Mid-holocene climate change over the Maritime Continent : impact, attribution and mechanisms

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NANYANG
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**MID-HOLOCENE CLIMATE CHANGE
OVER THE MARITIME CONTINENT:
IMPACT, ATTRIBUTION AND
MECHANISMS**

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ASIAN SCHOOL OF THE ENVIRONMENT

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

2018

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

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

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Summary

Speleothem oxygen stable isotope ($\delta^{18}\text{O}$) records suggest that precipitation over the Maritime Continent changed significantly during the Mid-Holocene, a time interval about 6,000 years ago. These precipitation proxy datasets show that climatic conditions over the northern part of the region (e.g., Borneo) became wetter, but drier over the southern part (e.g. Flores island). Such changes were hypothesised to be caused by the northward migration of the Inter-tropical Convergence Zone (ITCZ) and/or the weakening of the El Nino Southern Oscillation (ENSO). These two phenomena are mainly driven by changes in the incoming solar radiation (insolation) and also responding to the feedback of sea surface temperature (SST). However, the exact mechanisms through which the two physical quantities contributed to changing the precipitation over the Maritime Continent are still debated, and hence they are investigated here using General Circulation Models (GCMs).

The Community Climate System Model version 4 (CCSM4) attributes the higher precipitation in the Mid-Holocene to higher annual variability in its monthly climatology, especially over Borneo. The robustness of the Mid-Holocene annual precipitation signal is further confirmed by its persistence in climate experiments without the ENSO, using the Community Atmospheric Model version 4 (CAM4). The maxima of the change in precipitation monthly climatology fall within the July-October period, and they concentrate over the large islands of the Maritime Continent.

The Mid-Holocene higher landmass precipitation is caused by stronger insolation rather than SST feedbacks, as suggested by the numerical experiments using CAM4. The Mid-Holocene stronger insolation increases landmass surface temperature and triggers stronger convection. Then, the low-level southerly wind converges over the landmass and increases humidity, which further strengthens convection over the landmass, suppresses oceanic convection over the South China Sea, and weakens the local Hadley cell.

Chapter 1 – Introduction

Abstract

The Maritime Continent is famously known as the ‘boiler box’ of the atmosphere due to the immense production of rainfall and release of latent heat to the atmosphere in cumulonimbus convection over its many islands and coastlines (Simpson et al., 1993). The Intergovernmental Panel on Climate Change (IPCC) reported a wide range of uncertainty in projecting future climate over the region under various anthropogenic warming scenarios (IPCC AR5, 2014). The uncertainty is mostly due to the disagreement among models in simulating the response of the climate internal variability to the long-term forcings. One way to improve our understanding of the long-term behaviour of the climate system is to evaluate changes observed in the paleoclimate records. Paleoclimate provides critical knowledge to evaluate orbital to centennial scale climate variability in response to various climate forcings such as incoming solar radiation (insolation) and glacial/interglacial boundary conditions.

Mid-Holocene climate over the Maritime Continent was affected by significantly different insolation settings but similar glacial/interglacial boundary condition as compared to the present. Our knowledge about its variability is mainly derived from studies on speleothems and corals. Speleothem $\delta^{18}\text{O}$ records from several locations in the Maritime Continent indicate wetter north over Borneo coinciding with drier south over Flores island. Long-term variabilities of insolation and the feedback from the Western Pacific Warm Pool (WPWP) sea surface temperature (SST) are suggested to be responsible in driving the changes in the mid-Holocene rainfall. However, the detailed mechanisms remain elusive, in particular when we consider the complex land-sea interactions in the Maritime Continent. Therefore the motivation of this thesis is to

analyse the impacts of changing insolation and SST on the climate of the Maritime Continent during the mid-Holocene.

1.1. Background

1.1.1. 21st century global warming

Global warming in the early 21st century has gained global attention, and it is considered as humanity's biggest threat in modern history. Such attention is followed by the Noble Prize award presented to the Intergovernmental Panel on Climate Change (IPCC) in 2007 for the organization's continuous effort in investigating and underlining the main cause, impact and mitigation of the warming climate. The latest assessment report from the IPCC has concluded with high confidence that global warming is mainly caused by anthropogenic emission of greenhouse gases such as CO₂ (IPCC AR5, 2014). Moreover, the long-term prediction of global climate suggests a persistent warming of global surface temperature towards the end of 21st century (IPCC AR5, 2014) (Fig. 1.1). The ramification of such changes affects myriad aspects of human livelihood in the future such as degrading ecosystem and issues in food security (IPCC AR5, 2014). Despite such high importance, there is an uncertainty of ~2°C (0.3-1.7° warmer for the low-end scenario and 2.6-4.8°C warmer for the business-as-usual scenario) for the future projection of earth climate by multiple Global Climate Models (GCMs) involved in the Climate Model Inter-comparison Project (CMIP) (Fig. 1.1). The uncertainty might seem small but the regional warming around the globe can have warming extrema as much as ~11°C for the business-as-usual scenario, especially over the land mass (Fig. 1.2.a). Furthermore, the uncertainty of future warming partly contributes to the inter-model uncertainty in the precipitation trend in most places around the globe (Fig. 1.2.b).

Figure 1.1. Annual mean of global surface temperature anomalies (relative to 1986-2005) of CMIP5 concentration-driven experiments. The experiments for low-end scenario and business-as-usual scenario are based on Representative Concentration Pathways of 2.6 W/m^2 (RCP2.6) and 8.5 W/m^2 (RCP8.5), respectively. Solid line is multi-model mean and the vertical bars are the uncertainty for each scenario (IPCC AR5, 2014).

In general, there are many factors to cause model uncertainty; one factor is on the GCM representation of the climate system (e.g. clouds radiative feedback) in response to the long-term changes such as the increase of greenhouse gas concentration (IPCC AR5, 2014). Vecchi et al. (2008) displayed the various response of sea surface temperature (SST) over the Pacific Ocean to the increase of greenhouse gas emission from GCMs of CMIP3 during early 20th century. Furthermore, the warming trend of GCMs also shows a considerable disagreement with the observations.

Figure 1.2. Surface temperature (a), precipitation (b) change between 1986-2005 and 2081-2100 period of CMIP5 concentration-driven experiments (RCP2.6 as low-end scenario and RCP.8.5 as business-as-usual scenario). Stippling (dots) indicates regions where multi-model mean change is greater than two standard deviation of internal variability and where at least 90 % of models agree on the sign of change. Hatching indicates regions where the multi-model mean change is less than one standard deviation of internal variability (IPCC AR5, 2014).

1.1.2. 21st century warming in Southeast Asia

A comprehensive study for future climate projection over Southeast Asia (SEA) is reported by Marzin et al. (2015) as the Singapore 2nd National Climate Change Study – Phase-1. The downscaled GCMs of CMIP5 over the SEA shows future changes in precipitation ranging within $\sim\pm 1$ mm/day by the end of 21th century. However, regional changes such as over the South China Sea, Philippine and Java can reach to $\sim\pm 2$ mm/day (Fig. 1.3). The mixed phase in the range of uncertainty of the future precipitation over the SEA showcased the level of complexity of the regional climate system in response to the long-term forcing. Furthermore, the multi-dimensional impact of climate change in the future will be difficult to estimate at a high level of confidence.

The future precipitation over the local region of Singapore suffers a similar mixed trend as SEA region (Fig. 1.4). However, the climate variability in one of the GCMs with a positive trend shows spikes in several periods that signify an increase in extreme precipitation towards the end of 21th century. A positive trend in extreme precipitation is considered as the worst case scenario for Singapore since the urbanized country is more sensitive to the hydro-meteorological disaster (floods) as compare to draught events.

The impact of future climate change in Singapore will be experienced differently in the rest of SEA countries. Some countries in SEA as part of the Maritime Continent are sensitive to many other types of meteorological disasters -such as droughts- with direct impact to their agriculture economies. Therefore, constraining the future uncertainties of the precipitation over the Maritime Continent is a very important issue for the well-being of the people in the region.

Figure 1.3. Projected changes in average precipitation (mm/day) in the NDJ season for the 9 dynamically downscaled CMIP5 models (coloured bars in the regional boxes). The middle line represents the median and the ends the minimum and maximum values. Ranges of changes are displayed for the near-term (2010-2039) and the end of the century (2070-2099) compared to a baseline of 1980-2009 and for both RCP4.5 (blue) and RCP 8.5 (red). The orange shading represents the standard deviation of 30 year averages calculated from long historical runs of the GCMs used in IPCC AR5, and the hatching corresponds to regions where the projected median change is less than this standard deviation (Marzin et al., 2015).

Figure 1.4. Running 10-year annual mean rainfall totals. Purple: Model with strongest positive projected change signal (statistically significant upward trend) and Blue: Model with strongest negative projected change signal (statistically significant downward trend). The orange shading shows the spread of the 7 remaining models. Green: Model with median projected change signal (non-statistically significant trend). The solid lines show the least-squares fit through the annual data. 3 models have statistically significant upward trends; 4 models have statistically significant downward trends (Marzin et al., 2015).

1.2. Paleoclimate and natural archive

The aforementioned issues are challenging to overcome in order for GCMs to project future climate with a higher level of confidence. One way to approach such issues is to have a comprehensive understanding of the earth climate system not just at the present but also in the past where the climate conditions can be obtained independently through paleoclimate studies (see Appendix A.1.).

1.2.1 Climate natural archive

Paleoclimate observations are typically achieved by using proxy data extracted from natural archives whose chemical, physical and biological formations are sensitive to one or multiple climate parameters. Chronological controls of these paleoclimate records can be determined by geological dating techniques. Rainfall and sea surface temperature (SST) are the most extensively studied climate parameters. Over the Maritime Continent, proxy data for rainfall and SST can be extracted from natural archives, particularly speleothems and corals.

1.2.1.1 Speleothems

Speleothems form in limestone caves and are made of calcium carbonate (CaCO_3) precipitated from cave dripping water that originates from surface rains. The rain water becomes acidic after absorbing soil CO_2 and can then dissolve limestone bedrock. Later, the water enters caves and degases CO_2 ; carbonate chemicals become supersaturated in water and precipitation occurs to form speleothems. The oxygen isotope ratio of speleothems is one of the most commonly used proxies, and is often linked to regional rainfall intensity in the tropics (Cheng et al., 2012). Hence, speleothem oxygen isotope records are able to infer changes in monsoon rainfall intensity (Wang et al., 2001). The age control of speleothem is usually determined by using the high-precision uranium-

series dating method, which can date carbonate samples from 1 yr up to ~700,000 yrs old. The basic theory behind the speleothem oxygen isotope proxy and the uranium-series dating method are further described in Appendix A.2.

1.2.1.2 Corals

Shallow-water corals are mostly located near sea surface where sunlight is abundant and water temperature is warm enough (~18°C or higher), which limits their regional coverage to mostly within the tropics (Ruddiman, 2008). Coral skeleton is made of aragonite (CaCO_3) secreted from sea water during coral organism's biological activities (Felis and Patzold, 2004; Lough, 2010). The growth rate of coral's skeleton (calcification) is proportional to SST and nutrient content in seawater (Fairbanks and Dodge, 1979; Lough and Barnes, 1990). Coral samples can be extracted by drilling massive coral heads. In regions with strong seasonality of temperature and/or rainfall, annual bands can exist in a coral, which allows visual counting as sample chronological controls (dating). However, this is typically limited for living corals. The dating for fossil corals requires radioisotopic dating technique such as radiocarbon or uranium-series methods.

1.3. Paleoclimate of the Maritime Continent

1.3.1. Historical perspective

The Maritime Continent is a geographical terminology of the archipelagic region in Southeast Asia coined by Ramage (1968). The region comprises several islands nations including Indonesia, Malaysia, Singapore, Philippine, East Timor and Papua New Guinea. The Maritime Continent is located between Asian and Australia continents and also between the Pacific and Indian Oceans. The Maritime Continent is geologically comprises of two main shelves that are mostly underwater with major islands made up mostly from volcanic activity as part of the Pacific volcanic rim famously known as the

Ring of Fire, and also from accretionary wedge materials (Hutchison, 2007). The two main shelves are the Sunda and Sahul shelves which are located at the west and east of the Maritime Continent respectively. Sunda and Sahul shelves are an extension of the continental Southeast Asia and Australia respectively that have been colliding since ~15 Ma (Audley-Charles, 1982). The two shelves are separated by a lateral deep sea boundary located between Bali and Lombok and extending towards east of Borneo all the way up to the east of Philippine. The deep sea boundary is overlapped with the famously known Wallace line which is based on the different clusters of biodiversity discovered by Alfred Russel Wallace during his famous journey in 1859. The clusters are suggested to be related to the different evolutionary path due to the deep sea boundary (Mayr, 1944).

The Wallace line suggests a shared speciation throughout the Sunda Shelf region (Myers et al., 2000). Such common features in the ecosystem are shared among islands, indicating that species migration between islands should have taken place in the past (Gathorne-Hardy et al., 2002, Meijaard, 2004). The migration can only be done through the so-called 'land-bridges' between the major islands. Such 'land-bridges' are made possible through the exposure of the shallow bathymetry of the Sunda Shelf caused by receding sea level (Sathiamurty & Voris, 2006, Wurster et al., 2010). The exposed bathymetry increases tremendously the land area known as Sundaland. The exposure of Sundaland in the past is confirmed by the existence of submerged river systems which are extensions of the present day rivers from mainland Indo-China and the major islands (Voris, 2000). The drop of sea level is related to the changes of the earth climate regime, including the increase of the volumes of polar ice sheets and mountain glaciers on glacial-interglacial cycles. The exposure and submerge of Sundaland has occurred throughout at least for the last 2 million years.

1.3.2. Paleoclimate epochs

Paleoclimate records covering the past tens of thousands years (ka) or longer shows variability in centennial to orbital timescales. Long-term climate variability is mostly attributed to the two main paleoclimate forcings; they are Earth orbital configurations and glacial boundary conditions (Petit et al., 1999, Appendix A.1.). Among numerous climate variations recorded in the proxy datasets, those during certain time periods can be of specific research interest because of their unique climate conditions, such as temperature, sea-level, greenhouse gas contents in the atmosphere, etc. Hence, studying the climate conditions in these epochs –for instance, the Last Glacial Maximum and the mid-Holocene- can provide more insights of the dominant physical mechanisms of the climate system.

1.3.3. Last Glacial Maximum

The most recent exposure of Sundaland occurred during the Last Glacial Maximum (LGM) period ~21 ka which has become an important epoch to study. Hydroclimate proxy records from multiple natural archives over the Maritime Continent suggest an overall drying during the LGM (Fig. 1.5). The exposure of Sundaland reduces ocean coverage in the Maritime Continent, which is part of the Western Pacific Warm Pool (WPWP). The analyses on present day climate describe the WPWP as the most important large-scale oceanic feature in driving torrential rainfall over the region through the mechanism of moist-convection (Chiang et al., 2009). Thus, the modification of the WPWP is hypothesized as the main cause of the drying Maritime Continent during the LGM as a response to the large scale forcing (Vechi et al., 2006; Vechi & Soden, 2007).

DiNezio and Tierney (2013) investigated the drying Maritime Continent during the LGM by using multi GCMs involved in the Paleoclimate Model Inter-comparison Project (PMIP). The study carefully validated the performance of each model with the

proxy data and derived their conclusion only based on the better performed model. The study suggested a drier climate over the Maritime Continent during the LGM in association with suppressed convection and mostly as a response to the glacial maximum boundary conditions (Fig. 1.5). The suppressed convection is due to a colder Sundaland compare to its surrounding ocean that results in mass divergence and anomalous subsidence. The zonally overturning large-scale atmospheric circulation (Walker circulation) was reversed in comparison to pre-Industrial era (present day) climate, with an ascending motion and more active convection over the western Indian Ocean.

Figure 1.5 Network of terrestrial proxies (circles) of drier (brown), unchanged (white) or wetter (blue) climate during the LGM. Coastlines referred to a 120 m drop of sea level. Rainfall changes (ΔP) are LGM minus pre-Industrial era of HadCM3. Contour shows multi-model annual-mean precipitation during pre-Industrial era (DiNezio & Tierney, 2013).

A much larger landmass during the LGM resulted in a drier climate over the Maritime Continent and reversed Walker circulation. Such extreme changes in climate regime are very important to be properly represented by the GCM simulation. However, the LGM period actually experiences minor differences in insolation from present day so that the role of insolation could almost be negligible (Bush & Philander, 1999, DiNezio et al., 2011). The investigation on the role of insolation hence requires a different climate regime that has significant changes in insolation from today but with similar glacial/interglacial boundary conditions.

1.3.4. Mid-Holocene

Mid-Holocene (~6 ka) is a recent epoch that is known to be significantly different in insolation from the present day following the long-term cycle of the Earth's orbit (Appendix A.1). The largest difference in insolation occurs in the higher latitudes during boreal summer (northern hemisphere) and fall (southern hemisphere) as much as $\sim 30 \text{ W/m}^2$, which is one order of magnitude larger than the difference between the LGM and today (Bush & Philander 1999; Chiang et al., 2009). The glacial boundary conditions and atmospheric greenhouse gas concentrations during the mid-Holocene are similar to the present day (Indermühle et al., 1999). Such climate regime provides a unique scenario to test the response of the climate system to the significant changes in insolation (Fig. 1.6).

Figure 1.6. Insolation for the mid-Holocene, pre-Industrial and their differences based on Berger (1978).

The most prominent feature of the mid-Holocene climate change is the intensification of Asian Summer Monsoon shown by speleothem $\delta^{18}\text{O}$ records from Mainland China (Wang et al., 2001; Cheng et al., 2016). Monsoon comes with strong seasonal rainfall events in the tropics which help define tropical seasonality. The insolation driven climate phenomena take place in West Africa, Asia and Central America during boreal summer, while during boreal winter they occur in Australia, Southern Africa and South America (Ramage, 1971; Webster 1987; Nogues-Paegle et al., 2002; Sultan et al., 2003).

In Asian monsoon region, speleothem $\delta^{18}\text{O}$ is typically interpreted as a proxy of summer monsoon intensity based on the negative correlation between the weighted mean of rainfall $\delta^{18}\text{O}$ and annual rainfall, which is dominated by summer monsoon rainfall occurring in June, July and August (JJA). This relationship is also referred as the amount effect; hence on longer timescales, speleothem $\delta^{18}\text{O}$ indicates summer monsoon rainfall intensity. The lower $\delta^{18}\text{O}$ values are, the higher monsoon intensity, and vice versa. The proxy-derived Asian monsoon intensity varies coherently with the Northern Hemisphere summer insolation (July 21 at 65°N) (Fig. 1.7).

Figure 1.7. Speleothem $\delta^{18}\text{O}$ from Hulu cave (light blue) and Sanbao cave (other colors), overlaid with Northern Hemisphere (65°N) Summer Insolation (July 21st) (grey) (Wang et al., 2008).

The most notable rainfall proxy data from the Maritime Continent are speleothem $\delta^{18}\text{O}$ records from Gunung Buda, northern Borneo (4°N, 114°E) and Liang Luar, Flores (8° 32'S, 120° 26'E). These two caves are located in different hemispheres across the equator, which signifies a latitudinal variation of climate over the Maritime Continent.

1.3.4.1. Borneo speleothem $\delta^{18}\text{O}$ records

Partin et al., (2007) shows a coherency between Borneo speleothem $\delta^{18}\text{O}$ records and the equatorial insolation in September (Fig. 1.8.a, d, & Fig. 1.11), suggesting a role of intensified convection in response to the increased insolation during the second inter-monsoon season (September-October-November) (Fig. 1.6). The intensified convection during the 2nd inter-monsoon season suggests a regional scale climate activity in relation with the annual southward migration of the Inter-tropical Convergence Zone (ITCZ).

The annual variability of the local surface temperature (land and sea) as a response to the insolation is also important in driving the migration of ITCZ over the Maritime Continent. Lestari and Iwasaki (2006) suggested that the SST in the South China Sea plays a crucial role in driving the northward migration of the ITCZ over the Maritime Continent during the boreal summer. The warm SST increases the oceanic convection and becomes the main driver for the local Hadley cell of the season. However, Lestari & Iwasaki (2006) also suggested that the land response to the summer insolation suppresses the northward migration of the ITCZ. Thus, the opposite behaviours of the ITCZ migration due to the land and sea responses to the insolation are the two competing factors in influencing the ITCZ over the Maritime Continent.

The response of local SST and the land-mass in the Maritime Continent to the stronger boreal summer-to-fall insolation during the mid-Holocene can modify the ITCZ migration over the region. In turn, the modified ITCZ is likely modifies the annual-and-semi-annual precipitation over the region, including Borneo.

1.3.4.2. Flores speleothem $\delta^{18}\text{O}$ records

The complexity of the Maritime Continent climate is further displayed by a speleothem $\delta^{18}\text{O}$ record from the southern region of the Maritime Continent. Griffith et al. (2009) showed an incoherency between the speleothem $\delta^{18}\text{O}$ record from Flores and boreal fall (SON) equatorial insolation, which is different from the observation in Borneo $\delta^{18}\text{O}$ record (Fig. 1.8.b, d, & Fig. 1.11). Flores $\delta^{18}\text{O}$ is suggested to respond to the eustatic sea-level rise that was increasing at a higher rate during early Holocene, as compared to the rest of the Holocene period, which provided additional moisture supply and hence caused increased precipitation.

The regional nature of the climate variability can be inferred by the stark difference when the Flores record is compared with the speleothem records from Borneo and Kimberley, North-Coast of Australia (15°18'S, 128°37'E). Denniston et al. (2013) reported a wetter climate over the Kimberley region during the mid-Holocene, which is nearly in-phase with the Borneo record and the boreal fall (SON) equatorial insolation (Fig. 1.8.c, d). Mid-Holocene wetter climate over Borneo and Kimberley appears to contradict the drier climate at Flores Island, which could suggest that such drying feature occurred in a small spatial scale. Present day climate analyses suggest the importance of small scale features over the Maritime Continent such as mesoscale convective precipitation over islands (Yamanaka et al., 2016, 2018). The small scale features are associated with the complexity of land-sea distribution in the region.

Figure 1.8. Speleothem $\delta^{18}\text{O}$ records for Borneo (Partin et al., 2007) (a), Flores (Griffith et al., 2009) (b) and North-Coast of Australia (Denniston et al., 2013) (c) for the Holocene period (a,b and c are modified from Denniston et al., 2013). The lowest panel is the equatorial insolation for SON season calculated based on Berger (1978) (d). The black horizontal bars roughly indicate the mid-Holocene era.

1.3.4.3. Western Pacific Warm Pool variability

Large scale SST in the Indo-Pacific basin can play an equally important role in affecting Maritime Continent climate due to the existence of the WPWP, which is strongly related to the El Nino Southern Oscillation (ENSO). ENSO is the variability of the Walker Circulation occurring on the inter-annual time-scale (2-7 yrs) and it is the dominant atmospheric signals outside annual and semi-annual spectra (Lian & Chen, 2012). The variability occurs along with changes in the zonal SST gradient of the equatorial Pacific. One phase is known as El Nino, with a gentler SST gradient that significantly reduces the range of cold tongue to retreat eastward and give warmer SST at the coastal Peru (Fig. 1.9).

The gentler SST gradient is coupled by a weaker trade-wind through an air-sea feedback mechanism introduced by Bjerknes (1966, 1969). The positive feedback between equatorial Pacific SST and trade-wind is also responsible for its opposite phase (La Nina). La Nina extends the cold tongue further to the eastern side of the Maritime Continent which also shifts the warm pool and the convective region westward (Fig. 1.9). The inter-annual signal is named as ENSO in order to emphasize the main driver which is the air-sea interaction.

Figure 1.9. Surface temperature, circulation, convective zone and thermocline in the tropical Pacific during El Nino, Normal and La Nina conditions (Chiang, 2009 adapted from NOAA/PMEL/TAO Project Office, <https://www.pmel.noaa.gov/elnino/>).

1.3.4.4. Mid-Holocene rainfall and Western Pacific Warm Pool

Chen et al. (2016) further explored the possible influence of the ENSO on the Borneo rainfall during the mid-Holocene since the present day rainfall over the island has a weak seasonality and there is an inter-annual coherency with the ENSO (Moerman et al., 2013). Borneo $\delta^{18}\text{O}$ record shows a low inter-annual variability during the mid-Holocene correlated to weaker ENSO activity (Cobb et al., 2013; Carre et al., 2014). The dynamical explanation of such influence is via the changes in air-sea interaction as simulated by coupled-GCMs (Clement et al., 1999, 2000; Liu et al., 2000). Mid-Holocene insolation is suggested to strengthen the trade winds during boreal fall season which increased the upwelling and amplified the Eastern Equatorial Pacific ‘Cold Tongue’ leading to a weaker ENSO activity.

Coral records from the tropical Pacific suggest that the weakening of the mid-Holocene ENSO activity might also be related to changes in regional SST of the Maritime Continent. Abram et al. (2009) used coral Sr/Ca records from two locations at the longitudinal ‘perimeter’ of the WPWP (west Coast Sumatra and east Coast Papua) to show colder SST during the mid-Holocene (Fig. 1.10 & Fig. 1.11). Proxy records from the South China Sea however show no consensus on the changes of SST during this period (Steinke et al., 2001, Steven et al., 2009, Yokoyama et al., 2011). The colder SST at the ‘perimeter’ of the WPWP during the mid-Holocene is interpreted as a spatial contraction of the warm pool which could reduce the strength of the Walker circulation throughout the Indo-Pacific basin. Thus, it would weaken any related climate phenomena such as the ENSO.

Figure 1.10. Sr/Ca estimates of mean SST anomalies at the Mentawai (solid circles) and Muschu/Koil (open circles) sites since the mid-Holocene. Error bars show 2σ ranges on age determinations, and the combined error on mean SST estimates. Grey lines show weighted mean SST (and 2SE in grey shading) at the Mentawai site over intervals where a large number of Mentawai corals provide independent SST estimates (Abrams et al., 2009).

However, two critical issues remain to be addressed from Borneo $\delta^{18}\text{O}$ record in relation with ENSO:

1. The dynamical explanation about the mid-Holocene ENSO derived from GCM simulation has not been extended to the Maritime Continent (Clement et al., 1999, 2000; Liu et al., 2000). Thus, no direct mechanism is suggested that directly explains the mid-Holocene weaker ENSO as the main cause for wetter climate over Borneo.
2. The hypothesis of ENSO influence did not consider the complexity of the Maritime Continent (e.g. land-sea distribution, topography, etc.) that has been previously suggested to drive heterogeneous ENSO impact over the region (Chang et al., 2004; Jia et al., 2015; Jiang and Li, 2017).

Figure 1.11. Top panel : Fall season (SON) precipitation climatology (2000-2011) from TRMM satellite (plotted only > 0.1 mm/hour), overlaid by locations of speleothem $\delta^{18}O$ records at Borneo (Partin et al., 2007), Flores (Griffith et al., 2009) and North-Coast of Australia (Denniston et al., 2013). Also shown are sites of coral Sr/Ca records from Mentawai Islands (West-Coast Sumatra) and Muschu Islands (East-Coast Papua) (Abram et al., 2009). Bottom panel : Mid-Holocene climate change over the Maritime Continent illustrated from the rainfall and SST proxy records.

1.4. General motivation of this thesis

Spatio-temporal climate pattern over the Maritime Continent throughout the Holocene (especially the mid-Holocene) appears to be heterogeneous (Fig. 1.11). The rainfall proxy from Borneo suggests insolation and SST as the major factors that controlled the mid-Holocene climate over the region. However the exact mechanism needs further analysis. GCM simulations can be used for such analysis.

The coarse spatial resolution of the previous generation GCMs limits the discussions only to large scale climate phenomena such as monsoon and ENSO (Denniston et al., 2013; Ann & Choi, 2014). However, present day climate analyses shows that the complexity of land-sea distribution of the Maritime Continent also plays an important role in determining the climate of the region (Yamanaka et al., 2016, 2018). Therefore, GCM with sufficient spatial resolution is used in this study in order to resolve regional to mesoscale processes. Model description, validation and attribution are presented in Chapter 2.

