Hydroclimate changes in the Maritime Continent over the past 30,000 years recorded by speleothems from Sulawesi, Indonesia

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**NANYANG
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HYDROCLIMATE CHANGES IN THE
MARITIME CONTINENT OVER THE PAST
30,000 YEARS RECORDED BY SPELEOTHEMS
FROM SULAWESI, INDONESIA

YUAN SHUFANG

INTERDISCIPLINARY GRADUATE SCHOOL

2018

**Hydroclimate changes in the Maritime
Continent over the past 30,000 years
recorded by speleothems from Sulawesi,
Indonesia**

Yuan Shufang

Interdisciplinary Graduate School

A thesis submitted to the Nanyang
Technological University in fulfilment of the
requirement for the degree of Doctor of
Philosophy

2018

Candidate Statement of Originality

I hereby certify that the work embodied in this thesis is the result of original research, is free of plagiarised materials, and has not been submitted for a higher degree to any other University or Institution. I confirm that the investigations were conducted in accord with the ethics policies and integrity standards of Nanyang Technological University and that the research data are presented honestly and without prejudice.

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This thesis does not contain any publication documents. Therefore, this form does not apply for my thesis submission processes.

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Abstract

I use calcite speleothems from southwestern Sulawesi, Indonesia, to reconstruct changes in the regional hydroclimate and vegetation during the past 30,000 years. The reconstructions are based on multi-proxy records comprised of oxygen ($\delta^{18}\text{O}$) and carbon ($\delta^{13}\text{C}$) isotopic data and trace element ratios dated by high-precision uranium-thorium disequilibrium dating methods. The $\delta^{18}\text{O}$ record shows increased strength of moisture convection in the Maritime Continent from the last glacial period to the Holocene. The $\delta^{13}\text{C}$ and Mg/Ca records suggest a forest expansion coupled with increased soil CO_2 production. These glacial-interglacial changes in the regional hydroclimate and vegetation correspond to the rising of sea surface temperature and atmospheric CO_2 concentration, and Sunda Shelf flooding.

Comparisons between the Sulawesi $\delta^{18}\text{O}$ and other speleothem $\delta^{18}\text{O}$ records from the Maritime Continent suggest that the southern extent of the regional mean ITCZ has likely been between southwestern Sulawesi and Flores, which are fewer than 300 km apart, during the past 30,000 years. Based on the Sulawesi and Flores records, I have then derived an ITCZ index, which allows the characterization of the relative movements of the ITCZ. The index indicates that ITCZ migration is driven by precession on the orbital timescale, coupled with changes in climate boundary conditions and ocean circulations.

Further synthesis of tropical and subtropical speleothem $\delta^{18}\text{O}$ shows that, similarly to what can be observed today in rainfall $\delta^{18}\text{O}$, the continental rainout effect has primarily controlled the spatial distribution of speleothem $\delta^{18}\text{O}$

during the past 30,000 years, with isotopic fractionation having been dominated by temperature and continental water recycling. On the glacial-interglacial timescale, speleothem $\delta^{18}\text{O}$ has largely followed nearby sea surface temperature at upstream monsoonal and maritime sites, but at downstream monsoonal sites and sites located south of the ITCZ, it instead followed precessional insolation. Temporally, tropical-subtropical speleothem $\delta^{18}\text{O}$ variation has been predominately influenced by ITCZ migration since the LGM, with secondary influences from deep tropical convection during the Holocene.

Chapter 1

Introduction

1.1 The Maritime Continent: its climate significance and complexity

The ‘Maritime Continent’ (Ramage, 1968) (Figure 1.1), located between Asia to the north and Australia to the south, the Indian Ocean to the west and the Pacific Ocean to the east, is comprised of peninsulas, thousands of islands of various sizes, mountainous terrain, and many shallow seas. It also encompasses the largest area of the warmest ocean water in the world, known as the Indo-Pacific Warm Pool (IPWP). The Maritime Continent has the most intense convection and the highest tropopause on the planet. The regional convection in the atmosphere forms the rising limb of the Hadley and Walker Circulations, through which large amounts of heat and moisture are transported around the globe. Variations in the Hadley and Walker Circulations tremendously influence the earth climate system through connections with large-scale climate variations such as the Madden Julian Oscillation (MJO), Asian-Australian monsoons, Intertropical Convergence Zone (ITCZ) migration, El Niño-Southern Oscillation (ENSO), and Indian Ocean Dipole (IOD). Knowledge of the variation of regional convection strength and the corresponding driving mechanisms can help us better understand global climate changes, which are critical to people’s livelihood in the region and far beyond.

Figure 1.1 The major islands and seas of the Maritime Continent. Also shown is the topography of the region, with the white shaded areas marking the shallow seas that were exposed during the Last Glacial Maximum (LGM).

Presently, the Maritime Continent is characterized by a strong seasonal wind reversal that feeds the Asian-Australian summer monsoons, with southwesterly and northeasterly winds dominating during boreal and austral summers, respectively (Figure 1.2). Such wind reversal is accompanied by the seasonal latitudinal migration of the ITCZ and rainfall maximum. As a matter of fact, the ITCZ stays over the Maritime Continent during the majority of the year. This results in relatively abundant rainfall year-round (generally > 2000 mm) and therefore weak seasonality for most areas in the region. The surrounding surface ocean waters are not only warm, but also low in salinity (Figure 1.3). These warm and fresh waters are transferred from the South China Sea and the western tropical Pacific Ocean to the Indian Ocean through passageways across the Maritime Continent, which are known collectively as

the Indonesian Throughflow (ITF). Significant rainfall seasonality is only observed in the southern part of the region such as southern Sulawesi, Papua New Guinea, and the Lesser Sunda Islands. On inter-annual timescales, considerable rainfall variability in the region also exists, which is associated with fluctuations in the intensity of the Walker Circulation, as part of the ENSO phenomena.

Figure 1.2 Climatology over the Maritime Continent during June-July-August (top) and December-January-February (bottom). The precipitation data (the shaded colors, in mm/hr) is based on the Tropical Rainfall Measuring Mission (TRMM) 3B43 dataset (1998-present). The wind data (the arrows, in m/s) is based on the National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010) (Kalnay et al., 1996).

While the Maritime Continent and the Indo-Pacific Warm Pool play climatically critical roles in the globe during the present day, whether their roles remain the same in the past relies on better knowledge of past regional hydroclimate variation over various timescales, which is yet elusive. As existing records generally show widespread inconsistencies between marine

and/or terrestrial records (DiNezio and Tierney, 2013; Reeves et al., 2013), either on the glacial-interglacial timescale (Mohtadi et al., 2014; Russell et al., 2014) or on millennial timescales during North Atlantic abrupt climate events (Ayliffe et al., 2013; Partin et al., 2007).

Figure 1.3 Sea surface temperature (SST) and sea surface salinity over the Indo-Pacific Warm Pool during austral summer (February) and winter (August). Figure source: De Deckker (2016). The Indo-Pacific Warm Pool delineates the shallow ocean surrounding the Maritime Continent with temperature $>28^{\circ}$ and relatively low salinity. This warm and fresh water also migrates with seasons. The data were obtained from the World Ocean Atlas 2005 and were accessed at <http://ferret.pmel.noaa.gov/thredds/dods/PMEL/WOA5>.

One of the most pressing debates is on the glacial-interglacial hydroclimate change in the region: more specifically, whether it was indeed drier over the Maritime Continent during the Last Glacial Maximum (LGM) compared with the Holocene. Available paleoclimate proxy data from the region and modeling studies are generally inconclusive (Figure 1.4). For instance, on the marine side, ice-volume-corrected seawater oxygen isotope ratios ($\delta^{18}\text{O}_{\text{sw-ivc}}$) are commonly used to indicate the variation of sea surface salinity and hence of marine hydroclimate changes, with higher $\delta^{18}\text{O}_{\text{sw-ivc}}$ corresponding to higher sea surface salinity associated with lower rainfall amount, and vice versa. Off western coast of Sumatra (Mohtadi et al., 2014) and south of the Indonesian Archipelago (Levi et al., 2007), $\delta^{18}\text{O}_{\text{sw-ivc}}$ shows comparable values between the LGM and Holocene, suggesting no obvious glacial-interglacial hydroclimate change. However, the same proxy shows lower glacial values in the Sulu Sea (Linsley and Thunell, 1990; Rosenthal et al., 2003) and north of Papua New Guinea (de Garidel-Thoron et al., 2007), indicating freshening of surface water during the LGM. Yet, higher glacial $\delta^{18}\text{O}_{\text{sw-ivc}}$ from the Makassar Strait (Visser et al., 2003) and eastern edge of the Indonesian Archipelago (Stott et al., 2002) implies saltier seawater during the LGM compared with the Holocene.

Figure 1.4 Network of terrestrial (left) and marine (right) proxy records showing the hydroclimate change over the Indo-Pacific Warm Pool during the Last Glacial Maximum.

Figure source: DiNezio and Tierney (2013). The listed records are also compiled in the same reference.

Similar ambiguities also exist in the terrestrial proxy records. Sediment cores from the northwestern coast of Sumatra (Niedermeyer et al., 2014) and offshore northeast Borneo (Dubois et al., 2014) show comparable vegetation distribution between the LGM and Holocene based on almost constant values of leaf wax isotopic ratios (vascular plant fatty acid $\delta^{13}\text{C}$), suggesting no significant hydroclimate change. The same pattern is manifested by comparable values in elemental compositions and grain-size based proxies between the LGM and Holocene from offshore southern Java, indicating an equivalent austral summer monsoon strength between the two periods (Mohtadi et al., 2011). On the other hand, a drier LGM condition is suggested by higher leaf wax $\delta^{13}\text{C}$ and lower sedimentary Ti counts from a lake record in central Sulawesi (Russell et al., 2014), lower forest pollen counts and higher leaf wax $\delta^{13}\text{C}$ from coastal Sumba (Dubois et al., 2014), as well as lower Fe/Ca ratios from the Davao Gulf of the southern Philippines (Fraser et al., 2014).

Palynological records and lake level records generally suggest a drier LGM (Reeves et al., 2013). Two absolutely dated speleothem oxygen isotope records from northern Borneo and Flores show relatively more positive $\delta^{18}\text{O}$ during the LGM compared with the Holocene (Ayliffe et al., 2013; Carolin et al., 2013; Griffiths et al., 2009; Partin et al., 2007). Yet the Borneo record suggests a trivial glacial-interglacial hydroclimate change if we correct the effects of cave temperature and ocean water $\delta^{18}\text{O}$ reservoir (Partin et al., 2007), whereas the Flores record suggests that a comparably wet LGM relative to the Holocene (Ayliffe et al., 2013). In addition to the inconsistencies found in paleoclimate records, an isotope-enabled general circulation modeling study suggests a LGM climate wetter than the present day in the region, while the changes in rainfall $\delta^{18}\text{O}$ were overall negligible (Werner et al., 2016).

Controversies were also found in regional hydroclimate responses in North Atlantic millennial-scale climate events during the last glacial period. For instance, marine sediment records from the Makassar Strait (Schröder et al., 2016a), west coast of Sumatra (Mohtadi et al., 2014), and offshore South Java (Mohtadi et al., 2011) suggest drier conditions during the Younger Dryas (YD) and Heinrich Event 1 (H1), along with a wetter climate during the Bølling-Allerød (B/A). In contrast, another sediment record from southern Philippines (Fraser et al., 2014) and Flores speleothem $\delta^{18}\text{O}$ (Ayliffe et al., 2013; Griffiths et al., 2009) imply an opposite pattern, with the YD and H1 being wetter and B/A being drier. Borneo speleothem $\delta^{18}\text{O}$, despite suggesting dry periods during Heinrich Events, shows no obvious YD, B/A and other Dansgaard-Oeschger (DO) events (Carolin et al., 2013; Partin et al., 2007). Overall, the millennial scale features during the last deglaciation seem to be of relatively

small or even indiscernible magnitudes in the maritime-continent records (Reeves et al., 2013; Rosenthal et al., 2003; Steinke et al., 2006).

Such widespread inconsistencies over a range of timescales and spatial coverage described above could be due to the complex land and sea configuration over the Maritime Continent, which results in diversified hydroclimate responses to integrated forcing from different climate systems such as monsoons, the ITCZ and ENSO. Indeed, the Maritime Continent appears to be one of the most challenging regions for climate model simulations. The overall complicated atmospheric dynamics in the region can be attributed to multiple factors including its unique geographical setting in the deep tropics, latitudinally neighboring two major continents, and zonally surrounded by two massive oceans, complex land-sea interactions due to island topographic features, and the exposure and flooding of the Sunda and Sahul shelves during glacial-interglacial periods. All these factors lead to the overall complicated atmospheric circulation and water recycling in the region. A more comprehensive and less ambiguous understanding on regional hydroclimate thus requires a relatively widespread and dense coverage of climate reconstructions, particularly from climatically sensitive locations. Moreover, hydroclimatic information recorded in any archive is temporally and spatially limited: a proxy record may be biased toward a specific season or only sensitive in a specific time window. Additionally, a record may be influenced by local or regional integrated climatic signals. More importantly, different archives are usually dated through different dating techniques. Marine and lake sediments, for instance, are mostly dated by radiocarbon techniques, which often involve carbon reservoir corrections. Speleothems can be dated by high-precision

uranium-thorium dating techniques, provided that the speleothems contain relatively high U concentrations and low ($^{230}\text{Th}/^{232}\text{Th}$) initial values (Richards and Dorale, 2003). While comparisons are often practiced among the data from different proxies and even archives, it is more favorable to conduct comparisons on paleoclimate records that are based on the same type of proxy and well-constrained ages, and from the same geographical region. Hence, comparable proxy interpretation and chronology control can be guaranteed as a basis for more reliable knowledge of the spatial and temporal patterns of regional climate change.

1.2 Vegetation in the Maritime Continent: past and present

The hot and humid conditions throughout the Maritime Continent allow widespread closed lowland rainforests to thrive in the region. These rainforests, in turn, can support exceptionally rich biological diversity, and translate the Maritime Continent into one of the most important biodiversity hotspots for conservation (Myers et al., 2000; Underwood et al., 2014). At the present time, rainfall amount and seasonality largely determine the vegetation distribution over the region, with major vegetative types including evergreen rainforests, deciduous forests, montane forests, natural savannas, as well as more limited natural or disturbed vegetation (Heaney, 1991). Lowland evergreen forests and montane forests predominate on the islands of Borneo, Sulawesi, and Papua New Guinea, and to a lesser degree, the Malay Peninsula and Sumatra (Figure 1.5). These areas are characterized by high annual rainfall amount and weak rainfall seasonality. In contrast, seasonal forests and savannas are common in eastern Java, southeastern Borneo, western Philippines, and the Lesser Sunda

Islands, where annual rainfall amount is lower and rainfall seasonality is stronger.

Figure 1.5 Forest cover map of insular Southeast Asia. Figure Source: (Stibig et al., 2003).

The geographic extent of the rainforests in the Maritime Continent has varied due to global or regional climatic changes in sea level, temperature, and rainfall (Heaney, 1991). Earlier widespread palynological studies and more recent studies on biomarker records demonstrate that these exceptionally complex, highly-localized and diverse forests are quite sensitive and vulnerable to climatic changes. For instance, during the LGM, when sea level, atmospheric CO₂ concentration (pCO₂), and temperature differed dramatically, the glacio-eustatic depression of sea level by around 120 m fully exposed Sunda and Sahul shelves, resulting in the largest contrast of land-sea distribution across the whole tropics. Such changes might have caused enormous shifts in regional hydroclimate and vegetation (De Deckker, 2016; De Deckker et al., 2003). For example, the exposed Sunda Shelf possibly hosted a zonal rainforest belt (Hope et al., 2004) but a meridional ‘savanna corridor’ (Bird et al., 2005; Heaney, 1991) which may have facilitated the migration of *Homo sapiens* through more

continuous land connection and less forest coverage. Temperature change on glacial-interglacial timescales can also affect the large-scale vegetation shift and biomass changes through its influences on rainfall amount and seasonality (Schefuß et al., 2003). In addition, atmospheric concentration of CO₂ ($p\text{CO}_2$) might have influenced the distribution of C3 (trees) and C4 (tropical grass) plants during glacial-interglacial periods when large changes in $p\text{CO}_2$ occurred (Ehleringer, 2005; Ehleringer et al., 1997; Street-Perrott et al., 1997). Most paleo-vegetation records show pervasive grassland expansion over the Maritime Continent and other tropical regions during the LGM, suggesting generally a drier climate and/or enhanced rainfall seasonality (Reeves et al., 2013).

A thorough understanding of the responses of tropical vegetation to large changes in sea level, atmospheric CO₂ concentration, and temperature during the LGM and the following deglaciation can greatly improve our capability in predicting responses of tropical vegetation to the warming future. Most of the existing paleo-vegetation records over the Maritime Continent are, however, discontinuous or with low resolutions, and relatively poor-constrained in chronology. To help improve the situation requires vegetation reconstructions from key locations, with well-constrained ages, continuous coverage, and high resolution.

In a warming world, forest biomass will likely increase along with $p\text{CO}_2$ (Polley et al., 1993). Woodland encroachment on grasslands has also been related to $p\text{CO}_2$ rise since the preindustrial level (Polley et al., 1996). However, models still predict contraction of tropical rainforests in a warming world due to increased seasonal rainfall variability (Underwood et al., 2014) (Figure 1.6).

In addition to climate influences, human deforestation has substantially reduced the extent of rainforest cover (Kalamandeen, 2019), probably to its lowest level (Whitmore, 1984), threatening the entire rainforest biota (Heaney, 1991).

Figure 1.6 Ever-wet zones in Southeast Asia under current and future climate. Figure Source: (Underwood et al., 2014). The future climate is set for 2050 under representative concentration pathway (RCP) 8.5.

1.3 Speleothems as a climate archive

High-resolution paleoclimatic and environmental variations may be reconstructed from cave carbonate deposits known as speleothems, which grow under relatively stable cave environments with minimal air circulation where both relative humidity and partial pressure of CO_2 remain high and stable throughout the year (Hendy, 1971). One of the advantages of speleothems as a climate archive over other continental archives such as lake sediments and peat

bogs is that speleothems can offer an absolute chronological control with relatively high precision. Provided that the system remains chemically closed after carbonate formation, speleothems are nearly an ideal material for precise uranium-series dating, in particular U-Th dating, which is one of the most applicable dating tools for late Quaternary studies (Henderson, 2006). By measuring the radioactive decays in uranium, thorium, and/or protactinium disequilibrium systems (Edwards et al., 1987), U-series dating methods yield dates in calendar years, circumventing uncertainties inherent in radiocarbon calibration or orbital tuning when studying marine sediments and other terrestrial archives (McDermott, 2004). With improvements in the thermal ionization mass spectrometric (TIMS) and inductively coupled plasma mass spectrometric (ICP-MS) techniques, the U-Th dating method can date materials aged between only few years and up to ~700,000 years old, with much reduced sample sizes (e.g., <10 mg of sample with a ^{238}U content of ~1 ppm) (Cheng et al., 2016a; Wang, 2006).

The precise chronology of speleothem records offers unmatched opportunities for detailed reconstruction of climatic and environmental variations over a range of timescales, as well as opportunities to assess potential leads and lags between climate events to further understand corresponding driving mechanisms of climate changes (Cheng et al., 2009; Fleitmann et al., 2008; McDermott, 2004; Wang et al., 2001).

Speleothems can offer multiple paleoclimate proxies, including carbonate stable isotopes (most commonly O and C stable isotopes, but recently Ca and Mg isotopes, albeit still in their infancies), growth rate, trace elements, stable isotopes and noble gases of water inclusions, and organic compounds (Buhl et

al., 2007; Fairchild and Baker, 2012; Owen et al., 2016). Hence, if study sites are carefully chosen, these proxies in speleothems can provide an impressive toolbox to reconstruct a comprehensive history of rainfall, temperature, atmospheric circulation, soil and vegetation dynamics, as well as other environmental conditions (Fairchild and Treble, 2009; McDermott, 2004).

1.3.1 Speleothem oxygen isotope ratios ($\delta^{18}\text{O}$)

In a deep cave with vadose zone thicker than ~10 m, cave temperature typically remains relatively stable throughout the year, and approximates the mean annual temperature above the cave. When cave humidity is high enough so that evaporation is negligible, speleothem carbonates (mostly calcites) can be deposited under conditions close to an isotopic equilibrium (McDermott, 2004). Equilibrium conditions of calcite precipitation can be tested either by the so-called “Hendy Test” (Hendy, 1971), or more effectively, by replication between different speleothem specimens (Dorale et al., 2002; Wang et al., 2001). The oxygen isotope ($\delta^{18}\text{O}$) variations of speleothems deposited under equilibrium conditions reflect changes mainly in the isotopic compositions of meteoric water and cave temperature that controls the isotopic fractionation between drip water and calcite (Dorale et al., 2002; Fleitmann et al., 2008; McDermott, 2004). The temperature effect can become dominant in mid- and high-latitude caves or alpine caves, where relatively large temperature change can also exert a substantial isotopic fractionation in meteoric precipitation. In the tropics and part of the subtropics, where cave temperature only varies slightly (except during climate transition periods or extreme climatic events when tropical temperature might vary greatly, under which circumstances temperature effects need to be considered), calcite $\delta^{18}\text{O}$ mainly inherits changes

in rainfall $\delta^{18}\text{O}$, which reflect regional hydroclimate changes above the cave. In the recent decades, speleothem $\delta^{18}\text{O}$ has been widely adopted to infer tropical and subtropical monsoon changes over a range of timescales from both hemispheres (Cruz et al., 2005a; Fleitmann et al., 2007; Wang et al., 2006; Wang et al., 2007; Wang et al., 2005; Wang et al., 2008; Wang et al., 2001; Zhang et al., 2008).

Climatic interpretations of speleothem $\delta^{18}\text{O}$, however, are still complicated by numerous factors that control the variation of rainfall $\delta^{18}\text{O}$, such as altitude, continentality (latitude, longitude and distance from the sea, Figure 1.7), sea surface temperature, rainfall amount and seasonality (Dansgaard, 1964; Gat, 1996; Rozanski et al., 1993). While the Global Network of Isotopes in Precipitation (GNIP) has extensively documented modern rainfall $\delta^{18}\text{O}$ since the early 1960s, the instrumental data, however, is still too short for a thorough understanding on the rainfall $\delta^{18}\text{O}$ variability over longer timescales, which bears significance to long-term regional hydrological cycle, atmospheric circulation and global climate change. As speleothem $\delta^{18}\text{O}$ records become more abundant, they provide a potential to establish a long and spatially integrated view of precipitation and its $\delta^{18}\text{O}$ (Wang, 2006). Previous studies on the spatial gradient of speleothem $\delta^{18}\text{O}$ are sparse, conducted in parts of Europe (Deininger, 2013; Deininger et al., 2017; McDermott et al., 2011; McDermott et al., 1999), China (Hu et al., 2008b), New Zealand (Williams et al., 1999), and Central and South America (Lachniet et al., 2007; Wang et al., 2017). With a more extensive coverage of speleothem records, similar attempts on a global scale are now feasible and necessary.

Figure 1.7 Continental effects shown by rainfall $\delta^{18}\text{O}$ for selected European stations of the IAEA/WMO global network (Rozanski et al., 1993). The data is plotted as a function of the distance from the Atlantic coast. Long-term annual, summer (June, July, August) and winter (December, January, February) mean $\delta^{18}\text{O}$ values are indicated for each station. Note the increasing spatial gradient of rainfall $\delta^{18}\text{O}$ along distance from the coast, which is larger during winter than summer.

1.3.2 Speleothem carbon isotope ratios ($\delta^{13}\text{C}$)

Carbon isotope ratio ($\delta^{13}\text{C}$) of speleothem carbonates is an additional proxy that reflects environmental changes. Yet, it is more difficult to be interpreted in terms of climate signals (Baker et al., 1997). Speleothem $\delta^{13}\text{C}$ can be influenced by bedrock, vegetation shift and production, soil activity, temperature, precipitation, and in-cave processes such as CO_2 degassing and

prior calcite precipitation (Meyer et al., 2014). In an open system, speleothem $\delta^{13}\text{C}$ tends to be dominated by the isotopic signature of soil CO_2 due to continuous equilibration between seepage water and a nearly unlimited reservoir of soil CO_2 ; in comparison, in a closed system, the contribution of host bedrock becomes more prominent. In practice, most natural caves are likely to be partially open (McDermott, 2004). As a result, speleothem $\delta^{13}\text{C}$ variation has been commonly ascribed to climate-driven changes in vegetation, because soil-respired CO_2 in equilibrium with C3-dominated vegetation tends to be more depleted in ^{13}C compared with that equilibrated with C4 vegetation (Cerling, 1984; Cerling et al., 1991). Such distinct differences can be preserved in speleothem $\delta^{13}\text{C}$ values, ranging between -14‰~-6‰ for C3- and -6‰~-2‰ for C4-dominated plant assemblages (McDermott, 2004). Ideally, climatic interpretation of speleothem $\delta^{13}\text{C}$ should be accompanied by cave monitoring and ventilation studies to eliminate prominent effects from in-cave processes (Baldini et al., 2008). This, however, is not always practical, particularly for caves in remote locations. Therefore, a good replication of $\delta^{13}\text{C}$ records between stalagmites can instead form a strong argument for isotopic equilibrium and for common climatic or environmental signals.

1.3.3 Speleothem trace elements

Trace elements used to be an under-utilized climate proxy of speleothems due to the analytical limitation in acquiring large amount of data efficiently. The development of laser ablation inductively coupled plasma mass spectrometric (LA-ICP-MS) and X-ray fluorescence (XRF) techniques has enabled to precisely and rapidly analyze trace elements such as Mg, Sr, Ba, P, and U in annual or even monthly resolution. The primary source of most trace

elements in speleothems is bedrock and its overlying regolith (Fairchild and Treble, 2009). Modern drip water monitoring suggests that trace elements can reflect the infiltration rate of meteoric water in the vadose zone, and hence have the potential to reveal changes that might be climate related (Fairchild et al., 2001; Fairchild et al., 2000). Indeed, Mg has been shown to be sensitive to drought events, and has been used to record paleo-aridity, with supporting evidence from cave monitoring or its co-variation with oxygen and carbon isotopes (Fairchild and Treble, 2009; Hellstrom and McCulloch, 2000; Jo et al., 2010; McDonald et al., 2004). Covariance of Mg/Ca and Sr/Ca in speleothems has been attributed to carbonate precipitations in the upstream of cave drip water (Fairchild et al., 2006). This so-called prior calcite precipitation or PCP (to a much less degree in aragonite formation) is related to water residence time and recharge variation in the vadose zone (Cruz et al., 2007). Co-variations of speleothem Sr and Ba, and P and U over varied timescales have also been observed, indicating common controlling factors on these pairs of elements (Fairchild and Treble, 2009). However, because trace element compositions in speleothems can be controlled by multiple factors including source, dissolution process, karst system, PCP, metal substitution and absorption during carbonate crystallization (Fairchild and Treble, 2009), climatic interpretation of trace elements in speleothems should be accompanied by well characterized cave system.

1.4 Speleothem studies in the Maritime Continent

The Maritime Continent has an extensive, although not continuous, limestone distribution over its major islands, peninsulas, and archipelagos. Previous speleothem studies have been based mainly on samples from northern

Borneo, central Sumatra, Flores, and eastern Java (Figure 1.8), and mostly utilized proxies of speleothem oxygen isotopes, and to a lesser extent, of trace elements, carbon isotopes, and fluid inclusions.

1.4.1 Speleothem studies in northern Borneo

The first speleothem $\delta^{18}\text{O}$ record reported from northern Borneo was interpreted to reflect changes in the west tropical Pacific warm pool hydrology (Partin et al., 2007). The record spanned the last 27,000 years, and thus provided a detailed assessment of the responses of tropical Pacific convection to changes in radiative forcing, global climate boundary conditions, and abrupt climate changes (Partin et al., 2013). The record suggested that the warm pool convection weakened since the LGM and reached a minimum during H1, when temperature was rising since the early stages of the last deglaciation. The Borneo speleothem $\delta^{18}\text{O}$, however, showed no changes in regional convection strength during the YD and B/A, while it showed a plateau coinciding with the Antarctic Cold Reversal (ACR). The convection strength during the Holocene appeared to follow equinox insolation. Later on, the Borneo speleothem $\delta^{18}\text{O}$ was extended to cover the recent 160,000 years (Carolin et al., 2013; Carolin et al., 2016) and between around 200,000 and 550,000 years (Meckler et al., 2012). The extended records further confirmed that the Borneo speleothem $\delta^{18}\text{O}$, and hence the warm pool convection strength, was insensitive to glacial-interglacial boundary condition changes; instead $\delta^{18}\text{O}$ was dominated mainly by the influences of precession over the orbital timescale. On millennial timescales, long Borneo records confirmed that weaker regional convection during glacial terminations coincides with Heinrich events, while no obvious change of convection strength occurs during Dansgaard–Oeschger (DO) events.

Additionally, a high resolution (sub-annual) Borneo speleothem $\delta^{18}\text{O}$ record was used to study ENSO variability during the Holocene, suggesting a minimum ENSO variance during the mid-Holocene, coinciding with the strongest regional convection (Chen et al., 2016).

Carbon isotopes from Borneo speleothems showed replicable LGM-Holocene depletion of $\delta^{13}\text{C}$ of $\sim 1\text{‰}$, which was interpreted to reflect a vegetational transition from savannah (C4) to rainforest (C3) conditions (Partin et al., 2013). Mg/Ca and Sr/Ca from Borneo speleothems present a weak replication between stalagmites and their profiles show little resemblance to speleothem stable isotope records. Therefore, they were suggested to be poor indicators of hydroclimate conditions (Partin et al., 2013). Lastly, fluid inclusion water isotopes, noble gases, liquid-vapor homogenization, and clump isotopes from Borneo speleothems were also used to reconstruct glacial-interglacial temperature changes of $4\text{--}5^\circ\text{C}$ during marine isotope stage (MIS) 12-9 (Meckler et al., 2015).

Figure 1.8 The locations of speleothem records from northern Borneo (light blue), central Sumatra (yellow), Flores (blue), eastern Java (grey), and southwestern Sulawesi (red star).

1.4.2 Speleothem studies in central Sumatra, Flores, and Java

In comparison with Borneo speleothem $\delta^{18}\text{O}$ records, a 16,000-year-long speleothem $\delta^{18}\text{O}$ record from central Sumatra showed a clear deglacial structure, with drier YD and wetter B/A (Wurtzel et al., 2018). The Sumatra record was originally composed of a mixture of aragonite and calcite $\delta^{18}\text{O}$, but later converted to aragonite $\delta^{18}\text{O}$ by measuring the shifts of trace elements (Mg and Sr) between aragonite and calcite. Further south of the Maritime Continent, a 13,000-year-long speleothem $\delta^{18}\text{O}$ record from Flores was interpreted to reflect the variation of the Australia-Indonesia summer monsoon, showing wetter conditions during the YD followed by early-Holocene monsoon strengthening coinciding with Sunda Shelf flooding ~9500 years ago (Griffiths et al., 2010b; Griffiths et al., 2009). This interpretation of speleothem $\delta^{18}\text{O}$ as a hydroclimate proxy was supported by the covariation of speleothem $\delta^{18}\text{O}$ and Mg/Ca and Sr/Ca from the same stalagmite (Griffiths et al., 2010b). Covariation of speleothem $\delta^{18}\text{O}$, Mg/Ca, and also $\delta^{13}\text{C}$ in late-Holocene stalagmites from Flores further indicated that the variations of the three proxies were driven by changes in cave recharge due to varying monsoon strengths (Griffiths et al., 2012; Griffiths et al., 2016). This Flores speleothem $\delta^{18}\text{O}$ was later extended to cover the past 30,000 years, and further demonstrated drier conditions during the B/A and wetter conditions during H1, suggesting that ITCZ migration likely dominated regional hydroclimate changes during the last deglaciation (Ayliffe et al., 2013). Independently derived speleothem $\delta^{18}\text{O}$ records from western Flores and eastern Java show high degree of similarity in

both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ variations for the past 47,000 years (Westaway et al., 2007). $\delta^{13}\text{C}$ profile from the same eastern Java speleothem shows a much larger magnitude of glacial-interglacial variation than the corresponding speleothem $\delta^{18}\text{O}$, yet the two isotopic ratios generally covaried during the past 47,000 years, except for the decoupling during the deglaciation (Westaway et al., 2007). Lastly, fluid inclusion water isotope study of Flores speleothems suggested that cave temperature during the YD was 5°C cooler than the present (Griffiths et al., 2010c).

