The Role of Nonlinear Drying above the Boundary Layer in the Mid-Holocene African Monsoon

VISHAL DIXIT, STEVEN SHERWOOD, OLIVIER GEOFFROY, AND DAMIANOS MANTSIS

Climate Change Research Centre, University of New South Wales, Sydney, New South Wales, Australia

(Manuscript received 11 April 2017, in final form 29 September 2017)

ABSTRACT

Paleoclimatic proxies indicate that significant summertime rainfall reached the Sahara region during the mid-Holocene, presumably in response to stronger summertime heating in the Northern Hemisphere. Climate models generally do not replicate the enhanced precipitation. As a step toward understanding the response and possible role of model errors, a series of idealized experiments were conducted with the Community Earth System Model in which local atmospheric heat sources of increasing magnitude were applied in the boundary layer over the Sahel and Sahara. In response to this local heating, the cold and moist southwesterly monsoon inflow encroaches farther northward. A source strength of roughly 1 K day^{-1} produces responses similar to those in a simulation with mid-Holocene orbital forcing imposed globally, while that of 1.5 K day^{-1} produces a precipitation response similar to that from paleoproxies. The precipitation increases nonlinearly, with a jump at heating of around 1 K day^{-1} , even though the low-level monsoon inflow increases linearly. Competition at low-to-middle levels between drying by a shallow return flow just above the boundary layer and moistening by vertical advection within the layer affects convection and determines the northward extension of precipitation. When the heating becomes 1.5 K day^{-1} , the boundary layer flow encroaches sufficiently northward to weaken the shallow return flow, further aiding precipitation. This novel nonlinear mechanism operates without biogeophysical feedbacks, and suggests that poor representation of the local thermodynamic processes may hamper a model's ability to simulate dynamical feedbacks and hence the strength and poleward extension of monsoon rains under forcings like those during the mid-Holocene.

1. Introduction

Paleoclimatic proxies indicate that the African monsoon was stronger and extended farther northward over the Sahara during the early to mid-Holocene [$\sim 5\text{--}6 \text{ kyr ago}$ ($1 \text{ kyr} = 1000 \text{ yr}$); see review by deMenocal et al. (2000)]. While it is still debated whether the transition from the mid-Holocene moist and vegetated state to the present-day state of desert occurred rapidly or gradually (Kröplin et al. 2008; Brovkin and Claussen 2008), it is agreed that the equilibrium state during the mid-Holocene had a moist and vegetated “green” Sahara. Proxies suggest that a significant enhancement in the precipitation ($\sim 300\text{--}400 \text{ mm yr}^{-1}$, which amounts to increase of around $2\text{--}3 \text{ mm day}^{-1}$ during the monsoon

season) occurred over the Sahel–Sahara region during the mid-Holocene (Harrison et al. 2015). These changes have been shown to be associated with the increase in the summer top-of-atmosphere solar insolation that occurred in response to the changes in Earth's orbit during the mid-Holocene (Kutzbach 1981; Kutzbach and Liu 1997). The mechanism of how the modulations in the insolation manifest as changes in the monsoon circulation and precipitation has been an unresolved climate “enigma” since its discovery.

Researchers have used models of various complexities to investigate the processes responsible for Saharan greening. Initial studies argued that the green Sahara is a manifestation of the strong feedback between biogeophysical processes and atmospheric dynamics (Kutzbach et al. 1996; Broström et al. 1998; Claussen et al. 2002). Claussen and Gayler (1997) used a coupled atmosphere–vegetation GCM to show that the vegetation–albedo feedback is important, and the equilibrium state of mid-Holocene climate can be sensitive to the assumed initial conditions over land. Kutzbach and Liu (1997) and Hewitt

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-17-0234.s1>.

Corresponding author: Vishal Dixit, v2dixit@gmail.com

and Mitchell (1998) argued that ocean–atmospheric coupling could be important in the simulation of a green Sahara. Subsequently a number of researchers, conducting simulations of the mid-Holocene climate, have found support for either or both of these mechanisms. These studies motivated the Palaeoclimate Modelling Intercomparison Project phase 1 (PMIP; Joussaume and Taylor 1995). These models were atmosphere–land or atmosphere–land–slab ocean models and had a dynamic vegetation feedback included in them. The latter revisions of this effort (PMIP2 and CMIP5/PMIP3; Braconnot et al. 2007, 2012) included both atmosphere–land models as well as fully coupled atmosphere–ocean–land climate models with interactive vegetation dynamics. These models do not reproduce the expected enhancement in the strength and northward expansion of the precipitation zone over the Sahara (Harrison et al. 2015). This problem has been recognized since PMIP but still persists in the latest CMIP5/PMIP3 models.

Some studies have suggested more novel mechanisms. Chikira et al. (2006) argued that moist convection originating above the boundary layer over the desert is an important process, and that the strength of precipitation over the Sahara due to biogeophysical feedbacks is enhanced when this process is included in their climate model. Carrington et al. (2001) argued that the green Sahara had significant wetlands. They demonstrated the need to represent the wetlands and their effect on precipitation using a GCM. Swann et al. (2014) suggested that remote vegetation feedback due to Eurasian forest cover is important to enhance precipitation over the northward edge of the monsoon in their GCM. Pausata et al. (2016) reduced the concentration of dust over the Sahara consistent with the increased vegetation in the region in their GCM and found that this played role in enhancing the precipitation over the Sahara. They mentioned that the latest CMIP5/PMIP3 models do not incorporate the effect of reduced dust in their initial conditions. These studies identified factors that affected the mid-Holocene African monsoon in their simulations, but did not offer any mechanistic insight about how the circulation and associated moisture transports changed from the present-day climate to a green Sahara climate in their simulations.

Su and Neelin (2005) proposed a moist static energy (MSE) ventilation mechanism for Saharan greening. They argued that the northward extent of monsoons is limited by import of low MSE by flow from adjacent regions. Hales et al. (2006) further demonstrated that vegetation feedback is important to maintain a green Sahara once it has been produced, but is not able to produce it, while the Su and Neelin (2005) ventilation mechanism is important to produce it. These studies

used a medium-complexity quasi-equilibrium tropical climate model (QTCM) that had a simple vertical structure with only one baroclinic and one barotropic mode. Hence it is not clear how the import of low MSE is actually brought about by the atmospheric circulation, or how it could change in response to mid-Holocene insolation conditions. Patricola and Cook (2007) noted that in their regional model without interactive vegetation dynamics the specified vegetation played an important role and in the simulated green Sahara climate the low-level African easterly jet was completely eliminated. They noted that the changes in the flow decreased dry static energy (DSE) while increasing moisture, thereby maintaining a boundary layer MSE maximum over the desert in their mid-Holocene simulation; however, they did not diagnose if the complex vertical structure of the flows played an important role in the energy ventilation or if this ventilation mechanism changed under mid-Holocene forcing. Patricola and Cook (2008) observed that precipitation over the Sahel and Sahara increased nonlinearly in their regional model when they forced it with different latitudinal extent of the prescribed vegetation. They also noted that the precipitation change was associated with a nonlinear change in the boundary layer monsoonal flow and the low-level African easterly jet.

