

UCLA

UCLA Previously Published Works

Title

West African monsoon decadal variability and surface-related forcings: second West African Monsoon Modeling and Evaluation Project Experiment (WAMME II)

Permalink

<https://escholarship.org/uc/item/5np1h4vr>

Journal

Climate Dynamics, 47(11)

ISSN

0930-7575

Authors

Xue, Yongkang
De Sales, Fernando
Lau, William K-M
[et al.](#)

Publication Date


2016-12-01

DOI

10.1007/s00382-016-3224-2

Peer reviewed

West African monsoon decadal variability and surface-related forcings: second West African Monsoon Modeling and Evaluation Project Experiment (WAMME II)

Yongkang Xue¹  · Fernando De Sales² · William K.-M. Lau³ · Aaron Boone⁴ · Kyu-Myong Kim⁵ · Carlos R. Mechoso¹ · Guiling Wang⁶ · Fred Kucharski⁷ · Kathleen Schiro¹ · Masahiro Hosaka⁸ · Suosuo Li^{9,1} · Leonard M. Druyan¹⁰ · Ibrah Seidou Sanda¹¹ · Wassila Thiaw¹² · Ning Zeng³ · Ruth E. Comer¹³ · Young-Kwon Lim^{5,14} · Sarith Mahanama⁵ · Guoqiong Song¹ · Yu Gu¹ · Samson M. Hagos¹⁶ · Mian Chin⁵ · Siegfried Schubert⁵ · Paul Dirmeyer¹⁵ · L. Ruby Leung¹⁶ · Eugenia Kalnay³ · Akio Kitoh¹⁷ · Cheng-Hsuan Lu^{12,18} · Natalie M. Mahowald¹⁹ · Zhengqiu Zhang^{1,20}

Received: 21 July 2015 / Accepted: 5 June 2016 / Published online: 13 June 2016
© Springer-Verlag Berlin Heidelberg 2016

Abstract The second West African Monsoon Modeling and Evaluation Project Experiment (WAMME II) is designed to improve understanding of the possible roles and feedbacks of sea surface temperature (SST), land use land cover change (LULCC), and aerosols forcings

in the Sahel climate system at seasonal to decadal scales. The project's strategy is to apply prescribed observationally based anomaly forcing, i.e., "idealized but realistic" forcing, in simulations by climate models. The goal is to assess these forcings' effects in producing/amplifying seasonal and decadal climate variability in the Sahel between the 1950s and the 1980s, which is selected to characterize the great drought period of the last century. This is the first multi-model experiment specifically designed to simultaneously evaluate such relative contributions. The WAMME II models have consistently demonstrated that SST forcing is a major contributor to the twentieth century Sahel drought. Under the influence of the maximum possible SST forcing, the ensemble mean of WAMME II models can produce up to 60 % of the precipitation difference during the period. The present paper also addresses the role of SSTs

This paper is a contribution to the special issue on West African climate decadal variability and its modeling, consisting of papers from the West African Monsoon Modeling and Evaluation (WAMME) and the African Multidisciplinary Monsoon Analyses (AMMA) projects, and coordinated by Yongkang Xue, Serge Janicot, and William Lau. Provided Funding information has to be tagged.

Electronic supplementary material The online version of this article (doi:10.1007/s00382-016-3224-2) contains supplementary material, which is available to authorized users.

✉ Yongkang Xue
yxue@geog.ucla.edu

¹ University of California, Los Angeles, CA, USA

² San Diego State University, San Diego, CA, USA

³ University of Maryland, College Park, MD, USA

⁴ Centre National de Recherches Météorologiques, Météo-France, Toulouse, France

⁵ NASA Goddard Space Flight Center, Greenbelt, MD, USA

⁶ University of Connecticut, Storrs, CT, USA

⁷ The International Center for Theoretical Physics, Trieste, Italy

⁸ Meteorological Research Institute, Tsukuba, Japan

⁹ Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences, Lanzhou, China

¹⁰ NASA Goddard Institute for Space Studies and Columbia University, New York, NY, USA

¹¹ AGRHYMET Regional Center and Abdou Moumouni University, Niamey, Niger

¹² National Center for Environmental Prediction, College Park, MD, USA

¹³ Met Office Hadley Centre, Exeter, UK

¹⁴ Goddard Earth Sciences Technology and Research, I. M. Systems Group, Greenbelt, MD, USA

¹⁵ Center for Ocean-Land-Atmosphere Interactions Studies, George Mason University, Fairfax, VA, USA

¹⁶ Pacific Northwest National Laboratory, Richland, WA, USA

¹⁷ University of Tsukuba, Tsukuba, Japan

in triggering and maintaining the Sahel drought. In this regard, the consensus of WAMME II models is that both Indian and Pacific Ocean SSTs greatly contributed to the drought, with the former producing an anomalous displacement of the Intertropical Convergence Zone before the WAM onset, and the latter mainly contributes to the summer WAM drought. The WAMME II models also show that the impact of LULCC forcing on the Sahel climate system is weaker than that of SST forcing, but still of first order magnitude. According to the results, under LULCC forcing the ensemble mean of WAMME II models can produce about 40 % of the precipitation difference between the 1980s and the 1950s. The role of land surface processes in responding to and amplifying the drought is also identified. The results suggest that catastrophic consequences are likely to occur in the regional Sahel climate when SST anomalies in individual ocean basins and in land conditions combine synergistically to favor drought.

Keywords Sahel seasonal and decadal climate variability · Sahel drought, SST and land forcings · GCM

1 Introduction

The Sahel is a climatically diverse and economically fragile region. The dramatic changes that occurred over Sahelian Africa from wet conditions in the 1950s to much drier conditions in the 1970s–1980s to be followed by partial recovery after the 1980s represent one of the strongest interdecadal climate variabilities and the longest drought on the planet in the twentieth century (e.g., Redelsperger et al. 2007; Lebel and Ali 2009; Nicholson et al. 2012). A significant climate feature in Sahelian Africa is the West African monsoon (WAM), which exerts a strong control on climate variability in the Sahel. Observational evidence has supported the notion that there is strong decadal climate variability in the Sahel and surrounding areas from the 1950s to the 2000s, not only in precipitation, but also in vegetation conditions, land use and land cover (LULC) change, and aerosol loading.

Numerous papers have discussed SST variability in the last century and its links to Sahelian precipitation variability. Relationships between Sahel precipitation and SST anomalies in the Atlantic, Pacific, and Indian Oceans and

the Mediterranean Sea¹ have been studied extensively (see a comprehensive review in Rodriguez-Fonseca et al. 2015). Despite the dominant role of SST anomalies on Sahel climate variability, modeling studies have consistently shown that SST is not the sole forcing of significant Sahel climate variability and other sources of forcing should be considered.

Although a number of atmospheric general circulation models (AGCMs) are able to reproduce the twentieth century drying trend in West Africa and the Sahel dry conditions of the 1970–1980s and subsequent rainfall recovery using the observed SST as boundary conditions, simulated trends and drying are substantially weaker than in the observations (e.g., Giannini et al. 2003; Lu and Delworth 2005; Hoerling et al. 2006, 2010; Lau et al. 2006; Caminade and Terray 2010; Martin and Thorncroft 2014). For instance, the Climate of the twentieth century international project (C20C, Scaife et al. 2009) used 14 state-of-the-art GCMs with observed SSTs and other relevant data to study climate variations and changes over the last century. Only two C20C models simulated about half the magnitude of the Sahel rainfall changes between the 1950s and the 1980s. The results of the other models ranged between less than 30 % of the observed precipitation anomaly to even no negative anomalies at all. Scaife et al. (2009) concluded that the Sahel drought was only partly forced by SSTs in their experiment, which did not reproduce the magnitude of the Sahel drought. In spite of previous efforts to evaluate the SST contribution, a number of important questions remain unanswered. For instance, do different ocean basins provide a similar contribution to Sahelian precipitation variability or are some basins more influential than others? Rodriguez-Fonseca et al. (2015) pointed out that aspects on these questions remain open. The results obtained so far appear to be highly model dependent (e.g. Scaife et al. 2009; Biasutti et al. 2008). The possible nonlinearities in the atmospheric response to different SST anomalies and interactions among the basins complicate the problem (Hagos and Cook 2008).

In addition to SST effect, the Sahel has been identified as the region in the world where climate is most sensitive to land/atmosphere interactions through vegetation and soil processes (e.g., Koster et al. 2004; Xue et al. 2004b, 2010a; see also reviews in Xue et al. 2004a, 2012). The C20C experiment also noted that the two models that captured the large change in Sahel rainfall include land surface effects via either parameterized vegetation-climate interactions or specified land use and land cover changes (LULCC). The present study evaluates and compares the effects of both SST and land surface processes (mainly through LULCC

¹⁸ University at Albany, State University of New York, Albany, NY, USA

¹⁹ Cornell University, Ithaca, NY, USA

²⁰ Chinese Academy of Meteorological Sciences, Beijing, China

¹ For convenience, when we use the word “oceans” in this paper the Mediterranean Sea will be included.

experiments) on the Sahel decadal variability in precipitation, surface temperature, atmospheric circulation, and surface energy balance, especially during the drought of the 1980s. This comparison has not been conducted before. The Sahel decadal precipitation anomalies between the 1980s and the 1950s are used to characterize the Sahel 1980s drought in this study.

While a careful comparison of the contributions of SST, land, and other forcings to the Sahel drought is imperative to unravel the root causes of the long-term variations of the Sahel climate, it has not been performed in multi-modeling contexts. Our approach is based on the examination of the results from the West African Monsoon Modeling and Evaluation Project Experiment II (WAMME II). Since it is challenging to simulate the spatial dimensions, temporal evolution, triggering and strength of the West African Monsoon (WAM) system (e.g., Hourdin et al. 2010; Roehrig et al. 2013), in the first WAMME experiment (WAMME I), we comprehensively evaluated the performances of WAMME GCMs and RCMs in simulating the variability of West African monsoon (WAM) precipitation, surface temperature, and major circulation features at seasonal and intraseasonal scales (Xue et al. 2010a; Druyan et al. 2010; Boone et al. 2010). The results indicated that, despite deficiencies in many aspects, the WAMME models generally produce reasonable simulations of the spatial distribution of WAM seasonal mean precipitation and surface temperature and their temporal evolution, as well as some major circulation features (Xue et al. 2010a). The results from the WAMME I experiment provide a solid foundation for us to work further in the WAMME II experiment.

In this study, we examine whether the reported influence by the global SST and by individual Oceans' SST on the Sahel climate is simply an artifact of sampling variability or whether such an influence reflects genuine and interesting long-term changes in the climate system. The differences in seasonal WAM features at different decades are also evaluated. Another issue addressed is whether the drought was primarily due to natural multi-decadal modes of variability in the SST patterns or the results of anthropogenic effects. The objectives in this WAMME II paper are to provide a basic understanding of the issues related to SST; and to assess the relative contributions of forcings in SST and land surface processes in producing/amplifying Sahelian seasonal and decadal climate variability. In the following text, Sect. 2 introduces the GCMs and the design of the WAMME II experiment. Section 3 evaluates the impact of SST anomalies in the world ocean and individual ocean basins on Sahel climate variability, including precipitation, surface temperature, surface energy balances, and some major circulation features. The SST and LULCC effects are also compared. Section 4 summarizes results and discusses key issues for future studies.

