Tracking hydro-climate changes in Mainland Southeast Asia through the past 180,000 years

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NANYANG
TECHNOLOGICAL
UNIVERSITY

**TRACKING HYDRO-CLIMATE
CHANGES IN MAINLAND
SOUTHEAST ASIA THROUGH
THE PAST 180,000 YEARS**

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**INTERDISCIPLINARY GRADUATE
SCHOOL
2017**

**Tracking Hydro-Climate Changes in
Mainland Southeast Asia through the
past 180,000 years**

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Table of Contents

Acknowledgements	1
List of Tables	7
List of Figures.....	9
Summary.....	19
Chapter 1	21
Introduction.....	21
1.1. Regional hydroclimate	23
1.2. Regional paleo-hydroclimate	28
1.3. Speleothems as climate archives.....	33
1.3.1 U-Th Dating	33
1.3.2. Speleothem proxies	34
1.4. Field trips and speleothem samples	42
1.5. Thesis structure	47
1.7. References	52
Chapter 2	75
A 450-year-long Indian Summer Monsoon record from speleothem trace elements in Myanmar	75
Contributions	76
2.1. Abstract	77
2.2. Introduction	78
2.3. Cave setting and sample description.....	82

2.4. Analytical Methods.....	85
2.4.1. ²³⁰ Th dating.....	85
2.4.2. Trace element analysis	86
2.5. Results.....	88
2.5.1. ²³⁰ Th ages	88
2.5.2. Layer counting	89
2.5.3. Trace element analysis	92
2.5.4. Trace element oscillations	100
2.6. Discussion	103
2.6.1. Variations of alkaline earth elements	103
2.6.2. Variations of U in the sample.....	107
2.6.3. Low-frequency (interannual to decadal) climate variability	108
2.6.4. Driving mechanism of the interannual to decadal variabilities of ISM ...	110
2.7. Conclusions	114
2.8. Acknowledgements	115
2.9. References	116
Chapter 3	128
Indian Summer Monsoon during the Last Glacial Maximum	128
Contributions	129
3.1. Abstract	130
3.2. Introduction	131
3.3. Cave setting and Materials	138
3.4. Analytical Methods.....	139
3.5. Results.....	141

3.5.1. ^{230}Th ages	141
3.5.2. Isotopic equilibrium tests	144
3.6. Discussion	147
3.6.1. The climatic interpretation of speleothem $\delta^{18}\text{O}$	147
3.6.2. ISM variation	150
3.6.3. Driving mechanism of the ISM.....	153
3.6.4. Spatial response of speleothem $\delta^{18}\text{O}$ to glacial forcings.....	157
3.7. Conclusions	162
3.8. Acknowledgements	171
3.9. References	172
Chapter 4	183
On the glacial-interglacial variability of speleothem-inferred Asian Monsoon over the past 180 ka	183
Contributions	184
4.1. Abstract	185
4.2. Introduction	187
4.3. Cave setting	192
4.4. Materials and Analytical Methods.....	197
4.5. Results.....	199
4.6. Isotopic equilibrium tests.....	204
4.7. Discussion	209
4.7.1. Interpretation of speleothem $\delta^{18}\text{O}$ proxy	209
4.7.2. Spatial-temporal pattern of the ISM.....	215
4.7.3. Marine Isotope Stage 5.....	225

4.8. Conclusions	229
4.9. Acknowledgements	269
4.10 References	270
Chapter 5	285
Conclusions	285
5.1. Summary of key findings	285
5.2. Suggestions for future work	288
5.3. References	292
Complete List of References	296

List of Tables

Table 1. 1. Summary of terrestrial proxy-based paleo-hydroclimate records from MSEA.....	32
Table 1. 2. Summary of fieldtrips.	45
Table 1. 3. Speleothem samples from Mainland Southeast Asia.....	49
Table 2. 1. ^{230}Th dating results, error in 2σ (95% in confidence).....	91
Table 2. 2. Descriptive statistics for the LNT-03 LA-ICP-MS dataset. Mg/Ca are expressed as mmol/mol and the rest are $\mu\text{mol/mol}$	96
Table 2. 3. Correlation coefficients (r) for the low frequency record (thick red line in Fig. 2.4) and the detrended high frequency record (Fig. 2.5) (all p-values $<<0.05$).	96
Table 3. 1. ^{230}Th dating results, error in 2σ (95% in confidence).....	163
Table 3. 2. Stable isotope data for stalagmite samples from SMWT caves.....	165
Table 4. 1. ^{230}Th dating results for samples from LNT Cave, with errors in 2σ (95% in confidence).....	230
Table 4. 2. ^{230}Th dating results for samples from SE Yunnan Caves, with errors in 2σ (95% in confidence).....	234
Table 4. 3. Stable isotope composition of stalagmite samples from LNT Cave.	240

Table 4. 4. Stable isotope data for stalagmite sample from SE Yunnan Caves.

.....249

List of Figures

- Figure 1. 1. Left: the Asian Monsoon system (modified after Johnson. (2011)).
The black rectangle crops out MSEA. Right: Location of published paleo-
hydroclimate records in MSEA. See Table 1.1 for details.27
- Figure 1. 2. Projections on tropical land precipitation changes in two advanced
climate models show large discrepancies (Chadwick et al. (2015)).
Precipitation change for 2071–2100 minus 1971–2000, under the RCP8.5
emissions scenario for GFDL-ESM2M (a) and HadGEM2-ES (b).31
- Figure 1. 3. Schematic diagram of a cave system and carbonate dissolution and
deposition (Fairchild and Baker, 2012).40
- Figure 1. 4. The Chinese speleothem $\delta^{18}\text{O}$ records show different periodicities
compared with other major paleoclimate records. The compiled
speleothem $\delta^{18}\text{O}$ record from southern China (green) (Cheng et al., 2016)
follows the 65°N summer insolation (grey) tightly and is dominated by the
23 kyr precessional cycle. However, other major paleoclimate records,
such as atmospheric CO_2 concentration (red) (Lüthi et al., 2008) and
Antarctica surface temperature variation (Jouzel et al., 2007) retrieved
from ice core EPICA Dome C and benthic foraminifera $\delta^{18}\text{O}$ LR04 stack
(purple) (Lisiecki and Raymo, 2005) are dominated by the 100 kyr glacial-
interglacial cycle.41
- Figure 1. 5. Sampling strategy, with circled numbers indicating our sampling
locations. The lower right inset shows the AM system (modified after

Johnson (2011)). The black arrows represent the present-day averaged June-July-August (JJA) wind pattern. The data are from the National Centres for Environmental Prediction (NCEP) / National Centre for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010) (Kalnay et al., 1996). The numbered circles indicate the three sampling locations in Myanmar and China (see section 1.4).46

Figure 2. 1. The upper left map shows the Asian monsoon system (modified after Johnson, (2011)). The lower left plot shows monthly mean precipitation (blue bars) and temperature (red squares) at Taunggyi meteorological observatory (data from UNdata: <http://data.un.org/Explorer.aspx?d=CLINO>). Over 90% of the annual precipitation falls during the monsoon season (late April through early November). Right map shows the location of Lin Noe Twin Cave (LNT, red star) and Taunggyi City (red dot).81

Figure 2. 2. a). Polished fresh surface of speleothem LNT-03 shows clear annual banding. Top and bottom ages are marked. b). A close-up image of the annual banding with visible transparent and milky-white calcite couplets.84

Figure 2. 3. StalAge model (Scholz and Hoffmann, 2011) for LNT-03 suggests that the stalagmite grew steadily and continuously for about 450 years from ~450 AD to ~900 AD.....90

Figure 2. 4. Changes in Mg/Ca, Sr/Ca, Ba/Ca, and U/Ca along the entire 15 cm of the sample length. The left side represents the youngest end of the sample. The gray lines represent the original data after outliers were eliminated. The black and thick red lines are 10-point and 500-point running means of the data, respectively.....	95
Figure 2. 5. Comparisons of changes in Mg/Ca, Sr/Ca, Ba/Ca, and U/Ca, zoomed in on 10 mm of the sample length (here between 120 and 130 mm) for better visibility. All four elements show clear annual cyclicity, with roughly ~30 cycles. Only every fifth cycle in the Mg/Ca profile is marked. Mg is inversely correlated with Sr, Ba, and U. Sr and Ba show a strong positive correlation.....	97
Figure 2. 6. Annual means of Mg, Sr, Ba, and U, and annual growth rates of LNT-03 calculated from Mg peaks. From ~750 AD to ~900 AD, Mg fluctuations indicate frequent droughts in Myanmar.....	102
Figure 2. 7. The low R^2 value indicates that annual Sr/Ca mean does not correlate with annual growth rate of LNT-03.....	106
Figure 2. 8. Spectral analysis results of the Mg/Ca, Sr/Ca, and Ba/Ca time series. The spectrums were obtained using the program REDFIT (Schulz and Mudelsee, 2002).....	113
Figure 3. 1. a). Locations of Phabaung Gu Cave (PHA), Padamya Gu Cave (PDM) and Hin Tum No.2 Cave (HT2) (red stars). The red dot represents the location of the closest major city, Yangon. Locations of Mawmluh	

cave (orange star) (Berkelhammer et al., 2013; Dutt et al., 2015) and its closest major city, Shillong (orange dot) in Northeast India, and core KL126 (blue star) (Kudrass et al., 2001) in the BoB are also shown for comparison. The black arrows represent the present-day averaged June-July-August (JJA) wind pattern (data from the National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010) (Kalnay et al., 1996). b). Monthly mean precipitation (blue bars) and temperature (red squares) at Taunggyi and Shillong meteorological observatory stations. As shown, the majority of the annual precipitation in the region occurs during the monsoon season (data from UNdata: <http://data.un.org/Explorer.aspx?d=CLINO>). 137

Figure 3. 2. Plots of age versus depth for stalagmites HT2-1, HT2-2, PDM02, and PHA-02 derived from SMWT caves. All ages are reported in units of thousands of years before present (1950 AD) or ka BP. The age errors indicated in the plots are 2σ errors. The StalAge method (Scholz and Hoffmann, 2011) was used to establish the chronology for HT2-1, PDM-02, and PHA-02. The red lines indicate the 95% confidence limits calculated by a Monte Carlo simulation fitting. Weighted mean age and a linear extrapolation were used to set up the age model for HT2-2 (see the Results section for details). 142

Figure 3. 3. Scatter plot of $\delta^{18}\text{O}$ versus $\delta^{13}\text{C}$ for each stalagmite sample from the SMWT caves (see Fig. 3.1 for the location of the caves). The low

correlations (as indicated by R^2 values) of the plotted results indicate that carbon and oxygen isotopic ratios are not highly correlated as would be expected under conditions of kinetic fractionation.....	143
Figure 3. 4. Replications of stable isotope variation on sample PDM02. The low and high resolution measurements were performed at different locations of the growth layers of PDM02 in different laboratories. The low-resolution data were achieved in the EGL-EOS, while the high-resolution data were achieved in the Speleothem Isotope Laboratory, Nanjing Normal University.....	146
Figure 3. 5. The $\delta^{18}\text{O}$ time series and ^{230}Th ages of stalagmite samples from SMWT caves over the last 30 ka. The $\delta^{18}\text{O}$ profiles and dating error bars (2σ) are color coded as shown by the legend in the figure.	149
Figure 3. 6. The driving mechanism of a weakened ISM during the LGM. a). The combined speleothem records from SWMT (red) and Mawmluh cave (orange) (Berkelhammer et al., 2013; Dutt et al., 2015). The Mawmluh records were shifted by 2.4‰ negatively after correction of the temperature effect on calcite precipitation ($-0.23\text{‰}/^\circ\text{C}$, (O'Neil et al., 1969)), assuming the cave temperature is close to the annual mean temperature of the closest major cities (16.9°C at Shillong and 27.5°C at Yangon). b). 30°N summer (JJA) insolation. c). <i>G. ruber</i> $\delta^{18}\text{O}$ record, retrieved from sediment core KL126, as a proxy for ISM intensity (Kudrass et al., 2001). d). Tropical Indian Ocean SST reconstructed from sediment core SO189-29KL (Mohtadi et al., 2014). e). Timing of	

deglaciation in Tibet. The dark cyan line represents the oldest age for the onset of deglaciation, and the dark brown line represents the youngest age. The light color boxes represent the 1σ uncertainty range of the ages (Clark et al. (2009) and references therein). As shown, the local deglaciation on the Tibetan Plateau started no earlier than ~ 18 ka, indicating that the extension of snow/ice coverage was close to the maximum during the LGM (~ 19 -22 ka BP)..... 156

Figure 3. 7. Spatial-temporal comparison of speleothem $\delta^{18}\text{O}$ records from the coastal Indian Ocean and southern China. The red line represents the SWMT record in this study. The orange line represents the Mawmluh cave records (Berkelhammer et al., 2013; Dutt et al., 2015) after correcting for temperature effects. The green line is a compiled speleothem $\delta^{18}\text{O}$ record from Hulu Cave (Wang et al., 2001; Yuan et al., 2004), Dongge Cave (Dykoski et al., 2005; Wang et al., 2005), and Sanbao Cave, with the Sanbao record shifted by 1.6‰ to align with the other two cave records from southern China (Wang et al., 2008). The thick grey line shows the 65°N summer insolation. 161

Figure 4. 1. The location of Lin Noe Twin (LNT) Cave and the cave cluster in SE Yunnan (red stars). Black arrows represent present-day averaged June-July-August (JJA) wind patterns (Data from the National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010). (Kalnay et al.,

1996). The locations of other cave sites are indicated by blue stars: Sanbao Cave (Wang et al., 2008), Dongge Cave (Kelly et al., 2006; Yuan et al., 2004); Tianmen Cave (Cai et al., 2010) in China; Mawmluh Cave (Dutt et al., 2015) in India; southern Myanmar and western Thailand (SMWT) cave clusters close to the border of the two countries. The red line and red arrows indicate the 105°E designated boundary of the ISM and the EASM (Wang et al., 2003)..... 195

Figure 4. 2. Monthly mean precipitation (blue bars) and temperature (red squares) at Taunggyi and Kunming meteorological observatory stations (Data from UNdata: <http://data.un.org/Explorer.aspx?d=CLINO>)..... 196

Figure 4. 3. Plots of age versus depth for stalagmites LNT-01, LNT-02, LNT-03, LNT-04, LNT-05, LNT-09U, and LNT-09B from Lin Noe Twin cave. All ages are reported in units of thousands of years before present (1950 AD) or ka BP. The age errors indicated in the plots are 2 σ error. Linear interpolation is used to establish the chronology for sample LNT-01, LNT-03, LNT-04 and LNT-09U. StalAge method (Scholz and Hoffmann, 2011) is used to establish the chronology for sample LNT-02, LNT-05, and LNT-09B. The red lines indicated 95% confidence limits calculated by a Monte Carlo simulation fitting. 201

Figure 4. 4. Plots of age versus depth for stalagmites A-14, A-15, and A-16 from the anonymous cave; BF-02 from Baifen Cave; JJ-03, JJ-06, and JJ-08 from Jiangjun Cave; SL-08 and SL-10 from Shuanglong Cave; SY-10 from Shuiyuan Cave; and XR-05 from Xianren Cave from SE Yunnan. All

ages are reported in units of thousands of years before present (1950 AD) or ka BP. The age errors indicated in the plots are 2σ error. Linear interpolation is used to establish the chronology for sample A-14, A-15, A-16, BF-2, JJ-06, SL-08 and SY-10. StalAge method (Scholz and Hoffmann, 2011) is used to establish the chronology for sample SL-10, JJ-03, JJ-06 and XR-05. The red lines indicated 95% confidence limits calculated by a Monte Carlo simulation fitting.....203

Figure 4. 5. Scatter plot of $\delta^{18}\text{O}$ versus $\delta^{13}\text{C}$ from each stalagmite sample from Lin Noe Twin Cave. The low correlation (as indicated R^2 values) of the plotted results indicates that carbon and oxygen are not highly correlated as would be expected under the conditions of kinetic fractionation.206

Figure 4. 6. Scatter plot of $\delta^{18}\text{O}$ versus $\delta^{13}\text{C}$ from each stalagmite sample from SE Yunnan Caves. The low correlation (as indicated R^2 values) of the plotted results indicates that carbon and oxygen are not highly correlated as would be expected under the conditions of kinetic fractionation.207

Figure 4. 7. The “Hendy Test” (Hendy, 1971) on two different growth laminae from sample LNT-05 suggests that the speleothems most likely grew under isotopic equilibrium conditions.208

Figure 4. 8. The $\delta^{18}\text{O}$ time series and ^{230}Th ages of stalagmite samples from Lin Noe Twin Cave over the last 40 ka. The $\delta^{18}\text{O}$ profiles and dating error bars (2σ) are color coded as shown by the legend in the figure.212

Figure 4. 9. The $\delta^{18}\text{O}$ time series and ^{230}Th ages of stalagmite samples from SE Yunnan Cave over the last 180 ka. The $\delta^{18}\text{O}$ profiles and dating error bars

(2σ) are color coded as shown by the legend in the figure. Sample A-14 and the Holocene portion of A-15 are corrected for calcite-aragonite fractionation (see discussion). Contemporaneous $\delta^{18}\text{O}$ records from different stalagmites replicate within dating errors, suggesting equilibrium calcite precipitation. The summer (21 July) insolation at 30°N (grey curve) is also plotted for comparison.....214

Figure 4. 10. Spatial-temporal comparison of speleothem $\delta^{18}\text{O}$ records from Mainland Southeast Asia over the past 40 ka. Purple dots are from SE Yunnan caves (this study); pink dots are from Lin Noe Twin Cave (this study); cyan squares are from Mawmluh Cave (Dutt et al., 2015); and cyan dots are from the southern Myanmar and western Thailand (SMWT) caves (Chapter 3). Records from places other than SMWT were shifted accordingly to account for the oxygen isotopic fractionation that resulted from a difference in cave temperature between the sites and the Yunnan cave in the east (O'Neil et al., 1969). To highlight the one dimension spatial-temporal pattern of speleothem $\delta^{18}\text{O}$ records, we calculated the mean $\delta^{18}\text{O}$ values for the three strategic locations (Coastal BoB close to the moisture source: SMWT and Mawmluh; Middle: LNT; and Inland close to the boundary between ISM and EASM domain: SE Yunnan (Ding and Chan, 2005; Wang et al., 2003) for every thousand years (thick lines) after temperature effect corrected. The light shade areas are the 1σ uncertainty range of the $\delta^{18}\text{O}$ values.....224

Figure 4. 11. Spatial-temporal comparison of speleothem $\delta^{18}\text{O}$ records from Southern China over MIS 5. Upper left from Tianmen Cave (Cai et al., 2010); lower left from SE Yunnan Caves (this study); upper right from Sanbao Cave (Wang et al., 2008); and low right from Dongge Cave (Kelly et al., 2006). The summer (21 July) insolation at 30°N (grey curve) is also plotted for comparison. The black arrows indicated different temporal patterns of speleothem $\delta^{18}\text{O}$ in different caves during MIS 5.....228

Summary

The Asian Monsoon (AM) and its changes affect livelihoods of about half of the global population. Understanding AM variability and its impacts, therefore, is critical for local stakeholders to prepare appropriate climate mitigation and adaptation. In recent years, speleothems have been used as an important archive to reconstruct paleo-hydroclimate variability at different timescales. However, the climatic interpretation of speleothem proxies, in particular oxygen isotope ratio ($\delta^{18}\text{O}$), is highly debatable. In my thesis, I used speleothems from mainland Southeast Asia to reconstruct the annual to glacial-interglacial variability of the AM, and essentially reconciled the long-lasting debate on the interpretations of Chinese speleothems $\delta^{18}\text{O}$.

Three key findings have been achieved in my study. Firstly, I presented a 450-yr long record of trace element variations in an annually laminated stalagmite retrieved from a cave in central Myanmar. The record suggests that monsoon regime shifted from relatively wet to frequently dry conditions around the ninth century. Its timing coincides with the mysterious demise of the ancient Pyu Kingdom that flourished in central Myanmar over a thousand years. Secondly, I obtained a speleothem $\delta^{18}\text{O}$ record from the eastern coastal region of the Bay of Bengal, which covers both the Last Glacial Maximum (LGM) and the present day. The dramatic difference in $\delta^{18}\text{O}$ values between the two periods indicates that forced mainly by glacial boundary conditions, regional rain amount was

substantially low during the LGM, ~56% of modern values. Lastly, I established three sets of speleothem $\delta^{18}\text{O}$ records along a SW-NE transect of monsoon trajectory from southeastern coastal Myanmar to southwestern China. I found that speleothem $\delta^{18}\text{O}$ values vary significantly between glacial and interglacial periods at sites closer to monsoon moisture source, but much less further inland. This spatial variability suggests a stronger oxygen isotope fractionation in precipitation associated with progressive rainout of AM moisture during glacial times. The stronger fractionation during glacial periods is probably resulted from a larger temperature gradient between the sites and suppressed plant transpiration effect. I further propose that the muted glacial-interglacial variations shown in speleothem $\delta^{18}\text{O}$ records from southern China, located in the downwind region of the AM, were probably due to enhanced (weakened) isotope fractionation of transported moisture during glacial (interglacial) periods. This so-called transport pathway effect on water isotope fractionation counteracts the forcing of glacial boundary conditions.

These findings in my thesis not only contribute to our understanding of AM variability in the recent 180,000 years, but also provide an evaluation to climate models, and therefore, help to better predict hydroclimate change in the region.

Chapter 1

Introduction

For paleoclimate, the past two decades have been the age of the ice core. The next two may be the age of the speleothem.

—————Gideon M. Henderson, 2006 *Science* (313):622

Mainland Southeast Asia (MSEA), previously known as Indochina Peninsula or Indochina, is the continental portion of Southeast Asia (in contrast to the Maritime Southeast Asia). The majority of the peninsula has hot and humid climate throughout the year and plentiful rainfall during boreal summer, with exceptions in the northernmost part close to the Himalaya, where winter is cold and snowy, and the southernmost part close to the equator, where rain occurs throughout the year. Demographically, MSEA is home to over 300 million people living in seven countries. Unfortunately, a large percentage of these people live in poverty, heavily reliant on farming (e.g., rice and rubber production). In fact, three of the countries, Cambodia, Laos, and Myanmar, are on the United Nation's list of the 'Least Developed Countries'. With such weak economies and relatively poor infrastructure, the socio-economic health of much of MSEA is extremely vulnerable to various hydroclimate hazards, such as floods, droughts and storms, particularly in the three aforementioned

countries. For example, in 2008, the destructive Cyclone Nargis made landfall in Myanmar's highly populated Irrawaddy River Delta, claimed over 138,000 fatalities, and caused damage worth 10 billion USD, making it one of the deadliest cyclones in recorded history (Fritz et al., 2009). In 2011, Thailand was hit by severe floods, with losses and damages estimated at 45.7 billion USD. As an additional consequence, the 2011 flood led to an acute shortage of hard disks on the global market as Thailand is a major hard drive exporter (Tsai et al., 2015). Therefore, a good knowledge of hydroclimate variability in the region is invaluable, as it will allow local governments to plan appropriate mitigation and adaptation policies.

1.1. Regional hydroclimate

The hydroclimate of MSEA is influenced by the Indian Summer Monsoon (ISM), a major branch of the Asian Monsoon (AM) (Fig 1.1) (Ding and Chan, 2005; Wang et al., 2003), and many other important climate phenomena, such as the El Niño–Southern Oscillation (ENSO) (Kumar, 1999; Kumar et al., 2006; McPhaden et al., 2006) and the Indian Ocean Dipole (IOD) (Abram et al., 2008; Saji et al., 1999; Webster et al., 1999).

The AM (Fig. 1.1), which is a result of the different thermal-properties between the land and ocean (An et al., 2015), dominates the regional hydroclimate, producing distinctive rainy and dry seasons. During the rainy season (mostly May to September), fueled by the warm Indian Ocean, the AM brings to the region vast amounts of water mainly from the Bay of Bengal (BoB), and contributes to over 80% of the annual precipitation (Sein et al., 2015). During the dry season (mostly November to February), the cold and dry wind comes predominantly from the northern mid-to-high latitudes.

The ENSO involves variations in sea surface temperature (SST) and atmospheric pressure across the tropical Pacific (McPhaden et al., 2006), and profoundly influences rainfall across the Indo-Pacific. During the El Niño events, the west Pacific is abnormally cold, and the upward branch of the Walker Circulation shifts east, following the anomalous ocean warming, from

the Indo-Pacific Warm Pool, leaving more sinking motion in the ISM domain, resulting in less rain in MSEA (Kumar, 1999; Kumar et al., 2006; Mohtadi et al., 2016). It should be noted that the position of anomalous ocean warming in the Pacific, not just its strength, during an El Niño significantly influences the ISM rainfall (Kumar et al., 2006). El Niño events with the warmest SST anomalies in the central equatorial Pacific are more effective in producing droughts in the ISM domain than events with the warmest SSTs in the eastern equatorial Pacific. For example, the 2002 El Niño resulted in much more severe droughts in the ISM domain compared to the 1997 El Niño, although the latter was much stronger (Kumar et al., 2006). During La Niña events, in contrast, when convection strengthens in the Warm Pool, the regional monsoon generally brings more rain (Kumar, 1999).

The IOD is defined based on SST anomalies between the west (60° - 80° E, 10° S- 10° N) and east (90° E- 110° E, 10° S- 0° N) of the tropical Indian Ocean (Saji et al., 1999). The IOD strongly influences the Indian Ocean and its neighboring region (Saji et al., 1999; Webster et al., 1999). However, the exact mechanism of how the IOD modulates the rainfall in the ISM region is not well understood, with clear discrepancies between climate models and proxy data (Abram et al., 2008; Ashok et al., 2004; D'Arrigo et al., 2011).

A crucial question regarding the regional hydroclimate is how it will change in response to the ever-warming climate (Fig 1.2). The question has been

investigated intensively. Thermodynamically, humid MSEA is predicted to get even wetter (Held and Soden, 2006), with a higher CO₂ concentration than today and consequently a higher global temperature. This general prediction, however, may not be ubiquitously valid, as it was mostly based on oceanic data, while projections over continents show considerable complexities (Feng and Zhang, 2015; Greve et al., 2014). The most advanced climate models simulate significant changes in the regional rainfall distributions, however, also with considerable inconsistency (Fig 1.2) (Chadwick et al., 2015; Turner and Annamalai, 2012). Indeed, rather than monotonically increasing after the industrial revolution, instrumental data of monsoon rainfall show multidecadal variability, and more importantly, an overall decreasing trend in India since 1950 (Parthasarathy et al., 1994). A similar overall trend has been observed in MSEA (Manton et al., 2001), although with a large uncertainty. For example, in Myanmar, no significant trend in rainfall amount is observed (Misra and DiNapoli, 2013; Roy and Kaur, 2000), which is likely due to the instrumental data spanning only about 40 years (1961 to 1998). The decreasing trend of AM rainfall is proposed to result from aerosol forcing (Bollasina et al., 2011; Ramanathan et al., 2005; Turner and Annamalai, 2012). The local emission of aerosols is increasing following the industrialization of Asia. The aerosols scatter or absorb solar radiation and thus reduce energy from reaching the Earth's surface, thus countering the CO₂ forcing (Turner and Annamalai, 2012). However, the exact role of aerosol in influencing monsoonal rainfall is ambiguous, with studies (e.g., Wang, (2004)) suggesting that an increase in

aerosol will strengthen rather than weaken the monsoonal rainfall (see Turner and Annamalai (2012) for a review). In summary, research regarding the future regional hydroclimate under global warming has reached no consensus yet.

Figure 1. 1. Left: the Asian Monsoon system (modified after Johnson. (2011)). The black rectangle crops out MSEA. Right: Location of published paleo-hydroclimate records in MSEA. See Table 1.1 for details.

1.2. Regional paleo-hydroclimate

Given the uncertainties with respect to regional hydroclimate variabilities, long and continuous paleoclimate records from MSEA become invaluable, as they will help supplement the short instrumental data, evaluate the robustness of climate model simulations, examine rainfall behaviors in response to various climatic forcings in Earth's recent history, and advance our knowledge of teleconnections between the AM and other major climate components, such as the ENSO and the IOD. Furthermore, MSEA, particularly its western coast, is the first to embrace the AM in eastern Asia, and the local rainfall is almost solely supplied by moisture carried by the AM. A spatial-temporal comparison of hydroclimate records along the monsoonal trajectory from MSEA and the downstream AM, namely southern China, will allow us to trace the monsoon moisture transport under different climate regimes and improve our understanding of AM thermodynamics.

Various proxy-based paleoclimate reconstructions of the AM have been obtained from climate archives, such as lake and marine sediments (e.g., An et al., 2011; Kudrass et al., 2001; Rashid et al., 2007; Yancheva et al., 2007), loess deposits (e.g., Maher, 2016; Sun et al., 2015), speleothems (e.g., Cai et al., 2015; Cheng et al., 2016; Kathayat et al., 2016; Sinha et al., 2007; Y. Wang et al., 2005), and tree rings (e.g., Buckley et al., 2010; Cook et al., 2010; D'Arrigo et al., 2011). However, compared with India and southern China, where proxy-

based records are readily available, very few terrestrial paleoclimate records are available from MSEA. Most of the early works in MSEA focus on sediment palynology (e.g., Penny (2001) and White et al. (2004); for literature overviews see Chabangborn et al. (2014) and Chabangborn and Wohlfarth (2014)), which indicates local vegetation changes, thus indirectly reflecting the hydroclimate variability. Direct reconstruction of proxy-based hydroclimate history in MSEA from paleoclimate archives is a very recent endeavor, with only a handful of papers available, published almost entirely within the last decade. The paleoclimate archives studied include tree rings (Buckley et al., 2007; 2010; Cook et al., 2010), lake sediments (Chawchai et al., 2013; 2015a; 2015b; 2016; D'Arrigo et al., 2011; Leiberman, 2003; Sun et al., 2016; Wohlfarth et al., 2016; 2012; Yamoah et al., 2016a; 2016b; 2016c; 2017), speleothems (Cai et al., 2010; Chawchai et al., 2017; Hua et al., 2017; Muangsong et al., 2014; Yang, 2016), bivalve shells (Marwick and Gagan, 2011), and groundwater (Aggarwal, 2004), and they are summarized in Figure 1.1 and Table 1.1. Most of the studies' data sets, however, span no longer than the most recent two millennia (Table 1.1) and often bear poor chronology. The few records that cover Holocene or beyond suffer from low temporal resolution and, again, often poor chronology. Furthermore, paleo-hydroclimate interpretations inferred from these proxies frequently contradict each other. For example, on decadal scales, tree ring records (D'Arrigo et al., 2011) and lake sediments (Sun et al., 2016) revealed an inconsistent relationship between the IOD and monsoonal precipitation over Myanmar. Such discordance is likely due to our poor

understanding of the regional complexity of the AM precipitation and/or the incomplete interpretation of climate proxies and age uncertainties in the records. Thus, long, continuous, and well-dated hydroclimate records from continental archives are needed to understand the nature and cause of monsoonal rainfall variations.

In recent years, speleothems have been used as an important archive to reconstruct paleoclimate, especially paleo-rainfall variability at different timescales. However, only five short and isolated speleothem records from MSEA are available (Cai et al., 2010; Chawchai et al., 2017; Hua et al., 2017; Muangsong et al., 2014; Yang, 2016). A major portion of MSEA, such as its largest country, Myanmar, is unexplored, largely due to its political isolation during the past 60 years. Fortunately, with the recent improvement of the political situation, systematic speleothem research has become possible there.

Figure 1. 2. Projections on tropical land precipitation changes in two advanced climate models show large discrepancies (Chadwick et al. (2015)). Precipitation change for 2071–2100 minus 1971–2000, under the RCP8.5 emissions scenario for GFDL-ESM2M (a) and HadGEM2-ES (b).

Table 1. 1. Summary of terrestrial proxy-based paleo-hydroclimate records from MSEA.

Archives	Site NO.	Site Name	Country	Location	Time range	References
Lake sediment	1	Lake Twintaung	Myanmar	22°22'N, 95°02'E	Last ~500yr	Sun et al.(2016)
	2	Lake Pa Kho	Thailand	17°06'N, 102°56'E	Last ~2,000yr	Chawchai et al.(2015a) Yamoah et al.(2016b,2016c) Wohlfarth et al.(2016)
	3	Lake Kumpawapi	Thailand	17°11'N, 103°02'E	Last ~9,500 yr	Wohlfarth et al.(2012, 2016)
	4	Lake Nong Thale Prong	Thailand	7°11'N, 99°23'E	Last ~150 yr	Chawchai et al.(2013, 2015b, 2016) Yamoah et al.(2016a)
Tree ring	5	Mae Hong Son(MHS)	Thailand	Northwestern Thailand	Last ~500 yr	Buckley et al.(2007)
	6	Mu Cang Chai(MCC)	Vietnam	Northern Vietnam	Last ~500 yr	Leiberman.(2003)
	7	Bidoup Nui Ba National Park (BDNP)	Vietnam	Southern Vietnam	Last ~750 yr	Buckley et al.(2010)
	8	Maingtha Forest Reserve(MFR) Monsoon Asia Drought Atlas (MADA)	Myanmar MSEA	Central Myanmar	Last ~400 yr Last ~1000 yr	D'Arrigo et al.(2011) Cook et al.(2010)
Speleothem	9	NamJang Cave	Thailand	98°12'12"E, 19°40'30"N	Last ~400 yr	Cai et al.(2010) Muangsong et al.(2014)
Others	10	Unname Cave	Cambodia	10°38'40"N, 104°16'20"E	Last ~1,000 yr	Hua et al.(2017)
	11	Tham Mai Cave	Laos	20°45'N, 102°39'E	Last ~13,000 yr	Yang. (2016)
	12	Tham Nam Cave	Thailand	15°03'N,99°29'E	~85,000-~104,000 yr	Chawchai et al.(2017)
	13	Mae Hong Son Bivalve Groundwater	Thailand Thailand and Vietnam	Northwestern Thailand	~5,000-~35,000 yr last ~20,000 yr	Marwick and Gagan.(2011) Aggarwal et al.(2004)

1.3. Speleothems as climate archives

Speleothems, or cave deposits (from Greek *spelaiion*, cave, and *thema*, deposit), are almost entirely composed of calcite and/or aragonite. They form essentially due to carbonate dissolution in the soil and epikarst, and re-precipitation in the underlying vadose region (Fig 1.3). Speleothems were formerly considered to be too complex to provide reliable paleoclimate records, and until the late 1990s they were barely mentioned in paleoclimate textbooks (Fairchild and Baker, 2012). The turning point came in 2001, when Wang et al. (2001) published the first well-dated speleothem record from the now-famous Hulu cave, southern China. The paper has effectively demonstrated that speleothems, and particularly stalagmites, encapsulate detailed and accurate records of the past climate. Following this milestone, numerous speleothem records have been reconstructed across the globe. Those records have reaffirmed the robustness of speleothems as climate archives and significantly improved our understanding of the Earth's past climate and environment.

1.3.1 U-Th Dating

The most obvious advantage of speleothems as a climate archive is that the carbonate material can be absolutely and precisely dated by the U-Th disequilibrium dating technique. U is oxidized to hexavalent state (U(VI)) at or near surface conditions, and has much higher solubility in groundwater

compared to Th, which is in tetravalent state (Th(IV)). When speleothem calcareous minerals grow, U(VI) can enter their crystal lattice, while Th(IV) can hardly be incorporated, and in most cases, its initial amount in speleothem carbonate is considered negligible. In rare cases, when such initial Th amount is higher, isochron techniques can be applied to better determine the ages, (e.g., Carolin et al., 2013). If the calcareous mineral lattice remains closed (i.e., no further U or Th is gained or lost), which applies to most speleothem carbonates, and which can be determined by petrographic examination, the age of corresponding carbonate minerals can be calculated by measuring the isotopic ratios of U and Th. With recent improvements in the multi-collector inductively coupled plasma mass spectrometric (MC-ICP-MS) techniques (Cheng et al., 2013; Shen et al., 2012), speleothem samples as young as only several years old (e.g., Shen et al. (2013)) and as old as ~700,000 years (e.g., Cheng et al. (2016)) can be dated. Even in the case of very small sample sizes (e.g., < ~10 mg if samples contain ~1 ppm uranium), the typical analytical precision is now better than 0.5%.

1.3.2. Speleothem proxies

Speleothems form under complex interactions between carbonate rocks and groundwater, and they archive multiple climatic proxies. The three most studied proxies include $\delta^{18}\text{O}$ (e.g., Cheng et al., 2016; Wang et al., 2001); $\delta^{13}\text{C}$ (e.g.,

Genty et al., 2003), and trace element (e.g., Mg, Sr, Ba, U, P...) contents (e.g., Bernal et al., 2016; Roberts et al., 1998; Treble et al., 2003).

Among the three, speleothem $\delta^{18}\text{O}$ has been demonstrated to faithfully record the meteoric precipitation $\delta^{18}\text{O}$ (Cheng et al., 2012), and is the most extensively studied. Ever since Wang et al. (2001) set the benchmark, numerous well-dated speleothem $\delta^{18}\text{O}$ records across the globe (e.g., Asmerom et al., 2010; Carolin et al., 2016; Cheng et al., 2016; Fleitmann et al., 2003; Luetscher et al., 2015; Wang et al., 2004) have been published, and have greatly improved our understanding of the global hydrological cycle. However, the published speleothem $\delta^{18}\text{O}$ records, when compared to other hydroclimate records in the AM domain, have generated intense debate. The Chinese speleothem $\delta^{18}\text{O}$ records exhibit a dominant ~ 23 kyr precessional cycle, mainly forced by local summer insolation (Cheng et al., 2016; Y. Wang et al., 2005; 2008), whereas records from many other types of monsoonal archives, for example, loess deposits (An, 2000; An et al., 1990; Sun et al., 2015; 2006), marine sediments (Clemens and Prell, 2010; P. Wang et al., 2005; 2016), and lake sediments (An et al., 2011), reveal a dominant ~ 100 kyr cycle, suggesting a direct link to glacial boundary forcing. This inconsistency can in part be attributed to our incomplete interpretation of the speleothem $\delta^{18}\text{O}$ proxy. Since the absence of glacial-interglacial variation, the largest climate variation in the most recent million years, in speleothems has not been satisfactorily explained (Figure 1.4), the interpretation of speleothem $\delta^{18}\text{O}$ as monsoon intensity has been challenged

by many researches (e.g., Caley et al., 2014; Chiang et al., 2015; Clemens et al., 2010; Dayem et al., 2010; Johnson and Ingram, 2004; LeGrande and Schmidt, 2009; Liu et al., 2014; Maher, 2008; Pausata et al., 2011; Sun et al., 2015). Their alternative explanations tend to emphasize the moisture source effect on Chinese speleothem $\delta^{18}\text{O}$ (e.g., Maher, 2008; Pausata et al., 2011). To tackle this problem, the Chinese speleothem records can be compared with those from the upwind ISM domain, namely MSEA. The comparison will help to separate local and regional effects on the isotope records, and reveal the dynamics of atmospheric circulation patterns (Hu et al., 2008). The very debate on the interpretation of speleothem $\delta^{18}\text{O}$ records from southern China is one of the main foci of this thesis, and is discussed in detail in later chapters.

The next proxy, $\delta^{13}\text{C}$, has been humorously described as “always measured—rarely discussed”. $\delta^{13}\text{C}$ values are always provided as a byproduct of carbonate $\delta^{18}\text{O}$ measurement. However, climatic interpretation of $\delta^{13}\text{C}$ is difficult due to a multiplicity of processes and sources. Speleothem $\delta^{13}\text{C}$, if in equilibrium, registers isotope signals from two major carbon reservoirs: bedrock carbonate and soil CO_2 . Bedrock carbonate $\delta^{13}\text{C}$ values show small variations, and in general are close to 0‰ (Veizer and Hoefs, 1976). In contrast, soil reservoir $\delta^{13}\text{C}$ values are affected by surface vegetation density and/or composition (i.e., C3 or C4 plants). The soil respired CO_2 has $\delta^{13}\text{C}$ values of -26‰ to -20‰ in equilibrium with C3 plants, while $\delta^{13}\text{C}$ in CO_2 equilibrated with C4 plants is much heavier, ranging from -16‰ to -10‰, due to different pathways in which

CO₂ assimilates during photosynthesis (McDermott, 2004). This difference is prominent; therefore, speleothem $\delta^{13}\text{C}$ can be used to reconstruct the shift between C3 and C4 vegetation, when independent evidence, such as pollen records, corroborates the shift. However, in most cases, C4 vegetation is absent, and the variations of speleothem $\delta^{13}\text{C}$ are to a large extent determined by the relative openness of percolating water to the soil CO₂ (McDermott, 2004; see Genty et al. (2003) for an example), which is very hard to constrain. Other effects such as CO₂ degassing and prior calcite precipitation (PCP) (Fairchild and Treble, 2009; McDermott, 2004) will further complicate the interpretation. To comprehensively understand how $\delta^{13}\text{C}$ signal transfers from soil and bedrock to speleothems, monitoring of cave hydrology and precipitation above caves is invaluable (see Matthey et al. (2008) for an example).

