

**Rainfall and Vegetation of Indonesia
during the Late Pleistocene and Holocene**

by

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Wicaksono, S. A.; Russell, J. M.; Bijaksana, S. 2015. Compound-specific stable isotope records of vegetation and hydrologic change in central Sulawesi since 53,000 yr BP. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 430:47-56, DOI:10.1016/j.palaeo.2015.04.016

Hunt, G.; Wicaksono, S. A.; Brown, J. E.; and MacLeod, K. G. 2010. Climate-driven body-size trends in the ostracod fauna of the deep Indian Ocean, *Palaeontology*, 53(6):1255-1268

Sukmono, I.; Wicaksono, S. A.; Pramudya, Y.; and Riady, J. (eds.) 2010. *The East Coast Nuance: Thoughts and campus notes by Indonesian scholars in the East Coast*. New York: The Consulate General of the Republic of Indonesia.

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Speak to the Past and It Shall Teach Thee

—Inscription on the John Carter Brown Library, Brown University

Hujan Bulan Juni

Rain in June

<i>tak ada yang lebih tabah</i>	none the steadfast
<i>dari hujan bulan juni</i>	but rain in June
<i>dirahasiakannya rintik rindunya</i>	where the longing for flowering trees
<i>kepada pohon berbunga itu</i>	hidden among its drops

<i>tak ada yang lebih bijak</i>	none the wiser
<i>dari hujan bulan juni</i>	but rain in June
<i>dihapusnya jejak-jejak kakinya</i>	when its unsteady footsteps
<i>yang ragu-ragu di jalan itu</i>	deleted without hesitation

<i>tak ada yang lebih arif</i>	none so enlightened
<i>dari hujan bulan juni</i>	but rain in June
<i>dibiarkannya yang tak terucapkan</i>	who let the unspoken absorbed
<i>diserap akar pohon bunga itu</i>	by the roots of the flowering tree

Sapardi Djoko Damono (1989)

translated by George Sicillia

In memory of my loving grandfathers,
who dedicated their lives to society and family.

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CHAPTER ONE

Introduction

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1. Tropical paleoclimatology

A comprehensive discourse on global climate change should include an assessment of the tropical climate mean state and its variability. The tropics receive most of the Earth's incoming solar insolation and subsequently release the energy as latent heat that drives the global atmospheric circulation and energy cycle (Peixoto and Oort 1992, Qu et al. 2005). Warm waters in the tropics serve as a major source of global water vapor, energizing the Earth's water cycle (Meehl 1987). Further, climate anomalies in the tropics affect climate around the globe through atmospheric teleconnections (Cane and Clement 1999). A firmer understanding of tropical climatology will therefore shed light on feedbacks associated with global water and energy cycles, which will lead to better predictions of climate change regionally and globally. However, direct observations of climate in the tropics are limited to only the last two centuries, and often even less. To capture a more complete spectrum of tropical climate variations and expand our frame of reference regarding current anthropogenic climate change, we need to study the climates of time periods prior to instrumental observations (i.e. paleoclimatology) using indirect methods (i.e. climate proxies), such as fossil pollen and plankton in sediment cores, isotopic and elemental ratios in corals and cave stalagmites, and tree growth ring data.

Within the tropics, the Earth's three main convective centers are located over central Africa, the Amazon, and the Indonesian archipelago. The monsoon region of Indonesia is the largest of these, and is the most dominant source of convective activity on Earth (Meehl 1987). This region, situated over the Indo-Pacific Warm Pool, persistently has sea surface temperatures (SST) higher than 28°C, making its seawaters

the warmest in the world. The Indonesian archipelago, comprised of thousands of islands of various sizes, mountainous terrain, narrow peninsulas, and many shallow seas (Figure 1), is also known to climatologists, meteorologists, and oceanographers as the Maritime Continent, reflecting the extensive climatic interactions between sea and land occurring across the region. The warm waters of the Maritime Continent, the tropical heating of land surfaces, and the upward motion of warm air forced by mountainous terrains generate extensive buoyant and moist convection in this region (Ramage 1968, Dayem et al. 2007). This convection transfers large amounts of energy to the upper-atmosphere and contributes to both the meridional (north-south) circulation toward the extra-tropics (i.e. the Hadley Circulations) and the zonal (east-west) circulation within the tropics (i.e. the Walker Circulations; Neale and Slingo 2003). Indeed, the Maritime Continent has been described as an atmospheric boiler box for its central role in the global circulation system (Ramage 1968), highlighting the necessity to rigorously assess its past and present climates. A thorough understanding of key mechanisms controlling the archipelago's climate variations through time is particularly useful to constrain global climate model projections of future changes.

2. Indonesian rainfall: the past is the key to the present and future

The most important climate variable in Indonesia is rainfall. Given the large amount of convection, Indonesia has among the highest annual rates of precipitation in the world. However, significant variations in average rainfall, from 4 to more than 10 mm/day (Figure 2) are found on a range of temporal and spatial scales, reflecting the

complexity of regional convective activity. Present-day rainfall variability in Indonesia is influenced by various interrelated components, including the Intertropical Convergence Zone (ITCZ) , Australian-Indonesian Monsoon (AIM), regional SST, El Niño Southern Oscillation (ENSO), and Indian Ocean Dipole (IOD). On the annual scale, the lowest precipitation is found in southeastern Indonesia, i.e. above Nusa Tenggara, in part because the modern-day mean position of the ITCZ is slightly north of the equator. On the seasonal scale, the dominant control of rainfall variability is the AIM, which is highly influenced by the latitudinal position of the ITCZ (Aldrian and Susanto 2003). Most of Indonesia experiences a distinct rainfall peak during December-February, when seasonal heating of the southern hemisphere causes the ITCZ to shift south of the equator associated with a strengthening of northwesterly flow in the AIM (Figure 3). The northwest monsoon brings a considerable amount of moisture from the South China Sea, western Pacific Ocean, and eastern Indian Ocean to most of Indonesia. During June-August, most of Indonesia experiences a dry season, when the ITCZ lies north of the archipelago, and strong monsoonal winds blow from the southeast, carrying relatively dry air masses from Australia and its vicinity (Figure 3). Exceptions are found in a few regions, particularly in the northwestern part of Sumatra and eastern Kalimantan, which have two rainfall maxima due to the double passing of the ITCZ each year (September-November and March-May) (Aldrian and Susanto 2003). A local/antimonsoonal rainfall pattern is found in the Moluccas and parts of Papua, characterized by a single precipitation peak in June to July and a minimum peak in November to February, i.e. the opposite of the predominant monsoonal rainfall pattern, likely due to local orographic features (Aldrian and Susanto 2003).

The interannual variability of rainfall is mostly determined by ENSO (e.g. Hendon 2003, Kirono et al. 1999) and IOD (Saji and Yamagata 2003). ENSO is a quasiperiodic mode of variability that involves coupled changes in the SST of the tropical Pacific Ocean and air surface pressure in the tropical western Pacific via the Walker Circulation. The warm SST phase in the eastern Pacific, El Niño, accompanies high surface pressure in the western Pacific, while the cold SST phase, La Niña, accompanies low air surface pressure in the western Pacific. During an El Niño event, these changes in SST and pressure cause the zonal Pacific Walker Circulation to weaken, which reduces rainfall over Indonesia due to the formation of an atmospheric subsidence zone. The Indian Ocean Dipole also affects interannual rainfall variability in Indonesia, and is characterized by SST anomalies between the west and east/southeast Indian Ocean. During the positive episode of IOD, SST in the western part of the Indian Ocean is warmer than usual, while the east/southeast Indian Ocean becomes cooler. As such, rainfall tends to decrease over some parts of Indonesia. Any variation in these interrelated climate systems likely leads to steep regional gradients in rainfall, with potential negative repercussions to humans.

From the humanitarian and economic points of view, changes in the mean state and variability of rainfall in Indonesia affect the life and livelihoods of more than a quarter billion people living in the archipelago. Indeed, between 1980 and 2007, more than 174,000 Indonesians died in rainfall and drought-related disasters (CRED 2010). This death toll was the second highest in the world. Deviations from the twentieth-century averages of rainfall, and the length of dry and rainy seasons due to variations in the AIM, ENSO and IOD may impact a variety of economic domains, including

agriculture, forestry, fisheries, transportation, and energy (e.g. Boer and Subbiah 2005, Naylor et al. 2007, Oktaviani et al. 2011). For example, a significant increase in rainfall intensity and frequency during the rainy season may lead to a higher risk of flooding and landslides, whereas delayed rainfall and a longer dry season in rice-producing regions mean less yield and more import of rice. During an unusually strong 1997-1998 El Niño event, Indonesia experienced a major shortfall of rice production, forcing imports to a record 5.8 million tons in 1998, the largest quantity of rice ever imported by a single nation (Baldwin et al. 2012). The El Niño-induced drought also exacerbated forest fires in Kalimantan and Sumatra in 1997-1998, triggered primarily by illegal rainforest and peatland clearing. Smoke and haze from the fires shrouded much of Sumatra, Singapore, the Malaysian Peninsula, and Kalimantan, at a total cost of some US\$4.4 billion (Glover 2001). The total economic costs of the 2015 fires, which coincided with a strong El Niño, might have cost the Indonesian government US\$30 billion up until October 2015 (CIFOR 2015). Further, drought-driven fires reduced the nation's effort to cut total greenhouse gas emissions. During certain days, Indonesia's daily emissions from the forest fires were larger than the daily emissions from the entire United States economy (World Resources Institute 2015).

These negative impacts of rainfall changes underline the importance of delivering improved regional rainfall projections in order to enhance adaptive capacity and improve societal resilience to climate change. The recent Fifth Assessment Report by the Intergovernmental Panel on Climate Change relied heavily on climate simulations from more than thirty state-of-the-art Global Circulation Models (GCMs) to assess future changes under different scenarios of fossil fuel use. The simulation results indicate that

by 2100 the temperature across virtually all parts of Indonesia will increase by 2 to 4 °C and the annual precipitation may increase across the archipelago, except around Nusa Tenggara and part of the Moluccas (IPCC 2013). However, while most models agree on the trend and magnitude of temperature changes, there is little agreement on the trend and magnitude of rainfall changes (IPCC 2013). The fairly large spread among models in the prediction of Indonesian rainfall indicates uncertainties in the simulation of rain-generating convective processes and difficulties arising from the complexity of Indonesian precipitation and geographical details (e.g. Neale and Slingo 2003, Jourdain et al. 2013). The challenge of successfully simulating Maritime Continent precipitation therefore serves as an excellent test for GCMs, as successful simulations imply that the model can correctly represent atmospheric activity on a range of temporal and spatial scales. Long-term reconstructions across different parts of Indonesia and from different climate boundary conditions than the present provide a useful framework to tackle this challenge and improve the efficacy of GCMs.

The reconstructions of past rainfall from the Late Pleistocene-Holocene may provide the best benchmark to assess natural tropical climate variability. The Late Pleistocene includes the Last Glacial Maximum (LGM, ca. 26-19 kyr BP), when the global climate system was subject to large radiative perturbations and the Earth had very different climate boundary conditions than today. The last deglaciation (19-11 kyr BP) is one of the most dramatic periods of global climate change in the recent geologic past, and is the most recent time period when the Earth's climate warmed due to large changes in atmospheric greenhouse gas concentrations. The Holocene (post-11 kyr BP), in contrast, is relatively stable and serves as the baseline conditions prior to recent anthropogenic

climate change. While the temperature history of Indonesia is relatively well-known during these time periods, Indonesian rainfall history during the Late Pleistocene-Holocene remains unclear due to the limited number of proxy records and significant disagreement among proxy reconstructions in the direction and magnitude of regional precipitation change (Reeves et al. 2013 and references therein).

At the heart of this disagreement is the state of Indonesian rainfall during the LGM. Pollen records generally indicate drier conditions prevailed in the Indonesian archipelago during the LGM (Reeves et al. 2013), though some pollen records from some areas within the archipelago do indicate wetter conditions instead (Newsome and Flenley 1988, Kershaw et al. 2011, Reeves et al. 2013). Similarly, while proxies of terrigenous input in many areas indicate strongly reduced precipitation during the LGM (e.g. Russell et al. 2014), similar proxies from other parts of Indonesia indicate little evidence for large LGM-Holocene trends (e.g. Mohtadi et al. 2011), or even wetter LGM conditions (Shiau et al. 2011, Tachikawa et al. 2011). In recent years, many of the climate reconstructions in this region rely on proxies for changes in the isotopic composition of rainfall. Many past studies often conclude that the primary control of rainfall isotopes in the tropics is the ‘amount effect’, which describes the negative correlation between the isotopic composition of rainfall and rainfall amount, resulting from Rayleigh fractionation during rainout events (Dansgaard 1964, Rozanski et al. 1992). Speleothem-based rainfall $\delta^{18}\text{O}$ values during the LGM from Borneo and Flores suggest that the sites were either wetter or have similar precipitations to today (Partin et al. 2007, Griffiths et al. 2009). However, the interpretation of past variations in tropical rainfall isotopes as an indicator of changes in the rainfall amount is widely debated (Sachse et al. 2012 and references therein),

compounding the uncertainty in regional paleoclimate.

These discrepancies in the direction, magnitude, and interpretation of regional precipitation change have led to multiple competing hypotheses for the primary mechanisms controlling Indonesian rainfall on orbital-scales. These hypotheses include changes in atmospheric greenhouse gas levels (e.g. Tachikawa et al. 2013), orbital variations (e.g. Holbourn et al. 2005, Kershaw et al. 2011, Tierney et al. 2012, Carolin et al. 2013), shifts in sea surface temperatures (e.g. Hope et al. 1976, Hope 2007), ice volume changes (e.g. Lee et al. 2014, Russell et al. 2014), as well as regional changes related to the aerial exposure of the Sunda and Sahul Shelves (Griffiths et al. 2009, Partin et al. 2007, diNezio and Tierney 2013, Dubois et al. 2014, Niedermeyer et al. 2014). Similarly, at the other end of the temporal spectrum, Late Pleistocene-Holocene Indonesian rainfall is also affected by shorter-term millennial-scale climate oscillations and major volcanic eruptions, though the exact circumstances, amplitudes, and mechanisms through which these events influence regional climate are still widely debated (e.g. Robock et al. 2009, Anchukaitis et al. 2010, Muller et al. 2012). In light of existing uncertainties regarding Indonesian paleoclimatology, this thesis should help answer the following questions:

- 1) How has the Indonesian rainfall varied during the Late Pleistocene-Holocene and what climate forcing(s) caused those variations? (Chapters 2-4)
- 2) What are the primary climatic processes controlling rainfall isotopes in Indonesia on orbital scales? (Chapters 3-4)
- 3) Can a major volcanic eruption generate strong, long-term climatic impacts in the region? (Chapter 5)

3. Indonesian vegetation in the past, present, future

The distribution of vegetation in Indonesia is strongly regulated by the amount of rainfall and its seasonality, as well as by temperature, which varies with elevation (Peel et al. 2007). In the present, the high annual precipitation rates and warm temperatures, along with low precipitation seasonality observed across most of Indonesia, give rise to widespread lowland ever-wet or evergreen rainforests below the elevation of 1000 m (Göltenboth et al. 2006). This forest type grades to montane cloud forests and subalpine/alpine vegetation at colder, higher elevation regions. Such gradations in forest types are common on major Indonesian islands including Sumatra, Java, Kalimantan, Sulawesi and Papua. In waterlogged lowland areas, typical of Sumatra's eastern coastlands, southern Kalimantan, and southern Papua, the rainforests are replaced by swamp forests, and along sheltered shores on many Indonesian islands, the swamp forests and rainforests are replaced by mangroves (Göltenboth et al. 2006). In contrast to these "wet" vegetation assemblages, grasslands are regularly seen where rainfall is lower and precipitation seasonality is higher, such as on Nusa Tenggara, the Lesser Sunda Islands, southeastern Java, and southeast Papua (Peel et al. 2007). The strong linkage between precipitation and vegetation distribution in Indonesia is particularly evident in the Koppen-Geiger climate classification map, which is partly derived from the native vegetation cover across the archipelago (Figure 2).

Changes in land use have rapidly reduced the extent of pristine land cover in the country. Though humans have for centuries logged and cleared primary rainforests for

agriculture in the wetter, densely populated regions such as Java and Bali, forests still represented 84 percent of Indonesia's total land area as late as 1900 (Holmes 2000). However, by 2013, forest cover was only 51 percent of the total land area (World Bank 2015). The rate of deforestation has accelerated since the 1970s, particularly in Sumatra, Kalimantan, and Sulawesi, due to intensified logging and the burning of forests to clear land for cultivation or plantation, most of which were conducted illegally (Holmes 2000). The massive conversion, degradation, and fragmentation of Indonesian forests is causing biodiversity decline of the rainforest ecosystem. For many flora and fauna taxa inhabiting the forests, the movement towards their extinction might be irreversible, as there are few refugia left from which to replenish their species.

Climate change might also produce detrimental impacts on biodiversity, although many of the mechanisms through which climate change will affect Indonesian rainforests are still poorly understood. Studies in other parts of the tropics suggest that global warming will increase mortality rates of certain tropical forest species during increasingly common heat waves (Allen et al. 2010), encourage the spread of diseases among many taxa (Sturrock et al. 2011), and exacerbate the risk of extinction for high-elevation species and those with limited climatic ranges (Foster 2001). Further, rising temperatures might elevate the cloud base of tropical mountains, which could have a profound impact on montane rainforests in need of persistent cloud cover (Bruijnzeel 2001). A world with more El Niño events will induce droughts that may increase tree mortality and flammability, therefore exacerbating the effects of forest fires (Siebert et al. 2001). Indeed, increasing variability in future rainfall intensity and frequency is predicted to decrease the extent of ever-wet zones in Indonesia (Underwood et al. 2014). Importantly,

these vegetation changes might feed back to the climate system. A reduced extent of tropical rainforests may decrease evapotranspiration, with important implications for water and energy balance (Olchev et al. 2008). Given predicted changes in terrestrial vegetation due to anthropogenic activities and climate change, it is critical to understand the long-term stability and sensitivity of Indonesian rainforest ecosystems.

The nature of Late Pleistocene-Holocene terrestrial landscapes across Indonesia is largely unknown, and the limited evidence is contradictory, in particular for the LGM. Palynological or pollen records from southeastern Indonesia (Kershaw et al. 2011, van der Kaars et al. 2000) and Java (van der Kaars et al. 2001) indicate expansion of grasses and more open vegetation during the LGM, usually associated with a drier climate. However, evidence for continuous ever-wet or closed-canopy vegetation during the LGM exist from sites in Sumatra (Newsome and Flenley 1988), Papua (Hope and Tulip 1994), and Kalimantan/Borneo (Anshari et al. 2004, Wurster et al. 2010). These observations often contrast with paleoclimate studies; for instance, speleothem isotopic records (Ayliffe et al. 2013) suggest a wetter climate accompanied grassland expansion in southeast Indonesia. Modeling studies indicate an even greater extent of rainforest across Sundaland during the LGM due to sea level declines (Cannon et al. 2009, Woodruff 2010), leading to the hypothesis that Indonesian rainforests are currently in a geographic refugial state, due to the restricted island geography under present-day high sea levels. More vegetation records, especially from the regions where existing data are scarce, are needed to assess the nature of Indonesian vegetation change through time.

In view of these challenges, new data presented in thesis shall help answer the following questions:

- 4) How does Indonesian vegetation respond to Late Pleistocene-Holocene rainfall changes? (Chapters 2-4)
- 5) Are there any differences in vegetation response to precipitation change across spatial and altitudinal gradients? If so, why? (Chapters 2-4)

4. Thesis outline and highlights

This thesis is devoted to closing our knowledge gaps in the states of Indonesian rainfall and vegetation and their governing climatic mechanisms during the Late Pleistocene, in particular during the LGM, and Holocene. The geographic foci of this thesis are central and eastern Indonesia, where existing paleorainfall and paleovegetation reconstructions are sparse and often contradictory. In this thesis, chapters 2-5 have each been written in the form of self contained papers that have been published or will be submitted for peer-reviewed publication.

Chapter 2 presents a discussion of hydrological and vegetation records in central Sulawesi over the past 50 kyr BP based on lacustrine records from Lake Matano. Leaf wax carbon isotope records ($\delta^{13}\text{C}_{\text{wax}}$) indicate strong glacial-interglacial climate changes, with drying and forest opening around the catchments of Lake Matano and nearby Lake Towuti during Marine Isotope Stage 2 / LGM. This orbital-scale variation in central Sulawesi vegetation and hydroclimate appears most strongly correlated to ice volume changes. However, Lake Matano experienced less reduction in precipitation compared to Towuti, which sits at a lower elevation, suggesting the importance of altitudinal difference in controlling rainfall and vegetation changes in Indonesia. In particular, the

high mountains of the Maritime Continent likely served as glacial refugia for rainforests.

Chapter 3 reviews hydrological and vegetation changes in central Indonesia or the Wallacean region since the LGM, and presents new $\delta^{13}\text{C}_{\text{wax}}$ as well as hydrogen isotopic data from leaf waxes ($\delta\text{D}_{\text{wax}}$) from marine cores collected offshore of west Sulawesi.

These data were compared with existing data from the region to better understand changes in mean rainfall and seasonality during the LGM. The records strongly suggest widespread drying and possibly a more seasonal climate in the Wallacean region during the LGM, similar to results from Lake Matano. Further, the leaf wax proxy records help elucidate the climate processes controlling rainfall isotope variations through time.

Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$ was anti-correlated with $\delta\text{D}_{\text{wax}}$ during the LGM and the last deglaciation, but was positively correlated during most of the Holocene, indicating time-varying controls on the isotopic composition of rainfall. This result may help to reconcile conflicting interpretations of speleothem $\delta^{18}\text{O}$ and other proxy records over the Maritime continent. The rapid inundation of the Sunda Shelf between 11 and 9 kyr BP altered the region's atmospheric circulation and modified primary controls on precipitation isotope compositions, explaining long-standing issues in the interpretation of precipitation isotope data from the region.

Chapter 4 examines the spatial variability among Indonesian rainfall records in the past 16 kyr BP. In contrast to central Indonesia (Sulawesi), $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{wax}}$ data measured on sediment cores from Cenderawasih Bay, offshore from north Papua, are strongly positively correlated throughout the past 16 kyr BP, indicating that the amount of rainfall has been the chief mechanism controlling both northern Papuan vegetation and precipitation isotopes since the last deglaciation. In contrast to records from Sulawesi, the

significant reduction in precipitation between 11 and 6 kyr BP, accompanied by an expansion of grasses, indicates that northern Papua precipitation is primarily controlled on orbital scales by southern hemisphere summer insolation and the associated meridional ITCZ movement through the tropics. Further, these proxy records inform our understanding of the interactions between humans and the Papuan environment during the Holocene.

Chapter 5 describes the discovery of a thick tephra layer in Lake Towuti in central Sulawesi, currently dated to ~74 kyr BP, as evidence for a major volcanic eruption in the region at that time. Analyses of tephra geochemistry reject our initial hypothesis that the tephra originated from the Younger Toba supereruption ca. 75 kyr BP. Instead, the data point to a large eruption likely in northern Sulawesi. Analyses of lacustrine paleoproductivity, redox-sensitive and detrital sediment indicators, and vegetation ($\delta^{13}\text{C}_{\text{wax}}$) records on sediment samples deposited immediately above and below the ash suggest that there was limited or no long-term climatic and environmental response to this major eruption and the associated tephra emplacement.

5. Broader implications and perspectives for moving forward

This thesis addresses outstanding questions in tropical paleoclimatology related to the dynamics of Indonesian rainfall. New paleovegetation and paleorainfall records from central Sulawesi and northern Papua, along with other regional records, highlight the dominant role of the ITCZ, as opposed to ENSO-like variability, in controlling regional rainfall. The mean states of Indonesian rainfall have shifted between different modes of

equilibrium in the Late Pleistocene-Holocene due to perturbations to the ITCZ imposed by changes in northern hemisphere ice volume, insolation, and Sunda Shelf exposure. Most of Indonesia was likely drier than present in the LGM due to remote high latitude forcings associated with ice volume and thickness. However, the discrepancies in precipitation between Papua and central Indonesia imply strong spatial variations in convective structure and different regional sensitivities to these climate forcings and ITCZ position within Indonesia, with important implications to the projections of future climate in the region. Further, relatively abrupt transitions into and out of the dry period were observed in Papua, suggesting the potential for a non-linear response of the regional rainfall shift to gradual changes in climate forcing. The sensitivity and non-linearity of tropical climate responses to forcings and boundary conditions need to be investigated further through regional climate models.

A major volcanic eruption such as the Younger Toba supereruption ca. 75 kyr BP may also serve as a major climate forcing. While the exact mechanisms and temporal scales of eruption-induced climatic change are still widely debated (Williams 2012), environmental proxy records from a lake in central Sulawesi suggests that regional climatic changes generated by a major eruption in north Sulawesi ca. 74 kyr BP, if any, would have a short temporal scale below the centennial scale of the proxy records. This result suggests that future studies on this topic should consider proxies and/or sites with higher temporal resolutions.

Indonesian vegetation appears to be sensitive to major precipitation changes. Rapid shifts from rainforests to grasslands (and vice versa) may accompany relatively abrupt changes from wetter to drier conditions (and vice versa), as is the case of northern

Papuan vegetation. However, further research involving high-resolution paleovegetation records from other parts of Indonesia and other intervals of major climate change are needed to better understand the threshold sensitivities of tropical vegetation and how past climate changes have structured the present-day vegetation. The local topographic setting also has a significant impact on rainfall, and therefore plant type variations, as shown in the central Sulawesi studies. In particular, mountains served as rainforest refugia during less-than-optimal climate intervals, with important implications for the maintenance and management of regional biodiversity in the present and future. Higher elevation regions should become a top conservation priority amidst the double threat of climate change and deforestation on Indonesian montane rainforests. Based on these studies, we can develop a hypothetical map of inferred vegetation distribution across central Indonesia (Figure 4), highlighting the expansion of grasslands and open vegetation type and indicating the high elevation areas that likely remained forested during the LGM.

Finally, this thesis elucidates various aspects of the application of leaf wax proxies. Marine records of leaf wax data reflect a spatially broader regional view of hydroclimatic changes compared to those from lacustrine records. The far fewer millennial- and centennial-scale precipitation pattern variations in the marine records compared to the lacustrine records might indicate a “smoothing effect” by transported leaf waxes. Further, leaf wax proxies can be considered semi-quantitative, given the non-linear secondary parameters in the conversion of rainfall to proxy data. These secondary parameters are mostly related to plant physiology and soil conditions, suggesting that future work should focus on constraining these non-climate variables in the tropics. The coupled use of $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{wax}}$ may remove some of these uncertainties, as has been

attempted in this thesis, but this approach needs further improvement. The coupling of leaf wax proxy data is also particularly useful in clarifying the climate processes controlling rainfall isotope variations through time. The central Sulawesi data strongly indicate that the rainfall amount effect is not always the dominant process controlling rainfall isotopes in the tropics, with broad implications for the interpretation of tropical climate records. In the case of central Indonesia, changes in the boundary conditions, particularly the inundation of Sunda shelf, appear to modify the relationship between rainfall isotopes and the amount effect.

References

- Aldrian, E. and R. D. Susanto (2003), Identification of three dominant rainfall regions within Indonesia and their relationship to sea surface temperature. *Int. Jour. of Clim.*, 23, 1435-1452.
- Allen, C.D., A.K. Macalady, H. Chenchouni, et al. (2010). A global overview of drought and heat-induced tree mortality reveals emerging climate change risks for forests. *Forest Ecology and Management* 259, 660–684.
- Anchukaitis, K.J., B.M. Buckley, E.R. Cook, B.I. Cook, R.D. D'Arrigo, and C.M. Ammann, (2010), The influence of volcanic eruptions on the climate of the Asian monsoon region. *Geophys. Res. Lett.*, 37, L22703, doi:10.1029/2010GL044843.
- Anshari, G., A.P. Kershaw, S. van der Kaars, and G. Jacobsen (2004), Environmental change and peatland forest dynamics in the Lake Sentarum area, West Kalimantan, Indonesia. *Journal of Quaternary Science*, 19, 7, 637-656.
- Baldwin, K, J. Dyck, J. Hansen, and N. Childs (2012), *Southeast Asia's Rice Surplus*. ERS/USDA Electronic Outlook Report, Washington DC.
- Boer, R. and A.R. Subbiah (2005), Agriculture drought in Indonesia. In: *Monitoring and Predicting Agricultural Drought: A Global Study*, Boken, V.K., A.P. Cracknell and R.L. Heathcote (Eds.), Oxford University Press, New York.
- Bruijnzeel, L.A. (2001), Hydrology of tropical montane cloud forest: a reassessment. *Land Use and Water Resources Research* 1, 1.1-1.18.
- Cane, M. and A.C. Clement (1999), A role for the tropical Pacific coupled ocean-atmosphere system on Milankovitch and millennial timescales. Part II: global

- impacts. In: Clark, P.U., Webb, R.S., Keigwin, L.D. (Eds.). *Mechanisms of Global Climate Change at Millennial Time Scales*. American Geophysical Union, Washington, D.C, pp. 373-383.
- Cannon, C.H., R.J. Morley, and A.B.G. Bush (2009), The current refugial rainforests of Sundaland are unrepresentative of their biogeographic past and highly vulnerable to disturbance. *Proc Natl Acad Sci USA* 106, 27, 11188–11193.
- Carolin, S., K.M. Cobb, J.F. Adkins, B. Clark, and A. Tuen (2013), Varied response of western tropical Pacific hydrology to abrupt climate changes. *Science*, doi:10.1126/science.1233797.
- CIFOR (2015), Clearing the smoke: the causes and consequences of Indonesia's fires. Center for International Forestry Research Fact File, Bogor.
- CRED (2010), EM-DAT: International Disaster Database. Centre for Research on the Epidemiology of Disasters, Brussels. Online at www.emdat.net.
- Dansgaard, W. (1964), Stable isotopes in precipitation. *Tellus*, 16(4), 436–468, doi:10.1111/j.2153-3490.1964.tb00181.x.
- Dayem, K. E., D. C. Noone, and P. Molnar (2007), Tropical western Pacific warm pool and maritime continent precipitation rates and their contrasting relationships with the Walker Circulation. *J. Geophys. Res.*, 112, D06101, doi:10.1029/2006JD007870.
- DiNezio, P.N. and J.E. Tierney (2013), The effect of sea level on glacial Indo-Pacific climate *Nat. Geosci.*, 6, 485–491, <http://dx.doi.org/10.1038/NGEO1823>.
- Dubois N., D.W. Oppo, V. Galy, M. Mohtadi, S. van der Kaars, J. Tierney, Y. Rosenthal, T. Eglington, A. Lückge, and B. Linsley (2014), Indonesian vegetation response to

- changes in rainfall seasonality over the past 25,000 years. *Nature Geoscience* 7, 513-517.
- Foster, P. (2001) The potential negative impacts of global climate change on tropical montane cloud forests. *Earth Sci. Rev.* 55, 73-106.
- Glover, D. (2001), The Indonesian fires and haze of 1997: the economic toll. In: P. Eaton and M. Radojevic eds. *Forest fires and regional haze in Southeast Asia*, Nova Science Publishers, New York.
- Göltenboth, F., Langenberger, G., and Widmann, P (2006) Forest systems. In: Timotius, K.H., Milan, P.P., Margraf, J (eds.) *Ecology of Insular Southeast Asia: The Indonesian Archipelago*, Elsevier.
- Griffiths M. L., R. N. Drysdale, M. K. Gagan, J.-X. Zhao, L. K. Ayliffe, J. C. Hellstrom, W. S. Hantoro, S. Frisial, Y. -X. Feng, I. Cartwright, E. St. Pierre, M. J. Fischer, and B. W. Suwargadi (2009), Increasing Australian-Indonesian monsoon rainfall linked to early Holocene sea-level rise, *Nature Geoscience*, 2, 636-639.
- Hendon, H.H. (2003), Indonesian rainfall variability: Impacts of ENSO and local air-sea interaction. *Journal of Climate*, 16, 11, 1775-90.
- Holbourn, A., W. Kuhnt, H. Kawamura, Z. Jian, P. Grootes, H. Erlenkeuser and J. Xu (2005), Orbitally paced paleoproductivity variations in Timor sea and Indonesian throughflow variability during 460 kyr. *Paleoceanography*, 20, PA 3002.
- Holmes, D. (2000), *Deforestation in Indonesia: a view of the situation in 1999*. World Bank, Jakarta.
- Hope, G., and J. Tulip (1994), A long vegetation history from lowland Irian Jaya, Indonesia. *Palaeogeogr., Palaeoclimatol., Palaeoecol.* 109, 385-398.

- IPCC (2013), *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* (Stocker, T.F., D. Qin, G.-K. Plattner, M. Tignor, S.K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex and P.M. Midgley (eds.)) Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.
- Jourdain, N., A. Gupta, A. Taschetto, C. Ummenhofer, A. Moise, and K. Ashok (2013), The Indo-Australian monsoon and its relationship to ENSO and IOD in reanalysis data and the CMIP3/CMIP5 simulations. *Climate Dynamics* 41, 3073-3102.
- Kershaw, A.P., S. van der Kaars. and J.F. Flenley (2011), The Quaternary history of far eastern rainforests (2nd Edn.). In: Bush, M.B., Flenley, J.F. and Gosling, W.D. (Eds.) *Tropical Rainforest Responses to Climate Change*. Springer-Verlag, Berlin Heidelberg.
- Kirono, D., N. Tapper, and J. McBride, (1999), Documenting Indonesian rainfall in the 1997/1998 El Niño event. *Phys. Geogr.*, 20, 422–435.
- Lee, S.-Y., J.C.H. Chiang, and P. Chang (2014), Tropical Pacific Response to Continental Ice Sheet Topography, *Climate Dynamics*, DOI: 10.1007/s00382-014-2162-0.
- Meehl, G. A. (1987), The annual cycle and interannual variability in the tropical Pacific and Indian Ocean regions. *Mon. Wea. Rev.*, 115, 27-50.
- Mohtadi, M., D. W. Oppo, S. Steinke, J. W. Stuut, E. C. Hathorne, J. Groeneveld, D. Hebbeln, and A. Lückge (2011), Glacial to Holocene swings of the Australian-Indonesian monsoon, *Nature Geoscience*, 4, 540-544.
- Muller, J.F. McManus, D.W. Oppo, and R. Francois (2012), Strengthening of the northeast monsoon over the Flores Sea, Indonesia, at the time of Heinrich event 1.

Geology, 40, 635–638.

Naylor, R., D. Battisti, D. Vimont, W. Falcon, and M. Burke (2007), Assessing risks of climate variability and climate change for Indonesian rice agriculture.

Proceedings of the National Academy of Sciences of the United States of America, 104, 7752–7757.

Neale, R. B. and J.M. Slingo (2003), The Maritime Continent and its role in the global climate: a GCM study. *J. Climate*, 16, 834–848.

Newsome, J. and J. R. Flenley (1988), Late Quaternary vegetation history of the Central Highlands of Sumatra. II.: Palaeopalynology and Vegetational History. *J. Biogeog.*, 15, 555-578.

Niedermeyer E.M., A.L. Sessions, S.J. Feakins, and M. Mohtadi (2014), Hydroclimate of the western Indo-Pacific Warm Pool during the past 24,000 years. *Proceedings of the National Academy of Science* 111, 9402-9406.

Oktaviani, R., S. Amaliah, C. Ringler, M. Rosegrant, and T. Sulser (2011), The impact of global climate change on the Indonesian economy,
<http://www.ifpri.org/sites/default/files/publications/ifpridp01148.pdf>.

Olchev, A., A. Ibrom, J. Priess J, S. Erasmim, C. Leemhuis, A. Twele, K. Radler, H. Kreilein, O. Panferov, and G. Gravenhorst (2008) Effects of land-use changes on evapotranspiration of tropical rain forest margin area in Central Sulawesi (Indonesia): Modelling study with a regional SVAT mode. *Ecological Modelling* 212, 131–137.

- Partin J. W., K. M. Cobb, J. F. Adkins, B. Clark, and D. P. Fernandez (2007), Millennial-scale trends in west Pacific warm pool hydrology since the Last Glacial Maximum, *Nature*, 449, 452-455.
- Peel, M. C., B.L. Finlayson, and T.A. McMahon (2007), Updated world map of the Köppen– Geiger climate classification. *Hydrol. Earth Syst. Sci.*, 11, 1633–1644.
- Peixoto, J.P. and Oort, A.H. (1992), *Physics of climate*. American Institute of Physics, New York.
- Qu, T., Du, Y., Strachan, J., Meyers, G. & Slingo, J. (2005), Sea surface temperature and its variability in the Indonesian region. *Oceanography* 18, 50–61.
- Ramage, C. S. (1968), Role of a tropical “maritime continent” in the atmospheric circulation. *Mon. Wea. Rev.*, 96, 365-369.
- Reeves, J. M., H. C. Bostock, L. K. Ayliffe, T. T. Barrows, P. De Deckker, L. S. Devriendt, G. B. Dunbar, R. N. Drysdale, K. E. Fitzsimmons, M. K. Gagan, M. L. Griffiths, S. G. Haberle, J. D. Jansen, C. Krause, S. Lewis, H. V. McGregor, S. D. Mooney, P. Moss, G. C. Nanson, A. Purcell, S. van der Kaars (2013), Palaeoenvironmental change in tropical Australasia over the last 30,000 years – a synthesis by the OZ-INTIMATE group. *Quaternary Science Reviews*, <http://dx.doi.org/10.1016/j.quascirev.2012.11.027>
- Robock, A., C. M. Ammann, L. Oman, D. Shindell, S. Levis, and G. Stenchikov (2009), Did the Toba volcanic eruption of ~74 ka B.P. produce widespread glaciation? *J. Geophys. Res.*, 114, D10107, doi:10.1029/2008JD011652.

- Rozanski, K., L. Araguas-Araguas and R. Gonfiantini (1992), Relation between long-term trends of ^{18}O isotope composition of precipitation and climate, *Science*, 258, 981-985.
- Russell, J.M., H. Vogel, B.L. Konecky, S. Bijaksana, Y. Huang, M. Melles, N. Wattus, K. Costa, and J.W. King (2014), Glacial forcing of central Indonesian hydroclimate since 60,000 y B.P., *Proc Nat Ac Sci USA*, doi:10.1073/pnas.1402373111.
- Sachse, D, I. Billault, G.J. Bowen, Y. Chikaraishi, K.H. Freeman, C.R. Magill, F.A. McInerney, M.T.J. van der Meer, P. Polissar, R.J. Ribons, J.P. Sachs, H.-L. Schmidt, A.L. Sessions, J.W. White, J.B. West, and A. Kahmen (2012), Molecular paleohydrology: interpreting the hydrogen-isotopic composition of lipid biomarkers from photosynthesizing organisms. *Annu. Rev. Earth Planet. Sci.*, 40, 221–249.
- Saji, N. H. and T. Yamagata (2003), Structure of SST and surface wind variability during Indian Ocean Dipole Mode events: COADS observations. *Journal of Climate* 16, 16, 2735-51.
- Shiau, L.J., M.T. Chen, S.C. Clemens, C.A. Huh, M. Yamamoto, and Y. Yokoyama (2011), Warm pool hydrological and terrestrial variability near southern Papua New Guinea over the past 50 k. *Geophys. Res. Lett.* 38, L00F01.
- Siegert, F, G. Ruecker, A. Hinrichs, and A.A. Hoffmann (2001), Increased damage from fires in logged forests during droughts caused by El Niño. *Nature* 414, 437–440.
- Sturrock, R.N., S.J. Frankel, A.V. Brown, et al. (2011), Climate change and forest diseases. *Plant Pathol* 60, 133-149.

- Tachikawa, K., O. Cartapanis, L. Vidal, L. Beaufort, T. Barlyaeva, and E. Bard (2011), The precession phase of hydrological variability in the Western Pacific Warm Pool during the past 400 ka, *Quaternary Sci. Rev.*, 30, 3716–3727.
- Tachikawa, K., A. Timmermann, L. Vidal, C. Sonzogni, and O.E. Timm (2013) CO₂ radiative forcing and Intertropical Convergence Zone influences on western Pacific warm pool climate over the past 400 ka. *Quaternary Science Reviews*, 86, 24–34, DOI:10.1016/j.quascirev.2013.12.018.
- Tierney, J.E., D.W. Oppo, A.N. LeGrande, Y. Huang, Y. Rosenthal, and B.K. Linsley (2012), The influence of Indian Ocean atmospheric circulation on warm pool hydroclimate during the Holocene epoch. *J. Geophys. Res.*, 117, D19108, doi:10.1029/2012JD018060.
- Underwood, E.C., D. Olson, A. D. Hollander, and J. F. Quinn (2014), Ever-wet tropical forests as biodiversity refuges questions of bias in climate models. *Nature Climate Change*, 4, 9, 740–741.
- van der Kaars, S., D. Penny, J. Tibby, J. Fluin, R. Dam, and P. Suparan (2001) Late Quaternary palaeoecology, palynology and palaeolimnology of a tropical lowland swamp: Rawa Danau, West Java, Indonesia. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 171, 185–212.
- van der Kaars, S., X. Wang, A.P. Kershaw, F. Guichard, and D.A. Setiabudi (2000) A Late Quaternary palaeoecological record from the Banda Sea, Indonesia: Patterns of vegetation, climate and biomass burning in Indonesia and northern Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 155, 135–153.

- Williams, M. (2012), Did the 73 ka Toba super-eruption have an enduring effect? Insights from genetics, prehistoric archaeology, pollen analysis, stable isotope geochemistry, geomorphology, ice cores, and climate models. *Quaternary International* 269, 87–93.
- Woodruff, D. S (2010), Biogeography and conservation in Southeast Asia: how 2.7 million years of repeated environmental fluctuations affect today's patterns and the future of the remaining refugial-phase biodiversity. *Biodiversity and Conservation*, 19, 919–941.
- World Bank (2015), Forest area (% of land area). World Development Indicators – Food and Agriculture Organization, Washington, DC.
- World Resources Institute (2015), Indonesia's fire outbreaks producing more daily emissions than entire US Economy. World Resources Institute Insights, Washington, D.C.
- Wurster, C.M., M.I. Bird, I.D. Bull, F. Creed, C. Bryant, J.A.J. Dungait, and V. Paz (2010), Forest contraction in north equatorial Southeast Asia during the Last Glacial Period, *Proceedings of the National Academy of Sciences*, USA, 107, 15508–15511.

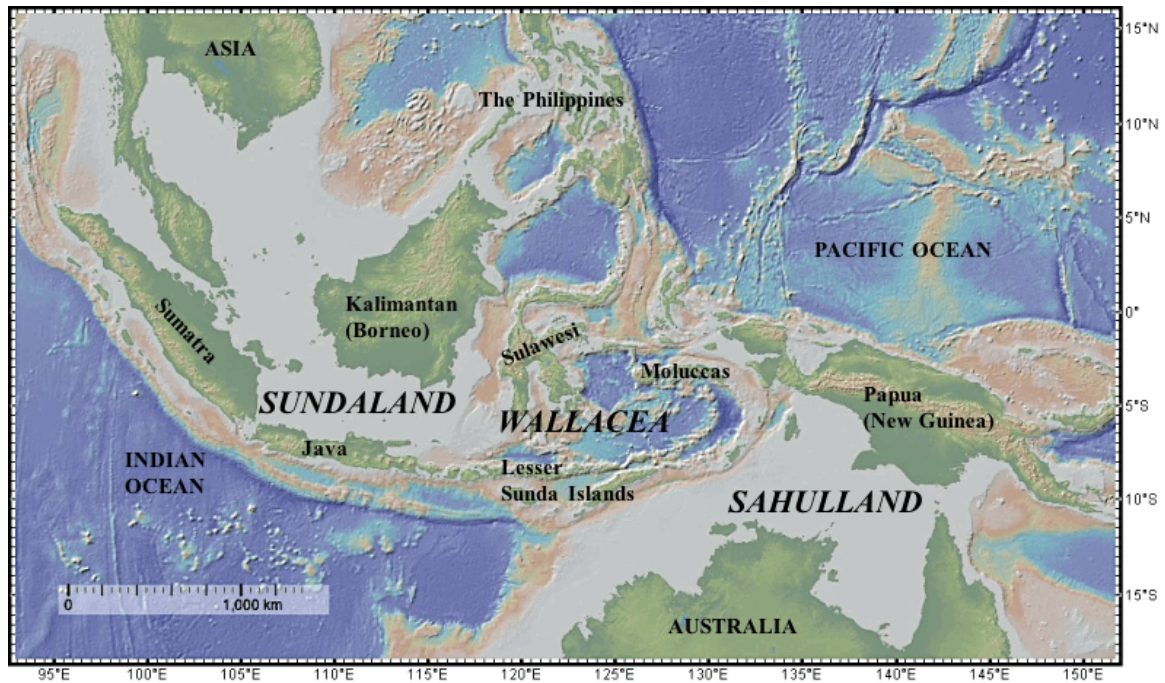


Figure 1 Map of the Indonesian archipelago (the Maritime Continent) with major islands labeled. During the Last Glacial Maximum, the exposed Sundaland was separated from Sahulland by an array of islands and deep, narrow straits, termed Wallacea (Figure from GeoMapApp). Areas exposed during the LGM are shaded in gray.

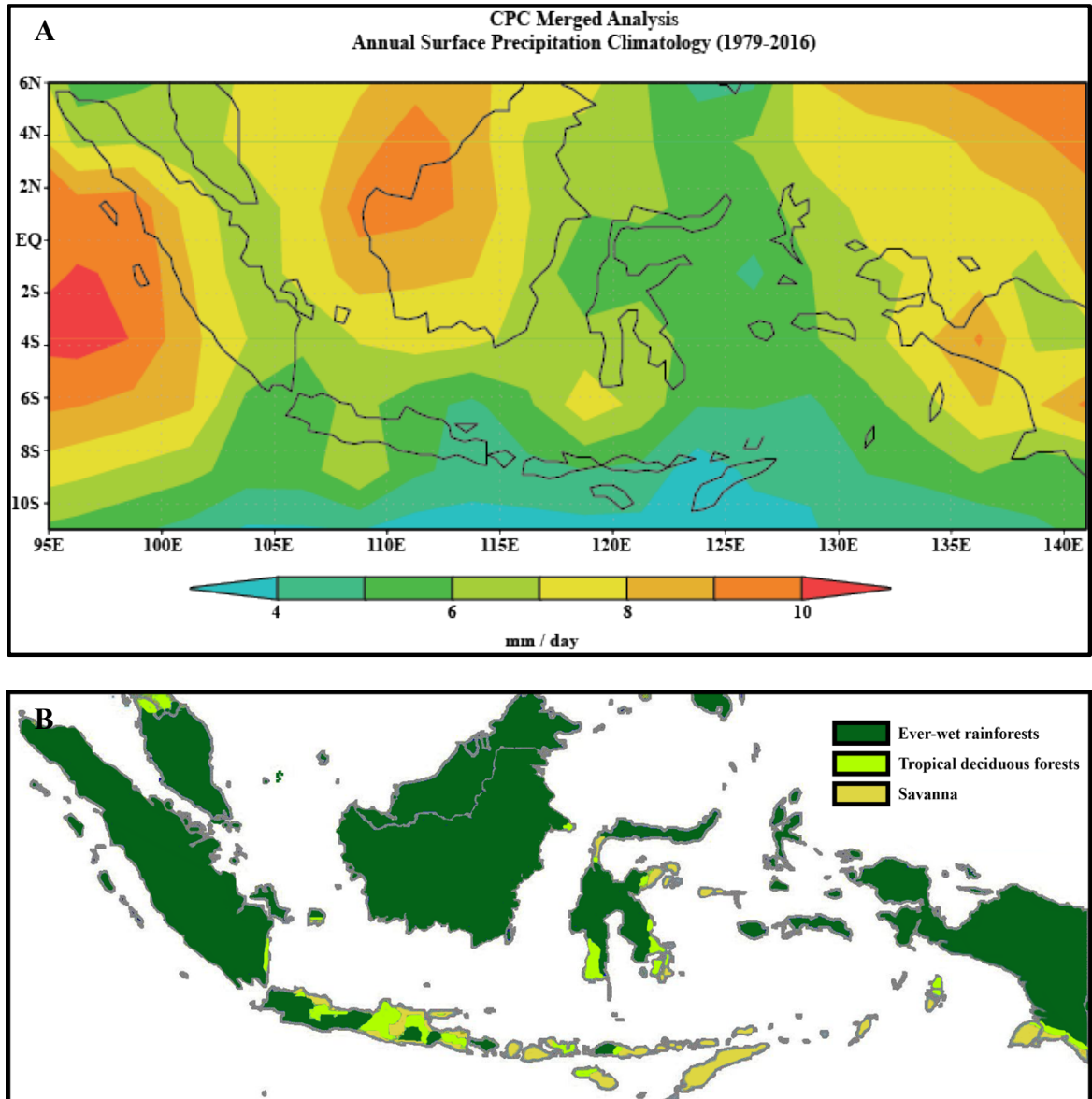


Figure 2 A) Annual surface precipitation climatology for Indonesia, from CPC Merged Analysis of Precipitation (Xie and Arkin 1997). The lowest precipitation rate occurs in the southeastern part of the archipelago. B) Koppen-Geiger climate-vegetation map for the Indonesian archipelago (adapted from Peel et al. 2007). Three types of climate are present in Indonesia, associated with the three types of native vegetation: ever-wet rainforests, tropical deciduous forests, and savannas. Ever-wet rainforests are found in the areas where all 12 months have average precipitation of at least 60 mm and where mean

daily rainfall amounts are high, whereas savannas are associated with areas whose climate has a pronounced dry season and where mean daily rainfall amounts are low, with the driest month having precipitation less than 60 mm and less than 1/25 of the total annual precipitation. Tropical deciduous rainforests represent a transitional climate and vegetation type; the driest month in the tropical deciduous rainforests area has rainfall less than 60 mm, but more than 1/25 the total annual precipitation.