2 WAMME models and WAMME II experimental design

The WAMME II project comprises 8 GCMs (Table 1) and 4 RCMs (Table 2) with a wide range of spatial resolutions and physical parameterizations. Among the GCMs, the JMA MRI (Japan Meteorological Administration Meteorological Research Institute, Mizuta et al. 2006) GCM has the highest horizontal resolution (20 km) while the ICTP-UMD (International Center for Theoretical Physics/University of Maryland, College Park; Kucharski et al. 2013) and the UCLA AGCM (University of California, Los Angeles; Mechoso et al. 2000) have the lowest resolutions (~200 km). The UCONN CAM5 (University of Connecticut/Community Earth System Model version 5. Neale et al. 2012) and GSFC GEOS-5 (the version 5 of the Goddard Earth Observing System Model of the Global Modeling and Assimilation Office at NASA/GSFC. Molod et al. 2012) include comprehensive aerosol schemes. All models include comprehensive biophysical models for land surface processes. The UCLA GFS (Kanamitsu et al. 2002b; Xue et al. 2004b) and the UCLA AGCM have the same land surface scheme. More information on the physical components of participating models, including land surface models, can be found in Tables 1 and 2. Among eight GCMs, the HadGEM2-A GCM (Pope et al. 2000) only participated in testing the LULCC effect. Two versions of the GSFC GEOS-5 have participated in the WAMME II experiments. In the SST experiments, the GSFC model includes the GOCART aerosol model (NASA Goddard Chemistry Aerosol Radiation and Transport, Chin et al. 2000, 2002). The GEOS-5 provides two sets of results with different aerosol optical properties (referred to as GEOS-5 I and GEOS-5 II in this paper). In the GEOS-5 LULCC experiment, a catchment model is used with the land surface scheme (Koster et al. 2000). Except for the ICTP-UMD GCM, other WAMME II models have been validated in the WAMME I for their ability in simulating WAM precipitation and circulation (Xue et al. 2010a). The WAMME II models' ability to model the West African climatology is outlined in the electronic supplement of this paper. The WAMME II also includes regional climate models (Table 2). Results from RCM in relation to monsoon response to LULCC and biomass burning are presented in separate papers (Hagos et al. 2014; De Sales et al. 2015; Wang et al. 2015).

The WAMME II control simulation (referred to as Case CTL) uses January 1, 2006, as the initial condition and time-varying climatological SSTs as boundary conditions. The models were integrated for 6 years and results that will be presented correspond to the average of the last 5 years. Since climatological SST is used and the first year's results are discarded, the effects of initial conditions

Table 1 List of WAMME II GCMs

Model	Contact person	Resolution	Rad. scheme	Conv. scheme	PBL/shallow cloud	Cloud	Surface	Aerosol/dust
ICTP/UMD GCM (Kucharski et al. 2013)	Fred Kucharski, N. Zeng, E. Kalnay	T30L8 or T47L8	Molteni (2003)	Mass-flux (Molteni 2003)	Molteni (2003)	Molteni (2003)	SL and-VEGAS (Zeng et al. 2000, 2003, 2005)	
Japan MRI/JMA AGCM (Mizuta et al. 2006)	Masahiro Hosaka, A. Kitoh	TL959L60	Shibata and Uchiyana (1992), Shibata and Aoki (1989)	Prognostic Arakawa-Schubert (1974)	Stratoscumulus param. Slingo (1987)	Smith (1999)	SiB with 4 soil layers, 3 snow layers (Sellers et al. 1996)	Sulfate aerosols, no dust
Modified NASA GMAO GEOS-5 (Molod et al. 2012)	K.-M. Lau, K. M. Kim, Y.-K. Kim, S. Schubert	$1^\circ \times 1.25^\circ \times L72$ layers	Chou and Suarez (1999); Chou et al. (2001), with <i>parameterization modification</i>	Relaxed Arakawa Schubert (1974) with stochastic Tokioka parameter	Lock (2000) and Louis (1982)	Bacmeister et al. (2000)		Aerosol from GOCART model (Chin et al. 2002; Colarco et al. 2010)
NASA GMAO GEOS-5 (Molod et al. 2012)	Sarith, Mahanama, R. Koster	$1^\circ \times 1.25^\circ \times L72$ layers	Chou and Suarez (1999), Chou et al. (2001)	Relaxed Arakawa Schubert (1974) with stochastic Tokioka parameter	Lock (2000) and Louis (1979)	Bacmeister et al. (2000)	Catchment model (Koster et al. 2000)	Aerosols from GOCART data
UCLA AGCM (Mechoso et al. 2000)	C. R. Mechoso, K. Schiro	$2.5^\circ \times 2^\circ \times L17$	Katayama (1972), Harshvardhan et al. (1987)	Arakawa and Schubert (1974), Lord et al. (1982)	Suarez et al. (1983)	Prognostic cloud water and ice (Kohler 1999)	SSIB-1 (Xue et al. 1991)	No aerosol or GOCART Data
UCLA GFS GCM (Kanamitsu et al. 2002a, b; Xue et al. 2004b)	Y. Xue, F. De Sales	T62L28	Chou (1992), Chou and Suarez (1999), Hou et al. (1996, 2002)	Moorthi and Suarez (1992)	Hong and Pan (1996)	Slingo (1987)	SSIB-1 (Xue et al. 1991)	No aerosol or GOCART Data
Univ. of Connecticut CAM5 (Neale et al. 2012)	G. Wang	T63L26	RRTMG Iacono et al. (2008) and Mlawer et al. (1997)	Zhang and McFarlane (1995), Richter and Rasch (2008), Raymond and Blyth (1992)	Bretherton and Park (2009), Park and Bretherton (2009)	Gottelman et al. (2010)	CLM4.0 (Oleson et al. 2010)	Easter et al. (2004); Neale et al. (2012)
HadGEM2-A (HadGEM2 Model Development Team 2011)	R. Comer	$1.25^\circ \times 1.875^\circ \times L38$	Edwards and Slingo (1996), Cusack et al. (1999)	Gregory and Rowntree (1990), Grant and Brown (1999), Derbyshire (2011)	Lock et al. (2000); Brown and Grant (1997)	Smith et al. (1999); Wilson and Ballard (1999)	Cox et al. (1999), hydrology scheme (Clark and Gedney 2008)	Bellouin et al. (2011); Collins et al. (2008); Woodward (2011)

Table 2 List of WAMME II RCMs

Model	Contactperson	Resolution	Rad. scheme	Conv. scheme	PBL/shallow cloud	Cloud	Surface	Aerosol/Dust
RegCM4.1 (with BATS)	Seidou Sanda Ibrah	50 km	CCM3 radiation (Kiel et al. 1996)	Grell (1994)	Holtlag et al. (1990)	Grell (1994)	BATS (Dickinson et al. 1986)	Solmon et al. (2006) Zakey et al. (2006)
UCLA WRF	Y. Xue F. De Sales	50 × 50 km, 38 levels	Fels and Schwartzkopf (1975)	Betts-Miller-Janjic (Janic 1990)	Mellor-Yamada 2.5 (Mellor and Yamada 1982; Janic 1996)	Zhao and Carr (1997)	SSiB-3 (Xue et al. 1991; Sun and Xue 2001)	
WRF 3.3 (CAM physics) (Skamarock et al. 2008)	Samson Hagos	25 km	RRTM	Zhang and McFarlane (1995)	UW shallow (Bretherton and Park 2004)	Morrison-Gottelman (CAM5)	NOAH (Chen, and Dudhia 2001; SSiB (Xue et al. 1991; Sun and Xue 2001), Pleim-Xiu (Pleim and Xiu 2003)	
RegCM4.1 (with CLM4) (Wang et al. 2015)	G. Wang	50 km	CCM3 radiation (Kiel et al. 1996)	Emanuel (1991)	Holtlag et al. (1990)	Emanuel (1991) and Pal et al. (2000)	CLM4 (Oleson et al. 2010)	(Not activated) Solmon et al. (2006) Zakey et al. (2006)

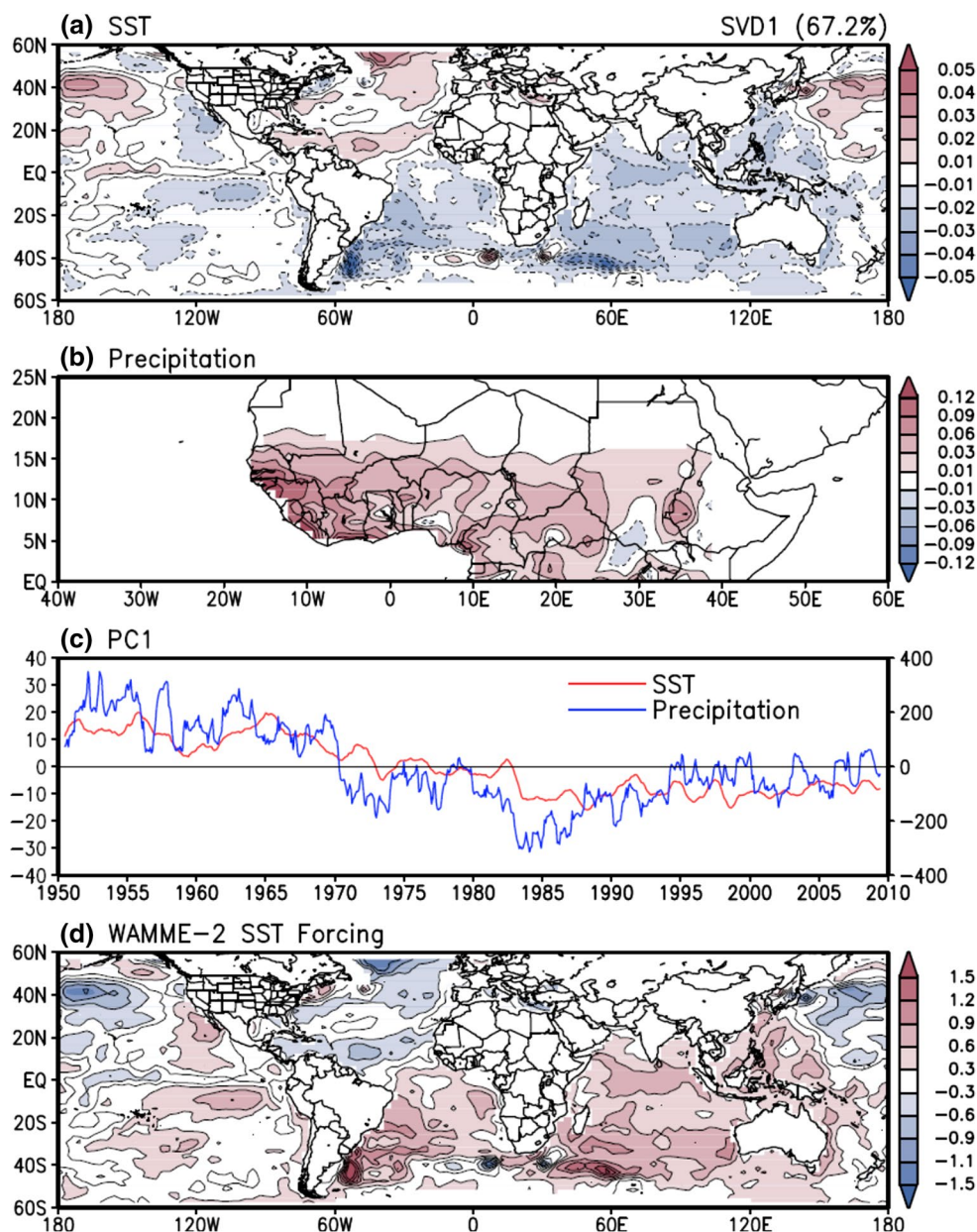
References cited in Tables 1 and 2 can be found at http://www.wamme.geog.ucla.edu/LOCAL/ref_gcm_wamme.pdf

are minimized. The ITCP-UMD GCM has a dynamic vegetation model that requires a 50-year model spin up; results presented here correspond to the last 50 years of simulation. By using the equilibrium type test, models should have less internal variability and produce clearer signals, which will facilitate understanding of the relevant mechanisms. The initial conditions for the atmosphere in the WAMME II experiment are from the NCEP/DOE (National Center for Environmental Prediction/Department of Energy) Reanalysis II (Kanamitsu et al. 2002a). This Reanalysis version includes corrections of human processing errors and incorporates upgrades to the forecast model and a diagnostic package that had been developed since the time the Reanalysis I was finalized.

