Land-ocean asymmetry of tropical precipitation changes in the mid-Holocene

Yang-Hui Hsu

Research Center for Environmental Changes, Academia Sinica, Taipei, Taiwan

*Chia Chou

*Research Center for Environmental Changes, Academia Sinica, and Department of Atmospheric
Sciences, National Taiwan University, Taipei, Taiwan*

Kuo-Yen Wei

Department of Geosciences, National Taiwan University, Taipei, Taiwan

*Corresponding author information:

Address: Chia Chou, Research Center for Environmental Changes, Academia Sinica, P.O. Box 1-48,
Taipei 11529, Taiwan, ROC.

E-mail: chiachou@rcec.sinica.edu.tw

Abstract

A series of model experiments were conducted using an intermediate ocean-atmosphere-land model for a better understanding of a distinct land-sea asymmetry in tropical precipitation differences between the mid-Holocene and pre-industrial climates. In austral (boreal) summer, most reduced (enhanced) precipitation occurs over continental convective regions, while most enhanced (reduced) precipitation occurs over oceanic convection zones. This land-sea asymmetry of tropical precipitation is particularly clear in austral summer. During the mid-Holocene, the solar forcing presents both spatial and seasonal asymmetric patterns. While the boreal summer insolation is stronger at high latitudes of the northern hemisphere in the mid-Holocene than at present, the austral summer insolation is weaker with a more spatially-symmetric distribution about the equator. Due to the slow response time of the ocean to forcing, the direct insolation forcing of the current season is cancelled by the ocean memory of an earlier insolation forcing, which in the case of the mid-Holocene is opposite to the current season insolation forcing. As a result, tropical sea surface temperature variation, as well as the tropical atmospheric temperature and moisture changes, is small, which gives rise to a different precipitation response from under the condition of stronger atmospheric temperature and moisture changes, such as in the case of post-industrial global warming induced by an increased concentration of atmospheric greenhouse gases. Thus, the cancellation between the direct and memory effects of the seasonally asymmetric insolation forcing leaves the net energy into the atmosphere to be responsible for the land-sea asymmetry of tropical precipitation changes.

1. Introduction

The climate of the mid-Holocene has been studied intensively via both geological paleoclimate records and model simulations. From the globally-distributed proxy record reconstruction, the mid-Holocene climate displays distinctly different patterns compared to pre-industrial climate, especially in the northern hemispheric monsoon regions, and was therefore chosen as a main target time slice for model simulations (Joussaume et al. 1999). General features of 18 Paleoclimate Modeling Intercomparison Project phase 1 (PMIP1) model simulations indicate a strengthening of monsoons in the northern hemisphere, which is roughly in agreement with proxy reconstructions (Joussaume et al. 1999). However, discrepancies between model simulations and data synthesis in North Africa have brought on more thorough model experiments on the regional vegetation and soil feedbacks in this region (e.g., Kutzbach et al. 1996; Texier et al. 1997; Brostrom et al. 1998; Jolly et al. 1998), and also some discussions on the influence of atmosphere-ocean coupling (Kutzbach and Liu 1997; Zhao et al. 2005). More recently, model experiments were conducted to access dynamical mechanisms that limit the northward extension of vegetation into the present Sahara Desert (Jolly et al. 1998) via an intermediate atmospheric model coupled with a simplified land surface model (Su and Neelin 2005). Several model studies were also performed in order to better understand the influence of atmosphere-ocean coupling to the East Asia summer monsoon during the mid-Holocene (e.g., Braconnot et al. 2000; Liu et al. 2003; Ohgaito and Abe-Ouchi 2007).

During the mid-Holocene, with surface conditions and greenhouse gas concentrations being similar to that of a pre-industrial condition, the insolation difference brought about by orbital configurations becomes the major forcing for the

climate, compared to pre-industrial modern conditions. The resulting insolation forcing of the combination of the three orbital parameters, namely obliquity, the precession of Earth's axis, and the eccentricity of Earth's orbit around the Sun, is shown in Fig. 1. Mean boreal summer (June, July, and August, or JJA) insolation is stronger during the mid-Holocene than at present, and therefore induces a warming effect with a spatially asymmetric pattern, with its highest value at high latitudes of the northern hemisphere. Austral summer (December, January, and Februarys, or DJF) averaged insolation, on the other hand, induces a cooling effect that is more symmetric to the equator. These opposite effects of seasonal insolation are termed "seasonally asymmetric insolation forcings" in our discussion. In other words, the insolation forcing tends to warm up the tropics, our area of interest, in JJA, and cool down the tropics in DJF. Overall, the mid-Holocene insolation forcing on the whole presents a temporal asymmetry, while the summer insolation forcing has a spatially asymmetric pattern.

This contrasting insolation forcing is also found earlier in the Holocene, about 9000 yr BP. Earlier modelers performed sensitivity tests on this time slice using a prescribed modern sea surface temperature (SST) in general circulation model (GCM) simulations considering the counteracting insolation that increases in the boreal summer and decreases in the austral summer (Kutzbach 1981; Kutzbach and Otto-Bliesner 1982). The enhanced northern hemispheric summer monsoon is simulated as a result of enhanced summer insolation, and is supported by proxy records. However, discussions on the interaction between seasonally-contrasting insolation forcings and geographic boundaries, such as land and ocean, were ignored. Most later model simulations incorporated ocean and vegetation feedbacks by coupling an atmospheric model with ocean and vegetation models. The simulated

climate pattern in comparison with proxy records (Qin et al. 1998; Joussaume et al. 1999; Liu et al. 2003), and the influence of the feedback processes in regional climate were discussed (Brostrom et al. 1998; Zhao et al. 2005). However, model experiment-based discussions on dynamical mechanisms that cause large-scale climatic variations have been left untouched. The purpose of this study therefore focuses on not only the mid-Holocene climate as a result of the seasonally-asymmetric insolation forcing but also the influences of a direct insolation forcing in different seasons on components of the earth's climate system, especially land and ocean, and how the seasonal insolation forcings interact with each other.

The description of the model and the experiment design are given in section 2. Section 3 presents the general feature of the mid-Holocene climate in model simulations and proxy records, a land-sea asymmetry of tropical precipitation changes in particular. Before further discussion on the cause of the precipitation differences between the mid-Holocene and current (pre-industrial) climate via model experiments, a budget analysis of the mid-Holocene simulation is performed in section 4. The impact of the direct insolation forcing and the ocean memory on tropical precipitation is discussed in section 5, followed by a general discussion on the mid-Holocene climate and conclusions.

2. Model and experiment design

a. The model

To understand mechanisms for tropical precipitation in the mid-Holocene, a climate model of intermediate complexity (Neelin and Zeng 2000; Zeng et al. 2000, ZNC hereafter) was used. Based on the analytical solutions derived from the Betts-Miller moist convective adjustment scheme (Betts and Miller 1993), typical

vertical structures of temperature, moisture and winds for deep convection are used as leading basis functions for a Galerkin expansion (Neelin and Yu 1994; Yu and Neelin 1994). The resulting primitive equation model makes use of constraints on the flow by quasi-equilibrium thermodynamic closures and is referred to as QTCM1 (quasi-equilibrium tropical circulation model with a single vertical structure of temperature and moisture for deep convection). Because the basis functions were based on vertical structures associated with the convective regions, these regions are expected to be well-represented and similar to a GCM with the Betts-Miller moist convective adjustment scheme. Far from the convective regions in the tropics, QTCM1 is a highly-truncated Galerkin representation, equivalent to a two-layer model. A cloud-radiation scheme (Chou and Neelin 1996; ZNC), simplified from the full GCM radiation schemes (Harshvardhan et al. 1987; Fu and Liou 1993), was included. Deep convection and cirrocumulus/cirrostratus cloud fraction was estimated by an empirical parameterization (Chou and Neelin 1999). A simple formula was used to obtain atmospheric boundary layer winds under assumptions of a steady state and a vertically-homogeneous mixed layer with fixed height (Stevens et al. 2002).