The fundamental mechanisms of climate change can be investigated by referring to the current understanding of the atmospheric dynamics. Thus, this thesis provide an adequate explanation to the mid-Holocene climate change over the Maritime Continent by investigating the dynamical response of the regional tropical atmosphere to the global changes of insolation and SST. The dynamical investigation through numerical experiments is presented in Chapters 3 and 4.

1.5. References

- Abram, N.J., McGregor, H.V., Gagan, M.K., Hantoro, W.S. and Suwargadi, B.W., 2009. Oscillations in the southern extent of the Indo-Pacific Warm Pool during the mid-Holocene. *Quaternary Science Reviews*, 28(25-26), pp.2794-2803.
- An, S.I. and Choi, J., 2014. Mid-Holocene tropical Pacific climate state, annual cycle, and ENSO in PMIP2 and PMIP3. *Climate dynamics*, 43(3-4), pp.957-970.
- Audley-Charles, M.G., 1981. Geological history of the region of Wallace's Line. *Wallace's line and plate tectonics*, pp.24-35.
- Berger, A., 1978. Long-term variations of daily insolation and Quaternary climatic changes. *Journal of the Atmospheric Sciences*, 35(12), pp.2362-2367.
- Bjerknes, J., 1966. A possible response of the atmospheric Hadley circulation to equatorial anomalies of ocean temperature. *Tellus*, 18(4), pp.820-829.
- Bjerknes, J., 1969. Atmospheric teleconnections from the equatorial Pacific. *Monthly Weather Review*, 97(3), pp.163-172.
- Bush, A.B. and Philander, S.G.H., 1999. The climate of the Last Glacial Maximum: Results from a coupled atmosphere-ocean general circulation model. *Journal of Geophysical Research: Atmospheres*, 104(D20), pp.24509-24525.
- Carré, M., Sachs, J.P., Purca, S., Schauer, A.J., Braconnot, P., Falcón, R.A., Julien, M. and Lavallée, D., 2014. Holocene history of ENSO variance and asymmetry in the eastern tropical Pacific. *Science*, 345(6200), pp.1045-1048.
- Chang, C.P., Wang, Z., Ju, J. and Li, T., 2004. On the relationship between western maritime continent monsoon rainfall and ENSO during northern winter. *Journal of Climate*, 17(3), pp.665-672.
- Cheng, H., Sinha, A., Wang, X., Cruz, F.W. and Edwards, R.L., 2012. The Global Paleomonsoon as seen through speleothem records from Asia and the Americas. *Climate Dynamics*, 39(5), pp.1045-1062.

- Cheng, H., Edwards, R.L., Sinha, A., Spötl, C., Yi, L., Chen, S., Kelly, M., Kathayat, G., Wang, X., Li, X. and Kong, X., 2016. The Asian monsoon over the past 640,000 years and ice age terminations. *Nature*, 534(7609), p.640.
- Chen, S., Hoffmann, S.S., Lund, D.C., Cobb, K.M., Emile-Geay, J. and Adkins, J.F., 2016. A high-resolution speleothem record of western equatorial Pacific rainfall: Implications for Holocene ENSO evolution. *Earth and Planetary Science Letters*, 442, pp.61-71.
- Chiang, J.C., 2009. The tropics in paleoclimate. *Annual Review of Earth and Planetary Sciences*, 37, pp.263-297.
- Clement, A.C., Seager, R. and Cane, M.A., 1999. Orbital controls on the El Niño/Southern Oscillation and the tropical climate. *Paleoceanography*, 14(4), pp.441-456.
- Clement, A.C., Seager, R. and Cane, M.A., 2000. Suppression of El Niño during the Mid-Holocene by changes in the Earth's orbit. *Paleoceanography*, 15(6), pp.731-737.
- Cobb, K.M., Westphal, N., Sayani, H.R., Watson, J.T., Di Lorenzo, E., Cheng, H., Edwards, R.L. and Charles, C.D., 2013. Highly variable El Niño–Southern Oscillation throughout the Holocene. *Science*, 339(6115), pp.67-70.
- Denniston, R.F., Wyrwoll, K.H., Polyak, V.J., Brown, J.R., Asmerom, Y., Wanamaker Jr, A.D., LaPointe, Z., Ellerbroek, R., Barthelmes, M., Cleary, D. and Cugley, J., 2013. A stalagmite record of Holocene Indonesian–Australian summer monsoon variability from the Australian tropics. *Quaternary Science Reviews*, 78, pp.155-168.
- DiNezio, P.N., Clement, A., Vecchi, G.A., Soden, B., Broccoli, A.J., Otto-Bliesner, B.L. and Braconnot, P., 2011. The response of the Walker circulation to Last Glacial Maximum forcing: Implications for detection in proxies. *Paleoceanography and Paleoclimatology*, 26(3).

- DiNezio, P.N. and Tierney, J.E., 2013. The effect of sea level on glacial Indo-Pacific climate. *Nature Geoscience*, 6(6), p.485.
- Fairbanks, R.G. and Dodge, R.E., 1979. Annual periodicity of the $^{18}\text{O}/^{16}\text{O}$ and $^{13}\text{C}/^{12}\text{C}$ ratios in the coral *Montastrea annularis*. *Geochimica et Cosmochimica Acta*, 43(7), pp.1009-1020.
- Felis, T. and Pätzold, J., 2004. Climate reconstructions from annually banded corals. *Global environmental change in the ocean and on land*, pp.205-227.
- Gathorne-Hardy, F.J., Davies, R.G., Eggleton, P. and Jones, D.T., 2002. Quaternary rainforest refugia in south-east Asia: using termites (Isoptera) as indicators. *Biological Journal of the Linnean Society*, 75(4), pp.453-466.
- Griffiths, M.L., Drysdale, R.N., Gagan, M.K., Zhao, J.X., Ayliffe, L.K., Hellstrom, J.C., Hantoro, W.S., Frisia, S., Feng, Y.X., Cartwright, I. and Pierre, E.S., 2009. Increasing Australian–Indonesian monsoon rainfall linked to early Holocene sea-level rise. *Nature Geoscience*, 2(9), p.636.
- Hutchison, C.S., 1989. *Geological evolution of South-east Asia* (Vol. 13, p. 368). Oxford: Clarendon Press.
- Indermühle, A., Stocker, T.F., Joos, F., Fischer, H., Smith, H.J., Wahlen, M., Deck, B., Mastroianni, D., Tschumi, J., Blunier, T. and Meyer, R., 1999. Holocene carbon-cycle dynamics based on CO_2 trapped in ice at Taylor Dome, Antarctica. *Nature*, 398(6723), p.121.
- IPCC, Contribution of Working Groups I, II and III to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (2014). *Climate Change 2014 : Synthesis Report*. IPCC, Geneva, Switzerland: Cambridge University Press.
- Jia, X., Ge, J. and Wang, S., 2015. Diverse impacts of ENSO on wintertime rainfall over the Maritime Continent. *International Journal of Climatology*, 36(9), pp.3384-3397.

- Jiang, L. and Li, T., 2017. Why rainfall response to El Niño over Maritime Continent is weaker and non-uniform in boreal winter than in boreal summer. *Climate Dynamics*, pp.1-19.
- Lestari, R.K. and Iwasaki, T., 2006. A GCM study on the roles of the seasonal marches of the SST and land-sea thermal contrast in the onset of the Asian summer monsoon. *Journal of the Meteorological Society of Japan. Ser. II*, 84(1), pp.69-83.
- Lian, T. and Chen, D., 2012. An evaluation of rotated EOF analysis and its application to tropical Pacific SST variability. *Journal of Climate*, 25(15), pp.5361-5373.
- Liu, Z., Kutzbach, J. and Wu, L., 2000. Modeling climate shift of El Nino variability in the Holocene. *Geophysical Research Letters*, 27(15), pp.2265-2268.
- Lough, J.M. and Barnes, D.J., 1990. Intra-annual timing of density band formation of Porites coral from the central Great Barrier Reef. *Journal of Experimental Marine Biology and Ecology*, 135(1), pp.35-57.
- Lough, J.M., 2010. Climate records from corals. *Wiley Interdisciplinary Reviews: Climate Change*, 1(3), pp.318-331.
- Marzin, C., Rahmat, R., Bernie, D., Bricheno, L., Buonomo, E., Calvert, D., Cannaby, H., Chan, S., Chattopadhyay, M., Cheong, W.K. and Hassim, M.E., 2015. Singapore's second national climate change study–Phase 1. *MET OFFICE, E., UK, CENTRE FOR CLIMATE RESEARCH SINGAPORE, S., NATIONAL OCEANOGRAPHY CENTRE, L., UK, CSIRO, A. & NEWCASTLE UNIVERSITY, N., UK (eds.)*.
- Mayr, E., 1944. Wallace's Line in the light of recent zoogeographic studies. *The Quarterly Review of Biology*, 19(1), pp.1-14.
- Meijaard, E., 2004. Biogeographic history of the Javan leopard *Panthera pardus* based on a craniometric analysis. *Journal of Mammalogy*, 85(2), pp.302-310.

- Moerman, J.W., Cobb, K.M., Adkins, J.F., Sodemann, H., Clark, B. and Tuen, A.A., 2013. Diurnal to interannual rainfall $\delta^{18}\text{O}$ variations in northern Borneo driven by regional hydrology. *Earth and Planetary Science Letters*, 369, pp.108-119.
- Myers, N., Mittermeier, R.A., Mittermeier, C.G., Da Fonseca, G.A. and Kent, J., 2000. Biodiversity hotspots for conservation priorities. *Nature*, 403(6772), p.853.
- Nogués-Paegle, J., Mechoso, C.R., Fu, R., Berbery, E.H., Chao, W.C., Chen, T.C., Cook, K., Diaz, A.F., Enfield, D., Ferreira, R. and Grimm, A.M., 2002. Progress in Pan American CLIVAR research: understanding the South American monsoon. *Meteorologica*, 27(12), pp.1-30.
- Partin, J.W., Cobb, K.M., Adkins, J.F., Clark, B. and Fernandez, D.P., 2007. Millennial-scale trends in west Pacific warm pool hydrology since the Last Glacial Maximum. *Nature*, 449(7161), p.452.
- Petit, J.R., Jouzel, J., Raynaud, D., Barkov, N.I., Barnola, J.M., Basile, I., Bender, M., Chappellaz, J., Davis, M., Delaygue, G. and Delmotte, M., 1999. Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature*, 399(6735), p.429.
- Ramage, C.S., 1968. Role of a tropical “maritime continent” in the atmospheric circulation. *Mon. Wea. Rev.*, 96(6), pp.365-370.
- Ruddiman, W.F., 2008. *Earth's climate: past and future*, 2nd Edition. W. H. Freeman.
- Sathiamurthy, E. and Voris, H.K., 2006. Maps of Holocene sea level transgression and submerged lakes on the Sunda Shelf. *Tropical Natural History*, (2), pp.1-44.
- Simpson, J., Keenan, T.D., Ferrier, B., Simpson, R.H. and Holland, G.J., 1993. Cumulus mergers in the maritime continent region. *Meteorology and Atmospheric Physics*, 51(1-2), pp.73-99.

- Steinke, S., Kienast, M., Pflaumann, U., Weinelt, M. and Stattegger, K., 2001. A high-resolution sea-surface temperature record from the tropical South China Sea (16,500–3000 yr BP). *Quaternary Research*, 55(3), pp.352-362.
- Stevenson, J., Siringan, F., Finn, J.A.N., Madulid, D. and Heijnis, H., 2010. Paoay Lake, northern Luzon, the Philippines: a record of Holocene environmental change. *Global Change Biology*, 16(6), pp.1672-1688.
- Sultan, B., Janicot, S. and Diedhiou, A., 2003. The West African monsoon dynamics. Part I: Documentation of intraseasonal variability. *Journal of Climate*, 16(21), pp.3389-3406.
- Vecchi, G.A., Soden, B.J., Wittenberg, A.T., Held, I.M., Leetmaa, A. and Harrison, M.J., 2006. Weakening of tropical Pacific atmospheric circulation due to anthropogenic forcing. *Nature*, 441(7089), p.73.
- Vecchi, G.A. and Soden, B.J., 2007. Global warming and the weakening of the tropical circulation. *Journal of Climate*, 20(17), pp.4316-4340.
- Vecchi, G.A., Clement, A. and Soden, B.J., 2008. Examining the tropical Pacific's response to global warming. *Eos, Transactions American Geophysical Union*, 89(9), pp.81-83.
- Voris, H.K., 2000. Maps of Pleistocene sea levels in Southeast Asia: shorelines, river systems and time durations. *Journal of Biogeography*, 27(5), pp.1153-1167.
- Wang, Y.J., Cheng, H., Edwards, R.L., An, Z.S., Wu, J.Y., Shen, C.C. and Dorale, J.A., 2001. A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *science*, 294(5550), pp.2345-2348.
- Wang, Y., Cheng, H., Edwards, R.L., Kong, X., Shao, X., Chen, S., Wu, J., Jiang, X., Wang, X. and An, Z., 2008. Millennial-and orbital-scale changes in the East Asian monsoon over the past 224,000 years. *Nature*, 451(7182), p.1090.

- Webster, P.J., 2001. The large-scale structure of the tropical atmosphere. In *Large-Scale Dynamical Processes in the Atmosphere* (p. 235).
- Wurster, C.M., Bird, M.I., Bull, I.D., Creed, F., Bryant, C., Dungait, J.A. and Paz, V., 2010. Forest contraction in north equatorial Southeast Asia during the Last Glacial Period. *Proceedings of the National Academy of Sciences*, 107(35), pp.15508-15511.
- Yamanaka, M.D., 2016. Physical climatology of Indonesian maritime continent: An outline to comprehend observational studies. *Atmospheric Research*, 178, pp.231-259.
- Yamanaka, M.D., Ogino, S.Y., Wu, P.M., Jun-Ichi, H., Mori, S., Matsumoto, J. and Syamsudin, F., 2018. Maritime continent coastlines controlling Earth's climate. *Progress in Earth and Planetary Science*, 5(1), p.21.
- Yokoyama, Y., Suzuki, A., Siringan, F., Maeda, Y., Abe-Ouchi, A., Ohgaito, R., Kawahata, H. and Matsuzaki, H., 2011. Mid-Holocene palaeoceanography of the northern South China Sea using coupled fossil-modern coral and atmosphere-ocean GCM model. *Geophysical Research Letters*, 38(8).

Chapter 2 – Mid-Holocene climate change over the Maritime Continent in the CCSM4

Abstract

I establish that the Community Climate System Model version 4 (CCSM4) is able to simulate the present-day climate. The validation was done for the Inter-tropical Convergence Zone (ITCZ), the El Nino Southern Oscillation (ENSO), and also their impact on the climate over the Maritime Continent. Model-proxy comparisons for the precipitation and sea surface temperature (SST) show that the CCSM4 is one of the models within the Paleoclimate Model Inter-comparison Project Phase-3 (PMIP3) that can successfully simulate the mid-Holocene climate change over the Maritime Continent.

Mid-Holocene precipitation change over the Maritime Continent as simulated by the CCSM4 is mostly attributed to the stronger annual signal of their monthly climatology. The climatological precipitation maxima during the mid-Holocene occurred during the July-October (JASO) period, and concentrated over the landmass of the Maritime Continent. Mid-Holocene higher precipitation over Borneo during JASO is coherent with a stronger insolation and warmer SST over the South China Sea, which signifies the importance of both physical quantities.

2.1. General Circulation Model (GCM) in paleoclimatology

The General Circulation Model (GCM) is an integral part in the study of paleoclimate. The GCM is a three-dimensional numerical representation of the earth's climate based on the conservation of mass, momentum, and energy. The conservation of these physical quantities is described mathematically using the concept of fluid dynamics (Chase & Barry, 2004). Climate internal dynamics is a state of equilibrium of the climate system as a response to climate forcings and storage capacity of the climate components including atmosphere, land surface, ocean and sea ice (Fig. 2.1).

Figure 2.1. Schematic diagram of the interactions between climate components represented by the GCM (Druryan et al., 1975).

The development of GCM has been one of the biggest achievements in modern climate science with a large contribution in improving the well-being of humanity, especially for its prediction capability on future climate. Climate prediction is made possible by the numerical system which solves the prognostic equations (differential form) of the atmospheric parameters in reaching equilibrium as a response to the applied forcings (Eq.2.1). Such numerical solution is sensitive to climate forcings such as insolation, glacial boundary conditions and greenhouse gas concentrations. So one would expect that a different state of equilibrium would be reached by a climate regime with a different combination of the climate forcings.

$$\frac{d(\text{storage of } X)}{dt} = \text{positive forcing} - \text{negative forcing} \quad \text{Eq.2.1}$$

The variety of earth's climate regime as shown by paleoclimate records can be numerically simulated by GCM through different configurations of the climate forcings and the boundary conditions; these configurations are known as climate scenarios. Paleoclimate scenarios are essentially a modification from the present day climate scenario which is the basis of the development of the GCM itself. Present day climate can be differentiated as before and after the influence of anthropogenic greenhouse gas emission. The absence of anthropogenic greenhouse gas ensures that the climate system is influenced by its own natural mechanism. The paleoclimate community labels the climate before 1850s as the 'pre-industrial' era, as there was a negligible anthropogenic influence before the industrial revolution. The configuration for the pre-Industrial scenario sets the atmosphere CO₂ concentration at ~280 ppm, expecting for a slightly colder global surface temperature compared to the later 20th century climate (Mahowald et al., 2006).

2.2. Community Climate System Model (CCSM)

The CCSM version 4 (CCSM4) is a Coupled-GCM (CGCM) developed by the National Center for Atmospheric Research (NCAR), USA that was made available publicly on the 1st April 2010. The CCSM4 has been closely involved in several international projects for future climate projection and paleoclimate study as reported by the IPCC Assessment Reports. Gent et al. (2011) highlights several improvements of the CCSM4 performances in the tropics to give the better representation of:

1. El Nino Southern Oscillation (ENSO) and the SST of the Pacific Ocean.
2. Frequency distribution of tropical precipitation and its intra-seasonal variability.
3. Mean and annual cycle of SST along the Equator in the eastern Pacific.
4. Surface land temperature.
5. Extreme events such as very heavy rainfall.

These improvements were made possible due to changes in the model specification such as higher spatial resolution ($\sim 1^\circ$ grid size) and physical parameterizations such as deep convection scheme. Improvement in higher latitude and polar region are also showcased by the CCSM4 such as the better representation of the North Atlantic Meridional Overturning Circulation (MOC) and the Arctic sea ice concentration.

Despite the significant improvements, the CCSM4 is still suffering from persistent issues such as double Inter-Tropical Convergence Zone (ITCZ) and biases in the cloud forcing and the mean surface temperature. Biases in the CCSM4 are suspected mainly result from the poor representation of aerosols among other causes.

2.2.1. CCSM4 validation for Tropical Pacific SST and ENSO

Double ITCZ is one of the persistent rainfall biases over the tropical Pacific appearing in most of the current state-of-the-art CGCMs. The excessive precipitation over the tropical Pacific causes a significant cold SST bias. The bias is explained by air-sea interaction processes known as the Bjerknes feedback, which simulates stronger trade winds, excessive latent heat flux, and insufficient shortwave flux (Lin, 2007; Li & Xie, 2014).

An excessive and narrow colder SST anomaly that extends into the western Pacific (cold tongue) produces a cold SST bias within the Western Pacific Warm Pool (WPWP) (Zheng et al., 2012). The SST bias in the WPWP contributes to the model representation of the ENSO and its impact over the Maritime Continent (Taschetto et al., 2014).

To check for any SST bias in the CCSM4, I used the tropical Indo-Pacific SST of the 20th Century (historical) simulation provided by the Climate Model Inter-Comparison Project phase-5 (CMIP5) and freely accessible in their database system (<https://esgf-node.llnl.gov/search/esgf-llnl/>). The 20th Century scenario in the CMIP5 imposes several changing conditions (Taylor et al., 2009):

- atmospheric composition change (including CO₂) due to both anthropogenic and volcanic influences
- solar forcing
- emissions or concentrations of short-lived species and natural and anthropogenic aerosols or their precursors and
- land use.

According to Gent et al. (2011), the 20th Century scenario in the CCSM4 is a simulation of Earth's climate between 1850 and 2005, with the spatial resolution (latitude

x longitude) of $1.25^\circ \times 0.94^\circ$, and 26 vertical layers for the atmosphere. The ocean model in the CCSM4 has its own coordinate system, which uses a spherical coordinate with 1° grid size for the Southern Hemisphere, and where the North Pole is displaced into Greenland at 80°N , 40°W . The horizontal grid has 320×384 points with 60 vertical levels, and a uniform resolution of 1.11° in the zonal direction. The meridional resolution is 0.27° around the equator, gradually increases to 0.54° at 33°N/S , from where it remains constant at higher latitudes.

To reveal a potential bias, the CCSM4 SST was then compared with the observational data such as the Characteristic of Global Sea Surface Temperature Analyses Data (COBE-SST) (Japan Meteorological Agency, 2006). The COBE-SST is an interpolated $1^\circ \times 1^\circ$ SST product for 1891 to the present, which combines SSTs from various datasets such as from the International Comprehensive Ocean-Atmosphere Data Set (ICOADS), the Japanese Kobe Collections, and reports from ships and buoys (Ishii et al., 2005). Since the two datasets do not share the same horizontal resolution, I first converted the horizontal resolution of the CCSM4 SST dataset to the horizontal resolution of the COBE-SST dataset, using the bilinear interpolation method provided by the Linux-based software Climate Data Operators (CDO) version 1.9.5 (see Appendix A.3). The statistics (mean, variance, etc.) of any physical quantities for all the years in the period of interest is labelled as ‘total’, e.g., total mean precipitation.

A shared period of 1891-2005 from the two datasets was selected for the bias check. The present-day climate scenario (20th century) in the CCSM4 shows an SST bias of 0-1K over the western tropical Pacific (Fig. 2.2), which is a much smaller bias than, for example, the -2.5K multi-model bias in Ackerley et al. (2016).

Figure 2.2. The total mean of the CCSM4 20th Century scenario SST minus COBE-SST for the period of 1891-2005. The dotted areas are not significant at 95 % confidence level based on the Welch's *t*-test (see Appendix A.4).

2.2.1.1. CCSM4 validation for ENSO

The long-term modulation of the ENSO is hypothesized to cause wetter Borneo during the mid-Holocene (Chen et al, 2016). Thus, it is important to validate ENSO performance in the CCSM4 by comparing it with the observed ENSO in COBE-SST dataset. Trenberth (1997) defined ENSO by using a Nino-3 index which is the area averaged of the monthly SST anomaly at 5°N-5°S and 150°-90°W. The El Nino (La Nina) phase is defined as Nino-3>0.5K (Nino-3<-0.5K). The large scale feature of each ENSO phase (El Nino and La Nina) is commonly displayed by averaging (compositing) the tropical Pacific SST for months that are identified as either one of the ENSO phase.

The two phases of ENSO (El Nino and La Nina) are well simulated by the CCSM4 with a slightly westward extension ('too-west') of the Pacific warm/cold tongue in comparison with the COBE-SST data (Fig. 2.3). The minor extension of ENSO-related cold/warm tongue is attributed to the minor SST bias in the WPWP simulated by the CCSM4 (Taschetto et al., 2014).

Figure 2.3. Composites of the SST anomaly based on the two phases of ENSO (El Niño and La Niña) from the CCSM4 20th Century scenario and COBE-SST based on Nino-3 index in 1891-2005. The dotted areas are not significant at 95% confidence level based on Student's *t*-test (see Appendix A.5).

The validation of ENSO impact on the Maritime Continent is done by performing correlation between Nino-3 and the precipitation (Chang et al., 2004). The correlation is done for the December-January-February (DJF) and June-July-August (JJA) season, respectively. The observed monthly precipitation is obtained from the Global Precipitation Climatology Project (GPCP) with 2.5° grid resolution (Adler et al., 2003). ENSO impact during DJF is known to be less homogeneous than the one in JJA (Chang et al., 2004; Jia et al., 2015; Jiang & Li, 2017). The CCSM4 displayed a similar spatial pattern despite of its higher spatial resolution compared to the GPCP dataset (Fig. 2.4). The CCSM4 shows slightly higher correlation scores during DJF season which signifies stronger response from local climate to the ENSO event.

Figure 2.4. The correlations between Nino-3 and precipitation over the Maritime Continent during DJF and JJA period as simulated by the CCSM4 20th Century scenario and observation (COBE-SST and GPCP) in 1979-2005. The dotted areas are not significant at 95% confidence level based on the correlation *t*-test (see Appendix A.6).

2.2.2. CCSM4 validation for the ITCZ and the Maritime Continent seasonal precipitation

The ITCZ is a global-scale tropical rain belt which provides the most prominent rainfall variability in the tropics such as in the Maritime Continent. The ITCZ is also a part of the large scale meridional overturning circulation known as the Hadley Cell that distribute solar energy from the tropics to the higher latitudes (Chiang, 2009a). The ITCZ oscillates annually in the meridional direction that is driven by the annual cycle of insolation. In my thesis, the validation of the ITCZ in the CCSM4 is done by comparing seasonal climatology of precipitation along the Indo-Pacific basin with the one from the GPCP.

Figure 2.5. Seasonal climatology (DJF, MAM, JJA and SON) of precipitation in the GPCP and the CCSM4 20th Century scenario in 1979-2005. The dotted areas are not significant at 95% confidence level based on Student's *t*-test (see Appendix A.5).

The double ITCZ in the CCSM4 is clearly shown by the seasonal climatology of the precipitation, especially over the tropical Pacific (Fig. 2.5). The possible effect of double ITCZ to the climate over the Maritime Continent is perhaps that the simulation underestimates the land precipitation during DJF season (Fig. 2.6). However, the modelled precipitation during JJA and SON seasons are generally valid within the range of 95% confidence level for the Maritime Continent and Borneo (Fig. 2.6). Furthermore, the month-to-month variation of the precipitation climatology in the CCSM4 is coherent with the GPCP. Thus, seasonal rainfall over the Maritime Continent is well-simulated by the CCSM4 despite the existence of double ITCZ.

Figure 2.6. Monthly climatologies of precipitation in the GPCP, the CCSM4 20th Century, and the pre-Industrial Control scenarios for Borneo (110°E to 118°E, 2°S to 5°N) and Maritime Continent (90°E to 140°E, 10°S to 10°N) plotted with uncertainty range of 95% confidence level based on the Student's *t*-distribution (see Appendix A.5).

2.3. CCSM4 validation for the mid-Holocene climate change over the Maritime Continent

The CCSM4 is closely involved in the Paleoclimate Model Inter-Comparison Project – Phase 3 (PMIP3), which is an international collaboration among paleoclimate scientists from different disciplines focusing on bridging the numerical-base theoretical study with the proxy-base observational study. The general purpose of PMIP is to perform a comprehensive model-proxy comparison and provide the underlying mechanism of the observed changes of past climate. PMIP was started in 1992 and it has reached their 3rd phase in 2009 (Crucifix, 2005; Haywood, 2011; Braconnot et al., 2011 & 2012). The 1st phase (PMIP1) focused on the mid-Holocene and LGM scenarios simulated by Atmospheric GCM with prescribed SST of present day. This project is also known as Atmospheric Modelling Inter-comparison Project (AMIP). Ocean and vegetation model components were introduced in the 2nd phase (PMIP2) and a well-structured model output

database was also developed. The 3rd phase (PMIP3) was in a close collaboration with CMIP5 by using similar GCMs and configuration. One important role established in PMIP3 is to provide the standard setup of paleoclimate scenarios, especially for the two main climate epochs, the mid-Holocene and LGM.