Recent development in theoretical understanding has found that poleward excursions of the large-scale precipitation zones (i.e., the ITCZ and monsoons) occur in association with meridional energy transports and are known to be sensitive to the interhemispheric difference in energy input in the atmosphere (Schneider et al. 2014). The zonal-mean precipitation has been shown to shift toward the differentially warmed hemisphere as a response to an imposed energy forcing at the surface (Kang et al. 2008). A recent study by Boos and Korty (2016) showed that because monsoons are regional, asymmetries due to zonal energy transports could be as important as meridional transports. They note that a simple energetic constraint may not be able to explain the mid-Holocene green Sahara. Moreover, they indicate that an additional large energy source is required to explain the expected increase in precipitation in a green Sahara. The energy input in terms of the heating in the atmospheric column over the desert is affected by surface fluxes, boundary layer turbulent processes, and differential heating due to aerosols and midlevel clouds. These processes mainly alter the net heating efficiency of the solar insolation in the boundary layer. Roehrig et al. (2013) demonstrated that many GCMs have systematic biases in the simulation of these processes over the Sahel–Sahara region. They specifically note biases in the simulation of the surface fluxes and midlevel clouds

that would strongly affect the efficiency of the boundary layer heating. A recent paper by [Chadwick et al. \(2017\)](#) demonstrated that biases in diabatic heat sources significantly affect the ability of climate models to simulate the northward extent of the African monsoonal circulation in the present climate. This suggests that the response in the mid-Holocene could also be highly sensitive to boundary layer heating efficiency and its changes.

In this study, we focus on the effect of regional differential atmospheric and surface heating (energy input) on the northward extent and strength of the African monsoon. In doing so we aim to diagnose the interaction between the energy input and changes in the circulation that brings about the precipitation change in the Sahel–Sahara region, while holding fixed other potentially important influences such as the vegetation, sea surface temperature, and aerosols. We impose idealized local heat sources that could represent a combination of external diabatic forcings and internally generated heat sources that might be missing from current models. We first describe the details of the model and experiments in [section 2](#). The results are presented in [section 3](#) and the mechanism of changes in circulation is discussed in [section 4](#). The important conclusions and discussion are presented in [section 5](#).

2. Model and simulation details

We have used the Community Earth System Model (CESM1.2.2) developed by the National Center for Atmospheric Research (NCAR). This model consists of the Community Atmosphere Model (CAM5.1; see [Neale et al. 2010](#)), Community Land Model (CLM4; see [Oleson et al. 2010](#)), and prescribed sea ice and sea surface temperature. The model uses the [Zhang and McFarlane \(1995\)](#) deep convection scheme and the [Bretherton and Park \(2009\)](#) shallow convection scheme. The Rapid Radiative Transfer Model for GCMs (RRTMG) radiation code is used to calculate the radiative fluxes. The model uses a finite-volume dynamical core at $1.9^\circ \times 2.5^\circ$ horizontal resolution and 31 vertical levels with hybrid coordinates. All simulations used fixed climatological SST (FC5 compset) derived from the merged Hadley Centre–NOAA sea surface temperature and sea ice concentration ([Hurrell et al. 2008](#)). The vegetation and land surface conditions are fixed to the present climate in all simulations.

To check the sensitivity of the results to the model physics, we also conducted experiments with CESM1.2.2 but with Community Atmosphere Model, version 4 (CAM4), physics [FC4 compset; see [Gent et al. \(2011\)](#) for more details]. [Neale et al. \(2013\)](#) have documented the

important features in the treatment of thermodynamic processes and the control climate simulated in this model. This setup uses the [Hack et al. \(1993\)](#) shallow convection scheme and the CAM4 radiation scheme (CAMRT). Apart from these important differences in the treatment of thermodynamic processes, we kept the methodology and experiments identical to the CAM5.1 simulations described below (see Table S1 in the supplemental material).

The series of simulations was as follows. A control simulation was conducted to simulate the strength and northward extent of African monsoons in present climate. Next, a mid-Holocene simulation was performed to assess the models response to mid-Holocene insolation forcing with present-day climatological SST maintained. The orbital parameters (eccentricity, longitude of perihelion, and obliquity) were changed following the PMIP3 protocol to simulate the mid-Holocene insolation ([Braconnot et al. 2012](#)). [Figure 1](#) shows the incoming solar insolation in the present day (Control) climate and the mid-Holocene anomalies (mid-Holo minus present-day). During the mid-Holocene, the longitude of perihelion occurred during Northern Hemisphere summer ([Braconnot et al. 1999](#); [Zhao et al. 2005](#)) in contrast to its present occurrence in winter. As a result the Northern Hemispheric tropics and subtropics received around 20 W m^{-2} more insolation during mid-Holocene summer than today. These simulated changes in the top of the atmosphere insolation are consistent with the previous PMIP simulations ([Braconnot et al. 2000](#)).

To study further the sensitivity of the African monsoons to mid-Holocene forcing, simulations with imposed regional heating were conducted. The heating was designed to represent possible localized errors in the boundary layer heating. We did not further consider remote influences. The mid-Holocene increase in the solar insolation in the region (20 W m^{-2}) roughly corresponds to an imposed heating of 1 K day^{-1} in the 1.5-km air layer near the surface. The heating was imposed throughout the simulation, in the radiative heating subroutine, within a rectangular box $15^\circ\text{--}35^\circ\text{N}$, $20^\circ\text{W}\text{--}30^\circ\text{E}$ (the numerical code to implement the changes is available in the supplemental material). The heating was added uniformly between the surface and 850 hPa. The strength of the regional heat source was increased from 0 to 2.5 K day^{-1} in increments of 0.5 K day^{-1} ; the mid-Holocene increase in the solar insolation in the region (20 W m^{-2}) roughly corresponds to the 1 K day^{-1} case in terms of anomalous energy input. Each simulation was run for 10 years starting from the same initial conditions (provided with the FC5 compset for the present climate). The details of these simulations can be found in [Table 1](#).

FIG. 1. The annual evolution of the zonal-mean top-of-atmosphere incoming insolation for (a) the Control simulation (W m^{-2}) and (b) the change for the mid-Holocene simulation (mid-Holo – Control; W m^{-2}).

To evaluate the ability of the model to simulate the present-day monsoonal circulation, we compare results to the monthly mean climatology (1979–2013) of the 3D wind field from the ERA-Interim dataset (Dee et al. 2011). The monthly mean GPCP data (1979–2013) were used to calculate the climatology of precipitation over the African region (Adler et al. 2003).

3. Results

a. Precipitation

The average precipitation during June–September (JJAS) in Control and its change in the mid-Holo simulation are analyzed to evaluate respectively the ability of the model to reproduce the current rainfall distribution

or that inferred for the mid-Holocene (Fig. 2a). The Control simulation reproduces all the qualitative features of the seasonal mean precipitation in the region observed in the present climate (Fig. 3). The peak in the precipitation (of magnitude 7 mm day^{-1}) is located at around 10°N . The model overestimates the precipitation over the Sahel–Sahara region but underestimates it to the south of this region. In both the mid-Holo and Control simulations, precipitation decreases rapidly on both the equatorward and poleward sides of the peak, similar to GPCP.

Consistent with expectations, the maximum precipitation increases slightly in the mid-Holo simulation and the northward flank of the peak expands toward the Sahara. A zonally coherent increase of around 1 mm day^{-1}

TABLE 1. The details of the numerical experiments performed with CAM5.1 in this study. All simulations were run for 10 years and monthly means were analyzed.