An intermediate land surface model with a single layer for both energy and water budgets was used to simulate interaction between land surface and the atmosphere (ZNC). There are three land surface types—forest, grass and desert—differing in soil moisture, field capacity, roughness length, and evapotranspiration characteristics (see Zeng et al. 2000). Prescribed surface albedo from the Darnell et al. (1992) climatology was used. The version of the model used here did not include topography, but mountain-like heating was used to mimic the thermal effect of topography. The topographic effect has strong influences on monsoon systems, so the assumption of no mechanical effect of topography implies caveats in discussing summer monsoons,

such as the Asian summer monsoon (Chou 2003; Chou and Neelin 2003). However, it will have little impact on the land-ocean asymmetry of tropical precipitation that we are interested in. To examine air-sea interaction, a mixed-layer ocean with 50-meter fixed depth was coupled with the atmospheric model (Chou et al. 2001). The prescribed seasonal divergence of ocean heat transport, Q-flux, in the mixed-layer ocean was estimated from a 10-year run with climatological SST. With the prescribed Q-flux, SST was determined by the energy balance between latent heat, sensible heat, surface radiative flux and Q-flux.

b. Experiment design

Three sets of experiments were performed to examine how the earth's climate responds to seasonally asymmetric insolation forcings in the mid-Holocene. With the insolation forcing being the focus of this study, a realistic insolation forcing was applied (Berger 1978), while the boundary conditions in all experiments were set to values identical to pre-industrial values, including the atmospheric CO₂ concentration (Raynaud et al. 1993). The results of each experiment shown in this study were calculated from the last 100 years of 200-year model simulations, while the first 100 years of spin up was disregarded. The details of the experiment design can be found in Table 1.

The first pair of experiments was used to examine the differences between the mid-Holocene (MH) and the current climate (CTRL). The PMIP2 (PMIP phase 2) modern pre-industrial condition was applied in the CTRL experiment, while the PMIP2 mid-Holocene condition was used in the MH experiment (Braconnot et al. 2007). In the MH experiment, the calendar effect (Joussaume and Braconnot 1997) was not included, similar to most of the PMIP2 simulations. This caused slight differences in insolation (not shown), with maxima in autumn (Braconnot et al. 2007).

However, it did not affect our results qualitatively, especially with the focus of this study being in June-July-August (JJA) and December-January-February (DJF), when the changes were relatively small.

To examine mechanisms of precipitation changes, the associated process was suppressed in paired experiments, EXP and EXP_CTRL. In this study, we only examined the effect of the net energy into the atmosphere F^{net} , which includes net solar and longwave radiation into the atmosphere, sensible heat and latent heat. The detailed derivative of F^{net} can be found in section 4. In this pair of experiments, all components of F^{net} except the net solar radiation were prescribed with the values obtained from the CTRL experiment. The net solar radiation of the EXP and EXP_CTRL experiments was calculated according to the mid-Holocene and the pre-industrial insolation parameters, respectively. Thus, most of the effect of F^{net} was suppressed in the results of the EXP minus the EXP_CTRL experiments, since the net solar radiation into the atmosphere is relatively small.

The third set of experiments, namely the direct insolation forcing experiment, was designed to examine effects of the insolation forcing in different seasons separately, especially boreal and austral summers, when the changes of the insolation forcing are the strongest. In the JJA_Dir (DJF_Dir) experiment, the insolation parameters of the mid-Holocene were assigned in JJA (DJF), while the pre-industrial insolation parameters were applied in the other 9 months. Thus, the insolation still varies for the entire year, with the mid-Holocene insolation in only one season. The results of both experiments were compared with the CTRL experiment. In other words, the difference between the paired JJA_Dir (DJF_Dir) and CTRL experiments were aroused solely by the JJA (DJF) insolation difference between the mid-Holocene and the pre-industrial periods. In the JJA_Dir experiment, for instance, changes in JJA was

induced by the direct insolation forcing, while changes in DJF was caused by a memory effect of the JJA insolation forcing, which was stored in the ocean. To simplify the experiments, the effects of the spring and autumn insolation forcings were omitted here. This simplification slightly modified the magnitudes of the changes, but did not alter the main conclusion discussed below.

3. The mid-Holocene climate

According to various geological records and model simulations during the mid-Holocene period, a very significant strengthening of the East Asian and African summer monsoons is documented (e.g., Kutzbach 1988; Kutzbach and Liu 1997; An et al. 2000). Both the PMIP phase 2 (PMIP2; Braconnot et al. 2007) model ensemble of 16 Ocean-Atmospheric GCMs (Fig. 2a) and the QTCM control pair results (Fig. 3a) show stronger precipitation north of the mean summer monsoonal rain belt of the pre-industrial climate (the thick contour) in North Africa and Asia, and weaker precipitation to the south located mostly over tropical to subtropical ocean areas, during the mid-Holocene than during the pre-industrial boreal summer.

The increase of surface temperature (Fig. 4a) over the northern hemisphere, which is a direct response to stronger boreal insolation in the mid-Holocene, is also in good agreement with the temperature trend indicated by pollen records (e.g., Xiao et al. 2004; Zhou et al. 2004; Zhou et al. 2005; Liew et al. 2006). Both simulated tropospheric temperature and moisture anomalies (Figs. 4b and 4c) vary with a rough correspondence to surface temperature anomalies. However, due to atmospheric wave dynamics (Wallace et al. 1998; Chiang and Sobel 2002; Su et al. 2003) they show a smoother pattern. Over the Eurasian continent, an enhanced meridional gradient of tropospheric temperature is found (Fig. 4b), which is believed to be associated with

the strengthening of the Asian summer monsoon in the mid-Holocene (Li and Yanai 1996).

As the northward shift and enhancement of the boreal summer monsoon rainfall is a prominent feature in both geological records and model simulations, the change of the mid-Holocene precipitation in austral summer is very different (Fig. 3b). Precipitation decreases over the summer monsoon-dominated regions in the southern hemisphere, including South Africa, North Australia and South America (Gasse 2000; Baker et al. 2001a; Baker et al. 2001b; Thevenon et al. 2002; Fritz et al. 2004; Garcin et al. 2006; Toledo et al. 2007). In contrast to the boreal summer warming of the continents by stronger summer insolation, weaker boreal winter insolation in the mid-Holocene causes surface cooling over most of the continental regions (Fig. 4d), which is roughly consistent with the PMIP2 ensemble (not shown). The cooling of the surface temperature in turn causes a slight cooling of the tropical troposphere and a decrease in the tropospheric moisture (Figs. 4e and 4f). The variation in the tropospheric temperature anomaly is relatively smaller in DJF than in JJA (Figs. 4b and 4d).

Although the climatological major precipitation areas in the tropics are more evenly distributed over the tropical continents and oceans (the thick line in Fig.3b), the simulated mid-Holocene precipitation anomaly of austral summer (DJF) over convective regions (the thick line) shows a clear land-sea asymmetry, with increased precipitation over oceanic convection zones, which contrasts with the decreased precipitation over continental convective regions at the same latitude. The pattern of increased precipitation over ocean and decreased precipitation on land is also clearly presented in the PMIP2 model ensemble (Braconnot et al. 2007). A similar but reversed land-sea asymmetry of precipitation changes is also observed in boreal

summer (JJA), although the northward shift of the major rainband is very prominent. Surface temperature differences show a similar tendency where land and ocean temperatures change in a more or less opposite direction to one another (Figs. 4a and 4d). As shown in both the PMIP2 ensemble and the QTCM results, the reversed pattern of boreal and austral summers under the influence of the seasonally asymmetric insolation forcing seems very obvious.

Due to the lack of ocean dynamics, some features might not be able to be simulated by the coupling system of QTCM1 and a mixed-layer ocean. This could create caveats in studying some aspects of the mid-Holocene climate. For instance, the weakening of the Atlantic ITCZ, which is found in the PMIP2 ensemble (Fig. 2a), is not well simulated by QTCM1 (Fig. 3a) even though negative precipitation anomalies, induced by a different mechanism (Chou and Neelin 2004), did occur near this region. Overall, this model system did a relatively good job in simulating the land-sea asymmetry of tropical precipitation change, which we are interested in. Thus, this coupling system should be good enough for studying mechanisms inducing this land-sea precipitation asymmetry.