Model-proxy comparison is of a great challenge mainly from the dimensional and scale incompatibility. Dimensional mismatch between the large grid size of the model output and the point base proxy data requires further spatial conversion (Brewer et al., 2007). However, the sparse location of the proxy prevents any spatial conversion to be implemented (Braconnot et al., 2012, DiNezio & Tierney, 2013). The second problem is on the scale mismatch between physical quantities produced by climate model and the proxy data. This problem is well illustrated by the incoherency between the month-to-month speleothem $\delta^{18}\text{O}$ records from Mulu cave, northern Borneo and the observed rainfall (Moerman et al., 2013). The interpretation of speleothem $\delta^{18}\text{O}$ as the rainfall proxy is generally robust but it undermines other factors that may reduce their signal-to-noise ratio. The scale problem further limits the model-proxy comparison to only focusing on the change of phase (e.g. drier, unchanged or wetter) instead of their actual magnitude (DiNezio & Tierney, 2013). In this thesis, all limitations in the model-proxy comparison are considered; thus, the change of phase in climate physical quantities such as precipitation and SST is used as the comparison method.

2.3.1. Mid-Holocene climate change over the Maritime Continent in PMIP3

The model-proxy comparison for the mid-Holocene climate change over the Maritime Continent involved the CCSM4 and eight other GCMs from the PMIP3 (Table 2.1). The outputs of all nine GCMs are freely available in their online database system (<https://esgf-node.llnl.gov/search/esgf-llnl/>). The parameters of interest from the atmospheric and oceanic components of the nine models were downloaded for the mid-Holocene and pre-Industrial scenarios. The comparison criteria of the nine models were based on their total mean changes in precipitation and SST over the Maritime Continent between the mid-Holocene and pre-Industrial simulations. The horizontal grid system of the SST of each GCM was converted to the grid system of the COBE-SST dataset to simplify their plotting algorithm (see Appendix A.3). The changes of precipitation and SST in the nine GCMs were then compared to the proxy data discussed in Chapter 1 (Fig. 1.12).

Models	Country	Grid System			Time (yrs)	
		Atmosphere	Ocean		piControl	midHolocene
CCSM4	USA	288x192 x L26	320x384 x L51		1050	300
CNRM-CM5	France	256x128 x L31	362x292 x L42		850	200
CSIRO-Mk3L-1-2	Australia	64x56 x L18	128x112 x L21		1000	500
IPSL-CM5A-LR	France	96x95 x L39	182x149 x L31		1000	500
MPI-ESM-P	Germany	196x98 x L47	256x220 x L40		1156	100
MRI-CGCM3	Japan	364x160 x L48	364x368 x L51		500	100
HadGEM2-ES	UK	192x145 x L38	360x216 x L40		497	102
MIROC-ESM	Japan	128x64 x L80	256x192 x L44		531	100
GISS-E2-R	USA	144x90 x L40	288x180 x L32		1000	500

Table 2.1. the descriptions of coupled-GCMs involved in the PMIP3, which are used in this study (<https://wiki.lsce.ipsl.fr/pmip3/doku.php/pmip3:database:status>).

All PMIP3 models for the mid-Holocene scenario are consistently simulating colder SST at the perimeter of the Maritime Continent which is comparable with the SST proxy data as reported by Abram et al. (2009) (Fig. 2.7 & Fig. 2.8). However, these models show various magnitudes in SST decrease, with a larger variance in the eastern Maritime Continent (Northern Papua). The colder SST of the CCSM4 falls at around the median position among all the PMIP3 models which suggest a non-extreme performance of the CCSM4.

SST variation among models at the eastern Maritime Continent is likely associated with the ‘cold pool’ bias in the Western Pacific as previously discussed in section 2.2.1. Ackerley et al. (2016) argued that the ‘cold pool’ bias in PMIP3 might contribute to the colder SST in the Western Pacific during the mid-Holocene. Thus, the contribution of ‘cold pool’ bias could be more severe in models with a much colder SST. However, I have shown previously that the CCSM4 does not suffer from cold pool bias (Fig. 2.2). Thus, the colder SST during the mid-Holocene simulated by the CCSM4 is a more valid feature than the other PMIP3 models that simulate a much colder SST.

All the PMIP3 models for the mid-Holocene scenario are consistently showing less precipitation over the colder oceanic region of the Maritime Continent following the Bjerknes feedback mechanism (Dayem et al., 2007; Chiang, 2009b). However, the changes in precipitation over the internal area of the Maritime Continent are highly variable within and among models. The possible cause of such spatial variation is likely related to the model specification such as spatial resolution, land-sea distribution and the SST on the internal oceanic region of the Maritime Continent such as the South China Sea.

The changes in precipitation over the internal area of the Maritime Continent can be visually put into three groups:

1. Dipole: Wetter (Drier) region in the western (eastern) Maritime Continent including some part of Borneo (Papua) island. Models : MIROC-ESM, MPI-ESM-P, CCSM4, CRNM-CM5.
2. All-wet: Wetter region in most of the Maritime Continent with higher concentration at the central area. Models : HadGEM2-ES.
3. Irregular: No large scale wetter/drier patches over the Maritime Continent. Models : GISS-E2-R, IPSL-CM5A-LR, MRI-CGCM3, CSIRO-Mk3.

The last group (irregular) is difficult to analyse since they displayed no clear large scale features which could be due to the under-representation of the land-sea distribution of the Maritime Continent as well as coarser spatial resolution. Thus, it will not be included in the follow-up discussion.

Figure 2.7. The total mean of the SST and precipitation for the mid-Holocene minus pre-Industrial scenario of the 5 GCMs of PMIP3. The other 4 GCMs are shown in Fig. 2.8. The dotted areas are not significant at 95% confidence level based on the Welch's *t*-test (see Appendix A.4).

Figure 2.8. Continue from Fig. 2.7 for the 4 GCMs of PMIP3.

2.3.2. Model-proxy comparison for the mid-Holocene precipitation change over the Maritime Continent

HadGEM2-ES is the only model in PMIP3 with an all-wet feature over the Maritime Continent, which is likely the response to a larger area of warmer SST as compared to the rest of the PMIP3 models. However, less precipitation over the northern coast of Australia is not consistent with the wetter rainfall proxy data as reported by Denniston et al. (2013).

There are 4 models in PMIP3 that exhibit a zonal dipole pattern in the changes of precipitation over the Maritime Continent (Fig. 2.7 & Fig. 2.8). All the 4 models except for MIROC-ESM show wetter climate over Borneo, drier/no-change climate over the Flores region and wetter climate over the northern coast of Australia which are consistent with all the rainfall proxy as previously summarized in Figure 1.12. MIROC-ESM shows a drier climate over the northern coast of Australia which perhaps is due to coarser spatial resolution compared to the other 3 models. Furthermore, MIROC-ESM and MPI-ESM models appear to have a much colder SST on the eastern Maritime Continent which is likely suffered from the ‘cold pool’ bias. Therefore, the CCSM4 and the CNRM-CM5 are the better performed GCMs in PMIP3 for the case of the mid-Holocene climate change over the Maritime Continent.

2.4. The main climate features and their impact on the mid-Holocene climate change over the Maritime Continent in the CCSM4

The CCSM4 simulation of the mid-Holocene climate change over the Maritime Continent has revealed positive change in the mean precipitation in several key regions such as Borneo, while the opposite occurred mostly over the eastern oceanic region. This model output can be further diagnosed for mechanisms underlying any changes between the two climate regimes. The first step is to identify the significant climatic features which contribute to the regional climate change.

2.4.1. Climate change captured in a shift in data distribution

To identify the significant climatic features which contribute to the regional climate change, we start by comparing the data distribution of two climate regimes. The shift in data distribution was originally used to evaluate the relationship between the changes of the mean and the extreme values. For example, the changes of the mean precipitation are coherent with the changes of extreme precipitation in most regions around the globe (Groisman et al., 1999; Easterling et al., 2000). This coherency is based on the shift in data distribution.

There are three types of shifts in data distribution, based on the changes in their mean and variance (Meehl et al., 2000). Type 1 is a change in the mean value with a constant variance. Type 2 is a change in the variance with a constant mean value. Type 3 exhibits changes in both the mean and variance. All three types are illustrated in IPCC Assessment Report 3 (2001), as shown in Figure 2.9.

The coherency between the mean and the extreme values is only relevant for types 1 and 3 (Fig. 2.9). Type 1 can be generalized as a complete shift of data distribution, which implies incoherency between the changes of the mean value and its variance. Thus, the changes of the variance in type 1 may not be related to the changes of their mean values. Type 3 might indicate a relationship between the change of the mean and the variance. One feature in type 3 that might indicate this relationship is the shared optima (minima and/or maxima) between the two distributions. One possible way to change the mean value of data distribution with fixed optima is to change its variance. Thus, the changes of the variance in type 3 could be important to the changes of the mean values. Nonetheless, more analyses are required to establish such a relationship.

To identify the type of distribution shift, precipitation data distribution for the mid-Holocene simulation can be compared with distribution for the pre-Industrial simulations.

If the resulting distribution shift is type 3, changes of the variance may be significant; thus, analyses of data variability will be important.

Figure 2.9. Illustration of the three general types of changes in temperature data distribution (IPCC, 2001).

2.4.2. CCSM4 precipitation distribution analyses

The CCSM4 simulations show a significant changes in the mean values of precipitation together with changes in the variance between the mid-Holocene and pre-Industrial scenario over some regions in the Maritime Continent, such as in Borneo and the northern coast of Papua (Fig. 2.10). The two areas show an opposite relationship between the two statistical parameters. The in-phase (anti-phase) relationship of mean and variance appears over Borneo (northern coast of Papua) and some of the western (eastern) Maritime Continent. Further confirmation on the shift of the precipitation distribution can be observed from the precipitation histogram of both areas (Fig. 2.11).

Figure 2.10. The total mean and variance (σ^2) of the CCSM4 precipitation for the mid-Holocene (mH) minus pre-Industrial (piC) scenarios. The dotted areas are not significant at 95% confidence level based on the Welch's t -test (see Appendix A.4). The red-dashed curves indicate the difference in $\sigma^2_{\text{prec.}}$ is 0 (mm/day) 2 .

The histogram of the area averaged precipitation over Borneo (110°E to 118°E, 2°S to 5°N) and the northern coast of Papua (135°E to 150°E, 4°S to 2°N) confirmed the type 3 change in distribution (Fig. 2.11). The increase in the mean precipitation over Borneo occurs together with the increase in their precipitation maxima and fixed minima. The

northern coast of Papua displayed fixed optima (maxima and minima) with the decrease of the mean precipitation and the increase in the frequency of their minima. Thus, the increase of the variance in both areas occurs with an opposite change in their mean values. Nevertheless, both regions are confirmed as a type 3 change of distribution that can indicate the significance of their changes in variability. Further analyses of the precipitation variability can be done by using spectral decomposition method.

Figure 2.11. The histogram of the CCSM4 monthly precipitation over all the years for the spatially averaged Borneo (110°E to 118°E, 2°S to 5°N) and the northern coast of Papua (135°E to 150°E, 4°S to 2°N) for the pre-Industrial (piC) and the mid-Holocene (mH) scenarios. The vertical lines are the mean values.

2.4.3. CCSM4 precipitation variability analyses

The variability analysis of the precipitation is by using Fast-Fourier Transform (FFT) which is a common method to analyse a univariate time series. FFT transforms signal from the time domain to the frequency domain. The main purpose of this spectral analysis is to identify the most dominant timescales shown by the largest power spectrum in the periodogram. Furthermore, the periodogram is smoothen by implementing Bartlett's method with a time window of 25-year (see Appendix A.7).

The periodogram of the monthly precipitation over Borneo and the northern coast of Papua is dominated by annual (yearly) and semi-annual (6-monthly) frequencies in both climate scenarios (Fig. 2.12). The inter-annual band (2-7 yrs) is stronger in the northern coast of Papua than in Borneo, although it is still weaker compared to its annual and semi-annual signals. Annual signal increases during the mid-Holocene for both locations. In Borneo there is a shift from semi-annual during the pre-Industrial to annual timescale in the mid-Holocene era. Further confirmation is by comparing the variance of the annual-and-semi-annual with the inter-annual signals for each grid in the Maritime Continent domain.

Figure 2.12. The power spectral of the CCSM4 monthly precipitation over the spatially averaged Borneo (110°E to 118°E, 2°S to 5°N) and the northern coast of Papua (135°E to 150°E, 4°S to 2°N) for the mid-Holocene and pre-Industrial scenarios.

The comparison between the two bands of frequencies should be conducted between their variances instead of their spectral amplitudes. Thus, the variances of a broad band spectrum in the annual-and-semi-annual and the inter-annual timescales are calculated and compared spatially for the Maritime Continent. The variance of a singular or a broad-band frequency is calculated after extracting the signal using an inverse-FFT out of its power spectral (see Appendix A.7).

Figure 2.13. The total variance of the CCSM4 precipitation (σ^2_{prec}) for the pre-Industrial (piC) (A) and the mid-Holocene (mH) (B) scenarios in the annual-and-semi-annual (0.5 & 1 yr) minus the inter-annual (2-7 yrs) timescales. The total variance of the CCSM4 precipitation for the mid-Holocene minus the pre-Industrial scenarios in the annual-and-semi-annual timescales (C), and also in the inter-annual timescales (D). The red-dashed curves indicate the difference in σ^2_{prec} is 0 $(\text{mm/day})^2$.

The annual-and-semi-annual precipitation signals show larger variances than the inter-annual signals over Borneo and the northern coast of Papua for both climate scenarios (Fig. 2.13. A, B). Furthermore, the variance changes in the annual-and-semi-annual signals over Borneo and the northern coast of Papua are much larger than the variance changes in the inter-annual signals (Fig. 2.13. C, D). Thus, the annual-and-semi-annual variance changes dominate the changes in precipitation variability between the mid-Holocene and the pre-Industrial era over the Maritime Continent.

The small changes of the variance in the inter-annual timescales indicate rather small changes of ENSO impact over Borneo during the mid-Holocene era. However, further

confirmation on the role of ENSO impact to the mid-Holocene climate over the Maritime Continent requires numerical experiments which will be discussed in Chapter 4.

2.4.4. Changes in precipitation monthly climatology over the Maritime continent

Climate signal in the annual and the semi-annual timescales are best represented by their monthly climatology. The changes in the precipitation monthly climatology show higher precipitation patches over the major islands of the Maritime Continent such as Sumatra, Borneo and Papua during the period of July-August-September-October (JASO) (Fig. 2.14). Borneo experienced a wetter climate during the mid-Holocene which starts from May and ends in October with the peak during September. The changes in the monthly climatology of the precipitation also confirms the increased annual signal during the mid-Holocene as shown by the spectral analyses as well as the shift of the precipitation maxima shown by the histogram (Fig. 2.11 & Fig. 2.12). Therefore, the changes of the monthly climatology between the two scenarios are the major contribution to change the all-year mean precipitation over the Maritime Continent and especially over Borneo.

Figure 2.14. Monthly climatology of the CCSM4 precipitation for the mid-Holocene minus the pre-Industrial scenarios. The dotted areas are not significant at 95 % confidence level based on the Welch's t -test (see Appendix A.4).

2.4.5. Mid-Holocene precipitation monthly climatology over Borneo

The monthly climatology of the precipitation over Borneo during the mid-Holocene maximizes during the Summer-to-Autumn season which suggests a contribution from the southward migration of the ITCZ. The well-known convergence zone is mainly forced by the annual insolation among other factors (Chiang et al., 2009a). Regional SST such as in the South China Sea is also identified as a crucial factor that affects the annual march of the ITCZ (Lestari & Iwasaki, 2006; 2012). Present-day climate analyses also show that the oceanic region around Borneo is important to the variability of precipitation over the island through the mechanisms of land-sea breezes (Qian et al., 2013). So in general, insolation and SST are the two factors that affect precipitation over Borneo. Thus, further investigation is required to quantify the contribution of the two physical quantities to the mid-Holocene precipitation over the Maritime Continent and especially over Borneo.

The equatorial insolation, SST in the South China Sea and Borneo precipitation for the mid-Holocene and the pre-Industrial scenarios are compared in order to observe their temporal coherencies. Apparently all three physical quantities maximize in September during the mid-Holocene (Fig. 2.15). Thus, both factors appear to be equally important to drive Borneo precipitation during the mid-Holocene. Numerical experiment by using GCM can dynamically quantify the contribution of each factor to the Maritime Continent precipitation. Experiments on insolation and SST can be done by using Atmospheric-GCM (AGCM) since both parameters are prescribed independently. The Community Atmospheric Model version 4 (CAM4) is the AGCM used in the CCSM4. Thus, it is technically reasonable to use the CAM4 to perform numerical experiment on the atmosphere. The detail experiments including the setup and operational of the CAM4 are discussed in the next chapter. The experimental results are further diagnosed to provide

the fundamental mechanisms behind the mid-Holocene climate change over the Maritime Continent.

Figure 2.15. Monthly climatology of the CCSM4 precipitation over the spatially averaged Borneo (110°E to 118°E, 2°S to 5°N) (A), SST in the South China Sea (110°E to 118°E, 10°N to 20°N) (B) and the equatorial insolation (2°S to 5°N) (C) for the mid-Holocene and the pre-Industrial scenarios. The precipitation and the SST are plotted with the uncertainty range of 95% confidence level based on the Student's *t*-distribution (see Appendix A.5).

2.5. References

- Adler, R.F., Huffman, G.J., Chang, A., Ferraro, R., Xie, P.P., Janowiak, J., Rudolf, B., Schneider, U., Curtis, S., Bolvin, D. and Gruber, A., 2003. The version-2 global precipitation climatology project (GPCP) monthly precipitation analysis (1979–present). *Journal of hydrometeorology*, 4(6), pp.1147-1167.
- Abram, N.J., McGregor, H.V., Gagan, M.K., Hantoro, W.S. and Suwargadi, B.W., 2009. Oscillations in the southern extent of the Indo-Pacific Warm Pool during the mid-Holocene. *Quaternary Science Reviews*, 28(25-26), pp.2794-2803.
- Ackerley, D., Reeves, J., Barr, C., Bostock, H., Fitzsimmons, K., Fletcher, M.S., Gouramanis, C., McGregor, H., Mooney, S., Phipps, S.J. and Tibby, J., 2016. Evaluation of PMIP2 and PMIP3 simulations of mid-Holocene climate in the Indo-Pacific, Australasian and Southern Ocean regions. *Climate of the Past*, 13(11), p.1661.
- An, S.I. and Choi, J., 2014. Mid-Holocene tropical Pacific climate state, annual cycle, and ENSO in PMIP2 and PMIP3. *Climate dynamics*, 43(3-4), pp.957-970.
- Braconnot, P., Harrison, S.P., Otto-Bliesner, B., Abe-Ouchi, A., Jungclaus, J., Peterschmitt, J.Y., 2011. The Paleoclimate Modeling Intercomparison Project contribution to CMIP5. *WCRP Coupled Model Intercomparison Project-Phase 5-CMIP5*, p.15.
- Braconnot, P., Harrison, S.P., Kageyama, M., Bartlein, P.J., Masson-Delmotte, V., Abe-Ouchi, A., Otto-Bliesner, B. and Zhao, Y., 2012. Evaluation of climate models using palaeoclimatic data. *Nature Climate Change*, 2(6), p.417.

- Chang, C.P., Wang, Z., Ju, J. and Li, T., 2004. On the relationship between western maritime continent monsoon rainfall and ENSO during northern winter. *Journal of Climate*, 17(3), pp.665-672.
- Chase, T.N. and Barry, R.G., 2004. Numerical models of the general circulation, climate and weather prediction. In *Atmosphere, Weather and Climate* (pp. 202-216). Routledge.
- Chen, S., Hoffmann, S.S., Lund, D.C., Cobb, K.M., Emile-Geay, J. and Adkins, J.F., 2016. A high-resolution speleothem record of western equatorial Pacific rainfall: Implications for Holocene ENSO evolution. *Earth and Planetary Science Letters*, 442, pp.61-71.
- Chiang, J.C., 2009a. The tropics in paleoclimate. *Annual Review of Earth and Planetary Sciences*, 37, pp.263-297.
- Chiang, J.C.H., Fang, Y. and Chang, P., 2009b. Pacific climate change and ENSO activity in the mid-Holocene. *Journal of Climate*, 22(4), pp.923-939.
- Crucifix, M., Braconnot, P., Harrison, S.P. and Otto-Bliesner, B., 2005. Second phase of paleoclimate modelling intercomparison project. *EOS, Transactions American Geophysical Union*, 86(28), pp.264-264.
- Dayem, K.E., Noone, D.C. and Molnar, P., 2007. Tropical western Pacific warm pool and maritime continent precipitation rates and their contrasting relationships with the Walker Circulation. *Journal of Geophysical Research: Atmospheres*, 112(D6).
- Denniston, R.F., Wyrwoll, K.H., Polyak, V.J., Brown, J.R., Asmerom, Y., Wanamaker Jr, A.D., LaPointe, Z., Ellerbroek, R., Barthelmes, M., Cleary, D. and Cugley, J., 2013. A Stalagmite record of Holocene Indonesian–Australian summer monsoon

- variability from the Australian tropics. *Quaternary Science Reviews*, 78, pp.155-168.
- DiNezio, P.N. and Tierney, J.E., 2013. The effect of sea level on glacial Indo-Pacific climate. *Nature Geoscience*, 6(6), p.485.
- Druyan, L.M., Somerville, R.C. and Quirk, W.J., 1975. Extended-range forecasts with the GISS model of the global atmosphere. *Monthly Weather Review*, 103(9), pp.779-795.
- Easterling, D.R., Evans, J.L., Groisman, P.Y., Karl, T.R., Kunkel, K.E. and Ambenje, P., 2000. Observed variability and trends in extreme climate events: a brief review. *Bulletin of the American Meteorological Society*, 81(3), pp.417-426.
- Gent, P.R., Danabasoglu, G., Donner, L.J., Holland, M.M., Hunke, E.C., Jayne, S.R., Lawrence, D.M., Neale, R.B., Rasch, P.J., Vertenstein, M. and Worley, P.H., 2011. The community climate system model version 4. *Journal of Climate*, 24(19), pp.4973-4991.
- Groisman, P.Y., Karl, T.R., Easterling, D.R., Knight, R.W., Jamason, P.F., Hennessey, K.J., Suppiah, R., Page, C.M., Wibig, J., Fortuniak, K. and Razuvaev, V.N., 1999. Changes in the probability of heavy precipitation: important indicators of climatic change. In *Weather and Climate Extremes* (pp. 243-283). Springer, Dordrecht.
- Haywood, A.M., Ramstein, G. and Abe-Ouchi, A., 2011. Comparing structurally different climate models in a paleoenvironmental context. *Eos, Transactions American Geophysical Union*, 92(21), pp.180-180.
- Ishii, M., Shouji, A., Sugimoto, S. and Matsumoto, T., 2005. Objective analyses of sea-surface temperature and marine meteorological variables for the 20th century using

- ICOADS and the Kobe collection. *International Journal of Climatology*, 25(7), pp.865-879.
- IPCC (Intergovernmental Panel on Climate Change), 2001. Climate change 2001: the scientific basis. *Contribution of Working Group I to the third assessment report of the Intergovernmental Panel on Climate Change*, p.881.
- Japan Meteorological Agency, 2006: Characteristics of Global Sea Surface Temperature Analysis Data (COBE-SST) for Climate Use. *Monthly Report on Climate System Separated Volume*, **12**, 116pp.
- Jia, X., Ge, J. and Wang, S., 2015. Diverse impacts of ENSO on wintertime rainfall over the Maritime Continent. *International Journal of Climatology*, 36(9), pp.3384-3397.
- Jiang, L. and Li, T., 2017. Why rainfall response to El Niño over Maritime Continent is weaker and non-uniform in boreal winter than in boreal summer. *Climate Dynamics*, pp.1-19.
- Lestari, R.K. and Iwasaki, T., 2006. A GCM study on the roles of the seasonal marches of the SST and land-sea thermal contrast in the onset of the Asian summer monsoon. *Journal of the Meteorological Society of Japan. Ser. II*, 84(1), pp.69-83.
- Lestari, R.K. and Iwasaki, T., 2012. An Empirical Study of the Response of the South China Sea Summer Monsoon to the Remote Forcing. *SOLA*, 8, pp.65-68.
- Lin, J.L., 2007. The double-ITCZ problem in IPCC AR4 coupled GCMs: Ocean–atmosphere feedback analysis. *Journal of Climate*, 20(18), pp.4497-4525.
- Li, G. and Xie, S.P., 2014. Tropical biases in CMIP5 multimodel ensemble: The excessive equatorial Pacific cold tongue and double ITCZ problems. *Journal of Climate*, 27(4), pp.1765-1780.

- Mahowald, N.M., Yoshioka, M., Collins, W.D., Conley, A.J., Fillmore, D.W. and Coleman, D.B., 2006. Climate response and radiative forcing from mineral aerosols during the last glacial maximum, pre-industrial, current and doubled-carbon dioxide climates. *Geophysical Research Letters*, 33(20).
- Meehl, G.A., Karl, T., Easterling, D.R., Changnon, S., Pielke Jr, R., Changnon, D., Evans, J., Groisman, P.Y., Knutson, T.R., Kunkel, K.E. and Mearns, L.O., 2000. An introduction to trends in extreme weather and climate events: observations, socioeconomic impacts, terrestrial ecological impacts, and model projections. *Bulletin of the American Meteorological Society*, 81(3), pp.413-416.
- Moerman, J.W., Cobb, K.M., Adkins, J.F., Sodemann, H., Clark, B. and Tuen, A.A., 2013. Diurnal to interannual rainfall $\delta^{18}\text{O}$ variations in northern Borneo driven by regional hydrology. *Earth and Planetary Science Letters*, 369, pp.108-119.
- Qian, J.H., Robertson, A.W. and Moron, V., 2013. Diurnal cycle in different weather regimes and rainfall variability over Borneo associated with ENSO. *Journal of Climate*, 26(5), pp.1772-1790.
- Schär, C., Vidale, P.L., Lüthi, D., Frei, C., Häberli, C., Liniger, M.A. and Appenzeller, C., 2004. The role of increasing temperature variability in European summer heatwaves. *Nature*, 427(6972), p.332.
- Taylor, K.E., 2009. A summary of the CMIP5 experiment design. http://cmip-pcmdi.llnl.gov/cmip5/docs/Taylor_CMIP5_design.pdf.
- Taschetto, A.S., Gupta, A.S., Jourdain, N.C., Santoso, A., Ummenhofer, C.C. and England, M.H., 2014. Cold tongue and warm pool ENSO events in CMIP5: mean state and future projections. *Journal of Climate*, 27(8), pp.2861-2885.

- Trenberth, K.E., 1997. The definition of el nino. *Bulletin of the American Meteorological Society*, 78(12), pp.2771-2778.
- Zheng, Y., Lin, J.L. and Shinoda, T., 2012. The equatorial Pacific cold tongue simulated by IPCC AR4 coupled GCMs: Upper ocean heat budget and feedback analysis. *Journal of Geophysical Research: Oceans*, 117(C5).
- Zheng, W., Wu, B., He, J. and Yu, Y., 2013. The East Asian summer monsoon at mid-Holocene: results from PMIP3 simulations. *Climate of the Past*, 9(1), pp.453-466.

Chapter 3 – Mid-Holocene climate change over the Maritime Continent in the CAM4 – Experiments, Attribution and Mechanisms

Abstract

The Community Atmospheric Model version 4 (CAM4) was successfully installed and tested in order to perform numerical experiment under the mid-Holocene and pre-Industrial scenarios. The atmosphere only experiments by CAM4 for both scenarios are mainly differentiated by the orbital parameters and the global sea surface temperatures (SST) simulated by the Community Climate System Model version 4 (CCSM4). The prescribed SST and the incoming solar radiation (insolation) are considered as two independent physical quantities in order to evaluate their individual impact on the climate over the Maritime Continent.

Numerical experiments suggest that during the mid-Holocene, warmer landmass as a response to the stronger insolation during July-October is the main driver of higher mean precipitation over Borneo and other landmass in the Maritime Continent via a stronger convection. This stronger convection over the landmass suppresses the convection over the South China Sea through a weaker and drier low-level southerly. Furthermore, the suppressed convection over the South China Sea in turn reduces the vertical velocity and weakens the local Hadley cell.

