Understanding the Australian Monsoon change during the Last Glacial Maximum with multi-model ensemble

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Abstract

The response of Australian monsoon to the external forcings and the related mechanisms during the Last Glacial Maximum (LGM) is investigated by multi-models' experiments in CMIP5/PMIP3. Although the annual mean precipitation over the Australian monsoon region decreases, the annual range, or the monsoonality, is enhanced. The precipitation increases in early austral summer and decreases in austral winter, resulting in the amplified annual range, but the main contribution comes from the decreased precipitation in austral winter. The decreased winter precipitation is primarily caused by weakened upward motion, although reduced water vapor has also a moderate contribution. The weakened upward motion is induced by the enhanced land–sea thermal contrast, which intensifies the divergence over the northern Australia. The increased Australian monsoon rainfall in early summer, on the other hand, is an integrated result of the positive effect of local dynamic processes (enhanced moisture convergence) and the negative effect of thermodynamics (reduced moisture content). The enhanced moisture convergence is caused by two factors: the strengthened northwest–southeast thermal contrast between the cooler Indochina–western Indonesia and the warmer northeastern Australia, and the east–west sea surface temperature gradients between the warmer western Pacific and cooler eastern Indian Ocean, both due to the alteration of land–sea configuration arising from the sea level drop. The enhanced Australian monsoonality in the LGM is not associated with global scale circulation change such as the shift of the ITCZ, rather, it is mainly due to the change of regional circulations around Australia arising from the changes in land-sea contrast and the east-west SST gradients over the Indo-western Pacific oceans. This finding should be taken into account when investigating its future change under global warming. Our findings may also explain why proxy records indicate different changes in Australian monsoon precipitation during the LGM.

1 Introduction

The changes of the Australian monsoon are crucial for human society and ecology in Australia (Reeves et al., 2013a), considering the socio-economic importance of monsoon rainfall (Wang et al., 2017). As the monsoons of the summer hemisphere are linked via outflows from the opposing winter hemisphere, the Australian monsoon can also influence the Asian–Indonesian–Australian monsoon system (Eroglu et al., 2016). It is important to understand how and why the Australian monsoon would change in response to global climate change.

With strong climatic forcings (including low greenhouse gas (GHG) concentrations, large ice-sheets, and low sea level, etc.), the Last Glacial Maximum (LGM) is one of the key periods that provides an opportunity to better understand the mechanisms of how global and regional climate respond to external forcings (Hewitt et al., 2001; Braconnot et al., 2007; Braconnot et al., 2011; Harrison et al., 2014). Previous studies have investigated how the external forcing and boundary conditions during the LGM affected the Intertropical Convergence Zone (ITCZ) (Broccoli et al., 2006; Donohoe et al., 2013; McGee et al., 2014), the Walker circulation (DiNezio et al., 2011), the Indo-Pacific climate (Xu et al., 2010; DiNezio and Tierney, 2013; DiNezio et al., 2016), the SH circulation (Rojas, 2013), and the global monsoon (Jiang et al., 2015; Yan et al., 2016). The Australian monsoon onset and variability during the post-glacial, the late deglaciation, and the Holocene have also been studied using proxy datasets (Ayliffe et al., 2013; De Deckker et al., 2014; Kuhnt et al., 2015; Bayon et al., 2017). However, due to the limitation of the scarce proxy datasets, the Australian monsoon change during the LGM is far from clearly understood.

There are different proxy evidences indicating different Australian monsoon change during the LGM. Here, the Australian monsoon intensity is represented by the seasonality of precipitation, i.e., a stronger monsoon means a wetter summer and/or drier winter. Some records show wet conditions over Australia during the LGM (Ayliffe et al., 2013), while other proxies indicate drier conditions (Denniston et al., 2013; Denniston et al., 2017; DiNezio and Tierney, 2013). The isotopes from eggshell of five regions across Australia affirms that Australia becomes drier in the LGM (Miller et al., 2016), while the archaeological record showed a refugia-type hunter-gatherer response over northwest and northeast Australia during the LGM (Williams et al., 2013), indicating that these areas may have had a wetter summer and were therefore preferred by people as refugia. The synthesis by the OZ-INTIMATE (Australian INTIMATE, INTegration of Ice core, MArine and TErrestrial records) project (Turney et al., 2006; Petherick et al., 2013) showed that the palaeoenvironment over Northern Australia during the LGM was characterized by drier conditions although wet periods were also noted in the fluvial records (Reeves et al., 2013a; Reeves et al., 2013b).

The change in the Australian monsoon was inconclusive during the LGM based on proxy data. Therefore, scholars started investigating the Australian monsoon change from numerical simulation perspectives. The sensitivity of Australian monsoon to forcings during the late Quaternary has been analyzed using simulations by Fast Ocean Atmosphere Model (Marshall and Lynch, 2006, 2008). Numerical experiments have been conducted to analyze the impacts of obliquity and precession with a coupled General Circulation Model (Wyrwoll et al., 2007) and orbital time-scale circulation with Community Climate Model (Wyrwoll et al., 2012) on the Australian monsoon. However, different models may have different responses to the same external forcings, such that the simulated results may have model dependence. Multi-model ensemble (MME) can reduce the model biases and therefore provide more reasonable results of how and why climate system responds to the external forcing changes. The MME can also provide a clearer perspective on model uncertainties.

Yan et al (2016) thus used the multi-model ensemble approach to examine the response of global monsoon to the LGM conditions. It was found that the global monsoon and most sub-monsoons weakened under the LGM conditions. Some brief hypothesis was made to explain the changes from global and hemispheric perspectives. The Australian monsoon was thought to be strengthened due to the southward shift of the ITCZ resulted from the hemispheric thermal contrast and due to the land-sea thermal contrast resulted from the land-configuration. However, this simulated result of strengthened monsoon or wet condition has not been proved yet. As mentioned above, it is inconclusive whether the Australian monsoon is strengthened or not during the LGM, due to the limitations of proxies' and models' uncertainties. Neither model outputs nor proxy records provide a "true" record of the LGM, as proxy records require interpretation and calibration and may be spatially incomplete, while models contain biases. Therefore, model-data and inter-model comparison are needed and studies on the mechanisms are required to better understand the Australian monsoon change during the LGM. Moreover, some studies show that the Australian climate during the last glacial period was modulated by additional mechanisms rather than simply the ITCZ（Bayon et al., 2017). Thus, single forcing runs are also required to figure out the contributions of different forcings.

This paper investigates the Australian monsoon change during the LGM and its mechanisms from both thermodynamics and dynamics perspectives, using the multi-model ensemble mean derived from models in the fifth phase of the Coupled Model Intercomparison Project (CMIP5) (Taylor et al., 2012) and the third phase of the Paleoclimate Modeling Intercomparison Project (PMIP3) (Braconnot et al., 2012). We are also trying to quantify the contributions of the thermodynamic and the dynamic processes to the Australian monsoon change during the LGM. Additionally, we are applying single forcing run to test the effect of land-configuration as mentioned in Yan et al. (2016). The models and experiments used in this paper are introduced in Sect. 2. Section 3 describes simulated results and the physical mechanisms. The comparison with proxies and other simulations is discussed in Sect. 4 and the conclusions are made in Sect. 5.

2 Methods

2.1 Model and experiments

Two experiments performed by models participating in CMIP5/PMIP3 are compared in this paper: the Last Glacial Maximum Experiment (LGME) and the pre-industrial (PI) control run (piControl). The models and experiments are listed in Table 1, including 7 models and 2 experiments. The models contributed to CMIP5 have been evaluated in the previous studies to have good performance in simulating the Australian monsoon precipitation seasonality or seasonal cycle (Jourdain et al., 2013; Brown et al., 2016), which is used to represent the Australian monsoon intensity in this study.

The last 100 years of the LGME and the last 500 years of the piControl from each model are used to illustrate the model climatology. To obtain the multi-model ensemble (MME), the model outputs were interpolated into a fixed 2.5° (latitude) × 2.5° (longitude) grid using the bilinear interpolation method.

The LGM external forcing and boundary conditions are listed in Table 2. More specific documentation can be found on the PMIP3 website (https://pmip3.lsce.ipsl.fr). Compared with the PI, during the LGM the Southern Hemisphere (SH) low latitudes (30°S-EQ) receive more insolation from January to August and less from August to December. The NH low latitudes (EQ-30°N) receive less insolation from June to October and more from November to May (Fig. S1). Due to the decreased sea level, the landmasses expanded during the LGM. A land bridge formed between Indochina and western Indonesia, and the Arafura Sea between New Guinea and Australia closed and became landmass (Fig. S2).

To illustrate the robust changes simulated by the different models, the signal-to-noise ratio (S2N) test is used. The S2N is defined by the ratio of the absolute mean of the MME (as the signal) to the averaged absolute deviation of the individual model against the MME (as the noise) (Yan et al., 2016). In the following sections, we only consider the areas in which the S2N ratio exceeds one when we examine the differences between the LGME and piControl derived from the MME.