The third proxy, speleothem trace elements, has been attracting a growing interest. With the help of techniques such as laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) (Desmarchelier et al., 2006) and X-ray fluorescence (XRF) (Duan et al., 2014), speleothem trace elements can be measured precisely and rapidly in ultra-high spatial resolution. Furthermore, compared to stable isotope measurements, which destroy carbonate samples, *in situ* trace element measurements cause almost no sample damage. However, in the past decades, the application of trace elements as environmental proxies was limited by their similarly complex nature, if not more complex, than $\delta^{13}\text{C}$ (Fairchild et al., 2000; Fairchild and Treble, 2009). Unlike corals or

foraminifera, whose trace elements precipitate from ocean water where the trace element–calcium ratio is fixed or very well understood, speleothems hardly show regional coherency in trace element patterns; sometimes, incoherent patterns can be found even within the same cave chamber, because individual speleothems are fed with drip water whose trace element contents vary due to complex climatic, geologic and geomorphic environments the water has passed through (Fairchild et al., 2000; Fairchild and Treble, 2009). Despite such difficulties, with the help of cave monitoring, we increasingly understand the key processes that regulate speleothem trace element concentration. For example, a new paradigm is emerging in which PCP is used to explain the well-documented co-variations of alkaline-earth elements (i.e., Mg, Sr, Ba) (e.g., Bernal et al., 2016; Griffiths et al., 2010; Sinclair et al., 2012; Tremaine and Froelich, 2013), and is further linked to wet or dry climate conditions (Fairchild et al., 2000; Fairchild and Treble, 2009).

Other speleothem petrographical and chemical proxies include luminescent laminae (e.g., Tan et al., 2006), fluorescent of organic matter (e.g., Blyth et al., 2016; McGarry and Baker, 2000), and growth rate (e.g., Baker et al., 1998; Duan et al., 2014). Recently, various other geochemical proxies in speleothems have been recognized to potentially indicate the surface environmental conditions, for example, strontium isotopes (e.g., Belli et al., 2017), lead isotopes (e.g., Yang et al., 2015), magnesium isotopes (e.g., Buhl et al., 2007; Galy et al., 2002), uranium isotopes (e.g., Zhou et al., 2005), sulphur isotopes

(e.g., Wynn et al., 2008), lithium isotopes (e.g., Pogge von Strandmann et al., 2017), carbonate clump isotopes (e.g., Meckler et al., 2015), organic compounds (e.g., Blyth et al., 2016), magnetic minerals (e.g., Bourne et al., 2015), and water and noble gases in speleothem inclusions (e.g., Affolter et al., 2015; Griffiths et al., 2010; Kluge et al., 2008; Labuhn et al., 2015; Meckler et al., 2012).

In addition, speleothem growth requires dripping water, which can be interrupted due to droughts, sub-zero temperatures, or cave inundation. Since speleothems can be dated precisely, their growth intervals can then constrain the timing of pluvial events (e.g., Wang et al., 2004), increased rainfall (e.g., Jo et al., 2014), thawing of permafrost (e.g., Vaks et al., 2013), or water level increase (e.g., Bard et al., 2002).

Figure 1. 3. Schematic diagram of a cave system and carbonate dissolution and deposition (Fairchild and Baker, 2012).

Figure 1. 4. The Chinese speleothem $\delta^{18}\text{O}$ records show different periodicities compared with other major paleoclimate records. The compiled speleothem $\delta^{18}\text{O}$ record from southern China (green) (Cheng et al., 2016) follows the 65°N summer insolation (grey) tightly and is dominated by the 23 kyr precessional cycle. However, other major paleoclimate records, such as atmospheric CO_2 concentration (red) (Lüthi et al., 2008) and Antarctica surface temperature variation (Jouzel et al., 2007) retrieved from ice core EPICA Dome C and benthic foraminifera $\delta^{18}\text{O}$ LR04 stack (purple) (Lisiecki and Raymo, 2005) are dominated by the 100 kyr glacial-interglacial cycle.

1.4. Field trips and speleothem samples

Karst landforms are present in two areas in Myanmar: the Shan-Plateau in central and northeastern Myanmar, and southern Myanmar. The Shan-Plateau is covered with tracts of limestone dating back to Permian age (Oo et al., 2002), and karst landscapes there are well developed. Southern Myanmar, a narrow strip bordering western Thailand, includes numerous scattered and isolated karst mountains that can be easily recognized from satellite images. A number of caves in the two areas have already been explored and documented by professional cavers from the Swiss Caving Society and the British Cave Research Association. They have built a website ¹, *Myanmar Cave Documentation Project*, which provided important guidance during our field trip planning. Furthermore, caves in Southeast Asia bear religious significance and thus can be easily located with the help of locals.

However, conducting field work in Myanmar is challenging as part of the country is still politically unstable (e.g., the north-eastern part of the country is within the “Golden Triangle”). Considering scientific merits, safety, and logistic issues, I decided to collect speleothem samples from three cave locations inside or close to Myanmar: (1) southern Myanmar and western Thailand close to the Thailand-Myanmar border, (2) central Myanmar, and (3) southeast Yunnan Province near the China-Myanmar border, with sites one and

¹ <http://www.myanmarcaves.com>

two inside Myanmar as priority locations (Fig 1.5). All these sites were selected based on the following considerations: The western side of mainland Southeast Asia directly faces the Bay of Bengal, where most of the Asian Monsoon moisture comes from. More importantly, these locations are all dominated by the ISM, share a similar climate (although the latter two locations are relatively cooler due to their high elevations), and form a southwest-northeast transect, which nearly follows the monsoon streamline. In fact, for the third location, scientifically, northeastern Myanmar would have been a better choice, as it would then have followed together with the other two locations more strictly along the SW-NE monsoon trajectory, and all locations would have been evenly spaced. However, northeastern Myanmar has remained not safe to travel to due to tribal insurgencies. We thus selected SE Yunnan as a reasonable alternative. Although the location is relatively distant downstream from the first two locations, the local climate is still dominated by the ISM, similar to northeastern Myanmar, and more importantly, samples are easily accessible in SE Yuan.

In collaboration with colleagues from the Myanmar Earthquake Committee, Myanmar, and Chulalongkorn University, Thailand, and with the help of locals, our team successfully conducted one fieldtrip to Thailand and three fieldtrips to Myanmar. I was the main participant in all these trips, and particularly the third trip to Myanmar, which was entirely organized by myself and funded by my

“Dr. Stephen Riady Geoscience Scholarship”. The four trips are summarized in Table 1.2.

In total, we explored over 60 caves and successfully retrieved over 60 stalagmite samples. We also collaborated with colleagues from Xi'an Jiaotong University, China, and shared stalagmite samples collected from caves in southeast Yunnan, China. All the speleothem samples collected are listed in Table 1.3. Over 300 ^{230}Th ages were determined, and over 3,000 stable isotope analyses were conducted on the samples in five different geochemistry laboratories. Over 100 machine hours were spent on trace element analyses. Among them, over 250 ^{230}Th dates, over 1,500 stable isotope analyses and all the trace element analyses were conducted by myself, which equaled to over 2,500 working hours in the laboratory. The details of the data analyses, such as sample preparations and machine particulars, will be described in detail in later chapters.

Table 1. 2. Summary of fieldtrips.

Fieldtrip	Duration	Team Members
Kayin and Mon State, Myanmar	11/20/2012-11/29/2012	Liu Guangxin, Wang Xianfeng and Chiang Hong-Wei from Earth Observatory of Singapore Lin Thu Aung from Myanmar Earthquake Committee
Shan State, Myanmar	05/01/2013-05/13/2014	Phyo Maung Maung from Myanmar Department of Meteorology and Hydrology Liu Guangxin, Wang Xianfeng, Chiang Hong-Wei and Joanne Lynne from Earth Observatory of Singapore (Joanne Lynne is a professional photographer) Win Naing Tun from Myanmar Earthquake Committee
Bai Rai District, Thailand	01/09/2016-01/13/2017	Phyo Maung Maung from Myanmar Department of Meteorology and Hydrology Liu Guangxin and Chiang Hong-Wei from Earth Observatory of Singapore Sakonvan Chawchai, Kampanart Jahkham and Warisa Paisonjunlongsri from Department of Geology, Chulalongkorn University Siripong Tonogto from Thailand PANDA CAMP
Shan State and Mandalay Region, Myanmar	11/11/2016-11/22/2017	Liu Guangxin, Wang Xianfeng and Lin Thu Aung from Earth Observatory of Singapore Chiang Hong-Wei from Department of Geosciences, National Taiwan University Win Naing Tun from Myanmar Earthquake Committee

Figure 1. 5. Sampling strategy, with circled numbers indicating our sampling locations. The lower right inset shows the AM system (modified after Johnson (2011)). The black arrows represent the present-day averaged June-July-August (JJA) wind pattern. The data are from the National Centres for Environmental Prediction (NCEP) / National Centre for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010) (Kalnay et al., 1996). The numbered circles indicate the three sampling locations in Myanmar and China (see section 1.4).

1.5. Thesis structure

The objectives of this thesis are: (1) to reconstruct the hydroclimate changes in Mainland Southeast Asia from interannual to glacial-interglacial timescales; (2) to evaluate the driving mechanisms of the Indian Summer Monsoon; and (3) to improve the understanding of the speleothem $\delta^{18}\text{O}$ proxy.

The subsequent chapters are structured as follows:

In Chapter 2, I present an ultra-high resolution trace element record from central Myanmar, covering ~450 years of recent Myanmar history. The anti-correlation between Mg and Sr in the record on seasonal to decadal time scales demonstrates that the trace elements are essentially controlled by monsoonal precipitation. I further evaluate natural variations of the inferred ISM precipitation in Myanmar.

In Chapter 3, I reconstruct ISM variability during the last glacial period, from the speleothem $\delta^{18}\text{O}$ records retrieved from southeastern Myanmar and western Thailand. I find that ISM precipitation was significantly reduced during this period, particularly during the Last Glacial Maximum, likely due to different glacial boundary conditions. A comparison with the Chinese speleothem records suggests that the sensitivity of speleothem $\delta^{18}\text{O}$ to major climate forcings varies spatially.

In Chapter 4, I trace the ISM moisture transport following a SW-NE transect from coastal Myanmar to SW China. I find that oxygen isotope fractionation in precipitation associated with progressive rainout over the ISM domain increased significantly during the glacial times, likely due to much colder temperatures and suppressed plant transpiration effect. I demonstrate that stronger isotopic fractionation from the ocean source region to cave sites is the main reason why the Chinese speleothem $\delta^{18}\text{O}$ records show unexpectedly low values during glacial maxima. I therefore further hypothesize that stronger (weaker) water isotope fractionation during glacial (interglacial) periods, dubbed transport pathway effect, is the key to the long-lasting debate on the orbital-scale pace of the AM.

In Chapter 5, I summarize the main conclusions of the thesis and discuss the potential future work stemming from the thesis.

Table 1. 3. Speleothem samples from Mainland Southeast Asia.

Cave	Location		Sample
Southern Myanmar			
Phabaung Gu	N 16°17.210’	E 97°54.092’	PHA01, PHA02, PHA03
Padamya Gu	N 16°49.725’	E 97°42.369’	PDM01, PMD02, PDM03
Ya Thea Pyan Gu	N 16°50.107’	E 97°34.245’	YTP01, YTP02, YTP03, YTP04
Saddan Gu	N 16°32.019’	E 97°42.895’	SADG01
Western Thailand			
Khao Pare Cave-1	N15° 05.175'	E99° 25.211'	KPC1-01
Khao Pare Cave-2	N15° 05.154'	E99° 25.155'	KPC2-01
Hin Tum-2	N15° 01.272'	E99° 26.757'	HT2-1, HT2-2
Tham Khao Dand	N15° 08.823'	E99° 27.784'	TKD01, TKD02, TKD03
Wat Tham Khao	N15° 01.780'	E99° 27.348'	WTKW01

Wong

Central Myanmar

Lin Noe Twin	N 21°14'13.4"	E 96°26'37.5"	LNT01, LNT02, LNT03, LNT04, LNT05, LNT06, LNT07, LNT08, LNT09, LNT10, LNT11, LNT12, LNT13, LNT14, LNT15, LNT16, LNT17, LNT18, LNT19, LNT20, LNT21
Happy Monk Cave	N 20°51'02.7"	E 97°14'23.0"	HYM01, HYM02, HYM03, HYM04, HYM05, HYM06
U Hmin Gu	N 20°43'22.0"	E 96°35'23.0"	UH01, UH02, UH03, UH04
Deedo Cave	N 21°42'44.1"	E 96°21'30.7"	DDT01, DDT02, DDT03, DDT04
Kyauk Sa Gu	N 20°49'04.1"	E 97°20'05.8"	KYS01, KYS02

Upper Kho Nan Shin	N 21°44'9.2"	E 96°20'13.7"	UNKS01
Maw Bar Gu	N 21°23'47.5"	E 96°21'13.6"	MGB01, MGB02
Leikte Gu	N 20°38'02.9"	E 96°33'04.9"	LAY01

Southeast Yunnan

Unname Cave	~N 23°20'	~E 104°10'	A14, A15, A16
Baifen Cave	~N 23°20'	~E 104°10'	BF02
Jiangjun Cave	~N 23°20'	~E 104°10'	JJ03, JJ06, JJ08
Shuiyuan Cave	~N 23°20'	~E 104°10'	SY10
Shuanglong Cave	~N 23°20'	~E 104°10'	SL08
Xianren Cave	~N 23°20'	~E 104°10'	XR05

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Chapter 2

A 450-year-long Indian Summer Monsoon record from speleothem trace elements in Myanmar

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Contributions

Liu Guangxin is the primary author. Liu Guangxin, Wang Xianfeng, and Zhu Zhiyong contributed to the design of this study. Liu Guangxin, Wang Xianfeng, Chiang Hong-Wei, and Phyo Maung Maung collected and prepared the samples, which were subsequently analyzed by Liu Guangxin and Zhu Zhiyong, and interpreted by Liu Guangxin, Wang Xianfeng, and He Shaoneng. Liu Guangxin wrote this paper.

Liu Guangxin: 55%

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2.1. Abstract

Despite the critical impact of the Indian Summer Monsoon (ISM) on human society, its past variability is still poorly understood. Paleoclimate records are particularly rare in Myanmar where the climate is dominantly controlled by ISM variability. In this chapter, we report the first speleothem study conducted in central Myanmar. We present trace element records spanning ~450 A.D. to ~900 A.D. from an annually laminated stalagmite. We demonstrate that strong high-frequency oscillations of Mg/Ca, Sr/Ca, Ba/Ca, and U/Ca in the sample are annual, and that their temporal variations can be attributed to changes in ISM rainfall. The trace element variations further indicate that local precipitation associated with the ISM is linked to solar cycles as well as influenced by the El Niño–Southern Oscillation (ENSO) and the Indian Ocean Dipole (IOD) on interannual to decadal time scales. In addition, our Mg/Ca record suggests that monsoonal rainfall in Myanmar shifted from a relatively wet state into a frequently dry regime around the ninth century, which could have disturbed the flourishing ancient Pyu society.

2.2. Introduction

Myanmar is the largest country in Mainland Southeast Asia (MSEA), yet it is also among the least developed ones in the world (UN DESA, 2015). Economically, it heavily relies on agriculture, which employs over 60 percent of the population and contributes to around a third of the country's GDP (UN OCHA, 2017). Myanmar's socio-economic health is, therefore, highly vulnerable to climate change, particularly to hydroclimate hazards. For example, in 2008, the cyclone *Nargis* brought severe flooding to the country and claimed over 138,000 fatalities, making it the second deadliest cyclone in recorded history (Fritz et al., 2009); in 2015, an unusually heavy monsoon resulted in severe flooding in western Myanmar, temporarily displacing approximately 1.7 million people and killing 172 (UN OCHA 2016); and in 2016, a 'Godzilla' El Niño (Kintisch, 2016) led to a heat wave and a severe drought in Myanmar, leaving roughly 300 villages in shortage of water (The Republic of The Union of Myanmar Ministry Of Information, 2016). With such extreme events likely becoming more frequent in a warming world (Cai et al., 2014a; 2014b; Knutson et al., 2010; Webster et al., 2005), knowledge of Myanmar's hydroclimate variability is urgently demanded by local governments in drafting policies for climate change mitigation and adaptation.

Myanmar, located by the Bay of Bengal, is the first MSEA country in the path of the Indian Summer Monsoon (ISM) (Fig. 2.1). Most parts of the country are

characterized by distinct dry and wet seasons, with the majority of the annual precipitation falling between late April and early November (Sein et al., 2015; Sein and Zhi, 2016; Shrivastava et al., 2016). Inter-annual variations of the ISM are driven by solar variabilities (Agnihotri et al., 2002; An et al., 2015; Boos and Storelvmo, 2016; Mohtadi et al., 2016) and modulated by El Niño–Southern Oscillation (ENSO) (An et al., 2015; Ashok et al., 2004; Kumar, 1999; Kumar et al., 2006; Mohtadi et al., 2016) and the Indian Ocean Dipole (IOD) (Ashok et al., 2004; 2001; Saji et al., 1999; Sein et al., 2015; Webster et al., 1999). The Pacific Decadal Oscillation (PDO) may also play a role on decadal time scales (D'Arrigo and Ummenhofer, 2015; Mohtadi et al., 2016; S. Sen Roy and N. Sen Roy, 2010). However, the tele-connections between Indian monsoonal precipitation and ENSO, IOD, and/or PDO are highly variable spatially and temporally (Kumar, 1999; Mishra et al., 2012; N. Sen Roy and Kaur, 2000; S. Sen Roy and N. Sen Roy, 2010), particularly in Myanmar, in part because local instrumental records are sparse and short.

While alternative records of hydroclimate variability exist in paleoclimate archives, such as speleothems (Sinha et al., 2007; 2015), tree rings (Buckley et al., 2010; Cook et al., 2010), ice cores (Kaspari et al., 2007), and lake sediments (Chawchai et al., 2015; Chu et al., 2011), Myanmar is neither the focus of these studies, nor can the information inferred from such records be extrapolated to Myanmar due to the considerable spatial variability of the ISM. To the best of our knowledge, only two published paleoclimate studies have been done in

Myanmar: one was derived from tree rings (D'Arrigo et al., 2011) and the other from lake sediments (Sun et al., 2016). However, these two studies are in conflict over the relationship between monsoonal precipitation and the IOD, which may have been caused by the spatial difference in rainfall distribution across the country, the interpretation of climate proxies and/or age uncertainties in the sediment record. Thus, Myanmar's paleoclimate remains largely unexplored.

To partially fill this gap, we present high-resolution trace element data, measured with laser ablation inductively coupled plasma mass spectrometric (LA-ICP-MS) techniques on a fast growing, annually laminated stalagmite, dated back to between ~450 A.D. and ~900 A.D. The sample was collected from Lin Noe Twin Cave in central Myanmar (Fig. 2.1). The stalagmite's trace element contents vary significantly, which we attribute to the local hydrology, controlled by monsoonal rainfall. Our trace element records capture both low- and high-frequency hydroclimate variability in central Myanmar. Our findings thus demonstrate that speleothem trace elements can allow us to reconstruct hydroclimate variations in high-resolution. Additionally, the records reflect a period of frequent droughts in the ninth century, which may have had far-reaching consequences for ancient societies.

Figure 2. 1. The upper left map shows the Asian monsoon system (modified after Johnson. (2011)). The lower left plot shows monthly mean precipitation (blue bars) and temperature (red squares) at Taunggyi meteorological observatory (data from UNdata: <http://data.un.org/Explorer.aspx?d=CLINO>). Over 90% of the annual precipitation falls during the monsoon season (late April through early November). Right map shows the location of Lin Noe Twin Cave (LNT, red star) and Taunggyi City (red dot).

2.3. Cave setting and sample description

Our sample, named as LNT-03, is a stalagmite collected from Lin Noe Twin Cave (21°14'13.4" N 96°26'37.5" E) (Fig. 2.1 and 2.2), which is located on eastern side of the Shan Plateau, central Myanmar. The cave develops within partially dolomitized Permian-Triassic limestone formations (Oo et al., 2002). The present climate of the area is dominated by the ISM, with warm and humid summers and dry and temperate winters. The annual average precipitation is ~1,700 mm, with more than 90% falling during the monsoonal season from late April to early November (Fig. 2.1). The closest city, Taunggyi, is located ~70 km southeast. Its annual mean temperature is ~19°C (Fig. 2.1).

Lin Noe Twin Cave consists of an inner and an outer chamber. The outer chamber is small and narrow since its roof has collapsed; it also has a steep slope extending ~20 m downwards. At its end, a narrow vertical drop of ~15 m leads to a large inner chamber, which is poorly ventilated, and where the relative humidity reaches ~100%. A broken stalagmite, LNT-03 was collected from the inner chamber. The sample is in a columnar shape, ~15 cm long and ~7 to 10 cm wide. After the sample was halved, no depositional hiatus was observed, which suggests an uninterrupted growth under a stable water drip. The sample consists mainly of transparent to light yellow, densely crystallized calcite with regular banding texture. The banding contains transparent and milky-white calcite couplets (Fig. 2.2). The banding texture, coupled with

seasonality of rainfall, strongly suggests that the lamination is annual, which is further supported by the ^{230}Th ages, band counting, and trace element variations. Additionally, five porous thin layers are visible at ~10, 22, 42, 75, and 89 mm from its top.

Figure 2. 2. a). Polished fresh surface of speleothem LNT-03 shows clear annual banding. Top and bottom ages are marked. b). A close-up image of the annual banding with visible transparent and milky-white calcite couplets.

2.4. Analytical Methods

2.4.1. ^{230}Th dating

LNT-03 was halved and polished. Seven subsamples ~ 100 mg were extracted by milling along growth horizons of the polished surface with a hand-held carbide dental drill. The subsamples were processed following the chemical protocol similar to Edwards et al. (1987) and Cheng et al. (2000) in a class-1,000 chemistry cleanroom at the Earth Observatory of Singapore (EOS), Nanyang Technological University (NTU), in order to separate and purify uranium and thorium. The uranium and thorium isotopes were analyzed on a Thermo-Finnigan Neptune Plus multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS) in the newly established Environmental Geochemistry Laboratory at the EOS, NTU. The analytical procedures were similar to those described in detail by Shen et al. (2012) and Cheng et al. (2013). All of the uranium and thorium isotopes, except ^{238}U and ^{232}Th , were measured with a peak-jumping mode on a secondary electron multiplier (SEM) equipped with a retarding potential quadrupole lens (RPQ) to improve abundance sensitivity. The major isotopes ^{238}U and ^{232}Th were measured on Faraday cups.

2.4.2. Trace element analysis

LNT-03 was further cut to a cuboid shape. In order for the sample to fit inside the laser ablation chamber, LNT-03 was halved diagonally across the 7.5-cm porous layer. The porous layer was later used as the marker to compile trace element data from the two halves. Trace elements were determined by a Resonetics M-50-LR 193 nm excimer laser coupled with an iCAP-Q Quadrupole ICP-MS housed at the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences, Wuhan, China, in the following manner: The polished stalagmite sections were fixed to the sample holder together with NIST-612 glass standards. The operation conditions of the instrument employed in this study followed Zhu et al. (2016). Five isotopes ($^{24}\text{Mg}^+$, $^{44}\text{Ca}^+$, $^{88}\text{Sr}^+$, $^{137}\text{Ba}^+$, and $^{238}\text{U}^+$) were analyzed, with dwell time for each mass set at 10 ms to enhance the number of sweeps and decrease the sweep time for trace elements. LNT-03 was first line-scanned at a speed of 200 $\mu\text{m/s}$ with a spot diameter of 150 μm for further cleaning of its surface and was then line-scanned at a speed of 20 $\mu\text{m/s}$ with a spot diameter of 60 μm for raster analysis. NIST-612 glass standards were analyzed in duplicates at the start and end of each line raster. Prior to each analysis, the first 100 s were used for background measurements. The total analysis time was approximately 2.5 h. Off-line analysis was performed using the software *Iolite* (Hellstrom et al., 2008; Paton et al., 2011; Woodhead et al., 2007), an open source add-in for *Igor Pro* (Wave Metrics) developed by the Melbourne Isotope Research Group at

the University of Melbourne. We adopted a semi-quantitative strategy to calibrate the trace elements of LNT-03 using the NIST-612 standard, and we applied Ca as the internal standard element. The precision of Ca calibrated element (i.e. Mg/Ca, Sr/Ca, Ba/Ca, and U/Ca) is $\sim 4\%$.

2.5. Results

To examine whether LNT-03's visual banding and trace element variations are annual and to achieve an optimal chronology, we have compared ^{230}Th ages with the number of visual bandings, and the number of high frequency trace element oscillations. While the accuracy of LNT-03's chronology is limited by ^{230}Th age uncertainties, the precision can be substantially improved by quantifying the positions of annual trace element peaks.

2.5.1. ^{230}Th ages

Seven ^{230}Th dates were obtained from the sample. U and Th isotopic compositions and ^{230}Th dates are in Table 2.1. The measured ^{238}U concentrations are on the order of 100 ppb. ^{232}Th concentrations are on the order of 1 ppb. ^{230}Th ages were calculated assuming an initial $^{230}\text{Th}/^{232}\text{Th}$ atomic ratio of $4.4 \pm 2.2 \times 10^{-6}$, the values for a material at secular equilibrium with the bulk Earth $^{232}\text{Th}/^{238}\text{U}$ value of 3.8. The relative errors (2σ) in ages before initial correction are about 1%. All the ages are in a stratigraphic order within errors. Unfortunately, the relative errors in ages after initial correction increased significantly, due to generally high ^{232}Th contents and low $^{230}\text{Th}/^{232}\text{Th}$ atomic ratios. To better constrain the LNT-03 chronology, a StalAge model (Scholz and Hoffmann, 2011) was applied (Fig. 2.3). While the precision after the application remains not ideal, the chronology indicates that

the sample grew continuously for about 450 years from ~450 A.D. to ~900 A.D, at a relatively stable growth rate.

2.5.2. Layer counting

To verify the accuracy of the ^{230}Th ages, we counted the annual banding of LNT-03 (Fig. 2.2). The annual banding was counted from a full-resolution scan photograph of the sample independently by two researchers, who agreed on $\sim 447 \pm 24$ annual bands. This number is in agreement with the ^{230}Th ages. Given that the length of the sample is ~ 152 mm and no hiatus is observed, the mean growth rate of LNT-03 is calculated to be $\sim 340 \mu\text{m yr}^{-1}$.

Figure 2. 3. StalAge model (Scholz and Hoffmann, 2011) for LNT-03 suggests that the stalagmite grew steadily and continuously for about 450 years from ~450 AD to ~900 AD.

Table 2. 1. ^{230}Th dating results, error in 2σ (95% in confidence).

Sample ID	^{238}U (ppb)	^{232}Th (ppt)	$^{230}\text{Th} / ^{232}\text{Th}$ (atomic $\times 10^{-6}$)	$\delta^{234}\text{U}^*$ (measured)	$^{230}\text{Th} / ^{238}\text{U}$ (activity)	^{230}Th Age (yr) (uncorrected)	^{230}Th Age (yr) (corrected)	$\delta^{234}\text{U}_{\text{initial}}^{**}$ (corrected)	^{230}Th Age (yr BP)*** (corrected)	^{230}Th Age (yr AD) (corrected)
LNT03-3mm	84 $\pm$ 0.1	486 $\pm$ 10	52 $\pm$ 1	520.8 $\pm$ 1.6	0.0184 $\pm$ 0.0003	1330 $\pm$ 19	1219 $\pm$ 81	523 $\pm$ 2	1152 $\pm$ 81	798 $\pm$ 81
LNT03-5mm	94 $\pm$ 0.1	1759 $\pm$ 36	18 $\pm$ 1	537.0 $\pm$ 2.0	0.0204 $\pm$ 0.0004	1458 $\pm$ 27	1102 $\pm$ 253	539 $\pm$ 2	1038 $\pm$ 253	912 $\pm$ 253
LNT03-39mm	89 $\pm$ 0.1	594 $\pm$ 12	52 $\pm$ 1	604.6 $\pm$ 1.7	0.0209 $\pm$ 0.0004	1431 $\pm$ 30	1311 $\pm$ 90	607 $\pm$ 2	1244 $\pm$ 90	706 $\pm$ 90
LNT03-82mm	134 $\pm$ 0.2	1425 $\pm$ 29	34 $\pm$ 1	541.8 $\pm$ 2.1	0.0219 $\pm$ 0.0002	1561 $\pm$ 15	1360 $\pm$ 143	544 $\pm$ 2	1293 $\pm$ 143	657 $\pm$ 143
LNT03-121mm	116 $\pm$ 0.2	966 $\pm$ 19	48 $\pm$ 1	617.1 $\pm$ 2.2	0.0243 $\pm$ 0.0002	1652 $\pm$ 17	1502 $\pm$ 107	620 $\pm$ 2	1435 $\pm$ 107	515 $\pm$ 107
LNT03-142mm	120 $\pm$ 0.2	514 $\pm$ 10	92 $\pm$ 2	587.9 $\pm$ 2.2	0.0238 $\pm$ 0.0002	1649 $\pm$ 17	1571 $\pm$ 58	591 $\pm$ 2	1504 $\pm$ 58	446 $\pm$ 58
LNT03-152mm	127 $\pm$ 0.2	1247 $\pm$ 25	41 $\pm$ 1	572.0 $\pm$ 2.0	0.0246 $\pm$ 0.0003	1720 $\pm$ 23	1538 $\pm$ 131	574 $\pm$ 2	1474 $\pm$ 131	476 $\pm$ 131

Sample ID is given by sample name and sub-sampling depth from its top in mm.

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).

$^*\delta^{234}\text{U} = ([^{234}\text{U}/^{238}\text{U}]_{\text{activity}} - 1) \times 1000 \%$. $^{**}\delta^{234}\text{U}_{\text{initial}}$ was calculated based on ^{230}Th age (T), i.e., $\delta^{234}\text{U}_{\text{initial}} = \delta^{234}\text{U}_{\text{measured}} \times e^{\lambda_{234} \times T}$.

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

Corrected ^{230}Th ages assume the initial $^{230}\text{Th}/^{232}\text{Th}$ 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 $^{232}\text{Th}/^{238}\text{U}$ 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.

2.5.3. Trace element analysis

The trace element records of LNT-03 are composed of over twenty thousand independent data points for each of the elements, resulting in a time-series with a sub-annual resolution. The data points are reported as ratios (Mg/Ca, Sr/Ca, Ba/Ca, and U/Ca; hereinafter referred to as Mg, Sr, Ba, and U). To further reduce noise while preserving the sub-annual resolution, we present the original data with 10-point running means. The results of the trace element analyses are plotted against distance along the analysis track in Fig. 2.4. Summary descriptive statistics for datasets are reported in Table 2.2. The U measured by laser scanning is comparable to the results from solution-based measurements (Tables 2.1 and 2.2), further supporting the accuracy of the method.

The Mg, Sr, Ba, and U fluctuate in both low- and high-frequency (annual timescale) (Figure 2.5). The figure shows ~30 cycles superimposed on the long-term trends in the plots of element ratios, the number of cycles being broadly consistent with the number of annual bands calculated based on the averaged growth rate ($\sim 340 \mu\text{m yr}^{-1}$). Elements are plotted in pairs to highlight their correlations.

In order to quantify the elements' correlations in both low- and high-frequency bands, we used a 500-point running mean (chosen here for a decadal trend, illustrated by the thick red line in Fig. 2.4) to represent the low-frequency trend,

and then de-trended the data by subtracting the 500-point running mean to enhance the high-frequency trend. A correlation matrix of the low-frequency (500-point running means) and high-frequency (detrended) records for the whole length of LNT-03 is presented in Table 2.3. In the high-frequency band, all the four elements show almost the same periodicity, but with Mg negatively correlated with Sr, Ba, and U (Fig. 2.4 and 2.5). These relationships are further confirmed by the correlation analysis (Table 2.3), which suggests that in the high-frequency band Mg and the other three elements are weakly but significantly negatively correlated. The negative correlations are even stronger in the low-frequency band (Table 2.3), in part because the high-frequency noises of water-rock interactions during calcite deposition could be smoothed out in the decadal-scale running mean. At both the low- and high-frequency bands, Sr and Ba are very strongly correlated with similar R-values: 0.91 (low frequency) and 0.84 (high frequency). This indicates that the processes controlling the distribution of the two elements were very similar and did not vary significantly on different time scales.

Six very large peaks in the Mg profiles occur at approximately 1 mm, 10 mm, 22 mm, 42 mm, 75mm, and 89 mm, with corresponding troughs in Sr, Ba, and U profiles (Fig. 2.3). The anomalous peak at the very top of the sample (~1 mm) is probably related to the muddy cover of the sample top, and the correlations in this layer between Mg and the other three elements are in fact not clear. The other five peaks are associated with porous layers. Thus, the

elemental ratios in these five layers may not be accurately quantifiable. However, we argue that the overall trends of the five peaks are still robust for the following reasons: First, the anti-correlation between Mg and Sr (also with Ba and U) is well preserved in all the five layers as Mg peaks/Sr (Ba and U) troughs, which suggests that the signals are not artificial; and second, the Mg (Sr) contents adjacent to the porous layers, particularly at the peaks of ~22 mm and 42 mm depths, are already much higher (lower) than the averaged ratios.

Figure 2. 4. Changes in Mg/Ca, Sr/Ca, Ba/Ca, and U/Ca along the entire 15 cm of the sample length. The left side represents the youngest end of the sample. The gray lines represent the original data after outliers were eliminated. The black and thick red lines are 10-point and 500-point running means of the data, respectively.

Table 2. 2. Descriptive statistics for the LNT-03 LA-ICP-MS dataset. Mg/Ca are expressed as mmol/mol and the rest are $\mu\text{mol/mol}$.

	Mean	S.D. (1σ)	Maximum	Minimum
Mg/Ca	15.7	3.0	36.7	6.9
Sr/Ca	21.6	3.4	43.2	6.9
Ba/Ca	2.8	0.9	7.4	0.7
U/Ca	0.059	0.017	0.109	0.008

Table 2. 3. Correlation coefficients (r) for the low frequency record (thick red line in Fig. 2.4) and the detrended high frequency record (Fig. 2.5) (all p-values $\ll 0.05$).

	Mg : Sr	Mg : Ba	Mg : U	Sr : Ba	Sr : U	Ba : U
High frequency	-0.34	-0.13	-0.11	0.84	0.13	-0.09
Low frequency	-0.48	-0.35	-0.51	0.91	0.35	0.08

Figure 2. 5. Comparisons of changes in Mg/Ca, Sr/Ca, Ba/Ca, and U/Ca, zoomed in on 10 mm of the sample length (here between 120 and 130 mm) for better visibility. All four elements show clear annual cyclicity, with roughly ~ 30 cycles. Only every fifth cycle in the Mg/Ca profile is marked. Mg is inversely correlated with Sr, Ba, and U. Sr and Ba show a strong positive correlation.

Figure 2.5. (continued)

Figure 2.5. (continued)

2.5.4. Trace element oscillations

To further confirm that the high-frequency (i.e., 10-point running mean) fluctuations are annual, and to improve the chronology precision, we counted the number of peaks in the Mg record. To define a peak, two mathematical parameters were needed: a minimum peak distance (i.e., a minimum horizontal distance between two discrete peaks) and a minimum topographic peak prominence (i.e., the smaller one of two vertical distances between a maximum and two adjacent minima to isolate a peak). We designated the two parameters based on the following considerations. First, the farthest distance between two possible peaks is around 145 points, indicating that the fastest annual growth rate was about three times that of the mean annual growth rate ($\sim 340 \mu\text{m yr}^{-1}$ or ~ 50 points). As suggested by the ^{230}Th StalAge model (Fig 2.3), LNT-03's growth rate was stable. Thus, its annual variations would unlikely have been larger than one order of magnitude. Accordingly, we assumed that the minimum annual growth rate was about one tenth of the possible maximum annual growth rate, arriving at 15 points in distance. Second, the prominence of a peak has to be larger than 0.208 mmol/mol (or approximately 50 ppm changes in Mg concentrations) to best fit the visual peaks agreed on by two observers. Thus, given the two parameters for principal peak identification, the Mg record yields 454 peaks, which is in agreement with the results from band counting and ^{230}Th ages. Consistent numbers of peaks are reached for other elements

after applying similar criteria. Therefore, the high frequency oscillations of the four trace elements in LNT-03 are annual in nature.

Since the annual Mg fluctuations are the most prominent, we used the positions of each Mg peak as the boundary of each year. Correspondingly, the distance between neighboring Mg peaks was taken to represent LNT-03's annual growth rate. We further calculated the annual Mg/Ca, by averaging all the Mg/Ca ratios between neighboring Mg peaks. The annual Sr/Ca, Ba/Ca, and U/Ca ratios were also calculated by averaging all the ratios between Mg peaks. The absolute chronology of the annual mean data was anchored to the bottom ^{230}Th StalAge, which has better precisions compared to the top of the sample. Therefore, we have attained a 453-year annually resolved trace element record from LNT-03 (Fig. 2.6).

Figure 2. 6. Annual means of Mg, Sr, Ba, and U, and annual growth rates of LNT-03 calculated from Mg peaks. From ~ 750 AD to ~ 900 AD, Mg fluctuations indicate frequent droughts in Myanmar.

2.6. Discussion

2.6.1. Variations of alkaline earth elements

An important observation in this study is the overall negative correlation between Mg and Sr (Ba) on all the timescales covered by our sample. Such negative correlation becomes even more significant on longer timescales after averaging out seasonal variations (Table 2.3). The anti-correlation clearly shows that Mg was differentiated from Sr (Ba) by a dominant geochemical process.

Many geochemical processes are known to affect Mg and Sr (Ba) in speleothems, including change in trace metal sources, temperature effects on Mg partitioning, growth rate effects on the Sr partitioning, prior calcite precipitation (PCP), or incongruent calcite dissolution (ICD) (Belli et al., 2017; Desmarchelier et al., 2006; Fairchild et al., 2000; Fairchild and Treble, 2009; Huang et al., 2001; Johnson et al., 2006; Sinclair et al., 2012; Treble et al., 2003; 2015; Tremaine and Froelich, 2013). However, these processes either significantly influence the distribution of a single element or drive an in-phased variation between Mg and Sr (Ba). The only known process that can result in the anti-phased trends between Mg and Sr (Ba) is water residence time (Fairchild et al., 2000; Roberts et al., 1998; 1999).

Water residence time in the karst system is controlled by effective precipitation (Fairchild et al., 2000; Roberts et al., 1998): in wet conditions, water recharge rate is high and water residence time is shortened. The reverse applies to dry conditions. In wet conditions, time for water-rock reaction decreases as water residence time is shortened, and hence proportionally less dolomite from bedrock is dissolved compared to calcite, since dolomite has a much slower dissolution rate compared to calcite (Fairchild et al., 2000). As dolomite has much higher Mg and generally low Sr (Ba) concentrations relative to calcite, the low proportion of dolomite dissolution will lead to relatively low Mg and high Sr (Ba) in karst water, and subsequently low Mg and high Sr (Ba) in speleothems. In dry conditions, a higher proportion of dolomite is dissolved, leading to relatively high Mg and low Sr (Ba) in cave deposits.

Another explanation for the negative correlation between Mg and Sr has been proposed by Treble et al. (2003). Their speleothem from western Australia comes from a cave in which dolomite is absent from the bedrock. They argue that Mg is affected by PCP under lower water recharge, and hence a dry climate results in a higher Mg/Ca ratio in speleothem. Moreover, Mg is not sensitive to the growth rate of calcite whereas Sr is (Huang and Fairchild, 2001). Treble et al. (2003) therefore argue that instead of the effect of PCP, Sr is mostly regulated by speleothem growth rate, showing a higher content when the growth rate is faster during the wet seasons and vice versa.

In our case, the mechanism proposed by Treble et al. (2003) is not applicable, because: (1) PCP is frequently used to explain the well-documented co-variations of alkaline-earth elements (i.e., Mg, Sr, Ba) (e.g., Bernal et al., 2016; Griffiths et al., 2010; Sinclair et al., 2012; Tremaine and Froelich, 2013), rather than the less common anti-correlation of Mg and Sr (Ba); (2) dolomite does widely exist in the bedrock of LNT cave (Oo et al., 2002); and (3) Sr (Ba) in LNT-03 does not show an obvious correlation with stalagmite growth rates calculated from Mg peaks (Fig. 2.7).