3.1. Community Atmospheric Model – version 4 (CAM4)

Mid-Holocene climate change over the Maritime Continent is hypothesized to be significantly affected by the change of orbital parameters that cause changes in annual insolation (Liu et al., 2000). Regional SST is another factor that influences the climate over the region which is directly influenced by the insolation and remotely driven by the El Nino Southern Oscillation (ENSO) (Abrams et al., 2009; Chen et al., 2016). The investigation of the influence of insolation and SST on the mid-Holocene climate change can be carried out by performing a numerical simulation that allows each of these physical quantities to independently influence the atmosphere. Such a numerical configuration can be achieved by following the framework of the Atmospheric General Circulation Model (AGCM). AGCM is a numerical representation of earth's atmosphere based on the principle of fluid dynamics in a discretized spherical domain with the surface properties as its boundary conditions (Arakawa & Lamb, 1981). The surface properties of the oceanic and sea-ice regions are prescribed in terms of SST time series which influences the thermodynamics of the atmosphere. In AGCM, the solar radiation interacts with the land and the atmosphere but not with the oceanic surface, thus heat sources from the land and sea can be addressed separately.

Community Atmospheric Model version 4 (CAM4), the AGCM chosen for this study, is the atmospheric component of the CCSM4. In Chapter 2, I showed that the CCSM4 is one of the better performing coupled-GCMs in PMIP3 in simulating the mid-Holocene climate change over the Maritime Continent. The CAM4 is the sixth generation of AGCM developed by the National Center for Atmospheric Research (NCAR) in collaboration with several international research institutions. The first generation was released in 1981 under the name of Community Climate Model 0-A (CCM0A) (Neale et al., 2010). The CAM4 used in this study is one of the optional atmospheric components in

the coupled climate model package named Community Earth System Model ver.1.2.2 (CESM1.2.2) released in June, 2014. CAM5, the latest version of CAM, is also available in CESM1.2.2 package, but the CAM4 is still chosen for this study in order to maintain consistency with the CCSM4 of PMIP3 as well as to reduce the computational resources demand.

AGCM discretizes the earth's atmosphere into a grid system with a fixed horizontal resolution comparable to the mesoscale weather events (~100 km), such as large-scale convection and squall-lines. Physical processes such as moist-thermal dynamics of water molecules that occur at sub-grid scale however need to be represented in a much larger scale domain, known as parameterization. Neale et al. (2010) described four groups of parameterizations in the CAM4:

- Moist Precipitation Processes (deep convection, shallow convection and large-scale condensation)
- Clouds and Radiation
- Surface Model (surface fluxes from land, ocean and sea-ice)
- Turbulent Mixing (planetary boundary layer, vertical diffusion and gravity wave drag)

The four parameterization groups are calculated in the above sequence starting from precipitation and ending with turbulent mixing. Precipitation from the deep convection processes is known to dominate the tropical region such as the Maritime Continent (Yamanaka 2016; Trismidianto et al., 2017). Deep convection scheme in the CAM4 uses a modified Zhang and McFarlane (1995) method, which is based on the vertical mass-flux of air parcels driven by a conditionally unstable air column. The upward mass-flux produces rainfall in order to balance the rate of cloud water formation due to net condensation against the detrainment of cloudy air. The magnitude of the cloud-base

upward mass-flux increases exponentially with a time scale of two hours and is scaled by the Convective Available Potential Energy (CAPE). CAPE is a measure of buoyancy energy between the level (z) of free convection (IL) and neutral buoyancy (EL). Such a metric is signified by the difference between the virtual temperature of the air parcel (θ_{vp}) and the environment (θ_v), where g is the acceleration due to gravity (Eq. 3.1) (Zhang and McFarlane, 1995). CAPE as a grid scale quantity acts as a closure to define the sub-grid scale quantity in the convective scheme.

$$CAPE = g \int_{IL}^{EL} \left(\frac{\theta_{vp} - \theta_v}{\theta_v} \right) dz. \quad (3.1)$$

The produced precipitation then interacts with the large scale processes by evaporating along its way to the surface. The CAM4 uses Sundqvist (1988) method to reduce the temperature and the precipitation amount due to the evaporated precipitation. Moist convection also affects the other physical parameterizations such as cloud fractions.

The heating rates in terms of longwave and shortwave fluxes are evaluated per-hour. The CAM4 used Berger (1978) method to calculate the insolation during a time window from 10^6 yrs before to 10^6 yrs after 1950 (Neale et al., 2010). The insolation signal consists of stationary cycles in diurnal (24 hrs) and annual (365 days) scales. The insolation cycles are declared by a fixed calendar year or by prescribing certain orbital parameters. The shortwave radiation interacts with the land to produce surface fluxes (momentum, sensible and latent heat) based on Monin-Obukhov similarity theory (Monin and Obukhov, 1954). Surface fluxes from the ocean are based on the near surface atmospheric stability calculated from the SST and near surface atmospheric parameters (potential temperature, specific humidity and 10-m wind speed).

The coupling of the physical parameterizations and the dynamical processes is done either concurrently or sequentially in each time-step. The concurrent coupling

approximation between the dynamical and parametrization processes is called ‘Process Split’ and is best applied for the spectral dynamical core. The sequential approximation is called ‘Time Split’ and is applied for the finite-volume (FV) dynamical core (Neale et al., 2010). FV dynamical core is chosen for this study due to its advantage over the spectral method in term of mass conservation, especially over regions with complex topography (Lin and Rodd, 1997).

The vertical coordinate is a terrain-following coordinate system that is gradually in-line with the pressure coordinate at the higher levels with a transition region in between. The finite-volume discretization scheme of the terrain-following coordinate system is based on the hydrostatic approximation which assumes that a vertical distribution of pressure (p) along the height (z) is in balance with the density (ρ) (Eq. 3.2) (Neale et al., 2010).

$$\frac{1}{\rho} \frac{\partial p}{\partial z} + g = 0. \quad (3.2)$$

The above approximation is the only vertical interaction between model layers; thus the dynamical transport is solved horizontally (2D) along the spherical domain.

The CAM4 is ported to the cluster computational centre in the Nanyang Technological University (NTU), Singapore. The technical aspect of the CAM4 installation, basic configuration and tests are described in the Appendix A.8.

3.2. The CAM4 simulation for the mid-Holocene and pre-Industrial scenarios

Mid-Holocene (~6 ka) is one of the climate scenarios included in the PMIP3 project. The main differences between the mid-Holocene and pre-Industrial settings is in the value of orbital parameters. Besides orbital forcing, there is also a small decrease in the mid-Holocene global methane (CH₄) concentration by ~100 ppbv (Table 3.1), which is one of the greenhouse gases (Blunier et al., 1995; Sowers, 2010; Singarayer et al., 2011). The orbital parameters and greenhouse gas concentrations are the main parameters to be configured in the model to simulate the mid-Holocene climate. Such configurations are standardized by PMIP3; thus they are adopted in the CAM4 simulations of this study (Table 3.1).

The prescription of the global greenhouse gas concentrations in the CAM4 is made by assuming a time invariant mixing ratio of the gases at the surface. This assumption is required to accommodate the global scale greenhouse gas concentrations such as CH₄ as mentioned above, and CO₂, extracted from Greenland and Antarctic Ice Sheets (Monnin et al., 2001; Axford et al., 2013). PMIP3 assigned a fixed global CO₂ concentration for the two scenarios with a value of ~280 ppmv as suggested by global CO₂ proxy records (e.g., Monnin et al., 2001). Therefore, greenhouse gas forcing for the mid-Holocene only differs in terms of a lower CH₄ concentration. Burns et al. (2011) suggested a possible mechanism for the mid-Holocene lower CH₄ concentration as a hydro-ecological response to the changes in orbital forcing via changes in precipitation. The water-dependent production of methane in tropical wetlands is suggested to be driven by monsoon precipitation, which could be weaker globally during the mid-Holocene relative to today. Thus, the lower concentration of global methane during the mid-Holocene is likely a passive response to the changes in orbital forcing.

The orbital parameters, introduced by Milankovitch, consist of three parameters: Eccentricity, Obliquity and Precession (see Appendix A.1). These orbital parameters occur on periodicities of ~100k, ~40k, and ~20k yrs respectively. Berger (1978) formulated each of the three orbital parameters and assigned each of them by a single value for each climate scenario. Joussaume et al. (1997) recalculated these values for the mid-Holocene and present day (1950 A.D.) orbital parameters based on Kepler's Law of the elliptical shape of the planet's orbit. The values of orbital parameters, as shown in Table 3.1, are accepted and standardized in PMIP3 for the mid-Holocene and pre-Industrial scenarios.

Conf.	Mid-Holocene	Pre-Industrial
Orbital	ecc = 0.018682	ecc = 0.016724
	obl = 24.105°	obl = 23.446°
	per-180° = 0.87°	per-180° = 102.04°
GHG	CH ₄ = 650 ppbv	CH ₄ = 760 ppbv
	CO ₂ = 280 ppmv	CO ₂ = 280 ppmv
	N ₂ O = 270 ppbv	N ₂ O = 270 ppbv
	CFC-11 & 12 = 0 ppbv	CFC-11 & 12 = 0 ppbv
SST	Mid-Holocene SST of the CCSM4 in the PMIP3 (300 yrs)	Pre-Industrial SST of the CCSM4 in the PMIP3 (300 yrs)

Table 3.1. Mid-Holocene and pre-Industrial scenarios as described by the PMIP3

(<https://pmip3.lsce.ipsl.fr/>).

Each time interval bears a certain combination of the three orbital parameters. Such configuration of orbital parameters is applied to the GCM simulation on the climatic timescale (~30 yrs) and is further extended to ~100 to 1000 yrs to guarantee climate equilibria as it was done in PMIP3. The orbital forcing provides annual and diurnal insolation cycles throughout the simulation time. The configuration of the CAM4 in this study sets orbital parameters as suggested by PMIP3 for the mid-Holocene and pre-Industrial scenarios. The simulation is conducted with a time range of 300 yrs and a spin-up time of 20 yrs; thus, 280 yrs is used for the follow-up analyses. The statistics (mean, variance, etc.) of any physical quantities for all the years in the period of interest is labelled as 'total', e.g., total mean precipitation.

The technical configuration of the CAM4 for this study uses the F1850 component set (compset) provided by the CESM1.2.2 package. Compset F1850 is a build-in configuration for pre-Industrial GCM simulation by using the CAM4 as its atmosphere and CLM4 as its land component with a 30-minute time-step. It is initialized using a pre-Industrial dataset of the year 1850. Mid-Holocene and pre-Industrial scenarios are implemented in F1850 by modifying the orbital and greenhouse gas parameters and also by prescribing the SST data with the one from the CCSM4 of PMIP3 (Table 3.1).

The CAM4 simulation uses f19 grid system, which is a 1.9 x 2.5 degree of horizontal resolution and 26 vertical layers (144 x 96 x 26L) discretized by finite-volume dynamical core (Vertenstein et al., 2013). The horizontal resolution of f19 is half the size of the spherical grid system of the CCSM24 SST dataset which has 288 x 192 grid numbers. Thus, the grid system of the CCSM4 SST dataset is first converted to be compatible with the f19 grid system for our CAM4 simulation (see Appendix A.3). Such a reduction in spatial resolution is also reducing the computational time without compromising the main features observed in the CCSM4 mid-Holocene simulations, for instance, climate change

over the tropics and specifically over the Maritime Continent. The main climate change feature such as wetter climate over Borneo, North Coast of Australia and Saharan regions will need to be validated in our CAM4 simulations.

3.3. The CAM4 and the CCSM4 performance comparison

The performance of the CAM4 in simulating the pre-Industrial and the mid-Holocene climate is validated with the CCSM4. The validity of the CAM4 towards the CCSM4 confirms model configuration for the two scenarios standardized in PMIP3. The comparison between the CAM4 and the CCSM4 is based on three aspects:

1. Mid-Holocene tropical climate change such as higher rainfall over Borneo, North Coast of Australia and Saharan regions.
2. A stronger annual-and-semi-annual than inter-annual variability of the precipitation over the Maritime Continent during the Mid-Holocene.
3. Higher precipitation monthly climatology during Mid-Holocene over the Maritime Continent with the maxima in the July-August-September-October (JASO) period.

These three aspects are based on the attribution of the mid-Holocene climate change over the Maritime Continent in the CCSM4 as previously discussed in Chapter 2.

3.3.1. Mid-Holocene tropical climate change in the CCSM4 and the CAM4

A good agreement exists between the CCSM4 and the CAM4 on the mid-Holocene precipitation change over the tropical Africa and Asia (Fig. 3.1). The two models are consistent in simulating several main features such as the wetter Borneo, northern coast of Australia and Saharan regions. Wetter climate over those particular regions is also consistent with the mid-Holocene proxy data (deMenocal et al., 2000; Partin et al., 2007; Tierney & deMenocal, 2013).

Figure 3.1. The total mean of the CCSM4 and the CAM4 precipitation for the mid-Holocene (mH) minus the pre-Industrial (piC) scenario. The dotted areas are not significant at 95% confidence level based on the Welch's t -test (see Appendix A.4).

Mid-Holocene precipitation change in the CAM4 managed to capture most features shown by the CCSM4 despite the lower spatial resolution. The consistency between the CAM4 and the CCSM4 signifies the large scale features in the change of the mean precipitation, especially over the Maritime Continent.

The Island of Flores is under-represented in our CAM4 simulations due to the coarse spatial resolution. Mid-Holocene drier climate over the Flores region is also suspected to be a small scale feature. Therefore, the climate change over the Flores region will not be included in the follow up analyses.

3.3.2. Changes in the mid-Holocene climate variability over the Maritime Continent in the CCSM4 and the CAM4

Mid-Holocene precipitation change over the Maritime Continent in the CCSM4 is dominated by the change in the annual-and-semi-annual variability as previously shown in Chapter 2 (Fig. 2.13). The comparison between the CCSM4 and the CAM4 simulations shows a consistency in the dominance of precipitation variability on the annual-and-semi-annual over inter-annual timescales (Fig.3.2).

Mid-Holocene precipitation in the annual-and-semi-annual timescales is consistent in both simulations with the large changes occurring over several areas in the Maritime Continent such as the one over Borneo (Fig. 3.2.A & C). Small changes in the mid-Holocene inter-annual precipitation are also consistent in both models (Fig. 3.2.B & D). The spatial consistency of the precipitation variability in both models suggests consistent mechanisms in driving the internal climate variability over the Maritime Continent.

Figure 3.2. The variance of the precipitation ($\sigma^2_{\text{prec.}}$) for the mid-Holocene (mH) minus pre-Industrial (piC) scenarios of the CCSM4 (A & B) and the CAM4 (C & D) in annual- and semi-annual (0.5 & 1 yr) and inter-annual (2-7 yrs) timescales. The red-dashed curves indicate the difference in $\sigma^2_{\text{prec.}}$ is 0 (mm/day)².

3.3.3. Changes in the mid-Holocene precipitation monthly climatology over the Maritime Continent in the CCSM4 and the CAM4

The annual-and-semi-annual variability is commonly represented by their monthly climatological pattern as previously shown in Chapter 2. Strong annual variability in the monthly climatology of the precipitation is the major contributor in increasing the total mean precipitation over the Borneo Island during the mid-Holocene. The monthly climatology of precipitation over the Borneo Island shows higher values during July-August-September-October (JASO) period (Fig. 2.14 & Fig. 3.3).

Figure 3.3. The monthly climatology of the CCSM4 (A) and the CAM4 (B) precipitation over Borneo for the mid-Holocene (mH) and pre-Industrial (piC) scenarios. The precipitation climatology is plotted with the uncertainty range of 95% confidence level based on the Student's t -distribution (see Appendix A.5).

The CAM4 managed to simulate the changes in precipitation climatology over the Borneo Island during JASO period, similar to the results from the CCSM4 (Fig. 3.3). Thus, the CAM4 can be used confidently for attributing the two main factors (insolation and SST) in changing the precipitation over the Maritime Continent during the mid-Holocene.

3.4. Numerical experiment for insolation and SST using the CAM4

The JASO period has been identified as the maxima of the mid-Holocene precipitation change over the Borneo Island. The maxima of the equatorial insolation and the SST in the South China Sea are also occurring during the same period as previously shown in Chapter 2 (Fig. 2.15). The roles of insolation and SST during the mid-Holocene in precipitation change over the Borneo Island can be individually quantified by performing numerical experiments in the CAM4.

3.4.1. Experimental design

Numerical experiment is a modified GCM simulation based on a control scenario in order to quantify the influence of a certain component in the model. Mid-Holocene climate simulation by the Coupled-GCM in PMIP3 is essentially an experiment of the pre-Industrial scenario with the modification of the main forcing, insolation and greenhouse gases (<https://pmip3.lsce.ipsl.fr/>). Similar modification is also implemented in our CAM4 simulation with the prescribed SST from the CCSM4 (Table 3.1). Thus, the numerical experiment to quantify the influence of one physical quantity can be done by setting the rest of the quantities to be the same as that in the control simulation. Furthermore, the quantification of the atmospheric changes to the particular quantity is by taking the difference between the control (pre-industrial) and the experimental simulation.

The numerical experiments for the two physical quantities are conducted by simulating two modified mid-Holocene scenarios. The first (second) experimental simulation is labelled as ‘stronger insolation’ or SSOLIN (‘warmer SST’ or WSST) which is a mid-Holocene scenario with the pre-Industrial actual SST (insolation) (Table 3.2). The subtraction of the SSOLIN (WSST) experiment with the pre-Industrial simulation is to quantify changes in the atmosphere towards the mid-Holocene insolation (SST) since the actual SST (insolation) is the same as that in the pre-Industrial scenario.

The attribution of each physical quantity in driving atmospheric changes at a particular location such as Borneo Island is done by comparing each experiment with their base scenarios (mid-Holocene and pre-Industrial). The numerical experiments are conducted with a time range of 300 yrs and a spin-up time of 20 yrs; thus, 280 yrs is used for the follow-up analyses.

Conf.	Mid-Holocene SSOLIN	Mid-Holocene WSST
Orbital	ecc = 0.018682	ecc = 0.016724
	obl = 24.105°	obl = 23.446°
	per-180° = 0.87°	per-180° = 102.04°
GHG	CH ₄ = 650 ppbv	CH ₄ = 650 ppbv
	CO ₂ = 280 ppmv	CO ₂ = 280 ppmv
	N ₂ O = 270 ppbv	N ₂ O = 270 ppbv
	CFC-11 & 12 = 0 ppbv	CFC-11 & 12 = 0 ppbv
SST	Pre-Industrial SST of the CCSM4 in the PMIP3 (300 yrs)	Mid-Holocene SST of the CCSM4 in the PMIP3 (300 yrs)

Table 3.2. Mid-Holocene ‘stronger insolation’ (SSOLIN) and mid-Holocene ‘warmer SST’ (WSST) experimental scenarios. See Table 3.1 for the original mid-Holocene and pre-Industrial scenarios as described by the PMIP3 (<https://pmip3.lsce.ipsl.fr/>).

3.4.2. The impact of insolation and SST on the mid-Holocene precipitation change over the Maritime Continent

The stronger tropical insolation in the mid-Holocene triggered a higher precipitation over the major islands in the Maritime Continent as shown by SSOLIN experiment (Fig. 3.4.A). Highest precipitation occurs around the equator within $\pm \sim 5^\circ$ latitude with a longitudinal orientation from the western Indian Ocean to the eastern Maritime Continent (Papua). The lowest precipitation occurs at the higher latitude adjacent to the highest precipitation region with a strong dipole feature between the Borneo Island and the South China Sea. The dipole feature is stronger during JASO and extends longitudinally throughout the period (Fig. 3.4.D).

Mid-Holocene warmer SST over the Maritime Continent triggered a higher precipitation over the oceanic region as shown by WSST experiment (Fig. 3.4.B). Higher precipitation occurred over the open oceanic region off both the northern and southern perimeter of the Maritime Continent. The equatorial region experience the lowest precipitation and mainly overlaps with the colder SST region at the east and west perimeters of the Maritime Continent. The central region of the Maritime Continent such as Borneo also exhibits a lower precipitation with a smaller magnitude as compared to the regions over a colder ocean. The maxima of the higher oceanic precipitation occur during JASO following the warmer mid-Holocene SST (Fig. 3.4.E).

Figure 3.4. The mean of the CAM4 precipitation change between experiments SSOLIN, WSST and SSOLIN+WSST with the pre-Industrial (piC) scenario for total (A-C) and JASO (D-F) period. The dotted areas are not significant at 95% confidence level based on the Welch's *t*-test (see Appendix A.4).

The two experiments simulate an opposite atmospheric response over the Maritime Continent. The equatorial higher precipitation only appeared in the SSOLIN simulation which suggests that insolation is the main factor to increase Borneo precipitation during the mid-Holocene. Likewise, lower precipitation over the northern coast of Papua is related to the colder SST as simulated by the WSST experiment. The land radiative response towards insolation in the CAM4 explains the differences in precipitation in both experiments. Increased precipitation is related to the stronger surface radiation over the landmass (ocean) that occurred only in the SSOLIN (WSST) experiment.

The opposite atmospheric response to the land and oceanic heating from the two experiments suggests that insolation and SST are competing factors in increasing the precipitation over the land/oceanic region in the Maritime Continent during the mid-Holocene. Further evidence comes by adding the two experiments which shows similar responses to the unmodified mid-Holocene scenario (Fig. 3.4.C & F). The differences in the mean precipitation between the SSOLIN+WSST and the pre-Industrial scenario are spatially similar to the differences between the mid-Holocene and the pre-Industrial scenarios (Fig. 3.4.C & Fig. 3.1 lower panel). The similarity between the aggregation of the two experiments and the mid-Holocene scenario suggests a substantial linear combination of the two physical quantities as the fundamental mechanisms to cause the mid-Holocene climate change over the Maritime Continent.

3.4.3. The impact of insolation and SST on the mid-Holocene convection change over Borneo and the South China Sea

The domination of convective precipitation is well-known over the tropical region especially over the Maritime Continent (Yamanaka 2016; Trismidianto et al., 2017). Likewise, convective precipitation appears dominantly in the tropics in the CAM4 experiments, which suggest that deep convection is the main mechanism of rainfall (Fig. 3.5). Thus, insolation and SST trigger a stronger deep convection as the atmospheric response to a warmer surface temperature on the land and over the ocean, respectively (Zhang 1993; Neale & Slingo, 2003).

Figure 3.5. The mean of the CAM4 convective (A) and large-scale (B) precipitation for the mid-Holocene minus pre-Industrial scenario during JASO. The dotted areas are not significant at 95% confidence level based on the Welch's *t*-test (see Appendix A.4).

Tropical convection is best represented by CAPE which is the positive buoyancy of an air parcel and an indicator of atmospheric instability (Blanchard 1998). CAPE was previously shown as the closure parameter in the Zhang & McFarlane deep convective scheme used in the CAM4 (Eq. 3.1). Stronger CAPE in JASO is consistently overlapped with the region of higher precipitation in the two experimental simulations (Fig. 3.4.D, E & Fig. 3.7.D, G). Stronger CAPE signifies a less stable air column that allows air parcel to rise following stronger buoyancy (Blanchard 1998; Riemann et al., 2009).

The vertical atmospheric structure of tropical convection requires the atmospheric parameters such as vertical velocity at 500 mb and horizontal moisture convergence to be spatially consistent with CAPE (Ye et al., 1998; Banacos & Schultz, 2005). A region with stronger (weaker) CAPE should correspond to stronger (weaker) vertical velocity at 500 mb and horizontal moisture convergence. Horizontal moisture convergence is the dominant factor to increase local humidity by collecting moisture from the surroundings due to the spatial gradient of the wind field especially at the lower troposphere (Banacos & Schultz, 2005). CAPE is sensitive to the lower tropospheric humidity mainly by defining the level of free convection (Eq. 3.1). The changes in all the three convective

parameters appear to be consistent in our CAM4 simulations which further signify stronger land convection as the main mechanisms to increase island precipitation over the Maritime Continent during the mid-Holocene (Fig. 3.6).

Figure 3.6. The changes of the mean of the CAM4 CAPE (1st row), moisture convergence (2nd row) and vertical velocity (Ω) at 500 mb (3rd row) between the mid-Holocene (A-C), SSOLIN (D-F) and WSST (G-I) with the pre-Industrial scenario during JASO. The dotted areas are not significant at 95% confidence level based on the Welch's *t*-test (see Appendix A.4), and $\langle . \rangle$ denotes vertical integration.

Stronger land (ocean) convection during the mid-Holocene in the SSOLIN (WSST) experiment can be attributed to the higher land-sea thermal contrast as a response to a stronger insolation (SST) and an unchanged SST (insolation) (Fig. 3.7.A, D, G). Furthermore, the SSOLIN experiment shows suppressed convection over the northern oceanic region in the Maritime Continent such as the South China Sea. The meridional dipole of the convection change between the major islands and northern oceanic region

shown by the SSOLIN experiment could explain the stronger convection over the Borneo Island in the mid-Holocene scenario.

The opposite influence of insolation and SST in changing the mid-Holocene precipitation over the Maritime Continent resembles the experimental results by Lestari and Iwasaki (2006) for the present day summer. Their experiments suggested the role of insolation (SST) in suppressing (promoting) the northward migration of the Inter-tropical Convergence Zone (ITCZ) during the onset of the Asian summer monsoon. Insolation (SST) is suggested to reduce (promote) the low-level southerly, the evaporation in the South China Sea and the cross equatorial moisture transport. Similar features and mechanisms are likely at play for the mid-Holocene climate change.

Suppressed convection over the South China Sea in the SSOLIN experiment is associated with a weaker and drier low-level wind over the region. Low-level southerly over the landmass is affected by higher CAPE which promotes stronger moisture convergence (Fig. 3.6.D & Fig. 3.7.E, F). Moisture convergence increases humidity over the landmass which has a positive feed-back to further increase the CAPE (Eq. 3.1). Moisture convergence over the landmass causes drier and weaker low-level southerly over the South China Sea (Fig. 3.6.E & Fig. 3.7.E, F). The latter in turn suppresses evaporation and CAPE; thus the South China Sea experiences less precipitation compared to the landmass (Fig. 3.6.A).

Figure 3.7. The changes of the mean of the CAM4 surface temperature (1st row), meridional wind at 925 mb (2nd row) and vertically integrated specific humidity (3rd row) between the mid-Holocene (A-C), SSOLIN (D-F) and WSST (G-I) with the pre-Industrial scenario. The 925 mb wind vector is from the pre-Industrial scenario during JASO. The dotted areas are not significant at 95% confidence level based on the Welch's *t*-test (see Appendix A.4), and <.> denotes vertical integration.

Higher CAPE over the South China Sea in the WSST experiment is due to warmer SST and an unchanged insolation. Similar insolation between the WSST experiment and the pre-Industrial scenario produces similar moisture convergence over the landmass, thus allowing stronger and moister low-level southerly to converge over the South China Sea (Fig. 3.6.H & Fig. 3.7.H, I). Stronger low-level southerly and moisture convergence increases evaporation and CAPE over the South China Sea; thus, producing higher precipitation compared to the landmass (Fig. 3.6.G, H).

Mid-Holocene experiment during JASO is forced by stronger insolation and warmer SST over the South China Sea. Stronger insolation increases CAPE over the landmass

due to warmer and moister low-level air. CAPE is slightly weaker over the South China Sea as a net result of a slower and drier low-level southerly and the warmer SST. Thus, the convection over the warmer and moister landmass should always be more active than the South China Sea in order to maintain higher CAPE and precipitation over the land such as Borneo Island. Therefore, changes in SST in the South China Sea are not relevant to the mid-Holocene higher land precipitation over the Maritime Continent.

3.4.4. The impact of insolation and SST on the mid-Holocene local Hadley circulation over the Maritime Continent

The two numerical experiments show an opposite atmospheric response over the South China Sea. Stronger insolation (SST) suppressed (promoted) convection over the South China Sea associated with the stronger (weaker) convection over the landmass. The meridional convective dipole between Borneo and the South China Sea also appears in the vertical velocity at 500 mb (Fig. 3.6.F). Thus, changes in the mid-tropospheric vertical motion are associated with the changes in the large scale meridional overturning circulation known as the Hadley cell (Lau & Yang, 2002).