1.4.3 A summary of speleothem $\delta^{18}\text{O}$ records in the Maritime Continent

Speleothem $\delta^{18}\text{O}$ records from Borneo and Sumatra located in central and western parts of the Maritime Continent, respectively, share similar variations over both glacial-interglacial and millennial timescales. In the southern portion of the region, Flores speleothem $\delta^{18}\text{O}$ record, albeit sharing a similar but smaller glacial-interglacial variation, has overall much higher $\delta^{18}\text{O}$ values and is opposite in the patterns of deglacial millennial-scale variation compared with Borneo and Sumatra speleothem $\delta^{18}\text{O}$. Another speleothem $\delta^{18}\text{O}$ from eastern Java, located also towards the south, resembles the Flores record in both speleothem $\delta^{18}\text{O}$ values and their variations. The spatial distribution of speleothem $\delta^{18}\text{O}$ in the Maritime Continent suggests the existence of broad, homogeneous moisture convection in the region, but different hydroclimate regimes between the northern and southern parts. The opposite patterns in the deglacial millennial scale events between the north and south were likely due to the latitudinal migration of the ITCZ (Aylinffe et al., 2013; Wang et al., 2007), the southern extent of which probably located between the northern and southern cave sites. To further confirm such meridional demarcation in

hydroclimate patterns, and potentially delineate past mean position of the ITCZ in the Maritime Continent, I propose to reconstruct a new speleothem $\delta^{18}\text{O}$ record from caves in southwestern Sulawesi, a location in-between Sumatra-Borneo and eastern Java-Flores.

1.5 Field trip and speleothem samples

During the summer of 2012, a team of researchers from the Earth Observatory of Singapore (EOS), Nanyang Technological University, Singapore and Brown University, USA, conducted a field survey of caves in both central and southwestern Sulawesi, through collaboration with colleagues from Bandung Institute of Technology and Hasanuddin University, Indonesia. The cave sites in central Sulawesi were located near Towuti Lake, in East Luwu, and North Konawe Regencies. Cave sites in southwestern Sulawesi are located near the Bantimurung-Bulusaraung National Park in the Maros-Pangkep karst region, which is the southern end of Tonasa Limestone Formation, deposited between the Eocene and the middle Miocene (Wilson and Bosence, 1996). The Maros-Pangkep karst covers an area of $\sim 450 \text{ km}^2$ parallel to the west coast of the island, with rivers cutting into the basal limestone, and forming clusters of plateau-like karst towers that rise abruptly from surrounding alluvial plains (Aubert et al., 2014; McDonald, 1976). The tower karst and the alluvial plain form a spectacular landscape, with extensive networks of footcaves at the tower bases.

In total, 15 caves and rock shelters were surveyed in the Towuti and Maros regions. In central Sulawesi, most caves were small and partially submerged in water and mud. Cave carbonate formations were poorly developed, and hence a

limited number of naturally fallen stalagmites were collected. In southwestern Sulawesi, most of the visited caves had been badly vandalized, and broken stalagmite pieces were randomly scattered in cave chambers. In total, 33 speleothem samples were collected from a total of 10 lowland caves from both regions (Table 1.1). The samples were carefully selected after considering both cave conservation and scientific research potential; most of the samples were already broken and simply collected from the cave floors. These samples typically had excellent calcite/aragonite crystals based on visual observation, and were mostly columnar. They were later brought to the EOS by Prof. Satria Bijaksana and Ms. Nurul Aufa from Bandung Institute of Technology. I, however, was not able to join the survey because it had taken place before the commencement of my Ph.D. program. Further cave surveys and cave monitoring in Sulawesi were, unfortunately, hindered by research visa restrictions due to a parallel field project in Indonesia and by highly-demanding bureaucratic paperwork. I thus naturally decided to work on the collected speleothem samples only as a more realistic strategy. I have analyzed the collected speleothem samples to constrain the ages of the speleothem carbonates, and also to obtain geochemical data such as stable isotope ratios and trace element concentrations. The Sulawesi speleothems have yielded over 200 dates and 1600 stable isotope data points. In this thesis, I only discuss the datasets that are used to reconstruct the multi-proxy record covering the past 30,000 years. Further analyses on the other samples will be continuously done when the project moves into the next phase.

In November 2016, as a main participant, I joined another cave survey in central Vietnam in collaboration with the Institute of Geological Survey,

Vietnam Academy of Science of Technology. The work on Vietnam speleothems was part of our group effort to isotopically map the hydroclimate changes in Southeast Asia. Till now, I have obtained over 80 U-Th dates along with 600 low-resolution stable isotope data points. The preliminary results are discussed in the Conclusion.

Table 1.1 Stalagmite samples collected from Sulawesi. The IDs of the stalagmites used to reconstruct the geochemical records spanning the past 30,000 years are highlighted in bold. The scanned images of these samples are also shown in the appendix A

Cave Name	Latitude	Longitude	Elevation	Sample ID	Length × Diameter (cm)
			(m)		
Gua Inang	S2°36'45.5"	E121°16'31.2"	420	INA01	11×7
				INA02	8×4 ×
				INA03	21×6
				INA04	19×6
				INA05	16×2
Gua Polihe	S2°55'5.3"	E121°38'10.6"	400	POL01	6×3
Gua Babangan Batu	S3°1'27.6"	E121°39'33.7"	440	BBG01	16×5
				BBG02	14×10
Gua Sariapa	S5°2'33.4"	E119°42'11.5"	50	SAR01	8.5×6
				SAR02	11.5×5
				SAR03	22.5×4
				SAR04	14×5
Gua Suleman	S5°2'45.0"	E119°42'33.3"	100	SLM01	26×7
				SLM02	14×7
				SLM03	28×4
				SLM04	20×3
Gua Salleh	S5°2'59.9"	E119°43'17.5"	150	SAL01	23×5
				SAL02	30×4
				SAL03	32×4
Leang Lomboa	S4°49'59.9"	E119°35'15.3"	20	LOM01	8.5×3
				LOM02	14×3
Leang Bumbung Parea	S4°50'4.9"	E119°35'21.2"	20	BUM01	19×4
				BUM02	19×4
				BUM03	20×3.5
Gua Kharisma	S5°1'34.7"	E119°44'33.5"	300	KHA01	12×7
				KHA02	18×8
				KHA03	15×2.5
				KHA04	38×4
Gua Batu Depan	S5°0'40.9"	E119°41'18.8"	180	BD01	61×4
				BD02	18×4
				BD03	17×6
				BD04	40×5

1.5 Thesis Outline

The objectives of this thesis are: (1) to reconcile the controversies regarding hydroclimate changes in the Maritime Continent, on both glacial-interglacial and millennial timescales, for a better understanding of the driving mechanism; (2) to understand how regional and global climate changes have influenced the vegetation in the region since the LGM; and (3) to further evaluate the driving mechanisms of tropical and subtropical hydroclimate changes since the LGM.

The subsequent chapters are structured as follows:

In Chapter 2, I present a high-resolution regional hydroclimate record for the past 30,000 years with speleothem $\delta^{18}\text{O}$ from southwestern Sulawesi, Indonesia. I demonstrate that regional hydroclimate changed on various time scales, due to both convection strength and meridional migration of the ITCZ over the Maritime Continent. By comparing the Sulawesi speleothem $\delta^{18}\text{O}$ data with existing speleothem records from the region, I further derive an index that can be used to quantify ITCZ migration over the past 30,000 years. The ITCZ index allows a careful evaluation on the driving forces of regional hydroclimate spanning different timescales.

In Chapter 3, I reconstruct the vegetation and soil CO_2 production changes over the Maritime Continent spanning the past 30,000 years, with high-resolution speleothem carbon isotope and trace elements records from southwestern Sulawesi. I show that tropical vegetation and soil CO_2 production was tightly coupled with regional rainfall amount during both glacial and

Holocene periods. During the last deglaciation, both vegetation and soil CO₂ production were primarily controlled by large changes in climate boundary conditions such as atmospheric CO₂ concentration and sea surface temperature, but were decoupled from regional rainfall amount that is associated with ITCZ migration dictated by changes in the Atlantic Meridional Ocean Circulation (AMOC).

In Chapter 4, I synthesize speleothem $\delta^{18}\text{O}$ records from the low latitudes to explore the driving forces of tropical and subtropical hydroclimate changes since the LGM. Similar to what has been observed in modern rainfall $\delta^{18}\text{O}$, I find that the continental rainout effect has driven the spatial variation of speleothem $\delta^{18}\text{O}$ during the past 30,000 years, with corresponding isotopic fractionation affected by land temperature and continental water recycling. I also find that speleothem $\delta^{18}\text{O}$ records largely followed the glacial-interglacial variation of sea surface temperature in the upstream monsoonal and maritime regions, while they closely tracked the latitudinal migration of the ITCZ in the regions downstream of the monsoons and south of the ITCZ. Using Empirical Orthogonal Function (EOF) analysis on compiled speleothem records, I further demonstrate that tropical and subtropical speleothem $\delta^{18}\text{O}$ variations have been primarily driven by ITCZ migration since the LGM. Finally, I highlight the potential use of speleothem $\delta^{18}\text{O}$ isoscapes to hint at patterns and mechanisms of past climate changes during various periods and on various timescales.

Chapter 5 gives the general conclusions for the thesis together with suggestions for future work.

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Chapter 2

Migrations of the Inter-tropical Convergence Zone since the Last Glacial Maximum

2.1 Abstract

Reconstruction of past Inter-tropical Convergence Zone (ITCZ) migrations can greatly help constrain future hydroclimate change in the tropics. Such reconstruction, however, remains rare, mainly due to the difficulty of extracting the ITCZ's dynamics from paleoclimate records. Here I reconstruct migrations of the mean ITCZ position during the past 30,000 years by comparing speleothem $\delta^{18}\text{O}$ -based hydroclimate records from northern Borneo, central Sumatra, southwestern Sulawesi, and Flores in the equatorial Maritime Continent. The newly resolved Sulawesi speleothem $\delta^{18}\text{O}$ has been comparable to other speleothem $\delta^{18}\text{O}$ from further north but continuously lower than the nearby Flores speleothem $\delta^{18}\text{O}$ from further south, with anti-phased millennial-scale variability between the northern and southern sites. Such north-south demarcation reveals that the mean ITCZ over the Maritime Continent has been shifting back and forth within a narrow zone of $\sim 3^\circ$ between southwestern Sulawesi and Flores during the past 30,000 years. The mean ITCZ position has shifted primarily in response to precession change on the orbital timescale, coupled with influences of glaciation and millennial-scale forcings. The ITCZ's convection strength has largely followed the variation of climate boundary conditions, with glacial-interglacial enhancement of regional moisture convection coinciding with surface ocean warming and regional sea level rising since the Last Glacial Maximum.

2.2 Introduction

The Inter-tropical Convergence Zone (ITCZ), as the ascending branch of the Hadley Circulation, is a narrow band of persistent low-pressure convergence, deep convective clouds, and intense rainfall (Waliser and Gautier, 1993). The global mean ITCZ is located around 6°N. Seasonally, the ITCZ migrates toward the summer hemisphere following the march of the sun, thus regulating rainfall variability in the tropics. On longer timescales, the mean ITCZ also migrates, typically towards a warming hemisphere, driven by variations of the interhemispheric temperature contrast (Chiang and Friedman, 2012; Schneider et al., 2014). For instance, the southward migration of the mean ITCZ was reconstructed as dry intervals in the Asian monsoon region (Wang et al., 2001) and the Cariaco Basin off the coast of Venezuela (Peterson et al., 2000), and as wet periods in the Amazon Basin (Cheng et al., 2013b; Wang et al., 2017) and Brazil (Wang et al., 2004; Wang et al., 2006) during millennial-scale events of the last glaciation when the northern high latitudes cooled. Similarly, the southward shift of the ITCZ was indicated by hemispheric seesaw patterns of hydroclimate changes during the Little Ice Age (LIA) (Haug et al., 2001; Sachs et al., 2009).

The ITCZ shifts for the above-mentioned periods can be as large as 5-7° on a regional scale, as suggested by syntheses of transects of marine hydroclimate records (Arbuszewski et al., 2013; Reimi and Marcantonio, 2016; Sachs et al., 2009). In contrast, $\leq 1^\circ$ shifts of the zonal mean ITCZ for the same periods have been estimated using a combination of past tropical sea surface temperature (SST) gradients and present-day linear relationship between ITCZ's position

and tropical SST gradients (McGee et al., 2014). The 5-7° shifts are likely seasonal rather than shifts in the mean ITCZ position, because sea surface productivity fluxes in marine sediments are seasonally biased (Arbuszewski et al., 2013), and because large mean ITCZ shifts may be difficult to maintain even under orbital forcing (Donohoe et al., 2013). Seasonal ITCZ migration might differ widely between specific locations due to substantial seasonal and zonal heterogeneity in the ITCZ's response to past climate changes (McGee et al., 2014), whereas the mean ITCZ shifts are likely to be zonally homogeneous due to comparable hydroclimate responses from different parts of the globe over orbital to centennial timescales (Schneider et al., 2014; Wang et al., 2006). In addition, the mean ITCZ migration also differs from seasonal ITCZ migration in terms of the forcing energetic framework. For instance, seasonal shifts are forced dominantly by changes of the Hadley Circulation, while migrations of the long-term mean ITCZ over the glacial-interglacial timescale are forced not only by the Hadley Circulation, but also by stationary and transient eddies associated with elevated greenhouse gas concentrations (Hill et al., 2015; Roberts et al., 2017).

Although ITCZ migration controls tropical rainfall variability on the global scale, locally, precipitation is also influenced by a variety of other factors, including the varying convection strength of the ITCZ, zonal circulation changes, and local hydrology (Boos and Korty, 2016; McGee et al., 2014; Roberts et al., 2017; Singarayer et al., 2017). For instance, the aforementioned hemispheric hydroclimate changes in paleorecords have been explained by equatorward contraction or poleward expansion of the ITCZ's coverage rather than by unidirectional ITCZ migration (Griffiths et al., 2016; Yan et al., 2015),

with the contraction/expansion likely associated with varying convection strengths of the ITCZ. Isolating the influence of ITCZ migration in hydroclimate reconstructions is hence challenging. To better constrain the mean ITCZ position and its dynamics in the past thus requires comparisons among tropical paleorecords that first, robustly represent regional rainfall change; second, record shifts in the mean position of the ITCZ rather than its seasonal positions; and third, originate from geographically nearby but climatically distinct locations.

Here I reconstruct past ITCZ variability, utilizing speleothem $\delta^{18}\text{O}$ records from the Maritime Continent. Speleothem $\delta^{18}\text{O}$ has been widely used to reconstruct past rainfall change in the tropics (Cheng et al., 2009; Cheng et al., 2016b; Hu et al., 2008a; Wang et al., 2008; Wang et al., 2001), since it mainly inherits weighted mean rainfall $\delta^{18}\text{O}$ (Cruz et al., 2005b; Moerman et al., 2014), the variation of which is controlled by integrated regional atmospheric moisture circulation (Dansgaard, 1964; Liu et al., 2014; Moerman et al., 2014; Pausata et al., 2011; Rozanski et al., 1993; Wang et al., 2001). The robustness of speleothem $\delta^{18}\text{O}$ as a hydroclimate proxy is further reinforced by replicable speleothem $\delta^{18}\text{O}$ variation on a regional scale (Cheng et al., 2012a).

Previously, three speleothem $\delta^{18}\text{O}$ records have been reported in the Maritime Continent, namely from northern Borneo, central Sumatra, and Flores (Figure 2.1). At present, the ITCZ passes over northern Borneo and central Sumatra twice a year, bringing double peaks of rainfall amount, with the annual rainfall exceeding 3,000 mm in northern Borneo and 2,000 mm in central Sumatra (Figure 2.2). Flores has an annual rainfall amount similar to central

Sumatra, but shows much stronger rainfall seasonality. Rainfall $\delta^{18}\text{O}$ from the three locations shows seasonality similar to that of individual rainfall amounts (Figure 2.3), with northern Borneo and central Sumatra exhibiting little seasonal variation in rainfall $\delta^{18}\text{O}$ compared with much amplified seasonal variation in Flores. Rainfall $\delta^{18}\text{O}$ from northern Borneo and central Sumatra is overall lower compared with rainfall $\delta^{18}\text{O}$ from Flores, showing a 3-4‰ difference in amount-weighted rainfall $\delta^{18}\text{O}$. A similar 3-4‰ offset exists between present-day Borneo/Sumatra and Flores speleothem $\delta^{18}\text{O}$. Reasons for these offsets could be potentially traced via satellite observations, which show that the large-scale ITCZ convective system annually passes over northern Borneo, central Sumatra, and areas as far south as SW Sulawesi, but barely reaches the Lesser Sunda Islands, where Flores is located (Figure 2.4 and 2.5). Rainfall $\delta^{18}\text{O}$ from SW Sulawesi during austral summer, as well as its weighted mean are indeed comparable to those of northern Borneo and central Sumatra but lower than rainfall $\delta^{18}\text{O}$ in Flores. Such agreement between the ITCZ's coverage and rainfall $\delta^{18}\text{O}$ distribution in the Maritime Continent strongly suggests that rainfall $\delta^{18}\text{O}$ from the region is modulated by the position and convection strength of the ITCZ.

Figure 2.1 Cave locations and deep convection in the Maritime Continent. Top, cave sites in Northern Borneo (light green), central Sumatra (purple), Southwestern Sulawesi (red), and Flores (blue). The background topographic map is based on grids data from Global Multi-Resolution Topography (GMRT) synthesis (Ryan et al., 2009), with the white area showing exposed continental shelf during the Last Glacial Maximum (LGM). Also shown are the major lands and seas in the Maritime Continent. Bottom, seasonal precipitation (shading, in millimeters per hour) and surface winds (vectors, in meters per second) over the Maritime Continent, averaged over June to August (JJA, left) and December to February (DJF, right) respectively. The precipitation data is based on the Tropical Rainfall Measuring Mission (TRMM) 3B43 dataset (1998-present). The wind data is based on the National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010) (Kalnay et al., 1996).

Figure 2.2 Monthly averaged precipitation for Northern Borneo (Mulu Meteorological Station, 30m a.s.l., 2007-2012, from (Moerman et al., 2013; Moerman et al., 2014)), central Sumatra (Gaw, 865m a.s.l., 2001-2006, Global Precipitation Climatology Project (GPCP)), Southwestern Sulawesi (Makassar, 14m a.s.l., 2002-2006, GPCP), and Bali (Denpasar, 1m a.s.l., 2002-2006, GPCP).

Figure 2.3 Amount-weighted monthly rainfall $\delta^{18}\text{O}$ for Northern Borneo (Mulu Meteorological Station, 30m a.s.l., 2007-2012, from (Moerman et al., 2013; Moerman et al., 2014)), central Sumatra (Gaw, 865m a.s.l., 2001-2006, (Kurita et al., 2009)), Southwestern Sulawesi (Makassar, 14m a.s.l., 2002-2006, (Kurita et al., 2009)), and Bali (Denpasar, 2002-2006, 1m a.s.l., (Kurita et al., 2009)). The dotted lines show the amount-weighted annual means for rainfall $\delta^{18}\text{O}$ at each site. The central Sumatra weather station of Gaw is located in relatively higher altitude, which likely causes more negative rainfall $\delta^{18}\text{O}$ than that of SW Sulawesi.

Figure 2.4 The annual mean ITCZ over the Maritime Continent as shown by the distribution of Highly Reflective Clouds (HRC) (between 1971-1985). The HRC covers the major lands and seas of the region, yet do not reach the small islands such as the Lesser Sunda Islands. The HRC data is provided by the National Oceanic & Atmospheric Administration (NOAA)/Earth System Research Laboratory (ESRL)/Physical Sciences Division (PSD), Boulder, Colorado, USA, from <https://www.esrl.noaa.gov/psd/data/gridded/data.noaa.hrc.html>. Also shown are the cave sites in Northern Borneo (light green), central Sumatra (purple), Southwestern Sulawesi (red), and Flores (blue).

Figure 2.5 The annual mean ITCZ over the Maritime Continent as shown by the distribution of precipitation. Relatively strong precipitation occurs over the majority of the Maritime Continent, excluding the Lesser Sunda Islands in the south. The precipitation data is based on the Tropical Rainfall Measuring Mission (TRMM) 3B43 dataset (1998-present). The wind data is based on the National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010) (Kalnay et al., 1996). Also shown are the cave sites in Northern Borneo (light green), central Sumatra (purple), Southwestern Sulawesi (red), and Flores (blue).

The N-S demarcation of rainfall/speleothem $\delta^{18}\text{O}$ and convection strength in the Maritime Continent suggests that the southern extent of the mean regional ITCZ is likely located between the northern Maritime Continent and the Lesser Sunda Islands. As a result, any northward (southward) shift of the mean ITCZ should be accompanied by lighter (heavier) rainfall/speleothem $\delta^{18}\text{O}$ over the northern Maritime Continent until areas as far south as SW Sulawesi, along with heavier (lighter) $\delta^{18}\text{O}$ in the Lesser Sunda Islands and further south. Patterns might have differed in the past: if this mean ITCZ migrated north of SW Sulawesi, then both Flores and Sulawesi would have been outside the ITCZ's coverage, and therefore shared comparably higher rainfall/speleothem $\delta^{18}\text{O}$; if the mean ITCZ migrated south of the Lesser Sunda Islands, both Sulawesi and Flores would have been covered by the mean ITCZ, and hence shared similarly lower rainfall/speleothem $\delta^{18}\text{O}$ with northern Borneo and central Sumatra. Considering the proximity of SW Sulawesi and Flores, ITCZ migration is likely the main forcing that differentiates the rainfall/speleothem $\delta^{18}\text{O}$ in the two locations. The mean position of the ITCZ and its dynamics in the past can thus be constrained by comparing Flores speleothem $\delta^{18}\text{O}$ with a new speleothem $\delta^{18}\text{O}$ record from SW Sulawesi.

2.3 Methods

SW Sulawesi speleothem $\delta^{18}\text{O}$ record is based on five broken stalagmites from three adjacent caves in the Maros Regency of SW Sulawesi, Indonesia. The cave bedrocks are composed of Eocene to middle Miocene limestone. I cut and polished the five stalagmites, which all show clear calcite laminae. I determined the chronology of SW Sulawesi speleothem record from a total of 67 U-Th dates obtained with a multi-collector inductively coupled plasma mass

spectrometric (MC-ICP-MS) dating technique. The chemical procedures for uranium and thorium separation and purification were similar to those described in (Cheng et al., 2000; Edwards et al., 1987). U-Th dating analyses were performed on a Neptune plus housed in the isotope geochemistry laboratory at the Earth Observatory of Singapore and the Asian School of the Environment, Nanyang Technological University, with the dating techniques largely following the methods described in (Chiang et al., 2019). Except for ^{238}U and ^{232}Th , all the other U and Th isotopes were measured on a secondary electron multiplier (SEM) equipped with a retarding potential quadruple lenses (RPQ) for abundance sensitivity improvement in a peak-jumping mode. Sulawesi stalagmites typically have a uranium concentration of several hundred parts per billion (ppb), and their measured $^{230}\text{Th}/^{232}\text{Th}$ ranges from tens to thousands of parts per million (ppm). The corrected ^{230}Th ages are calculated assuming an initial $^{230}\text{Th}/^{232}\text{Th}$ ratio of 4.4 ± 2.2 ppm, a value for a material at secular equilibrium with the bulk Earth $^{232}\text{Th}/^{238}\text{U}$ value of 3.8. The uncertainties for the U-Th dates are at 2σ level. The age model for each stalagmite was established by linear interpolation (Figure 2.6). Most of the dates have a precision better than 1%, except for several dates obtained from the top and/or bottom of the broken stalagmites, where detritus contents were relatively high.

Oxygen isotope ratios ($\delta^{18}\text{O}$) of the Sulawesi speleothems were analyzed at the School of Geography Science, Nanjing Normal University, China. Subsample powders were drilled in 1-5 mm resolution by dental drill bits, and then analyzed on a Finnigan MAT-253 mass spectrometer equipped with a Kiel Carbonate Device III, which shows a long-term reproducibility of $\sim 0.1\text{‰}$ (1σ).

A total of 1412 oxygen isotope data points were obtained, comprising 69, 309, 45, 533, and 456 analyses for samples SAL03, SAL02, SLML04, BD01, and BD02, respectively. These $\delta^{18}\text{O}$ data points yielded an average temporal resolution of ~ 30 years. Additional $\delta^{18}\text{O}$ analyses were done on several subsamples using an on-line carbonate preparation system (GasBench II) interfaced with an isotope ratio mass spectrometer (Delta V) in the Environmental Geochemistry Laboratory at the Earth Observatory of Singapore and the Asian School of the Environment, Nanyang Technological University. The long-term reproducibility is $\sim 0.1\text{‰}$ (1σ). The data obtained are consistent with those from Nanjing Normal University. Stable isotopic data are reported in per mil (‰), relative to the Vienna Pee Dee Belemnite (VPDB) standard: ($\delta^{18}\text{O} = [({}^{18}\text{O}/{}^{16}\text{O})_{\text{sample}} / ({}^{18}\text{O}/{}^{16}\text{O})_{\text{standard}} - 1] \times 1,000\text{‰}$).

Figure 2.6 Age models for stalagmites SAL3, SAL2, BD1, BD2, and SLML4. The chronology of the stalagmites is established by linear interpolation, with dates and age uncertainties (2σ) shown in black dots and red bars respectively.

2.4 Results

Among the five stalagmites, two specimens (BD01 and BD02) from Batu Depan Cave (S5°1', E119°41') grew during 26.0–11.0 thousand years before present (kyr BP, present is 1950 A.D.) and 29.8–14.9 kyr BP, respectively (Figure 2.7). Two other samples (SAL02 and SAL03) from Salleh Cave (S5°3', E119°43') grew during 10.7–1.1 kyr BP and 4.0–0.1 kyr BP, respectively. The fifth sample, SLML04 from Suleman Cave (S5°3', E119°42'), grew discontinuously during 19.7–18.8 kyr BP and 6.3–4.1 kyr BP. A spliced $\delta^{18}\text{O}$ record can then be established for the past 30,000 years, with a short gap between 11.0 and 10.7 kyr BP. Overall, SW Sulawesi speleothem $\delta^{18}\text{O}$ shows higher values during the last glacial period, offset by $\sim 3.5\text{‰}$ compared to lower values in the Holocene. During the glacial period, $\delta^{18}\text{O}$ values oscillated between -6.5‰ and -7.5‰ between ~ 30 and 20 kyr BP, with positive millennial-scale excursions centered ~ 28 , 27, and 24.8 kyr BP. Afterwards, $\delta^{18}\text{O}$ gradually rose towards a maximum value of -5.8‰ ~ 15 kyr BP. $\delta^{18}\text{O}$ then broadly decreased during the last deglaciation, interrupted only by a positive excursion ~ 12.5 kyr BP. The decreasing trend lasted until the mid-Holocene (~ 6 –4 kyr BP), reaching -9.2‰ . In the late Holocene, $\delta^{18}\text{O}$ values sharply increased by 1‰ at ~ 3.5 kyr BP, accompanied by a nearly halted growth of samples SAL02 and SAL03 until ~ 2 kyr BP. Since then, the $\delta^{18}\text{O}$ has oscillated between -8.2‰ and -9‰ until the present day. In general, the SW Sulawesi speleothem $\delta^{18}\text{O}$ is characterized by relatively gradual shifts on both orbital and millennial time scales, with a high degree of overlap in $\delta^{18}\text{O}$ values between different samples over contemporaneous growth periods.

Figure 2.7 SW Sulawesi speleothem $\delta^{18}\text{O}$ in comparisons with insolation, Indo-Pacific Warm Pool (IPWP) SST, and relative sea level. Top, the Sulawesi record is composed of data from five stalagmite samples, with different colors assigned for different stalagmite. U-Th dates and age uncertainties are shown by colored error bars (2σ). The red dot along the y axis show the calcite $\delta^{18}\text{O}$ calculated based on present-day amount weighted rainfall $\delta^{18}\text{O}$ from SW Sulawesi (Kurita et al., 2009) and mean annual cave temperature of 25°C following $1,000 \times \ln\alpha(\text{calcite-H}_2\text{O}) = 17.66 \times (10^3 \times \text{T}^{-1}) - 30.16$, where α is the isotopic fractionation factor between calcite and water and T is the temperature (in kelvin) (Johnston et al., 2013). The agreement between speleothem calcite $\delta^{18}\text{O}$ and the value predicted from modern rainfall $\delta^{18}\text{O}$ supports the arguments that cave calcite from SW Sulawesi is deposited under equilibrium conditions, and the speleothem $\delta^{18}\text{O}$ robustly records rainfall $\delta^{18}\text{O}$ signal. Also shown is the adjusted speleothem $\delta^{18}\text{O}$ (grey) after accounting for ice-volume-related changes in global mean seawater $\delta^{18}\text{O}$ from (Waelbroeck et al., 2002) in comparison with September-October-November (SON) insolation (purple) from 5°S (Laskar et al., 2004). Center, Mg/Ca-based IPWP sea surface temperature reconstructions GeoB10069-3 (light blue line) (Gibbons et al., 2014) and SO217-18159 (blue scatter-line) and -18522 (dark blue scatter-line) (Schröder et al., 2018). Bottom, relative sea level reconstructed based on benthic $\delta^{18}\text{O}$ (cyan shading) (Waelbroeck et al., 2002), global distributed corals (dark cyan dots) (Thompson and Goldstein, 2006), and regional offshore sediments (purple dots) (Geyh et al., 1979; Hanebuth et al., 2000; Hanebuth et al., 2009).

2.5 Discussion

2.5.1 The interpretation of SW Sulawesi speleothem $\delta^{18}\text{O}$

The remarkable replications of our SW Sulawesi speleothem $\delta^{18}\text{O}$ profiles between different stalagmites and caves strongly suggest that cave calcite deposited under equilibrium conditions with negligible kinetic isotope fractionation (Hendy, 1971). Comparisons among speleothem records from SW Sulawesi, northern Borneo, and central Sumatra also show strikingly similar calcite $\delta^{18}\text{O}$ values throughout the past 30,000 years despite rather different rainfall seasonality and moisture sources between the locations (Figure 2.8), with the Borneo record being slightly lower compared with the other two records, probably due to the mountainous terrain and relatively higher rainfall amount at the cave site. The Flores speleothem $\delta^{18}\text{O}$, although sharing a broadly similar glacial-interglacial pattern with the other three records, positively offset from the other records by 1-2‰ during the glacial period and by 3-4‰ during the Holocene. Such a spatial pattern of speleothem $\delta^{18}\text{O}$ is supported by modern rainfall $\delta^{18}\text{O}$ observations from corresponding locations, with the weighted mean rainfall $\delta^{18}\text{O}$ from SW Sulawesi, northern Borneo, and central Sumatra being consistently lower by 3-4‰ than the weighted mean rainfall $\delta^{18}\text{O}$ from Bali in the Lesser Sunda Islands (Figure 2.3). Present-day SW Sulawesi speleothem $\delta^{18}\text{O}$ also agrees with calcite $\delta^{18}\text{O}$ predicted from the weighted mean rainfall $\delta^{18}\text{O}$ in SW Sulawesi. Together, these lines of evidence demonstrate that the SW Sulawesi speleothem $\delta^{18}\text{O}$ record, along with the other three speleothem $\delta^{18}\text{O}$ records from the Maritime Continent, faithfully records

changes in the weighted mean rainfall $\delta^{18}\text{O}$, and is dominated by regional climate signal rather than site-specific influences.

Additionally, I argue that, in accordance with previous speleothem $\delta^{18}\text{O}$ interpretations (Carolin et al., 2013; Cobb et al., 2007; Partin et al., 2007; Wurtzel et al., 2018), that the four speleothem $\delta^{18}\text{O}$ records reflect mainly the varying strengths of moisture convection and hence rainfall amount in the region. First, the spatial distributions of speleothem and rainfall $\delta^{18}\text{O}$ in the Maritime Continent coincide with the distributions of large-scale convective systems and regional rainfall amount, as shown by satellite observations (Figure 2.4 and 2.5). And second, a strong correlation between rainfall $\delta^{18}\text{O}$ and regional rainfall amount has been confirmed by cave monitoring studies and modern rainfall $\delta^{18}\text{O}$ observations across the region, with seasonal minima of drip water and rainfall $\delta^{18}\text{O}$ coinciding with the passage of the ITCZ (Belgaman et al., 2017; Cobb et al., 2007; Moerman et al., 2013; Suwarman et al., 2013; Wurtzel et al., 2018). On intra-seasonal timescales, anomalies of drip water/rainfall $\delta^{18}\text{O}$ also commonly coincide with rainfall anomalies associated with the variability of the Madden-Julian Oscillation (MJO), with a similar relationship existing for the El Niño/Southern Oscillation (ENSO) on interannual timescales (Cobb et al., 2007; Moerman et al., 2013; Moerman et al., 2014).