In summary, we conclude that Mg and Sr (Ba) are primarily controlled by varying water residence time in the karst system, which reflects effective precipitation. Given that over 90% of local precipitation is brought in by the ISM, we further interpret the Mg and Sr (Ba) as proxies of ISM precipitation, with low Mg and high Sr (Ba) indicating stronger ISM precipitation and vice versa.

Figure 2. 7. The low R^2 value indicates that annual Sr/Ca mean does not correlate with annual growth rate of LNT-03.

2.6.2. Variations of U in the sample

U can be an effective paleo-hydrological indicator, similarly to Sr and Ba, since its partition coefficient is lower than unity in calcite (Fairchild and Treble, 2009). However, U as a proxy in speleothems has not been extensively studied, largely due to our poor understanding of the sources and geochemical processes that control U distributions. U as a potential climate proxy has been favoured, for example, by Treble et al. (2003), who observed that U, similar to Sr, tends to decline with reduced rainfall. Our observation, however, has revealed different relationships between U and alkaline earth elements at low and high frequency bands, suggesting that U affinity is controlled by rather complex geochemical processes on different timescales. On interannual timescales, U is broadly similar to Sr (Ba) and is anti-correlated with Mg, indicating that U variation is likely also controlled by varying water residence time. On seasonal timescales, U is weakly correlated with alkaline earth elements, probably because U is sensitive to temperature, pH of karst water, and/or the availability of P (Desmarchelier et al., 2006; Jamieson et al., 2016; Treble et al., 2005). Moreover, the solubility of U in epikarst water is also controlled by its valence state and complexation with carbonate ions and organic compounds (Treble et al., 2003). To further evaluate how the seasonal variability of U is controlled by the aforementioned factors, drip water chemistry needs to be monitored, particularly regarding U concentration, speciation, and its correlation with other alkaline earth elements.

2.6.3. Low-frequency (interannual to decadal) climate variability

The aforementioned variations in trace elements in LNT-03, in particular the alkaline earth elements, largely reflect the change in ISM precipitation. Therefore, the annual mean values of Mg, Sr, and Ba covering ~450 A.D. to ~900 A.D allow us to assess ISM variability under modern climate conditions, yet without the influence of a strong anthropogenic forcing.

The Mg record shows generally low values of ~17 mmol/mol and weak variability during the first ~300 years (Figure 2.6). From ~750 to ~900 A.D., Mg variability dramatically increased and Mg/Ca ratio reached up to ~25 mmol/mol. While the variability of Sr, Ba, and U is less prominent due to their low concentrations, the Mg record clearly suggests that over the course of ~450 years, the monsoonal rainfall in the region had shifted from a relatively wet state into a frequently dry regime (Figure 2.6). Similar shifts have also been shown in the East Asian Summer Monsoon domain, where they have been linked to the fall of the Tang Dynasty in China (Zhang et al., 2008).

Interestingly, around the same time of this weakened monsoon period, the Pyu City States in Myanmar also started to decline (Stargardt, 2016). The Pyu Kingdom had flourished in Myanmar for over 1,000 years since about the second century BC. Pyu people built remarkable cities along the upper and central Irrawaddy River, employed a sophisticated irrigation system, were able

to process iron, produced domestic ceramics, developed complex rituals, were literate, and practiced Buddhism (Stargardt, 2016). Around the ninth century, however, they abandoned their cities mysteriously. While at that time, according to the sketchy Chinese Tang chronicles, the Pyu people were under a military threat from a neighboring kingdom in the north, the decline of Pyu society due to climate change remains an intriguing speculation, especially considering that monsoon climate is one of the major drivers of the waxing and waning of ancient civilizations in the MSEA (Buckley et al., 2010; 2014; Cook et al., 2010).

2.6.4. Driving mechanism of the interannual to decadal variabilities of ISM

To investigate the interannual to decadal cyclicity of the ISM rainfall inferred from trace elements, and to evaluate its possible driving mechanisms, we performed a power spectrum analysis on the time series of Mg, Sr, and Ba using the software REDFIT (Schulz and Mudelsee, 2002). Several low-frequency (inter-annual to decadal) quasi-periodicities (2.0-3.3, 3.5-4.5, 5-5.5, 7.7, and 11.1 years) at a confidence level greater than 95% (Fig. 2.8) were detected from the Mg time series. Similar low-frequency quasi-periodicities were also detected at a confidence level greater than 95% in the Sr time series (2.5-3.2, 4.0, 4.4, 6.1, 6.8-7.0, and 10.1 years), and the Ba time series (2.2-3.2, 6.1, and 6.8-7.0 years) (Fig. 2.8).

The inter-annual variability (2 to 7 years) in the Mg, Sr, and Ba time series of LNT-03 closely resembles the quasi-periodicity of modern ENSO, indicating a causal link. This link can be supported by instrumental data from Myanmar, which indicate a strong correlation between rainfall and ENSO. El Niño events typically cause less rainfall in the region, whereas La Niña events generally bring in more rains (N. Sen Roy and Kaur, 2000; S. Sen Roy and N. Sen Roy, 2010; Sein et al., 2015; Sein and Zhi, 2016; Shrivastava et al., 2016). ENSO-like variability has also been observed in tree ring records and lacustrine sediment data from Myanmar (D'Arrigo et al., 2011; Sun et al., 2016) and other

regions under the ISM influence (Buckley et al., 2010; Liu et al., 2014; Xu et al., 2013).

The interannual variability in our sample may also be linked to the IOD, which similarly exhibits quasi-periodicities of seven years or less (Abram et al., 2008), and which has a strong influence over the Indian Ocean and its neighboring continents (Abram et al., 2008; Saji et al., 1999; Webster et al., 1999). However, the mechanism of how the IOD modulates the rainfall in the ISM region, particularly in Myanmar, is not well understood. Generally, during a positive IOD phase, the SST gradient across the tropical Indian Ocean enhances the southerly monsoonal wind over the northern Indian Ocean, which brings more rainfall into India (Abram et al., 2008; Saji et al., 1999; Webster et al., 1999). A recent lacustrine study in central Myanmar also suggested that the IOD could drive the interannual-to-decadal variability of regional rainfall (Sun et al., 2016). However, the instrumental records from Myanmar only indicate a weak relationship between monsoonal rainfall and the IOD index (Sein et al., 2015). A weak relationship between the rainfall in Myanmar and the IOD mode is also suggested by tree ring records (D'Arrigo et al., 2011). Nevertheless, the coherence of periodicities in our trace element records and the IOD favors a connection between monsoonal rainfall variability and the IOD. The discrepancies in instrumental and proxy records may suggest a transient correlation between the monsoon and the IOD, or may be explained by their limited lengths.

In addition to the inter-annual variability in the range of 2 to 7 years, both the Mg and Sr time series also exhibit spectral peaks at ~11 years (Fig. 2.8). The Ba series exhibits a spectral peak at ~11 years at confidence levels greater than 90% (Fig. 2.8). Such decadal variability is coherent with the 11-year solar cycle, which has been demonstrated to significantly influence the monsoon activity despite its relatively small fluctuations (Kodera, 2004; Meehl et al., 2009). Similar 11-year solar cycles have also been detected in other high-resolution proxy-based paleo-climate records from tropical regions (Burns et al., 2002; Kaspari et al., 2007).

Our speleothem trace element records therefore reveal regional to large-scale climate variability in Myanmar. Given the considerable spatial complexity of ISM precipitation, more high-resolution paleo-climate data covering a wider geographical range are needed to evaluate regional rainfall variability and its dynamic links to atmospheric and oceanic processes.

Figure 2. 8. Spectral analysis results of the Mg/Ca, Sr/Ca, and Ba/Ca time series. The spectrums were obtained using the program REDFIT (Schulz and Mudelsee, 2002).

2.7. Conclusions

We present trace element records that span ~450 years in recent history from an annually laminated stalagmite collected from central Myanmar. The strong high frequency oscillations of Mg/Ca, Sr/Ca, Ba/Ca, and U/Ca are annual, as the Mg peak counting matches speleothem growth layer counting as well as ^{230}Th ages. The clear anti-correlation between Mg and Sr (Ba) on intra-annual to decadal timescales suggests that variations of the three elements are essentially caused by changes of ISM rainfall. The long-term variability of U is also likely to be controlled by local hydrology. The cause of the observed seasonal variability of U is yet not clear, and needs to be further investigated by monitoring of drip water chemistry.

By analyzing variations of the annual mean Mg, we show that local precipitation associated with the ISM is forced by solar cycles, and strongly influenced by ENSO and possibly the IOD on interannual to decadal time scales. The Mg record also suggests that monsoonal rainfall in the region shifted from a relatively wet state into a frequently dry regime around the ninth century, which could have disturbed the ancient Pyu society that flourished in central Myanmar. Given the considerable spatial complexity of ISM precipitation, more high-resolution paleo-climate data covering a wider geographical range are needed to evaluate regional rainfall variability and its dynamic links to atmospheric and oceanic processes.

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Chapter 3

Indian Summer Monsoon during the Last Glacial

Maximum

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Contributions

Liu Guangxin was the primary author. Liu Guangxin and Wang Xianfeng contributed to the design of this study. Liu Guangxin, Wang Xianfeng, Sakonvan Chawchai, Chiang Hong-Wei, Lin Thu Aung, and Phyo Maung Maung collected and prepared the samples, which were subsequently analyzed by Liu Guangxin and interpreted by Liu Guangxin, Wang Xianfeng, Sakonvan Chawchai, and He Shaoneng. Liu Guangxin wrote this paper.

Liu Guangxin: 55%

Wang Xianfeng: 15%

Sakonvan Chawchai: 10%

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3.1. Abstract

In this chapter, we present a speleothem $\delta^{18}\text{O}$ record obtained from southern Myanmar and western Thailand (SMWT), where the local rainfall is dominantly brought in by the Indian Summer Monsoon (ISM). Our record covers two characteristic time intervals, the Last Glacial Maximum (LGM) and the last 600 years, and shows a large, positive shift in $\delta^{18}\text{O}$ ($> 3.5\%$) during the LGM relative to today, which is more than twice that of the Chinese speleothem records. Similar shifts are recorded in speleothems from India and the Middle East, where the ISM also dominates the local hydroclimate. Based on the large $\delta^{18}\text{O}$ shift, we calculated that during the LGM, the regional ISM precipitation was only about a half amount of today's. Given that the two time intervals share a similar Northern Hemisphere summer insolation intensity, the change in the ISM was likely forced by glacial boundary conditions. The discrepancy between our record and the Chinese records reveals that the sensitivity of speleothem $\delta^{18}\text{O}$ to insolation and glacial forcing varies spatially. The muted glacial-interglacial difference in the Chinese speleothem records must have resulted from a counteraction of a larger (smaller) isotope fractionation of precipitation $\delta^{18}\text{O}$ during glacial (interglacial) times.

3.2. Introduction

The Asian Monsoon (AM) brings rainfall to the most densely populated regions in the world, and its future variation will impact the livelihood of billions of people. To predict the AM changes in response to global warming therefore is a major challenge for climate scientists. The question has been addressed retrospectively, that is, by examining past monsoon variabilities, particularly during a cold state of the Earth in the past, such as the Last Glacial Maximum (LGM, ~19-22 ka), when the atmospheric CO₂ level was ~100 ppmv lower, and the average temperature of the Earth was over 5°C colder compared to the recent time before the Industrial Revolution (thereafter pre-industrial period or today) (Shakun et al., 2012). The changes of monsoon between then and today have been used as an analogue for the future (e.g., Sun et al. (2015)).

Our knowledge of the past AM variabilities has improved significantly over the past two decades with the contribution from studies on speleothem $\delta^{18}\text{O}$ proxy. Tropical speleothem $\delta^{18}\text{O}$ has been demonstrated to faithfully record the $\delta^{18}\text{O}$ values of the precipitation that recharges the cave system (Cheng et al., 2012), with a very minor influence from cave temperature ($\sim -0.23\text{‰}/^{\circ}\text{C}$ (Kim and O'Neil, 1997; O'Neil et al., 1969)). The speleothem $\delta^{18}\text{O}$ records obtained from China were originally linked to East Asia Summer Monsoon (EASM) intensity, based on the fact that $\delta^{18}\text{O}$ values in today's precipitation associated with the EASM are relatively low, with more negative $\delta^{18}\text{O}$ values corresponding to a

higher proportion of EASM precipitation in annual total, hence stronger monsoon intensity (Wang et al., 2001; 2008). More recent studies have shown that $\delta^{18}\text{O}$ records from caves thousands of kilometres apart across southern China are similar on decadal to orbital timescales (Cheng et al., 2012; Wang et al., 2001; 2008; Yuan et al., 2004). These studies demonstrated that speleothem $\delta^{18}\text{O}$ can effectively record regional hydroclimate changes, and register a monsoonal rainfall amount integrated from tropical ocean sources to Chinese cave sites (Yuan et al., 2004). Naturally, given its regional climatic significance, Chinese speleothem $\delta^{18}\text{O}$ was further interpreted as a proxy of the AM (Cheng et al., 2009; 2016; Wang et al., 2005; Yuan et al., 2004). The studies indicate that past variations of the AM were dominantly forced by the Northern Hemisphere summer insolation during the recent glacial cycles, with a few exceptions associated with abrupt climate events, for example, Heinrich events and Dansgaard-Oeschger (D-O) events, which originated in the North Atlantic region (Cheng et al., 2016; Wang et al., 2008).

However, speleothem $\delta^{18}\text{O}$ -based AM reconstructions (Wang et al., 2008) are inconsistent with most other AM records retrieved from marine (e.g. Kudrass et al. (2001); Prell and Kutzbach (1987)) and terrestrial sediments (e.g., An, (2000); An et al. (2011); Sun et al. (2015)). The former indicate an LGM-monsoon intensity very similar to today's, whereas the latter suggest a significantly weakened AM during the LGM. This discrepancy probably results from our incomplete interpretation of the Chinese speleothem $\delta^{18}\text{O}$ proxy

(Caley et al., 2014; Chiang et al., 2015; Clemens et al., 2010; Dayem et al., 2010; Johnson and Ingram, 2004; LeGrande and Schmidt, 2009; Liu et al., 2014; Maher, 2008; Pausata et al., 2011; Sun et al., 2015). A key question, then, is: what is the dominant process controlling speleothem $\delta^{18}\text{O}$?

In theory, the precipitation $\delta^{18}\text{O}$ inferred from speleothem $\delta^{18}\text{O}$ from southern China can be controlled by four processes: changes in the Indian Ocean moisture source; rainout effect and moisture recycling during moisture transport; integrated rainfall amount; and mixture of locally sourced moisture. Instead of emphasizing integrated rainfall amount (hence monsoon intensity), studies that challenge the monsoon-intensity-oriented interpretation turn to the other three aspects. For example, it was pointed out that over 20% of the annual precipitation in southern China occurs in winter, and therefore the contribution of precipitation in non-monsoon seasons to the annual precipitation-weighted $\delta^{18}\text{O}$ signals needs to be re-evaluated (Caley et al., 2014; Clemens et al., 2010; Wang et al., 2001). Another body of research emphasizes the source effect (LeGrande and Schmidt, 2009; Maher, 2008; Pausata et al., 2011). Instead of local or regional precipitation amount, the high $\delta^{18}\text{O}$ values in Chinese speleothems are interpreted to reflect an ^{18}O enrichment in the major moisture source over the Indian basin or a higher proportion of monsoon moisture from the western Pacific, which is characterized by less isotopically depleted values. Recently, by studying the speleothems in South America, Wang et al. (2017) also realized that moisture recycling through plant transpiration during transport

is important in regulating the downwind rainfall $\delta^{18}\text{O}$ values. A similar mechanism could potentially be important for speleothem $\delta^{18}\text{O}$ from southern China, as this region is also heavily vegetated.

We suggest that the discrepancies in interpreting speleothem $\delta^{18}\text{O}$ may be reconciled by comparing records from southern China with those from cave sites strategically located in the upwind Indian Summer Monsoon (ISM) domain, for example, the coastal region of the Bay of Bengal (BoB). Due to their proximity to the Indian Ocean, the precipitation at these locations is almost entirely dominated by the ocean source, with minimum water recycling. The $\delta^{18}\text{O}$ of the precipitation, therefore, carries a nearly pure signal of monsoon rainfall. Speleothem- $\delta^{18}\text{O}$ -based ISM records retrieved from this region will thus bear fewer uncertainties in interpreting $\delta^{18}\text{O}$ signals. Their comparison with the speleothem $\delta^{18}\text{O}$ records from southern China will help us evaluate the relative importance of different processes regulating the changes of $\delta^{18}\text{O}$.

Speleothem $\delta^{18}\text{O}$ records from such locations, however, are still very limited. In this chapter, we present such records developed from speleothems collected from southern Myanmar and western Thailand (SMWT) (Fig. 3.1). The speleothem $\delta^{18}\text{O}$ records cover two major time intervals, the LGM and the recent 600 years. In contrast to the Chinese speleothem $\delta^{18}\text{O}$ records, our SMWT records document dramatic changes in the strength of the ISM when

compared to the modern day. A spatial-temporal comparison between the newly established records and those from southern China can assess different factors controlling the speleothem $\delta^{18}\text{O}$ proxy.

Figure 3. 1. a). Locations of Phabaung Gu Cave (PHA), Padamya Gu Cave (PDM) and Hin Tum No.2 Cave (HT2) (red stars). The red dot represents the location of the closest major city, Yangon. Locations of Mawmluh cave (orange star) (Berkelhammer et al., 2013; Dutt et al., 2015) and its closest major city, Shillong (orange dot) in Northeast India, and core KL126 (blue star) (Kudrass et al., 2001) in the BoB are also shown for comparison. The black arrows represent the present-day averaged June-July-August (JJA) wind pattern (data from the National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010) (Kalnay et al., 1996). b). Monthly mean precipitation (blue bars) and temperature (red squares) at Taunggyi and Shillong meteorological observatory stations. As shown, the majority of the annual precipitation in the region occurs during the monsoon season (data from UNdata: <http://data.un.org/Explorer.aspx?d=CLINO>).

3.3. Cave setting and Materials

The SMWT record is based on four columnar stalagmites collected from three caves (PHA-02 from Phabaung Gu Cave (16°17.210' N, 97°54.092' E), PDM-02 from Padamya Gu Cave (16°49.725' N, 97°42.369' E), and HT2-1 and HT2-2 from Hin Tum No.2 Cave (15°01.272' N, 99°26.757'E)), located at both sides of the Myanmar-Thailand border, about 200 km SE from Yangon, Myanmar (Fig. 3.1). The local climate at the cave sites is approximately similar to Yangon, with an annual mean temperature of ~27.5°C and very minor annual seasonal fluctuations. Over 90% of the annual total precipitation (~2,000 mm) occurs between mid-May and late-October, indicating a dominance of summer monsoon rainfall at the cave sites (Fig. 3.1). Geographically, the local precipitation is mostly sourced from the BoB, and brought to the region by the ISM (Roy and Kaur, 2000). All the four caves have multiple chambers with limited air circulation and high relative humidity close to 100% in the inner chambers. To minimize the potential kinetic effect on carbonate stable isotopes, we only collected samples from the inner chambers. Moreover, in consideration of cave preservation, we almost exclusively collected stalagmites that were already broken. The only standing stalagmite we collected was PHA-02, with active dripping water. It has been used to evaluate the equilibrium conditions of carbonate precipitation. All the samples are composed of calcite.

3.4. Analytical Methods

All stalagmites were cut into halves along the growth axes, and their surfaces were then polished. A total of 25 subsamples were drilled along the growth axes for ^{230}Th dating. The detailed chemical procedures of U and Th separation and purification are described by Edwards et al. (1987) and Cheng et al. (2000). The analyses of stalagmite samples were performed mostly at the Environmental Geochemistry Laboratory (EGL), Earth Observatory of Singapore (EOS), Nanyang Technological University, Singapore. Before the EOS-EGL was set-up, the ages of the top and bottom subsamples of PHA-01 and PDM-02 had been screened at the High-Precision Mass Spectrometry and Environment Change Laboratory (HISPEC), Department of Geosciences, National Taiwan University, Taipei. The ^{230}Th dating techniques at the two laboratories were identical. All measurements were made on Thermo-Finnigan Neptune Plus multi-collector inductively coupled plasma mass spectrometers (MC-ICP-MS) using recently improved techniques (Cheng et al., 2013; Shen et al., 2012).

A total of 474 subsamples were drilled with dental burrs ranging in size from 0.3 to 0.7 mm perpendicularly to the extension axes of the stalagmite samples. The oxygen and carbon stable isotopic compositions ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) of the stalagmite samples were analyzed at two laboratories—the EOS-EGL, Nanyang Technological University, Singapore and the Speleothem Isotope Laboratory,

Nanjing Normal University, Nanjing, China. Results are reported in per mil (‰), relative to the Vienna PeeDee Belemnite (VPDB) standard. Stable isotope samples measured at the EOS-EGL were analyzed using an on-line carbonate preparation system (Gasbench II) interfaced with an isotope ratio mass spectrometer (Delta V). The long-term reproducibility is $\sim 0.1\text{‰}$ (1σ). The stable isotope measurements at Nanjing Normal University were performed on Thermo-Finnigan MAT-253 mass spectrometers fitted with a Kiel Device III. Duplicate measurements of NBS19 and in-house TTB1 standards show a long-term reproducibility of $\sim 0.1\text{‰}$ or better (1σ). The data obtained from these two laboratories are highly comparable.

3.5. Results

3.5.1. ^{230}Th ages

The U and Th isotopic compositions of the subsamples and their corresponding ^{230}Th dates are listed in Table 3.1. ^{238}U concentrations range from 50 to 400 ppb. ^{232}Th concentrations range from a few ppb to tens of ppb. ^{230}Th ages were calculated assuming an initial $^{230}\text{Th}/^{232}\text{Th}$ atomic ratio of $4.4 \pm 2.2 \times 10^{-6}$, the value for a material at secular equilibrium with the bulk Earth $^{232}\text{Th}/^{238}\text{U}$ value of 3.8. We consider that it is reasonable to adopt this value for initial ^{230}Th correction because the value is also comparable to the measured $^{230}\text{Th}/^{232}\text{Th}$ atomic ratio of the top subsample of stalagmite PHA-02, which is presumably modern. All the dates are in stratigraphic order within age uncertainties. However, a few ages have relatively large uncertainties due to high ^{232}Th contents. To improve chronology, the StalAge approach (Scholz and Hoffmann, 2011) was used to establish depth-age models for stalagmites PHA-02, PDM02, and HT2-1 (Fig. 3.2). For stalagmite HT2-2, the three dates at 37, 76, and 145 mm are all around 20,000 years before present (ka BP, with the present being 1950 AD), but with very large uncertainties. To reduce uncertainty, we used the weighted mean of the three dates as the age of the middle point (i.e., 86 mm), and then used this weighted mean age and the measured age at 216 mm to set up a linear extrapolated age model for HT2-2 (Fig. 3.2). The subsample depths, modeled ages, and oxygen isotope compositions are listed in Table 2.

Figure 3. 2. Plots of age versus depth for stalagmites HT2-1, HT2-2, PDM02, and PHA-02 derived from SMWT caves. All ages are reported in units of thousands of years before present (1950 AD) or ka BP. The age errors indicated in the plots are 2σ errors. The StalAge method (Scholz and Hoffmann, 2011) was used to establish the chronology for HT2-1, PDM-02, and PHA-02. The red lines indicate the 95% confidence limits calculated by a Monte Carlo simulation fitting. Weighted mean age and a linear extrapolation were used to set up the age model for HT2-2 (see the Results section for details).

Figure 3. 3. Scatter plot of $\delta^{18}\text{O}$ versus $\delta^{13}\text{C}$ for each stalagmite sample from the SMWT caves (see Fig. 3.1 for the location of the caves). The low correlations (as indicated by R^2 values) of the plotted results indicate that carbon and oxygen isotopic ratios are not highly correlated as would be expected under conditions of kinetic fractionation.

3.5.2. Isotopic equilibrium tests

The measured $\delta^{18}\text{O}$ values show large shifts, ranging from -8.5 to -2.5‰ over the profile (Fig. 3.3). Various non-climatic processes, such as kinetic isotope fractionation (McDermott, 2004), might contribute to the carbonate $\delta^{18}\text{O}$ variations. Therefore, before seeking a climatic interpretation of the stalagmite $\delta^{18}\text{O}$, it is important to confirm that the stalagmite calcite was deposited under isotopic equilibrium conditions. We first calculated the correlation coefficients between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ for each sample (Fig. 3.3). Their low R^2 values (<0.15) argue against the dominance of kinetic isotope fractionation during speleothem growth according to the traditional Hendy criteria (Hendy, 1971). Secondly, stable isotope analyses were performed twice on sample PDM-02, with different subsampling resolutions along two transects parallel to the growth axis. The $\delta^{18}\text{O}$ variations were very well reproduced (Fig. 3.4), which suggests that stable isotope variations along the same growth layer were very small, and therefore further argue against any kinetic fractionation during the speleothem growth (Hendy, 1971). Thirdly, the $\delta^{18}\text{O}$ values of the modern sample PHA-02 are consistent with the annual mean modern meteoric precipitation $\delta^{18}\text{O}$ values measured in Yangon (Araguás-Araguás et al., 1998), the closest major city that shares the same climatology as the cave locations, assuming an equilibrium calcite-water isotope fractionation at 27.5°C (O'Neil et al., 1969). A more robust equilibrium test involved replication of isotopic records across caves (Cheng et al., 2006; 2016; Dorale, 1998; Dykoski et al., 2005). Our speleothem

records from SMWT can be broadly correlated over their contemporaneous growth periods on millennial to orbital timescales with those from caves several thousand kilometers away in the Asian Monsoon region (Wang et al., 2001, 2005, 2008; Cai et al., 2015; Cheng et al., 2016). Similar correlations also exist between the SMWT record and those from Mawmluh (Berkelhammer et al., 2013; Dutt et al., 2015) and Bittoo caves (Kathayat et al., 2016) in northern India. With the above lines of evidence, we conclude that the speleothem samples collected in our study most likely grew under isotopic equilibrium conditions. The carbonate $\delta^{18}\text{O}$ is therefore mostly controlled by the $\delta^{18}\text{O}$ of the meteorological precipitation and cave temperature at the time of carbonate precipitation.

Figure 3. 4. Replications of stable isotope variation on sample PDM02. The low and high resolution measurements were performed at different locations of the growth layers of PDM02 in different laboratories. The low-resolution data were achieved in the EGL-EOS, while the high-resolution data were achieved in the Speleothem Isotope Laboratory, Nanjing Normal University.

3.6. Discussion

3.6.1. The climatic interpretation of speleothem $\delta^{18}\text{O}$

The temperature at our cave sites have very likely varied by less than 8°C within the last 24,000 years (Braconnot et al., 2007), and possibly by less than 4°C if we assume that the change in local temperature is similar to that in the nearby BoB (Rashid et al., 2007). Given the small temperature-dependent calcite-water oxygen isotope fractionation ($\sim -0.23\text{‰}/^\circ\text{C}$, (O'Neil et al., 1969)), the cave temperature effect on the speleothem $\delta^{18}\text{O}$ is likely to be about 1‰ or less, much smaller than the total $\delta^{18}\text{O}$ variation ($\sim 6\text{‰}$) observed. We thus argue that the variations in the speleothem calcite $\delta^{18}\text{O}$ of our SMWT records essentially reflect the changes in precipitation $\delta^{18}\text{O}$ in the region.

The precipitation over the coastal lowland of Mainland Southeast Asia is primarily fed by the moisture from the BoB, brought by the monsoonal wind (Fig. 3.1). Since our cave sites are very close to the coastal line of the BoB, the local precipitation $\delta^{18}\text{O}$ signal has been altered only minimally by the continental effect, which includes water recycling through evapotranspiration and latitude/altitude effects. $\delta^{18}\text{O}$, therefore, primarily records an isotope fractionation process that can be described by the Rayleigh fractionation model. Our cave sites are thus among the best locations in Mainland Southeast Asia to

study the changes in precipitation associated with the monsoon. Meteorological data from Mainland Southeast Asia show that, after averaging out the seasonal complexity, the annual precipitation $\delta^{18}\text{O}$ is predominantly controlled by the regional precipitation amount, and broadly reflects ISM intensity, with lower precipitation $\delta^{18}\text{O}$ corresponding to stronger ISM intensity, and vice versa (Araguás-Araguás et al., 1998; Breitenbach et al., 2010; Cai et al., 2015; Kurita et al., 2009; Zhang et al., 2006). Given a minimal change in the coastal line of this region through the past glacial cycles, the process that controls the variation in precipitation $\delta^{18}\text{O}$ during glacial times should resemble that of today (Aggarwal, 2004; Cai et al., 2015). The variation of SMWT speleothem $\delta^{18}\text{O}$, which registers the precipitation $\delta^{18}\text{O}$ at our cave sites, can serve as an index of precipitation over the monsoonal region. The shift of SMWT speleothem $\delta^{18}\text{O}$ therefore primarily reflects the varying strength of the AM, or more specifically, of the ISM.

Figure 3. 5. The $\delta^{18}\text{O}$ time series and ^{230}Th ages of stalagmite samples from SMWT caves over the last 30 ka. The $\delta^{18}\text{O}$ profiles and dating error bars (2σ) are color coded as shown by the legend in the figure.

3.6.2. ISM variation

The ISM strength inferred from the SMWT speleothem $\delta^{18}\text{O}$ has changed dramatically from ~ 29 ka to ~ 11 ka. The $\delta^{18}\text{O}$ record shows relatively high values ($\sim -3\text{‰}$ to $\sim -4\text{‰}$) during the late Marine Isotope Stage 3 (MIS 3, ~ 29 ka to ~ 22 ka) and the LGM (~ 22 ka to ~ 19 ka), and achieves the most positive values after the LGM, between ~ 18.5 ka and ~ 15.5 ka ($\sim -2.5\text{‰}$ to $\sim -3.5\text{‰}$), likely linked to Heinrich Event 1. $\delta^{18}\text{O}$ then shifts to negative values close to today's ($\sim -6\text{‰}$ to $\sim -8\text{‰}$) from ~ 14 ka to ~ 12 ka, probably linked to the Bølling-Allerød warming period (Fig. 3.5). The overall speleothem $\delta^{18}\text{O}$ shift in our record is in good agreement with other speleothem records from the ISM domain, for example from India (Dutt et al., 2015; Kathayat et al., 2016; Raza et al., 2017), southwest China (Cai et al., 2015), and Oman and Yemen (Burns et al., 2001; Shakun et al., 2007). It also broadly resembles the speleothem records from the EASM domain (Wang et al., 2008). In particular, the speleothem $\delta^{18}\text{O}$ shift in our record is very similar to the speleothem $\delta^{18}\text{O}$ record from Mawmluh cave in northeast India (Berkelhammer et al., 2013; Dutt et al., 2015), which is also located very close to the BoB, and shares similar climate characteristics with our cave sites (Fig. 3.1 and 3.6). We therefore combined the records from the two locations to represent the temporal variation of ISM precipitation. The combined record shows a striking similarity with foraminifer *G. ruber* $\delta^{18}\text{O}$ record retrieved from the northern BoB near the mouth of the Ganges-Brahmaputra rivers (Fig. 3.6), which is also interpreted as

a proxy of fluvial input or ISM intensity (Kudrass et al., 2001). Hence, their resemblance not only demonstrates the records' robustness, but also further supports our interpretation of speleothem $\delta^{18}\text{O}$ as a proxy of ISM strength.

The most prominent feature of our record is the large difference in speleothem $\delta^{18}\text{O}$ between the LGM and last few hundred years (Fig. 3.5 and 3.6). The $> 3.5\text{‰}$ offset between the two intervals is one of the largest known $\delta^{18}\text{O}$ offsets among all the speleothem records in the Asian Monsoon region. The observed isotopically heavier $\delta^{18}\text{O}$ during the LGM relative to today's is broadly in accordance with a $> 2\text{‰}$ increase in groundwater $\delta^{18}\text{O}$ recovered from Bangladesh during the LGM (Aggarwal, 2004). The ^{18}O -enrichment in precipitation suggests that the ISM significantly weakened during the LGM, in line with implications from other paleo-ISM records (An et al., 2011; Burns et al., 2001; Cai et al., 2015; Prell and Kutzbach, 1987; Rashid et al., 2007).

Since the monsoonal moisture over the SMWT region is almost solely supplied from the tropical Indian Ocean, the decrease in precipitation amount during the LGM can be estimated using the Rayleigh distillation model following the method described in Wang et al. (2017). We assume a 4°C surface cooling during the LGM relative to today (27.5°C) (Rashid et al., 2007), while relative humidity of coastal Myanmar remained close to 100% through time. The LGM speleothem $\delta^{18}\text{O}$ was calculated using values ($\sim -3.8\text{‰}$, $n=136$) between 22 ka BP and 19 ka BP. The precipitation amount integrated from the ocean source to

the cave sites during the LGM is, therefore, estimated to have been about 56% of today's value. This estimation is comparable to the observation in eastern Amazon lowlands, where the LGM precipitation was about 58% relative to modern levels (Wang et al., 2017). Together, these estimations suggest a much-weakened hydrological cycle in the deep tropics when the Earth was in a full glacial state.

3.6.3. Driving mechanism of the ISM

The relative strength of the ISM has been proposed to be mainly determined by solar radiation, continental ice/snow coverage and sea surface temperature (SST) in the tropical Indian Ocean (An et al., 2015; Wang et al., 2014). This then leads to differences in sensible heating and/or tropospheric latent heating between land and ocean (Clemens et al., 1991). Summer insolation heats the land and ocean differently, and initiates the summer monsoon; it is therefore often invoked as the major driving force of monsoon variability. Paleoclimate studies and climate simulations have also demonstrated the dominant role of Northern Hemisphere summer insolation on regulating the AM (Battisti et al., 2014; Cheng et al., 2009; 2016; Dutt et al., 2015; Kathayat et al., 2016; Wang et al., 2001; 2005). However, in the SMWT record, the elevated speleothem $\delta^{18}\text{O}$ values during the LGM can hardly be explained by insolation forcing alone, as the summer insolation during this period was very close to today's in values (Fig. 3.6). Instead, we suggest that the reduced ISM strength is mainly attributable to a combination of lower SST over the tropical Indian Ocean during the LGM and increased continental ice/snow coverage, particularly on the Tibet Plateau and the Himalaya region associated with full glacial conditions (Clark et al., 2009; Shakun et al., 2012) (Fig. 3.6). During the LGM, tropical SST dropped by $\sim 3\text{-}4^\circ\text{C}$ (Mohtadi et al., 2014; Rashid et al., 2007), resulting in a weakened convection over the region and a decrease in precipitation with increased $\delta^{18}\text{O}$ values over the Indian Monsoon domain

(Pausata et al., 2011). At the same time, the expansion of ice/snow coverage on the continent, particularly in the Himalaya and Tibetan Plateau region (Barnett et al., 1988; Clark et al., 2009), led to increased albedo and reduced sensible heating on the continent; therefore it is also responsible for weakened monsoonal precipitation (Clemens et al., 1991; Kudrass et al., 2001).

Figure 3. 6. The driving mechanism of a weakened ISM during the LGM. a). The combined speleothem records from SWMT (red) and Mawmluh cave (orange) (Berkelhammer et al., 2013; Dutt et al., 2015). The Mawmluh records were shifted by 2.4‰ negatively after correction of the temperature effect on calcite precipitation ($-0.23\text{‰}/^{\circ}\text{C}$, (O'Neil et al., 1969)), assuming the cave temperature is close to the annual mean temperature of the closest major cities (16.9°C at Shillong and 27.5°C at Yangon). b). 30°N summer (JJA) insolation. c). *G. ruber* $\delta^{18}\text{O}$ record, retrieved from sediment core KL126, as a proxy for ISM intensity (Kudrass et al., 2001). d). Tropical Indian Ocean SST reconstructed from sediment core SO189-29KL (Mohtadi et al., 2014). e). Timing of deglaciation in Tibet. The dark cyan line represents the oldest age for the onset of deglaciation, and the dark brown line represents the youngest age. The light color boxes represent the 1σ uncertainty range of the ages (Clark et al. (2009) and references therein). As shown, the local deglaciation on the Tibetan Plateau started no earlier than ~ 18 ka, indicating that the extension of snow/ice coverage was close to the maximum during the LGM ($\sim 19\text{-}22$ ka BP).

3.6.4. Spatial response of speleothem $\delta^{18}\text{O}$ to glacial forcings

The SMWT records differ significantly from the Chinese records with respect to speleothem $\delta^{18}\text{O}$ offsets between the LGM and today. For example, the combined Hulu/Dongge records show less than 1.5‰ in offset between the two time intervals (Yuan et al., 2004). The offset is less than 1‰ for the Sanbao records (Wang et al., 2008), and in the Jiuxian records it almost disappears (Cai et al., 2010). The observed difference in the $\delta^{18}\text{O}$ offset reveals fundamental differences in the sensitivities of speleothem $\delta^{18}\text{O}$ proxy to orbital-scale climate forcings between the two regions. Within the upwind ISM domain, the variation of speleothem $\delta^{18}\text{O}$ is sensitive to both the insolation forcing and glacial forcing, whereas in the downwind EASM domain, the speleothem $\delta^{18}\text{O}$ is apparently dominated by insolation forcing.

The larger $\delta^{18}\text{O}$ offset observed in the records from the coastal BoB ($> 3.5\text{‰}$) suggests that the upstream depletion mechanism (Maher, 2008; Pausata et al., 2011) cannot fully account for the variations of the Chinese speleothem $\delta^{18}\text{O}$. Contrary to the observations, similar or larger isotope offsets would be expected in the Chinese records because the downwind signal would inherit or even amplify the variability of changes in the upwind (Cai et al., 2012; 2015).

The observation also disagrees with the local source mechanism, which proposed that a significant amount of the locally-sourced moisture from the South China Sea and the Western Pacific had altered the precipitation $\delta^{18}\text{O}$ signal (Caley et al., 2014; Clemens et al., 2010; Dayem et al., 2010; Johnson et al., 2006). The locally sourced moisture is more enriched in ^{18}O compared to the moisture remotely sourced from the Indian Ocean (Aggarwal, 2004). Moreover, the contribution of locally-sourced moisture to the annual precipitation in southern China should have increased during the LGM (Cai et al., 2015). This should have led to more positive speleothem $\delta^{18}\text{O}$ values during the LGM, and an even larger $\delta^{18}\text{O}$ offset between the LGM and today in southern China relative to the signal in the upwind. However, the observed smaller $\delta^{18}\text{O}$ offsets in southern China clearly suggest a minimal contribution of locally-sourced moisture.

Instead, we believe that the mechanism behind the spatially different response of the speleothem $\delta^{18}\text{O}$ to glacial forcing should lie in the moisture transport. A robust way to evaluate the change of moisture transport through time is to compare speleothem $\delta^{18}\text{O}$ records covering a wide geographical range (Wang et al., 2017). We made such a spatial-temporal comparison between the well-dated speleothem records from Hulu, Dongge, and Sanbao caves from the EASM domain (Dykoski et al., 2005; Kelly et al., 2006; Wang et al., 2001; 2005; 2008; Yuan et al., 2004) and our newly-resolved SMWT record together with speleothem records from Mawmluh cave (Berkelhammer et al., 2013; Dutt et

al., 2015) from the ISM domain (Fig. 3.7). The most prominent feature in the comparison is a much larger isotope gradient ($> 3\text{‰}$) between the records from the two regions during the LGM or broadly the last glacial time, relative to today or broadly the Holocene. The larger isotopic gradient suggests an essentially stronger oxygen isotope fractionation of the rain process associated with the AM during glacial times. A recent study on the comparison of speleothem records from SW China and SE China (Cai et al., 2015) revealed different glacial-interglacial isotope offsets between the two records, which is consistent with our observation. Therefore, we argue that the isotope fractionation during the moisture transport from the Indian Ocean to the cave sites in southern China during glacial periods is much larger relative to interglacial periods. The larger fractionation leads to more isotopically depleted speleothem $\delta^{18}\text{O}$ values in the downwind, even with potentially weakened AM precipitation. Therefore, we further hypothesize that the lack of response in speleothem $\delta^{18}\text{O}$ to glacial-interglacial variations in southern China is related to the counteraction of stronger moisture isotope fractionation from source to cave sites during glacial times.