2.2 Decomposition method

For attribution of precipitation changes, we use a simplified relation based on the linearized equation of moisture budget used in the previous works (Chou et al., 2009; Seager et al., 2010; Huang et al., 2013; Endo and Kitoh, 2014; Liu et al., 2016). Considering a quasi-equilibrium state, the vertical integrated moisture conservation can be written as:

$$-\int_{1000}^{0} \nabla \cdot (q\vec{v})dp = P - E \qquad (1)$$

where q is specific humidity, $\vec{v}$ is horizontal velocity, p is pressure, P is precipitation, and E the surface evaporation. Since water vapor is concentrated in the lower troposphere, the vertical integrated total column moisture divergence can be approximately replaced by the integration from the surface to 500 hPa. Define the Δ (.) as the change from PI to the LGM, i.e.,

$$\Delta(.) = (.)\mathrm{LGM} - (.)\mathrm{PI} \qquad (2)$$

Then the precipitation change ΔP can be calculated as follows:

$$\Delta \mathrm{P} = -\int_{p1000}^{p500} \Delta(\mathrm{q} \cdot \nabla\vec{v})dp - \int_{p1000}^{p500} \Delta(\vec{v} \cdot \nabla q)\, dp + \Delta E \qquad (3)$$

To further simplify the equation, we use $-\omega_{500}$ to represent vertical integrated $\nabla\vec{v}$, and q at the surface to represent vertical integrated specific humidity (Huang et al., 2013). Thus, the precipitation change (ΔP) can be represented as

$$\Delta \mathrm{P} \propto \overline{\omega}_{500} \cdot \Delta \mathrm{q} + \bar{\mathrm{q}} \cdot \Delta\omega_{500} + \Delta E - \Delta T_{adv} \quad (4)$$

where $\overline{\omega}_{500}$ is 500 hPa vertical velocity in PI, $\overline{\mathrm{q}}$ is surface specific humidity in PI, ΔT_{adv} is the changes due to the moisture advection ($\int_{p0}^{p500} \Delta(\vec{v} \cdot \nabla q)\, dp$).

The first term in the right-hand side of (4) ($\overline{\omega}_{500} \cdot \Delta\mathrm{q}$) represents thermodynamic effect (due to the change of q), and the second term ($\overline{\mathrm{q}} \cdot \Delta\omega_{500}$) represents dynamic effect (due to the change of circulation).

2.3 Monsoon domain

The monsoon domain is defined following hydroclimate definition, i.e., a contrast between wet summer and dry winter (Wang and Ding 2008). The monsoon domain is defined by the area where the annual range (local summer minus local winter) exceeds 2.0 mm/day, and the local summer precipitation exceeds 55% of the annual total precipitation. Here in the southern hemisphere, summer means November to March and winter means May to September. Since the domains derived from different models are different, and the changes of domain are also different, we use the fixed domain derived from the merged Climate Prediction Center Analysis of Precipitation (CMAP, Xie and Arkin, 1997) and Global Precipitation Climatology Project (GPCP, Huffman et al., 2009) data.

Note that the monsoon domain is shown to give a general view of precipitation change, but not the purpose of this study.

3 Results

We defined the difference of precipitation rate between austral summer (DJF) and austral winter (JJA) as the annual range, i.e. the seasonality, to measure the monsoonality of the Australian monsoon. An increased annual range (or seasonality) means a strengthened monsoonality. Unlike the South African and South American monsoon regions (not shown), the monsoonality of the Australian monsoon derived from the seven models' multi-model ensemble (7MME) is strengthened during the LGM (Fig. 1a). This amplified annual range is the result of increased precipitation in austral summer and decreased precipitation in austral winter (Fig. 1b). Note that the largest decrease in precipitation occurred from April to July (late autumn to early winter), not exactly in austral winter; and the largest increase of precipitation occurred in November and December (ND), i.e., austral early summer. Since the amount of autumn–winter reduction of precipitation exceeds the increased precipitation in early summer, the annual mean precipitation over the strengthened annual range region decreases (by 0.36 mm/day). In summary, while the total annual precipitation decreases in the LGM, the annual range (or the seasonality) of the Australian monsoon rainfall is amplified due to seasonal redistribution of the precipitation, especially the drying in austral autumn (April–May) and winter (JJA) over Australia.

Although there are model biases, most of the models (except MPI-ESM-P) simulate an enhanced annual range (or seasonality/monsoonality) in the central Australian monsoon region (20°S-5°S, 120°E-145°E) (Table 3 and Fig. 1c). Most of the models (except CNRM-CM5 and MPI-ESM-P) also simulate an increased summer precipitation over that region. All the models simulate decreased precipitation from April to September (Fig. 1c). On the other hand, the simulated annual mean precipitation is decreased in most models, except GISS-E2-R. The model uncertainties will be discussed later in Sec. 4.

3.1 Reasons for the decreased precipitation during the LGM in austral winter (JJA)

During the LGM, the lower GHG concentration and the large ice-sheets are the primary causes for the decreased global temperature and the humidity. The global surface specific humidity is reduced by 2 g/kg (or 20 %) in JJA during the LGM, compared with the PI. For the SH monsoon regions, the surface specific humidity is reduced more over the Australian monsoon region than over the other two monsoon regions of South Africa and South America (Fig. 2).

As suggested by the Clausius–Clapeyron relation (C-C relation), one degree of temperature decrease would lead to about a 7 % decrease in the saturation water vapor (Held and Soden, 2006), or roughly the same scale of decrease in the low tropospheric specific humidity. If the circulation, evaporation and advection remains unchanged, the precipitation should also be reduced by 7 % with regard to the equation (4). During the LGM, the simulated near surface air temperature over the central Australian monsoon region (20°S-5°S, 120°E-145°E) decreases significantly by 2.5 K in JJA, which implies a decrease of about 17 % resulted from the C-C relation. However, the simulated precipitation in the LGM is reduced by 1.45 mm/day or 44 % comparing to the PI, which is far beyond the value suggested by the thermodynamic effect (approximately 17 %). This suggests that the majority of the reduction in winter precipitation is due to the changes of rest terms in equation (4), including the circulation change (dynamics), the evaporation change and the change due to the advection term. The changes of each terms show that the circulation change plays a dominant role in the precipitation change over Australia (Fig. S3). The change due to the evaporation is also important. The change due to the advection term is negligible.

The change of the surface wind field shows a strengthened divergence pattern over the Australian monsoon region (Fig. 3a, vector), which is consistent with the strengthened descending flow over the Australian monsoon region (Fig. 4) and thus reduced precipitation (Fig. 3a, shading). The JJA mean near surface air temperature shows that the land is cooler than the adjacent ocean around northern Australia (Fig. 5a), which illustrates a strengthened land–sea thermal contrast because the land cools more than the ocean surface during the LGM. This strengthened land–sea contrast leads to a higher sea-level pressure (SLP) over land and lower SLP over ocean in general (Fig. 5b, shading), and thus the outflows from land (Fig. 5b, vector). The geopotential height at 850 hPa also shows the relative pattern that matches the wind change (Fig. S4a). The difference of divergence/convergence field (Fig. 5c) also indicates that the divergence at 850 hPa over the northern Australia is strengthened during the LGM. The vertical velocity at 500 hPa over the central Australian monsoon region (20°S-5°S, 120°E-145°E) illustrates that the descending flow strengthened by about 48 %.

In conclusion, both the dynamic process (increased subsidence) and the thermodynamic process (reduced water vapor content) contribute to the drier winter in the Australian monsoon region, but the local dynamic processes play a dominant role in the reduction of Australian winter precipitation.

3.2 Why the precipitation increased in austral early summer (ND)

During ND, the LGM minus PI surface wind difference field shows a strengthened convergence pattern over the central northern Australian monsoon region (Fig. 6a, vector), which is consistent with the increased precipitation (Fig. 6a, shading). The vertical velocity at 500 hPa also shows a strengthened ascending flow over this area (Fig. 7). The increased precipitation over the central Australian monsoon region is clearly against the thermodynamic effects of the low GHG concentration and the presence of the ice-sheets, which tends to reduce the precipitation. The 2-m air temperature was decreased by 2.2 K and the surface specific humidity reduced by 2.6 g/kg (or 16.0 %) over the Australian monsoon region (Fig. 8). The precipitation would decrease by 15.4 % according to the thermal effect without the circulation change. However, the precipitation over the Australian monsoon region increased by about 13.0 %. Therefore, the changes in dynamic processes must induce a 29 % increase of precipitation, so that the net increase in precipitation reaches 13 %.

There is a cyclonic wind anomaly associated with an anomalous low pressure over the northwest Australia (Fig. 6a and Fig. 9b, vector), accompanied by a strengthened low-level convergence (Fig. 9c), which favors increased precipitation in the Australian monsoon region. The change of the moisture transport (moisture flux) also indicated increased moisture transport into northern Australia (not shown). The cyclonic vorticity in northwest Australia is partially caused by the enhanced strong low-level westerlies that prevail north of Australia.

We now seek to determine why there was a strengthened low-level westerly with maximum over north of Australia. We first consider the temperature change. The ND mean 2-m air temperature during the LGM shows that the two enlarged landmasses over the Indo-Pacific warm pool region (resulting from the lower sea level) change differently (Fig. 9a). It is cooler over the northwest landmass (western Indonesia–Indochina) and relatively warmer over the southeast landmass (eastern Indonesia–northern Australia). This temperature variation forms a southeast–northwest temperature gradient (Fig. 9a, Fig. S5a, S5b), accompanied by a northwest–southeast SLP gradient (Fig. 9b, Fig. S5c, S5d). The northwest–southeast pressure gradient is stronger in the geopotential height change at 850 hPa (Fig. S4b). The high pressure in the western Indonesia–Indochina is a part of the larger scale enhanced winter monsoon over the South China Sea. This enhanced winter monsoon flows cross the equator from the NH to the SH and turn into strong westerlies due to deflection induced by the Coriolis force. The 850 hPa convergence strengthens over the Australian monsoon region (Fig. 9c), and the corresponding ascending motion at 500 hPa also increases over the Australian monsoon region.