Figure 2.8 Comparisons of speleothem $\delta^{18}\text{O}$ records from the maritime continent, top) during the last 30,000 years; bottom left) during the Holocene; bottom right) during the last deglaciation. The records are from Snail Shell (darkgreen) and Bukit Assam Caves (lightgreen) in northern Borneo (Carolin et al., 2013; Partin et al., 2007), Tangga Cave in central Sumatra (purple) (Wurtzel et al., 2018), Salleh, Batu Depan, Suleman Caves in southwestern Sulawesi (red) (this study) and Liang Lua Cave in Flores (blue) (Ayliffe et al., 2013; Griffiths et al., 2009). Vertical bars manifest common millennial scale events shared by the records, with dry and wet episodes registered in Borneo, Sumatra and Sulawesi records marked by grey and yellow bars respectively. All the speleothem $\delta^{18}\text{O}$ are shown in original calcite $\delta^{18}\text{O}$ except for the Sumatra record, which was shifted -1‰ to transfer original aragonite $\delta^{18}\text{O}$ to calcite $\delta^{18}\text{O}$ (Wurtzel et al., 2018).

2.5.2 Glacial-interglacial variation of deep convection in the Maritime Continent

The speleothem records from northern Borneo, central Sumatra, and SW Sulawesi, and to a lesser degree, from Flores, share a consistent glacial-interglacial trend of $\delta^{18}\text{O}$ decrease. Such glacial-interglacial variation, however, has to be adjusted for the varying sea water $\delta^{18}\text{O}$ (hence rainfall and speleothem $\delta^{18}\text{O}$) and the influences of temperature changes on calcite/water oxygen isotopic fractionation, before interpreting hydroclimate changes. The Sulawesi speleothem $\delta^{18}\text{O}$ has shown a decrease of $\sim 3\text{‰}$ between the LGM and the present day: a 1‰ decrease can be attributed to lighter seawater $\delta^{18}\text{O}$ as a result of ice melting (Schrag et al., 1996); another $\leq 0.8\text{‰}$ decrease can be attributed to calcite/water $\delta^{18}\text{O}$ fractionation based on $\leq 3.5\text{ °C}$ of surface ocean warming over the Maritime Continent (Mohtadi et al., 2014; Rosenthal et al., 2003; Stott et al., 2002; Visser et al., 2003). The remaining $\delta^{18}\text{O}$ decrease attributable to hydroclimate changes was probably no more than 1‰ , if we further remove a small contribution from regional seawater $\delta^{18}\text{O}$ variation (Gibbons et al., 2014; Mohtadi et al., 2014; Rosenthal et al., 2003; Schröder et al., 2016b; Stott et al., 2002). The less than 1‰ speleothem $\delta^{18}\text{O}$ decrease in SW Sulawesi suggests a slightly intensified regional ITCZ convection from the LGM to the Holocene, coinciding with the variation of climate boundary conditions such as rising sea surface temperature (Mohtadi et al., 2011; Visser et al., 2003), increasing atmospheric CO_2 concentration (Carolin et al., 2013; Monnin et al., 2001), and Sunda Shelf flooding (Bird et al., 2005; Hanebuth et al., 2000; Hanebuth et al., 2009; Hanebuth et al., 2011).

Alternative explanations have been suggested, linking the orbital-scale variation of speleothem $\delta^{18}\text{O}$ and regional convection strength to either equinox insolation (Carolin et al., 2013; Partin et al., 2007), or rejecting any connection between regional hydroclimate and Sunda Shelf flooding based on leaf wax $\delta^{13}\text{C}$ record from a lake sediment core (Russell et al., 2014). I have found that the correlation between speleothem $\delta^{18}\text{O}$ and equinox insolation was rather weak in the Maritime Continent. For instance, the LGM and the present day share a similar equinox insolation, but their speleothem $\delta^{18}\text{O}$ values differ considerably. On the other hand, when equinox insolation varied regularly from 30 to 20 kyr BP, both speleothem $\delta^{18}\text{O}$ and SST were relatively stable. The linkage between regional convection and Sunda Shelf flooding cannot be fully supported or rejected as my record is not long enough, although leaf wax $\delta^{13}\text{C}$ may more likely record vegetation changes associated not only with rainfall amount but also with other factors such as rainfall seasonality, temperature, and atmospheric $p\text{CO}_2$ (see Chapter 3 for more discussions).

2.5.3 The ITCZ in the Maritime Continent

On millennial timescales, the Maritime-Continent speleothem $\delta^{18}\text{O}$ records also shared common variability during the last deglaciation, with the records from SW Sulawesi, northern Borneo, and central Sumatra documenting relatively higher $\delta^{18}\text{O}$ values during H1 and the YD, but slightly lower values during the Bølling-Allerød event (B/A). The magnitudes of these millennial episodes were attenuated compared to those registered in eastern China speleothem records (e.g., Wang et al., 2001), suggesting the deep convection in the Maritime Continent was less sensitive to the North Atlantic millennial-scale forcing (Carolin et al., 2013; Partin et al., 2007). The same millennial-scale

episodes registered in the Flores speleothem $\delta^{18}\text{O}$ were, however, anti-phased, showing relatively lower $\delta^{18}\text{O}$ values during H1 and the YD, but slightly higher $\delta^{18}\text{O}$ during the B/A. Such an anti-phased relationship between the Borneo/Sumatra/Sulawesi records on the one hand and Flores speleothem $\delta^{18}\text{O}$ on the other also existed in the mid-to-late Holocene millennial scale features. The anti-phased millennial-scale variabilities, coupled with a continuous $\delta^{18}\text{O}$ offset between the Borneo/Sumatra/Sulawesi records and the Flores speleothem record, indicates that the southern extent of the mean ITCZ in the Maritime Continent has been located between SW Sulawesi and the Lesser Sunda Islands (within $\sim 3^\circ$ in latitude) during the past 30,000 years. This southern extent might extend westward to the eastern Java Sea and eastward to the Banda Sea based on modern satellite observations. Another speleothem $\delta^{18}\text{O}$ time series from eastern Java replicates the Flores record in both $\delta^{18}\text{O}$ variation patterns and values (Westaway et al., 2007), corroborating a mean ITCZ position north of the Java-Flores Seas.

2.5.4 ITCZ migration during the past 30,000 years

Given the similar climatology and geographical proximity between SW Sulawesi and Flores, meridional migration of the regional mean ITCZ likely represents the single main forcing that has differentiated the hydroclimate and speleothem $\delta^{18}\text{O}$ between SW Sulawesi and Flores. In comparison, the other major forcings (e.g., insolation, glacial boundary conditions, shelf flooding, as well as those associated with zonal atmospheric circulations) probably influenced the hydroclimate of the two locations comparably, and hence these influences can be largely cancelled out by subtracting the Flores speleothem $\delta^{18}\text{O}$ from the SW Sulawesi speleothem $\delta^{18}\text{O}$. The residue's ($\Delta\delta^{18}\text{O}_{\text{SW-FL}}$)

variation is almost solely attributable to the latitudinal shift of the regional mean ITCZ.

The robustness of $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ as an ITCZ migration index can be tested by comparing it with well-dated ITCZ-sensitive subtropical monsoon records from southern China (Cheng et al., 2016a) and Brazil (Wang et al., 2007) which are responsive to the meridional migration of the ITCZ mean position. The two subtropical monsoon records strongly correlate with the $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ profile from the Maritime Continent ($R=0.57$, $p<0.001$ and $R=-0.67$, $p<0.001$, for southern China and Brazil monsoons, respectively) (Figure 2.9). Such correlation not only confirms the strong sensitivity of subtropical monsoon systems to the latitudinal migration of the ITCZ, but most importantly, also validates that $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ is not biased toward either of the original speleothem $\delta^{18}\text{O}$ from the Maritime Continent, and hence is a reliable index of the latitudinal displacement of the ITCZ.

Figure 2.9 ITCZ migration index ($\Delta\delta^{18}\text{O}_{\text{SW-FL}}$) in comparison with subtropical monsoons and climate forcings. The $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ is derived by subtracting Flores speleothem $\delta^{18}\text{O}$ from SW Sulawesi speleothem $\delta^{18}\text{O}$, with both records transformed to non-overlapping 200-year binned averages before the subtraction. An uncertainty of 1σ is also given in a shaded envelope. $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ (pink) is compared with a) Chinese speleothem $\delta^{18}\text{O}$ (olive) (Cheng et al., 2016a), b) speleothem $\delta^{18}\text{O}$ record from southern Brazil (green) (Wang et al., 2007), c) precession (purple) (Laskar et al., 2004), d) reconstructed AMOC strength in $^{231}\text{Pa}/^{230}\text{Th}$ from the North Atlantic (Böhm et al., 2014; McManus et al., 2004).

$\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ shows active movements of the ITCZ during the past 30,000 years. Broadly, the ITCZ slowly migrated southward between the early MIS2 and ~15 kyr BP, after which it shifted northward in a rapid two-step motion towards the early Holocene. Later, the ITCZ gradually migrated southward again during the Holocene. Superimposed on this glacial-interglacial migratory trend are several prominent millennial-scale excursions of the ITCZ, such as southerly excursions during H1, the YD, ~6.5 kyr BP, and ~3 kyr BP, as well as northerly excursions during the BA and ~11.5 kyr BP. $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ during the LGM was slightly higher than the corresponding value in the present day, suggesting a slightly southward position of the ITCZ. This conclusion agrees well with modeling studies which suggest that the Laurentide ice sheet at its maximum size displaced the ITCZ more southward (Chiang et al., 2003; Chiang and Bitz, 2005). This southward displacement would cause drier conditions in the northern Maritime Continent, while bringing wetter climate to the Lesser Sunda Islands. Nevertheless, these hydroclimate changes were likely minor, given the small offset of $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$. Indeed, the overall small glacial-interglacial variation of the SW Sulawesi and the other speleothem $\delta^{18}\text{O}$ records from the Maritime Continent confirmed that regional convection strength only changes slightly between the LGM and the present day. Such small regional hydroclimate changes could be obscured by local hydrological changes, which might explain the controversies regarding the regional hydroclimate changes during the LGM inferred from existing paleorecords (DiNezio and Tierney, 2013; Russell et al., 2014).

2.5.5 The driving mechanisms of ITCZ migration

Paleoclimate records suggest that interhemispheric thermal contrast, associated with precession cycle, glaciation, and changes in North Atlantic ocean circulations, might have driven the meridional shift of the ITCZ by influencing extra-tropical temperature gradients and the energy export out of the tropics on seasonal scales (Chiang and Friedman, 2012; Schneider et al., 2014). Extra-tropical forcings have also been shown in model simulations to alter the cross-equatorial atmospheric energy transport (AET) (Broccoli et al., 2006; Chiang and Friedman, 2012; Frierson and Hwang, 2012; Kang et al., 2009; Kang et al., 2008). Our ITCZ index validates the above explanations as $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ broadly follows the precession change on orbital timescales, and coincides with reconstructed strength of the Atlantic Meridional Ocean Circulation (AMOC) on millennial timescales (Böhm et al., 2014; McManus et al., 2004), particularly during the deglaciation. Notably, the magnitude of ITCZ displacement on millennial timescales associated with deglacial AMOC variability was larger than the magnitudes driven by boundary condition changes. This agrees with the strong sensitivity of tropical hydroclimate to deglacial millennial-scale forcing found in a wide range of paleoclimate records (Ayliffe et al., 2013; Carolin et al., 2013; Denniston et al., 2013b; Fleitmann et al., 2007; Genty et al., 2003a; Griffiths et al., 2009; Muller et al., 2012; Partin et al., 2007; Shakun et al., 2007; Wang et al., 2007; Wang et al., 2001) as well as with the results of modeling studies and other estimations (McGee et al., 2014; Roberts et al., 2017; Singarayer et al., 2017).

2.5.6 The ITCZ and regional hydroclimate

Comparison of the original SW Sulawesi speleothem $\delta^{18}\text{O}$ record with the ITCZ migration index ($\Delta\delta^{18}\text{O}_{\text{SW-FL}}$) profile suggests that ITCZ migration in fact influenced regional deep convection only weakly (Figure 2.10), with influences most dominant during the millennial-scale events of the last deglaciation. For Flores and areas beyond the coverage of the mean ITCZ, regional hydroclimate closely tracked ITCZ migration. Notably, the Flores speleothem $\delta^{18}\text{O}$ decreased prominently and abruptly during the early Holocene (Griffiths et al., 2009), coinciding with a gradual southward migration of the ITCZ. A similar response is seen in speleothem $\delta^{18}\text{O}$ from northwestern Australia (Denniston et al., 2013b), and has been explained as post-glacial strengthening of the Indonesian-Australian monsoon in response to a southward ITCZ migration. Such amplified hydroclimate responses were probably associated with enhanced water vapor feedback and stronger moisture convection caused by shelf flooding and higher SST during the early Holocene (Denniston et al., 2013b; Griffiths et al., 2009).

Figure 2.10 $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ (grey) and speleothem $\delta^{18}\text{O}$ records from SW Sulawesi (top in red) and Flores (bottom in blue) (Ayliffe et al., 2013; Griffiths et al., 2009)

2.6 Implications for modern climate

As the ITCZ is very sensitive to changes of atmospheric energy balance, our reconstruction of past ITCZ migrations can be used to constrain hydroclimate predictions under future global warming, particularly for regions where slight shifts in seasonal ITCZ can change the hydroclimate drastically (Broecker and Putnam, 2013; Schneider et al., 2014). If I extrapolate the patterns of past ITCZ migration and hydroclimate changes observed in the Maritime Continent into the future warming world, stronger ITCZ convection can be expected due to the dramatic increase in tropical temperatures (Marcott et al., 2013), supporting the ‘wet becomes wetter’ paradigm (Pachauri et al., 2015). On the other hand, increased greenhouse gas concentrations may lead to differential warming of the Northern Hemisphere, and shift the ITCZ northward, bringing a wetter climate to regions north of the ITCZ and drying regions further south (Broecker and Putnam, 2013; Schneider et al., 2014). My study here also highlights the importance of synthesizing paleoclimate records within a regional or even global context for better knowledge of large-scale climate changes (Roberts et al., 2017; Singarayer et al., 2017).

2.7 Conclusion

A new speleothem $\delta^{18}\text{O}$ record from SW Sulawesi, Indonesia, is compared with the existing speleothem $\delta^{18}\text{O}$ records from northern Borneo, central Sumatra, and Flores to study the moisture convection and ITCZ migrations over the Maritime Continent. Both speleothem and rainfall $\delta^{18}\text{O}$ have been persistently lower in northern Borneo, central Sumatra, and SW Sulawesi but higher in Flores. Additionally, speleothem $\delta^{18}\text{O}$ variation on millennial

timescales has been anti-phased between Borneo/Sumatra/Sulawesi and Flores. These pieces of evidence point to a regional ITCZ whose mean position has been migrating between the northern Maritime Continent and the Lesser Sunda Islands during the past 30,000 years. The geographically nearby but distinct speleothem $\delta^{18}\text{O}$ from SW Sulawesi and Flores were utilized to reconstruct the ITCZ migration in the past, confirming the precession control of ITCZ dynamics on the orbital timescale, with superimposed influences of glaciation and millennial AMOC variability.

2.8 References

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Table 2.1 ^{230}Th dating results. The uncertainty is coded on 2σ level.

Sample-depth (mm)	^{238}U (ppb)	^{232}Th (ppt)	$^{230}\text{Th} / ^{232}\text{Th}$ (atomic $\times 10^{-6}$)	$d^{234}\text{U}^*$ (measured)	$^{230}\text{Th} / ^{238}\text{U}$ (activity)	^{230}Th Age (yr) (uncorrected)	^{230}Th Age (yr) (corrected)	$d^{234}\text{U}_{\text{Initial}}^{**}$ (corrected)	^{230}Th Age (yr BP) *** (corrected)
SLML-4									
SLML4-5.5	159 $\pm$ 0.3	1436 $\pm$ 29	84 $\pm$ 3	177.6 $\pm$ 1.9	0.0461 $\pm$ 0.0012	4351 $\pm$ 112	4128 $\pm$193	180 $\pm$ 2	4063 $\pm$193
SLML4-55.5	183 $\pm$ 0.4	206 $\pm$ 4	735 $\pm$ 16	153.1 $\pm$ 1.8	0.0502 $\pm$ 0.0003	4851 $\pm$ 26	4823 $\pm$33	155 $\pm$ 2	4758 $\pm$33
SLML4-75	193 $\pm$ 0.8	3266 $\pm$ 67	60 $\pm$ 1	154.3 $\pm$ 3.5	0.0616 $\pm$ 0.0006	5971 $\pm$ 61	5545 $\pm$307	157 $\pm$ 4	5480 $\pm$307
SLML4-153.5	154 $\pm$ 0.3	77 $\pm$ 2	1937 $\pm$ 51	145.9 $\pm$ 1.7	0.0589 $\pm$ 0.0003	5744 $\pm$ 33	5731 $\pm$35	148 $\pm$ 2	5666 $\pm$35
SLML4-157.5	137 $\pm$ 0.8	582 $\pm$ 12	238 $\pm$ 6	138.9 $\pm$ 4.4	0.0615 $\pm$ 0.0007	6050 $\pm$ 71	5942 $\pm$104	141 $\pm$ 4	5877 $\pm$104
SLML4-173.5	47 $\pm$ 0.3	474 $\pm$ 10	289 $\pm$ 6	99.9 $\pm$ 4.9	0.1775 $\pm$ 0.0016	19139 $\pm$ 214	18872 $\pm$284	105 $\pm$ 5	18807 $\pm$284
SLML4-213.5	101 $\pm$ 0.9	114 $\pm$ 4	2832 $\pm$ 110	169.7 $\pm$ 6.3	0.1928 $\pm$ 0.0021	19554 $\pm$ 256	19526 $\pm$256	179 $\pm$ 7	19461 $\pm$256
SAL-3									
SAL3-9	198 $\pm$ 0.3	228 $\pm$ 5	23 $\pm$ 2	20.4 $\pm$ 1.6	0.0016 $\pm$ 0.0001	168 $\pm$ 13	135 $\pm$27	20 $\pm$ 2	70 $\pm$27
SAL3-45.5	296 $\pm$ 0.6	2025 $\pm$ 41	17 $\pm$ 0	21.7 $\pm$ 1.5	0.0071 $\pm$ 0.0001	766 $\pm$ 9	570 $\pm$138	22 $\pm$ 1	505 $\pm$138
SAL3-82.5	214 $\pm$ 0.6	448 $\pm$ 11	77 $\pm$ 3	22.2 $\pm$ 2.1	0.0098 $\pm$ 0.0003	1047 $\pm$ 36	988 $\pm$55	22 $\pm$ 2	923 $\pm$55
SAL3-162	199 $\pm$ 0.5	528 $\pm$ 12	81 $\pm$ 3	22.3 $\pm$ 2.3	0.0129 $\pm$ 0.0003	1390 $\pm$ 37	1315 $\pm$65	22 $\pm$ 2	1250 $\pm$65
SAL3-217.5	216 $\pm$ 0.5	985 $\pm$ 20	57 $\pm$ 1	18.5 $\pm$ 1.7	0.0159 $\pm$ 0.0001	1712 $\pm$ 15	1581 $\pm$93	19 $\pm$ 2	1516 $\pm$93
SAL3-245	204 $\pm$ 0.5	3480 $\pm$ 70	23 $\pm$ 1	18.8 $\pm$ 2.0	0.0239 $\pm$ 0.0005	2584 $\pm$ 52	2097 $\pm$349	19 $\pm$ 2	2032 $\pm$349
SAL3-261.5	132 $\pm$ 0.3	3144 $\pm$ 63	25 $\pm$ 1	21.1 $\pm$ 2.1	0.0365 $\pm$ 0.0007	3965 $\pm$ 73	3286 $\pm$486	21 $\pm$ 2	3221 $\pm$486
SAL3-344	4516 $\pm$ 9.4	22729 $\pm$ 457	123 $\pm$ 3	20.6 $\pm$ 1.8	0.0376 $\pm$ 0.0002	4096 $\pm$ 24	3952 $\pm$104	21 $\pm$ 2	3887 $\pm$104
SAL-2									
SAL2-3.5	327 $\pm$ 0.8	1303 $\pm$ 27	63 $\pm$ 2	144.1 $\pm$ 2.1	0.0151 $\pm$ 0.0002	1450 $\pm$ 20	1348 $\pm$74	145 $\pm$ 2	1283 $\pm$74
SAL2-3.5	335 $\pm$ 0.4	8139 $\pm$ 30	14 $\pm$ 0	147.2 $\pm$ 2.1	0.0207 $\pm$ 0.0005	1987 $\pm$ 47	1425 $\pm$285	148 $\pm$ 2	1360 $\pm$285
SAL2-36	162 $\pm$ 0.3	635 $\pm$ 14	173 $\pm$ 4	159.5 $\pm$ 1.5	0.0412 $\pm$ 0.0006	3928 $\pm$ 64	3830 $\pm$94	161 $\pm$ 1	3765 $\pm$94
SAL2-36	163 $\pm$ 0.3	949 $\pm$ 20	118 $\pm$ 3	157.8 $\pm$ 1.6	0.0417 $\pm$ 0.0006	3992 $\pm$ 62	3845 $\pm$120	160 $\pm$ 2	3780 $\pm$120
SAL2-61.5	359 $\pm$ 1.0	559 $\pm$ 11	532 $\pm$ 11	154.2 $\pm$ 1.9	0.0503 $\pm$ 0.0003	4850 $\pm$ 34	4811 $\pm$43	156 $\pm$ 2	4746 $\pm$43

Table 2.1 (Continued)

Sample-depth (mm)	^{238}U (ppb)	^{232}Th (ppt)	$^{230}\text{Th} / ^{232}\text{Th}$ (atomic $\times 10^{-6}$)	$d^{234}\text{U}^*$ (measured)	$^{230}\text{Th} / ^{238}\text{U}$ (activity)	^{230}Th Age (yr) (uncorrected)	^{230}Th Age (yr) (corrected)	$d^{234}\text{U}_{\text{Initial}}^{**}$ (corrected)	^{230}Th Age (yr BP) *** (corrected)
SAL2-73.5	370 $\pm$ 1.3	1856 $\pm$ 38	184 $\pm$ 4	154.9 $\pm$ 2.3	0.0562 $\pm$ 0.0003	5433 $\pm$ 31	5307 $\pm$95	157 $\pm$ 2	5242 $\pm$95
SAL2-113	297 $\pm$ 1.0	2362 $\pm$ 48	122 $\pm$ 3	159.2 $\pm$ 2.8	0.0591 $\pm$ 0.0005	5693 $\pm$ 56	5493 $\pm$152	162 $\pm$ 3	5428 $\pm$152
SAL2-113	306 $\pm$ 0.8	2800 $\pm$ 57	110 $\pm$ 2	157.7 $\pm$ 2.1	0.0611 $\pm$ 0.0005	5897 $\pm$ 51	5667 $\pm$170	160 $\pm$ 2	5602 $\pm$170
SAL2-141	409 $\pm$ 1.1	493 $\pm$ 10	882 $\pm$ 18	147.8 $\pm$ 1.8	0.0644 $\pm$ 0.0002	6294 $\pm$ 26	6264 $\pm$34	150 $\pm$ 2	6199 $\pm$34
SAL2-158.5	471 $\pm$ 2.2	3399 $\pm$ 70	166 $\pm$ 3	139.8 $\pm$ 2.7	0.0728 $\pm$ 0.0004	7185 $\pm$ 46	7001 $\pm$138	143 $\pm$ 3	6936 $\pm$138
SAL2-158.5	391 $\pm$ 1.3	2889 $\pm$ 59	165 $\pm$ 3	137.8 $\pm$ 2.0	0.0738 $\pm$ 0.0003	7305 $\pm$ 38	7117 $\pm$138	141 $\pm$ 2	7052 $\pm$138
SAL2-200	543 $\pm$ 6.7	978 $\pm$ 23	677 $\pm$ 17	134.6 $\pm$ 9.8	0.0740 $\pm$ 0.0010	7344 $\pm$ 124	7298 $\pm$128	137 $\pm$ 10	7233 $\pm$128
SAL2-200	571 $\pm$ 1.3	1409 $\pm$ 29	503 $\pm$ 11	139.0 $\pm$ 1.6	0.0754 $\pm$ 0.0004	7455 $\pm$ 47	7392 $\pm$65	142 $\pm$ 2	7327 $\pm$65
SAL2-241	514 $\pm$ 1.6	4755 $\pm$ 97	147 $\pm$ 3	139.9 $\pm$ 1.9	0.0824 $\pm$ 0.0003	8166 $\pm$ 36	7930 $\pm$170	143 $\pm$ 2	7865 $\pm$170
SAL2-241	538 $\pm$ 1.3	5413 $\pm$ 109	136 $\pm$ 3	140.6 $\pm$ 1.7	0.0829 $\pm$ 0.0003	8212 $\pm$ 32	7956 $\pm$184	144 $\pm$ 2	7891 $\pm$184
SAL2-268	594 $\pm$ 64.6	2473 $\pm$ 269	304 $\pm$ 35	164.4 $\pm$ 80.3	0.0769 $\pm$ 0.0088	7434 $\pm$ 1032	7330 $\pm$1021	168 $\pm$ 82	7265 $\pm$1021
SAL2-268	228 $\pm$ 0.4	915 $\pm$ 19	346 $\pm$ 8	146.6 $\pm$ 1.6	0.0844 $\pm$ 0.0005	8320 $\pm$ 56	8218 $\pm$91	150 $\pm$ 2	8153 $\pm$91
SAL2-315	466 $\pm$ 1.4	9318 $\pm$ 189	78 $\pm$ 2	136.6 $\pm$ 2.0	0.0949 $\pm$ 0.0005	9493 $\pm$ 52	8982 $\pm$366	140 $\pm$ 2	8917 $\pm$366
SAL2-328	560 $\pm$ 1.2	50131 $\pm$ 1009	24 $\pm$ 0	154.0 $\pm$ 1.9	0.1324 $\pm$ 0.0006	13259 $\pm$ 65	10989 $\pm$1608	159 $\pm$ 2	10924 $\pm$1608
SAL2-328	586 $\pm$ 0.7	89599 $\pm$ 743	15 $\pm$ 0	153.1 $\pm$ 1.9	0.1403 $\pm$ 0.0031	14131 $\pm$ 338	10585 $\pm$1833	158 $\pm$ 2	10520 $\pm$1833
BD-1									
BD1-(-2.5)	77 $\pm$ 0.2	463 $\pm$ 12	276 $\pm$ 8	19.3 $\pm$ 2.8	0.1007 $\pm$ 0.0009	11356 $\pm$ 109	11200 $\pm$134	20 $\pm$ 3	11135 $\pm$134
BD1-(-1)	59 $\pm$ 0.1	356 $\pm$ 8	281 $\pm$ 7	24.6 $\pm$ 1.8	0.1024 $\pm$ 0.0011	11480 $\pm$ 126	11309 $\pm$174	25 $\pm$ 2	11244 $\pm$174
BD1-5	82 $\pm$ 0.2	248 $\pm$ 7	585 $\pm$ 17	24.8 $\pm$ 1.8	0.1067 $\pm$ 0.0010	11985 $\pm$ 123	11900 $\pm$137	26 $\pm$ 2	11835 $\pm$137
BD1-12	70 $\pm$ 0.1	311 $\pm$ 7	407 $\pm$ 9	38.2 $\pm$ 1.7	0.1094 $\pm$ 0.0008	12136 $\pm$ 95	12012 $\pm$129	40 $\pm$ 2	11947 $\pm$129
BD1-31.5	71 $\pm$ 0.1	265 $\pm$ 6	485 $\pm$ 11	24.7 $\pm$ 1.7	0.1099 $\pm$ 0.0006	12365 $\pm$ 79	12259 $\pm$109	26 $\pm$ 2	12194 $\pm$109
BD1-31.5	89 $\pm$ 0.1	363 $\pm$ 8	448 $\pm$ 10	25.2 $\pm$ 1.6	0.1107 $\pm$ 0.0006	12459 $\pm$ 74	12343 $\pm$111	26 $\pm$ 2	12278 $\pm$111
BD1-59.5	76 $\pm$ 0.1	628 $\pm$ 13	230 $\pm$ 5	22.3 $\pm$ 1.9	0.1145 $\pm$ 0.0011	12943 $\pm$ 137	12709 $\pm$215	23 $\pm$ 2	12644 $\pm$215

Table 2.1 (Continued)