Two mechanisms have been proposed to explain such changes in moisture isotope fractionation. Cai et al. (2015) suggests that a combination of circulation changes and reduced evaporation over the exposed Sunda Shelf during glacial times is responsible for isotopically depleted speleothem $\delta^{18}\text{O}$ in southern China. Wang et al. (2017) suggests that reduced water recycling

through plant transpiration during the last glacial period could significantly increase the isotope gradient over the continent. Our current study cannot provide enough evidence to evaluate the two mechanisms. A better spatial coverage of speleothem records, particularly along the monsoon moisture transport corridor is needed to thoroughly understand the monsoon moisture cycle.

Figure 3. 7. Spatial-temporal comparison of speleothem $\delta^{18}\text{O}$ records from the coastal Indian Ocean and southern China. The red line represents the SWMT record in this study. The orange line represents the Mawmluh cave records (Berkelhammer et al., 2013; Dutt et al., 2015) after correcting for temperature effects. The green line is a compiled speleothem $\delta^{18}\text{O}$ record from Hulu Cave (Wang et al., 2001; Yuan et al., 2004), Dongge Cave (Dykoski et al., 2005; Wang et al., 2005), and Sanbao Cave, with the Sanbao record shifted by 1.6‰ to align with the other two cave records from southern China (Wang et al., 2008). The thick grey line shows the 65°N summer insolation.

3.7. Conclusions

The SMWT speleothem $\delta^{18}\text{O}$ record documents past changes of ISM intensity, and covers a significant portion of the past 30 ka. Our data indicate that the ISM was much weaker during the LGM, with the regional rain amount possibly being ~56% of the today's. This observation is consistent with proxy-based reconstructions from other climate archives. A weakened ISM during the LGM was mainly driven by changes in glacial boundary conditions. The difference between the SMWT records and Chinese speleothem records reveals that the sensitivity of speleothem $\delta^{18}\text{O}$ to insolation and glacial forcing varies spatially. The muted glacial-interglacial differences in the Chinese speleothem records are likely a result of the counteracted effect of a larger (smaller) isotope fractionation of precipitation $\delta^{18}\text{O}$ during glacial (interglacial) times.

Table 3. 1. ^{230}Th dating results, error in 2σ (95% in confidence).

Sample ID	^{238}U (ppb)	^{232}Th (ppt)	$^{230}\text{Th} / ^{232}\text{Th}$ (atomic $\times 10^{-6}$)	$\delta^{234}\text{U}$ ^a (measured)	$^{230}\text{Th} / ^{238}\text{U}$ (activity)	^{230}Th Age (yr) (uncorrected)	^{230}Th Age (yr) (corrected)	$\delta^{234}\text{U}_{\text{Initial}}$ ^b (corrected)	^{230}Th Age (yr BP) (corrected) ^c
HT2-1									
HT2-1-1mm	106.8 $\pm$ 0.3	2784 $\pm$ 56	68.0 $\pm$ 1.4	61.0 $\pm$ 2.0	0.1074 $\pm$ 0.0006	11628 $\pm$ 77	10913 $\pm$ 512	63 $\pm$ 2	10847 $\pm$ 512
HT2-1-56mm	51.0 $\pm$ 0.1	3698 $\pm$ 74	30.9 $\pm$ 0.6	68.3 $\pm$ 2.0	0.1360 $\pm$ 0.0009	14822 $\pm$ 105	12833 $\pm$ 1411	71 $\pm$ 2	12766 $\pm$ 1411
HT2-1-66mm	66.1 $\pm$ 0.1	700 $\pm$ 14	203.4 $\pm$ 4.4	69.9 $\pm$ 1.8	0.1307 $\pm$ 0.0011	14183 $\pm$ 127	13895 $\pm$ 240	73 $\pm$ 2	13828 $\pm$ 240
HT2-1-156mm	70.0 $\pm$ 0.1	1687 $\pm$ 34	101.6 $\pm$ 2.2	92.6 $\pm$ 1.8	0.1485 $\pm$ 0.0011	15895 $\pm$ 127	15254 $\pm$ 471	97 $\pm$ 2	15187 $\pm$ 471
HT2-1-191mm	75.3 $\pm$ 0.3	3718 $\pm$ 75	53.3 $\pm$ 1.1	91.3 $\pm$ 2.7	0.1596 $\pm$ 0.0009	17202 $\pm$ 116	15884 $\pm$ 939	95 $\pm$ 3	15818 $\pm$ 939
HT2-2									
HT2-2-37mm	105.8 $\pm$ 0.3	15496 $\pm$ 312	26.4 $\pm$ 0.6	147.6 $\pm$ 2.5	0.2346 $\pm$ 0.0016	24812 $\pm$ 202	21065 $\pm$ 2660	157 $\pm$ 3	20999 $\pm$ 2660
HT2-2-76mm	116.2 $\pm$ 0.2	23341 $\pm$ 468	19.7 $\pm$ 0.4	136.4 $\pm$ 1.8	0.2398 $\pm$ 0.0012	25726 $\pm$ 156	20499 $\pm$ 3707	145 $\pm$ 2	20432 $\pm$ 3707
HT2-2-145mm	93.9 $\pm$ 0.1	29747 $\pm$ 597	13.3 $\pm$ 0.3	145.9 $\pm$ 1.7	0.2548 $\pm$ 0.0012	27279 $\pm$ 151	18994 $\pm$ 5885	154 $\pm$ 3	18927 $\pm$ 5885
HT2-2-215mm	99.0 $\pm$ 0.2	1879 $\pm$ 38	194.4 $\pm$ 4.0	138.1 $\pm$ 2.0	0.2237 $\pm$ 0.0011	23761 $\pm$ 140	23279 $\pm$ 368	147 $\pm$ 2	23213 $\pm$ 368
PHA02									
PHA02-3mm	401.6 $\pm$ 0.9	15527 $\pm$ 49	4.0 $\pm$ 0.1	-20.2 $\pm$ 2.2	0.0094 $\pm$ 0.0003	1053 $\pm$ 39	4 $\pm$ 528	-20 $\pm$ 2	-59 $\pm$ 528
PHA02-3mm	395.1 $\pm$ 0.5	12543 $\pm$ 53	2.6 $\pm$ 0.1	-20.9 $\pm$ 1.5	0.0051 $\pm$ 0.0003	568 $\pm$ 31	-293 $\pm$ 433	-21 $\pm$ 1	-356 $\pm$ 433
PHA02-45mm	339.5 $\pm$ 0.3	9945 $\pm$ 44	3.2 $\pm$ 0.2	-41.0 $\pm$ 1.2	0.0057 $\pm$ 0.0003	647 $\pm$ 39	-164 $\pm$ 409	-41 $\pm$ 1	-227 $\pm$ 409
PHA02-47mm	354.3 $\pm$ 0.4	6026 $\pm$ 22	6.1 $\pm$ 0.2	-40.7 $\pm$ 1.5	0.0063 $\pm$ 0.0002	714 $\pm$ 25	244 $\pm$ 237	-41 $\pm$ 1	181 $\pm$ 237
PHA02-100mm	286.9 $\pm$ 0.3	6009 $\pm$ 21	6.5 $\pm$ 0.2	-37.8 $\pm$ 1.4	0.0082 $\pm$ 0.0003	935 $\pm$ 31	358 $\pm$ 291	-38 $\pm$ 1	295 $\pm$ 291
PHA02-127mm	358.1 $\pm$ 1.2	2998 $\pm$ 13	15.3 $\pm$ 0.4	-34.2 $\pm$ 3.1	0.0078 $\pm$ 0.0002	881 $\pm$ 24	652 $\pm$ 115	-34 $\pm$ 3	589 $\pm$ 115
PMD02									
PMD02-2mm	149.9 $\pm$ 0.4	2241 $\pm$ 13	119.2 $\pm$ 1.2	64.9 $\pm$ 2.7	0.1080 $\pm$ 0.0010	11660 $\pm$ 114	11289 $\pm$ 218	67 $\pm$ 3	11227 $\pm$ 218
PMD02-2mm	115.9 $\pm$ 0.1	775 $\pm$ 11	261.5 $\pm$ 4.2	65.7 $\pm$ 1.3	0.1059 $\pm$ 0.0008	11420 $\pm$ 88	11254 $\pm$ 121	68 $\pm$ 1	11191 $\pm$ 121
PMD02-60mm	61.4 $\pm$ 0.1	504 $\pm$ 12	273.6 $\pm$ 6.8	69.6 $\pm$ 1.9	0.1363 $\pm$ 0.0012	14840 $\pm$ 138	14617 $\pm$ 209	73 $\pm$ 2	14554 $\pm$ 209
PMD02-160mm	45.3 $\pm$ 0.1	1391 $\pm$ 28	92.4 $\pm$ 2.0	73.2 $\pm$ 1.9	0.1719 $\pm$ 0.0014	19001 $\pm$ 171	18169 $\pm$ 613	77 $\pm$ 2	18104 $\pm$ 613
PMD02-160mm	53.2 $\pm$ 0.1	1639 $\pm$ 33	94.3 $\pm$ 2.2	81.9 $\pm$ 2.4	0.1762 $\pm$ 0.0019	19348 $\pm$ 231	18519 $\pm$ 629	86 $\pm$ 3	18454 $\pm$ 629

PDM02-245mm	54.0 ±0.1	2445 ±11	69.9 ±0.9	76.3 ±1.7	0.1914 ±0.0022	21334 ±274	20221 ±622	81 ±2	20158 ±622
PDM02-295mm	82.7 ±0.1	4368 ±14	57.8 ±0.7	79.1 ±2.0	0.1849 ±0.0021	20477 ±260	19179 ±701	83 ±2	19116 ±701
PDM02-320mm	48.9 ±0.1	2741 ±55	63.8 ±1.3	78.5 ±2.3	0.2167 ±0.0014	24394 ±189	22879 ±1087	84 ±2	22814 ±1087
PDM02-320mm	51.0 ±0.1	2778 ±56	64.5 ±1.3	77.9 ±2.2	0.2131 ±0.0013	23958 ±171	22485 ±1056	83 ±2	22420 ±1056
PDM02-325mm	68.0 ±0.1	2742 ±56	80.0 ±1.7	74.6 ±2.3	0.1957 ±0.0015	21865 ±188	20772 ±795	79 ±2	20709 ±795

Sample ID is given by sample name and sub-sampling depth from its top in mm.

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).

^a $\delta^{234}\text{U} = ([^{234}\text{U}/^{238}\text{U}]_{\text{activity}} - 1) \times 1000 \text{ ‰}$.

^b $\delta^{234}\text{U}_{\text{initial}}$ was calculated based on ^{230}Th age (T), i.e., $\delta^{234}\text{U}_{\text{initial}} = \delta^{234}\text{U}_{\text{measured}} \times e^{\lambda_{234} \times T}$.

Corrected ^{230}Th ages assume the initial $^{230}\text{Th}/^{232}\text{Th}$ 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 $^{232}\text{Th}/^{238}\text{U}$ value of 3.8. The errors are arbitrarily assumed to be 50%.

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

Table 3. 2. Stable isotope data for stalagmite samples from SMWT caves.

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
HT2-1				136	14.89	-6.94	-5.75	167	22.13	-6.12	-4.05
1	10.75	-1.19	-4.95	141	14.97	-5.72	-6.29	174	22.28	-5.16	-4.02
6	11.00	-1.39	-4.67	146	15.05	-6.25	-4.62	179	22.38	-8.55	-5.83
11	11.24	-3.00	-5.88	151	15.13	-6.08	-6.87	184	22.48	-7.23	-4.79
16	11.49	-4.45	-6.80	156	15.22	-7.11	-5.55	190	22.60	-8.44	-4.33
21	11.78	-6.11	-6.57	161	15.30	-6.46	-6.05	195	22.70	-7.96	-4.81
26	12.15	-4.95	-7.00	166	15.38	-7.25	-7.36	202	22.85	-6.97	-4.06
31	12.80	-5.63	-6.21	171	15.47	-6.10	-5.09	208	22.97	-7.42	-3.79
36	13.00	-5.43	-7.84	176	15.54	-7.66	-7.54	PHA02			
41	13.12	-5.35	-7.83	181	15.63	-8.04	-7.74	10	-0.04	-6.30	-7.46
46	13.23	-6.57	-6.84	186	15.71	-8.08	-5.09	20	0.01	-8.94	-8.03
51	13.32	-6.50	-7.19	191	15.80	-7.87	-5.89	30	0.06	-9.40	-7.72
56	13.44	-4.91	-8.55	196	15.87	-6.52	-4.88	40	0.11	-8.04	-7.83
61	13.60	-6.79	-7.05	HT2-2				50	0.16	-7.44	-7.02
66	13.75	-8.62	-8.30	28	20.00	-6.86	-3.98	60	0.21	-8.08	-7.27
71	13.85	-7.76	-8.17	34	20.09	-6.66	-3.71	70	0.26	-8.27	-7.51
76	13.91	-8.08	-6.99	40	20.19	-6.74	-3.82	80	0.31	-8.39	-7.50
81	13.99	-7.43	-6.56	45	20.26	-5.50	-3.31	90	0.36	-9.63	-7.61
86	14.07	-6.37	-6.85	57	20.29	-7.39	-3.82	100	0.41	-10.01	-7.76
91	14.15	-5.83	-5.94	76	20.53	-6.93	-3.56	110	0.46	-9.14	-7.54
96	14.24	-6.61	-6.67	84	20.66	-6.17	-4.05	120	0.51	-8.84	-6.98
101	14.32	-5.19	-7.06	112	21.12	-6.83	-4.78	YTP01			
106	14.40	-7.69	-6.81	123	21.32	-5.40	-4.73	10	34.47	-6.85	-5.72
111	14.48	-5.96	-6.67	133	21.48	-6.91	-3.75	20	34.61	-7.10	-5.03
116	14.56	-3.32	-5.08	139	21.59	-6.79	-2.89	30	34.75	-7.97	-5.51
121	14.65	-7.87	-6.71	144	21.69	-8.46	-4.06	40	34.88	-7.52	-5.05
126	14.73	-7.56	-6.96	156	21.93	-7.59	-4.71	50	35.02	-7.58	-5.34
131	14.81	-7.17	-5.30	161	22.02	-7.06	-4.09	60	35.16	-7.25	-6.08

Table 3.2. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
70	35.29	-8.28	-5.78	14	13.05	-11.56	-6.60	42	14.00	-11.57	-6.23
80	35.43	-8.39	-6.06	15	13.09	-12.04	-6.86	43	14.04	-10.78	-6.38
90	35.57	-8.56	-6.06	16	13.12	-12.41	-7.36	44	14.07	-12.12	-6.20
100	35.70	-7.46	-5.73	17	13.16	-11.84	-5.89	45	14.11	-11.30	-6.40
110	35.84	-7.72	-5.92	18	13.19	-12.49	-6.99	46	14.14	-10.23	-6.10
120	35.98	-7.60	-5.24	19	13.23	-12.57	-6.65	47	14.18	-9.34	-6.43
130	36.12	-7.32	-4.30	20	13.26	-12.59	-6.84	48	14.21	-9.49	-6.56
140	36.25	-7.40	-4.74	21	13.30	-12.20	-5.77	49	14.24	-11.03	-6.66
150	36.39	-7.21	-5.00	22	13.33	-12.39	-6.30	50	14.28	-9.75	-6.17
160	36.53	-7.50	-5.00	23	13.36	-12.28	-7.09	51	14.31	-10.64	-6.37
170	36.66	-6.74	-4.69	24	13.40	-12.56	-6.96	52	14.34	-10.30	-5.93
180	36.80	-6.14	-4.54	25	13.43	-12.40	-6.83	53	14.38	-11.05	-5.98
190	36.94	-6.58	-4.85	26	13.47	-13.12	-6.94	54	14.41	-11.72	-6.12
200	37.08	-7.15	-4.83	27	13.50	-12.72	-6.81	55	14.44	-11.07	-6.09
210	37.21	-6.53	-4.82	28	13.53	-12.73	-6.71	56	14.48	-11.29	-5.86
220	37.35	-6.66	-4.91	29	13.57	-12.88	-6.74	57	14.51	-11.67	-5.29
230	37.49	-6.56	-5.07	30	13.60	-12.89	-6.53	58	14.55	-11.46	-4.97
240	37.62	-7.01	-5.20	31	13.63	-12.68	-5.97	59	14.58	-11.74	-4.77
PDM02				32	13.67	-11.57	-6.16	60	14.61	-12.07	-4.70
5	12.75	-9.94	-6.66	33	13.70	-11.41	-6.51	61	14.65	-11.72	-5.71
6	12.78	-10.84	-6.40	34	13.73	-12.79	-6.78	62	14.68	-11.58	-5.59
7	12.81	-10.16	-6.38	35	13.76	-12.47	-6.81	63	14.72	-11.82	-5.57
8	12.84	-9.91	-6.01	36	13.80	-12.17	-6.29	64	14.75	-11.64	-5.47
9	12.87	-8.76	-5.35	37	13.83	-11.19	-5.97	65	14.78	-11.42	-5.38
10	12.91	-9.15	-6.59	38	13.87	-12.06	-5.81	66	14.82	-11.93	-5.61
11	12.95	-9.43	-6.68	39	13.90	-12.43	-6.02	67	14.85	-11.76	-5.18
12	12.99	-11.30	-6.81	40	13.94	-12.08	-5.81	68	14.89	-12.21	-5.06
13	13.02	-11.54	-6.82	41	13.97	-12.36	-6.63	69	14.92	-10.39	-4.99

Table 3.2. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
70	14.96	-11.91	-4.83	98	15.94	-10.32	-3.54	126	16.97	-11.47	-2.96
71	14.99	-12.24	-4.64	99	15.98	-9.17	-3.27	127	17.01	-11.07	-2.66
72	15.03	-11.30	-4.71	100	16.01	-8.98	-3.48	128	17.05	-9.43	-2.59
73	15.06	-11.35	-4.46	101	16.05	-9.25	-3.52	129	17.08	-9.35	-3.07
74	15.10	-11.68	-4.71	102	16.09	-9.87	-3.69	130	17.12	-7.19	-2.70
75	15.13	-9.49	-4.47	103	16.12	-10.19	-3.55	131	17.16	-10.50	-3.24
76	15.17	-9.34	-4.36	104	16.16	-10.21	-3.37	132	17.19	-10.58	-3.24
77	15.20	-9.41	-4.54	105	16.19	-11.48	-3.59	133	17.23	-9.46	-2.71
78	15.24	-9.51	-4.31	106	16.23	-11.46	-3.62	134	17.27	-8.86	-2.58
79	15.27	-10.49	-4.18	107	16.27	-11.44	-3.68	135	17.30	-10.12	-2.68
80	15.31	-8.21	-3.72	108	16.30	-11.49	-3.93	136	17.34	-11.12	-2.75
81	15.34	-9.24	-3.62	109	16.34	-11.05	-3.68	137	17.38	-12.10	-2.85
82	15.38	-8.62	-3.56	110	16.38	-11.46	-3.20	138	17.41	-11.91	-2.56
83	15.41	-10.69	-3.71	111	16.42	-10.11	-3.64	139	17.45	-11.96	-2.56
84	15.45	-12.24	-3.81	112	16.45	-8.41	-3.25	140	17.48	-11.77	-2.48
85	15.48	-11.92	-4.23	113	16.49	-10.23	-2.85	141	17.52	-11.12	-2.47
86	15.52	-10.11	-4.44	114	16.52	-11.15	-2.91	142	17.55	-10.35	-2.57
87	15.55	-9.89	-4.09	115	16.56	-10.60	-2.87	143	17.59	-9.28	-2.74
88	15.59	-10.93	-3.27	116	16.60	-10.10	-3.24	144	17.62	-10.05	-2.54
89	15.62	-11.13	-3.28	117	16.64	-11.61	-3.62	145	17.66	-10.65	-2.52
90	15.65	-10.94	-3.51	118	16.68	-10.95	-3.21	146	17.69	-10.44	-2.73
91	15.69	-10.94	-3.51	119	16.71	-11.03	-3.18	147	17.72	-9.93	-2.52
92	15.72	-10.89	-4.26	120	16.75	-10.09	-2.68	148	17.75	-10.10	-2.31
93	15.76	-10.69	-3.41	121	16.79	-9.61	-2.92	149	17.79	-7.11	-2.58
94	15.80	-11.22	-3.38	122	16.83	-9.01	-3.54	150	17.82	-10.00	-3.31
95	15.83	-9.83	-3.62	123	16.86	-11.43	-3.59	151	17.86	-11.51	-2.82
96	15.87	-8.88	-3.69	124	16.90	-10.85	-3.09	152	17.89	-11.86	-3.27
97	15.91	-9.73	-3.36	125	16.93	-11.85	-3.21	153	17.92	-11.90	-3.02

Table 3.2. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
154	17.95	-11.16	-3.18	182	18.66	-6.47	-3.84	210	19.36	-7.11	-3.00
155	17.98	-10.81	-3.20	183	18.68	-6.98	-3.56	211	19.37	-8.93	-3.66
156	18.01	-10.77	-3.21	184	18.71	-6.54	-3.81	212	19.39	-8.62	-3.29
157	18.04	-9.74	-3.41	185	18.74	-6.25	-4.13	213	19.40	-8.63	-3.43
158	18.07	-9.09	-3.58	186	18.77	-6.13	-3.52	214	19.41	-7.94	-3.31
159	18.08	-10.27	-3.59	187	18.80	-6.02	-3.32	215	19.43	-7.72	-3.33
160	18.08	-10.90	-3.74	188	18.83	-5.98	-3.49	216	19.44	-6.06	-3.36
161	18.09	-11.00	-3.21	189	18.86	-5.76	-3.58	217	19.45	-5.89	-3.61
162	18.09	-11.20	-3.28	190	18.89	-7.70	-3.34	218	19.47	-9.66	-3.50
163	18.11	-11.11	-3.64	191	18.92	-8.64	-3.41	219	19.48	-9.74	-3.48
164	18.14	-10.07	-3.44	192	18.95	-8.52	-3.34	220	19.49	-9.32	-3.14
165	18.17	-7.59	-3.57	193	18.98	-7.73	-3.73	221	19.51	-8.84	-3.59
166	18.20	-9.34	-3.88	194	19.01	-6.12	-3.67	222	19.52	-8.00	-3.35
167	18.24	-7.90	-3.69	195	19.03	-9.96	-3.68	223	19.53	-9.77	-3.90
168	18.27	-10.04	-3.13	196	19.06	-10.02	-3.64	224	19.54	-8.85	-3.12
169	18.30	-9.60	-3.77	197	19.08	-9.59	-3.39	225	19.55	-8.02	-3.58
170	18.32	-8.43	-3.79	198	19.11	-8.34	-3.64	226	19.56	-5.83	-3.84
171	18.35	-7.58	-3.74	199	19.14	-6.72	-3.61	227	19.57	-8.47	-3.91
172	18.38	-5.86	-3.57	200	19.17	-6.56	-3.66	228	19.57	-9.41	-3.65
173	18.41	-5.63	-3.84	201	19.19	-7.04	-4.50	229	19.58	-7.84	-3.69
174	18.44	-6.31	-3.64	202	19.22	-7.50	-3.78	230	19.58	-7.61	-3.68
175	18.47	-6.48	-3.73	203	19.24	-7.79	-3.49	231	19.59	-8.68	-4.58
176	18.50	-8.29	-3.67	204	19.27	-7.49	-3.82	232	19.59	-7.64	-4.02
177	18.52	-8.51	-3.83	205	19.29	-6.83	-3.69	233	19.60	-8.63	-3.75
178	18.55	-8.27	-3.68	206	19.30	-7.04	-3.81	234	19.61	-10.33	-3.68
179	18.58	-7.48	-3.11	207	19.32	-8.05	-3.94	235	19.61	-10.49	-3.59
180	18.60	-7.32	-3.40	208	19.33	-5.11	-3.65	236	19.61	-10.78	-3.30
181	18.63	-8.14	-3.39	209	19.35	-5.86	-3.66	237	19.61	-11.12	-3.61

Table 3.2. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
238	19.61	-10.93	-3.59	266	19.67	-7.27	-3.67	294	19.80	-5.89	-3.90
239	19.61	-10.04	-3.72	267	19.67	-5.58	-3.55	295	19.80	-5.80	-3.67
240	19.61	-8.73	-3.76	268	19.67	-6.16	-3.51	296	19.82	-5.67	-3.78
241	19.62	-8.94	-3.81	269	19.67	-6.12	-3.74	297	19.85	-6.10	-3.72
242	19.62	-9.63	-5.12	270	19.67	-6.79	-3.81	298	19.89	-6.00	-3.61
243	19.62	-10.04	-3.57	271	19.67	-6.65	-4.36	299	19.94	-4.46	-3.86
244	19.62	-10.49	-3.79	272	19.67	-6.13	-3.60	300	19.99	-4.29	-3.90
245	19.62	-10.43	-4.00	273	19.67	-5.86	-3.83	301	20.05	-3.90	-4.61
246	19.63	-10.60	-3.95	274	19.67	-5.84	-3.89	302	20.12	-3.61	-4.32
247	19.63	-9.93	-3.18	275	19.67	-9.07	-3.73	303	20.21	-3.67	-4.00
248	19.63	-9.73	-3.33	276	19.67	-8.65	-3.72	304	20.34	-4.52	-3.72
249	19.63	-9.51	-3.92	277	19.67	-8.12	-3.85	305	20.47	-5.40	-4.11
250	19.63	-8.52	-3.44	278	19.68	-7.81	-3.78	306	20.60	-5.46	-4.11
251	19.63	-7.57	-3.62	279	19.68	-6.78	-3.41	307	20.74	-5.82	-4.17
252	19.63	-7.03	-3.69	280	19.69	-6.10	-3.86	308	20.88	-4.93	-3.46
253	19.64	-7.54	-4.53	281	19.69	-5.17	-4.02	309	21.01	-4.37	-3.42
254	19.64	-7.25	-3.61	282	19.69	-5.12	-3.63	310	21.15	-4.39	-3.41
255	19.64	-10.80	-3.45	283	19.69	-5.43	-3.11	311	21.29	-4.33	-3.36
256	19.65	-11.07	-3.27	284	19.73	-5.63	-3.92	312	21.42	-4.30	-3.87
257	19.65	-11.42	-3.48	285	19.74	-4.82	-4.08	313	21.56	-4.26	-3.88
258	19.65	-11.20	-4.87	286	19.75	-5.70	-3.55	314	21.69	-4.76	-3.92
259	19.65	-9.79	-4.01	287	19.76	-6.85	-3.99	315	21.83	-5.07	-3.87
260	19.65	-9.31	-4.48	288	19.77	-6.24	-3.57	316	21.96	-4.57	-3.91
261	19.66	-8.52	-3.91	289	19.78	-5.89	-3.94	317	22.10	-5.18	-4.12
262	19.67	-7.99	-3.82	290	19.78	-5.42	-3.66	318	22.25	-4.80	-3.80
263	19.67	-6.53	-4.26	291	19.78	-4.95	-3.95	319	22.38	-5.22	-3.68
264	19.67	-6.10	-3.58	292	19.78	-4.98	-3.57	320	22.45	-5.08	-3.12
265	19.67	-7.68	-3.69	293	19.80	-5.27	-3.79	321	22.52	-5.31	-4.45

Table 3.2. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
322	22.64	-5.19	-3.17	350	26.28	-7.77	-4.01	95		-11.33	-3.50
323	22.79	-7.55	-3.69	351	26.42	-5.31	-3.60	105		-9.96	-3.71
324	22.93	-7.51	-3.88	352	26.55	-5.82	-3.67	115		-11.15	-3.84
325	23.05	-7.18	-3.56	353	26.67	-6.65	-3.72	125		-10.99	-3.62
326	23.17	-6.77	-3.63	354	26.80	-8.09	-4.19	135		-9.81	-3.06
327	23.28	-6.46	-3.84	355	26.93	-8.48	-3.75	145		-12.06	-2.99
328	23.41	-6.52	-3.54	356	27.05	-7.18	-3.78	155		-10.24	-2.90
329	23.54	-7.25	-3.87	357	27.17	-5.79	-3.66	165		-10.62	-3.68
330	23.66	-4.63	-3.96	358	27.30	-4.62	-3.56	175		-9.58	-3.52
331	23.79	-4.32	-3.92	359	27.44	-5.29	-3.84	185		-8.22	-3.77
332	23.93	-3.82	-3.52	360	27.57	-6.56	-4.27	195		-7.18	-3.55
333	24.06	-3.80	-3.42	361	27.70	-6.12	-3.43	205		-6.88	-3.88
334	24.19	-3.52	-3.50	362	27.83	-6.09	-3.78	215		-8.81	-4.60
335	24.32	-3.35	-3.75	363	27.96	-6.21	-3.65	225		-9.62	-3.54
336	24.45	-3.38	-3.87	364	28.09	-6.51	-3.62	235		-9.81	-3.33
337	24.57	-4.13	-3.93	365	28.22	-6.75	-4.01	245		-9.48	-3.90
338	24.70	-5.06	-3.57	PDM02				255		-9.34	-3.66
339	24.83	-3.87	-4.12	Low resolution				265		-10.65	-4.68
340	24.95	-3.97	-4.03	5				275		-6.53	-4.79
341	25.08	-5.46	-3.80	10				285		-8.39	-3.97
342	25.21	-7.77	-3.86	15				295		-6.83	-3.92
343	25.35	-8.29	-4.13	25				305		-5.28	-4.74
344	25.49	-8.12	-4.13	35				315		-4.29	-4.10
345	25.62	-8.23	-4.08	45				325		-5.11	-3.50
346	25.76	-9.07	-4.38	55				335		-5.36	-4.34
347	25.89	-9.74	-4.17	65				345		-6.93	-3.85
348	26.02	-8.87	-3.60	75				355		-6.23	-4.63
349	26.15	-9.14	-3.55	85				365		-5.91	-4.89

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Chapter 4

On the glacial-interglacial variability of speleothem- inferred Asian Monsoon over the past 180 ka

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Contributions

Liu Guangxin was the primary author. Liu Guangxin, Li Xianglei, Wang Xianfeng and Cheng Hai contributed to the design of this study. Liu Guangxin, Li Xianglei, Wang Xianfeng, Chiang Hong-Wei, Sakonvan Chawchai, Lin Thu Aung and Phyo Maung Maung collected and prepared the samples, which were subsequently analyzed by Liu Guangxin and Li Xianglei and interpreted by Liu Guangxin, Li Xianglei, Wang Xianfeng, Cheng Hai, He Shaoneng. Liu Guangxin wrote the paper.

Liu Guangxin: 50%

Li Xianglei: 20%

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4.1. Abstract

The Chinese speleothem $\delta^{18}\text{O}$ record characterizes changes in the Asian Monsoon (AM) for the past 640 ka (Cheng et al., 2016). However, due to the lack of glacial-interglacial variations of Chinese speleothem $\delta^{18}\text{O}$, it is highly controversial whether the proxy actually documents variability of East Asian Summer Monsoon (EASM) rainfall, or of the moisture from the Indian Summer Monsoon (ISM), or of other factors. The correct climatic interpretation of the $\delta^{18}\text{O}$ proxy requires investigations of hydroclimate records with a broader spatial coverage in the region, particularly from sites located in the upwind of the moisture trajectory from the Indian Ocean to inland China. In this chapter, we present two such speleothem $\delta^{18}\text{O}$ records from central Myanmar and southeastern Yunnan, China, both located in the ISM domain, covering a majority of the last 40 ka and 180 ka, respectively. Spatial-temporal comparisons with available speleothem $\delta^{18}\text{O}$ records show that the inferred precipitation $\delta^{18}\text{O}$ signals have strong glacial-interglacial variability close to the Indian Ocean moisture source and less so further inland. The spatial variability of speleothem $\delta^{18}\text{O}$ on glacial-interglacial timescales likely results from a stronger oxygen isotope fractionation in precipitation associated with progressive rainout of the AM moisture during glacial times, possibly due to much colder global temperatures together with suppressed plant transpiration effect. We thus believe that the muted glacial-interglacial variations in the

inland speleothem $\delta^{18}\text{O}$ records from southern China are due to enhanced (weakened) isotope fractionation of transported moisture during glacial (interglacial) periods, dubbed transport pathway effect, which counteracts the forcing of glacial boundary conditions.

4.2. Introduction

The oxygen isotope ratio ($\delta^{18}\text{O}$) of speleothem is considered as a powerful hydroclimate proxy and has been widely used to reconstruct paleomonsoon changes (Cai et al., 2015; Cheng et al., 2009; 2016; Cruz et al., 2005; Fleitmann et al., 2003; Kathayat et al., 2016; X. Wang et al., 2004; Y. Wang et al., 2001; 2005; 2008). The intensive study on speleothem $\delta^{18}\text{O}$ proxy over the past two decades or so has significantly contributed to our understanding of the global monsoon, particularly the East Asian Summer Monsoon (EASM) and Indian Summer Monsoon (ISM), or collectively, the Asian Monsoon (AM). Although numerous and complex processes can influence speleothem $\delta^{18}\text{O}$ (Fairchild et al., 2006; Lachniet, 2009), the broad replication of $\delta^{18}\text{O}$ records from caves thousands of kilometers apart demonstrates that on decadal to orbital timescales, speleothem $\delta^{18}\text{O}$ can effectively record the changes in regional climate (Cheng et al., 2012).

The climatic interpretation of speleothem $\delta^{18}\text{O}$ records is, however, highly debatable, specifically for those from southern China in the EASM domain (Caley et al., 2014; Chiang et al., 2015; Clemens et al., 2010; Dayem et al., 2010; Johnson and Ingram, 2004; LeGrande and Schmidt, 2009; Z. Liu et al., 2014; Maher, 2008; Pausata et al., 2011; Sun et al., 2015). The debate centers on the cause of a different periodicity in the Chinese speleothem $\delta^{18}\text{O}$ records relative to most of the other Asian Monsoon records. The Chinese speleothem

$\delta^{18}\text{O}$ records exhibit a dominant ~ 23 kyr precessional cycle, mainly forced by local summer insolation (Cheng et al., 2016; Wang et al., 2005; 2008), whereas the records from many other monsoonal archives, for examples, the loess (An, 2000; An et al., 1990; Sun et al., 2006; 2015), marine (Clemens and Prell, 1990; P. Wang et al., 2005; 2016) and lake sediments (An et al., 2011), reveal a dominant ~ 100 kyr cycle, suggesting a direct link to the glacial boundary forcing.

Early studies suggested that Chinese speleothem $\delta^{18}\text{O}$ signals characterize EASM intensity (Wang et al., 2001). Such interpretation is based on the fact that the $\delta^{18}\text{O}$ values in today's precipitation associated with the EASM are relatively low, with more negative (isotopically depleted) $\delta^{18}\text{O}$ values corresponding to a higher proportion of EASM precipitation in annual total, hence stronger monsoon intensity. The $\delta^{18}\text{O}$ -inferred monsoon largely responds to changes in Northern Hemisphere summer insolation on orbital timescales, with exceptions on millennial timescales which are associated with North Atlantic abrupt climate events (Cheng et al., 2016; Wang et al., 2001; 2008). However, this monsoon-intensity-oriented interpretation cannot explain two major observations within the last glacial-interglacial cycle when the signals are in the highest resolution. First, the Chinese speleothem $\delta^{18}\text{O}$ records show similar values between present day and the last glacial maximum (LGM, ~ 19 - 22 thousand years ago, ka) (Wang et al., 2008), indicating subtle change in EASM intensities between the two periods. However, the Earth's climate

during the LGM differed substantially from present in several aspects, including atmospheric CO₂ concentration, sea surface temperature, sea level, continental albedo, and the extension of ice mass at high latitude and high altitude (Clark et al., 2009; Denton et al., 2010). Even though the insolation strengths during the two periods are very similar, various paleoclimate records favor a weakened EASM (Lu et al., 2013; Morley and Heusser, 1997; Sun et al., 2015) forced by glacial boundary conditions, which is confirmed by sensitivity tests or climate model simulations (Braconnot et al., 2007; Kutzbach and Guetter, 1986; Kutzbach and Street-Perrott, 1985; Prell and Kutzbach, 1987; 1992; Sun et al., 2015). And second, the speleothem $\delta^{18}\text{O}$ suggests a relatively weaker EASM during the penultimate interglacial time or marine isotope stage (MIS) 5e, compared to insolation peak periods during MIS 5a and MIS 5c (Wang et al., 2008). This is particularly astonishing given that both the glacial forcing and insolation forcing tend toward the opposite.

Alternatively, the low $\delta^{18}\text{O}$ values of the monsoonal precipitation are interpreted to reflect changes in precipitation amount integrated between tropical ocean sources and Chinese cave sites (the so-called ‘upstream depletion mechanism’), following the Rayleigh fractionation model (Yuan et al., 2004). Isotope modeling studies for the Holocene (LeGrande and Schmidt, 2009) and idealized Heinrich Event 1 (Pausata et al., 2011) favor this upstream depletion mechanism and lead to a hypothesis that Chinese speleothem $\delta^{18}\text{O}$ signals reflect the isotope depletion from the upstream instead of local precipitation

amount. The recently-resolved speleothem $\delta^{18}\text{O}$ records from the upstream ISM domain, however, are clearly responsive to the glacial forcing (Cai et al., 2015; and chapter III). Interestingly, the Xiaobailong $\delta^{18}\text{O}$ record follows the insolation even closer during the MIS 5 when compared to the record from Sanbao Cave (Cai et al., 2010). These new observations from the ISM domain argue that the upstream depletion mechanism is also not fully responsible for the Chinese speleothem $\delta^{18}\text{O}$ signals. More importantly, such observations reveal that the speleothem $\delta^{18}\text{O}$ signals exhibit both temporal and spatial distribution patterns, and that the spatial distribution can change through time.

The modern-day precipitation over East Asia already shows a spatial distribution in its $\delta^{18}\text{O}$ signal (Araguás-Araguás et al., 1998; J. Liu et al., 2014; Vuille et al., 2005), which generally undergoes isotopical depletion towards inland following the monsoon trajectory. However, little attention has been paid to how this pattern changed in response to various climate forcings in the past. In fact, Hu et al. (2008) is, to the best of our knowledge, the only speleothem research in East Asia that tried to quantify Holocene Asian monsoon rainfall from spatially separated cave records. Subtle but significant difference among speleothem $\delta^{18}\text{O}$ records that represent past precipitation $\delta^{18}\text{O}$, has often been overlooked because these researches largely focused on the correlation or similarity of speleothem $\delta^{18}\text{O}$ records. For example, unlike the Sanbao records mentioned above, the Dongge $\delta^{18}\text{O}$ record has indistinguishable values between MIS 5c and MIS 5e (Kelly et al., 2006; Yuan et al., 2004), even though most of

the later studies tend to compile them together with the Hulu records without further questioning (Battisti et al., 2014; Cai et al., 2015; Caley et al., 2014; Cheng et al., 2013b; Jo et al., 2014; Kathayat et al., 2016; Meckler et al., 2012).

Apparently, a larger spatial coverage of speleothem $\delta^{18}\text{O}$ records, specifically from the relatively understudied ISM domain, is needed for better understanding of the spatial-temporal distribution of speleothem $\delta^{18}\text{O}$ signals, and thus for better elucidation of speleothem $\delta^{18}\text{O}$ proxy. In this chapter, we present two speleothem $\delta^{18}\text{O}$ records: one from Lin Noe Twin (LNT) Cave in Central Myanmar that covers a major portion of the last 40 ka, and the other one from a cluster of caves in SE Yunnan province, China that covers the last 180 ka. Among these two, the record from SE Yunnan is particularly important, as it is situated at a strategic location close to the boundary between the EASM domain and ISM domain (Wang et al., 2003). A previous study has shown that speleothem $\delta^{18}\text{O}$ in this region is sensitive to both insolation forcing and glacial forcing (Cai et al., 2015). However, key time intervals (Holocene and MIS 5) in the record are either missing or have a low temporal resolution. Our new records have filled these gaps with better resolved speleothem records. In combination with the previous established records, our newly resolved speleothem $\delta^{18}\text{O}$ records allow us to reconstruct the spatial pattern of precipitation $\delta^{18}\text{O}$ in the Asian Monsoon domain in the past, particularly during Holocene and MIS 5, and to assess potential forcing mechanisms.

4.3. Cave setting

Lin Noe Twin Cave (LNT, 21°14'13.4" N, 96°26'37.5" E) (Fig. 4.1) is located on the eastern side of the Shan Plateau, central Myanmar. The cave has developed within the partially dolomitized Permian-Triassic limestone formations (Oo et al., 2002). The present climate of the cave area is dominated by the ISM, with warm and humid summer and dry and temperate winter. The annual average temperature of the closest city, Taunggyi (~70 km from the cave site), is ~19°C, and the annual precipitation is ~1,400 mm with more than 90% occurring during the monsoon season from late April to early November (Fig. 4.2). LNT Cave consists of an inner and an outer chamber. The outer chamber is small and narrow since its roof has collapsed; it is also steeply sloping ~20 m downwards. At its end, a narrow vertical drop of ~15 m leads to a large inner chamber, which is poorly ventilated, and the relative humidity reaches ~100%.