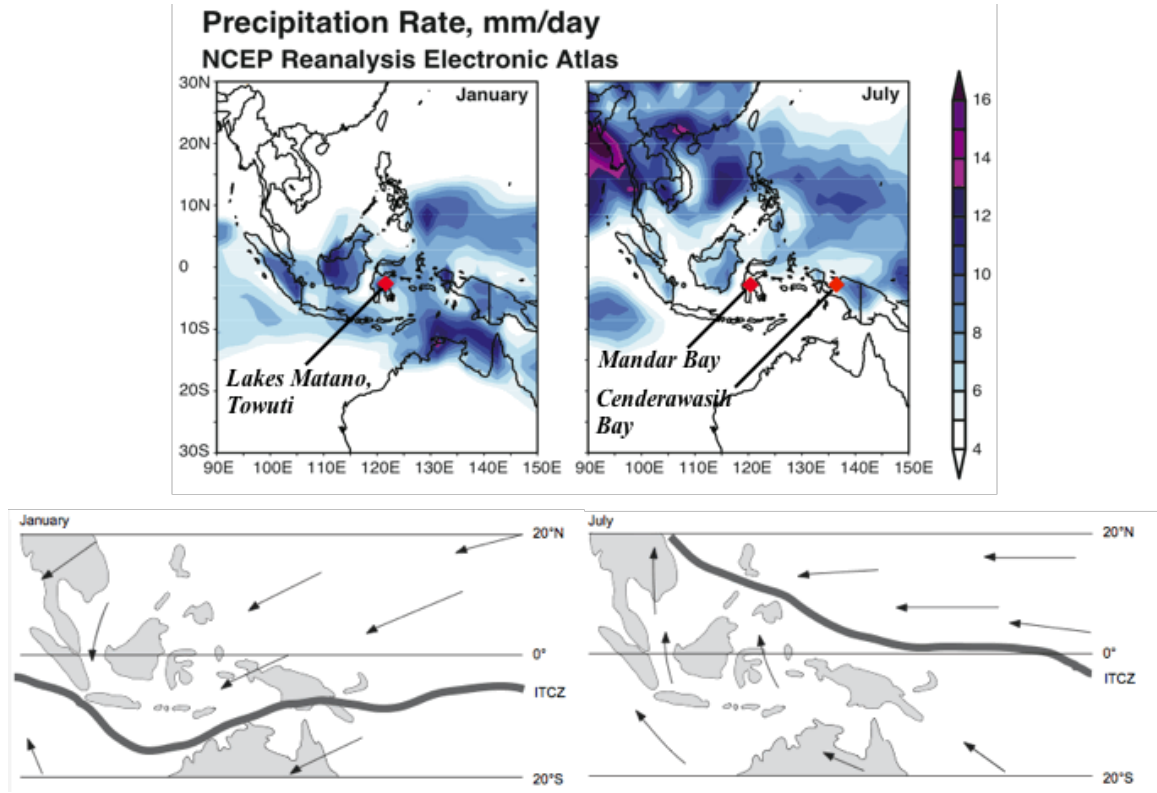


Figure 3 Top panel indicates long-term monthly means of precipitation above the Indonesian archipelago, derived from NCEP/NCAR Reanalysis data for years 1981-2010 for January (left panel) and July (right panel) highlighting the seasonality of precipitation. Precipitation in the region peaks in austral summer and is lowest in austral winter, associated with the seasonal migration of the ITCZ (bottom panel, from Williams et al. 2009). The red dots indicate study sites discussed in this thesis: Lakes Matano and Towuti (central Sulawesi), Mandar Bay (off western Sulawesi), and Cenderawasih Bay (off northern Papua).

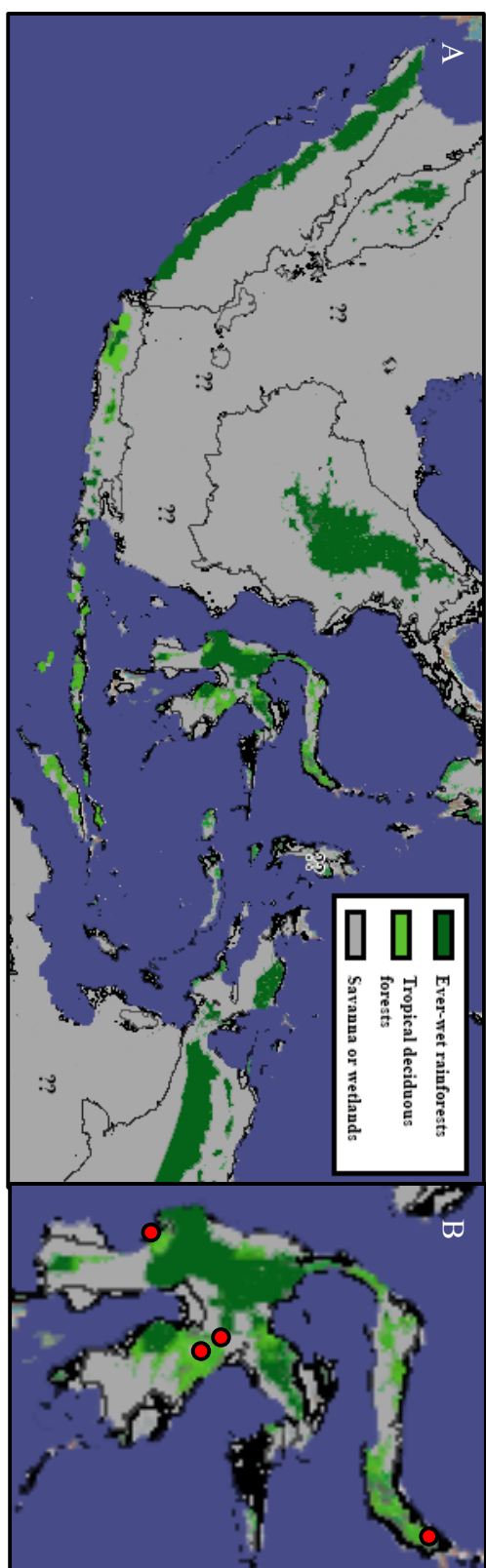


Figure 4 A) Map of Indonesia during the LGM ca. 21 kyr BP. Areas not in blue indicate land areas, including exposed shelf. Areas in dark green indicate high montane regions that likely served as the refugia for C₃ rainforests during the LGM, whereas those in light green indicate transitional areas that might have been partially forested, perhaps with deciduous trees. Savannas or C₄ grasslands, as well as wetlands, likely dominate the rest of the landscape, particularly in the lowlands, although we still do not know for certain the type of vegetation covering the exposed shelf areas. B) Map of Sulawesi during the LGM ca. 21 kyr BP. The red dots indicate several sites in Sulawesi that were used to infer vegetation types during the LGM. From top to bottom: Lake Tondano (Dam et al. 2001), Lake Matano (Wicaksono et al. 2015), Lake Towuti (Russell et al. 2014), and Mandar Bay (Chapter 3, this dissertation). Maps were modified from GeoMap App.

CHAPTER TWO

Compound-specific carbon isotope records of vegetation and hydrologic change in central Sulawesi, Indonesia, since 53,000 yr BP

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Abstract

Lake Matano (2°29.1'S 121°20'E) in central Sulawesi, the deepest and one of the largest lakes in Indonesia, provides an excellent opportunity to reconstruct long-term hydrological and vegetation changes at the heart of the Maritime Continent within the tropical Western Pacific Warm Pool. We measured the carbon isotopic composition of terrestrial leaf waxes ($\delta^{13}\text{C}_{\text{wax}}$) in new sediment cores from Lake Matano to understand regional variations in vegetation and their interactions with local and high latitude climate forcings during the past 53 kyr. Variations in $\delta^{13}\text{C}_{\text{wax}}$ record the expansion and contraction of C_3 versus C_4 vegetation as well as changes in plants' carbon and water use efficiency and, by inference, regional hydroclimate. During Marine Isotope Stages 1 and 3, more negative $\delta^{13}\text{C}_{\text{wax}}$ values indicate that closed-canopy C_3 rainforests predominated in Sulawesi, reflecting a wetter, less seasonal climate. In contrast, more positive $\delta^{13}\text{C}_{\text{wax}}$ values record a terrestrial ecosystem with more abundant open canopy vegetation and possible expansion of C_4 grasses between 29 and 14 ka BP, indicating that during Marine Isotope Stage 2, including the Last Glacial Maximum, the climate was more arid, likely with increased precipitation seasonality, in central Sulawesi. Comparison of our record with other paleovegetation and paleohydrological records from Sulawesi and nearby islands suggests coherent long-term hydrological changes in central Indonesia that appear to be sensitive to high latitude climate across glacial-interglacial boundaries. However, variations in the amplitude of the response of vegetation to these glacial-interglacial forcings highlight the importance of regional topographic control on climate and

vegetation and suggest an important role for higher elevations in maintaining rainforest refugia during regionally arid time intervals when C₄ savannas and grasslands expanded at lower elevations.

Keywords:

Precipitation; Monsoon; Vegetation; Biomarker; Carbon; Sulawesi; Maritime Continent; Orbital-scale

1. Introduction

The Australian-Indonesian Monsoon (AIM) is the main variable controlling Indonesia's hydroclimate (Aldrian and Susanto 2003), and is a globally important source of energy and water vapor for the Earth's atmosphere (Qu et al. 2005). Understanding the history of the Indonesian hydrology and the AIM is therefore of paramount importance to understanding the role of the tropics in global climate change. In recent years a number of new reconstructions have begun to illuminate the hydrologic history of Indonesia, also known as the Maritime Continent. These records are based upon a variety of different proxies including geochemical data from lake and marine sediment, proxies for past vegetation, and speleothems (e.g. Carolin et al. 2013, Dubois et al. 2014, Gibbons et al. 2014, Niedermeyer et al. 2014, Russell et al. 2014). Unfortunately, many of these records exhibit strong contrasts between different proxies and between different regions, and implicate a number of different forcing mechanisms of past hydroclimatic variations. For

instance, some speleothem (Griffiths et al. 2009), foraminiferal records (Holbourn et al. 2005), and palynological data (e.g. Kershaw et al. 2006, Kershaw et al. 2011) suggest that regional precipitation is controlled by precessional orbital forcing, whereby the AIM oscillated between strong states and weak states during intervals of high and low seasonal insolation. In contrast, recent plant biomarker study from Lake Towuti, situated in central Sulawesi, exhibits little evidence of orbital precession and instead suggests that Northern Hemisphere high latitude climate forcings, in particular changes in ice volume, control regional precipitation changes across glacial-interglacial stages (Russell et al. 2014). While these differences could be attributed to variable sensitivities of different parts of Indonesia to climate forcings, at least in some cases such differences arise even over short spatial scales. For instance, speleothem records from Flores, one of the Lesser Sunda Islands in southeastern Indonesia, suggest a relatively wet Last Glacial Maximum (LGM) (Ayliffe et al. 2013), despite previous palynological evidence for aridity in southern Indonesia at that time (van der Kaars et al. 2000).

Understanding these differences will require detailed evaluation of individual proxy records and their sensitivities within a dense network of proxy reconstructions. On the proxy interpretation side, for instance, there is a growing realization that proxy records of precipitation isotopes, such as speleothems, often do not record changes in local rainfall amount, complicating hydroclimatic interpretation for the Maritime Continent (e.g. LeGrande and Schmidt. 2009, Lewis et al. 2010, Moerman et al. 2013). Similarly, vegetation records, which form the basis for many Indonesian climate reconstructions, exhibit complex relationships with hydroclimate. Both precipitation amount and precipitation seasonality, both linked to the AIM, are important determinants

of vegetation distribution within the tropical Maritime Continent (Peel et al. 2007). In the present, the high annual precipitation rates –including high precipitation rates during the dry season– observed across most of the archipelago give rise to widespread lowland ever-wet rainforests and montane cloud forests at higher elevations (Göltenboth et al. 2006). These forest types predominate within the major Indonesian islands including Sumatra, Kalimantan, Sulawesi and Papua. In contrast, C₄ grasslands and savannas are common in lowland regions where average annual rainfall is lower and seasonal variability in AIM rainfall (i.e. seasonality) is high, for example in Nusa Tenggara or the Lesser Sunda Islands and southeast Papua. In regions with moderate precipitation seasonality, such as southwestern Sulawesi and central and eastern Java, open-canopy C₃ monsoon forests are often observed (Göltenboth et al. 2006, Peel et al. 2007).

Recent plant biomarker records have highlighted important changes in the distribution of these ecosystems during and since the LGM, generally reflected by the expansion of drier, more seasonal vegetation regimes and climates during the LGM (Wurster et al. 2010, Dubois et al. 2014, Russell et al., 2014). However, floristic analyses of modern plants and several paleoenvironmental studies of floral and faunal assemblages indicate the presence of disjunct rainforest areas in some parts of the Maritime Continent during the LGM, highlighting the possible role of rainforest refugia in establishing the region's ecosystem diversity (Slik et al. 2003, Barton 2005, Moyle et al. 2011). Evidence for such refugia, coupled with robust evidence for regionally pervasive aridity, suggests significant heterogeneity in past changes in climate and vegetation and highlights important gaps in our understanding of the long-term stability of the regional rainforest ecosystems. Recent work has also highlighted the role that regional topography might

play in controlling long-term changes in both rainfall amount and the types of vegetation present in different parts of the Maritime Continent since 25 kyr BP (Wurster et al. 2010, Niedermeyer et al. 2014). In light of this evidence, dense networks of proxy records are needed to determine robust signals of paleoclimate changes and to evaluate the processes that govern climate and vegetation heterogeneity in the Maritime Continent.

Here we present new compound-specific stable isotope records from Late Pleistocene and Holocene-aged sediments of Lake Matano in central Sulawesi, situated at the heart of the Maritime Continent within the tropical Western Pacific Warm Pool. The close proximity between Lake Matano and Lake Towuti allow us to evaluate the robustness of local to regional paleohydrologic signals from both lakes, as well as possible causes for differences between these records, with important implications for the interpretation of other local to regional reconstructions of Indonesian climate. Our reconstructions highlight linkages between the AIM and high-latitude forcings at orbital timescales. In addition, the heterogeneity of local hydrological and vegetation settings among records from central Indonesia during the past 50 kyr, including between Lakes Matano and Towuti, highlight the importance of orography in interpreting past vegetation and paleohydrologic records and provide new information about rainforest refugia during arid episodes in Indonesia.

2. Study Site

Lake Matano (2°29.1'S 121°20'E, 394 m asl) is located on the island of Sulawesi in central Indonesia (Figures 1, 2). The annual average precipitation near Lake Matano

ranges from 2500 to 3000 mm (Figure 2) and occurs year-round, placing the region into the tropical rainforest climate category in the Köppen climate classification (Peel et al. 2007). Precipitation is highest in December-May, with peak precipitation of ~400 mm/month in April, and lower in June-November, with minimum precipitation in September of ~100 mm/month (Figures 1, 2). This seasonality is largely set by the AIM and its interactions with the regional seas (Aldrian and Susanto 2003). During the rainy season (Dec-May), the Intertropical Convergence Zone (ITCZ) migrates southward to northern Australia and the northwesterly monsoon blows south across the equator, bringing moisture from the Northern Hemisphere sector of the Indian Ocean and the South Pacific to most of Indonesia (Aldrian and Susanto 2003, Gimeno et al. 2012). During the dry season (June-November), the ITCZ shifts north over Asia, and strong southeasterly flow cools the local oceans of the southern maritime continent, limiting convective activity (Aldrian and Susanto 2003). Precipitation in our study region increases with elevation; weather stations at higher altitudes record higher annual rainfall amounts (with the exception of one outlier; Figure 2). Precipitation seasonality also decreases with elevation, such that higher elevations experience year-round wetness.

Lake Matano is one of the five lakes forming the ancient Malili Lakes system, a set of large tectonic lakes in central Sulawesi. With a maximum depth of 590 m, this ultra-oligotrophic lake (Haffner et al. 2001) is the eighth deepest lake in the world (Lehmusluoto and Machbub 1997). Lake Towuti, the largest tectonic lake in Indonesia, lies at a lower elevation (319 m asl) approximately 10 km south of Lake Matano (Figure 2). Both lakes lie mostly in an iron-rich ultramafic ophiolite formation, except in the southwestern border of Lake Matano, where a limestone formation is located (Figure 2).

The catchment area of Lake Matano, at 276 km², is steep and mountainous (maximum altitude: ~1340 m, mean altitude: ~650 m) as compared with that of Lake Towuti, and vegetation in Matano's catchment is dominated by closed canopy lowland C₃ rainforest (van Balgooy and Tantra 1986). At altitudes above 850 m, cloud forests including submontane and montane plant types are present, characterized by abundant Fagaceae including *Lithocarpus* and *Castanopsis*, as well as mosses, ferns, orchids, and other epiphytic plants (Hope 2001, Culmsee et al. 2010).

3. Methods

Core IDLE-MAT10-2B (Figure 1, 2°31'10.52"S, 121°24'41.72"E, core length: 9.07 m) was collected from 137 m water depth using a Kullenberg piston corer in July 2010. The core was air freighted to the LacCore facility, University of Minnesota, where it was split, described, and logged for its major lithofacies. The age model for this core is based on radiocarbon (¹⁴C) analyses of eight bulk sediment samples and two fragments of wood/plant macrofossils (Table 1). Comparison of bulk organic carbon and macrofossil ages from the same stratigraphic horizons yields an average ¹⁴C lacustrine reservoir effect of approximately 975 ¹⁴C years (Figure 3, Table 1), assuming that the plant materials reflect the atmospheric ¹⁴C age at the time of their deposition. This reservoir effect is likely due to dissolved old carbon inputs resulting from the weathering of limestone deposits in Lake Matano's catchment. After correcting for this reservoir and converting corrected radiocarbon ages to calendar ages by applying the IntCal09 curve using the

CALIB 6.0 program, we fit a third-order polynomial through the calibrated age points to construct the age model (Figure 3).

We measured 68 samples from the core for the carbon isotopic composition of long-chain *n*-alkanoic acids, terrestrial leaf waxes that are abraded from terrestrial biomass and transported to and preserved in lake sediments. Sediment samples were freeze-dried, homogenized with a mortar and pestle, and lipids were extracted using a Dionex Accelerated Solvent Extractor. Alkanoic acids were separated from the total lipid extracts using aminopropyl column chromatography. The separated acids were methylated with acidified methanol of a known isotopic composition, and further purified via silica gel chromatography. The resulting fatty acid methyl esters (FAMES) were analyzed in at least duplicate via gas chromatography isotope ratio monitoring-mass spectrometry (GC-IR-MS) using a Thermo Delta V Plus at Brown University. We targeted long chain fatty acid homologues (between C₂₆ and C₃₂) because they are largely derived from terrestrial plants (Sachse et al. 2012) In this study, we report $\delta^{13}\text{C}_{\text{wax}}$ values of the C₂₈ homologue (Figure 3), the most abundant homologue in Matano sediments, after correcting for the isotopic composition of the methyl group added during methylation.

4. Results

The core is composed primarily of dark olive to reddish-brown slightly silty clay with no obvious disconformities, and radiocarbon dating indicates fairly continuous sedimentation rates over the past ~53 kyr BP of ~14 cm/kyr, similar to sedimentation

rates in nearby Lake Towuti (Figure 3). During this time, $\delta^{13}\text{C}_{\text{wax}}$ varies between -44 and -33 ‰, averaging -38.5 ‰ (Figure 3). These values are similar to, but slightly more negative than waxes deposited over the same time period at Lake Towuti (Figure 4). The average difference between duplicates (n=56) in the $\delta^{13}\text{C}_{\text{wax}}$ of C_{28} was 0.18 ‰, and the average standard error on samples measured in triplicate or greater (n=12) was 0.34 ‰ (Figure 3). The $\delta^{13}\text{C}_{\text{wax}}$ of C_{28} and C_{26} *n*-acid are strongly correlated ($r^2 = 0.79$, $p < 0.001$), as are the $\delta^{13}\text{C}_{\text{wax}}$ of C_{28} and C_{30} *n*-acid ($r^2 = 0.85$, $p < 0.001$), indicating that measurements on C_{28} are broadly representative of the isotopic composition of the long-chain waxes. The mean values of the $\delta^{13}\text{C}_{\text{wax}}$ for C_{26} and C_{30} *n*-acid are -36.6 and -34.4 ‰ respectively.

From ~53 to 31 kyr BP, i.e. much of Marine Isotope Stage 3 (MIS 3), $\delta^{13}\text{C}_{\text{wax}}$ averages -40.5 ‰. At 31 kyr BP $\delta^{13}\text{C}_{\text{wax}}$ rapidly transitions to more positive (^{13}C -enriched) values, and between ~28 to 14 kyr BP (i.e. during MIS 2) the average $\delta^{13}\text{C}_{\text{wax}}$ was -34.9 ‰. $\delta^{13}\text{C}_{\text{wax}}$ then shifts rapidly toward more negative (^{13}C -depleted) values and averages -38.8 ‰ in the Holocene, similar to the values observed during the later part of MIS 3 (Figure 4). The depleted modern $\delta^{13}\text{C}_{\text{wax}}$ values in Lake Matano are comparable to the $\delta^{13}\text{C}$ of modern soil samples (n = 5) taken around the Malili Lakes (average: -36 ‰, maximum: -38.6 ‰, minimum: -32.6 ‰). The long-term isotopic changes record a series of shifts in vegetation that show both similarities to and differences among other regional records, as discussed below.

5. Discussion

5.1 Interpretation of the $\delta^{13}\text{C}_{\text{wax}}$ proxy for vegetation and climate

In the tropics, $\delta^{13}\text{C}_{\text{wax}}$ is primarily interpreted to reflect changes in the relative abundance of plants using the C_3 and C_4 photosynthetic pathways (Vogts et al. 2009). The $\delta^{13}\text{C}$ of n-alkanes leaf waxes from C_4 plants, which include most tropical grasses, typically range from -14 ‰ to -27 ‰, while waxes from C_3 plants range between -29‰ and -42 ‰ (Bi et al. 2005, Chikaraishi and Naraoka 2007, Vogts et al. 2009). For n-alkanoic acid leaf waxes, the reported $\delta^{13}\text{C}_{\text{wax}}$ values for C_4 plants range from -19 ‰ to -26 ‰ and the values for C_3 plants range from -30 ‰ to -45 ‰ (Chikaraishi and Naraoka 2007). $\delta^{13}\text{C}_{\text{wax}}$ can also vary due to changes in the efficiency of carbon and water utilization within C_3 and C_4 plants, often related to the effects of edaphic factors, such as soil, precipitation, humidity (Diefendorf et al. 2010). Globally, C_3 biomass exhibit a ~6 ‰ enrichment across precipitation gradients from closed-canopy rainforest to more xeric ecosystems (Diefendorf et al. 2010). The average n-alkanes $\delta^{13}\text{C}_{\text{wax}}$ for C_3 plants in African rainforests are more negative (-36.9 ± 2.5 ‰) relative to C_3 plants in more xeric African ecosystems, at -33.5 ± 2.8 ‰, with waxes from some rainforest trees reaching $\delta^{13}\text{C}_{\text{wax}}$ as negative as -45‰ (Vogts et al. 2009).

Analyses of the $\delta^{13}\text{C}$ of n-alkanoic acids from surface marine sediments across the Maritime Continent and northwestern Australia suggest that $\delta^{13}\text{C}_{\text{wax}}$ is correlated strongly with rainfall seasonality as well as with average monthly precipitation (Dubois et al. 2014). $\delta^{13}\text{C}_{\text{wax}}$ values observed in marine surface sediments from southwestern Indonesia,

where C₄ grasslands and savannas are common, are enriched compared to those from wetter, less seasonal regions including Sulawesi. Southwestern Indonesia experiences a much more highly seasonal climate, marked by strong precipitation in boreal winter associated with a strengthened winter monsoon and southward migration of the ITCZ. These data show that $\delta^{13}\text{C}_{\text{wax}}$ tracks ecoclimate/biome classifications for the Maritime Continent at least partly, if not dominantly, through dry season water stress, which also would affect carbon use efficiency in C₃ forests. Thus, we interpret Lake Matano $\delta^{13}\text{C}_{\text{wax}}$ record as a proxy for changes in vegetation type, which is largely controlled by regional hydroclimate.

The long-term pattern of $\delta^{13}\text{C}_{\text{wax}}$ preserved in Lake Matano likely reflects catchment-wide transitions between closed-canopy rainforest and more open ecosystems, possibly incorporating C₄ grasses. During the MIS 3 and MIS 1, $\delta^{13}\text{C}_{\text{wax}}$ remained at about -38 ‰ (Figure 4), similar to the values recorded for closed-canopy rainforests in other regions (Vogts et al., 2009) and consistent with the present vegetation within the catchment. The 5.5 ‰-enrichment observed during MIS 2 indicates a shift toward more drought-tolerant ecosystems, and a drier, more seasonal climate after 30 kyr BP (Figure 4). The absolute values of $\delta^{13}\text{C}_{\text{wax}}$ we observe during the LGM (-35 ‰) do not require incorporation of C₄ grasses and instead could reflect more open, C₃ ecosystems with increased water stress. However, it is likely that this transition also saw at least limited expansion of C₄ grasses within the catchment. Discontinuous pollen data from mires located only 1.5 km from the southwestern shore of Lake Matano, indicate a substantial increase in Poaceae (grass) pollen between 35 and 17 ka (Hope 2001). Moreover, $\delta^{13}\text{C}_{\text{wax}}$ reconstructions from Lake Towuti suggest more enriched values during the LGM than

observed at Lake Matano (-25 vs. -35 ‰, respectively) that are enriched relative to typical C₃ end-member $\delta^{13}\text{C}_{\text{wax}}$ values, implying expanded C₄ grasslands around Towuti only 10 km from our study area (Figure 4). Thus, we interpret the $\delta^{13}\text{C}_{\text{wax}}$ record at Lake Matano to reflect the prevalence of closed-canopy rainforest during MIS 3 and the Holocene, interrupted by the development of a more open C₃ forest with scattered C₄ grasses during late MIS3 and the LGM.

What might have caused these transitions? Atmospheric pCO₂ is known to affect vegetation type and $\delta^{13}\text{C}$, with low CO₂ concentrations favoring C₄ grasses relative to C₃ plants (Ehleringer et al. 1997). However, the patterns of changes from MIS 3 through the Holocene observed at our site is unlikely to be explained solely by changes in CO₂ concentrations, as CO₂ concentrations were quite low during MIS 3 relative to the Holocene (Figure 6), yet our data suggest similar, if not more depleted $\delta^{13}\text{C}_{\text{wax}}$ in MIS 3 than in the Holocene. Temperature also likely influenced vegetation changes, and while high-altitude pollen taxa have been observed in relatively low elevation settings in the Maritime Continent (Kershaw et al. 2011), variations in Sulu Sea sea surface temperatures during the past 50 kyr, which might be indicative of synoptic temperature change in the region, also show relatively little contrast during MIS 2 and 3 followed by a much warmer Holocene (Figure 6), in contrast to the temporal pattern in our $\delta^{13}\text{C}_{\text{wax}}$ records. Furthermore, modern temperature gradients across the Maritime Continent are small and not correlated with changes in the relative abundance of C₃ and C₄ vegetation (Dubois et al. 2014). Lastly, sedimentary Ti concentrations in Lake Towuti, which indicate fluvial erosion and hence precipitation, show a temporal pattern that is strikingly similar to $\delta^{13}\text{C}_{\text{wax}}$ data, with high Ti concentrations indicating strong runoff during MIS 3

and the Holocene, and low concentrations reflecting drier conditions during MIS 2 (Russell et al. 2014). Therefore, we interpret the more enriched values between 29 and 14 kyr BP as indicative of a drier and more seasonal MIS 2.

5.2 Regional vegetation changes in central Sulawesi during the LGM

The long-term trend in $\delta^{13}\text{C}_{\text{wax}}$ in Matano is similar to that observed at Lake Towuti (Figure 4, Russell et al. 2014), demonstrating coherent vegetation and hydrological changes in central Sulawesi since 50 kyr BP. The two records do differ in their absolute values, particularly during the last glacial maximum, MIS 2, as well as during parts of MIS 3. During MIS 2, both lakes recorded a substantial ^{13}C -enrichment relative to the Holocene and MIS 3, but Towuti experienced a much larger enrichment compared to Matano. During MIS 3 and the Holocene, $\delta^{13}\text{C}_{\text{wax}}$ values are also more depleted at Lake Matano, although less so than at the LGM. The differences at the LGM reflect quite different terrestrial ecosystems likely related to regional climatic variations, perhaps controlled by topographic differences between the two lakes' catchments.

Despite the close proximity between Lake Towuti and Lake Matano, altitudinal and geomorphological differences led to a number of floristic and climatic contrasts between their catchments that could explain the different amplitudes of variation in $\delta^{13}\text{C}_{\text{wax}}$. Today, the catchment of Lake Towuti (area: $\sim 1200 \text{ km}^2$, maximum elevation: $\sim 1250 \text{ m}$, mean elevation: $\sim 500 \text{ m}$) is dominated by lowland rainforest, similar to what is found in Lake Matano's catchment below 850 m, and many of the flat river valleys surrounding Lake Towuti today are inhabited by mixed forest and C_4 grassland

(BBKSDA Sulsel 2011). In contrast, the steeper and higher mountain ranges encircling Lake Matano are home to lowland rainforests below 650 m, and submontane and montane rainforests above, in keeping with the increase in precipitation with elevation observed regionally (Fig. 2). Pockets of anthropogenic C₄ grasses are found near Lake Matano (Hope 2001). The large differences in $\delta^{13}\text{C}_{\text{wax}}$ between Matano and Towuti during the LGM suggest that the grasslands in the valleys around Towuti today expanded, contributing to the substantial enrichment (~13 ‰) of around Towuti (Figure 4), while Matano remained surrounded largely by C₃ ecosystems. We also note 1-2 kyr differences in the timing of $\delta^{13}\text{C}_{\text{wax}}$ transitions, in which Lake Towuti variations generally lead those of Lake Matano. During the last glacial termination, for instance, $\delta^{13}\text{C}_{\text{wax}}$ shifts toward more depleted values at Lake Towuti beginning at ~16 kyr BP, while values become more enriched beginning ~14 kyr BP at Matano. While this could reflect age model errors, it is also possible that this is due to the topographic differences. The presence of the many flat river valleys surrounding Lake Towuti could facilitate a relatively rapid expansion and contraction of C₄ grasses, while transitions at Matano are controlled by slower shifts related to both C₄ grass expansion and changes in C₃ tree composition.

Indeed, pollen-based vegetation reconstructions from the Wanda site near Lake Matano indicate that the period between 35 and 17 kyr was marked by an increase in Poaceae pollen relative to MIS 3 and the Holocene, supporting our inference of regionally drier and more seasonal conditions during MIS 2 (Hope 2001). In addition, the Wanda site also records increased *Castanopsis* and *Lithocarpus* pollen during the glacial MIS 2 (Hope 2001). While Poaceae indicates increasing aridity, which in itself might have driven the enrichment of $\delta^{13}\text{C}_{\text{wax}}$ during the glacial period, *Castanopsis* and

Lithocarpus might indicate that colder, higher-altitude tree taxa migrated to lower elevation areas closer to Lake Matano, presumably due to cooler LGM temperatures. Fagaceae, including *Lithocarpus* and *Castanopsis*, is highly abundant in the high mountains around Lake Matano, as has been observed in other mountain forests across Sulawesi (Culmsee et al. 2010, Culmsee and Leuschner 2013). Interestingly, Fagaceae plants from settings comparable to our study site (i.e. having similar altitudes and precipitation regimes) from the tropics and subtropics have lower $\delta^{13}\text{C}$ values than Fagaceae and other evergreen angiosperms (Read and Farquhar 1991, Wei et al. 2010, Du et al. 2014). The larger carbon isotope fractionation (by approximately 2 ‰) has been attributed to the relatively low water use efficiency requirements of Fagaceae in cool, moist submontane forest ecosystems in China (Du et al. 2014).

While the Wanda pollen record is highly discontinuous (Hope 2001), the long-term trends in $\delta^{13}\text{C}_{\text{wax}}$ in Matano are compatible with the pollen reconstruction from the Wanda site. Enriched $\delta^{13}\text{C}_{\text{wax}}$ during the LGM is compatible with grassland expansion observed at Wanda, and the more depleted $\delta^{13}\text{C}_{\text{wax}}$ values observed during the MIS 2/LGM at Lake Matano relative to Towuti suggest that the mountains surrounding Matano remained home to submontane and montane taxa. Downslope migration of these taxa during the LGM could have somewhat offset the enrichment in $\delta^{13}\text{C}_{\text{wax}}$ related to C_4 expansion. This suggests that the cooler temperatures at these higher elevations may have reduced water stress, and provided refugia to rainforests during the regionally dry conditions.

5.3. Altitudinal gradients in Indonesian vegetation and climate during the LGM

Given that precipitation is a primary factor controlling $\delta^{13}\text{C}_{\text{wax}}$ variations among modern C_3 ecosystems (Diefendorf et al. 2010) as well as variations between C_3 and C_4 ecosystems in the Maritime Continent (Dubois et al. 2014), we suggest that the differences in Matano and Towuti $\delta^{13}\text{C}_{\text{wax}}$ records may also be explained by rainfall variations with altitude within central Sulawesi. In the present, large-scale sea breezes are thought to play an important role in maintaining modern montane cloud forests across the interior of Sulawesi and other Indonesian islands, which together make up the largest area of cloud forests in the world (Bruijnzeel 2001, van der Molen et al. 2006). Radar and TRMM observations above Sulawesi demonstrate that the sea breeze carrying moisture from the seas around the island often results in higher precipitation in mountainous central Sulawesi in the afternoon, especially during the wet season (Renggono 2011). Today, Matano's higher altitude and steeper slopes cause intense cloud cover development, resulting in higher rainfall amounts and less seasonality than near Towuti (Figure 2) and more extensive active montane cloud forests (Culmsee et al. 2010). The higher supply of soil water, combined with the more humid, cloudier microclimate near Matano is likely to result in relatively low year-round water use efficiency and a relatively large carbon isotope fractionation in vegetation near Matano compared to Towuti (e.g. Cernusak et al. 2013).

During the last glacial period, lowered sea level (i.e. increasing distance toward the island's center) combined with cooler sea surface temperature might be expected to reduce the strength of the sea breezes. However, Sulawesi's coastline is relatively steep,

resulting in little shoreline migration during sea level lowstands as compared to other Indonesian islands. The large mountainous area in the center of the island would still have generated a strong orographic effect and attracted sea breezes, thus ensuring a relatively wetter region and cloud generation during all months of the year. At the same time, the relatively cooler surface and tropospheric temperatures during the LGM (e.g. Tripathi et al. 2014) should have depressed the condensation height. This is in keeping with downward migration of high-altitude taxa observed in pollen records from intermediate elevations in central Sulawesi (Hope 2001) and other Indonesian islands (van der Kaars et al. 2000, van der Kaars et al. 2001). Within this context, the highlands of Sulawesi, including the higher altitude areas of Lake Matano's catchment, were thus likely to maintain elements of C₃ rainforests during the last glacial cycle.

The shifts in tropical rainfall variability above Sulawesi, as well as their altitudinal gradients, during the past 50 kyr are in agreement with the hydrological interpretation of many pollen and plant biomarker records from marine and terrestrial sites across the Maritime Continent. Marine sediment cores from the Aru Sea (Kershaw et al. 2006) and Banda Sea (van der Kaars et al. 2000) record reduced abundances of Pteridophyta and increased concentrations of charcoal from ~30 to 13.8 kyr, indicating less rainfall than present during MIS 2 in the eastern part of the archipelago. Highest pteridophyte and rainforest pollen concentrations occur during the Holocene and MIS 3, indicating that the interglacials were much wetter than glacial periods. The local vegetation around Rawa Danau, a lowland lake in Java, was dominated by grasses from ~28.7 to 16.7 kyr, indicating drier climate around the site during MIS 2 (van der Kaars et al. 2001). Records of n-alkane $\delta^{13}\text{C}$ measured from cave guano profiles from low

elevation Batu cave in Peninsula Malaysia, and Makangit and Gangub caves on Palawan Island (Figure 5, Wurster et al. 2010), display substantial MIS 2 enrichment compared to Holocene values, indicating drier LGM/MIS 2 similar to our records. The much more enriched C_{29} n-alkane $\delta^{13}C_{\text{guano}}$ values for the two Palawan caves compared to the C_{28} n-alkanoic acid $\delta^{13}C_{\text{wax}}$ from central Sulawesi (Figure 5) is consistent with previous work demonstrating that long-chain n-alkane are in general enriched in ^{13}C (up to $\sim 7\text{‰}$) relative to the corresponding long-chain n-alkanoic acids (Chikaraishi and Naraoka 2007). The enriched values for Palawan are also expected given the open-canopy characteristic of Palawan rainforests (Peel et al. 2007). Trends in our carbon isotope records, coupled with published pollen evidence in Central Sulawesi, are highly consistent with the $\delta^{13}C_{\text{wax}}$ records collected in Savu Sea, a basin surrounded by the Nusa Tenggara islands in southeastern Indonesia (Figure 4, Dubois et al. 2014). While the Savu record has reduced amplitude and appears to have less “noise” in $\delta^{13}C_{\text{wax}}$ compared to Matano and Towuti (Figure 4), possibly due to the much larger wax source areas and thus a more mixed signal, all three records indicate enriched $\delta^{13}C_{\text{wax}}$ during the LGM compared to the Holocene (Figure 4). The amplitude of $\delta^{13}C_{\text{wax}}$ change between the LGM and the Holocene in Savu is 2-3 ‰, similar to the shift observed at Matano (3-4 ‰), both much smaller than the shift observed at Lake Towuti (13 ‰) (Figure 4). However, $\delta^{13}C_{\text{wax}}$ values at Lake Towuti during MIS 2 were only 1 ‰ more enriched than the LGM values of Savu. Thus, we infer that grasslands or patches of grasslands dominated lowland areas in the Sulawesi-Nusa Tenggara corridor during the drier, more seasonal climate of the LGM, while the highlands supported C_3 ecosystems under cooler and moister conditions. In this context, it seems likely that the large modern gradients in

precipitation and its seasonality that exist between Sulawesi and Nusa Tenggara developed during the last glacial termination. Importantly, our results also highlight the need for numerous records at regional to continental spatial scales to disentangle large-scale climatological signals trends from more regional patterns that can be influenced by orography.

In contrast to these low elevation records, there is also evidence for continuous ever-wet vegetation from many of Indonesia's high elevation lakes (above 500 m asl). Palynological data from Danau di Atas in Sumatra (1535 m asl, Newsome and Flenley 1988), Lake Hordorli in West Papua (780 m asl, Hope and Tulip 1994), as well as from peatlands in Kalimantan (35 m asl but flanked by mountain ranges as high as 2000 m, Anshari et al. 2004) suggest that the highlands and the interior of the Maritime Continent remained relatively wet during MIS 2. Similarly, n-alkanoic acid $\delta^{13}\text{C}_{\text{wax}}$ records from west Sumatra (Niedermeyer et al. 2014) and Tarakan Delta off east Kalimantan (Dubois et al. 2014), which are associated with catchments dominated by high elevation (>2000 m asl) mountain ranges, indicate wet conditions during the LGM. Interestingly, despite its low altitude, records from Niah cave in northern Borneo display little changes in $\delta^{13}\text{C}_{\text{guano}}$ values from the LGM through early Holocene (Wurster et al. 2010), suggesting that rainfall variability over Borneo during regionally dry LGM was not sufficient to result in the reduction of C_3 rainforests in Borneo –or northern Sundaland– lowlands (Dubois et al. 2014). Despite this observation, taken together, these low and high elevation records show that the vegetation gradients we observe in central Sulawesi reflect larger-scale altitudinal patterns observed across much of the Maritime Continent.

Based on the Maritime Continent pollen and biomarker records, we surmise that the phytogeography of the region's rainforests across glacial-interglacial boundaries is closely linked with both climatic and altitudinal factors. C_4 grasslands, which in present-day Maritime Continent islands occupy drier and highly seasonal areas in the lowlands (Whitten et al. 1987, Peel et al. 2007), were expanded into larger areas at the expense of lowland C_3 rainforests during the last glacial period. Meanwhile, higher elevation regions cooled and remained relatively wet, supporting C_3 ecosystems, and at least some montane taxa were also able to expand to intermediate altitudes. The maintenance of wetter, C_3 ecosystems at higher elevations indicates the significance of orographic precipitation and temperatures in reducing dry season water stress, thus providing favorable conditions to C_3 rainforest taxa. These data are in contrast to biome modelling studies suggesting that rainforests perhaps expanded into low elevation, exposed shelves during the LGM due to the availability of additional land area, which could support those rainforests given sufficient precipitation (Cannon et al. 2009), and instead highlight an important role for the higher elevations as rainforest refugia.

5.4. Climatic controls on central Indonesian rainfall since 53 kyr BP

Our data suggest that over the last 53 kyr BP, regional climate varied between wet conditions during MIS 3 and the Holocene, to dry and more seasonal conditions from 30 to ~14 kyr BP. This history is very similar to recent reconstructions from Lake Towuti (Figure 4, Russell et al. 2014), as well as other Sulawesi records including Lake Tondano (Dam et al. 2001), where lower lake levels record drying during the LGM. Together with

vegetation-based biomarker records from Makangit and Gangub caves (Wurster et al. 2010) as well as biomarker and pollen data from Savu Sea (Dubois et al. 2014), we surmise that a drier and more seasonal climate than present was a robust MIS 2 feature across Sulawesi and the islands north and south of Sulawesi. Indeed, the strong similarity between $\delta^{13}\text{C}_{\text{wax}}$ data from Lakes Matano and Towuti indicate a regionally robust response to climate forcing.

The records from Sulawesi, however, do not entirely agree with other regional paleoclimate data derived from $\delta^{18}\text{O}$ of speleothems and planktic foraminifera. Variations in speleothem $\delta^{18}\text{O}$ from Gunung Buda in northern Borneo (Figure 5, Partin et al. 2007, Carolin et al. 2013) have been interpreted to demonstrate that regional precipitation is controlled primarily by precessional orbital forcing and was only weakly affected by glacial-interglacial transitions. Similarly, shorter speleothem $\delta^{18}\text{O}$ records from Liang Luar in Flores suggest a wet LGM in contrast to other southwest Indonesian records (Figure 5, Ayliffe et al. 2013). Planktic foraminifera-derived $\delta^{18}\text{O}$ of seawater ($\delta^{18}\text{O}_{\text{sw}}$) proxy records from offshore western Sumatra (Mohtadi et al. 2014), off south Java (Mohtadi et al. 2011), and the Savu Basin (Gibbons et al. 2014), indicate relatively little glacial-interglacial hydrologic changes during the Late Pleistocene-Holocene. These $\delta^{18}\text{O}_{\text{sw}}$ records were interpreted to reflect basin-scale sea surface salinity changes driven by either river runoff from the surrounding islands or changes in regional currents or atmospheric circulation.

These discrepancies highlight the possibility that vegetation-based $\delta^{13}\text{C}$ and speleothem/seawater $\delta^{18}\text{O}$ proxy data may record different seasonal hydroclimatic signals of the AIM (Dubois et al. 2014). Given that $\delta^{13}\text{C}_{\text{wax}}$ is a function of both mean

precipitation and dry season intensity and duration (Dubois et al. 2014, this study), it is possible that the $\delta^{13}\text{C}_{\text{wax}}$ data in Savu Basin records a stronger dry season (Dubois et al. 2014), while $\delta^{18}\text{O}$ data from Liang Luar records a stronger wet season (Griffiths et al. 2009, Ayliffe et al. 2013). Meanwhile, the foraminifera used to generate the $\delta^{18}\text{O}_{\text{sw}}$ records, *G. ruber*, has been shown to record annual instead of seasonal mixed layer conditions (Mohtadi et al. 2011), such that the mean average conditions measured by $\delta^{18}\text{O}_{\text{sw}}$ might reflect little glacial-interglacial change. While these factors might explain these discrepancies, the $\delta^{13}\text{C}_{\text{wax}}$ at Matano and Towuti record expanded C_4 ecosystems signifying a stronger/longer dry season at the same time that Ti concentrations in Towuti sediments decline (Russell et al. 2014). As surface runoff and fluvial erosion are primarily driven by peak precipitation (i.e. wet season), our data suggest reductions in both wet and dry season, and therefore mean annual, precipitation. Alternatively, the discrepancies might be explained by additional factors controlling $\delta^{18}\text{O}$ proxy data in this region, in particular the possibility that different moisture sources and precipitation trajectories change the isotopic composition of the rainfall (LeGrande and Schmidt 2009, Lewis et al. 2010, Carolin et al. 2013). Finally, it is also possible that the apparent discrepancies might point to strong regional hydroclimatic gradients that existed in Indonesia during MIS 2/LGM. The northern Borneo cave and sites used in the $\delta^{18}\text{O}_{\text{sw}}$ studies are situated at the peripheries of the Maritime Continent, and the southern sites are situated near the southernmost limit of present-day ITCZ migration across Indonesia, which might be more sensitive to precessional orbital forcing. In contrast, the $\delta^{13}\text{C}_{\text{wax}}$ records from Palawan and Sulawesi are located near the center of the Maritime Continent where ever-wet conditions and proximity to the Pacific Ocean and its overlying

atmosphere might make the area more sensitive to global temperatures and boundary conditions. Additional records from northern Indonesia could help to evaluate this possibility.

The new data from Matano, combined with other biomarker records from Savu (Dubois et al 2014) and Palawan (Wurster et al. 2010) (Figure 5), corroborate the finding of Russell et al. (2014) that the variations in central Indonesian hydroclimate are most strongly correlated to changes in ice volume (Figure 6). There is little relationship between our proxy data and insolation forcing in the tropics as highlighted by the lack of 21-kyr precessional insolation cycles in our Matano data. However, the $\delta^{13}\text{C}_{\text{wax}}$ records from both Lakes Matano and Towuti appear to show correlations with the regional sea surface temperature (SST), the concentration of atmospheric greenhouse gases (GHG), the Sunda Shelf areal exposure, and the extent of northern hemisphere ice sheets (Figure 6). Among these, the Matano records most closely track the timing of significant Laurentide ice sheet growth during late MIS 3 (Figure 6). Regional SST records show similar temperatures during MIS 3 and MIS 2 and the Sunda Shelf had nearly similar exposed area between 53 and 12 kyr BP, whereas changes in the GHG concentrations between portions of MIS 3 and MIS 2 cannot wholly explain the variations observed in our $\delta^{13}\text{C}_{\text{wax}}$ (Figure 6), suggesting the relatively minor influence of these three forcings compared to that of ice sheet volume and topography. Indeed, modeling studies show that a relatively small change in ice sheet thickness might significantly increase the Hadley mass transport above the Pacific and shift the mean position of the ITCZ southward (e.g. Lee et al. 2014). In this light, the regional MIS 2 drying observed across central Maritime Continent could result from orographic forcing by the Laurentide ice sheet, which

perturbed the tropical Pacific atmospheric circulation and subsequently drove the migration of intense precipitation to the south of central Indonesia.

6. Conclusions

Late Pleistocene and Holocene sediments from Lake Matano in central Sulawesi reveal a paleovegetation and paleohydrological history that is consistent with similar records from the region, in particular the vegetation record from Lake Towuti, which lies at a lower elevation approximately 10 km south of Lake Matano. Together, $\delta^{13}\text{C}_{\text{wax}}$ from Lakes Matano and Towuti indicate that grassland and open canopy forest ecosystems expanded during a more arid and seasonal MIS 2, including during the LGM, whereas closed-canopy rainforest ecosystems predominated in central Sulawesi during MIS 1 and MIS 3, when the climate was wetter and less seasonal. The distinct glacial-interglacial shifts observed in central Sulawesi records during the past 50 kyr are likely due to the sensitivity of Maritime Continent hydroclimate to changes in global climate boundary conditions, and particularly remote high latitude forcings associated with ice volume and thickness. Given the importance of Lake Towuti as the site of an upcoming International Continental Drilling Project, which will attempt to reconstruct regional hydroclimate and vegetation history during the past 650 kyr or longer (Russell and Bijaksana 2012), the similar histories between the two lakes during the past 50 kyr strongly suggest that the climate proxy records of Lake Towuti are representative of the regional hydroclimate at the heart of the Western Pacific Warm Pool.

Within this framework, $\delta^{13}\text{C}_{\text{wax}}$ from Lake Matano are more depleted than those of Lake Towuti, and Lake Matano experienced a lesser enrichment in $\delta^{13}\text{C}_{\text{wax}}$ than Lake Towuti during late MIS 3 and MIS 2. We suggest that the local topographic setting in Sulawesi had a significant impact on local climatic and/or plant type variations across glacial-interglacial timescales, whereby the cooler, wetter climates at the higher elevations surrounding Lake Matano supported C_3 ecosystems during a regionally drier time period. This is consistent with the interpretation of multiple vegetation-based reconstructions from high and low altitude regions within the Maritime Continent, and suggests an important role for mountains in maintaining rainforest elements in the Maritime Continent during glacial-interglacial cycles.