Sample-depth (mm)	^{238}U (ppb)	^{232}Th (ppt)	$^{230}\text{Th} / ^{232}\text{Th}$ (atomic $\times 10^{-6}$)	$d^{234}\text{U}^*$ (measured)	$^{230}\text{Th} / ^{238}\text{U}$ (activity)	^{230}Th Age (yr) (uncorrected)	^{230}Th Age (yr) (corrected)	$d^{234}\text{U}_{\text{Initial}}^{**}$ (corrected)	^{230}Th Age (yr BP) *** (corrected)
BD1-73.5	98 $\pm$ 0.2	526 $\pm$ 11	351 $\pm$ 7	24.2 $\pm$ 2.0	0.1143 $\pm$ 0.0006	12892 $\pm$ 72	12740 $\pm$130	25 $\pm$ 2	12675 $\pm$130
BD1-107	93 $\pm$ 0.1	266 $\pm$ 6	675 $\pm$ 15	23.1 $\pm$ 1.6	0.1172 $\pm$ 0.0007	13263 $\pm$ 83	13181 $\pm$100	24 $\pm$ 2	13116 $\pm$100
BD1-144.5	109 $\pm$ 0.3	191 $\pm$ 15	1106 $\pm$ 89	21.7 $\pm$ 2.0	0.1182 $\pm$ 0.0021	13404 $\pm$ 250	13354 $\pm$252	22 $\pm$ 2	13289 $\pm$252
BD1-150	103 $\pm$ 0.2	872 $\pm$ 18	239 $\pm$ 5	23.4 $\pm$ 1.8	0.1222 $\pm$ 0.0007	13854 $\pm$ 93	13614 $\pm$193	24 $\pm$ 2	13549 $\pm$193
BD1-192	96 $\pm$ 0.2	350 $\pm$ 7	567 $\pm$ 12	19.2 $\pm$ 1.7	0.1254 $\pm$ 0.0005	14307 $\pm$ 72	14203 $\pm$103	20 $\pm$ 2	14138 $\pm$103
BD1-193	120 $\pm$ 0.2	1313 $\pm$ 26	193 $\pm$ 4	23.4 $\pm$ 1.7	0.1280 $\pm$ 0.0005	14562 $\pm$ 66	14250 $\pm$230	24 $\pm$ 2	14185 $\pm$230
BD1-235	124 $\pm$ 0.2	179 $\pm$ 6	1440 $\pm$ 46	24.7 $\pm$ 1.9	0.1263 $\pm$ 0.0009	14334 $\pm$ 108	14293 $\pm$112	26 $\pm$ 2	14228 $\pm$112
BD1-243	116 $\pm$ 0.2	1579 $\pm$ 32	159 $\pm$ 3	23.9 $\pm$ 1.5	0.1319 $\pm$ 0.0006	15036 $\pm$ 74	14648 $\pm$284	25 $\pm$ 2	14583 $\pm$284
BD1-264.5	146 $\pm$ 0.4	2497 $\pm$ 50	129 $\pm$ 3	22.8 $\pm$ 2.0	0.1331 $\pm$ 0.0006	15199 $\pm$ 77	14713 $\pm$352	24 $\pm$ 2	14648 $\pm$352
BD1-264.5	147 $\pm$ 0.3	2186 $\pm$ 44	148 $\pm$ 3	25.1 $\pm$ 1.5	0.1330 $\pm$ 0.0005	15151 $\pm$ 68	14729 $\pm$306	26 $\pm$ 2	14664 $\pm$306
BD1-275	95 $\pm$ 0.3	580 $\pm$ 12	360 $\pm$ 7	26.8 $\pm$ 2.1	0.1337 $\pm$ 0.0006	15200 $\pm$ 83	15026 $\pm$148	28 $\pm$ 2	14961 $\pm$148
BD1-275	78 $\pm$ 0.2	454 $\pm$ 9	379 $\pm$ 8	25.1 $\pm$ 1.9	0.1335 $\pm$ 0.0006	15209 $\pm$ 81	15044 $\pm$142	26 $\pm$ 2	14979 $\pm$142
BD1-278	94 $\pm$ 0.2	2566 $\pm$ 52	84 $\pm$ 2	22.6 $\pm$ 1.7	0.1389 $\pm$ 0.0009	15915 $\pm$ 120	15134 $\pm$565	24 $\pm$ 2	15069 $\pm$565
BD1-278	94 $\pm$ 0.2	2523 $\pm$ 51	86 $\pm$ 2	24.5 $\pm$ 1.8	0.1399 $\pm$ 0.0009	16006 $\pm$ 116	15239 $\pm$555	26 $\pm$ 2	15174 $\pm$555
BD1-347.5	172 $\pm$ 0.4	492 $\pm$ 10	820 $\pm$ 18	25.2 $\pm$ 2.0	0.1422 $\pm$ 0.0010	16274 $\pm$ 129	16193 $\pm$141	26 $\pm$ 2	16128 $\pm$141
BD1-347.5	171 $\pm$ 5.2	373 $\pm$ 15	1046 $\pm$ 44	31.3 $\pm$ 24.1	0.1381 $\pm$ 0.0045	15666 $\pm$ 677	15605 $\pm$676	33 $\pm$ 25	15540 $\pm$676
BD1-395.5	135 $\pm$ 0.3	1387 $\pm$ 28	246 $\pm$ 5	26.7 $\pm$ 1.7	0.1537 $\pm$ 0.0013	17678 $\pm$ 164	17385 $\pm$264	28 $\pm$ 2	17320 $\pm$264
BD1-440.5	213 $\pm$ 0.4	1375 $\pm$ 28	405 $\pm$ 8	27.8 $\pm$ 1.7	0.1589 $\pm$ 0.0006	18300 $\pm$ 78	18117 $\pm$151	29 $\pm$ 2	18052 $\pm$151
BD1-440.5	216 $\pm$ 0.4	1151 $\pm$ 23	495 $\pm$ 10	29.9 $\pm$ 1.5	0.1601 $\pm$ 0.0006	18413 $\pm$ 75	18262 $\pm$130	31 $\pm$ 2	18197 $\pm$130
BD1-484	175 $\pm$ 0.4	2422 $\pm$ 49	209 $\pm$ 4	31.0 $\pm$ 1.6	0.1755 $\pm$ 0.0009	20333 $\pm$ 121	19943 $\pm$301	33 $\pm$ 2	19878 $\pm$301
BD1-553	168 $\pm$ 0.4	4778 $\pm$ 96	112 $\pm$ 2	35.3 $\pm$ 1.8	0.1925 $\pm$ 0.0006	22411 $\pm$ 94	21612 $\pm$573	38 $\pm$ 2	21547 $\pm$573
BD1-596	264 $\pm$ 0.3	1102 $\pm$ 22	751 $\pm$ 15	32.9 $\pm$ 1.3	0.1900 $\pm$ 0.0005	22137 $\pm$ 72	22020 $\pm$110	35 $\pm$ 1	21955 $\pm$110
BD1-596	252 $\pm$ 0.3	1134 $\pm$ 23	696 $\pm$ 14	32.7 $\pm$ 1.2	0.1900 $\pm$ 0.0004	22154 $\pm$ 65	22027 $\pm$110	35 $\pm$ 1	21962 $\pm$110
BD1-637	195 $\pm$ 0.2	399 $\pm$ 9	1597 $\pm$ 34	32.4 $\pm$ 1.3	0.1984 $\pm$ 0.0005	23251 $\pm$ 78	23193 $\pm$88	35 $\pm$ 1	23128 $\pm$88

Table 2.1 (Continued)

Sample-depth (mm)	^{238}U (ppb)	^{232}Th (ppt)	$^{230}\text{Th} / ^{232}\text{Th}$ (atomic $\times 10^{-6}$)	$d^{234}\text{U}^*$ (measured)	$^{230}\text{Th} / ^{238}\text{U}$ (activity)	^{230}Th Age (yr) (uncorrected)	^{230}Th Age (yr) (corrected)	$d^{234}\text{U}_{\text{Initial}}^{**}$ (corrected)	^{230}Th Age (yr BP) *** (corrected)
BD1-637	190 $\pm$ 0.2	543 $\pm$ 11	1148 $\pm$ 24	33.0 $\pm$ 1.3	0.1990 $\pm$ 0.0005	23314 $\pm$ 71	23233 $\pm$91	35 $\pm$ 1	23168 $\pm$91
BD1-741	170 $\pm$ 0.4	3190 $\pm$ 64	195 $\pm$ 4	37.7 $\pm$ 2.0	0.2222 $\pm$ 0.0008	26236 $\pm$ 125	25710 $\pm$392	41 $\pm$ 2	25645 $\pm$392
BD-2									
BD2-3	138 $\pm$ 0.3	1710 $\pm$ 35	179 $\pm$ 4	20.8 $\pm$ 2.0	0.1339 $\pm$ 0.0006	15331 $\pm$ 80	14978 $\pm$262	22 $\pm$ 2	14913 $\pm$262
BD2-3	142 $\pm$ 0.2	2141 $\pm$ 43	146 $\pm$ 3	20.8 $\pm$ 1.7	0.1335 $\pm$ 0.0006	15275 $\pm$ 81	14844 $\pm$315	22 $\pm$ 2	14779 $\pm$315
BD2-54	141 $\pm$ 0.2	2410 $\pm$ 48	139 $\pm$ 3	19.6 $\pm$ 1.7	0.1445 $\pm$ 0.0006	16655 $\pm$ 78	16167 $\pm$354	20 $\pm$ 2	16102 $\pm$354
BD2-54	139 $\pm$ 0.2	2229 $\pm$ 45	148 $\pm$ 3	19.5 $\pm$ 1.7	0.1440 $\pm$ 0.0006	16599 $\pm$ 78	16142 $\pm$333	20 $\pm$ 2	16077 $\pm$333
BD2-105.5	140 $\pm$ 0.3	4106 $\pm$ 83	94 $\pm$ 2	23.1 $\pm$ 1.8	0.1678 $\pm$ 0.0007	19521 $\pm$ 92	18683 $\pm$599	24 $\pm$ 2	18618 $\pm$599
BD2-105.5	170 $\pm$ 0.3	4074 $\pm$ 82	112 $\pm$ 2	21.8 $\pm$ 1.8	0.1632 $\pm$ 0.0006	18965 $\pm$ 86	18279 $\pm$492	23 $\pm$ 2	18214 $\pm$492
BD2-136	207 $\pm$ 0.2	3740 $\pm$ 75	154 $\pm$ 3	25.3 $\pm$ 1.3	0.1693 $\pm$ 0.0005	19661 $\pm$ 67	19148 $\pm$369	27 $\pm$ 1	19083 $\pm$369
BD2-136	202 $\pm$ 0.3	3717 $\pm$ 75	150 $\pm$ 3	26.4 $\pm$ 1.5	0.1671 $\pm$ 0.0005	19358 $\pm$ 72	18836 $\pm$376	28 $\pm$ 2	18771 $\pm$376
BD2-180.5	156 $\pm$ 0.3	1290 $\pm$ 26	361 $\pm$ 7	27.7 $\pm$ 1.6	0.1808 $\pm$ 0.0008	21079 $\pm$ 104	20846 $\pm$195	29 $\pm$ 2	20781 $\pm$195
BD2-180.5	154 $\pm$ 0.3	1681 $\pm$ 34	290 $\pm$ 6	26.6 $\pm$ 1.7	0.1918 $\pm$ 0.0008	22526 $\pm$ 113	22217 $\pm$245	28 $\pm$ 2	22152 $\pm$245
BD2-212	176 $\pm$ 0.3	2453 $\pm$ 49	225 $\pm$ 5	27.0 $\pm$ 2.0	0.1903 $\pm$ 0.0011	22332 $\pm$ 150	21937 $\pm$317	29 $\pm$ 2	21872 $\pm$317
BD2-212	209 $\pm$ 0.2	2997 $\pm$ 60	219 $\pm$ 4	26.5 $\pm$ 1.4	0.1898 $\pm$ 0.0005	22272 $\pm$ 71	21867 $\pm$295	28 $\pm$ 1	21802 $\pm$295
BD2-245.5	172 $\pm$ 0.4	1641 $\pm$ 33	348 $\pm$ 7	25.8 $\pm$ 2.0	0.2015 $\pm$ 0.0008	23832 $\pm$ 115	23560 $\pm$223	28 $\pm$ 2	23495 $\pm$223
BD2-245.5	188 $\pm$ 0.4	1869 $\pm$ 38	331 $\pm$ 7	24.9 $\pm$ 2.1	0.2003 $\pm$ 0.0008	23693 $\pm$ 117	23410 $\pm$231	27 $\pm$ 2	23345 $\pm$231
BD2-259	207 $\pm$ 0.2	1394 $\pm$ 28	511 $\pm$ 10	31.3 $\pm$ 1.2	0.2083 $\pm$ 0.0005	24576 $\pm$ 69	24387 $\pm$150	34 $\pm$ 1	24322 $\pm$150
BD2-259	234 $\pm$ 0.3	2987 $\pm$ 60	265 $\pm$ 5	28.2 $\pm$ 1.5	0.2051 $\pm$ 0.0007	24240 $\pm$ 101	23878 $\pm$275	30 $\pm$ 2	23813 $\pm$275
BD2-307	216 $\pm$ 0.3	3636 $\pm$ 73	212 $\pm$ 4	27.1 $\pm$ 1.6	0.2167 $\pm$ 0.0006	25808 $\pm$ 93	25331 $\pm$350	29 $\pm$ 2	25266 $\pm$350
BD2-307	229 $\pm$ 0.3	4101 $\pm$ 82	200 $\pm$ 4	27.9 $\pm$ 1.4	0.2169 $\pm$ 0.0004	25822 $\pm$ 73	25314 $\pm$366	30 $\pm$ 2	25249 $\pm$366
BD2-350	207 $\pm$ 0.3	1600 $\pm$ 32	477 $\pm$ 10	24.9 $\pm$ 1.5	0.2243 $\pm$ 0.0006	26907 $\pm$ 98	26687 $\pm$184	27 $\pm$ 2	26622 $\pm$184
BD2-350	230 $\pm$ 0.3	1996 $\pm$ 40	427 $\pm$ 9	26.1 $\pm$ 1.4	0.2242 $\pm$ 0.0006	26857 $\pm$ 93	26611 $\pm$197	28 $\pm$ 2	26546 $\pm$197

Table 2.1 (Continued)

Sample-depth (mm)	²³⁸ U (ppb)	²³² Th (ppt)	²³⁰ Th / ²³² Th (atomic x10 ⁻⁶)	d ²³⁴ U* (measured)	²³⁰ Th / ²³⁸ U (activity)	²³⁰ Th Age (yr) (uncorrected)	²³⁰ Th Age (yr) (corrected)	d ²³⁴ U _{initial} ** (corrected)	²³⁰ Th Age (yr BP)*** (corrected)
BD2-398	152 ±0.1	1321 ±27	448 ±9	25.1 ±1.3	0.2358 ±0.0007	28466 ±110	28220 ±206	27 ±1	28155 ±206
BD2-398	1484 ±1.6	17210 ±345	339 ±7	26.2 ±1.4	0.2381 ±0.0007	28760 ±104	28431 ±254	28 ±2	28366 ±254
BD2-450	162 ±0.2	1638 ±33	403 ±8	28.1 ±1.4	0.2472 ±0.0008	29948 ±121	29662 ±235	31 ±2	29597 ±235

U decay constants: $\lambda_{238} = 1.55125 \times 10^{-10}$ (Jaffey et al., 1971) and $\lambda_{234} = 2.82206 \times 10^{-6}$ (Cheng et al., 2013).

Th decay constant: $\lambda_{230} = 9.1705 \times 10^{-6}$ (Cheng et al., 2013).

*d²³⁴U = ($^{234}\text{U}/^{238}\text{U}$)_{activity} - 1)x1000. ** d²³⁴U_{initial} was calculated based on ²³⁰Th age (T), i.e., $d^{234}\text{U}_{\text{initial}} = d^{234}\text{U}_{\text{measured}} \times e^{\lambda_{234}T}$.

Corrected ²³⁰Th ages assume the initial ²³⁰Th/²³²Th atomic ratio of $4.4 \pm 2.2 \times 10^{-6}$. Those are the values for a material at secular equilibrium, with the bulk earth ²³²Th/²³⁸U value of 3.8. The errors are arbitrarily assumed to be 50%.

***B.P. stands for “Before Present” where the “Present” is defined as the year 1950 A.D.

Chapter 3

Tropical vegetation and soil $p\text{CO}_2$ changes in central Indonesia during the past 30,000 years

3.1 Abstract

Present-day vegetation composition in Southeast Asia is largely controlled by rainfall amount and its seasonality. How the relationship maintained in the past remains controversial. Here, I reconstruct tropical vegetation and soil CO_2 production changes with speleothem $\delta^{13}\text{C}$ and trace element ratios from southwestern Sulawesi, Indonesia. Comparing these records with the hydroclimate record based on calcite $\delta^{18}\text{O}$ from the same stalagmites demonstrates that tropical vegetation and soil CO_2 production changes have largely followed hydroclimate changes during the Last Glacial Maximum and Holocene periods, when temperature and atmospheric CO_2 remained relatively stable. During the last deglaciation, tropical vegetation and soil CO_2 production responded to hydroclimate changes only weakly; instead, they largely followed changes in climate boundary conditions such as temperature and atmospheric CO_2 concentration. My results support the use of tropical vegetation records to indicate relative changes in rainfall amount for time periods when temperature and atmospheric CO_2 did not vary substantially, but I urge caution regarding similar interpretations for periods of dramatic climate transition, such as the last deglaciation. The study also highlights the resistance of rainforest refugia to relatively large changes in temperature and atmospheric CO_2 concentration, as well as millennial-scale ENSO-like climate

variability that strongly influenced tropical vegetation and soil CO₂ production during the mid-late Holocene.

3.2 Introduction

Tropical vegetation distribution in the Maritime Continent is largely determined by rainfall amount and rainfall seasonality. A high annual precipitation rate and lack of dry seasons across most of the archipelago have sustained widespread lowland rainforests (with montane forests in higher elevations) in Sumatra, Borneo, Sulawesi, and Papua (Figure 1.1). In comparison, tropical deciduous or monsoonal forests are observed in regions further south such as southwestern Sulawesi, central-eastern Java, and the Lesser Sunda Islands, where seasonal rainfall variability is higher (Göltenboth et al., 2006; Wicaksono et al., 2015). The sustainability of these forests is highly dependent on climatic changes. However, the specific impacts of rising temperatures and atmospheric CO₂ concentration ($p\text{CO}_2$) on these forests remain uncertain. Answering this question requires knowledge from past periods with similarly large-scale changes in temperature and atmospheric $p\text{CO}_2$, for instance, during the transition from the Last Glacial Maximum (LGM) to the Holocene.

The history of climate and ecosystem in the Indo-Australian region has mostly been derived from palynological and plant biomarker records. Pollen records have generally shown that grasslands expanded across the Indonesian terrestrial landscape during the LGM (Dubois et al., 2014; Hope, 2001; Moss and Kershaw, 2000; van der Kaars et al., 2001; van der Kaars et al., 2000; Van der Kaars, 1991), indicating a drier glacial climate. Carbon isotopic compositions ($\delta^{13}\text{C}$) of cave guano samples and terrestrial leaf waxes in coastal and lacustrine sediments also documented higher values during the LGM,

indicating savannah expansion and hence a relatively drier climate in Peninsula Malaysia, Palawan, Sulawesi, and the Lesser Sunda Islands (Dubois et al., 2014; Russell et al., 2014; Wicaksono et al., 2015; Wicaksono et al., 2017; Wurster et al., 2010). In contrast, evidence of persistent rainforests during the LGM from Sumatra (Newsome and Flenley, 1988; Niedermeyer et al., 2014), Borneo (Dubois et al., 2014; Wurster et al., 2010), Java (Stuijts, 1993), and Papua (Hope and Tulip, 1994) suggests little change in the glacial-interglacial hydroclimate. In addition, regional paleo-vegetation records sometimes conflict with non-vegetation hydroclimate records from almost identical locations. For instance, pollen and biomarker records (paleo-vegetation records) from the Lesser Sunda Islands suggest grass expansion during the LGM (Dubois et al., 2014), whereas speleothem $\delta^{18}\text{O}$ records (non-vegetation hydroclimate records) from Flores and northwestern Australia suggest a wetter climate for the same period (Ayliffe et al., 2013; Denniston et al., 2013b). Such discrepancies could be due to that different proxies represent different aspects of hydroclimate changes; on the other hand, the discrepancies also raise questions regarding the relationship between vegetation changes and rainfall, as well as the robustness of using vegetation records to indicate hydroclimate changes over periods such as the glacial-interglacial timescale, when climate boundary conditions changed drastically.

Disentangling the driving mechanisms of tropical vegetation changes over various timescales requires long and continuous records with precise and comparable chronology constraints from both the vegetation and climatic forcing sides. Currently, most vegetation and hydroclimate records are typically obtained from different archives with often different dating techniques. Here, I

reconstruct a multi-proxy record with speleothems from southwestern (SW) Sulawesi, Indonesia. I compare the speleothem stable isotopes and trace element ratios to shed light on tropical vegetation and soil CO₂ production changes as well as the corresponding driving mechanisms on different timescales during the past 30,000 years.

3.3 Study sites

The study site is located in the Maros Regency of SW Sulawesi, Indonesia. The local climate is strongly influenced by the Indo-Australian summer monsoon, which transports moisture to the region from the northeast (Figure 3.1). The annual rainfall amount exceeds 2,500 mm, almost 80% of which occurs during the rainy season, which typically lasts from May to October (Figure 3.2). The vegetation in SW Sulawesi is a mix of closed-canopy forests, tropical deciduous forests, and grasslands, as compared to predominantly ever-wet rainforests associated with abundant year-round rainfall and little seasonality prevailing over the northern Maritime Continent, and grasslands and savannas with lower rainfall amount and strong seasonality in the Lesser Sunda Islands (Goltenboth et al., 2006; Heaney, 1991; Peel et al., 2007).

Figure 3.1 The location of cave sites in southwestern Sulawesi. The background topographic map is based on grids data from Global Multi-Resolution Topography (GMRT) synthesis (Ryan et al., 2009), with the white area showing exposed continental shelf during the Last Glacial Maximum (LGM). Also shown are the wind fields during June to August (JJA, left) and December to February (DJF, right) respectively. The wind data is based on the National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010) (Kalnay et al., 1996).

Figure 3.2 Monthly averaged temperature, precipitation, and rainfall $\delta^{18}\text{O}$ over SW Sulawesi. Temperature is averaged over 1993-2014 from Global Surface Summary of the Day (GSOD), National Oceanic and Atmospheric Administration (NOAA). Precipitation data is based on the Tropical Rainfall Measuring Mission (TRMM) 3B43 dataset (1998-present). Rainfall $\delta^{18}\text{O}$ is based on monitoring from Makassar (14m a.s.l.) between 2002 and 2006 (Kurita et al., 2009).

3.4 Methods

The SW Sulawesi speleothem record is based on five stalagmites (SAL3, SAL2, BD1, BD2, and SLML4) collected from Cave Salleh (S5°3', E119°43'), Cave Batu Depan (S5°1', E119°41'), and Cave Suleman (S5°3', E119°42'), respectively. All the stalagmites show clear calcite laminae, with BD1, BD2, and SLML4 showing relatively cleaner calcite compared with SAL2 and SAL3, characterized by frequent muddy laminae or even continuous muddy deposits associated with growth cession. The record comprises 1,413 stable isotope measurements (these measurements include both $\delta^{18}\text{O}$ discussed in Chapter 2 and $\delta^{13}\text{C}$ discussed in this chapter) constrained by a total 67 U-Th dates. The stalagmites were cut into half and polished before drilling. The U-Th dating was performed on a multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS) (Neptune Plus) in the Isotope Geochemistry Laboratory at the Earth Observatory of Singapore and the Asian School of the Environment, Nanyang Technological University, with the chemistry procedure and measurements largely following methods described elsewhere (Cheng et al., 2013a; Cheng et al., 2016a; Edwards et al., 1987; Shen et al., 2012). Uranium and thorium isotopes, except for ^{238}U and ^{232}Th , were analyzed with a peak-jumping mode on a secondary electron multiplier (SEM) equipped with a retarding potential quadrupole (RPQ) lens to improve abundance sensitivity. The chronology profiles and U-Th dates can be found in Table 2.1 and Figure 2.7. The stable isotope analyses were conducted on a Finnigan MAT 253 mass spectrometer equipped with a Kiel Carbonate Device III at the School of Geography Science, Nanjing Normal University, with a long-term reproducibility of $\sim 0.10\text{‰}$ based on duplicate measurements of standard

NBS19 and arbitrary samples. The analyses were conducted on SAL2, BD1, and BD2 with a 1-mm resolution, and 5-mm resolution analyses were conducted on SAL3 and SLML4. Stable isotope results are reported in per mil (‰) relative to the Vienna Pee Dee Belemnite (VPDB) standard.

Trace elements were measured on a laser ablation inductively coupled plasma mass spectrometer (LA-ICP-MS) with an iCAP-Q Quadrupole ICP-MS and a Resonetics M-50-LR 193 nm excimer laser at the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences, Wuhan, China. Stalagmites SAL2, BD1, and BD2 were analyzed for isotopes of ^{24}Mg , ^{27}Al , ^{28}Si , ^{31}P , ^{42}Ca , ^{86}Sr , ^{138}Ba , ^{232}Th , and ^{238}U . The three stalagmites were first scanned with a round 200 μm -diameter spot to clean their surfaces, and later scanned with a 50 μm -spot for seasonal temporal resolution. Before and after each transect analysis, reference materials and background intensity were measured to check for and correct any machine drift. The operation parameters used in this study followed (Zhu et al., 2016). Off-line data were analyzed with Iolite (Hellstrom et al., 2008; Paton et al., 2011) on Igor Pro. Absolute concentrations of elements were calculated by using ^{42}Ca as an internal standard, and NIST 612 glass and MACS3 carbonate as reference materials.

3.5 Results

The ages of samples BD1 and BD2 from Cave Batu Depan and SAL2 and SAL3 from Cave Salleh cover most of the last 30,000 years (Figure 3.3). Both oxygen and carbon isotope records replicate remarkably in contemporaneous growth periods of stalagmites, for example, between SAL2 and SAL3 as well

as between BD1 and BD2, except for their minor deviations in $\delta^{13}\text{C}$. While the $\delta^{18}\text{O}$ profile of SLML4 from Cave Suleman replicates the other $\delta^{18}\text{O}$ during both deglacial and Holocene periods, its deglacial $\delta^{13}\text{C}$ is offset from the existing BD1 and BD2 profiles. Yet its $\delta^{13}\text{C}$ replicates the Holocene portion of the other datasets and presents a similar decreasing trend over the glacial-interglacial timescale. On the glacial-interglacial timescale, SW Sulawesi speleothem $\delta^{13}\text{C}$ shows a broad trend of ^{13}C depletion from the last glacial period to the Holocene. Specifically, between $\sim 30,000$ years ago and the LGM, the $\delta^{13}\text{C}$ oscillated between -8‰ and -9‰ with no obvious long-term trend. It then gradually decreased to $\sim -12.7\text{‰}$ until 5 thousand years before the present (kyr BP, where the present = 1950 A.D.). Superimposed on the broad glacial-interglacial $\delta^{13}\text{C}$ decrease are several positive millennial-scale excursions centered around 26, 20.6, 17, 15, 6.5, and 3 kyr BP, with the $\sim 3\text{‰}$ shift at ~ 3 kyr BP as the most prominent $\delta^{13}\text{C}$ excursion in the record.

The SW Sulawesi speleothem $\delta^{18}\text{O}$ shows an overall similar decreasing glacial-interglacial trend and millennial-scale excursions synchronous with those of $\delta^{13}\text{C}$. The two stable isotopic records are, however, strikingly different in their structures during the last deglaciation. While the $\delta^{18}\text{O}$ did not decrease until ~ 15 kyr BP, the decrease of the deglacial $\delta^{13}\text{C}$ values initiated as early as 20.6 kyr BP. Also noticeable are several prominent millennial-scale $\delta^{18}\text{O}$ fluctuations during the last deglaciation, whereas $\delta^{13}\text{C}$ decreased almost monotonically through the same period.

Figure 3.3 Speleothem $\delta^{13}\text{C}$ (top) and $\delta^{18}\text{O}$ (bottom) from southwestern Sulawesi. Five stalagmites are SAL03 (light blue), SAL02 (orange), SLML04 (pink for $\delta^{13}\text{C}$ and magenta for $\delta^{18}\text{O}$), BD01 (green) and BD02 (dark blue) in sequence. U-Th dates for each stalagmite are plotted in corresponding colors along the top.

The trace element ratios (M/Ca, or simply M) in SW Sulawesi speleothems show relatively stable values for the past 30,000 years, except for several dramatic excursions in SAL2 (Figure 3.4). Indeed, the muddy Holocene stalagmite exhibits a much higher variability than the glacial stalagmites for all the trace elements. Both the glacial and Holocene stalagmites, however, shared a roughly consistent baseline of ~0.4 ppm for Ba/Ca, ~10 ppm for Sr/Ca, ~40 ppm for P/Ca, and ~0.08 ppm for U/Ca. The baseline for Mg/Ca is much higher than for the other ratios and differs between stalagmites from different caves, with ~750 ppm for the Holocene sample and ~100 ppm for the glacial samples.

Among all the trace element ratios, Sr/Ca and Ba/Ca, and P/Ca and U/Ca exhibit a rather strong co-variation for all the stalagmites (Table 3.1). A less strong co-variation is found between $\delta^{13}\text{C}$, Mg/Ca, and U/Ca for the glacial stalagmites, with all proxies sharing a relatively small variation between 30 kyr BP and 20 kyr BP, followed by a continuous decrease until 11.4 kyr BP. For the Holocene sample, $\delta^{13}\text{C}$ and Mg/Ca also share a similar variation, with a minor long-term trend reaching a minimum at ~5 kyr BP and a prominent millennial-scale excursion at ~3 kyr BP. Sr/Ca and Ba/Ca, and P/Ca and U/Ca resemble $\delta^{13}\text{C}$ and Mg/Ca very weakly, with also little resemblance between the two pairs of elements themselves. Some resemblance, however, was found between Sr/Ca and $\delta^{18}\text{O}$.

Figure 3.4 Trace element ratios and stable isotopes of SW Sulawesi speleothem for the past 30,000 years (left) and for the deglacial and glacial periods only (right). From top to bottom are Ba/Ca, Sr/Ca, $\delta^{18}\text{O}$, P/Ca, U/Ca, Mg/Ca, and $\delta^{13}\text{C}$, with orange, purple, and green colors marking stalagmites SAL2, BD1, and BD2 respectively. The P data for stalagmite BD1 is spoiled, likely due to systematic problem during the analyses. All the trace element ratios are composed of original data in lighter colors and its 1000-point moving means in darker colors.

Table 3.1 Correlation coefficients for stable isotopes and trace elements. All the correlation with p-value <0.001 is shown. Both trace element and stable isotope data are interpolated to 10-year resolution for correlation.

SAL2	Mg/Ca	P/Ca	Sr/Ca	Ba/Ca	U/Ca	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
Mg/Ca	1.00						
P/Ca	-0.38	1.00					
Sr/Ca	0.21	-	1.00				
Ba/Ca	-	0.28	0.79	1.00			
U/Ca	-0.31	0.85	-	-	1.00		
$\delta^{18}\text{O}$	0.13	0.12	0.22	0.20	-	1.00	
$\delta^{13}\text{C}$	0.46	-0.13	0.18	-	-0.13	0.75	1.00
BD1	Mg/Ca	P/Ca	Sr/Ca	Ba/Ca	U/Ca	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
Mg/Ca	1.00						
P/Ca	NAN	NAN					
Sr/Ca	-	NAN	1.00				
Ba/Ca	-	NAN	0.53	1.00			
U/Ca	0.23	NAN	-0.24	-0.13	1.00		
$\delta^{18}\text{O}$	-0.17	NAN	-	-0.15	-	1.00	
$\delta^{13}\text{C}$	0.50	NAN	-	-	0.30	0.21	1.00
BD2	Mg/Ca	P/Ca	Sr/Ca	Ba/Ca	U/Ca	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
Mg/Ca	1.00						
P/Ca	-0.12	1.00					
Sr/Ca	0.11	-0.22	1.00				
Ba/Ca	0.11	-	0.52	1.00			
U/Ca	-	0.61	-0.31	-	1.00		
$\delta^{18}\text{O}$	-0.38	-	-	0.10	-	1.00	
$\delta^{13}\text{C}$	0.51	-0.11	-	-	0.17	-0.21	1.00

3.6 Discussion

3.6.1 Interpretation of speleothem stable isotopes and trace elements

In Chapter 2, speleothem $\delta^{18}\text{O}$ from SW Sulawesi has been interpreted as reflecting regional hydroclimate change, or more specifically, the strength of deep convection over the Maritime Continent, due to the remarkable agreements between cave $\delta^{18}\text{O}$ records across the region. Here, therefore, I use the Sulawesi speleothem $\delta^{18}\text{O}$ as a proxy of regional rainfall amount.

The generally well-replicated $\delta^{13}\text{C}$ of the SW Sulawesi stalagmites, except for the minor deviations of $\delta^{13}\text{C}$ in SAL3 and the bottom portion of BD1 (Figure 3.3), strongly suggests that the cave calcite precipitation occurred under equilibrium conditions with negligible kinetic isotopic fractionation (Hendy, 1971; Wang et al., 2001). The $\delta^{13}\text{C}$ of SAL3, although generally replicating the $\delta^{13}\text{C}$ variation of SAL2, is much lower than in SAL2, which is likely due to the kinetic isotopic fractionation caused by 2–3 times higher growth rate of SAL3. Deviated $\delta^{13}\text{C}$ for the bottom portion of BD1 is probably associated with obvious crystal dissolution that altered the original calcite $\delta^{13}\text{C}$, although corresponding $\delta^{18}\text{O}$ still replicates well with BD2's $\delta^{18}\text{O}$ as it is less susceptible to post-depositional alteration. In the following discussion, I therefore skip these $\delta^{13}\text{C}$ deviations and only consider the $\delta^{13}\text{C}$ records that are well replicated between samples.

The overall replication of the SW Sulawesi $\delta^{13}\text{C}$ between different stalagmites from the same or different caves suggests a common environmental control rather than cave- or sample-specific effects on speleothem $\delta^{13}\text{C}$. Indeed,

speleothem $\delta^{13}\text{C}$ is a sensitive proxy for climate-driven changes in ecosystems (Fleitmann et al., 2009). Its variation is commonly related to changes in vegetation types and density as well as soil microbial activity above the cave (Baker et al., 1997; Breecker et al., 2012; Genty et al., 2003b; Meyer et al., 2014), as the $\delta^{13}\text{C}$ of seepage water is strongly influenced by the uptake of soil biogenic CO_2 and its carbon isotopic composition (Bar-Matthews et al., 1998), which are further controlled by changes in vegetation and climatic conditions such as temperature and precipitation (Fleitmann et al., 2009; Genty et al., 2003b). Speleothem $\delta^{13}\text{C}$ values of -14‰ - -6‰ and -6‰ - 2‰ have been suggested to reflect the dominance of C3 (e.g., forest plants) and C4 (e.g., tropical grasses) vegetation above a cave system (Baker et al., 1997; Dreybrodt, 1980; McDermott, 2004). The $\delta^{13}\text{C}$ between -14‰ and -7‰ in SW Sulawesi stalagmites therefore indicates a continuous dominance of C3 plants for the landscape above our cave sites during the past 30,000 years, consistent with the forested woodland thriving in the study area during the present day. C4 plants probably existed as a secondary vegetation type in the past, with higher $\delta^{13}\text{C}$ values indicating expanded C4 vegetation.