To test the impacts of SST and land forcings, the WAMME II strategy is to apply observational data-based prescribed anomaly forcing, i.e., “idealized but realistic” forcing, in GCM and RCM simulations. To analyze the effect of SST anomalies on Sahel precipitation, long-term variations of global SST associated with the decadal dry and wet spells of Sahel rainfall are reconstructed using SVD analysis. Annual cycles are removed based on 60-year monthly averages for 1950–2009. To minimize the effect of interannual variation of SST, the ENSO signal is removed based on linear regression between SST anomalies at each grid and the NINO3.4 SST anomaly index. Possible residuals associated with the ENSO signal as well as shorter time-scale variations are further reduced by applying a 13-month moving average. SVD is then applied to ENSO-free global SST and Sahel rainfall anomalies.

Figure 1a and b show the first singular value decomposition mode (SVD1) for global SST and WAM rainfall respectively, while Fig. 1c display the times series of the first principal components (PC1s). SVD analysis depicts the common patterns between two variables (Wallace et al. 1992; Weng et al. 1999). SST and rainfall data used in the figure are from the Hadley Centre Sea Ice and Sea Surface Temperature data set (Rayner et al. 2003) and the Climate Prediction Center’s global gauge-based analysis of precipitation (CPC GTS, Chen et al. 2002), respectively. The leading SVD mode, which explains a significant covariance (67.2 %) between SST and rainfall, captures the well-known patterns of inter-decadal variation of SST and associated change in rainfall over the Sahel. The inter-hemispheric asymmetry of SST anomalies in the Atlantic Ocean and associated SST trends in the Indian Ocean and the Mediterranean Sea as displayed in Fig. 1a, c are consistent with results of previous studies (e.g. Lamb 1978; Folland et al. 1991; Carton et al. 1996; Janicot et al. 1998; Rowell 2001; Joly et al. 2007; Hastenrath and Polzin 2010). The second SST mode (8.7 %) shows a globally homogeneous spatial pattern associated with rainfall change in the southern Sahel (5°N–12°N) and the PC2 has

Fig. 1 **a, b** SST and precipitation SVD1 spatial patterns; **c** SVD PC1 time series during the 1950–2010 period; **d** WAMME SST forcing difference between 1980s and 1950s (°C)



a weak trend only after the mid-1970s. The third SST mode only explains 4.5 % variance. The second and third SVDs are not considered in our construction of the SST forcing as discussed below.

In WAMME II, the amplitude of the SST forcing field is required to be both realistic and strong enough to achieve a robust model response, and to be consistent with that used in other sensitivity experiments (e.g., LULCC and aerosol). To emphasize the effect of decadal SST change on the Sahel rainfall between the 1950s and 1980s, the maximum value in the time series of PC1 in the 1950s and the minimum value in the 1980s are used to reconstruct the SST anomalies for those decades. This design is consistent with the “idealized but realistic” strategy of WAMME

II. Figure 1d shows the difference between reconstructed SST anomalies for the 1980s and 1950s. The distribution and amplitude of SST anomalies are both comparable to those in previous studies (e.g., Lu and Delworth 2005; Mohino et al. 2011). These anomalies are added to the 60-year monthly averages of 1950–2009 to construct an annual cycle of SST for the 1950s and the 1980s. Since the primary interest is on decadal variability, constant SST anomalies in time are superimposed on the SST seasonal cycle. In WAMME II, model runs similar to Case CTL but with the prescribed SST anomalies for the 1950s and the 1980s are referred to as Case 50SST and Case 80SST, respectively. In these two runs, except for the SST conditions, other initial atmospheric and boundary conditions are

the same as in Case CTL. The model results are bi-linearly interpolated to the 0.5° CPC GTS (Chen et al. 2002) grid for comparison. We use Reanalysis I for comparisons with simulation in the following text because only this reanalysis dataset dates back to the 1950s.

The WAMME II also investigates the impact of SST anomalies in different ocean basins. The SST effect from one ocean basin over the Sahel could be direct (i.e., via teleconnections) or indirect (i.e., through interaction with atmospheric anomalies caused by SST anomalies in other ocean basins; then this interaction affects the Sahel climate variability via teleconnections) (e.g., Hagos and Cook 2008). If we simply added one ocean's anomaly, this indirect effect may be missing (because other ocean's anomalies are not imposed). Our approach was specifically designed to account for both direct and indirect effects of each ocean's SST anomalies. The approach we use in WAMME II consists of starting with the global 1950 SST anomalies and the global 1980 SST anomalies in the cases that we will refer to as Case 50SST and Case 80SST, respectively, which preserves the influence due to interactions of atmospheric circulation caused by different oceans' SST anomalies; next, we eliminate the anomalies of SST forcing from a single basin (Atlantic, Indian, Mediterranean, or Pacific, respectively) in the global 1950 SST/1980 SST anomalies and perform simulations with the remaining SST anomalies. To obtain the effect of, for instance, the 1980s Atlantic SST anomaly, we calculate the difference between Case 80SST and the case, which SST anomalies is based on the global 1980s SST anomaly but without Atlantic Ocean's 1980s SST anomaly. This results in consideration of both direct and indirect effects of SST anomalies in the Atlantic, because when the forcing of a particular ocean SST is removed, both direct and indirect effects from that ocean's anomaly are eliminated. How large the difference might be or whether there is substantial difference between this and other traditional approach (i.e., adding each ocean's SST anomaly one by one) needs to be further investigated.

The WAMME II also includes an experiment to evaluate the sensitivity of the West African climate to LULCC and to compare the effect of LULCC on the Sahel rainfall reduction with that of SST anomalies. A similar strategy as in the SST experimental design, i.e., "idealized but realistic" anomalies, is applied to the LULCC experiment, in which a recently available land use map is employed (Hurtt et al. 2006). This map, which combines crop and pasture area change in past centuries over West Africa, was derived from global land-use history products (Goldewijk 2001; Ramankutty and Foley 1999) and employed for Phase 6 of the Coupled Model Intercomparison Project (CMIP6) LULCC experiment (Hurtt et al. 2011). The land map shows a clear pattern of substantial land use change over

the Sahel in the past century but LULCC largely levels off after the 1980s (Song 2013), which is generally consistent with the significant population increase, climate effect, and conservation measures in the late last century there. This part of the work is comprehensively analyzed in Boone et al. (2016, in this special issue). The initial atmospheric and boundary conditions, including SST, in the LULCC experiment are the same as in Case CTL in order to isolate the LULCC effect. A brief introduction of the LULCC experimental design and a comparison between the LULCC experiment runs (referred to as Case LULCC) and the SST runs will be discussed in Sect. 3.3.

3 Results

The impact of SST on the Sahel decadal variability and drought and its comparison with the effects of LULCC are analyzed in this section. Drought has a number of generally accepted definitions, such as meteorological, hydrological, agricultural, etc. The Sahel drought has already been studied extensively and there is no debate concerning whether or not this was a severe drought regardless of the chosen definitions. In this paper we are primarily interested in the meteorological aspects of the Sahel drought, and use the precipitation reduction from the 1950s to the 1980s to characterize the drought in the paper.

3.1 SST impact on the WAM spatial distribution and temporal evolution

3.1.1 Spatial anomaly patterns

The observed June–September (JJAS) precipitation, which is based on the CPC GTS, has a clear zonal pattern with three areas associated with high precipitation: one in the West African coastal area, one in the Ethiopian highlands, and another in Cameroon and Nigerian coastal areas (Fig. 2a and contour lines in Fig. 2b). The 3-mm day⁻¹ isohyet crosses the Sahel around 13°N, where Lake Chad is located. The WAMME II models (Fig. 2d–j, Figure ES1 in the Electrical Supplement) produce the precipitation highs over the West African coastal area and the Ethiopian highlands, as well as, except the UCONN CAM5, the high amount of precipitation near the Nigerian and Cameroon coastal area but with some differences in their geographical locations. In most models, the 3-mm day⁻¹ zonal isohyets accurately cross Lake Chad. The UCONN CAM5's 3-mm day⁻¹ isohyet, however, is about 2°–3° north of 13°N; while the GSFC GEOG-5 produces the 3-mm day⁻¹ isohyet along 13°N over West Africa. Over the eastern Sahel, however, this isohyet is about 2°–3° to the south of 13°N. In the ensemble mean (Fig. 2k), the two precipitation highs

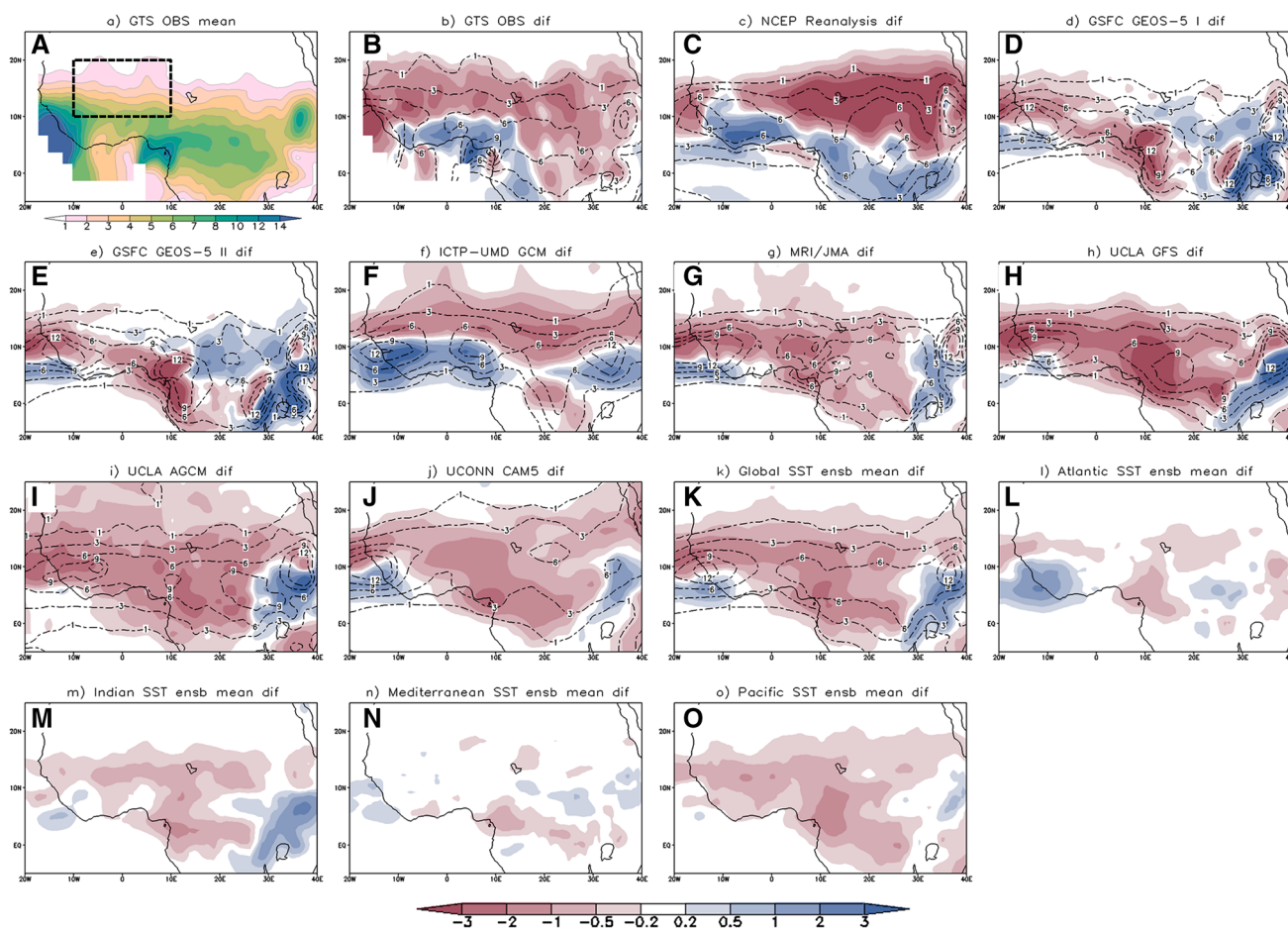


Fig. 2 JJAS mean observed and simulated precipitation (mm day^{-1}): **a** GTS climatology; **b** GTS precipitation difference between the 1980s and the 1950s; **c** Reanalysis I precipitation difference between the 1980s and the 1950s; **d–j** individual model simulated difference