Another reason for circulation change is the sea surface temperature (SST) gradient change. The SST anomaly in ND shows a warmer Western Pacific and cooler Eastern Indian Ocean pattern (Fig. 10), indicating a westward temperature gradient (Fig. S5e), and thus an eastward pressure gradient which, in the equatorial region, can directly enhance westerly winds near the northern Australian monsoon region (Fig. 9b). Li et al. (2012) also found that a cold state of the Wharton Basin (100°E–130°E, 20°S–5°S) was accompanied by anomalous westerlies and cyclonic circulation anomalies in the Australian monsoon region, which were associated with a strong tropical Australian summer monsoon and enhanced rainfall over northeast Australia.

In summary, during ND, the enlarged land area due to sea-level drop enhances the land–sea thermal contrast, and forms a northwest–southeast thermal contrast which induces low pressure over northern Australia but high pressure over the adjacent ocean and the Indochina–western Indonesia, leading to enhanced convergence over northern Australia and thus increasing the early summer monsoon rainfall. The SST gradients between the warm equatorial western Pacific and relatively cool eastern Indian Ocean during the pre-summer monsoon season also contribute to the strengthened equatorial westerlies and the cyclonic wind anomaly over northern Australia. These dynamic mechanisms have a positive contribution to the early summer precipitation. The thermal effects have negative contribution to the precipitation change, but with smaller impact. Therefore, the early summer precipitation over northern Australia increases. We can also tell from the changes of the decomposed terms that the dynamics plays much more important role in the precipitation change over Australia, especially the distribution pattern (Fig. S6). Again, the impacts of evaporation and advection terms are small.

4 Discussion

The intensification of the Australian monsoon in this study is measured by the enhanced seasonal difference (or the seasonality) of precipitation, and is particularly attributed to the decreased austral winter precipitation. This is consistent with the reconstructed results by Mohtadi et al. (2011), which indicated that it was not significantly drier in austral summer during the LGM, while (the winter monsoon is weakened). Whereas the annual mean precipitation is decreased, which means the Australian monsoon would be weakened during the LGM when it is measured by the annual mean precipitation. The modeling study by DiNezio et al. (2013) suggests a decreasing rainfall across northern Australia during the LGM, consistent with the proxy synthesis by stalagmite (Denniston et al., 2017). The decreased rainfall in their work represents the annual mean precipitation, which also consists with our work in this point of view. On the other hand, the increased rainfall in austral summer in this study is consistent with what has been revealed in the reconstructed work by Liu et al. (2015). The decreased annual mean precipitation and the intensified seasonality of precipitation over the Australian monsoon region are in agreement with the synthesis from the simulated result by Tharammal et al. (2013) using a set of experiments.

For the forcings and mechanisms of the Australian monsoon change during the LGM, there are large changes in four external forcings during the LGM, including the insolation change resulting from the orbital change, the land–sea configuration change, the GHG change and the presence of ice-sheets. The lower GHG concentrations and the presence of ice-sheets are likely to be contributors to the thermal effect leading to the reduced water vapor and thus the decreased rainfall both in austral winter and early summer. The enlarged the landmasses over western Indonesia and northeastern Australia are essential to the local dynamic processes that influence the rainfall. The low obliquity and high precession during the LGM may be another factor that can affect the rainfall (Liu et al., 2015). However, the impact of the insolation change caused by the orbital change remains unknown.

During the LGM, the insolation over tropical region increased from December to June and decreased from July to November (Fig. 11a). In the annual variation, precipitation responds to the lower tropospheric moisture convergence. The moisture change depends on temperature change while the circulation change depends on surface temperature gradients change. The change of the surface temperature lags insolation changes because of the ocean and land surfaces have heat capacity (thermal inertial). In other words, insolation is a heating rate which equals to temperature change (tendency) but not the temperature itself. Thus the precipitation change would lag the insolation change by about two months, due to the ocean–atmosphere interaction without other processes. For example, the change of seasonal distribution of NH monsoon precipitation lagged the change of the NH insolation by one month (Yan et al., 2016). Whereas in this study, the Australian monsoon precipitation decreased from March to September and increased from November to February (Fig. 11b), quite different from what it would be (i.e., decrease from September to January and increase from February to August). Meanwhile, the insolation over SH increased during the LGM from April to August, when Australia is in late autumn and winter. An increased insolation might make land warmer than ocean thus against the climatology, which may be described by cooler land and warmer ocean in winter. However, the simulated surface temperature reduced more over Australia than the adjacent oceans (Fig. 5a). On the other hand, the synthesis of Wyrwoll et al. (2007) and Liu et al. (2015) indicates the strong convergence rain belt (ITCZ) stays in the north, resulting in more rainfall over Papua New Guinea and less rainfall over North Australia during those times with low obliquity and high precession. The rain belt stays a little more northerly than that stays in our study, which means the effect of orbital change and thus the insolation change might be suppressed by other factors.

Moreover, the paleoclimate records suggest that it was dry and cool in the Indo-Pacific Warm Pool region during the LGM (Xu et al., 2010). The simulated SST is consistent with the reconstructions. Although in the early austral summer, over the Indian Ocean warm pool, it is cooler over the SH, while over the Pacific warm pool, it is cooler over the NH (Fig. S7). Such anomalous SST asymmetry may favor the southward shift of the ITCZ over Australia and the southwest Pacific, which might be related to the enhanced austral summer monsoon precipitation. However, the 7MME shows no significant ITCZ shift during the LGM, particularly over the central Australian monsoon region (Fig. S8). The reconstructions and simulations by McGee et al. (2014) and Mohtadi et al. (2014) also suggested that there was no significant shift of ITCZ position during the LGM.

Therefore, it is the local dynamical processes, instead of the large-scale circulation such as the position of the ITCZ induced by the NH-SH thermal contrast, that might be the key factor influencing the early summer mean precipitation change over the Australian monsoon region during the LGM. To test this synthesis, we conducted two experiments using a newly developed earth system model, the Nanjing University of Information Science and Technology Earth System model version 1 (NESM v1, Cao et al., 2015). The PI control run is designed the same as PMIP3 protocol. The land sea configuration sensitive run (LSM) is the same as the piControl but with the LGM land sea configuration. The two experiments are run 500 years after spin-up, and the last 100 years are used. The changes of the ND mean precipitation and wind field at 1000 hPa between LSM and piControl are similar to the changes derived from the 7MME, i.e. the precipitation is increased and the convergence is strengthened over northern Australia (Fig. 12a). The changes of the 2-m air temperature, SLP and 850 hPa wind field (Fig. 12b, c) are also similar to those results in the 7MME (Fig. 9). It is also cooler over the northwest landmass (western Indonesia–Indochina) and relatively warmer over the southeast landmass (eastern Indonesia–northern Australia) (Fig. 12b). This temperature variation is also accompanied by a northwest–southeast SLP gradient and the strengthened cross equatorial flow converging to north

Australia (Fig. 12c). This sensitive simulation confirms that the local dynamical process induced by the land sea configuration to be essential to the Australian monsoonality change.

Although the 7MME simulates strengthened Australian monsoonality, there are uncertainties among individual models. The most notable uncertainty is the increased austral summer (DJF) precipitation. Five out of the 7 models simulate increased DJF mean precipitation over the Australian monsoon region during the LGM (CCSM4, GISS-E2-R, IPSL-CM5A-LR, MIROC-ESM and MRI-CGCM3), while the other two simulate decreased precipitation (CNRM-CM5 and MPI-ESM-P) (Fig. 13), especially over the land area. The wind filed at 850hPa geopotential height shows a cyclonic anomaly pattern over northern Australia in the five models (Fig. 14a), accompanied with a strengthened ascending flow (not shown). While in the other two models, there is no cyclonic wind anomaly over Australia region (Fig. 14b), and the ascending flow is weakened (not shown). The different changes of wind field indicate the different precipitation responses to the LGM boundary conditions in the two model groups.

The austral spring and summer mean 2m-air temperature and SST also change differently in these two model groups. The main differences are located over the tropic Pacific Ocean and the North Atlantic Ocean. It is cooler over high-latitude Northern Atlantic Ocean in the five models, whereas warmer in the two models, mainly in the austral spring (Fig. 15a, 15b). In the austral summer, there is an East-Pacific El Nino like patter in the five models, while there is a Central-Pacific El Nino (CP-El Nino) like pattern in the two models (Fig. 15c, 15d). Studies have shown that the CP-El Nino is related to the Asian-Australian monsoon system (Yu et al. 2009), and would lead to a markedly decreased precipitation in December (Taschetto et al. 2009).

Therefore, the different SST response over Pacific Oceans and North Atlantic Ocean in austral spring and summer in different models might be the key factor that leads to different wind anomalies and thus different Australian monsoon precipitation changes.

5 Conclusions

The global mean temperature and water vapor have an overall decrease under the LGM forcings (lower GHG and large ice-sheets). Nevertheless, the simulated Australian monsoon seasonality derived from CMIP5/PMIP3 multi-model ensemble has a distinctive amplification (or the monsoonality is intensified) against the weakened global monsoons elsewhere during the

LGM. This study then investigated the possible reasons for this strengthened Australian monsoonality from both a thermodynamic and dynamic perspective.

The conclusions are as follows:

1) The Australian monsoon seasonality is strengthened as a result of the enhanced seasonal difference between austral summer and winter, i.e., the increased early summer (ND) mean rainfall and the reduced winter (JJA) mean rainfall. Both the dynamic processes and thermal effects contribute to the precipitation change; however, the dynamic processes have a much stronger contribution than the thermal effects.

2) The Australian winter (JJA mean) precipitation derived from 7MME decreased during the LGM relative to the preindustrial control experiment. The dynamic processes, induced by the enhanced land–ocean thermal contrast, contribute more to the decreased rainfall through the strengthened divergence over northern Australia (Fig. 16a), whereas the thermodynamic effect (i.e., the reduced atmospheric water vapor due to the lower temperature induced by lower GHGs and present ice-sheets) and evaporation have moderate contribution.