The covariance between $\delta^{13}\text{C}$ and Mg/Ca across different timescales suggests a common process controlling their variations. Prior calcite precipitation (PCP) has been frequently invoked to interpret the covariance of $\delta^{13}\text{C}$ and Mg/Ca in speleothems (Fairchild and Treble, 2009; Hellstrom and McCulloch, 2000). PCP typically occurs due to degassing in a vadose zone, when relatively low $p\text{CO}_2$ leads to super-saturation of CaCO_3 in water solutions and hence to precipitation. The process removes Ca more readily than Mg, thus Mg is enriched in solutions and subsequently in speleothems. The CO_2

degassing process in the vadose zone also enriches ^{13}C in speleothems, as ^{12}C is preferentially released during carbon isotope fractionation. The occurrence of PCP, however, requires relatively dry conditions that enhance ventilation and evaporation in the vadose zone. The process may dominantly control the variations of Mg/Ca as well as of other trace element ratios in speleothems from relatively dry regions. SW Sulawesi, however, is located in the deep tropics with abundant annual rainfall. The continuously low $\delta^{13}\text{C}$ values in speleothems also suggest that no significant CO_2 degassing occurred in the vadose zone. I, therefore, consider PCP as less likely the main process that controlled the trace element ratios in the SW Sulawesi speleothem. However, PCP might still occur in the vadose zone as a secondary process contributing to the variations of the trace element ratios, especially during dry seasons or periods with less abundant rainfall and stronger evaporation due to lower humidity.

The SW Sulawesi speleothem trace element ratios show slight or almost no long-term trends during the past 30,000 years, except for the relatively large excursions in the Holocene stalagmite, which are likely associated with the existing muddy laminae. As a matter of fact, during the past 30,000 years, Sr/Ca, Ba/Ca, P/Ca, and U/Ca, and to a lesser degree, Mg/Ca, have mostly remained close to, but never exceeding, the baselines, which were likely constrained by the bedrock compositions. Such a pattern suggests that the cave system has maintained a relatively high recharge, and that overall strong dissolution in the bedrock dominates the variations of the speleothem trace element ratios. Once the trace element ratios achieved these baselines due to a sufficiently high rainfall recharge, the ratios were no longer sensitive to any

further increase in rainfall recharge, but would instead maintain the minimized levels until the occurrence of dry periods with lower recharge, which would then deviate the ratios from the baselines as a result of weaker bedrock dissolution. Similar patterns of “trimmed” speleothem trace element ratios have been observed in northern Borneo and Flores (Griffiths et al., 2010a; Partin et al., 2013), supporting the existence of relatively strong bedrock dissolution due to abundant rainfall, the degree of which was largely determined by water acidity and hence the partial pressure of CO₂ in the soil (Fairchild and Baker, 2012).

While Sr/Ca, Ba/Ca, P/Ca, and U/Ca are much lower in SW Sulawesi speleothems and hence in the bedrock, their variations associated with bedrock dissolution might be originally less noticeable, or can easily be obscured by other element contributions such as detritus flushing during the formation of muddy speleothem laminae (Fairchild and Treble, 2009; Wong et al., 2011) or even simply the instrumental memory effect during analysis. Other processes such as changes in speleothem growth rate might contribute additional variations in Sr and Ba (Treble et al., 2003). These influences help explain the lack of coherence between Sr/Ca–Ba/Ca and P/Ca–U/Ca, as well as between these element ratios and $\delta^{13}\text{C}$. I hence leave aside these elements, and focus only on Mg/Ca, which covaries with contemporary $\delta^{13}\text{C}$. The SW Sulawesi speleothem Mg/Ca has been much higher than other trace element ratios, most likely due to the higher content of Mg in the bedrock, which allowed the ratio to be more sensitive to dissolution changes as long as the ratio did not decrease to its baseline of bedrock composition. Similar as speleothem $\delta^{13}\text{C}$, water acidity, which controls limestone dissolution and consequently Mg/Ca in

speleothems, is essentially tied with soil $p\text{CO}_2$. We therefore suggest that the covariation of speleothem $\delta^{13}\text{C}$ and Mg/Ca can be linked to the production and uptake of soil CO_2 as a function of temperature and rainfall amount.

3.6.2 Glacial-interglacial changes in vegetation and soil CO_2 production

Speleothem $\delta^{13}\text{C}$ of $\sim 8\text{‰}$ during the glacial period in comparison with $\sim 13\text{‰}$ during the Holocene periods suggests expansion of C4 plants during the LGM. Such a pattern agrees with regional pollen and biomarker records from not only Sulawesi (Dam et al., 2001; Hope, 2001; Russell et al., 2014; Wicaksono et al., 2017), but also the Savu Sea and Banda Sea basins (Dubois et al., 2014; van der Kaars et al., 2000), supporting expanded savannah and grassland during the last glacial period. Similar with the $\delta^{13}\text{C}$, the SW Sulawesi speleothem Mg/Ca also shift from higher numbers during the glacial period towards lower values during the Holocene. Such covariation indicates an increased contribution of ^{13}C -depleted biogenic soil CO_2 and stronger bedrock dissolution, likely caused by enhanced soil CO_2 production and then more biogenic CO_2 dissolved in the soil water, which can be associated with increasing temperature and rainfall amount over the glacial-interglacial timescale (Genty et al., 2003b; McDermott, 2004) (Figure 3.5). Indeed, an increase of 3–4 °C in regional sea surface temperature since the LGM has been suggested by marine sediment records (Gibbons et al., 2014; Schröder et al., 2016b; Stott et al., 2007; Visser et al., 2003). The SW Sulawesi speleothem $\delta^{18}\text{O}$ record also suggests strengthening moisture convection and hence higher regional rainfall since the last glacial periods (discussed in detail in Chapter 2), supporting both forest expansion and increase of soil CO_2 production. In addition, the atmospheric $p\text{CO}_2$ increase from much lower glacial levels could

also stimulate the growth of C4 plants over C3 plants (Ehleringer, 2005; Ehleringer et al., 1997; Polley et al., 1993; Street-Perrott et al., 1997), which might also contribute to raise the soil CO₂ production. Overall, the multi-proxy records from the SW Sulawesi speleothems show a picture of climate amelioration accompanied by forest expansion, enhanced soil CO₂ production, as well as stronger cave bedrock dissolution from the last glacial period to the Holocene.

Figure 3.5 SW Sulawesi speleothem $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$, Mg/Ca, and their comparisons with other vegetation records from the Maritime Continent and records of climate boundary conditions. a. SW Sulawesi speleothem $\delta^{13}\text{C}$ (red) and $\delta^{18}\text{O}$ (blue), b. SW Sulawesi speleothem $\delta^{13}\text{C}$ and 1000-point moving-mean of Mg/Ca (cyan). c. $\delta^{13}\text{C}$ of C30 n-alkanoic fatty acid ($\delta^{13}\text{C}_{\text{wax}}$) in marine core GeoB10069-3 from offshore southern Indonesia (Dubois et al., 2014). d. $\delta^{13}\text{C}_{\text{wax}}$ from Lake Towuti in central Sulawesi (Russell et al., 2014). e. Mg/Ca-based sea surface temperature reconstructions from MD98-2181 (Stott et al., 2007). f. Antarctica ice core based atmosphere CO_2 concentration (Monnin et al., 2001). p.p.m.v, parts per million by volume. The yellow bar marks the deglacial shift of regional vegetation and boundary conditions.

During the last deglaciation, the SW Sulawesi speleothem $\delta^{13}\text{C}$ and Mg/Ca show almost continuous shifts in vegetation type and soil CO_2 production from the LGM to the early Holocene. Contemporary speleothem $\delta^{18}\text{O}$, however, demonstrates two-step changes in the regional rainfall amount: a substantial decrease during H1 before an increase towards the B/A, followed by another decrease during the YD and an increase towards the early Holocene. The deglacial changes of the regional rainfall amount have been attributed to latitudinal shifts of the ITCZ forced by millennial-scale variabilities of the Atlantic Meridional Ocean Circulation (AMOC) (discussed in detail in Chapter 2). Such dramatic hydroclimate changes, however, have influenced the shifts in vegetation and biological soil activity only mildly on millennial timescales, as shown by the evident but small excursions of speleothem $\delta^{13}\text{C}$ and Mg/Ca. Both proxies were instead dominated by the continuous trend of boundary condition changes (namely temperature and atmospheric $p\text{CO}_2$) during glacial-interglacial transition periods. Within the glacial or Holocene periods, when temperature and atmospheric $p\text{CO}_2$ remained relatively stable, a high degree of coincidence among speleothem $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and Mg/Ca suggests that the rainfall amount indeed dominated changes of vegetation and soil CO_2 production. Similarly, speleothem $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records from eastern Java show tight coupling during both the Holocene and glacial periods but decoupling during the deglaciation, with a continuous decrease of the deglacial $\delta^{13}\text{C}$ also coinciding with boundary condition changes (Westaway et al., 2007), corroborating the various climate forcings of tropical vegetation during the last glacial-interglacial period.

Such decoupling between tropical vegetation and rainfall amount during the deglaciation, or boundary condition control on tropical vegetation, however, has exceptions in the region. For instance, several pollen and carbon isotope composition records from Sumatra, Borneo, and western Java have suggested persistent dominance of rainforests from the LGM to the Holocene (Dubois et al., 2014; Newsome and Flenley, 1988; Niedermeyer et al., 2014; Stuijts, 1993; Wurster et al., 2010), which manifests little sensitivity of these rainforests to glacial-interglacial boundary condition changes. These records, instead, appear to mainly follow rainfall changes during the deglaciation, with drastic changes during H1, the B/A, and the YD (Dubois et al., 2014; Niedermeyer et al., 2014; Wurster et al., 2010). Very small glacial-interglacial variation in vegetation, and likely of soil CO₂ production, was also documented in speleothem $\delta^{13}\text{C}$ from northern Borneo (Partin et al., 2013), with prominent millennial-scale excursions coinciding with those in rainfall variability (see details in Chapter 2). Low sensitivities of the rainforests from these locations (also known as rainforest ‘refugia’) to glacial-interglacial boundary condition changes (in temperature and atmospheric $p\text{CO}_2$) were likely due to lack of dry seasons in mountainous locations facing the rain-bearing winds from nearby seas. Such low sensitivities also probably imply some resistance of these rainforests to global warming.

During the Holocene, prominent millennial-scale variability in both the SW Sulawesi speleothem $\delta^{13}\text{C}$ and Mg/Ca, indicates episodes of drastic changes of vegetation and soil CO₂ production, probably associated with hydroclimate changes. Indeed, the most prominent positive anomalies of $\delta^{13}\text{C}$ and Mg/Ca ~ 3 kyr BP are characterized by a significant slow-down or even complete cessation

of stalagmite growth, both of which coincide with a prolonged positive $\delta^{18}\text{O}$ excursion centered ~ 3 kyr in the speleothem $\delta^{18}\text{O}$ records from northern Borneo (Partin et al., 2007), SW Sulawesi (Chapter 2), and Flores (Griffiths et al., 2009). The slow-down in stalagmite growth and the positive $\delta^{18}\text{O}$ excursions indicate extensive dry conditions across the Maritime Continent. A severe dry condition ~ 3 kyr is also suggested by coral-based geochemical studies from the tropical western Pacific (Gagan et al., 2004; McGregor and Gagan, 2004; Tudhope et al., 2001; Woodroffe et al., 2003; Woodroffe and Gagan, 2000). Such widespread dry conditions across the lands and seas in the Maritime Continent resembled a regional dry phase in response to the present-day active El Niño. These El Niño-like millennial events probably caused forest collapse, weakening of soil CO_2 production and bedrock dissolution, enhanced PCP, and thus increases in speleothem $\delta^{13}\text{C}$ and Mg/Ca.

3.7 Conclusion

The SW Sulawesi speleothem stable isotopes and trace element ratios have demonstrated a tight coupling between tropical vegetation and soil CO_2 production on the one hand and regional hydroclimate changes on the other hand on different timescales during the past 30,000 years. Such coupling, however, temporarily broke down during the last deglaciation, with changes in vegetation and soil CO_2 production largely following changes in climate boundary conditions such as temperature and atmospheric pCO_2 . During the late Holocene, strong millennial-scale rainfall variability likely associated with the ENSO conditions drastically changed forest biomass and soil CO_2 production in the Maritime Continent. My study highlights the application of speleothem $\delta^{13}\text{C}$ and Mg/Ca as proxies of tropical vegetation and soil CO_2

production changes. I also caution against simplistic climatic interpretation of tropical vegetation records across different periods and timescales.

3.8 References

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Chapter 4

The spatial and temporal variations of tropical and subtropical speleothem $\delta^{18}\text{O}$ records since the Last Glacial Maximum

4.1 Abstract

Tropical and subtropical speleothem $\delta^{18}\text{O}$ records spanning the past 30,000 years are synthesized to explore the control factors of speleothem $\delta^{18}\text{O}$ over orbital and millennial timescales. By comparing speleothem $\delta^{18}\text{O}$ records within a regional context and utilizing a new ITCZ migration index (Chapter 2), I have highlighted here three aspects of speleothem $\delta^{18}\text{O}$ from low latitudes regarding its driving mechanisms: 1) the continental rainout effect and its variation on the glacial-interglacial timescale; 2) the differential sensitivity to SST and land heating between coastal sites and continental interior sites; 3) the dominant influences of ITCZ migration since the LGM, and the secondary influence of deep tropical convection in the Holocene. Finally, I highlight preliminary isoscapes of speleothem $\delta^{18}\text{O}$ and their temporal variation to further hint on tropical-subtropical hydroclimate changes during various periods and timescales.

4.2 Introduction

Tropical hydroclimate bears a significant role on the prosperity of ~40% of global population. Understanding its variability and driving mechanisms on various timescales is thus necessary and important. Instrumental records on precipitation provide insights on hydroclimate dynamics and rainfall change over inter-annual and decadal time scales. Climate archives, on the other hand, record hydroclimate changes over decadal to orbital timescales, and hence can provide invaluable information on long-term climate changes. Among existing terrestrial climate archives, speleothems stand out due to robust replication of their oxygen isotope ratio ($\delta^{18}\text{O}$) records from the same or different caves across a region despite numerous complex factors that affect carbonate deposition. When speleothems are deposited under equilibrium conditions in low-latitude caves, speleothem $\delta^{18}\text{O}$ primarily reflects amount-weighted average of meteoric rainfall $\delta^{18}\text{O}$, which is controlled by large-scale changes in atmospheric circulation across a region together with varying rainfall amount, temperature, seasonality, as well as moisture source and trajectories (Cheng et al., 2016a; Cheng et al., 2012a; Cheng et al., 2016b; McDermott, 2004). As a result, speleothem $\delta^{18}\text{O}$ has been widely used as a proxy for tropical and subtropical hydroclimate changes. More importantly, speleothem carbonates can be precisely dated by uranium-thorium dating methods, which can effectively circumvent some chronology problems of radiocarbon dating commonly seen in marine and lacustrine records. The absolute chronology of speleothem $\delta^{18}\text{O}$ record and its representability for changes in regional climate can provide comparability between different sites of the globe. With the emergence of more and more high-quality speleothem records from a dense

network across the tropics and subtropics (Figure 4.1), it is now possible to evaluate the spatial and temporal distribution of tropical and subtropical speleothem $\delta^{18}\text{O}$, and further hint at the mechanical processes that drive changes in regional hydroclimate and atmospheric circulation.

One of the important control factors for tropical and extra-tropical climate over a variety of timescales is the variation of the mean latitudinal position and structure of the Intertropical Convergence Zone (ITCZ) (Fleitmann et al., 2007) (Figure 4.2). Yet considerable discrepancies exist regarding ITCZ dynamics, its mean positions, and convection strength in the past. This greatly limits our knowledge on the extents to which the ITCZ controls tropical and subtropical hydroclimate changes. Existing reconstruction on ITCZ variation is mostly based on a single location, therefore inevitably containing additional local environmental and climatic influences, which are difficult to be constrained. A newly derived ITCZ index based on two nearby speleothem $\delta^{18}\text{O}$ records from the maritime continent ($\Delta\delta^{18}\text{O}_{\text{SW-FL}}$) allows detailed characterization of ITCZ variability for the past 30,000 years (see details in Chapter 2). The independent chronology of this ITCZ index allows careful comparisons between itself and the existing speleothem $\delta^{18}\text{O}$ -based hydroclimate records, which can shed light on the influences of the ITCZ on tropical and subtropical hydroclimate over varied timescales.

Previously a synthesis study on tropical and subtropical speleothem $\delta^{18}\text{O}$ records enables a good knowledge on the global paleomonsoon as well as its driving mechanisms on a range of timescales (Cheng et al., 2012). But a detailed analysis on the spatial gradient and sensitivity of speleothem $\delta^{18}\text{O}$ from

different hydroclimate domain has not been done. Also absent is a thorough test on the relationship between tropical/subtropical hydroclimate changes and ITCZ dynamics. Here I synthesize tropical and subtropical speleothem $\delta^{18}\text{O}$ records covering the past 30,000 years. I compare speleothem $\delta^{18}\text{O}$ from the same region to detect a spatial gradient of speleothem $\delta^{18}\text{O}$ and the difference of $\delta^{18}\text{O}$ sensitivity between sites. I further compare these records with the ITCZ migration index from Chapter 2 and sea surface temperature (SST) records from nearby oceans to test to what degrees ITCZ migration and SST influence regional hydroclimate. I also perform Empirical Orthogonal Function (EOF) analysis on the speleothem $\delta^{18}\text{O}$ records for the last deglacial and the Holocene periods to understand the dominant forcing mechanism of tropical and subtropical hydroclimate changes since the LGM. Finally, I make preliminary isoscapes of speleothem $\delta^{18}\text{O}$ and its variation during different periods.

4.3 Methods

I selected existing speleothem $\delta^{18}\text{O}$ records from both the tropics and subtropics to study the hydroclimate changes over the low latitudes. I included speleothem $\delta^{18}\text{O}$ records growing between the Last Glacial Maximum (LGM) and the late Holocene from 30°S to 30°N, as well as other ones from higher latitudes whose interpretations are dominated by changes in hydroclimate and atmospheric circulation (Table 4.1). I avoided speleothem records spanning only recent one or two millennia as these records are mostly based on aragonite carbonates, with $\delta^{18}\text{O}$ not necessarily comparable with older speleothem records. For selected records that also cover periods older than the LGM, I included part of the glacial portion till as old as 30,000 years for better comparison with climatic forcings. All the records used are reported with

original calcite $\delta^{18}\text{O}$ and in calendar years before present of 1950 AD (BP). The majority of speleothem $\delta^{18}\text{O}$ records used in this study are available from <https://www.ncdc.noaa.gov/paleo-search> by National Center for Environmental Information (NCEI) of National Oceanic and Atmospheric Administration (NOAA). The rest of used records are from the supplemental materials attached to original publications.

I divided the selected records into four domains based primarily on site distributions across continents or ocean basins: the Indian Ocean Domain, the Mediterranean Westerly Domain, the Western Pacific Domain, and the American Domain. I selected relatively long and continuous speleothem $\delta^{18}\text{O}$ records from the same domain on comparable scales, and aligned them primarily by their latitude and secondarily by their moisture pathways.

I applied EOF analysis to compute dominant modes of speleothem $\delta^{18}\text{O}$ variability. As the dominant climatic forcing for the Holocene and last deglaciation could be quite different, I did this analysis separately for the two periods, with 24 records included for each period. To maximize the numbers of records included and also ensure the length of the interested periods, I selected relatively continuous speleothem $\delta^{18}\text{O}$ records covering 0.55-11.5 thousand years (ka) or 11.5-21 ka with no single hiatus lasting longer than 2.5 ka. These records were later interpolated to 50-years resolution and standardized to avoid biases towards any records with larger-than-average variability. The principal components (PCs) generated are not sensitive to factor-of-ten changes in sampling resolution and also the numbers of input records.

Figure 4.1 Tropical and subtropical speleothem $\delta^{18}\text{O}$ records under seasonal climatology between June and August (JJA, up), and between December and February (DJF, bottom). The precipitation data (shading, in millimeters per hour) is based on the Tropical Rainfall Measuring Mission (TRMM) 3B43 dataset (1998-present, http://disc2.gesdisc.eosdis.nasa.gov/data/TRMM_L3/TRMM_3B43.7). The surface winds data (vectors, in meters per second) is based on the National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010, <https://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis.derived.surface.html>) (Kalnay et al., 1996).

Figure 4.2 The annual mean ITCZ as shown by the Highly Reflective Clouds (HRC) based on long-term monthly mean between 1971-1985. The HRC data is provided by the National Oceanic & Atmospheric Administration (NOAA)/Earth System Research Laboratory (ESRL)/Physical Sciences Division (PSD), Boulder, Colorado, USA, from <https://www.esrl.noaa.gov/psd/data/gridded/data.noaa.hrc.html>. Also shown are the speleothem $\delta^{18}\text{O}$ records in green circles.

Table 4.1 Speleothem $\delta^{18}\text{O}$ records included in this study.

Location	Longitude	Latitude	Elevation (m)	References
Bells	-110.8	31.8	1600	Wagner et al., 2010
Fort Stanton	-105.3	33.3	1800	Asmerom et al., 2010
Pink Panther	-105.2	32.1	1240	Asmerom et al., 2007
Diablo	-99.9	18.2	1030	Bernal et al., 2011
Juxtlahuaca	-99.2	17.4	927	Lachniet et al., 2013
Natural Bridge	-98.3	29.7	300	Wong et al., 2015
Macal	-89.1	16.9	530	Akers et al., 2016
Chen Ha	-89.1	16.7	550	Pollock et al., 2016
Terciopelo	-85.3	10.2	370	Lachniet et al., 2009
Venado	-84.8	10.6	380	Lachniet et al., 2004
Buckeye Creek	-80.4	38	600	Hardt et al., 2010
Chilibrillo	-79.7	9.2	60	Lachniet et al., 2004
Santiago	-78.1	-3	980	Mosblech et al., 2012
Diamante	-77.5	-5.7	960	Cheng et al., 2013
Tigre Perdido	-77.3	-5.9	1000	van Breukelen et al., 2008
Condor	-77.3	-5.9	860	Cheng et al., 2013
Abaco	-77.2	26.2	-17	Arienzo et al., 2015; Arienzo et al., 2017
Pacupahuain	-75.8	-11.2	3800	Kanner et al., 2012
Huagapo	-75.8	-11.3	3850	Kanner et al., 2013
Paraiso	-55.5	-4.1	60	Wang et al., 2017
Botuverá	-49.2	-27.2	230	Cruz et al., 2005; Wang et al., 2007
Rio Grande do Norte	-37.7	-5.6	100	Cruz et al., 2009
Cold Air	29.1	-24	1340	Holmgren et al., 2003
Soreq	35	31.5	400	Bar-Matthews et al., 2003
Jerusalem	35.2	31.8	700	Frumkin et al., 1999
Jeita	35.7	34	100	Cheng et al., 2015
Moomi	54	12.5	700	Shakun et al., 2007
Qunf	54.3	17.2	650	Fleitmann et al., 2007
Hoti	57.4	23.1	800	Neff, et al., 2001
Tonnelnaya	67.2	38.4	2450	Cheng et al., 2016
Bittoo	77.8	30.8	3000	Kathayat et al., 2016
Kesang	81.8	42.9	2000	Cheng et al., 2012; Cheng et al., 2016
Tianmen	90.1	30.9	4800	Cai et al., 2012
Mawmluh	91.9	25.3	1000	Berkelhammer et al., 2012; Dutt et al., 2015
Tangga	100.8	-0.4	600	Wurtzel et al., 2018
Xiaobailong	103.4	24.2	1500	Cai et al., 2015 ; Tan et al., 2017
S.Yunnan	104.2	23.3	1500	Liu et al., submitted
Dongge	108.1	25.3	680	Dykoski et al., 2005; Wang et al., 2005
Jiuxian	109.1	33.6	1495	Cai et al., 2010
Heshang	110.3	30.3	294	Hu et al., 2008
Sanbao	110.4	31.7	1900	Wang et al., 2008
Cape Range	113.9	-22.1	100	Denniston et al., 2013
SnailShell	114.8	4	150	Partin et al., 2007
BukitAssam	114.8	4	150	Partin et al., 2007; Carolin et al., 2013
Hulu	119.2	32.5	1500	Wang et al., 2001
S.W.Sulawei	119.5	-5.1	150	Chapter 2
Liang Lua	120.4	-8.5	550	Griffiths et al., 2009; Ayliffe et al., 2013
Ball Gown	124.1	-17.3	100	Denniston et al., 2013; Denniston al., 2017
KNI-51	128.4	-15.2	100	Denniston et al., 2013; Denniston al., 2017
Maboroshi	133.2	34.8	450	Shen at al., 2010
Kiriana	136.6	34.4	620	Mori et al., 2018
Ohtaki	137	35.7	400	Mori et al., 2018
Fukugaguchi	137.8	37	170	Sone et al., 2013
Hollywood	171.5	-42	130	Whittaker et al., 2011
NW SouthIsland	172	-42	170	Williams et al., 2010
Mt.Arthur	172.7	-41.3	390	Hellstrom et al., 1998

4.4 Results

4.4.1 Indian Ocean Domain

Most areas within the Indian Ocean Domain are influenced by the Indian Summer Monsoon, with only the very southern portion characterized by mixed influences of boreal winter precipitation brought by wind reversal and the westerly wind (Figure 4.1). Records in the Indian Ocean Domain include those from Cold Air Cave in South Africa (Holmgren et al., 2003), Qunf and Hoti Caves from Oman (Fleitmann et al., 2007), Moomi Cave from Socotra Island of Yemen (Shakun et al., 2007), Mawmluh Cave from north India (Berkelhammer et al., 2012; Dutt et al., 2015; Lechleitner et al., 2017), Bittoo Cave from northeast India (Kathayat et al., 2016), Tianmen Cave from western China (Cai et al., 2012), and Xiaobailong Cave plus several nearby caves from south Yunnan in southwestern China (Cai et al., 2006; Liu et al., in preparation; Tan et al., 2017) (Figure 4.3). All the speleothem records generally show a glacial-interglacial trend of $\delta^{18}\text{O}$ decrease, with smaller magnitudes for records from South Africa and Arabian Peninsula (3-4‰), but larger magnitudes for records from India and southwestern China (6-7‰). The Indian Ocean Domain records consistently exhibit positive $\delta^{18}\text{O}$ excursions during Heinrich stadial 2 (H2), Heinrich stadial 1 (H1), and the Younger Dryas (YD), and a negative $\delta^{18}\text{O}$ excursion during the Bølling-Allerød (B/A), with a similar spatial pattern of the magnitudes across the domain. The majority of the Indian Ocean Domain records show an early-Holocene $\delta^{18}\text{O}$ minimum, while the South African record shows a mid-late Holocene $\delta^{18}\text{O}$ minimum. Along the moisture pathway, speleothem $\delta^{18}\text{O}$ generally decreases with increasing distance from the coast

throughout the past 30,000 years. Such a spatial $\delta^{18}\text{O}$ gradient probably also varies with time, with an $\sim 9\text{‰}$ $\delta^{18}\text{O}$ gradient between the more upstream Mawmluh record and the more downstream Xiaobailong and southern Yunnan records during the glacial periods compared with $\sim 6\text{‰}$ during the Holocene.

Figure 4.3 Speleothem $\delta^{18}\text{O}$ records over the past 30,000 years from the Indian Ocean Domain. Also shown is the domain topography map for the corresponding cave sites based on ETOPO5 5-minute gridded elevation data from “Data Announcement 88-MGG-02, Digital relief of the Surface of the Earth. NOAA, National Geophysical Data Center, Boulder, Colorado, 1988”. The colored vertical bars from the left to the right highlight the millennial-scale excursions coinciding with the YD, B/A, H1, and H2, with the cyan and yellow shading indicating wet and dry periods respectively.

4.4.2 Western Pacific Domain

Most areas within the Western Pacific Domain are influenced by the Asian-Australian Summer Monsoons, with southwesterly and northeasterly winds dominating the boreal and austral summer precipitation, respectively. Records in the Western Pacific Domain include those from Fukugaguchi, Ohtaki, Maboroshi, and Kiriana Caves from central Japan (Mori et al., 2018; Shen et al., 2010; Sone et al., 2013), Jiuxian, Sanbao, and Heshang Caves from central China (Cai et al., 2010; Hu et al., 2008a; Wang et al., 2008), Hulu Cave from eastern China (Wang et al., 2001), and Dongge Cave from southern China (Dykoski et al., 2005) (Figure 4.4). Towards the equatorial Maritime Continent, records originate from Snail Shell and Bukit Assam caves in northern Borneo (Carolin et al., 2013; Partin et al., 2007), Tangga Cave in central Sumatra (Wurtzel et al., 2018), several caves in southwestern Sulawesi (Chapter 2), and Liang Lua Cave in Flores (Ayliffe et al., 2013; Griffiths et al., 2009). Further in the Southern Hemisphere, KNI-51, Ball Gown, and Cape Range from northwest and west coasts of Australia (Denniston et al., 2013a; Denniston et al., 2017; Denniston et al., 2013b) and caves in Mount Arthur, Hollywood Cave, and several other caves from South Island, New Zealand (Hellstrom et al., 1998; Whittaker et al., 2011; Williams et al., 2010) are also included in the comparisons.

The speleothem $\delta^{18}\text{O}$ records from central, eastern, and southern China largely resembled the southwestern China speleothem records from the downstream Indian Monsoon region for the past 30,000 years, and shared a 4-5‰ glacial-interglacial variation, except for higher $\delta^{18}\text{O}$ values likely due to

closer moisture source from the western Pacific and the South China Sea. The Jiuxian and Sanbao records from higher elevations in central China were isotopically more depleted compared with the Heshang record from central Chinese lowlands and the Dongge and Hulu records from southern and eastern China. All of these Chinese speleothem records share prominent millennial scale variability, such as the positive excursions during H2, H1, the YD, and around 3ka, along with the negative excursion during the B/A. The speleothem $\delta^{18}\text{O}$ from maritime Japan generally shows comparable or higher $\delta^{18}\text{O}$ than the Dongge and Hulu records, with seemingly less prominent millennial-scale variability.

The Maritime-Continent speleothem $\delta^{18}\text{O}$ also shows a glacial-interglacial decrease trend, with a magnitude of 3-4‰ for the speleothem $\delta^{18}\text{O}$ from Borneo, Sumatra, and Sulawesi and ~2‰ for the Flores speleothem $\delta^{18}\text{O}$. The Borneo, Sumatra, and Sulawesi speleothem $\delta^{18}\text{O}$ records are highly consistent, but more isotopically depleted than the Flores record. The three speleothem records are comparable with the Dongge-Hulu records in the range of $\delta^{18}\text{O}$, and in-phase with the Dongge-Hulu records regarding the millennial-scale $\delta^{18}\text{O}$ variability during the glacial, deglacial, and Holocene periods, except with much subdued amplitudes (Figure 4.5). All Maritime-Continent records generally show relatively weak long-term trends during the glacial and Holocene periods, with a minor Holocene $\delta^{18}\text{O}$ minimum during the mid-Holocene, which contrasts with a much more prominent early-Holocene $\delta^{18}\text{O}$ minimum in the Chinese speleothem records. The Flores speleothem $\delta^{18}\text{O}$ shows anti-phased millennial-scale variabilities in comparison with the Chinese and northern Maritime-Continent speleothem $\delta^{18}\text{O}$ records. Further south, the

Australian and New Zealand records show even smaller glacial-interglacial trends (1-2‰ in magnitude), with relatively isotopically enriched $\delta^{18}\text{O}$, except for the records from the western coast of Australia. All the Australian and New Zealand records are in-phase with the Flores record on the glacial and deglacial millennial-scale variability, and anti-phased with the records from further north.

Figure 4.4 Speleothem $\delta^{18}\text{O}$ records over the past 30,000 years from the northern (left) and southern (right) parts of the Western Pacific Ocean Domain. Also shown is the domain topography map for the corresponding cave sites. The colored vertical bars from the left to the right highlight the millennial scale excursions coinciding with the YD, B/A, H1, and H2, with the green and yellow shading indicating wet and dry periods respectively.

Figure 4.5 Millennial-scale variation in speleothem $\delta^{18}\text{O}$ from south China (top) and SW Sulawesi (bottom) for the past 30,000 years. The shaded bars in yellow and cyan mark isotopically enriched and depleted episodes respectively.