between Case 1980SST and Case 1950SST; **k** ensemble mean difference for global SST effect; **l–o** ensemble mean for each ocean basin effect. The corresponding climatology from observation and simulations for each model are displayed as *contour lines* in **b–k**

in eastern and western Sahel, as well as the 3-mm day^{-1} isohyet along 13°N , are adequately reproduced. The high precipitation over the Nigerian and Cameroon coastal area, however, is slightly weaker mostly because the precipitation high of each model have a slightly different geographic location there (Fig. 2k, Figure ES1L). The main WAM precipitation band reaching sufficiently north is important in testing the models' response to SST and LULCC forcing. In Fig. 2, we display the 1-mm and 3-mm day^{-1} isohyet positions to examine the extent of the boreal summertime northward migration of the primary rainfall.

The observed JJAS precipitation differences between the 1980s and the 1950s show a severe deficit that represents the drought in the Sahel during the 1980s (Fig. 2b). The observation clearly shows a meridional dipole pattern. The areas with the most extensive precipitation deficits are over the latitude band between about 10°N and 20°N , mostly to the north of the 6-mm day^{-1} isohyet in the climatology.

To the south of this region, i.e., along the coastal area of the Bay of Guinea, there are positive anomalies (Fig. 2b), which indicates that the rainfall band in the 1980s was shifted to the south (Xue and Shukla 1993); thus the northern part of the monsoon precipitation band suffered the largest precipitation deficit associated with the drought. According to Fig. 2b, the maximum rainfall deficit was over West Africa. The average precipitation deficit between 10°N and 20°N and 10°W and 10°E , which covers most of the West African severe drought area and is referred to as the Reference Area in this paper (marked by a rectangle area in Fig. 2a), was 1.41 mm day^{-1} , which represents about a reduction of about 44 % compared with the climatology of the past 60 years (Table 3). Meanwhile, both positive and negative anomaly areas show strong spatial heterogeneity with many small peak cells. This pattern has been confirmed by the analysis based on rain gauge data and the small cells may reflect the natural variability of mesoscale

Table 3 Observed and model-simulated variable differences over the Reference Area

	Names	Precipitation (P)		Ts	Evapotranspiration (E)		VIMFC	
		Difference ^a (ΔP)	Relative difference (%) ^b		Difference ^a (ΔE)	$\Delta E/\Delta P$ (%)	Difference ^a ($\Delta VIMFC$)	$\Delta VIMFC/\Delta P$ (%)
Observation and Reanalysis	GTS	-1.41	-44					
	UDEL	-1.21	-35	0.96				
	CRU	-1.11	-33	1.11				
Models	Reanalysis I	-1.43	-51	0.85	-0.29	20	-1.14	80
	GSFC GEOS-5	-0.18	-7	0.29	-0.09	47	-0.10	53
	ICTP-UMD	-0.79	-20	0.62	-0.26	33	-0.53	67
	MRI/JMA	-0.93	-31	0.8	-0.28	30	-0.65	70
	UCLA GFS	-1.39	-53	1.17	-0.46	33	-0.93	67
	UCLA AGCM	-1.08	-35	1.16	-0.54	50	-0.54	50
	UCONN CAM5	-0.77	-20	0.67	-0.17	22	-0.60	78
	Ensemble mean	-0.86	-27	0.78	-0.30	35	-0.56	65

Units: precipitation: mm day⁻¹; surface temperature: K; Evapotranspiration: mm day⁻¹; Vertical Integrated Moisture Flux Convergence (VIMFC): mm day⁻¹

Reference Area: 10°N–20°N and 10°W–10°E

^a Differences for observation and Reanalysis are based on the differences between the 1980s and the 1950s. Simulated differences are based on 1980s SST and 1950s SST. The GSFC GEOS-5 results are based on the average of GEOS-5 I and GEOS-5 II

^b Precipitation relative difference: for observational data: observed difference/observational climatology; for simulations: simulated difference by one model/this model's climatology

Table 4 Observed and simulated differences over the Reference Area due to different forcings

Scenarios	Precipitation (P)		Ts	Evapotranspiration (E)		VIMFC	
	Difference ^a (ΔP)	Relative difference to observation (%) ^b		Difference (ΔE)	$\Delta E/\Delta P$ (%)	Difference ($\Delta VIMFC$)	$\Delta VIMFC/\Delta P$ (%)
GTS	-1.41		0.96 ^d				
Global SST	-0.86	61	0.78	-0.30	35	-0.56	65
India SST	-0.32	23	0.34	-0.13	41	-0.19	59
Pacific SST	-0.50	35	0.41	-0.12	24	-0.38	76
LULCC ^c	-0.60	43	0.50	-0.32	54	-0.23	46

Units: precipitation: mm day⁻¹; surface temperature: K; Evapotranspiration: mm day⁻¹; Vertical Integrated Moisture Flux Convergence (VIMFC): mm day⁻¹

Reference Area: 10°N–20°N and 10°W–10°E

^a Differences for observation and Reanalysis are based on the differences between the 1980s and the 1950s. Simulated differences are based on 1980s SST and 1950s SST. The GSFC GEOS-5 results are based on the average of GEOS-5 I and GEOS-5 II

^b Precipitation relative difference for simulations: simulated difference between Case 80SST and Case 50SST/observed difference between the 1980s and the 1950s

^c Land use land cover changes (LULCC) are based on the degraded vegetation map and control vegetation map

^d Data are based on UDEL observational data

convective activity (Le Barbé et al. 2002) that is missed due to the relatively low resolution of the GCMs.

Table 3 also consists of the statistics for other observational data sets, which include the University of Delaware (UDEL, Willmott and Matsuura 1995) and the Climate Research Unit (CRU, New et al. 1999) monthly mean

precipitation data sets. These two data sets represent very similar spatial precipitation differences between the 1980s and the 1950s (Appendix Fig. 15) but with less intensity (Table 3). Since we used the GTS data to conduct the SVD analysis, we will mainly compare the simulations with the GTS data in this paper.

The NCEP/NCAR Reanalysis I (Fig. 2c) shows a strong precipitation deficit ($-1.43 \text{ mm day}^{-1}$ over the Reference Area), comparable to the observed. However, despite the generally proper spatial coverage and intensity of the precipitation deficit over the Reference Area in the Reanalysis climatology (except for the missing precipitation high in the Cameroon and Nigerian coastal areas), the dominant drought areas are over the eastern Sahel and there is a large positive anomaly in eastern Central Africa and a positive anomaly in some northern West African areas, which were not found in the observation (Fig. 2b, Figures A1b and c). The relative reduction over the Reference Area is 51 % compared with the Reanalysis climatology, higher than the observed relative reduction (44 %, Table 3).

The simulated precipitation differences between Case 80SST and Case 50SST capture reasonably well the zonal rainfall deficit patterns associated with the drought with maximum strength over West Africa, but with different intensities and different features of spatial distribution of anomalies (Fig. 2d–j). For the WAMME II models, the precipitation reductions compared to their own climatology range from 7 to 53 % (Table 3). Among the models, the MRI/JMA GCM, UCLA AGCM, and UCLA GFS GCM, which adequately simulate the 3 mm day^{-1} isohyet (Fig. 2) and intensity in model precipitation climatology (for the first two models), show the larger impacts; while the NASA GEOS-5 GCM produces the least precipitation reduction over the Reference Area because its main rainfall reduction area is shifted to the south of that area. Nevertheless, the models show overall consistency in the positive effect of SST on rainfall reduction, especially over West Africa. Due to the coarse model resolution and the lack of mesoscale convective activity in GCMs, the drought patterns are much more spatially homogeneous than observed. The ensemble mean (Fig. 2k) indicates that with maximum SST forcing, the models reproduce about 60 % of the absolute amount of drought, 0.86 mm day^{-1} (Table 4), which is about a 27 % reduction of the model climatology (compared with 44 % reduction in observation) (Table 3). It should be pointed out that the climatology of the ensemble mean's precipitation is actually equal to the GTS's climatology (Table 3). In Fig. 2 and the following Fig. 3, the absolute differences larger than 0.5 mm day^{-1} are statistically significant at $\alpha < 0.1$ level with a 2-tailed T-test for the SST ensemble means. When we compute the ensemble mean, the GSFC GEOS-5 I and II results are averaged first to become one sample member in the ensemble mean.

The areas with positive anomalies are much smaller in most models as well as the multi-model ensemble mean than observed, mostly over the southwestern West African coast. Only the ICTP-UMD model, which has the 1 mm day^{-1} isohyet reach sufficiently north and simulates three precipitation highs in its climatology (Fig. 2f),

produces a strong dipole pattern, albeit the positive anomaly is too strong in eastern Central Africa compared to the observation. However, the UCLA AGCM, which also produces these features in its climatology, fails to produce the dipole in West Africa at all. The SST anomaly mode imposed on the SST forcing is based on an SVD analysis and the precipitation SVD1 does not show a strong positive anomaly along the coastal area (Fig. 1b). In addition, the SST variability associated with the interannual variability was filtered as discussed in Sect. 2. These factors may also contribute to the relatively weak positive anomaly in the simulation.

It should be pointed out that, since we also expect other external forcings such as LULCC to play a role in the decadal variability as will be discussed in Sect. 3.3, the rainfall reduction produced by SST forcing should be weaker than observed and the simulated spatial anomaly patterns may not be exactly the same as observed. However, the model results from the WAMME II experiments seem to support a stronger SST role in the Sahel drought when compared with the results from previous model inter-comparisons with SST forcing (e.g., Scaife et al. 2009).