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References

- Aldrian, E. and R. D. Susanto (2003), Identification of three dominant rainfall regions within Indonesia and their relationship to sea surface temperature, *Int. Jour. of Clim.*, 23, 1435-1452.
- Anshari, G., A.P. Kershaw, S. van der Kaars, and G. Jacobsen (2004), Environmental change and peatland forest dynamics in the Lake Sentarum area, West Kalimantan, Indonesia. *Journal of Quaternary Science*, 19, 7, 637-656.
- Ayliffe, L., M. Gagan, J-x Zhao, R.N. Drysdale, J.C. Hellstrom, W.S. Hantoro, M.L. Griffiths, H. Scott-Gagan, E St Pierre, J.A. Cowley, and B.W. Suwargadi (2013), Rapid interhemispheric climate links via the Australasian monsoon during the last deglaciation, *Nature Communications*, 4, 1-6.
- Barton, H. 2005. The case for rainforest foragers: the starch record at Niah Cave, Sarawak. *Asian Perspectives*, 44, 56–72.
- BBKSDA Sulsel (2011), Statistik BBKSDA Sulsel, Ditjen PHKA Kementerian Kehutanan. Makassar.
- Bi, X., G. Sheng., X. Liu, C. Li, and J. Fu (2005), Molecular and carbon and hydrogen isotopic composition of n-alkanes in plant leaf waxes, *Organic Geochemistry* 36, 1405–1417.
- Bruijnzeel, L.A. (2001), Hydrology of tropical montane cloud forest: a reassessment. *Land Use and Water Resources Research* 1, 1.1-1.18.

Cannon, C.H., R.J. Morley, and A.B.G. Bush (2009), The current refugial rainforests of Sundaland are unrepresentative of their biogeographic past and highly

vulnerable to disturbance, *Proc Natl Acad Sci USA* 106(27), 11188–11193

Carolin, S., K.M. Cobb, J.F. Adkins, B. Clark, and A. Tuen (2013), Varied response of western tropical Pacific hydrology to abrupt climate changes, *Science*, doi:10.1126/science.1233797.

Cernusak, L.A., N. Ubierna, K. Winter, J.A.M. Holtum, J.D. Marshall, and G.D.

Farquhar (2013), Environmental and physiological determinants of carbon isotope discrimination in terrestrial plants, *New Phytologist* doi: 10.1111/nph.12423.

Chikaraishi, Y. and H. Naraoka (2007), $\delta^{13}\text{C}$ and δD relationships among three n-alkyl compound classes (n-alkanoic acid, n-alkane and n-alkanol) of terrestrial higher plants. *Org. Geochem.* 38, 198-215

Culmsee, H. and C. Leuschner (2013), Consistent patterns of elevational change in tree taxonomic and phylogenetic diversity across Malesian mountain forests, *Journal of Biogeography* 40, 10, 1997-2010.

Culmsee, H., C. Leuschner, G. Moser, and R. Pitopang (2010), Forest aboveground biomass along an elevational transect in Sulawesi, Indonesia, and the role of Fagaceae in tropical montane rain forests, *Journal of Biogeography*, 37, 960-974.

Dam, R., J. Fluin, P. Suparan, and S. van der Kaars (2001), Palaeoenvironmental developments in the Lake Tondano area (N. Sulawesi, Indonesia) since 33,000 yr BP. *Palaeogeography, Palaeoclimatology, Palaeoecology* 171, 147–183.

- Diefendorf, A.F., K.E. Mueller, S.L. Wing, P.L. Koch, and K.H. Freeman (2010) Global patterns in leaf ^{13}C discrimination and implications for studies of past and future climate. *Proc Natl Acad Sci USA* 107(13), 5738–5743.
- Du, B., C. Liu, H. Kang, P. Zhu, S. Yin, G. Shen, J. Hou, and H. Ilvesniemi (2014), Climatic Control on Plant and Soil $\delta^{13}\text{C}$ along an Altitudinal Transect of Lushan Mountain in Subtropical China: Characteristics and Interpretation of Soil Carbon Dynamics, *PLOS One*. DOI: 10.1371/journal.pone.0086440
- Dubois N., D.W. Oppo, V. Galy, M. Mohtadi, S. van der Kaars, J. Tierney, Y. Rosenthal, T. Eglington, A. Lückge, and B. Linsley (2014), Indonesian vegetation response to changes in rainfall seasonality over the past 25,000 years. *Nature Geoscience* 7, 513-517.
- Ehleringer, J. R., T.E. Cerling, and B.R. Helliker (1997), C_4 photosynthesis, atmospheric CO_2 , and climate. *Oecologia* 112, 285–299.
- Gibbons F., D.W. Oppo, M. Mohtadi, Y. Rosenthal, J. Cheng, Z. Liu, and B.K. Linsley (2014), Deglacial $\delta^{18}\text{O}$ and hydrologic variability in the tropical Pacific and Indian Oceans. *Earth and Planetary Science Letters* 387, 240-251.
- Gimeno, L., A. Stohl, R. M. Trigo, F. Dominguez, K. Yoshimura, L. Yu, et al. (2012), Oceanic and terrestrial sources of continental precipitation. *Rev. Geophys.* 50, RG4003. doi: 10.1029/2012RG000389
- Göltenboth, F., Langenberger, G., and Widmann, P (2006) Forest systems. In: Timotius, K.H., Milan, P.P., Margraf, J (eds.) *Ecology of Insular Southeast Asia: The Indonesian Archipelago*, Elsevier.

- Griffiths M. L., R. N. Drysdale, M. K. Gagan, J.-X. Zhao, L. K. Ayliffe, J. C. Hellstrom, W. S. Hantoro, S. Frisial, Y. -X. Feng, I. Cartwright, E. St. Pierre, M. J. Fischer, and B. W. Suwargadi (2009), Increasing Australian-Indonesian monsoon rainfall linked to early Holocene sea-level rise, *Nature Geoscience*, 2, 636-639.
- Haffner, G. D., P. E. Hehanussa, and D. Hartoto (2001), The biology and physical processes of large lakes of Indonesia, p. 183–194. In M. Munawar and R. E. Hecky [eds.], *The great lakes of the world: Food web, health and integrity*. Backhuys.
- Holbourn, A., W. Kuhnt, H. Kawamura, Z. Jian, P. Grootes, H. Erlenkeuser and J. Xu (2005), Orbitally paced paleoproductivity variations in Timor sea and Indonesian throughflow variability during 460 kyr. *Paleoceanography*, 20, PA 3002.
- Hope, G. S. (2001), Environmental change in the Late Pleistocene and later Holocene at Wanda Site, Soroako, South Sulawesi, Indonesia, *Palaeogeogr. Palaeoclim. Palaeoecol.*, 171, 129-145.
- Hope, G. and Tulip, J. (1994), A long vegetation history from lowland Irian Jaya, Indonesia. *Palaeogeogr., Palaeoclimatol., Palaeoecol.* 109, 385-398.
- Kalnay et al. (1996), The NCEP/NCAR 40-year reanalysis project, *Bull. Amer. Meteor. Soc.*, 77, 437-470.
- Kershaw, A. P., S. van der Kaars, P. Moss, B. Opdyke, F. Guichard, S. Rule, and C. Turney (2006) Environmental change and the arrival of people in the Australian region. Before Farming (online version) 2006/1 article 2.

- Kershaw, A.P., S. van der Kaars. and J.F. Flenley (2011), The Quaternary history of far eastern rainforests (2nd Edn.). In: Bush, M.B., Flenley, J.F. and Gosling, W.D. (Eds.) *Tropical Rainforest Responses to Climate Change*. Springer-Verlag, Berlin Heidelberg, pp. 85-123.
- Lee, S.-Y., J.C.H. Chiang, and P. Chang (2014), Tropical Pacific Response to Continental Ice Sheet Topography, *Climate Dynamics*, DOI: 10.1007/s00382-014-2162-0.
- LeGrande, A.N. and G.A. Schmidt (2009), Sources of Holocene variability of oxygen isotopes in paleoclimate archives, *Clim Past*, 5, 441-455.
- Lehmusluoto, P. and B. Machbub (1997), National Inventory of the Major Lakes and Reservoirs in Indonesia, Expedition Indodanau Technical Report, Revised Ed., Research. Inst. For Water Res. Dev., Min. of Public Works, Rep. of Indonesia, and Dept. of Limnology and Env. Protection, University of Helsinki, Finland, Helsinki.
- Lewis, S. C., A. N. LeGrande, M. Kelley, and G. A. Schmidt (2010), Water vapour source impacts on oxygen isotope variability in tropical precipitation during Heinrich events, *Climate of the Past*, 6, 325– 343.
- Moerman, J.W., K.M. Cobb, J.F. Adkins, H. Sodemann, B. Clark, and A. Tuen (2013), Diurnal to interannual rainfall oxygen isotope variations in northern Borneo driven by regional hydrology. *Earth and Planetary Sciences Letter*, 369, 108-119.

- Mohtadi, M., D. W. Oppo, S. Steinke, J. W. Stuut, E. C. Hathorne, J. Groeneveld, D. Hebbeln, and A. Lückge (2011), Glacial to Holocene swings of the Australian-Indonesian monsoon, *Nature Geoscience*, 4, 540-544.
- Mohtadi, M., M. Prange, D.W. Oppo, R. De Pol-Holz, U. Merkel, X. Zhang, S. Steinke, and A. Lückge (2014), North Atlantic forcing of tropical Indian Ocean climate. *Nature* 509, 76-80.
- Monnin, E., A. Indermühle, A. Dällenbach, J. Flückiger, B. Stauffer, T.F. Stocker, D. Raynaud, and J.-M. Barnola. 2001. Atmospheric CO₂ concentrations over the last glacial termination. *Science*, 291, 112-114.
- Moyle, R. G., S.S. Taylor, C.H. Oliveros, H.C. Lim, C.L Haines, M.A. Rahman, and F.H. Sheldon. 2011. Diversification of an endemic Southeast Asian genus: phylogenetic relationships of the spiderhunters (Nectariniidae), *The Auk*, 128, 777–788.
- Newsome, J. and J. R. Flenley (1988), Late Quaternary vegetation history of the Central Highlands of Sumatra. II.: Palaeopalynology and Vegetational History. *J. Biogeog.*, 15, 555-578.
- Niedermeyer E.M., A.L. Sessions, S.J. Feakins, and M. Mohtadi (2014), Hydroclimate of the western Indo-Pacific Warm Pool during the past 24,000 years. *Proceedings of the National Academy of Science* 111, 9402-9406.
- Partin J. W., K. M. Cobb, J. F. Adkins, B. Clark, and D. P. Fernandez (2007), Millennial-scale trends in west Pacific warm pool hydrology since the Last Glacial Maximum, *Nature*, 449, 452-455.

- Peel, M. C., B.L. Finlayson, and T.A. McMahon (2007), Updated world map of the Köppen– Geiger climate classification, *Hydrol. Earth Syst. Sci.*, 11, 1633–1644.
- Petit, J.R., J. Jouzel, D. Raynaud, N.I. Barkov, J.-M. Barnola, I. Basile, M. Benders, J. Chappellaz, M. Davis, G. Delayque, M. Delmotte, V.M. Kotlyakov, M. Legrand, V.Y. Lipenkov, C. Lorius, L. Pépin, C. Ritz, E. Saltzman, and M. Stievenard. (1999), Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica, *Nature* 399, 429-436.
- PT Vale Indonesia (unpubl.) Data from weather station near the Malili Lakes.
- Qu, T., Y. Du, J. Strachan, G. Meyers and J. Slingo (2005), Sea surface temperature and its variability in the Indonesian region. *Oceanography* 18, 50–61.
- Read, J. and G. Farquhar (1991), Comparative studies in Nothofagus (Fagaceae). I. Leaf carbon isotope discrimination, *Functional Ecology* 5, 684-695.
- Renggono, F. (2011) Pola Sebaran Hujan di DAS Larona (in Indonesian). *Jurnal Sains dan Teknologi Modifikasi Cuaca*, 12, 1, 17-24.
- Russell, J. and S. Bijaksana (2012), The Towuti Drilling Project: Paleoenvironments, Biological Evolution, and Geomicrobiology of a Tropical Pacific Lake., *Scientific Drilling*, 14, 68-71.
- Russell, J.M., H. Vogel, B.L. Konecky, S. Bijaksana, Y. Huang, M. Melles, N. Wattrus, K. Costa, and J.W. King (2014), Glacial forcing of central Indonesian hydroclimate since 60,000 y B.P., *Proc Nat Ac Sci USA*, doi:10.1073/pnas.1402373111.

- Sachse, D, I. Billault, G.J. Bowen, Y. Chikaraishi, K.H. Freeman, C.R. Magill, F.A. McInerney, M.T.J. van der Meer, P. Polissar, R.J. Ribons, J.P. Sachs, H.-L. Schmidt, A.L. Sessions, J.W. White, J.B. West, and A. Kahmen (2012), Molecular paleohydrology: interpreting the hydrogen-isotopic composition of lipid biomarkers from photosynthesizing organisms. *Annu. Rev. Earth Planet. Sci.*, 40, 221–249.
- Slik, J. W. F., A.D. Poulsen. et al. (2003), A floristic analysis of the lowland dipterocarp forests of Borneo, *Journal of Biogeography*, 30, 1517–1531.
- Stott, L., A. Timmermann, and R. Thunell (2007), Southern Hemisphere and deep-seawarming led deglacial atmospheric CO₂ rise and tropical warming. *Science* 318(5849), 435–438.
- Tierney, J.E., D.W. Oppo, A.N. LeGrande, Y. Huang, Y. Rosenthal, and B.K. Linsley (2012), The influence of Indian Ocean atmospheric circulation on warm pool hydroclimate during the Holocene epoch. *J. Geophys. Res.*, 117, D19108, doi:10.1029/2012JD018060.
- Tripathi, A.K., S. Sahany, D. Pittman, R.A. Eagle, J.D. Neelin, J.L. Mitchell, and L. Beaufort (2014), Modern and glacial tropical snowlines controlled by sea surface temperature and atmospheric mixing. *Nature Geoscience*, 7, 205–209 doi:10.1038/ngeo2082.
- van Baalgoy, M.M.J and I.G.M. Tantra (1986) The vegetation in two areas in Sulawesi, Indonesia. *Buletin Penelitian Hutan*, Special Edition, pp. 1-61.
- van der Kaars, S., D. Penny, J. Tibby, J. Fluin, R. Dam, and P. Suparan (2001) Late Quaternary palaeoecology, palynology and palaeolimnology of a tropical

- lowland swamp: Rawa Danau, West Java, Indonesia. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 171, 185–212.
- van der Kaars, S., X. Wang, A.P. Kershaw, F. Guichard, and D.A. Setiabudi (2000) A Late Quaternary palaeoecological record from the Banda Sea, Indonesia: Patterns of vegetation, climate and biomass burning in Indonesia and northern Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 155, 135–153.
- van der Molen, M.K., A.J. Dolman, M.J. Waterloo and L.A. Bruijnzeel (2006), Climate is affected more by maritime than by continental land use change: a multiple-scale analysis. *Global and Planetary Change*, 54, 128-149.
- Vogts, A., H. Moossen, F. Rommerskirchen, and J. Rullkötter (2009), Distribution patterns and stable carbon isotopic composition of alkanes and alkan-1-ols from plant waxes of African rain forest and savanna C₃ species. *Organic Geochemistry*, 40, 1037-1054.
- Waelbroeck, C., L. Labeyrie, E. Michel, J.C. Duplessy, J. McManus, K. Lambeck, E. Balbon, and M. Labracherie (2002), Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records. *Quaternary Science Reviews*, 21, 295-305.
- Wei, Z., X. Jibin., C. Jixiu, Z. Yanming, M. Qiaohong, O. Jun, C. Ying, Z. Zhiguo, and L. Wei (2010), Bulk organic carbon isotopic record of lacustrine sediments in Dahu swamp, eastern Nanling mountains in South China: implication for catchment environmental and climatic changes in the last 16,000 years, *Journal of Asian Earth Sciences*, 38, 162–169.

Whitten, T., M. Mustafa and G.S. Henderson (1987), *The Ecology of Sulawesi*. Gadjah
Mada University Press, Yogyakarta, Indonesia. Republished by Periplus, Hong
Kong, 2002.

Wurster, C.M., M.I. Bird, I.D. Bull, F. Creed, C. Bryant, J.A.J. Dungait, and V. Paz
(2010), Forest contraction in north equatorial Southeast Asia during the Last
Glacial Period, *Proceedings of the National Academy of Sciences*, USA, 107,
15508–15511.

ID	Material	Depth (m)	¹⁴ C Age	± ¹⁴ C	Reservoir correction (¹⁴ C years)	Reservoir corrected ¹⁴ C Age	Calibrated age (yr BP)
2B-1-107.5	bulk	0.40	4610	30	977.53	3632	3944
2B-1-109	wood/plant	0.41	3680	80	n/a	3680	4020
2B-2-51.5	bulk	1.26	10250	50	977.53	9272	10453
2B-3-1.5	wood/plant	2.23	15900	60	n/a	15900	19168
2B-3-5.5	bulk	2.27	17100	70	977.53	16122	19457
2B-3-94	bulk	3.15	21600	90	977.53	20622	24835
2B-4-47.5	bulk	4.18	28000	130	977.53	27022	31064
2B-5-23	bulk	5.30	33900	280	977.53	32922	37008
2B-6-54.5	bulk	6.62	42100	330	977.53	41122	44629
2B-7-16	bulk	7.74	46000	460	977.53	45022	48391

Table 1. Calibrated and reservoir-corrected ages used to determine the age model for core MAT10-2B. The reservoir correction is based on the average of two interpolated paired ages.

NCEP/NCAR Reanalysis 1980-2010 Climatology Surface Precipitation Rate (mm/day)

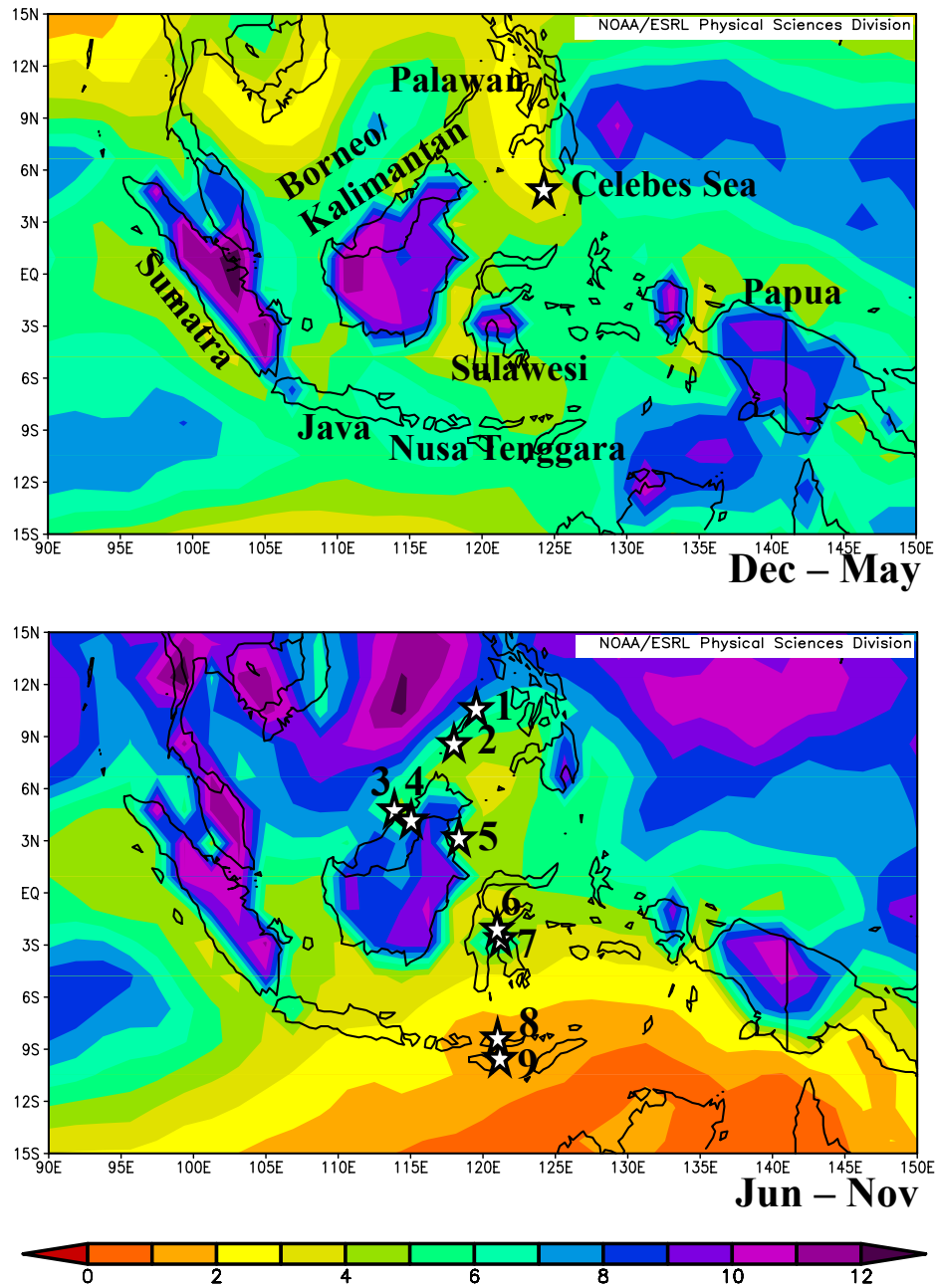


Figure 1. Surface precipitation rates for December to May and June to November from NCEP/NCAR Reanalysis (Kalnay et al. 1996). Locations of our study site and other sites/records discussed in the text are shown on the map: 1) Makangit Cave, 2) Gangub

Cave, 3) Niah Cave, 4) Gunung Buda Cave, 5) Tarakan Delta, 6) Lake Matano, 7) Lake Towuti, 8) Liang Luar Cave, 9) Savu Basin.

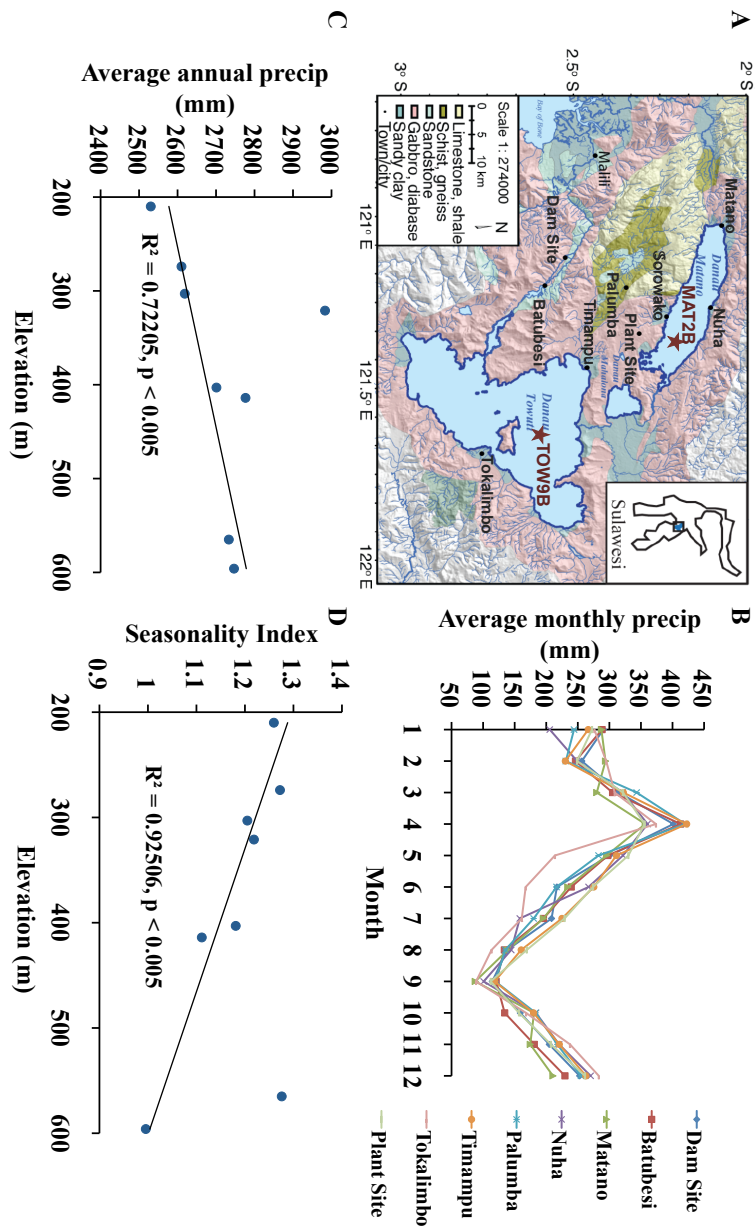


Figure 2. (A) Geological map of the Malili Lakes region. Core locations in both lakes are indicated with red stars. Locations of weather stations near the lakes are shown on the map. (B) Monthly mean precipitation for eight weather stations near Lakes Matano and Towuti (PT Vale Indonesia, unpublished) from 1997 to 2006. (C) Average annual

precipitation for those weather stations plotted against their elevations. A trend line is fitted to the data, omitting one outlier (Timampu). **(D)** Seasonality index (difference between the average rainfall during the wettest and driest months divided by the mean monthly rainfall) for those weather stations were computed and plotted against their elevations. A trend line is fitted to the data, omitting one outlier (Palumba).

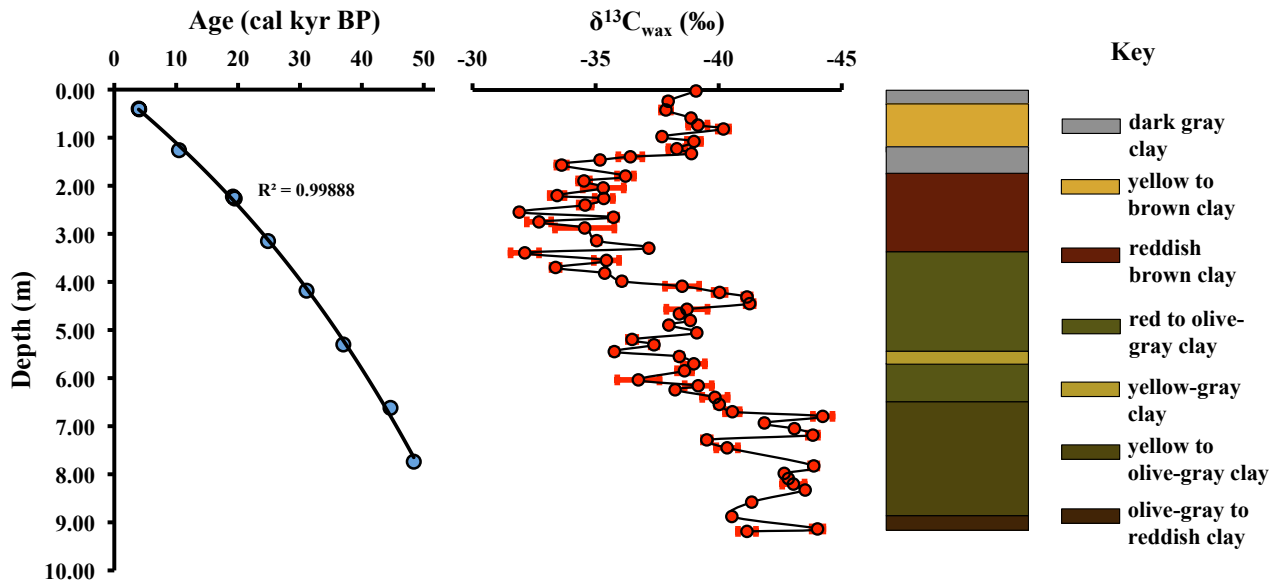


Figure 3. Age-depth profile, $\delta^{13}\text{C}$ of C_{28} n-alkanoic acids ($\delta^{13}\text{C}_{\text{wax}}$) data, and lithostratigraphic description of Lake Matano's Core 2B. 1-sigma errors for $\delta^{13}\text{C}_{\text{wax}}$ measurements are plotted.

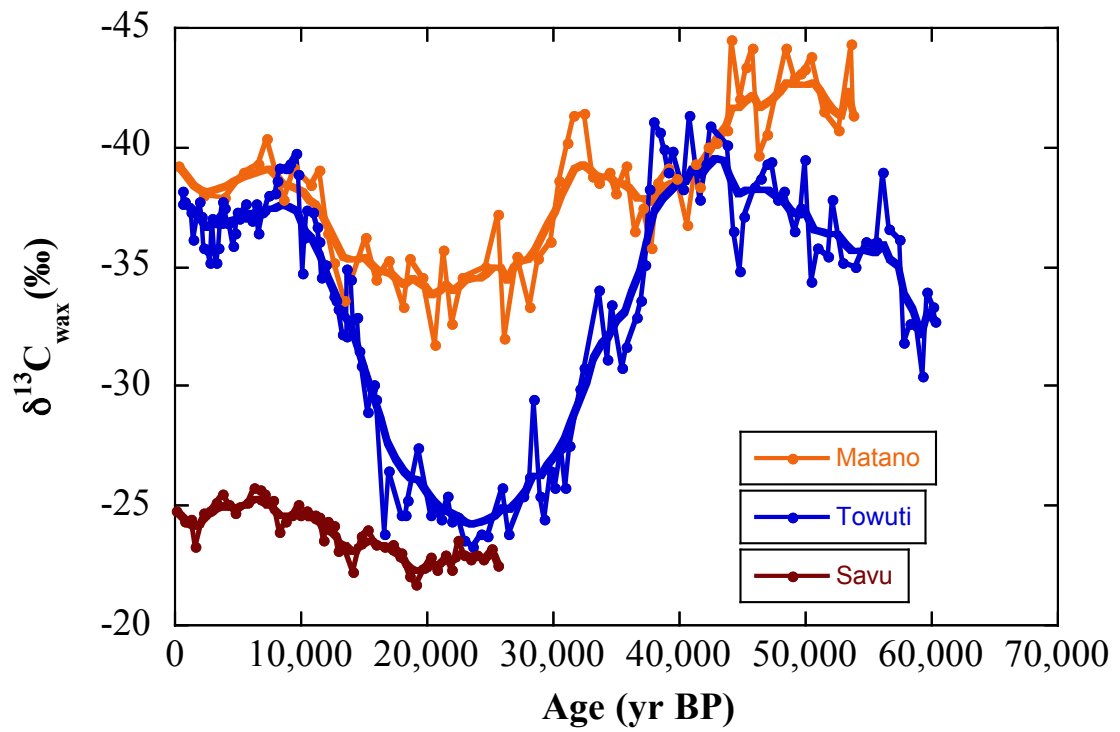


Figure 4. $\delta^{13}\text{C}$ from Lakes Matano and Towuti in central Sulawesi and Savu Basin near the Nusa Tenggara islands. Dashed lines represent a Gaussian smoothing of the data to emphasize long-term trends.

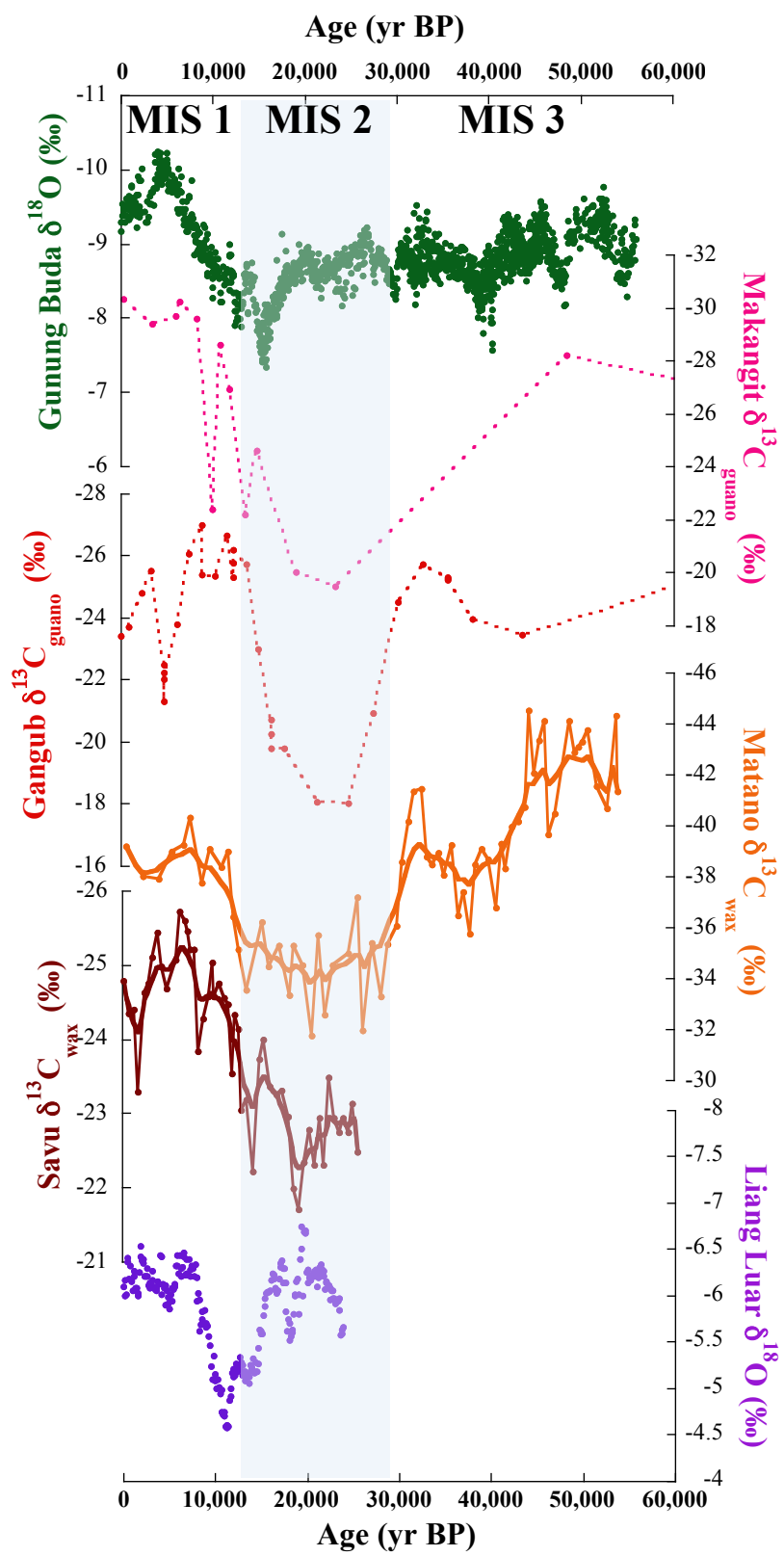


Figure 5. Comparison of Lake Matano $\delta^{13}\text{C}_{\text{wax}}$ record with other $\delta^{13}\text{C}$ records of regional vegetation. From top to bottom: 1) Northern Borneo/Gunung Buda $\delta^{18}\text{O}$ record, ice volume-corrected (Partin et al. 2007 and Carolin et al. 2013), 2) C_{29} n-alkane $\delta^{13}\text{C}_{\text{guano}}$ from Makangit on Palawan Island (Wurster et al. 2010), 3) C_{29} n-alkane $\delta^{13}\text{C}_{\text{guano}}$ from Gangub on Palawan Island (Wurster et al. 2010), 4) Matano $\delta^{13}\text{C}_{\text{wax}}$ (this study), 5) Savu $\delta^{13}\text{C}_{\text{wax}}$ (Dubois et al. 2014), 6) Flores/Liang Luar $\delta^{18}\text{O}$ record, ice volume-corrected (Griffiths et al. 2009, Ayliffe et al. 2013).

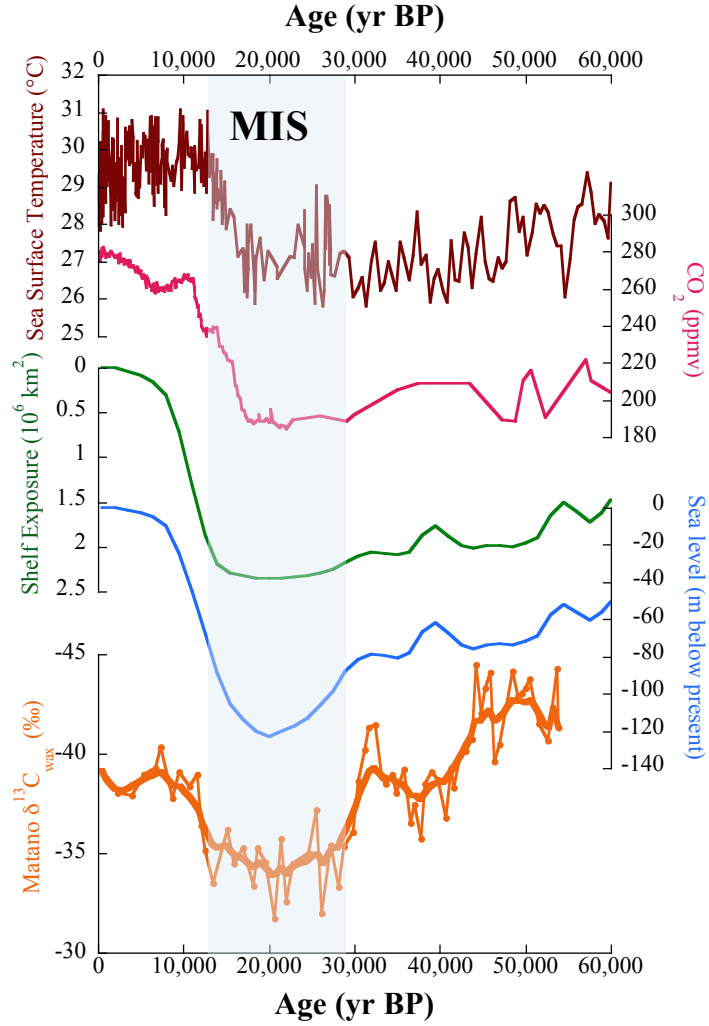


Figure 6. Comparison of Lake Matano $\delta^{13}\text{C}_{\text{wax}}$ records with climate forcings. From top to bottom: 1) Mg/Ca sea surface temperature record from Celebes Sea (Stott et al. 2007), 2) CO_2 concentration record from Dome C and Vostok, Antarctica (Monnin et al. 2001, Petit et al. 1999), 3) areal shelf exposure of Sunda Shelf, obtained by applying sea level fluctuations data to Sunda Shelf bathymetry, as described in Russell et al. (2014), 4) sea level fluctuations, as a proxy for global ice volume, derived from benthic foraminifera oxygen isotopes, as described in Waelbroeck et al. (2002), 5) Matano $\delta^{13}\text{C}_{\text{wax}}$ (this study).

CHAPTER THREE

Hydrological and vegetation shifts in the Wallacean region of central Indonesia since the Last Glacial Maximum

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Abstract

Precipitation is the most important variable of Indonesian climate, yet there are substantial uncertainties about past and future hydroclimate dynamics over the region. This study explores central Indonesian vegetation and rainfall and associated changes in atmospheric circulation during the past 26,000 years in Wallacea, a biogeographical area in central Indonesia, wedged between the Sunda and Sahul shelves and known for its exceptionally high rainforest biodiversity. We use terrestrial plant biomarkers from sediment cores retrieved from Mandar Bay, off west Sulawesi, to reconstruct changes in Wallacean vegetation and climate since the Last Glacial Maximum (LGM). Enriched leaf wax carbon isotope ($\delta^{13}\text{C}_{\text{wax}}$) values recorded in Mandar Bay during the LGM, together with other regional vegetation records, document grassland expansion, implying a regionally dry, and possibly more seasonal, glacial climate. Depleted leaf wax deuterium isotope ($\delta\text{D}_{\text{wax}}$) values in Mandar Bay during the LGM, and low reconstructed precipitation isotope compositions from nearby sites, reveal a southward shift of the mean position of the Intertropical Convergence Zone and an intensified Austral-Asian monsoon circulation, likely due to strong southern hemisphere summer insolation and the presence of large northern hemisphere ice sheets. Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$ was anti-correlated with $\delta\text{D}_{\text{wax}}$ during the LGM and the last deglaciation, but was positively correlated during most of the Holocene, indicating time-varying controls on the isotopic composition of rainfall in this region. The inundation event of the Sunda Shelf and in particular the opening of the Java Sea and Karimata Strait between 9.4 and 11.1 thousand years ago might have provided new moisture sources for regional convection and/or influenced

moisture source trajectories, providing the trigger for shifts in atmospheric circulation and the controls on precipitation isotope compositions from the LGM to the Holocene.

Keywords

Precipitation; Vegetation; Biomarker; Isotope; Indonesia; Last Glacial Maximum

1. Introduction

Alterations in modern-day Indonesian rainfall patterns influence the stability of its rich and diverse tropical rainforests (Underwood et al. 2014) and impact the livelihoods of more than a quarter billion people living in the region (Naylor et al. 2007, Oktaviani et al. 2011). However, there is relatively weak agreement among Intergovernmental Panel on Climate Change climate models on the trend and magnitude of future precipitation changes over Indonesia (Knutti and Sedláček 2013), reflecting substantial uncertainty about hydroclimate dynamics in the archipelago under future climate boundary conditions. Similarly, there is a long-running debate on changes in Indonesian hydroclimate during the Last Glacial Maximum (LGM), when atmospheric CO₂ concentrations were approximately 80 ppm lower than pre-industrial values and Sundaland and Sahulland were two, separate contiguous landmasses (Figure 1). Nearly all terrestrial- and marine-based climate proxy records indicate cooler temperatures in Indonesia during the LGM, but regional precipitation patterns are still poorly understood (Reeves et al. 2013 and references therein). Vegetation responses to these rainfall variations are even less clear (Wurster et al. 2014). Given the importance of LGM climate

to climate model validation and testing (e.g. diNezio and Tierney 2013), new proxy data of precipitation and vegetation response during the LGM are needed to close these knowledge gaps.

At the heart of Indonesia lies Wallacea, a biogeographical region comprised of Sulawesi, the Maluku Islands, and the Lesser Sunda Islands (Figure 1). Wallacea is separated from Sundaland (the Malay Peninsula, Sumatra, Borneo, Java, and Bali) to the west and from Sahulland (Papua and Australia) to the east by deep seaways, and thus had a different environmental history from the neighboring islands connected to the exposed Sunda and Sahul shelves during the LGM. Wallacea is a global biodiversity hotspot due to its high species diversity and high risk of habitat loss (Myers et al. 2000). Here we assess Wallacean climate and vegetation within the larger context of climate changes across Indonesia since 26 kyr BP to improve our understanding of regional vegetation and climate history, with a particular emphasis on the LGM.

The LGM climate of the Indonesian archipelago has proved challenging to understand, due to significant disagreement among proxy reconstructions in the direction and magnitude of regional precipitation change. This uncertainty could be partly due to the use of different precipitation proxies in different archives, which are sensitive to different elements of paleohydrologic change. For example, proxy records of precipitation isotopes such as speleothem oxygen isotopic compositions (e.g. Griffiths et al. 2009, Ayliffe et al. 2013) and seawater oxygen isotope data derived from foraminifera (Mohtadi et al. 2011) have been interpreted to reflect changes in the amount of annual precipitation. This interpretation is underpinned by the dilution of isotopically-enriched seawater with isotopically-depleted rainfall. The latter is influenced by the so-called

“amount effect”, in which local rainfall amount is considered as the primary control of rainfall isotopes (e.g. Dansgaard 1964 and Rozanski et al. 1992). However, recent modeling studies and observations of modern precipitation isotope data indicate that precipitation isotopes can record a range of atmospheric circulation processes including moisture sources and trajectories (e.g. LeGrande and Schmidt. 2009, Kurita et al. 2009, Lewis et al. 2010, Moerman et al. 2013). Better constraints on the interpretation of rainfall isotope proxy data through comparisons of multiple climate proxies will therefore help to elucidate not only the regional hydroclimate sensitivities to global climate forcings, including atmospheric greenhouse gas levels (e.g. Tachikawa et al. 2013), orbital variations (e.g. Holbourn et al. 2005, Kershaw et al. 2011, Tierney et al. 2012, Carolin et al. 2013), ice volume changes (e.g. Lee et al. 2014, Russell et al. 2014, Wicaksono et al. 2015), and aerial exposure of the Sunda and Sahul Shelves (Griffiths et al. 2009, Partin et al. 2007, diNezio and Tierney 2013, Dubois et al. 2014, Niedermeyer et al. 2014), but also the atmospheric circulation changes that govern the regional climate response.

Here we employ stable isotope analyses of terrestrial plant biomarkers to reconstruct paleoclimate and paleovegetation in Wallacea. Carbon and hydrogen isotope ratios of terrestrial plant leaf wax compounds (e.g. *n*-alkanes, *n*-acids, and others) can record changes in vegetation type and the isotopic composition of precipitation, providing valuable paleoclimate data (Sachse et al. 2012). We analyzed the carbon and hydrogen isotopic composition of long-chain *n*-alkanoic acids extracted from a new marine sediment core collected from Mandar Bay, offshore central Sulawesi, and compare these data with several published hydroclimate and vegetation records from Wallacea and

other parts of Indonesia. Our record builds upon and extends previous reconstructions from Mandar Bay (Tierney et al. 2012) and helps to: 1) determine the patterns of Wallacean hydroclimate change since the LGM, 2) understand the response of vegetation to long-term hydroclimate variations in this high biodiversity region, and 3) elucidate the climate processes controlling rainfall isotope variations through time.

2. Study Region

Wallacea lies in central Indonesia and extends from 5 °N to 12 °S (Figure 1). Precipitation in the northern part of Wallacea (the majority of Sulawesi and the Maluku Islands, from 5 °N to 6 °S averages more than 2000 mm/year, with all months experiencing rainfall of more than 50 mm/month. Despite the lack of a dry season, there is a distinct wet season from November through May, when strong northwesterly trade winds associated with the Austral-Asian Monsoon flow over the region, carrying moisture originating in the northern hemisphere (Aldrian and Susanto 2003). In contrast, the southern part of Wallacea (the Lesser Sunda Islands from 7 °S to 12 °S) is notably drier, with average annual precipitation between 500-1500 mm. This area experiences a long dry season between April and November, when the southeast trade winds bring dry air from the Australian land mass (Monk et al. 1997). Both regions, but in particular in central Sulawesi, exhibit strong orographic controls on precipitation, with higher elevation areas having up to twice the rainfall of the coastal zones (Whitten et al. 1987, Monk et al. 1997).

Vegetation distribution in the Wallacean region tracks variations in precipitation (Figure 2). Wetter, northern Wallacea is dominated by evergreen rainforests, including lowland, lower montane, upper montane, and subalpine rainforests that vary as function of altitude (Whitten et al. 1987). Conversely, a significant portion of the Lesser Sunda Islands is covered in tropical deciduous forests and savanna woodlands. Ever-wet rainforests in the Lesser Sunda Islands are found only at high elevations and in areas facing the rain-bearing winds (Monk et al. 1997).

Our leaf wax-based hydroclimate and vegetation records were reconstructed from a marine sediment core retrieved from Mandar Bay during the 2011 Sonne-217 “MAJA” expedition. Mandar Bay is west of central Sulawesi adjacent to the Makassar Strait (Figure 2). Core SO-217-18515-3 (3.63 °S, 119.36 °E, water depth: 688 m, core length: 12.175 m) was obtained 12 km off the Saddang (or Sa’dan) river mouth. The Saddang River catchment has a mean elevation of 1173 m, a maximum altitude of 3428 m, a length of 182 km, and a total watershed area of approximately 6,380 km², covering much of the interior of west central Sulawesi including Tana Toraja. The Saddang River lies at a vegetation divide with evergreen C₃ rainforests in the catchment to the north (Whitten et al. 1987), and tropical deciduous forests with substantial C₄ grasses to the south (Figure 2). These more open vegetation types are typical of the natural vegetation in the southern arm of Sulawesi, which has a transitional climate between the wetter central Sulawesi and drier Lesser Sunda Islands (Whitten et al. 1987).

3. Methods

We analyzed the carbon and hydrogen isotopic composition of long-chain *n*-alkanoic acids from terrestrial leaf waxes ($\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{wax}}$) on sediment samples from core SO-217-18515-3. We extracted lipids from homogenized, freeze-dried sediment samples using a Dionex Accelerated Solvent Extractor and purified acids from the lipid extracts using aminopropyl column chromatography. The fatty acids were methylated with acidified methanol of a known isotopic composition, then further purified using silica gel chromatography. The resulting fatty acid methyl esters were analyzed via gas chromatography isotope ratio monitoring-mass spectrometry (GC-IR-MS) using Thermo Delta V Plus (for $\delta^{13}\text{C}_{\text{wax}}$) and Thermo Delta Plus XL (for $\delta\text{D}_{\text{wax}}$) at Brown University.