3) For the increased precipitation in early summer (ND) in the 7MME, the local dynamic processes have a positive contribution and the thermodynamic effect has a negative contribution. Both the decomposition method and the sensitive simulations show that the dynamic effect plays most important role for the increased rainfall. The local dynamic processes are mainly induced by the northwest–southeast thermal contrast between Indochina–western Indonesia and northeastern Australia. The east Indian Ocean–west Pacific Ocean thermal gradient also contributes to these processes (Fig. 16b).

4) The sensitive simulation illustrates that the change in circulation over Australia is very likely to be rooted in the enlarged landmasses over the Indochina–western Indonesia and New Guinea, and northern Australia. Another factor contributes to the circulation change might be the asymmetric change between western Pacific Ocean and eastern Indian Ocean. These have critical impacts on the thermal gradients that induce changes in the low-level circulation pattern and convergence/divergence.

Note that models have uncertainties, i.e. not all the models simulate an intensified seasonality of Australian monsoon. The different SST responses over Pacific Ocean and Atlantic Ocean in different models to the same external forcings are essential for the model uncertainties. More model-data comparison and inter-model comparison are required to improve model performance.

Our results are based on the equilibrium simulation, representing a mean state of the Australian monsoon change and its possible mechanisms during the LGM. More simulations with single forcing (such as the SST asymmetry change, the insolation change) are required to further understand the effect of each factor and to specifically quantify the contribution of each forcing to the Australian monsoon change.

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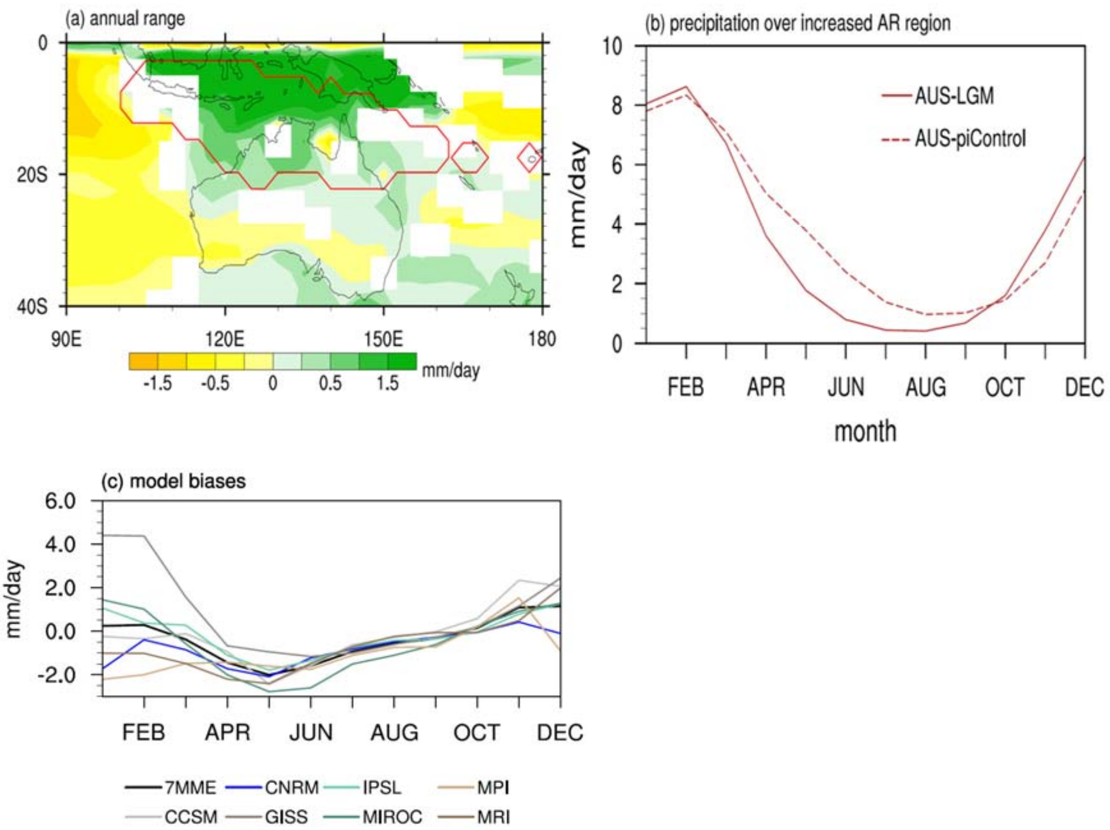

Figure 1 (**a**) Spatial distribution of changes in the annual range (AR) of precipitation measured by the difference between LGME and piControl, (**b**) seasonal distribution of the precipitation in the increased AR area (20°S-5°S, 120°E-145°E), and (c) seasonal distribution of the precipitation differences in the increased AR area derived from 7 MME (black line) and each model (colored lines). The red solid line in (**a**) encloses the Australian monsoon rainfall domain. The dashed (solid) line in (**b**) denotes the seasonal distribution of precipitation derived from the piControl (LGME) run. Only those areas where signal-to-noise ratio exceeds one are plotted in (**a**).

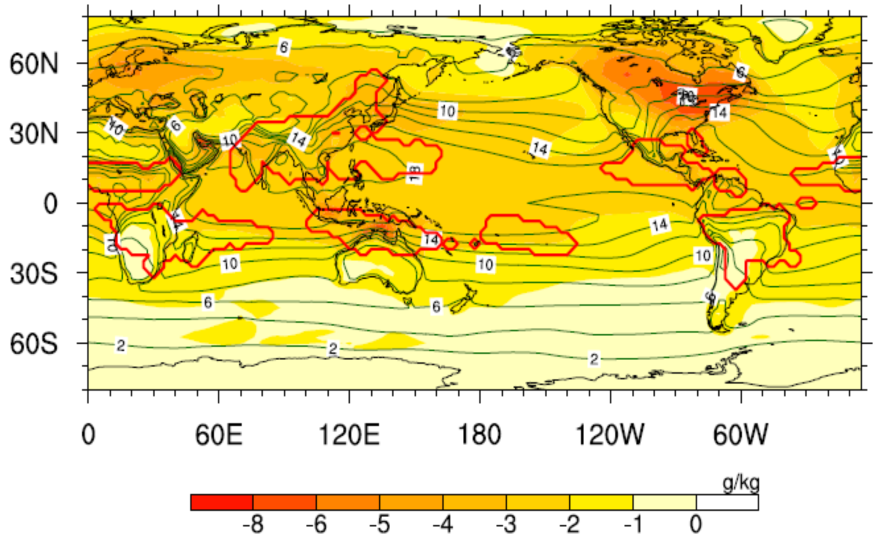

Figure 2 Difference of JJA mean surface specific humidity between LGME and piControl (shaded). The green contours denote the climatology derived from piControl. The red lines enclose the monsoon domains. Only those areas where signal-to-noise ratio exceeds one are plotted.

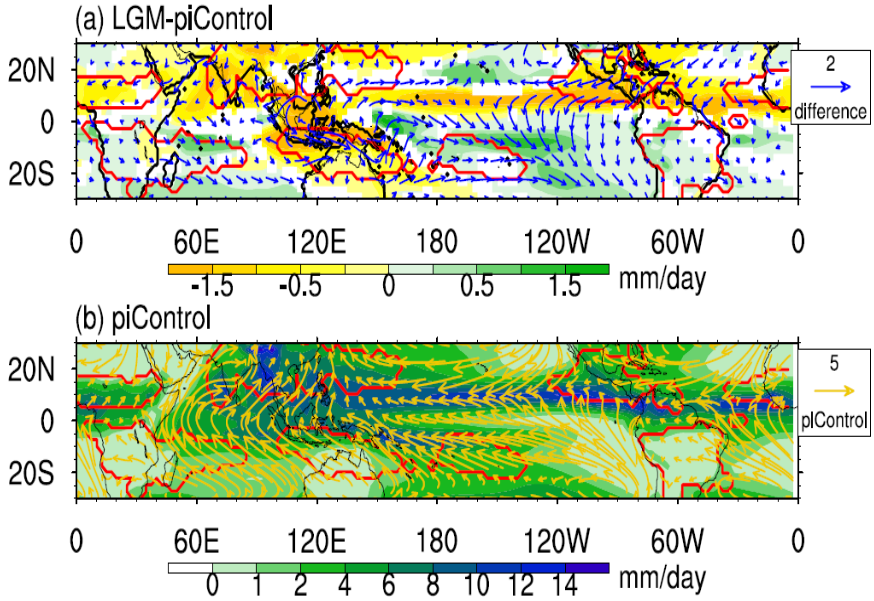

Figure 3 (**a**) JJA mean precipitation (shading) difference and surface wind (vectors) difference between LGME and piControl, and (**b**) the climatology of JJA mean precipitation (shading) and surface wind (vectors) derived from piControl. The red lines enclose the monsoon domains. The thick black lines in (**a**) denote the coastal lines in LGME, and the thin black lines denote the coastal lines in piControl. Only those areas where signal-to-noise ratio exceeds one are plotted in (**a**).