The WAMME II also examines the effects of individual oceans. The relative importance of SST anomalies in different basins on Sahel rainfall has still been greatly debated (Rodriguez-Fonseca et al. 2015). Due to the scope of this paper, we mainly report the consensus and major discrepancies of the WAMME II results on the oceans' effects. The ensemble-mean simulated contributions of individual ocean basins are shown in Fig. 2l–o. The Pacific Ocean SST generates the largest impact (Table 4; Fig. 2o). In fact, all WAMME models produce a pattern with negative anomalies (not shown) due to the Pacific SST forcing similar to their corresponding ones for the global SST effects shown in Fig. 2, but with less intensity. SST anomalies in the Pacific alone produce about 35 % of the observed precipitation reduction, with strongest effects in West Africa and the western part of Central Africa. The Indian Ocean SST has the second largest impact, which accounts for about 23 % of the observed drought, with the main impact area over the central Sahel and also the western part of Central Africa (Table 4; Fig. 2m). There is an area with substantial and positive precipitation anomalies over eastern Central Africa, which does not have a counterpart in the observed differences between the 80 s and the 50 s. The WAMME models show little discrepancy in the Indian Ocean SST effect. Most models produce the drought with the Indian Ocean SST anomaly, with the NASA GEOS-5 being the only exception. The important role of the warming trends of the Pacific and Indian SSTs in contributing to the Sahel drought during the late twentieth century has been identified in a number of studies (e.g., Tippett and Giannini 2006; Caminade and Terray 2010; Mohino et al. 2011; Bader and

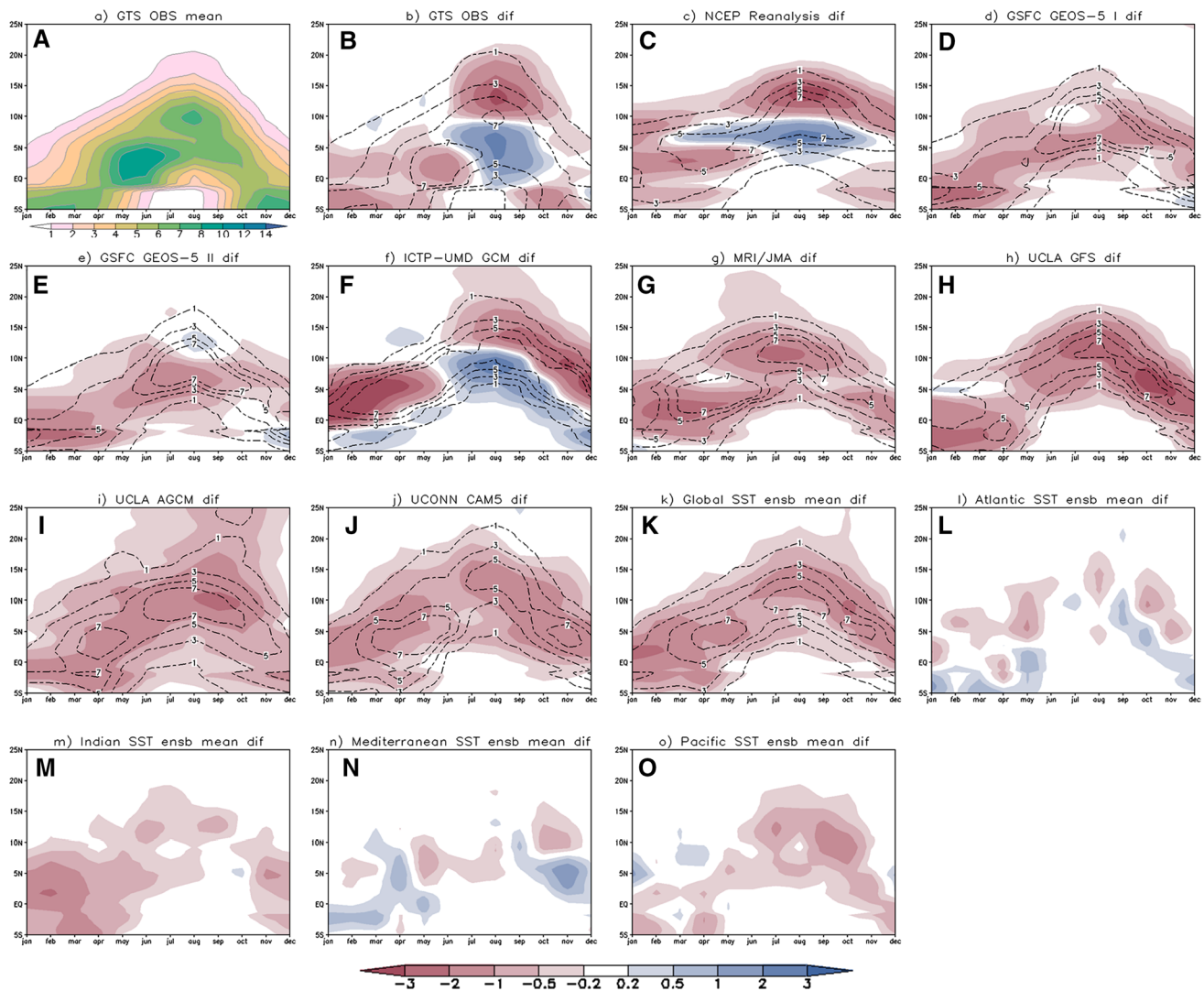


Fig. 3 Observed and simulated precipitation temporal evolution (mm day^{-1}): **a** GTS climatology; **b** GTS precipitation difference between the 1980s and the 1950s; **c** Reanalysis I precipitation difference between the 1980s and the 1950s; **d–j** individual model simulated difference between Case 1980SST and Case 1950SST;

k ensemble mean difference for global SST effect; **l–o** ensemble mean difference for each ocean basin effect. The corresponding climatology from observation and simulations for each model are displayed as *contour lines* in **b–k**

Latif 2011), and the WAMME II results are generally in line with these studies.

The Atlantic SST anomalies (Fig. 1d), which resemble the Atlantic Niño pattern (e.g., Carton et al. 1996), result in precipitation reduction over West Africa, the central Sahel, and parts of Nigeria and Cameroon and a very pronounced positive anomaly in the southwestern coastal area of West Africa in the WAMME II simulation (Fig. 2l). Several papers using different methodologies have shown that positive SST anomalies (Atlantic Niños) induce a dipole pattern of precipitation anomalies consisting of positive values along the coast of Guinea and negative ones over the Sahel (e.g., Horel et al. 1986; Wagner and Da Silva 1994; Fontaine and Janicot 1996; Ward 1998). For a warming in

the Atlantic, some studies show how this dipole is the result of a weakening in the sea level pressure gradient between ocean and land and, hence, in a southward shift of the ITCZ from its climatological position, which translates into more rainfall over Guinea and less rainfall over the Sahel. However, substantial discrepancies in simulating the Atlantic SST effect among AGCM models have been reported in Losada et al. (2010). The WAMME models also show quite large discrepancies in simulating Atlantic SST effects: while the GSFC GEOS-5, the MRI/JMA, and the UCLA AGCM show a large impact over West Africa, the ICTP-UMD GCM, the UCLA GFS, and the UCONN CAM5 show reduction in precipitation only over the eastern Sahel and some positive precipitation anomalies over West Africa

(Appendix Fig. 16). Moreover, each model shows negative anomalies but with large variability in the locations of dry areas over the Reference Area, such that overall the Atlantic SST produces only a weak precipitation reduction in the ensemble mean.

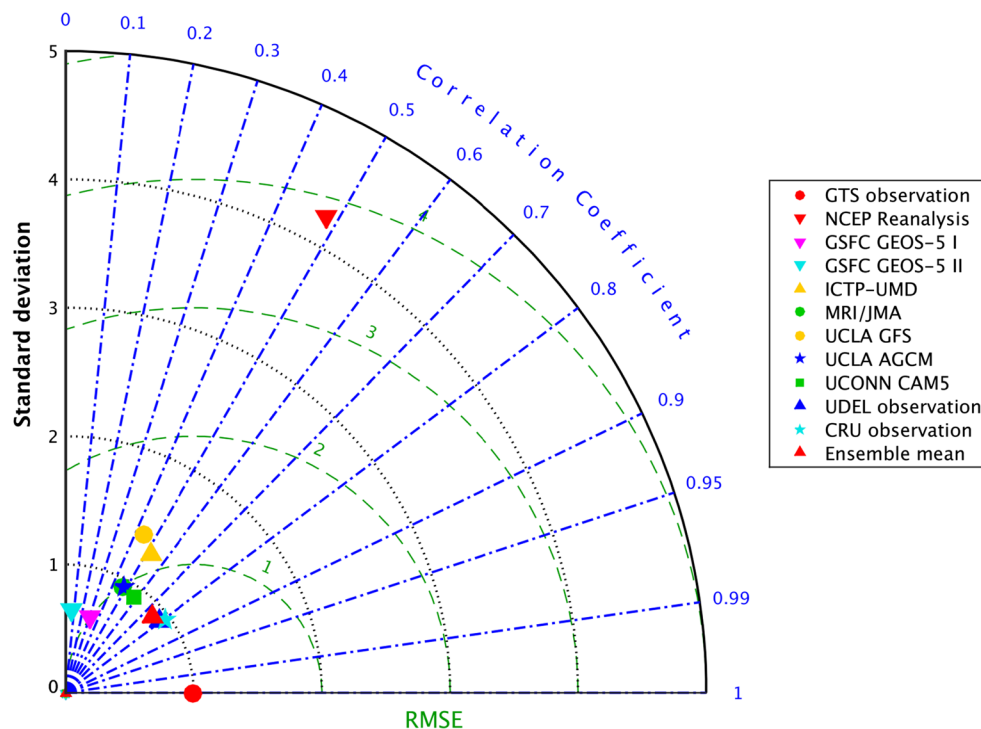
The simulated effects of Mediterranean Sea SST forcing are quite diverse among the models. While the ICTP-UMD, the MRI/JMA, the UCLA AGCM, and the UCONN CAM5 show negative precipitation anomalies with the Mediterranean SST, the GSFC GEOS-5 GCM and the UCLA GFS show strong positive anomalies (Appendix Fig. 17), which largely cancel out the negative anomalies produced by the other models. Overall, the ensemble mean does not capture any signal (Fig. 2n). Rowell (2003) demonstrates that increased evaporation over positive SST anomalies over the Mediterranean Sea leads to increased moisture content in the lower troposphere, which is advected southwards into the Sahel by the low-level flow across the eastern Sahara, resulting in stronger moisture convergence and precipitation over the Sahel. The negative SST anomaly there should produce the opposite results. Fontaine et al. (2010) also confirmed this relationship. We checked each model's results and found that models that produce the dry conditions with the Mediterranean SST anomaly have such a feature and models that do not show the Mediterranean SST effect produced the opposite wind flow. We report the WAMME II results' major discrepancies for these two ocean SST effects in this paper for a record to show the current status of the state-of-the-art WAMME GCMs' performance. A comprehensive analysis of the causes of these discrepancies is beyond the scope of this paper and will be the subjects of further study.