4. Mechanisms of tropical precipitation change

a. Moisture and moist static energy budgets

To understand the mechanisms that determine changes in precipitation in all the experiments, the vertically integrated moisture budget in a steady state is first diagnosed to gain an insight on how the atmospheric moisture is balanced. The precipitation differences between the mid-Holocene and the pre-industrial climate can be estimated by

$$P' \approx -\langle \bar{\omega} \partial_p q' \rangle - \langle \omega' \partial_p \bar{q} \rangle - \langle \mathbf{v} \cdot \nabla q \rangle' + E', \quad (1)$$

where E is evaporation, ω is vertical velocity on the pressure scale, and $\mathbf{v}$ is horizontal velocity. The specific humidity q is in energy units by absorbing the latent heat per unit mass, L . The precipitation is in energy units (Wm^{-2}), which divided by 28 become mm day^{-1} . $\langle \rangle$ denotes a mass integration through the troposphere with p_T as the depth of troposphere:

$$\langle x \rangle = -\frac{1}{g} \int_{p_s}^{p_s - p_T} x dp, \quad (2)$$

where p_s is surface pressure, and g is gravity, $(\overline{})$ represents climatology of the pre-industrial climate, and $()'$ is the difference between the pre-industrial and the mid-Holocene climates, i.e., the mid-Holocene minus the pre-industrial climate. All terms in the moisture budget (1) are in energy units.

In the tropics, the vertical velocity anomalies ω' can be estimated via the moist static energy (MSE) budget since tropical convection is the most dominant process. The vertically-integrated MSE budget can be written as

$$\langle \omega' \partial_p \overline{h} \rangle \approx -\langle \overline{\omega} \partial_p h' \rangle - \langle \mathbf{v} \cdot \nabla (q + T) \rangle' + F^{net'}, \quad (3)$$

where T is atmospheric temperature that absorbs the heat capacity at constant pressure C_p ; MSE is $h = q + s$. The dry static energy is $s = \varphi + T$, and φ is the geopotential. F^{net} is defined as the net energy input to the atmospheric column:

$$F^{net} = S_t^\downarrow - S_t^\uparrow - S_s^\downarrow + S_s^\uparrow - R_t^\uparrow - R_s^\downarrow + R_s^\uparrow + E + H. \quad (4)$$

Here, subscripts s and t on the solar ($S_t^\downarrow$ and $S_t^\uparrow$) and longwave ($R_t^\downarrow$ and $R_t^\uparrow$) radiative terms denote surface and model top; $R_t^\downarrow \approx 0$ has been used, and H is the sensible heat flux.

b. Budget interpretation in the quasi-equilibrium tropical circulation model

We can interpret the above budgets by using the approximation in QTCM. Under the constraint of quasi-equilibrium convective closures discussed in Chou and Neelin (2004), the vertical moisture advection $-\langle \omega \partial_p q \rangle$, with the definition of a typical vertical profile of vertical motion $\Omega(p)$ of deep convection, can be rewritten as

$$-\langle \omega \partial_p q \rangle = M_q \nabla \cdot \mathbf{v}_1, \quad (5)$$

where $\mathbf{v}_1$ is the wind component associated with the baroclinic structure under the convective quasi-equilibrium constraint. The gross moisture stratification M_q (Neelin and Yu 1994; Yu et al. 1998) is given by

$$M_q = \langle \Omega \partial_p q \rangle. \quad (6)$$

In the convective regions, M_q measures the basic-state moisture available for precipitation. Note that $\nabla \cdot \mathbf{v}_1 > 0$ indicates low-level convergence and upper-level divergence, hence, upward motion. $\Omega(p)$ is positive, so M_q is positive as well. The moisture budget (1) then becomes

$$P' \approx \bar{M}_q \nabla \cdot \mathbf{v}_1' + M_q' \nabla \cdot \bar{\mathbf{v}}_1 - \langle \mathbf{v} \cdot \nabla q \rangle' + E'. \quad (7)$$

The second-order term $M_q' \nabla \cdot \mathbf{v}_1'$, including the nonlinear effect and transients, is neglected in this study. Equation (7) can be taken for the diagnostic analysis of precipitation changes between two model runs, that is, precipitation changes are balanced by processes on the right hand side (r.h.s.).

Similar to the moisture advection in (5), the vertical MSE advection $-\langle \omega \partial_p h \rangle$ can be written as

$$-\langle \omega \partial_p h \rangle = -M \nabla \cdot \mathbf{v}_1, \quad (8)$$

where the gross moist stability, M , (Neelin and Yu 1994; Yu et al. 1998) is defined as

$$M = -\langle \Omega \partial_p h \rangle. \quad (9)$$

In the convective regions, M represents effective static stability for large-scale motions. Thus, the vertically-integrated MSE budget (3) can be rewritten as

$$\overline{M}\nabla \cdot \mathbf{v}'_1 = -M\nabla \cdot \bar{\mathbf{v}}_1 - \langle \mathbf{v} \cdot \nabla (q + T) \rangle' + F^{net'}. \quad (10)$$

The vertical velocity $\nabla \cdot \mathbf{v}'_1$ can be estimated by (10), so the moisture budget (7) is further modified as

$$\begin{aligned} P' \approx & \frac{\overline{M}_q}{\overline{M}} [-\langle \mathbf{v} \cdot \nabla T \rangle' + F^{net'}] + E' - \left(\frac{\overline{M}_q}{\overline{M}} + 1 \right) \times \langle \mathbf{v} \cdot \nabla q \rangle' \\ & + \left(-\frac{\overline{M}_q}{\overline{M}} M' + M'_q \right) \nabla \cdot \bar{\mathbf{v}}_1. \end{aligned} \quad (11)$$

c. Diagnosis

The precipitation differences between the MH and CTRL runs (Figs. 3a and 3b for JJA and DJF averages) are compared with four terms on the r.h.s. of (7), the moisture budget (Figs. 5 and 6). For both JJA and DJF climatologies, the overall patterns of the four terms on the r.h.s. of (7) show that the precipitation changes are mostly dominated by moisture convergence induced by anomalous vertical motion, namely $\overline{M}_q \nabla \cdot \mathbf{v}'_1$ (Figs. 5a and 6a). While the enhanced moisture convergence coincides with the increased precipitation, the weakened moisture convergence occurs in regions with decreased precipitation. The second term, $M'_q \nabla \cdot \bar{\mathbf{v}}_1$ (Figs. 5b and 6b), changes most pronouncedly when the tropospheric moisture changes dramatically (Figs. 4c and 4f), such as in the Asian-African summer monsoon region, and is also slightly magnified by the mean vertical motion. However, these changes are not significant enough to become a major contributor to the precipitation changes. The horizontal moisture advection anomaly (Figs. 5c and 6c) has a more disorderly pattern. The dry and moist advection is somewhat stronger in boreal summer than in austral summer. In areas with a strong moisture gradient, the horizontal moisture advection

changes are strong. For instance, the Asian summer monsoon region is dominated by a dry advection (Fig. 5c), which is associated with a positive meridional gradient of moisture changes (Fig. 4c). Evaporation changes are relatively weak in general, with some exceptions in summer monsoon-dominated regions where strong precipitation changes occur. The continental evaporation changes are mainly controlled by surface temperature and soil moisture content. The strong evaporation changes are concentrated in limited continental regions where strong summer monsoon precipitation occurs, which increases the soil water content (not shown) that in turn affects the evaporation in those regions not only in boreal but also later in austral summer. On the other hand, evaporation over ocean varies mainly with sea surface temperature (SST) changes (Figs. 4a and 4d), which has opposite signs in boreal and austral summers; therefore, the evaporation changes also show an opposite response (Figs. 5d and 6d). We note that relatively strong negative evaporation anomalies are found over the northern Indian Ocean in JJA. The reduced evaporation is associated with both colder SST and weaker surface wind speed (not shown) in the mid-Holocene.