Hadley cell can be quantified as the mass transport within the mean meridional circulation. Cook (2003) represented atmospheric mass transport with the Stokes streamfunction (ψ) as:

$$\psi(\phi, p) = \frac{2\pi a \cos\phi}{g} \int_p^{p_{top}} [v] dp, \quad (3.3)$$

where ϕ is latitude, a is the earth's radius, $[v]$ is the time-averaged zonal-mean meridional wind, p_o is the surface air pressure, and p_{top} is the pressure at the top of the atmosphere. The positive (negative) streamfunctions signify clockwise (anti-clockwise) direction of the meridional circulation along the south-to-north vertical cross section.

Besides the Stokes streamfunction Hadley cell can also be displayed by the meridional-vertical (v - w) wind vector of the longitudinally averaged wind component in each vertical level along the tropical latitudes (Kajikawa et al., 2003).

Figure 3.8 is the local Hadley cell over the Maritime Continent (90°E-150°E) for JASO during the pre-Industrial, the mid-Holocene and the other two experiments which all have common features as follow:

1. Meridional circulation in the anti-clockwise direction based on the negative streamfunction and the v - w wind vector.
2. Upward motion over the mid-to-northern Maritime Continent especially over the oceanic region such as the South China Sea.
3. Latitudinal coverage of the Hadley cell is within $\sim\pm 20^\circ$ latitudes.

The general agreements among all the three features suggest a persistent Hadley cell in all four simulations. However, there are slight changes following the changes in the vertical velocity between the landmass and the oceanic region.

The two numerical experiments show larger changes in the Hadley cell than the mid-Holocene scenario. The Hadley cell in the two experiments undergoes two main changes:

1. Southward (northward) contraction (extension) over the mid-to-northern Maritime Continent in the SSOLIN (WSST) experiment (Fig. 3.8.A, B, D).
2. Weakening (Strengthening) over the mid-to-northern Maritime Continent in the SSOLIN (WSST) experiment with the maxima at ~ 300 mb (Fig. 3.9.B, C). These changes are associated with a clock-wise (anti-clockwise) modification of the Hadley cell due to weaker (stronger) vertical velocity over the South China Sea.

The opposite changes of the Hadley cell due to the insolation and the SST are expected as a result of the opposite atmospheric response on the surface. The net result of these two changes is the changes in the Hadley cell in the mid-Holocene scenario.

Figure 3.8. The mean of the CAM4 longitudinally (90°E-150°E) averaged streamfunction and v - w vector for the pre-Industrial (piC, A), the mid-Holocene (mH, C), SSOLIN (B) and WSST (D) simulations during JASO. v - w vector consists of the meridional wind in 1 m/s and the vertical velocity in 10^2 Pa/s following Kajikawa et al., (2003). The dotted areas are not significant at 95% confidence level based on the Student's t -test (see Appendix A.5).

Mid-Holocene atmospheric responses at the surface are more relevant with the result of experiment SSOLIN. Likewise, a slightly weaker Hadley cell during JASO is the response of the tropical troposphere to the mid-Holocene scenario (Fig. 3.9.A). The weakening of the Hadley cell is therefore caused by a stronger convection over the landmass such as Borneo Island accompanied by a weaker convection over the South China Sea. It can therefore be concluded that the stronger insolation during the mid-Holocene produces three main responses to the atmosphere over the Maritime Continent:

1. Stronger convection and precipitation over the landmass such as Borneo Island
2. Suppressed convection over the South China Sea.
3. Weaker local Hadley cell.

The combination of stronger insolation on the Maritime Continent landmass is the most important factor to drive the mid-Holocene climate change over the region. Warmer SST is a competing factor to balance the effect of stronger insolation since the two physical quantities exhibit an opposite atmospheric responses especially over the oceanic region. It can therefore be inferred that, the contraction of the Western Pacific Warm Pool (WPWP) does not play a significant role in bringing higher precipitation over Borneo during the mid-Holocene.

Figure 3.9. The changes of the mean of the CAM4 longitudinally (90°E-150°E) averaged streamfunction and v - w vector between the mid-Holocene (A), SSOLIN (B) and WSST (C) with the pre-Industrial scenario during JASO. v - w vector consists of the meridional wind in 1 m/s and the vertical velocity in $5 \cdot 10^2$ Pa/s. The dotted areas are not significant at 95% confidence level based on the Welch's t -test (see Appendix A.4).

3.5. References

- Abram, N.J., McGregor, H.V., Gagan, M.K., Hantoro, W.S. and Suwargadi, B.W., 2009. Oscillations in the southern extent of the Indo-Pacific Warm Pool during the mid-Holocene. *Quaternary Science Reviews*, 28(25-26), pp.2794-2803.
- Arakawa, A. and Lamb, V.R., 1981. A potential enstrophy and energy conserving scheme for the shallow water equations. *Monthly Weather Review*, 109(1), pp.18-36.
- Banacos, P.C. and Schultz, D.M., 2005. The use of moisture flux convergence in forecasting convective initiation: Historical and operational perspectives. *Weather and Forecasting*, 20(3), pp.351-366.
- Berger, A., 1978. Long-term variations of daily insolation and Quaternary climatic changes. *Journal of the Atmospheric Sciences*, 35(12), pp.2362-2367.
- Blanchard, D.O., 1998. Assessing the vertical distribution of convective available potential energy. *Weather and Forecasting*, 13(3), pp.870-877.
- Blunier, T., J. Chappellaz, J. Schwander, B. Stauffer, and D. Raynaud. "Variations in atmospheric methane concentration during the Holocene epoch." *Nature* 374, no. 6517 (1995): 46.
- Burns, S.J., 2011. Speleothem records of changes in tropical hydrology over the Holocene and possible implications for atmospheric methane. *The Holocene*, 21(5), pp.735-741.
- Chen, S., Hoffmann, S.S., Lund, D.C., Cobb, K.M., Emile-Geay, J. and Adkins, J.F., 2016. A high-resolution speleothem record of western equatorial Pacific rainfall: Implications for Holocene ENSO evolution. *Earth and Planetary Science Letters*, 442, pp.61-71.
- Cook, K.H., 2003. Role of continents in driving the Hadley cells. *Journal of the atmospheric sciences*, 60(7), pp.957-976.

- deMenocal, P., Ortiz, J., Guilderson, T., Adkins, J., Sarnthein, M., Baker, L. and
 Yarusinsky, M., 2000. Abrupt onset and termination of the African Humid Period: rapid climate responses to gradual insolation forcing. *Quaternary Science Reviews*, 19, pp.347-361.
- Joussaume, S. and Braconnot, P., 1997. Sensitivity of paleoclimate simulation results to season definitions. *Journal of Geophysical Research: Atmospheres*, 102(D2), pp.1943-1956.
- Kajikawa, Y., Yasunari, T. and Kawamura, R., 2003. The Role of the Local Hadley Circulation over the Western Pacific on the Zonally Asymmetric Anomalies over the Indian Ocean. *Journal of the Meteorological Society of Japan. Ser. II*, 81(2), pp.259-276.
- Lau K-M, and S. Yang, 2002: Walker Circulation. *Encyclopedia of Atmospheric Sciences*
- Lestari, R.K. and Iwasaki, T., 2006. A GCM study on the roles of the seasonal marches of the SST and land-sea thermal contrast in the onset of the Asian summer monsoon. *Journal of the Meteorological Society of Japan. Ser. II*, 84(1), pp.69-83.
- Liu, Z., Kutzbach, J. and Wu, L., 2000. Modeling climate shift of El Nino variability in the Holocene. *Geophysical Research Letters*, 27(15), pp.2265-2268.
- Lin, S.J. and Rood, R.B., 1997. An explicit flux-form semi-Lagrangian shallow-water model on the sphere. *Quarterly Journal of the Royal Meteorological Society*, 123(544), pp.2477-2498.
- Monin, A.S. and Obukhov, A.M.F., 1954. Basic laws of turbulent mixing in the surface layer of the atmosphere. *Contrib. Geophys. Inst. Acad. Sci. USSR*, 151(163), p.e187.
- Monnin, E., Indermühle, A., Dällenbach, A., Flückiger, J., Stauffer, B., Stocker, T.F., Raynaud, D. and Barnola, J.M., 2001. Atmospheric CO₂ concentrations over the last glacial termination. *Science*, 291(5501), pp.112-114.

- Neale, R. and Slingo, J., 2003. The Maritime Continent and its role in the global climate: A GCM study. *Journal of Climate*, 16(5), pp.834-848.
- Neale, R.B., Richter, J.H., Conley, A.J., Park, S., Lauritzen, P.H., Gettelman, A., Rasch, P.J. and Vavrus, J., 2010. S., Taylor, MA, C., Zhang, M., and Lin, S.: Description of the NCAR Community Atmosphere Model (CAM4). *National Center for Atmospheric Research Tech. Rep. NCAR/TN+ STR*.
- Partin, J.W., Cobb, K.M., Adkins, J.F., Clark, B. and Fernandez, D.P., 2007. Millennial-scale trends in west Pacific warm pool hydrology since the Last Glacial Maximum. *Nature*, 449(7161), p.452.
- Riemann-Campe, K., Fraedrich, K. and Lunkeit, F., 2009. Global climatology of convective available potential energy (CAPE) and convective inhibition (CIN) in ERA-40 reanalysis. *Atmospheric Research*, 93(1-3), pp.534-545.
- Singarayer, J.S., Valdes, P.J., Friedlingstein, P., Nelson, S. and Beerling, D.J., 2011. Late Holocene methane rise caused by orbitally controlled increase in tropical sources. *Nature*, 470(7332), p.82.
- Sowers, T., 2010. Atmospheric methane isotope records covering the Holocene period. *Quaternary Science Reviews*, 29(1-2), pp.213-221.
- Sundqvist, H., 1988. Parameterization of condensation and associated clouds in models for weather prediction and general circulation simulation. In *Physically-based modelling and simulation of climate and climatic change* (pp. 433-461). Springer, Dordrecht.
- Tierney, J.E. and deMenocal, P., 2013. Abrupt shifts in Horn of Africa hydroclimate since the Last Glacial Maximum. *Science*, 342(6160), pp.843-846.
- Trismidianto, Yulihastin, E., Satyawardhana, H., Nugroho, J.T. and Ishida, S., 2017, January. The Contribution of the Mesoscale Convective Complexes (MCCs) to total

- rainfall over Indonesian Maritime Continent. In *IOP Conference Series: Earth and Environmental Science* (Vol. 54, No. 1, p. 012027). IOP Publishing.
- Vertenstein, M., Bertini, A., Craig, T., Edwards, J., Levy, M., Mai, A. and Schollenberger, J., 2013. Cesm user's guide (cesm1. 2 release series user's guide). *NCAR technical note*.
- Yamanaka, M.D., 2016. Physical climatology of Indonesian maritime continent: An outline to comprehend observational studies. *Atmospheric Research*, 178, pp.231-259.
- Ye, B., Del Genio, A.D. and Lo, K.K., 1998. CAPE variations in the current climate and in a climate change. *Journal of Climate*, 11(8), pp.1997-2015.
- Zhang, C., 1993. Large-scale variability of atmospheric deep convection in relation to sea surface temperature in the tropics. *Journal of Climate*, 6(10), pp.1898-1913.
- Zhang, G.J. and McFarlane, N.A., 1995. Sensitivity of climate simulations to the parameterization of cumulus convection in the Canadian Climate Centre general circulation model. *Atmosphere-ocean*, 33(3), pp.407-446.

Chapter 4 – ENSO impact on mid-Holocene climate change over the Maritime Continent

Abstract

The Community Atmospheric Model version 4 (CAM4) is used to quantify the impact of the El Nino Southern Oscillation (ENSO) on the climate over the Maritime Continent between the mid-Holocene and the pre-Industrial era. In the numerical experiments, the climate scenarios are modified by removing the anomalous sea surface temperature (SST) of the Community Climate System Model version 4 (CCSM4) in the eastern Pacific, and then the outputs are compared with unmodified scenarios. The experimental results show that the mid-Holocene ENSO impacted climate change over the Maritime Continent insignificantly. However, a mid-Holocene simulation without the ENSO shows a significant influence on the climate. Even then, though, the spatio-temporal climatology of the mid-Holocene precipitation over the region is not appreciably affected. Thus, my CAM4 numerical experiments suggest that the climate over the Maritime Continent is responsive more to the long-term annual variability of the Western Pacific Warm Pool (WPWP) than to the ENSO.

4.1. El Nino Southern Oscillation

The El Nino Southern Oscillation (ENSO) is the dominant mode of variability in the tropical Pacific sea surface temperature (SST) anomalies (Lian & Chen, 2012). The ENSO is a large-scale coupled ocean-atmosphere system that drives the dynamic of the warmer SST in the Western Pacific Warm Pool (WPWP). The WPWP is closely linked with a colder eastern tropical Pacific known as the Pacific cold tongue, together creating a zonal SST gradient. The zonal SST gradient drives the large-scale zonal overturning circulation known as the Walker circulation (Chiang, 2009).

The most prominent ENSO phase is known as the El Nino, which is the eastward migration of the positive SST anomaly in the tropical Pacific (Fig. 4.1). This migration shifts the dominant convective region away from the west (and also results in a prolonged drought over the Maritime Continent) (Ropelewski & Halpert, 1987; Trenberth & Caron, 2000). The coupling between the positive SST anomaly and convective rainfall is made possible by the Bjerknes feedback mechanism through the weakening of the trade winds (Bjerknes, 1966; 1969). Thus, eastern Pacific SST, the trade winds, and precipitation over the Maritime Continent are the main physical quantities identifying the ENSO.

Figure 4.1. Sea surface temperature, circulation, convective zone, and thermocline in the tropical Pacific during El Nino and during normal conditions (Chiang, 2009 adapted from NOAA/PMEL/TAO Project Office, <https://www.pmel.noaa.gov/elnino/>).

4.1.1 Mid-Holocene ENSO in proxy records

Our knowledge of ENSO behaviour during the mid-Holocene comes mainly from paleoclimate proxy records such as coral $\delta^{18}\text{O}$ datasets. Both marine and terrestrial ENSO proxy records report a weaker and less-frequent variability of the ENSO during the mid-Holocene throughout the tropical Pacific. For example, Cobb et al. (2013) report a ~60% reduction of the coral $\delta^{18}\text{O}$ inter-annual signal in the equatorial central Pacific (Fig. 4.2). Chen et al. (2016) reported a reduction in the magnitude and frequency of the speleothem $\delta^{18}\text{O}$ inter-annual signal during the mid-Holocene in northern Borneo, which is located within the WPWP. In addition, Koutavas and Joanides (2012) and Carré et al. (2014) reported a weaker ENSO signal during the mid-Holocene based on the seafloor sediments from the Galapagos and fossil mollusc shells from Peru, in the eastern tropical Pacific.

Figure 4.2. Relative ENSO variance changes in fossil coral $\delta^{18}\text{O}$ time series in the past 7.5 Kyr from seven sites on the Northern Line Islands, the central Pacific (Cobb et al., 2013).

The physical mechanisms that caused the weaker ENSO during the mid-Holocene have been suggested by experiments using the Coupled General Circulation Model (CGCM): the mid-Holocene summer insolation was the main driver producing colder SST over the tropical Pacific, and suppressed the eastward development of the WPWP (Clement et al., 1999, 2000; Liu et al., 2000). The colder SST was produced by the Bjerknes feedback mechanism when the summer insolation created a larger zonal thermal gradient and triggered stronger trade winds.

4.1.2. Mid-Holocene ENSO change in the PMIP3

An & Choi (2014) analysed the mid-Holocene ENSO using an ensemble of 11 CGCMs involved in the Paleoclimate Inter-comparison Project phase-3 (PMIP3). The results show a weakening of the ENSO associated with the basin-wide cooling and stronger trade winds in the tropical Pacific, as suggested by the earlier studies (e.g., Clement et al., 1999, 2000; Liu et al., 2000). However, the ensemble models show that the ENSO was reduced by only ~5.4%, which is significantly less than the ~60% reduction recorded in paleoclimate proxy data (Cobb et al., 2013).

An & Choi (2014) also reported an increase in the frequency of the ENSO in the central tropical Pacific (i.e., in the ENSO Modoki, Ashok et al., 2007), and a decrease in the eastern tropical Pacific. However, this conclusion is still preliminary since the changes are not statistically significant. To further confirm the increase in the central Pacific ENSO during the mid-Holocene, I analyse the spatio-temporal variability of the tropical Pacific SST from the nine PMIP3 models (Table 2.1).

I calculate the power spectrum of the SST within the inter-annual timescales (2-7 years) for each grid along the tropical Pacific. The power spectrum is smoothened by using the Bartlett's method (see Appendix A.7). The peak of the power spectrum can be identified as the core domain of the ENSO spectrum. Thus, its shift in the longitude and/or frequency domain can be observed.

The power spectrum along the tropical Pacific SST displays a variety of spatio-temporal patterns among nine PMIP3 models (Fig. 4.3). The peak inter-annual signals (core ENSO spectra) vary between higher, middle, and lower frequencies and also between eastern and central Pacific domains. The nine PMIP3 models can be clustered in three major groups based on the shift of the core ENSO spectrum between the two climate scenarios:

1. Nearby shift: A shift of the ENSO spectrum to a nearby longitude-frequency domain. Models: CCSM4, MPI-ESM-P, CNRM-CM5, HadGEM2-ES, MIROC-ESM, CSIRO-Mk3.
2. Frequency shift: A substantial shift of the ENSO spectrum along the frequency domain. Models: IPSL-CM5A, GISS-E2-R.
3. Longitude-frequency shift: A substantial shift of the ENSO spectrum along the longitude and frequency domain. Models: MRI-CGCM3.

Figure 4.3. The power spectrum of the SST along the equatorial Pacific ($120^{\circ}\text{E} - 280^{\circ}\text{E}$ and averaged at $\pm 5^{\circ}$ latitudes) for the mid-Holocene (colour) and the pre-Industrial era (contour) of the nine PMIP3 models. The power spectrum is smoothened by using Bartlett's method with a time window of 20-yr (see Appendix A.7).

4.1.2.1 Model-proxy comparison for the mid-Holocene ENSO change in the CCSM4 and PMIP3

The shifts of the core ENSO spectrum in the PMIP3 models are mostly towards nearby longitude-frequency domains except for three models (Fig. 4.3). The MRI-CGCM3 is the only PMIP3 model that exhibits a westward shift of the core ENSO spectrum, leading to the increase in the central Pacific ENSO during the mid-Holocene, as suggested by An and Choi (2014). However, the MRI-CGCM3 shows a stronger core spectrum of the ENSO during the mid-Holocene throughout the tropical Pacific, which is not consistent with the proxy data. Therefore, there is no evidence for a substantial shift of the major ENSO events from the eastern Pacific ENSO in the pre-Industrial scenario to the central Pacific ENSO in the mid-Holocene scenario for these nine PMIP3 models.

The IPSL-CM5A and GISS-E2-R show frequency shifts of the core ENSO spectra. However, each model suffers from biases. The IPSL-CM5A exhibits a higher power of the core ENSO spectrum during the mid-Holocene, which is not consistent with the ENSO proxy data. The GISS-E2-R suffers from a longitudinal bias of the core ENSO spectrum, since it is located at the central instead of the eastern tropical Pacific (Fig. 4.3).

The CCSM4 and the other five PMIP3 models exhibit a variety of the core ENSO spectra along the longitude-frequency domain. Four models exhibit individual problems in their core ENSO spectra:

1. The MPI-ESM-P exhibits a stronger core ENSO spectrum during the mid-Holocene, which is not consistent with the proxy data.
2. The CSIRO-Mk3 exhibits a core ENSO spectrum on the central instead of the typical eastern tropical Pacific in both scenarios.
3. The HadGEM2-ES exhibits a strong and dominant decadal signal in both scenarios, which is not within the ENSO timescale.

4. The MIROC-ESM exhibits a weaker core ENSO spectrum during the mid-Holocene, as suggested by the proxy data, but without any shift in the frequency domain.

The CCSM4 and CNRM-CM5 are the only PMIP3 models that are consistent with the proxy data, having weaker and less-frequent ENSO spectra during the mid-Holocene. The two models show ~20 % weaker magnitudes of their core ENSO spectra during the mid-Holocene (Fig. 4.3). The percentage of a weaker mid-Holocene ENSO in the two models underestimates what was shown by the ENSO proxy dataset (Fig. 4.2). The two models also exhibit ~1-year longer ENSO periodicity during the mid-Holocene. Nonetheless, the two models have previously been shown to perform well in simulating the mid-Holocene climate change over the Maritime Continent (Fig. 2.7).

The power spectrum of the SST along the tropical Pacific provides a comprehensive framework to analyse the spatio-temporal variability of the ENSO. The variety of the core ENSO spectra among the PMIP3 models shows the complexity of air-sea interaction processes simulated by the Coupled-GCMs. Each PMIP3 model simulates the ENSO differently, since there is no first-order consistency in their core ENSO spectra. The only general consistency (six out of the nine PMIP3 models) is in the longitudinal position of the core ENSO spectra, which occurs in the eastern tropical Pacific. Multi-model consistency in the ENSO spectrum along the tropical Pacific should get more attention by the climate modelling community to improve the performance of the CGCMs in simulating the ENSO.

4.2. Numerical experiments using the CAM4 to estimate the impact of the mid-Holocene weaker ENSO on the mean precipitation over the Maritime Continent

In Chapter 2, I have shown that the changes of variance on inter-annual timescales (2-7 years) of the CCSM4 precipitation over the Maritime Continent between the mid-Holocene and the pre-Industrial scenarios are smaller than the changes of variance on annual-and-semi-annual timescales (Fig. 2.13). Furthermore, the changes in the total mean precipitation over Borneo between the two scenarios are clearly attributed to the changes in their monthly climatology, especially during the July-August-September-October (JASO) period (Fig. 2.15). The above indicates minor differences between ENSO impacts on precipitation over the Maritime Continent in the two scenarios. However, the exact role of ENSO impact in the region remains unclear, and is investigated in this chapter.

In one climate regime, the ENSO can contribute to the total mean of the regional precipitation if the atmospheric response is non-linear. However, empirically estimating the non-linearity of ENSO impact is challenging. For example, the two phases of ENSO impact (El Nino and La Nina) are symmetrical (Jiang & Li, 2017), not asymmetrical, over the Maritime Continent. Furthermore, the ENSO affects several atmospheric phenomena on various spatio-temporal scales, such as seasonal and decadal variabilities (Wang & Mehta, 2008; Jiang & Li, 2017), and it is influenced by local features such as topography or land-sea distribution (Chang et al., 2004). An alternative to empirical investigation is the use of numerical experiments, since different conditions of the ENSO can be simulated by modifying the tropical Pacific SST (Kosaka & Xie, 2013).

Numerical experiments can highlight how changes in ENSO impact may affect the total mean precipitation. For this purpose, I have developed two approaches to specifically compare the mid-Holocene and the pre-Industrial scenarios:

1. By first quantifying ENSO impact on the climate in the two scenarios, and then by subtracting one from the other.
2. By quantifying the atmospheric response in the absence of the ENSO in the mid-Holocene scenario, which is an extreme case of the mid-Holocene weaker ENSO scenario.

The numerical simulations were conducted using the Community Atmospheric Model version 4 (CAM4), which is the atmospheric GCM of the CCSM4. The CAM4 simulations required global SST from the CCSM4 for each scenario, and also modified values for the tropical Pacific for the experiments. The grid system of the CCSM4 SST dataset was first converted to make it compatible with the f19 grid system (1.9 x 2.5 degrees of horizontal resolution) used in our CAM4 configuration (see Appendix A.3).

4.2.1. Experimental design

The design of our numerical experiments in the CAM4 must be able to ‘remove’ the main feature of the ENSO, which is the inter-annual variability of the tropical Pacific SST, especially from the central to eastern region. The inter-annual ENSO signal is known to dominate the anomaly of the monthly SST in the central to eastern tropical Pacific (Lian & Chen, 2012). An anomaly is a deviation from the monthly climatology, which automatically retains climate signals outside the annual-and-semi-annual timescales (Chapter 2, Fig. 2.12). Thus, the tropical Pacific SST is modified by removing its anomaly signal. In the other word, the CAM4 simulation is imposed by the global CCSM4 SST with the monthly climatology in the tropical Pacific (Table 4.1).

Figure 4.4. The total variance of the CCSM4 SST anomaly for the pre-Industrial scenario (A) and its differences with the mid-Holocene scenario (B). The red curves indicate the difference in variance is 0 K^2 . The black rectangle (160°E-280°E and 20°S-20°N) is the modified SST region to suppressed ENSO.

Previously I have shown that the core ENSO spectrum of the CCSM4 SST, which was located in the eastern Pacific (~240°E) during the mid-Holocene, was weaker during the pre-Industrial era (Fig. 4.3), while the CCSM4 SST anomaly had a higher variance during the mid-Holocene in the eastern Pacific from ~250°E to the coastal Peru (Fig. 4.4.B). The higher variance of the mid-Holocene SST anomaly in the eastern Pacific does not match the weaker ENSO during the mid-Holocene found in proxy datasets from the eastern Pacific (Carré et al., 2014). This disagreement suggests that ENSO behaviour during the mid-Holocene in the CCSM4 SST may not be represented well in the eastern tropical Pacific, which could affect its impact over the Maritime Continent.

Alternatively, as previously mentioned, the absent ENSO experiment for the mid-Holocene scenario is considered as an extreme case of a weaker ENSO scenario. This extreme scenario could set the upper bound of the weaker ENSO impact during the mid-Holocene.

The modified SST in the central to eastern tropical Pacific covers the area within 160°E-280°E and 20°S-20°N (Fig. 4.4). The area captures the largest variance of the SST anomaly in the tropical Pacific including the central Pacific. A similar area is used by Kosaka and Xie (2013) to perform numerical experiments in a coupled-GCM to study the influence of the tropical Pacific on the global warming hiatus in the present-day climate. Chiang et al. (2009) modified the orbital forcing over a similar area to study its influence on the ENSO during the mid-Holocene. Xie et al. (2009) performed a similar modification on the Indian Ocean SST to study the influence of the Indian Ocean Dipole (IOD) on the climate over the Indo-Pacific. In the above studies, the SST was modified to better serve the main purpose of the experiment. Likewise, the removal of the SST anomaly in the central to eastern tropical Pacific removes the ENSO in the CAM4, because the SST in that area will then only exhibit the climatological signal. The mid-Holocene and the pre-Industrial scenarios with the monthly climatology of the SST in the tropical Pacific is labelled as 'NoENSO' (Table 4.1). All the CAM4 simulations are conducted with a time range of 300 yrs and a spin-up time of 20 yrs; thus, 280 yrs is used for the follow-up analyses. The statistics (mean, variance, etc.) of any physical quantities for all the years in the period of interest is labelled as 'total', e.g., total mean precipitation.

Conf.	mH_NoENSO	piC_NoENSO
Orbital	ecc = 0.018682	ecc = 0.016724
	obl = 24.105°	obl = 23.446°
	per-180° = 0.87°	per-180° = 102.04°
GHG	CH ₄ = 650 ppbv	CH ₄ = 760 ppbv
	CO ₂ = 280 ppmv	CO ₂ = 280 ppmv
	N ₂ O = 270 ppbv	N ₂ O = 270 ppbv
	CFC-11 & 12 = 0 ppbv	CFC-11 & 12 = 0 ppbv
SST	Mid-Holocene SST of the CCSM4 in the PMIP3 (300 yrs), with monthly climatology in the tropical Pacific	Pre-Industrial SST of the CCSM4 in PMIP3 (300 yrs), with monthly climatology in the tropical Pacific

Table 4.1. The absent ENSO experimental scenarios for the mid-Holocene (mH_NoENSO) and the pre-Industrial (piC_NoENSO) era. The monthly climatology of the SST covers the central to eastern tropical Pacific region (160°E-280°E and 20°S-20°N, see Figure 4.4). See Table 3.1 for the original mid-Holocene and pre-Industrial scenarios as described by the PMIP3 (<https://pmip3.lsce.ipsl.fr/>).

