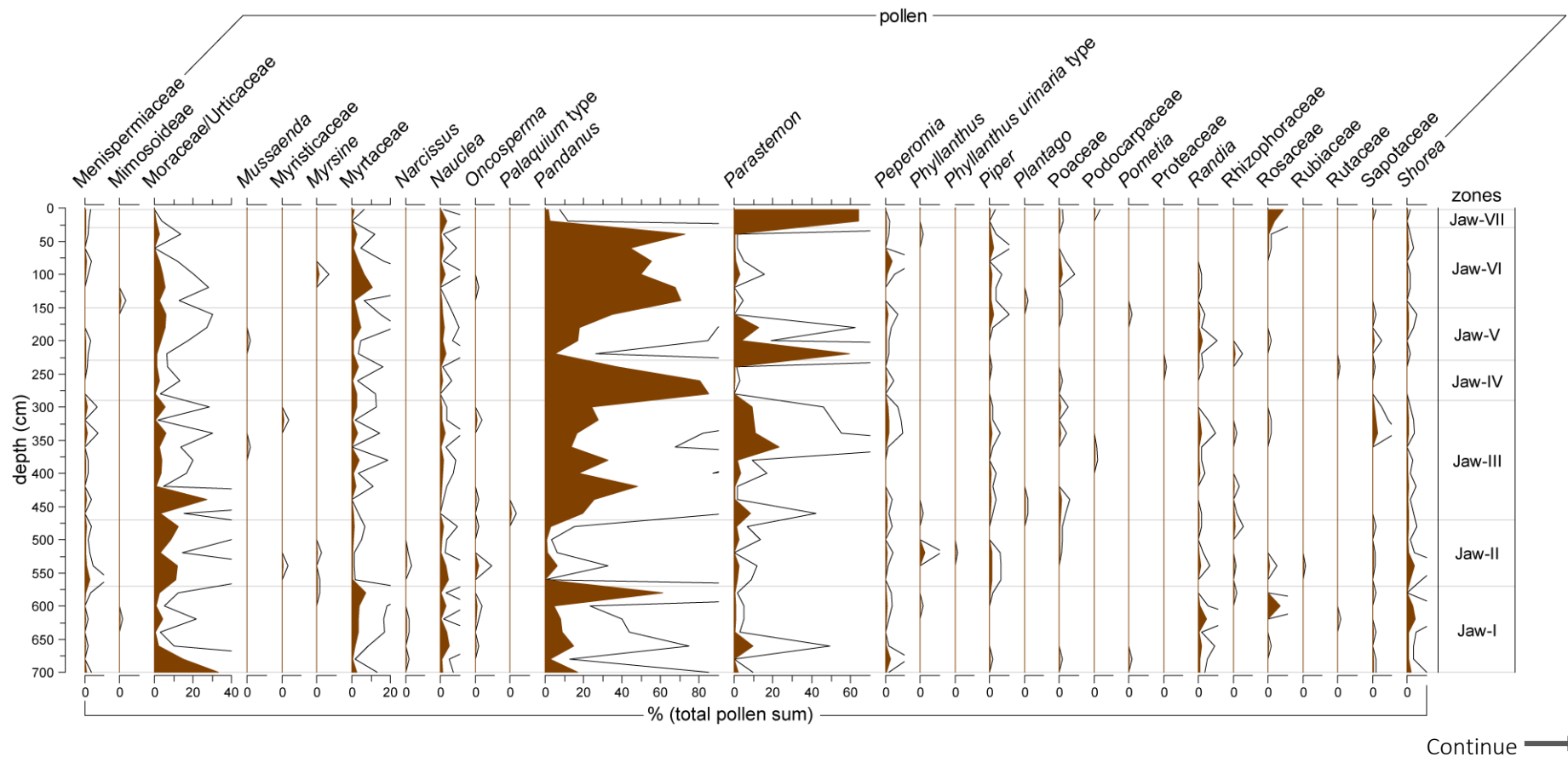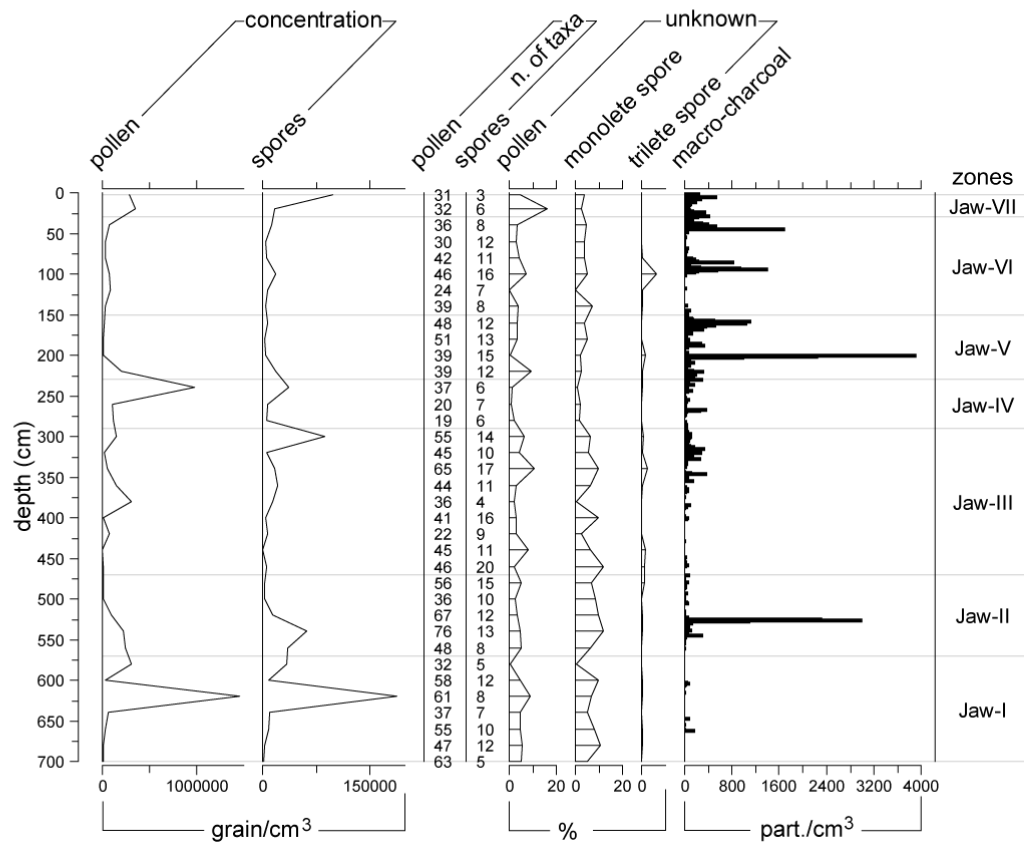
Declaration of Academic Integrity

I hereby confirm that the present dissertation is solely the work of myself. All scientific collaborators appear as co-authors of the manuscripts. If any passages or figures/diagrams from books, papers, the Web or other sources have been copied or in any other way used, all references, including those found in electronic media, have been acknowledged and fully cited.

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Siria Biagioni

Göttingen, May 5th, 2016