4.4.3 Mediterranean Westerly Domain

At the present day, the Mediterranean Westerly Domain is characterized by westerly moisture from the Atlantic Ocean, and the Mediterranean and Caspian Seas, with boreal winter precipitation prevailing over Israel and central Asia, and summer precipitation prevailing over western China. The records in the Westerly Domain are from Soreq (Bar-Matthews et al., 2003), Jerusalem West (Frumkin et al., 2000), Jeita Caves in Israel (Cheng et al., 2015), Tonnelnaya Cave in Uzbekistan (Cheng et al., 2016b), and Kesang Cave in western China (Cheng et al., 2012b) (Figure 4.6). These records generally resemble Indian and Chinese speleothem records in their orbital and millennial-scale variations. The speleothem $\delta^{18}\text{O}$ generally becomes lower following the moisture pathway from upstream Israel to downstream western China. The $\delta^{18}\text{O}$ gradient shifted from 2‰ between Israel and central Asia and 4‰ between central Asia and western China during the early Holocene to 0‰ and 1‰, respectively, during the late Holocene.

Figure 4.6 Speleothem $\delta^{18}\text{O}$ records over the past 30,000 years from the Mediterranean Westerly Domain. Also shown is the domain topography map for the corresponding cave sites. The colored vertical bars from the left to the right highlight the millennial-scale excursions coinciding with the YD, B/A, H1, and H2, with the green and yellow shading indicating wet and dry periods, respectively.

4.4.4 American Domain

The American Domain comprises both the North American and South American continents. North America is dominated by easterly and southeasterly moisture sources during boreal summer. During boreal winter, the Pacific westerly also contributes a significant amount of precipitation in the southwestern United States. South America is characterized by austral summer precipitation, known as the South American Summer Monsoon, with tropical Atlantic moisture transported by equatorial trade winds. This moisture later turns southeastward along the Andes, and forms a low-level jet towards central and southeastern Brazil.

The records in the American Domain are from Bells (Wagner et al., 2010), Fort Stanton (Asmerom et al., 2010), Pink Panther (Asmerom et al., 2007), and Buckeye Creek Caves (Hardt et al., 2010) in the United States, Juxtlahuaca (Lachniet et al., 2013) and Diablo Caves (Bernal et al., 2011) in Mexico, Terciopelo (Lachniet et al., 2009) and Venado Caves (Lachniet et al., 2004a) in Costa Rica, Chilibrillo Cave in Panama (Lachniet et al., 2004b), Abaco Island Cave in the Bahamas (Arienzo et al., 2015), Paraíso (Wang et al., 2017) and Tigre Perdido Caves in the Amazon, Condor (Cheng et al., 2013b), Diamante, Huagapo (Kanner et al., 2013), and Pacupahuain Caves (Kanner et al., 2012) in Peru, Santiago Cave in Ecuador (Mosblech et al., 2012), several caves in Rio Grande do Norte in northeastern Brazil (Cruz et al., 2009), and Botuverá Cave in southern Brazil (Cruz Jr et al., 2005; Wang et al., 2007) (Figure 4.7).

The southwestern United States records show variations that appear generally anti-phased with variations in other records from North and Central

Americas over both orbital and millennial timescales, with lower speleothem $\delta^{18}\text{O}$ during the glacial periods and during stadials such as H2, H1, and the YD, and higher speleothem $\delta^{18}\text{O}$ during the Holocene and B/A. In contrast, other records from North and Central Americas show a glacial-interglacial trend of $\delta^{18}\text{O}$ decrease, with generally positive excursions during H2 and H1, in contrast with a negative excursion during the B/A. The records from Mexico, located more downstream of the easterly wind, are generally more isotopically depleted compared with the upstream records from the East Coast of the United States and the Bahamas, with the spatial gradient getting smaller from the glacial period to the Holocene (from $\sim 6\text{‰}$ to $\sim 3\text{‰}$). During the Holocene, all the North and Central American records show prominent positive excursions during the 8.2 ka event.

In South America, the speleothem $\delta^{18}\text{O}$ from eastern Amazonian Paraíso Cave and northeastern Brazilian Rio Grande do Norte still shows large magnitudes of glacial-interglacial $\delta^{18}\text{O}$ decrease (5-6‰). The $\delta^{18}\text{O}$ gradient between the eastern Amazon and northeastern Brazil records increases from 0‰ in the mid-Holocene to $\sim 3\text{‰}$ in the late-Holocene. Towards the western Amazon, glacial-interglacial variation of speleothem $\delta^{18}\text{O}$ becomes minor, while further in the Andes and southern Brazil, speleothem $\delta^{18}\text{O}$ shows a glacial-interglacial trend of increase. The spatial gradient of speleothem $\delta^{18}\text{O}$ also decreased from the glacial period to the Holocene, shifting from $\sim 5\text{‰}$ to 0‰ between both the eastern and western Amazon, and from $\sim 10\text{‰}$ to $\sim 6\text{‰}$ between the lowland western Amazon and the Andes. Despite different glacial-interglacial patterns, all the South American records show isotope depletion during H2, H1, and the YD but enrichment during the B/A.

Figure 4.7 Speleothem $\delta^{18}\text{O}$ records over the past 30,000 years from North (left) and South (right) America in the American Domain. Also shown is the domain topography map for the corresponding cave sites. The colored vertical bars from the left to the right highlight the millennial-scale excursions coinciding with the 8.2 ka event, YD, B/A, H1, and H2, with the green and yellow shading indicating wet and dry periods, respectively.

4.5 Discussion

4.5.1 Spatial gradient of speleothem $\delta^{18}\text{O}$ and its temporal evolution

The spatial pattern of selected tropical and subtropical speleothem $\delta^{18}\text{O}$ generally resembles that of rainfall $\delta^{18}\text{O}$ (Bowen and Wilkinson, 2002; Rozanski et al., 1993), confirming the robustness of speleothem $\delta^{18}\text{O}$ as a proxy of rainfall $\delta^{18}\text{O}$ in the tropics and parts of the subtropics. The spatial pattern of speleothem $\delta^{18}\text{O}$ suggests the continental and altitude rainout effect as shown by rainfall $\delta^{18}\text{O}$, with the $\delta^{18}\text{O}$ gradient varying through the glacial-interglacial timescale. A more precise quantification of the $\delta^{18}\text{O}$ gradient needs to consider the distance between cave sites as well as their altitude difference. Generally larger $\delta^{18}\text{O}$ gradients during the last glacial period highlight larger isotopic fractionation across the land when the temperature was lower. In contrast, reduced $\delta^{18}\text{O}$ gradients between cave sites might be attributed to negligible isotopic fractionation due to strong vegetational transpiration (Wang et al., 2017).

The selected speleothem $\delta^{18}\text{O}$ records show larger magnitudes of glacial-interglacial variation for the upstream monsoonal sites but smaller for the maritime and downstream monsoonal sites. Such differences highlight diverse sensitivities of speleothem $\delta^{18}\text{O}$ for large continents and maritime islands, probably associated with different degrees of land heating. Additionally, they also highlight the difference between the upstream and downstream monsoonal records. The difference has been suggested to be caused by a stronger continental rainout effect and weaker vegetation transpiration, termed as

moisture pathway effect, which have cancelled out the large magnitude of glacial-interglacial $\delta^{18}\text{O}$ variation registered in the upstream sites, leading to smaller magnitudes in the downstream (Liu et al., in preparation).

In the Mediterranean Westerly Domain, the larger $\delta^{18}\text{O}$ gradient in the early Holocene was likely associated with larger continental isotopic fractionation, which was caused by land cooling due to glacial melting at high altitudes in central Asia and western China (Marcott et al., 2013). The increasing $\delta^{18}\text{O}$ gradient from the mid- to late Holocene between northeastern Brazil and the eastern Amazon, and between the eastern and western Amazon may have been caused by weakened vegetation transpiration due to forest contraction in drier climate (Wang et al., 2017).

4.5.2 The mean position of the ITCZ

The speleothem $\delta^{18}\text{O}$ variation over glacial-interglacial and millennial timescales allows us to demarcate probable hydroclimate regimes within each domain, with each regime indicated by shifts in the pattern of orbital-millennial $\delta^{18}\text{O}$ variation between nearby cave records. For instance, in the deep tropics of the Western Pacific Domain, shifts in the speleothem $\delta^{18}\text{O}$ between the northern Maritime Continent and the Lesser Sunda Islands suggest different hydroclimate regimes separated by the regional mean ITCZ (Chapter 2). This demarcation is confirmed by satellite and rainfall observations, showing the annual mean ITCZ covers the majority of the Maritime Continent north of the Lesser Sunda Islands.

Similar demarcation probably also exists in the tropical Amazon, with zonal distinction of orbital-scale $\delta^{18}\text{O}$ variation between the eastern Amazon records on the one hand and the western Amazon records as well as those from further downstream on the other hand; yet all of these South American records show a consistent pattern of millennial-scale variability. Such shifts of record correlation across different timescales could be interpreted that the mean position of the ITCZ was located between the two regions on the orbital scale, but moved further north of all the South American cave sites on millennial timescales. This explanation is however problematic as the ITCZ is known to have frequently migrated southward during millennial-scale cold events (Chapter 2). Also, the present-day annual mean ITCZ in the region appears to include both the eastern and western Amazon cave sites in the strong convection. Alternatively, the orbital antiphase between the eastern and western Amazon cave sites can be explained by a zonal dipole mechanism essentially forced by insolation (Cruz et al., 2009). This explanation, however, still requires a mean ITCZ north of all the South American cave sites to explain the anti-phased millennial scale variability between central and South American cave records.

4.5.3 ITCZ migration and speleothem $\delta^{18}\text{O}$

Here, I compared the speleothem $\delta^{18}\text{O}$ records with ITCZ migration to estimate the influences of ITCZ migration in each domain (Figure 4.8-4.11). To sum up, during the past 30,000 years, upstream monsoonal and maritime speleothem $\delta^{18}\text{O}$ appears to have been dominantly driven by SST changes in the nearby oceans on the glacial-interglacial timescale, with the influences of

ITCZ migration only dominating during millennial timescales. In contrast, the subtropical downstream monsoonal speleothem $\delta^{18}\text{O}$ was predominantly controlled by ITCZ migration, which was essentially driven by precession changes on the orbital timescale (see details in Chapter 2). Thus, such a spatial difference in speleothem $\delta^{18}\text{O}$ variation can essentially be linked to land heating by ocean SST at coastal sites but local summer insolation in continental interiors. Other Southern Hemisphere speleothem $\delta^{18}\text{O}$ records, including those from South Africa (Figure 4.8), Flores, Australia, and New Zealand (Figure 4.9), followed ITCZ migration reversely. They however present more amplified responses in $\delta^{18}\text{O}$ during the Holocene in comparison with the glacial periods, which were likely associated with enhanced water vapor feedback caused by Holocene ocean warming.

Figure 4.8 Speleothem $\delta^{18}\text{O}$ records from the Indian Ocean Domain in comparison with ITCZ index of $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ and tropical Indian Ocean SST. The grey lines with no scales are $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ (Chapter 2); more upward direction of $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ indicating more northward mean position of the ITCZ, except with reversed pattern for the last panel. The red lines are SST based on Mg/Ca ratio from marine sediment Core SK237-GC04 (10°58.65'N, 74°59.96'E) from equatorial Indian Ocean (Saraswat et al., 2013). The downstream Indian Monsoon records generally followed ITCZ migration during both glacial-interglacial and millennial timescales, while the upstream Indian Monsoon records only followed ITCZ migration during the deglaciation and part of the Holocene, but decoupled with ITCZ migration during the last glacial periods and the early Holocene. Instead, the upstream monsoon records largely followed nearby northern Indian Ocean sea surface temperature (SST) during the past 30,000 years. The South African record also coincided with ITCZ migration for the past 30,000 years, but in an opposite way, with probably amplified $\delta^{18}\text{O}$ response to ITCZ migration during the early Holocene.

Figure 4.9 Speleothem $\delta^{18}\text{O}$ records from the Western Pacific Domain in comparison with ITCZ index $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ and tropical Pacific Ocean SST. The grey lines with no scales are $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ (Chapter 2); more upward direction of $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ indicating more northward mean position of the ITCZ for the first three panels and more southward ITCZ position for the last three panels. The red lines are SSTs based on $\text{U}^{\text{K}'}_{37}$ from a South China Sea sediment core 17940 (20°7.0'N, 117°23'E) (Müller et al., 1998; Pelejero et al., 1999) and *G. ruber* Mg/Ca ratio from marine sediment Core SO217-18526 (3°37'S 118°10'E) (Schröder et al., 2018). Similar with the downstream Indian Monsoon records, Chinese speleothem records also tightly tracked ITCZ migration during the past 30,000 years. However, both the maritime Japan and northern Maritime Continent records decoupled from ITCZ migration during the glacial periods and the early Holocene, and largely tracked the tropical Western Pacific SST instead. Further south, the Flores, Australian, and New Zealand records, similar to the South African record, also generally followed ITCZ migration, but with amplified $\delta^{18}\text{O}$ responses during the early Holocene.

Figure 4.10 Speleothem $\delta^{18}\text{O}$ records from the Mediterranean Westerly Domain in comparison with ITCZ index $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ and Mediterranean SST. The grey lines with no scales are $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ (Chapter 2); more upward direction of $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ indicating more northward mean position of the ITCZ. The red line is SST based on $\text{U}^{\text{K}'}_{37}$ from marine sediment Core MD95-2043 (36° 8.6'N, 2°37.3'W) in the western Mediterranean (Cacho et al., 2001). The upstream records from Israel also largely tracked the Mediterranean SST during the past 30,000 years, whereas the downstream records generally followed ITCZ migration during the Holocene.

Figure 4.11 Speleothem $\delta^{18}\text{O}$ records from the American Domain in comparison with ITCZ index $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ and nearby SST. The grey lines with no scales are $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ (Chapter 2); more upward direction of $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ indicating more southward mean position of the ITCZ for the first one and the last three panels and more northward ITCZ position for the rest of panels. The first two red lines are SSTs based on UK'37 from marine sediment Core LPAZ 21P (22°59.02'N, 109°28.00'W) offshore California (Herbert et al., 2001) and Mg/Ca ratio from marine sediment Core KNR166-2-26JPC (24°19.61'N, 83°15.14'W) offshore Florida Straits (Schmidt and Lynch-Stieglitz, 2011; Schmidt et al., 2012; Them et al., 2015). The remaining red lines are based on alkenone from marine sediment Core GeoB 3910-2 (4°14.7'S, 36°20.7'W) (Jaeschke et al., 2007). In the American Domain, the southwestern American records reversely tracked ITCZ migration during the deglaciation, and offshore California SST on the glacial-interglacial timescale. The upstream easterly records in central America also followed Florida Strait SST during the past 30,000 years, while the downstream easterly records largely coincided with ITCZ migration. In the South American Monsoon region, while upstream northeastern Brazil and eastern Amazon records tracked tropical Atlantic SST for the past 30,000 years, the western Amazon and Andes records tracked tropical Atlantic SST reversely. The downstream southern Brazil record, however, tightly followed ITCZ migration.

4.5.4 Driving mechanisms of tropical-subtropical speleothem $\delta^{18}\text{O}$ during the last deglaciation and Holocene

4.5.4.1 The last deglaciation

The last deglaciation marks a period when major global climate changes occurred. Selected deglacial speleothem $\delta^{18}\text{O}$ records generate the first two of EOFs contributing 62.2% of the total variance, between which EOF1 and EOF2 explain 43.4% and 18.8%, respectively. Both EOFs show relatively evenly distributed negative and positive factor loading in speleothem records (Figure 4.12). The remaining EOFs, each explaining less than 10% of total variance, are mostly dominated by individual records and hence less significant. EOF1 has clearly divided selected speleothem records into north and south hemispheric patterns by negative and positive factor loadings, suggesting a dominating bipolar seesaw response in the speleothem records during the last deglaciation. On the other hand, EOF2 appears to distribute more randomly, with different records from the same locations showing completely opposite patterns.

Corresponding PC1 shows overall increasing trend disrupted by two drops. More specifically, PC1 was relatively stable between 21 ka and 19 ka, followed by decreasing towards around 15ka; it then rose till 13 ka and decreased again before an increase from around 12.5 ka onwards (Figure 4.13). Such variation resembles the interhemispheric temperature gradient (IHTG), the ITCZ, and the strength of Atlantic Meridional Ocean Circulation (AMOC) strength during the last deglaciation, indicating the dominant influences of ITCZ migration on tropical and subtropical hydroclimate, with ITCZ dynamics forced by varying

IHTG associated with precession and AMOC strength changes as well as glaciation (Schneider et al., 2014). PC2 marks a feature of $\delta^{18}\text{O}$ reversal in the selected speleothem records, which is most prominent in speleothem records whose millennial-scale excursions during the deglaciation was in less clear structure or smaller magnitude, such as some maritime-continent records or upstream continental records.

Figure 4.12 The bi-plot and maps of EOF1 and EOF2 for the deglaciation. The red and blue shading and circles mark the records with respectively positive and negative factor loading on EOF1.

Figure 4.13 The PCs and climatic forcings during the last deglaciation. a, PC1 (magenta) with precession (grey) (Laskar et al., 2004). b, Relative sea level reconstructed based on benthic $\delta^{18}\text{O}$ (Waelbroeck et al., 2002) c, Global Interhemispheric temperature gradient (Shakun et al., 2012). d, $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ from the maritime continent as an ITCZ migration index (Chapter 2). e, $^{231}\text{Pa}/^{230}\text{Th}$ from North Atlantic sediment core OCE326-GGC5 as reconstructed AMOC strength (McManus et al., 2004). f, Modeled AMOC strength (He, 2011; Shakun et al., 2012). g, PC2.

4.5.4.2 The Holocene

Selected Holocene speleothem $\delta^{18}\text{O}$ records generate the first two of EOFs contributing 67.6% of the total variance, between which EOF1 and EOF2 explaining respectively 44.2% and 23.4%. The remaining EOFs, each explaining less than 10% of total variance, are mostly dominated by individual records and hence less significant. EOF1 shows relatively evenly distributed negative and positive factor loading in speleothem records (Figure 4.14), also dividing speleothem records into north and south hemispheric patterns by negative and positive factor loadings, similar with the deglacial EOF1. EOF2 shows positive loadings for almost all the records, with the largest loading in the deep tropics.

Corresponding PC1 is characterized by overall decreasing trend during the Holocene. PC1 shows a relatively stable peak after the rise from 11.5ka, followed by gradual decrease from around 9.5 ka till 0.7 ka (Figure 4.15), after which it started to rise again. Such variation resembles the Northern Hemisphere extra-tropical temperature variation and ITCZ migration, both of which essentially follow the Northern Hemisphere summer insolation. Overall, during the Holocene, ITCZ migration driven by precessional insolation still dominated in driving tropical and subtropical hydroclimate changes.

PC2 is characterized by a continuous decrease since the early Holocene until a trough during the mid-Holocene (around 4-6ka), after which it increased till the late Holocene. PC2 coincides with zonal-averaged tropical temperature variation (Marcott et al., 2013) as well as ENSO variability during the Holocene (Carré et al., 2014; Cobb et al., 2013). The mid-Holocene transition

in equatorial paleoclimate records is likely associated with contemporary equinox insolation shifts in the deep tropics, causing stronger deep convection and enhanced Walker circulation accompanied by dampened ENSO activity.

Noticeably, the EOF bi-plots and maps can effectively distinguish anti-phased hydroclimate responses to ITCZ dynamics in speleothem records for both periods. While most geographically nearby speleothem records are also close to each other in the EOF bi-plots and maps, the two aforementioned speleothem records from Sulawesi and Flores in the maritime continent, based on which the ITCZ migration index is derived from, are relatively further separated in both EOF bi-plots and maps. With previous study suggesting that the two sites are located on different sides of the ITCZ (see details in Chapter 2), such separation between geographically nearby speleothem records in the EOF bi-plots implies a possible hydroclimate boundary between the sites.

Figure 4.14 The bi-plot and maps of EOF1 and EOF2 for the Holocene. The red and blue shading and circles mark the records with respectively positive and negative factor loading on EOF1.

Figure 4.15 The PCs and climatic forcings during the Holocene. a, PC1 (magenta) and North Hemisphere summer insolation (grey) (Laskar et al., 2004). b, Zonal mean temperature between 90°N and 30°N (Marcott et al., 2013). The anomaly is relative to 1961-1990 instrumental mean. c, Bulk Ti content of Cariaco Basin sediments from ODP Site 1002 (Haug et al., 2001). d, $\Delta\delta^{18}\text{O}_{\text{SW-FL}}$ from the maritime continent as an ITCZ migration index (Chapter 2). e, PC2 (blue) and equatorial September insolation (pink) (Laskar et al., 2004). f, Zonal temperature anomaly between 30°N and 30°S (Marcott et al., 2013). The anomaly is relative to 1961-1990 instrumental mean. g, The variation of El Niño frequency indicated by botryococcene concentration in El Junco Lake sediment (log scale, in line) (Zhang et al., 2014) and ENSO variance indicated by changes in the standard deviation of 2-7-year variance of fossil coral $\delta^{18}\text{O}$ values in central Pacific islands (in scatter) relative to modern coral $\delta^{18}\text{O}$ (Cobb et al., 2013).

4.5.5 A step further: preliminary speleothem $\delta^{18}\text{O}$ isoscapes

The existing compilation allows further synthesis efforts such as preliminary isoscapes of speleothem $\delta^{18}\text{O}$, so as to better illustrate the spatial distribution of tropical and subtropical speleothem $\delta^{18}\text{O}$ over different time periods including the Last Glacial maximum (21-19 ka), Early Holocene (11-9 ka), Mid-Holocene (4-7 ka), and Late Holocene (0-2 ka) (Figure 4.16). Although the interpolation of isoscapes might be biased and appear unrealistic for areas with a limited data coverage, but the large-scale speleothem $\delta^{18}\text{O}$ distribution is likely reliable, as verified by the general consistence of the spatial pattern between the modern rainfall $\delta^{18}\text{O}$ isoscape (Bowen and Wilkinson, 2002) and the reconstructed speleothem $\delta^{18}\text{O}$ isoscape for the late Holocene. Such series of speleothem $\delta^{18}\text{O}$ isoscapes allow us to investigate the temporal changes of tropical and subtropical speleothem $\delta^{18}\text{O}$ and hence hydroclimate changes.

In addition, speleothem $\delta^{18}\text{O}$ during any climatically critical period can be compared with the late Holocene value to estimate overall hydroclimate changes during the interested time periods (Figure 4.17). For instance, a broadly positive offset between the LGM and the late Holocene probably reveals generally drier conditions prevalent in the low latitudes during the LGM, whereas the speleothem $\delta^{18}\text{O}$ difference relative to the late Holocene indicates a gradually southward movement of wetness in the low latitudes from the early to mid-Holocene. The magnitude of speleothem $\delta^{18}\text{O}$ variation during a climatic event or a transition period can also be mapped to study various

responses of speleothem $\delta^{18}\text{O}$ to particular climatic forcings. For instance, while speleothem $\delta^{18}\text{O}$ difference between the early Holocene and the LGM suggests an overall increase of wetness in the tropics during the last deglaciation (with exception for several Southern Hemisphere locations), the deep tropics seem to be less sensitive to the millennial-scale forcing during the last deglaciation (Figure 4.18). Similarly, speleothem $\delta^{18}\text{O}$ difference between H1 and the LGM reveals a clear seesaw pattern in hydroclimate changes during H1. More careful evaluation of hydroclimate changes requires consideration and calibration of other global or site-specific influences including varying seawater $\delta^{18}\text{O}$ caused by ice volume change, changes of temperature, high elevation, and among others. Nevertheless, such mapping of speleothem $\delta^{18}\text{O}$ variation on a global scale plays irreplaceable roles on hinting at the mechanisms of climate changes over different timescales.

Figure 4.16 Speleothem $\delta^{18}\text{O}$ isoscapes for different periods.

Figure 4.17 Isoscapes for speleothem $\delta^{18}\text{O}$ difference between specific periods and the late Holocene.

Figure 4.18 Isoscapes for speleothem $\delta^{18}\text{O}$ variation during the last deglaciation and Heinrich event 1. The magnitude of variation is based on speleothem $\delta^{18}\text{O}$ difference between the early Holocene and the LGM for the last deglaciation and between H1 and the LGM for H1.

4.6 Conclusion

Speleothem $\delta^{18}\text{O}$ from the low latitudes are synthesized on regional to global scales to investigate its spatial and temporal variation and corresponding driving mechanisms during the past 30,000 years. The spatial gradient of speleothem $\delta^{18}\text{O}$ has been primarily controlled by continental rainout effects, with isotopic fractionation having been dominated by temperature and continental water recycling. On the glacial-interglacial timescale, monsoonal speleothem $\delta^{18}\text{O}$ largely followed SST in the upstream but ITCZ migration in the downstream. On millennial timescales, tropical and subtropical speleothem $\delta^{18}\text{O}$ were dominated by ITCZ migration, particularly during the last deglaciation, with the magnitudes of variation smaller for the maritime region but larger for the continents. Statistical analysis on selected speleothem $\delta^{18}\text{O}$ records on a global scale confirmed the dominant control of ITCZ migration during both the deglaciation and Holocene, driven by the varying interhemispheric temperature gradient resulted from changes in precession, continental ice volume, and ocean circulations. During the Holocene, deep tropical forcing, probably associated with ENSO activity, also contributed prominently to speleothem $\delta^{18}\text{O}$ variances. Further synthesizing efforts of global isoscapes of speleothem $\delta^{18}\text{O}$ and its variation can effectively hint at the forcing mechanisms of past climate changes during various periods and on various timescales.

4.7 References

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Chapter 5

Conclusion

5.1 Summary of key findings

In the previous three chapters, I have established speleothem stable isotopes and trace elements records from southwestern Sulawesi, and synthesized global tropical and subtropical speleothem oxygen isotope records since the Last Glacial Maximum (LGM). The multi-proxy speleothem records from Sulawesi provide a comprehensive picture of regional climate and environmental conditions during the past 30,000 years. Comparison of the Sulawesi speleothem oxygen isotope record with other speleothem oxygen isotope records from the Maritime Continent constrains the regional mean position of the Intertropical Convergent Zone (ITCZ) within the Flores Sea, based on which I have developed an ITCZ index indicative of its past dynamics. Direct comparisons between the ITCZ index and synthesized speleothem oxygen isotope records from low latitudes, along with Empirical Orthogonal Function (EOF) analysis of these records, reveal dominant driving mechanisms of tropical and subtropical hydroclimate changes since the LGM.

In Chapter 2, I have demonstrated that the discrepancies in existing paleoclimate records over varied timescales across the Maritime Continent can be partially attributed to the migration of the ITCZ. I have constrained the regional mean position of the ITCZ to be within the Flores Sea. The mean ITCZ position highlights the demarcation of distinct hydroclimate regimes. North of the Flores Sea, the deep convection of the ITCZ results in highly

consistent and generally lower speleothem $\delta^{18}\text{O}$ over Sumatra, Borneo and Sulawesi. South of the Flores Sea, beyond the reach of the ITCZ, speleothem $\delta^{18}\text{O}$ becomes higher in Flores and southwestern Australia, with millennial-scale variability anti-phased with those north of the Flores Sea. The distinct hydroclimate responses between Sulawesi and Flores, as well as their geographical proximity, allowed me to derive an ITCZ index based on speleothem $\delta^{18}\text{O}$ from the two locations. By analyzing the index, I have demonstrated that the ITCZ was located further south during the LGM compared to today. As a result, during the LGM, the northern portion of the Maritime Continent experienced weakened ITCZ convection and drier conditions, whereas the southern portion was wetter. Last but not least, I have inferred that the migration of the ITCZ is driven by precession coupled with glaciation and the strength of ocean circulation.

In Chapter 3, by analyzing SW Sulawesi speleothem stable isotopes and trace elements, I have found a robust correlation between tropical vegetation, soil CO_2 production, and climate variability: expanded forests (grasses) and enhanced (weakened) soil CO_2 production coincided with increased (decreased) regional rain amount, all of which generally followed the glacial-interglacial variation of regional sea surface temperature (SST). During the deglaciation, climate boundary conditions, including rising SST and atmospheric CO_2 concentration, dominated continuously forest expansion and soil CO_2 production enhancement; however, increase in regional annual rainfall amount lagged a few thousand years behind, due to the southward migration of the ITCZ during the Heinrich event 1. On the other hand, millennial-scale ENSO-like rainfall anomalies during the Holocene can also cause forest collapse and

drastically decrease in soil CO₂ production in SW Sulawesi, highlighting the strong sensitivity of tropical forests and soil biomes to hydroclimate variability.

In Chapter 4, I have compared speleothem $\delta^{18}\text{O}$ along the moisture pathway, and detected the continental rainout effect previously seen in rainfall $\delta^{18}\text{O}$, thus verifying that speleothem $\delta^{18}\text{O}$ from low latitudes has mainly inherited rainfall $\delta^{18}\text{O}$ variation. This rainout effect has varied with time, with isotopic fractionation controlled by changes of land temperature and continental water recycling. In addition, I have found that the speleothem $\delta^{18}\text{O}$ has been mainly influenced by nearby SST in the monsoonal upstream and ITCZ migration in the downstream. Finally, I have verified the dominant role of ITCZ migration, with secondary influences from deep tropical convection, in driving the variances of speleothem $\delta^{18}\text{O}$ and regional hydroclimate since the LGM.

5.2 Suggestions for future work

5.2.1 Trace elements over inter-annual timescales: ENSO variability during the Holocene

As described in Chapter 3, I have obtained seasonal resolution of trace elements data from the SW Sulawesi speleothems using the LA-ICP-MS technique. In addition, I have conducted preliminary research beyond the scope of Chapter 3. My preliminary examination of the higher-resolution trace element data from the Holocene stalagmite reveals strong inter-annual variability, with a strong correlation between Mg, Sr, and Ba on the one hand and P and U on the other, while at the same time a strong anti-correlation exists between the two groups (Figure 5.1). The in-phase covariance of Mg, Sr, and

Ba can arise from either enhanced selective leaching effects with prolonged water-rock interaction time or prior calcite precipitation (PCP) along the flow path (Fairchild et al., 2000). The correlation between P and U could be related to the amount of organic matter, determined by soil microbial activity and vegetation biomass (Baldini et al., 2002; Fairchild and Treble, 2009).

Considering the modern rainfall seasonality in SW Sulawesi, I attribute the anti-correlation between the two groups of trace elements to the varying strength of bedrock dissolution, which is further controlled by inter-annual variation of rainfall amount and seasonality. Since the largest inter-annual climatic variability in the Maritime Continent is the El Niño–Southern Oscillation (ENSO) at the present day, I hypothesize that ENSO variability in the past was also responsible for the inter-annual variability in SW Sulawesi speleothem trace elements. .

Due to the relatively low contents of Sr, Ba, P, and U, I have amplified the common inter-annual signal of bedrock dissolution and the ENSO by division between comparable Sr and P as well as Ba and U. The resulting Sr/P and Ba/U ratios show prominent millennial-scale variability that is much less obvious in the raw data, and that also coincides with corresponding excursions in Mg and stable isotopes (Figure 5.2). These prominent millennial-scale excursions are very likely associated with enhanced millennial-scale ENSO activity during the Holocene.

I will further test the inter-annual correlations among trace elements in glacial stalagmites and perform spectral analysis for more detailed information about the variances over the inter-annual spectrum. Potentially, I may be able to

reconstruct the ENSO variability since the LGM, and analyze its evolution during the last glacial-interglacial period.

Figure 5.1 The inter-annual variability of trace elements for a portion of stalagmite SAL2.

Figure 5.2 The millennial scale variability of trace elements for stalagmite SAL2.

5.2.2 The glacial-interglacial hydroclimate variation over the South China Sea Basin

To characterize the past ITCZ dynamics over the Southeast Asia in a more comprehensive way, I propose to establish another speleothem record from central Vietnam, which forms a N-S transect with speleothem records from SW Sulawesi. We have conducted a field survey in central Vietnam in November 2016, during which 26 speleothem samples were collected. To date I have obtained ~80 uranium-thorium dates and 600 stable isotope data from the collected speleothems (Figure 5.6). I have established a preliminary stable isotope record that discontinuously covers the past 80,000 years, with majority of the samples growing during the last glacial period. Stable isotope data from different caves and stalagmites replicate well during their contemporaneous periods. The speleothem $\delta^{18}\text{O}$ during the LGM was found comparable with the modern value. I plan to measure trace elements on Vietnam speleothems for a more comprehensive picture of regional climatic and environmental changes over the glacial-interglacial timescale. I will also compare central Vietnam and SW Sulawesi speleothem records for a better understanding on ITCZ dynamics during the last glacial-interglacial periods.