3.1.2 Temporal evolution

SST effects are also manifested in the annual evolution of precipitation. Figure 3a, b show the annual evolutions of the observed climatological precipitation and of the difference between the 1980s and the 1950s, respectively, averaged over 10°W and 10°E. The annual evolution of precipitation and its anomaly follows the ITCZ movement. The anomaly starts developing in winter and spring. After the West African monsoon jump in late June (Fig. 3a, b), a strong dry anomaly peak appears during the summer monsoon season and the early monsoon retreat; meanwhile, a positive anomaly to the south after the monsoon jump is also evident. Later, in the fall, there is no substantial anomaly. The areas associated with large precipitation anomalies are mostly located to the south of the 1 mm day⁻¹ isohyet. Past studies have suggested that differential heating of the northern and southern hemisphere may lead to a meridional shift of the ITCZ, creating the dipole rainfall anomaly in the WAM/Sahel (Xue and Shukla 1993; Zhang and

Delworth 2006; Knight et al. 2006; Ting et al. 2011). The WAMME II models generally capture this anomaly evolution, albeit with significant different spatial and temporal characteristics among individual models, including the Reanalysis I (Fig. 3c, j). The Reanalysis I shows little latitudinal movement for both its climatology and anomalies during different seasons, with negative anomalies persisting throughout the entire year over West Africa and positive anomalies over southern West Africa over spring and summer (Fig. 3c). The evolution of the GEOS-5 II is mostly confined to the south of 10°N, while that of the UCLA AGCM extends much further north of 20°N. In general, the model-simulated anomalies are located to the south of the 1 mm day⁻¹ isohyet. The model-produced anomalies are stronger than the observed during the periods before and after the monsoon season (Fig. 3d–k), which is probably due to our experimental design, i.e., imposing the same SST anomaly for every month (Sect. 2). Another noticeable difference is that most WAMME models, except for ICTP-UMD, do not show a positive precipitation anomaly in summer to the south of the Sahel. By and large, the anomaly evolution process is properly produced as shown in the ensemble mean (Fig. 3k), but the maximum rainfall suppression during the monsoon season is weaker than observed. We expect that other external forcings also contribute to the drought. Most pre-monsoon and monsoon onset studies focus on May or, at the earliest, April (e.g., Sultan and Janicot 2000). This study reveals that the SST forcing produces an anomalous displacement of the ITCZ before the WAM onset; the precipitation anomaly starts to develop much earlier when the ITCZ is still near the equator under the SST effect. Since the WAMME II simulated annual evolution processes, the results suggest a potential predictability for WAM precipitation linked to SST.

The effects of SST anomalies in different oceans on the anomaly evolution show different characteristics. The Pacific Ocean's effect is strong mainly during the monsoon season and after the retreat (Fig. 3o). The average precipitation reduction over West Africa (8°N–20°N and 10°W–10°E) during June to October is 0.54 mm day⁻¹. In particular, during September and October, it is 0.72 mm day⁻¹. During other seasons, the reduction over West Africa varies from 0.01 to 0.13 mm day⁻¹. The effects are small during the ITCZ northward movement from January to May. The precipitation reduction averaged over 5°S to 8°N and –10°W to 10°E is only 0.16 mm day⁻¹. The Atlantic Ocean's effect (Fig. 3l) acts throughout the entire year with distinct dipole patterns and intermittent pulses but is relatively weak. The role of Atlantic SST in producing this dipole has been discussed in several studies (e.g., Knight et al. 2006; Mohino et al. 2011; Dieppois et al. 2013). The Indian Ocean SST (Fig. 3m) has the largest impact in the winter and early spring during the period

Fig. 4 Taylor diagram for JJAS precipitation anomaly

when the ITCZ moves northward. During January–May, the mean precipitation reduction is $-0.97 \text{ mm day}^{-1}$ over 5°S to 8°N and 10°W to 10°E , then reduces to only 0.15 mm day^{-1} during JJAS. It has substantial impact during the WAM season; the average reduction over West Africa is 0.30 mm day^{-1} . The results here suggest that variability in the Indian and probably the Atlantic Ocean plays an important role in triggering the West African monsoon anomaly by contributing to the ITCZ anomalous displacement and modulating atmospheric circulation well before the monsoon onset, while the Pacific Ocean SST anomaly greatly contributes to the summer drought. We will discuss this issue further in the mechanisms section. The Pacific and the Indian Oceans' contributions over the Reference Area, on which the WAMME II models have consensus, are summarized in Table 4.

3.1.3 Taylor analyses

To further quantitatively evaluate the WAMME models' performance in simulating the difference between the 1980s and the 1950s precipitation, we use a Taylor diagram (Taylor 2001) that encapsulates statistical comparisons of the WAMME model runs' spatial anomaly pattern as well as temporal variability with the observed anomalies (Fig. 4). For comparison, UDEL and CRU are also shown in this figure. The results in the figure are based on the June-through-September monthly mean differences over the 5 years at every grid cell over the land points within 20°W – 30°E

and 8°N – 25°N . The JJAS mean difference between Case 80SST and Case 50SST for each model is removed when calculating root-mean-square-error (RMSE). Therefore, this diagram does not directly show the model bias from the observed difference between the 1980s and the 1950s as listed in Table 3. The position of each symbol appearing on the plot quantifies how closely model's simulated and observed precipitation differences and their variability match CPC GTS observations (Fig. 4). The standard deviation, correlation coefficient, and RMSE are used to evaluate the model's performance (see Xue et al. (2010a) for more detailed explanations for the Taylor diagram analysis). The closer the model's symbol to the observation point, the better the simulation is. The point representing GTS is located at the unit distance of the horizontal radius (red dot in Fig. 4),

The figure shows that models standard deviations and RMSE are close to observation. The models with the best standard deviation results include the UCLA AGCM, UCONN CAM5, and MRI/JMA. The GSFC GEOS-5 models exhibit the lowest standard deviations and low correlation coefficients among the models because their simulated droughts areas shift to the south compared with observed. The correlation coefficients for most models are in the range 0.4–0.6, which is less than the correlations between GTS and other observational data sets, ~ 0.8 . The best correlated results are those generated by the ICTP-UMB and the UCONN CAM5 models, with coefficients equal to 0.52 and 0.58, respectively. However, the ensemble mean shows

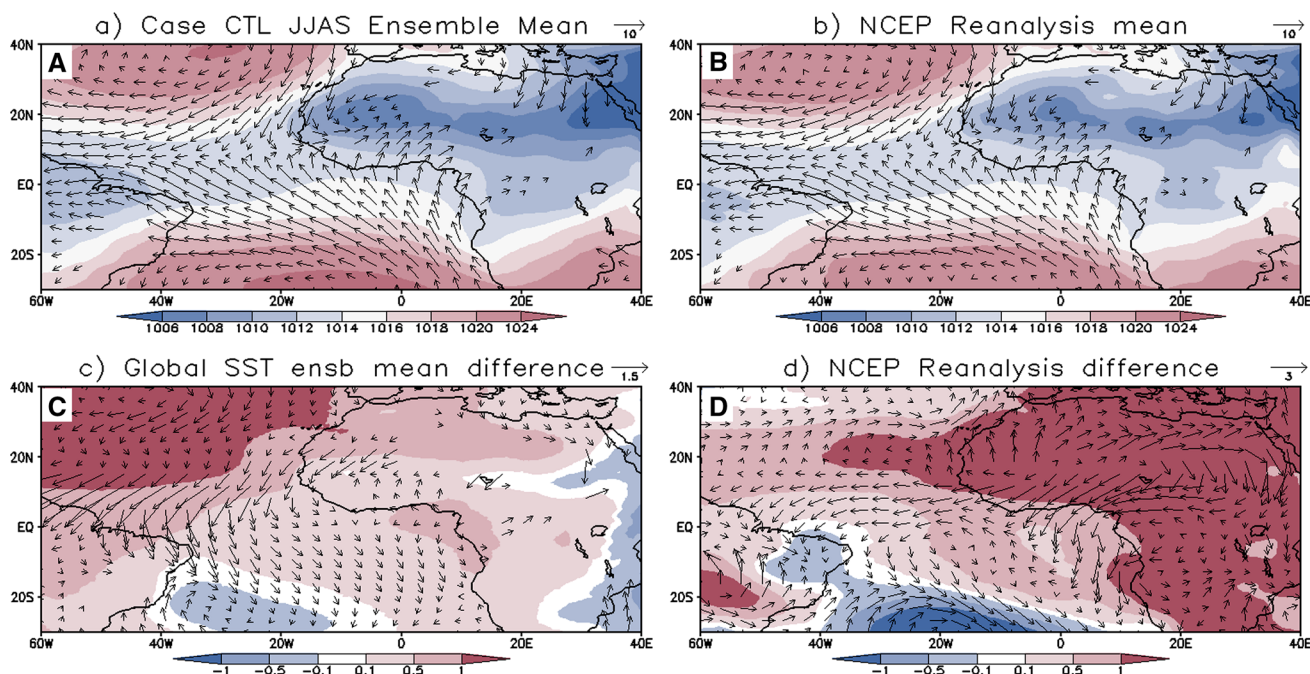


Fig. 5 Simulated JJAS mean and anomalies of 950 hPa wind vectors (m s^{-1}) and SLP (hPa): **a** Case CTL ensemble mean; **b** NCEP Reanalysis climatology; **c** Case 80SST–Case 50SST ensemble mean; **d** NCEP Reanalysis difference between the 1980s and the 1950s

superior performance and its difference from GTS is consistent with other observational data sets. The ensemble mean's RMSE, standard deviation, and correlation coefficients are 0.67, 0.9, and 0.75, respectively, which is comparable to UDel's 0.62, 0.92, 0.79 and CRU's 0.62, 0.80, and 0.92, respectively. Meanwhile, the Reanalysis I shows large discrepancy with the observations. Although its correlation coefficient, 0.48, is similar to other models, the Reanalysis I exhibits much larger RMSE (3.85) and standard deviation (4.23), as it overestimates the precipitation difference and places maximum rainfall deficit to the east of the observed location. By and large, the Taylor diagram suggests that the WAMME II models' results are generally quite consistent and the impacts produced by the ensemble mean probably properly represent the best approximation to observations.

3.2 Impact of SST on circulation and surface energy balance

SST anomalies remotely affect WAM precipitation by modulating the regional circulation. This section focuses on this modulation and changes in associated processes. As discussed in the previous section, the ensemble mean represents the WAMME models' consensus on the SST effect and shows its difference from the CPC GTS (after removing the bias in simulating the difference between the 80 s and the 50 s) is comparable to other observational data sets (Fig. 4); hence we mainly discuss the results

from the ensemble mean. In fact, WAMME I also shows that WAMME multi-model ensembles have produced good WAM seasonal mean precipitation and surface temperature spatial distribution, intensity, and variability, better than reanalyses in many respects when compared with observations (Xue et al. 2010a, b). Therefore, the impact characteristics shown in the ensemble mean probably are more suitable to apply for interpretation of the simulated impact in this study. Different characteristics of different models will be discussed only when the model(s) exhibit abnormal features compared with the ensemble mean. The discussions in this section emphasize the global SST results and identify the oceans that the WAMME models have consensus and contribute to the major characteristics of the global SST effects. The mechanisms discussed in this section will be compared with those from the LULCC effect, which will be presented in the following section.