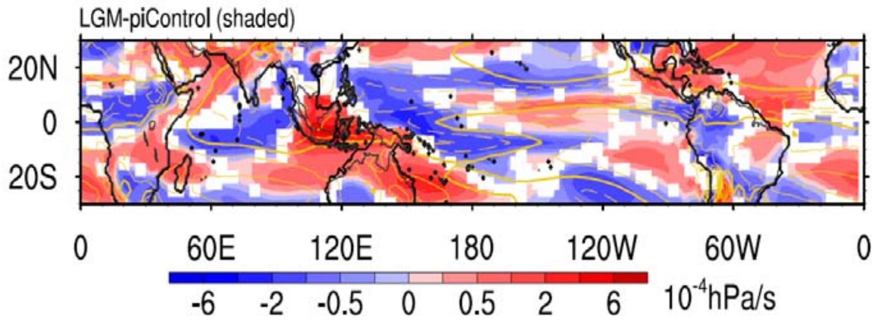

Figure 4 The difference of JJA mean vertical velocity at 500 hPa between LGME and piControl (in shading) and the corresponding climatology derived from piControl (yellow contours). The thick black lines denote the coastal lines in LGME, and the thin black lines denote the coastal lines in piControl. Only those areas where signal-to-noise ratio exceeds one are plotted in the difference pattern.

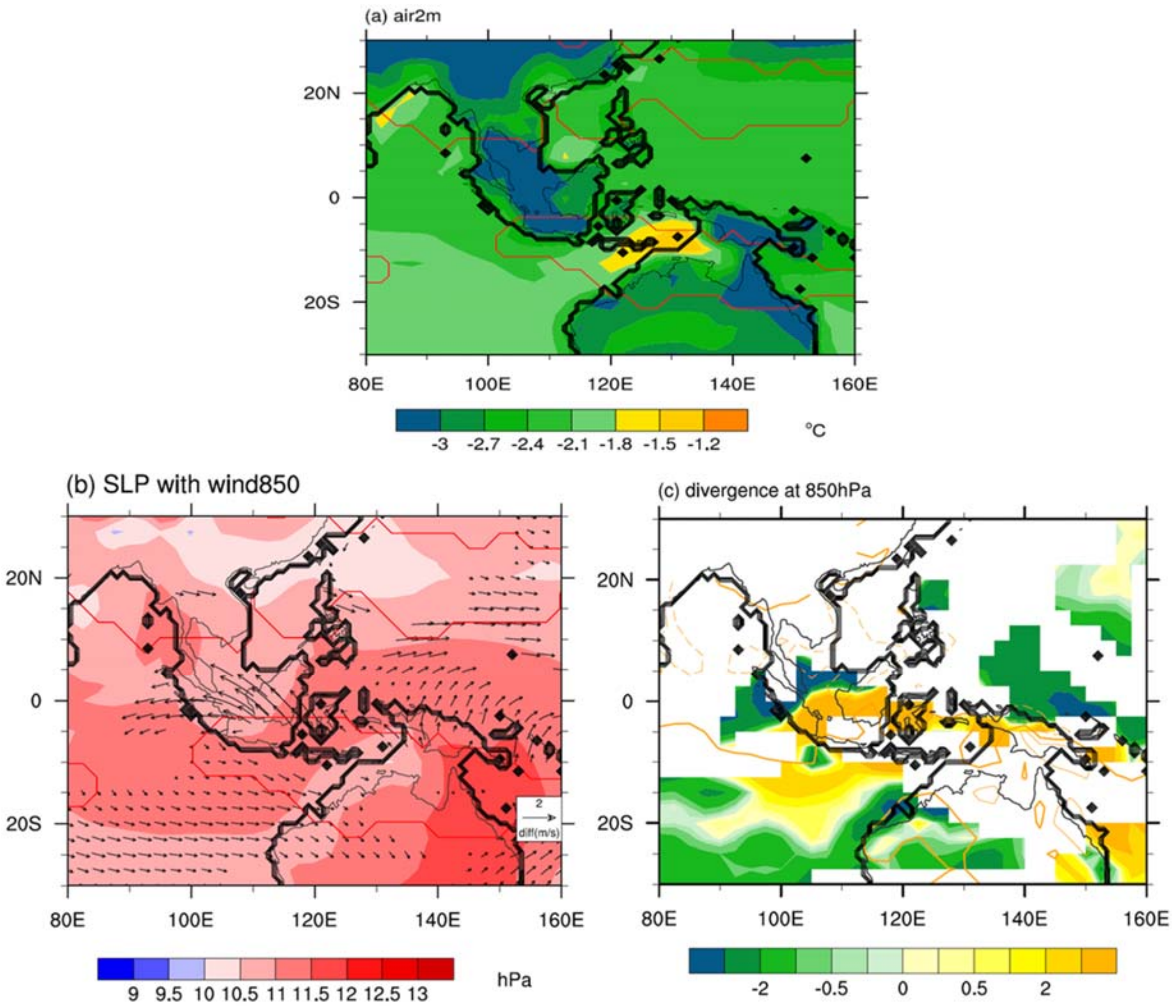

Figure 5 JJA mean (**a**) surface air temperature, (**b**) sea level pressure (shading) with 850 hPa wind (vector), and (**c**) 850 hPa divergence differences between LGME and piControl. The red lines in (**a**) and (**b**) enclose the monsoon domains. The orange lines in (**c**) represent the climatology derived from piControl. The thick black lines denote the coastal lines in LGME, and the thin black lines denote the coastal lines in piControl. Only those areas where signal-to-noise ratio exceeds one are plotted.

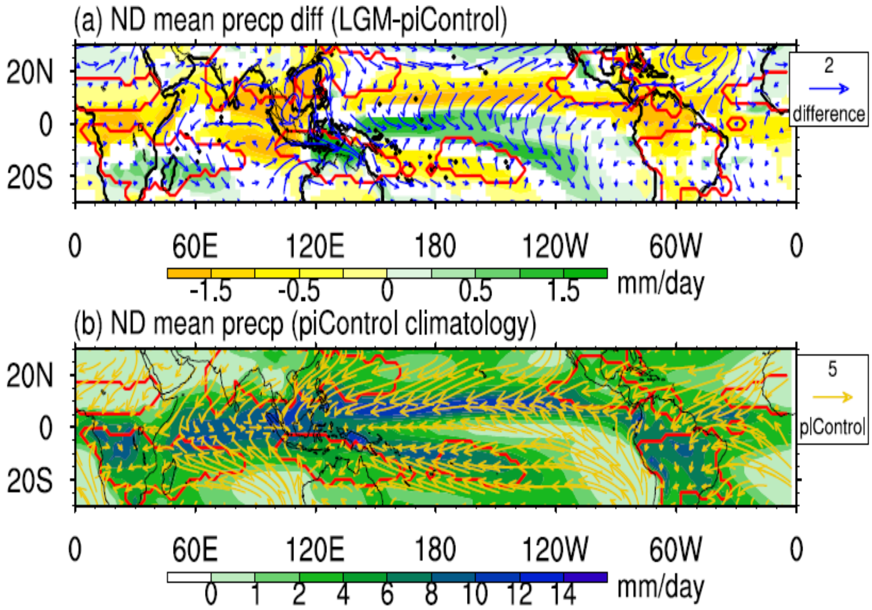

Figure 6 (**a**) ND mean precipitation (shading) difference and surface wind (vectors) difference between LGME and piControl, and (**b**) the climatology of ND mean precipitation (shading) and surface wind (vectors) derived from piControl. The red lines enclose the monsoon domains. The thick black lines in (**a**) denote the coastal lines in LGME, and the thin black lines denote the coastal lines in piControl. Only those areas where signal-to-noise ratio exceeds one are plotted in (**a**).

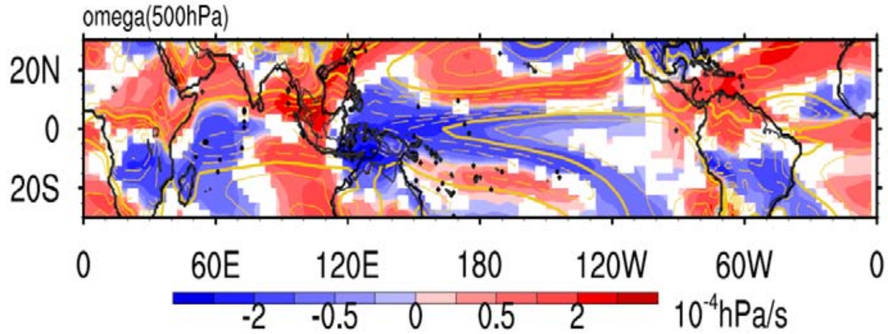

Figure 7 The difference of the ND mean vertical velocity at 500 hPa between LGME and piControl (in shading) and the corresponding climatology derived from piControl (yellow contours). The thick black lines denote the coastal lines in LGME, and the thin black lines denote the coastal lines in piControl. Only those areas where signal-to-noise ratio exceeds one are plotted in the difference pattern.

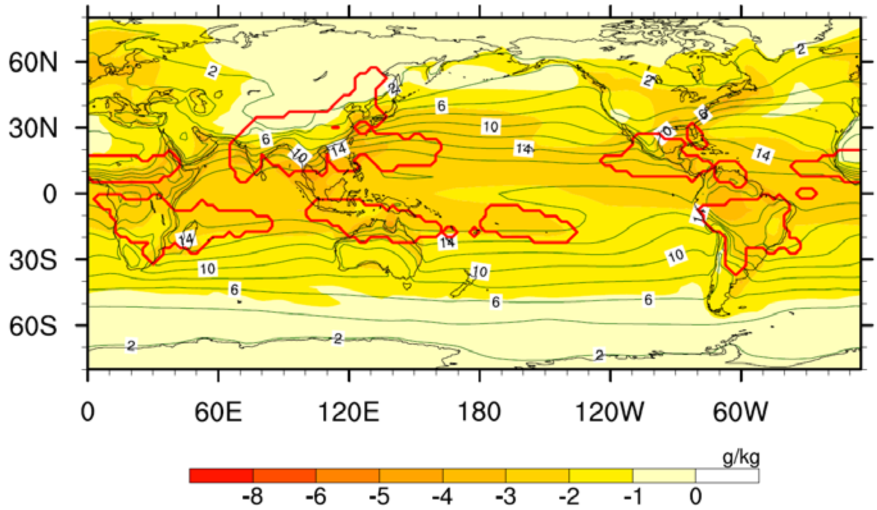

Figure 8 Difference of ND mean surface specific humidity between LGME and piControl (shaded). The green contours denote the climatology derived from piControl. The red lines enclose the monsoon domains. Only those areas where signal-to-noise ratio exceeds one are plotted.