Based on the spatial patterns and the amplitudes of the moisture budget shown in Figs. 5 and 6, the anomalous moisture convergence associated with the anomalous vertical motion ($\overline{M}_q \nabla \cdot \mathbf{v}'_1$) is the primary term contributing to the anomalous precipitation. The MSE budget (10) is therefore analyzed to understand factors causing the vertical velocity changes because convection is the main process producing precipitation in the tropics. On the l.h.s. of (10), $\overline{M} \nabla \cdot \mathbf{v}'_1$ (Figs. 7a and 8a), which has a very similar pattern to $\overline{M}_q \nabla \cdot \mathbf{v}'_1$ (Figs. 5a and 6a), is balanced by $\overline{M} \nabla \cdot \overline{\mathbf{v}}_1$, the horizontal MSE advection $-\langle \mathbf{v} \cdot \nabla (q + T) \rangle'$ and $F^{\text{net},r}$ on the r.h.s. of (10). It is clear that the net energy absorbed by the atmosphere $F^{\text{net},r}$ in Figs. 7d and 8d

shows the most resemblance to the distribution of $\overline{M}\nabla \cdot \mathbf{v}'_1$, as shown in both boreal and austral cases where the land anomaly is opposite to the ocean anomaly. In continental regions, surface temperature responds rapidly, as well as longwave radiation, sensible and latent heat fluxes, to compensate the change in solar radiation for both boreal and austral summers. As the upward longwave radiation and the sensible heat flux from Earth's surface are controlled by surface temperature, the continental surface temperature therefore varies in the same direction as the insolation forcing, which indicates an enhancement in boreal summer and a reduction in austral summer. On the other hand, in tropical ocean regions, the SST varies contrarily to the continental surface temperature, which is inconsistent with the insolation forcing. The evaporation changes over ocean also show a tendency opposite to the latitudinal shortwave radiation changes and are the dominant contributor to the F^{net} changes over ocean. The term $M\nabla \cdot \bar{\mathbf{v}}_1$ (Figs. 7b and 8b) contributes very little to the anomalous vertical motion in the tropics (Figs. 7a and 8a). Most larger values of the horizontal MSE advection occur in higher latitudes (Figs. 7c and 8c), and these values are related to mid-latitude storm activities. The strong horizontal advection over the Asian summer monsoon region, however, is associated with strong meridional gradients of temperature (Fig. 4b) and moisture (Fig. 4c). This strong advection is responsible for the dipole pattern of the Asian summer monsoon rainfall changes (Fig. 3a). In other words, the northward shift of the Asian summer monsoon rainband is associated with the horizontal advection of MSE, which is consistent with previous studies (e.g., Hewitt and Mitchell 1998; Joussaume et al. 1999; Braconnot et al. 2002).

To directly test the significance of $F^{\text{net}'}$ in determining the mid-Holocene precipitation change (MH minus CTRL), the $F^{\text{net}'}$ term was suppressed in the hypothesis-testing experiments, i.e., the EXP and EXP_CTRL experiments. In other

words, the climate responded only to the mid-Holocene insolation forcing without contributions from the net longwave radiation into the atmosphere, latent heat, or sensible heat flux components in the F^{net} term. The resulting precipitation difference over convective regions (EXP minus EXP_CTRL) for both boreal and austral summers (Fig. 9) is significantly diminished when compared with the control experiment (MH-CTRL) results (Fig. 3). We note that the meridionally varied summer insolation forcing is still important, so the reduction of the precipitation difference is more evident in DJF than in JJA. Outside convective regions, other effects, such as horizontal moisture and temperature advections, can still induce precipitation changes. The disappearance of a major precipitation anomaly pattern over convective regions after suppressing F^{net} suggests that the F^{net} term is quite effective in determining precipitation changes, consistent with the analysis of the MSE budget, which was discussed above.

5. Impacts of direct insolation and ocean memory

Based on the budget analysis of the control experiments, a strong land-sea asymmetry in precipitation changes and F^{net} suggests a possible effect due to different properties of ocean and land under the mid-Holocene insolation forcing. To examine this effect, the direct insolation forcing experiments (DJF_Dir and JJA_Dir) were conducted. While the model is forced only by the JJA insolation forcing (JJA_Dir minus CTRL), the boreal summer insolation (Fig. 1) causes significant surface warming over all tropical regions (Fig. 10b), with slightly stronger warming over land than ocean. Moreover, influenced by the spatially-asymmetric insolation forcing, that is, one which is stronger in the northern hemisphere, the surface warming also shows a similar spatial pattern with more pronounced warming in the north. The tropospheric

temperature changes (Fig. 10c) show a corresponding warming pattern, with slightly stronger warming in the northern hemisphere but smoother than the surface temperature. The tropospheric moisture (Fig. 10d), which is mostly associated with the tropospheric temperature, therefore also increases more substantially in the northern hemisphere.

On the other hand, significant tropical cooling occurs when the model is forced only by the DJF insolation forcing (DJF_Dir minus CTRL). In austral summer (DJF), the surface temperature responds directly to the reduced insolation, with slightly stronger cooling on land than ocean (Fig. 11b). Whereas the surface temperature changes somewhat reflect the land-sea distribution, the tropospheric cooling presents a smoother spatial pattern (Fig. 11c). The tropospheric moisture content, which is influenced by tropospheric cooling, drops as a result. However, while the tropospheric temperature pattern shows a rather small variation in spatial distribution (Fig. 11c), moisture changes concentrate in the convective regions near the equator (Fig. 11d). The discrepancy between the anomalous moisture and the temperature distributions is due to the following reason: as tropospheric temperature decreases, moisture would have to drop more pronouncedly in convective regions due to suppressed convection. Outside the convective regions, other processes, such as evaporation, dominate the moisture content. Therefore the moisture content does not change as much as in the convective regions (Fig. 11d).

Precipitation changes in the direct insolation forcing experiments have a different pattern compared to that in the control experiment, especially in the case of the austral summer. In boreal summer, the direct response of precipitation (Fig. 10a) to the insolation forcing is somewhat similar to the results of the control experiment pair (Fig. 3a). The northward shifting of the summer monsoonal rain belt as in the control

experiments shown in Fig. 3a is also quite pronounced in the JJA_Dir experiment, though the strength of the anomalous precipitation is not as strong. Over ocean, the precipitation decrease is also weaker than in the result of the control experiment pair (MH-CTRL). In the DJF_Dir experiment (Fig. 11a), austral summer precipitation decreases either over land or ocean in the convective regions, but increases in the margins of the convective regions. The moisture budget (not shown) analysis indicates a significant contribution to the precipitation changes not only from the moisture convergence induced by the anomalous vertical motion ($\overline{M}_q \nabla \cdot v'_1$), but also from $M'_q \nabla \cdot \bar{v}_1$, which results from the decrease in tropical tropospheric moisture. The term $\nabla \cdot v'_1$, as analyzed by the MSE budget, is mostly contributed to by $M \nabla \cdot \bar{v}_1$ in the convective centers and the horizontal MSE advection in the convective margins (not shown).

By analyzing the direct effect of the insolation forcing, it is clear how different the results (Figs. 10 and 11) are from the control experiment (Figs. 3 and 4); this indicates the existence of some contributions from feedback mechanisms of the climate system other than merely from the direct insolation forcing. Some clues can be deduced from the analysis of other seasons. The DJF climatologies of the JJA_Dir experiment show that the warming signal of the boreal summer insolation forcing remains (Fig. 12). Although there is no direct insolation forcing in this season, tropical surface temperature remains warmer (Fig. 12b). The maximum warming areas are mostly in the northern hemisphere, which reflects stronger insolation in the north. It is noteworthy that warming also occurs over most ocean surfaces. The tropospheric temperature and moisture also respond similarly to surface temperature, with more tropospheric warming and increased moisture in the tropical northern hemisphere (Figs. 12c and 12d).

The cooling effect of the austral summer insolation forcing in the DJF_Dir experiment also persists until boreal summer (JJA), as clearly observed in the JJA climatologies (Fig. 13). The surface temperature changes show tropical cooling as in DJF climatology (Fig. 11b) but with weaker amplitude (Fig. 13b), a pattern very similar to DJF surface temperature in the JJA_Dir experiment with opposite signs. The tropospheric temperature responds rather similarly to the surface temperature (Figs. 13b and 13c), causing corresponding tropospheric moisture changes in the tropics (Fig. 13d).

The precipitation changes (Figs. 12a and 13a) are also governed by the same mechanisms, which result in the precipitation variation directly forced by insolation. While increased (decreased) precipitation occurs at the center of the convective regions as a result of a significant increase (decrease) in moisture and a reduced (enhanced) effective static stability, diminished (enhanced) precipitation is found at the margins of the convective region due to the horizontal dry (moist) advection of surrounding air (Figs. 12a and 13a, respectively). We note that by an add-on of the DJF (JJA) precipitation anomalies of the JJA_Dir and DJF_Dir experiments together (Figs. 11a and 12a), the precipitation variation pattern is slightly different from the control experiments (MH-CTRL) shown in Fig. 3b. It is because the contribution of ocean memory from other seasons, spring and autumn, is also important.

From the budget analysis discussed above, the main contributor to the vertical velocity change in the direct insolation forcing experiments (JJA_Dir and DJF_Dir experiments) is quite different from that in the control experiment (MH-CTRL). While $M \nabla \cdot \bar{v}_1$ and the horizontal MSE advection terms contribute mostly to the precipitation difference in the direct insolation forcing experiments, F^{net_r} dominates the control experiment precipitation changes (Figs. 7 and 8). These differences

between experiments arise from their diverse responses to the atmospheric moisture profile. Here, the austral summer case is discussed below for its relatively meridionally-symmetric insolation forcing.