Simulation No.	Model	Simulation name	Orbital conditions	Heating box
1	CAM5.1	Control (heating 0 K day^{-1})	Present	—
2	CAM5.1	Mid-Holo	Mid-Holocene	—
3	CAM5.1	Heat 0.5 K day^{-1}	Present	$15^\circ\text{--}35^\circ\text{N}$, $20^\circ\text{W--}0^\circ$, 850 hPa–surface
4	CAM5.1	Heat 1.0 K day^{-1}	Present	$15^\circ\text{--}35^\circ\text{N}$, $20^\circ\text{W--}30^\circ\text{E}$, 850 hPa–surface
5	CAM5.1	Heat 1.5 K day^{-1}	Present	$15^\circ\text{--}35^\circ\text{N}$, $20^\circ\text{W--}30^\circ\text{E}$, 850 hPa–surface
6	CAM5.1	Heat 2 K day^{-1}	Present	$15^\circ\text{--}35^\circ\text{N}$, $20^\circ\text{W--}30^\circ\text{E}$, 850 hPa–surface
7	CAM5.1	Heat 2.5 K day^{-1}	Present	$15^\circ\text{--}35^\circ\text{N}$, $20^\circ\text{W--}30^\circ\text{E}$, 850 hPa–surface

FIG. 2. (a) The climatology of precipitation (mm day^{-1}) in GPCP data and its simulation during JJAS for the Control and mid-Holo simulations over the African region (20°W – 30°E) and the difference between them. (b) The average precipitation anomaly over the African region for different heating simulations. (c) The mean precipitation anomaly over the Sahel and Sahara as a function of imposed boundary layer heating rate.

on the northern flank as far north as 25°N can be observed (see Fig. S5 in supplemental material). Although the sign of the difference is as expected, the magnitude and extent are not consistent with the paleoproxies. In this sense, the mid-Holo experiment suffers from a similar deficiency as previously reported in PMIP3 simulations (Harrison et al. 2015). The paleoproxies suggest a marked increase of around 300–400 mm in the seasonally accumulated precipitation during mid-Holocene summers compared to today (Harrison et al. 2015). This amounts to approximately 2 – 3 mm day^{-1} (assuming JJAS as the length of the rainy season) of additional precipitation

to the north of the main precipitation zone (i.e., between 15° and 25°N), well in excess of that simulated.

The regional heating simulations show an increase in precipitation to the north of the main precipitation zone (Fig. 2b), similar to the mid-Holo experiment. As the heating is increased from 0 to 2.5 K day^{-1} , the precipitation to the north of the precipitation zone increases. The peak increase occurs between 15° and 20°N , close to the latitudes estimated from paleoproxies (Harrison et al. 2015). The amplitude of the rainfall anomaly is consistent with paleo-proxy estimates (2 mm day^{-1}) for the 1.5 and 2 K day^{-1} heating rates but not the 1.0 K day^{-1} heating rate.

Climatology : JJAS Precipitation (mm/day)

FIG. 3. The comparison of the spatial distribution of the precipitation climatology (mm day^{-1}) for JJAS over the African region simulated in (left) the Control simulation and (right) the GPCP data.

The precipitation in the region of interest increases monotonically with heating rate, but in a nonlinear fashion (Fig. 2c). When the heating is below 1 K day^{-1} , the increase is relatively slow, and then jumps at around $1\text{--}1.5 \text{ K day}^{-1}$. The response in the mid-Holo simulation is comparable to that of the 1 K day^{-1} heating simulation, albeit with some differences south of the precipitation zone (around $5^{\circ}\text{--}10^{\circ}\text{N}$) where precipitation decreases in idealized heating experiments. Since 1 K day^{-1} is roughly equivalent to the local heat input from mid-Holocene forcing, and produces a similar Sahel–Sahara precipitation response, we conclude that the northward extension of precipitation is driven mainly by local heating, and that errors in this extension could be due to errors in local heat sources.

b. Circulation

To understand the nonlinear response of precipitation north of the main rainband, we examine carefully the circulation accompanying these shifts. First, we evaluate the circulation in the Control simulation by comparing the zonal average meridional cross section of the vertical and meridional winds over the African region to reanalysis data (Fig. 4). The simulation captures all the key features of the zonal winds over African region. The westerly monsoonal flow in the boundary layer extends to 20°N in the reanalysis and the simulation. Previous studies (Thorncroft et al. 2011; Nicholson 2013; Pu and Cook 2010, 2012) have reported a monsoon jetlike structure in the boundary layer. A closed contour with a strength of 6 m s^{-1} is simulated indicating the presence of a jet in the simulation. Above the boundary layer in the lower troposphere, a low-level African easterly jet

(AEJ) is simulated at around 500–700 hPa at 15°N . In the upper troposphere, a tropical easterly jet (TEJ) is simulated.

The anomalies in the mid-Holocene simulation (Fig. 5b) show that the monsoonal boundary layer flow extends northward and upward. This is consistent with findings of previous mid-Holocene simulations (e.g., Patricola and Cook 2007, 2008). The low-level African easterly jet weakens and the upper tropospheric easterlies strengthen. Similarly, as the strength of heating is increased (Figs. 5c–f), the vertical and northward extent of the monsoonal boundary layer flow increases in rough proportion to the heating. For a heating of less than 1 K day^{-1} , the anomalies in the monsoonal boundary layer flow are less than in the mid-Holocene simulation. For a heating of 1.5 K day^{-1} or more, the monsoonal flow increases significantly and is at least 2 m s^{-1} more than in the mid-Holocene simulation.

The meridional cross section of meridional winds in the Control simulation (Fig. 6a) shows a southerly monsoonal flow extending to 15°N where it meets the northerly winds (the latter are referred to as Harmattan winds in the literature; Thorncroft et al. 2011). The location where the southerlies and northerlies meet is referred to as the ITCZ by Nicholson (2013). A secondary circulation forms over the Sahara where northerlies rise above the monsoonal southerlies. This forms a second branch of the circulation that penetrates into the monsoon precipitation zone from the north in the lower troposphere, and reaches a maximum at around 750 hPa (Zhang et al. 2008). We refer to the monsoon rainband as the precipitation zone, and the circulation to its north, which involves the monsoon and the Harmattan winds,

FIG. 4. The comparison of climatology of circulation during JJAS over African region (20°W – 30°E) simulated in (left) the Control simulation and (right) the ERA-Interim data. The vertical velocity is multiplied by 100 so that vertical motion is seen. The shading shows the zonal winds (m s^{-1}) and the vectors show the meridional circulation. The reference vector shows a meridional wind of 4 m s^{-1} and a vertical pressure velocity of 0.004 Pa s^{-1} .

as the shallow meridional circulation (SMC). The northerly flow above the boundary layer we refer to as the shallow return flow. In the Control simulation these features are captured well (Fig. 4). The depth of the shallow circulation is overestimated in Control (below 350 hPa) as compared to the reported depth (below 500 hPa) in reanalysis (Zhang et al. 2008) and observations (Zhang et al. 2006).

The mid-Holocene simulation (Fig. 6b) shows a slight strengthening of the southerly monsoon flow in the boundary layer (by 1 m s^{-1}) and the return flow in the upper troposphere. In the heating simulations (Figs. 6c–f), the strength and northward extent of the monsoonal southerlies increases with the strength of heating. Interestingly, when the heating is 1 K day^{-1} , the monsoon flow is stronger than in the mid-Holocene simulation even though both show similar increases in precipitation (as seen in Fig. 2). This suggests that the strength and extent of the monsoon flow may not be the only factor determining precipitation anomalies, and that the circulation response may be reduced by remote heating in mid-Holo that is absent in the regional-heating simulations.