The SE Yunnan record is retrieved from a suite of caves (Baifen Cave, BF; Jiangjun Cave, JJ; Shuanglong Cave, SL; Shuiyuan Cave, SY; Xianren Cave, XR and an unnamed cave, marked as A) located between Honghe and Wenshan cities (~23.5°N, ~104°E), about 200 km SE from Kunming, China. Local climate is distinguished by warm and wet summers and cool and dry winters. About 80% of annual total precipitation (~1,000 mm) occurs between June and September, indicating the dominance of summer monsoon rainfall at the cave sites (Fig. 4.2). The local precipitation is mostly sourced from the ISM, even

though geographically, the clustered sites are already very close to the artificial delineation (105°E) between the ISM and EASM (Fig. 4.1).

Figure 4. 1. The location of Lin Noe Twin (LNT) Cave and the cave cluster in SE Yunnan (red stars). Black arrows represent present-day averaged June-July-August (JJA) wind patterns (Data from the National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) Reanalysis Monthly Mean (1981-2010). (Kalnay et al., 1996). The locations of other cave sites are indicated by blue stars: Sanbao Cave (Wang et al., 2008), Dongge Cave (Kelly et al., 2006; Yuan et al., 2004); Tianmen Cave (Cai et al., 2010) in China; Mawmluh Cave (Dutt et al., 2015) in India; southern Myanmar and western Thailand (SMWT) cave clusters close to the border of the two countries. The red line and red arrows indicate the 105°E designated boundary of the ISM and the EASM (Wang et al., 2003).

Figure 4. 2. Monthly mean precipitation (blue bars) and temperature (red squares) at Taunggyi and Kunming meteorological observatory stations (Data from UNdata: <http://data.un.org/Explorer.aspx?d=CLINO>).

4.4. Materials and Analytical Methods

Stalagmites samples (6, 1, 3, 2, 1, 1, 3 stalagmite(s) from LNT, BF, JJ, SL, SY, XR and A respectively) were all collected from the inner chambers of the caves with very limited air ventilation and humidity close to 100%. Except for stalagmites A-14 and the top portion of A-15, which are composed of aragonite, all the other samples are made of calcite.

All stalagmites were cut into halves along the growth axes and their surfaces were polished. A total of 147 subsamples were drilled along growth axes for ^{230}Th dating. We followed similar chemical procedures as described by Edwards et al. (1987) and Cheng et al. (2000) to separate the uranium and thorium for ^{230}Th dating. The analysis was performed at the Environmental Geochemistry Laboratory, Earth Observatory of Singapore, Nanyang Technological University, Singapore (Yunnan Samples and Myanmar Samples) and the Institute of Global Environmental Change, Xi'an Jiaotong University, Xi'an, China (Yunnan Samples). The ^{230}Th dating techniques are essentially identical in the two laboratories. All measurements were made on Thermo-Finnigan Neptune Plus multi-collector inductively coupled plasma mass spectrometers (MC-ICP-MS) using the recently improved techniques (Cheng et al., 2013a; Shen et al., 2012).

A total of 2413 of subsamples were drilled perpendicularly to the extension axes of the stalagmite samples at 1 to 5 mm increments. The oxygen and carbon stable isotopic compositions ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) of stalagmite samples were analyzed at three laboratories with different analytical systems — the Environmental Geochemistry Laboratory (EGL), Earth Observatory of Singapore (EOS), Nanyang Technological University, Singapore, the Speleothem Isotope Laboratory (SIL), Nanjing Normal University, Nanjing, China, and the Institute of Global Environmental Change (IGEC), Xi'an Jiaotong University, Xi'an, China. Results are reported in per mil (‰), relative to the Vienna PeeDee Belemnite (VPDB) standard. At EGL, stable isotope samples were analyzed using an on-line carbonate preparation system (Gasbench II) interfaced with an isotope ratio mass spectrometer (Delta plus XL). The stable isotope measurements at SIL were performed on two Thermo-Finnigan MAT-253 mass spectrometers fitted with a Kiel Carbonate Device III and the stable isotope measurements at IGEC were made on a Thermo-Finnigan MAT-253 mass spectrometer equipped with a Kiel Carbonate Device IV. At all the three labs, duplicate measurements of NBS19 and in-house TTB1 standards show a long-term reproducibility of $\sim 0.1\text{‰}$ or better (1σ).

4.5. Results

U and Th isotopic compositions and ^{230}Th dates are given in Table 1 and 2. ^{238}U concentrations mostly range from 20 to 200 ppb with few exceptions (~3-13 ppm) mainly from aragonite sample A-14. ^{232}Th concentrations mostly range from less than ten ppt to a few ppb. ^{230}Th ages were calculated assume an initial $^{230}\text{Th}/^{232}\text{Th}$ atomic ratio of $(4.4 \pm 2.2) \times 10^{-6}$, the values for a material at secular equilibrium with the bulk Earth $^{232}\text{Th}/^{238}\text{U}$ value of 3.8. We consider that it is reasonable to adopt this value for initial ^{230}Th correction as most samples have relatively high $^{230}\text{Th}/^{232}\text{Th}$ and initial correction is negligible. Indeed, the typical relative errors (2σ) in ages after initial ^{230}Th correction are better than 1%, and all ^{230}Th dates are in stratigraphic order within errors. Linear extrapolation or the probabilistic Monte Carlo approach method (Scholz and Hoffmann, 2011) were used to establish depth-age models of all stalagmites (Fig. 3 and 4). The $\delta^{18}\text{O}$ values show large shifts, ranging from -9.0 to -4.7‰ in central Myanmar and from -16.1 to -6.1‰ in SE Yunnan. The subsample depths, modeled ages and oxygen isotope composition are listed in Table 3 and 4.

Figure 4. 3. Plots of age versus depth for stalagmites LNT-01, LNT-02, LNT-03, LNT-04, LNT-05, LNT-09U, and LNT-09B from Lin Noe Twin cave. All ages are reported in units of thousands of years before present (1950 AD) or ka BP. The age errors indicated in the plots are 2σ error. Linear interpolation is used to establish the chronology for sample LNT-01, LNT-03, LNT-04 and LNT-09U. StalAge method (Scholz and Hoffmann, 2011) is used to establish the chronology for sample LNT-02, LNT-05, and LNT-09B. The red lines indicated 95% confidence limits calculated by a Monte Carlo simulation fitting.

Figure 4. 4. Plots of age versus depth for stalagmites A-14, A-15, and A-16 from the anonymous cave; BF-02 from Baifen Cave; JJ-03, JJ-06, and JJ-08 from Jiangjun Cave; SL-08 and SL-10 from Shuanglong Cave; SY-10 from Shuiyuan Cave; and XR-05 from Xianren Cave from SE Yunnan. All ages are reported in units of thousands of years before present (1950 AD) or ka BP. The age errors indicated in the plots are 2σ error. Linear interpolation is used to establish the chronology for sample A-14, A-15, A-16, BF-2, JJ-06, SL-08 and SY-10. StalAge method (Scholz and Hoffmann, 2011) is used to establish the chronology for sample SL-10, JJ-03, JJ-06 and XR-05. The red lines indicated 95% confidence limits calculated by a Monte Carlo simulation fitting.

4.6. Isotopic equilibrium tests

The stalagmite samples from Yunnan and Myanmar are mostly composed of calcite, except for A-14 and the upper portion of A-15, which is aragonite. It is important to confirm that the calcite is deposited under isotopic equilibrium conditions before seeking the climatic interpretation of the stalagmite $\delta^{18}\text{O}$. We first calculated the correlation factors between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ for each sample (Figs. 5 and 6). Their low r^2 values (<0.32) argue against kinetically driven fractionation during speleothem growth according to traditional Hendy criteria (Hendy, 1971). We also applied the ‘‘Hendy Test’’ on two individual growth laminae on stalagmite LNT-05 (Fig. 7). The stable isotope variations along the same layer were very small and no systematic isotope value increase was found toward the edges of the sample.

A more robust equilibrium test involves replication of isotopic records across caves (Cheng et al., 2006; 2016; Dorale, 1998; Dykoski et al., 2005). Our speleothem records from Yunnan and Central Myanmar can be broadly correlated on millennial to orbital timescales to those from caves several thousand kilometers away in the Asian Monsoon region over their contemporaneous growth periods (Y. Wang et al., 2001, 2005, 2008; Cai et al., 2015; Cheng et al., 2016). Similar correlations also exist between records from southern Myanmar Caves and those from Thailand caves (Chapter III) and

Mawmluh (Berkelhammer et al., 2013; Dutt et al., 2015) and Bittoo caves (Kathayat et al., 2016) in northern India.

Regarding sample A-14 and the upper portion of A-15, we applied a 0.76‰ aragonite-calcite correction (Kim and O'Neil, 1997; Kim et al., 2007), assuming the cave temperature to be close to the annual mean temperature of the nearby city, Kunming (~14.9°C) without significant fluctuation through the Holocene. The corrected stalagmite $\delta^{18}\text{O}$ values are also in a very good agreement during the overlapping periods with the data of sample A-16, which is from the same unnamed cave and of samples JJ-08 and SL-10 from Jiangjun Cave and Shuanglong Cave, respectively.

With the above lines of evidence, we reasonably conclude that the speleothem samples collected in our study most likely grew under isotopic equilibrium conditions. The variation of carbonate $\delta^{18}\text{O}$ is therefore mostly controlled by the $\delta^{18}\text{O}$ of meteoric precipitation and cave temperature at the time of carbonate precipitation.

Figure 4. 5. Scatter plot of $\delta^{18}\text{O}$ versus $\delta^{13}\text{C}$ from each stalagmite sample from Lin Noe Twin Cave. The low correlation (as indicated R^2 values) of the plotted results indicates that carbon and oxygen are not highly correlated as would be expected under the conditions of kinetic fractionation.

Figure 4. 6. Scatter plot of $\delta^{18}\text{O}$ versus $\delta^{13}\text{C}$ from each stalagmite sample from SE Yunnan Caves. The low correlation (as indicated R^2 values) of the plotted results indicates that carbon and oxygen are not highly correlated as would be expected under the conditions of kinetic fractionation.

Figure 4. 7. The “Hendy Test” (Hendy, 1971) on two different growth laminae from sample LNT-05 suggests that the speleothems most likely grew under isotopic equilibrium conditions.

4.7. Discussion

4.7.1. Interpretation of speleothem $\delta^{18}\text{O}$ proxy

Temperature fluctuations in caves might have contributed to changes in speleothem $\delta^{18}\text{O}$. However, if the temperature were the sole cause, an over 30°C change in cave temperature would be required to account for the large $\delta^{18}\text{O}$ shift ($>8\text{‰}$) observed in the Yunnan records given the carbonate-water isotope fractionation coefficient of $\sim -0.23\text{‰}/^\circ\text{C}$ (O'Neil et al., 1969) (Fig. 4.8). In Southeast Asia, the temperature change through glacial cycles was estimated to be within 8°C (Braconnot et al., 2007) or possibly within 4°C (Cai et al., 2015). For central Myanmar (Fig. 4.9), the contribution from temperature is likely even less since the region is closer to the deep tropics with presumably smaller temperature variation through glacial cycles. Therefore, the $\delta^{18}\text{O}$ variation in each speleothem record should mainly record the changes of local precipitation $\delta^{18}\text{O}$. In Chapter III, we have demonstrated that the coastal Bay of Bengal (BoB) is the ideal location where the variability of speleothem $\delta^{18}\text{O}$ faithfully represents the changes of ISM intensity. For samples from central Myanmar and SE Yunnan, we follow the argument by (Cai et al., 2015) that fluctuation of speleothem $\delta^{18}\text{O}$ also broadly serves as an index of precipitation over the monsoonal region, which primarily reflects varying strength of the Asian Monsoon or more specifically the ISM.

The ISM strength inferred by the speleothem $\delta^{18}\text{O}$ resembles Northern Hemisphere summer insolation (NHSI) very well on orbital timescale, with lower $\delta^{18}\text{O}$ values, heavier precipitation, or stronger ISM corresponding to higher NHSI, and vice versa. On glacial-interglacial timescale, the SE Yunnan $\delta^{18}\text{O}$ record shows relatively positive values ($\sim 9\text{‰}$ to $\sim 11\text{‰}$) during deep glacial periods (~ 160 ka to ~ 135 ka, and ~ 75 ka to ~ 20 ka) but relatively negative $\delta^{18}\text{O}$ values ($\sim 11\text{‰}$ to $\sim 15\text{‰}$) in interglacial periods (~ 125 ka and after ~ 10 ka). A similar glacial-interglacial $\delta^{18}\text{O}$ contrast also appears in the central Myanmar records. This is in a good agreement with the observations from the southern Myanmar and western Thailand (SMWT) speleothem records (Chapter III) and other paleo-ISM records (An et al., 2011; Burns et al., 2001; Cai et al., 2015; Prell and Kutzbach, 1987; Rashid et al., 2007). These observations suggest a generally weakened ISM during glacial periods and that glacial boundary conditions are important to changes of ISM strength.

The Myanmar-Yunnan records are also punctuated by several millennial-scale abrupt events, in correlation with the North Atlantic abrupt events, particularly during Younger Dryas and Heinrich event 1, suggesting a teleconnection between the ISM and climate change in the North Atlantic as seen in model simulations (e.g., Pasatau et al., 2011). The millennial-scale abrupt events are however less clear back in time in part due to the coarse resolution of the records.

In the previous chapter, we demonstrated that the speleothem-inferred ISM strength is forced by both the NHSI and glacial boundary conditions on orbital timescales, while in the downwind southern China, EASM monsoon strength seems to be insensitive to glacial boundary conditions. We suggested that a detailed spatial-temporal comparison of the speleothem $\delta^{18}\text{O}$ records from both monsoon domains is required to delineate the different monsoon sensitivity in response to changes in insolation and glacial boundary conditions. With the newly resolved speleothem records from SE Yunnan and central Myanmar, in combination with existing records, such a comparison is finally achievable.

Figure 4. 8. The $\delta^{18}\text{O}$ time series and ^{230}Th ages of stalagmite samples from Lin Noe Twin Cave over the last 40 ka. The $\delta^{18}\text{O}$ profiles and dating error bars (2σ) are color coded as shown by the legend in the figure.

Figure 4. 9. The $\delta^{18}\text{O}$ time series and ^{230}Th ages of stalagmite samples from SE Yunnan Cave over the last 180 ka. The $\delta^{18}\text{O}$ profiles and dating error bars (2σ) are color coded as shown by the legend in the figure. Sample A-14 and the Holocene portion of A-15 are corrected for calcite-aragonite fractionation (see discussion). Contemporaneous $\delta^{18}\text{O}$ records from different stalagmites replicate within dating errors, suggesting equilibrium calcite precipitation. The summer (21 July) insolation at 30°N (grey curve) is also plotted for comparison.

4.7.2. Spatial-temporal pattern of the ISM

To understand surface moisture transport from the BoB towards inland China and spatial changes in ISM precipitation, we compared our newly resolved speleothem records with existing records from Mawmluh Cave in North India (Dutt et al., 2015), and Southern Myanmar caves (Chapter III) (Figs. 4.1 and 4.10). These caves are located along the ISM trajectory across a distance about 1,000 km, with the southern Myanmar caves and Mawmluh caves close to the moisture source of the BoB, the LNT Cave in the middle, and SE Yunnan caves near the boundary between the ISM and EASM domains (Ding and Chan, 2005; Wang et al., 2003). To better portrait the spatial pattern of meteoric precipitation $\delta^{18}\text{O}$ values, the speleothem calcite $\delta^{18}\text{O}$ from places other than SMWT were corrected for the temperature-dependent isotope fractionation resulted from differences in cave temperature, based on two assumptions: (1) cave temperature is similar to the annual mean temperature of a nearby city: SE Yunnan Caves, 14.9°C (Kunming); Lin Noe Twin (LNT) Cave, 19.1°C (Taunggyi); Mawmluh Cave, 16.9°C (Shillong); SMWT Caves, 27.5°C (Yangon); and (2) the temperature differences among caves do not change through time (Fig. 4.10). The second assumption, however, is not completely true, because in reality the temperature gradient was likely to be larger particularly during the LGM, due to (1) global meridional temperature gradient was generally larger, and (2) tropical lapse rate was possibly steepened (Loomis et al., 2017). Temperature gradient changes across the region is difficult to

estimate, but should have reached its maximum during glacial maximums. During the LGM, the temperature of Myanmar coast was likely dropped between 2 to 4 °C (Cai et al., 2015; Kudrass et al., 2001; Rashid et al., 2007), whereas, the temperature of SE Yunnan was likely dropped between 4 to 8 °C (Braconnot et al., 2007; Cai et al., 2015; Cook et al., 2011). Therefore, given the small temperature-dependent isotope fractionation factor ($\sim -0.23\text{‰}/^{\circ}\text{C}$ (Kim and O'Neil, 1997; O'Neil et al., 1969)), even a highly unlikely $6^{\circ}\text{C}/1000\text{km}$ increase in temperature gradient during the LGM will account for less than 1.4‰ discrepancy between our corrected SE Yunnan records and coastal records, which is dwarfed by the total speleothem $\delta^{18}\text{O}$ variations (up to $\sim 15\text{‰}$) across the three sites. A more realistic $\sim 2^{\circ}\text{C}/1000\text{km}$ increase in regional temperature gradient during the LGM was implied from a simulation by Cai et al. (2015). Moreover, the amplified temperature gradient during the LGM would cause speleothem $\delta^{18}\text{O}$ values in the downwind region to shift positively. Correction of this effect would result in an even larger speleothem $\delta^{18}\text{O}$ offset between the records. Therefore, we suggest that our second assumption is close to reality or at least provides a conservative estimation. To highlight the one dimension spatial-temporal pattern of speleothem $\delta^{18}\text{O}$ records, after temperature effect corrected, we further calculated the mean $\delta^{18}\text{O}$ values for the three strategic locations for every thousand years (Fig. 4.10).

The precipitation $\delta^{18}\text{O}$ values across the mainland Southeast Asia inferred by the speleothem records show an isotope depletion trend following the SW-NE

monsoon trajectory (Fig. 4.10). This broad pattern of precipitation isotope depletion from coast towards inland is consistent with the progressive rainout or Rayleigh distillation effect on water isotopes. It implies a relatively stable trajectory of monsoon moisture for at least the recent ~40 ka, in a good agreement with climate model simulations (Braconnot et al., 2007).

While the isotope depletion trend persists through time, the offset of speleothem $\delta^{18}\text{O}$ values has shifted through time (Fig. 4.10). The changes in the offsets between speleothem $\delta^{18}\text{O}$ values essentially describe the changes in the precipitation $\delta^{18}\text{O}$ gradient along the ISM trajectory. During the most recent few hundred years, the speleothem $\delta^{18}\text{O}$ values are reduced by ~6‰ across the mainland Southeast Asia. The spatial isotope depletion gradient, ~6‰/1000km northeastward from the coastal BoB to Southwest China, is in an excellent agreement with the modern-day observations (Araguás-Araguás et al., 1998). This ~6‰/1000km gradient is to a large extent determined by the temperature gradient (~12.6°C/1000km) across the catchment, which is due to a combined altitude and latitude effect.

Our records show that the isotope gradient was much larger during the last glacial time (~8-10‰/1000km) relative to today or broadly the Holocene (~5-7‰/1000km), suggesting a stronger-than-today isotope fractionation of precipitation $\delta^{18}\text{O}$ during the last glacial period. This change in isotope gradient through time is particularly important as it reasonably explains the apparently

different sensitivities of speleothem $\delta^{18}\text{O}$ to glacial forcing. Shown in Figure 10, speleothem records from the three locations contain different $\delta^{18}\text{O}$ offsets between the LGM and the present day, with $\sim 3.5\text{‰}$, $\sim 2\text{‰}$ and $< 1\text{‰}$ in coastal Myanmar, central Myanmar and SE Yunnan, respectively. Analyses of individual records could have led to inconsistent conclusions on the $\delta^{18}\text{O}$ sensitivities to climate forcings. For example, the coastal Myanmar speleothem $\delta^{18}\text{O}$ record suggests that the ISM is strongly forced by glacial boundary conditions while the SE Yunnan speleothem records imply that the monsoon seems to be insensitive to glacial forcings. We propose that the ISM is indeed strongly forced by glacial boundary conditions, consistent with previous conjectures (An et al., 2011; Burns et al., 2001; Cai et al., 2015; Prell and Kutzbach, 1987; Rashid et al., 2007). The changes in climate boundary conditions, in particular continental snow/ice coverage and tropical ocean sea surface temperature, can lead to a large negative shift in precipitation $\delta^{18}\text{O}$ from the LGM to present at sites close to the moisture source through the so-called amount effect, such as recorded in cave speleothems in southern Myanmar (a $\sim 3.5\text{‰}$ offset). However, further inland, the offset of speleothem $\delta^{18}\text{O}$ values between the LGM and present becomes gradually muted (e.g., $< 1\text{‰}$ in SE Yunnan), because a larger-than-today isotope gradient along moisture transport resulted in a counterbalancing negative shift in precipitation $\delta^{18}\text{O}$ values during the LGM.

We believe that temperature should be one of primary drivers of the changes of isotope gradient between the glacial time and present. The isotopic distillation during the progressive rainout of moist air mass (i.e. Rayleigh distillation) is temperature dependent. During the last glacial period, the global temperature was much lower than today's, in particular on the continent ($\sim 4^{\circ}\text{C}$ or more), resulting from an enhanced ice albedo and low atmospheric CO_2 (Clark et al., 2009; Denton et al., 2010), and temperature gradient was larger as discussed above. The lower temperature and larger temperature gradient led to a stronger-than-modern isotopic distillation (Annan and Hargreaves, 2013; Jasechko, 2016; Jasechko et al., 2015), and hence progressively lower $\delta^{18}\text{O}$ values of precipitation toward the ISM downstream.

To test if temperature effect alone could account for isotope gradient changes, we calculated the approximate isotope gradient based a pure function of temperature-driven rainout using the Rayleigh equation:

$$R = R_i f^{(\alpha-1)}$$

where R and R_i are $^{18}\text{O}/^{16}\text{O}$ ratio of the remaining water vapor and its initial value, respectively. The fraction of the water vapor remaining (f) is determined by the change in saturated vapor pressure of water, and α is equilibrium liquid-vapor isotope ratio fractionation. Therefore, f and α are temperature dependent. To approximate the mean condition, α is calculated using the average

temperature between Myanmar coast and SE Yunnan. Today, the annual mean temperature is 27.5°C at Myanmar coast and 14.9 °C at SE Yunnan. We assume that during the LGM, Myanmar coast was 4°C cooler (or 23.5°C) and temperature gradient increased modestly by ~2-3°C/1000km (or 8.9 to 7.9°C at SE Yunnan). Thus, the theoretical isotope gradient increased ~1.5-2‰/1000km from today to the LGM. Note that we do not consider here the additional temperature influence on speleothem $\delta^{18}\text{O}$ during calcite precipitation caused by this temperature gradient change, which could cause speleothem $\delta^{18}\text{O}$ in the downwind to shift positively to counterbalance a small portion of rainfall $\delta^{18}\text{O}$ decrease during the LGM. Such effect however can essentially be cancelled if we shift the equivalent amount of $\delta^{18}\text{O}$ change toward negative values when reconstructing rainfall $\delta^{18}\text{O}$ from speleothem records. Similarly, the ocean reservoir effect (Schrag et al., 2002) can also be removed from the comparison of contemporaneous speleothem $\delta^{18}\text{O}$ values from different cave sites.

Such increase is still insufficient to explain the observed ~3‰/1000km changes, indicating that idealized temperature driven Rayleigh distillation is not enough to account for the observed changes. Three other physical processes, known to drive away from idealized Rayleigh distillation and thus influence the continental isotope gradient, are eddy diffusion, moisture recycling through evaporation and moisture recycling through plant transpiration (Wang et al., 2017; Winnick et al., 2014). For a given degree of rainout, eddy transport results in shallower gradients in precipitation isotopes, while moisture recycling

through evaporation leads to steeper gradients. However, their influence in the humid tropics is rather weak (Winnick et al., 2014). Moisture recycling through plant transpiration can also result in shallower gradients (Wang et al., 2017). For example, the isotope gradient across the Amazon Basin is largely dampened due to the strong moisture recycling through rainforest transpiration (Salati et al., 1979; Wang et al., 2017). Evidence shows that during the last glacial period, the isotope gradient across the Amazon Basin increased substantially with reduced water recycling through plant transpiration (Wang et al., 2017). Thus, similar to temperature, plant transpiration can also regulate the isotope offset in rainfall between the last glacial period and present, and therefore, is partially responsible for the observed changes in isotope gradient in the heavily forested ISM region. Indeed, various studies have indicate a reduction in forest in the now-heavily-vegetated Southeast Asia (Bird et al., 2005; Cook and Jones, 2012; Wurster et al., 2010).

In short, the temperature effect together with possibly suppressed plant transpiration effect is likely responsible for an enlarged isotope gradient in rainfall along the moisture trajectory of the ISM during the last glacial period. The larger-than-today isotope gradient in rainfall leads to a gradually smaller speleothem $\delta^{18}\text{O}$ offset between the LGM and present from tropical coasts to inland. We thus further hypothesize that the lack of glacial-interglacial variations in the inland speleothem $\delta^{18}\text{O}$ records are due to strengthened (weakened) isotope fractionation of transported moistures during glacial

(interglacial) periods, which counteracts the forcing of glacial boundary conditions.

Figure 4. 10. Spatial-temporal comparison of speleothem $\delta^{18}\text{O}$ records from Mainland Southeast Asia over the past 40 ka. Purple dots are from SE Yunnan caves (this study); pink dots are from Lin Noe Twin Cave (this study); cyan squares are from Mawmluh Cave (Dutt et al., 2015); and cyan dots are from the southern Myanmar and western Thailand (SMWT) caves (Chapter 3). Records from places other than SMWT were shifted accordingly to account for the oxygen isotopic fractionation that resulted from a difference in cave temperature between the sites and the Yunnan cave in the east (O'Neil et al., 1969). To highlight the one dimension spatial-temporal pattern of speleothem $\delta^{18}\text{O}$ records, we calculated the mean $\delta^{18}\text{O}$ values for the three strategic locations (Coastal BoB close to the moisture source: SMWT and Mawmluh; Middle: LNT; and Inland close to the boundary between ISM and EASM domain: SE Yunnan (Ding and Chan, 2005; Wang et al., 2003) for every thousand years (thick lines) after temperature effect corrected. The light shade areas are the 1σ uncertainty range of the $\delta^{18}\text{O}$ values.

4.7.3. Marine Isotope Stage 5

To further test our hypothesis that transport pathway effect can substantially modulate moisture $\delta^{18}\text{O}$ values, we compared speleothem $\delta^{18}\text{O}$ records from the ISM region with those from southern China, where the records generally have weak glacial-interglacial variations. We focus specifically on MIS 5 and selected four speleothem $\delta^{18}\text{O}$ records from strategic cave locations and with high data resolution for this spatial-temporal comparison. They include records from Tianmen Cave which is located relatively closer to the moisture source and is dominantly influenced by the ISM (Cai et al., 2010; 2012), SE Yunnan caves (this study) and Dongge Cave (Kelly et al., 2006; Yuan et al., 2004) which are separated by the designated boundary of the ISM and EASM (105°E longitude, Wang et al. (2003)), and Sanbao Cave located deep inland in the EASM region (Wang et al., 2008) (Figs. 1 and 11).

Two major spatial features appear in the comparison (Fig. 11). First, the four speleothem records do not show a simple isotope depletion trend towards inland after temperature effect being corrected. This is actually consistent with the pattern of modern-day precipitation (J. Liu et al., 2014), even though the caves are broadly located along the Asian Monsoon trajectory. And second, striking difference exists among the records when comparing their $\delta^{18}\text{O}$ values during MIS 5e to those of the subsequent two interstadials MIS 5a and 5c. The speleothem $\delta^{18}\text{O}$ from the Tianmen and SE Yunnan caves shows prominently

the lowest values during MIS 5e. In Dongge Cave, the speleothems contain similar $\delta^{18}\text{O}$ values during MIS 5e and 5c. Further inland, the $\delta^{18}\text{O}$ values in Sanbao speleothems however become higher during MIS 5e than in MIS 5c and 5a, which is at odds with an enhanced insolation and stronger interglacial climate boundary conditions during the MIS 5e.

Two mechanisms may have contributed to the complex spatial pattern of precipitation $\delta^{18}\text{O}$ observed in the four cave records. For example, Tianmen Cave is located $\sim 4,800\text{m}$ above sea level. The local precipitation $\delta^{18}\text{O}$ is subject to a strong topographic effect and hence is extremely depleted in ^{18}O (Cai et al., 2010; 2012). Then, precipitation over southern China can be brought in remotely from the tropical Indian Ocean by the ISM or sourced adjacently from the South China Sea and Western Pacific (Clemens et al., 2010; Dayem et al., 2010; Johnson et al., 2006). The latter source is isotopically heavier and can result in relatively higher $\delta^{18}\text{O}$ values in Dongge and Sanbao cave records.

However, neither of the above mechanisms can satisfyingly explain the temporal pattern observed in the Sanbao record, which is largely confirmed by a dataset from a nearby cave (Li et al., 2014). The altitude effect should not have a dramatic influence on the temporal variability of speleothem $\delta^{18}\text{O}$, and in fact does not induce the similar change even in the Tianmen record. The relative contribution of isotopically heavier moisture from the Pacific to the annual total precipitation in southern China is suggested to be reduced during

MIS 5e when compared to that during glacial times (Cai et al., 2015). We hereby argue that the relatively positive $\delta^{18}\text{O}$ values in Sanbao speleothems (and to a certain degree in the Dongge record) during MIS 5e can mainly be attributed to weakened isotope distillation from the ocean sources to inland cave sites during the interglacial periods, possibly due to an enhanced plant transpiration effect and hence consistent with our aforementioned hypothesis.

Figure 4. 11. Spatial-temporal comparison of speleothem $\delta^{18}\text{O}$ records from Southern China over MIS 5. Upper left from Tianmen Cave (Cai et al., 2010); lower left from SE Yunnan Caves (this study); upper right from Sanbao Cave (Wang et al., 2008); and low right from Dongge Cave (Kelly et al., 2006). The summer (21 July) insolation at 30°N (grey curve) is also plotted for comparison. The black arrows indicated different temporal patterns of speleothem $\delta^{18}\text{O}$ in different caves during MIS 5.

4.8. Conclusions

We provide a speleothem $\delta^{18}\text{O}$ record covering a large portion of the last 40 ka retrieved from central Myanmar and a speleothem $\delta^{18}\text{O}$ record spanning the last 180 ka retrieved from southeast Yunnan, China. Together with speleothem records in southern Myanmar/western Thailand, we have traced the ISM moisture transport, following the SW-NE transect from southeastern coastal Myanmar to southwestern China. We found that oxygen isotope fractionation in precipitation associated with progressive rainout over the ISM domain increased significantly during glacial times, likely due to much colder temperatures and suppressed plant transpiration effect. By applying a spatial-temporal comparison of speleothem $\delta^{18}\text{O}$ records from both ISM and EASM regions, we have further demonstrated that a stronger isotope fractionation from the ocean source region to cave sites is the main reason why the Chinese speleothem $\delta^{18}\text{O}$ records show unexpected low values during glacial times compared with interglacial periods, such as MIS 5c relative to MIS 5e. We have therefore, further hypothesized that the moisture transport pathway effect is the key to the long-lasting debate on the orbital-scale pace of the AM. During glacial periods, transported moisture has endured stronger isotope fractionation; and on the contrary, a weaker isotope fractionation of rainwater exists during interglacial periods.

Table 4. 1. ^{230}Th dating results for samples from LNT Cave, with errors in 2σ (95% in confidence).

Sample ID	^{238}U (ppb)	^{232}Th (ppt)	$^{230}\text{Th} / ^{232}\text{Th}$ (atomic $\times 10^{-6}$)	$\delta^{234}\text{U}^a$ (measured)	$^{230}\text{Th} / ^{238}\text{U}$ (activity)	^{230}Th Age (yr) (uncorrected)	^{230}Th Age (yr) (corrected)	$\delta^{234}\text{U}_{\text{initial}}^b$ (corrected)	^{230}Th Age (yr BP) (corrected) ^c
LNT01									
LNT01-2mm	33.5 $\pm$ 0.1	4 $\pm$ 4	45864 $\pm$ 46611	416 $\pm$ 3	0.3444 $\pm$ 0.0027	29965 $\pm$ 275	29963 $\pm$ 275	453 $\pm$ 3	29900 $\pm$ 275
LNT01-68mm	51.6 $\pm$ 0.1	28 $\pm$ 8	11303 $\pm$ 3274	455 $\pm$ 3	0.3689 $\pm$ 0.0030	31391 $\pm$ 297	31380 $\pm$ 298	497 $\pm$ 3	31317 $\pm$ 298
LNT01-68mm	7.69 $\pm$ 0.01	261 $\pm$ 5	181 $\pm$ 4	438 $\pm$ 2	0.3735 $\pm$ 0.0014	32285 $\pm$ 147	31615 $\pm$ 496	479 $\pm$ 2	31552 $\pm$ 496
LNT02									
LNT02-2mm	36.6 $\pm$ 0.1	4 $\pm$ 3	56757 $\pm$ 47574	409 $\pm$ 3	0.3453 $\pm$ 0.0016	30233 $\pm$ 176	30231 $\pm$ 176	445 $\pm$ 3	30168 $\pm$ 176
LNT02-5mm	28.66 $\pm$ 0.05	8 $\pm$ 4	23094 $\pm$ 11501	456 $\pm$ 3	0.3713 $\pm$ 0.0030	31600 $\pm$ 298	31595 $\pm$ 298	499 $\pm$ 3	31531 $\pm$ 298
LNT02-20mm	31.1 $\pm$ 0.1	14 $\pm$ 4	15488 $\pm$ 4963	475 $\pm$ 4	0.4088 $\pm$ 0.0040	34747 $\pm$ 408	34738 $\pm$ 408	524 $\pm$ 5	34674 $\pm$ 408
LNT02-40mm	31.5 $\pm$ 0.1	2 $\pm$ 4	114043 $\pm$ 229354	482 $\pm$ 4	0.4272 $\pm$ 0.0034	36381 $\pm$ 354	36380 $\pm$ 354	534 $\pm$ 4	36316 $\pm$ 354
LNT02-65mm	57.5 $\pm$ 0.1	23 $\pm$ 4	18086 $\pm$ 3490	492 $\pm$ 3	0.4377 $\pm$ 0.0036	37132 $\pm$ 368	37124 $\pm$ 368	546 $\pm$ 4	37060 $\pm$ 368
LNT02-95mm	54.3 $\pm$ 0.1	12 $\pm$ 3	33623 $\pm$ 9100	548 $\pm$ 2	0.4569 $\pm$ 0.0025	37341 $\pm$ 245	37337 $\pm$ 245	609 $\pm$ 2	37273 $\pm$ 245
LNT02-103mm	49.6 $\pm$ 0.1	47 $\pm$ 4	7926 $\pm$ 604	550 $\pm$ 3	0.4556 $\pm$ 0.0018	37151 $\pm$ 191	37134 $\pm$ 192	610 $\pm$ 3	37071 $\pm$ 192
LNT03									
LNT03-5mm	93.6 $\pm$ 0.1	1759 $\pm$ 36	17.9 $\pm$ 0.5	537 $\pm$ 2	0.0204 $\pm$ 0.0004	1458 $\pm$ 27	1102 $\pm$ 253	539 $\pm$ 2	1039 $\pm$ 253
LNT03-145mm	126.7 $\pm$ 0.2	1247 $\pm$ 25	41 $\pm$ 1	572 $\pm$ 2	0.0246 $\pm$ 0.0003	1720 $\pm$ 23	1538 $\pm$ 131	574 $\pm$ 2	1475 $\pm$ 131

LNT04									
LNT04-2mm	47.4 ±0.1	354 ±8	427 ±11	776 ±2	0.1933 ±0.0016	12467 ±112	12347 ±140	803 ±3	12283 ±140
LNT04-154mm	34.43 ±0.05	418 ±9	247 ±6	617 ±3	0.1820 ±0.0021	12923 ±161	12707 ±222	640 ±3	12643 ±222
LNT04-154mm	39.0 ±0.1	450 ±10	265 ±6	642 ±3	0.1851 ±0.0016	12939 ±118	12737 ±185	666 ±3	12673 ±185
LNT04-154mm	38.9 ±0.1	450 ±10	265 ±6	646 ±6	0.1854 ±0.0017	12938 ±136	12737 ±196	669 ±7	12673 ±196
LNT05									
LNT05-0mm	65.86 ±0.1	15 ±5	22985 ±7756	480 ±2	0.3167 ±0.0020	25926 ±188	25922 ±188	516 ±2	25858 ±188
LNT05-1mm	68.6 ±0.1	4 ±5	94560 ±104977	593 ±2	0.3456 ±0.0019	26258 ±166	26257 ±166	639 ±2	26193 ±166
LNT05-1mm	64.6 ±0.1	2 ±1	163823 ±90204	590 ±2	0.3458 ±0.0012	26327 ±113	26326 ±113	636 ±2	26260 ±113
LNT05-50mm	85.3 ±0.2	9 ±3	46292 ±14158	426 ±2	0.3110 ±0.0013	26509 ±134	26507 ±134	459 ±2	26441 ±134
LNT05-55mm	57.7 ±0.1	1 ±1	208370 ±95213	447 ±2	0.3168 ±0.0011	26597 ±109	26596 ±109	482 ±2	26530 ±109
LNT05-55mm	46.0 ±0.1	6 ±1	39448 ±7227	477 ±2	0.3294 ±0.0012	27133 ±120	27130 ±120	515 ±2	27064 ±120
LNT05-65mm	43.0 ±0.1	22 ±1	10911 ±435	543 ±2	0.3438 ±0.0013	27080 ±119	27071 ±119	586 ±2	27005 ±119
LNT05-65mm	40.3 ±0.1	24 ±1	9378 ±335	547 ±2	0.3449 ±0.0012	27103 ±111	27092 ±112	590 ±2	27026 ±112
LNT05-93mm	49.0 ±0.1	6 ±3	44682 ±19115	434 ±2	0.3323 ±0.0020	28352 ±197	28349 ±197	470 ±3	28283 ±197
LNT05-100mm	30.7 ±0.1	1 ±1	128001 ±69306	443 ±2	0.3322 ±0.0012	28146 ±119	28145 ±119	479 ±2	28079 ±119
LNT05-100mm	34.8 ±0.1	4 ±1	43750 ±7226	439 ±2	0.3339 ±0.0012	28397 ±122	28394 ±122	476 ±2	28328 ±122
LNT05-125mm	49.0 ±0.1	3 ±1	97233 ±24093	507 ±2	0.3567 ±0.0013	28993 ±128	28991 ±128	551 ±3	28925 ±128
LNT05-130mm	73.0 ±0.1	11 ±3	38628 ±10195	429 ±2	0.3385 ±0.0014	29088 ±143	29085 ±143	465 ±2	29019 ±143
LNT05-156mm	55.1 ±0.1	3 ±1	94066 ±19961	428 ±2	0.3343 ±0.0012	28690 ±129	28689 ±129	464 ±3	28623 ±129
LNT05-156mm	50.9 ±0.1	3 ±1	101415 ±26515	429 ±2	0.3359 ±0.0010	28819 ±107	28818 ±107	466 ±2	28752 ±107
LNT05-181mm	83.7 ±0.1	6 ±3	73775 ±31366	408 ±2	0.3332 ±0.0013	29054 ±136	29053 ±136	443 ±2	28987 ±136

LNT05-201mm	58.0	±0.1	3	±3	116959	±113291	413	±2	0.3473	±0.0015	30331	±160	30330	±160	450	±2	30264	±160
LNT05-202mm	37.7	±0.1	2	±1	96098	±29447	466	±2	0.3593	±0.0011	30199	±122	30198	±122	507	±2	30132	±122
LNT05-202mm	41.1	±0.1	3	±1	86247	±21673	467	±2	0.3627	±0.0011	30495	±112	30494	±112	509	±2	30428	±112
LNT05-214mm	53.6	±0.1	603	±12	519	±11	428	±2	0.3541	±0.0013	30628	±139	30405	±210	466	±3	30339	±210

LNT09U

LNT09U-1mm	134.6	±0.2	749	±15	6	±1	631	±2	0.0020	±0.0003	136	±18	36	±73	631	±2	-28	±73
LNT09U-1mm	92.3	±0.2	610	±12	10	±1	632	±2	0.0041	±0.0003	277	±18	159	±85	632	±2	93	±85
LNT09U-35mm	120.7	±0.3	876	±18	20	±1	588	±2	0.0090	±0.0001	621	±9	487	±95	589	±2	421	±95
LNT09U-35mm	94.6	±0.2	670	±14	19	±1	586	±3	0.0084	±0.0002	577	±11	447	±93	587	±3	381	±93
Hiatus-40mm																		
LNT09U-48mm	126.6	±0.2	686	±14	93	±2	546	±2	0.0306	±0.0002	2176	±18	2074	±74	550	±2	2008	±74
LNT09U-48mm	111.7	±0.3	609	±12	91	±2	548	±2	0.0302	±0.0002	2147	±17	2044	±74	551	±2	1978	±74
LNT09U-75mm	83.2	±0.2	999	±20	46	±1	485	±2	0.0333	±0.0003	2468	±22	2233	±168	488	±2	2167	±168
LNT09U-75mm	84.9	±0.2	1018	±21	44	±1	483	±2	0.0324	±0.0003	2403	±20	2167	±168	486	±2	2101	±168

LNT09B

LNT09B-3mm	37.06	±0.04	690	±14	238	±5	532	±2	0.2686	±0.0013	20785	±111	20438	±269	564	±2	20372	±269
LNT09B-3mm	35.52	±0.05	633	±13	248	±5	536	±2	0.2679	±0.0014	20676	±122	20344	±264	567	±2	20278	±264
LNT09B-31mm	32.65	±0.03	2201	±44	69	±1	518	±2	0.2830	±0.0013	22240	±116	20967	±908	550	±2	20901	±908
LNT09B-31mm	36.56	±0.06	2533	±51	67	±1	516	±3	0.2813	±0.0012	22123	±115	20812	±934	547	±3	20746	±934
LNT09B-56mm	36.25	±0.04	2826	±57	60	±1	517	±2	0.2840	±0.0013	22343	±117	20869	±1050	548	±3	20803	±1050
LNT09B-56mm	35.65	±0.07	2799	±56	59	±1	515	±3	0.2832	±0.0013	22308	±118	20821	±1059	546	±3	20755	±1059
LNT09B-75mm	55.48	±0.08	1043	±21	240	±5	485	±2	0.2736	±0.0012	21963	±108	21601	±277	516	±2	21535	±277

LNT09B-75mm	55.0	±0.1	1059	±21	232	±5	482	±2	0.2711	±0.0012	21791	±111	21420	±285	512	±3	21354	±285
LNT09B-96mm	57.95	±0.09	2103	±42	127	±3	490	±2	0.2800	±0.0011	22441	±102	21745	±503	521	±2	21679	±503
LNT09B-96mm	37.25	±0.06	1413	±28	122	±2	489	±2	0.2801	±0.0010	22472	±98	21744	±524	520	±3	21678	±524

Table 4. 2. ^{230}Th dating results for samples from SE Yunnan Caves, with errors in 2σ (95% in confidence).