4.2.2. The role of a weaker ENSO in changing the mean precipitation over the tropical Pacific and the Maritime Continent during the mid-Holocene

ENSO impact on the present-day precipitation over the tropical Pacific is recognized as the zonal dipole of an anomalous convective zone along the tropical Pacific (Ropelewski & Halpert, 1987; Trenberth & Caron, 2000). The zonal dipole during the El Nino is characterized by an anomalous convective zone located in the eastern tropical Pacific, which thus results in severe droughts in the western tropical Pacific, for example, the 1997/98 droughts in Indonesia (Kirono et al., 1999).

ENSO impact in one climate scenario on the tropical Pacific, including the Maritime Continent, as simulated by my CAM4, is quantified by subtracting the NoENSO simulation output from the unmodified one, for example, piC minus piC_NoENSO (Fig 4.5.A). The subtraction result then equals ENSO impact on the atmosphere. The quantified ENSO impact on the total mean precipitation along the tropical Pacific for both scenarios (Fig. 4.5.A, B) resembles El Nino impact (Jiang & Li, 2017), suggesting possible dominance of El Nino impact in my CAM4 simulations. This dominance could be explained, for example, by higher frequency or magnitude of El Nino over La Nina events, contribution from other climate events modulated by the ENSO (e.g., Pacific Decadal Oscillation), and the influence of local features (e.g., topography, land-sea distribution, etc.).

In this CAM4 experiment, ENSO impact on the total mean precipitation over the Maritime Continent of the mid-Holocene and pre-Industrial scenarios are significant (non-zero), allowing me to quantify the difference between ENSO impacts in both scenarios.

Figure 4.5. The total mean of the CAM4 precipitation for the pre-Industrial (piC) minus piC_NoENSO (A), mid-Holocene (mH) minus mH_NoENSO (B), and (mH minus mH_NoENSO) minus (piC minus piC_NoENSO) (C). The dotted areas are not significant at the 95% confidence level based on Welch's *t*-test (see Appendix A.4).

There is no significant change in ENSO impact on the total mean precipitation over the Maritime Continent between the two climate scenarios except in the Papua region (Fig. 4.5.C), where a weaker ENSO contributes to a slightly wetter climate. In Chapter 2, I have shown that the climate over the northern coast of Papua became drier in the mid-Holocene scenario (Fig. 2.10). In Chapter 3, I have discussed that the drier climate was mainly influenced by the colder SST in the adjacent sea during the boreal autumn season (Fig. 3.4.B & 3.7.A). The opposite impacts of the colder SST and the weaker ENSO on the Papua region suggest that it is the dynamics of the WPWP on annual-and-semi-annual timescales, rather than the ENSO itself that influence the long-term climate over the region more significantly.

The changes of ENSO impact over the western part of the Maritime Continent, including Borneo, are mostly small and statistically insignificant. Therefore, the impact of a weaker ENSO during the mid-Holocene cannot explain the changes in the total mean precipitation over the Maritime Continent, including Borneo. The results further confirm the dominant role of stronger insolation during the July-August-September-October (JASO) period in causing more abundant precipitation over the Maritime Continent during the mid-Holocene.

However, the insignificant impact of a weaker ENSO over the Maritime Continent shown in this study might be due to a relatively small ENSO weakening during the mid-Holocene (by ~20% according to the CCSM4), as compared to ~60% shown by coral $\delta^{18}\text{O}$ records in the central equatorial Pacific (Cobb et al., 2013). Another possible factor is the higher variance of the mid-Holocene SST anomaly in the eastern tropical Pacific (Fig. 4.4.B). These biases remain a challenge in the climate modelling community.

A weaker ENSO also likely modulated JASO precipitation over the Maritime Continent during the mid-Holocene. Nevertheless, the weaker ENSO was a secondary

factor, and would not have negated the mechanism of higher JASO precipitation over the Maritime Continent, as previously discussed in Chapter 3.

4.2.3. Mid-Holocene precipitation change in the NoENSO scenario and its monthly climatology over the Maritime Continent

The absence of the ENSO during the mid-Holocene can be considered as the extreme case of a weaker ENSO, and its simulation provides the upper bound of ENSO impact on the climate over the Maritime Continent. This simulation is essentially the mH_NoENSO experiment from the previous discussion, but here the mH_NoENSO experiment is directly compared with the pre-Industrial scenario. This comparison serves two purposes: first, to analyse the precipitation impact of ENSO absence, and second, to highlight the monthly climatological impact on the precipitation during ENSO absence.

Figure 4.6. The total (A) and JASO (B) mean of the CAM4 precipitation for the pre-Industrial (piC) minus the mH_NoENSO scenario. The dotted areas are not significant at the 95% confidence level based on Welch's *t*-test (see Appendix A.4). The CAM4 monthly climatology of the precipitation over Borneo (110°E to 118°E, 2°S to 5°N) for the pre-Industrial (piC), piC_NoENSO, mid-Holocene (mH), and mH_NoENSO scenarios (C), are plotted with the range of uncertainties at the 95% confidence level based on Student's *t*-distribution (see Appendix A.5).

The mH_NoENSO experiment has highlighted several points:

1. Island precipitation over the Maritime Continent increases compared to the pre-Industrial scenario, without any substantial changes in the spatial distribution (Fig. 4.6.A). The similar spatial distribution suggests similar climatology throughout the Maritime Continent.
2. The annual signal in the monthly climatology of the precipitation remains dominant, with similar maxima in the JASO period (Fig. 4.6.B, C).
3. The climatology of Borneo precipitation is amplified at most by ~10% compared to the mid-Holocene scenario. The amplification occurs in all-seasons except for the March-June period (Fig. 4.6.C).

The mH_NoENSO experiment shows that a weaker ENSO significantly amplified mid-Holocene precipitation over the region, while annual variability remained the dominant factor of the mid-Holocene climate. Thus, annual variability of the mid-Holocene precipitation is a robust feature regardless of the ENSO.

4.3. References

- An, S.I. and Choi, J., 2014. Mid-Holocene tropical Pacific climate state, annual cycle, and ENSO in PMIP2 and PMIP3. *Climate dynamics*, 43(3-4), pp.957-970.
- Ashok, K., Behera, S.K., Rao, S.A., Weng, H. and Yamagata, T., 2007. El Niño Modoki and its possible teleconnection. *Journal of Geophysical Research: Oceans*, 112(C11).
- Barry, R.G. and Chorley, R.J., 2003. *Atmosphere, weather and climate*, 8th Edition. Routledge.
- Bjerknes, J., 1966. A possible response of the atmospheric Hadley circulation to equatorial anomalies of ocean temperature. *Tellus*, 18(4), pp.820-829.
- Bjerknes, J., 1969. Atmospheric teleconnections from the equatorial Pacific. *Monthly Weather Review*, 97(3), pp.163-172.
- Carré, M., Sachs, J.P., Purca, S., Schauer, A.J., Braconnot, P., Falcón, R.A., Julien, M. and Lavallée, D., 2014. Holocene history of ENSO variance and asymmetry in the eastern tropical Pacific. *Science*, 345(6200), pp.1045-1048.
- Chang, C.P., Wang, Z., Ju, J. and Li, T., 2004. On the relationship between western maritime continent monsoon rainfall and ENSO during northern winter. *Journal of Climate*, 17(3), pp.665-672.
- Chen, S., Hoffmann, S.S., Lund, D.C., Cobb, K.M., Emile-Geay, J. and Adkins, J.F., 2016. A high-resolution speleothem record of western equatorial Pacific rainfall: Implications for Holocene ENSO evolution. *Earth and Planetary Science Letters*, 442, pp.61-71.
- Chiang, J.C., 2009. The tropics in paleoclimate. *Annual Review of Earth and Planetary Sciences*, 37, pp.263-297.

- Clement, A.C., Seager, R. and Cane, M.A., 1999. Orbital controls on the El Niño/Southern Oscillation and the tropical climate. *Paleoceanography*, 14(4), pp.441-456.
- Clement, A.C., Seager, R. and Cane, M.A., 2000. Suppression of El Niño during the mid-Holocene by changes in the Earth's orbit. *Paleoceanography*, 15(6), pp.731-737.
- Cobb, K.M., Westphal, N., Sayani, H.R., Watson, J.T., Di Lorenzo, E., Cheng, H., Edwards, R.L. and Charles, C.D., 2013. Highly variable El Niño–Southern Oscillation throughout the Holocene. *Science*, 339(6115), pp.67-70.
- Jiang, L. and Li, T., 2017. Why rainfall response to El Niño over Maritime Continent is weaker and non-uniform in boreal winter than in boreal summer. *Climate Dynamics*, pp.1-19.
- Kirono, D.G., Tapper, N.J. and McBride, J.L., 1999. Documenting Indonesian rainfall in the 1997/1998 El Niño event. *Physical Geography*, 20(5), pp.422-435.
- Kosaka, Y. and Xie, S.P., 2013. Recent global-warming hiatus tied to equatorial Pacific surface cooling. *Nature*, 501(7467), p.403.
- Koutavas, A. and Joannides, S., 2012. El Niño–Southern Oscillation extrema in the Holocene and Last Glacial Maximum. *Paleoceanography and Paleoclimatology*, 27(4).
- Lian, T. and Chen, D., 2012. An evaluation of rotated EOF analysis and its application to tropical Pacific SST variability. *Journal of Climate*, 25(15), pp.5361-5373.
- Liu, Z., Kutzbach, J. and Wu, L., 2000. Modeling climate shift of El Niño variability in the Holocene. *Geophysical Research Letters*, 27(15), pp.2265-2268.
- Ropelewski, C.F. and Halpert, M.S., 1987. Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation. *Monthly weather review*, 115(8), pp.1606-1626.

- Trenberth, K.E., 1997. The definition of el nino. *Bulletin of the American Meteorological Society*, 78(12), pp.2771-2778.
- Trenberth, K.E. and Caron, J.M., 2000. The Southern Oscillation revisited: Sea level pressures, surface temperatures, and precipitation. *Journal of Climate*, 13(24), pp.4358-4365.
- Wang, H. and Mehta, V.M., 2008. Decadal variability of the Indo-Pacific warm pool and its association with atmospheric and oceanic variability in the NCEP–NCAR and SODA reanalyses. *Journal of Climate*, 21(21), pp.5545-5565.
- Xie, S.P., Hu, K., Hafner, J., Tokinaga, H., Du, Y., Huang, G. and Sampe, T., 2009. Indian Ocean capacitor effect on Indo–western Pacific climate during the summer following El Niño. *Journal of Climate*, 22(3), pp.730-747.

Chapter 5 - Conclusions and Suggestions

The investigation of the mid-Holocene climate change over the Maritime Continent in this thesis was carried out in three general steps. In the first step, I validated that the CCSM4 robustly simulated the Mid-Holocene climate change represented by climate proxies and present-day observations. In the second step, I determined the most significant climate change features simulated by the CCSM4 by analysing the change of climate data variability. After performing spectral analyses, I highlighted the most dominant climate signals. In the third step, I suggested the possible mechanisms of climate change in the mid-Holocene based on the two main factors (insolation and SST). I performed numerical experiments on the atmosphere under the influence of each factor using the CAM4, which is the atmospheric model of the CCSM4. After diagnosing the experimental results, I proposed dynamical mechanisms and discussed their impact. The robustness of the main features and the mechanisms were further confirmed by quantifying the impact of the mid-Holocene weaker ENSO. The summary of these three steps, their significance, and suggestions for future work are elaborated below.

5.1. Step 1 - Validation

In Chapter 2, I started with a two-way validation of whether the CCSM4 could be regarded as one of the better performing GCMs under the PMIP3 umbrella in simulating the mid-Holocene climate change over the Maritime Continent. First, I compared the Inter-tropical Convergence Zone (ITCZ) and El Nino Southern Oscillation (ENSO) in the CCSM4 present-day climate scenario with satellite observations. While the CCSM4-simulated SST did not suffer from the typical western Pacific ‘cold pool’ bias, the CCSM4 still suffered from the typical ‘too-west’ bias in the ENSO and the ‘double’ ITCZ. I showed that the ‘too-west’ bias of the ENSO did not significantly impact the Maritime Continent. The CCSM4 simulated ‘double’ ITCZ especially during the boreal

winter, which did not affect any of the follow-up analyses since the focus period of this thesis is July-October.

Second, I compared a proxy-model of the mid-Holocene climate change in the CCSM4 with nine other models under the PMIP3 umbrella. The CCSM4 was one of the three models capable of simulating the mid-Holocene wetter climate over Borneo, the drier/no-change climate over the Flores region, and the wetter climate over the northern coast of Australia, as shown by the rainfall proxy (Partin et al., 2007; Griffith et al., 2010; Denniston et al., 2013). The CCSM4 was also one of the nine models capable of simulating the mid-Holocene colder SST in the east-west perimeter of the Maritime Continent, as shown by the SST proxy data (Abram et al., 2009). Both validation steps then showed that the CCSM4 was one of the better performing GCMs.

5.1.1. Suggestions for step 1

The GCM validation method applied here, while sufficient for this study's purpose, is not the most rigorous method developed to date. The highest standard of model-proxy comparison was showcased by DiNezio and Tierney (2013), who validated a GCM simulation for the LGM scenario using Cohen's κ statistic. That method can be applied to precipitation and SST proxies for a more comprehensive evaluation of the GCM performance (e.g. biases, etc.). In addition, that method is needed for the mid-Holocene scenario when the number of hydroclimate proxies extracted from the Maritime Continent increases significantly (Konecky et al., 2016; Wurtzel et al., 2018) in order to better represent the complexity of the land-sea distribution of the Maritime Continent.

The representation of land-sea distribution, especially the coastline, is important to properly represent the averaged amount of rainfall in an archipelagic landscape, such as the Maritime Continent, through various small-scale mechanisms (Yamanaka et al., 2018). The coastline can be properly represented in the climate model by improving the

spatial resolution of the GCM to $\ll 100$ km. However, spatial resolution improvement to $\ll 100$ km is premised on the availability of computational resources.

A common solution to limited computational resources is downscaling the GCM output using a regional climate model (RCM). For example, numerical downscaling using an RCM has been used and developed to study the present-day climate over the Maritime Continent, especially to study higher frequency spectra such as the intra-seasonal to diurnal timescales (Teo et al., 2011; Fonseca et al., 2018).

Collaborative effort to continuously improve the performance of RCMs over the Maritime Continent has resulted in the creation of the framework of the Southeast Asia Regional Climate Downscaling (SEACLID) supported by a global partnership project named the Coordinated Regional Climate Downscaling Experiment (CORDEX) for Southeast Asia (SEACLID/CORDEX-SEA) (Cruz et al., 2017; Ngo-Duc et al., 2017). The project involves most countries in Southeast Asia and climate-related international institutions such as the CSIRO (Australia), the Hadley Centre (United Kingdom), and the APEC Climate Center (South Korea). SEACLID/CORDEX-SEA downscaled the GCM output from $1.5^\circ \times 1.5^\circ$ resolution up to ~ 36 -km resolution for the Southeast Asian domain, including the Maritime Continent (Fig. 5.1). A similar downscaling can be applied to improve the spatial resolution of the simulated mid-Holocene and other paleoclimate scenarios over the Maritime Continent.

Figure 5.1. The downscaling domain of the SEACLID/CORDEX-SEA (Ngo-Duc et al., 2017).

5.2. Step 2 - Attribution

In order to attribute the main cause of the differences between the two climatic regimes, their dominant variability of the simulated climate over Borneo and the Maritime Continent in general was identified first. The annual-and-semi-annual precipitation change over the Maritime Continent, especially Borneo, dominates the spectrum, including the inter-annuals. The maximum increase in the mid-Holocene monthly climatology of the precipitation over Borneo was during the July-August-September-October (JASO) period, coherent with the peak of the mid-Holocene insolation over Borneo and SST of the South China Sea. Insolation and SST have been hypothesized as

the main physical quantities of the changing mid-Holocene climate (Abram et al., 2009; Cheng et al., 2012). The role of each quantity was identified by numerical experiments in step 3.

5.2.1. Suggestions for step 2

The attribution of climate change is an attempt to identify features that change significantly between two climate regimes. The identification of each feature is based on the currently established knowledge about a climate system of a region, which is limited to the present-day climate. A climate feature is identified by variability analyses through a variety of methods. In this thesis, I implemented a spectral analysis, while multivariate statistics (e.g., the Empirical Orthogonal Function (EOF)) would also have been an option.

The EOF method serves two important purposes which are relevant to an atmospheric system: One is to accommodate atmospheric non-linearity since the EOF method does not presume a sinusoidal signal; the other is to increase a signal-to-noise ratio by identifying noises as signals with the lowest covariance (Aldrian & Djamil, 2008). A notable study using the EOF method over the Maritime Continent re-confirms three rainfall zones there (Aldrian & Susanto (2003), Fig. 5.2). The zonation of rainfall is identified by the EOF method as the annual and semi-annual signals, consistent with our spectral analyses. Furthermore, the EOF decomposes the annual signals into the monsoonal and anti-monsoonal signals, which are not differentiated in our spectral analyses. Thus, changes of the EOF components between two climate regimes can provide more details about the climate features that drive changes in the mean climate state.

Figure 5.2. The three rainfall zones (A, B & C) identified by the EOF method (the top panel). Each zone exhibits annual (A & C) and semi-annual (B) signals (the lower panels), with the dashed lines as the signals $\pm$ their standard deviations (Aldrian & Susanto, 2003).

5.3. Step 3 - Mechanisms

The CAM4 simulates a wet phase of the mid-Holocene annual precipitation over Borneo during the JASO period. This period was much wetter during the mid-Holocene compared to the same period during the pre-Industrial era due to stronger seasonality of their monthly climatology during the mid-Holocene. The mid-Holocene maximum climatological precipitation occurred in September, which coincided with equatorial insolation and a warmer regional SST especially of the South China Sea. Insolation and SST have been hypothesised as the main causes of increased precipitation over Borneo (Abram et al., 2009; Cheng et al., 2012), which points to the fundamental mechanisms behind the mid-Holocene climate change (the main focus of Chapter 3). The numerical experiments using the CAM4 were conducted to investigate the atmospheric responses to each physical quantity (insolation and SST). Two mid-Holocene experiments were conducted in the CAM4 with either quantity replaced by its pre-Industrial era value. The effect of each quantity was quantified by subtracting the pre-Industrial era values from the mid-Holocene experimental values.

The CAM4 mid-Holocene experiments showed opposite atmospheric responses over the Maritime Continent for insolation and SST. Mid-Holocene insolation increased precipitation over the landmass in the Maritime Continent (e.g., Borneo), but decreased precipitation over the oceanic region (e.g., the South China Sea). The higher precipitation during the mid-Holocene was due to stronger convection, which was a response to a warmer surface temperature either on land or at sea. The impact of stronger landmass convection suppressed convection over the South China Sea due to a slower and drier cross equatorial moisture transport. A further implication of suppressed convection over the South China Sea was a weaker local Hadley cell. Thus, the mid-Holocene warmer

landmass as a respond to the equatorial insolation caused a wetter climate over Borneo and weaker the local Hadley cell.

The impact of the weaker mid-Holocene ENSO was also investigated by CAM4 experiments (Chapter 4). The CAM4 experiments excluded the ENSO (labelled as NoENSO), which represented the extreme case of a weaker ENSO. The CAM4 mid-Holocene NoENSO experiment showed increased precipitation over the Maritime Continent, with a persistent annual signal. The mid-Holocene ENSO experiment also showed the robustness of the annual precipitation in their monthly climatology as the main feature of the mid-Holocene climate change over the Maritime Continent.

5.3.1. Suggestions for step 3

The effort to unveil the fundamental mechanisms behind the mid-Holocene climate change over the Maritime Continent consisted of two parts: the numerical experiments and the diagnostics. The importance of landmass in increasing mid-Holocene tropical convection can be further investigated by a numerical experiment on the land-sea distribution. Yamanaka (2016) reviews studies on the important role of landmass and coastlines in the Maritime Continent in comparison with an all-ocean Coupled-GCM simulation (aqua-planet) (Fig. 5.3). A mid-Holocene simulation with an aqua-planet could provide a similar result as our mid-Holocene experiment with the pre-Industrial era SST (WSST), since the subtraction of the pre-Industrial era simulation from the WSST experiment removed the atmospheric responses to the land radiative feedback.

Figure 5.3. Cloud distribution by a meteorological satellite (left) and an aqua-planet simulation (right) (Yamanaka, 2016).

The diagnostic of the mid-Holocene wetter landmass in the Maritime Continent can be further detailed by investigating the changes in the moisture budget. Chen and Bordoni (2016) expanded the changes in moisture flux convergence into the changes in thermodynamics (temperature, specific humidity, etc.) and circulation (winds) as well as their covariance. One of the derived terms is about the changes in saturation specific humidity as a function of a changing temperature, which is known as the ‘wet-get-wetter’ pattern. The detailed mechanism of the mid-Holocene stronger convection could be based on this particular term. Chen and Bordoni (2016) also approximated a vertical velocity at 500 mb as the ratio between the energy input and the moist static stability. The approximation can be used to investigate the contribution of the mid-Holocene stronger insolation to the dipole changes of the vertical velocity between the land and sea in the Maritime Continent.

The large-scale implication of the mid-Holocene climate change over the Maritime Continent could be extended to the Asian Monsoon region. Lestari and Iwasaki (2006) explained the relationship between the summer southerly and the monsoonal westerly, suggesting that the summer southerly drives the monsoonal westerly. Thus, a weaker local Hadley cell over the Maritime Continent during the mid-Holocene could slow down the monsoonal westerly, and affect the mid-Holocene summer ITCZ. Furthermore, the slower westerly could also affect the zonal overturning circulation known as the Walker cell, which is associated with the ENSO and other tropical inter-annual climate phenomena. DiNezio and Tierney (2013) demonstrate a slower Walker cell due to a weaker convection over the Maritime Continent in the Last Glacial Maximum (LGM) scenario. The dynamical relationship between the low-level meridional flux over the Maritime Continent and a large-scale zonal flux could generally explain the changes of the Walker cell throughout various climate epochs.

5.4. References

- Abram, N.J., McGregor, H.V., Gagan, M.K., Hantoro, W.S. and Suwargadi, B.W., 2009. Oscillations in the southern extent of the Indo-Pacific Warm Pool during the mid-Holocene. *Quaternary Science Reviews*, 28(25-26), pp.2794-2803.
- Aldrian, E. and Dwi Susanto, R., 2003. Identification of three dominant rainfall regions within Indonesia and their relationship to sea surface temperature. *International Journal of Climatology*, 23(12), pp.1435-1452.
- Aldrian, E. and Djamil, Y.S., 2008. Spatio-temporal climatic change of rainfall in East Java Indonesia. *International Journal of Climatology: A Journal of the Royal Meteorological Society*, 28(4), pp.435-448.
- Chen, J. and Bordoni, S., 2016. Early summer response of the East Asian summer monsoon to atmospheric CO₂ forcing and subsequent sea surface warming. *Journal of Climate*, 29(15), pp.5431-5446.
- Cheng, H., Sinha, A., Wang, X., Cruz, F.W. and Edwards, R.L., 2012. The Global Paleomonsoon as seen through speleothem records from Asia and the Americas. *Climate Dynamics*, 39(5), pp.1045-1062.
- Cruz, F.T., Narisma, G.T., Dado, J.B., Singhruck, P., Tangang, F., Linarka, U.A., Wati, T., Juneng, L., Phan-Van, T., Ngo-Duc, T. and Santisirisomboon, J., 2017. Sensitivity of temperature to physical parameterization schemes of RegCM4 over the CORDEX-Southeast Asia region. *International Journal of Climatology*, 37(15), pp.5139-5153.
- Denniston, R.F., Wyrwoll, K.H., Polyak, V.J., Brown, J.R., Asmerom, Y., Wanamaker Jr, A.D., LaPointe, Z., Ellerbroek, R., Barthelmes, M., Cleary, D. and Cugley, J., 2013. A Stalagmite record of Holocene Indonesian–Australian summer monsoon

- variability from the Australian tropics. *Quaternary Science Reviews*, 78, pp.155-168.
- DiNezio, P.N. and Tierney, J.E., 2013. The effect of sea level on glacial Indo-Pacific climate. *Nature Geoscience*, 6(6), p.485.
- Fonseca, R., Koh, T.Y. and Teo, C.K., 2018. Multi-scale interactions in a high-resolution tropical-belt experiment and observations. *Climate Dynamics*, pp.1-30.
- Griffiths, M.L., Drysdale, R.N., Gagan, M.K., Zhao, J.X., Ayliffe, L.K., Hellstrom, J.C., Hantoro, W.S., Frisia, S., Feng, Y.X., Cartwright, I. and Pierre, E.S., 2009. Increasing Australian–Indonesian monsoon rainfall linked to early Holocene sea-level rise. *Nature Geoscience*, 2(9), p.636.
- Konecky, B., Russell, J. and Bijaksana, S., 2016. Glacial aridity in central Indonesia coeval with intensified monsoon circulation. *Earth and Planetary Science Letters*, 437, pp.15-24.
- Lestari, R.K. and Iwasaki, T., 2006. A GCM study on the roles of the seasonal marches of the SST and land-sea thermal contrast in the onset of the Asian summer monsoon. *Journal of the Meteorological Society of Japan. Ser. II*, 84(1), pp.69-83.
- Ngo-Duc, T., Tangang, F.T., Santisirisomboon, J., Cruz, F., Trinh-Tuan, L., Nguyen-Xuan, T., Phan-Van, T., Juneng, L., Narisma, G., Singhruck, P. and Gunawan, D., 2017. Performance evaluation of RegCM4 in simulating extreme rainfall and temperature indices over the CORDEX-Southeast Asia region. *International Journal of Climatology*, 37(3), pp.1634-1647.
- Partin, J.W., Cobb, K.M., Adkins, J.F., Clark, B. and Fernandez, D.P., 2007. Millennial-scale trends in west Pacific warm pool hydrology since the Last Glacial Maximum. *Nature*, 449(7161), p.452.

- Teo, C.K., Koh, T.Y., Chun-Fung Lo, J. and Chandra Bhatt, B., 2011. Principal component analysis of observed and modeled diurnal rainfall in the Maritime Continent. *Journal of Climate*, 24(17), pp.4662-4675.
- Wurtzel, J.B., Abram, N.J., Lewis, S.C., Bajo, P., Hellstrom, J.C., Troitzsch, U. and Heslop, D., 2018. Tropical Indo-Pacific hydroclimate response to North Atlantic forcing during the last deglaciation as recorded by a speleothem from Sumatra, Indonesia. *Earth and Planetary Science Letters*, 492, pp.264-278.
- Yamanaka, M.D., 2016. Physical climatology of Indonesian maritime continent: An outline to comprehend observational studies. *Atmospheric Research*, 178, pp.231-259.
- Yamanaka, M.D., Ogino, S.Y., Wu, P.M., Jun-Ichi, H., Mori, S., Matsumoto, J. and Syamsudin, F., 2018. Maritime continent coastlines controlling Earth's climate. *Progress in Earth and Planetary Science*, 5(1), p.21.

Complete List of References

- Abram, N.J., McGregor, H.V., Gagan, M.K., Hantoro, W.S. and Suwargadi, B.W., 2009. Oscillations in the southern extent of the Indo-Pacific Warm Pool during the mid-Holocene. *Quaternary Science Reviews*, 28(25-26), pp.2794-2803.
- Adler, R.F., Huffman, G.J., Chang, A., Ferraro, R., Xie, P.P., Janowiak, J., Rudolf, B., Schneider, U., Curtis, S., Bolvin, D. and Gruber, A., 2003. The version-2 global precipitation climatology project (GPCP) monthly precipitation analysis (1979–present). *Journal of hydrometeorology*, 4(6), pp.1147-1167.
- Aldrian, E. and Dwi Susanto, R., 2003. Identification of three dominant rainfall regions within Indonesia and their relationship to sea surface temperature. *International Journal of Climatology*, 23(12), pp.1435-1452.
- Aldrian, E. and Djamil, Y.S., 2008. Spatio-temporal climatic change of rainfall in East Java Indonesia. *International Journal of Climatology: A Journal of the Royal Meteorological Society*, 28(4), pp.435-448.
- Ackerley, D., Reeves, J., Barr, C., Bostock, H., Fitzsimmons, K., Fletcher, M.S., Gouramanis, C., McGregor, H., Mooney, S., Phipps, S.J. and Tibby, J., 2016. Evaluation of PMIP2 and PMIP3 simulations of mid-Holocene climate in the Indo-Pacific, Australasian and Southern Ocean regions. *Climate of the Past*, 13(11), p.1661.
- An, S.I. and Choi, J., 2014. Mid-Holocene tropical Pacific climate state, annual cycle, and ENSO in PMIP2 and PMIP3. *Climate dynamics*, 43(3-4), pp.957-970.
- Arakawa, A. and Lamb, V.R., 1981. A potential enstrophy and energy conserving scheme for the shallow water equations. *Monthly Weather Review*, 109(1), pp.18-36.