When the seasonal insolation asymmetry is removed in the direct insolation forcing experiment (DJF_Dir), the tropical atmospheric temperature decreases significantly as well as the moisture (Figs. 11b, 11c and 11d). The climate response in the DJF_Dir experiment has a high resemblance to a double CO₂ global warming scenario discussed in Chou and Neelin (2004), Tan et al. (2008) and Chou et al. (2009). While the insolation forcing causes a uniform cooling over the entire tropical region in the DJF_Dir experiment, a doubled atmospheric CO₂ concentration causes global warming. For both cases, the vertical moisture profiles change significantly; as a result, the anomalous vertical motion, which induces precipitation changes, is dominated by $-M\nabla \cdot \bar{v}_1$ in the convective centers (the “rich-get-richer” mechanism) whereas the horizontal moisture advection term, which is the dominant term in $-\langle v \cdot \nabla(q + T) \rangle'$, contributes to the convective margins (the “upped-ante” mechanism), except with opposite signs due to the contrast in temperature responses to radiative forcing. The use of a similar mechanism in explaining the precipitation changes in both cooling and warming scenarios indicates that the dynamical response in the tropical atmosphere has an internal consistency under uniform global forcing, such as radiative warming or cooling.

On the other hand, when the air temperature variation in the control experiment is relatively small, the response of atmospheric moisture is diminished. Consequently, the variations in the atmospheric stability and the horizontal moisture gradient are diminished. While the former reduces the contribution of $-M\nabla \cdot \bar{v}_1$ (Fig. 8b), the latter moderates the contribution of the horizontal MSE advection (Fig. 8c) to the

vertical velocity anomalies. The abatement of the first two terms contributing to $-\overline{M}\nabla \cdot \mathbf{v}'_1$ in the MSE budget (10) leaves $F^{\text{net}'}$ to be a major contributor to the vertical velocity anomaly, which in turn causes precipitation changes. In DJF, the ocean retains the energy induced by the increased insolation from previous seasons, so SST changes are still positive over most of the tropical oceans (Fig. 4d) even though the DJF insolation is reduced. This creates positive $F^{\text{net}'}$, mainly through evaporation, so precipitation over oceanic convection zones is enhanced. Over land, on the other hand, colder surface induced by the reduced insolation suppresses convection, and the associated precipitation is then reduced. The dominance of $F^{\text{net}'}$ in the precipitation changes is sensitive to the intensity of the atmospheric moisture variation, which is controlled mainly by the magnitude of average tropospheric temperature variation under a large-scale forcing.

Therefore, the above discussion obviously shows that the warming (cooling) signals of last season's insolation remains in the ocean system during later months, which might affect or cancel the direct insolation cooling (warming) of the current season, especially for the ocean area. This result suggests that the slight opposite variation of the ocean temperature to the direct mid-Holocene insolation forcing in the control experiments and in the PMIP2 simulations may be caused by the cancellation between the direct insolation forcing and the ocean memory via $F^{\text{net}'}$ under the configuration of the seasonally-asymmetric insolation forcing. Also, the precipitation responses can readily be expressed as the combined outcome of the direct insolation forcing and the ocean memory effects. In short, these experimental results show the importance of the ocean's capability of retaining heat on a longer time scale than annual cycles as well as its influence on long-term climate changes (Ohgaito and Abe-Ouchi 2007; Braconnot et al. 2008). In this study, we considered a fixed

mixed-layer depth, so the magnitude of the land-ocean precipitation contrast might differ from those in the PMIP II simulations. A further study should be done in order to understand the impacts of the mixed-layer depth on the land-ocean precipitation asymmetry.

6. Conclusions

Three sets of model experiments were conducted to understand the mid-Holocene climate variability, a land-sea asymmetry of precipitation in particular, from different aspects by an intermediate atmosphere model coupled with a simple land surface model and a mixed-layer ocean model. Since the mechanical effect of topography was not included in this model, the associated summer monsoons were slightly weaker than those in the PMIP II simulations. Overall, the simulated general climate pattern of the mid-Holocene agrees well with the GCM simulations and paleoclimatic proxy reconstructions, with some small discrepancies due to the lack of ocean dynamics. A distinct asymmetry in precipitation change is found over convective regions between tropical continents and oceans. In boreal summer, the increased precipitation occurs mostly over the tropical continents; however, decreased precipitation occurs over the tropical ocean regions. On the other hand, austral summer precipitation increases over the ocean regions but decreases over the continents, a land-sea asymmetry of precipitation changes.

The mechanism causing this distinct land-sea asymmetry in the precipitation changes lies in two properties of the earth's system and the insolation forcing, respectively. First, due to the different thermal inertia of continents and oceans, the response time of continents and oceans to the insolation forcing differs. Second, the mid-Holocene insolation forcing presents a seasonally asymmetric pattern, which

causes a warming effect in one half of the year and a cooling effect in the other half. In the continental regions, the temperature, and in turn the precipitation, respond as soon as the solar radiation changes. Therefore, precipitation changes over continents vary in the same direction as the mid-Holocene insolation forcing. However, the ocean responds to radiative forcing slowly; as a result, the memory of last season's insolation forcing remains and cancels out the later insolation forcing, which carries an opposite sign to the earlier forcing, leading to no or slight SST changes. Thus, the process associated with the net energy into the atmosphere becomes dominant in determining the land-sea precipitation asymmetry.

The small variation of the SST as a result of the cancellation between the direct insolation forcing and the ocean memory effect not only enables reasonable results from earlier SST-prescribed model simulations (Kutzbach 1981; Kutzbach and Otto-Bliesner 1982), but also implies that a SST-prescribed simulation is virtually sufficient to account for the energy balance between the atmosphere and the mixed-layer ocean. By comparing coupled atmosphere-ocean GCM simulations with SST-prescribed simulations, the influence of ocean feedbacks to regional monsoon climate has been discussed (e.g., Hewitt and Mitchell 1998; Vettoretti et al. 1998; Braconnot et al. 2000; Liu et al. 2004). Our results suggest that the SST-prescribed simulation tends to obscure the effect of the ocean-atmosphere energy balance to the climate responses, and therefore would underestimate the ocean memory effect on climate in the discussions of ocean feedback (Braconnot et al. 2007).

In conclusion, our study emphasizes the importance of the different responses of land and ocean based on their different characteristics under a seasonally-asymmetric insolation forcing. The cancelling effect of the direct insolation forcing and the ocean memory of the previous season forcing can moderate the temperature and

precipitation changes. Since both the obliquity and precession variations of the Earth's orbit induce a seasonal contrast in insolation change, a more thorough investigation of the influence of a seasonally-varied forcing on orbital time-scale climate changes may provide some significant clues to mechanisms that have governed the glacial-interglacial cycles.

Acknowledgements:

We acknowledge the international modeling groups for providing their data for analysis, the Laboratoire des Sciences du Climat et de l'Environnement (LSCE) for collecting and archiving the model data. The PMIP2/MOTIF Data Archive is supported by CEA, CNRS, the EU project MOTIF (EVK2-CT-2002-00153) and the Programme National d'Etude de la Dynamique du Climat (PNEDC). More information is available on <http://pmip2.lsce.ipsl.fr/> and <http://motif.lsce.ipsl.fr/>. We thank three anonymous reviewers' valuable comments for improving the quality of this paper. This work was supported under the National Science Council Grant NSC98-2628-M-001-001.