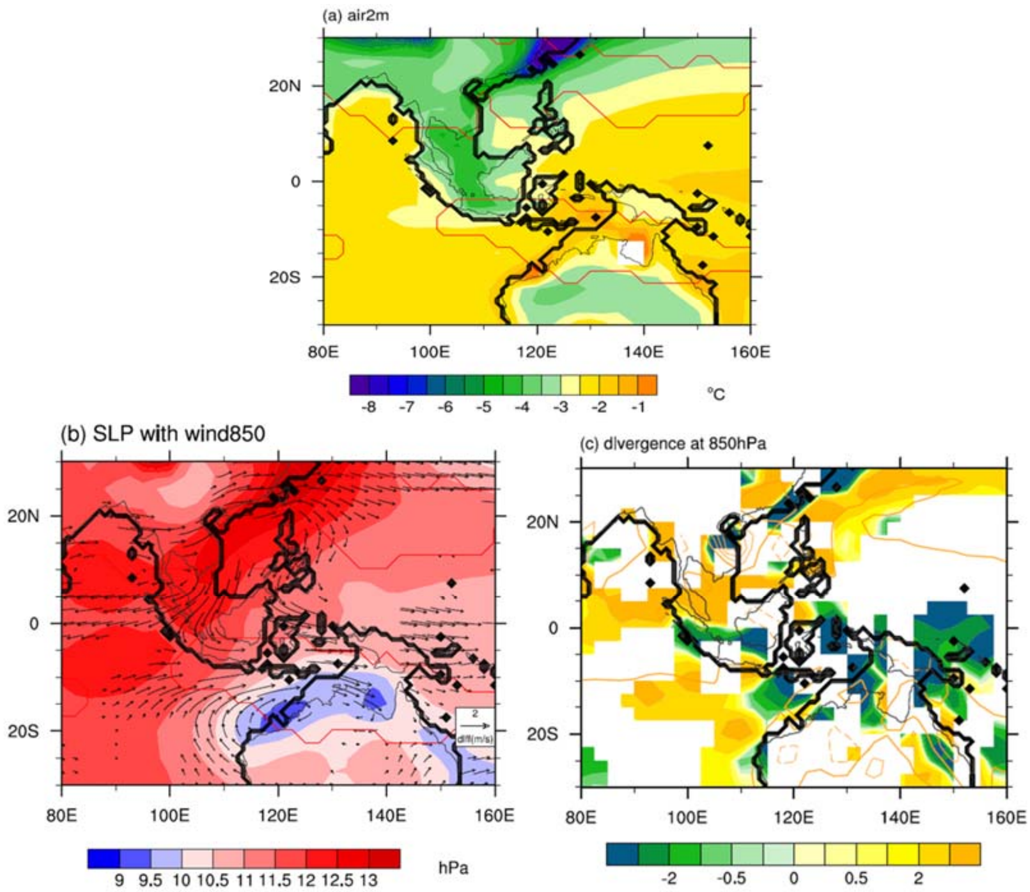

Figure 9 ND mean (**a**) surface air temperature, (**b**) sea level pressure (shading) with 850 hPa wind (vector), and (**c**) 850 hPa divergence difference between LGME and piControl (shading). The red lines in (**a**) and (**b**) enclose the monsoon domains. The orange lines in (**c**) represents the climatology derived from piControl. The thick black lines denote the coastal lines in LGME, and the thin black lines denote the coastal lines in piControl. Only those areas where signal-to-noise ratio exceeds one are plotted.

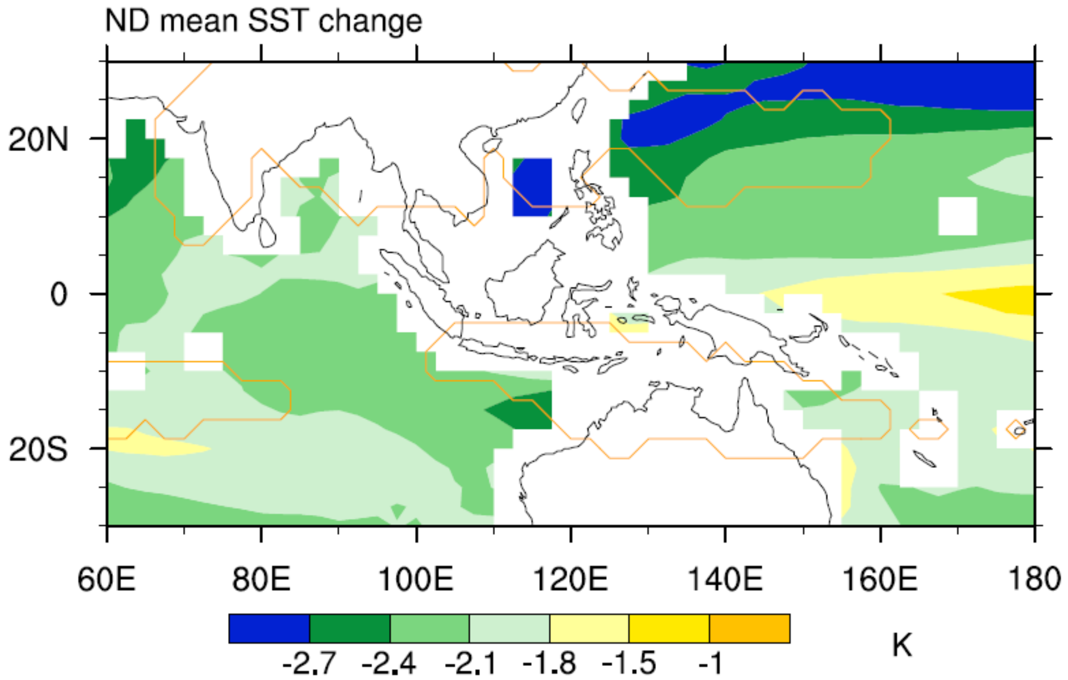

Figure 10 ND mean SST difference between LGME and piControl. The red lines enclose the monsoon domains. Only those areas where signal-to-noise ratio exceeds one are plotted.

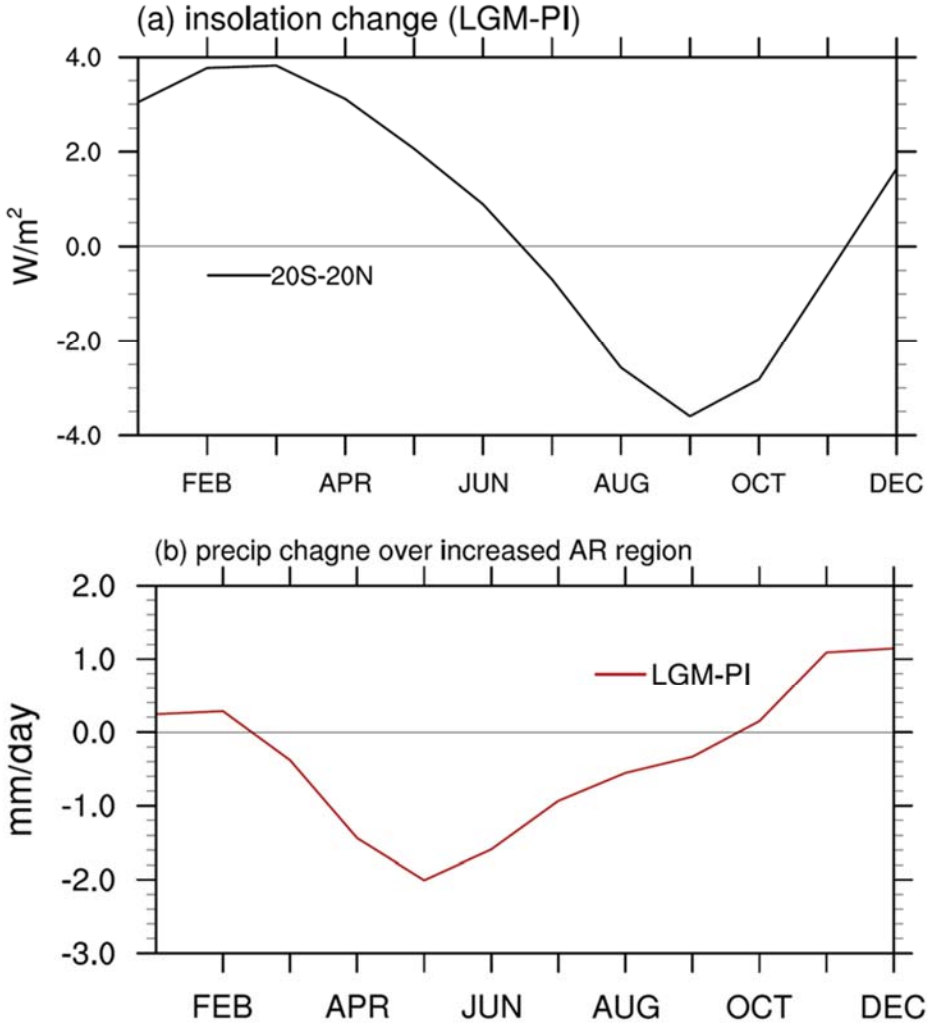

Figure 11 Seasonal distribution of (**a**) insolation change between 20°S and 20°N, and (**b**) precipitation change over the increased AR region as indicated in Fig. 1b (20°S-5°S, 120°E-145°E). The changes are calculated by the LGM value minus the PI value.