Figure 5.3 Speleothem stable isotopes from central Vietnam. Different colors represent different stalagmites.

5.2.3 Reconstruct the temperature during the Last Glacial Maximum with speleothems from Southeast Asia

Current estimates of climate sensitivity have a relatively large uncertainty ($\sim 1.5\text{--}4.5^{\circ}\text{C}$). Temperature reconstruction from the past with comparable magnitude of changes in atmospheric CO_2 concentration can provide a robust hindcast for model's performances (Hargreaves et al., 2012). Existing paleo-temperature reconstructions, particularly in the Indo-Pacific region, are however, very scarce and even controversial (Annan and Hargreaves, 2013). In speleothems, mean annual surface temperature can be derived from isotopic ratios in inclusion water and coeval carbonate. I plan to reconstruct the LGM temperature using speleothem water inclusion from a wide spatial coverage of Southeast Asia including SW Sulawesi, central Vietnam and also central Myanmar. Such reconstruction has been rare due to technical difficulties. With recent breakthroughs in fluid extracting method and cavity ring-down spectroscopic (CRDS) technique on isotopic analysis (Uemura et al., 2016), I propose to quantify the glacial-interglacial temperature change in a spatial context of Southeast Asia. My preliminary results from water inclusion analysis on SW Sulawesi speleothems suggest $3.9 \pm 0.6^{\circ}\text{C}$ cooling during the LGM, which shows the potential of obtaining more precise LGM temperature estimates with speleothems from Southeast Asia. A detailed description on this work can be found in Appendix D.

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Appendices

Appendix A: Scanned images of SW Sulawesi speleothems

Appendix B: Age models of SW Sulawesi speleothems by StalAge

Figure 1. Age models for SW Sulawesi speleothems by StalAge (Scholz and Hoffmann, 2011). StalAge models generally agree with the age models based on linear interpolation, except for wrongly treating the SLML4 which grew discontinuously during the mid-Holocene and the Last Glacial Maximum as continuous growth.

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Appendix C: Stable isotope data for SW Sulawesi speleothems

SAL3-Age (kyr BP)	SAL3- $\delta^{18}\text{O}$ (‰ VPDB)	SAL3- $\delta^{13}\text{C}$ (‰ VPDB)	SAL3-Age (kyr BP)	SAL3- $\delta^{18}\text{O}$ (‰ VPDB)	SAL3- $\delta^{13}\text{C}$ (‰ VPDB)
0.087	-8.736	-12.918	2.011	-9.480	-13.334
0.135	-8.717	-13.086	2.097	-8.577	-12.793
0.184	-8.841	-13.276	2.493	-8.763	-13.186
0.232	-8.700	-13.367	2.890	-8.372	-10.552
0.280	-8.941	-13.512	3.286	-8.312	-11.070
0.329	-8.529	-13.123	3.325	-8.563	-11.792
0.377	-9.530	-14.035	3.365	-8.336	-12.442
0.425	-9.081	-13.518	3.404	-8.682	-12.219
0.474	-9.202	-13.849	3.443	-8.434	-11.875
0.522	-8.867	-13.824	3.482	-8.837	-12.373
0.570	-8.329	-13.505	3.521	-8.880	-12.551
0.630	-8.225	-12.514	3.560	-8.246	-12.344
0.690	-9.055	-13.189	3.600	-8.627	-12.889
0.749	-8.732	-12.602	3.639	-8.503	-13.121
0.809	-8.573	-12.341	3.678	-9.452	-13.040
0.868	-8.680	-12.424	3.717	-8.741	-12.849
0.928	-8.478	-12.610	3.756	-8.770	-13.010
0.988	-8.808	-13.513	3.795	-8.546	-13.414
1.009	-8.646	-13.297	3.835	-9.264	-13.243
1.031	-8.828	-13.257	3.874	-8.951	-12.906
1.053	-8.217	-12.301	3.913	-8.278	-12.777
1.075	-8.884	-13.289	3.952	-8.863	-12.944
1.097	-8.964	-13.032			
1.118	-8.353	-12.854			
1.140	-7.839	-12.472			
1.162	-8.703	-12.820			
1.184	-8.953	-12.936			
1.206	-8.294	-12.999			
1.227	-8.835	-13.346			
1.249	-8.825	-13.334			
1.271	-8.710	-13.770			
1.293	-8.931	-13.463			
1.315	-8.873	-13.373			
1.339	-9.268	-13.420			
1.363	-8.617	-12.892			
1.387	-9.572	-13.397			
1.412	-8.929	-13.495			
1.460	-8.653	-13.075			
1.484	-9.004	-12.839			
1.509	-8.280	-12.473			
1.533	-8.744	-13.371			
1.557	-8.947	-12.865			
1.581	-8.740	-12.251			
1.667	-8.810	-13.264			
1.753	-9.155	-13.078			
1.839	-8.746	-12.743			
1.925	-9.102	-12.782			

SAL2-Age (kyr BP)	SAL2- $\delta^{18}\text{O}$ (‰ VPDB)	SAL2- $\delta^{13}\text{C}$ (‰ VPDB)	SAL2-Age (kyr BP)	SAL2- $\delta^{18}\text{O}$ (‰ VPDB)	SAL2- $\delta^{13}\text{C}$ (‰ VPDB)
1.128	-8.377	-11.266	4.623	-9.229	-12.702
1.203	-8.640	-11.726	4.661	-9.369	-12.742
1.278	-8.677	-12.407	4.736	-9.139	-12.786
1.353	-8.340	-12.139	4.773	-8.875	-12.545
1.429	-8.612	-11.937	4.811	-9.003	-12.837
1.654	-8.617	-11.429	4.850	-9.084	-12.730
1.729	-8.233	-10.801	4.890	-9.294	-13.198
1.805	-8.690	-12.172	4.929	-9.139	-13.535
1.880	-8.609	-12.451	4.968	-9.117	-13.078
1.955	-7.947	-11.267	5.008	-8.884	-12.481
2.030	-8.343	-11.781	5.047	-9.223	-12.875
2.106	-8.183	-11.504	5.086	-9.028	-12.733
2.181	-8.647	-12.078	5.126	-9.248	-12.481
2.256	-8.559	-12.198	5.165	-9.140	-12.939
2.331	-8.801	-12.398	5.204	-8.980	-12.374
2.482	-8.977	-11.957	5.244	-9.371	-12.388
2.557	-8.589	-11.943	5.252	-9.011	-12.448
2.632	-8.582	-10.618	5.260	-8.786	-12.828
2.783	-8.304	-10.034	5.268	-8.913	-12.405
2.858	-7.959	-9.578	5.276	-8.984	-12.412
2.933	-8.669	-11.061	5.284	-9.048	-12.509
3.008	-8.525	-11.918	5.293	-8.869	-12.749
3.084	-8.528	-11.032	5.301	-8.871	-12.433
3.159	-8.599	-11.670	5.309	-9.281	-12.459
3.234	-9.235	-12.314	5.317	-9.166	-12.239
3.309	-9.060	-11.888	5.325	-9.341	-12.600
3.385	-8.633	-11.546	5.342	-8.851	-12.565
3.460	-8.806	-12.321	5.358	-8.701	-12.544
3.535	-8.979	-11.848	5.366	-8.868	-12.389
3.610	-8.602	-11.658	5.374	-9.159	-12.390
3.685	-8.538	-11.218	5.382	-9.182	-12.936
3.761	-8.895	-12.498	5.390	-9.049	-12.611
3.836	-8.977	-12.314	5.399	-9.054	-12.406
3.873	-8.900	-12.032	5.407	-9.226	-12.313
3.948	-9.476	-12.522	5.415	-9.308	-12.703
3.986	-9.387	-12.443	5.423	-9.040	-12.804
4.023	-9.308	-12.351	5.431	-8.897	-12.682
4.061	-9.381	-12.557	5.439	-8.850	-12.188
4.098	-9.336	-12.298	5.448	-9.126	-12.500
4.136	-9.199	-12.535	5.456	-9.143	-12.404
4.173	-9.198	-13.042	5.464	-8.994	-12.718
4.211	-9.267	-12.748	5.472	-9.159	-12.611
4.248	-9.296	-12.497	5.480	-9.202	-12.671
4.286	-9.297	-12.073	5.488	-9.022	-12.788
4.323	-9.267	-12.509	5.497	-9.198	-12.658
4.361	-9.177	-13.062	5.505	-9.255	-12.791
4.398	-8.963	-12.450	5.513	-9.216	-12.720
4.436	-8.659	-11.737	5.521	-9.244	-12.894
4.473	-8.809	-11.338	5.529	-9.059	-12.538
4.548	-8.932	-12.423	5.537	-8.986	-12.735
4.586	-9.062	-12.555	5.546	-9.293	-12.988

SAL2-Age (kyr BP)	SAL2- $\delta^{18}\text{O}$ (‰ VPDB)	SAL2- $\delta^{13}\text{C}$ (‰ VPDB)	SAL2-Age (kyr BP)	SAL2- $\delta^{18}\text{O}$ (‰ VPDB)	SAL2- $\delta^{13}\text{C}$ (‰ VPDB)
5.554	-9.272	-12.638	7.042	-9.161	-12.668
5.562	-9.485	-12.615	7.051	-9.150	-12.880
5.595	-8.720	-11.660	7.060	-9.111	-12.584
5.620	-8.932	-11.514	7.069	-8.543	-12.627
5.644	-9.301	-12.318	7.079	-8.682	-12.591
5.669	-9.010	-12.403	7.088	-8.618	-12.457
5.694	-8.509	-12.143	7.097	-9.100	-12.532
5.719	-9.297	-12.879	7.106	-8.797	-12.431
5.743	-9.197	-13.027	7.115	-8.544	-12.435
5.768	-9.093	-12.599	7.124	-9.037	-12.364
5.793	-9.134	-12.887	7.134	-8.945	-12.403
5.818	-9.281	-12.686	7.143	-9.087	-12.651
5.843	-9.192	-12.552	7.152	-9.169	-12.358
5.867	-9.321	-12.945	7.161	-8.787	-12.373
5.892	-9.158	-12.812	7.170	-9.024	-11.900
5.917	-9.311	-12.756	7.180	-9.101	-12.284
5.942	-9.427	-12.650	7.189	-9.115	-12.452
5.966	-9.513	-12.819	7.198	-8.900	-12.252
5.991	-9.400	-12.868	7.207	-9.089	-12.482
6.016	-9.315	-12.709	7.216	-9.065	-12.109
6.041	-9.158	-12.594	7.226	-8.891	-12.431
6.065	-9.302	-12.351	7.235	-8.798	-12.303
6.090	-9.229	-12.133	7.244	-8.489	-12.109
6.115	-9.357	-12.597	7.253	-8.796	-12.090
6.140	-9.113	-12.000	7.262	-8.816	-12.000
6.165	-8.950	-11.811	7.271	-8.763	-12.170
6.189	-9.077	-12.079	7.281	-8.874	-12.301
6.214	-9.073	-12.199	7.290	-9.313	-12.414
6.239	-9.337	-12.404	7.299	-8.852	-12.444
6.264	-9.129	-12.208	7.308	-8.804	-12.241
6.304	-8.826	-12.364	7.317	-8.831	-12.575
6.345	-9.019	-12.524	7.327	-8.976	-12.646
6.426	-8.970	-12.590	7.336	-8.960	-12.445
6.467	-8.810	-12.454	7.345	-8.804	-12.353
6.508	-8.788	-12.094	7.354	-8.887	-12.579
6.548	-8.707	-11.677	7.363	-9.062	-12.312
6.589	-8.875	-11.543	7.373	-8.987	-12.293
6.630	-8.732	-11.325	7.385	-9.004	-12.066
6.670	-8.907	-11.598	7.397	-9.001	-12.427
6.711	-8.646	-11.688	7.410	-9.205	-12.345
6.752	-8.798	-12.291	7.422	-8.649	-11.964
6.792	-8.831	-12.065	7.434	-8.999	-12.101
6.833	-8.875	-12.269	7.459	-8.817	-12.207
6.874	-8.832	-11.795	7.471	-8.698	-11.853
6.914	-8.941	-12.480	7.496	-9.006	-12.700
6.955	-8.736	-12.268	7.508	-9.250	-12.589
6.996	-9.058	-12.140	7.521	-9.063	-12.467
7.005	-9.155	-12.688	7.533	-8.835	-12.276
7.014	-9.106	-12.497	7.545	-8.704	-11.900
7.023	-8.917	-12.518	7.558	-8.842	-12.038
7.033	-9.132	-12.520	7.570	-8.913	-12.168

SAL2-Age (kyr BP)	SAL2- $\delta^{18}\text{O}$ (‰ VPDB)	SAL2- $\delta^{13}\text{C}$ (‰ VPDB)	SAL2-Age (kyr BP)	SAL2- $\delta^{18}\text{O}$ (‰ VPDB)	SAL2- $\delta^{13}\text{C}$ (‰ VPDB)
7.582	-8.884	-12.319	8.227	-8.677	-12.072
7.595	-8.952	-12.223	8.244	-8.490	-12.134
7.607	-8.791	-12.097	8.260	-8.773	-11.891
7.620	-8.777	-12.292	8.277	-8.828	-11.889
7.632	-8.968	-12.170	8.293	-8.849	-11.993
7.644	-9.326	-12.120	8.309	-8.957	-12.384
7.657	-8.911	-12.168	8.326	-8.842	-12.343
7.669	-8.817	-12.045	8.342	-8.700	-12.219
7.681	-9.052	-12.357	8.359	-8.626	-12.089
7.694	-9.079	-12.104	8.375	-8.789	-12.234
7.706	-8.879	-12.051	8.391	-8.619	-11.346
7.718	-9.129	-12.091	8.408	-8.987	-12.333
7.731	-8.837	-12.151	8.424	-8.492	-12.121
7.743	-8.838	-12.490	8.441	-8.343	-11.868
7.755	-8.907	-12.304	8.457	-8.784	-11.744
7.768	-8.980	-12.516	8.473	-8.551	-11.822
7.780	-8.752	-12.332	8.490	-8.642	-12.252
7.792	-9.182	-12.205	8.506	-8.630	-11.810
7.805	-8.789	-12.310	8.523	-8.722	-11.687
7.817	-8.806	-12.124	8.539	-8.440	-11.878
7.829	-8.836	-12.007	8.555	-8.340	-11.700
7.842	-9.052	-12.151	8.572	-8.452	-11.926
7.854	-9.046	-12.100	8.588	-8.511	-12.014
7.867	-8.866	-12.163	8.605	-8.766	-11.782
7.879	-9.030	-11.935	8.621	-8.291	-11.721
7.891	-8.553	-12.306	8.637	-8.364	-11.751
7.903	-8.643	-12.176	8.654	-8.316	-12.139
7.916	-8.774	-11.960	8.687	-8.811	-12.097
7.928	-8.744	-12.199	8.703	-8.969	-12.099
7.940	-8.288	-11.928	8.736	-8.518	-12.514
7.953	-8.430	-12.096	8.752	-8.647	-12.349
7.965	-8.677	-12.146	8.769	-8.801	-11.784
7.977	-8.689	-12.232	8.785	-8.532	-12.234
8.002	-9.133	-12.403	8.801	-9.029	-12.120
8.014	-9.024	-12.572	8.818	-8.532	-11.627
8.026	-8.982	-12.279	8.834	-8.340	-11.596
8.039	-8.827	-12.204	8.850	-8.484	-12.046
8.051	-8.815	-12.309	8.867	-8.680	-12.051
8.063	-8.863	-12.302	8.883	-8.602	-11.910
8.076	-8.969	-12.162	8.900	-8.752	-12.078
8.088	-9.065	-12.391	8.916	-8.712	-12.029
8.100	-8.763	-11.780	8.932	-8.494	-11.947
8.113	-9.060	-12.203	8.949	-8.553	-12.278
8.125	-9.072	-12.197	8.965	-8.188	-11.784
8.137	-8.709	-11.961	8.982	-8.268	-12.381
8.150	-9.039	-12.266	9.123	-8.726	-12.041
8.162	-8.855	-12.036	9.404	-8.380	-11.689
8.174	-8.656	-11.774	9.686	-8.318	-11.850
8.186	-8.609	-11.474	9.827	-8.787	-11.953
8.199	-8.536	-11.641	9.968	-8.484	-11.957
8.211	-8.720	-11.916	10.250	-8.430	-11.285

SAL2-Age	SAL2- $\delta^{18}\text{O}$	SAL2- $\delta^{13}\text{C}$	SLML4-Age	SLML4- $\delta^{18}\text{O}$	SLML4- $\delta^{13}\text{C}$
(kyr BP)	(‰ VPDB)	(‰ VPDB)	(kyr BP)	(‰ VPDB)	(‰ VPDB)
10.391	-8.320	-11.584	4.128	-8.844	-11.986
10.532	-8.072	-10.866	4.197	-8.322	-10.292
10.673	-8.267	-10.851	4.267	-9.505	-12.439
			4.336	-9.427	-13.453
			4.406	-9.064	-12.393
			4.475	-9.026	-12.924
			4.545	-8.824	-12.445
			4.614	-9.122	-13.219
			4.684	-8.389	-11.866
			4.753	-8.804	-12.553
			4.823	-8.924	-12.411
			5.004	-8.950	-12.598
			5.184	-8.819	-11.546
			5.365	-9.306	-12.641
			5.545	-8.500	-12.150
			5.557	-8.293	-12.517
			5.569	-8.069	-10.741
			5.581	-8.465	-12.411
			5.593	-8.502	-11.992
			5.605	-8.748	-12.835
			5.616	-8.644	-12.218
			5.628	-9.007	-12.286
			5.640	-9.145	-13.018
			5.652	-8.438	-12.050
			5.664	-9.227	-13.374
			5.676	-8.872	-12.599
			5.688	-9.559	-13.086
			5.699	-8.857	-12.579
			5.711	-8.961	-12.779
			5.723	-8.983	-13.272
			5.735	-9.626	-13.473
			5.942	-8.960	-13.239
			6.148	-9.104	-13.602
			6.355	-8.715	-11.615
			18.872	-6.816	-2.800
			18.954	-6.252	-3.061
			19.036	-6.302	-3.110
			19.117	-6.351	-2.952
			19.199	-6.867	-4.281
			19.281	-6.827	-3.719
			19.363	-6.902	-3.601
			19.445	-6.403	-2.976
			19.526	-6.778	-3.864
			19.608	-6.508	-3.403
			19.690	-6.824	-3.875

BD1-Age	BD1- $\delta^{18}\text{O}$	BD1- $\delta^{13}\text{C}$	BD1-Age	BD1- $\delta^{18}\text{O}$	BD1- $\delta^{13}\text{C}$
(kyr BP)	(‰ VPDB)	(‰ VPDB)	(kyr BP)	(‰ VPDB)	(‰ VPDB)
10.981	-7.912	-12.096	12.476	-6.833	-10.363
11.036	-8.054	-12.059	12.490	-6.722	-10.621
11.090	-7.598	-12.184	12.505	-6.460	-9.348
11.145	-7.797	-12.259	12.519	-6.708	-10.432
11.200	-7.878	-12.187	12.534	-6.409	-9.979
11.254	-7.987	-12.086	12.549	-6.576	-10.613
11.309	-7.740	-12.101	12.563	-6.165	-10.382
11.407	-7.553	-11.739	12.578	-6.459	-10.121
11.506	-7.209	-11.084	12.592	-6.428	-10.361
11.604	-7.373	-10.266	12.607	-6.963	-10.684
11.703	-6.851	-10.688	12.622	-6.501	-10.623
11.801	-6.673	-10.871	12.636	-6.684	-10.471
11.900	-6.671	-10.658	12.651	-6.711	-10.553
11.916	-7.011	-10.631	12.665	-7.031	-10.904
11.932	-6.864	-10.865	12.680	-6.845	-10.823
11.948	-6.783	-11.153	12.694	-6.579	-10.290
11.964	-7.470	-11.497	12.709	-6.384	-10.337
11.980	-6.812	-11.070	12.711	-6.656	-10.712
11.996	-6.735	-10.788	12.713	-6.819	-10.535
12.012	-6.559	-10.624	12.716	-6.778	-10.386
12.027	-6.776	-10.729	12.718	-6.370	-10.000
12.041	-7.182	-11.258	12.720	-6.925	-10.678
12.056	-6.441	-10.412	12.722	-6.769	-10.321
12.070	-6.532	-10.963	12.724	-6.720	-10.358
12.084	-6.753	-10.943	12.727	-6.605	-10.100
12.099	-6.444	-10.625	12.729	-6.419	-9.529
12.113	-6.353	-11.022	12.731	-6.814	-10.794
12.128	-6.590	-11.114	12.735	-6.856	-10.999
12.142	-7.884	-10.994	12.740	-7.058	-10.509
12.156	-6.384	-10.012	12.753	-6.649	-10.586
12.171	-6.812	-10.314	12.766	-6.854	-10.626
12.185	-7.325	-10.826	12.780	-6.443	-9.600
12.200	-6.522	-10.570	12.793	-6.609	-9.877
12.214	-6.580	-10.637	12.807	-6.532	-9.820
12.229	-6.479	-10.320	12.820	-6.678	-10.085
12.243	-6.601	-10.331	12.833	-7.382	-10.390
12.257	-7.175	-10.465	12.847	-7.053	-10.267
12.272	-6.639	-10.060	12.860	-6.671	-10.252
12.286	-6.998	-10.733	12.873	-6.844	-10.609
12.301	-6.598	-10.543	12.887	-6.593	-10.416
12.315	-6.446	-10.343	12.900	-6.673	-9.944
12.330	-6.475	-10.205	12.914	-6.667	-10.317
12.344	-6.642	-10.844	12.927	-6.246	-10.274
12.359	-6.691	-9.817	12.940	-6.635	-10.397
12.374	-6.537	-10.199	12.954	-6.179	-9.598
12.388	-7.179	-10.677	12.981	-6.638	-10.322
12.403	-6.953	-10.934	12.994	-6.944	-10.211
12.417	-6.760	-10.720	13.007	-6.645	-9.978
12.432	-7.519	-10.967	13.021	-6.766	-10.571
12.446	-6.737	-10.590	13.034	-6.627	-9.770
12.461	-7.665	-10.971	13.047	-6.469	-9.496

BD1-Age (kyr BP)	BD1- $\delta^{18}\text{O}$ (‰ VPDB)	BD1- $\delta^{13}\text{C}$ (‰ VPDB)	BD1-Age (kyr BP)	BD1- $\delta^{18}\text{O}$ (‰ VPDB)	BD1- $\delta^{13}\text{C}$ (‰ VPDB)
13.061	-6.358	-9.523	13.628	-7.008	-9.852
13.074	-6.091	-8.795	13.642	-7.077	-10.021
13.088	-6.082	-8.627	13.656	-7.018	-10.074
13.101	-6.069	-8.802	13.670	-6.689	-9.534
13.114	-6.562	-9.398	13.684	-6.783	-10.024
13.128	-6.609	-9.882	13.698	-7.479	-10.051
13.141	-6.407	-9.704	13.712	-7.197	-10.184
13.154	-6.633	-9.876	13.726	-7.107	-10.133
13.168	-6.807	-10.118	13.740	-6.962	-10.273
13.181	-6.627	-9.888	13.754	-6.503	-10.004
13.186	-6.787	-10.069	13.768	-6.510	-10.379
13.190	-7.107	-10.168	13.782	-6.035	-10.003
13.195	-6.741	-9.642	13.796	-6.241	-9.807
13.199	-7.031	-9.903	13.810	-6.611	-10.193
13.209	-6.805	-9.652	13.824	-6.591	-10.298
13.213	-6.581	-9.878	13.838	-6.620	-9.925
13.218	-6.374	-9.404	13.852	-6.860	-10.106
13.222	-6.451	-9.580	13.866	-6.818	-10.137
13.231	-6.530	-10.013	13.880	-6.763	-8.869
13.236	-6.399	-9.698	13.894	-6.680	-9.430
13.240	-6.111	-9.546	13.908	-6.269	-9.358
13.245	-6.634	-9.640	13.922	-6.617	-9.937
13.249	-6.752	-9.756	13.936	-6.404	-9.244
13.254	-7.092	-10.555	13.951	-6.448	-9.577
13.259	-7.080	-10.345	13.965	-6.486	-9.900
13.263	-7.059	-10.126	13.979	-6.580	-9.574
13.268	-6.974	-10.176	13.993	-6.592	-9.858
13.272	-6.780	-10.524	14.007	-6.627	-9.752
13.277	-7.252	-10.173	14.021	-6.564	-9.900
13.281	-7.144	-10.306	14.035	-6.612	-9.983
13.286	-7.161	-10.235	14.049	-6.622	-9.897
13.290	-7.790	-10.358	14.063	-6.576	-9.567
13.295	-7.158	-10.222	14.077	-6.505	-9.844
13.300	-7.565	-10.362	14.091	-6.576	-9.641
13.304	-7.211	-10.252	14.105	-6.153	-9.765
13.309	-6.994	-10.090	14.119	-6.194	-9.490
13.313	-6.683	-10.604	14.133	-6.425	-9.324
13.318	-7.036	-10.288	14.147	-6.714	-10.184
13.322	-7.576	-10.197	14.161	-6.124	-9.729
13.327	-7.717	-10.549	14.175	-6.184	-9.494
13.331	-7.379	-10.706	14.189	-6.467	-9.500
13.336	-7.227	-10.526	14.203	-6.597	-9.624
13.340	-7.369	-10.608	14.250	-6.590	-10.087
13.345	-7.219	-10.130	14.251	-6.699	-10.122
13.350	-7.110	-10.420	14.252	-6.081	-9.512
13.354	-7.059	-10.664	14.253	-6.238	-9.562
13.406	-6.809	-10.231	14.254	-6.559	-9.364
13.458	-7.068	-10.119	14.256	-6.527	-9.390
13.510	-6.998	-9.759	14.257	-6.290	-9.985
13.562	-6.675	-9.571	14.258	-6.204	-10.017
13.614	-7.020	-10.293	14.259	-6.279	-9.732

BD1-Age (kyr BP)	BD1- $\delta^{18}\text{O}$ (‰ VPDB)	BD1- $\delta^{13}\text{C}$ (‰ VPDB)	BD1-Age (kyr BP)	BD1- $\delta^{18}\text{O}$ (‰ VPDB)	BD1- $\delta^{13}\text{C}$ (‰ VPDB)
14.260	-6.329	-9.679	14.677	-6.204	-9.761
14.261	-5.892	-9.709	14.698	-6.212	-9.388
14.262	-5.929	-9.495	14.720	-6.509	-9.445
14.263	-5.977	-9.168	14.741	-6.597	-9.365
14.264	-5.969	-9.335	14.762	-6.151	-9.288
14.265	-6.061	-9.608	14.784	-6.157	-9.192
14.266	-7.223	-10.124	14.805	-6.024	-9.200
14.267	-6.322	-9.812	14.826	-6.073	-9.277
14.268	-6.022	-9.197	14.848	-6.280	-9.661
14.269	-6.004	-9.511	14.869	-5.954	-9.216
14.270	-6.201	-9.946	14.890	-5.921	-9.548
14.271	-6.138	-8.921	14.912	-5.884	-9.513
14.272	-6.039	-9.155	14.933	-5.891	-9.731
14.273	-6.185	-9.677	14.954	-6.017	-9.254
14.274	-6.739	-9.568	14.975	-6.290	-9.315
14.275	-6.667	-9.911	14.997	-5.997	-9.098
14.276	-6.381	-9.670	15.018	-5.916	-9.196
14.277	-5.905	-9.154	15.039	-6.124	-9.606
14.278	-6.132	-9.682	15.060	-5.802	-8.834
14.279	-6.739	-10.020	15.082	-5.750	-9.399
14.280	-6.675	-9.609	15.103	-5.887	-9.292
14.281	-6.398	-9.785	15.145	-5.626	-9.481
14.282	-6.465	-9.701	15.166	-6.032	-9.565
14.283	-6.688	-10.158	15.188	-6.140	-9.037
14.284	-6.825	-10.145	15.203	-6.119	-9.627
14.285	-6.531	-9.712	15.218	-6.407	-10.130
14.286	-6.498	-10.016	15.233	-5.995	-9.684
14.287	-6.331	-9.741	15.248	-6.323	-9.158
14.288	-6.252	-9.018	15.263	-6.593	-10.145
14.289	-6.578	-9.425	15.278	-6.365	-9.859
14.290	-6.181	-9.309	15.293	-6.260	-9.906
14.291	-7.165	-11.040	15.308	-6.428	-10.470
14.292	-5.914	-9.283	15.324	-6.516	-10.160
14.293	-6.207	-9.562	15.339	-6.045	-9.282
14.314	-6.102	-9.268	15.354	-6.324	-8.564
14.336	-6.398	-9.785	15.369	-6.137	-8.875
14.357	-6.880	-9.428	15.384	-6.108	-8.778
14.378	-6.447	-9.573	15.399	-6.026	-8.740
14.400	-6.330	-9.398	15.414	-6.208	-9.382
14.421	-6.139	-8.942	15.444	-5.734	-9.004
14.442	-6.502	-9.730	15.459	-5.879	-9.407
14.464	-5.932	-9.024	15.489	-6.315	-9.386
14.485	-6.032	-8.923	15.520	-6.335	-9.877
14.506	-6.297	-9.434	15.535	-6.274	-9.642
14.528	-5.744	-8.903	15.550	-6.265	-9.363
14.549	-5.991	-9.925	15.565	-6.565	-9.307
14.570	-6.003	-9.379	15.595	-6.592	-9.459
14.592	-5.712	-8.989	15.610	-6.251	-8.951
14.613	-5.648	-8.831	15.640	-6.261	-9.082
14.634	-5.593	-8.392	15.671	-5.752	-8.926
14.656	-6.121	-9.148	15.686	-6.147	-9.109

BD1-Age	BD1- $\delta^{18}\text{O}$	BD1- $\delta^{13}\text{C}$	BD1-Age	BD1- $\delta^{18}\text{O}$	BD1- $\delta^{13}\text{C}$
(kyr BP)	(‰ VPDB)	(‰ VPDB)	(kyr BP)	(‰ VPDB)	(‰ VPDB)
15.701	-6.422	-9.032	16.972	-6.662	-8.961
15.791	-6.586	-9.622	16.996	-6.668	-8.961
15.806	-6.224	-9.303	17.020	-6.563	-8.481
15.821	-6.324	-9.532	17.045	-6.421	-8.666
15.837	-6.404	-9.346	17.069	-6.290	-8.730
15.852	-6.387	-9.408	17.093	-6.763	-8.640
15.867	-5.415	-6.062	17.118	-6.816	-9.187
15.897	-6.421	-8.641	17.142	-6.526	-8.443
15.912	-6.496	-9.379	17.166	-6.388	-8.694
15.927	-6.187	-9.697	17.191	-6.556	-8.629
15.942	-6.480	-9.011	17.215	-6.631	-8.852
15.957	-6.272	-9.460	17.239	-6.397	-8.730
15.972	-6.793	-9.316	17.264	-6.280	-8.542
15.987	-6.419	-9.571	17.288	-6.456	-9.010
16.002	-6.099	-9.558	17.312	-6.672	-8.955
16.033	-6.632	-9.414	17.337	-6.235	-8.644
16.048	-6.395	-9.365	17.361	-6.370	-9.199
16.063	-6.015	-8.282	17.385	-6.582	-9.301
16.078	-6.357	-9.521	17.404	-6.112	-8.583
16.093	-6.462	-9.732	17.442	-6.379	-9.157
16.108	-7.056	-9.591	17.461	-6.532	-8.927
16.123	-7.086	-9.892	17.480	-6.208	-8.788
16.153	-6.833	-9.927	17.499	-6.305	-8.855
16.168	-6.943	-9.146	17.518	-6.594	-9.028
16.193	-6.695	-9.390	17.537	-6.308	-8.713
16.241	-6.477	-9.391	17.556	-6.479	-9.029
16.266	-6.295	-9.416	17.575	-6.602	-8.773
16.290	-6.481	-9.424	17.594	-6.192	-8.817
16.314	-6.677	-8.996	17.613	-6.559	-8.523
16.339	-6.504	-9.151	17.632	-6.449	-8.948
16.363	-6.987	-9.194	17.651	-6.348	-8.757
16.388	-6.644	-10.049	17.689	-6.842	-9.068
16.412	-6.426	-9.981	17.708	-6.474	-8.957
16.461	-6.568	-9.582	17.726	-6.709	-9.031
16.485	-6.154	-9.414	17.745	-6.616	-8.947
16.509	-6.558	-9.532	17.764	-6.884	-9.039
16.534	-6.664	-9.679	17.783	-6.685	-8.909
16.558	-7.091	-9.630	17.802	-6.808	-8.804
16.582	-6.944	-9.624	17.821	-6.741	-9.010
16.631	-6.911	-9.602	17.840	-6.330	-9.018
16.655	-7.004	-9.833	17.859	-6.357	-8.413
16.680	-7.410	-9.947	17.878	-7.036	-8.974
16.753	-7.051	-9.610	17.897	-6.983	-8.939
16.777	-6.605	-9.079	17.916	-7.032	-8.946
16.801	-6.525	-8.291	17.935	-7.052	-9.025
16.826	-5.983	-9.062	17.954	-6.549	-8.778
16.850	-6.392	-9.206	17.973	-6.702	-8.328
16.874	-6.322	-9.357	17.992	-7.157	-9.173
16.899	-6.462	-8.956	18.011	-6.819	-8.933
16.923	-6.564	-8.920	18.030	-6.657	-8.718
16.947	-6.390	-8.968	18.049	-6.554	-8.974