References

- An, Z. S., S. C. Porter, J. E. Kutzbach, X. H. Wu, S. M. Wang, X. D. Liu, X. Q. Li, and W. J. Zhou, 2000: Asynchronous Holocene optimum of the East Asian monsoon. *Quat. Sci. Rev.*, **19**, 743-762.
- Baker, P. A., C. A. Rigsby, G. O. Seltzer, S. C. Fritz, T. K. Lowenstein, N. P. Bacher, and C. Veliz, 2001a: Tropical climate changes at millennial and orbital timescales on the Bolivian Altiplano. *Nature*, **409**, 698-701.
- Baker, P. A., G. O. Seltzer, S. C. Fritz, R. B. Dunbar, M. J. Grove, P. M. Tapia, S. L. Cross, H. D. Rowe, and J. P. Broda, 2001b: The history of South American tropical precipitation for the past 25,000 years. *Science*, **291**, 640-643.
- Berger, A. L., 1978: Long-term variations of daily insolation and quaternary climatic changes. *J. Atmos. Sci.*, **35**, 2362-2367.
- Betts, A. K., and M. J. Miller, 1993: The Betts–Miller scheme. The representation of cumulus convection in numerical models. Meteor Monogr 46, Amer Meteor Soc, pp 107–121.
- Braconnot, P., O. Marti, S. Joussaume, and Y. Leclainche, 2000: Ocean feedback in response to 6 kyr BP insolation. *J. Climate*, **13**, 1537-1553.
- Braconnot, P., M.-F. Loutre, B. Dong, S. Joussaume, P. Valdes, and PMIP participating groups, 2002 : How the simulated change in monsoon at 6 ka BP is related to the simulation of the modern climate : results from the Paleoclimate Modeling Intercomparison Project. *Climate Dyn.*, **19**, 107-121.
- Braconnot, P., and Coauthors, 2007: Results of PMIP2 coupled simulations of the Mid-Holocene and Last Glacial Maximum—Part 1: Experiments and large-scale features. *Climate Past*, **3**, 261-277.
- Braconnot, P., C. Marzin, L. Grégoire, E. Mosquet, and O. Marti, 2008 : Monsoon response to changes in earth’s orbital parameters: comparisons between simulations of the Emian and of the Holocene. *Climate Past*, **4**, 281-294.
- Brostrom, A., M. Coe, S. P. Harrison, R. Gallimore, J. E. Kutzbach, J. Foley, I. C. Prentice, and P. Behling, 1998: Land surface feedbacks and palaeomonsoons in northern Africa. *Geophys. Res. Lett.*, **25**, 3615-3618.
- Chiang, J. C. H., and A. H. Sobel, 2002: Tropical tropospheric temperature variations caused by ENSO and their influence on the remote tropical climate. *J. Climate*, **15**, 2616-2631.

- Chou, C., and J. D. Neelin, 1996: Linearization of a longwave radiation scheme for intermediate tropical atmospheric models. *J. Geophys. Res.*, **101**, 15129-15145.
- Chou, C., and J. D. Neelin, 1999: Cirrus detrainment-temperature feedback. *Geophys. Res. Lett.*, **26**, 1295-1298.
- Chou, C., J. D. Neelin, and H. Su, 2001: Ocean-atmosphere-land feedbacks in an idealized monsoon. *Quart. J. Roy. Meteor. Soc.*, **127**, 1869-1891.
- Chou, C., 2003: Land-sea heating contrast in an idealized Asian summer monsoon. *Climate Dyn.*, **21**, 11-25.
- Chou, C., and J. D. Neelin, 2003: Mechanisms limiting the northward extent of the northern summer monsoons over North America, Asia, and Africa. *J. Climate*, **16**, 406-425.
- Chou, C., and J. D. Neelin, 2004: Mechanisms of global warming impacts on regional tropical precipitation. *J. Climat*, **17**, 2688-2701.
- Chou, C., J. D. Neelin, C.-A. Chen, and J.-Y. Tu, 2009: Evaluating the “rich-get-richer” mechanism in tropical precipitation change under global warming. *J. Climate*, **22**, 1982-2005.
- Darnell, W. L., W. F. Staylor, S. K. Gupta, N. A. Ritchey, and A. C. Wilber, 1992: Seasonal-variation of surface radiation budget derived from international satellite cloud climatology Project C1 Data. *J. Geophys. Res.*, **97**, 15741-15760.
- Fritz, S. C., P. A. Baker, T. K. Lowenstein, G. O. Seltzer, C. A. Rigsby, G. S. Dwyer, P. M. Tapia, K. K. Arnold, T. L. Ku, and S. D. Luo, 2004: Hydrologic variation during the last 170,000 years in the southern hemisphere tropics of South America. *Quat. Res.*, **61**, 95-104.
- Fu, Q., and K. N. Liou, 1993: Parameterization of the radiative properties of cirrus clouds. *J. Atmos. Sci.*, **50**, 2008-2025.
- Garcin, Y., D. Williamson, M. Taieb, A. Vincens, P. E. Mathe, and A. Majule, 2006: Centennial to millennial changes in maar-lake deposition during the last 45,000 years in tropical Southern Africa (Lake Masoko, Tanzania). *Paleogeogr Paleoclimatol Paleoecol* **239**, 334-354.
- Gasse, F., 2000: Hydrological changes in the African tropics since the Last Glacial Maximum. *Quat. Sci. Rev.*, **19**, 189-211.
- Harshvardhan, R. D., D. A. Randall, and T. G. Corsetti, 1987: A fast radiation parameterization for atmospheric circulation models. *J. Geophys. Res.*, **92**, 1009-1016.
- Hewitt, C. D., and J. F. B. Mitchell, 1998: A fully coupled GCM simulation of the

- climate of the mid-Holocene. *Geophys. Res. Lett.*, **25**, 361–364.
- Jolly, D., S. P. Harrison, B. Damnati, and R. Bonnefille, 1998: Simulated climate and biomes of Africa during the late quaternary: Comparison with pollen and lake status data. *Quat. Sci. Rev.*, **17**, 629–657.
- Joussaume, S., and coauthors, 1999: Monsoon changes for 6000 years ago: results of 18 simulations from the Paleoclimate Modeling Intercomparison Project (PMIP). *Geophys. Res. Lett.*, **26**, 859–862.
- Kutzbach, J. E., 1981: Monsoon climate of the early Holocene - climate experiment with the earth's orbital parameters for 9000 years ago. *Science*, **214**, 59–61.
- Kutzbach, J. E., and B. L. Otto-Bliesner, 1982: The sensitivity of the African-Asian monsoonal climate to orbital parameter changes for 9000 years bp in a low-resolution general-circulation model. *J. Atmos. Sci.*, **39**, 1177–1188.
- Kutzbach, J. E., 1988: Climatic changes of the last 18,000 years - observations and model simulations. *Science*, **241**, 1043–1052.
- Kutzbach, J. E., G. Bonan, J. Foley, and S. P. Harrison, 1996: Vegetation and soil feedbacks on the response of the African monsoon to orbital forcing in the early to middle Holocene. *Nature*, **384**, 623–626.
- Kutzbach, J. E., and Z. Liu, 1997: Response of the African monsoon to orbital forcing and ocean feedbacks in the middle Holocene. *Science*, **278**, 440–443.
- Li, C. F., and M. Yanai, 1996: The onset and interannual variability of the Asian summer monsoon in relation to land sea thermal contrast. *J. Climate*, **9**, 358–375.
- Liew, P. M., C. Y. Lee, and C. M. Kuo, 2006: Holocene thermal optimal and climate variability of East Asian monsoon inferred from forest reconstruction of a subalpine pollen sequence, Taiwan. *Earth Planet Sci. Lett.*, **250**, 596–605.
- Liu, Z., B. L. Otto-Bliesner, J. E. Kutzbach, L. Li, and C. Shields, 2003: Coupled climate simulation of the evolution of global monsoons in the Holocene. *J. Climate*, **16**, 2472–2490.
- Liu, Z., S. P. Harrison, J. E. Kutzbach, and B. L. Otto-Bliesner, 2004: Global monsoons in the mid-Holocene and oceanic feedback. *Climate Dyn.*, **22**, 157–182.
- Neelin, J. D., and J. Y. Yu, 1994: Modes of tropical variability under convective adjustment and the Madden-Julian oscillation. Part I: Analytical theory. *J. Atmos. Sci.*, **51**, 1876–1894.
- Neelin, J. D., and Z. Zeng, 2000: A quasi-equilibrium tropical circulation