In summary, the meridional cross section of zonal and meridional winds shows four important features of the African monsoon circulation: the southerlies, and the westerlies (sometimes jetlike), in the boundary layer

that form the classical monsoon flow; the shallow return flow; and the tropical easterly jet just above the boundary layer. We now investigate if the nonlinear dependence of precipitation on heating rate is associated with any of these features (Fig. 7).

There is approximately a linear relationship between the boundary layer westerly flow over the Sahel and Sahara (15° – 25°N , 20°W – 30°E ; north of precipitation zone) and the imposed heating (Fig. 7a). Each increase in heating by 0.5 K day^{-1} increases the zonal flow by around 0.7 m s^{-1} . The boundary layer southerly flow also increases linearly with the heating (Fig. 7b).

The strength of the AEJ decreases in response to the imposed heating. Interestingly, the nonlinear nature of the precipitation anomaly is similar to the nonlinear decrease in the strength of the AEJ (Fig. 7c, red dots). The initial increase in strength from 0 to 0.5 K day^{-1} is due to the choice of averaging latitudes and the corresponding shift in monsoon circulation. When the latitude of averaging is centered on the jet core (Fig. 7c, blue asterisks), a monotonic decrease is observed. Thorncroft and Blackburn (1999) and Cornforth et al. (2009) have shown that the AEJ forms as a response to the meridional geopotential gradient set by the combination of dry convection to the north and moist convection in the monsoon precipitation zone. The AEJ and

W Africa (20W–30E) mean U / U_anom (m/s)

FIG. 5. (a) The climatological meridional cross section of the mean zonal wind (m s^{-1}) for the Control simulation and (b)–(f) zonal wind anomalies (simulation – Control) averaged over African region (20°W–30°E) for JJAS for different simulations indicated by the legend. [The black (green) box in (a) marks the region chosen for averaging in Fig. 7a (Fig. 7c).]

shallow meridional circulation are part of same Saharan heat-low circulation (Thorncroft et al. 2011; Shekhar and Boos 2017), calling attention to the shallow return flow above the boundary layer. This flow also changes nonlinearly with the imposed heating (Fig. 7d). As the heating is increased from 0 to 1 K day^{-1} , the strength of this flow increases linearly, but when the heating crosses the 1 K day^{-1} threshold the flow begins rapidly decreasing (this nonlinearity is not an artifact of choice of averaging latitudes, as shown by the similarity between red and blue marks). This suggests that the precipitation anomaly in the Sahel–Sahara region could be strongly modulated by the shallow return flow and/or the AEJ.

In the CAM4 simulations we again found a nonlinear response to the imposed heating in the precipitation, shallow return flow, and AEJ, except that the nonlinearity was weak (see Fig. S6 in the supplemental material and Fig. 8). The climatology of the African

monsoon circulation was significantly better in CAM5.1 than in CAM4 (see Fig. S7 in the supplemental material). This may explain the differences in the nonlinearity in these models.

We now focus on the mechanism behind the nonlinear precipitation increase in the CAM5.1 simulations. It is possible that the Saharan heat low, shallow circulation, and AEJ are all part of the same circulation system that is modulated by the forced heating in the simulations. This possibility is explored in subsequent analysis through energy and moisture budgets.

4. Mechanism of nonlinear precipitation increase

a. Energetics

The energetic theory of monsoons suggests that the northward limit of precipitation zone is associated with the maximum of the boundary layer MSE (Privé and

W Africa (20W–30E) mean V / V_{anom} (m/s)

FIG. 6. As in Fig. 5, but for the mean meridional wind and meridional wind anomalies. [The black (green) box in (a) marks the region chosen for averaging in Fig. 7b (Fig. 7d).]

Plumb 2007). The boundary layer MSE maximum is pushed northward at around 20°N in both the mid-Holo and heating simulations (Fig. 8, panels 2d and 3d), which is consistent with the energetic theory. We now diagnose the individual contributions to the MSE change by focusing on temperature, moisture, and geopotential changes.

In the mean state of the Control simulation (not shown), the boundary layer temperature is highest in the desert region north of 15°N. Consistent with observations, temperatures over the desert are at least 6 K higher than over the equatorial Atlantic. The mid-Holocene (Fig. 8, panel 1a) free troposphere is warmer than in Control by at least 1 K. However, although the summer solar insolation is greater in the mid-Holocene, the boundary layer north of the precipitation zone (between 15° and 25°N) is, counterintuitively, colder. The regional heating simulations (Fig. 8, panels 2a and

3a) produce these same responses, with the boundary layer cooling and free-tropospheric warming each increasing with heating strength. The boundary layer cooling shifts and strengthens northward, and the cold anomaly tilts southward with height, consistent with the penetration of the monsoonal low that brings cold air in the boundary layer over the Sahara (Figs. 5 and 6). The changes in geopotential height are consistent with the temperature changes (Fig. 8, panels 1c, 2c, and 3c). The mid-Holocene simulation shows a decrease in geopotential height over the Sahel–Sahara region. As the boundary layer monsoon flow penetrates northward, it cools the layer, and hence decreases the geopotential near the layer top.

The Control boundary layer moisture (not shown) is greatest in the monsoon zone, consistent with observations. The specific humidity decreases rapidly to the north of the precipitation zone and more gradually to

FIG. 7. The circulation in the Sahel–Sahara region (15°–25°N, 20°W–30°E; red dots) for (a) the boundary layer zonal wind (averaged between 800 and 1000 hPa), (b) the boundary layer meridional wind (averaged between 800 and 1000 hPa), (c) the AEJ (averaged between 500 and 700 hPa), and (d) the shallow return flow (averaged between 500 and 700 hPa). The blue asterisks in (c) and (d) show the values averaged with reference to the AEJ (10°–20°N).

the south. Consistent with the structure of precipitation and temperature, the strength and extent of the moisture anomaly increase with heating rate (Fig. 8, panels 2b and 3b). The mid-Holocene moisture anomalies fall between those of the 1 and 1.5 K day^{−1} heating simulations (Fig. 8, panel 1b).

Local heating in the boundary layer over the Sahara region deepens the surface low (pressure) linearly (Fig. S1 in the supplemental material), accounting for the linear increase in strength of the monsoon flow. Associated with this is feedback through the advection from the adjacent regions (from the south and west), which cools and moistens the boundary layer. The reduced MSE imported due to cold advection is less than the increased MSE imported due to moisture advection; thus, in the net, the boundary layer MSE increases to the north of the precipitation zone (Fig. 8,

panels 1d, 2d, and 3d). The maximum MSE changes occur over the Sahara region (20°–30°N), pushing the boundary layer MSE maximum northward. The qualitative similarity in MSE changes noted in the mid-Holocene and heating simulations again suggests that the changes in circulation are captured in the idealized heating simulations.

In summary, the MSE maximum shifted northward both in the mid-Holo and regional heating simulations consistent with the theory. At the latitude of maximum MSE increase, both temperature and geopotential tend to decrease while humidity strongly increases. Hence, the shift in MSE maximum is mainly due to increase in the moisture to the north of the main precipitation zone. We now investigate the role of moist processes in controlling the precipitation using the moisture budget.