Sample ID	^{238}U (ppb)	^{232}Th (ppt)	$^{230}\text{Th} / ^{232}\text{Th}$ (atomic $\times 10^{-6}$)	$\delta^{234}\text{U}$ ^a (measured)	$^{230}\text{Th} / ^{238}\text{U}$ (activity)	^{230}Th Age (yr) (uncorrected)	^{230}Th Age (yr) (corrected)	$\delta^{234}\text{U}_{\text{initial}}$ ^b (corrected)	^{230}Th Age (yr BP) (corrected) ^c
A16									
A16-30mm	91.9 $\pm$ 0.3	3558 $\pm$ 72	18.1 $\pm$ 0.5	189 $\pm$ 4	0.0426 $\pm$ 0.0007	3973 $\pm$ 63	3023 $\pm$ 675	191 $\pm$ 4	2957 $\pm$ 675
A16-70mm	89.7 $\pm$ 0.2	1878 $\pm$ 38	35 $\pm$ 1	180 $\pm$ 3	0.0448 $\pm$ 0.0005	4218 $\pm$ 48	3701 $\pm$ 369	182 $\pm$ 3	3635 $\pm$ 369
A16-3mm	74.6 $\pm$ 0.1	1783 $\pm$ 36	37 $\pm$ 1	184 $\pm$ 2	0.0543 $\pm$ 0.0005	5114 $\pm$ 51	4526 $\pm$ 419	186 $\pm$ 2	4460 $\pm$ 419
A16-254mm	62.6 $\pm$ 0.1	1193 $\pm$ 24	48 $\pm$ 1	178 $\pm$ 2	0.0552 $\pm$ 0.0006	5232 $\pm$ 56	4761 $\pm$ 338	180 $\pm$ 2	4695 $\pm$ 338
A15									
A15-2mm	94.3 $\pm$ 0.3	714 $\pm$ 15	16 $\pm$ 3	263 $\pm$ 4	0.0075 $\pm$ 0.0012	653 $\pm$ 102	479 $\pm$ 160	263 $\pm$ 4	416 $\pm$ 160
A15-24mm	61.70 $\pm$ 0.09	148 $\pm$ 3	330 $\pm$ 10	170 $\pm$ 1	0.0481 $\pm$ 0.0010	4576 $\pm$ 94	4516 $\pm$ 103	173 $\pm$ 1	4450 $\pm$ 103
A15-63.5mm	72.0 $\pm$ 0.1	1513 $\pm$ 30	61 $\pm$ 1	168 $\pm$ 2	0.0780 $\pm$ 0.0010	7522 $\pm$ 103	6999 $\pm$ 384	171 $\pm$ 2	6933 $\pm$ 384
A15-90mm	70.3 $\pm$ 0.1	2159 $\pm$ 43	51 $\pm$ 1	198 $\pm$ 2	0.0943 $\pm$ 0.0007	8927 $\pm$ 73	8181 $\pm$ 533	202 $\pm$ 2	8115 $\pm$ 533
A15-124.5mm	56.10 $\pm$ 0.09	571 $\pm$ 12	166 $\pm$ 3	208 $\pm$ 2	0.1024 $\pm$ 0.0006	9640 $\pm$ 65	9396 $\pm$ 185	213 $\pm$ 2	9330 $\pm$ 185
A15-155mm	55.39 $\pm$ 0.09	451 $\pm$ 9	206 $\pm$ 4	188 $\pm$ 2	0.1016 $\pm$ 0.0006	9725 $\pm$ 63	9526 $\pm$ 154	193 $\pm$ 2	9461 $\pm$ 154
A15-176.5mm	59.63 $\pm$ 0.10	75 $\pm$ 2	2588 $\pm$ 66	297 $\pm$ 2	0.1971 $\pm$ 0.0010	17860 $\pm$ 102	17832 $\pm$ 103	312 $\pm$ 2	17766 $\pm$ 103
A15-185mm	54.02 $\pm$ 0.07	442 $\pm$ 9	476 $\pm$ 10	189 $\pm$ 2	0.2363 $\pm$ 0.0010	24024 $\pm$ 118	23826 $\pm$ 183	202 $\pm$ 2	23761 $\pm$ 183
A15-185mm	54.09 $\pm$ 0.08	525 $\pm$ 11	401 $\pm$ 8	188 $\pm$ 2	0.2363 $\pm$ 0.0008	24041 $\pm$ 95	23806 $\pm$ 192	201 $\pm$ 2	23741 $\pm$ 192
A15-217.5mm	33.80 $\pm$ 0.05	121 $\pm$ 3	1567 $\pm$ 35	273 $\pm$ 2	0.3401 $\pm$ 0.0021	33503 $\pm$ 249	33423 $\pm$ 255	300 $\pm$ 2	33357 $\pm$ 255

A15-242.5mm	36.85	±0.05	483	±10	460	±10	221	±2	0.3655	±0.0020	38353	±262	38047	±340	246	±2	37981	±340
A15-270mm	33.82	±0.05	22	±1	10499	±554	255	±2	0.4161	±0.0022	43288	±288	43273	±288	288	±2	43207	±288
A15-290mm	25.93	±0.03	39	±1	4572	±148	185	±2	0.4165	±0.0024	46621	±349	46585	±350	211	±2	46519	±350
A15-335mm	39.89	±0.06	144	±3	2058	±51	163	±3	0.4504	±0.0034	52743	±546	52654	±549	189	±4	52591	±549
A14																		
A14-2mm	208	±1	74	±2	771	±20	1088	±3	0.0167	±0.0002	874	±12	869	±13	1091	±2	803	±13
A14-9mm	12593	±49	77	±2	84553	±2608	1061	±3	0.0315	±0.0001	1676	±8	1676	±8	1066	±3	1610	±8
A14-30mm	2624	±2	277	±12	5522	±240	1024	±2	0.0354	±0.0003	1921	±15	1919	±15	1030	±2	1853	±15
A14-135mm	4706	±42	49	±1	64482	±1699	954	±7	0.0403	±0.0004	2271	±24	2271	±24	960	±7	2205	±24
BF02																		
BF02-8mm	148.8	±0.1	29	±1	40092	±1126	293	±1	0.4692	±0.0007	48274	±108	48270	±108	335	±2	48204	±108
BF02-33mm	141.6	±0.1	248	±5	5096	±103	301	±1	0.5422	±0.0009	57395	±142	57357	±145	354	±2	57291	±145
BF02-83mm	185.9	±0.2	42	±1	49743	±1179	301	±2	0.6838	±0.0010	78496	±215	78491	±215	376	±2	78425	±215
BF02-155mm	162.8	±0.2	124	±3	16175	±337	312	±1	0.7458	±0.0011	87788	±245	87772	±245	400	±2	87706	±245
BF02-160mm	165.1	±0.2	348	±7	5756	±116	287	±1	0.7353	±0.0010	88672	±239	88627	±241	369	±2	88561	±241
BF02-205mm	174.8	±0.2	112	±2	19828	±416	310	±1	0.7688	±0.0010	92115	±242	92101	±243	402	±2	92035	±243
JJ08																		
JJ08-75mm	85.9	±0.2	1486	±30	29	±1	141	±2	0.0304	±0.0006	2940	±56	2499	±317	142	±2	2433	±317
JJ08-140mm	97	±1	739	±19	72	±2	112	±14	0.0334	±0.0006	3321	±77	3122	±159	113	±14	3056	±159
JJ06																		

JJ06-12mm	24.47	±0.03	1818	±36	148	±3	349	±2	0.6688	±0.0032	72065	±510	70539	±1191	426	±3	70473	±1191
JJ06-20mm	53.32	±0.05	1357	±27	450	±9	181	±1	0.6952	±0.0014	94012	±361	93411	±556	235	±2	93345	±556
JJ06-115mm	26.70	±0.02	1082	±22	321	±6	232	±2	0.7892	±0.0022	106792	±549	105889	±839	312	±2	105823	±839
JJ06-245mm	49.50	±0.06	5832	±117	115	±2	204	±2	0.8216	±0.0017	119229	±544	116528	±1985	284	±3	116462	±1985
JJ06-295mm	56.61	±0.07	1993	±40	415	±8	251	±2	0.8858	±0.0016	126355	±534	125598	±753	357	±3	125532	±753
JJ06-325mm	44.96	±0.04	2037	±41	300	±6	166	±1	0.8242	±0.0015	128102	±526	127031	±919	237	±2	126965	±919
JJ06-455mm	34.46	±0.03	325	±7	1455	±29	176	±1	0.8319	±0.0018	127789	±595	127570	±614	253	±2	127504	±614
JJ06-572mm	44.13	±0.05	950	±19	643	±13	180	±2	0.8396	±0.0018	129138	±627	128639	±716	259	±2	128573	±716
JJ06-655mm	1562	±1	7244	±145	2929	±59	152	±1	0.8240	±0.0014	131202	±530	131092	±535	220	±2	131026	±535
JJ06-723mm	39.29	±0.04	2610	±52	211	±4	154	±2	0.8523	±0.0020	139533	±790	137946	±1365	227	±3	137880	±1365
JJ06-820mm	58.20	±0.05	116	±2	6843	±144	140	±1	0.8300	±0.0016	136162	±600	136113	±601	205	±2	136047	±601
JJ03																		
JJ03-9mm	78.9	±0.1	3000	±60	331	±7	166	±2	0.7641	±0.0025	111990	±718	111085	±957	227	±3	111021	±957
JJ03-22mm	37.84	±0.04	1794	±36	276	±6	100	±2	0.7922	±0.0043	134640	±1482	133429	±1697	146	±3	133365	±1697
JJ03-34mm	48.30	±0.06	350	±7	1790	±38	94	±2	0.7865	±0.0035	134385	±1234	134199	±1239	138	±3	134135	±1239
JJ03-54mm	43.29	±0.05	1498	±30	372	±8	77	±2	0.7801	±0.0035	137191	±1296	136282	±1437	112	±3	136218	±1437
JJ03-79mm	43.07	±0.05	2886	±58	196	±4	80	±2	0.7963	±0.0027	141731	±1104	139975	±1650	119	±3	139911	±1650
JJ03-107mm	35.39	±0.04	1836	±37	254	±5	81	±2	0.7993	±0.0039	142553	±1497	141198	±1761	121	±3	141134	±1761
JJ03-134mm	13.40	±0.01	1141	±23	156	±3	85	±2	0.8078	±0.0074	144528	±2767	142307	±3134	127	±3	142243	±3134
JJ03-182mm	21.31	±0.02	1810	±36	157	±3	85	±2	0.8065	±0.0033	144042	±1325	141826	±2035	126	±3	141762	±2035
JJ03-237mm	20.70	±0.02	1397	±28	199	±4	93	±2	0.8124	±0.0065	143734	±2425	141996	±2685	138	±4	141932	±2685
JJ03-270mm	28.75	±0.03	993	±20	387	±8	88	±2	0.8107	±0.0048	144410	±1812	143519	±1904	133	±3	143455	±1904
JJ03-290mm	17.84	±0.02	1085	±22	222	±5	85	±2	0.8181	±0.0064	148254	±2537	146675	±2739	129	±4	146611	±2739

JJ03-349mm	15.48	±0.02	235	±5	888	±21	89	±2	0.8174	±0.0083	146831	±3144	146441	±3145	134	±4	146377	±3145
JJ03-372mm	19.50	±0.02	1413	±28	194	±4	97	±2	0.8523	±0.0057	157210	±2360	155363	±2663	151	±3	155299	±2663
JJ03-407mm	19.46	±0.02	779	±16	351	±7	97	±2	0.8510	±0.0059	156874	±2451	155857	±2532	150	±3	155793	±2532
JJ03-443mm	22.62	±0.03	591	±12	559	±12	105	±2	0.8870	±0.0053	168870	±2460	168217	±2488	168	±3	168153	±2488
JJ03-468mm	9.56	±0.01	143	±3	954	±27	88	±3	0.8671	±0.0126	167127	±5718	166744	±5705	141	±5	166680	±5705
JJ03-483mm	12.38	±0.01	132	±4	1421	±45	117	±3	0.9169	±0.0138	177233	±6575	176973	±6562	192	±6	176911	±6562

SY10

SY10-5mm	44.3	±0.1	2000	±40	30	±2	826	±4	0.0821	±0.0042	4996	±263	4278	±572	836	±5	4215	±572
SY10-15mm	1422	±3	113064	±2273	40	±1	818	±4	0.1928	±0.0009	12128	±66	10867	±896	843	±4	10801	±896
SY10-26mm	43.8	±0.1	780	±16	213	±5	716	±2	0.2301	±0.0028	15551	±202	15254	±291	747	±2	15189	±291
SY10-38mm	60.9	±0.1	868	±17	313	±6	715	±3	0.2708	±0.0011	18529	±90	18292	±190	752	±3	18226	±190
SY10-85mm	53.2	±0.1	922	±19	225	±5	325	±2	0.2369	±0.0017	21311	±171	20935	±316	344	±3	20870	±316
SY10-238mm	70.3	±0.1	281	±6	1111	±23	476	±2	0.2691	±0.0011	21712	±102	21635	±116	506	±2	21570	±116
SY10-262mm	58.3	±0.2	42	±2	7148	±298	583	±6	0.3139	±0.0020	23776	±190	23763	±190	623	±6	23698	±190
SY10-310mm	52.9	±0.1	467	±10	673	±16	600	±4	0.3604	±0.0037	27380	±329	27224	±346	648	±5	27161	±346

SL10

SL10-6mm	87.0	±0.2	1979	±40	54	±1	176	±3	0.0748	±0.0009	7158	±91	6595	±408	179	±3	6529	±408
SL10-28mm	82.6	±0.1	2716	±55	44	±1	184	±2	0.0885	±0.0008	8457	±79	7649	±577	188	±2	7583	±577
SL10-77mm	94.0	±0.2	2524	±51	54	±1	186	±2	0.0872	±0.0005	8311	±57	7652	±469	190	±2	7586	±469
SL10-94mm	98.7	±0.3	6666	±135	28	±1	182	±3	0.1139	±0.0009	11024	±97	9357	±1184	187	±4	9291	±1184
SL10-120mm	85.6	±0.3	3824	±77	41	±1	181	±3	0.1106	±0.0007	10694	±78	9593	±782	186	±3	9527	±782
SL10-172mm	90.4	±0.3	6240	±126	27	±1	177	±3	0.1116	±0.0007	10842	±76	9128	±1215	181	±3	9062	±1215

SL08									
SL08-33mm	41.9 ±0.2	5885 ±120	29 ±1	201 ±6	0.2476 ±0.0033	25012 ±405	21596 ±2450	214 ±7	21531 ±2450
SL08-58mm	19.4 ±0.1	566 ±12	183 ±5	369 ±4	0.3235 ±0.0069	29027 ±709	28420 ±826	400 ±4	28355 ±826
SL08-71mm	31 ±4	14368 ±291	721 ±17	204 ±6	0.4881 ±0.0061	55769 ±967	55507 ±982	238 ±7	55442 ±982
SL08-78mm	24.8 ±0.1	474 ±10	446 ±12	247 ±10	0.5176 ±0.0093	57328 ±1474	56896 ±1498	290 ±12	56831 ±1498
SL08-84mm	17.7 ±0.1	262 ±6	582 ±17	253 ±3	0.5250 ±0.0100	57994 ±1444	57660 ±1459	298 ±4	57595 ±1459
SL08-239mm	26.4 ±0.1	50 ±2	4894 ±218	232 ±3	0.5641 ±0.0052	65253 ±843	65210 ±843	279 ±4	65145 ±843
SL08-330mm	43.9 ±0.1	795 ±16	503 ±11	205 ±4	0.5525 ±0.0053	65559 ±913	65135 ±957	246 ±5	65070 ±957
XR05									
XR05-3mm	22.29 ±0.04	23 ±2	7029 ±598	224 ±2	0.4388 ±0.0023	47675 ±321	47652 ±321	257 ±2	47587 ±321
XR05-5mm	35.70 ±0.05	389 ±8	703 ±16	250 ±3	0.4645 ±0.0045	49827 ±630	49580 ±653	287 ±4	49517 ±653
XR05-15mm	13.71 ±0.02	127 ±3	1518 ±33	180 ±2	0.8537 ±0.0039	133249 ±1270	133034 ±1276	262 ±3	132968 ±1276
XR05-18mm	23.43 ±0.04	98 ±3	3461 ±102	185 ±2	0.8800 ±0.0047	139767 ±1535	139671 ±1535	275 ±3	139606 ±1535
XR05-105mm	14.48 ±0.03	43 ±2	4800 ±245	138 ±3	0.8559 ±0.0046	145278 ±1756	145207 ±1756	207 ±4	145142 ±1756
XR05-135mm	20.39 ±0.04	34 ±2	8439 ±480	98 ±2	0.8462 ±0.0047	154758 ±1967	154716 ±1966	151 ±3	154651 ±1966
XR05-135mm	22.12 ±0.04	29 ±2	10886 ±768	80 ±2	0.8553 ±0.0045	165040 ±2102	165006 ±2102	128 ±3	164941 ±2102
XR05-154mm	15.12 ±0.02	10 ±1	22976 ±1373	113 ±2	0.8963 ±0.0029	169601 ±1411	169585 ±1411	182 ±3	169519 ±1411
XR05-225mm	17.42 ±0.03	60 ±2	4299 ±162	106 ±2	0.8912 ±0.0046	170183 ±2145	170098 ±2144	171 ±3	170033 ±2144
XR05-472mm	23.68 ±0.03	21 ±1	17803 ±576	127 ±1	0.9367 ±0.0025	181926 ±1353	181904 ±1353	212 ±2	181838 ±1353
XR05-513mm	43.40 ±0.06	105 ±2	6241 ±136	103 ±2	0.9153 ±0.0025	182890 ±1466	182829 ±1466	173 ±3	182763 ±1466
XR05-513mm	44.35 ±0.08	69 ±2	9648 ±221	101 ±2	0.9151 ±0.0032	183633 ±1849	183593 ±1849	170 ±3	183527 ±1849
XR05-540mm	70.8 ±0.2	215 ±5	4874 ±109	88 ±2	0.9002 ±0.0038	182601 ±2059	182523 ±2059	148 ±3	182458 ±2059

Sample ID is given by sample name and sub-sampling depth from its top in mm.

U decay constants: $\lambda_{238}=1.55125\times10^{-10}$ (Jaffey et al., 1971) and $\lambda_{234}=2.82206\times10^{-6}$ (Cheng et al., 2013). Th decay constant:

$\lambda_{230}=9.1705\times10^{-6}$ (Cheng et al., 2013).

^a $\delta^{234}\text{U} = ([^{234}\text{U}/^{238}\text{U}]_{\text{activity}} - 1) \times 1000\text{‰}$.

^b $\delta^{234}\text{U}_{\text{initial}}$ was calculated based on ^{230}Th age (T), i.e., $\delta^{234}\text{U}_{\text{initial}} = \delta^{234}\text{U}_{\text{measured}} \times e^{\lambda_{234} \times T}$.

Corrected ^{230}Th ages assume the initial $^{230}\text{Th}/^{232}\text{Th}$ 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 $^{232}\text{Th}/^{238}\text{U}$ value of 3.8. The errors are arbitrarily assumed to be 50%.

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

Table 4. 3. Stable isotope composition of stalagmite samples from LNT Cave.

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
LNT01				30	30.50	-1.59	-6.96	58	31.10	-0.82	-6.70
3	29.92	-1.56	-5.52	31	30.52	-1.57	-6.66	59	31.12	-1.23	-6.52
4	29.94	-1.26	-5.76	32	30.54	-1.40	-6.45	60	31.15	-1.38	-6.86
5	29.96	-1.28	-5.74	33	30.57	-1.09	-6.69	61	31.17	-1.06	-6.07
6	29.99	-1.33	-5.69	34	30.59	-1.15	-6.63	62	31.19	-1.22	-6.20
7	30.01	-1.44	-5.84	35	30.61	-1.11	-6.65	63	31.21	-1.51	-6.42
8	30.03	-1.50	-6.06	36	30.63	-1.20	-6.70	64	31.23	-1.69	-6.26
9	30.05	-1.40	-5.63	37	30.65	-1.31	-6.85	65	31.25	-1.78	-6.24
10	30.07	-0.91	-4.98	38	30.67	-0.72	-6.61	66	31.27	-2.00	-6.35
11	30.09	-1.19	-5.45	39	30.69	-0.71	-6.85	67	31.30	-1.86	-6.70
12	30.11	-1.22	-5.21	40	30.72	-0.81	-6.24	68	31.32	-2.19	-7.66
13	30.14	-1.85	-5.82	41	30.74	-1.04	-6.36	LNT02			
14	30.16	-1.60	-5.64	42	30.76	-0.80	-6.36	1	30.14	-1.76	-5.76
15	30.18	-1.60	-6.00	43	30.78	-0.75	-6.21	2	30.38	-1.80	-6.17
16	30.20	-1.54	-6.21	44	30.80	-0.77	-6.15	3	30.63	-1.75	-6.06
17	30.22	-1.42	-5.71	45	30.82	-0.50	-5.96	4	30.87	-1.73	-5.61
18	30.24	-1.49	-6.01	46	30.84	-0.75	-6.44	5	31.11	-1.58	-5.44
19	30.26	-1.50	-6.54	47	30.87	-1.14	-6.44	6	31.36	-0.83	-5.68
20	30.29	-1.57	-6.48	48	30.89	-1.20	-7.06	7	31.61	-0.25	-7.35
21	30.31	-1.99	-6.02	49	30.91	-1.25	-6.67	8	31.85	-0.79	-6.37
22	30.33	-2.23	-6.12	50	30.93	-1.02	-6.21	9	32.09	-0.85	-6.24
23	30.35	-2.15	-6.05	51	30.95	-0.99	-6.38	10	32.36	-1.23	-6.97
24	30.37	-2.06	-6.16	52	30.97	-1.04	-6.16	11	32.62	-1.04	-7.26
25	30.39	-2.08	-6.90	53	31.00	-0.88	-5.90	12	32.89	-0.82	-6.63
26	30.42	-1.61	-7.10	54	31.02	-0.50	-6.58	13	33.17	-1.04	-6.92
27	30.44	-1.53	-6.70	55	31.04	-0.89	-7.04	14	33.48	-0.67	-5.85
28	30.46	-1.33	-6.01	56	31.06	-0.95	-7.16	16	34.10	-0.20	-7.24
29	30.48	-1.57	-6.65	57	31.08	-0.90	-6.44	17	34.37	-0.46	-7.11

Table 4.3. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
18	34.56	-0.32	-7.23	48	36.52	0.45	-6.57	77	36.99	-1.33	-7.93
19	34.72	-0.49	-7.23	49	36.54	0.32	-6.46	78	37.00	-0.99	-7.57
20	34.87	-1.24	-7.84	50	36.57	0.09	-6.67	79	37.02	-1.29	-8.25
21	35.02	-1.13	-7.08	51	36.60	-0.44	-6.80	80	37.03	-1.58	-8.50
23	35.27	-1.13	-7.69	52	36.63	-0.73	-6.81	81	37.04	-0.98	-6.81
24	35.41	-0.46	-6.67	53	36.65	-1.15	-6.64	82	37.05	-1.53	-8.50
25	35.52	-0.69	-7.31	55	36.71	-0.44	-6.80	83	37.06	-0.70	-7.65
26	35.56	-0.45	-6.62	56	36.73	-0.80	-6.79	84	37.07	-0.60	-6.93
27	35.57	-0.35	-5.81	57	36.75	-0.22	-6.55	86	37.09	-0.54	-6.34
28	35.61	-0.82	-6.19	58	36.78	-0.47	-7.45	87	37.10	-0.74	-6.49
29	35.72	0.13	-5.69	59	36.80	-0.79	-7.60	88	37.11	-1.21	-7.32
31	36.08	-0.74	-5.83	60	36.83	-0.87	-7.39	89	37.12	-1.50	-7.75
32	36.17	-1.11	-5.94	61	36.84	-0.97	-7.95	92	37.14	-0.85	-7.07
33	36.20	-1.33	-6.66	62	36.86	-1.28	-8.16	93	37.14	-1.54	-7.77
34	36.24	-0.94	-6.07	63	36.87	-0.79	-7.17	95	37.16	-0.74	-6.51
35	36.27	-0.60	-6.01	64	36.88	-0.91	-7.50	96	37.17	-1.11	-7.07
36	36.29	-0.38	-5.74	65	36.88	-0.57	-7.34	97	37.18	-1.44	-7.34
37	36.31	-0.14	-6.10	66	36.89	-0.77	-6.83	99	37.19	-0.77	-7.65
38	36.32	-0.25	-5.98	67	36.90	-1.26	-7.96	100	37.19	-0.76	-7.61
39	36.34	0.24	-5.76	68	36.91	-1.07	-7.63	101	37.20	-0.83	-7.54
40	36.36	-1.16	-6.93	69	36.92	-0.68	-7.29	102	37.21	-0.97	-7.70
41	36.37	-0.27	-6.07	70	36.92	-0.75	-6.81	105	37.22	-1.85	-8.99
42	36.39	-0.91	-6.49	71	36.93	-1.30	-7.63	LNT03			
43	36.41	-0.49	-6.10	72	36.94	-0.89	-7.18	5	1.04	-8.58	-7.16
44	36.43	-0.78	-6.85	73	36.95	-0.73	-7.26	10	1.05	-8.53	-7.44
45	36.45	-1.13	-7.00	74	36.96	-0.77	-7.17	15	1.07	-9.03	-7.30
46	36.47	-1.69	-7.78	75	36.97	-1.12	-7.66	20	1.09	-8.75	-7.36
47	36.49	-0.14	-6.31	76	36.98	-1.16	-7.13	25	1.10	-9.08	-7.75

Table 4.3. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
30	1.12	-9.19	-7.39	2	12.28	-3.58	-6.40	30	12.35	-3.48	-5.86
35	1.13	-8.71	-7.85	3	12.29	-3.46	-6.49	31	12.36	-3.41	-5.94
40	1.15	-7.83	-7.20	4	12.29	-3.09	-5.98	32	12.36	-3.13	-5.67
45	1.16	-8.72	-7.70	5	12.29	-3.28	-6.13	33	12.36	-3.21	-5.65
50	1.18	-9.55	-8.51	6	12.29	-3.18	-6.03	34	12.37	-2.98	-5.55
55	1.20	-8.80	-7.69	7	12.30	-3.22	-5.93	35	12.37	-3.41	-5.68
60	1.21	-9.45	-7.68	8	12.30	-2.97	-5.90	36	12.37	-3.58	-6.50
65	1.23	-9.47	-7.63	9	12.30	-3.07	-6.12	37	12.37	-3.65	-6.72
70	1.24	-9.45	-7.33	10	12.30	-2.95	-6.01	38	12.38	-3.44	-6.29
75	1.26	-9.79	-7.53	11	12.31	-3.02	-6.18	39	12.38	-3.12	-5.95
80	1.27	-9.39	-7.46	12	12.31	-2.94	-5.91	40	12.38	-3.09	-6.19
85	1.29	-9.75	-7.23	13	12.31	-3.08	-6.05	41	12.38	-3.15	-5.97
90	1.30	-9.60	-7.46	14	12.31	-3.07	-5.99	42	12.39	-3.73	-6.18
95	1.32	-9.55	-7.52	15	12.32	-3.49	-6.33	43	12.39	-3.46	-6.27
100	1.34	-9.36	-7.43	16	12.32	-3.47	-6.20	44	12.39	-3.68	-6.80
105	1.35	-10.03	-7.30	17	12.32	-3.61	-6.22	45	12.39	-3.42	-6.44
110	1.37	-9.30	-7.45	18	12.32	-3.39	-6.05	46	12.40	-3.52	-6.00
115	1.38	-9.57	-7.49	19	12.33	-3.26	-6.07	47	12.40	-3.20	-6.44
120	1.40	-9.37	-7.13	20	12.33	-2.91	-6.96	48	12.40	-3.29	-6.36
125	1.41	-9.63	-7.81	21	12.33	-3.46	-6.43	49	12.40	-3.20	-6.53
130	1.43	-9.47	-7.10	22	12.33	-3.40	-6.78	50	12.41	-3.39	-6.70
135	1.44	-9.85	-7.96	23	12.34	-3.21	-6.40	51	12.41	-3.33	-6.36
140	1.46	-9.79	-7.54	24	12.34	-3.09	-6.70	52	12.41	-3.44	-6.48
145	1.48	-9.69	-7.43	25	12.34	-3.23	-6.32	53	12.41	-3.59	-6.45
150	1.49	-9.35	-7.42	26	12.34	-3.28	-6.20	54	12.42	-3.56	-7.09
LNT04				27	12.35	-3.33	-6.42	55	12.42	-3.56	-6.66
0	12.28	-3.21	-6.08	28	12.35	-3.04	-5.85	56	12.42	-3.90	-6.25
1	12.28	-3.52	-6.44	29	12.35	-3.54	-6.26	57	12.42	-3.96	-6.88

Table 4.3. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
58	12.43	-3.48	-5.96	86	12.50	-3.22	-6.30	114	12.57	-3.02	-5.78
59	12.43	-3.68	-6.32	87	12.50	-3.36	-6.17	115	12.57	-3.16	-5.95
60	12.43	-3.57	-6.41	88	12.50	-3.50	-6.29	116	12.58	-3.24	-5.97
61	12.43	-3.74	-6.41	89	12.51	-3.06	-5.83	117	12.58	-3.20	-6.00
62	12.44	-4.19	-6.69	90	12.51	-3.31	-5.95	118	12.58	-3.21	-6.17
63	12.44	-3.84	-6.66	91	12.51	-3.08	-5.75	119	12.58	-3.07	-6.02
64	12.44	-3.70	-6.51	92	12.51	-3.21	-5.90	120	12.59	-3.08	-6.18
65	12.44	-3.72	-6.51	93	12.52	-3.06	-6.29	121	12.59	-2.99	-5.95
66	12.45	-3.65	-6.47	94	12.52	-3.01	-5.97	122	12.59	-3.59	-6.42
67	12.45	-3.84	-6.42	95	12.52	-3.08	-5.87	123	12.59	-3.56	-6.16
68	12.45	-3.87	-6.33	96	12.52	-3.51	-6.25	124	12.60	-3.43	-6.28
69	12.45	-3.78	-6.44	97	12.53	-3.48	-6.43	125	12.60	-3.53	-6.72
70	12.46	-3.72	-6.55	98	12.53	-3.63	-6.76	126	12.60	-3.49	-6.78
71	12.46	-3.49	-6.17	99	12.53	-3.55	-5.89	127	12.60	-3.55	-6.30
72	12.46	-3.53	-6.33	100	12.53	-3.53	-5.83	128	12.61	-3.49	-6.43
73	12.47	-3.84	-6.64	101	12.54	-3.63	-6.06	129	12.61	-3.38	-6.24
74	12.47	-3.68	-6.78	102	12.54	-3.62	-5.92	130	12.61	-2.94	-6.18
75	12.47	-3.60	-6.66	103	12.54	-3.63	-5.72	131	12.61	-3.57	-6.44
76	12.47	-3.42	-6.14	104	12.54	-3.80	-6.14	132	12.62	-3.28	-6.33
77	12.48	-3.40	-6.43	105	12.55	-3.64	-6.32	133	12.62	-3.72	-6.29
78	12.48	-3.73	-6.40	106	12.55	-3.76	-6.68	134	12.62	-3.81	-6.88
79	12.48	-3.15	-6.16	107	12.55	-3.77	-6.58	135	12.62	-3.55	-6.24
80	12.48	-3.37	-5.85	108	12.56	-3.70	-6.39	136	12.63	-3.15	-5.77
81	12.49	-3.14	-6.00	109	12.56	-3.48	-6.17	137	12.63	-3.11	-5.91
82	12.49	-3.53	-6.15	110	12.56	-3.48	-6.07	138	12.63	-3.03	-5.87
83	12.49	-3.19	-6.57	111	12.56	-3.32	-6.17	139	12.63	-3.10	-5.91
84	12.49	-2.93	-6.20	112	12.57	-3.01	-5.80	140	12.64	-2.89	-5.42
85	12.50	-3.13	-6.10	113	12.57	-3.00	-5.71	141	12.64	-3.16	-6.02

Table 4.3. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
142	12.64	-3.09	-5.72	13	26.29	-2.47	-6.02	41	26.41	-4.13	-6.52
143	12.65	-3.08	-5.81	14	26.29	-2.66	-6.14	42	26.42	-4.21	-6.36
144	12.65	-3.15	-5.84	15	26.30	-2.82	-6.78	43	26.42	-4.04	-6.14
145	12.65	-3.15	-5.82	16	26.30	-2.66	-6.81	44	26.42	-4.01	-6.47
146	12.65	-3.20	-5.87	17	26.30	-2.72	-5.45	45	26.42	-4.06	-6.63
147	12.66	-3.71	-6.51	18	26.31	-2.77	-5.63	46	26.42	-4.19	-6.71
148	12.66	-3.94	-6.15	19	26.31	-2.96	-5.77	47	26.43	-4.05	-6.71
149	12.66	-3.61	-6.27	20	26.32	-3.27	-5.92	48	26.43	-4.23	-6.87
150	12.66	-3.50	-6.12	21	26.32	-3.15	-6.53	49	26.43	-4.00	-6.79
151	12.67	-3.75	-6.57	22	26.33	-3.32	-5.93	50	26.43	-4.33	-6.53
152	12.67	-4.18	-6.93	23	26.33	-3.37	-5.94	51	26.44	-4.37	-6.25
153	12.67	-4.06	-6.20	24	26.34	-3.58	-6.16	52	26.45	-3.91	-5.98
154	12.67	-4.39	-6.07	25	26.34	-3.57	-6.26	53	26.46	-3.06	-5.65
155	12.68	-4.06	-5.75	26	26.35	-3.35	-6.47	54	26.47	-3.36	-5.76
156	12.68	-3.80	-5.60	27	26.35	-3.42	-6.10	55	26.74	-3.52	-5.84
LNT05				28	26.36	-3.69	-6.58	56	26.98	-3.49	-6.30
1	26.10	-2.38	-5.92	29	26.36	-3.69	-6.19	57	27.00	-3.50	-6.25
2	26.23	-2.85	-6.10	30	26.37	-3.82	-6.32	58	27.00	-3.75	-6.16
3	26.25	-2.70	-6.38	31	26.37	-3.83	-6.45	59	27.00	-3.91	-6.45
4	26.25	-2.61	-5.79	32	26.38	-3.69	-6.67	60	27.00	-3.88	-6.55
5	26.25	-2.44	-5.50	33	26.38	-3.79	-6.35	61	27.00	-3.69	-5.98
6	26.26	-2.61	-5.86	34	26.38	-3.97	-6.02	62	27.00	-2.05	-5.68
7	26.26	-2.66	-6.07	35	26.39	-4.09	-6.31	63	27.00	-1.90	-5.69
8	26.26	-2.75	-6.62	36	26.39	-4.00	-6.12	64	27.02	-2.95	-6.18
9	26.27	-2.94	-6.46	37	26.40	-3.92	-6.16	65	27.04	-4.25	-6.20
10	26.27	-2.94	-6.39	38	26.40	-3.93	-6.35	66	27.06	-3.72	-6.09
11	26.28	-2.91	-6.49	39	26.41	-3.69	-6.05	67	27.10	-4.17	-6.87
12	26.28	-2.53	-5.88	40	26.41	-4.15	-6.72	68	27.15	-4.12	-6.93

Table 4.3. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
69	27.20	-4.16	-6.11	97	28.17	-1.87	-6.36	125	28.76	-2.70	-5.55
70	27.24	-2.38	-6.17	98	28.18	-2.03	-6.48	126	28.76	-2.53	-5.68
71	27.29	-2.81	-6.62	99	28.19	-2.13	-6.18	127	28.76	-3.10	-6.00
72	27.33	-2.07	-6.66	100	28.20	-1.78	-6.13	128	28.76	-3.35	-6.38
73	27.38	-1.49	-6.09	101	28.23	-2.05	-6.45	129	28.76	-3.51	-6.43
74	27.42	-1.83	-6.78	102	28.26	-2.04	-6.27	130	28.76	-3.51	-6.89
75	27.47	-1.80	-6.18	103	28.28	-1.88	-6.19	131	28.76	-3.15	-6.63
76	27.51	-2.07	-6.49	104	28.31	-2.01	-6.37	132	28.76	-3.37	-7.06
77	27.56	-2.08	-6.39	105	28.34	-2.26	-6.16	133	28.76	-3.63	-7.15
78	27.61	-2.02	-6.41	106	28.37	-2.39	-6.48	134	28.76	-3.25	-6.91
79	27.65	-1.81	-6.29	107	28.41	-2.65	-6.40	135	28.76	-3.40	-6.35
80	27.70	-2.48	-6.49	108	28.44	-3.06	-7.00	136	28.76	-3.32	-6.38
81	27.75	-2.00	-6.67	109	28.47	-3.18	-6.55	137	28.76	-3.44	-6.35
82	27.80	-2.21	-7.02	110	28.50	-3.43	-6.60	138	28.76	-3.18	-6.17
83	27.84	-1.99	-6.87	111	28.52	-3.25	-6.44	139	28.76	-3.53	-5.95
84	27.89	-2.12	-7.14	112	28.55	-3.32	-6.43	140	28.76	-3.31	-6.20
85	27.94	-2.00	-7.82	113	28.58	-3.53	-6.46	141	28.76	-3.25	-6.30
86	27.98	-1.90	-7.60	114	28.61	-3.31	-6.54	142	28.76	-3.47	-6.35
87	28.03	-1.96	-7.33	115	28.64	-3.14	-6.91	143	28.76	-3.37	-6.19
88	28.07	-1.79	-7.71	116	28.67	-3.03	-6.55	144	28.76	-3.19	-6.19
89	28.11	-1.36	-6.58	117	28.70	-2.80	-6.11	145	28.76	-3.37	-6.13
90	28.13	-2.03	-6.98	118	28.73	-2.95	-6.34	146	28.76	-3.25	-6.01
91	28.15	-2.25	-6.58	119	28.74	-2.29	-6.19	147	28.76	-3.18	-6.03
92	28.16	-2.25	-6.05	120	28.75	-2.27	-6.05	148	28.76	-3.28	-5.98
93	28.17	-2.01	-6.35	121	28.75	-2.74	-6.11	149	28.76	-3.44	-6.37
94	28.17	-2.16	-6.21	122	28.76	-2.90	-6.66	150	28.76	-3.48	-6.53
95	28.17	-1.91	-6.05	123	28.76	-2.88	-6.22	151	28.76	-3.22	-6.56
96	28.17	-1.87	-6.61	124	28.76	-3.16	-6.40	152	28.76	-3.49	-6.65