In total, we analyzed 100 samples for $\delta^{13}\text{C}_{\text{wax}}$ and 147 samples for $\delta\text{D}_{\text{wax}}$ in duplicate or greater. The average analytical error for $\delta^{13}\text{C}$ analysis of C_{28} *n*-acid was 0.14 ‰, whereas for $\delta\text{D}_{\text{wax}}$ it was 1.96 ‰. Measurements on C_{28} *n*-acid are broadly representative of the isotopic composition of the long-chain waxes. The correlations of $\delta^{13}\text{C}$ values between *n*-acid homologues were strong, with $r^2 = 0.77$, $p < 0.001$ for $\delta^{13}\text{C}_{\text{wax}}$ of C_{28} and C_{26} , and $r^2 = 0.93$, $p < 0.001$ for $\delta^{13}\text{C}_{\text{wax}}$ of C_{28} and C_{30} . The correlations were weaker for δD measurements (0.47 and 0.20 for C_{28} and C_{26} and for C_{28} and C_{30} , respectively). We report $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{wax}}$ ($\delta\text{D}_{\text{precip}}$) values of the C_{28} homologue, as it is the most abundant homologue in our sediment samples and has the lowest analytical error. Except where otherwise noted, we interpret the $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{wax}}$ values of C_{28} as representative isotope compositions of the waxes preserved in our core.

The age model for core SO-217-18515-3 was obtained by linearly interpolating sixteen ^{14}C age control points (2 from woods and 14 from planktonic foraminifera), which covered the past 26 kyr BP (Table 1). ^{14}C ages were calibrated with the IntCal 13.14C curve using Calib 7.0.1 (Reimer et al. 2013). The age control points indicate fairly continuous sedimentation rates, averaging 45 cm/kyr (Figure 3). More detailed discussion on the age model can be found in Schroeder et al. (in review).

4. Ecological and climatic interpretations of leaf wax isotopes

$\delta^{13}\text{C}_{\text{wax}}$ is a proxy for changes in vegetation type, and in particular the relative abundance of C_3 versus C_4 plants. $\delta^{13}\text{C}_{\text{wax}}$ for C_4 plants range from -19 ‰ to -26 ‰, whereas values for C_3 plants range from -30 ‰ to -45 ‰ (Chikaraishi and Naraoka 2007). In the tropics, the relative abundance of C_3 versus C_4 plants is largely controlled by precipitation and related hydroclimate variables, with dry environments favoring C_4 grasses, and wet environments favoring C_3 trees (Scheffuss et al. 2005). Changes in the efficiency of carbon and water utilization within C_3 and C_4 plants may also affect $\delta^{13}\text{C}_{\text{wax}}$, with more positive $\delta^{13}\text{C}$ values in drier climates (Diefendorf et al. 2010). This process generally has a smaller effect on $\delta^{13}\text{C}_{\text{wax}}$ than the shifts between C_3 and C_4 plants; global C_3 biomass exhibits a ~6 ‰ enrichment across precipitation gradients ranging from closed-canopy rainforest to desert ecosystems (Diefendorf et al. 2010).

Within Indonesia, $\delta^{13}\text{C}_{\text{wax}}$ values observed in marine surface marine sediments are consistent with climatically-driven vegetation gradients in the region, with strong correlations between $\delta^{13}\text{C}_{\text{wax}}$ and average monthly precipitation and rainfall seasonality

(Dubois et al. 2014). $\delta^{13}\text{C}_{\text{wax}}$ values from surface marine sediments retrieved near the Lesser Sunda Islands, where C_4 grasslands and savannas are common, are enriched compared to those from wetter, less seasonal regions offshore from Sulawesi including Mandar Bay. Thus, we assume that our $\delta^{13}\text{C}_{\text{wax}}$ record reflects the types of vegetation that existed in west-central Sulawesi, and particularly the catchment of the Saddang River, and, by inference, the region's hydroclimate.

$\delta\text{D}_{\text{wax}}$ has been shown to reliably track the δD of local precipitation ($\delta\text{D}_{\text{precip}}$) in a range of terrestrial biomes and climatic conditions (Sachse et al. 2012 and references therein). This includes many sites in the tropics, where $\delta\text{D}_{\text{precip}}$ is influenced by many factors, including local/regional precipitation amount via “the amount effect” (Kurita et al. 2009), convective intensity/rain recycling (Worden et al. 2007), local SST interactions (Moerman et al. 2013), downdrafts and large-scale subsidence (Brown et al. 2008), cloud types and processes (Kurita et al. 2011), and horizontal moisture advection (Brown et al. 2008). Any spatial and temporal changes in the atmospheric circulation and precipitation are thus likely to have at least some impact on $\delta\text{D}_{\text{precip}}$. While $\delta\text{D}_{\text{wax}}$ can thus provide information on a variety of hydroclimatic processes, the isotopic fractionation between $\delta\text{D}_{\text{precip}}$ and $\delta\text{D}_{\text{wax}}$ (i.e. apparent fractionation) is sensitive to plant physiology (Sachse et al. 2012 and references therein). Here we report $\delta\text{D}_{\text{precip}}$ after correcting $\delta\text{D}_{\text{wax}}$ for changes in vegetation type following Konecky et al. (2015). In brief, $\delta^{13}\text{C}_{\text{wax}}$ values from the same wax samples were used to infer the relative abundance of C_3 and C_4 vegetation. We then used apparent fractionation values for C_3 and C_4 end-members obtained from the literature (Sachse et al. 2012, Chikaraishi and Naraoka 2007) to correct $\delta\text{D}_{\text{wax}}$. Due to differences in the number of samples measured for carbon and hydrogen isotopes, we

interpolated $\delta^{13}\text{C}_{\text{wax}}$ values to match the $\delta\text{D}_{\text{wax}}$ data points. To isolate atmospheric signals in our proxy data we applied a correction for global ice-volume changes to the $\delta\text{D}_{\text{precip}}$ reconstruction based on benthic foraminiferal oxygen isotope data (Waelbroeck et al. 2002) converted to δD values using the global meteoric water line (Craig 1961).

To facilitate comparison to $\delta^{18}\text{O}$ records from regional speleothems, we also converted ice volume-corrected $\delta^{18}\text{O}_{\text{speleothem}}$ from Liang Luar Cave, Flores and Gunung Buda Cave, Borneo (Figure 2, Figure S1) to $\delta\text{D}_{\text{precip}}$. Because carbonate precipitation is temperature-dependent, we use the carbonate-water paleotemperature equation (Fairchild and Baker 2012) to calculate estimates of dripwater $\delta^{18}\text{O}$ from the $\delta^{18}\text{O}_{\text{speleothem}}$ values measured on cave calcite. For Flores, assuming that Liang Luar cave temperatures have always been ~ 3 degrees cooler than sea surface temperatures (Griffiths et al. 2010), we subtracted 3 degrees from the closest available sea surface temperature records (Lombok Basin, Levi et al. 2007) to estimate past temperatures of formation of cave calcite (Figure S1). We then subtracted 1‰ from the computed dripwater $\delta^{18}\text{O}$, as Griffiths et al. (2010) found a ~ 1 -permil enrichment in $\delta^{18}\text{O}$ of the fluid inclusions within Liang Luar speleothems relative to local rainfall. Finally, we converted Flores $\delta^{18}\text{O}_{\text{precip}}$ to $\delta\text{D}_{\text{precip}}$ using modern local meteoric water line computed by Griffiths et al. (2009) (Figure S1). Similar computations and corrections were done on Gunung Buda $\delta^{18}\text{O}_{\text{speleothem}}$ (Figure S1) based on sea surface temperature records from the Celebes Sea (Stott et al. 2007) and measurements of local meteorology and in-situ rainfall and dripwater isotopes (Cobb et al. 2007, Moerman et al. 2013).

5. Results

In the past 26 kyr BP, Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$ varied between -24.5 and -29.6 ‰ (Figure 4). From ~26 to 19 kyr BP, $\delta^{13}\text{C}_{\text{wax}}$ was ^{13}C -enriched, averaging -25.0 ‰, followed by a gradual transition to more negative values during deglaciation, until it reaches relatively stable values averaging -28.3 ‰ between 13 and 7 kyr BP. Afterwards, $\delta^{13}\text{C}_{\text{wax}}$ becomes even more negative, reaching its most negative value (-29.6 ‰) ca. 5 kyr BP, before rapidly shifting toward more positive values of -27.0 ‰ in the late Holocene.

Mandar Bay $\delta\text{D}_{\text{wax}}$ varied between -160 and -172 ‰ during the past 26 kyr BP (Figure 5). Applying our vegetation corrections to estimate $\delta\text{D}_{\text{precip}}$, we observe a slightly larger range of values (from -88 to -70 ‰), but similar overall trends to the uncorrected $\delta\text{D}_{\text{wax}}$ (Figure 5). The average difference of ~92 ‰ between $\delta\text{D}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ represents the average apparent fractionation specific to C_{28} *n*-acid (Konecky et al. 2015). More negative values for $\delta\text{D}_{\text{precip}}$ were found during the LGM and early in the last deglaciation, between 26 and ~17.5 kyr BP, averaging -82.5 ‰, with the most negative value (-88 ‰) at ca. 20 kyr BP. Starting at 18 kyr BP, $\delta\text{D}_{\text{precip}}$ becomes more positive, and averages -73 ‰ between 14 and 10 kyr BP. At the start of the Holocene, $\delta\text{D}_{\text{precip}}$ becomes significantly more negative toward the mid-Holocene, and averages -78 ‰ between 6 and 3 kyr BP. Afterwards, $\delta\text{D}_{\text{precip}}$ shifts to more positive values of approximately -73 ‰ in the late Holocene.

The reconstructed late Holocene $\delta\text{D}_{\text{precip}}$ value of ~-73 ‰ is slightly more negative than $\delta\text{D}_{\text{precip}}$ values measured on precipitation collected in recent years at two

Sulawesi sites, Timampu (Figure 6) and Makassar (Figure 7). The amount-weighted annual average δD_{precip} are $\sim -61\text{‰}$ in Timampu (Konecky et al. 2015) and $\sim -44\text{‰}$ in Makassar (Kurita et al. 2009). The excess depletion observed in Mandar Bay might be due to elevation effects, given the 1170 m mean elevation of the Saddang River catchment, or due to inaccurate estimates of the biosynthetic fractionation of deuterium vs. hydrogen in the formation of leaf waxes by Indonesian vegetation. Whatever the case, our reconstructed δD_{precip} is within the seasonal range of isotopic values observed regionally.

6. Discussion

6.1 Wallacean vegetation and rainfall histories since the LGM

Vegetation records covering the LGM to present from Wallacea are limited to pollen records from the Banda Sea (van der Kaars et al. 2000), Lake Tondano on the northeastern tip of Sulawesi (Dam et al. 2001), and the Wanda site near Lake Matano in central Sulawesi (Hope 2001), and to plant biomarker records from Lake Towuti (Russell et al. 2014), Lake Matano (Wicaksono et al. 2015), and Savu Sea (Dubois et al. 2014; see Figure 2). These records span latitudes extending from 1°N to 8°S , and despite the scarcity of data over such a large region, they exhibit spatially and temporally coherent changes that fit the vegetation and precipitation gradients observed in Wallacea today, as discussed below.

In the present, Wallacea's southernmost island, Sumba, has the driest, most seasonal climate in the region, resulting in the dominance of tropical deciduous forests in northern Sumba and savannas in its southern lowlands (Peel et al. 2007). Correspondingly, the core-top $\delta^{13}\text{C}_{\text{wax}}$ values at Savu Sea, adjacent to Sumba, are the most positive in the region (-25 ‰) and suggest up to 20 % C_4 vegetation (Dubois et al. 2014). On the other hand, the modern $\delta^{13}\text{C}_{\text{wax}}$ values recorded from Lake Towuti and Lake Matano (Figure 8) in the mountainous region of central Sulawesi, average -39 ‰ and -37 ‰, respectively, reflecting closed-canopy C_3 rainforests and montane rainforests with limited to no C_4 input (Russell et al. 2014, Wicaksono et al. 2015). Compared to these records, our Mandar Bay core-top $\delta^{13}\text{C}_{\text{wax}}$ values ($\sim$ -27 ‰) are similar to, but more negative than, Savu Sea, and much more positive than the records from Lakes Matano and Towuti (Figure 8), reflecting the mix of C_4 grassland and tropical deciduous forest in southwestern Sulawesi. This vegetation type likely stems from the area's transitional precipitation regime between the drier and seasonal Lesser Sunda Islands and the ever-wet interior central Sulawesi (Whitten et al. 1987).

Variations in $\delta^{13}\text{C}_{\text{wax}}$ in Mandar Bay records and other Wallacean biomarker records (Figure 8) provide insight into the nature of vegetation changes during and since the LGM. During the LGM, $\delta^{13}\text{C}_{\text{wax}}$ values from all four Wallacean sites were more positive compared to their modern day/Late Holocene values (Figure 8), suggesting rainforest opening and, at some sites, grassland expansion, during the LGM. This trend is in keeping with pollen-based LGM climate reconstructions from the region. Savu Sea pollen records indicate that grasses were abundant during the LGM, reaching up to 50 % of Sumba's vegetation (Dubois et al. 2014), an inference supported by $\delta^{13}\text{C}_{\text{wax}}$ analyses

from the same core. Hope (2001) observed grass pollen concentrations of ~30% in late glacial sediments compared to ~15% in the Holocene at Wanda Site, near Lakes Matano and Towuti, and palynological analyses of cores from the Banda Sea, from eastern Wallacea, also suggest expanded grasslands during the LGM (van der Kaars et al. 2000).

While vegetation can be controlled by numerous environmental and climatic variables, temperature is unlikely to be the primary driver of the vegetation changes observed in Wallacea, as changes in the relative abundance of trees and grasses are not strongly temperature sensitive. Moreover, the longer $\delta^{13}\text{C}_{\text{wax}}$ data from Matano and Towuti do not appear to track changes in Sulu Sea surface temperatures, in particular during MIS 2 and 3 (Russell et al. 2014, Wicaksono et al. 2015). Similarly, although atmospheric CO_2 concentrations can influence the relative abundances of C_3 vs. C_4 grasses (Ehleringer et al. 1997), the expansion of grasses relative to tree taxa observed in $\delta^{13}\text{C}_{\text{wax}}$ and palynological data in Wallacea implies changes in rainfall, rather than CO_2 -driven shifts within grassland communities. These proxy data corroborate our interpretation of high $\delta^{13}\text{C}_{\text{wax}}$ values as indicative of lowered mean annual precipitation and increased dry season water stress during the LGM.

A variety of proxy data supports the use of $\delta^{13}\text{C}_{\text{wax}}$ and other paleovegetation records as an indicator for changes in Wallacean hydroclimate. Sediment cores from Lake Tondano in North Sulawesi indicate lake level lowering during Marine Isotope Stage (MIS) 2/LGM, attributed to reduced precipitation (Dam et al. 2001). Sedimentary Ti concentrations in Lake Towuti were low during the LGM, reflecting lower fluvial erosion and hence less precipitation (Russell et al. 2014), and sedimentological data indicates lower lake levels at Towuti during the LGM, also reflecting drier conditions

(Vogel et al. 2015). The dry LGM climate observed in Wallacea is also coherent with other proxy records across Southeast Asia (Reeves et al. 2013). Cave guano samples measured from Peninsula Malaysia and Palawan Island in the Philippines display enrichment in $\delta^{13}\text{C}$ during the LGM, indicating a dry environment (Wurster et al. 2010). In Java, grasses dominated the local vegetation around Rawa Danau during the LGM, indicating a drier climate (van der Kaars et al. 2001). However, LGM rainforest refugia have been detected in the high mountains of Sulawesi as well as other highland areas within Indonesia. $\delta^{13}\text{C}_{\text{wax}}$ records from a marine core off west Sumatra's high mountain ranges (Niedermeyer et al. 2014), along with pollen records from a high-altitude lake in west Sumatra (Newsome and Flenley 1988) indicate the presence of rainforests and little change in climate during the LGM. These higher elevation records demonstrate the long-standing importance of orographic precipitation in Indonesia amidst large-scale changes in atmospheric regimes, similar to the patterns we observe in Wallacea.

Gradients in the $\delta^{13}\text{C}_{\text{wax}}$ values suggest that the present-day climate gradient of Wallacea, with drier climates to the south, has remained largely in place over the past 26 kyr. Savu Sea $\delta^{13}\text{C}_{\text{wax}}$ remained more positive than records from Sulawesi throughout the last 26 kyr, likely indicating that the Lesser Sunda Islands were always drier than Sulawesi. It is particularly striking that our new $\delta^{13}\text{C}_{\text{wax}}$ record from Mandar Bay exhibits a nearly constant offset from that of the Savu Sea, suggesting that the two regions experienced very similar changes in climate through time. This similarity occurs despite the significantly drier climate during the LGM inferred from these records and the possibility of southward migration of the ITCZ during the LGM (e.g. Chiang et al. 2003, Chiang and Bitz 2005, Lee et al. 2014, Russell et al. 2014, Wicaksono et al. 2015).

In contrast, variations in the amplitudes and absolute values of central Sulawesi $\delta^{13}\text{C}_{\text{wax}}$ data indicate more localized and heterogeneous vegetation changes, though with a similar trend from a dry LGM to a wetter climate in the Holocene reflecting the larger-scale hydrologic changes. Lake Towuti $\delta^{13}\text{C}_{\text{wax}}$ registers the largest change, marked by a 12-permil depletion of $\delta^{13}\text{C}_{\text{wax}}$ from the LGM to the present. This change must have involved major vegetation shifts, from open forests and C_4 -dominated grasslands such as exist in present-day Sumba to the closed-canopy rainforests that dominate the region today. Indeed, the $\delta^{13}\text{C}_{\text{wax}}$ values of LGM sediment at Towuti are similar to the core-top Savu $\delta^{13}\text{C}_{\text{wax}}$, while Holocene $\delta^{13}\text{C}_{\text{wax}}$ exhibits much more negative values similar to those observed in nearby Lake Matano and in regional soils today (Wicaksono et al. 2015). In this light, the vegetation around Lake Towuti has oscillated between end-member vegetation communities in present-day Wallacea. $\delta^{13}\text{C}_{\text{wax}}$ in Lake Matano tracks the changes observed elsewhere, but with more depleted values, suggesting that the higher-altitude catchment of Lake Matano remained relatively wet, even during the LGM (Wicaksono et al. 2015), likely due to sustained orographically-induced precipitation at higher elevations. At Mandar Bay, $\delta^{13}\text{C}_{\text{wax}}$ data indicate much more modest changes than observed in the lake records, with open forest and grassland ecosystems similar to present-day Sumba at the LGM transitioning to a mix of closed-canopy forests, tropical deciduous forests, and grasslands in the Holocene, similar to the vegetation mosaic inhabiting southwest Sulawesi today.

Comparison of $\delta^{13}\text{C}_{\text{wax}}$ records from Wallacea may also provide insight into the application of the leaf wax proxy itself. Although these records are not from identical regions, the marine $\delta^{13}\text{C}_{\text{wax}}$ records from Savu Sea and Mandar Bay exhibit far fewer

millennial- and centennial-scale variations compared to the lacustrine records (Figure 7), have different mean values, and have lower-amplitude glacial-interglacial signals. While these differences could reflect real differences in regional vegetation, they may also reflect the different sizes of leaf wax source areas and transport mechanisms between offshore marine and inland lake records. Both the Savu Sea and Mandar Bay records have much larger leaf wax source areas compared to Lakes Matano and Towuti. For instance, the catchment of Saddang River, associated with the Mandar Bay records, has an area of 6380 km², compared to the smaller catchment areas for Lakes Matano and Towuti (287 and 983 km², Wicaksono et al. 2015). Palynological studies have shown that larger catchments often incorporate different vegetation types, resulting in more averaging and smaller amplitude changes (i.e. a “smoothing effect”) than in smaller catchments (Sugita 2007). Moreover, although most leaf waxes found in Mandar Bay cores likely originate from the catchment, some waxes might also come from northerly far away sources, brought into the site by oceanic currents, in particular the strong currents of the Indonesian Throughflow (Gordon et al. 2003). This long distance oceanic transport is likely to carry waxes more enriched in ¹³C given the drier climate of the regions north of Wallacea, possibly contributing to the relatively enriched $\delta^{13}\text{C}_{\text{wax}}$ values in the Savu Sea and Mandar Bay samples. Our data thus suggest that $\delta^{13}\text{C}_{\text{wax}}$ from marine records reflect a spatially broader view of environmental changes compared to lacustrine records, at least in this region. Observations on marine and terrestrial $\delta\text{D}_{\text{wax}}$ records, as discussed in the following section, suggest similar patterns.

6.2 Climatic controls on Wallacean rainfall isotopes since the LGM

Our new record of δD_{precip} from Mandar Bay builds upon and extends a reconstruction of δD_{wax} from a sediment core collected close to our study site from Makassar Strait (Tierney et al. 2012). The two records, as well as δD_{wax} measurements of different *n*-acids from our Mandar Bay core, exhibit generally coherent trends (Figure 9). Tierney et al. (2012) do not report δD_{wax} values for C_{28} *n*-acid, but the Makassar Strait C_{30} *n*-acid δD_{wax} are very similar to Mandar Bay C_{30} *n*-acid δD_{wax} , and both records exhibit similar D-depletion from the late glacial to the Holocene (Figure 9). There are, however, differences in the two records. Following the shift to more enriched values during the last deglaciation, δD_{wax} becomes more negative in our core at ~ 13 kyr BP. This depletion in Mandar is slightly later than in the Makassar Strait record, which displays D-depletion since ~ 14 kyr BP. During the Holocene, our records indicate a minimum in δD_{wax} at ~ 5 kyr BP, while the Makassar Strait δD_{wax} is relatively stable through the Holocene. There could be many reasons for the differences between the Mandar Bay and Makassar Strait datasets. Limitations in dating may have caused some of the differences between the two records. For example, although the Makassar Strait age model is generally well constrained, there is a relatively long interval between two ^{14}C ages of 11.223 and 13.990 yr BP (Tierney et al. 2012), during which time the two records display different trends. It is also possible that, despite the proximity of the two records, they experience subtle differences in leaf wax source areas, though this hypothesis is very difficult to evaluate. Lastly, we note that many of the differences in the records approach the analytical

uncertainty of the δD_{wax} measurements of ~ 2 ‰. While we cannot fully explain these differences, the two records agree in the D-depletion observed during the late deglacial, and our δD_{wax} record from Mandar Bay shows strong similarities to isotopic records from other sites in Wallacea.

Indeed, long-term changes in δD_{precip} data from Mandar Bay, Lake Towuti, and Flores Cave show striking similarities (Figure 10). During the LGM, these data sets display more negative reconstructed δD_{precip} . This D-depletion is followed by D-enrichment during the deglaciation, and then a trend toward D-depletion beginning ~ 11 kyr BP. There are variations among the reconstructed δD_{precip} data in terms of absolute values, but present-day reconstructed values generally match the modern δD_{precip} gradients in the region (Figure 7). Both measured and reconstructed present-day δD_{precip} reveal more negative values in central Sulawesi compared to northern Borneo and Bali (Figure 7). Though inaccuracies in the conversion of leaf wax- and cave-based data sets to δD_{precip} could contribute to the differences in the reconstructed absolute values, the reasonable fit between reconstructed and observed δD_{precip} gradients therefore likely reflect real regional variations in the factors controlling precipitation isotopes, including rainfall amount (i.e. the ‘amount effect’), moisture sources, and trajectories.

The coherent long-term isotopic evolution observed in Towuti, Mandar Bay, and Flores suggests that the sites and region experienced similar atmospheric circulation changes at the orbital scale since 26 kyr BP. In this context, the depletion of δD_{precip} in these records during the LGM is challenging to explain in light of the widespread drying observed in the region. The ‘amount effect’ is often assumed to be the primary control of past changes in δD_{precip} and $\delta^{18}\text{O}_{\text{precip}}$ in the tropics. The amount effect, i.e. the negative

correlation between the isotopic composition of precipitation and precipitation rates, was first observed by Dansgaard (1964) and later confirmed in subsequent studies (e.g. Rozanski et al. 1992 and Bowen 2008). The removal of isotopically-enriched water vapor as condensation proceeds in a strong convection system leaves the isotopic content of the remaining vapor increasingly lighter and the rainwater more depleted (Vuille et al. 2003). Further, the reduced evaporation of falling raindrops in a wetter atmospheric column and near active convection may reduce re-enrichment of the rainwater (Dansgaard 1964, Lawrence et al. 2004). While these physical processes are relatively well-understood, applying the amount effect to the interpretation of negative δD_{precip} data observed in Wallacea would indicate a regionally wet climate during the LGM (e.g. Griffiths et al. 2009, Ayliffe et al. 2013), in contrast to virtually all other paleoenvironmental proxy data from the region.

What shifts in the atmospheric regime could have caused this LGM isotopic depletion? The northwest trade winds are known to carry D-depleted air masses from the Indian and Pacific Oceans relative to air masses originated locally or from the south (Kurita et al. 2009, Griffiths et al. 2009, Moerman et al. 2013). Indeed, the effects of monsoon circulation and seasonal ITCZ movement on rainfall isotopes are visible in central Sulawesi in the present. Data from Timampu suggest “extra depletion” of δD_{precip} when the northwest Austral-Asian Monsoon dominates and transports rain-bearing air masses from the northern Pacific in November-January (Figure 6, Konecky et al. 2015). Meanwhile, convective and stratiform rainfall within the ITCZ, which characterizes most rainfall in central Sulawesi in March-May, appears to have “limited depletion” compared to the trade wind-related precipitation (Figure 6), likely due to shorter transport paths at

this time. The “extra depletion” in central Sulawesi might be predominantly a function of moisture source and rainout along the long path taken by the trade wind-steered air mass, highlighting the importance of advection and “upstream” atmospheric processes in controlling δD_{precip} along long transport paths (e.g. Brown et al. 2008, Moerman et al. 2013, Konecky et al. 2015). More negative δD_{precip} values are also observed between December and March in Denpasar, the closest monitoring station to Flores, when the northwest trade winds are strong and the ITCZ passes over the region (Griffiths et al. 2009, Figure 7).

Thus, a stronger northwest monsoon could have brought more depleted precipitation during the LGM to Wallacea, as registered by depleted δD_{precip} observed at Mandar Bay, Lake Towuti, and Flores. In contrast, reconstructed δD_{precip} from northern Borneo (Partin et al. 2007) indicates LGM precipitation was isotopically enriched relative to the Holocene. The D-enrichment in Borneo is consistent with decreased regional convective activity and proxy evidence for drier conditions during the LGM (Wurster et al. 2010). In the present, precipitation amounts and isotopic composition at Mulu, northern Borneo, exhibit relatively weak seasonality, with little D-depletion during boreal winter when the northwest monsoon strengthens over southern and central Indonesia (Moerman et al. 2013, Figure 7). The strong similarity of the reconstructions over Wallacea, where southward ITCZ migration and the northwest monsoon play a critical role during the modern-day wet season, relative to Borneo, where these processes are less important, strongly implicate ITCZ migration and monsoon intensification in the D-depletion observed across Wallacea during the LGM (Ayliffe et al. 2013).

Global climate boundary conditions during the LGM might have perturbed Wallacean atmospheric circulation, the monsoons, and the ITCZ via multiple pathways. For example, expanded ice sheets and northern hemisphere cooling is thought to have strengthened the East Asian Winter Monsoon (EAWM; Porter and An 1995, Shi et al. 2011). Given the modern-day correlation between the EAWM and the Austral-Asian northwest monsoon (Suppiah and Wu 1998, Hong and Li 2009), the strengthening of the EAWM under glacial boundary conditions could have intensified the Austral-Asian northwest monsoon. Modeling studies further indicate the large volume and high topography of the Laurentide ice sheets drove a net southward shift of the mean ITCZ position (e.g. Chiang et al. 2003, Chiang and Bitz 2005, Lee et al. 2014), consistent with a strengthened northwest monsoon. Southward ITCZ migration and strengthened monsoon circulation, however, did not lead to increased rainfall in Wallacea, as observed from our vegetation reconstructions and other proxy data (e.g. Reeves et al. 2013).

A southward shift in the ITCZ, a strengthened northwest monsoon, and D-depleted rainfall may also be caused by orbital precession and seasonal insolation forcing. Indeed, Wallacean δD_{precip} records appear to be correlated to southern hemisphere summer insolation throughout the past 26 kyr BP (Figure 10), suggesting a dynamical linkage between precession, ITCZ displacement, and monsoon strength (e.g. Holbourn et al. 2005, Tachikawa et al. 2011, Kuhnt et al. 2015). Strong southern hemisphere summer insolation during the LGM might have pulled the caloric equator toward the southern tropics, shifting the seasonal and mean annual ITCZ positions southward. Moreover, weak southern hemisphere summer insolation ca. 12 kyr BP is synchronous with the maximum D-enrichment observed in the Wallacean δD_{precip}

records, including Mandar Bay, Flores, and Towuti, when the ITCZ shifted to the north (Wang et al. 2001, Wang et al. 2008). As the ITCZ shifts northward, most of Wallacea was subject to increased influence of the isotopically-enriched southeasterlies, therefore increasing the overall δD_{precip} values. The strengthening of the northwest monsoon as the ITCZ shifted to the south during the Holocene is also recorded in the marine record from Timor Sea, to the south of the Lesser Sunda Islands (Kuhnt et al. 2015). Despite the correlation between Wallacean δD_{precip} and southern hemisphere summer insolation, the D-depletion we observe at Mandar Bay and other Wallacean sites during the LGM relative to the Holocene cannot be explained solely by insolation forcing, implicating glacial boundary conditions in forcing the inferred changes in monsoon strength and ITCZ position.

Our records may further explain the limited glacial-interglacial changes observed in planktic foraminifera-derived $\delta^{18}\text{O}$ of seawater ($\delta^{18}\text{O}_{\text{sw}}$) proxy records from offshore western Sumatra (Mohtadi et al. 2014), off south Java (Mohtadi et al. 2011), and the Savu Basin (Gibbons et al. 2014). The $\delta^{18}\text{O}_{\text{sw}}$ has been interpreted to primarily reflect mean annual local sea surface salinity, likely controlled by shifts in regional precipitation/evaporation and river runoff from nearby islands, and to reflect some influence from oceanic advection and mixing of water with different isotopic compositions. In this context, the authors explained the small LGM-present $\delta^{18}\text{O}_{\text{sw}}$ as indicative of relative similarities in precipitation amount during the LGM and present. This interpretation, however, is in contrast to the substantial drying recorded by regional paleoenvironmental records, as discussed above. Instead, it is possible that the isotope

depletion we observe offset changes in precipitation amount and runoff in regional $\delta^{18}\text{O}_{\text{sw}}$ records.

6.3 Time-varying aspect of precipitation- $\delta\text{D}_{\text{precip}}$ relationship in Wallacea

A prominent feature of our Mandar Bay data is a switch in the relationship between $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ (Figure 4). $\delta\text{D}_{\text{precip}}$ is negatively correlated with $\delta^{13}\text{C}_{\text{wax}}$ during the LGM and the last deglaciation, but positively correlated during the Holocene. Assuming $\delta^{13}\text{C}_{\text{wax}}$ records changes in precipitation amount, this switch could suggest that seasonal changes in moisture source and atmospheric circulation dominate the $\delta\text{D}_{\text{precip}}$ signal when glacial boundary conditions prevailed, while $\delta\text{D}_{\text{precip}}$ records primarily changes in precipitation amount during the Holocene. Therefore, this switch likely signifies a major shift in atmospheric circulation near the beginning of the Holocene.

By calculating a moving correlation window, i.e. searching for the time of the switch that optimizes R-squared values for both the correlation and anti-correlation, we found that the highest R-squared values are found between 11.1 and 9.4 kyr BP (Figure 11), suggesting that rising sea level and flooding of the regional shelves exert an important control on atmospheric circulation and precipitation isotopes (e.g. Griffiths et al., 2011). At 11.1 kyr BP, the rapid inundation of the Sunda Shelf, associated with the Meltwater Pulse 1B, had just been completed (Fairbanks 1989), resulting in the initial opening of the South China Sea and the Java Sea (Sathiamurthy and Voris 2006, Hanebuth et al. 2009). Between 11.5 and 9.5 kyr BP, sea level rose from ~50 to ~20 meter below present level and flooded the Karimata Strait and, more importantly, the rest

of the Java Sea (Sathiamurthy and Voris 2006, Hanebuth et al. 2009), thereby creating a large moisture source region within the Indonesian archipelago during this time window (Figure 11). Thus, by 9.4 kyr BP, the channels and straits separating Java, Kalimantan, Sumatra, and the Malay Peninsular were already becoming significant water bodies (Hanebuth et al. 2009).

While uncertainty remains regarding the effects of Sunda Shelf exposure on regional atmospheric circulation (DiNezio and Tierney 2012), the presence of large, warm seas in western Indonesia appears to have increased the relative influence of the amount effect during the Holocene, likely through major reductions in the length of moisture transport during the wetter months and at the time of southward ITCZ transit. Interestingly, the latter part of the time window of the switch is coincident with an abrupt, non-linear switch in the monsoon system above Flores at ~9.5 kyr BP observed in the correlations between $\delta^{18}\text{O}_{\text{speleothem}}$, Mg/Ca, and Sr/Ca data from Flores cave and attributed to the rapid inundation of the Sunda shelf at that time (Griffiths et al. 2011). Our data, however, suggest that the switch in the atmospheric circulation or the monsoon system occurred gradually, in line with the relatively gradual inundation of the Sunda Shelf. In particular, the flooding between 11.1 and 9.4 kyr BP likely represent a key interval for the establishment of the modern-day relationship between precipitation and δD_{precip} in the region.

During the Holocene, following the flooding of the Java Sea, records from Wallacea exhibit much more heterogeneity. Borneo exhibits the most depleted δD_{precip} at ~4.5 kyr BP, Towuti shows more depleted δD_{precip} at ~3.5 kyr BP, and Flores δD_{precip} is relatively stable (Figure 10). Our record from Mandar Bay shows more depleted δD_{precip}

at ~4.5 kyr BP, similar to Borneo, though this finding contrasts with nearby δD_{precip} records from western Sulawesi (Tierney et al. 2012); thus, the trends in our Mandar Bay data must be evaluated with caution. The heterogeneity in these records, in the absence of strong glacial forcings, could reflect different sensitivities among the sites to insolation-forced changes in the ITCZ and the monsoons (Partin et al. 2007, Griffiths et al. 2009) and/or to coupled ocean-atmosphere processes operating in the Pacific and Indian Oceans during the Holocene. The variations in Borneo rainfall isotopes during the Holocene have been linked to insolation during boreal fall, which controls the intensity of precipitation as the ITCZ passes above the site in the fall (Partin et al. 2007, Carolin et al. 2013). As the mean position of the ITCZ shifts southwards over the course of the Holocene in response to fall precessional forcing (Carolin et al. 2013), the peak of intense precipitation and the isotopic minima also shift southward across the region (Partin et al. 2007). Combined with the depletion observed at Towuti ca. 3.5 kyr BP, these records may jointly indicate the crossing of the ITCZ above the northern part of Indonesian archipelago during mid-Holocene. The absence of this ITCZ-forced isotopic depletion above Flores likely indicates that precipitation over the Lesser Sunda Islands was less influenced by the shift in the ITCZ during the Holocene, possibly because of other dominant controlling factors such as sea surface temperatures and Sunda shelf exposure.

7. Conclusions

$\delta^{13}C_{\text{wax}}$ measured from the Mandar Bay core indicate that regional vegetation is highly sensitive to hydroclimatic changes, and that the largely meridional present-day

climate gradient of Wallacea has existed since at least 26 kyr BP. Wallacean paleovegetation records document grassland expansion into many currently forested areas in the region, implying a drier climate and increased seasonality during the LGM. Wallacean δD_{precip} records have similar trends during the past 26 kyr BP, with significant isotopic depletion during a dry LGM implying a different rainfall isotope dynamic compared to today. Our Mandar Bay records demonstrate that δD_{precip} was anti-correlated with $\delta^{13}C_{\text{wax}}$ during the LGM and the last deglaciation, but was positively correlated during most of the Holocene. Based upon the similarity of the isotopic depletion in records from equatorial and southern Wallacea, where the Austral-Asian northwest monsoon dominates present-day precipitation patterns, we argue that the LGM depletion was caused by a southward shift of the ITCZ mean position and an intensified, yet drier northwest Austral-Asian monsoon. This circulation pattern, in turn, was likely forced by stronger southern hemisphere summer insolation and the presence of large northern hemisphere ice sheets. At Mandar Bay, $\delta^{13}C_{\text{wax}}$ and δD_{precip} become positively correlated during most of the Holocene. The switch in correlation likely occurred gradually between 11.1 and 9.4 kyr, when the Java Sea and Karimata Strait were flooded by the rising sea level, suggesting that these new moisture sources and/or changes in moisture transport pathways increased the influence of the amount effect in the region.

To further understand the LGM depletion of rainfall isotopes and better characterize the extent and impact of ITCZ movements and seasonal variations on Wallacean δD_{precip} across multiple timescales, we will need to couple a larger spatial coverage of precipitation isotope records from the region with a systematic study of high-resolution isotope-enabled climate model simulations under different scenarios of

forcings and boundary conditions. Importantly, the time-dependent and varying controls on the isotopic composition of rainfall suggest more caution is needed when interpreting long-term rainfall isotope proxy data in this climatically complex region.

References

- Aldrian, E. and R. D. Susanto (2003), Identification of three dominant rainfall regions within Indonesia and their relationship to sea surface temperature. *Int. Jour. of Clim.*, 23, 1435-1452.
- Ayliffe, L., M. Gagan, J-x Zhao, R.N. Drysdale, J.C. Hellstrom, W.S. Hantoro, M.L. Griffiths, H. Scott-Gagan, E St Pierre, J.A. Cowley, and B.W. Suwargadi (2013), Rapid interhemispheric climate links via the Australasian monsoon during the last deglaciation. *Nature Communications*, 4, 1-6.
- Bowen, G. J. (2008), Spatial analysis of the intra-annual variation of precipitation isotope ratios and its climatological corollaries. *J. Geophys. Res.*, 113, D05113, doi:10.1029/2007JD009295.
- Brown, D., J. Worden, and D. Noone (2008), Comparison of atmospheric hydrology over convective continental regions using water vapor isotope measurements from space, *J. Geophys. Res.*, 113, D15124, doi:10.1029/2007JD009676.
- Carolin, S., K.M. Cobb, J.F. Adkins, B. Clark, and A. Tuen (2013), Varied response of western tropical Pacific hydrology to abrupt climate changes. *Science*, doi:10.1126/science.1233797.

- Chiang, J.C.H and C.M. Bitz (2005), The influence of high latitude ice on the position of the marine Intertropical Convergence Zone. *Climate Dynamics*, DOI 10.1007/s00382-005-0040-5.
- Chiang, J.C.H., M. Biasutti, and D.S. Battisti (2003), Sensitivity of the Atlantic Intertropical Convergence Zone to Last Glacial Maximum boundary conditions. *Paleoceanography*, 18 (4), 1094, <http://dx.doi.org/10.1029/2003PA000916>
- Chikaraishi, Y. and H. Naraoka (2007), $\delta^{13}\text{C}$ and δD relationships among three n-alkyl compound classes (n-alkanoic acid, n-alkane and n-alkanol) of terrestrial higher plants. *Org. Geochem.* 38, 198-215.
- Cobb, K.M., J.F. Adkins, J.W. Partin, and B. Clark (2007), Regional-scale climate influences on temporal variations of rainwater and cave dripwater oxygen isotopes in northern Borneo. *Earth Planet. Sci. Lett.* 263, 207–220.
- Craig, H. (1961), Isotopic variations in meteoric waters. *Science*, 133, 1702–1703.
- Dam, R., J. Fluin, P. Suparan, and S. van der Kaars (2001), Palaeoenvironmental developments in the Lake Tondano area (N. Sulawesi, Indonesia) since 33,000 yr BP. *Palaeogeography, Palaeoclimatology, Palaeoecology* 171, 147–183.
- Dansgaard, W. (1964), Stable isotopes in precipitation. *Tellus*, 16(4), 436–468, doi:10.1111/j.2153-3490.1964.tb00181.x.
- Diefendorf, A.F., K.E. Mueller, S.L. Wing, P.L. Koch, and K.H. Freeman (2010), Global patterns in leaf ^{13}C discrimination and implications for studies of past and future climate. *Proc Natl Acad Sci USA* 107(13), 5738–5743.
- DiNezio, P.N. and J.E. Tierney (2013), The effect of sea level on glacial Indo-Pacific climate *Nat. Geosci.*, 6, 485–491, <http://dx.doi.org/10.1038/NGEO1823>.

- Dubois N., D.W. Oppo, V. Galy, M. Mohtadi, S. van der Kaars, J. Tierney, Y. Rosenthal, T. Eglinton, A. Lückge, and B. Linsley (2014), Indonesian vegetation response to changes in rainfall seasonality over the past 25,000 years. *Nature Geoscience* 7, 513-517.
- Ehleringer, J. R., T.E. Cerling, and B.R. Helliker (1997), C₄ photosynthesis, atmospheric CO₂, and climate. *Oecologia* 112, 285–299.
- Fairchild, I.J. and A. Baker (2012), *Speleothem Science. From Process to Past Environment*. Wiley-Blackwell.
- Gibbons F., D.W. Oppo, M. Mohtadi, Y. Rosenthal, J. Cheng, Z. Liu, and B.K. Linsley (2014), Deglacial $\delta^{18}\text{O}$ and hydrologic variability in the tropical Pacific and Indian Oceans. *Earth and Planetary Science Letters* 387, 240-251.
- Griffiths M. L., R. N. Drysdale, M. K. Gagan et al. (2012), Abrupt increase in east Indonesian rainfall from flooding of the Sunda Shelf ~9500 years ago, *Quaternary Science Reviews*, 74, 273-279. DOI: 10.1016/j.quascirev.2012.07.006
- Griffiths M. L., R. N. Drysdale, M. K. Gagan, J.-X. Zhao, L. K. Ayliffe, J. C. Hellstrom, W. S. Hantoro, S. Frisial, Y. -X. Feng, I. Cartwright, E. St. Pierre, M. J. Fischer, and B. W. Suwargadi (2009), Increasing Australian-Indonesian monsoon rainfall linked to early Holocene sea-level rise, *Nature Geoscience*, 2, 636-639.
- Griffiths, M. L., et al. (2010), Evidence for Holocene changes in Australian-Indonesian monsoon rainfall from stalagmite trace element and stable isotope ratios, *Earth Planet Sci. Lett.*, 292(1–2), 27–38.
- Hanebuth, T.J.J., K. Stattegger, and A. Bojanows-ki (2009), Termination of the Last Glacial Maximum sea-level lowstand: the Sunda sea- level record revisited. *In*:

- Camoin, G., Droz-ler, A., Miller, K., and Fulthorpe, C. (Eds.), Records of Quaternary sea-level changes: *Global and Planetary Change*, 66, p.76-84.
- Holbourn, A., W. Kuhnt, H. Kawamura, Z. Jian, P. Grootes, H. Erlenkeuser and J. Xu (2005), Orbitally paced paleoproductivity variations in Timor sea and Indonesian throughflow variability during 460 kyr. *Paleoceanography*, 20, PA 3002.
- Hong, C.-C. and T Li (2009), The extreme cold anomaly over southeast Asia in February 2008: roles of ISO and ENSO, *Journal of Climate* 22, 3787e3801.
- Hope, G. S. (2001), Environmental change in the Late Pleistocene and later Holocene at Wanda Site, Soroako, South Sulawesi, Indonesia, *Palaeogeogr. Palaeoclim. Palaeocl.*, 171, 129145.
- Kershaw, A.P., S. van der Kaars. and J.F. Flenley (2011), The Quaternary history of far eastern rainforests (2nd Edn.). In: Bush, M.B., Flenley, J.F. and Gosling, W.D. (Eds.) *Tropical Rainforest Responses to Climate Change*. Springer-Verlag, Berlin Heidelberg, pp. 85-123.
- Knutti, R. and J. Sedlacek (2013), Robustness and uncertainties in the new CMIP5 coordinated climate model projections, *Nature Climate Change*, 3, 369-373, doi:10.1038/nclimate1716.
- Konecky, B.L., Russell, J.M., Bijaksana, S. (2015), Glacial aridity in Indonesia coeval with intensified monsoon circulation, *Earth and Planetary Science Letters*, 437, 15-24, doi: 10.1016/j.epsl.2015.12.037.
- Kuhnt, W., A. Holbourn, J. Xu, B. Opdyke, P. De Decker, U. Röhl, M. Mudelsee (2015), Southern Hemisphere control on Australian monsoon variability during the late deglaciation and Holocene, *Nat Comm*, 6, 5916.

- Kurita, N., D. Noone, C. Risi, G.A. Schmidt, H. Yamada, and K. Yoneyama (2011), Intraseasonal isotopic variation associated with the Madden-Julian Oscillation. *J. Geophys. Res.*, 116, D24101
- Kurita, N., K. Ichiyanagi, J. Matsumoto, M. D. Yamanaka, and T. Ohata (2009), The relationship between the isotopic content of precipitation and the precipitation amount in tropical regions, *J. Geochem. Exp.*, 102, 113–122.
- Laskar, J., P. Robutel, F. Joutel, M. Gastineau, A.C.M. Correia, and B. Levrard (2004), A long-term numerical solution for the insolation quantities of the Earth. *Astron. Astrophys.*, 428, 261–285.
- Lawrence, J.R., S.D. Gedzelman, D. Dexheimer, H.K. Cho, G.D. Carrie, R. Gasparini, C.R. Anderson, K.P. Bowman, and M.I. Biggerstaff (2004), Stable isotopic composition of water vapor in the tropics. *J. Geophys. Res. Atmos.* 109, D06115.
- Lee, S.-Y., J.C.H. Chiang, and P. Chang (2014), Tropical Pacific Response to Continental Ice Sheet Topography, *Climate Dynamics*, DOI: 10.1007/s00382-014-2162-0.
- LeGrande, A.N. and G.A. Schmidt (2009), Sources of Holocene variability of oxygen isotopes in paleoclimate archives, *Clim Past*, 5, 441-455.
- Levi, C., L. Labeyrie, F. Bassinot, F. Guichard, E. Cortijo, C. Waelbroeck, N. Caillon, J. Duprat, T. de Garidel-Thoron, and H. Elderfield (2007), Low-latitude hydrological cycle and rapid climate changes during the last deglaciation, *Geochem. Geophys. Geosyst.*, 8 (5), p. Q05N12, <http://dx.doi.org/10.1029/2006GC001514>.

- Lewis, S. C., A. N. LeGrande, M. Kelley, and G. A. Schmidt (2010), Water vapour source impacts on oxygen isotope variability in tropical precipitation during Heinrich events, *Climate of the Past*, 6, 325–343.
- Moerman, J.W., K.M. Cobb, J.F. Adkins, H. Sodemann, B. Clark, and A. Tuen (2013), Diurnal to interannual rainfall oxygen isotope variations in northern Borneo driven by regional hydrology. *Earth and Planetary Sciences Letter*, 369, 108-119.
- Mohtadi, M., D. W. Oppo, S. Steinke, J. W. Stuut, E. C. Hathorne, J. Groeneveld, D. Hebbeln, and A. Lückge (2011), Glacial to Holocene swings of the Australian-Indonesian monsoon, *Nature Geoscience*, 4, 540-544.
- Mohtadi, M., M. Prange, D.W. Oppo, R. De Pol-Holz, U. Merkel, X. Zhang, S. Steinke, and A. Lückge (2014), North Atlantic forcing of tropical Indian Ocean climate. *Nature* 509, 76-80.
- Monk, K. A., Y. de Fretes, and G. Reksodiharjo-Lilley (1997), *The ecology of Nusa Tenggara and Maluku*. The Ecology of Indonesia Series, Vol. 5. Hong Kong: Periplus.
- Muller, J.F. McManus, D.W. Oppo, and R. Francois (2012), Strengthening of the northeast monsoon over the Flores Sea, Indonesia, at the time of Heinrich event 1. *Geology*, 40, 635–638.
- Myers, N., R.A. Mittermeier, C.G. Mittermeier, G.A.B. da Fonseca, J. Kent (2000), Biodiversity hotspots for conservation priorities. *Nature* 403, 853–858.
- Naylor, R., D. Battisti, D. Vimont, W. Falcon, and M. Burke (2007), Assessing risks of climate variability and climate change for Indonesian rice agriculture.