Africa (20W–30E, 15N–25N) Climatology for JJAS

FIG. 8. (a) The climatological meridional cross section of anomalous temperature (K), (b) moisture (kg kg^{-1}), (c) geopotential ($\text{m}^2 \text{s}^{-2}$), and (d) MSE (J kg^{-1}) averaged over the African region (20°W – 30°E). (top)–(bottom) The anomalies are shown for the mid-Holo (row 1), Heat 1 K day^{-1} (row 2), and Heat 1.5 K day^{-1} (row 3) simulations, respectively.

b. Moisture budget and nonlinear drying above the boundary layer

We calculated the vertically integrated moisture budget over the African region. The time-mean budget can be written

$$u \frac{\partial q}{\partial x} + v \frac{\partial q}{\partial y} + \omega \frac{\partial q}{\partial p} = E - P + \text{Residual}, \quad (1)$$

where q is specific humidity, E is evaporation, P is precipitation, and u , v , and ω are the zonal, meridional, and vertical pressure velocity, respectively. In the precipitation zone where $P - E > 0$, the dominant balance

occurs between the moistening due to the vertical advection and evaporation and drying due to the precipitation ($\omega \partial q / \partial p = E - P$; see Fig. 9). On the northward edge of the precipitation zone where $P = E$, the moistening due to vertical advection is largely balanced by drying due to meridional advection ($\omega \partial q / \partial p = -v \partial q / \partial y$). This shows that the moistening due to vertical advection is ineffective in producing precipitation if the drying due to meridional advection counterbalances it.

We further calculated the vertical profiles of mass-weighted moisture tendencies due to advective terms. On the northward edge of the precipitation zone, the

FIG. 9. The climatological mean terms of vertically integrated moisture budget [W m^{-2} ; calculated in energy units from Eq. (1)] averaged over the African region (20°W – 30°E) for JJAS.

column moistening due to vertical advection is balanced mainly by drying due to meridional advection between 500 and 700 hPa (Fig. S4 in supplemental material). These are the same levels where we have previously noted the nonlinear evolution of the shallow return flow. The vertical advection in this layer is affected by the import of moisture in the boundary layer by the monsoon flow. As the monsoon flow encroaches northward, the balance alters between relative moistening in the boundary layer (due to low-level monsoon jet or the low-level westerly jet; Pu and Cook 2010, 2012) and drying above it due to the shallow return flow. The strength of boundary layer flow depends on the depth of the surface low, which changes linearly with the heating (Fig. S1). Hence, the monotonic increase in the precipitation can be attributed to the linear increase in the moisture import. However, the nonlinearity is associated with the shallow return flow response and associated drying.

To investigate in detail the relative effects of drying due to shallow return flow and moistening due to vertical advection, we analyzed the relative magnitude of total advective moisture tendency between 500 and 700 hPa in different simulations (Fig. 10). A net drying tendency [positive values, total value of the left-hand side in Eq. (1)] can be observed until the heating magnitude reaches 1 K day^{-1} , beyond which the advective tendency becomes moistening (negative values). This transition is abrupt and is related to the nonlinear drying associated with the shallow return flow. This advective nonlinearity manifests in nonlinearity of the precipitation response.

FIG. 10. The variation of total moisture advection tendency (W m^{-2} ; $[u\partial q/\partial x + v\partial q/\partial y + \omega\partial q/\partial p]$, mass-weighted average between 500 and 700 hPa, as indicated by square brackets, converted to energy units) as a function of heating in different heating simulations.

The northerly shallow return flow is driven by the geopotential difference between the desert and the precipitation zone. In the following subsection, a potential temperature budget is calculated to diagnose the processes leading to temperature and geopotential changes.

c. Potential temperature budget and nonlinear evolution of shallow return flow

The steady state potential-temperature budget averaged zonally over the African region can be written as

$$u \frac{\partial \theta}{\partial x} + v \frac{\partial \theta}{\partial y} + \omega \frac{\partial \theta}{\partial p} = \text{DTCOND} + \text{RAD} + \text{Residual}, \quad (2)$$

where θ denotes the potential temperature, DTCOND is the heating due to all moist processes, and RAD is the net radiative heating. The residual includes all the terms not considered explicitly (turbulent diffusion, effect of transient eddies, and the applied heating perturbation). In the Control simulation (Fig. 11) the main balance in this budget occurs between the heating due to moist processes (Fig. 9e) and adiabatic cooling due to vertical advection (Fig. 11c) in the free troposphere. In the region of descent (south of the equator and north of 25°N), adiabatic warming due to vertical advection is balanced by radiative cooling (Fig. 9d).

In the boundary layer below 700 hPa, advection and residual processes also play important roles. Zonal (Fig. 11a)

Ctrl : W Africa (20W–30E) Theta budget Climatology

FIG. 11. (a)–(f) The climatological meridional cross section of the important terms from Eq. (2) of the equivalent potential temperature budget (K s^{-1}) averaged over the African region (20°W – 30°E) for JJAS for the Control simulation.

and meridional (Fig. 11b) advection each cool the boundary layer below 850 hPa, while the shallow return flow (700 hPa; Fig. 11b) warms the layer just above this near 20°N . The residual processes are most prominent near the surface layer where turbulent diffusion is important. As idealized heating is introduced, the horizontal advection in the boundary layer changes. Meridional advection by the southerly flow cools the boundary layer and warms the layer above it when heating is less than 1 K day^{-1} (Fig. S2 in supplemental material). When the heating surpasses 1 K day^{-1} these advective tendencies change sign (Fig. S3 in supplemental material), suggesting a complex coupling between temperature advection and geopotential and hence on zonal and meridional velocity in and above the heating layer.

As the monsoon flow encroaches northward in response to the heating, the associated cooling reduces the layer-mean temperature and hence the geopotential from 500 to 700 hPa in the precipitation zone, which

increases the geopotential difference driving the shallow flow (as seen in Fig. 7 and illustrated in Fig. 12b). When the heating exceeds 1 K day^{-1} , the encroaching monsoon flow reaches the latitude of the origin of the northerly dry shallow flow in the desert at around 20°N , the associated cooling of the layer mean temperature reduces the geopotential difference and likewise the strength of the shallow circulation significantly (Figs. 7 and 12c). This explains the nonlinear evolution of the shallow flow and its effect on the evolution of precipitation.

5. Discussion and conclusions

The mid-Holocene greening of the Sahara is a long-standing problem, as evidenced by its stubbornness across several generations of PMIP. Most models produce positive Sahel–Sahara precipitation anomalies in mid-Holocene simulations but do not capture the

FIG. 12. A pictorial representation of the mechanism explaining the increase in precipitation over the Sahel–Sahara region. Meridional section of the African monsoon (ocean, land, and desert are indicated in blue, brown, and yellow bars, respectively, and the increase in the vegetation over the desert is indicated by the green bar) showing (a) important features of the climatology of circulation and its changes in (b) Heat 1 K day^{-1} and (c) Heat 1.5 K day^{-1} simulations. The brown dotted arrow indicates the shallow return flow, and the blue arrow shows the southwesterly boundary layer monsoon flow. The light green circle with cross indicates the African easterly jet and dark green circle with cross indicates the tropical easterly jet. The magenta line indicates the geopotential gradients driving the shallow return flow. The red bar indicates the northward edge of the precipitation band where the drying due to meridional advection balances the moistening due to vertical advection. The light blue shaded region in (b),(c) is a region of cold anomaly.

magnitude and extent of the increase implied by proxy data. In this global model study we perturbed the African monsoon system by adding an artificial heat source of varying amplitude to the boundary layer to inspect

the simulated changes in circulation that occur in conjunction with Saharan precipitation changes of varying magnitude. We find that this experiment produces local precipitation responses similar to those of global mid-Holocene orbital forcing, when the local heating rate matches that of the orbital forcing, but more generally allows us to examine the sensitivity of the response to local diabatic effects that may not be fully captured by the model. Our study is thus relevant to the mid-Holocene greening but also to understanding the present-day African monsoon. Note that we did not allow for remote influences (e.g., sea surface temperature changes) or local vegetation changes, which would likely amplify precipitation shifts in the mid-Holocene case.