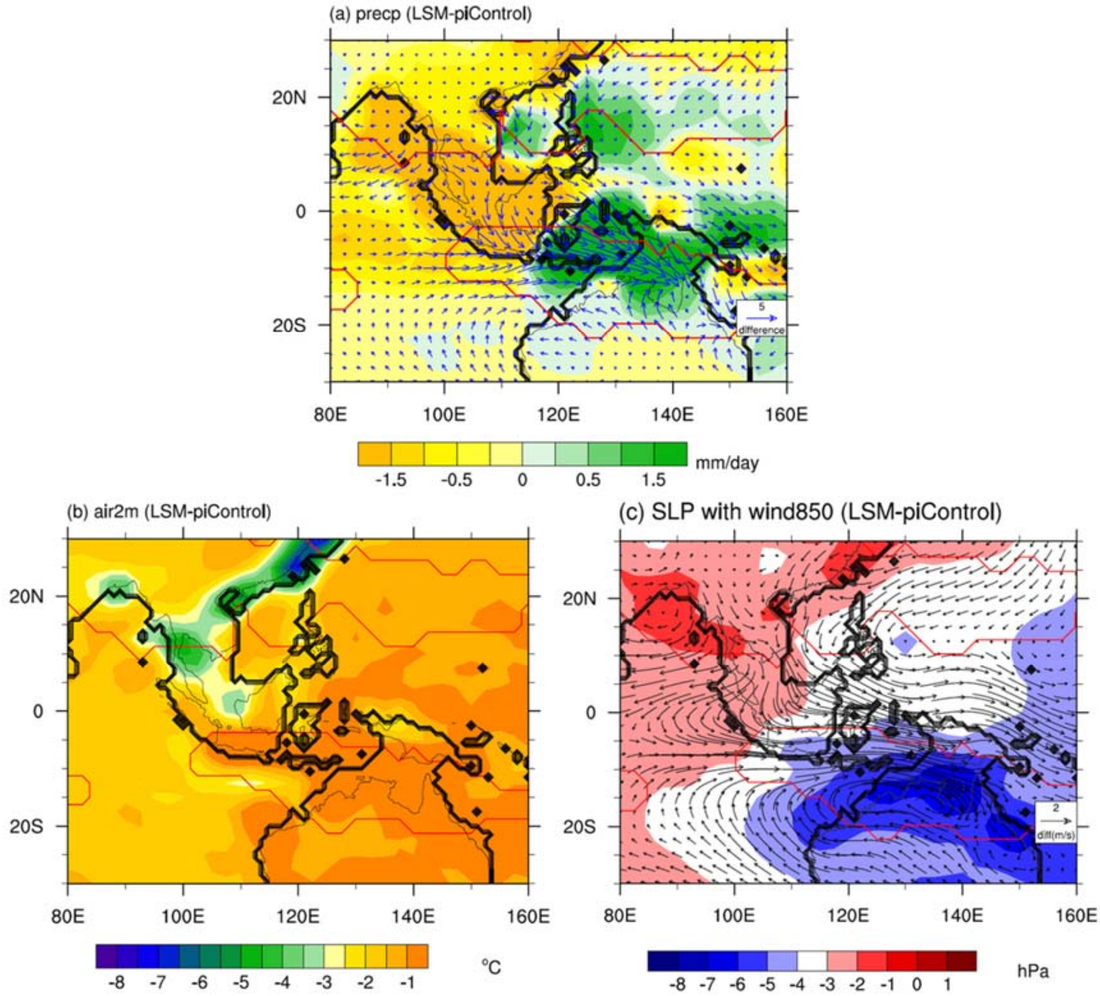

Figure 12 ND mean (a) precipitation (shading) with 1000 hPa wind (vector), (b) surface air temperature, and (c) sea level pressure (shading) with 850 hPa wind (vector) difference between land sea configuration experiment (LSM) and piControl. The red lines enclose the monsoon domains. The thick black lines denote the coastal lines in LSM, and the thin black lines denote the coastal lines in piControl.

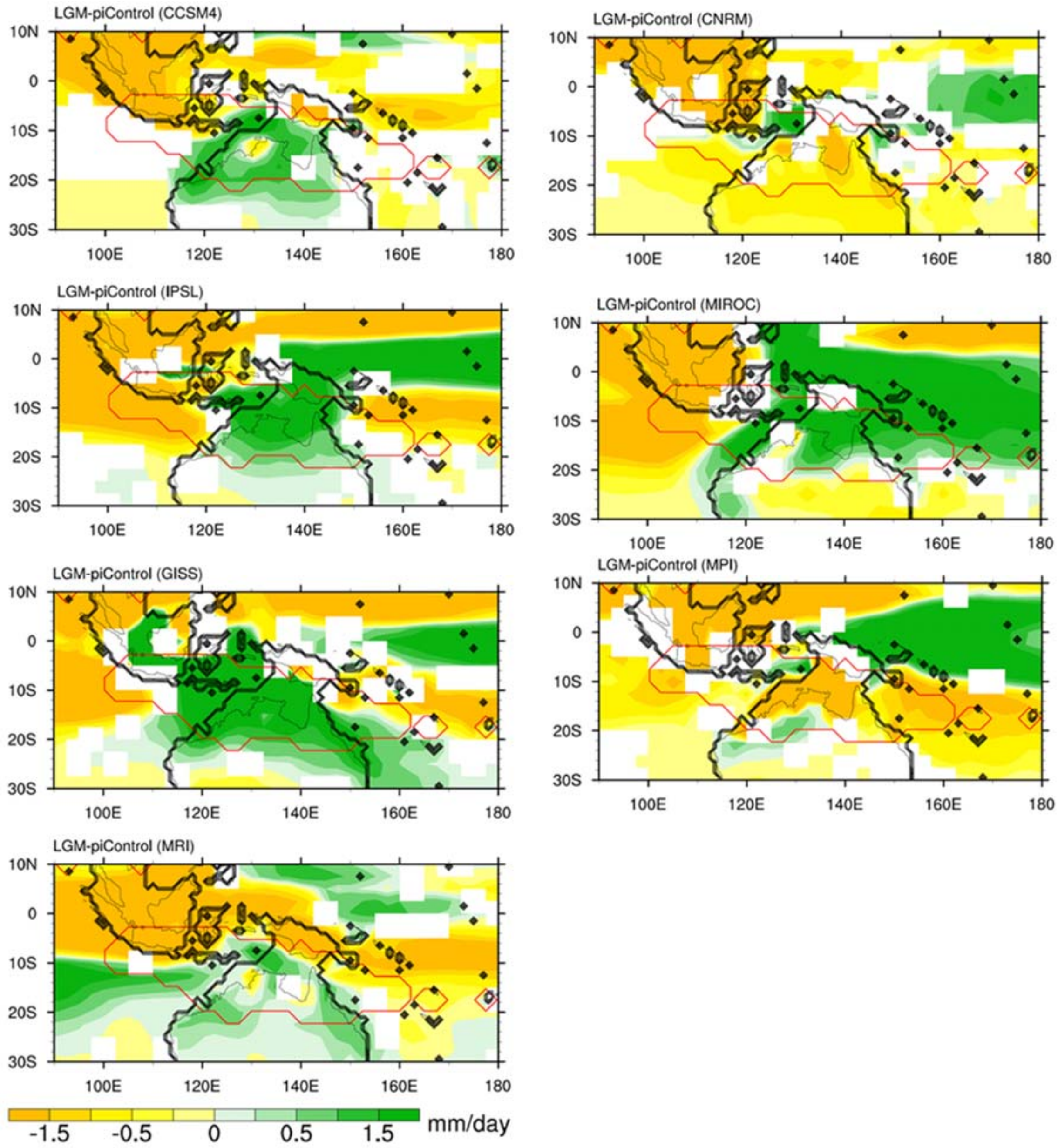

Figure 13 DJF mean precipitation difference between LGME and piControl derived from each model. The red lines enclose the monsoon domains. Only those areas where signal-to-noise ratio exceeds one are plotted.