Low-level moisture transfer is an important feature in WAM development. The JJAS low-level wind vectors and sea level pressure from the ensemble mean in Case CTL are shown in Fig. 5a. The relatively low sea level pressure over $\sim 0^\circ\text{W}$ to 5°E and $\sim 20^\circ\text{N}$ to 30°N is well simulated. Westerly and southwesterly flow from the Atlantic brings moisture into West Africa, which merges with northeasterly Harmattan winds, generating convergence in the Sahel during the monsoon season. Figure 5a accurately produces the major features as shown in the NCEP Reanalysis I climatology (Fig. 5b). The difference between Case 80SST and

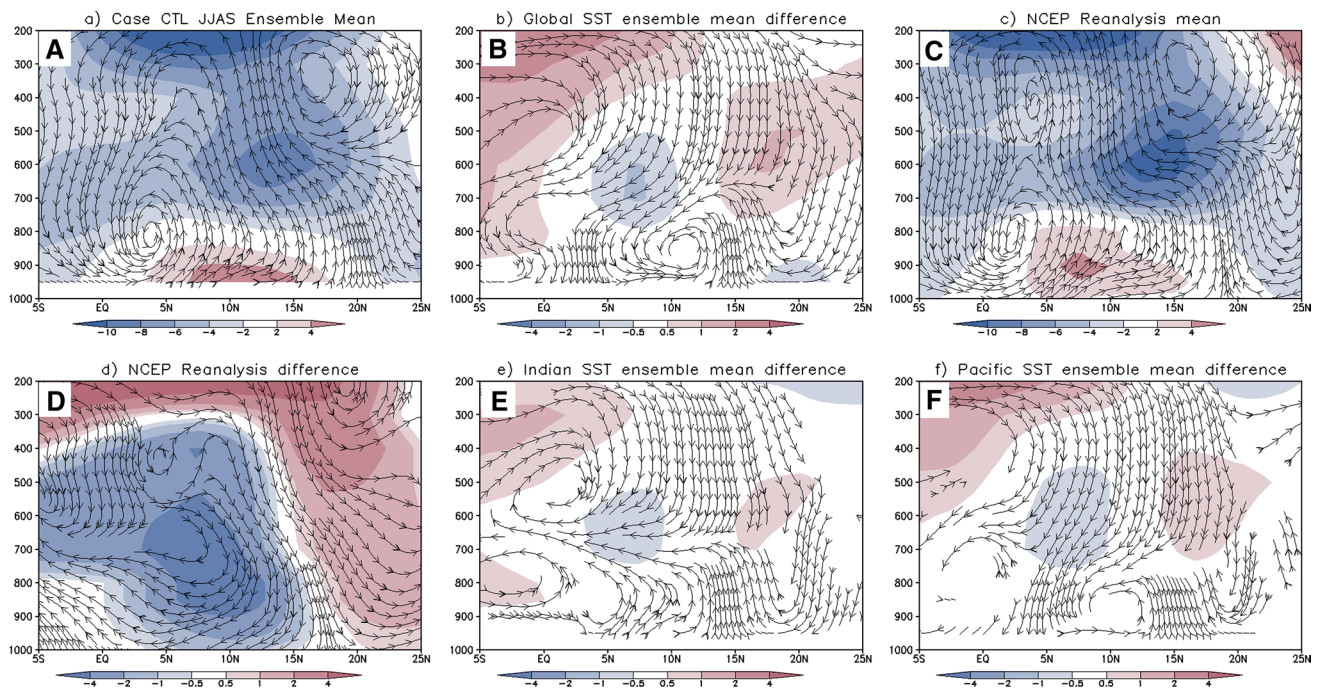


Fig. 6 Latitude-height cross section of JJAS mean and anomaly streamline (v , $-100 \times w$) averaged between 10°W and 10°E : **a** Case CTL ensemble; **b** Case 80SST–Case 50SST ensemble mean; **c** NCEP Reanalysis climatology; **d** NCEP Reanalysis difference between the

1980s and the 1950s; **e** Indian Ocean SST effect; **f** Pacific Ocean SST effect. *Note* The average of zonal wind or zonal wind differences are in *shade* (m s^{-1})

Case 50SST due to global SST effect shows that southwesterly flow in the West African coastal area becomes weaker, northeasterly flow becomes stronger, and sea level pressure over the Sahara becomes higher (Fig. 5c). The NCEP Reanalysis I also shows similar features (Fig. 5d), but with a much larger wind and sea level pressure difference over the African continent, which may be due to NCEP Reanalysis I data issue associated with sea level pressure data in 1950's and early 1960's (Yang et al. 2002). To more clearly delineate the circulation change, Fig. 6 shows the latitude-height JJAS mean streamlines and their differences between the Case 80SST and Case 50SST ensemble means. The average of zonal wind or zonal wind difference are shown in shading in Fig. 6 to help comprehend the following discussions.

Due to higher temperature over the land surfaces in the warm season, the monsoon inflow is seen near the surface between 5°N and 20°N , with ascending motion across these latitudes (Fig. 6a), contributing to the major WAM precipitation. Meanwhile, the northerly flows that occur in the middle troposphere between $\sim 15^\circ\text{N}$ and 22°N around 600 hPa as well as over the region of deep convection ($\sim 5^\circ\text{N}$ – 10°N) at ~ 200 – 250 hPa correspond to the African easterly jet (AEJ) and to the tropical easterly jet (TEJ) (Fig. 6a), respectively. Those simulated features are quite consistent with the Reanalysis I (Fig. 6c). Compared to Case 50SST, Case 80SST has a shallow layer with relatively stronger rising

air between $\sim 13^\circ\text{N}$ to 20°N (Fig. 6b), which is associated with high surface temperature there as will be discussed later. This rising motion converges with the widely spread relative sinking motion above 800 hPa, which is associated with the relative divergence over most parts of the Sahel and the reduction in precipitation, enhancing the northerly flow around 700 hPa. The southern flank of AEJ is enhanced and the northern flank of AEJ weakens. Near the equator, the relative sinking motion induces a relative southerly flow near ~ 200 to 300 hPa, weakening the TEJ, which is consistent with climate features in the Sahel dry years: a stronger AEJ or a southward shift of its location as well as a weaker TEJ favoring more precipitation over the Guinea coastal region and less precipitation over the Sahel (Xue and Shukla 1993; Cook 1999). The WAMME II ensemble mean of Case 80SST minus Case 50SST correctly reproduces these climate features associated with the Sahel drought. The Reanalysis I, on other hand, does not capture these features and, instead, exhibits widely spread and very strong zonal wind changes (Fig. 6d) that have not been reported in any analyses on WAM variability. Among the WAMME II models, most produce a circulation change similar to the ensemble mean, while the ICTP-UMD GCM has a distinct response with a homogeneous sinking to the north of 12°N and uniform rising motion to the south; no horizontal wind change appears. Among the four oceans examined, the circulation

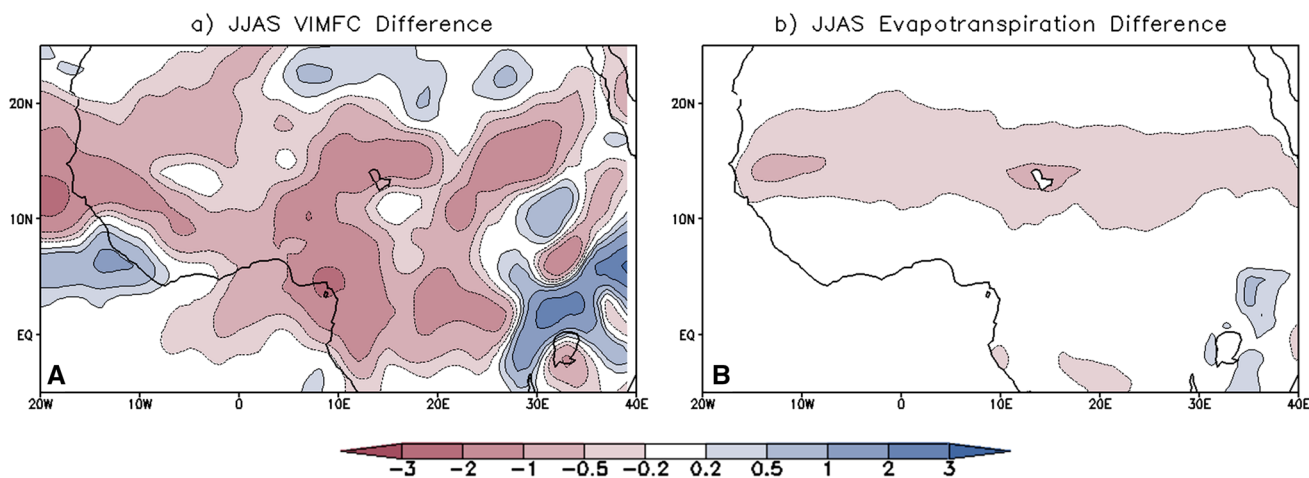


Fig. 7 JJAS ensemble mean difference between Case 80SST and Case 50SST (mm day^{-1}), **a** vertical integrated moisture flux convergence (VIMFC) and **b** evapotranspiration

changes due to the Pacific SST and Indian Ocean SST seem to be more similar to the aforementioned climate features (Fig. 6b, e, f).

Moisture flux convergence and surface evapotranspiration (ET) are the main moisture sources for precipitation. Reanalysis I shows the largest change in vertically integrated moisture flux convergence (VIMFC), which amounts to about 80 % of the simulated changes in precipitation between the 80 s and the 50 s, and relatively low ET change amounting to about 20 % (Table 3). Both VIMFC and ET are model products in the Reanalysis and depend on its land model. Our previous WAMME I study shows that the surface heat fluxes, including the latent heat flux, in the Reanalysis products have large discrepancies with the products of the AMMA land-surface model intercomparison project (ALMIP) based on the AMMA observation over West Africa (Boone et al. 2010). The low ET to precipitation ratio here may result from the simple land model in Reanalysis I. Our earlier studies suggest that the GCM coupled with this simple land scheme has difficulties with land/atmosphere interactions (Xue et al. 2004b, 2010b). In addition, Fig. 6d shows that the southerly wind at ~ 600 hPa is enhanced in Reanalysis I in the 1980s, which is at odds with the climate features during the dry years (Fig. 6b). Discrepancies in producing the shallow meridional circulations in the Reanalyses data in the West African region have been reported (Zhang et al. 2008).

The reductions between Case 80SST and Case 50SST of the VIMFC and ET under global SST forcing range from about 50–80 and 50–20 %, respectively, of each WAMME II model's precipitation difference (Table 3). For the model ensemble mean, VIMFC accounts for 65 % of its precipitation difference and ET for 35 % (Table 3). Among the models, the GSFC GEOS-5 GCM and UCLA AGCM show high

ET/precipitation ratio, about 50 %, and the UConn CAM5 shows the lowest ET/precipitation ratio, (~ 22 %). The SST anomalies affect Sahel precipitation through modulating the large-scale circulation. The change in moisture transport is apparently the dominant factor contributing to the precipitation difference in the WAMME models' simulations and in the ensemble mean. Figure 7a, b show the JJAS VIMFC and ET differences, respectively, between Case 80SST and Case 50SST for the ensemble mean. The VIMFC changes between Case 80SST and Case 50SST are consistent with the circulation changes (Figs. 5, 6). The wide spread sinking motion between the equator and 15°N (Fig. 6b) are consistent with the negative VIMFC over these areas (Fig. 7a); within $\sim 12^{\circ}\text{N}$ and 16°N and 10°W and 10°E , the relative shallow rising motion (Fig. 6b) compensates for this change. The spatial distributions of the VIMFC difference of the ensemble mean for the global SST effect (Fig. 7a) and each ocean SST effect (not shown) are generally consistent with the regions of large precipitation differences as shown in Fig. 2 but with lower magnitude. These regions include the Sahel, West African coastal area, and central Africa; while the ET changes are mostly confined within the semi-arid Sahelian zone (Fig. 7b). The averaged ET changes due to each ocean's SST effect are around 0.1 mm day^{-1} (Table 4) and are not shown here. The annual temporal evolutions of VIMFC (Fig. 8a–c) for the global and Pacific and Indian ocean's SST effect are also very much consistent with the precipitation evolution, which moves from 5°S to 20°N then retreats in the annual evolution. For instance, the Indian SST effect shows large impact before the monsoon season while the Pacific SST effect shows large impact during the monsoon season and after the monsoon retreat. The changes of ET rates show less latitudinal movement, limited to within the semi-arid area between 5°N to 20°N