Table 4.3. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
153	28.76	-3.05	-6.74	181	28.99	-2.50	-6.51	209	30.33	-1.61	-6.25
154	28.76	-3.45	-6.56	182	29.02	-2.53	-6.26	210	30.35	-1.59	-6.50
155	28.76	-3.62	-6.83	183	29.06	-2.51	-6.62	211	30.36	-1.70	-6.68
156	28.76	-3.58	-6.47	184	29.10	-2.77	-6.52	212	30.37	-1.85	-6.50
157	28.77	-3.89	-6.49	185	29.14	-2.67	-6.92	213	30.39	-1.93	-6.76
158	28.77	-3.67	-6.62	186	29.18	-2.72	-6.81	214	30.40	-1.59	-6.44
159	28.77	-3.78	-6.75	187	29.22	-2.54	-7.12	215	30.41	-1.61	-6.27
160	28.77	-3.52	-6.70	188	29.28	-1.76	-6.44	LNT09U			
161	28.77	-3.31	-6.83	189	29.37	-1.64	-6.35	0	-0.04	-11.23	-7.61
162	28.78	-3.30	-7.14	190	29.46	-1.30	-5.59	1	-0.03	-10.68	-7.30
163	28.78	-3.28	-6.68	191	29.53	-1.70	-5.35	2	-0.02	-10.78	-7.15
164	28.79	-3.61	-6.66	192	29.60	-1.86	-5.25	3	0.00	-10.63	-7.07
165	28.79	-3.57	-6.78	193	29.66	-1.78	-5.65	4	0.01	-10.39	-7.11
166	28.81	-3.54	-6.44	194	29.73	-1.21	-5.25	5	0.02	-10.63	-7.49
167	28.82	-3.48	-6.50	195	29.79	-1.21	-5.18	6	0.03	-10.28	-7.38
168	28.83	-3.41	-6.78	196	29.86	-1.43	-5.34	7	0.04	-10.27	-7.23
169	28.84	-3.37	-6.75	197	29.92	-1.09	-4.90	8	0.06	-10.31	-7.10
170	28.85	-3.35	-6.80	198	29.98	-1.59	-5.33	9	0.07	-10.09	-7.10
171	28.86	-3.59	-7.04	199	30.04	-1.37	-5.65	10	0.08	-10.55	-7.27
172	28.88	-3.21	-6.77	200	30.10	-1.34	-5.74	11	0.09	-10.39	-7.31
173	28.89	-3.32	-6.48	201	30.17	-1.39	-5.53	12	0.10	-10.45	-7.20
174	28.89	-2.95	-6.46	202	30.23	-1.51	-6.09	13	0.12	-10.20	-7.37
175	28.89	-3.00	-6.12	203	30.26	-1.59	-6.17	14	0.13	-10.31	-7.73
176	28.90	-2.90	-6.12	204	30.27	-1.77	-6.47	15	0.14	-10.43	-7.94
177	28.90	-3.07	-6.42	205	30.29	-1.86	-6.67	16	0.15	-10.18	-7.97
178	28.90	-2.85	-6.28	206	30.30	-1.88	-6.58	17	0.16	-9.90	-7.78
179	28.92	-2.81	-6.04	207	30.31	-1.84	-6.69	18	0.18	-9.99	-7.68
180	28.95	-2.61	-6.64	208	30.32	-1.76	-6.64	19	0.19	-9.62	-7.53

Table 4.3. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
20	0.20	-9.53	-7.79	47	2.00	-9.19	-7.75	75	2.10	-8.66	-8.13
21	0.21	-9.29	-7.61	48	2.01	-9.16	-7.97	76	2.10	-8.57	-7.66
22	0.23	-9.67	-7.67	49	2.01	-8.72	-8.15	77	2.11	-8.12	-7.51
23	0.24	-9.89	-7.76	50	2.01	-9.04	-8.21	78	2.11	-7.01	-7.62
24	0.25	-9.15	-7.78	51	2.02	-8.45	-8.04	LNT09B			
25	0.26	-9.47	-7.71	52	2.02	-8.57	-7.63	1	20.29	-10.19	-7.93
26	0.27	-9.46	-7.71	53	2.03	-9.00	-7.58	2	20.31	-2.57	-5.74
27	0.29	-9.65	-7.88	54	2.03	-8.58	-7.43	3	20.32	-2.81	-5.70
28	0.30	-9.69	-7.99	55	2.03	-7.99	-7.48	4	20.33	-2.69	-5.75
29	0.31	-9.77	-7.88	56	2.04	-8.60	-7.23	5	20.34	-2.71	-5.24
30	0.32	-9.87	-7.73	57	2.04	-8.43	-7.35	6	20.36	-2.73	-5.41
31	0.33	-10.13	-7.52	58	2.04	-7.71	-6.94	7	20.37	-2.64	-5.45
32	0.35	-9.83	-7.55	59	2.05	-8.84	-7.47	8	20.39	-2.75	-5.34
33	0.36	-10.49	-7.93	60	2.05	-9.07	-7.83	9	20.40	-2.52	-5.38
34	0.37	-10.26	-7.94	61	2.05	-8.99	-7.89	10	20.42	-2.60	-5.94
35	0.38	-9.96	-7.61	62	2.06	-8.77	-7.92	11	20.43	-2.78	-5.65
36	0.39	-10.06	-7.47	63	2.06	-8.90	-7.75	12	20.45	-2.80	-5.75
37	0.41	-9.48	-7.80	64	2.06	-9.04	-7.81	13	20.46	-2.90	-6.36
38	0.42	-9.46	-7.67	65	2.07	-8.84	-7.93	14	20.48	-3.13	-5.65
39	0.43	-9.69	-7.71	66	2.07	-8.50	-8.07	15	20.49	-2.93	-5.60
hiatus				67	2.07	-8.40	-7.84	16	20.51	-2.99	-5.82
40	1.98	-9.87	-7.71	68	2.08	-8.31	-7.64	17	20.52	-2.91	-5.48
41	1.98	-9.06	-7.63	69	2.08	-8.16	-7.74	18	20.54	-2.46	-4.98
42	1.99	-9.45	-7.61	70	2.08	-8.86	-7.87	19	20.55	-2.22	-5.32
43	1.99	-9.70	-7.50	71	2.09	-8.88	-8.13	20	20.56	-2.08	-5.15
44	1.99	-9.18	-7.31	72	2.09	-8.32	-7.88	21	20.58	-2.93	-5.68
45	2.00	-9.05	-7.17	73	2.09	-8.47	-7.50	22	20.59	-2.92	-5.25
46	2.00	-9.15	-7.58	74	2.10	-8.24	-7.75	23	20.61	-2.63	-4.76

Table 4.3. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
24	20.62	-2.74	-5.44	52	21.00	-2.71	-5.70	80	21.47	-2.63	-5.60
25	20.63	-2.98	-5.78	53	21.02	-2.92	-5.93	81	21.48	-2.76	-5.52
26	20.65	-2.83	-5.36	54	21.03	-3.10	-5.69	82	21.50	-3.26	-5.86
27	20.66	-2.67	-5.05	55	21.05	-2.88	-5.71	83	21.51	-3.37	-5.72
28	20.68	-2.72	-4.69	56	21.07	-2.84	-5.52	84	21.53	-3.67	-5.95
29	20.69	-2.89	-5.72	57	21.08	-2.71	-5.74	85	21.55	-3.54	-5.45
30	20.71	-2.51	-5.25	58	21.10	-2.57	-5.22	86	21.56	-3.48	-6.08
31	20.72	-2.62	-4.96	59	21.12	-2.57	-5.48	87	21.58	-3.54	-6.06
32	20.73	-2.84	-5.30	60	21.13	-2.68	-5.61	88	21.60	-3.64	-5.86
33	20.73	-2.74	-5.80	61	21.15	-2.51	-6.40	89	21.61	-3.43	-5.75
34	20.74	-2.72	-5.46	62	21.16	-3.09	-5.73	90	21.63	-3.34	-5.52
35	20.76	-2.51	-5.33	63	21.18	-3.06	-6.30	91	21.64	-2.91	-5.67
36	20.77	-2.97	-5.58	65	21.22	-3.39	-5.81	92	21.66	-3.17	-5.87
37	20.79	-2.16	-5.03	65	21.22	-3.18	-5.79	93	21.67	-3.17	-5.85
38	20.80	-2.13	-5.40	66	21.24	-3.38	-5.72				
39	20.82	-2.26	-5.42	67	21.25	-3.47	-5.43				
40	20.83	-2.25	-5.44	68	21.27	-3.64	-5.55				
41	20.85	-2.22	-5.24	69	21.29	-3.72	-5.89				
42	20.86	-2.17	-5.37	70	21.31	-3.77	-5.97				
43	20.87	-2.22	-5.13	71	21.33	-3.31	-5.72				
44	20.89	-1.56	-5.03	72	21.35	-3.38	-6.02				
45	20.90	-1.85	-5.19	73	21.36	-3.41	-5.90				
46	20.92	-2.14	-5.22	74	21.38	-3.22	-5.32				
47	20.93	-2.43	-5.38	75	21.40	-3.13	-5.47				
48	20.94	-2.50	-5.68	76	21.41	-3.15	-5.62				
49	20.96	-2.56	-5.46	77	21.42	-2.84	-5.49				
50	20.97	-2.75	-5.52	78	21.43	-2.88	-5.84				
51	20.99	-3.14	-5.90	79	21.45	-3.04	-5.85				

Table 4. 4. Stable isotope data for stalagmite sample from SE Yunnan Caves.

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
A14				40	1.89	-7.82	-8.97	96	2.07	-8.62	-10.09
aragonite				42	1.89	-7.84	-9.08	98	2.08	-8.53	-9.97
0	0.57	-10.08	-9.85	44	1.90	-8.38	-9.31	100	2.09	-8.62	-9.90
2	0.80	-11.03	-9.64	46	1.91	-8.42	-9.38	102	2.09	-8.82	-9.58
3	0.92	-11.17	-9.74	48	1.91	-8.49	-9.41	104	2.10	-8.89	-9.28
4	1.03	-11.35	-9.63	50	1.92	-8.60	-9.51	106	2.11	-8.48	-9.09
5	1.15	-10.75	-9.46	52	1.93	-8.52	-9.42	108	2.11	-8.12	-9.14
6	1.26	-11.79	-10.07	54	1.93	-8.29	-9.57	110	2.12	-8.02	-9.41
7	1.38	-11.45	-9.90	56	1.94	-8.16	-9.49	112	2.13	-8.09	-9.61
8	1.49	-10.87	-9.87	58	1.95	-8.22	-9.53	114	2.13	-8.19	-9.46
9	1.61	-9.52	-9.22	60	1.95	-7.73	-9.37	116	2.14	-7.50	-9.01
10	1.62	-9.46	-8.86	62	1.96	-7.85	-9.93	118	2.15	-8.08	-9.03
11	1.63	-9.86	-9.14	64	1.97	-8.06	-10.03	120	2.15	-7.95	-9.09
12	1.64	-9.95	-9.50	66	1.97	-8.25	-9.87	122	2.16	-8.38	-9.27
13	1.66	-9.80	-9.39	68	1.98	-8.31	-9.67	126	2.17	-8.25	-9.20
14	1.67	-9.20	-9.68	70	1.99	-8.95	-9.83	128	2.18	-8.18	-9.04
16	1.69	-9.31	-9.76	72	1.99	-8.70	-9.22	130	2.19	-8.28	-9.08
18	1.71	-9.00	-9.67	74	2.00	-8.66	-9.34	132	2.19	-7.98	-9.23
20	1.74	-8.98	-9.64	76	2.01	-8.55	-9.34	134	2.20	-8.25	-9.68
22	1.76	-9.06	-9.74	78	2.01	-8.52	-9.57	136	2.21	-8.39	-9.65
24	1.78	-8.59	-9.65	80	2.02	-8.68	-9.77	138	2.22	-7.56	-9.12
26	1.81	-8.27	-9.87	82	2.03	-8.48	-9.89	140	2.22	-8.47	-9.38
28	1.83	-8.13	-9.70	84	2.03	-8.51	-9.84	A15			
30	1.85	-7.99	-9.17	86	2.04	-8.62	-9.52	aragonite			
32	1.86	-7.88	-9.28	88	2.05	-8.59	-9.40	1	0.23	-6.97	-9.24
34	1.87	-7.95	-9.02	90	2.05	-8.15	-9.29	2	0.42	-7.39	-9.76
36	1.87	-7.75	-9.15	92	2.06	-8.50	-9.56	3	0.60	-8.79	-10.12
38	1.88	-7.65	-9.13	94	2.07	-8.38	-9.78	4	0.78	-9.21	-10.86

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
5	0.97	-7.78	-9.89	33	5.02	-7.58	-9.70	61	6.78	-6.24	-11.11
6	1.15	-7.77	-9.99	34	5.08	-8.18	-9.80	62	6.84	-6.16	-11.19
7	1.33	-8.17	-10.32	35	5.14	-8.40	-9.82	63	6.90	-6.50	-11.29
8	1.52	-8.92	-10.83	36	5.20	-8.33	-9.90	64	6.95	-6.76	-10.83
9	1.70	-7.56	-10.22	37	5.27	-8.36	-10.35	65	7.00	-5.98	-11.21
10	1.88	-6.99	-9.81	38	5.33	-8.64	-10.41	66	7.04	-5.95	-11.54
11	2.07	-7.61	-10.27	39	5.39	-6.66	-10.84	67	7.09	-6.26	-11.70
12	2.25	-6.47	-10.14	40	5.46	-5.99	-10.76	68	7.13	-5.93	-11.53
13	2.43	-5.17	-9.91	41	5.52	-5.68	-10.56	69	7.18	-6.90	-11.38
14	2.62	-5.18	-9.76	42	5.58	-5.38	-10.02	70	7.22	-6.08	-11.57
15	2.80	-6.89	-9.89	43	5.64	-5.37	-10.30	71	7.27	-6.50	-12.01
16	2.98	-6.72	-9.71	44	5.71	-6.71	-10.39	72	7.31	-6.24	-12.26
17	3.17	-7.04	-9.48	45	5.77	-6.40	-10.58	73	7.36	-6.31	-11.91
18	3.35	-6.88	-9.51	46	5.83	-9.47	-10.73	74	7.40	-6.40	-11.77
19	3.53	-6.76	-9.23	47	5.90	-5.82	-10.44	75	7.45	-6.99	-11.46
20	3.72	-6.65	-9.49	48	5.96	-5.74	-10.78	76	7.49	-6.11	-11.94
21	3.90	-7.07	-9.45	49	6.02	-5.72	-11.06	77	7.53	-6.38	-12.49
22	4.08	-7.10	-9.64	50	6.08	-5.47	-10.32	78	7.58	-6.26	-11.84
23	4.27	-8.94	-10.52	51	6.15	-5.78	-10.47	79	7.62	-6.38	-11.44
24	4.45	-10.11	-10.71	52	6.21	-5.60	-10.64	80	7.67	-6.26	-12.09
25	4.51	-5.97	-10.12	53	6.27	-5.32	-10.62	81	7.71	-7.00	-11.64
26	4.58	-6.14	-10.11	54	6.34	-5.41	-10.58	82	7.76	-6.67	-11.86
27	4.64	-5.97	-10.35	55	6.40	-5.38	-10.85	83	7.80	-6.42	-11.88
28	4.70	-7.37	-10.35	56	6.46	-5.96	-11.14	84	7.85	-6.11	-11.81
29	4.76	-6.70	-9.95	57	6.52	-5.46	-11.04	85	7.89	-5.53	-12.25
30	4.83	-7.46	-10.12	58	6.59	-5.66	-10.81	86	7.94	-5.87	-12.14
31	4.89	-7.48	-9.64	59	6.65	-5.55	-11.15	87	7.98	-6.48	-12.05
32	4.95	-7.52	-10.13	60	6.71	-5.37	-11.03	88	8.03	-6.65	-12.53

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
89	8.07	-6.52	-12.03	117	9.07	-5.50	-11.53	145	9.42	-8.73	-13.86
90	8.11	-6.71	-12.00	118	9.10	-7.57	-12.44	146	9.42	-8.12	-12.64
91	8.15	-6.23	-12.67	119	9.14	-6.03	-12.44	147	9.43	-7.42	-12.45
92	8.18	-6.66	-12.39	120	9.17	-5.71	-12.05	148	9.43	-7.59	-12.85
93	8.22	-6.54	-12.25	121	9.21	-7.25	-12.26	149	9.44	-7.60	-12.83
94	8.26	-5.32	-12.29	122	9.24	-7.86	-12.19	150	9.44	-6.44	-11.99
95	8.29	-5.97	-11.67	123	9.28	-8.20	-12.82	151	9.44	-7.28	-12.46
96	8.33	-6.26	-12.11	124	9.31	-7.53	-12.94	152	9.45	-8.94	-13.07
97	8.36	-7.06	-11.91	125	9.33	-7.00	-12.70	153	9.45	-8.31	-13.08
98	8.40	-7.03	-13.20	126	9.34	-5.52	-12.23	154	9.46	-5.79	-10.53
99	8.43	-6.21	-12.28	127	9.34	-5.61	-11.99	A15			
100	8.47	-6.57	-11.61	128	9.34	-7.29	-12.75	180	22.28	-8.72	-9.81
101	8.50	-6.62	-11.82	129	9.35	-6.08	-12.24	181	22.58	-7.52	-9.35
102	8.54	-6.08	-11.61	130	9.35	-7.65	-12.22	182	22.88	-4.67	-8.97
103	8.57	-5.90	-11.64	131	9.36	-7.67	-13.09	183	23.17	-9.39	-9.35
104	8.61	-7.16	-12.41	132	9.36	-7.08	-12.91	184	23.47	-8.26	-9.50
105	8.64	-7.24	-12.43	133	9.37	-7.09	-12.97	185	23.76	-7.93	-10.13
106	8.68	-8.17	-12.77	134	9.37	-7.26	-13.24	186	24.06	-6.26	-9.83
107	8.71	-7.77	-12.60	135	9.37	-7.71	-12.64	187	24.35	-9.72	-10.14
108	8.75	-6.37	-12.52	136	9.38	-8.57	-13.65	188	24.65	-7.12	-9.98
109	8.78	-7.11	-12.02	137	9.38	-8.53	-12.92	189	24.94	-7.61	-9.62
110	8.82	-7.20	-13.13	138	9.39	-8.54	-12.48	190	25.24	-5.85	-9.34
111	8.85	-5.62	-12.47	139	9.39	-9.23	-12.72	191	25.53	-7.02	-9.61
112	8.89	-5.88	-12.30	140	9.40	-9.20	-12.73	192	25.83	-7.98	-9.79
113	8.92	-5.71	-11.78	141	9.40	-9.49	-12.80	193	26.12	-7.21	-9.23
114	8.96	-6.79	-12.27	142	9.41	-9.09	-13.55	194	26.42	-6.79	-9.06
115	9.00	-7.50	-11.88	143	9.41	-7.69	-12.83	195	26.71	-8.44	-10.00
116	9.03	-5.38	-12.06	144	9.41	-8.68	-12.93	196	27.01	-5.31	-9.05

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
197	27.30	-6.34	-9.86	225	34.74	-7.04	-10.73	253	39.98	-10.14	-11.49
198	27.60	-5.47	-9.68	226	34.93	-7.77	-10.31	254	40.17	-10.43	-11.23
199	27.89	-5.60	-9.56	227	35.11	-7.05	-9.94	255	40.36	-10.80	-11.86
200	28.19	-9.05	-10.77	228	35.30	-6.33	-9.72	256	40.55	-10.61	-12.14
201	28.49	-9.84	-10.90	229	35.48	-8.15	-10.75	257	40.74	-10.74	-11.47
202	28.78	-8.47	-10.57	230	35.67	-7.19	-10.30	258	40.93	-10.61	-11.04
203	29.08	-8.12	-10.38	231	35.85	-7.72	-10.59	259	41.12	-10.74	-11.38
204	29.37	-8.70	-10.50	232	36.04	-7.99	-10.30	260	41.31	-10.50	-12.09
205	29.67	-8.33	-10.35	233	36.22	-9.01	-10.53	261	41.50	-10.37	-11.92
206	29.96	-8.26	-10.40	234	36.41	-8.46	-10.14	262	41.69	-10.26	-12.07
207	30.26	-8.75	-10.70	235	36.59	-8.55	-10.55	263	41.88	-7.79	-10.95
208	30.55	-8.33	-10.53	236	36.78	-9.96	-11.27	264	42.07	-7.54	-10.49
209	30.85	-8.25	-10.54	237	36.96	-9.71	-11.79	265	42.26	-6.45	-10.18
210	31.14	-8.33	-10.29	238	37.15	-8.97	-11.07	266	42.45	-4.63	-8.39
211	31.44	-7.37	-10.21	239	37.33	-8.66	-10.83	267	42.64	-6.42	-8.83
212	31.73	-6.94	-10.33	240	37.52	-9.23	-10.85	268	42.83	-5.82	-8.45
213	32.03	-9.60	-10.80	241	37.70	-6.41	-10.25	269	43.02	-7.32	-8.33
214	32.32	-6.64	-9.92	242	37.89	-6.44	-10.40	270	43.21	-6.29	-8.33
215	32.62	-5.72	-9.30	243	38.08	-5.09	-9.26	271	43.37	-8.10	-8.47
216	32.91	-6.17	-9.02	244	38.27	-5.95	-9.56	272	43.54	-7.00	-8.13
217	33.21	-5.20	-8.50	245	38.46	-7.85	-10.53	273	43.70	-6.86	-9.12
218	33.45	-6.05	-9.98	246	38.65	-8.86	-10.93	274	43.87	-4.82	-9.04
219	33.63	-6.40	-10.28	247	38.84	-9.97	-11.10	275	44.04	-6.40	-9.88
220	33.82	-7.54	-10.65	248	39.03	-9.81	-11.57	276	44.20	-5.74	-10.14
221	34.00	-8.65	-10.60	249	39.22	-9.53	-11.33	277	44.37	-5.56	-10.36
222	34.19	-7.96	-10.49	250	39.41	-9.48	-11.21	278	44.53	-8.32	-10.32
223	34.37	-6.14	-10.42	251	39.60	-9.32	-11.60	279	44.70	-5.72	-10.18
224	34.56	-7.62	-11.01	252	39.79	-9.65	-11.76	280	44.86	-6.53	-10.37

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
281	45.03	-8.20	-10.69	309	49.08	-5.49	-9.67	A16			
282	45.19	-6.66	-10.08	310	49.22	-5.70	-9.69	0	2.45	-8.01	-9.91
283	45.36	-7.87	-10.28	311	49.35	-5.80	-9.73	2	2.48	-8.19	-9.84
284	45.53	-9.98	-10.49	312	49.49	-5.80	-9.64	4	2.52	-8.24	-10.22
285	45.69	-6.37	-9.30	313	49.62	-5.62	-9.54	6	2.55	-8.37	-10.51
286	45.86	-6.87	-10.04	314	49.76	-5.52	-9.58	8	2.58	-8.68	-10.27
287	46.02	-7.61	-12.20	315	49.89	-5.63	-9.61	10	2.62	-8.28	-9.98
288	46.19	-9.51	-11.49	316	50.03	-6.12	-9.88	12	2.65	-8.96	-9.88
289	46.35	-9.29	-11.35	317	50.16	-7.00	-9.95	14	2.69	-9.26	-10.68
290	46.52	-10.41	-10.98	318	50.30	-7.38	-10.56	16	2.72	-9.60	-10.65
291	46.65	-3.91	-10.05	319	50.43	-7.36	-10.45	18	2.75	-9.44	-10.64
292	46.79	-7.38	-11.45	320	50.57	-7.65	-10.52	20	2.79	-10.10	-11.07
293	46.92	-5.13	-9.21	321	50.70	-7.80	-10.55	22	2.82	-8.49	-10.36
294	47.06	-4.71	-9.34	322	50.84	-5.60	-9.88	24	2.85	-8.51	-10.19
295	47.19	-6.20	-10.29	323	50.97	-4.17	-9.55	26	2.89	-9.05	-10.60
296	47.33	-7.02	-10.50	324	51.11	-4.51	-9.30	28	2.92	-9.06	-10.13
297	47.46	-4.39	-9.74	325	51.24	-4.10	-9.26	30	2.96	-10.45	-10.85
298	47.60	-4.63	-9.43	326	51.38	-5.34	-8.63	32	2.99	-9.34	-11.45
299	47.73	-6.81	-9.78	327	51.51	-8.03	-9.16	34	3.02	-9.16	-11.22
300	47.87	-6.67	-10.03	328	51.65	-9.27	-9.37	36	3.06	-11.14	-11.69
301	48.00	-6.44	-9.97	329	51.78	-8.01	-9.16	38	3.09	-9.42	-11.25
302	48.14	-4.79	-9.11	330	51.92	-5.45	-8.22	40	3.13	-10.61	-11.16
303	48.27	-6.07	-9.41	331	52.05	-5.72	-8.32	42	3.16	-11.07	-10.76
304	48.41	-6.16	-9.40	332	52.19	-7.41	-8.18	44	3.19	-8.69	-10.98
305	48.54	-5.10	-9.57	333	52.32	-7.04	-8.51	46	3.23	-9.88	-10.79
306	48.68	-3.79	-9.18	334	52.46	-8.25	-8.71	48	3.26	-10.97	-11.56
307	48.81	-4.07	-9.29	335	52.59	-8.17	-8.32	50	3.30	-10.63	-11.90
308	48.95	-5.09	-9.79					52	3.33	-10.43	-10.72

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
54	3.36	-10.14	-11.53	112	4.00	-10.33	-12.15	172	4.48	-10.45	-11.56
56	3.40	-9.28	-11.22	114	4.02	-10.75	-10.73	176	4.49	-10.45	-11.48
58	3.43	-10.54	-12.96	116	4.04	-9.81	-11.17	178	4.50	-10.53	-11.24
60	3.47	-9.86	-10.66	118	4.06	-8.83	-10.42	180	4.50	-10.84	-11.89
62	3.50	-9.94	-11.16	120	4.07	-9.67	-11.17	182	4.51	-10.92	-10.92
64	3.53	-9.97	-11.00	122	4.09	-10.97	-11.81	184	4.51	-10.62	-11.44
66	3.57	-10.30	-10.17	124	4.11	-10.44	-11.06	186	4.52	-10.17	-11.18
68	3.60	-10.09	-10.76	126	4.13	-9.71	-11.43	188	4.52	-9.76	-11.62
70	3.63	-9.56	-11.73	128	4.14	-11.55	-12.19	190	4.53	-10.33	-12.07
72	3.65	-10.58	-10.85	130	4.16	-10.95	-12.17	192	4.53	-10.47	-12.78
74	3.67	-10.49	-11.77	132	4.18	-9.52	-10.21	194	4.54	-9.58	-12.23
76	3.69	-10.55	-10.53	134	4.20	-10.64	-12.08	196	4.54	-10.68	-12.53
78	3.70	-10.03	-11.97	136	4.21	-11.49	-11.66	198	4.55	-10.67	-12.10
80	3.72	-10.14	-11.00	138	4.23	-10.53	-11.10	200	4.55	-10.30	-11.22
82	3.74	-10.27	-11.70	140	4.25	-11.21	-11.06	202	4.56	-10.95	-11.55
84	3.76	-11.21	-11.50	142	4.27	-9.96	-11.50	204	4.56	-10.99	-12.00
86	3.78	-11.31	-11.18	144	4.28	-9.84	-11.87	206	4.57	-9.79	-11.18
88	3.79	-9.96	-10.86	146	4.30	-10.81	-12.63	208	4.57	-10.47	-10.95
90	3.81	-10.13	-10.42	148	4.32	-11.05	-12.30	210	4.58	-10.47	-10.81
92	3.83	-10.82	-11.79	150	4.34	-11.22	-11.61	212	4.59	-10.14	-11.09
94	3.85	-11.11	-12.04	152	4.35	-10.61	-12.16	214	4.59	-10.00	-11.52
96	3.86	-11.20	-12.45	154	4.37	-9.28	-11.05	216	4.60	-10.48	-10.15
98	3.88	-10.40	-11.71	156	4.39	-10.37	-11.43	218	4.60	-10.61	-11.96
102	3.92	-9.80	-11.83	160	4.42	-10.28	-11.97	220	4.61	-10.45	-11.64
104	3.93	-9.97	-11.16	162	4.44	-9.80	-11.25	222	4.61	-10.58	-11.59
106	3.95	-11.13	-13.38	164	4.46	-10.12	-11.54	224	4.62	-11.13	-11.72
108	3.97	-10.86	-11.07	166	4.47	-10.89	-11.43	226	4.62	-11.13	-12.06
110	3.99	-10.12	-12.25	168	4.47	-10.07	-11.38	228	4.63	-10.57	-12.81

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
230	4.63	-10.88	-12.12	286	4.78	-10.33	-11.96	26	54.75	-10.15	-9.80
232	4.64	-10.62	-11.07	288	4.78	-9.56	-12.90	28	55.47	-9.86	-10.00
234	4.64	-11.00	-12.20	290	4.79	-10.14	-11.81	30	56.20	-10.71	-11.00
236	4.65	-10.36	-11.57	292	4.79	-10.83	-11.72	32	56.93	-8.37	-10.82
238	4.65	-10.09	-11.97	294	4.80	-10.24	-12.62	34	57.71	-7.64	-11.62
240	4.66	-10.87	-11.53	296	4.80	-10.19	-11.35	36	58.56	-6.68	-11.41
242	4.66	-10.93	-11.77	298	4.81	-9.67	-10.81	38	59.40	-5.85	-11.65
244	4.67	-9.87	-11.73	300	4.82	-9.95	-10.38	40	60.25	-5.97	-11.75
246	4.67	-10.35	-12.51	302	4.82	-9.48	-11.74	42	61.10	-4.51	-10.56
248	4.68	-9.75	-12.03	304	4.83	-10.39	-11.27	44	61.94	-3.84	-10.81
250	4.68	-9.77	-11.42	308	4.84	-10.31	-10.89	46	62.79	-4.83	-9.50
252	4.69	-10.29	-11.48	310	4.84	-10.48	-11.92	48	63.63	-4.52	-9.22
254	4.70	-9.19	-11.42	312	4.85	-10.48	-12.53	50	64.48	-3.65	-8.91
256	4.70	-10.29	-11.25	314	4.85	-10.70	-11.78	52	65.32	-6.07	-9.48
258	4.71	-10.37	-11.77	BF2				54	66.17	-8.21	-9.77
260	4.71	-10.56	-12.12	0	45.30	-9.68	-10.00	56	67.01	-7.05	-9.71
262	4.72	-10.89	-12.28	2	46.02	-9.38	-9.15	58	67.86	-10.49	-10.13
264	4.72	-10.42	-11.68	4	46.75	-10.37	-9.94	60	68.70	-10.62	-10.72
266	4.73	-10.37	-10.82	6	47.48	-9.81	-9.32	62	69.55	-10.33	-10.83
268	4.73	-10.08	-10.53	8	48.20	-9.70	-9.19	64	70.39	-10.40	-10.85
270	4.74	-10.93	-11.32	10	48.93	-10.88	-9.41	66	71.24	-10.58	-10.65
272	4.74	-10.24	-11.52	12	49.66	-10.40	-9.74	68	72.09	-10.07	-9.61
274	4.75	-10.22	-10.71	14	50.38	-10.91	-9.68	70	72.93	-9.90	-9.83
276	4.75	-11.00	-10.70	16	51.11	-10.91	-10.31	72	73.78	-9.77	-9.89
278	4.76	-10.98	-10.61	18	51.84	-10.84	-9.69	74	74.62	-9.90	-11.61
280	4.76	-10.26	-11.11	20	52.57	-10.73	-10.01	76	75.47	-9.99	-11.58
282	4.77	-10.74	-11.10	22	53.29	-10.52	-9.25	78	76.31	-10.95	-12.00
284	4.77	-10.97	-12.41	24	54.02	-10.08	-10.23	80	77.16	-10.44	-12.56

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
82	78.00	-10.75	-12.64	138	85.52	-5.07	-9.15	194	91.19	-10.67	-10.28
84	78.55	-8.48	-12.43	140	85.77	-5.25	-9.40	196	91.34	-10.09	-9.97
86	78.81	-2.99	-11.31	142	86.03	-5.07	-9.83	198	91.49	-9.84	-9.49
88	79.07	-2.73	-9.13	144	86.29	-5.38	-9.61	202	91.80	-9.78	-9.82
90	79.33	-5.08	-9.98	146	86.55	-5.16	-9.36	204	91.96	-9.10	-9.74
92	79.59	-5.04	-10.42	148	86.80	-4.81	-9.65	206	92.11	-9.92	-9.12
94	79.84	-4.92	-9.62	150	87.06	-4.92	-9.65	208	92.27	-10.51	-10.04
96	80.10	-5.88	-10.49	152	87.32	-5.45	-9.64	210	92.42	-10.53	-9.62
98	80.36	-7.77	-10.66	154	87.58	-6.34	-10.62	212	92.58	-10.79	-9.97
100	80.62	-8.16	-10.43	156	87.88	-5.95	-10.31	214	92.73	-10.56	-10.11
102	80.87	-7.30	-10.15	158	88.22	-5.85	-10.25	216	92.88	-10.70	-10.29
104	81.13	-6.91	-10.22	160	88.56	-5.59	-10.58	218	93.04	-10.24	-9.73
106	81.39	-6.69	-10.49	162	88.72	-11.03	-11.49	220	93.19	-10.85	-10.38
108	81.65	-6.93	-10.26	164	88.87	-10.55	-11.49	1108			
110	81.91	-6.63	-10.60	166	89.02	-10.74	-12.20	2	1.73	-9.21	-9.76
112	82.16	-6.24	-10.20	168	89.18	-10.71	-12.26	4	1.75	-8.90	-10.46
114	82.42	-7.12	-10.32	170	89.33	-10.76	-11.90	6	1.77	-9.87	-12.72
116	82.68	-6.32	-9.86	172	89.49	-10.30	-11.69	8	1.79	-8.80	-10.95
118	82.94	-6.45	-10.23	174	89.64	-10.25	-11.63	10	1.81	-8.67	-11.48
120	83.19	-6.82	-10.99	176	89.80	-10.09	-11.21	12	1.83	-9.43	-11.55
122	83.45	-6.34	-10.93	178	89.95	-10.39	-11.03	14	1.85	-8.99	-10.78
124	83.71	-6.43	-10.39	180	90.11	-9.89	-10.65	16	1.87	-8.95	-11.34
126	83.97	-5.96	-10.35	182	90.26	-9.47	-10.89	18	1.89	-8.60	-10.65
128	84.23	-5.77	-9.87	184	90.41	-9.92	-10.79	20	1.91	-8.59	-10.60
130	84.48	-5.74	-9.37	186	90.57	-9.57	-10.74	22	1.92	-7.93	-11.08
132	84.74	-6.14	-10.15	188	90.72	-10.06	-10.52	24	1.94	-8.18	-10.96
134	85.00	-6.25	-10.17	190	90.88	-10.24	-10.20	26	1.96	-7.94	-10.12
136	85.26	-5.26	-10.25	192	91.03	-10.31	-10.19	28	1.98	-8.19	-11.23

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
30	2.00	-7.81	-11.35	86	2.54	-9.33	-10.32	142	3.08	-10.28	-11.60
32	2.02	-8.37	-11.26	88	2.56	-8.51	-10.24	144	3.09	-10.04	-10.60
34	2.04	-8.28	-10.33	90	2.58	-8.73	-10.47	146	3.11	-8.42	-10.88
36	2.06	-7.86	-11.25	92	2.60	-9.71	-10.62	148	3.13	-9.05	-11.61
38	2.08	-7.46	-10.17	94	2.61	-8.82	-11.24	150	3.15	-8.31	-10.63
40	2.10	-7.47	-10.08	96	2.63	-8.10	-11.04	152	3.17	-10.00	-11.27
42	2.12	-8.13	-10.84	98	2.65	-9.64	-11.49	154	3.19	-7.61	-11.43
44	2.14	-8.37	-10.59	100	2.67	-9.75	-12.32	156	3.21	-8.55	-11.37
46	2.15	-8.63	-11.38	102	2.69	-8.47	-11.90	158	3.23	-10.02	-9.79
48	2.17	-8.22	-10.42	104	2.71	-8.68	-11.78	160	3.25	-10.14	-11.09
50	2.19	-7.97	-11.10	106	2.73	-9.72	-11.86	162	3.27	-7.57	-11.74
52	2.21	-7.93	-11.22	108	2.75	-9.11	-10.99	164	3.29	-8.82	-12.29
54	2.23	-8.57	-11.13	110	2.77	-8.34	-11.44	166	3.31	-9.69	-12.67
56	2.25	-7.87	-11.42	112	2.79	-7.28	-10.97	168	3.32	-9.58	-11.43
58	2.27	-8.15	-11.72	114	2.81	-6.71	-10.42	J106			
60	2.29	-9.18	-12.03	116	2.83	-7.50	-10.82	0	87.72	-7.37	-10.03
62	2.31	-8.65	-11.29	118	2.84	-7.93	-10.11	5	88.51	-7.74	-8.54
64	2.33	-8.98	-11.27	120	2.86	-8.33	-11.05	10	89.30	-6.40	-9.19
66	2.35	-8.26	-10.99	122	2.88	-9.40	-11.88	15	90.92	-9.11	-7.75
68	2.37	-9.00	-10.22	124	2.90	-8.16	-10.96	20	93.41	-8.19	-9.45
70	2.38	-9.82	-11.79	126	2.92	-9.34	-11.40	25	93.99	-9.32	-10.26
72	2.40	-9.50	-10.81	128	2.94	-10.14	-11.32	30	94.60	-9.17	-9.32
74	2.42	-10.01	-11.00	130	2.96	-8.27	-11.38	35	95.19	-7.58	-9.64
76	2.44	-10.61	-12.13	132	2.98	-7.32	-10.96	40	95.79	-7.83	-8.59
78	2.46	-11.29	-11.30	134	3.00	-8.81	-11.42	45	96.41	-6.70	-10.10
80	2.48	-9.39	-10.08	136	3.02	-8.24	-11.58	50	97.06	-7.71	-9.23
82	2.50	-10.00	-11.55	138	3.04	-7.96	-10.39	55	97.71	-7.46	-9.43
84	2.52	-9.94	-11.58	140	3.06	-10.22	-11.22	60	98.33	-10.20	-9.85

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
65	99.00	-7.76	-10.50	205	115.07	-8.26	-8.81	345	126.62	-11.50	-13.10
70	99.65	-10.42	-10.73	210	115.63	-8.98	-8.74	350	126.66	-10.75	-13.28
75	100.26	-10.41	-11.76	215	116.18	-10.00	-10.02	355	126.71	-10.52	-13.04
80	100.89	-11.00	-13.96	220	116.73	-10.08	-9.31	360	126.75	-9.38	-12.29
85	101.55	-9.31	-11.87	225	117.30	-9.94	-9.03	365	126.79	-5.46	-12.52
90	102.20	-10.04	-10.83	230	117.85	-9.83	-8.39	370	126.83	-7.12	-12.35
95	102.82	-8.54	-11.66	235	118.40	-9.36	-9.66	375	126.88	-6.04	-11.49
100	103.46	-9.52	-12.34	240	118.96	-6.33	-8.28	380	126.92	-7.41	-11.71
105	104.11	-8.56	-13.68	245	119.51	-7.90	-8.67	385	126.95	-6.94	-11.40
110	104.73	-9.05	-11.60	250	120.06	-11.30	-8.67	390	126.99	-6.42	-11.12
115	105.33	-9.23	-12.40	255	120.64	-10.34	-8.32	395	127.02	-6.10	-10.76
120	105.89	-5.87	-10.55	260	121.23	-7.82	-13.35	400	127.06	-8.03	-11.11
125	106.46	-6.99	-10.92	265	121.85	-10.46	-13.89	405	127.10	-7.03	-11.17
130	107.02	-5.76	-9.37	270	122.50	-10.63	-13.83	410	127.13	-8.62	-11.38
135	107.58	-9.19	-11.04	275	123.18	-7.85	-13.36	415	127.17	-7.54	-11.26
140	108.13	-10.05	-11.55	280	123.87	-8.97	-14.05	420	127.21	-6.70	-10.04
145	108.66	-10.12	-12.29	285	124.52	-8.87	-13.45	425	127.25	-7.22	-10.74
150	109.20	-10.46	-10.29	290	125.16	-10.08	-13.16	430	127.30	-6.15	-9.38
155	109.71	-7.69	-10.59	295	125.60	-11.14	-13.99	435	127.35	-7.90	-11.01
160	110.22	-9.65	-11.63	300	125.89	-12.12	-14.17	440	127.40	-8.09	-11.41
165	110.71	-6.20	-8.14	305	126.16	-9.46	-12.45	445	127.45	-8.72	-11.43
170	111.24	-9.63	-9.13	310	126.43	-9.36	-12.27	450	127.50	-8.97	-11.03
175	111.77	-7.59	-7.85	315	126.52	-9.57	-12.15	455	127.54	-9.11	-10.72
180	112.32	-8.19	-9.35	320	126.53	-10.64	-12.54	460	127.59	-8.11	-10.14
185	112.86	-8.38	-8.75	325	126.53	-11.32	-14.13	465	127.64	-8.58	-9.09
190	113.41	-8.97	-8.52	330	126.54	-11.86	-15.09	470	127.69	-8.74	-8.39
195	113.96	-8.43	-8.70	335	126.56	-9.62	-12.66	475	127.74	-8.30	-7.76
200	114.51	-8.34	-8.94	340	126.58	-10.19	-12.94	480	127.80	-6.58	-8.10