BD1-Age	BD1- $\delta^{18}\text{O}$	BD1- $\delta^{13}\text{C}$	BD1-Age	BD1- $\delta^{18}\text{O}$	BD1- $\delta^{13}\text{C}$
(kyr BP)	(‰ VPDB)	(‰ VPDB)	(kyr BP)	(‰ VPDB)	(‰ VPDB)
18.068	-6.590	-8.666	20.008	-7.174	-8.651
18.087	-6.192	-8.390	20.030	-7.055	-8.218
18.105	-6.300	-8.844	20.052	-7.275	-8.352
18.124	-6.397	-9.283	20.074	-7.074	-8.514
18.143	-6.576	-8.899	20.096	-7.018	-8.471
18.162	-6.529	-8.672	20.118	-7.006	-8.442
18.181	-6.380	-8.584	20.140	-6.746	-8.532
18.200	-6.354	-8.660	20.161	-7.154	-8.131
18.239	-6.300	-8.460	20.183	-7.275	-8.282
18.278	-6.243	-8.370	20.205	-7.461	-8.538
18.316	-6.452	-9.052	20.227	-7.301	-8.803
18.355	-6.733	-8.880	20.249	-6.651	-8.284
18.394	-6.574	-8.752	20.271	-6.672	-8.357
18.433	-6.502	-8.446	20.293	-6.597	-8.084
18.471	-6.671	-9.033	20.315	-6.706	-8.525
18.549	-6.435	-8.868	20.336	-7.030	-8.639
18.587	-6.508	-8.826	20.358	-6.802	-8.092
18.626	-6.502	-8.944	20.380	-7.060	-8.519
18.665	-6.736	-8.876	20.402	-6.557	-7.733
18.704	-7.118	-9.207	20.424	-6.787	-8.236
18.742	-6.878	-9.375	20.446	-6.920	-8.304
18.781	-6.719	-8.586	20.468	-7.292	-8.134
18.820	-6.680	-8.647	20.490	-7.512	-8.252
18.897	-6.480	-8.510	20.511	-7.257	-8.314
18.975	-6.786	-8.654	20.533	-7.162	-7.920
19.013	-6.626	-8.500	20.577	-6.628	-7.034
19.052	-6.798	-8.791	20.599	-6.977	-8.356
19.091	-6.690	-8.332	20.621	-7.576	-8.411
19.130	-6.505	-8.310	20.643	-7.148	-8.366
19.168	-6.233	-8.254	20.664	-7.218	-8.495
19.207	-6.532	-8.184	20.686	-7.062	-8.336
19.246	-6.479	-8.189	20.708	-6.905	-8.267
19.284	-6.796	-8.268	20.752	-6.960	-8.012
19.323	-6.744	-8.404	20.774	-6.926	-8.314
19.362	-6.770	-7.685	20.796	-7.035	-7.551
19.401	-6.611	-7.958	20.818	-6.980	-8.194
19.439	-7.011	-8.440	20.839	-7.428	-8.571
19.517	-6.699	-8.121	20.861	-7.113	-7.991
19.555	-6.904	-8.308	20.883	-7.073	-8.371
19.594	-6.580	-8.448	20.905	-6.911	-7.952
19.633	-6.595	-8.379	21.218	-7.499	-8.539
19.672	-6.714	-8.418	21.345	-7.111	-8.923
19.710	-6.775	-8.364	21.473	-7.310	-8.468
19.749	-6.596	-8.454	21.600	-7.030	-8.464
19.788	-6.852	-8.373	21.647	-7.054	-8.045
19.827	-6.870	-8.003	21.694	-6.691	-7.836
19.865	-6.793	-8.686	21.741	-7.395	-8.390
19.904	-6.540	-7.789	21.788	-7.443	-8.602
19.943	-6.787	-8.282	21.835	-6.806	-8.492
19.965	-6.675	-8.270	21.882	-7.345	-8.575
19.986	-7.304	-8.611	21.929	-7.137	-8.572

BD1-Age	BD1- $\delta^{18}\text{O}$	BD1- $\delta^{13}\text{C}$	BD2-Age	BD2- $\delta^{18}\text{O}$	BD2- $\delta^{13}\text{C}$
(kyr BP)	(‰ VPDB)	(‰ VPDB)	(kyr BP)	(‰ VPDB)	(‰ VPDB)
21.976	-6.751	-7.871	14.851	-6.132	-9.723
22.023	-7.294	-8.460	14.875	-6.819	-9.608
22.172	-7.228	-8.646	14.899	-6.165	-9.491
22.321	-7.196	-7.964	14.924	-5.920	-8.751
22.469	-7.492	-8.813	14.948	-6.027	-9.871
22.618	-7.152	-8.622	14.972	-5.617	-9.120
22.767	-7.600	-9.003	14.996	-6.178	-8.762
22.916	-7.156	-8.902	15.020	-6.455	-8.746
23.064	-6.243	-8.946	15.044	-6.294	-8.891
23.213	-7.546	-9.117	15.068	-6.278	-8.499
23.344	-7.207	-8.832	15.092	-5.856	-8.994
23.476	-7.332	-8.888	15.116	-6.295	-9.362
23.607	-7.060	-8.674	15.141	-6.670	-9.362
23.739	-7.723	-9.229	15.165	-6.572	-9.270
23.870	-6.648	-8.474	15.189	-6.061	-9.096
24.002	-7.011	-8.965	15.213	-5.901	-8.993
24.133	-7.102	-8.989	15.237	-6.399	-9.137
24.264	-7.145	-8.961	15.261	-6.423	-8.522
24.396	-7.186	-8.494	15.285	-5.612	-8.469
24.527	-6.670	-7.765	15.309	-6.527	-10.186
24.659	-6.808	-9.295	15.334	-6.324	-9.735
24.790	-6.632	-9.036	15.358	-6.042	-9.572
24.921	-6.628	-9.495	15.382	-6.359	-9.568
25.053	-6.915	-9.276	15.406	-5.978	-9.267
25.184	-6.838	-9.535	15.430	-6.047	-9.157
25.316	-6.614	-9.274	15.454	-6.364	-9.270
25.447	-6.518	-9.456	15.478	-6.267	-9.080
25.579	-6.703	-9.169	15.526	-7.414	-9.657
25.710	-7.075	-9.726	15.551	-6.101	-9.480
25.841	-7.238	-10.235	15.575	-6.413	-9.523
25.973	-6.561	-9.332	15.599	-6.323	-9.453
			15.623	-6.749	-10.376
			15.647	-6.702	-9.573
			15.671	-6.937	-9.083
			15.695	-6.172	-9.454
			15.719	-6.898	-10.200
			15.743	-6.453	-9.016
			15.768	-6.090	-9.545
			15.792	-6.308	-9.834
			15.816	-6.220	-8.926
			15.840	-6.222	-9.152
			15.864	-6.162	-9.410
			15.888	-6.320	-9.425
			15.912	-6.426	-9.078
			15.936	-6.528	-9.189
			15.961	-6.508	-9.139
			15.985	-6.526	-9.388
			16.009	-6.968	-9.713
			16.033	-7.029	-9.247
			16.057	-6.492	-9.208
			16.081	-6.218	-8.345

BD2-Age (kyr BP)	BD2- $\delta^{18}\text{O}$ (‰ VPDB)	BD2- $\delta^{13}\text{C}$ (‰ VPDB)	BD2-Age (kyr BP)	BD2- $\delta^{18}\text{O}$ (‰ VPDB)	BD2- $\delta^{13}\text{C}$ (‰ VPDB)
16.105	-6.744	-9.658	18.352	-6.161	-8.746
16.129	-7.124	-9.428	18.397	-6.483	-8.955
16.153	-6.601	-9.376	18.442	-6.710	-8.980
16.198	-6.597	-9.473	18.460	-6.788	-8.888
16.243	-6.783	-9.543	18.478	-6.874	-9.042
16.288	-6.367	-9.218	18.496	-6.773	-9.006
16.333	-6.596	-9.120	18.513	-6.707	-8.775
16.378	-6.985	-9.305	18.531	-6.639	-8.880
16.423	-6.736	-9.912	18.549	-7.153	-9.157
16.468	-6.481	-9.778	18.567	-7.167	-9.334
16.512	-6.646	-9.416	18.585	-6.573	-8.543
16.557	-6.811	-10.033	18.602	-7.023	-8.901
16.602	-6.787	-9.414	18.620	-7.108	-8.989
16.647	-6.838	-9.594	18.638	-6.779	-8.881
16.692	-6.965	-9.529	18.656	-6.867	-8.856
16.737	-7.169	-9.081	18.674	-7.085	-8.798
16.782	-6.909	-8.659	18.692	-6.774	-8.494
16.827	-6.575	-8.975	18.709	-6.768	-8.194
16.871	-6.310	-8.996	18.727	-6.681	-8.449
16.916	-6.247	-9.172	18.745	-6.682	-8.534
16.961	-6.485	-8.712	18.763	-6.831	-8.419
17.006	-6.765	-8.242	18.781	-6.891	-8.577
17.051	-6.559	-8.712	18.799	-6.582	-8.651
17.096	-6.771	-8.704	18.816	-6.782	-8.714
17.141	-6.471	-8.866	18.834	-6.723	-8.292
17.186	-6.482	-9.004	18.852	-7.248	-8.371
17.230	-6.644	-8.756	18.870	-6.875	-8.686
17.275	-6.613	-8.736	18.888	-7.080	-8.611
17.320	-6.862	-8.931	18.906	-6.726	-8.705
17.365	-6.922	-8.991	18.923	-6.691	-8.820
17.410	-7.004	-9.208	18.941	-6.679	-8.612
17.455	-6.922	-8.822	18.959	-6.718	-8.576
17.500	-6.670	-9.090	18.977	-6.572	-8.305
17.545	-6.788	-8.951	18.995	-7.080	-8.071
17.589	-6.354	-8.519	19.048	-6.981	-8.572
17.634	-6.597	-8.428	19.101	-7.125	-8.845
17.679	-6.847	-8.415	19.153	-6.604	-8.468
17.724	-6.851	-8.872	19.206	-6.661	-8.252
17.769	-7.107	-8.940	19.259	-7.331	-8.437
17.814	-6.836	-8.856	19.312	-6.942	-8.489
17.859	-6.482	-8.716	19.365	-7.009	-8.350
17.904	-6.679	-8.721	19.418	-6.712	-8.462
17.948	-6.925	-8.513	19.471	-6.665	-8.170
17.993	-6.696	-8.352	19.524	-6.725	-8.653
18.038	-6.825	-8.582	19.577	-7.424	-8.473
18.083	-6.407	-8.517	19.630	-6.553	-8.064
18.128	-6.362	-8.743	19.683	-6.396	-7.906
18.173	-6.947	-8.748	19.735	-6.652	-8.143
18.218	-6.827	-8.687	19.788	-6.713	-8.600
18.263	-6.380	-8.764	19.841	-7.273	-8.340
18.307	-6.711	-8.502	19.894	-7.139	-8.510

BD2-Age (kyr BP)	BD2- $\delta^{18}\text{O}$ (‰ VPDB)	BD2- $\delta^{13}\text{C}$ (‰ VPDB)	BD2-Age (kyr BP)	BD2- $\delta^{18}\text{O}$ (‰ VPDB)	BD2- $\delta^{13}\text{C}$ (‰ VPDB)
19.947	-6.911	-7.883	21.781	-6.777	-8.264
20.000	-6.819	-8.414	21.798	-6.888	-8.105
20.053	-6.859	-8.453	21.815	-6.941	-8.405
20.106	-6.597	-8.197	21.832	-6.950	-8.525
20.159	-6.787	-8.448	21.849	-6.990	-8.474
20.212	-6.862	-8.355	21.866	-7.446	-8.553
20.265	-6.779	-8.418	21.883	-7.019	-8.695
20.317	-6.739	-7.944	21.899	-7.107	-8.814
20.370	-6.475	-7.612	21.946	-6.928	-8.460
20.423	-6.715	-8.136	21.993	-7.428	-8.802
20.476	-7.412	-8.101	22.040	-7.088	-8.450
20.529	-7.361	-8.222	22.086	-7.145	-8.669
20.582	-6.886	-7.788	22.133	-7.041	-8.155
20.635	-6.861	-8.094	22.180	-6.921	-8.525
20.688	-7.012	-8.031	22.226	-7.007	-8.615
20.741	-6.842	-8.079	22.273	-7.129	-8.359
20.794	-7.017	-7.983	22.320	-7.190	-8.519
20.847	-6.944	-8.414	22.367	-7.183	-8.450
20.899	-6.997	-8.097	22.413	-7.338	-8.571
20.952	-7.102	-8.212	22.460	-7.251	-8.408
21.005	-6.863	-8.190	22.507	-7.348	-8.781
21.058	-7.172	-8.441	22.554	-7.047	-8.732
21.111	-7.229	-8.328	22.600	-7.106	-8.765
21.164	-7.473	-8.319	22.647	-7.148	-8.575
21.217	-7.506	-8.480	22.694	-7.275	-8.700
21.270	-7.316	-8.575	22.740	-6.834	-7.710
21.323	-7.438	-8.598	22.787	-7.254	-8.465
21.376	-7.526	-8.646	22.834	-7.692	-8.433
21.393	-7.107	-8.193	22.881	-7.221	-8.343
21.409	-7.057	-8.501	22.927	-7.458	-8.531
21.426	-7.117	-8.573	22.974	-7.402	-8.507
21.443	-7.081	-8.482	23.021	-7.014	-8.482
21.460	-6.789	-7.978	23.067	-7.175	-8.356
21.477	-6.868	-8.044	23.114	-7.363	-8.560
21.494	-7.544	-8.162	23.161	-7.175	-8.785
21.511	-6.874	-8.259	23.208	-7.451	-8.655
21.528	-6.740	-8.330	23.254	-7.307	-8.561
21.545	-7.157	-8.524	23.301	-7.294	-8.469
21.562	-6.983	-8.720	23.348	-7.325	-8.828
21.578	-7.135	-8.390	23.394	-7.443	-8.720
21.595	-7.466	-8.302	23.441	-7.363	-8.970
21.612	-7.571	-8.300	23.488	-7.025	-8.288
21.629	-6.694	-7.746	23.518	-7.071	-9.070
21.646	-7.273	-8.124	23.548	-7.541	-8.195
21.663	-6.995	-8.232	23.578	-7.159	-8.616
21.680	-7.356	-8.415	23.608	-7.529	-8.928
21.697	-7.206	-8.036	23.638	-7.304	-8.131
21.714	-7.225	-8.250	23.668	-7.244	-8.458
21.730	-7.065	-8.391	23.698	-7.220	-8.347
21.747	-7.218	-8.548	23.729	-6.929	-8.392
21.764	-6.897	-8.424	23.759	-6.934	-8.285

BD2-Age	BD2- $\delta^{18}\text{O}$	BD2- $\delta^{13}\text{C}$	BD2-Age	BD2- $\delta^{18}\text{O}$	BD2- $\delta^{13}\text{C}$
(kyr BP)	(‰ VPDB)	(‰ VPDB)	(kyr BP)	(‰ VPDB)	(‰ VPDB)
23.789	-7.081	-8.879	25.323	-6.688	-7.749
23.819	-7.035	-8.573	25.354	-6.956	-8.161
23.849	-7.344	-8.830	25.385	-7.274	-8.550
23.879	-7.031	-8.735	25.416	-7.057	-8.313
23.909	-7.047	-8.814	25.447	-7.177	-8.219
23.939	-6.965	-8.298	25.477	-7.135	-8.424
23.969	-7.252	-8.691	25.508	-7.182	-8.461
23.999	-7.430	-9.250	25.539	-7.211	-8.194
24.029	-7.079	-9.084	25.570	-7.246	-8.369
24.059	-7.302	-8.605	25.601	-7.144	-7.598
24.090	-7.271	-8.651	25.632	-7.169	-8.452
24.120	-7.261	-8.659	25.663	-6.984	-8.537
24.150	-6.903	-8.547	25.694	-6.804	-8.134
24.180	-7.028	-8.761	25.725	-6.826	-8.142
24.210	-6.765	-8.528	25.756	-6.928	-8.214
24.240	-7.143	-8.494	25.787	-7.090	-7.907
24.270	-6.928	-8.957	25.817	-7.079	-8.066
24.300	-6.595	-8.303	25.848	-6.894	-8.058
24.330	-6.934	-8.985	25.879	-6.802	-8.460
24.360	-7.184	-9.078	25.910	-6.757	-7.657
24.390	-6.812	-8.677	25.941	-7.045	-7.639
24.420	-6.844	-8.899	25.972	-6.637	-8.038
24.451	-6.923	-8.683	26.003	-6.892	-7.049
24.481	-7.091	-8.967	26.034	-6.521	-8.091
24.511	-6.574	-8.582	26.065	-7.668	-8.551
24.541	-6.967	-9.100	26.096	-7.005	-8.344
24.571	-6.751	-8.818	26.127	-7.257	-8.276
24.601	-6.902	-8.588	26.157	-7.378	-8.629
24.631	-6.888	-8.469	26.188	-7.207	-8.422
24.661	-6.966	-8.543	26.219	-6.963	-8.063
24.691	-7.056	-8.800	26.250	-7.405	-9.047
24.721	-7.072	-8.553	26.281	-7.050	-8.375
24.751	-7.107	-8.841	26.312	-7.051	-8.595
24.781	-7.069	-8.671	26.343	-7.046	-8.474
24.812	-7.153	-8.800	26.374	-7.118	-8.549
24.842	-7.079	-8.959	26.405	-7.407	-8.029
24.872	-7.127	-9.046	26.436	-7.086	-8.448
24.902	-7.097	-8.880	26.467	-7.462	-8.646
24.932	-7.080	-8.662	26.497	-7.194	-8.734
24.962	-7.070	-8.650	26.528	-6.832	-8.483
24.992	-6.909	-8.299	26.559	-7.100	-8.366
25.022	-6.849	-8.099	26.590	-7.491	-8.431
25.052	-6.836	-8.620	26.621	-7.471	-8.849
25.082	-6.726	-7.981	26.652	-7.236	-8.790
25.112	-6.851	-8.101	26.686	-6.791	-8.131
25.142	-7.274	-8.516	26.721	-7.461	-8.307
25.173	-7.182	-8.645	26.755	-6.700	-8.284
25.203	-6.995	-8.355	26.790	-6.578	-8.357
25.233	-7.279	-8.360	26.824	-6.739	-8.197
25.263	-7.205	-8.441	26.858	-6.063	-7.956
25.293	-6.757	-8.384	26.893	-6.796	-8.538

BD2-Age (kyr BP)	BD2- $\delta^{18}\text{O}$ (‰ VPDB)	BD2- $\delta^{13}\text{C}$ (‰ VPDB)	BD2-Age (kyr BP)	BD2- $\delta^{18}\text{O}$ (‰ VPDB)	BD2- $\delta^{13}\text{C}$ (‰ VPDB)
26.927	-7.205	-8.452	28.591	-6.550	-8.271
26.962	-6.724	-8.254	28.617	-7.046	-8.600
26.996	-6.995	-8.189	28.643	-6.923	-8.261
27.030	-7.010	-8.502	28.669	-6.735	-8.075
27.065	-6.577	-7.896	28.695	-7.061	-8.617
27.099	-6.914	-7.971	28.721	-7.006	-8.759
27.134	-6.922	-8.509	28.748	-6.472	-8.113
27.168	-7.027	-8.129	28.774	-6.659	-8.127
27.202	-7.354	-8.816	28.800	-7.469	-8.746
27.237	-7.400	-8.282	28.826	-7.096	-8.505
27.271	-7.632	-8.341	28.852	-7.451	-9.173
27.306	-6.983	-8.958	28.878	-7.324	-8.730
27.340	-6.903	-8.481	28.904	-7.028	-8.016
27.375	-6.891	-8.664	28.930	-7.590	-8.924
27.409	-7.124	-8.257	28.957	-6.909	-8.539
27.443	-7.517	-8.488	28.983	-6.396	-8.370
27.478	-7.376	-8.707	29.009	-6.465	-8.218
27.512	-7.032	-8.180	29.035	-6.658	-8.810
27.547	-7.155	-8.460	29.061	-7.180	-8.960
27.581	-7.588	-8.254	29.087	-6.399	-8.050
27.615	-7.030	-8.302	29.113	-6.639	-8.242
27.650	-7.563	-8.876	29.139	-7.007	-8.367
27.684	-7.443	-9.095	29.166	-6.597	-7.828
27.719	-7.121	-8.683	29.192	-7.215	-8.350
27.753	-7.603	-8.742	29.218	-6.911	-8.524
27.787	-7.150	-9.213	29.244	-6.857	-8.444
27.822	-7.359	-8.331	29.270	-6.775	-8.656
27.856	-7.105	-8.430	29.296	-6.516	-8.172
27.891	-7.009	-8.724	29.322	-6.846	-8.698
27.925	-6.835	-8.245	29.348	-6.610	-8.623
27.959	-7.361	-8.586	29.375	-6.637	-8.133
27.994	-6.570	-8.612	29.401	-6.809	-8.257
28.028	-6.573	-8.554	29.427	-6.769	-8.342
28.063	-6.529	-8.279	29.453	-6.539	-8.593
28.097	-6.545	-8.345	29.479	-7.224	-8.253
28.131	-6.601	-8.698	29.505	-7.054	-8.347
28.166	-6.607	-8.791	29.531	-7.250	-8.724
28.200	-6.591	-8.546	29.557	-7.246	-9.250
28.235	-6.787	-9.007	29.584	-6.825	-8.617
28.269	-6.549	-8.697	29.610	-6.926	-8.764
28.303	-7.294	-8.982	29.636	-7.069	-8.379
28.330	-7.334	-8.950	29.662	-6.792	-8.598
28.356	-6.731	-8.287	29.688	-6.634	-8.451
28.382	-6.856	-8.437	29.714	-6.465	-8.185
28.408	-7.363	-8.892	29.740	-6.568	-8.162
28.434	-6.648	-9.043	29.766	-6.744	-8.105
28.460	-6.870	-9.074	29.793	-6.844	-8.082
28.486	-6.790	-8.199	29.819	-6.814	-8.511
28.512	-7.054	-8.498			
28.539	-7.278	-8.405			
28.565	-6.995	-8.246			

Appendix D: Reconstruct the temperature during the Last Glacial Maximum with speleothems from Southeast Asia

1. Relevant prior work and scientific questions

Temperature records from the tropics

The current estimate of the climate sensitivity (1.5-4.5°C) has dauntingly large uncertainty, which poses a serious challenge in constraining future projections of climate change. To help improve the situation, reconstruction on the temperature of relatively stable period in the past is needed¹. The Last Glacial Maximum (LGM, ~19,000 to 23,000 years ago), for instance, is such a period with the atmospheric CO₂ concentration ($p\text{CO}_2$) as low as ~190 parts per million (ppm). The knowledge of how the land and ocean temperature change in response to this amount of decrease in $p\text{CO}_2$ and corresponding radiate forcing, is yet controversial. The existing temperature reconstruction during the LGM is largely biased towards the marine records, especially for the tropical region. For instance, across the Southeast Asia, the majority of temperature reconstructions are sea surface temperature (SST) estimates based on the abundances of microfossil species or geochemical paleothermometers (Mg/Ca or Sr/Ca ratios in carbonate shells or corals, alkenone unsaturation ratio U_{37}^{K} and TEX_{86}). These records together suggest regional ocean cooling during the LGM from as low as 1°C to as high as 5°C, showing large discrepancies¹⁻⁶ (Fig. 1). Existing terrestrial temperature records from the region, mostly based on the elevation change of snowline and pollen assemblage change, are limited in spatial coverage and relatively poor in chronology. How much does tropical temperature change in response to ~100 ppm decrease in $p\text{CO}_2$ during the LGM, is still in debate. Such situation needs to be improved by efforts in more precise temperature reconstruction from different locations in the region, particularly over the land side.

Speleothem as a temperature archive

Among existing terrestrial archives, speleothem, stalagmites in particular, are shown to be quite sensitive to climate variability and can be well constrained in chronology by high precision uranium-thorium dating. Its stable isotopes (oxygen and carbon isotopes) in the calcite are commonly used as proxies of

hydroclimate and ecosystem change in the past. During the stalagmite formation, ancient drip water can also be trapped in the calcite, the oxygen isotope of which ($\delta^{18}\text{O}_w$), when combined with the calcite oxygen isotope ($\delta^{18}\text{O}_{cc}$), can be used to estimate the temperature of calcite formation. Previously several water inclusion studies on speleothem have been done in the Southeast Asia. Two are from Flores, which suggests 5°C ($\pm 1.5^\circ\text{C}$) cooling during Younger Dryas (YD, ~11,700 to 12,900 years ago) ⁷ along with $2\text{--}3^\circ\text{C}$ ($\pm 1.4^\circ\text{C}$) warming during Dansgaard–Oeschger event 21 (~87,000–84,000 years ago) ⁸. The other one from northern Borneo shows glacial-interglacial temperature shift of $4\text{--}5^\circ\text{C}$ ($\pm 0.9\text{--}2.0^\circ\text{C}$) for Marine Isotope Stage (MIS) 12-9. Yet no LGM temperature has been reported by using tropical speleothems.

Fluid inclusion techniques

Earlier speleothem water inclusion analysis typically requires relatively large volume of inclusion water (from around half to more than one microliter) from the calcite and requires the conversion of water molecular to H_2 and CO_2 for isotope ratio mass spectrometric (IRMS) measurement ^{9,10}. A recently developed technique using cavity ring-down spectroscopy (CRDS) avoids the molecular conversion step and enables the direct measurements of water $\delta^{18}\text{O}$ and δD by detecting the isotopologues of water molecules with an infrared laser ^{11,12}. With a newly developed crushing devices, a refined sample extraction line and careful evaluation of water/carbonate adsorption effect, Uemura et al. managed to precisely and accurately measure relatively small volume (less than $0.3\ \mu\text{L}$) of inclusion water from relatively small speleothem sample size by CRDS ¹².

Speleothem from Southeast Asia

We have collected speleothem samples from caves in central Myanmar, central Vietnam and Sulawesi, Indonesia. Some of these samples were found to be deposited during the LGM and the recent millennial by preliminary uranium-thorium dating (Figure 2). These speleothems, if containing enough amount of well-preserved inclusion water, can be used to relatively precisely determine the glacial-interglacial temperature change over a wide spatial coverage of Southeast Asia by analyzing its inclusion water $\delta^{18}\text{O}$ with improved isotopic technique based on CRDS and coeval calcite $\delta^{18}\text{O}$. The derived temperature

estimates can be further used to evaluate their influences on regional hydroclimate based on calcite $\delta^{18}\text{O}$.

2. Working plan

In November 2017, we have done some preliminary water inclusion $\delta^{18}\text{O}$ analysis on calcite speleothem from southwestern Sulawesi, Indonesia. Based on the $\delta^{18}\text{O}$ of inclusion water and coeval calcite of speleothem grown during the LGM and the recent millennial (Figure 3), we have calculated $3.9 \pm 0.6^\circ\text{C}$ cooling during the LGM for this location. This number shows the potential of obtaining more precise LGM temperature estimates with speleothem from Southeast Asia. We propose, in the next two fiscal years (FY18 and FY19), to measure the $\delta^{18}\text{O}$ of inclusion water in calcite speleothem from Southeast Asia, to quantify the regional temperature change between the LGM and the recent millennial on a regional scale by coupling the $\delta^{18}\text{O}$ of inclusion water and coeval calcite with stalagmites from different parts of the region, and to study the potential influences of temperature change on hydroclimate variability.

1). U/Th dating and isotopic analysis on carbonates

In FY2018, we will further constrain the sample ages of selected speleothems using the high-precision U/Th disequilibrium-dating methods developed in the lab, with the separation and purification of actinide elements done in a wet chemistry clean room and their isotopes subsequently measured on an MC-ICP-MS (a Neptune Plus). To know about the regional hydroclimate change, we will also measure the calcite $\delta^{18}\text{O}$ values on a stable isotope ratio mass spectrometer (a Delta V equipped with a Gas Bench II system).

2). Water inclusion analysis

For water inclusion study, we will first carefully study the petrography of the stalagmites to assess the nature of fluid inclusions. Later on, we will estimate the water content to decide the amount of stalagmite subsample necessary for the analysis. We will then start the isotopic analysis of inclusion water using the newly developed inclusion water extract line and cavity ring-down spectroscopic techniques by collaborating with Dr. Ryu Uemura from University of the Ryukyus. Each sequence of inclusion water measurements is comprised of one stalagmite sample, at least one working water reference

before and three references after the sample for calibration use, costing 4-7 hours in total. After the water inclusion analysis, we will collect the crushed sample and analyze its $\delta^{18}\text{O}_{\text{cc}}$. In the preliminary analysis, we found that the temperature calculated based on inclusion water and coeval calcite $\delta^{18}\text{O}$ is very sensitive to even small changes in calcite $\delta^{18}\text{O}$, hence it is critical to have the crushed calcite samples thoroughly homogenized and later on at least triple-analyzed for stable isotopes.

Once we obtain the isotope ratios of both inclusion water and its coeval enclosing calcite, we will calculate the carbonate formation temperature based on the fractionation factor of oxygen isotopes between the two phases, provided that the calcite is precipitated from drip water under equilibrium conditions.

Note that, although the LGM temperature reconstruction will be our main target, we include the samples growing during the past millennium in this study. We will compare the temperature obtained from the most recent samples with modern instrumental observations, hence to evaluate the robustness and potential limitations of the isotopic thermometer.

3). FY2019: Continuation of analysis, data interpretation and paper writing

We will continue the geochemical analysis on U/Th dating, $\delta^{18}\text{O}$ of speleothem calcite and inclusion water in FY2019. In the meantime, we will evaluate the obtained data sets including their reproducibility and quality. We will also conduct data interpretations accordingly. In the end, we will reconstruct the LGM temperature from three locations across the Southeast Asia, and evaluate the influences of temperature changes on hydroclimate variability. Outcomes of this project will be submitted to international scientific journals in FY2019.

3. Broader impact

This study will be the first to quantify the tropical temperature change between the LGM and the recent millennial on a regional scale by using stalagmites from Southeast Asia. The proposed project can provide much needed knowledge on tropical temperature change between the LGM and present day, hence help constrain the equilibrium climate sensitivity. The reconstructed temperature is also valuable for testing the performance of climate models and

improving the projections for future climate change. As a research project, it involves training of postdocs, graduate and undergraduate students.

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Figure 1. LGM temperature reconstruction in Southeast Asia. The left panel shows the distribution of temperature estimates from marine and terrestrial parts of the region, the locations of which corresponding to the magenta diamonds and blue circles in the map to the right. The proposed cave sites are also shown in red stars. The data is from references ^{4,13-30}.

Figure 2. Number of stalagmites grown in the LGM and the preindustrial period (PI) from SW Sulawesi, central Vietnam and central Myanmar. Also shown are the scanned images of stalagmites with ages of interest. ‘ka’ stands for thousand years before present.

Figure 3. Preliminary data on water inclusion analysis of LGM and pre-industrial (PI) stalagmites from southwestern Sulawesi, Indonesia. The temperature is calculated by equation from Coplen et al., 2007³¹.