We find that although the boundary layer southerly monsoon flow changes linearly with heating rate (Fig. 7), the precipitation varies nonlinearly, with a significant increase at heating rates above those most closely approximating realistic mid-Holocene forcing (Fig. 2). The jump in precipitation is associated with rapid shifts in the zonal (African easterly jet) and meridional flow (shallow return flow) above the boundary layer but confined to the lower troposphere (Fig. 7). Below a 1 K day^{-1} threshold, the northward circulation shifts and precipitation increases are linear, while above this threshold changes accelerate. The linear part of this response resembles that observed during the anomalous wet African monsoon years as documented by Shekhar and Boos (2017). The nonlinear changes included the weakening of the shallow return flow that dries the key area above the boundary layer, which might be key to reproducing the correct mid-Holocene response in climate models.

We conclude that relatively cold and moist boundary layer air from the equatorial Atlantic affects precipitation in a manner that varies nonlinearly with the strength of the wind speed. As this air encroaches farther northward, it favors convection thermodynamically by increasing low-level moisture and moist static energy, but when it is sufficiently strong it also favors it dynamically by weakening the shallow return flow through its reduction of temperature and consequently the geopotential gradients driving this flow. As a result, precipitation increases rapidly when a threshold in heating is crossed. Note that although the stronger monsoon flow reduces boundary layer temperature hence dry static energy in the monsoon zone, the increase in specific humidity is more than sufficient to offset this and maintain the MSE maximum northward to support precipitation (Fig. 8). This is consistent with the findings of Patricola and Cook (2007) and previous studies on the role of boundary layer MSE in deciding the northward extent of monsoons (Privé and Plumb 2007; Nie et al.

2010). Figure 12 shows a pictorial representation showing this mechanism.

The northward extent of the African monsoon is limited by the ventilation of moist static energy by dry advection from the shallow return flow above the boundary layer. This result highlights the importance of vertical structure of the flows in ventilating the energy and is consistent with findings of [Peyrillé and Lafore \(2007\)](#). Further, the simulations suggest that the changes in precipitation depend nonlinearly on forcing, on account of the complex role of monsoon flow in moistening the boundary layer and influencing the shallow return flow. This view is also consistent with [Nie et al. \(2010\)](#), who demonstrated that the African monsoon does not follow a strict quasi-equilibrium assumption. Probably the offset in boundary layer versus free-tropospheric equivalent potential temperature changes is brought about by the import of hot, dry air with low equivalent potential temperature above the boundary layer by the shallow return flow. This mechanism would not be captured by simple models that include only one baroclinic mode or by linear/wave models that do not include nonlinear advection processes. Moreover, we expect that the response may be sensitive to the model treatment of convection and cloud processes due to the mechanisms reliance on a shallow circulation and the sensitivity of convection to the import of different forms of moist static energy at different levels.

A number of previous studies have examined the linearity of monsoon responses. [Patricola and Cook \(2008\)](#) reported nonlinear changes to the AEJ and precipitation in response to the prescribed vegetation cover in their regional model. They noted that the soil moisture associated with prescribed vegetation introduced nonlinear feedbacks in their model. The changes in AEJ that we reported are consistent with this study. The major difference is that we found a nonlinear response of the circulation and rainfall to heating, even though atmosphere–vegetation feedbacks were not included. Another important difference is the emphasis on the role of meridional circulation [as opposed to the zonal flows as proposed by [Patricola and Cook \(2008\)](#)] in the moisture budget. As shown in [Fig. 10](#), the drying due to the zonal mean meridional advection is much stronger than zonal advection over the Sahel and Sahara. The meridional flows induce stronger drying than zonal flows due to large moisture gradients in meridional direction in our simulations. A nonlinear response of monsoon precipitation has also been proposed previously by [Levermann et al. \(2009\)](#) on the basis of a simple model. Further analysis by [Schewe et al. \(2012\)](#) suggested that the nonlinear transitions in monsoons could occur on account of a specific humidity threshold due to complex

moisture advection feedbacks. [Boos and Storelmo \(2016\)](#) and [Seshadri \(2017\)](#) criticized the simple model since it did not accurately represent the dominant balance of the dry static energy budget in tropical atmosphere. [Boos and Storelmo \(2016\)](#) used a global model to show that the Asian monsoon responds linearly to wide variety of radiative forcing.

The present results do not support or oppose the previous findings, but suggest a new, nonlinear mechanism that may be specific to the African monsoon. The complex vertical structure of the flows involved makes this mechanism different from those proposed in previous work (e.g., [Levermann et al. 2009](#)). While some studies suggest an important role of atmosphere–vegetation feedbacks in the nonlinearity, our study exhibits nonlinearity without these feedbacks. We find a nonlinear interaction between boundary layer advection and thermally driven shallow return flows, which could potentially explain why the African monsoon may have gone through abrupt changes in the past or why few models project nonlinear changes in Sahel precipitation ([Schewe and Levermann 2017](#); [Tierney et al. 2017](#); [Neupane and Cook 2013](#)). In particular we find that proxy-estimated mid-Holocene rain signal can be reproduced in CAM5.1 by increasing the orbitally driven local heat source by only 50.

Several caveats of this study should be noted. First, the heating perturbations added here were artificial. In reality, the complex interaction between radiative and convective processes would control the strength of boundary layer heating over the desert. These are not represented accurately in current climate models ([Roehrig et al. 2013](#)), which produce large bias in simulated surface fluxes and amount of clouds over the desert. These processes along with complex feedback of natural aerosols shape the differential atmospheric heating over the desert. In this sense, the present results expose only qualitatively the nature of nonlinear changes that may occur. The quantitative threshold for nonlinearity may depend on many factors such as the radiation, boundary layer, or convection schemes in the model, or on each model's ability to produce the basic state of the African monsoon. We believe that the similarity in response between idealized heating and mid-Holo simulations suggests that the qualitative nature of changes in monsoon circulation is physically plausible. The PMIP mid-Holocene simulations may not be simulating strong enough boundary layer heating over the desert as compared to over the precipitation zone to trigger the nonlinear response. Further work is required to diagnose the systematic bias in different PMIP models in simulating this differential heating.

Acknowledgments. We thank ECMWF for sharing the ERA-Interim dataset and NOAA/OAR/ESRL PSD, Boulder, Colorado, for sharing GPCP dataset (available online at <https://www.esrl.noaa.gov/psd/data/gridded/data.gpcp.html>). This work was funded by the Australian Research Council Grants FL150100035 and DP140101104, and computations were performed at NCI.