- Ashok, K., Behera, S.K., Rao, S.A., Weng, H. and Yamagata, T., 2007. El Niño Modoki and its possible teleconnection. *Journal of Geophysical Research: Oceans*, 112(C11).
- Audley-Charles, M.G., 1981. Geological history of the region of Wallace's Line. *Wallace's line and plate tectonics*, pp.24-35.
- Barry, R.G. and Chorley, R.J., 2003. *Atmosphere, weather and climate*, 8th Edition. Routledge.
- Banacos, P.C. and Schultz, D.M., 2005. The use of moisture flux convergence in forecasting convective initiation: Historical and operational perspectives. *Weather and Forecasting*, 20(3), pp.351-366.
- Bartlett, M.S., 1948. Smoothing Periodograms from Time-Series with Continuous Spectra. *Nature*, 161, pp.686-687.
- Berger, A., 1978. Long-term variations of daily insolation and Quaternary climatic changes. *Journal of the Atmospheric Sciences*, 35(12), pp.2362-2367.
- Bjerknes, J., 1966. A possible response of the atmospheric Hadley circulation to equatorial anomalies of ocean temperature. *Tellus*, 18(4), pp.820-829.
- Bjerknes, J., 1969. Atmospheric teleconnections from the equatorial Pacific. *Monthly Weather Review*, 97(3), pp.163-172.
- Blanchard, D.O., 1998. Assessing the vertical distribution of convective available potential energy. *Weather and Forecasting*, 13(3), pp.870-877.
- Blunier, T., J. Chappellaz, J. Schwander, B. Stauffer, and D. Raynaud. "Variations in atmospheric methane concentration during the Holocene epoch." *Nature* 374, no. 6517 (1995): 46.
- Braconnot, P., Harrison, S.P., Otto-Bliesner, B., Abe-Ouchi, A., Jungclauss, J., Peterschmitt, J.Y., 2011. The Paleoclimate Modeling Intercomparison Project

- contribution to CMIP5. *WCRP Coupled Model Intercomparison Project-Phase 5-CMIP5*, p.15.
- Burns, S.J., 2011. Speleothem records of changes in tropical hydrology over the Holocene and possible implications for atmospheric methane. *The Holocene*, 21(5), pp.735-741.
- Bush, A.B. and Philander, S.G.H., 1999. The climate of the Last Glacial Maximum: Results from a coupled atmosphere-ocean general circulation model. *Journal of Geophysical Research: Atmospheres*, 104(D20), pp.24509-24525.
- Carré, M., Sachs, J.P., Purca, S., Schauer, A.J., Braconnot, P., Falcón, R.A., Julien, M. and Lavallée, D., 2014. Holocene history of ENSO variance and asymmetry in the eastern tropical Pacific. *Science*, 345(6200), pp.1045-1048.
- Chang, C.P., Wang, Z., Ju, J. and Li, T., 2004. On the relationship between western maritime continent monsoon rainfall and ENSO during northern winter. *Journal of Climate*, 17(3), pp.665-672.
- Chase, T.N. and Barry, R.G., 2004. Numerical models of the general circulation, climate and weather prediction. In *Atmosphere, Weather and Climate* (pp. 202-216). Routledge.
- Cheng, H., Sinha, A., Wang, X., Cruz, F.W. and Edwards, R.L., 2012. The Global Paleomonsoon as seen through speleothem records from Asia and the Americas. *Climate Dynamics*, 39(5), pp.1045-1062.
- Cheng, H., Edwards, R.L., Sinha, A., Spötl, C., Yi, L., Chen, S., Kelly, M., Kathayat, G., Wang, X., Li, X. and Kong, X., 2016. The Asian monsoon over the past 640,000 years and ice age terminations. *Nature*, 534(7609), p.640.
- Chen, J. and Bordoni, S., 2016. Early summer response of the East Asian summer monsoon to atmospheric CO₂ forcing and subsequent sea surface warming. *Journal of Climate*, 29(15), pp.5431-5446.

- Chen, S., Hoffmann, S.S., Lund, D.C., Cobb, K.M., Emile-Geay, J. and Adkins, J.F., 2016. A high-resolution speleothem record of western equatorial Pacific rainfall: Implications for Holocene ENSO evolution. *Earth and Planetary Science Letters*, 442, pp.61-71.
- Chiang, J.C., 2009a. The tropics in paleoclimate. *Annual Review of Earth and Planetary Sciences*, 37, pp.263-297.
- Chiang, J.C.H., Fang, Y. and Chang, P., 2009b. Pacific climate change and ENSO activity in the mid-Holocene. *Journal of Climate*, 22(4), pp.923-939.
- Clement, A.C., Seager, R. and Cane, M.A., 1999. Orbital controls on the El Niño/Southern Oscillation and the tropical climate. *Paleoceanography*, 14(4), pp.441-456.
- Clement, A.C., Seager, R. and Cane, M.A., 2000. Suppression of El Niño during the Mid-Holocene by changes in the Earth's orbit. *Paleoceanography*, 15(6), pp.731-737.
- Cobb, K.M., Westphal, N., Sayani, H.R., Watson, J.T., Di Lorenzo, E., Cheng, H., Edwards, R.L. and Charles, C.D., 2013. Highly variable El Niño–Southern Oscillation throughout the Holocene. *Science*, 339(6115), pp.67-70.
- Cook, K.H., 2003. Role of continents in driving the Hadley cells. *Journal of the atmospheric sciences*, 60(7), pp.957-976.
- Crucifix, M., Braconnot, P., Harrison, S.P. and Otto-Bliesner, B., 2005. Second phase of paleoclimate modelling intercomparison project. *EOS, Transactions American Geophysical Union*, 86(28), pp.264-264.
- Cruz, F.T., Narisma, G.T., Dado, J.B., Singhruck, P., Tangang, F., Linarka, U.A., Wati, T., Juneng, L., Phan-Van, T., Ngo-Duc, T. and Santisirisomboon, J., 2017. Sensitivity of temperature to physical parameterization schemes of RegCM4 over the CORDEX-Southeast Asia region. *International Journal of Climatology*, 37(15), pp.5139-5153.

- Dayem, K.E., Noone, D.C. and Molnar, P., 2007. Tropical western Pacific warm pool and maritime continent precipitation rates and their contrasting relationships with the Walker Circulation. *Journal of Geophysical Research: Atmospheres*, 112(D6).
- deMenocal, P., Ortiz, J., Guilderson, T., Adkins, J., Sarnthein, M., Baker, L. and Yarusinsky, M., 2000. Abrupt onset and termination of the African Humid Period: rapid climate responses to gradual insolation forcing. *Quaternary Science Reviews*, 19, pp.347-361.
- Denniston, R.F., Wyrwoll, K.H., Polyak, V.J., Brown, J.R., Asmerom, Y., Wanamaker Jr, A.D., LaPointe, Z., Ellerbroek, R., Barthelmes, M., Cleary, D. and Cugley, J., 2013. A stalagmite record of Holocene Indonesian–Australian summer monsoon variability from the Australian tropics. *Quaternary Science Reviews*, 78, pp.155-168.
- DiNezio, P.N., Clement, A., Vecchi, G.A., Soden, B., Broccoli, A.J., Otto-Bliesner, B.L. and Braconnot, P., 2011. The response of the Walker circulation to Last Glacial Maximum forcing: Implications for detection in proxies. *Paleoceanography and Paleoclimatology*, 26(3).
- DiNezio, P.N. and Tierney, J.E., 2013. The effect of sea level on glacial Indo-Pacific climate. *Nature Geoscience*, 6(6), p.485.
- Druyan, L.M., Somerville, R.C. and Quirk, W.J., 1975. Extended-range forecasts with the GISS model of the global atmosphere. *Monthly Weather Review*, 103(9), pp.779-795.
- Easterling, D.R., Evans, J.L., Groisman, P.Y., Karl, T.R., Kunkel, K.E. and Ambenje, P., 2000. Observed variability and trends in extreme climate events: a brief review. *Bulletin of the American Meteorological Society*, 81(3), pp.417-426.

- Fairbanks, R.G. and Dodge, R.E., 1979. Annual periodicity of the $^{18}\text{O}/^{16}\text{O}$ and $^{13}\text{C}/^{12}\text{C}$ ratios in the coral *Montastrea annularis*. *Geochimica et Cosmochimica Acta*, 43(7), pp.1009-1020.
- Felis, T. and Pätzold, J., 2004. Climate reconstructions from annually banded corals. *Global environmental change in the ocean and on land*, pp.205-227.
- Fonseca, R., Koh, T.Y. and Teo, C.K., 2018. Multi-scale interactions in a high-resolution tropical-belt experiment and observations. *Climate Dynamics*, pp.1-30.
- Gathorne-Hardy, F.J., Davies, R.G., Eggleton, P. and Jones, D.T., 2002. Quaternary rainforest refugia in south-east Asia: using termites (Isoptera) as indicators. *Biological Journal of the Linnean Society*, 75(4), pp.453-466.
- Gent, P.R., Danabasoglu, G., Donner, L.J., Holland, M.M., Hunke, E.C., Jayne, S.R., Lawrence, D.M., Neale, R.B., Rasch, P.J., Vertenstein, M. and Worley, P.H., 2011. The community climate system model version 4. *Journal of Climate*, 24(19), pp.4973-4991.
- Griffiths, M.L., Drysdale, R.N., Gagan, M.K., Zhao, J.X., Ayliffe, L.K., Hellstrom, J.C., Hantoro, W.S., Frisia, S., Feng, Y.X., Cartwright, I. and Pierre, E.S., 2009. Increasing Australian–Indonesian monsoon rainfall linked to early Holocene sea-level rise. *Nature Geoscience*, 2(9), p.636.
- Groisman, P.Y., Karl, T.R., Easterling, D.R., Knight, R.W., Jamason, P.F., Hennessey, K.J., Suppiah, R., Page, C.M., Wibig, J., Fortuniak, K. and Razuvaev, V.N., 1999. Changes in the probability of heavy precipitation: important indicators of climatic change. In *Weather and Climate Extremes* (pp. 243-283). Springer, Dordrecht.
- Haywood, A.M., Ramstein, G. and Abe-Ouchi, A., 2011. Comparing structurally different climate models in a paleoenvironmental context. *Eos, Transactions American Geophysical Union*, 92(21), pp.180-180.

- Hutchison, C.S., 1989. *Geological evolution of South-east Asia* (Vol. 13, p. 368). Oxford: Clarendon Press.
- Indermühle, A., Stocker, T.F., Joos, F., Fischer, H., Smith, H.J., Wahlen, M., Deck, B., Mastroianni, D., Tschumi, J., Blunier, T. and Meyer, R., 1999. Holocene carbon-cycle dynamics based on CO₂ trapped in ice at Taylor Dome, Antarctica. *Nature*, 398(6723), p.121.
- IPCC (Intergovernmental Panel on Climate Change), 2001. Climate change 2001: the scientific basis. *Contribution of Working Group I to the third assessment report of the Intergovernmental Panel on Climate Change*, p.881.
- IPCC, Contribution of Working Groups I, II and III to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (2014). *Climate Change 2014 : Synthesis Report*. IPCC, Geneva, Switzerland: Cambridge University Press.
- Japan Meteorological Agency, 2006: Characteristics of Global Sea Surface Temperature Analysis Data (COBE-SST) for Climate Use. *Monthly Report on Climate System Separated Volume*, **12**, 116pp.
- Jia, X., Ge, J. and Wang, S., 2015. Diverse impacts of ENSO on wintertime rainfall over the Maritime Continent. *International Journal of Climatology*, 36(9), pp.3384-3397.
- Jiang, L. and Li, T., 2017. Why rainfall response to El Niño over Maritime Continent is weaker and non-uniform in boreal winter than in boreal summer. *Climate Dynamics*, pp.1-19.
- Joussaume, S. and Braconnot, P., 1997. Sensitivity of paleoclimate simulation results to season definitions. *Journal of Geophysical Research: Atmospheres*, 102(D2), pp.1943-1956.
- Kajikawa, Y., Yasunari, T. and Kawamura, R., 2003. The Role of the Local Hadley Circulation over the Western Pacific on the Zonally Asymmetric Anomalies over the

- Indian Ocean. *Journal of the Meteorological Society of Japan. Ser. II*, 81(2), pp.259-276.
- Kirono, D.G., Tapper, N.J. and McBride, J.L., 1999. Documenting Indonesian rainfall in the 1997/1998 El Nino event. *Physical Geography*, 20(5), pp.422-435.
- Konecky, B., Russell, J. and Bijaksana, S., 2016. Glacial aridity in central Indonesia coeval with intensified monsoon circulation. *Earth and Planetary Science Letters*, 437, pp.15-24.
- Kosaka, Y. and Xie, S.P., 2013. Recent global-warming hiatus tied to equatorial Pacific surface cooling. *Nature*, 501(7467), p.403.
- Koutavas, A. and Joannides, S., 2012. El Niño–Southern Oscillation extrema in the Holocene and Last Glacial Maximum. *Paleoceanography and Paleoclimatology*, 27(4).
- Lau K-M, and S. Yang, 2002: Walker Circulation. *Encyclopedia of Atmospheric Sciences*
- Lian, T. and Chen, D., 2012. An evaluation of rotated EOF analysis and its application to tropical Pacific SST variability. *Journal of Climate*, 25(15), pp.5361-5373.
- Liu, Z., Kutzbach, J. and Wu, L., 2000. Modeling climate shift of El Nino variability in the Holocene. *Geophysical Research Letters*, 27(15), pp.2265-2268.
- Lestari, R.K. and Iwasaki, T., 2006. A GCM study on the roles of the seasonal marches of the SST and land-sea thermal contrast in the onset of the Asian summer monsoon. *Journal of the Meteorological Society of Japan. Ser. II*, 84(1), pp.69-83.
- Lestari, R.K. and Iwasaki, T., 2012. An Empirical Study of the Response of the South China Sea Summer Monsoon to the Remote Forcing. *SOLA*, 8, pp.65-68.
- Li, G. and Xie, S.P., 2014. Tropical biases in CMIP5 multimodel ensemble: The excessive equatorial Pacific cold tongue and double ITCZ problems. *Journal of Climate*, 27(4), pp.1765-1780.

- Lin, J.L., 2007. The double-ITCZ problem in IPCC AR4 coupled GCMs: Ocean–atmosphere feedback analysis. *Journal of Climate*, 20(18), pp.4497-4525.
- Lin, S.J. and Rood, R.B., 1997. An explicit flux-form semi-Lagrangian shallow-water model on the sphere. *Quarterly Journal of the Royal Meteorological Society*, 123(544), pp.2477-2498.
- Liu, Z., Kutzbach, J. and Wu, L., 2000. Modeling climate shift of El Nino variability in the Holocene. *Geophysical Research Letters*, 27(15), pp.2265-2268.
- Lough, J.M. and Barnes, D.J., 1990. Intra-annual timing of density band formation of Porites coral from the central Great Barrier Reef. *Journal of Experimental Marine Biology and Ecology*, 135(1), pp.35-57.
- Lough, J.M., 2010. Climate records from corals. *Wiley Interdisciplinary Reviews: Climate Change*, 1(3), pp.318-331.
- Mahowald, N.M., Yoshioka, M., Collins, W.D., Conley, A.J., Fillmore, D.W. and Coleman, D.B., 2006. Climate response and radiative forcing from mineral aerosols during the last glacial maximum, pre-industrial, current and doubled-carbon dioxide climates. *Geophysical Research Letters*, 33(20).
- Marzin, C., Rahmat, R., Bernie, D., Brichenno, L., Buonomo, E., Calvert, D., Cannaby, H., Chan, S., Chattopadhyay, M., Cheong, W.K. and Hassim, M.E., 2015. Singapore's second national climate change study–Phase 1. *MET OFFICE, E., UK, CENTRE FOR CLIMATE RESEARCH SINGAPORE, S., NATIONAL OCEANOGRAPHY CENTRE, L., UK, CSIRO, A. & NEWCASTLE UNIVERSITY, N., UK (eds.)*.
- Mayr, E., 1944. Wallace's Line in the light of recent zoogeographic studies. *The Quarterly Review of Biology*, 19(1), pp.1-14.
- Meehl, G.A., Karl, T., Easterling, D.R., Changnon, S., Pielke Jr, R., Changnon, D., Evans, J., Groisman, P.Y., Knutson, T.R., Kunkel, K.E. and Mearns, L.O., 2000. An introduction to trends in extreme weather and climate events: observations,

socioeconomic impacts, terrestrial ecological impacts, and model projections.

Bulletin of the American Meteorological Society, 81(3), pp.413-416.

Meijaard, E., 2004. Biogeographic history of the Javan leopard *Panthera pardus* based on a craniometric analysis. *Journal of Mammalogy*, 85(2), pp.302-310.

Moerman, J.W., Cobb, K.M., Adkins, J.F., Sodemann, H., Clark, B. and Tuen, A.A., 2013. Diurnal to interannual rainfall $\delta^{18}\text{O}$ variations in northern Borneo driven by regional hydrology. *Earth and Planetary Science Letters*, 369, pp.108-119.

Monin, A.S. and Obukhov, A.M.F., 1954. Basic laws of turbulent mixing in the surface layer of the atmosphere. *Contrib. Geophys. Inst. Acad. Sci. USSR*, 151(163), p.e187.

Monnin, E., Indermühle, A., Dällenbach, A., Flückiger, J., Stauffer, B., Stocker, T.F., Raynaud, D. and Barnola, J.M., 2001. Atmospheric CO₂ concentrations over the last glacial termination. *Science*, 291(5501), pp.112-114.

Myers, N., Mittermeier, R.A., Mittermeier, C.G., Da Fonseca, G.A. and Kent, J., 2000. Biodiversity hotspots for conservation priorities. *Nature*, 403(6772), p.853.

Neale, R. and Slingo, J., 2003. The Maritime Continent and its role in the global climate: A GCM study. *Journal of Climate*, 16(5), pp.834-848.

Neale, R.B., Richter, J.H., Conley, A.J., Park, S., Lauritzen, P.H., Gettelman, A., Rasch, P.J. and Vavrus, J., 2010. S., Taylor, MA, C., Zhang, M., and Lin, S.: Description of the NCAR Community Atmosphere Model (CAM4). *National Center for Atmospheric Research Tech. Rep. NCAR/TN+ STR*.

Ngo-Duc, T., Tangang, F.T., Santisirisomboon, J., Cruz, F., Trinh-Tuan, L., Nguyen-Xuan, T., Phan-Van, T., Juneng, L., Narisma, G., Singhruck, P. and Gunawan, D., 2017. Performance evaluation of RegCM4 in simulating extreme rainfall and temperature indices over the CORDEX-Southeast Asia region. *International Journal of Climatology*, 37(3), pp.1634-1647.

- Nogués-Paegle, J., Mechoso, C.R., Fu, R., Berbery, E.H., Chao, W.C., Chen, T.C., Cook, K., Diaz, A.F., Enfield, D., Ferreira, R. and Grimm, A.M., 2002. Progress in Pan American CLIVAR research: understanding the South American monsoon. *Meteorologica*, 27(12), pp.1-30.
- Partin, J.W., Cobb, K.M., Adkins, J.F., Clark, B. and Fernandez, D.P., 2007. Millennial-scale trends in west Pacific warm pool hydrology since the Last Glacial Maximum. *Nature*, 449(7161), p.452.
- Petit, J.R., Jouzel, J., Raynaud, D., Barkov, N.I., Barnola, J.M., Basile, I., Bender, M., Chappellaz, J., Davis, M., Delaygue, G. and Delmotte, M., 1999. Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature*, 399(6735), p.429.
- Qian, J.H., Robertson, A.W. and Moron, V., 2013. Diurnal cycle in different weather regimes and rainfall variability over Borneo associated with ENSO. *Journal of Climate*, 26(5), pp.1772-1790.
- Ramage, C.S., 1968. Role of a tropical “maritime continent” in the atmospheric circulation. *Mon. Wea. Rev.*, 96(6), pp.365-370.
- Riemann-Campe, K., Fraedrich, K. and Lunkeit, F., 2009. Global climatology of convective available potential energy (CAPE) and convective inhibition (CIN) in ERA-40 reanalysis. *Atmospheric Research*, 93(1-3), pp.534-545.
- Ropelewski, C.F. and Halpert, M.S., 1987. Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation. *Monthly weather review*, 115(8), pp.1606-1626.
- Ruddiman, W.F., 2008. *Earth's climate: past and future*, 2nd Edition. W. H. Freeman.
- Rudin, W., 1959. Some theorems on Fourier coefficients. *Proceedings of the American Mathematical Society*, 10(6), pp.855-859.

- Sathiamurthy, E. and Voris, H.K., 2006. Maps of Holocene sea level transgression and submerged lakes on the Sunda Shelf. *Tropical Natural History*, (2), pp.1-44.
- Schär, C., Vidale, P.L., Lüthi, D., Frei, C., Häberli, C., Liniger, M.A. and Appenzeller, C., 2004. The role of increasing temperature variability in European summer heatwaves. *Nature*, 427(6972), p.332.
- Simpson, J., Keenan, T.D., Ferrier, B., Simpson, R.H. and Holland, G.J., 1993. Cumulus mergers in the maritime continent region. *Meteorology and Atmospheric Physics*, 51(1-2), pp.73-99.
- Singarayer, J.S., Valdes, P.J., Friedlingstein, P., Nelson, S. and Beerling, D.J., 2011. Late Holocene methane rise caused by orbitally controlled increase in tropical sources. *Nature*, 470(7332), p.82.
- Steinke, S., Kienast, M., Pflaumann, U., Weinelt, M. and Stattegger, K., 2001. A high-resolution sea-surface temperature record from the tropical South China Sea (16,500–3000 yr BP). *Quaternary Research*, 55(3), pp.352-362.
- Stevenson, J., Siringan, F., Finn, J.A.N., Madulid, D. and Heijnis, H., 2010. Paoay Lake, northern Luzon, the Philippines: a record of Holocene environmental change. *Global Change Biology*, 16(6), pp.1672-1688.
- Sultan, B., Janicot, S. and Diedhiou, A., 2003. The West African monsoon dynamics. Part I: Documentation of intraseasonal variability. *Journal of Climate*, 16(21), pp.3389-3406.
- Sundqvist, H., 1988. Parameterization of condensation and associated clouds in models for weather prediction and general circulation simulation. In *Physically-based modelling and simulation of climate and climatic change* (pp. 433-461). Springer, Dordrecht.

- Taschetto, A.S., Gupta, A.S., Jourdain, N.C., Santoso, A., Ummenhofer, C.C. and England, M.H., 2014. Cold tongue and warm pool ENSO events in CMIP5: mean state and future projections. *Journal of Climate*, 27(8), pp.2861-2885.
- Teo, C.K., Koh, T.Y., Chun-Fung Lo, J. and Chandra Bhatt, B., 2011. Principal component analysis of observed and modeled diurnal rainfall in the Maritime Continent. *Journal of Climate*, 24(17), pp.4662-4675.
- Tierney, J.E. and deMenocal, P., 2013. Abrupt shifts in Horn of Africa hydroclimate since the Last Glacial Maximum. *Science*, 342(6160), pp.843-846.
- Trenberth, K.E., 1997. The definition of el nino. *Bulletin of the American Meteorological Society*, 78(12), pp.2771-2778.
- Trenberth, K.E. and Caron, J.M., 2000. The Southern Oscillation revisited: Sea level pressures, surface temperatures, and precipitation. *Journal of Climate*, 13(24), pp.4358-4365.
- Trismidianto, Yulihastin, E., Satyawardhana, H., Nugroho, J.T. and Ishida, S., 2017, January. The Contribution of the Mesoscale Convective Complexes (MCCs) to total rainfall over Indonesian Maritime Continent. In *IOP Conference Series: Earth and Environmental Science* (Vol. 54, No. 1, p. 012027). IOP Publishing.
- Vecchi, G.A., Soden, B.J., Wittenberg, A.T., Held, I.M., Leetmaa, A. and Harrison, M.J., 2006. Weakening of tropical Pacific atmospheric circulation due to anthropogenic forcing. *Nature*, 441(7089), p.73.
- Vecchi, G.A. and Soden, B.J., 2007. Global warming and the weakening of the tropical circulation. *Journal of Climate*, 20(17), pp.4316-4340.
- Vecchi, G.A., Clement, A. and Soden, B.J., 2008. Examining the tropical Pacific's response to global warming. *Eos, Transactions American Geophysical Union*, 89(9), pp.81-83.

- Vertenstein, M., Bertini, A., Craig, T., Edwards, J., Levy, M., Mai, A. and Schollenberger, J., 2013. Cesm user's guide (cesm1. 2 release series user's guide). *NCAR technical note*.
- Voris, H.K., 2000. Maps of Pleistocene sea levels in Southeast Asia: shorelines, river systems and time durations. *Journal of Biogeography*, 27(5), pp.1153-1167.
- Wang, H. and Mehta, V.M., 2008. Decadal variability of the Indo-Pacific warm pool and its association with atmospheric and oceanic variability in the NCEP–NCAR and SODA reanalyses. *Journal of Climate*, 21(21), pp.5545-5565.
- Wang, Y.J., Cheng, H., Edwards, R.L., An, Z.S., Wu, J.Y., Shen, C.C. and Dorale, J.A., 2001. A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *science*, 294(5550), pp.2345-2348.
- Wang, Y., Cheng, H., Edwards, R.L., Kong, X., Shao, X., Chen, S., Wu, J., Jiang, X., Wang, X. and An, Z., 2008. Millennial-and orbital-scale changes in the East Asian monsoon over the past 224,000 years. *Nature*, 451(7182), p.1090.
- Webster, P.J., 2001. The large-scale structure of the tropical atmosphere. In *Large-Scale Dynamical Processes in the Atmosphere* (p. 235).
- Wurster, C.M., Bird, M.I., Bull, I.D., Creed, F., Bryant, C., Dungait, J.A. and Paz, V., 2010. Forest contraction in north equatorial Southeast Asia during the Last Glacial Period. *Proceedings of the National Academy of Sciences*, 107(35), pp.15508-15511.
- Wurtzel, J.B., Abram, N.J., Lewis, S.C., Bajo, P., Hellstrom, J.C., Troitzsch, U. and Heslop, D., 2018. Tropical Indo-Pacific hydroclimate response to North Atlantic forcing during the last deglaciation as recorded by a speleothem from Sumatra, Indonesia. *Earth and Planetary Science Letters*, 492, pp.264-278.