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
485	127.84	-6.82	-8.15	625	130.17	-7.30	-6.48	770	135.19	-11.50	-8.04
490	127.89	-7.70	-8.38	630	130.32	-7.88	-7.00	775	135.35	-11.80	-7.96
495	127.95	-8.85	-8.40	635	130.48	-11.81	-8.50	780	135.52	-9.39	-7.44
500	128.00	-9.14	-8.30	640	130.64	-9.98	-7.68	785	135.69	-10.83	-8.50
505	128.05	-8.96	-8.82	645	130.81	-7.71	-6.60	790	135.85	-12.41	-8.61
510	128.10	-6.79	-7.87	650	130.98	-7.56	-6.09	795	136.02	-11.33	-7.90
515	128.14	-9.56	-8.90	655	131.16	-10.45	-8.29	800	136.19	-5.82	-7.47
520	128.18	-7.31	-8.44	660	131.33	-8.65	-7.05	805	136.36	-9.02	-8.38
525	128.22	-7.37	-7.52	665	131.49	-8.18	-7.13	810	136.53	-11.10	-8.17
530	128.25	-7.11	-7.81	670	131.66	-10.44	-7.64	815	136.70	-11.37	-10.14
535	128.29	-7.04	-7.75	675	131.82	-11.56	-6.42	820	136.86	-10.48	-9.48
540	128.32	-8.26	-8.58	680	132.00	-11.01	-6.84	825	137.03	-10.60	-9.10
545	128.36	-9.38	-9.41	685	132.18	-11.63	-6.67	830	137.19	-9.20	-9.91
550	128.41	-9.53	-8.38	690	132.37	-10.47	-7.23	835	137.36	-11.46	-10.14
555	128.47	-9.36	-8.50	695	132.55	-8.44	-6.50	840	137.53	-9.80	-9.23
560	128.53	-9.15	-7.86	700	132.76	-10.05	-7.83	845	137.69	-8.88	-8.78
565	128.59	-9.60	-9.15	705	132.96	-10.76	-7.47	850	137.86	-9.49	-9.39
570	128.68	-9.69	-8.89	710	133.15	-9.04	-7.20	855	138.04	-9.54	-8.75
575	128.76	-9.50	-9.18	715	133.33	-9.70	-7.68	860	138.21	-8.20	-8.81
580	128.86	-6.32	-7.63	720	133.50	-11.25	-8.45	1103			
585	128.97	-9.14	-6.58	725	133.68	-9.43	-7.18	12.1	131.40	-4.06	-7.22
590	129.09	-8.07	-7.33	730	133.85	-8.69	-7.22	13.8	131.70	-4.17	-7.45
595	129.22	-8.35	-7.23	735	134.01	-11.67	-8.62	15	131.94	-4.53	-6.86
600	129.38	-10.34	-8.76	740	134.18	-11.98	-7.89	16.5	132.22	-6.04	-6.27
605	129.54	-8.24	-7.23	745	134.35	-12.55	-9.68	17.3	132.35	-5.54	-6.33
610	129.69	-11.68	-9.53	750	134.52	-8.47	-6.53	19	132.57	-4.02	-6.12
615	129.85	-9.33	-7.81	755	134.68	-9.18	-7.82	20	132.69	-11.15	-7.69
620	130.01	-8.75	-6.94	765	135.02	-10.85	-8.40	22	132.91	-10.49	-7.65

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
23	133.02	-10.26	-7.44	65.1	137.67	-8.76	-7.70	113.3	141.41	-7.27	-7.81
24	133.13	-7.86	-6.55	67	137.92	-8.77	-8.23	116.6	141.52	-9.33	-8.76
25	133.24	-9.16	-7.60	69	138.17	-9.88	-7.79	118.2	141.53	-8.19	-8.07
26	133.35	-9.31	-7.88	71	138.41	-6.80	-6.54	121	141.54	-8.55	-10.13
27	133.47	-8.69	-6.90	72.3	138.57	-10.77	-7.37	123	141.54	-8.05	-9.56
28.3	133.62	-9.39	-7.61	74	138.76	-7.80	-6.32	124.4	141.54	-7.98	-9.40
29.5	133.76	-9.66	-6.61	76	138.96	-7.80	-9.44	126	141.54	-8.75	-9.51
31	133.95	-10.19	-6.74	77.5	139.11	-4.76	-7.50	127.5	141.55	-10.30	-9.26
32.6	134.14	-7.44	-7.01	78.7	139.22	-5.63	-8.26	129.5	141.57	-8.86	-9.86
34	134.29	-4.67	-7.25	81	139.43	-5.77	-8.84	131	141.59	-8.44	-9.57
35.1	134.41	-5.59	-7.17	82.8	139.59	-6.38	-9.23	131.1	141.59	-8.11	-9.12
37.4	134.65	-9.06	-7.07	84.8	139.76	-7.48	-8.95	132	141.60	-8.67	-8.59
39.1	134.83	-6.59	-6.77	86.5	139.89	-7.79	-8.82	133	141.62	-10.07	-9.23
41	135.02	-10.52	-7.99	88	140.00	-8.41	-8.27	135	141.63	-10.00	-9.35
42.5	135.18	-11.95	-8.84	89	140.07	-10.22	-9.54	137	141.64	-11.58	-9.65
43	135.24	-11.05	-7.29	90	140.15	-7.83	-8.80	139	141.64	-9.15	-9.56
44.3	135.38	-10.76	-7.49	91	140.22	-10.33	-8.43	140	141.64	-10.91	-9.20
46	135.57	-10.54	-7.50	93	140.35	-7.63	-7.76	142	141.64	-10.48	-9.92
47.5	135.73	-10.80	-7.83	95	140.48	-8.52	-9.83	144	141.64	-11.00	-10.27
49.1	135.91	-11.83	-8.53	96	140.53	-8.75	-9.42	145.5	141.64	-10.54	-10.42
51.9	136.21	-11.56	-8.43	97.9	140.64	-5.59	-8.14	147	141.64	-10.93	-10.41
53.5	136.38	-9.83	-7.71	99.5	140.73	-7.01	-7.70	148.2	141.64	-11.72	-10.76
55	136.54	-10.69	-7.84	101.3	140.84	-7.29	-8.88	150	141.64	-11.11	-10.23
58	136.86	-11.03	-7.85	103	140.92	-7.44	-7.90	151.8	141.65	-11.38	-10.33
59.9	137.06	-10.57	-7.76	104.2	140.98	-5.45	-7.65	153	141.65	-11.49	-10.07
61	137.18	-9.64	-7.46	108	141.17	-6.94	-9.57	154	141.65	-10.97	-9.26
62	137.30	-5.74	-6.48	110	141.27	-8.91	-9.25	155	141.65	-10.19	-8.20
63.8	137.50	-8.57	-6.78	112	141.36	-7.51	-9.08	156	141.66	-10.54	-9.93

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
158.3	141.68	-10.34	-10.31	212	142.51	-8.86	-9.41	268	143.80	-9.39	-9.08
159.5	141.70	-9.61	-10.11	214	142.54	-9.53	-9.62	270	143.87	-9.65	-10.13
162	141.73	-7.17	-9.25	216	142.57	-9.52	-10.41	272	143.96	-7.26	-9.14
164.3	141.76	-8.75	-10.20	218	142.60	-10.31	-10.49	274	144.03	-9.22	-10.80
166	141.78	-8.80	-9.22	220	142.63	-9.38	-10.78	276	144.08	-10.88	-11.57
168.2	141.81	-9.68	-10.07	222	142.66	-8.53	-9.42	278	144.15	-11.20	-11.24
170	141.84	-10.25	-9.98	224	142.70	-9.04	-10.85	280	144.24	-10.79	-10.46
172.5	141.88	-9.95	-10.61	226	142.73	-9.77	-10.68	282	144.33	-9.98	-10.93
175	141.92	-10.49	-7.96	228	142.77	-10.23	-11.00	284	144.41	-8.81	-10.19
177	141.95	-11.11	-10.10	230	142.81	-10.18	-11.41	286	144.52	-9.00	-10.95
178.5	141.97	-10.35	-10.40	232	142.85	-10.78	-11.39	288	144.64	-4.90	-9.94
180	142.00	-10.37	-9.69	234	142.90	-10.76	-11.77	290	144.75	-7.22	-9.74
182	142.03	-9.15	-11.57	236	142.94	-9.60	-12.15	292	144.91	-7.19	-9.94
183.5	142.06	-4.83	-9.56	238	142.99	-9.64	-10.62	294	144.97	-7.92	-9.70
185	142.08	-7.96	-10.16	240	143.03	-7.11	-9.41	296	145.02	-6.93	-9.76
187	142.12	-10.89	-10.50	242	143.07	-7.95	-10.02	298	145.11	-5.49	-9.67
188.2	142.14	-7.08	-9.99	244	143.11	-8.84	-10.83	300	145.19	-7.83	-9.76
190.3	142.17	-8.26	-10.36	246	143.16	-8.99	-9.83	302	145.28	-7.69	-9.84
192	142.20	-7.74	-10.06	248	143.20	-9.01	-9.93	304	145.33	-7.92	-10.33
194	142.24	-7.32	-10.21	250	143.25	-9.41	-10.08	306	145.40	-7.97	-9.66
196	142.27	-8.73	-10.08	252	143.30	-10.50	-10.47	308	145.46	-8.59	-9.96
198	142.31	-10.00	-10.54	254	143.35	-10.62	-10.83	310	145.51	-8.75	-9.82
200	142.33	-10.38	-10.35	256	143.41	-10.71	-10.92	312	145.54	-8.90	-9.92
202	142.36	-9.10	-10.90	258	143.47	-11.05	-11.07	314	145.54	-9.30	-10.16
204	142.39	-9.48	-10.31	260	143.52	-10.66	-10.17	316	145.58	-9.36	-10.22
206	142.42	-8.65	-9.74	262	143.59	-11.13	-9.94	318	145.60	-8.67	-10.39
208	142.45	-9.20	-9.81	264	143.66	-9.49	-10.25	320	145.62	-9.58	-10.73
210	142.48	-9.37	-10.70	266	143.72	-10.13	-9.76	322	145.62	-9.70	-10.29

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
324	145.62	-7.53	-9.28	380	153.33	-6.54	-8.43	436	164.72	-7.39	-8.70
326	145.62	-10.49	-9.67	382	153.72	-6.51	-9.48	438	165.15	-8.11	-9.89
328	145.63	-10.50	-9.88	384	154.01	-6.49	-9.34	440	165.57	-8.75	-8.63
330	145.67	-10.76	-9.87	386	154.34	-7.28	-9.81	442	165.99	-4.53	-8.79
332	145.74	-10.16	-9.74	388	154.67	-7.70	-9.36	444	166.43	-3.80	-7.84
334	145.84	-10.89	-9.96	390	154.97	-7.60	-9.39	446	166.84	-8.76	-8.59
336	146.01	-10.02	-9.82	392	155.27	-8.75	-9.54	448	167.28	-7.23	-9.56
338	146.16	-9.90	-9.87	394	155.61	-9.33	-8.84	450	167.68	-9.21	-9.76
340	146.34	-10.96	-10.10	396	155.93	-10.20	-10.36	452	168.07	-9.60	-10.10
342	146.54	-10.24	-10.43	398	156.26	-11.29	-9.89	454	168.49	-6.96	-8.46
344	146.72	-11.32	-11.63	400	156.62	-9.93	-9.00	456	168.88	-7.41	-9.21
346	146.96	-10.87	-11.14	402	157.01	-9.54	-9.78	458	169.28	-7.16	-8.70
348	147.19	-9.89	-10.88	404	157.39	-9.96	-9.77	460	169.65	-4.91	-7.60
350	147.52	-6.19	-9.07	406	157.79	-7.08	-9.81	462	170.06	-4.46	-8.49
352	147.82	-8.98	-11.26	408	158.17	-8.64	-9.15	464	170.45	-6.57	-8.95
354	148.13	-10.28	-11.71	410	158.58	-10.22	-11.05	466	170.82	-5.69	-8.67
356	148.43	-7.98	-10.52	412	158.98	-8.63	-9.77	470	171.60	-6.75	-7.61
358	148.80	-5.60	-9.43	414	159.37	-9.67	-9.91	472	171.98	-4.43	-7.42
360	149.19	-7.90	-10.79	416	159.83	-11.11	-10.14	474	172.39	-8.61	-9.04
362	149.54	-10.28	-11.88	418	160.27	-10.97	-10.15	476	172.78	-6.88	-8.80
364	149.88	-9.63	-11.21	420	160.75	-6.32	-9.49	478	173.17	-10.90	-8.98
366	150.31	-9.54	-11.51	422	161.25	-8.43	-9.48	480	173.55	-6.40	-9.31
368	150.79	-6.88	-10.40	424	161.79	-8.50	-9.25	482	173.93	-9.33	-9.56
370	151.20	-2.78	-8.25	426	162.30	-8.73	-10.08	484	174.34	-6.27	-8.66
372	151.75	-6.12	-9.86	428	162.80	-9.38	-9.55	486	174.73	-5.38	-8.91
374	152.24	-6.95	-10.22	430	163.29	-8.65	-9.54	488	175.14	-7.75	-9.18
376	152.60	-7.72	-9.37	432	163.81	-6.64	-8.67	SY10			
378	152.96	-5.41	-8.75	434	164.24	-7.99	-8.72	8	8.01	-7.88	-12.16

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
10	8.81	-6.64	-11.88	44	18.56	-7.03	-9.83	72	20.14	-9.90	-10.89
11	9.20	-6.45	-12.80	45	18.62	-8.95	-9.71	73	20.20	-10.05	-11.62
12	9.60	-6.09	-12.84	46	18.68	-7.50	-9.54	74	20.25	-10.52	-9.57
13	10.00	-6.47	-13.25	47	18.73	-7.50	-9.81	75	20.31	-10.00	-9.89
14	10.40	-5.62	-12.78	48	18.79	-6.79	-9.38	76	20.36	-9.71	-11.00
15	10.80	-6.34	-12.75	49	18.84	-7.71	-9.76	77	20.42	-9.61	-11.13
16	11.20	-5.32	-12.16	50	18.90	-8.36	-9.72	78	20.48	-10.42	-10.62
18	12.00	-6.35	-9.34	51	18.96	-8.76	-10.45	79	20.53	-10.83	-10.85
20	12.80	-6.02	-9.55	52	19.01	-8.93	-9.88	80	20.59	-9.85	-11.10
22	13.59	-6.07	-10.05	53	19.07	-9.22	-10.01	81	20.65	-9.91	-11.33
24	14.39	-6.43	-11.02	54	19.13	-9.75	-10.22	82	20.70	-9.37	-10.68
25	14.79	-6.56	-9.80	55	19.18	-9.52	-10.49	83	20.76	-9.01	-10.61
26	15.19	-6.02	-7.98	56	19.24	-10.36	-10.27	84	20.81	-8.29	-10.16
27	15.44	-8.83	-7.43	57	19.30	-9.67	-9.89	85	20.87	-10.37	-10.97
28	15.70	-9.93	-7.85	58	19.35	-9.32	-10.51	86	20.93	-9.72	-10.29
30	16.20	-7.25	-8.49	59	19.41	-9.34	-10.14	87	20.88	-9.76	-10.81
31	16.45	-5.63	-7.51	60	19.46	-9.29	-10.63	88	20.88	-9.49	-10.80
32	16.71	-4.48	-7.37	61	19.52	-9.97	-11.11	89	20.89	-9.82	-11.55
33	16.96	-6.72	-8.56	62	19.58	-7.80	-9.97	90	20.89	-9.62	-10.93
34	17.21	-8.61	-8.79	63	19.63	-9.99	-10.11	91	20.90	-10.17	-10.98
35	17.47	-6.54	-9.85	64	19.69	-9.46	-9.72	92	20.90	-9.71	-10.27
36	17.72	-7.57	-9.99	65	19.75	-9.77	-9.90	93	20.91	-9.60	-10.81
37	17.97	-7.25	-9.86	66	19.80	-9.70	-10.63	94	20.91	-10.05	-9.44
38	18.23	-7.30	-9.96	67	19.86	-10.35	-11.00	95	20.92	-10.06	-10.40
39	18.28	-9.83	-9.99	68	19.91	-9.63	-10.39	96	20.92	-9.91	-10.93
40	18.34	-8.90	-9.82	69	19.97	-10.14	-10.77	97	20.93	-10.21	-10.77
42	18.45	-6.38	-9.43	70	20.03	-10.48	-10.53	98	20.93	-10.31	-10.55
43	18.51	-5.89	-9.57	71	20.08	-9.53	-10.24	99	20.93	-10.45	-10.46

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
100	20.94	-9.90	-10.15	128	21.07	-10.53	-10.04	174	21.28	-9.67	-9.72
101	20.94	-9.74	-10.98	129	21.07	-10.58	-10.27	176	21.29	-10.11	-10.27
102	20.95	-9.74	-10.45	131	21.08	-10.40	-10.98	178	21.30	-9.97	-10.05
103	20.95	-10.44	-10.51	132	21.09	-10.60	-9.92	180	21.30	-8.29	-9.67
104	20.96	-10.08	-10.06	133	21.09	-10.73	-10.42	182	21.31	-9.88	-9.91
105	20.96	-10.53	-10.58	134	21.09	-10.32	-10.28	184	21.32	-10.32	-10.43
106	20.97	-9.75	-11.05	135	21.10	-10.09	-10.50	186	21.33	-9.57	-9.90
107	20.97	-10.17	-10.86	136	21.10	-9.46	-9.33	188	21.34	-10.29	-10.52
108	20.98	-10.26	-9.90	137	21.11	-9.63	-9.71	190	21.35	-9.39	-10.16
109	20.98	-9.86	-9.67	138	21.11	-9.55	-9.93	192	21.36	-8.90	-9.62
110	20.98	-9.94	-9.83	139	21.12	-9.67	-9.39	194	21.37	-9.75	-10.53
111	20.99	-9.76	-10.94	140	21.12	-9.88	-8.81	196	21.38	-9.87	-9.69
112	20.99	-9.91	-11.25	142	21.13	-10.27	-9.37	198	21.39	-9.94	-10.24
113	21.00	-10.59	-12.11	144	21.14	-10.46	-9.83	200	21.40	-10.54	-10.87
114	21.00	-10.28	-10.38	146	21.15	-8.99	-9.55	202	21.41	-10.29	-10.58
115	21.01	-10.12	-10.56	148	21.16	-10.15	-10.88	204	21.41	-10.11	-10.29
116	21.01	-9.88	-10.26	150	21.17	-9.95	-10.03	206	21.42	-10.07	-10.55
117	21.02	-9.74	-11.23	152	21.18	-10.02	-10.28	208	21.43	-10.25	-10.49
118	21.02	-9.83	-9.49	154	21.19	-9.86	-10.24	210	21.44	-9.91	-10.36
119	21.03	-9.52	-9.81	156	21.19	-9.89	-10.48	212	21.45	-10.28	-9.52
120	21.03	-9.79	-10.43	158	21.20	-9.45	-9.42	214	21.46	-9.53	-9.45
121	21.03	-10.16	-10.94	160	21.21	-9.20	-9.95	216	21.47	-9.78	-9.65
122	21.04	-10.03	-10.83	162	21.22	-10.04	-9.54	218	21.48	-9.85	-9.91
123	21.04	-9.83	-10.55	164	21.23	-10.22	-9.18	220	21.49	-9.49	-10.21
124	21.05	-9.58	-10.53	166	21.24	-10.11	-9.20	222	21.50	-9.64	-9.81
125	21.05	-9.39	-10.48	168	21.25	-10.11	-8.90	224	21.51	-8.66	-9.68
126	21.06	-10.07	-11.08	170	21.26	-9.72	-9.53	226	21.52	-8.45	-10.13
127	21.06	-10.93	-10.18	172	21.27	-10.30	-10.10	228	21.52	-9.10	-10.62

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
230	21.53	-9.37	-10.55	288	25.57	-5.14	-9.57	32	7.13	-10.25	-12.83
232	21.54	-9.18	-9.91	290	25.72	-6.68	-9.72	34	7.16	-10.76	-12.74
234	21.55	-9.95	-9.16	292	25.86	-6.07	-10.17	36	7.19	-11.12	-12.38
236	21.56	-9.90	-11.00	294	26.01	-5.57	-9.92	38	7.22	-11.11	-13.61
238	21.57	-9.43	-10.78	296	26.15	-6.44	-10.49	40	7.24	-10.82	-13.22
240	21.75	-9.58	-10.62	298	26.30	-6.71	-10.53	42	7.27	-10.54	-14.43
242	21.92	-9.67	-10.86	300	26.44	-6.83	-10.40	44	7.30	-10.95	-13.44
244	22.10	-8.36	-9.92	302	26.58	-5.49	-10.57	46	7.32	-11.34	-12.56
246	22.28	-7.38	-9.89	304	26.73	-5.85	-10.50	48	7.33	-11.50	-15.06
248	22.46	-7.15	-10.03	306	26.87	-6.40	-10.41	52	7.36	-10.95	-13.63
250	22.63	-7.15	-9.91	308	27.02	-7.41	-10.78	54	7.39	-10.68	-13.16
252	22.81	-8.34	-9.92	310	27.16	-7.12	-10.51	56	7.39	-11.25	-13.72
254	22.99	-6.22	-9.89	SL10				58	7.39	-11.34	-14.77
256	23.17	-8.48	-10.17	2	6.74	-11.12	-13.17	60	7.40	-11.56	-16.07
258	23.34	-8.51	-10.06	4	6.77	-11.03	-12.12	62	7.42	-10.88	-13.87
260	23.52	-7.88	-9.89	6	6.79	-11.22	-11.98	64	7.44	-10.65	-13.51
262	23.70	-6.08	-9.56	8	6.82	-11.16	-13.69	66	7.46	-9.69	-14.63
266	23.99	-6.25	-9.52	10	6.84	-10.36	-12.73	68	7.50	-10.92	-12.71
268	24.13	-6.93	-8.88	12	6.87	-10.91	-12.09	70	7.54	-11.09	-12.17
270	24.28	-6.02	-10.41	14	6.89	-9.65	-13.01	72	7.60	-10.94	-12.92
272	24.42	-6.06	-10.16	16	6.92	-11.00	-12.93	74	7.66	-11.14	-13.14
274	24.56	-5.70	-10.05	18	6.94	-11.36	-13.11	76	7.70	-10.80	-12.09
276	24.71	-7.06	-10.04	20	6.97	-10.55	-13.39	77	7.73	-11.10	-12.12
278	24.85	-5.84	-9.78	22	6.99	-11.12	-14.04	78	7.77	-11.14	-12.62
280	25.00	-6.14	-10.37	24	7.02	-10.61	-12.95	79	7.80	-10.87	-12.86
282	25.14	-6.24	-10.13	26	7.05	-11.28	-13.67	80	7.84	-10.60	-12.72
284	25.29	-6.38	-10.12	28	7.08	-11.13	-15.03	81	7.87	-9.99	-12.73
286	25.43	-5.63	-9.55	30	7.10	-10.80	-12.41	82	7.90	-10.01	-13.59

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
83	7.94	-9.92	-14.08	114	9.41	-10.20	-12.97	15	16.62	-7.39	-8.19
84	7.98	-10.17	-12.73	118	9.52	-11.05	-12.82	18	17.44	-9.08	-7.95
85	8.03	-11.22	-13.66	120	9.55	-11.60	-14.00	21	18.26	-6.87	-9.32
86	8.07	-10.50	-13.25	122	9.58	-10.85	-14.11	24	19.07	-7.43	-10.58
87	8.11	-10.49	-13.68	124	9.62	-11.10	-12.85	27	19.89	-8.71	-9.79
88	8.16	-10.52	-12.81	126	9.64	-12.01	-14.65	30	20.71	-10.38	-9.54
89	8.21	-10.76	-13.03	128	9.65	-11.11	-14.06	33	21.53	-11.06	-10.09
90	8.26	-10.41	-11.54	130	9.65	-11.60	-13.87	36	22.35	-7.34	-9.87
91	8.31	-9.95	-13.11	132	9.65	-11.63	-13.84	39	23.17	-8.36	-9.01
92	8.35	-10.47	-13.25	134	9.65	-11.75	-14.49	42	23.99	-8.20	-9.84
93	8.37	-11.25	-14.30	136	9.66	-11.05	-13.21	45	24.81	-7.51	-9.45
94	8.44	-11.48	-14.17	138	9.66	-11.87	-12.63	48	25.63	-9.68	-10.32
95	8.57	-10.71	-13.06	140	9.66	-11.51	-12.12	51	26.44	-9.34	-10.00
97	8.76	-10.89	-12.67	142	9.66	-10.92	-13.91	54	27.26	-7.23	-9.22
98	8.79	-11.24	-13.31	144	9.66	-11.28	-13.28	57	28.08	-9.58	-10.33
99	8.82	-11.22	-13.67	146	9.66	-11.10	-13.47				
100	8.87	-11.01	-13.18	148	9.66	-10.88	-11.30	72	55.64	-10.17	-11.35
101	8.92	-10.81	-14.83	150	9.67	-11.56	-12.74	75	56.24	-10.11	-11.33
102	8.96	-10.33	-13.62	154	9.69	-10.85	-13.98	78	56.83	-10.07	-12.13
103	9.00	-10.83	-13.82	156	9.69	-11.21	-12.97	80	57.09	-7.55	-12.03
104	9.04	-10.88	-12.38	158	9.70	-11.01	-13.44	85	57.63	-10.17	-11.60
105	9.07	-11.37	-13.48	160	9.70	-10.73	-13.01	90	57.78	-7.42	-11.49
106	9.11	-10.85	-12.77	SL08				95	57.93	-8.90	-11.79
107	9.15	-10.61	-13.30	0	12.52	-9.83	-10.65	100	58.08	-9.86	-11.35
108	9.19	-11.37	-14.39	3	13.34	-10.31	-11.03	105	58.23	-9.74	-11.08
109	9.23	-11.37	-13.67	6	14.16	-8.24	-9.15	110	58.39	-8.05	-9.16
110	9.26	-10.79	-13.00	9	14.98	-7.66	-8.38	115	58.54	-8.44	-8.93
112	9.34	-10.71	-13.44	12	15.80	-10.29	-9.34	120	58.69	-10.17	-10.37

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
125	58.84	-10.36	-10.77	265	63.10	-11.08	-10.66	70	142.57	-9.44	-8.47
130	58.99	-10.51	-10.66	270	63.25	-11.19	-10.75	75	143.13	-9.38	-7.99
135	59.14	-8.73	-9.49	275	63.40	-10.70	-11.52	80	143.72	-9.39	-7.51
140	59.30	-10.72	-9.99	280	63.55	-11.22	-11.20	85	144.30	-7.94	-7.33
145	59.45	-9.87	-9.12	285	63.70	-11.29	-10.78	90	144.84	-6.34	-7.69
150	59.60	-11.79	-11.17	290	63.85	-11.68	-11.15	95	145.33	-6.98	-7.46
155	59.75	-10.22	-11.06	295	64.01	-9.90	-10.77	100	145.87	-8.23	-7.37
160	59.90	-10.23	-10.63	300	64.16	-9.30	-10.51	105	146.43	-7.07	-7.49
165	60.06	-10.47	-10.73	305	64.31	-11.26	-10.69	110	146.98	-8.99	-8.28
170	60.21	-10.24	-11.68	310	64.46	-10.64	-11.13	115	147.53	-9.56	-9.60
175	60.36	-11.79	-11.89	315	64.61	-8.61	-10.35	120	148.08	-8.27	-9.18
180	60.51	-11.26	-11.12	320	64.77	-11.43	-11.36	125	148.65	-6.14	-9.79
185	60.66	-11.78	-11.11	325	64.92	-11.21	-11.20	130	149.25	-6.34	-10.17
190	60.82	-9.64	-11.18	330	65.07	-10.14	-10.54				
195	60.97	-10.89	-11.27	335	65.22	-10.96	-11.02	140	168.21	-5.76	-9.19
200	61.12	-10.78	-11.08	XR05				145	168.50	-7.13	-8.79
205	61.27	-9.43	-10.21	15	133.56	-7.51	-8.78	150	168.65	-6.73	-8.52
210	61.42	-11.24	-11.70	19	136.43	-6.45	-9.67	155	168.85	-7.50	-8.93
215	61.58	-11.02	-10.97	23	137.25	-8.36	-7.91	160	169.04	-6.08	-9.60
220	61.73	-10.96	-11.61	25	137.50	-8.63	-7.64	165	169.24	-6.48	-10.16
225	61.88	-9.47	-11.49	30	138.05	-10.16	-8.22	170	169.44	-6.97	-9.49
230	62.03	-10.07	-11.60	35	138.60	-9.26	-8.62	175	169.62	-9.73	-9.55
235	62.18	-9.64	-10.87	40	139.15	-9.36	-7.86	180	169.78	-6.11	-10.50
240	62.34	-9.26	-10.88	45	139.75	-8.98	-7.78	185	169.92	-7.04	-11.88
245	62.49	-8.83	-10.22	50	140.33	-9.62	-7.87	190	170.04	-7.65	-12.32
250	62.64	-7.99	-10.24	55	140.82	-10.24	-8.60	195	170.17	-8.06	-11.66
255	62.79	-11.01	-10.93	60	141.39	-10.41	-7.82	200	170.33	-8.36	-11.69
260	62.94	-10.20	-11.32	65	141.98	-11.38	-9.43	205	170.50	-8.34	-12.30

Table 4.4. (continued)

Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Depth (mm)	Age (kyr, B.P.)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)
210	170.66	-8.41	-11.46	355	176.57	-10.44	-12.86	490	182.10	-11.97	-8.92
215	170.85	-11.57	-10.81	360	176.78	-10.39	-13.16	495	182.28	-12.34	-9.13
220	171.06	-9.32	-11.53	365	176.99	-10.04	-12.89	505	182.66	-12.20	-10.61
225	171.25	-11.98	-13.10	370	177.19	-10.18	-12.10	510	182.84	-11.72	-9.36
230	171.46	-12.44	-13.54	375	177.40	-10.31	-10.46	515	182.85	-12.23	-10.64
235	171.66	-10.56	-14.60	380	177.60	-10.28	-9.78	520	183.04	-11.93	-9.69
240	171.86	-10.28	-13.14	385	177.81	-10.63	-11.26	525	183.21	-12.32	-9.84
245	172.06	-10.70	-12.18	390	178.01	-10.40	-11.10	530	183.38	-12.05	-9.81
250	172.27	-11.16	-12.99	395	178.21	-10.26	-11.35	535	183.56	-11.81	-9.87
255	172.47	-11.17	-13.01	400	178.42	-9.81	-10.85	540	183.74	-11.94	-8.52
260	172.67	-10.33	-12.28	405	178.62	-10.38	-10.69	545	183.91	-12.03	-10.10
265	172.88	-10.75	-12.20	410	178.83	-11.07	-9.96	550	184.09	-11.78	-9.51
270	173.07	-9.68	-13.57	415	179.04	-11.40	-10.42	555	184.24	-10.50	-9.41
275	173.27	-9.33	-12.31	420	179.24	-11.91	-9.86	560	184.43	-11.04	-9.37
280	173.48	-9.74	-13.50	425	179.44	-11.60	-9.81				
285	173.68	-8.63	-12.94	430	179.65	-11.54	-9.50				
290	173.88	-9.76	-13.09	435	179.85	-11.77	-8.85				
295	174.09	-10.19	-11.83	440	180.06	-11.88	-8.72				
300	174.30	-9.80	-11.10	443	180.18	-11.71	-8.94				
305	174.51	-10.13	-11.39	447	180.35	-11.80	-10.03				
310	174.71	-10.39	-12.02	450	180.47	-12.24	-9.15				
315	174.92	-10.70	-12.67	455	180.67	-12.22	-7.94				
320	175.12	-11.08	-12.41	460	180.88	-12.11	-9.25				
325	175.33	-10.72	-13.67	465	181.09	-12.36	-9.60				
330	175.53	-11.06	-13.05	470	181.30	-12.28	-10.03				
340	175.95	-10.45	-12.31	475	181.50	-12.32	-9.34				
345	176.16	-10.41	-12.36	480	181.70	-11.65	-8.76				
350	176.37	-10.83	-13.16	485	181.90	-11.78	-9.07				

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Chapter 5

Conclusions

In the previous three chapters, I have established speleothem proxy-based paleo-hydroclimate records on annual to glacial-interglacial timescales from Mainland Southeast Asia (MSEA), addressed the likely driving mechanisms of the Indian Summer Monsoon (ISM), and provided plausible explanations for the lack of glacial-interglacial variations of Chinese speleothem $\delta^{18}\text{O}$. In this chapter, I summarize the key findings, and propose how to extend the research in the future.

5.1. Summary of key findings

In Chapter 2, I have presented speleothem trace element records from central Myanmar that cover ~450 years in recent history. The four elements—Mg, Sr, Ba, and U—all show strong annual oscillations. Among the trace elements, Mg is the most prominent. It is, overall, anti-correlated with Sr(Ba), indicating that the variations of the elements were controlled by water residence time, which can be linked to local rainfall changes. Therefore, the annual mean Mg record reflects ISM variability.

The Mg record suggests that the climate in the region shifted from a relatively wet state into a frequently dry regime around the ninth century, which could have disturbed the ancient Pyu society that flourished in central Myanmar. By spectrum analysis of the Mg record, I have identified solar cycles, together with the El Niño–Southern Oscillation (ENSO) and Indian Ocean Dipole (IOD), as the main forcings of the ISM precipitation on interannual to decadal timescales.

In Chapter 3, I have reconstructed the ISM history during the late stage of the last glacial time, based on speleothem $\delta^{18}\text{O}$ records retrieved from southern Myanmar and western Thailand. During the Last Glacial Maximum (LGM), the ISM was much weaker, with the regional rain amount possibly reduced to ~56% of modern values. The weakened ISM was driven mainly by a combination of reduced tropical Indian Ocean sea surface temperature and increased continental albedo due to a larger snow/ice coverage on the continent, that is, the glacial boundary conditions during the LGM. The difference between our records and the Chinese speleothem records reveals that the sensitivity of speleothem $\delta^{18}\text{O}$ to insolation and glacial forcing varies spatially.

In Chapter 4, I have provided a speleothem $\delta^{18}\text{O}$ record from central Myanmar covering a large portion of the last 40 ka and a speleothem $\delta^{18}\text{O}$ record from southeastern Yunnan, China, spanning the last 180 ka. Combining these records with the speleothem record reported in Chapter 3, I have traced the ISM moisture transport along a SW-NE transect from coastal Myanmar to

southwestern China. The trajectory of monsoon moisture has been stable for at least the recent ~40 ka. But, the oxygen isotope fractionation in precipitation associated with progressive rainout during moisture transport increased significantly during the glacial times, likely due to a much colder temperature and suppressed plant transpiration effect. A further spatial-temporal comparison of speleothem records from the ISM and East Asian Summer Monsoon regions has demonstrated that the stronger isotope fractionation from the ocean source region to cave sites is the main reason why the Chinese speleothem $\delta^{18}\text{O}$ records show unexpectedly low values during glacial maxima. I have therefore hypothesized that the stronger (weaker) isotope fractionation during glacial (interglacial) periods, dubbed transport pathway effect, is the key to the long-lasting debate about the orbital pace of the Asian Monsoon.

5.2. Suggestions for future work

In my research presented in Chapter 2, I have discussed two intriguing questions:

- 1) What controls trace element variations in Lin Noe Twin (LNT) cave?
- 2) Did climate change play an important role in Myanmar's history?

Although I have offered my interpretations, the topics merit further attentions, both due to their theoretical significance in climate science and their potential implication for government policy planning.

In Chapter 2, I suggested that the key process controlling trace elements variations is the changing water residence time in the karst system, which results from changes in ISM precipitation. This interpretation is derived from my trace element data from an annually laminated stalagmite in LNT cave that show a strong anti-correlation between Mg/Ca and Sr/Ca from seasonal to multidecadal timescales. However, I am also aware that my interpretation contrasts with another recently-identified key process, prior calcite precipitation (PCP), which has been used to explain the changes in alkaline-earth elements (e.g., Mg, Sr, and Ba), particularly the co-variations of Mg and Sr (Bernal et al., 2016; Griffiths et al., 2010; Sinclair et al., 2012; Tremaine and Froelich, 2013).

Under dry climate conditions, the PCP mechanism leads to an increase in both Mg and Sr concentrations (Fairchild et al., 2000; Fairchild and Treble, 2009) and systematic rises in $\delta^{13}\text{C}$ values in speleothem deposits (Griffiths et al., 2010; Johnson et al., 2006).

Thus, to test my interpretation and further improve the understanding of the major driving mechanism of speleothem trace elements, I propose to compare speleothem trace elements with $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in coeval calcite minerals, as the stable isotope signals can provide additional information on the regional climate and karst hydrology.

In Chapter 2, I have also discussed the possible impact of climate change on ancient Pyu society. Was that just an isolated case, or did climate change repeatedly play an important role in Myanmar's history? A recurrent influence in the last 2,000 years is possible, given the documented significant monsoonal variations on decadal to centennial time scales (Anderson, 2002; Anderson et al., 2010; Buckley et al., 2010; Chawchai et al., 2015; Cook et al., 2010; Sinha et al., 2011; 2015; Yamoah et al., 2016a; Yancheva et al., 2007; Zhang et al., 2008). Exceptional examples have been reported on the monsoonal waxing and waning that are likely responsible for the prosperity and demise of ancient civilizations in Asia (e.g., Buckley et al., 2014; Zhang et al., 2008).

To answer the two above-mentioned questions, more proxy records will be required. The speleothem trace element records used to reconstruct ISM precipitation variation in MSEA only cover ~450yr. A more complete evaluation should be based on longer climate records that last one thousand years or longer (e.g., covering at least one dramatic climate event, such as the Little Ice Age). Similar attempts have been made by studying tree rings (Buckley et al., 2014; 2010; Cook et al., 2010; D'Arrigo et al., 2011) and lake sediments (Chawchai et al., 2015; Sun et al., 2016; Yamoah et al., 2016a; 2016b). While notable successes have been achieved, these records also suffer from low resolution, relatively poor chronology, and/or problematic proxy interpretation. Alternatively, a more convincing evaluation of ISM history may be obtained through speleothem multi-proxy (such as stable isotopes and trace elements) studies.

A robust speleothem record can be attainable from LNT cave in central Myanmar—one of caves where my samples were collected. LNT cave has the potential to provide high-resolution, well-dated hydroclimate records, since some of the stalagmites grow very fast ($>300 \mu\text{m yr}^{-1}$), thus allowing sub-annual sampling. While these stalagmites have relatively low U concentrations (~100 ppb), they also have incredibly low Th concentrations (<100 ppt), allowing them to be dated with precision potentially better than 0.5%.

Acquiring better proxy records will, ultimately, allow me to answer also two broader questions, only alluded to in Chapter 2: What is the monsoonal variability in MSEA over the recent few millennia? What drives its changes?

The knowledge of Asian Monsoon dynamics over the past several millennia is important for climate scientists to reliably predict future monsoon changes in a warming world. The historical lessons learned from the climate-related waxing and waning of civilizations are also crucial if we are to make the right decision when facing new climate-related challenges.

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