REFERENCES

- Adler, R. F., and Coauthors, 2003: The version-2 Global Precipitation Climatology Project (GPCP) monthly precipitation analysis (1979–present). *J. Hydrometeorol.*, **4**, 1147–1167, [https://doi.org/10.1175/1525-7541\(2003\)004<1147:TVGPCP>2.0.CO;2](https://doi.org/10.1175/1525-7541(2003)004<1147:TVGPCP>2.0.CO;2).
- Boos, W. R., and R. L. Korty, 2016: Regional energy budget control of the intertropical convergence zone and application to Mid-Holocene rainfall. *Nat. Geosci.*, **9**, 892–897, <https://doi.org/10.1038/ngeo2833>.
- , and T. Storelvmo, 2016: Near-linear response of mean monsoon strength to a broad range of radiative forcings. *Proc. Natl. Acad. Sci. USA*, **113**, 1510–1515, <https://doi.org/10.1073/pnas.1517143113>.
- Braconnot, P., S. Joussaume, O. Marti, and N. de Noblet, 1999: Synergistic feedbacks from ocean and vegetation on the African monsoon response to Mid-Holocene insolation. *Geophys. Res. Lett.*, **26**, 2481–2484, <https://doi.org/10.1029/1999GL006047>.
- , —, N. de Noblet, and G. Ramstein, 2000: Mid-Holocene and Last Glacial Maximum African monsoon changes as simulated within the Paleoclimate Modelling Intercomparison Project. *Global Planet. Change*, **26**, 51–66, [https://doi.org/10.1016/S0921-8181\(00\)00033-3](https://doi.org/10.1016/S0921-8181(00)00033-3).
- , and Coauthors, 2007: Results of PMIP2 coupled simulations of the Mid-Holocene and Last Glacial Maximum—Part 2: Feedbacks with emphasis on the location of the ITCZ and mid- and high latitudes heat budget. *Climate Past*, **3**, 279–296, <https://doi.org/10.5194/cp-3-279-2007>.
- , S. P. Harrison, M. Kageyama, P. J. Bartlein, V. Masson-Delmotte, A. Abe-Ouchi, B. Otto-Bliesner, and Y. Zhao, 2012: Evaluation of climate models using palaeoclimatic data. *Nat. Climate Change*, **2**, 417–424, <https://doi.org/10.1038/nclimate1456>.
- Bretherton, C. S., and S. Park, 2009: A new moist turbulence parameterization in the Community Atmosphere Model. *J. Climate*, **22**, 3422–3448, <https://doi.org/10.1175/2008JCLI2556.1>.
- Broström, 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, <https://doi.org/10.1029/98GL02804>.
- Brovkin, V., and M. Claussen, 2008: Comment on “Climate-driven ecosystem succession in the Sahara: The past 6000 years.” *Science*, **322**, 1326b, <https://doi.org/10.1126/science.1163381>.
- Carrington, D. P., R. G. Gallimore, and J. E. Kutzbach, 2001: Climate sensitivity to wetlands and wetland vegetation in Mid-Holocene North Africa. *Climate Dyn.*, **17**, 151–157, <https://doi.org/10.1007/s003820000099>.
- Chadwick, R., G. M. Martin, D. Copsey, G. Bellon, M. Caian, F. Codron, C. Rio, and R. Roehrig, 2017: Examining the West African monsoon circulation response to atmospheric heating in a GCM dynamical core. *J. Adv. Model. Earth Syst.*, **9**, 149–167, <https://doi.org/10.1002/2016MS000728>.
- Chikira, M., A. Abe-Ouchi, and A. Sumi, 2006: General circulation model study on the Green Sahara during the Mid-Holocene: An impact of convection originating above boundary layer. *J. Geophys. Res.*, **111**, D21103, <https://doi.org/10.1029/2005JD006398>.
- Claussen, M., and V. Gayler, 1997: The greening of the Sahara during the Mid-Holocene: Results of an interactive atmosphere-biome model. *Global Ecol. Biogeogr. Lett.*, **6**, 369–377, <https://doi.org/10.2307/2997337>.
- , V. Brovkin, and A. Ganopolski, 2002: Africa: Greening of the Sahara. *Challenges of a Changing Earth*, Springer-Verlag, 125–128.
- Cornforth, R. J., B. J. Hoskins, and C. D. Thorncroft, 2009: The impact of moist processes on the African easterly jet–African easterly wave system. *Quart. J. Roy. Meteor. Soc.*, **135**, 894–913, <https://doi.org/10.1002/qj.414>.
- Dee, D. P., and Coauthors, 2011: The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, <https://doi.org/10.1002/qj.828>.
- deMenocal, P., J. Ortiz, T. Guilderson, J. Adkins, M. Sarnthein, L. Baker, and M. Yarusinsky, 2000: Abrupt onset and termination of the African Humid Period: Rapid climate responses to gradual insolation forcing. *Quat. Sci. Rev.*, **19**, 347–361, [https://doi.org/10.1016/S0277-3791\(99\)00081-5](https://doi.org/10.1016/S0277-3791(99)00081-5).
- Gent, P. R., and Coauthors, 2011: The Community Climate System Model version 4. *J. Climate*, **24**, 4973–4991, <https://doi.org/10.1175/2011JCLI4083.1>.
- Hack, J. J., B. A. Boville, B. P. Briegleb, J. T. Kiehl, P. J. Rasch, and D. L. Williamson, 1993: Description of the NCAR Community Climate Model (CCM2). NCAR Tech. Note NCAR/TN-382+STR, 108 pp., <http://dx.doi.org/10.5065/D6QZ27XV>.
- Hales, K., J. D. Neelin, and N. Zeng, 2006: Interaction of vegetation and atmospheric dynamical mechanisms in the Mid-Holocene African monsoon. *J. Climate*, **19**, 4105–4120, <https://doi.org/10.1175/JCLI3833.1>.
- Harrison, S. P., P. J. Bartlein, K. Izumi, G. Li, J. Annan, J. Hargreaves, P. Braconnot, and M. Kageyama, 2015: Evaluation of CMIP5 palaeo-simulations to improve climate projections. *Nat. Climate Change*, **5**, 735–743, <https://doi.org/10.1038/nclimate2649>.
- 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, <https://doi.org/10.1029/97GL03721>.
- Hurrell, J. W., J. J. Hack, D. Shea, J. M. Caron, and J. Rosinski, 2008: A new sea surface temperature and sea ice boundary dataset for the Community Atmosphere Model. *J. Climate*, **21**, 5145–5153, <https://doi.org/10.1175/2008JCLI2292.1>.
- Joussaume, S., and K. E. Taylor, 1995: Status of the Paleoclimate Modeling Intercomparison Project (PMIP). *Proc. of the First Int. AMIP Scientific Conf.*, Monterey, CA, WMO Tech. Doc. WMO/TD-732, W. L. Gates, Ed., 425–430.
- Kang, S. M., I. M. Held, D. M. W. Frierson, and M. Zhao, 2008: The response of the ITCZ to extratropical thermal forcing: Idealized slab-ocean experiments with a GCM. *J. Climate*, **21**, 3521–3532, <https://doi.org/10.1175/2007JCLI2146.1>.
- Kröpelin, S., and Coauthors, 2008: Climate-driven ecosystem succession in the Sahara: The past 6000 years. *Science*, **320**, 765–768, <https://doi.org/10.1126/science.1154913>.
- 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, <https://doi.org/10.1126/science.214.4516.59>.
- , and Z. Liu, 1997: Response of the African monsoon to orbital forcing and ocean feedbacks in the Middle Holocene. *Science*, **278**, 440–443, <https://doi.org/10.1126/science.278.5337.440>.