- Xie, S.P., Hu, K., Hafner, J., Tokinaga, H., Du, Y., Huang, G. and Sampe, T., 2009. Indian Ocean capacitor effect on Indo–western Pacific climate during the summer following El Niño. *Journal of Climate*, 22(3), pp.730-747.
- Yamanaka, M.D., 2016. Physical climatology of Indonesian maritime continent: An outline to comprehend observational studies. *Atmospheric Research*, 178, pp.231-259.
- Yamanaka, M.D., Ogino, S.Y., Wu, P.M., Jun-Ichi, H., Mori, S., Matsumoto, J. and Syamsudin, F., 2018. Maritime continent coastlines controlling Earth's climate. *Progress in Earth and Planetary Science*, 5(1), p.21.
- Ye, B., Del Genio, A.D. and Lo, K.K., 1998. CAPE variations in the current climate and in a climate change. *Journal of Climate*, 11(8), pp.1997-2015.
- Yokoyama, Y., Suzuki, A., Siringan, F., Maeda, Y., Abe-Ouchi, A., Ohgaito, R., Kawahata, H. and Matsuzaki, H., 2011. Mid-Holocene palaeoceanography of the northern South China Sea using coupled fossil-modern coral and atmosphere-ocean GCM model. *Geophysical Research Letters*, 38(8).
- Zhang, C., 1993. Large-scale variability of atmospheric deep convection in relation to sea surface temperature in the tropics. *Journal of Climate*, 6(10), pp.1898-1913.
- Zhang, G.J. and McFarlane, N.A., 1995. Sensitivity of climate simulations to the parameterization of cumulus convection in the Canadian Climate Centre general circulation model. *Atmosphere-ocean*, 33(3), pp.407-446.
- Zheng, Y., Lin, J.L. and Shinoda, T., 2012. The equatorial Pacific cold tongue simulated by IPCC AR4 coupled GCMs: Upper ocean heat budget and feedback analysis. *Journal of Geophysical Research: Oceans*, 117(C5).
- Zheng, W., Wu, B., He, J. and Yu, Y., 2013. The East Asian summer monsoon at mid-Holocene: results from PMIP3 simulations. *Climate of the Past*, 9(1), pp.453-466.

Appendix

A.1. Major climate forcings in paleoclimatology

Land-sea distribution is known to vary significantly in the past due to the rise and fall of sea level caused by many factors such as tectonic processes, ocean thermal expansion, glaciers and ice sheet formation, etc. (Lambeck & Chappell, 2001; Bradley, 2015). Among these factors, the most climate-related one is the variability known as glacial-interglacial cycles with a periodicity $\sim 100,000$ years (Petit et al., 1999; Lisiecki & Raymo, 2005; Jouzel et al., 2007) (Fig. A.1.1). The driving mechanism of the glacial-interglacial cycle is suggested due to the combined factors between the long-term variability of insolation, atmospheric CO_2 concentration and also important internal feedback mechanisms such as ice-albedo feedback (Hays et al., 1976; Petit et al., 1999; Gildor & Tziperman, 2000; Shackleton, 2000; Ganopolski & Calov, 2011). However, the clear distinction between glacial-interglacial cycle and long-term insolation time series has made them mostly considered as a separate climate forcing (Fig. A.1.1).

Figure A.1.1. Time series of CO₂, Temperature, CH₄ and δ¹⁸O_{atm} from Vostok ice core, East Antarctica and June insolation (Wm⁻²) at 65° lat. (Petit et al., 1999).

The long-term variability of insolation is based on the variation of Earth's orbit around the sun. Earth's long-term orbital cycle was famously developed by Milankovitch in 1940s, which consists of three dominant cycles (~100K, ~40K and ~20K yrs) (Clift & Plumb, 2008; Bradley, 2015) (Fig. A.1.2). The orbital cycles also drive the long-term variability of annual insolation which directly affects seasonal climate phenomena such as monsoon and the Inter-tropical Convergence Zone (ITCZ) (Wu et al., 2016a, b). The long-term variability of Asian Monsoon intensity has been identified to be strongly associated with July insolation at 65°N for the last ~650K yrs (Cheng et al., 2016). Such changes in insolation also modulate the inter-hemispheric thermal gradient which shifts the meridional position of the ITCZ (Raymo & Nisancioglu, 2003; Koutavas & Stieglitz, 2004). Annual ITCZ migration follows the warmer hemisphere as a response to the annual insolation. Likewise, similar mechanism is also applied for ITCZ migration on the orbital time scale (Donohoe et al., 2013; Marshall et al., 2014).

Figure A.1.2. Three Earth's orbital parameters: Eccentricity, Obliquity and Precession (Clift & Plumb, 2008).

A.2. Speleothem oxygen isotopes ratio as a rainfall proxy

Oxygen isotopic ratios ($\delta^{18}\text{O}$) in speleothems are commonly used as a rainfall proxy, particularly in the tropics. Speleothems are a collection of secondary calcium carbonates formed inside karst cave systems, including stalagmites, stalactites and flowstones. The oxygen isotope ratios of calcium carbonates inherit the signals of drip water, which can be traced back to the meteoric-water (rainfall) (Fairchild et al., 2006a, b). Thus, speleothem oxygen isotopic ratios are mostly controlled by isotope fractionation of oxygen occurring during atmosphere processes such as evaporation and condensation which originates from open oceans, particularly in the tropics, as well as moisture condensation. In the next few paragraphs I will describe the step-by-step processes of oxygen isotopes fractionation starting from the ocean surface until speleothem formation.

A.2.1. Meteoric-water oxygen isotopes

Oxygen has three stable isotopes (^{16}O , ^{17}O and ^{18}O) while hydrogen has two (^1H and ^2H or D). According to geochemistry notation, isotope ratio is presented as *delta*-(δ) values. So the ratio for oxygen isotopes is formulated as:

$$\delta^{18}\text{O} = \left[\frac{(^{18}\text{O}/^{16}\text{O})_{\text{sample}} - (^{18}\text{O}/^{16}\text{O})_{\text{standard}}}{(^{18}\text{O}/^{16}\text{O})_{\text{standard}}} \right] * 1000 \text{ ‰} \quad \text{Eq. A.2.1}$$

Water molecule with heavier isotopes (e.g., ^{18}O or D) exerts a lower vapour pressure than the lighter isotopes (e.g., ^{16}O or H); thus heavy isotopes become depleted in vapour phase during surface water evaporation, a process dominated by kinetic fractionation mechanism. The heavy isotopes in moist air mass will keep getting depleted throughout a moisture transport trajectory due to rainout processes (Craig & Gordon, 1965). During the rainout processes in a colder environment, water vapour condenses with preferable enrichment of heavy isotopes (Fig. A.2.1).

Figure A.2.1. Oxygen and Hydrogen isotopes fractionation of water in the atmosphere (Xi, 2014 adapted from SAHRA).

Daansgard (1964) summarized several important observations on water isotope:

1. The exponential depletion of heavy isotopes along decreasing temperature can be described by a Rayleigh condensation model.
2. An anti-phased relationship between δ -values of water isotopes and rainfall amount is observed in the tropics, which is known as 'amount effect'.

Rindsberger et al. (1990) observed that convective and/or intense rainfall with weak evaporation tends to lower δ -values than those formed under other conditions.

A.2.2. Speleothem oxygen isotopes

Karst area consists of limestone caves which allow a variety of formations of speleothems. Speleothems growing upward from cave floor (stalagmites) are less subject to post-deposition perturbation and hence most favourable for climate reconstruction. Calcium carbonate in the speleothem inherits oxygen isotopic signatures from surface rainfall, although several chemical processes occur in between rainfall and speleothem

formation (Fig. A.2.2.b). Thus, $\delta^{18}\text{O}$ of calcium carbonate extracted from speleothems has been used extensively as a rainfall proxy.

Figure A.2.2. a. Illustration of slabbed speleothem with the sampling orientation (black lines) along its growth line (Fairchild et al, 2006a). b. Chemical reactions of carbonate dissolution and deposition from the soil layer into a cave system (Fairchild et al, 2006b).

Speleothem can grow in layers with various thicknesses (Fig. A.2.2.a). The growth rate variability can be as frequent as seasonal and annual (Baker et al., 1993; Shopov et al., 1994). Such variability has been attributed to climate influence in terms of rainfall and/or temperature (Railsback et al., 1994). Brook et al. (1999) observed an inter-annual change in growth layer of a speleothem sample from Madagascar, which is attributed to the influence of the El Nino-Southern Oscillation (ENSO). The thicker layers are associated with higher rainfall deposition during El Nino which is the positive phase of ENSO. During El Nino phase, positive SST anomaly appears in the central Indian Ocean

which increases cyclone activity and in turn increases the rainfall over Madagascar (Jury et al., 1995). Speleothem with such high temporal resolution can be important to study rapid changes and multi-scale interaction of climate events (Fairchild et al., 2006a).

A.2.3. Uranium-series dating method

The chronological control (age) of speleothems can mostly be obtained through high-precision uranium-series (U-series) dating method. The method is based on radioisotopic disequilibrium decay of Uranium-234 (^{234}U) to Thorium-230 (^{230}Th) (Richards and Dorale, 2003), hence also named as $^{230}\text{Th}/^{234}\text{U}$ or U-Th dating method. The U-Th method is a preferable tool for age estimation of carbonates up to 600-700 Kyr old if the initial value of $^{234}\text{U}/^{238}\text{U}$ ($\delta^{234}\text{U}$) (often denoted as $\delta^{234}\text{U}$ -initial) is favourable.

On the order of millions of years or longer, the activity ratio of ^{234}U and ^{230}Th in nature material reaches unity, the so-called state of secular equilibrium. The formation of carbonate crystal disturbs the secular equilibrium condition by precipitating U that is dissolved in groundwater but not Th since it is water insoluble. The decaying ^{234}U in the crystal then has a tendency to reach a new state of secular equilibrium with its daughter isotope ^{230}Th . The time to approach the new state of secular equilibrium can be used to determine the age of carbonate samples (Ivanovich and Harmon, 1982). Depending on the initial $\delta^{234}\text{U}$ values of samples, the ultimate time to reach a state of secular equilibrium can vary dramatically (Fig. A.2.3).

Figure A.2.3 $^{230}\text{Th}/^{238}\text{U}$ activity ratio versus time and initial value of $\delta^{234}\text{U}$ by assuming a closed system and initial $^{230}\text{Th} = 0$ (Richards and Dorale, 2003).

The amount of each isotope in carbonate samples can be counted by using mass spectrometers machines. The latest generation of the machine for U-series dating is known as multicollector inductively coupled plasma mass spectrometers (MC-ICP-MS), which was introduced in 1997 (Spötl and Boch, 2012). Dramatic improvements on both accuracy and precision have been achieved in the U-series dating techniques (Shen et al., 2012; Cheng et al., 2013). For example, with a sample size as small as a few mg of calcite, Shen et al. (2012) was able to date modern speleothem or coral samples with a precision of 0.5% or better.

A.3. Grid system conversion

In this thesis, the spherical grid system of the CCSM4 SST dataset has been converted into another grid system, for example, the horizontal lat/lon grid system of the COBE-SST dataset. The conversion method applied was a bilinear interpolation that is available as one of the interpolation modules in the Climate Data Operator (CDO) version 1.9.5 named `remapbil` (Schulzweida, 2018).

The implementation of a bilinear interpolation method in the CDO required a targeted grid system as a reference. The targeted grid system was extracted as metadata (e.g., from the COBE-SST dataset) using another sub-routine in the CDO named `griddes`. The extracted metadata was then used as a reference input to convert the grid system of the CCSM4 SST using `repambil`.

A.4. Welch's t -test

Welch's t -test evaluates the significance of the difference of the mean value between the two datasets. This method is available as the `ttest2` function in the Matlab Statistical Toolbox.

Welch's t -test is a two-sample t -test to test whether the two data samples are independent random samples from normal distribution with equal means (null hypothesis) (Welch, 1947). The evaluation is done by comparing the probability of the 'test statistic' (p-value) to be as extreme as, or more extreme than, the probability under the null hypothesis (100% - confidence level) in the Student's t -distribution. A small p value allows the null hypothesis to be rejected.

The test statistic is:

$$t = \frac{\bar{x} - \bar{y}}{\sqrt{\frac{s_x^2}{n} + \frac{s_y^2}{m}}}, \quad (\text{A.4.1})$$

where $\bar{x}$ and $\bar{y}$ are the sample means, s_x and s_y are the sample standard deviations, and n and m are the sample sizes.

Welch's t -test is the case where it is not assumed that the two data samples are from populations with equal variances, and the test statistic under the null hypothesis has an approximate Student's t -distribution with a number of degrees of freedom (ν) given by Satterthwaite's approximation:

$$\nu \approx \frac{\left(\frac{s_x^2}{n} + \frac{s_y^2}{m}\right)^2}{\frac{s_x^4}{nv_x} + \frac{s_y^4}{mv_y}}. \quad (\text{A.4.2})$$

Here $\nu_x = n - 1$, the degree of freedom associated with the variance of x .

$\nu_y = m - 1$, the degree of freedom associated with the variance of y .

A.5. One-sample Student's t -test

One-sample Student's t -test evaluates the significance of the mean value of a dataset (univariate). The test used zero mean t -distribution as the null hypothesis to be evaluated against the t -statistic of the univariate dataset (Student, 1908). This method is available as the `ttest` function in the Matlab Statistical Toolbox.

The t -statistic is:

$$t = \frac{\bar{x} - \mu}{s/\sqrt{n}}, \quad (\text{A.5.1})$$

where $\bar{x}$ is the sample mean, μ is the hypothesized population mean, s is the sample standard deviation, and n is the sample size. Under the null hypothesis, the test statistic has Student's t distribution with $n - 1$ degrees of freedom.

The test compares the probability of the t -statistic (p-value) to be as extreme as, or more extreme than, the probability under the null hypothesis (100% - confidence level) in the Student's t -distribution. A small p value allows the null hypothesis to be rejected.

A.5.1. Confidence interval of a mean value based on the Student's t -distribution

Student's t -distribution estimates the population distribution of the data sample based on its degrees of freedom. Population distribution can be used to estimate the confidence interval (error bar) of the population mean. The confidence interval is the range of values around the population mean of a certain confidence level (e.g., probability of 95%) in the Student's t -distribution. The confidence interval of the Student's t -distribution is stated as the t -values. The confidence interval of the population mean is the conversion from the t -values based on the t -statistic.

A.6. Pearson's correlation

Pearson's correlation measures the level of relationship between two datasets (bivariate). The level of relationship is quantified by normalizing the covariance of the two datasets with their standard deviations (Wilks, 1995). Pearson's correlation is available as the `corrcoef` function in the Matlab core module.

Pearson's correlation coefficient (r) from two datasets (x & y) with sample size N is:

$$r(x, y) = \frac{1}{N-1} \sum_{i=1}^N \left(\frac{x_i - \bar{x}}{s_x} \right) \left(\frac{y_i - \bar{y}}{s_y} \right), \quad (\text{A.6.1})$$

where $\bar{x}$ and $\bar{y}$ are the sample means, and s_x and s_y are the sample standard deviations.

The range of the correlation coefficient is $[-1, 1]$. A positive (negative) correlation indicates an in-phase (anti-phase) relationship of the two datasets.

A.6.1. Significance test for Pearson's correlation coefficient

The significance of the Pearson's correlation coefficient can be evaluated using a t -test. The t -distribution with a zero correlation coefficient (r) is used as the null hypothesis to evaluate the t -statistic of r from the sample. The t -statistic is:

$$t = r \sqrt{\frac{1-r^2}{n-2}}. \quad (\text{A.6.3})$$

The t -test compares the probability of the t -statistic (p-value) to be as extreme as, or more extreme than, the probability under the null hypothesis (100% - confidence level) in the t -distribution based on the degrees of freedom ($n - 2$). A small p value allows the null hypothesis to be rejected.

A.7. Fast Fourier transform

The fast Fourier transform (FFT) is an algorithm to efficiently compute the discrete Fourier transform (DFT). The DFT converts a finite discrete sample such as a time series into its representation in the frequency domain (Wilks, 1995). This algorithm is available as the `fft` function in the Matlab core module, and its inverse as the `ifft` function.

The DFT of a time-series $X(j)$ with sample size N into the frequency spectra $Y(k)$ are defined as:

$$Y(k) = \sum_{j=0}^{N-1} X(j) \cdot e^{-\frac{i2\pi}{N}kj} \quad (\text{A.7.1})$$

with

$$e^{-\frac{i2\pi}{N}kj} = \cos(2\pi kj/N) - i \cdot \sin(2\pi kj/N). \quad (\text{A.7.2})$$

The inverse transform from $Y(k)$ to $X(j)$ is defined as:

$$X(j) = \frac{1}{N} \sum_{k=0}^{N-1} Y(k) \cdot e^{i2\pi kj/N}. \quad (\text{A.7.3})$$

In the case of $X(j)$ as a time series, j is the temporal index, and k is the frequency (time scale) index. The squared-magnitude component of the DFT (power spectrum) is called periodogram.

The inverse-DFT can be used to perform a temporal filtering of a time series by putting zeros for the Y values that correspond to the undesired k . Thus, signals in different frequency bands from a single time series can be extracted, and their magnitudes (variances) can be compared. The mathematical evidence to relate the variance and the DFT component for a singular or broadband frequency is known as the Parseval's theorem (Rudin et al., 1959).

A.7.1. Bartlett's method

Bartlett's method smooths a periodogram in three general steps (Bartlett, 1948):

1. Splitting the time-series into several non-overlapping data segments with a consistent length (time window);
2. Calculating the periodogram for each data segment; and
3. Calculating the mean power spectra from all the periodograms.

This procedure can smooth the power spectra, albeit at the cost of reduced frequency resolution due to a shorter data length.

A.8. CAM4 installation and test

The operation of CAM4 begins by conducting an installation and performing speed tests. CESM1.2.2 as the main numerical package was installed in the High Performance Computing Center (HPCC) of Nanyang Technological University (NTU), Singapore. HPCC is a cluster based parallel computational facility that was established in 2008 and has been periodically upgraded since then (HPCC, 2012). The system consists of 3000 core processors and 64 Gb of Memory with Red Hat Enterprise Linux 6.3 as its server operating system (ts.ntu.edu.sg/sites/hpc). The access is commercialized with a computational price charged on the number of core per unit time. The price ranges from 0.6866 SGD for 1 core per-day to 59041.92 SGD for 512 cores per-6 months. The system has been proven reliable in hosting regional climate models such as the Weather Research and Forecast (WRF) that was also developed by NCAR (Fonseca et al., 2018).

The installation of CESM1.2.2 uses GNU.4.8.4 as the compiler and openMPI.1.10.2 as its message passing interface for the parallel computational function. The rest of the supporting software and libraries such as netCDF.4.4.0 depends on the version compatibility. Several attempts in a trial-and-error basis were conducted to find the compatible softwares and libraries. Such an attempt was done by compiling a so-called ‘null’ configuration (compset X) of CESM1.2.2. This is the first step to test the compatibility of the compiler, supporting softwares and libraries without any codes of the model component are compiled.

The configuration of CESM1.2.2 can be grouped into four types based on the number of model components involved (Vertenstein et al., 2013), which are as follows:

1. Framework compatibility (compset X)
2. AGCM (compset F)
3. Coupled-GCM with slab ocean model (compset E)

4. Coupled-GCM with 3D ocean model (compset B)

The second test is the AGCM configuration (compset F) with the purpose to identify the optimum number of cores that will be involved. The test was done by performing a short-run of CAM4 in 1.9 x 2.5 degrees of the horizontal resolution. The model-time was purposely changed between 2, 5 and 11 days. The number of cores also changed with several independent runs done for each of them. They are:

- 20 runs for 64 and 128 cores
- 15 runs for 256 cores

The real-time spans for the simulation were recorded, and together with the model-time they were extrapolated into the number of years in model-time per one day in real-time. Figure A.3.1 shows the number of years that were obtained by the model in one real-time day in three different groups based on the numbers of cores. The model seems to have an inconsistent performance from slow to relatively fast computation in all the three groups. However, it is obvious that the peak performance occurs at the group of 128 cores which can reach ~27 yrs of model-time in one day real-time. The inconsistency is a drawback of the system that might be caused by different memory loads at each core which can slow down the synchronization of the parallel processes. Such an issue remains in the system and is considered unavoidable. Therefore, the solution done in practice was to closely check the speed of each simulation at the start, and re-started the simulation if the speed was considered too slow.

The short-run test managed to identify the optimum performance of CAM4 configuration installed in the HPCC machine. The performance of the model can reach as much as ~27 yrs of model-time simulated in one day real-time by using 128 cores. This number of cores is used in the CAM4 configuration for the paleoclimate simulations.

Figure A.8.1. Model-years accomplished by a certain number of cores per real-time day

A.9. References

- Baker, A., Smart, P.L., Edwards, R.L. and Richards, D.A., 1993. Annual growth banding in a cave stalagmite. *Nature*, 364(6437), p.518.
- Bartlett, M.S., 1948. Smoothing Periodograms from Time-Series with Continuous Spectra. *Nature*, 161, pp.686-687.
- Bradley, R.S., 2015. *Paleoclimatology: reconstructing climates of the Quaternary*, 3rd Edition. Elsevier.
- Brook, G.A., Rafter, M.A., Railsback, L.B., Sheen, S.W. and Lundberg, J., 1999. A high-resolution proxy record of rainfall and ENSO since AD 1550 from layering in stalagmites from Anjohibe Cave, Madagascar. *The Holocene*, 9(6), pp.695-705.
- Cheng, H., Edwards, R.L., Shen, C.C., Polyak, V.J., Asmerom, Y., Woodhead, J., Hellstrom, J., Wang, Y., Kong, X., Spötl, C. and Wang, X., 2013. Improvements in ²³⁰Th dating, ²³⁰Th and ²³⁴U half-life values, and U–Th isotopic measurements by multi-collector inductively coupled plasma mass spectrometry. *Earth and Planetary Science Letters*, 371, pp.82-91.
- Cheng, H., Edwards, R.L., Sinha, A., Spötl, C., Yi, L., Chen, S., Kelly, M., Kathayat, G., Wang, X., Li, X. and Kong, X., 2016. The Asian monsoon over the past 640,000 years and ice age terminations. *Nature*, 534(7609), p.640.
- Clift, P.D. and Plumb, R.A., 2008. *The Asian monsoon: causes, history and effects* (Vol. 270). Cambridge: Cambridge University Press.
- Craig H, Gordon L (1965) Deuterium and oxygen-18 variations in the ocean and the marine atmosphere. In: Symposium on marine geochemistry. Graduate School of Oceanography, University of Rhode Island, OCC Publ No 3: 277
- Dansgaard, W., 1964. Stable isotopes in precipitation. *Tellus*, 16(4), pp.436-468.

- Donohoe, A., Marshall, J., Ferreira, D. and Mcgee, D., 2013. The relationship between ITCZ location and cross-equatorial atmospheric heat transport: From the seasonal cycle to the Last Glacial Maximum. *Journal of Climate*, 26(11), pp.3597-3618.
- Fairchild, I.J., Frisia, S., Borsato, A. and Tooth, A.F. 2006a. Speleothems. In: *Geochemical Sediments and Landscapes* (ed. Nash, D.J. and McLaren, S.J.), Blackwells, Oxford
- Fairchild, I.J., Smith, C.L., Baker, A., Fuller, L., Spötl, C., Matthey, D. and McDermott, F., 2006b. Modification and preservation of environmental signals in speleothems. *Earth-Science Reviews*, 75(1-4), pp.105-153.
- Fonseca, R., Koh, T.Y. and Teo, C.K., 2018. Multi-scale interactions in a high-resolution tropical-belt experiment and observations. *Climate Dynamics*, pp.1-30.
- Ganopolski, A. and Calov, R., 2011. The role of orbital forcing, carbon dioxide and regolith in 100 kyr glacial cycles. *Climate of the Past*, 7(4), pp.1415-1425.
- Gildor, H. and Tziperman, E., 2000. Sea ice as the glacial cycles' climate switch: Role of seasonal and orbital forcing. *Paleoceanography and Paleoclimatology*, 15(6), pp.605-615.
- Hays, J.D., Imbrie, J. and Shackleton, N.J., 1976, December. Variations in the Earth's orbit: pacemaker of the ice ages. Washington, DC: American Association for the Advancement of Science.
- High Performance Computing, Nanyang Technological University, 2012: <http://www.ntu.edu.sg/cits/hpc/Pages/default.aspx>
- Jury, M.R., Parker, B.A., Raholijao, N. and Nassor, A., 1995. Variability of summer rainfall over Madagascar: climatic determinants at interannual scales. *International journal of climatology*, 15(12), pp.1323-1332.
- Jouzel, J., Masson-Delmotte, V., Cattani, O., Dreyfus, G., Falourd, S., Hoffmann, G., Minster, B., Nouet, J., Barnola, J.M., Chappellaz, J. and Fischer, H., 2007. Orbital

- and millennial Antarctic climate variability over the past 800,000 years. *science*, 317(5839), pp.793-796.
- Koutavas, A. and Lynch-Stieglitz, J., 2004. Variability of the marine ITCZ over the eastern Pacific during the past 30,000 years. In *The Hadley Circulation: Present, Past and Future* (pp. 347-369). Springer, Dordrecht.
- Lambeck, K. and Chappell, J., 2001. Sea level change through the last glacial cycle. *Science*, 292(5517), pp.679-686.
- Lisiecki, L.E. and Raymo, M.E., 2005. A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography*, 20(1).
- Marshall, J., Donohoe, A., Ferreira, D. and McGee, D., 2014. The ocean's role in setting the mean position of the Inter-Tropical Convergence Zone. *Climate dynamics*, 42(7-8), pp.1967-1979.
- Petit, J.R., Jouzel, J., Raynaud, D., Barkov, N.I., Barnola, J.M., Basile, I., Bender, M., Chappellaz, J., Davis, M., Delaygue, G. and Delmotte, M., 1999. Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature*, 399(6735), p.429.
- Railsback, L.B., Brook, G.A., Chen, J., Kalin, R. and Fleisher, C.J., 1994. Environmental controls on the petrology of a late Holocene speleothem from Botswana with annual layers of aragonite and calcite. *Journal of Sedimentary Research*, 64(1).
- Raymo, M.E. and Nisancioglu, K.H., 2003. The 41 kyr world: Milankovitch's other unsolved mystery. *Paleoceanography*, 18(1).
- Richards, D.A. and Dorale, J.A., 2003. Uranium-series chronology and environmental applications of speleothems. *Reviews in Mineralogy and Geochemistry*, 52(1), pp.407-460.

- Rindsberger, M., Jaffe, S., Rahamim, S. and Gat, J.R., 1990. Patterns of the isotopic composition of precipitation in time and space: data from the Israeli storm water collection program. *Tellus B*, 42(3), pp.263-271.
- Rudin, W., 1959. Some theorems on Fourier coefficients. *Proceedings of the American Mathematical Society*, 10(6), pp.855-859.
- Schulzweida, U., 2018. CDO User Guide.
- Shackleton, N.J., 2000. The 100,000-year ice-age cycle identified and found to lag temperature, carbon dioxide, and orbital eccentricity. *Science*, 289(5486), pp.1897-1902.
- Shen, C.C., Wu, C.C., Cheng, H., Edwards, R.L., Hsieh, Y.T., Gallet, S., Chang, C.C., Li, T.Y., Lam, D.D., Kano, A. and Hori, M., 2012. High-precision and high-resolution carbonate ^{230}Th dating by MC-ICP-MS with SEM protocols. *Geochimica et Cosmochimica Acta*, 99, pp.71-86.
- Shopov, Y.Y., Ford, D.C. and Schwarcz, H.P., 1994. Luminescent microbanding in speleothems: High-resolution chronology and paleoclimate. *Geology*, 22(5), pp.407-410.
- Spötl, C. and Boch, R., 2012. Uranium series dating of speleothems. In *Encyclopedia of Caves (Second Edition)* (pp. 838-844).
- Student, 1908. The probable error of a mean. *Biometrika*, pp.1-25.
- Verstein, M., Bertini, A., Craig, T., Edwards, J., Levy, M., Mai, A. and Schollenberger, J., 2013. Cesm user's guide (cesm1. 2 release series user's guide). *NCAR technical note*.
- Wilks, D. S., 1995. *Statistical methods in the atmospheric sciences*. Oxford ; Waltham, MA: Academic Press.

- Welch, B.L., 1947. The generalization of 'Student's' problem when several different population variances are involved. *Biometrika*, 34(1/2), pp.28-35.
- Wu, C.H., Chiang, J.C., Hsu, H.H. and Lee, S.Y., 2016a. Orbital control of the western North Pacific summer monsoon. *Climate dynamics*, 46(3-4), pp.897-911.
- Wu, C.H., Lee, S.Y., Chiang, J.C. and Hsu, H.H., 2016b. The influence of obliquity in the early Holocene Asian summer monsoon. *Geophysical Research Letters*, 43(9), pp.4524-4530.
- Xi, X., 2014. A review of water isotopes in atmospheric general circulation models: recent advances and future prospects. *International Journal of Atmospheric Sciences*, 2014.