- Proceedings of the National Academy of Sciences of the United States of America* 104: 7752–7757.
- Newsome, J. and J. R. Flenley (1988), Late Quaternary vegetation history of the Central Highlands of Sumatra. II.: Palaeopalynology and Vegetational History. *J. Biogeog.*, 15, 555-578.
- Niedermeyer E.M., A.L. Sessions, S.J. Feakins, and M. Mohtadi (2014), Hydroclimate of the western Indo-Pacific Warm Pool during the past 24,000 years. *Proceedings of the National Academy of Science* 111, 9402-9406.
- Oktaviani, R., S. Amaliah, C. Ringler, M. Rosegrant, and T. Sulser (2011), The impact of global climate change on the Indonesian economy, <http://www.ifpri.org/sites/default/files/publications/ifpridp01148.pdf>
- Partin J. W., K. M. Cobb, J. F. Adkins, B. Clark, and D. P. Fernandez (2007), Millennial-scale trends in west Pacific warm pool hydrology since the Last Glacial Maximum, *Nature*, 449, 452-455.
- Peel, M. C., B.L. Finlayson, and T.A. McMahon (2007), Updated world map of the Köppen– Geiger climate classification, *Hydrol. Earth Syst. Sci.*, 11, 1633–1644.
- Porter, S. C. and An, Z. S. (1995), Correlation between climate events in the North Atlantic and China during the last glaciation, *Nature*, 375, 305–308.
- Reeves, J. M., H. C. Bostock, L. K. Ayliffe, T. T. Barrows, P. De Deckker, L. S. Devriendt, G. B. Dunbar, R. N. Drysdale, K. E. Fitzsimmons, M. K. Gagan, M. L. Griffiths, S. G. Haberle, J. D. Jansen, C. Krause, S. Lewis, H. V. McGregor, S. D. Mooney, P. Moss, G. C. Nanson, A. Purcell, S. van der Kaars (2013), Palaeoenvironmental change in tropical Australasia over the last 30,000 years – a

- synthesis by the OZ-INTIMATE group. *Quaternary Science Reviews*,
<http://dx.doi.org/10.1016/j.quascirev.2012.11.027>
- Reimer, P. J., E. Bard, A. Bayliss, et al. (2013), IntCal13 and Marine13 Radiocarbon Age Calibration Curves 0-50,000 Years cal BP. *Radiocarbon*, 55(4).
- Rozanski, K., L. Araguas-Araguas and R. Gonfiantini (1992), Relation between long-term trends of ^{18}O isotope composition of precipitation and climate, *Science*, 258, 981-985.
- Russell, J.M., H. Vogel, B.L. Konecky, S. Bijaksana, Y. Huang, M. Melles, N. Wattrus, K. Costa, and J.W. King (2014), Glacial forcing of central Indonesian hydroclimate since 60,000 y B.P., *Proc Nat Ac Sci USA*,
doi:10.1073/pnas.1402373111.
- Sachse, D, I. Billault, G.J. Bowen, Y. Chikaraishi, K.H. Freeman, C.R. Magill, F.A. McInerney, M.T.J. van der Meer, P. Polissar, R.J. Ribons, J.P. Sachs, H.-L. Schmidt, A.L. Sessions, J.W. White, J.B. West, and A. Kahmen (2012), Molecular paleohydrology: interpreting the hydrogen-isotopic composition of lipid biomarkers from photosynthesizing organisms. *Annu. Rev. Earth Planet. Sci.*, 40, 221–249.
- Sathiamurthy, E. and H. K. Voris (2006), Maps of Holocene sea level transgression and submerged lakes on the Sunda Shelf. Nat Hist J Chulalongkorn University, Supplement 2:1–43. Maps available at
http://fmnh.org/research_collections/zoology/zoo_sites/seamaps/.
- Schefuß E., Schouten S. and Schneider R. R. (2005), Climatic controls on central African hydrology during the past 20,000 years. *Nature* 437, 1003–1006.

- Shi, Z., X. Liu, Y. Sun, Z. An, Z. Liu, and J. Kutzbach (2011), Distinct responses of East Asian summer and winter monsoons to astronomical forcing, *Clim. Past*, 7, 1363–1370, doi:10.5194/cp-7-1363-2011.
- Stott, L., A. Timmermann, and R. Thunell (2007), Southern Hemisphere and deep-seawarming led deglacial atmospheric CO₂ rise and tropical warming. *Science* 318(5849), 435–438.
- Sugita, S. (2007), Theory of quantitative reconstruction of vegetation I: pollen from large sites REVEALS regional vegetation. *The Holocene* 17, 229–241.
- Suppiah, R., and X. Wu (1998), Surges, cross-equatorial flows and their links with the Australian summer monsoon circulation and rainfall, *Aust. Meteorol. Mag.*, 47, 113–130.
- Tachikawa, K., A. Timmermann, L. Vidal, C. Sonzogni, and O.E. Timm (2013) CO₂ radiative forcing and Intertropical Convergence Zone influences on western Pacific warm pool climate over the past 400 ka. *Quaternary Science Reviews*, 86, 24–34, DOI:10.1016/j.quascirev.2013.12.018.
- Tachikawa, K., O. Cartapanis, L. Vidal, L. Beaufort, T. Barlyaeva, and E. Bard (2011), The precession phase of hydrological variability in the Western Pacific Warm Pool during the past 400 ka, *Quaternary Sci. Rev.*, 30, 3716–3727.
- Tierney, J.E., D.W. Oppo, A.N. LeGrande, Y. Huang, Y. Rosenthal, and B.K. Linsley (2012), The influence of Indian Ocean atmospheric circulation on warm pool hydroclimate during the Holocene epoch. *J. Geophys. Res.*, 117, D19108, doi:10.1029/2012JD018060.

- Underwood, E.C., D. Olson, A. D. Hollander, and J. F. Quinn (2014) Ever-wet tropical forests as biodiversity refuges questions of bias in climate models, *Nature Climate Change*, 4, 9, 740–741.
- van der Kaars, S., D. Penny, J. Tibby, J. Fluin, R. Dam, and P. Suparan (2001) Late Quaternary palaeoecology, palynology and palaeolimnology of a tropical lowland swamp: Rawa Danau, West Java, Indonesia. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 171, 185–212.
- van der Kaars, S., X. Wang, A.P. Kershaw, F. Guichard, and D.A. Setiabudi (2000) A Late Quaternary palaeoecological record from the Banda Sea, Indonesia: Patterns of vegetation, climate and biomass burning in Indonesia and northern Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 155, 135–153.
- Vuille, M., R. S. Bradley, M. Werner, R. Healy, and F. Keimig (2003), Modeling $\delta^{18}\text{O}$ in precipitation over the tropical Americas: 1. Interannual variability and climatic controls, *J. Geophys. Res.*, 108, 4174, doi:10.1029/2001JD002038.
- Waelbroeck, C., L. Labeyrie, E. Michel, J.C. Duplessy, J. McManus, K. Lambeck, E. Balbon, and M. Labracherie (2002), Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records. *Quaternary Science Reviews*, 21, 295-305.
- Wang, Y.J., H. Cheng, R.L. Edwards, X.G. Kong, X.H. Shao, S.T. Chen, J.Y. Wu, X.Y. Jiang, X.F. Wang, and Z.S. An (2008), Millennial- and orbital-scale changes in the East Asian monsoon over the past 224,000 years, *Nature*, 451, 1090-1093.

- Wang, Y.J., H. Cheng, R.L. Edwards, Z.S. An, J.Y. Wu, C.C. Shen, and J.A. Dorale (2001), A high-resolution absolute-dated Late Pleistocene monsoon record from Hulu Cave, China, *Science*, 294, 2345-2348.
- Whitten, T., M. Mustafa and G.S. Henderson (1987), *The Ecology of Sulawesi*. Gadjah Mada University Press, Yogyakarta, Indonesia. Republished by Periplus, Hong Kong, 2002.
- Wicaksono, S.A., J.M. Russell, J. M., and S. Bijaksana (2015), Compound-specific stable isotope records of vegetation and hydrologic change in central Sulawesi since 53,000 yr BP. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 430, 47-56, DOI:10.1016/j.palaeo.2015.04.016
- Worden, J., D. Noone, K. Bowman, and T. E. Spect (2007), Importance of rain evaporation and continental convection in the tropical water cycle, *Nature*, 445(7127), 528–532, doi:10.1038/nature05508.
- Wurster, C.M. and M.I. Bird (2014) Barriers and bridges: early human dispersals in equatorial SE Asia. In: Harff, J., Bailey, G., and Lüth, F., (eds.) *Geology and Archaeology: submerged landscapes of the continental shelf*. Special Publications, Geological Society of London, London, UK. doi 10.1144/SP411.2
- Wurster, C.M., M.I. Bird, I.D. Bull, F. Creed, C. Bryant, J.A.J. Dungait, and V. Paz (2010), Forest contraction in north equatorial Southeast Asia during the Last Glacial Period, *Proceedings of the National Academy of Sciences*, USA, 107, 15508–15511.

Sample	Depth (mbsf)	Type	Conventional Age (years BP)	1 σ Error	Reservoir age correction (years)	Calibrated age (years BP)	1 σ Error
KIA 47048 3a	0.98	mixed planktonic forams	3065	30	200	2999	212
KIA 48078	1.88	wood	4945	55	na	5665	56
KIA 48079	3.06	wood	7330	40	na	8113	92
KIA 48403	3.96	<i>G. ruber</i>	8775	60	200	9650	248
KIA 50367	4.62	<i>G. ruber</i> + <i>G. sacculifer</i>	9980	55	200	11234	295
KIA 50361	5.06	<i>G. ruber</i>	10905	50	600	12104	142
KIA 50362	5.24	<i>G. ruber</i> + <i>G. sacculifer</i>	11605	60	600	12873	108
KIA 50363	5.38	<i>G. ruber</i>	12725	60	600	13967	144
KIA 50364	5.58	<i>G. ruber</i>	13570	70	600	15502	183
KIA 50365	5.96	<i>G. ruber</i>	14500	70	600	16842	194
KIA 50366	6.40	<i>G. ruber</i> + <i>G. sacculifer</i>	15650	80	1100	17708	347
KIA 50397	6.96	<i>G. ruber</i> + <i>G. sacculifer</i>	15860	90	1100	17966	335
KIA 50398	7.54	mixed planktonic forams	16870	100	1100	19068	300
KIA 50383	8.56	<i>G. ruber</i> + <i>G. sacculifer</i>	18240	110	750	21137	337
KIA 50399	9.76	mixed planktonic forams	20110	140	750	23339	354
KIA 50385	10.56	<i>G. ruber</i> + <i>G. sacculifer</i>	21190	150	750	24618	348
KIA 50384	11.76	mixed planktonic forams	22520	180	750	26029	272

Table 1 Raw and calibrated ^{14}C Data for core SO-217-18515-3. Calibrated and reservoir-corrected ages were used to determine the age model.

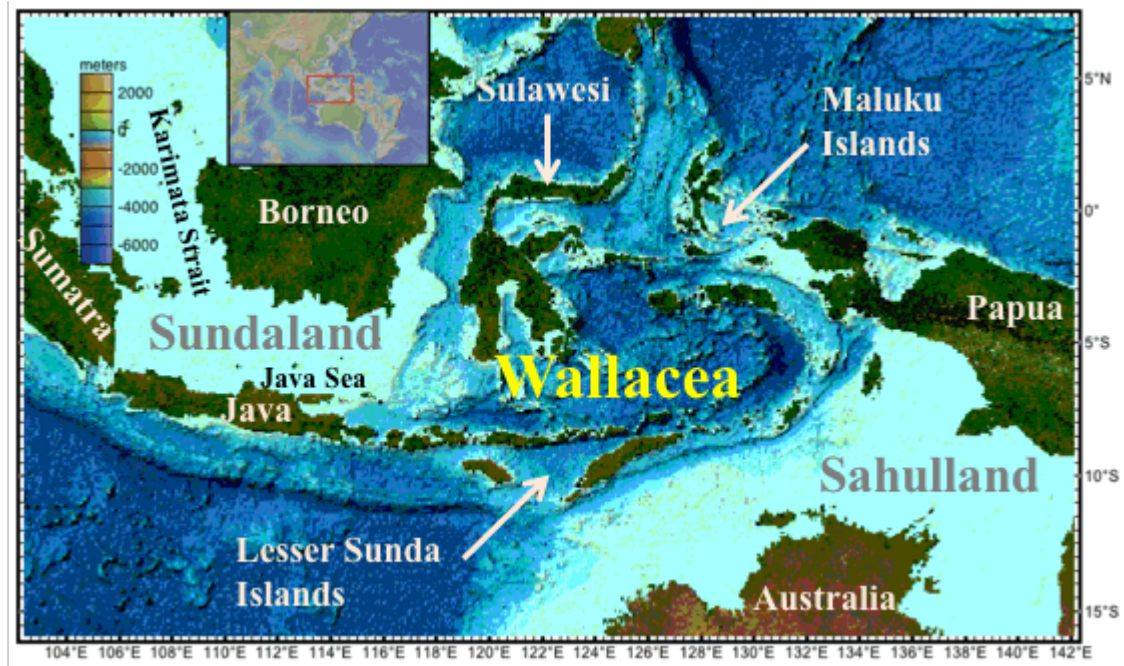


Figure 1 Bathymetric map of the Indonesian archipelago. During the Last Glacial Maximum, the exposed Sundaland was separated from Sahulland by deep and narrow straits, along with a group of islands termed Wallacea, comprised of Sulawesi, Maluku islands, and Lesser Sunda Islands (from GeoMapApp, bathymetric data from General Bathymetric Chart of the Ocean 2008).

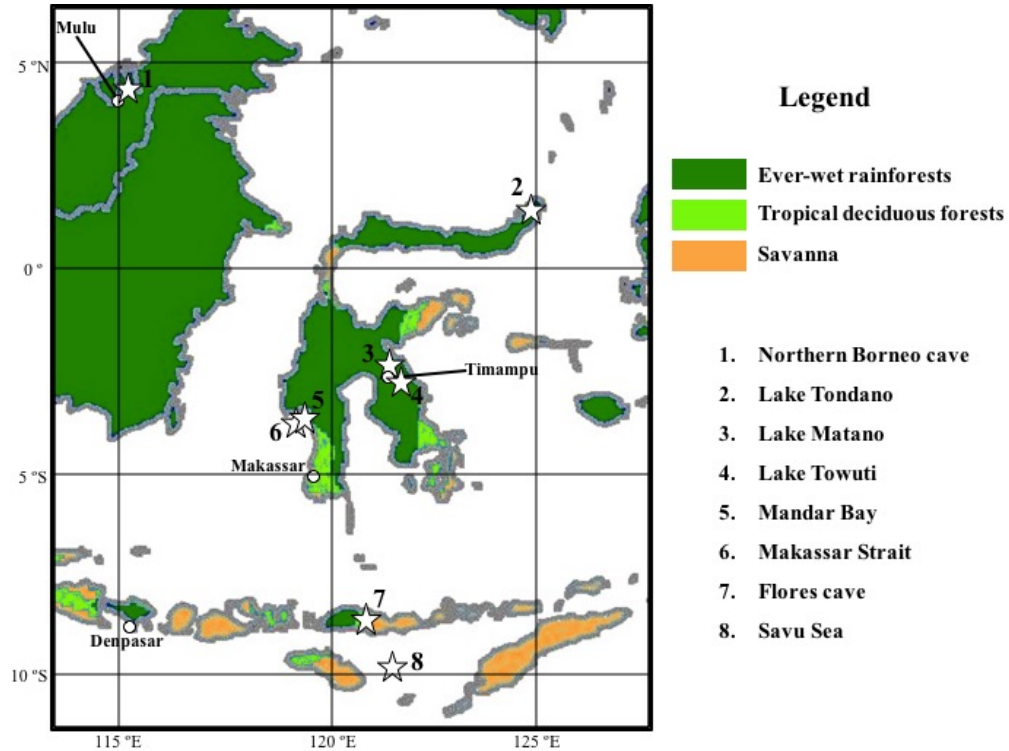


Figure 2 Map of Köppen–Geiger climate-vegetation classification for Wallacea (adapted from Peel et al. 2007). Ever-wet rainforests dominate central Sulawesi, whereas monsoon forests and savannas are common in southern Wallacea. Locations of our study site (Mandar Bay) and other proxy records discussed in the text are marked with white stars. White dots indicate the locations of rainfall and/or rainfall isotope stations discussed in the text.

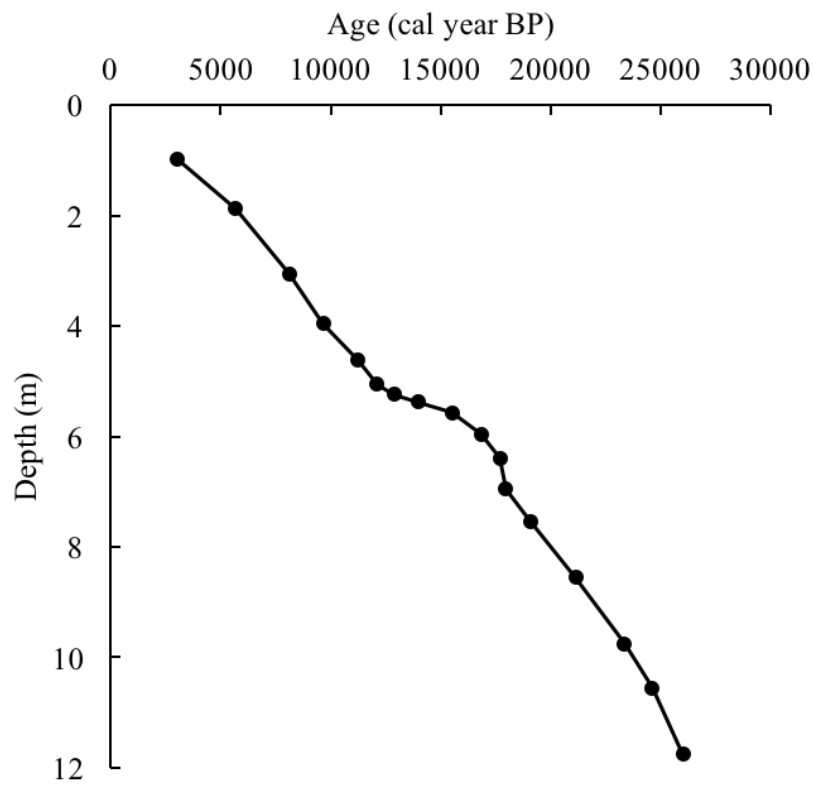


Figure 3 Age-depth relationship for core SO-217-18515-3.

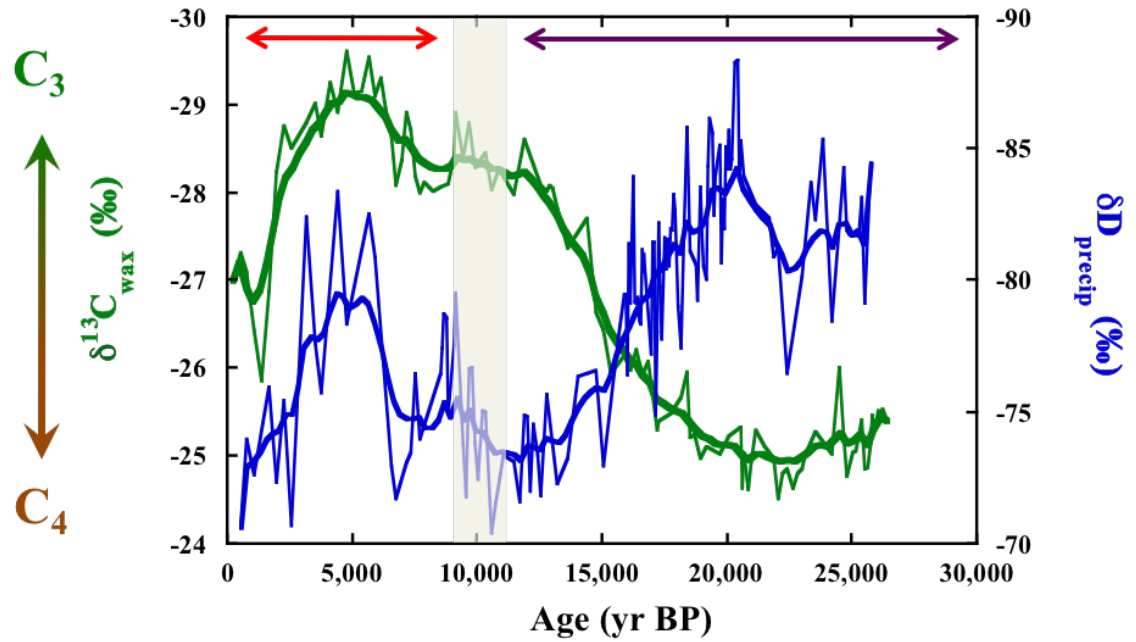


Figure 4 Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ since 26 kyr BP are plotted in this same time series plot. Dashed lines represent a Gaussian smoothing of the data to emphasize long-term trends. $\delta\text{D}_{\text{precip}}$ was largely anti-correlated with $\delta^{13}\text{C}_{\text{wax}}$ during the LGM and the last deglaciation (indicated by the purple arrow), but was positively correlated during most of the Holocene (red arrow). Shaded area indicates the transitional period whereby the switch in correlation likely occurred.

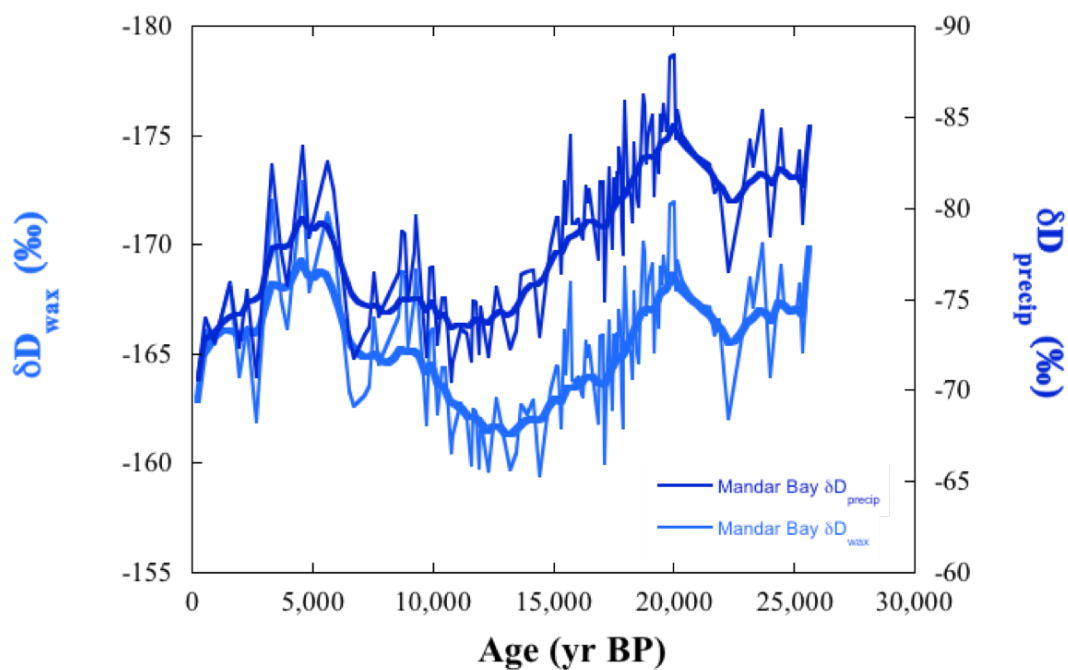


Figure 5 Comparison between raw Mandar Bay δD_{wax} records (light blue) with those of vegetation- and ice volume-corrected δD_{precip} (dark blue).

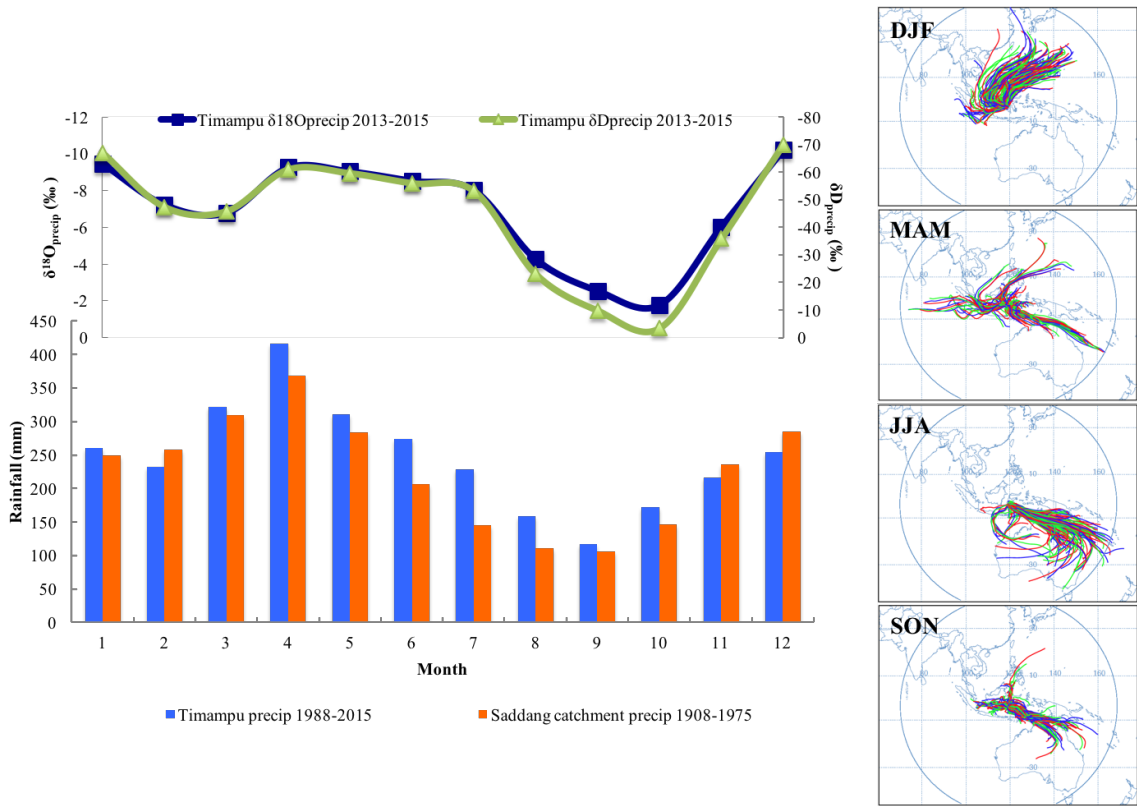


Figure 6 Monthly mean precipitation for Timampu near Lake Towuti from 1988 to 2015 (light blue bars; PT Vale Indonesia, unpublished), as well as for Saddang catchment close to Mandar Bay (orange bars) show that the two regions in central Sulawesi have a similar precipitation seasonality. The Saddang catchment data are the averages from three WMO stations within the catchment of Saddang River, namely Rantepao (34 year dataset), Mamasa (37 year dataset), and Enrekang (39 year dataset). The data were downloaded from the GHCN (2013) data via the KNMI explorer interface, and their locations are indicated in Figure 2. Above, light green and dark blue lines display $\delta\text{D}_{\text{precip}}$ and $\delta^{18}\text{O}_{\text{precip}}$, respectively, that have been measured from Timampu rainfall since 2013 (Konecky et al., 2015). On the right are back-trajectory analyses for Mamasa. The solid lines represent HYSPLIT (Draxler and Rolph 2013)-calculated 168-h back-trajectories of air parcels that generated rainfall during four trimesters (December-February, March-

May, June-August, September-November) of July 2003–June 2004, a 1-year interval that did not experience El Niño Southern Oscillation and Indian Ocean Dipole-related anomalies. Different colors represent the height at which the air parcel originates. Back trajectory analyses for Timampu yield very similar patterns.

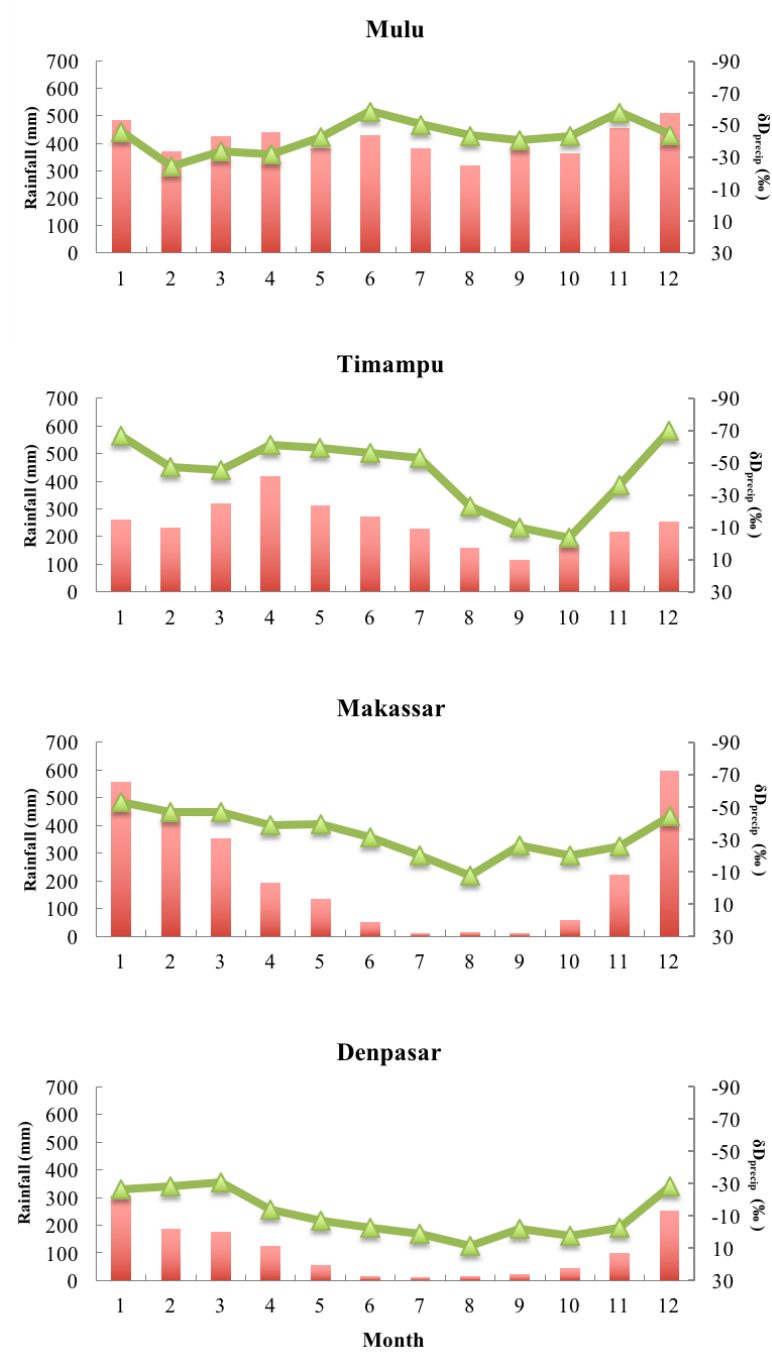


Figure 7 Monthly mean precipitation and δD_{precip} for (from top to bottom) Mulu, Timampu, Makassar, and Denpasar (bottom). The data from Timampu are the same as in Figure 3, whereas Makassar and Denpasar data are from JAMSTEC stations, collected in

2003-2006 and published in Kurita et al. (2009). Data from Mulu were collected in 2006-2011 and published by Moerman et al. (2013). The locations for these stations are shown in Figure 2.

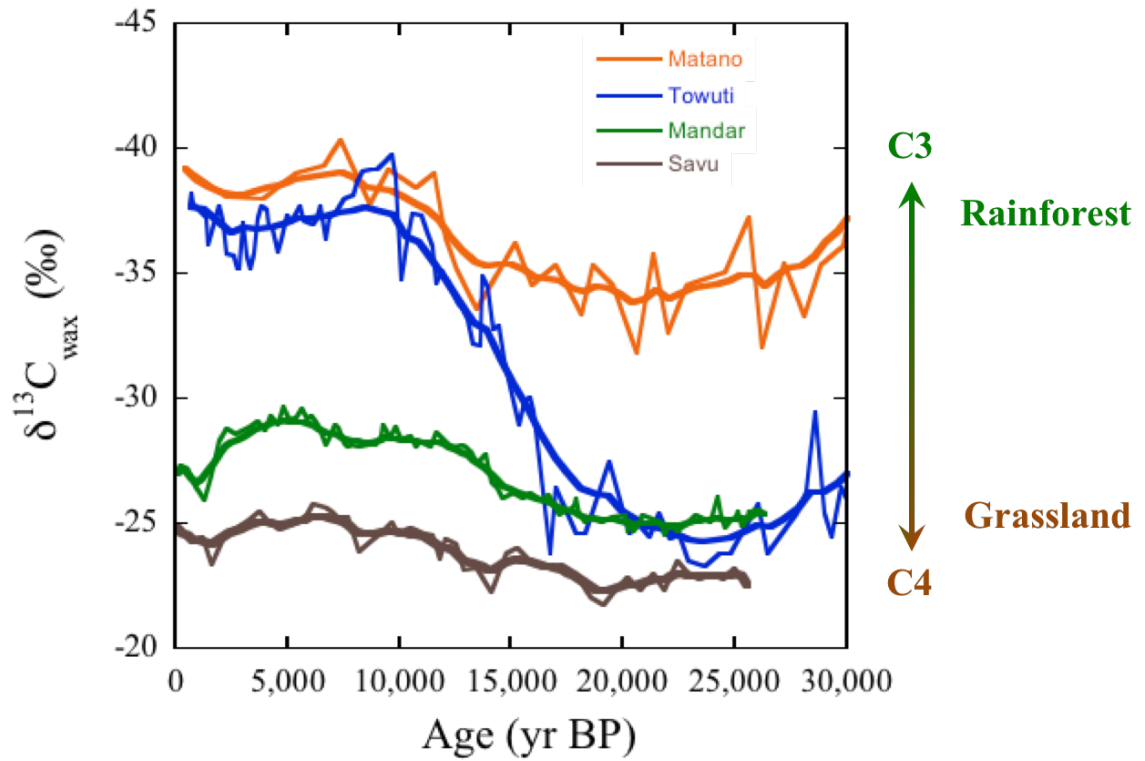


Figure 8 Comparison of Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$ record with other Wallacean $\delta^{13}\text{C}$ records.

From top to bottom: 1) Lake Matano $\delta^{13}\text{C}_{\text{wax}}$ (Wicaksono et al. 2015), 2) Lake Towuti $\delta^{13}\text{C}_{\text{wax}}$ (Russell et al. 2014), 3) Mandar Bay (this study), and 4) Savu Sea $\delta^{13}\text{C}_{\text{wax}}$ (Dubois et al. 2014).

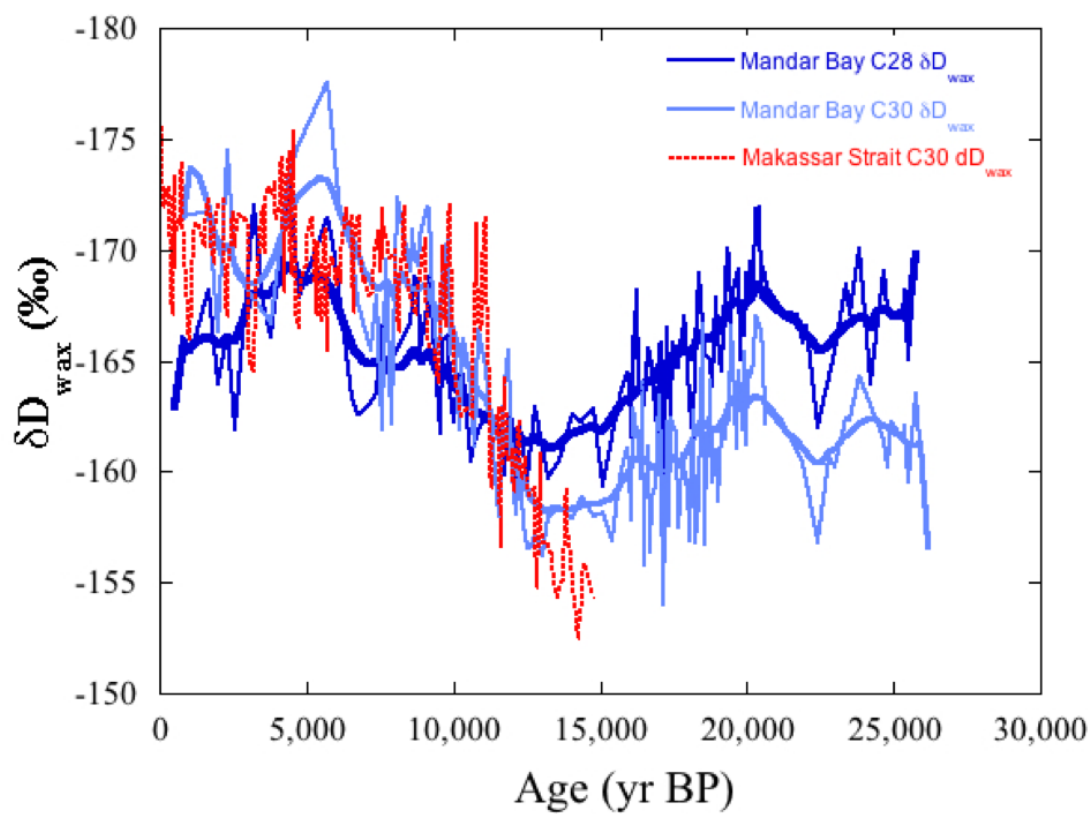


Figure 9 Comparison between C28 and C30 *n*-acid Mandar Bay δD_{wax} records (dark blue and purple respectively) with Makassar Strait raw, C30 *n*-acid δD_{wax} data (red, Tierney et al. 2012).

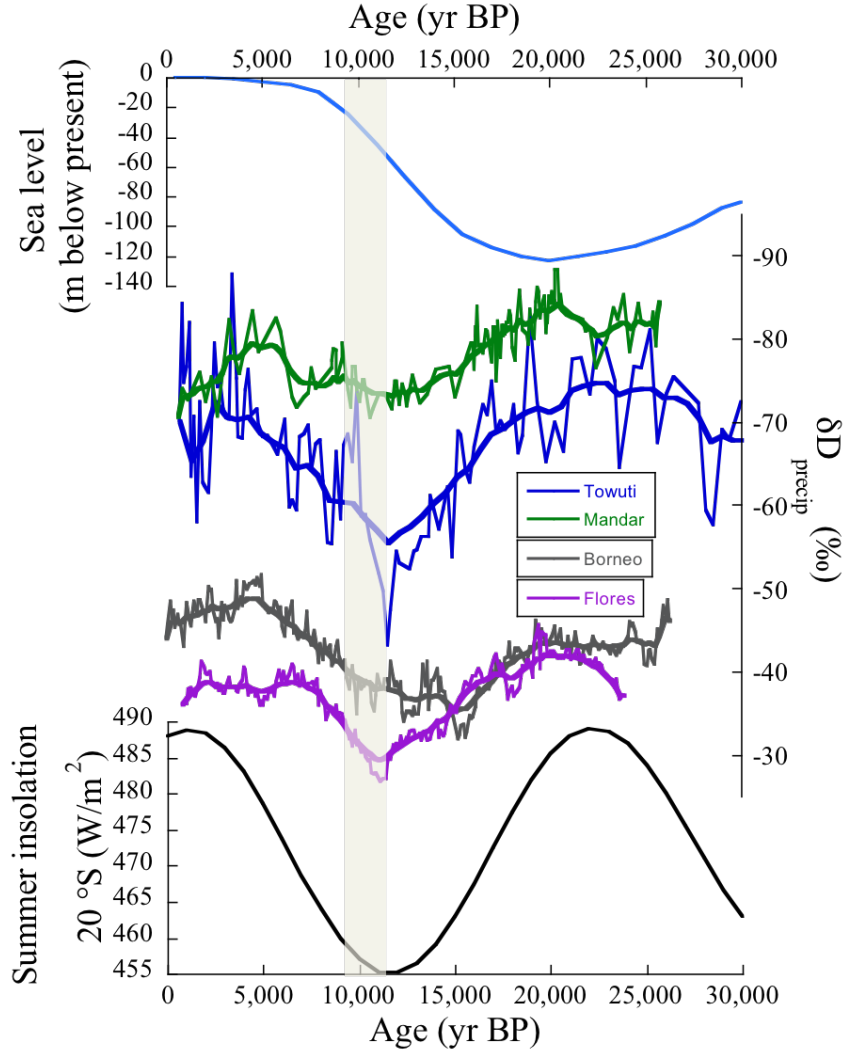


Figure 10 Comparison of Mandar Bay δD_{precip} records with climate forcings and other Wallacean δD_{precip} . From top to bottom: 1) sea level fluctuations, as a proxy for global ice volume, derived from benthic foraminiferal oxygen isotopes, as described in Waelbrock et al. (2002), 2) Mandar Bay δD_{precip} (this study), 3) Lake Towuti δD_{precip} (Konecky et al., in review), 4) Northern Borneo Cave δD_{precip} (computed from $\delta^{18}\text{O}_{\text{speleothem}}$ published in Partin et al. 2007), 5) Flores Cave δD_{precip} (computed from $\delta^{18}\text{O}_{\text{speleothem}}$ published in Griffiths et al. 2009 and Ayliffe et al. 2013), 5) southern hemisphere summer insolation

(calculated for 20°S, after Laskar et al. 2004). Shaded area marks the interval between 9.4 and 11.1 kyr BP when atmospheric regime switch likely occurred.

Time break (age yr BP)	R ² prior to time break	R ² after time break	Average R ²
8188.65	0.287	0.541	0.414
8795.95	0.282	0.540	0.411
9103.15	0.283	0.530	0.407
9277.35	0.282	0.543	0.412
9465.20	0.296	0.607	0.451
9742.45	0.292	0.591	0.441
10002.35	0.293	0.615	0.454
10325.00	0.292	0.598	0.445
10654.50	0.290	0.595	0.443
11111.50	0.292	0.556	0.424
11569.50	0.298	0.520	0.409
12011.00	0.294	0.507	0.401
12468.50	0.298	0.437	0.367
12943.00	0.294	0.414	0.354
13370.50	0.304	0.311	0.307
13717.00	0.310	0.208	0.259
14073.00	0.293	0.148	0.221
14314.50	0.228	0.091	0.160
14784.00	0.126	0.007	0.067
15526.00	0.064	0.001	0.032

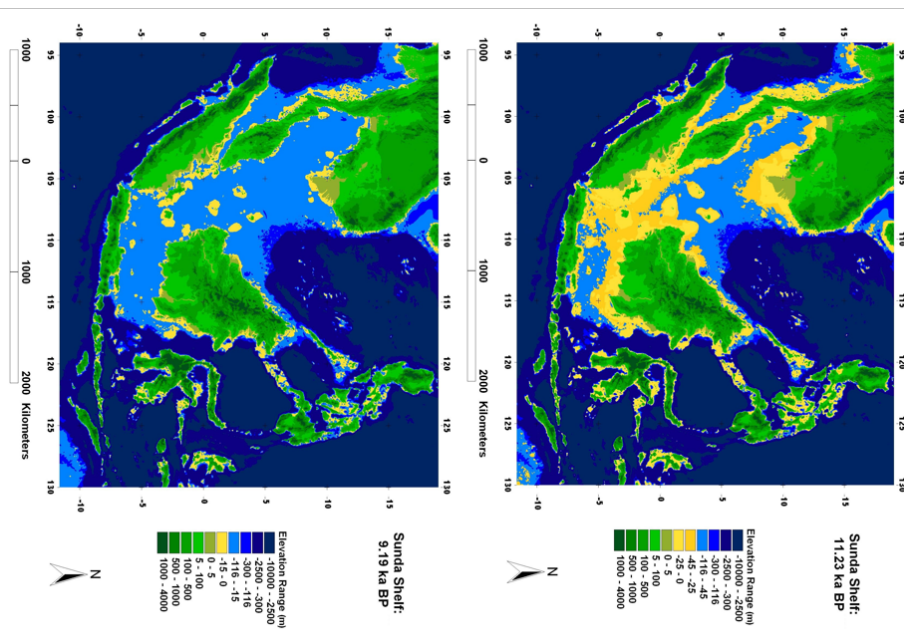


Figure 11 The R-squared values for both the correlation and anti-correlation between Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ are shown in the table (on the left), along with maps of exposed Sunda Shelf right before and after the key interval during which the switch in

correlation occurred (on the right, from Sathiamurthy and Voris 2006). The highest averages of R-squared values for the correlation and anti-correlation were found between 11.1 and 9.4 kyr BP, indicating that the correlation switch occurred at this interval, coincident with the rapid inundation of the Sunda Shelf. Indeed, as shown on the maps, both the Java Sea and Karimata Strait became significant bodies of water during this interval, thereby creating a new, large moisture source region within the Indonesian archipelago.

Supplementary Information

To convert ice volume-corrected $\delta^{18}\text{O}_{\text{speleothem}}$ from Liang Luar Cave, Flores and Gunung Buda Cave, Borneo to $\delta\text{D}_{\text{precip}}$ (Figure S1), a series of calculations and corrections were performed.

We first used the following carbonate-water paleotemperature equation (Fairchild and Baker 2012) to calculate estimates of dripwater $\delta^{18}\text{O}$ from the $\delta^{18}\text{O}_{\text{speleothem}}$ values measured on cave calcite:

$$T (^{\circ}\text{C}) = 19.4 - (5.3 * \delta^{18}\text{O}_{\text{speleothem}} \text{VPDB}) + (5.3 * \delta^{18}\text{O}_{\text{dripwater}} \text{VSMOW})$$

Paleotemperature data were derived from the closest available sea surface temperature records (SST) from both sites. For Flores, assuming that Liang Luar cave temperatures have always been ~ 3 degrees cooler than sea surface temperatures (Griffiths et al. 2010), we subtracted 3 degrees the Lombok Basin SST data (Levi et al. 2007). Similarly, we used SST data from the Celebes Sea (Stott et al. 2007) to estimate Gunung Buda cave temperatures, also assumed to be ~ 3 degrees cooler than sea surface temperatures given the cave's elevation (Cobb et al. 2007).

We then subtracted 1‰ from the $\delta^{18}\text{O}_{\text{dripwater}} \text{VSMOW}$ data computed via the paleotemperature equation, as both Griffiths et al. (2010) and Moerman et al. (2013) reported a ~ 1 -permil enrichment in $\delta^{18}\text{O}$ of the fluid inclusions within their speleothems relative to local rainfall.

Finally, the computed $\delta^{18}\text{O}_{\text{precip}}$ values were converted into $\delta\text{D}_{\text{precip}}$ using modern local meteoric water line (LMWL) equations. The LMWL equation for Liang Luar is $\delta\text{D} = 6.8 * \delta^{18}\text{O} + 4.8$ (Griffiths et al. 2009), whereas for Gunung Buda the LMWL equation is $\delta\text{D} = 7.9 * \delta^{18}\text{O} + 10.3$ (Moerman et al. 2013).

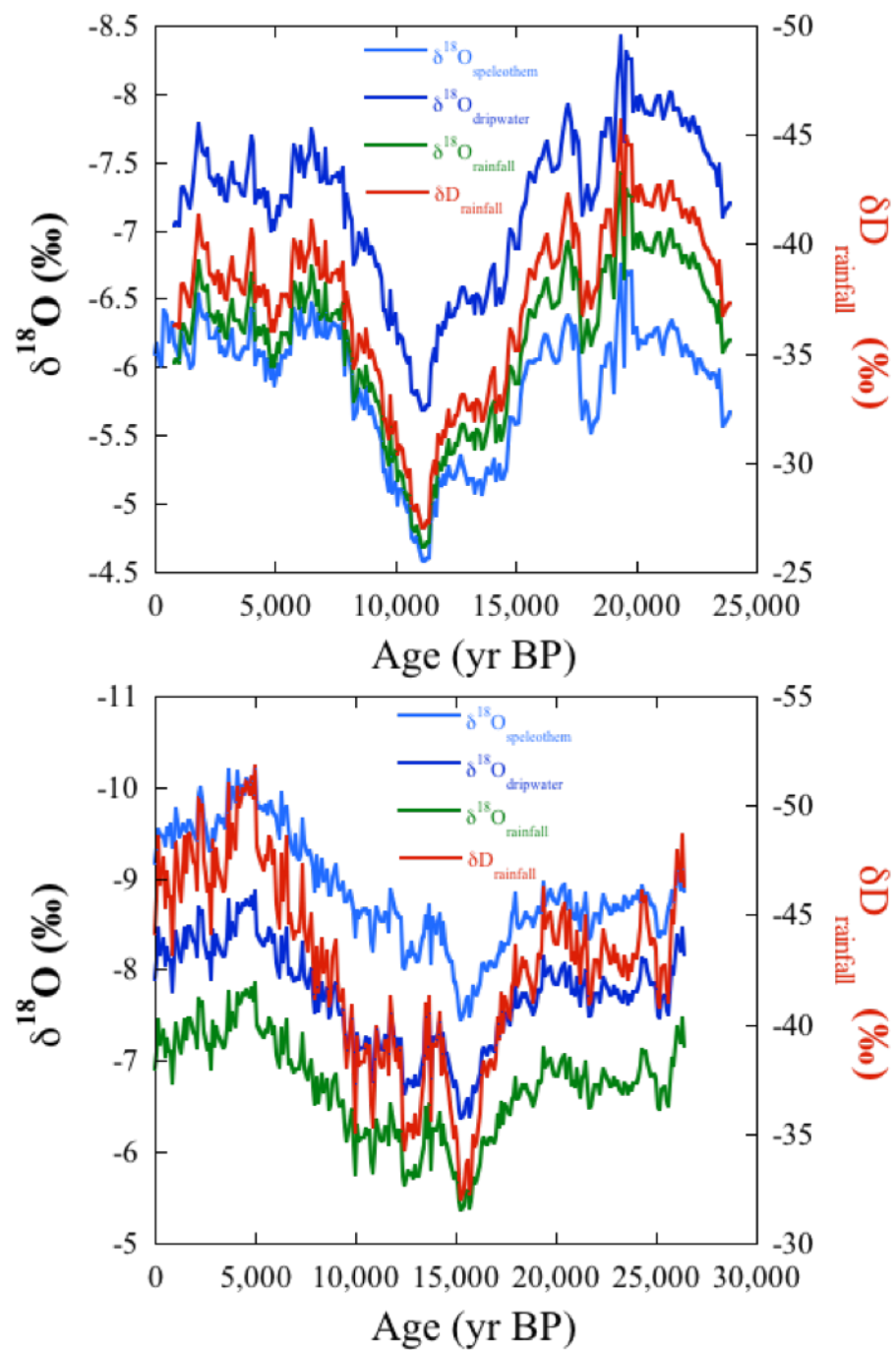


Figure S1 Conversion of $\delta^{18}\text{O}_{\text{speleothem}}$ to $\delta\text{D}_{\text{precip}}$ for Liang Luar Cave, Flores (top) and Gunung Buda Cave, Borneo (bottom).