- , 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, <https://doi.org/10.1038/384623a0>.
- Levermann, A., J. Schewe, V. Petoukhov, and H. Held, 2009: Basic mechanism for abrupt monsoon transitions. *Proc. Natl. Acad. Sci. USA*, **106**, 20 572–20 577, <https://doi.org/10.1073/pnas.0901414106>.
- Neale, R. B., and Coauthors, 2010: Description of the NCAR Community Atmosphere Model (CAM5.0). NCAR Tech. Note NCAR/TN-486+STR, 268 pp., www.cesm.ucar.edu/models/cesm1.1/cam/docs/description/cam5_desc.pdf.
- , J. Richter, S. Park, P. H. Lauritzen, S. J. Vavrus, P. J. Rasch, and M. Zhang, 2013: The mean climate of the Community Atmosphere Model (CAM4) in forced SST and fully coupled experiments. *J. Climate*, **26**, 5150–5168, <https://doi.org/10.1175/JCLI-D-12-00236.1>.
- Neupane, N., and K. H. Cook, 2013: A nonlinear response of Sahel rainfall to Atlantic warming. *J. Climate*, **26**, 7080–7096, <https://doi.org/10.1175/JCLI-D-12-00475.1>.
- Nicholson, S. E., 2013: The West African Sahel: A review of recent studies on the rainfall regime and its interannual variability. *ISRN Meteor.*, **2013**, 453521, 32 pp., <http://dx.doi.org/10.1155/2013/453521>.
- Nie, J., W. R. Boos, and Z. Kuang, 2010: Observational evaluation of a convective quasi-equilibrium view of monsoons. *J. Climate*, **23**, 4416–4428, <https://doi.org/10.1175/2010JCLI3505.1>.
- Oleson, K. W., and Coauthors, 2010: Technical description of version 4.0 of the Community Land Model (CLM). NCAR Tech. Note NCAR/TN-478+STR, 257 pp., doi:10.5065/D6FB50WZ.
- Patricola, C. M., and K. H. Cook, 2007: Dynamics of the West African monsoon under Mid-Holocene precessional forcing: Regional climate model simulations. *J. Climate*, **20**, 694–716, <https://doi.org/10.1175/JCLI4013.1>.
- , and —, 2008: Atmosphere/vegetation feedbacks: A mechanism for abrupt climate change over northern Africa. *J. Geophys. Res.*, **113**, D18102, <https://doi.org/10.1029/2007JD009608>.
- Pausata, F. S. R., G. Messori, and Q. Zhang, 2016: Impacts of dust reduction on the northward expansion of the African monsoon during the Green Sahara period. *Earth Planet. Sci. Lett.*, **434**, 298–307, <https://doi.org/10.1016/j.epsl.2015.11.049>.
- Peyrillé, P., and J.-P. Lafore, 2007: An idealized two-dimensional framework to study the West African monsoon. Part II: Large-scale advection and the diurnal cycle. *J. Atmos. Sci.*, **64**, 2783–2803, <https://doi.org/10.1175/JAS4052.1>.
- Privé, N. C., and R. A. Plumb, 2007: Monsoon dynamics with interactive forcing. Part I: Axisymmetric studies. *J. Atmos. Sci.*, **64**, 1417–1430, <https://doi.org/10.1175/JAS3916.1>.
- Pu, B., and K. H. Cook, 2010: Dynamics of the West African westerly jet. *J. Climate*, **23**, 6263–6276, <https://doi.org/10.1175/2010JCLI3648.1>.
- , and —, 2012: Role of the West African westerly jet in Sahel rainfall variations. *J. Climate*, **25**, 2880–2896, <https://doi.org/10.1175/JCLI-D-11-00394.1>.
- Roehrig, R., D. Bouniol, F. Guichard, F. Hourdin, and J.-L. Redelsperger, 2013: The present and future of the West African monsoon: A process-oriented assessment of CMIP5 simulations along the AMMA transect. *J. Climate*, **26**, 6471–6505, <https://doi.org/10.1175/JCLI-D-12-00505.1>.
- Schewe, J., and A. Levermann, 2017: Non-linear intensification of Sahel rainfall as a dynamic response to future warming. *Earth Syst. Dyn.*, **8**, 495–505, <https://doi.org/10.5194/esd-8-495-2017>.
- , —, and H. Cheng, 2012: A critical humidity threshold for monsoon transitions. *Climate Past*, **8**, 535–544, <https://doi.org/10.5194/cp-8-535-2012>.
- Schneider, T., T. Bischoff, and G. H. Haug, 2014: Migrations and dynamics of the intertropical convergence zone. *Nature*, **513**, 45–53, <https://doi.org/10.1038/nature13636>.
- Seshadri, A. K., 2017: Energetics and monsoon bifurcations. *Climate Dyn.*, **48**, 561–576, <https://doi.org/10.1007/s00382-016-3094-7>.
- Shekhar, R., and W. R. Boos, 2017: Weakening and shifting of the Saharan shallow meridional circulation during wet years of the West African monsoon. *J. Climate*, **30**, 7399–7422, <https://doi.org/10.1175/JCLI-D-16-0696.1>.
- Su, H., and J. D. Neelin, 2005: Dynamical mechanisms for African monsoon changes during the Mid-Holocene. *J. Geophys. Res.*, **110**, D19105, <https://doi.org/10.1029/2005JD005806>.
- Swann, A. L. S., I. Y. Fung, Y. Liu, and J. C. H. Chiang, 2014: Remote vegetation feedbacks and the mid-Holocene green Sahara. *J. Climate*, **27**, 4857–4870, <https://doi.org/10.1175/JCLI-D-13-00690.1>.
- Thorncroft, C. D., and M. Blackburn, 1999: Maintenance of the African easterly jet. *Quart. J. Roy. Meteor. Soc.*, **125**, 763–786, <https://doi.org/10.1002/qj.49712555502>.
- , H. Nguyen, C. Zhang, and P. Peyrillé, 2011: Annual cycle of the West African monsoon: Regional circulations and associated water vapour transport. *Quart. J. Roy. Meteor. Soc.*, **137**, 129–147, <https://doi.org/10.1002/qj.728>.
- Tierney, J. E., F. S. R. Pausata, and P. B. deMenocal, 2017: Rainfall regimes of the Green Sahara. *Sci. Adv.*, **3**, e1601503, <https://doi.org/10.1126/sciadv.1601503>.
- Zhang, C., P. Woodworth, and G. Gu, 2006: The seasonal cycle in the lower troposphere over West Africa from sounding observations. *Quart. J. Roy. Meteor. Soc.*, **132**, 2559–2582, <https://doi.org/10.1256/qj.06.23>.
- , D. S. Nolan, C. D. Thorncroft, and H. Nguyen, 2008: Shallow meridional circulations in the tropical atmosphere. *J. Climate*, **21**, 3453–3470, <https://doi.org/10.1175/2007JCLI1870.1>.
- Zhang, G. J., and N. A. McFarlane, 1995: Sensitivity of climate simulations to the parameterization of cumulus convection in the Canadian Climate Centre general circulation model. *Atmos.–Ocean*, **33**, 407–446, <https://doi.org/10.1080/07055900.1995.9649539>.
- Zhao, Y., and Coauthors, 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, <https://doi.org/10.1007/s00382-005-0075-7>.