- model—Formulation. *J. Atmos. Sci.*, **57**, 1741-1766.
- Ohgaito, R., and A. Abe-Ouchi, 2007: The role of ocean thermodynamics and dynamics in Asian summer monsoon changes during the mid-Holocene. *Climate Dyn.*, **29**, 39-50.
- Qin, B. Q., S. P. Harrison, and J. E. Kutzbach, 1998: Evaluation of modelled regional water balance using lake status data: A comparison of 6 ka simulations with the NCAR CCM. *Quat. Sci. Rev.*, **17**, 535-548.
- Raynaud, D., J. Jouzel, J. M. Barnola, J. Chappellaz, R. J. Delmas, and C. Lorius, 1993: The ice record of greenhouse gases. *Science*, **259**, 926-934.
- Stevens, B., J. J. Duan, J. C. McWilliams, M. Munnich, and J. D. Neelin, 2002: Entrainment, Rayleigh friction, and boundary layer winds over the tropical Pacific. *J. Climate*, **15**, 30-44.
- Su, H., J. D. Neelin, and J. E. Meyerson, 2003: Sensitivity of tropical tropospheric temperature to sea surface temperature forcing. *J. Climate*, **16**, 1283-1301.
- Su, H., and J. D. Neelin, 2005: Dynamical mechanisms for African monsoon changes during the mid-Holocene. *J. Geophys. Res. Atmos.*, **110**, D19105, Doi: 10.1029/2005JD005806.
- Tan, P.-H., C. Chou, and J.-Y. Tu, 2008: Mechanisms of global warming impacts on robustness of tropical precipitation asymmetry. *J. Climate*, **21**, 5585-5602.
- Texier, D., N. de Noblet, S. P. Harrison, A. Haxeltine, D. Jolly, S. Joussaume, F. Laarif, I. C. Prentice, and P. E. Tarasov, 1997: Quantifying the role of biosphere–atmosphere feedbacks in climate change: coupled model simulation for 6000 years BP and comparison with palaeodata for northern Eurasia and northern Africa. *Climate Dyn.*, **13**, 865–882.
- Thevenon, F., D. Williamson, and M. Taieb, 2002: A 22 kyr BP sedimentological record of Lake Rukwa (8 degrees S, SW Tanzania): environmental, chronostratigraphic and climatic implications. *Paleogeogr Paleoclimatol Paleoecol*, **187**, 285-294.
- Toledo, F. A. L., K. B. Costa, and M. A. G. Pivel, 2007: Salinity changes in the western tropical South Atlantic during the last 30 kyr. *Glob. Planet Change*, **57**, 383-395.
- Vettoretti, G., W. R. Peltier, and N. A. McFarlane, 1998: Simulations of mid-Holocene climate using an atmospheric general circulation model. *J.*

Climate, **11**, 2607-2627.

- Wallace, J.M., E. M. Rasmusson, T. P. Mitchell, V. E. Kousky, E. S. Sarachik, and H. von Storch, 1998: The structure and evolution of ENSO-related climate variability in the tropical Pacific: Lessons from TOGA. *J. Geophys. Res. Oceans.*, **103**, 14241-14259.
- Xiao, J. L., Q. H. Xu, T. Nakamura, X. L. Yang, W. D. Liang, and Y. Inouchi, 2004: Holocene vegetation variation in the Daihai Lake region of north-central China: a direct indication of the Asian monsoon climatic history. *Quat. Sci. Rev.*, **23**, 1669-1679.
- Yu, J. Y., and J. D. Neelin, 1994: Modes of tropical variability under convective adjustment and the Madden-Julian oscillation. Part II: Numerical results. *J. Atmos. Sci.*, **51**, 1895-1914.
- Yu, J. Y., C. Chou, and J. D. Neelin, 1998: Estimating the gross moist stability of the tropical atmosphere. *J. Atmos. Sci.*, **55**, 1354-1372.
- Zeng, N., J. D. Neelin, and C. Chou, 2000: A quasi-equilibrium tropical circulation model—Implementation and simulation. *J. Atmos. Sci.*, **57**, 1767-1796.
- Zhao, Y., P. Braconnot, O. Marti, S. P. Harrison, C. Hewitt, A. Kitoh, Z. Liu, U. Mikolajewicz, B. L. Otto-Bliesner, and S. L. Weber, 2005: A multi-model analysis of the role of the ocean on the African and Indian monsoon during the mid-Holocene. *Climate Dyn.*, **25**, 777-800.
- Zhou, W. J., X. F. Yu, A. J. T. Jull, G. Burr, J. Y. Xiao, X. F. Lu, and F. Xian, 2004: High-resolution evidence from southern China of an early Holocene optimum and a mid-Holocene dry event during the past 18,000 years. *Quat. Res.*, **62**, 39-48.
- Zhou, W. J., S. C. Xie, P. A. Meyers, and Y. H. Zheng, 2005: Reconstruction of late glacial and Holocene climate evolution in southern China from geolipids and pollen in the Dingnan peat sequence. *Org. Geochem.*, **36**, 1272-1284.

TABLE CAPTIONS

TABLE 1: The experimental design. Since the other boundary conditions are set as pre-industrial condition, this table only focuses on the settings of insolation. For MH run and EXP runs, insolation is set as realistic 6 ka insolation parameters, while those in CTRL and EXP_CTRL runs are set as realistic 0ka insolation parameters. The difference between MH (CTRL) and EXP (EXP_CTRL) runs is the use of fully interactive boundary condition and prescribed F^{net} obtained from the CTRL experiment. In JJA_Dir (DJF_Dir) run, 6ka insolation is used in JJA (DJF), as emphasized by the gray bar in JJA_Dir run, while 0ka insolation is used in the other 9 months.

FIGURE CAPTIONS

FIGURE 1: Average boreal and austral summer insolation difference (Wm^{-2}) between 6 ka and the present. Solid line is for boreal summer (JJA) average and dashed line is for austral summer (DJF) average.

FIGURE 2: Precipitation anomaly of 16 PMIP2 models average between the mid-Holocene and the preindustrial modern for (a) JJA and (b) DJF climatology. Contour interval is 0.5 mm day^{-1} . Dark shading is above 0.5 mm day^{-1} and light shading is below -0.5 mm day^{-1} .

FIGURE 3: Precipitation difference between the mid-Holocene and pre-industrial control runs (**MH-CTRL**) for (a) JJA and (b) DJF climatology. Contour interval is 0.5 mm day^{-1} . Dark shading is above 0.5 mm day^{-1} and light shading is below -0.5 mm day^{-1} . Thick contour denotes **CTRL** run precipitation climatology of 7 mm day^{-1} precipitation.

FIGURE 4: Control experiment climatology difference (**MH-CTRL**) for JJA (a) surface temperature, (b) tropospheric temperature and (c) tropospheric moisture. (d), (e), (f) as in (a), (b), (c), respectively, but for difference in DJF. Contour interval for (a), (b), (d) and (e) is $0.2 \text{ }^{\circ}\text{C}$ with dark shading above $0.2 \text{ }^{\circ}\text{C}$ and light shading below $-0.2 \text{ }^{\circ}\text{C}$. Contour interval for (c) and (f) is 0.1 g Kg^{-1} , dark shading above 0.1 g Kg^{-1} and light shading below -0.1 g Kg^{-1} .

FIGURE 5: Moisture budget terms of the control experiment pair difference (**MH-CTRL**) in JJA: (a) $(\overline{M}_q \nabla \cdot \mathbf{v}'_1) \frac{P_T}{g}$, (b) $(M'_q \nabla \cdot \overline{\mathbf{v}}_1) \frac{P_T}{g}$, (c) $-\langle \mathbf{v} \cdot \nabla q \rangle' \frac{P_T}{g}$, and (d) evaporation difference. Contour interval is 10 W m^{-2} for (a) and 2 W m^{-2} for (b) with dark shading above 20 W m^{-2} and 4 W m^{-2} and light shading below -20 W m^{-2} and -4 W m^{-2} , respectively. Contour interval for (c) and (d) is 5 W m^{-2} , with dark shading above 10 W m^{-2} and light shading below -10 W m^{-2} . All terms on the right of

the moisture budget (7) shown here are in energy unit (Wm^{-2}).

FIGURE 6: As in Fig. 5, except for DJF.

FIGURE 7: As in Fig. 5, except for moist static energy budget terms: (a)

$$(\overline{M}\nabla \cdot \mathbf{v}_1') \frac{p_T}{g}, \text{ (b) } -(M\nabla \cdot \bar{\mathbf{v}}_1) \frac{p_T}{g}, \text{ (c) } -\langle \mathbf{v} \cdot \nabla (q + T) \rangle' \frac{p_T}{g}, \text{ and (d) } F^{\text{net}}.$$

Contour interval is 3 W m^{-2} with dark shading above 6 W m^{-2} and light shading below -6 W m^{-2} for (a), (b) and (d). Contour interval for (c) is 6 W m^{-2} with dark shading above 12 W m^{-2} and light shading below -12 W m^{-2} .

FIGURE 8: As in Fig. 7, except for DJF.

FIGURE 9: Precipitation difference under the mid-Holocene insolation forcing but with fixed longwave, sensible heat and latent heat fluxes (**EXP** minus **EXP_CTRL**) for (a) JJA and (b) DJF. Contour and shading are as in Fig. 3.