CHAPTER FOUR

Shifts in rainfall and vegetation of northern Papua during the past 16 kyr BP

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Abstract

The Western Pacific Warm Pool (WPWP) is a major center of tropical convection that strongly influences global climate. Two competing hypotheses have been proposed to explain orbital-scale variations in WPWP convective activity: precession-forced interhemispheric Intertropical Convergence Zone (ITCZ) migration *and* zonal circulation shifts associated with the Walker circulation. Here we use new plant biomarker records measured from sediment cores retrieved from Cenderawasih Bay, off northern Papua, Indonesia at the heart of WPWP, to reconstruct changes in terrestrial vegetation and precipitation to understand WPWP convective history since 16 kyr BP. These records indicate a major reduction in precipitation between 12 and 6 kyr BP, accompanied by an expansion of C₄ taxa, suggesting that orbital-scale northern Papua precipitation is regulated by southern hemisphere summer insolation and ITCZ movement. Our coupled leaf wax carbon isotope ($\delta^{13}\text{C}_{\text{wax}}$) and hydrogen isotope ($\delta\text{D}_{\text{wax}}$) records also implicate precipitation amount as the primary control on vegetation changes in the catchment, reconciling long-standing questions existing charcoal records from Papua and Papua New Guinea elucidates our understanding of the interactions between humans and the Papuan environment during the Holocene.

1. Introduction

The Western Pacific Warm Pool (WPWP) is the major source of moisture and heat to the extra-tropics and therefore plays a critical role in global climate (Meehl 1987,

Cane and Clement 1999). Climate proxy records from the WPWP indicate large changes in regional rainfall during the Late Pleistocene and Holocene (e.g. Deser and Wallace 1990, Reeves et al. 2013), but the primary forcings and mechanisms that govern WPWP precipitation variability on orbital time-scales are disputed. While some records suggest that regional hydrologic variations were caused by interhemispheric shifts in the position and strength of convection within the Intertropical Convergence Zone (ITCZ), forced by orbital precession (Tachikawa et al. 2011, McGee et al. 2014) and/or obliquity (Liu et al. 2015), climate models and data syntheses suggest that orbital-scale changes in regional hydrology could occur due to orbitally-forced changes in atmospheric circulation associated with the Walker circulation (diNezio and Tierney 2013), with important implications on El Niño–Southern Oscillation (ENSO) variability (Tudhope et al. 2001). These climate processes intersect on the island of New Guinea, where paleoceanographic records suggest a strong control of insolation on paleoprecipitation (Tachikawa et al. 2011, Liu et al. 2015), while terrestrial vegetation records have been interpreted to reflect orbitally-driven changes in the Walker Circulation and ENSO variability (Gagan et al., 2004).

New Guinea is the second largest island in the world, and is comprised of Papua New Guinea (PNG) to the east and the Papuan province of Indonesia to the west (hereafter Papua). It is located just south of the WPWP, resulting in vigorous regional atmospheric convection and great expanses of evergreen tropical rainforest that support extremely high biodiversity. Paleohydrologic reconstructions from the island are scant, and their interpretations are often equivocal. For instance, a record of terrigenous runoff in a marine sediment core retrieved north of PNG indicates increased rainfall during the

late Holocene, interpreted to reflect southern hemisphere precessional insolation forcing of the position of the ITCZ (Tachikawa et al. 2011). In contrast, charcoal reconstructions from New Guinea suggest increased fire frequency and a drier climate during the late Holocene, interpreted to reflect changes in the Walker Circulation and intensification of El Niño-Southern Oscillation variability (Haberle et al. 2001, Gagan et al. 2004). Unfortunately, the interpretation of vegetation records based on fossil pollen and charcoal is complicated by the long history of anthropogenic impacts on Papuan vegetation (e.g. Hope 2007, Denham et al. 2004, Haberle 2007). New records that can independently resolve vegetation and hydrologic change could clarify the primary mode of regional rainfall variability and distinguish the impacts of past climate changes versus anthropogenic forest clearance in Papua.

Here we develop new leaf wax stable isotope records of climate and vegetation change from a terrestrially sourced marine sediment core retrieved off northern Papua. Our data cover the last 16 kyr BP, capturing part of the last deglaciation and the Holocene –key intervals in climate and human history– thus allowing us to evaluate the response of regional precipitation and atmospheric circulation to climate forcings as well as the influence of local climate and human populations on vegetation.

2. Study Site

Core MD01-2382 (2.70 °S, 135.78 °E; core length: 33.5 m, Figure 1) was retrieved in Cenderawasih Bay approximately 7 km off the Derewo (Degeuwo)-Wapoga River Basin. This catchment covers a large area in northern Papua (~10,234 km²), whose

topography ranges from the tall mountains of the Central Range (> 3,500 m above sea level) to the lowland delta area near Cenderawasih Bay. Waren and Nabire (Figure 1), the closest stations to Derewo-Wapoga Basin, receive more than 3,700 mm of precipitation annually (Figure 2) driven by strong year-round convection over the WPWP (Prentice and Hope 2007). Monthly precipitation exhibits a minimum from May to October and a maximum between December and April (Figure 2). This seasonality is associated with moisture flows from the north from December-April, as the Austral-Asian monsoon strengthens and the ITCZ migrates southward over the region (Figure 2).

Tropical rainforest dominates Derewo-Wapoga Basin. Lowland evergreen rainforests, swamp forests, and mangroves are abundant below 300 m asl, with montane rainforests between 300 m and 3200 m asl, and subalpine and alpine zones present above 3200 m asl. Both C₃ and C₄ taxa are found in almost all elevations and biomes, except above 3200 m asl, where C₃ grasslands and shrublands are the only vegetation types present (Bird et al. 1994, Johns et al. 2007). C₄ taxa are typically found within the swamp grasslands prevalent around the rivers (Morrison 2002), in relatively dry areas in the montane zone (Johns et al. 2007), and in areas where clear-cutting and fires recently occurred (Marshall and Beehler 2007).

3. Methods

MD01-2382 (Figure 3) was retrieved during the 2001 International Marine Past Global Changes Study (IMAGES) VII- Western Pacific Margin expedition. We constructed an age model based on linear interpolation of six ¹⁴C age control points (5

from planktonic foraminifera and 1 from wood), covering the past 16 kyr BP (Table 1). ^{14}C ages were calibrated with the IntCal 13.14C curve using Calib 7.0.1 (Reimer et al. 2013). No reservoir correction was applied to ^{14}C ages because the measurement on foraminifera samples 19 cm below the sea floor yields a modern age (Table 1). The age control points indicate high and continuous sedimentation rates, averaging 5.8 yr/cm (Figure 3).

We measured the carbon and hydrogen isotopic composition of long chain *n*-alkanoic fatty acids (C_{26} , C_{28} , and C_{30}), derived from terrestrial plants (Sachse et al. 2012). Fatty acids were extracted and purified according to protocol outlined in Hou et al. (2008). We extracted lipids from homogenized, freeze-dried sediment samples using a Dionex Accelerated Solvent Extractor. The acids were purified using aminopropyl column chromatography, and then methylated with acidified methanol of a known isotopic composition. The fatty acid methyl esters were purified using silica gel chromatography and analyzed in at least duplicate via gas chromatography isotope ratio monitoring-mass spectrometry (GC-IR-MS) using a Thermo Delta V Plus (for $\delta^{13}\text{C}_{\text{wax}}$) and Thermo Delta Plus XL (for $\delta\text{D}_{\text{wax}}$) at Brown University. We measured 69 samples for $\delta^{13}\text{C}_{\text{wax}}$ and 86 samples for $\delta\text{D}_{\text{wax}}$.

Here we report the $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{wax}}$ ($\delta\text{D}_{\text{precip}}$) values of the C_{28} homologue. C_{28} is the most abundant homologue in our sediment samples, has the lowest analytical error, and its isotopic compositions correlate well with other homologues. The average analytical errors for C_{28} *n*-acid were 0.12 ‰ for $\delta^{13}\text{C}_{\text{wax}}$ and 1.98 ‰ for $\delta\text{D}_{\text{wax}}$. The correlations of $\delta^{13}\text{C}_{\text{wax}}$ values between *n*-acid homologues were strong, with $r^2 = 0.95$, $p < 0.001$ for $\delta^{13}\text{C}_{\text{wax}}$ of C_{28} and C_{26} , and $r^2 = 0.81$, $p < 0.001$ for $\delta^{13}\text{C}_{\text{wax}}$ of C_{28} and C_{30} . The

correlations were weaker but still significant for δD_{wax} measurements (0.67 for C_{28} and C_{26} and 0.18 for C_{28} and C_{30} , $p < 0.001$). We corrected both $\delta^{13}C_{\text{wax}}$ and δD_{wax} for the isotopic composition of the methyl group added during methylation, and converted δD_{wax} into δD_{precip} based on apparent fractionation factors between C_{28} *n*-acid and precipitation for C_3 and C_4 plants (Chikaraishi and Naraoka 2007). We also applied a correction for global ice-volume changes to the reconstructed δD_{precip} based on benthic foraminiferal oxygen isotope data (Waelbroeck et al. 2002), which were converted to δD values using the global meteoric water line (Craig 1961).

4. Hydroclimatic interpretations of leaf wax isotopes

$\delta^{13}C_{\text{wax}}$ is a proxy for changes in vegetation type and in particular the ratio of C_3 versus C_4 plants, which in the tropics is controlled by hydroclimate (Schefuss et al. 2005, Diefendorf et al. 2010). C_3 plants –common in wetter, weakly seasonal climates– have $\delta^{13}C_{\text{wax}}$ of -30 ‰ to -45 ‰, whereas C_4 plants –favored in drier, more seasonal climates– have more positive $\delta^{13}C_{\text{wax}}$ ranging from -19 ‰ to -26 ‰ (Chikaraishi and Naraoka 2007). $\delta^{13}C_{\text{wax}}$ may also vary due to changes in the efficiency of carbon and water utilization within C_3 and C_4 plants, with more enriched $\delta^{13}C$ values in drier climates, although this process has a smaller effect on $\delta^{13}C_{\text{wax}}$ than the shifts between C_3 and C_4 plants (Diefendorf et al. 2010). While $\delta^{13}C_{\text{wax}}$ has not previously been studied on Papua, a study of the $\delta^{13}C$ of Papuan soils clearly indicates a decrease in soil $\delta^{13}C$ ($\delta^{13}C_{\text{soil}}$) with increasing C_3/C_4 ratio in the landscape, with moisture availability as the key control on $\delta^{13}C$ and C_3/C_4 abundance (Bird et al. 1994). We therefore interpret Cenderawasih Bay

$\delta^{13}\text{C}_{\text{wax}}$ record as a proxy for changes in the proportion of C_3/C_4 vegetation within Derewo-Wapoga Basin, which tracks precipitation.

$\delta\text{D}_{\text{wax}}$ records the δD of precipitation ($\delta\text{D}_{\text{precip}}$) used in plant photosynthesis, with a series of isotopic offsets during wax biosynthesis. The apparent isotopic fractionation between $\delta\text{D}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ is dependent upon plant physiology, including carbon fixation pathways (Sachse et al. 2012 and references therein). In the tropics, $\delta\text{D}_{\text{precip}}$ can be primarily controlled by local/regional precipitation amount via ‘the amount effect’ (Dansgaard 1964, Rozanski et al. 1992, Bowen 2008, Kurita et al. 2009). The amount effect describes the negative correlation between the isotopic composition of precipitation and precipitation rates, driven by the removal of isotopically-enriched water vapor as condensation proceeds in deep convective clouds during rainout (Rozanski et al. 1992, Vuille et al. 2003), and changes in evaporative enrichment as rain moves through dry vs. humid air (Lawrence et al. 2004). Long-term $\delta\text{D}_{\text{precip}}$ data from Madang and Jayapura (Figure 2), the two closest Global Network of Isotopes in Precipitation (GNIP) stations to Derewo-Wapoga Basin, indicate negative correlations between $\delta\text{D}_{\text{precip}}$ values and precipitation amount (-0.82 at Madang and -0.50 at Jayapura; Permana 2011). While many atmospheric processes can control $\delta\text{D}_{\text{precip}}$ (e.g. Moerman et al. 2013, Konecky et al. 2014), the strong correlation between modern $\delta\text{D}_{\text{precip}}$ and rainfall amount illustrates the utility of Cenderawasih Bay $\delta\text{D}_{\text{precip}}$ record as a proxy for northern Papuan precipitation amount since 16 kyr BP, a hypothesis we will test through comparison of reconstructed $\delta\text{D}_{\text{precip}}$ and $\delta^{13}\text{C}_{\text{wax}}$.

5. Results

During the past 16 kyr BP, $\delta^{13}\text{C}_{\text{wax}}$ varies between -26.9 ‰ and -32 ‰, with an average value of -30 ‰ (Figure 4). $\delta^{13}\text{C}_{\text{wax}}$ averages -31.3 ‰ between 16 and 12 kyr BP, then shifts toward more positive values, reaching the most positive values ($\sim$ -27 ‰) at $\sim$ 6.5 kyr BP. After 6 kyr BP, $\delta^{13}\text{C}_{\text{wax}}$ rapidly shifts to more negative values that are slightly more depleted than those observed between 16 and 12 kyr BP. Between 5 kyr BP and present, $\delta^{13}\text{C}_{\text{wax}}$ averages -31 ‰. Overall, millennial-scale variability in Cenderawasih Bay $\delta^{13}\text{C}_{\text{wax}}$ is largely absent, with amplitudes that are within or close to the analytical error. Instead, the ^{13}C -depletion between 12 and 5 kyr BP constitutes the most striking feature in our record.

After applying vegetation and ice volume corrections (Figure 5), Cenderawasih Bay $\delta\text{D}_{\text{precip}}$ displays similar patterns that strongly correlate with the observed changes in $\delta^{13}\text{C}_{\text{wax}}$ ($R^2 = 0.78$; Figure 4). $\delta\text{D}_{\text{precip}}$ values are relatively stable between 16 and 12 kyr BP, averaging -43 ‰, then become more negative between 12 and 5.7 kyr BP, averaging -18 ‰. After 5.7 kyr BP, $\delta\text{D}_{\text{precip}}$ became more negative, averaging -39.7 ‰ during the late Holocene. The shifts into and out of the mid-Holocene period of positive $\delta\text{D}_{\text{precip}}$ are abrupt.

6. Discussion

A variety of climatic, environmental, and biological processes can control vegetation distribution and therefore $\delta^{13}\text{C}_{\text{wax}}$. Low atmospheric CO_2 concentrations favor

C₄ grasses relative to C₃ plants (Ehleringer et al. 1997); however, atmospheric $p\text{CO}_2$ increased during the last deglaciation and was relatively high and stable through the Holocene (Monnin et al. 2001), in contrast to our $\delta^{13}\text{C}_{\text{wax}}$. Similarly, while temperature may partly explain the decrease in C₄ species relative to C₃ species observed today in Papua with altitude, sea surface temperatures recorded at sites close to New Guinea (e.g. MD97-2140, deGaridel-Thoron et al. 2005; MD05-2920, Tachikawa et al. 2013; and MD05-2928, Shiao et al. 2011) do not appear to track changes observed in Mandar Bay $\delta^{13}\text{C}_{\text{wax}}$, suggesting temperature is not the major control on $\delta^{13}\text{C}_{\text{wax}}$. Consequently, precipitation changes are the most plausible mechanism to account for the C₃/C₄ shift, similar to inferences from other Western Pacific $\delta^{13}\text{C}_{\text{wax}}$ records (e.g. Russell et al. 2014, Dubois et al. 2014, Wicaksono et al. 2015).

The strong correlation between $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ also implies that both proxies are controlled largely by precipitation amount –the climate parameter most likely to influence both $\delta\text{D}_{\text{precip}}$ and C₃/C₄ shifts in Papuan vegetation. On monthly time-scales the amount effect exerts a strong influence on $\delta\text{D}_{\text{precip}}$ in northern Papua (Permana 2011), but on longer time-scales the amount effect could change and/or be opposed by other processes controlling precipitation isotopes (Konecky et al. 2014, Moerman et al. 2013, Wicaksono et al. in prep). For instance, changes in the sign of the correlation between $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ records from central Sulawesi suggests that seasonal changes in moisture source and atmospheric circulation dominate the $\delta\text{D}_{\text{precip}}$ signal in central Indonesia during the LGM, while $\delta\text{D}_{\text{precip}}$ during the Holocene primarily records changes in precipitation amount (Wicaksono et al. in prep). In contrast, the strong, stable

correlation between $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ in Papua indicates a stable relationship between $\delta\text{D}_{\text{precip}}$ and precipitation amount during the past 16 kyr BP.

Thus, the 20-permil shift in $\delta\text{D}_{\text{precip}}$ after 12 kyr BP and the concomitant 3-permil increase in $\delta^{13}\text{C}_{\text{wax}}$, and similar shifts but with the opposite sign at 6 kyr BP, imply a drier climate between 12 and 6 kyr BP which favored forest opening, with wetter conditions prior to 12 and after 6 kyr BP that favored the rainforests observed on the landscape today. Our evidence for early to mid-Holocene drying resembles proxy records for terrigenous input from marine sites off PNG (Figure 4), which have been interpreted to reflect changes in erosion and terrestrial hydroclimate. Ti/Ca from MD05-2920 (Tachikawa et al. 2011) is generally low during the early to mid-Holocene, implying reduced precipitation and erosion during the dry interval indicated by our $\delta\text{D}_{\text{precip}}$ and $\delta^{13}\text{C}_{\text{wax}}$ records. There are differences between the records; for instance, Ti/Ca peaks between 9 and 8 kyr BP, indicating a wet period not recorded by our datasets. This could be related to the offshore and more eastern location of MD05-2920, subjecting the site to winnowing by oceanic currents (Tachikawa et al. 2011) and/or to additional terrigenous inputs from the volcanically and seismically active region of eastern PNG (Chappell 1993). Lower resolution proxy records from MD05-2928, located offshore from southern PNG, also demonstrate reduced ^{232}Th during the Holocene, indicating reduced terrigenous input and a decrease in southern PNG rainfall. While this supports our finding for drying during the late deglaciation, late Holocene data are missing from MD05-2928 (Shiau et al. 2011).

The dry interval in Papua between 12 and 6 kyr BP coincides with drying observed in many records from the southern tropics (Figure 6), such as Botuvera cave in

Brazil (27°S, Wang et al. 2007), and is contemporaneous with inferred wetter conditions in the northern tropics as recorded by elemental data from Cariaco Basin in Venezuela (~10°N, Haug et al. 2001) and $\delta^{18}\text{O}$ records from Sanbao and Hulu caves in China (31-32°N, Wang et al. 2001, Wang et al. 2008). This indicates that the orbital-scale interhemispheric antiphasing observed in these regions extends to the WPWP, implicating global shifts in the mean position of the ITCZ as the primary cause of changing hydrologic conditions in Papua. Furthermore, our records are well correlated to southern hemisphere seasonal insolation during the summer months (December to March) (Figure 6). At the start of the last deglaciation and in the late Holocene, perihelion coincides with the southern hemisphere summer, shifting the mean location ITCZ southward and intensifying the wet Austral-Asian summer monsoon over Papua. In contrast, the strengthening of the northern hemisphere summer insolation relative to the southern hemisphere summer insolation between 12 and 6 kyr BP results in a more northern mean location of the ITCZ and/or a shorter period annually during which the ITCZ passes above northern Papua. These precession-dominated changes in seasonal insolation affecting northern Papua have been shown to also influence, along with obliquity, the long-term hydroclimate of southeastern PNG (Liu et al. 2015).

It is hypothesized that precession-forced changes in the Walker circulation and ENSO behavior cause shifts in northern Papuan rainfall by changing the mean state of the tropical Pacific (e.g. Koutavas et al. 2002) or changing the amplitude and/or frequency of the ENSO events (Clement et al. 2000, Moy et al. 2002). If the Walker Circulation were the primary driver of orbital-scale hydrological shifts on Papua, then we should observe orbital-scale antiphasing between records from the eastern and western sides of the

Pacific Ocean. This E-W antiphasing is not apparent in current reconstructions. Records from El Junco Lake in the Galapagos, for instance, indicate lake level rose from the early to lake Holocene, similar to our observations at Papua (Conroy et al. 2008). Precipitation proxy records from Peru and Bolivia in the southern tropics (e.g. Baker et al. 2005, Bird et al. 2011) indicate reduced precipitation in the early and mid-Holocene, similar to the records in northern Papua, therefore suggesting that El Niño-like or La Niña-like conditions are not the primary control of orbital-scale rainfall in Papua.

Interestingly, our δD_{precip} record displays relatively abrupt transitions into and out of the dry period, suggesting a non-linear response of regional rainfall to the insolation forcing. It is possible that these abrupt changes are highly localized, as terrigenous runoff records from MD05-2920 do not appear as abrupt as at our site. However, the abrupt onset of dry conditions at ~12 kyr BP we observe on Papua is synchronous with abrupt changes in precipitation observed in many parts of the tropics, e.g. over equatorial Africa (Garcin et al. 2007, Otto-Bliesner et al. 2014) and South America (Hughen et al. 1996, Haug et al. 2001). Abrupt changes in tropical precipitation at this time coincide with the end of cool conditions in the North Atlantic associated with the Younger Dryas (YD), and rapid resumption of the Atlantic meridional overturning circulation (McManus et al. 2004). We do not see clear evidence for the YD in northern Papua in our leaf wax records, reinforcing previous findings that YD cooling could not perturb western Pacific convection and/or the position of the tropical Pacific ITCZ, in contrast to Heinrich events (Carolin et al. 2013). However, our results suggest that rapid north Atlantic warming AMOC resumption, and the abrupt intensification of northern hemisphere monsoons and precipitation at the end of the YD observed over Africa (Otto-Bliesner et al. 2014), Asia

(Wang et al. 2007), and South America (Haug et al. 2001) were linked to abrupt drying in the southwest tropical Pacific as the ITCZ shifted northward.

Perhaps more interestingly, the abrupt transition to wetter conditions at ~6 kyr BP is synchronous with the abrupt termination of the African Humid Period observed in some northern African records (deMenocal 2015). This transition has been attributed to positive feedbacks between vegetation, surface albedo, and precipitation operating in the Sahara, such that small reductions in precipitation forced by seasonal insolation change catalyzed rapid climate and vegetation changes from a green, wetter Sahara to a desert state (e.g. Claussen et al. 1999, deMenocal 2015). Although strong correlation between our $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ records suggest strong coupling between precipitation and vegetation in Papua, it seems unlikely that the changes in vegetation were large enough to drive such feedbacks. The 5‰ enrichment in $\delta^{13}\text{C}_{\text{wax}}$ we observe in the mid-Holocene suggest modest expansion of grasses rather than complete ecosystem turnover, limiting the potential role of surface albedo effects. Whereas rainforest expansion and contraction might alter regional transpiration and, subsequently, precipitation, it is also possible that the increase in precipitation at this time reflects a locally abrupt response to the gradual southward migration of the ITCZ. Indeed, this mechanism may explain the locally abrupt but time-transgressive drying observed across North Africa during the mid-Holocene (Shanahan et al. 2015).

Vegetation and charcoal records from Papua and PNG have been previously interpreted to reflect both the climatic impacts of changes in ENSO and the Walker Circulation, and anthropogenic impacts associated with agricultural expansion (Haberle 2007). For example, Hope and Tulip (1994) interpreted an increase in charcoal from the

low elevation Lake Hordorli at 11.2 kyr BP to reflect the early development of agriculture in the lowlands, with other vegetation records indicating agricultural intensification in highland locations between 11.5 and 5 kyr BP (Hope 1983, Haberle et al. 1991, Haberle 1996, Denham et al. 2004). In contrast, a compilation of charcoal records from the New Guinean highlands suggest relatively low regional occurrence of fire from 9 to 5 kyr BP (Haberle et al. 2001), interpreted to reflect a wetter and less variable early Holocene climate driven by weaker ENSO variability than today (Gagan et al. 2004). While it can be difficult to disentangle paleoclimatic from paleoanthropologic effects on charcoal and pollen records, human activities should not affect our δD_{precip} record. Thus, our coupled δD_{precip} and $\delta^{13}C_{\text{wax}}$ records strongly suggest a dominant control of climate on vegetation at the scale of large river catchments such as Derewo-Wagopa. In this context, the early Holocene vegetation clearance and fire could be related to a drying climate, while the increased regional occurrence of charcoal after 5 kyr BP (Haberle et al. 2001) represents an intensification of human impacts on the island under a wetter climate in the late Holocene.

7. Conclusions

Coupled $\delta^{13}C_{\text{wax}}$ and δD_{precip} records from Mandar Bay indicate abrupt drying and C_4 expansion in northern Papua at 12 kyr BP, as the Northern Hemisphere warmed at the end of the Younger Dryas. An abrupt return to wetter conditions and rainforest expansion occurred at 6 kyr BP, as the ITCZ migrated southward due to precessional orbital forcing. Our coupled precipitation and vegetation proxy records indicate that

precipitation amount is the primary mechanism controlling northern Papua precipitation isotopes on the orbital scale, highlighting local human impacts on vegetation during the late Holocene.

References

- Bird, M.I., S.G. Haberle, and A.R. Chivas (1994), Effect of altitude on the carbon isotope composition of forest and grassland soils from Papua New Guinea. *Global Biogeochem. Cycles*, 8, 13–22.
- Bowen, G. J. (2008), Spatial analysis of the intra-annual variation of precipitation isotope ratios and its climatological corollaries. *J. Geophys. Res.*, 113, D05113, doi:10.1029/2007JD009295.
- Bowman, D. M. (2000), *Australian rainforests: islands of green in the land of fire*. Cambridge, U.K: Cambridge Univ. Press.
- Cane, M., Clement, A.C. (1999), A role for the tropical Pacific coupled ocean-atmosphere system on Milankovitch and millennial timescales. Part II: global impacts. In: Clark, P.U., Webb, R.S., Keigwin, L.D. (Eds.), *Mechanisms of Global Climate Change at Millennial Time Scales*. American Geophysical Union, Washington, D.C, pp. 373-383.
- Chappell, J. (1993) Contrasting Holocene sedimentary geologies of lower Daly River, northern Australia, and lower Sepik-Ramu, Papua New Guinea, *Sediment. Geol.*, 83, 339–358.
- Chikaraishi, Y. and H. Naraoka (2007), $\delta^{13}\text{C}$ and δD relationships among three n-alkyl compound classes (n-alkanoic acid, n-alkane and n-alkanol) of terrestrial higher plants. *Org. Geochem.* 38, 198-215.
- Claussen, M. (2009) Late Quaternary vegetation-climate feedbacks. *Climate of the Past*, 5, 203–216.

- Claussen, M., C. Kubatzki, V. Brovkin, A. Ganopolski, P. Hoelzmann, and H.-J Pachur (1999), Simulation of an abrupt change in Saharan vegetation in the mid-Holocene. *Geophysical Research Letters*, 26, 2037–2040.
- Conroy, J.L., J.T. Overpeck, J.E. Cole, T.M. Shanahan, and M. Steinitz-Kannan (2008), Holocene changes in eastern tropical Pacific climate inferred from a Galápagos lake sediment record, *Quaternary Science Reviews*, 27(11-12), 1166-1180.
- Craig, H. (1961), Isotopic variations in meteoric waters. *Science*, 133, 1702–1703.
- Dansgaard, W. (1964), Stable isotopes in precipitation. *Tellus*, 16(4), 436–468, doi:10.1111/j.2153-3490.1964.tb00181.x.
- de Garidel-Thoron, T., Y. Rosenthal, F. Bassinot, and L. Beaufort (2005), Stable sea surface temperatures in the west Pacific warm pool over the past 5 million years. *Nature*, 433, 294–298, doi:10.1038/nature03189.
- deMenocal, P.B. (2015), The End of the Humid Period. *Nature Geoscience* 8, 86–87.
- Denham T.P, S.G. Haberle, and C. Lentfer (2004). New evidence and revised interpretations of early agriculture in highland New Guinea. *Antiquity*, 78, 839–857.
- Diefendorf, A.F., K.E. Mueller, S.L. Wing, P.L. Koch, and K.H. Freeman (2010) Global patterns in leaf ^{13}C discrimination and implications for studies of past and future climate. *Proc Natl Acad Sci USA* 107(13), 5738–5743.
- DiNezio, P.N., and J.E. Tierney, J.E. (2013), The effect of sea level on glacial Indo-Pacific climate. *Nat Geosci* 6, 485–491.
- Draxler, R. R. & Rolph, (2013), G. D. HYSPLIT—HYbrid Single-Particle Lagrangian Integrated Trajectory Model <<http://www.arl.noaa.gov/ready/hysplit4.html>>.

- Dubois N., D.W. Oppo, V. Galy, M. Mohtadi, S. van der Kaars, J. Tierney, Y. Rosenthal, T. Eglinton, A. Lückge, and B. Linsley (2014), Indonesian vegetation response to changes in rainfall seasonality over the past 25,000 years. *Nature Geoscience* 7, 513-517.
- Ehleringer, J. R., T.E. Cerling, and B.R. Helliker (1997), C₄ photosynthesis, atmospheric CO₂, and climate. *Oecologia* 112, 285–299.
- Gagan, M.K., E.J. Henty, S.G. Haberle, W.S. Hantoro (2004), Post-glacial evolution of the Indo-Pacific Warm Pool and El Niño-southern oscillation. *Quatern. Int.*, 118–119, 127–143.
- Garcin, Y., A. Vincens, D. Williamson, G. Buchet, and J. Guiot (2007), Abrupt resumption of the African Monsoon at the Younger Dryas-Holocene climatic transition. *Quaternary Science Reviews*, 26, 690-704, doi:10.1016/j.quascirev.2006.10.014.
- Global Historical Climatology Network (2013), GHCN Monthly version 3.2.0. data set. Available on-line from KNMI Climate Explorer <http://climexp.knmi.nl>.
- Haberle S.G, G.S. Hope, and W.A. Van der Kaars (2001), Biomass burning in Indonesia and Papua New Guinea: natural and human induced fire events in the fossil record. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 171, 259–268.
- Haberle S.G. (1996), Palaeoenvironmental changes in the eastern highlands of Papua New Guinea. *Archaeol. Oceania*, 31, 1–11
- Haberle S.G., G.S. Hope, and Y. DeFretes (1991), Environmental changes in the Baliem Valley, montane Irian Jaya, Republic of Indonesia. *J. Biogeogr.* 18, 25–40.

- Haberle, S.G. (2007), Prehistoric human impact on rainforest biodiversity in highland New Guinea. *Philosophical Transactions of the Royal Society B: Biological Sciences*, 362, 219– 228.
- Haug, G., K. Hughen, D. Sigman, L. Peterson, and U. Rohl (2001), Southward migration of the Intertropical Convergence Zone through the Holocene, *Science*, 293(5533), 1304–1308.
- Hoffmann, W.A., E.L. Geiger, S. Gotsch, D.R. Rossatto, L.C.R. Silva, O.L. Lau, M. Haridasan, and A.C. Franco (2012), Ecological thresholds at the savanna-forest boundary: How plant traits, resources and fire govern the distribution of tropical biomes. *Ecology Letters*, 15, 759-768.
- Hope G.S. (1983), The vegetational changes of the last 20000 years at Telefomin, Papua New Guinea, *J. Trop. Geogr.*, 4, 25–33.
- Hope, G. and Tulip, J. (1994), A long vegetation history from lowland Irian Jaya, Indonesia. *Palaeogeogr., Palaeoclimatol., Palaeoecol.* 109, 385-398.
- Hope, G. S. (2007), Palaeoecology and palaeoenvironments of Papua. In Marshall, A. J., and Beehler, B. M. (eds.), *The Ecology of Papua*. Singapore: Periplus Editions.
- Hou, J., W. J. D'Andrea, and Y. Huang (2008), Can sedimentary leaf waxes record D/H ratios of continental precipitation? Field, model, and experimental assessments, *Geochim. Cosmochim. Acta*, 72(14), 3503–3517.
- International Atomic Energy Agency (IAEA) (2001). *GNIP Database, Maps and Animations*, International Atomic Energy Agency, Vienna. <http://isohis.iaea.org>

- Johns, R.J., G.A. Shea, and P. Puradyatmika (2007), Montane Vegetation of Papua. In Marshall, A. J., and Beehler, B. M. (eds.), *The Ecology of Papua*. Singapore: Periplus Editions.
- Konecky, B., J.M. Russell, M. Vuille, and K. Rehfeld (2014), Invited Review: The Indian Ocean Zonal Mode over the past millennium in observed and modeled precipitation isotopes. *Quaternary Science Reviews*, 103, 1-18, doi: 10.1016/j.quascirev.2014.08.019
- Koutavas, A., J. Lynch-Stieglitz, T.M. Marchitto, and J.P. Sachs (2002), El Nino-like pattern in ice age tropical Pacific sea surface temperature. *Science* 297 (5579
- Kurita, N., K. Ichiyanagi, J. Matsumoto, M. D. Yamanaka, and T. Ohata (2009), The relationship between the isotopic content of precipitation and the precipitation amount in tropical regions, *J. Geochem. Exp.*, 102, 113–122.
- Laskar, J., P. Robutel, F. Joutel, M. Gastineau, A.C.M. Correia, and B. Levrard (2004), A long-term numerical solution for the insolation quantities of the Earth. *Astron. Astrophys*, 428, 261–285.
- Lawrence, J.R., S.D. Gedzelman, D. Dexheimer, H.K. Cho, G.D. Carrie, R. Gasparini, C.R. Anderson, K.P. Bowman, and M.I. Biggerstaff (2004), Stable isotopic composition of water vapor in the tropics. *J. Geophys. Res. Atmos.* 109, D06115.
- Liu, Y., L. Lo, Z. Shi, et al. (2015), Obliquity pacing of the western Pacific Intertropical Convergence Zone over the past 282,000 years. *Nature Communications* 6, 10018, doi: 10.1038/ncomms10018.
- Marshall, A. J., and B. M. Beehler (eds.) (2007), *The Ecology of Papua*. Singapore: Periplus Editions.

- McGee, D., A. Donohoe, J. Marshall, and D. Ferreira (2014), Changes in ITCZ location and cross-equatorial heat transport at the Last Glacial Maximum, Heinrich Stadial 1, and the mid-Holocene, *Earth Planet. Sci. Lett.*, 390, 69–79.
- McGregor, H. V., and M. K. Gagan (2004), Western Pacific coral $\delta^{18}\text{O}$ records of anomalous Holocene variability in the El Niño – Southern Oscillation. *Geophys. Res. Lett.*, 31, L11204, doi:10.1029/2004GL019972.
- McManus, J.F., R. Francois, J.-M. Gherardi, L.D. Keigwin, and S. Brown-Ledger (2004), Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes, *Nature*, 428, 834–837, doi: 10.1038/nature02494.
- Meehl, G. A., (1987), The tropics and their role in the global climate system. *Geographical Journal*, 153, 21–36.
- Moerman, J.W., K.M. Cobb, J.F. Adkins, H. Sodemann, B. Clark, and A. Tuen (2013), Diurnal to interannual rainfall oxygen isotope variations in northern Borneo driven by regional hydrology. *Earth and Planetary Sciences Letter*, 369, 108-119.
- Monnin, E., A. Indermühle, A. Dällenbach, J. Flückiger, B. Stauffer, T.F. Stocker, D. Raynaud, and J.-M. Barnola (2001), Atmospheric CO_2 concentrations over the last glacial termination: *Science*, v. 291/5501, p. 112-114.
- Morrill C., J.T. Overpeck, and J.E. Cole (2003), A synthesis of abrupt changes in the Asian summer monsoon since the last deglaciation: *The Holocene*, v. 13, p. 465–476.
- Morrison, J. (2002), Details of Ecoregion Delineation. In E. D. Wikramanayake (Ed.), *Terrestrial Ecoregions of the Indo-Pacific: A Conservation Assessment*. Island Press

- Moy, C., G. O. Seltzer, D. T. Rodbell, and D. M. Anderson (2002), Variability of El Nino/Southern Oscillation activity at millennial timescales during the Holocene epoch, *Nature*, 420, 162–165.
- Otto-Bliesner, B.M., J.M. Russell, P.U. Clark, Z. Liu, J.T. Overpeck, B.L. Konecky, P.B. DeMenocal, S.E. Nicholson, F. He, and Z. Lu (2014), Coherent changes of southeastern equatorial and northern African rainfall during the last deglaciation, *Science*, 346, 1223–1227.
- Permana, D.S. (2011), *Climate, Precipitation Isotopic Composition and Tropical Ice Core Analysis of Papua*, Indonesia. M.Sc Thesis at the School of Earth Sciences, The Ohio State University, Ohio, USA.
http://rave.ohiolink.edu/etdc/view?acc_num=osu1313480990
- Prentice, M. and G.S. Hope (2007), Climate of Papua. In: A. J. Marshall and B. M. Beehler, (Eds.), *The Ecology of Papua*. Singapore, Periplus Editions. pp 177-195.
- Ratnam, J., W.J. Bond., R.J. Fensham, W.A. Hoffmann, S. Archibald, C.E.R. Lehmann, M.T. Andersen, S.I. Higgins, M. Sankaran (2011), When is a "forest" a savanna, and why does it matter? *Global Ecology and Biogeography*, 20, 653-660.
- Reimer, P. J., E. Bard, A. Bayliss, et al. (2013), IntCal13 and Marine13 Radiocarbon Age Calibration Curves 0-50,000 Years cal BP. *Radiocarbon*, 55, 4.
- Reeves, J. M., H. C. Bostock, L. K. Ayliffe, et al. (2013), Palaeoenvironmental change in tropical Australasia over the last 30,000 years – a synthesis by the OZ-INTIMATE group. *Quaternary Science Reviews*,
<http://dx.doi.org/10.1016/j.quascirev.2012.11.027>

- Rozanski, K., L. Araguas-Araguas and R. Gonfiantini (1992), Relation between long-term trends of ^{18}O isotope composition of precipitation and climate, *Science*, 258, 981-985.
- Russell, J.M., H. Vogel, B.L. Konecky, S. Bijaksana, Y. Huang, M. Melles, N. Wattus, K. Costa, and J.W. King (2014), Glacial forcing of central Indonesian hydroclimate since 60,000 y B.P., *Proc Nat Ac Sci USA*, doi:10.1073/pnas.1402373111.
- Russell-Smith, J. (2004). Rain forest invasion of eucalypt-dominated woodland savanna, iron range, north-eastern Australia: I. Successional processes. *J. Biogeogr.*, 31
- Sachse, D, I. Billault, G.J. Bowen, Y. Chikaraishi, et al. (2012), Molecular paleohydrology: interpreting the hydrogen-isotopic composition of lipid biomarkers from photosynthesizing organisms. *Annu. Rev. Earth Planet. Sci.*, 40, 221–249
- Schefuß E., S. Schouten and R.R. Schneider (2005), Climatic controls on central African hydrology during the past 20,000 years. *Nature* 437, 1003–1006.
- Shanahan, T.M. et al. (2015), The time-transgressive termination of the African Humid Period, *Nat Geosci* 8(2), 140–144.
- Shiau, L.J., M.T. Chen, S.C. Clemens, C.A. Huh, M. Yamamoto, and Y. Yokoyama (2011), Warm pool hydrological and terrestrial variability near southern Papua New Guinea over the past 50 k. *Geophys. Res. Lett.* 38, L00F01.
- Tachikawa, K., A. Timmermann, L. Vidal, C. Sonzogni, and O.E. Timm (2013) CO_2 radiative forcing and Intertropical Convergence Zone influences on western Pacific warm pool climate over the past 400 ka. *Quaternary Science Reviews*, 86,

24–34, DOI:10.1016/j.quascirev.2013.12.018.

Tachikawa, K., O. Cartapanis, L. Vidal, L. Beaufort, T. Barlyaeva, and E. Bard (2011),

The precession phase of hydrological variability in the Western Pacific Warm Pool during the past 400 ka, *Quaternary Sci. Rev.*, 30, 3716–3727.

Talbot, M. R., M. L. Filippi, N. B. Jensen, and J.-J. Tiercelin (2007), An abrupt change in the African monsoon at the end of the Younger Dryas, *Geochem. Geophys. Geosyst.*, 8, Q03005.

Tudhope, A. W., C. P. Chilcott, M. T. McCulloch, E. R. Cook, J. Chappell, R. M. Ellam,

D. W. Lea, J. M. Lough, and G. B. Shimmield (2001), Variability in the El Nino Southern Oscillation through a glacial-interglacial cycle. *Science*, 291, 1511–1517.

Vuille, M., R. S. Bradley, M. Werner, R. Healy, and F. Keimig (2003), Modeling $\delta^{18}\text{O}$ in precipitation over the tropical Americas: 1. Interannual variability and climatic controls, *J. Geophys. Res.*, 108, 4174, doi:10.1029/2001JD002038.

Waelbroeck, C., L. Labeyrie, E. Michel, J.C. Duplessy, J. McManus, K. Lambeck, E.

Balbon, and M. Labracherie (2002), Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records. *Quaternary Science Reviews*, 21, 295–305.

Wang, X., A.S. Auler, R.L. Edwards, H. Cheng, E. Ito, Y. Wang, X. Kong, and M.

Solheid (2007), Millennial-scale precipitation changes in southern Brazil over the past 90,000 years, *Geophysical Research Letters*, 34, L23701, doi:10.1029/2007GL031149.

Wang, Y.J., H. Cheng, R.L. Edwards, X.G. Kong, X.H. Shao, S.T. Chen, J.Y. Wu, X.Y.

Jiang, X.F. Wang, and Z.S. An (2008), Millennial- and orbital-scale changes in the East Asian monsoon over the past 224,000 years, *Nature*, 451, 1090-1093.

Wang, Y.J., H. Cheng, R.L. Edwards, Z.S. An, J.Y. Wu, C.C. Shen, and J.A. Dorale

(2001), A high-resolution absolute-dated Late Pleistocene monsoon record from Hulu Cave, China, *Science*, 294, 2345-2348.

Wicaksono, S.A., J.M. Russell, J. M., and S. Bijaksana (2015), Compound-specific stable isotope records of vegetation and hydrologic change in central Sulawesi since

53,000 yr BP. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 430, 47-56.

Depth (cmbsf)	Sample Description	Conventional Age (years BP)	1σ Error	Calibrated Age (years BP)	1σ Error
19	mixed planktonic	365	30	0	
251	mixed planktonic	995	40	580	66
526	mixed planktonic	3400	75	3266	85
951	mixed planktonic	5220	40	5967	60
1651	mixed planktonic	10150	85	11229	231.5
3351	wood	13700	90	16444	248.5

Table 1 Calibrated ages used to determine the age model for core MD01-2382.

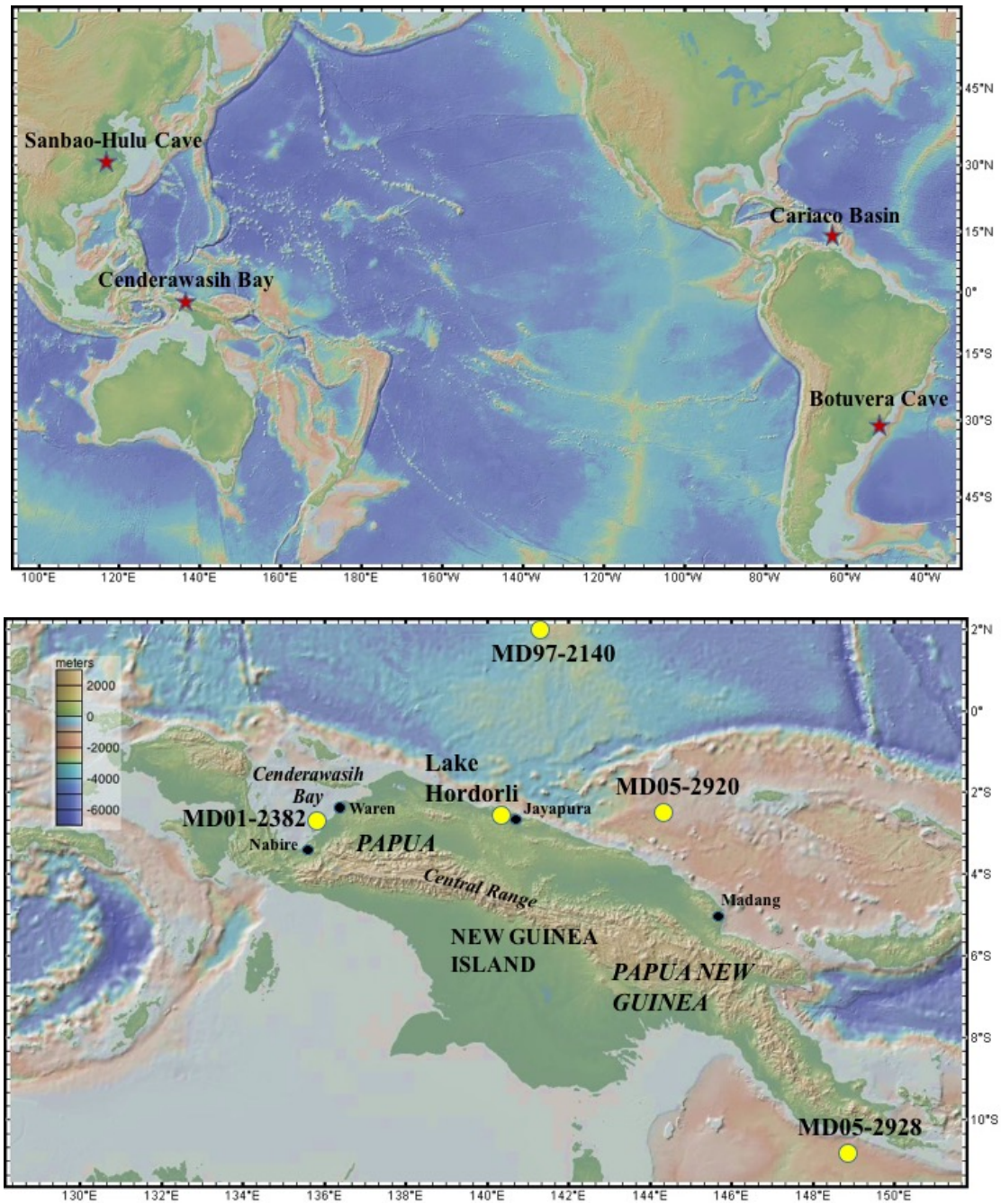


Figure 1 Our study site, Cenderawasih Bay, is indicated on both maps. Cenderawasih Bay, off northern Papua, part of the New Guinea island, is where core MD01-2382 was retrieved. Locations of other sites/climate records/weather stations discussed in the text are shown on the maps. Both maps were modified from GeoMapApp.

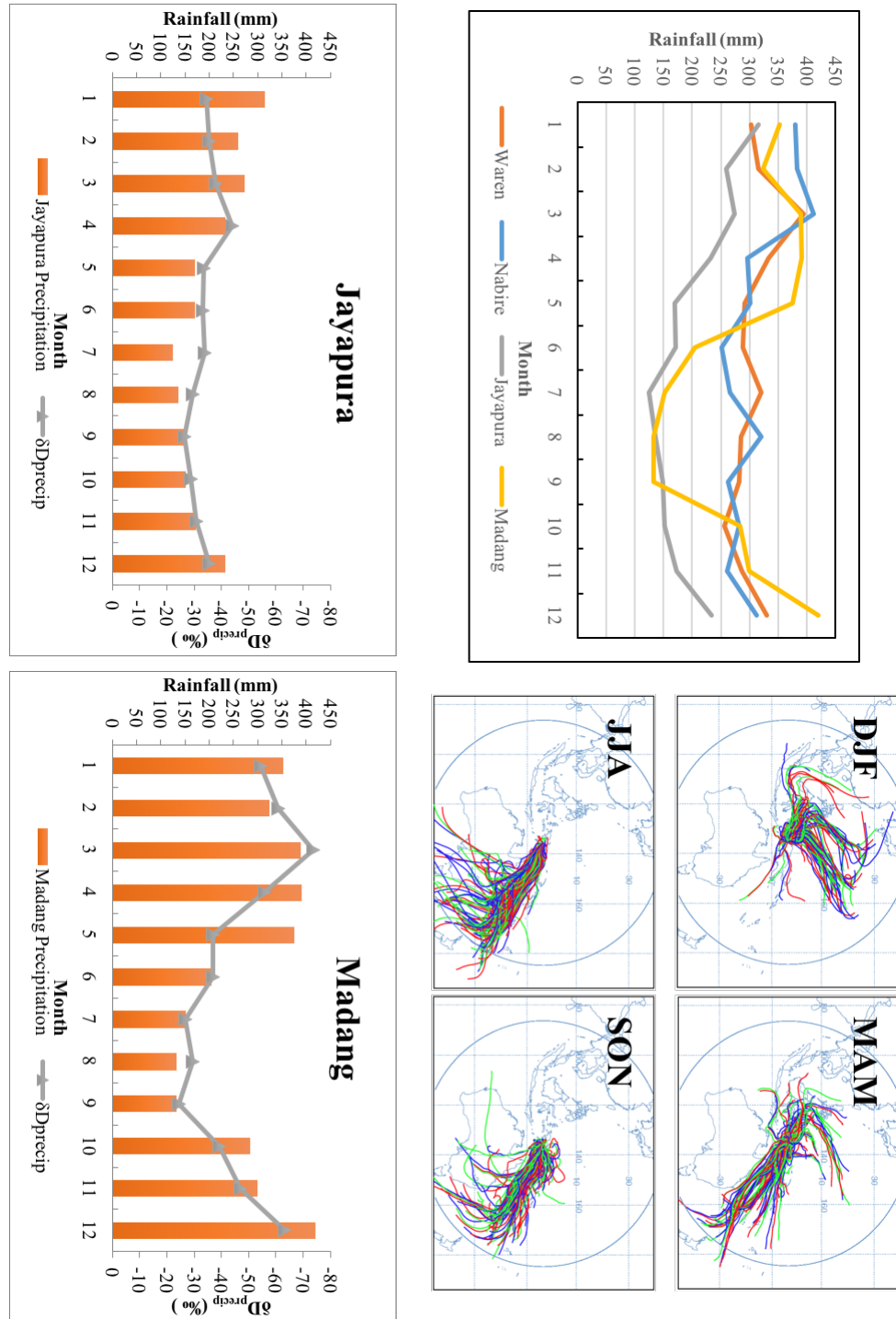


Figure 2 Climatological data for the study region. Top left: monthly mean precipitation for four weather stations in northern Papua; the monthly data for Waren (1950-194) and

Nabire (1952-1999) are from GHCN database (GHCN 2013), whereas the monthly data for Jayapura (1957-1991) and Madang (1968-1982) are from GNIP IAEA/WMO database (IAEA 2001). Top right: back-trajectory analyses for Nabire; the solid lines represent HYSPLIT (Draxler and Rolph 2013)-calculated 168-h back-trajectories of air parcels that generated rainfall during four trimesters (December-February, March-May, June-August, September-November) of July 2003–June 2004, a 1-year interval that did not experience El Niño Southern Oscillation anomalies; different colors represent the height at which the air parcel originates. Bottom: monthly mean precipitation and δD_{precip} for Jayapura and Madang; the data are from GNIP IAEA/WMO stations (IAEA 2001), collected in 1957-1991 (at Jayapura) and 1968-1982 (at Madang).