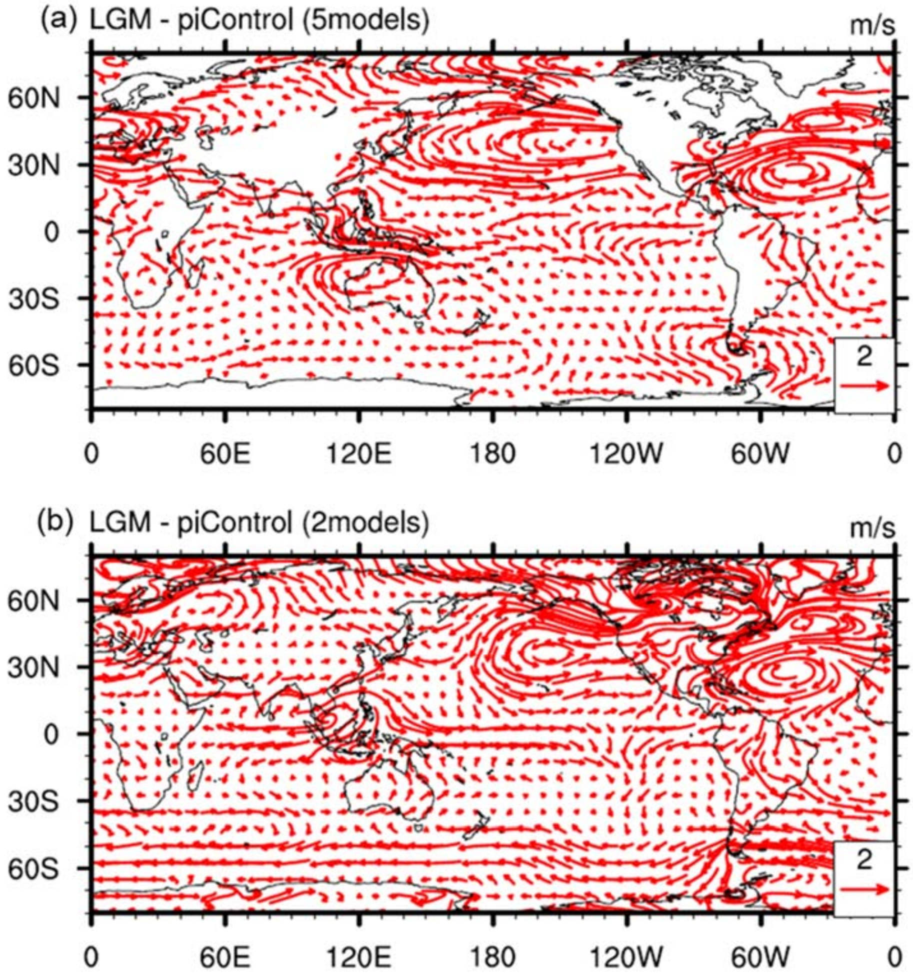

Figure 14 DJF mean 850hPa wind differences between LGME and piControl derived from (a) the five models and (b) the two models. Only those areas where signal-to-noise ratio exceeds one are plotted.

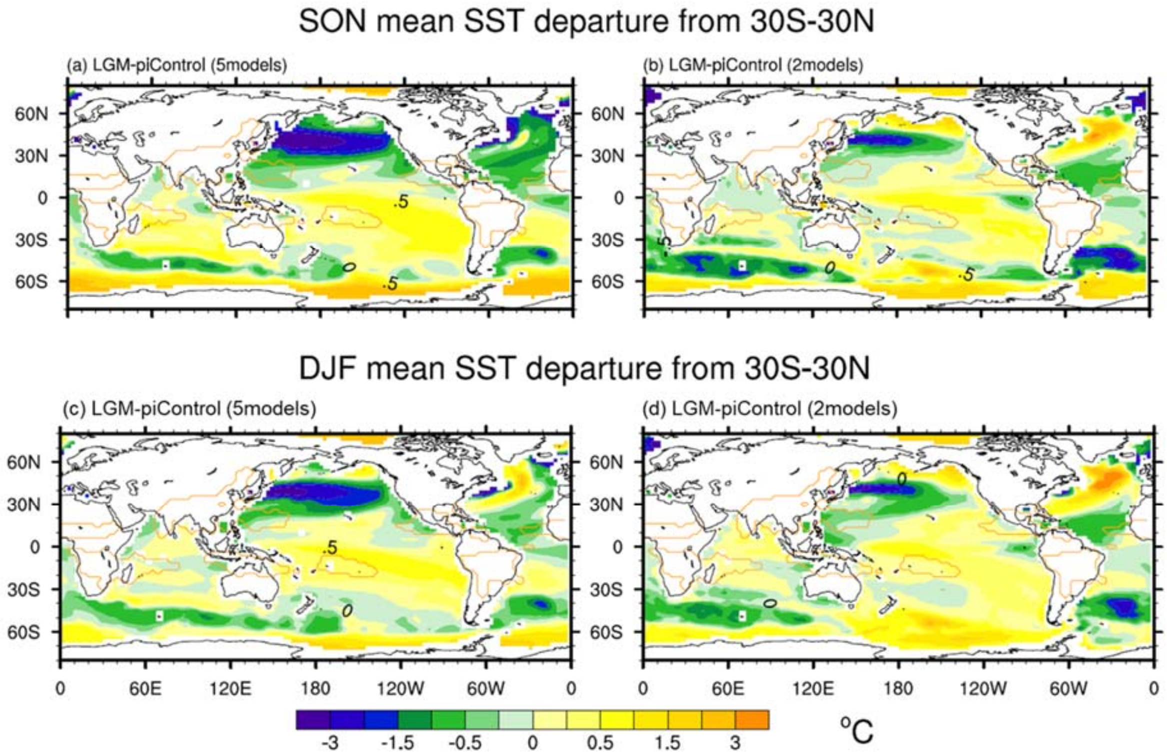

Figure 15 SON mean (a)-(b) and DJF mean (c)-(d) SST differences between LGME and piControl derived from (a), (c) the five models and (b), (d) the two models. Only those areas where signal-to-noise ratio exceeds one are plotted. The area average of tropical (30°S-30°N) SST change is distracted to make it clearer to illustrate the regional differences.

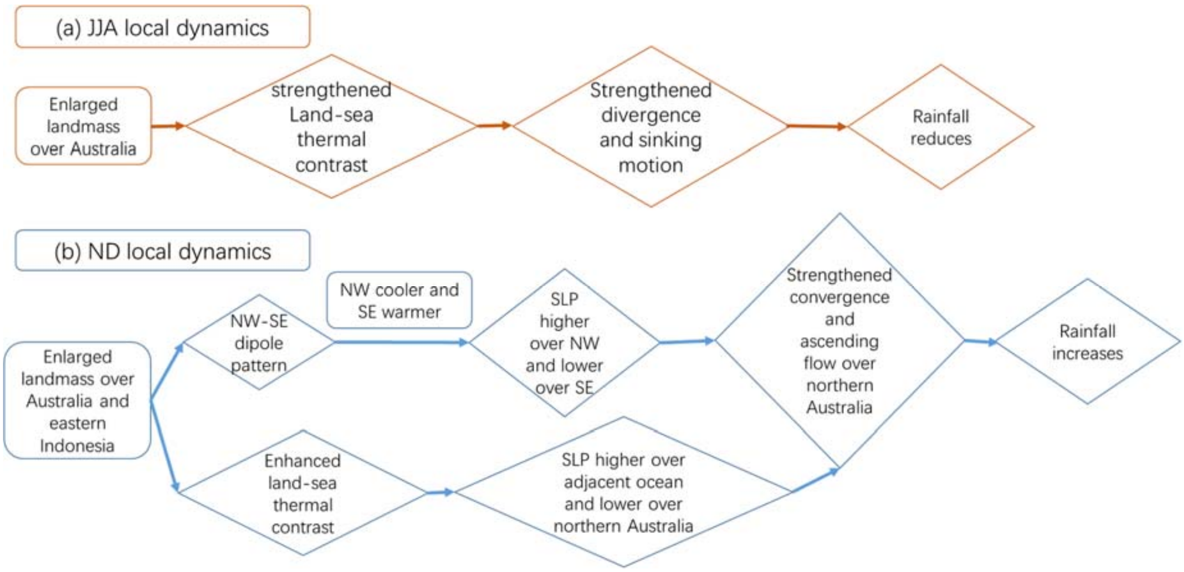

Figure 12 Mechanisms of Australian monsoon precipitation change (**a**) in JJA, and (**b**) in ND during the LGM, in the local dynamics perspective.

Table 1 CMIP5/PMIP3 models and experiments used in this study.

Model	Institution	piControl Time span (years)	LGME Time span (years)	Spatial resolution for atmospheric module Lon × Lat Grids	Spatial resolution for oceanic module Lon × Lat Grids
CCSM4	National Centre for Atmospheric Research (NCAR)	501	101	288 × 192	320×384
CNRM-CM5	Centre National de Recherches Meteorologiques/Centre Europeen de Recherche et Formation Avancees en Calcul Scientifique (CNRM-CERFACS)	850	200	256 × 128	362×292
GISS-E2-R	NASA Goddard Institute for Space Studies (NASA GISS)	1200	100	144 × 90	288×180
IPSL-CM5A-LR	Institute Pierre-Simon Laplace (IPSL)	1000	200	96 × 95	182×149
MIROC-ESM	Atmosphere and Ocean Research Institute, University of Tokyo, National Institute for Environmental studies, and Japan Agency for Marine-Earth Science and Technology	531	100	128 × 64	256×192
MPI-ESM-P	Max Planck Institute for Meteorology	1156	100	196 × 98	256×220
MRI-CGCM3	Meteorological Research Institute (MRI)	500	100	320 × 160	364×368

Table 2 Main changed boundary conditions used for the piControl and LGME experiments.

	piControl	**LGME**
Orbital parameters	Eccentricity = 0.016724	Eccentricity = 0.018994
	Obliquity = 23.446°	Obliquity = 22.949°
	Angular precession = 102.04°	Angular precession = 114.42°
Trace gases	CO_2 = 280 ppm	CO_2 = 185 ppm
	CH_4 = 650 ppb	CH_4 = 350 ppb
	N_2O = 270 ppb	N_2O = 200 ppb
Ice sheets	Modern	Provided by ICE-6G v2 (Peltier, 2009)
Land surface elevation and coastlines	Modern	Provided by PMIP3

Table 3 Annual mean, austral summer (DJF) mean and annual range of precipitation change over the region of (20°S-5°S, 120°E-145°E). The area averaged value is calculated based on the areas where S2N ratio exceed one.

Model	**Annual mean (mm/day)**	**Summer mean (mm/day)**	**Annual range (mm/day)**
CCSM4	-0.14	0.49	1.36
CNRM-CM5	-0.78	-0.74	0.12
GISS-E2-R	0.79	3.74	4.66
IPSL-CM5A-LR	-0.17	0.90	1.82
MIROC-ESM	-0.53	1.25	3.17
MPI-ESM-P	-1.02	-1.71	-0.52
MRI-CGCM3	-0.68	-0.01	0.85
7MME	-0.36	0.56	1.61