FIGURE 10: Direct JJA insolation effect: JJA difference between **JJA_Dir** run and **CTRL** run (**JJA_Dir** minus **CTRL**) for (a) precipitation (mm day^{-1}), (b) surface temperature, (c) tropospheric temperature and (d) tropospheric moisture. Contour interval and shading for precipitation are as that in Fig. 3. Contour interval for (b) and (c) is $0.2 \text{ }^\circ\text{C}$ with dark shading above $0.8 \text{ }^\circ\text{C}$ for (b) and $1.2 \text{ }^\circ\text{C}$ for (c). The entire tropical region has a positive temperature anomaly. Contour interval for (d) is 0.1 g Kg^{-1} with dark shading above 0.3 g Kg^{-1} .

FIGURE 11: As in Fig 10. but for direct DJF insolation effect: DJF difference between **DJF_Dir** run and **CTRL** run (**DJF_Dir** minus **CTRL**). Contour interval is as in Fig.10 with light shading below $-1 \text{ }^\circ\text{C}$ for (b), $-1.4 \text{ }^\circ\text{C}$ for (c) and -0.3 g Kg^{-1} for (d).

FIGURE 12: As in Fig. 10, but for the ocean memory effect of JJA insolation forcing in DJF climatology, namely, DJF difference between **JJA_Dir** run and **CTRL** run (**JJA_Dir** minus **CTRL**). Contour interval is as in Fig. 10 with dark shading above

0.6 °C for (b), 0.8 °C for (c) and 0.2 g Kg⁻¹ for (d).

FIGURE 13: As in Fig. 10, but for the ocean memory effect of DJF insolation forcing in JJA climatology, namely, JJA difference between **DJF_Dir** run and **CTRL** run (**DJF_Dir** minus **CTRL**). Contour interval is as in Fig.10 with light shading below -0.8 °C for (b), -1 °C for (c) and -0.3 g Kg⁻¹ for (d).

		MH	CTRL	EXP	EXP_CTRL	JJA_D		DJF_D	
		6 ka	0 ka	6 ka	0 ka	6 ka	0 ka	6 ka	0 ka
insolation	Jan.	v	v	v	v		v	v	
	Feb.	v	v	v	v		v	v	
	Mar.	v	v	v	v		v		v
	Apr.	v	v	v	v		v		v
	May	v	v	v	v		v		v
	Jun.	v	v	v	v	v			v
	Jul.	v	v	v	v	v			v
	Aug.	v	v	v	v	v			v
	Sep.	v	v	v	v		v		v
	Oct.	v	v	v	v		v		v
	Nov.	v	v	v	v		v		v
	Dec.	v	v	v	v		v	v	
Experiment boundary condition		Fully interactive	Fully interactive	Climatological F^{act}	Climatological F^{act}	Fully interactive		Fully interactive	

Table 1. The experimental design. Since the other boundary conditions are set as pre-industrial condition, this table only focus on the settings of insolation. For MH run and EXP runs, insolation are set as realistic 6 ka insolation parameters, while those in CTRL and EXP_CTRL runs are set as realistic 0ka insolation parameters. The difference between MH (CTRL) and EXP (EXP_CTRL) runs are the use of fully interactive boundary condition and a climatological F^{act} . In JJA_D and DJF_D runs, 6ka insolation are used in JJA and DJF, respectively, while 0ka insolation are used in the other 9 months, as emphasized by the gray bar in JJA_D run.

Fig. 1 Average boreal and austral summer insolation difference (Wm^{-2}) between 6 ka and the present. Solid line is for boreal summer (JJA) average and dashed line is for austral summer (DJF) average.

Fig. 2 Precipitation anomaly of 16 PMIP2 models average between the mid-Holocene and the preindustrial modern for (a) JJA and (b) DJF climatology. Contour interval is 0.5 mm/day. Dark shading is above 0.5 mm/day and light shading is below -0.5 mm/day.

Fig. 3 Precipitation difference between the mid-Holocene and pre-industrial control runs (**MH-CTRL**) for (a) JJA and (b) DJF climatology. Contour interval is 0.5 mm/day. Dark shading is above 0.5 mm/day and light shading is below -0.5 mm/day. Thick contour denotes **CTRL** run precipitation climatology of 7 mm/day precipitation.

Fig. 4 Control experiment climatology difference (**MH-CTRL**) for JJA (a) surface temperature, (b) tropospheric temperature and (c) tropospheric moisture. (d), (e), (f) as in (a), (b), (c), respectively, but for difference in DJF. Contour interval for (a), (b), (d) and (e) is 0.2 °C with dark shading above 0.2 °C and light shading below -0.2 °C. Contour interval for (c) and (f) is 0.1 g Kg⁻¹, dark shading above 0.1 g Kg⁻¹ and light shading below -0.1 g Kg⁻¹.

Fig. 5 Moisture budget terms of the control experiment pair difference (MH-CTRL) in JJA: (a) $(\bar{M}_q \nabla \cdot v'_1) \frac{p_T}{g}$, (b) $(M'_q \nabla \cdot \bar{v}_1) \frac{p_T}{g}$, (c) $-\langle v \cdot \nabla q \rangle' \frac{p_T}{g}$, and (d) evaporation difference. Contour interval is 10 W m^{-2} for (a) and 2 W m^{-2} for (b) with dark shading above 20 W m^{-2} and 4 W m^{-2} and light shading below -20 W m^{-2} and -4 W m^{-2} , respectively. Contour interval for (c) and (d) is 5 W m^{-2} , with dark shading above 10 W m^{-2} and light shading below -10 W m^{-2} . All terms on the right of the moisture budget (7) shown here are in energy unit (W m^{-2}).

Fig. 6 As in Fig. 5, except for DJF.

Fig. 7 As in Fig. 5, except for moist static energy budget terms:

(a) $(\bar{M}\nabla \cdot \bar{v}_1) \frac{p_T}{g}$, (b) $-(M\nabla \cdot \bar{v}_1) \frac{p_T}{g}$, (c) $-\langle \bar{v} \cdot \nabla (q+T) \rangle' \frac{p_T}{g}$, and (d) F^{net} .

Contour interval is 3 W m^{-2} with dark shading above 6 W m^{-2} and light shading below -6 W m^{-2} for (a), (b) and (d). Contour interval for (c) is 6 W m^{-2} with dark shading above 12 W m^{-2} and light shading below -12 W m^{-2} .

Fig. 8 As in Fig. 7, except for DJF.

Fig. 9 Precipitation difference under the mid-Holocene insolation forcing but with fixed longwave, sensible heat and latent heat fluxes (**EXP** minus **EXP_CTRL**) for (a) JJA and (b) DJF. Contour and shading are as in Fig. 3.

Fig. 10 Direct JJA insolation effect: JJA difference between **JJA_Dir** run and **CTRL** run (**JJA_Dir** minus **CTRL**) for (a) precipitation (mm day^{-1}), (b) surface temperature, (c) tropospheric temperature and (d) tropospheric moisture. Contour interval and shading for precipitation are as that in Fig. 3. Contour interval for (b) and (c) is $0.2\text{ }^{\circ}\text{C}$ with dark shading above $0.8\text{ }^{\circ}\text{C}$ for (b) and $1.2\text{ }^{\circ}\text{C}$ for (c). The entire tropical region has a positive temperature anomaly. Contour interval for (d) is 0.1 g Kg^{-1} with dark shading above 0.3 g Kg^{-1} .

Fig. 11 As in Fig 10. but for direct DJF insolation effect: DJF difference between **DJF_Dir** run and **CTRL** run (**DJF_Dir** minus **CTRL**). Contour interval is as in Fig.10 with light shading below -1°C for (b), -1.4°C for (c) and -0.3 g Kg^{-1} for (d).

Fig. 12 As in Fig. 10, but for the ocean memory effect of JJA insolation forcing in DJF climatology, namely, DJF difference between **JJA_Dir** run and **CTRL** run (**JJA_Dir** minus **CTRL**). Contour interval is as in Fig. 10 with dark shading above 0.6 °C for (b), 0.8 °C for (c) and 0.2 g Kg⁻¹ for (d).

Fig. 13 As in Fig. 10, but for the ocean memory effect of DJF insolation forcing in JJA climatology, namely, JJA difference between **DJF_Dir** run and **CTRL** run (**DJF_Dir** minus **CTRL**). Contour interval is as in Fig.10 with light shading below -0.8°C for (b), -1°C for (c) and -0.3 g Kg^{-1} for (d).