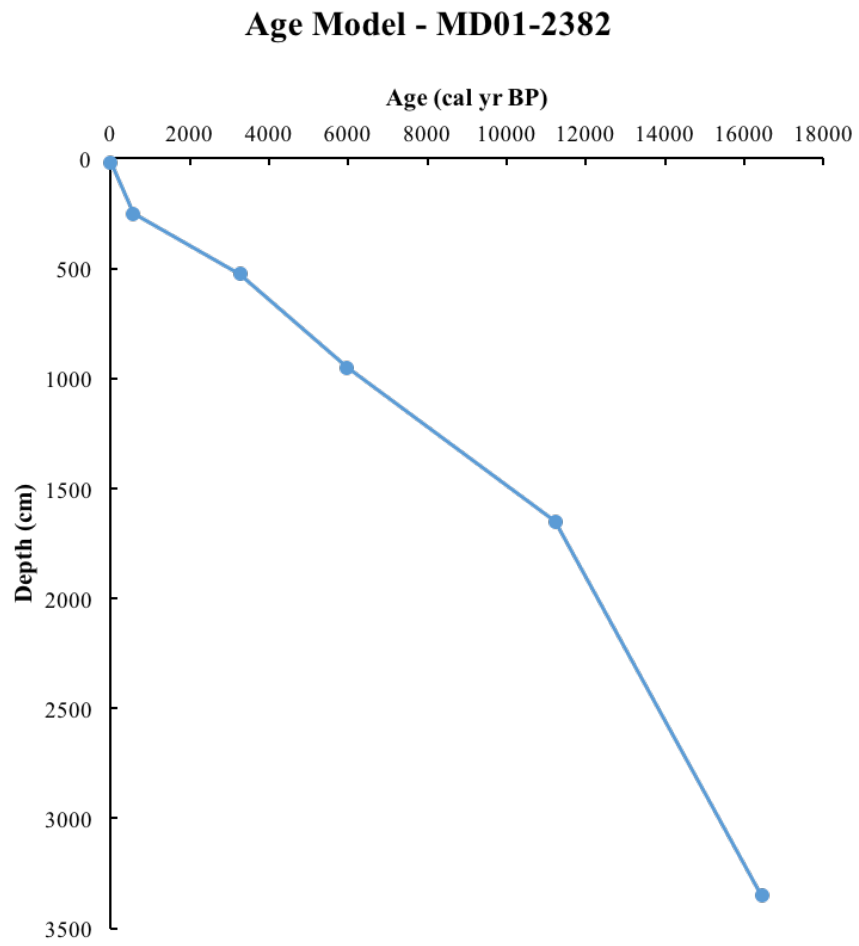


Figure 3 Age-depth profile for core MD01-2382.

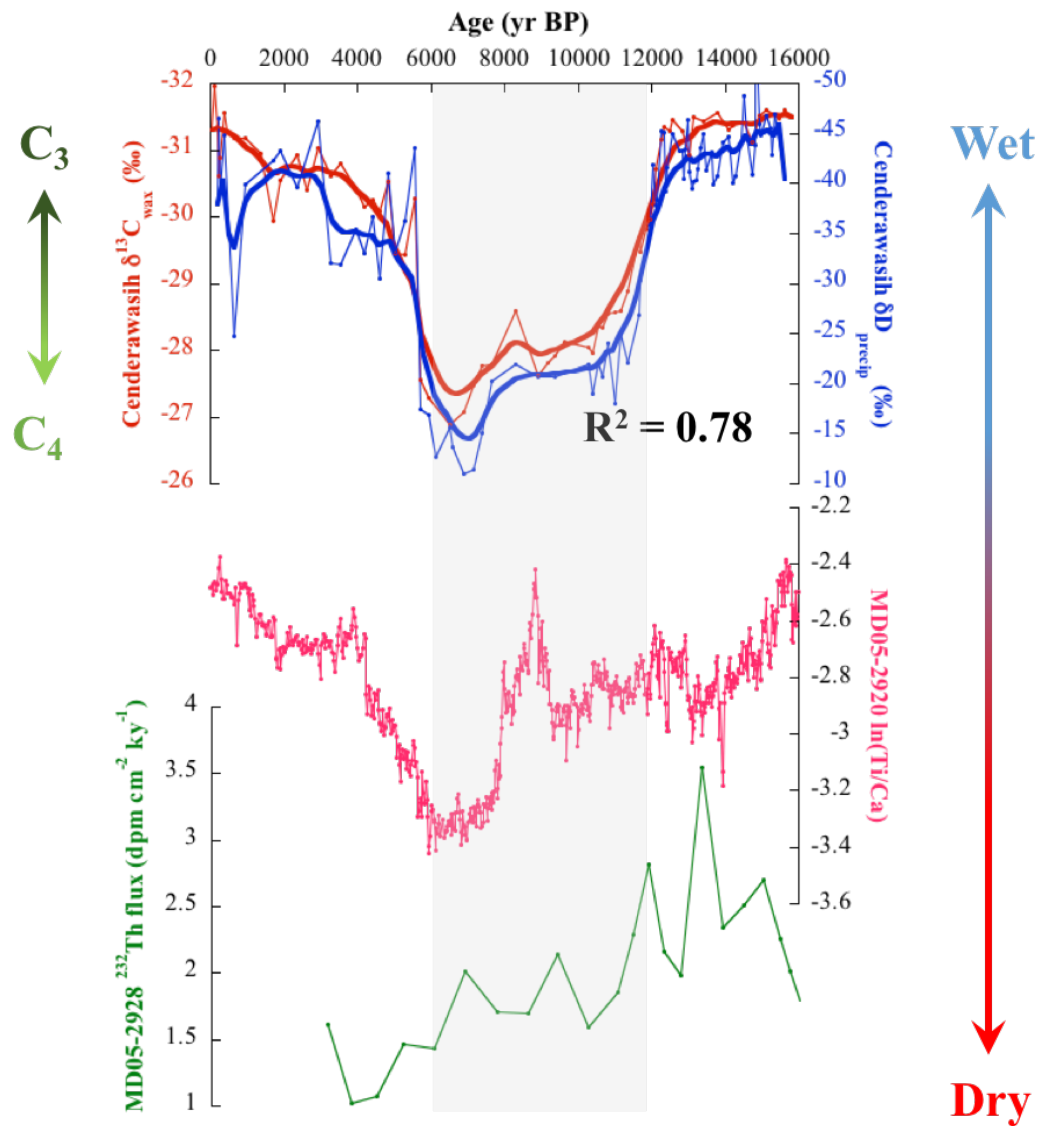


Figure 4 Comparison of Cenderawasih Bay $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ record with regional proxy records of paleoprecipitation. From top to bottom: 1) Cenderawasih Bay/MD01-2382 $\delta^{13}\text{C}_{\text{wax}}$ and $\delta\text{D}_{\text{precip}}$ (this study, solid lines represent a Gaussian smoothing of the data to emphasize long-term trends, and the R-squared value for the correlation between the two data sets is indicated), 2) Log-scale XRF intensity ratio of titanium and calcium of core MD05-2920 (Tachikawa et al. 2011) 3) ^{232}Th activity flux data of core MD05-2928 (Shiau et al. 2011).

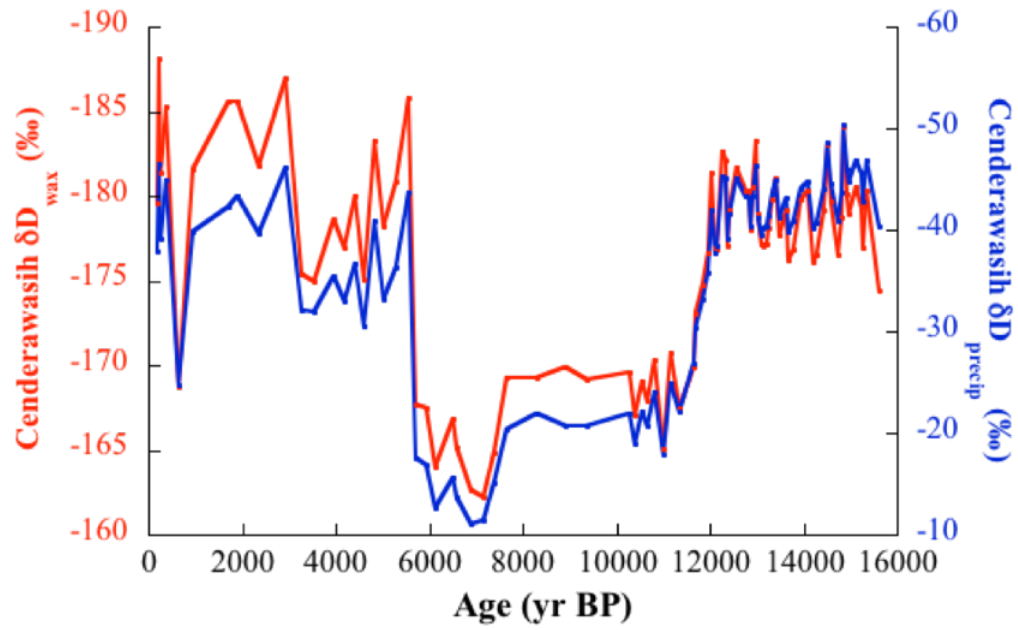


Figure 5 Comparison between raw Cenderawasih Bay δD_{wax} records (purple) with those of vegetation- and ice volume-corrected δD_{precip} (blue).

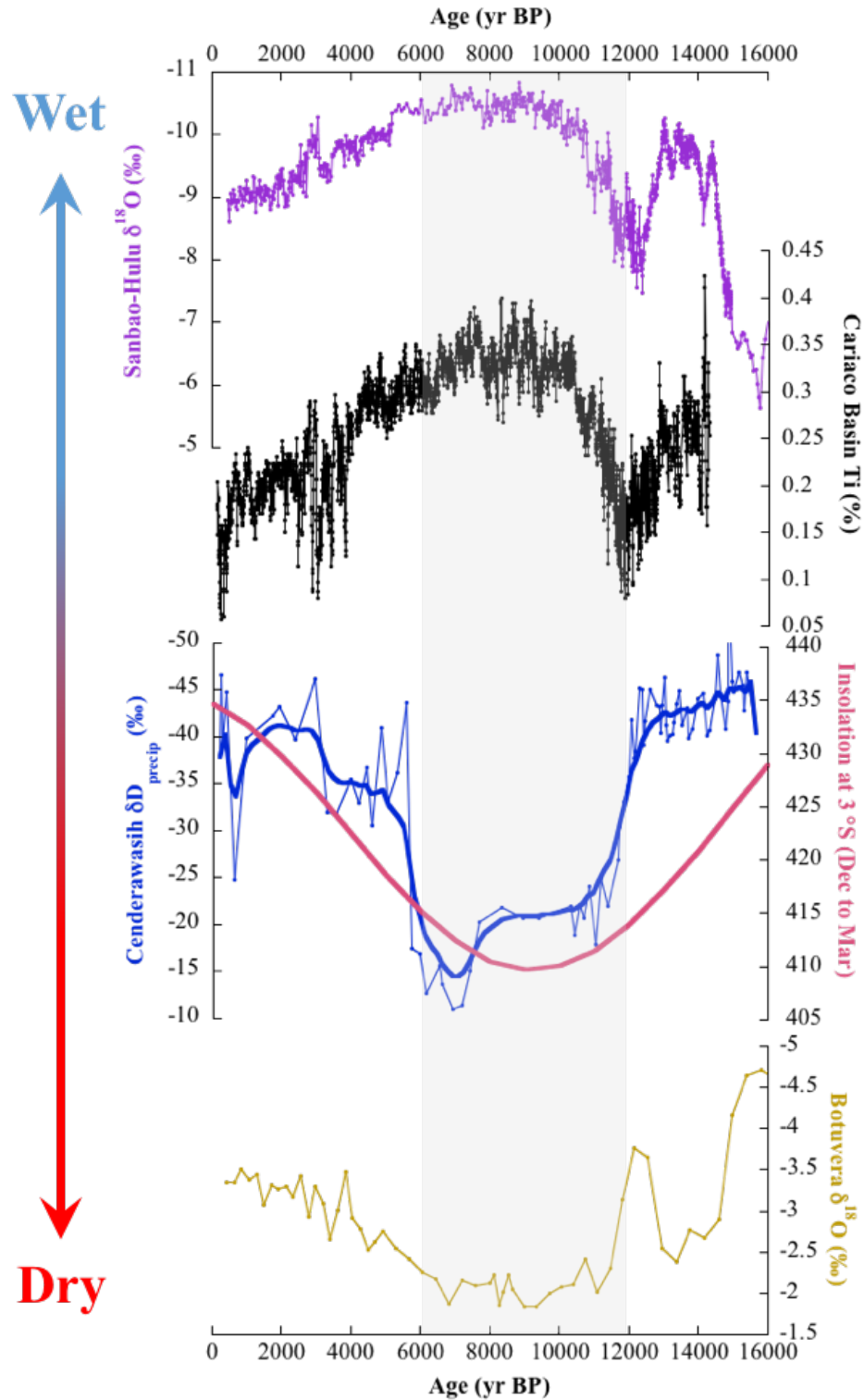


Figure 6 Comparison of Cenderawasih Bay δD_{precip} records with other tropical hydroclimate records. From top to bottom: 1) Sanbao and Hulu caves $\delta^{18}\text{O}$ (non-ice

volume corrected, Wang et al. 2001, Wang et al. 2008), 2) Cariaco Basin Ti concentration (Haug et al. 2001), 3) southern tropics average insolation for December through March (calculated for 3°S, after Laskar et al. 2004) superimposed on Cenderawasih Bay δD_{precip} (this study), 4) Botuvera cave $\delta^{18}\text{O}$ (non-ice volume corrected, Wang et al. 2007).

CHAPTER FIVE

Limited climatic and environmental responses to a major Late Pleistocene volcanic eruption recorded in tephra deposit from Lake Towuti, central Sulawesi, Indonesia

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Abstract

Large volcanic eruptions can cause major climatic shifts and environmental changes, including changes in both terrestrial and aquatic ecosystems. Major eruptions in the past in the tropics are particularly critical because of their potentially disproportionate global climate forcing relative to eruptions in the high latitudes. Here we report the detection of a ~6 cm-thick tephra layer in the sediments of Lake Towuti in central Sulawesi, Indonesia, which is the thickest tephra in the ~100-kyr-long sediment cores from the lake. Geochemical fingerprinting of tephra shards matches this deposit to tephra layers found in the Celebes Sea. Magnetostratigraphic data suggest an age of 74 ka for the tephra in the Towuti sediment core, very similar to the independently dated ~72 ka age for the tephra found in Celebes Sea. Based on the chemistry, thicknesses, and the locations of tephra emplacement, we hypothesize that the source of Towuti tephra was a major volcanic eruption in northeastern Sulawesi. We analyzed potential environmental and climatic impacts of the eruption by examining lacustrine paleoproductivity, vegetation, and elemental concentration data in the Towuti sediments before and after tephra deposition. Increased phosphorous concentrations and better diatom preservation above the tephra indicate that emplacement of this tephra perturbed the lake's biogeochemistry and ecology, likely via increased nutrient availability. However, other proxy data strongly suggest no medium- to long-term terrestrial ecological changes in central Sulawesi, nor regional drying and/or cooling resulted from this eruption. The duration of climatic impact induced by this particular eruption, if any, is likely less than a few decades.

Keywords

Volcano; Tephra; Eruption; Limnology; Climate; Sulawesi; Indonesia

1. Introduction

Volcanic eruptions can cause large changes in Earth's climate and are an important agent of environmental change. The climatic and environmental impacts of major volcanic eruptions in the last 200 years are well-documented, and include significant global cooling following the eruptions of Mt. Pinatubo (the Philippines, 1991), El Chichon (Mexico, 1982), Mt. Agung (Indonesia, 1963), Santa Maria (Guatemala, 1902), Krakatoa (Indonesia, 1883), and Tambora (Indonesia, 1815) (e.g. McGee et al. 1997, Robock 2000, Trenberth and Dai 2007, Robock et al. 2009). The global climatic impacts of these eruptions occur in part because of their tropical location, which enhances vertical transport of sulfur aerosols into the stratosphere and facilitates aerosol dispersion into both hemispheres. Four of the largest volcanically-induced climate changes occurring in the last two centuries were triggered by eruptions in Southeast Asia, highlighting the potentially disproportionate climate impact of the region's volcanoes.

Despite Southeast Asia's importance, the eruption history and climate impacts of the region's volcanoes during prehistoric times are relatively unknown. More than 99 % of all known eruptions in Indonesia have been dated using historical techniques, emphasizing the need for more studies of the pre-historic volcanic records in this region (Siebert et al. 2010). The largest known volcanic eruption in the Late Quaternary, Sumatra's Younger Toba supereruption at 75 ± 0.9 kyr BP (Mark et al. 2014), has been

the subject of numerous studies, yet the magnitude, duration, and spatial extent of the cooling associated with this supereruption are still widely debated. Estimates range from just a few degrees of cooling to a state of volcanic winter lasting more than a millennium (e.g. Rampino and Self 1992, Oppenheimer 2002, Robock et al. 2009, Williams 2012, Lane et al. 2013). The history of other major caldera-forming eruptions, such as those on the islands of Sulawesi and Sangehe Islands in central Indonesia (Figure 1), have received scant attention. Nevertheless, geomorphological evidence suggests that several large, explosive eruptions had occurred in Sulawesi and the Sangehe Islands during the Late Quaternary (van Padang 1983, Newhall and Dzurisin 1988, Dam et al. 2001, Poedjoprajitno 2012), with potentially important paleoclimatic and geochronologic significance.

Volcanic events can have enormous impacts on terrestrial and aquatic ecosystems, both indirectly, via their effects on climate, as well as through the direct effects of tephra deposition. Terrestrial vegetation can be destroyed by very large eruptions (Whittaker et al. 1992, Gomez-Romero et al. 2006), and tephra inputs into streams, lakes, and rivers may induce relatively short-lived changes in pH, nutrients, water salinity, and chemistry (Barker et al. 2000, Birks and Lotter 1994, Eastwood et al. 2002, Heinrichs et al. 1999). The duration of these effects may be short, but longer-term weathering of tephra deposited on landscapes can serve as an important source of silica and phosphorus, which may produce important changes in algal communities and benthic fauna in aquatic systems (Birks and Lotter 1994, Abella 1988, Telford et al. 2004). Despite the abundance of volcanoes in Indonesia, very few studies have been conducted on the consequences of volcanic activity to terrestrial ecosystems (e.g. Whittaker et al. 1992, Suryanto et al.

2010, Sutomo 2013), and none of these studies have investigated lacustrine ecosystems and their watersheds.

Lake Towuti is a large tectonic lake on the island of Sulawesi that has provided some of the longest continuous environmental records from Indonesia (e.g. Russell et al. 2014, Costa et al. 2015). An ~100 kyr-long sediment core from the lake (Figure 1) contains a thick layer of tephra (4-8 cm), the only visually significant ash in Towuti's sedimentary record. The lake is situated hundreds of kilometers from the nearest tephra sources, implying that this tephra likely represents a regionally if not globally important eruption. Moreover, as the Towuti catchment is composed predominantly of ophiolitic, silica-poor bedrock, this tephra layer provides an interesting case study to analyze the effects of silica-rich (rhyolitic) tephra inputs on lake ecosystems.

Here we utilize magnetostratigraphic techniques to date the tephra and employed geochemical techniques to correlate the Towuti tephra with other tephra beds from the region. We examine multiple climate and environmental proxies in the Towuti sediments before and after tephra deposition to assess potential medium- to long-term impacts of the eruption. In addition to improving our understanding of regional volcanism and its consequences, our analyses of the Towuti tephra could provide a new geochronologic marker to link paleoclimate records in Sulawesi and further afield.

2. Study Site

Lake Towuti (elevation: 319 m asl, area: 561.1 km², maximum water depth: 200 m) is Indonesia's largest tectonic lake. At present the lake is thermally and chemically

stratified, with nearly constant bottom water temperatures of 28.4 °C and anoxic water below 140 m water depth (Costa et al. 2015). The upper 90 m of the water column mixes during the cooler, dry season (Costa et al. 2015), implying thermally-driven density inversions as the dominant mechanism for lake mixing, similar to nearby Lake Matano (Katsev et al. 2010). The anoxic lake bottom is rich in metals such as Fe, Cr, and Mn due to reductive dissolution of metals derived from intense chemical weathering of ultramafic bedrock and lateritic soils in the catchment basin (Haffner et al. 2001). Climate-driven changes in Lake Towuti's thermal stratification and mixing in the past have been shown to affect the lake's biogeochemistry, inducing the precipitation of these metals, with important implications to biota in the lake (Costa et al. 2015).

Lake Towuti is ultraoligotrophic, with low phytoplankton biomass and relatively clear water (Lehmusluoto 1997, Haffner et al. 2001, Sabo et al. 2008). Both total nitrogen and total phosphorous at the surface are low, less than 0.8 ppm and 0.3 ppm respectively (Nasution et al. 2015). Despite the ultraoligotrophic conditions, the lake has exceptional endemism for numerous taxa, including diatoms (Bramburger et al. 2008, von Rintelen et al. 2012). Annual precipitation near Lake Towuti is high (2500-3000 mm) (Russell et al. 2014), supporting the dominant closed canopy C₃ rainforest vegetation in its catchment. However, during the drier and more seasonal Marine Isotope Stage (MIS) 2, C₄ grasses were abundant in the catchment (Russell et al. 2014), suggesting that the history of vegetation surrounding Lake Towuti is closely intertwined with the history of regional precipitation amount.

3. Materials and Methods

Sediment core IDLE-TOW10-3B-1K was collected in 2010 from 56 m water depth using a modified Kullenberg piston corer, and was shipped to LacCore, University of Minnesota, where it was split, described, and logged following the protocol outlined in Schnurrenberger et al. (2003). Core TOW10-3B is one of 11 piston cores recovered during the Indonesian Deep Lakes Expedition (IDLE). During splitting, a 4 cm-long light gray ash (Figure 2) was found 6.67 m below the top of core TOW10-3B. This ash was also found in nearby cores, with thickness ranging from 4-8 cm.

Thus far, we have been unable to directly date this tephra due to low feldspar concentrations, and the tephra is in sediments older than the practical upper limit of ^{14}C dating. Here we used magnetostratigraphic techniques to constrain the age of the tephra and the rates at which the sediments around the tephra were deposited. In particular, we employed magnetic paleointensity variations recorded in the Fe-rich, highly magnetic sediments of TOW10-3B, compared to local and regional magnetic paleointensity data. The relative paleointensity (RPI) signal for TOW10-3B was constructed by normalizing the natural remanent magnetization (NRM) with the laboratory-induced anhysteretic remanent magnetization (ARM). Both the NRM and ARM were measured at an interval of 1 cm in the 150 mT demagnetization state using a 2G-Enterprises Cryogenic Magnetometer at the University of Rhode Island. The RPI was normalized by dividing the values with the maximum RPI value recorded. We were able to correlate variations in RPI in TOW10-3B with those of TOW10-9B, a sediment core located ~9 km southeast from TOW10-3B (Tamuntuan et al. 2015), and to correlate the Towuti RPI records to an

RPI dataset from core PC83 in the South China Sea, which was dated using radiocarbon and matching marine $\delta^{18}\text{O}$ stratigraphy (Xiaoqiang et al. 2009). Previous work has shown excellent agreement between RPI data from radiocarbon-dated cores from Lake Towuti and PC 83 (Kirana et al. in prep).

We sampled the tephra at 1 cm, 2 cm, and 3 cm from its top and measured major element concentrations of individual tephra shards via wavelength dispersive spectrometry using a Cameca SX-100 electron microprobe analyzer (EPMA), following the procedure described in Gatti et al. (2014). We analyzed 11-17 glass shards per sample. The instrument was calibrated for each set of beam conditions using a suite of mineral standards and verified using additional secondary standards. To allow for universal comparisons, shard analyses were normalized to a water-free basis, i.e. recalculated to sum of 100%.

We sampled every 1 cm of the sediments in the 10 cm below and above the ash, with additional lower-resolution (~16 cm) samples spread along the length of the core. Samples were freeze-dried and homogenized prior to geochemical analyses. We measured total carbon and total nitrogen (TN) concentrations using a CE Instruments Model NC2100 Elemental Analyzer, and obtained total organic carbon (TOC) concentrations by subtracting total inorganic carbon, measured using a UIC, Inc. CM240 TIC Analyzer paired with the CM5240 automated sample acidification apparatus, from total carbon. Ratios of atomic carbon to nitrogen (C/N) and the carbon isotopic composition of organic matter ($\delta^{13}\text{C}_{\text{OM}}$) were measured using a Thermo DeltaVPlus MS coupled to an Agilent 6890 GC. These analyses provide insight into changes in primary productivity and organic matter sources through time. We measured the carbon isotopic

composition of long chain fatty acids (leaf waxes, hereafter $\delta^{13}\text{C}_{\text{wax}}$) following the methods described in Wicaksono et al. (2015). $\delta^{13}\text{C}_{\text{wax}}$ reflects the relative abundance of terrestrial plants using the C_3 and C_4 photosynthetic pathways, and has been shown to be a sensitive indicator of vegetation and climate in central Sulawesi (Russell et al. 2014). Major and trace element concentrations were measured by digesting sediments via flux fusion (Murray et al. 2000), followed by elemental analysis using an inductively coupled plasma atomic emission spectroscopy (ICP-AES), as described in Russell et al. (2014). These data complement X-ray fluorescence (XRF) core scanning conducted with an ITRAX XRF core scanner equipped with a Cr tube at 0.2 cm interval for comparison to ICP determinations.

4. Results

The age model for TOW10-3B is constructed primarily using four tie points with PC83 paleointensity curves indicating low RPI values at ~8 kyr BP, ~41 kyr BP (i.e. Laschamp excursion), ~64 kyr BP, and ~97 kyr BP (Figure 3). Assuming a linear sedimentation rate between the latter two tie points, the age of the Towuti tephra is estimated to be ~74 kyr BP, and the sedimentation rate pre- and post-tephra is ~190 years/cm. Additional tie points were obtained by directly comparing magnetic susceptibility data for TOW10-3B to a radiocarbon-dated magnetic susceptibility stratigraphy from TOW10-9B (Figure S1).

The major and trace element composition of Towuti tephra shards are shown in Table 1. The average total prior to normalization was 95.3 %, indicating magmatic water

present in the tephra. After normalizing to 100% for all elements, the ~74.6% SiO₂ composition implies that the Towuti tephra is high-silica rhyolite (73.0-80.8 % SiO₂, Ku et al. 2009). Variations in the major element concentrations are within analytical error in the top, middle, and bottom sections of the tephra (Table 1) and indicate the tephra is substantially enriched in Ca, Si, Al, Ti, and K, and much lower in Fe, than surround sediment (Figure 6).

TOC, TN, C/N, $\delta^{13}\text{C}_{\text{OM}}$, and $\delta^{13}\text{C}_{\text{wax}}$ data do not change dramatically in sediments deposited before and after the eruption (Figure 4), especially in comparison to the amplitude of changes observed over the entire core (Figure 5). The largest and most consistent changes in these data were observed during MIS 2, when, for example, both $\delta^{13}\text{C}_{\text{OM}}$ and $\delta^{13}\text{C}_{\text{wax}}$ became much more enriched (Figure 5). No similar changes in $\delta^{13}\text{C}_{\text{OM}}$ and $\delta^{13}\text{C}_{\text{wax}}$ were observed around the time of the eruption (Figure 5). Two-tailed student's t-tests demonstrate statistically significant changes in the mean values of TOC, TN, C/N, and $\delta^{13}\text{C}_{\text{OM}}$ pre- and post-eruption, (Table 2), marked by a decrease in TOC and TN, an increase in C/N, and a small, 0.2 ‰ depletion in $\delta^{13}\text{C}_{\text{OM}}$.

Similar t-tests performed on low-resolution and high-resolution elemental concentration data (Na, P, Ca, Si, Al, Ti, K, and Fe) also demonstrated statistically significant differences in pre- and post-tephra data (Table 2, Figure 6). However, comparison of these proxy data with long-term data from the same core indicates that these changes are small relative to the variations observed during MIS 2, especially for Ti, K, and Fe (Figure 7).

5. Discussion

Tephra geochemistry

Comparison of the chemistry of the tephra preserved in Towuti core TOW10-3B with geochemical data from tephra layers preserved in marine sediment cores from the Celebes Sea, north of Sulawesi (Table 3, Figure 1), indicates strong similarity between our tephra and a 10-cm thick tephra found at 660 cm depth in MD01-2387 (Ku et al. 2009), a 5-cm thick tephra at 368 cm depth in ODP 767 A (Pouclet et al. 1999), and a 4-cm thick tephra at 376 cm depth in ODP 767 B (Pouclet et al. 1999). The glass shards in the Celebes Sea cores have high $\text{Na}_2\text{O}+\text{K}_2\text{O}$ (> 7.1 wt %) and high SiO_2 (> 74 wt %), similar to the Towuti tephra (Table 3). Using the similarity coefficient calculation after Borchardt et al. (1972), we found high similarity coefficients between the Towuti tephra and the tephras from MD01-2387, ODP 767 A, and ODP 767 B (0.86, 0.87, and 0.84 respectively), strongly suggesting that these tephras are derived from the same eruption (Table 3). The similarities between Towuti tephra and the paired MD01-2387 and ODP 767 tephra layers are also evident in binary plots oxide elements against SiO_2 . High matches are found for most elements, e.g. TiO_2 , Al_2O_3 , MnO , MgO , and Na_2O (Figure 8). The Towuti tephra has slightly different values of FeO , CaO , and K_2O , but these differences are within the analytical error and could be related to small inter-laboratory differences in analytical calibration and protocols (Hunt and Hill 1996, Lowe et al. 2011) and/or the possibility of distinct fingerprints at different stages of an eruption event or multiple closely-spaced eruptions, which can affect ash-fall deposits depending on wind

directions (Davies et al. 2010). The catchment of Lake Towuti provides ultrabasic and ferruginous metal substrates that could potentially inflate the concentration of FeO in the measured Towuti tephra relative to marine tephtras through mixing or clay authigenesis, although no discernible physical mixing or chemical alteration of the tephra and sediments is observed (Figure 2). Ku et al. (2009) estimated the age of this particular pair to ca. 72 kyr BP assuming a linear sedimentation rate between the present-day core-top and the measured age of the first foraminiferal occurrence in ODP 767, thus placing its age within the magnetostratigraphic age constraints provided for the Towuti tephra. Based on these lines of evidence, we argue that the two tephtras are likely derived from the same source.

The similarity coefficient between the Towuti tephra and the Younger Toba Tephra (YTT), computed using the compiled EPMA data for Younger Toba tephtras (Gatti et al. 2014) is much lower (0.58), therefore rejecting Toba as the tephra's source. The ashes from the Toba eruption were likely blown primarily to the west (Williams 2012, Costa et al. 2014), apparently with little tephra deposition as far east as Sulawesi. To our knowledge, east of Sumatra and Peninsular Malaysia, the YTT has thus far been discovered only in the southern section of the South China Sea (Buhring and Sarthein 2000). Our finding of a major regional tephra of similar age to the YTT urges caution in attributing tephtras deposited around this time without a full examination of shard chemistry.

Similar to the Towuti tephra, the ~72 ka tephra in the Celebes Sea cores is the thickest tephra in the sediment record of the past 100 kyr, suggesting this was a very large regional eruption. Apart from the Younger Toba supereruption, no major explosive

eruptions of the Sunda Arc volcanoes between 70 and 80 kyr BP are known. Comparison of major elemental compositions and Sr isotopic ratios from the Celebes Sea tephra with those of regional Late Quaternary volcanic rocks suggests that the tephra likely originated from the volcanic province comprised of Mindanao Island, Sangihe Islands, Halmahera Island, or northeastern Sulawesi (Ku et al. 2009, Figure 1). It is difficult to conclusively eliminate these various sources, yet northeastern Sulawesi emerges as the most likely candidate. Despite differences in the deposition and preservation modes of tephtras in lacustrine and marine systems, both the Towuti tephra and Celebes Sea tephra have similar thicknesses (4-8 cm in Towuti and 4-10 cm in Celebes, suggesting that the volcano producing the tephra layer is likely situated in between the two sites, i.e. in northeastern Sulawesi. Within North Sulawesi, potential candidates include several caldera-forming volcanoes (Pangalombian, Sempu, and Tondano) and stratovolcanoes (e.g. Soputan, Lokon, Mahawu, Klabat, Karangetang, Tampusu, Lengkuhan, Kasuratan), some of which are still active. These volcanoes are roughly ~ 400 km from the Celebes Sea sites and at ~600 km from Lake Towuti (Figure 1), implying that the tephra layers must have been produced by a massive eruption. As a basis for comparison, the farthest 5-cm isopach of the phoenix cloud ashfall of the 1815 Tambora eruption (Volcanic Explosivity Index = 7) includes southeastern Kalimantan and the southwestern tip of Sulawesi, ~600 km from Mount Tambora (Oppenheimer 2003), suggesting that the purported northeastern Sulawesi eruption was at least as explosive as that of Tambora.

Ecological impacts of tephra emplacement

The emplacement of a 4-8-cm thick Towuti tephra could have directly impacted Lake Towuti's biogeochemistry and ecology. Increased diatom concentrations and/or shifts in diatom assemblages have been observed elsewhere after in-lake deposition of tephra (e.g. Harper et al. 1986, Barker et al. 2000, Telford et al. 2004) due to tephra dissolution and leaching, which increases the availability of Si and/or P in the water column (Harper et al. 1986, Eastwood et al. 2002, Telford et al. 2004), providing competitive advantage for certain diatom taxa (Harper et al. 1986, Barker et al. 2000, Eastwood et al. 2002). While the fertilization potential of volcanic ash in a lacustrine environment is dependent on the type of eruption, particulate size, deposition rates and environmental buffering capacities, an overall increase in lacustrine productivity has often been linked with the emplacement of tephra (e.g. Harper et al. 1986, Telford et al. 2004, Jones and Gislason 2008). Lake Towuti is an interesting new system in which to investigate these changes, as the ultraoligotrophic environment contains a unique and highly endemic diatom flora (Rintelen et al. 2011), and the ophiolitic bedrock results in low dissolved silica concentrations.

The tephra is massive with relatively sharp basal and upper contact (Figure 2), and the concentrations of Na, Ca, and K, which are all abundant in the tephra, do not increase in sediments above the tephra (Figures 4 and 6, Table 2), indicating very limited physical reworking of tephra shards into overlying sediments. Thus, any chemical changes in the overlying sediments, and aquatic ecosystem responses to tephra deposition are likely caused by either the chemical release of Si, P, and other elements from

weathering of tephra deposited on land, or internal processes, including leaching of Si, P, and other elements into the water column and alteration of pore-water chemistry in Towuti's sediments.

Volcanic eruptions have been shown to fertilize primary production in many lakes, and given Towuti's ultramafic and Si- and P-poor chemistry, one might expect similar changes to occur here. The values of TOC and TN in the sediments near the tephra were among the lowest throughout the whole record (Figure 5), and interestingly, TOC and TN became significantly lower after the eruption (Table 2). C/N increases and becomes more variable after the eruption, although the post-eruption variability is within the range of variability observed in the ~2 m above and below the tephra (Figure 5). TOC in lake sediments may be derived from terrestrial organic matter that has been transported into the lake or from autochthonous production (e.g. Hu et al. 2003, Bechtel et al. 2007). Both TOC and TN are low in TOW10-3B (Figure 5), highlighting Towuti's ultraoligotrophic state, and the values of C/N range from 14 to 32 (Figure 5), suggesting the dominance of both terrestrially-sourced and aquatic-sourced organic matter at different times in the past. In general, terrestrial vascular plants exhibit C/N values higher than 20 due to their low protein content and abundance of cellulose (Meyers 1994, Talbot and Laerdal 2000), while C/N ratios below 10 are indicative of protein-rich, lignin poor organic matter sourced from phytoplankton. Meanwhile, intermediate C/N values between 10 and 20 often indicate organic matter of a mixed terrestrial and phytoplankton source (Meyers 1994). However, in lakes with N-limited productivity, elevated C/N ratios (~15-20) are common, reflecting nutrient-poor condition and organic matter predominantly of aquatic origin (Talbot and Laerdal 2000, Olsen et al. 2013). Further,

Lake Towuti $\delta^{13}\text{C}_{\text{OM}}$ values are quite depleted, ranging from -28 to -32 ‰, and there is no major change in the carbon isotope of organic matter ($\delta^{13}\text{C}_{\text{OM}}$) pre- and post-eruption (Figure 4). In particular, we do not observe any clear shift toward enriched compositions that might record increase primary production.

The increased variability of C/N above the tephra, with generally higher C/N ratios, coupled with lower TOC and TN, suggests a reduction in lacustrine biomass relative to terrestrial sources, contrary to the effects observed in other lakes. It is possible that in an N-limited environment such as Lake Towuti, the chemical release of Si and P from tephra do not stimulate strong increases in primary production. Alternatively, rapid chemical weathering in the hot, humid environment of central Sulawesi could quickly remove nutrients from volcanic ash, such any changes in primary productivity in the lake were too short to be captured by our cm-scale sampling resolution.

While the carbon cycle appears not to have been strongly affected by tephra deposition, the concentrations of Si and P within the Towuti tephra are higher relative to surrounding sediments (Figure 6). Interestingly, while P increases after tephra deposition (Table 2), Si declines. This is somewhat surprising in light of elevated Si concentrations in the tephra and clear evidence from some lakes for increased diatom productivity following tephra deposition (Harper et al. 1986, Telford et al. 2004). That said, although the changes in Si are significant, they are very small, with a reduction in average Si of only 0.34%. The increase in P concentration above the tephra (Figure 4) suggests that Towuti tephra emplacement did provide fresh P to the lake and might have induced changes to phytoplankton communities that lasted several centuries after the eruption. While we see little evidence for increased net primary productivity, interspecies

competition for nutrients determines the species composition (Tilman and Kilham 1976), and different Si:P ratios can determine the types of diatom assemblage that may thrive in a lake (Saros et al. 2005).

Microscopic examination of pre- and post-tephra sediments indicates that partially dissolved diatoms are common in the two centimeters below the tephra, whereas well-preserved diatoms from various taxa (e.g. *Gomphonema wolterecki*, *Caloneis silicula s.l.*, *Diploneis sp.*, and *Surirella sp.*) are found in the first two centimeters above the tephra, despite the decline in Si concentrations after tephra deposition. Meaningful analyses of possible assemblage shift in Lake Towuti are precluded by the poor preservation of diatom frustules in pre-tephra sediments. The presence of diatom concentrations above the tephra likely results from an increase in silica preservation due to leaching of Si from the tephra into pore water, despite the lower overall Si concentrations. The overall decline in Si concentrations after tephra deposition thus merits additional attention, as it suggests a net decrease in the benthic silica flux for unknown reasons. Regardless, while our analyses indicate that the presence of tephra might strongly improve diatom preservation, we observe very little change in primary productivity or diatom productivity in response to tephra deposition.

Climatic impacts

The recent major eruptions of Krakatau, Tambora, and Pinatubo produced both significant amount of ash and sulfate aerosols, causing global, albeit relatively short-lived, cooling events (McGee et al. 1997, Robock 2000, Robock et al. 2009). The largest

event among the three, the 1815 Tambora eruption, produced ~ 110 Tg of sulfate aerosols (Self et al. 2004) and induced a global cooling that lasted for three years (Briffa et al. 1998). It is difficult to determine the amount of sulfate aerosols produced by older eruptions; the volume of sulfate aerosol released by the Younger Toba supereruption is widely debated (Rampino and Self 1982, McGee et al. 1997). Several studies estimated large amounts of aerosols (>1000 Tg) that might have induced global cooling for a millennium (e.g. Rampino and Self 1992, Zielinski et al. 1996, Robock et al. 2009), yet recent melt inclusion studies suggest that sulfate aerosols ejected by the YTT eruption were only ~ 100 Tg, less than Tambora (Chesner and Luhr 2010). In addition to causing temperature changes, major volcanic eruptions in the last millennium have been shown to induce droughts in some monsoon regions for up to five years, likely related to the stronger cooling over land than over the ocean, which weakens monsoon circulation (Trenberth and Dai 2007, Robock et al. 2009, Zhuo et al. 2014). However, short-term volcanic forcing may also potentially be transformed into long-term climatic changes by dynamical ice-ocean feedbacks, in particular via the rearrangement of ocean circulation in the North Atlantic following a large eruption, which could weaken the Atlantic Meridional Overturning Circulation and lead to a prolonged global cooling that lasts for several decades and centuries (Zhong et al. 2011, Miller et al. 2012, Swingedou 2015). While we cannot know the amount of sulfate released during the eruption that fed ash to Lake Towuti, the notable thickness of the tephra in both Lake Towuti and the Celebes Sea suggest it was the largest in the region over the past 100 kyr BP.

Previous work has shown that elemental tracers of terrigenous sediment inputs, redox sensitive metals, and the carbon isotope composition of terrestrial leaf waxes are all

sensitive indicators of climate in Lake Towuti. During MIS 2, the concentration of Ti and other elemental tracers of rainfall-driven erosion all decline strongly, reflecting reduced precipitation and runoff (Russell et al. 2014). The concentrations of Fe, a redox-sensitive element, increase significantly (Figure 7) indicating precipitation of iron oxides due to increased lake bottom oxygenation and more vigorous water-column mixing as the lake cooled (Costa et al. 2015). $\delta^{13}\text{C}_{\text{wax}}$ became much more enriched, reflecting a transition from the present-day closed-canopy rainforest to more open vegetation with abundant C₄ grasses (Russell et al. 2014).

Examining these proxies in pre- and post-tephra sediment, the concentrations of Fe in post-tephra sediment became lower relative to the pre-tephra values (Table 2, Figure 3), suggesting that the eruption did not trigger long-term changes in lake mixing linked to surface cooling. Similarly, we do not observe coherent reductions in the concentrations of detrital elements, such as Al, Ti, and K, due to reduced precipitation and weathering in the Lake Towuti catchment (Figure 7). While Al, Ti, and K concentrations pre- and post-tephra were statistically different, their concentrations are consistent with trends that had already started a few hundred years prior to the eruption. Drying following a major eruption might increase the proportion of C₄ grasses in the landscape because C₄ taxa are better suited to drier environments (Vogts et al. 2009). Further, grassland might be expected from observations of plant community response to deep tephra cover (more than 15 cm) as grasses are early colonizing taxa (Whittaker et al. 1992, Gomez-Romero et al. 2006). Yet we did not observe shifts in the $\delta^{13}\text{C}_{\text{wax}}$ data post-tephra deposition (Figure 4), suggesting that no significant expansion of C₄ grasses due to

a major drying event or a massive destruction of vegetation occurred around Towuti after the eruption.

Together, our data strongly indicate that no medium- to long-term terrestrial ecological and environmental changes in central Sulawesi nor regional drying or cooling resulted from a major volcanic eruption in north Sulawesi ca. 74 kyr BP. Moreover, despite the ongoing debate regarding the timing of the YTT eruption, its latest age estimate, ca. 75 ± 0.9 kyr BP (Mark et al. 2014), suggests that evidence for the presence of the Toba tephra and regional climatic ramifications from the Toba supereruption should be observed within our high-resolution sampling interval below the Towuti tephra. We did not observe any glass shards or physical changes in the sediment indicative of the Toba tephra, nor did we detect any rapid shifts in the paleoclimatic proxy data, corroborating the conclusion of recent limnological studies about the limited spatiotemporal scope and magnitude of the Younger Toba supereruption climate impacts (e.g. Lane et al. 2013, Jackson et al. 2015). Thus, any possible regional climatic changes caused by the Toba supereruption and the more local northern Sulawesi eruption were probably short-lived, below the centennial resolution of our Lake Towuti record. Further investigation of the climatic eruption to the north Sulawesi eruption and the Toba supereruption should utilize higher (annual or decadal) resolution paleoenvironmental data.

6. Conclusions

The 4-8 cm-long white ash in the sedimentary record of Lake Towuti, dated using magnetostratigraphic techniques to ca. 74 kyr BP, provides a new marker horizon linking paleoarchives in Sulawesi and its vicinity. The tephra is geochemically similar to tephra layers found in the Celebes Sea sedimentary records, suggesting a volcanic source in northeastern Sulawesi. An expanded regional tephra data set are needed to better characterize this particular eruption, which could have been the largest in central Indonesia in the past 100 kyr BP. Increased P concentration and better diatom preservation immediately post-tephra deposition indicate that emplacement of this tephra directly perturbed the ecology and biochemistry within Lake Towuti, likely by changing nutrient availability. However, pre- and post-tephra elemental and leaf wax isotope data strongly suggest that no medium- to long-term terrestrial ecological changes in central Sulawesi nor regional drying and/or cooling had resulted from this eruption. Our study suggests that the typical duration of cooling induced by major tropical eruptions, including the Younger Toba supereruption, is less than a few decades.

References

Abella, S.E.B. (1988), The effect of the Mt. Mazama ashfall on the planktonic diatom community of Lake Washington, *Limnology and Oceanography*, 33, 6, 1376–1385.

- Barker, P., R. Telford, O. Merdaci, D. Williamson, M. Taieb, A. Vincens, E. Gibert (2000), The sensitivity of a Tanzanian crater lake to catastrophic tephra input and four millennia of climate change, *The Holocene*, 10, 303–310.
- Bechtel, A., M. Woszczyk, D. Reischenbacher, R.F. Sachsenhofer, R. Gratzner, W. Püttmann and W. Spychalski (2007), Biomarkers and geochemical indicators of Holocene environmental changes in coastal Lake Sarbsko (Poland). *Org. Geochem.*, 38, 1112-1131.
- Birks, H. J. B. and A. F. Lotter (1994), The impact of the Laacher See Volcano (11 000 yr B.P.) on terrestrial vegetation and diatoms. *J. Palim.*, 11, 313-322.
- Borchardt, G.A., P.J. Aruscavage, and H.T. Millard (1972), Correlation of the Bishop ash, a Pleistocene marker bed, using instrumental neutron activation analysis. *Journal of Sedimentary Petrology*, 42, 301–306.
- Bramburger, A.J., P.B. Hamilton, P.E. Hehanussa, G.D. Haffner (2008), Processes regulating the community composition and relative abundance of taxa in the diatom communities of the Malili Lakes, Sulawesi Island, Indonesia. *Hydrobiologia*, 615, 215-24.
- Briffa, K. R., P. D. Jones, F.H. Schweingruber, and T.J. Osborn, T. J (1988), Influence of volcanic eruptions on Northern Hemisphere summer temperature over the past 600 years, *Nature*, 393, 450–454.
- Chesner, C.A. and J.F. Luhr (2010) A melt inclusion study of the Toba Tuffs, Sumatra, Indonesia. *Journal of Volcanology and Geothermal Research* 197, 259-278.

- Costa, A., V.C. Smith, G. Macedonio, M. Naomi et al. (2014), The magnitude and impact of the Youngest Toba Tuff super-eruption, *Frontiers in Earth Science* 2, 16.
<http://dx.doi.org/10.3389/feart.2014.00016>.
- Costa, K., J.M. Russell, S. Bijaksana, H. Vogel (2015), Hydrological connectivity and mixing of Lake Towuti, Indonesia, in response to paleoclimatic changes of the past 60,000 years, *Palaeogeogr Palaeoclimatol Palaeoecol*, 417, 467–475.
- Dam, R., J. Fluin, P. Suparan, and S. van der Kaars (2001), Palaeoenvironmental developments in the Lake Tondano area (N. Sulawesi, Indonesia) since 33,000 yr BP. *Palaeogeography, Palaeoclimatology, Palaeoecology* 171, 147–183.
- Davies, S.M., G. Larsen, S. Wastegård, C.S.M. Turney, V.A. Hall, L. Coyle, T. Thordarson (2010), Widespread dispersal of Icelandic tephra: how does the Eyjafjöll eruption of 2010 compare to past Icelandic events? *Journal of Quaternary Science*, 25, 605–611.
- Eastwood, W.J., J. Tibby, N. Roberts, H.J.B. Birks, H.F. Lamb (2002), The environmental impact of the Minoan eruption of Santorini (Thera): statistical analysis of palaeoecological data from Golhisar, southwest Turkey, *The Holocene* 12, 431–444.
- Gatti, E., I.M. Villa, H. Achyuthan, P.L. Gibbard, C. Oppenheimer (2014), Geochemical variability in distal and proximal glass from the Youngest Toba Tuff eruption. *Bulletin of Volcanology* 76: 1–16.
- Gomez-Romero, M. et al. (2006), Effect of tephra depth on vegetation development in areas affected by volcanism, *Plant Ecol.* 183, 207–213.

- Haffner, G.D., P.E. Hehanussa, D. Hartoto (2001), The biology and physical processes of large lakes of Indonesia: lakes Matano and Towuti. In: Munawar M, Hecky RE, eds, *The great lakes of the world: food-web health and integrity*. Leiden: Backhuys, pp. 183–192.
- Harper, M. A., R. Howorth, M. Mcleod (1986), Late Holocene diatoms in Lake Poukawa: Effects of airfall tephra and changes in depth, *New Zealand Journal of Marine and Freshwater Research*, 20(1), 107-118, doi: 10.1080/00288330.1986.9516135.
- Heinrichs, M.L., I.R. Walker, R.W. Mathewes, and R.J. Hebda (1999), Holocene chironomid-inferred salinity and paleovegetation reconstruction from Kilpoola Lake, British Columbia, *Geographie Physique et Quaternaire*, 53, 211–221
- Hu, F.S., D. Kaufman, S. Yoneji, D. Nelson, A. Shemesh, Y.S. Huang, J. Tian, G. Bond, B. Clegg, and T. Brown (2003), Cyclic variation and solar forcing of Holocene climate in the Alaskan subarctic, *Science* 301, 1890-1893.
- Hunt, J.B., P.G. Hill (1996), An inter-laboratory comparison of the electron probe microanalysis of glass geochemistry. *Quaternary International*, 34–36, 229–241.
- Jackson, L.J., J.R. Stone, A. S. Cohen, and C. L. Yost (2015), High-resolution paleoecological records from Lake Malawi show no significant cooling associated with the Mount Toba supereruption at ca. 75 ka, *Geology*, G36917.1, doi:10.1130/G36917.1.
- Jones, M.T. and S.R. Gislason (2008), Rapid releases of metal salts and nutrients following the deposition of volcanic ash into aqueous environments. *Geochim. Cosmochim. Ac.*, 72, 3661–3680.
- Katsev, S., S.A. Crowe, A. Mucci, B. Sundby, S. Nomosatryo, G.D. Haffner, D.A. Fowle

- (2010), Mixing and its effects on biogeochemistry in the persistently stratified, deep, tropical Lake Matano, Indonesia. *Limnol. Oceanogr.*, 55, 763–776.
- Ku, Y.-P., C.-H. Chen, S.-R. Song, Y. Iizuka., J.J.-S. Shen (2009), Late Quaternary explosive volcanic activities of the Mindanao–Molucca Sea Collision Zone in the western Pacific as inferred from marine tephrostratigraphy in the Celebes Sea, *Terrestrial, Atmospheric and Oceanic Sciences* 20, 4, 587–605. doi:10.3319/TAO.2008.07.30.01.
- Lane, C.S., B.T. Chorn, T.C. Johnson (2013), Ash from the Toba supereruption in Lake Malawi shows no volcanic winter in East Africa at 75 ka, *Proc. Natl. Acad. Sci. U. S. A.*, 110, 8025–8029.
- Lehmusluoto, P. (1997), National inventory of the major lakes and reservoirs in Indonesia. Bandung, Indonesia: Research Institute for Water Resources Development, Ministry of Public Works, Agency for Research and Development.
- Lowe, D.J. (2011), Tephrochronology and its application: A review, *Quaternary Geochronology*, 6(2), 107-153.
- Mark, D.F., M. Petraglia, V.C. Smith, L.E. Morgan, D.N. Barfod, B.S. Ellis, N.J. Pearce, J.N. Pal and R. Korisettar (2014), A high-precision $^{40}\text{Ar}/^{39}\text{Ar}$ age for the Young Toba Tuff and dating of ultra-distal tephra: Forcing of Quaternary climate and implications for hominin occupation of India. *Quat Geochronol.*, 10.1016/j.quageo.2012.12.004.
- McGee, K.A., M.P. Doukas, R. Kessler, and T. Gerlach (1997), Impacts of volcanic gases on climate, the environment, and people, *U.S. Geological Survey Open-File Report* 97-262, 2 p.

- Meyers, P.A. (1994), Preservation of elemental and isotopic source identification of sedimentary organic matter, *Chemical Geology*, 144, 289–302.
- Miller, G. H., et al. (2012), Abrupt onset of the Little Ice Age triggered by volcanism and sustained by sea-ice/ocean feedbacks, *Geophys. Res. Lett.*, 39, L02708, doi:10.1029/2011GL050168.
- Murray, R., D.J. Miller, K. Kryc (2000), Analysis of major and trace elements in rocks, sediments, and interstitial waters by inductively coupled plasma–atomic emission spectrometry (ICP–AES). *ODP Tech. Note* 29, 1–27.
- Nasution S. H., S. Sulastri, Z.A. Muchlisin (2015), Habitat characteristics of Lake Towuti, South Sulawesi, Indonesia - the home of endemic fishes. *AACL Bioflux* 8(2), 213-223.
- Newhall, C.G., and D. Dzurisin (1988), Historical unrest at large calderas of the world, *U.S. Geological Survey Bulletin*, 1855-1108.
- Olsen, J., N.C. Anderson, and M. J. Leng (2013). Limnological controls on stable isotope records of Late Holocene palaeoenvironmental change in SW Greenland: a paired lake study. *Quaternary Science Review* 88, 85-95
- Oppenheimer, C. (2002), Limited global change due to the largest known Quaternary eruption, Toba ~74kyr BP? *Quat. Sci. Rev.* 81, 1593–1609.
- Oppenheimer, C. (2003), Climatic, environmental and human consequences of the largest known historic eruption: Tambora volcano (Indonesia) 1815. *Progress in Physical Geography* 27 (2), 230–259.

- Poedjoprajitno, S (2012), Morphostructure Control Towards the Development of Mahawu Volcanic Complex, North Sulawesi. *Indonesian Journal of Geology*, 7, 39-54
- Pouclet, A., M. Pubellier, P. Spadea (1991), Volcanic ash from Celebes and Sulu Sea basins off the Philippines (Leg 124): Petrography and geochemistry. In: Silver, E. A., C. Rangin, M. T. von Breyman, and Leg 124 Scientific Party (Eds.), *Proceedings of the Ocean Drilling Program, Scientific Results*, 124, 467-487.
- Rampino, M.R. and S. Self (1992), Volcanic winter and accelerated glaciation following the Toba super-eruption. *Nature* 359, 50–52.
- Rintelen, T.v., K.v. Rintelen, M. Glaubrecht, C. Schubart, and F. Herder (2011), Aquatic biodiversity hotspots in Wallacea – the species flocks in the ancient lakes of Sulawesi, Indonesia. In Gower, D.J., Johnson, K.G., Richardson, J.E., Rosen, B.R., Rüber, L., and Williams, S.T. (Eds.), *Biotic Evolution and Environmental Change in Southeast Asia*: Cambridge (Cambridge University Press).
- Robock, A. (2000), Volcanic eruptions and climate. *Rev. Geophys.*, 38, 191-219.
- Robock, A., C.M. Ammann, L. Oman, D. Shindell, S. Levis, and G. Stenchikov (2009), Did the Toba volcanic eruption of ~74k BP produce widespread glaciation? *J. Geophys. Res.*, 114, D10107, doi:10.1029/2008JD011652.
- Russell, J.M., H. Vogel, B.L. Konecky, S. Bijaksana, Y. Huang, M. Melles, N. Wattus, K. Costa, and J.W. King (2014), Glacial forcing of central Indonesian hydroclimate since 60,000 y B.P., *Proc Nat Ac Sci USA*, doi:10.1073/pnas.1402373111.

- Sabo, E., D. Roy, P.B. Hamilton, P.E. Hehanussa, R. McNeely, G.D. Haffner (2008), The plankton community of Lake Matano: factors regulating plankton composition and relative abundance in an ancient, tropical lake of Indonesia. *Hydrobiologia*, 615, 225-35.
- Saros, J. E., T. J. Michel, S. J. Interlandi, and A. P Wolfe (2005), Resource requirements of *Asterionella formosa* and *Fragilaria crotonensis* in oligotrophic alpine lakes: implications for recent phytoplankton community reorganizations. *Can. J. Fish. Aquat. Sci.*, 62, 1681–1689.
- Schnurrenberger, D.W, J.M. Russell, K.R. Kelts (2003), Classification of lacustrine sediments based on sedimentary components. *Journal of Paleolimnology* 29: 141-154.
- Self, S., R. Gertisser, T. Thordarson, M.R. Rampino, J.A. Wolff (2004), Magma volume, volatile emissions, and stratospheric aerosols from the 1815 eruption of Tambora. *Geophys. Res. Lett.* 31, L20608, doi:1029/2004GL020925.
- Siebert, L., T. Simkin, P. Kimberly (2010), *Volcanoes of the World* (3rd edition). Berkeley: Univ California Press, 551 p.
- Suryanto, P., M.H. Zaki, M.A. Azani, and M. Azmy (2010), Species diversity of Gunung Merapi National Park, Java, Indonesia following 2006 eruption. *Res. Environ. Life Sci. Research in Environmental and Life Science*, 3 (1), 1-6.
- Sutomo, T and D. Fardilla (2013), Floristic Composition of Groundcover Vegetation after the 2010 Pyroclastic Fire on Mount Merapi. *Jurnal Manajemen Hutan Tropika*, 19, 54-62.

- Swingedouw, D., P. Ortega, J. Mignot, E. Guilyardi, V. Masson-Delmotte, P.G. Butler, M. Khodri, R. S  f  rian (2015), Bidecadal North Atlantic ocean circulation variability controlled by timing of volcanic eruptions. *Nature Communications*, 6, 6545, doi:10.1038/ncomms7545.
- Talbot, M.R. and T. Laerdal (2000), The late Pleistocene-Holocene paleoclimatology of Lake Victoria, East Africa, baed upon elemental and isotopic analyses of sedimentary organic matter. *J. Paleoclimat*, 23, 141-164.
- Tamuntuan, G., S. Bijaksana, J. King, J. Russell, U. Fauzi, K. Maryunani, N. Aufa, L.O. Safiuddin (2015), Variation of magnetic properties in sediments from lake Towuti, Indonesia, and its paleoclimatic significance. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 420,163–172.
- Telford, R.J., P. Barker, S. Metcalf, A. Newton (2004), Lacustrine responses to tephra deposition: examples from Mexico. *Quaternary Science Reviews*, 23, 2337–2353.
- Tilman, D. and S. S. Kilham (1976), Phosphate and silicate growth and uptake kinetics of the diatoms *Asterionella formosa* and *Cyclotella meneghiniana* in batch and semicontinuous culture. *Journal of Phycology*, 12, 375–383.
- Trenberth, K.E. and A. Dai (2007), Effects of Mount Pinatubo volcanic eruption on the hydrological cycle as an analog of geoengineering. *Geophys Res Lett* 34, L15702. doi: 10.1029/2007GL030524
- van Padang, N. (1983), History of volcanology in the East Indies *Scripta Geol.* 71.
- Vogts, A., H. Moossen, F. Rommerskirchen, J. Rullk  tter (2009), Distribution patterns and stable carbon isotopic composition of alkanes and alkan-1-ols from plant waxes of African rain forest and savanna C₃ species. *Organic Geochemistry* 40,

1037–1054.

Whittaker, R.J., J. Walden and J. Hill (1992), Post- 1883 ash fall on Panjang and Sertung and its ecological impact, *Geojournal*, 28(2), 153–71.

Whittaker, R.J., M.B. Bush and K. Richards (1989), Plant recolonization and vegetation succession on the Krakatau Islands, Indonesia, *Ecological Monographs*, 59, 59-123.

Wicaksono, S.A., J.M. Russell, and S. Bijaksana (2015), Compound-specific stable isotope records of vegetation and hydrologic change in central Sulawesi since 53,000 yr BP. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 430, 47-56, doi:10.1016/j.palaeo.2015.04.016

Williams, M. (2012), Did the 73 ka Toba super-eruption have an enduring effect? Insights from genetics, prehistoric archaeology, pollen analysis, stable isotope geochemistry, geomorphology, ice cores, and climate models, *Quaternary International* 269:87–93.

Xiaoqiang, Y., F. Heller, N. Y. Wu, J. Yang, and Z. H. Su (2009), Geomagnetic paleointensity dating of South China Sea sediments for the last 130 kyr, *Earth Planet. Sci. Lett.*, 284, 258–266.

Zhong, Y., G. H. Miller, B. Otto-Bliesner, M. Holland, D. Bailey, D. Schneider, and A. Geirsdottir (2010), Centennial-scale climate change from decadal- paced explosive volcanism: A coupled sea ice-ocean mechanism. *Clim. Dyn.*, 37, 1–15, doi:10.1007/s00382-010-0967-z.

Zhuo, Z., C. Gao, Y. Pan (2014), Proxy evidence for China's monsoon precipitation response to volcanic aerosols over the past seven centuries, *J. Geophys. Res.* 119, 10.1002/2013JD021061.

Zielinski, G. A., P. A. Mayewski, L. D. Meeker, S. Whitlow, M. S. Twickler, and K. Taylor (1996), Potential atmospheric impact of the Toba mega-eruption ~71,000 years ago, *Geophys. Res. Lett.*, 23, 837-840.

	Top			Middle			Bottom			Average		
	Measured	Std Dev	Normalized	Std Dev	Measured	Std Dev	Normalized	Std Dev	Measured	Std Dev	Normalized	Std Dev
SiO2	71.32	0.60	74.99	1.01	71.33	0.35	74.78	0.51	70.72	0.50	74.20	0.74
TiO2	0.34	0.07	0.36	0.07	0.34	0.03	0.36	0.03	0.37	0.03	0.39	0.03
Al2O3	12.63	0.41	13.28	0.35	12.82	0.23	13.43	0.19	13.06	0.24	13.70	0.25
FeO	1.89	0.28	1.99	0.28	2.00	0.09	2.10	0.09	2.09	0.10	2.19	0.10
MnO	0.07	0.03	0.08	0.03	0.07	0.03	0.07	0.03	0.07	0.02	0.08	0.02
MgO	0.33	0.08	0.35	0.08	0.36	0.03	0.38	0.03	0.39	0.05	0.41	0.05
CaO	1.50	0.21	1.58	0.21	1.61	0.08	1.69	0.08	1.72	0.11	1.81	0.11
Na2O	4.38	0.56	4.61	0.56	4.31	0.54	4.52	0.55	4.36	0.88	4.57	0.91
K2O	2.63	0.16	2.77	0.18	2.55	0.08	2.67	0.08	2.52	0.07	2.65	0.09
SUM	95.11		100.00		95.39		100.00		95.31		100.00	

Table 1 Glass geochemistry data recorded for different depths (top, middle, and bottom sections) of the Towuti tephra measured using an electron microprobe analyzer (EMPA). The average values are also presented.

Variable	Post-tephra average	Pre-tephra average	P-value
TOC (%)	1.02	1.42	7.55287E-06
TN (%)	0.05	0.08	4.43204E-08
C/N	20.97	16.87	0.000969091
$\delta^{13}\text{C}_{\text{om}}(\text{‰})$	-31.39	-31.01	0.002749997
$\delta^{13}\text{C}_{\text{wax}}(\text{‰})$	-33.90	-33.73	0.25252073*
Na (%)	0.10	0.15	6.30622E-10
P (%)	0.04	0.03	0.000112764
Ca (%)	0.17	0.24	0.005411497
Si (%)	19.88	20.22	0.002925418
Al (%)	5.40	5.07	0.005482335
Ti (%)	0.24	0.27	2.30457E-05
K (%)	0.49	0.54	2.27504E-06
Fe (%)	11.80	13.93	9.75638E-14

Table 2 Averages for pre- and post-Towuti tephra productivity, isotopic, and elemental data and the associated p-values from the student's t-tests comparing pre- and post-tephra averages. The text not in bold in the upper half of the table indicate the use of low-resolution data (every 1 cm for the 10 cm pre-tephra and post-tephra), whereas the text in bold indicate the use of high-resolution data (every 0.2 cm for the 3 cm pre-tephra and post-tephra) in the student's t-test analyses. The asterisk symbol in the p-value column indicates that the difference is not significant (i.e. p-values above 0.05).

	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O
MD01-2387	75.27	0.33	13.86	1.50	0.10	0.34	1.43	3.93	3.23
ODP 767A	74.64	0.36	14.06	1.58	0.05	0.39	1.36	4.13	3.44
ODP 767B	74.70	0.32	13.85	1.55	0.15	0.39	1.31	4.27	3.47
TOW10-3B	74.63	0.37	13.49	2.10	0.07	0.38	1.70	4.56	2.69

	MD01-2387	ODP 767A	ODP 767B	TOW10-3B
MD01-2387	1			
ODP 767A	0.90	1		
ODP 767B	0.91	0.90	1	
TOW10-3B	0.86	0.87	0.84	1

Table 3 Averages of normalized elemental geochemistry data for tephra layers that might be generated from the same eruption (above). These tephra layers are found in the sedimentary layers of Celebes Sea (MD01-2387, Ku et al. 2009; ODP767A and B, Pouclet et al. 1999) and of Towuti (TOW10-3B, this study). Similarity coefficients between these tephra layers are also shown (below).

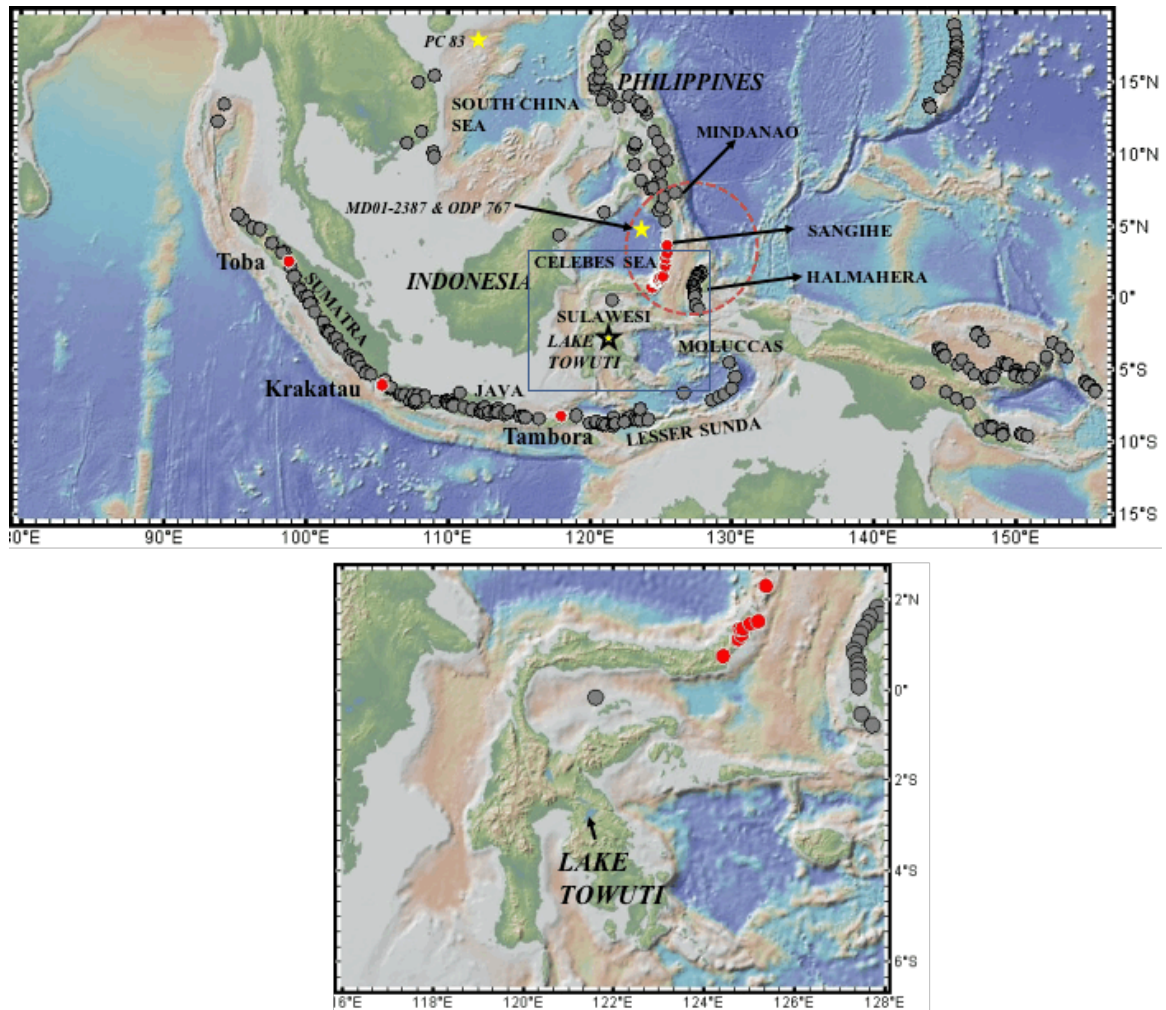


Figure 1 Map of Southeast Asia indicating major islands mentioned in the text and volcanoes in Southeast Asia (in grey circles). The locations of Toba, Krakatau, Tambora, and volcanoes in northeastern Sulawesi and Sangihe islands closest to Lake Towuti are shown in red circles. The location of Lake Towuti in central Sulawesi is marked with a yellow star, whereas other yellow stars indicate the locations of marine cores discussed in the text. The big red dashed circle indicates a volcanic province comprised of northeast Sulawesi, southern Mindanao, and Sangihe and Halmahera islands, where the volcano generating the Towuti tephra might be situated. The blue rectangle indicates the area

blown up in the inset map, focusing on Sulawesi. Maps were modified from GeoMapApp.

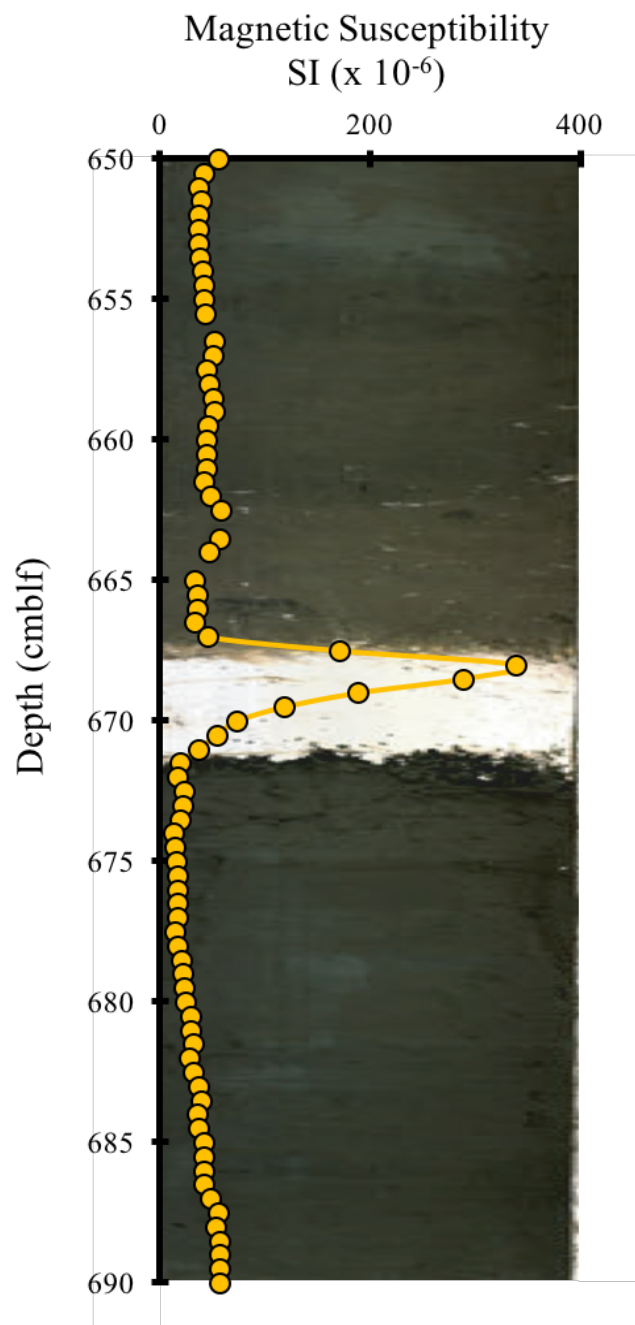


Figure 2 Magnetic susceptibility data overlaid on a scanned image of a section from core TOW10-3B showing the highly visible Towuti tephra at 668-672 cm below lake floor.

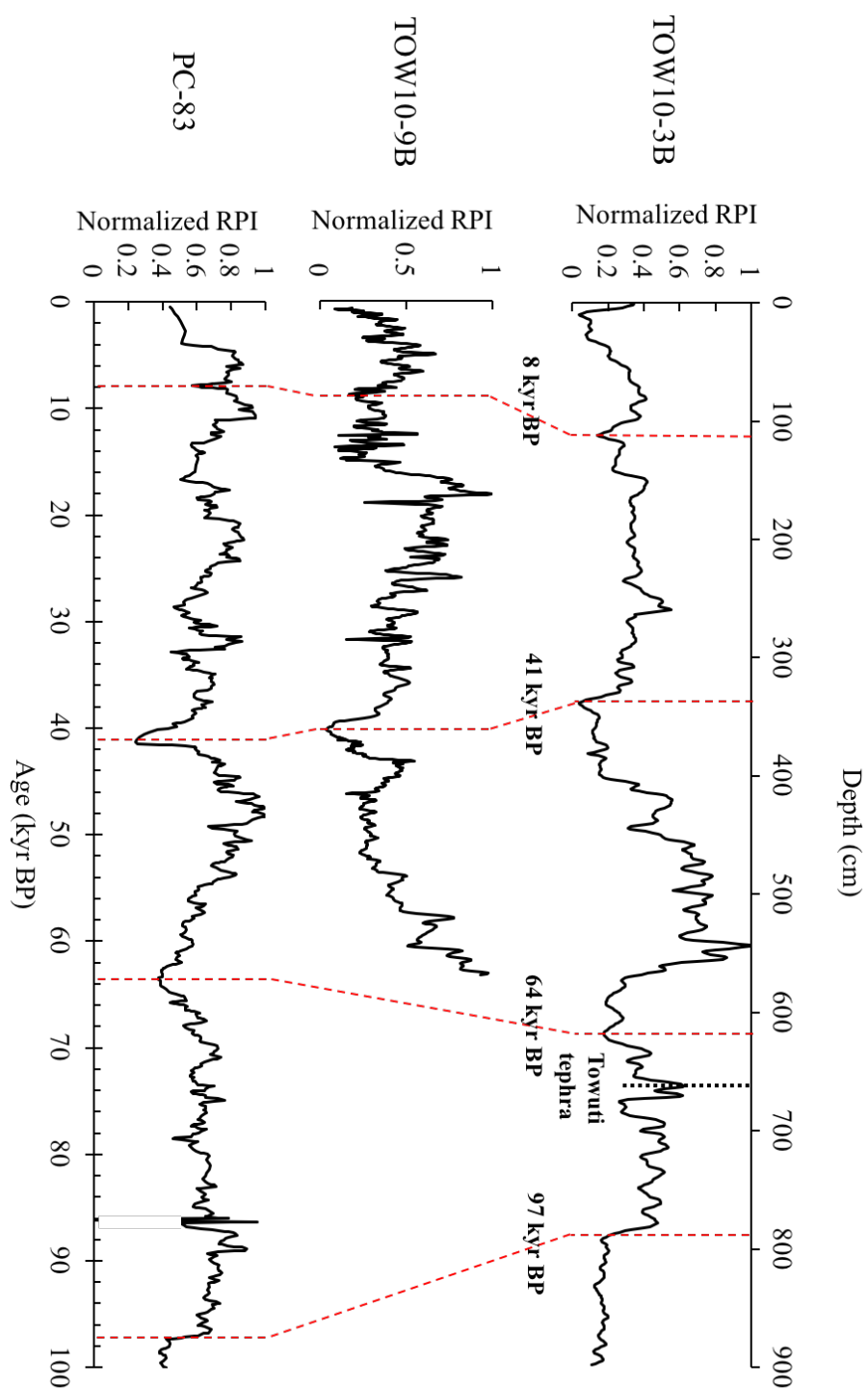


Figure 3 Magnetic paleointensity data for (from top to bottom) TOW10-3B, TOW 10-9B. and PC-83. Relative palointensity (RPI) records have been developed and published

for the well-dated TOW10-9B (Tamuntuan et al. 2015), a Lake Towuti core situated ~9 km southeast from our TOW10-3B core. Note that TOW10-9B has the same y-axis as PC-83, indicating age (kyr BP). The RPI data set from Lake Towuti has been correlated by Kirana et al. (in prep) with RPI from PC83 in South China Sea (Xiaoqiang et al. 2009). The tie points between these RPI data are indicated in red dashed lines. The depth of the tephra layer in TOW10-3B is indicated by a black dashed line.

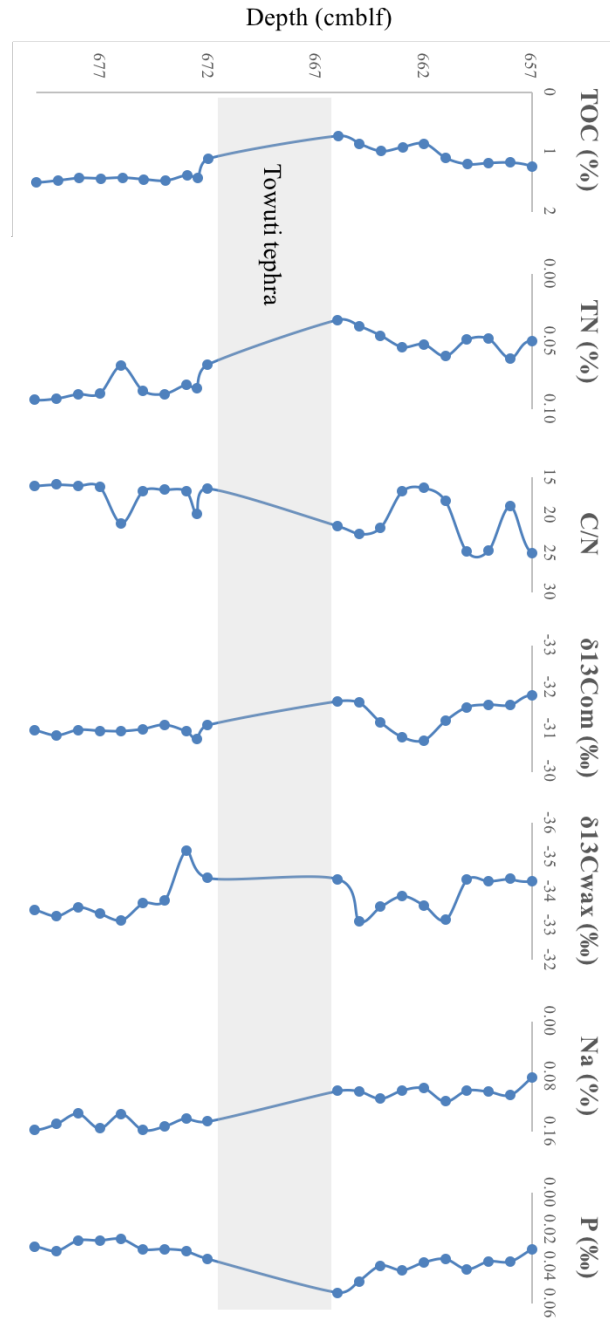


Figure 4 Low (cm-scale) resolution productivity, isotopic, and elemental data for the 10 cm below and the 10 cm above the Towuti tephra. The data include total organic carbon (TOC), total nitrogen (TN), carbon-to-nitrogen (C/N) ratio, carbon isotope of organic matter ($\delta^{13}\text{C}_{\text{om}}$), carbon isotope of leaf wax ($\delta^{13}\text{C}_{\text{wax}}$), sodium (Na), and phosphorus (P).

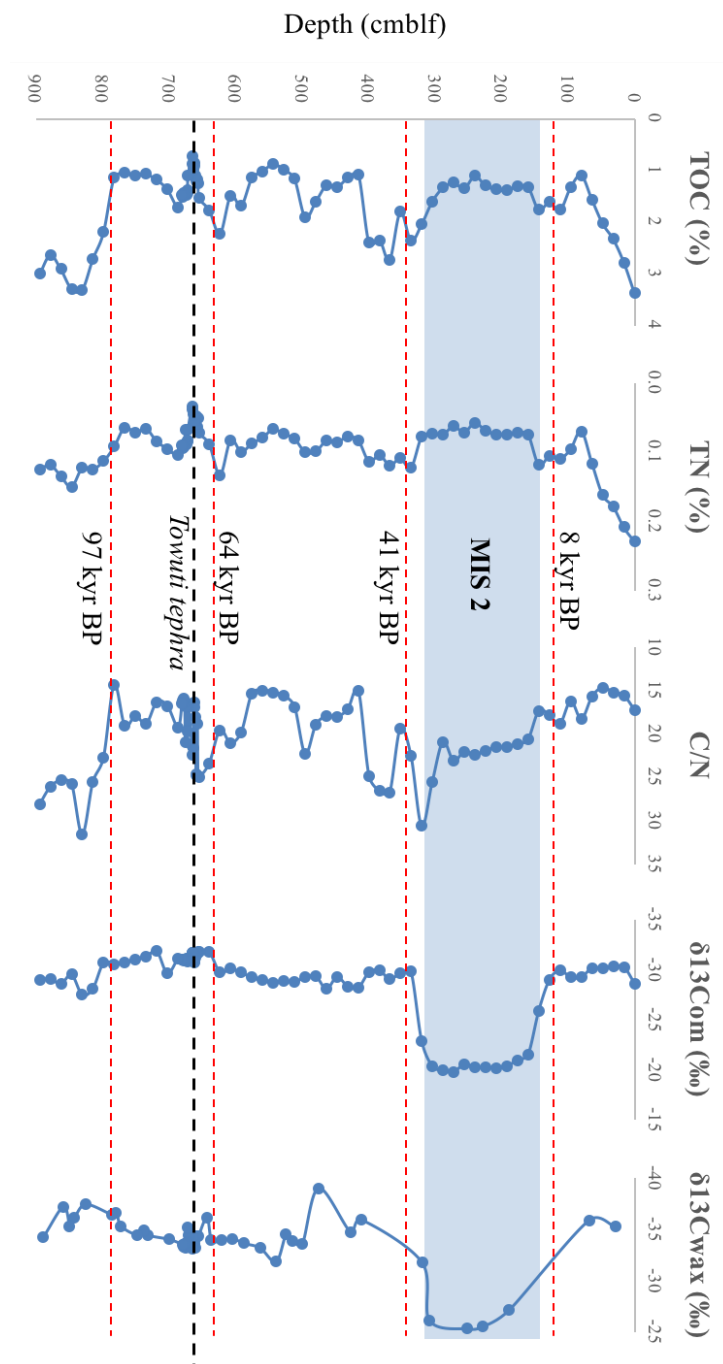


Figure 5 Similar to Figure 4, but for the whole length of TOW10-3B core, which are measured at a lower resolution (~ 16 cm). The ages of certain depths are indicated in red dashed lines based on the tie points in Figure 3. Core-length Na and P data are not available.

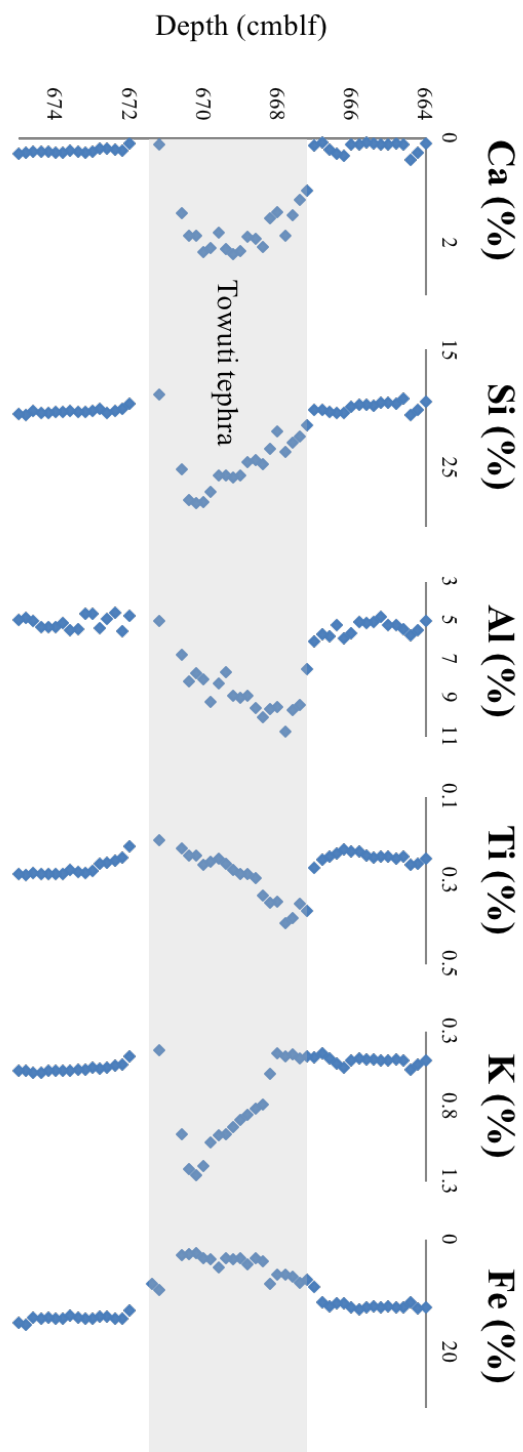


Figure 6 High resolution (every 0.2 cm) elemental data for the 3 cm below and the 3 cm above the Towuti tephra as well as for the tephra itself. The data include calcium (Ca), silicon (Si), aluminum (Al), titanium (Ti), potassium (K), and iron (Fe).

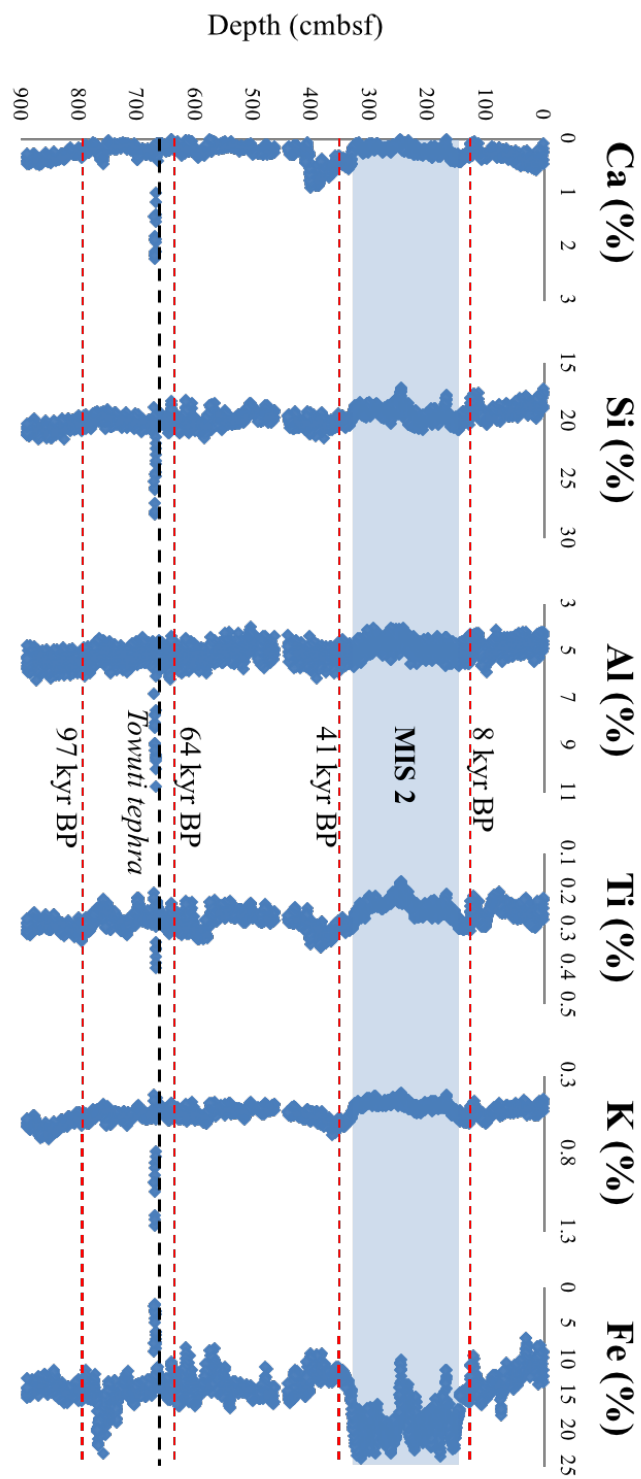


Figure 7 Same as Figure 6, but for the whole length of TOW10-3B core. The ages for certain depths are indicated in red dashed lines based on the tie points in Figure 3.

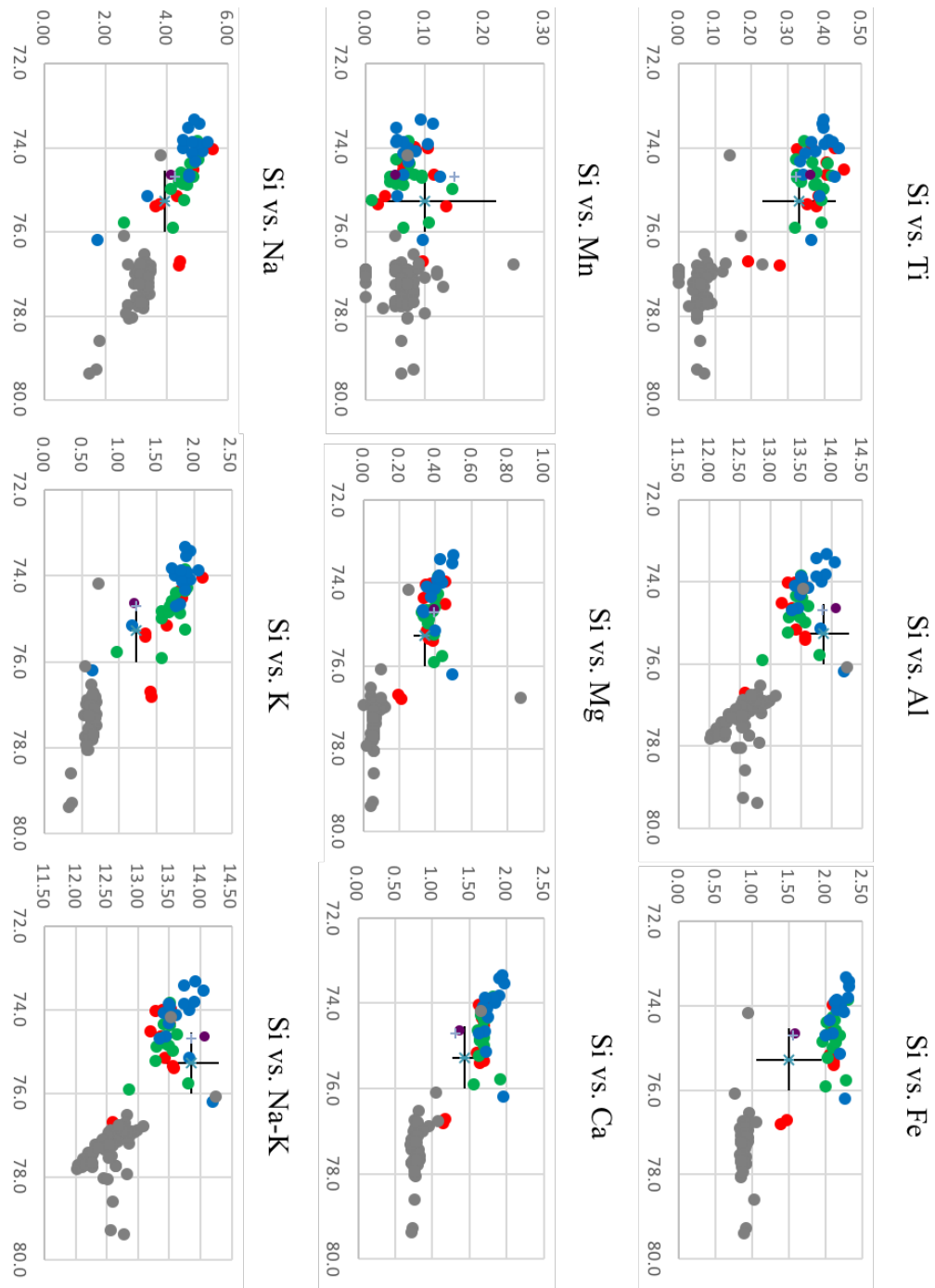


Figure 8 Major element bivariate plots showing normalized compositions of volcanic glass (in %) in the Towuti tephra and other regional tephra layers indicated in different shapes and colors: red circle (Towuti – top), green circle (Towuti – middle), blue circle

(Towuti – bottom), grey circle (Toba), green asterisk (MD01-2387), purple circle (ODP 767A), light blue cross (ODP 767B).

Supplementary Information

Additional age data points for TOW10-3B are obtained by directly comparing or “wobble-matching” magnetic susceptibility data of TOW10-3B and TOW10-9B.

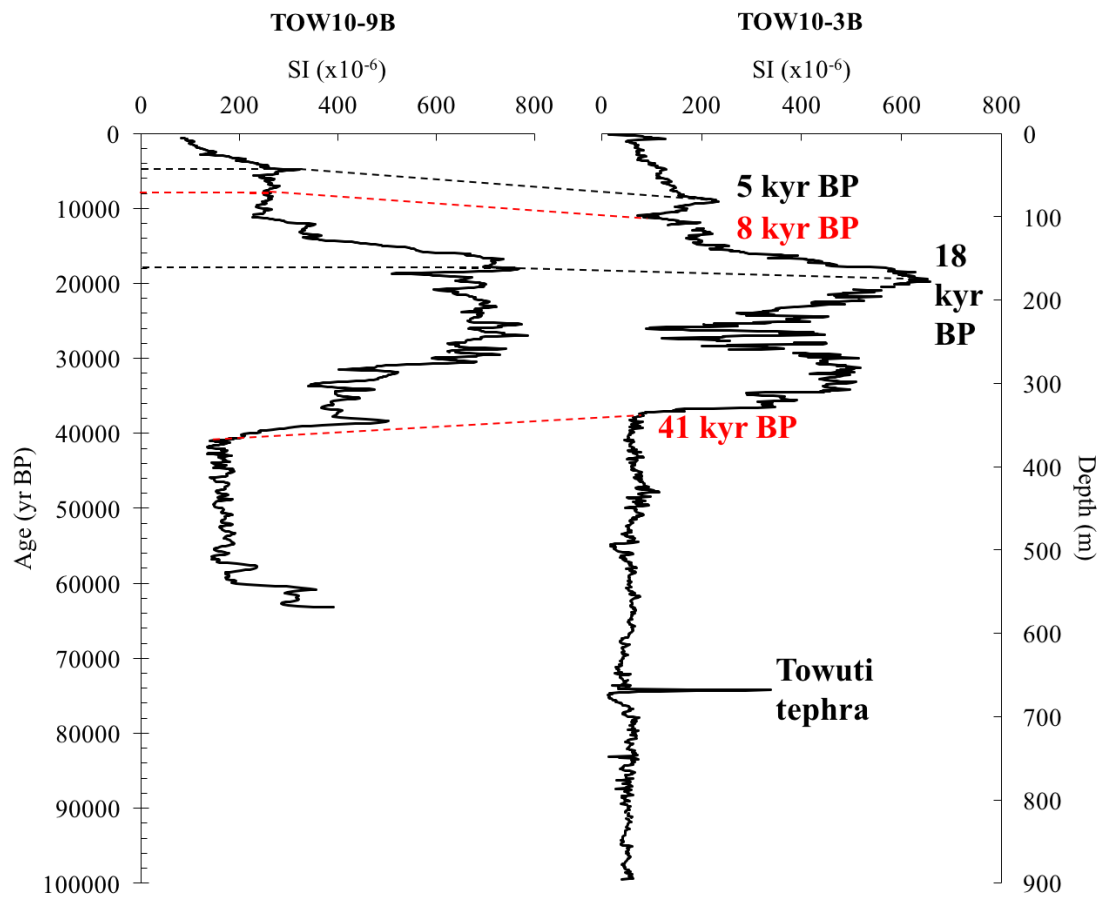


Figure S1 Magnetic susceptibility data for TOW 10-9B and TOW10-3B. The tie points based on magnetic susceptibility data are shown in black dashed lines, whereas the tie points based on RPI data (Figure 3) are indicated in red dashed lines. The Towuti tephra is marked by a spike in magnetic susceptibility in TOW10-3B.