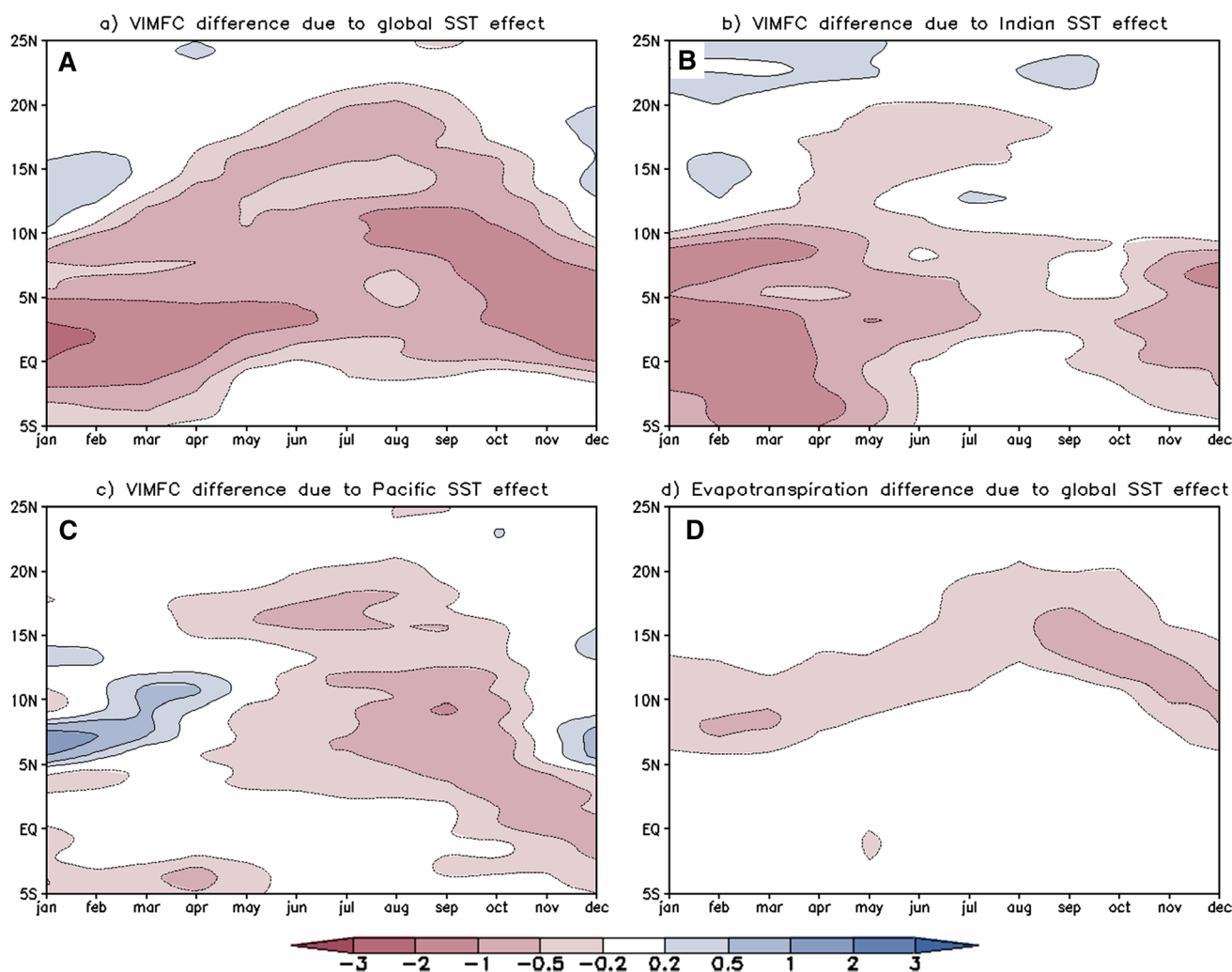


Fig. 8 **a** Simulated VIMFC difference temporal evolution from Case 80SST–Case 50SST ensemble mean; **b** same as **a** but for Indian Ocean SST effect; **c** same as **a** but for Pacific Ocean SST effect; **d** simulated evapotranspiration evolution. Units: mm day^{-1}

(Fig. 8d), and large reduction occurs after the monsoon season. This confirms that the change in moisture flux convergence, which is closely related to the circulation change, is probably the main mechanism causing the rainfall change in these experiments.

The SST influences the large scale circulation and MFC, which interact with cloud then affecting the surface energy balances. About 90 % of Sahel rainfall comes from intense, organized convective activity (Mathon et al. 2002). In addition, low-level continental stratus clouds frequently occur over large regions of southern West Africa during the WAM wet season (Stein et al. 2011; Schrage and Fink 2012). Associated with the widespread relative sinking motion (Fig. 6b), the cloud cover in the ensemble mean is reduced in Case 80SST, especially during the early spring and during the monsoon season (Fig. 9a), in association with enhanced downward shortwave radiation (Fig. 9b). Corresponding to this enhancement, the upward shortwave

radiation is increased by a fraction depending on the surface albedo over the land area (north of 5°N) (Fig. 9c). Over the ocean (south of 5°N in the figure), the change in upward shortwave radiation is very small because of the low albedo. The reduction of downward longwave radiation in the area to the north (~10°N–15°N, Fig. 9d) is primarily due to a decrease in cloudiness. To the south of this area, there is not much change in downward longwave radiation as enhanced humidity due to the rising motion near the surface layer may compensate for the effect of reduced cloud cover. The upward longwave radiation changes (Fig. 9e) are consistent with the surface temperature changes and will be discussed next. The overall net radiation changes (Fig. 9f) exhibit a dipole pattern: to the south of ~10°N, the increased downward short wave radiation is dominant, while to the north of this latitude, the reduced downward long wave radiation and enhanced upward long wave radiation are dominant. Over the Reference Area in JJAS, the net radiation reduction is

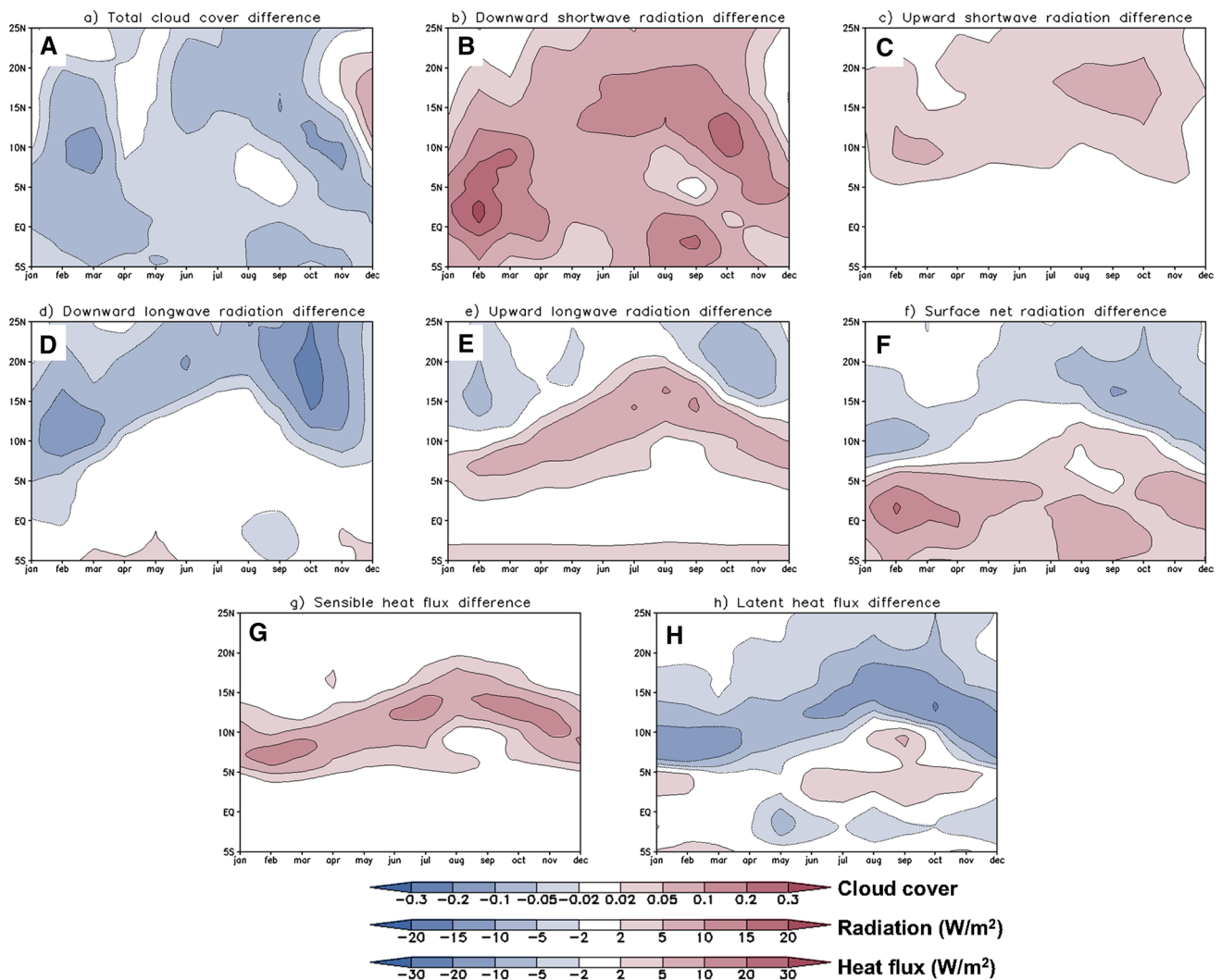


Fig. 9 Simulated temporal evolution of surface energy components differences between Case 80SST and Case 50SST, **a** total cloud cover; **b** downward short wave radiation; **c** upward short wave radiation; **d** downward long wave radiation; **e** upward long wave radiation; **f** surface net radiation; **g** sensible heat flux; **h** latent heat flux. Units: Total Cloud Cover: %; Others: W m^{-2}

2.6 W m^{-2} for the ensemble mean, which is very different from Reanalysis I, which has an increase of 8.65 W m^{-2} . Table 5 summarizes these changes in the JJAS radiation component averaged over the Reference Area. The changes in annual radiation are similar to those in JJAS.

The temporal evolution of surface temperature changes shows a dipole pattern (Fig. 10a, b). To the north of 15°N , the reduction in net radiation is dominant such that the surface temperature becomes cooler, mainly in fall, winter, and early spring. To the south, the changes in surface energy balance are dominant (Table 5); the increase in surface temperature is consistent with the temporal evolution of increased downward shortwave radiation (Fig. 9b) and reduced latent heat flux (Fig. 9h). This feature appears to be quite robust as every WAMME II model produces a warming over this area (Table 3), consistent with the observed

warming that is based on the CPC GTS data in the Sahel during the 1980s (Table 5). Based on the analyses of surface temperature modes and their relationship with precipitation evolution in the WAMME I experiment, we identify two summer temperature modes (Xue et al. 2010a): the Sahara mode (mostly covering the Sahara area between 20°N and 30°N) and the Sahel Mode (mostly covering the Sahel area between 10°N and 20°N). The crucial role of the Sahara Heat Low in WAM development is well known (Lavaysse et al. 2009; Biasutti et al. 2009; Evan et al. 2015) and corresponds to the Sahara mode. Meanwhile, the WAMME I results also reveal the important role of another mode, the Sahel mode, and show that the summer WAM precipitation northward movement/retreat is closely associated with an enhanced/weakened Sahara mode and a weakened/enhanced Sahel mode. In this study, the simulated

Table 5 Surface energy balance differences over the Reference Area

	CLD	SWD	SWU	LWD	LWU	Net rad	LH	SH	Temperature
Observation									0.96
Reanalysis I	-0.13	26.24	6.95	-4.93	5.71	8.65	-8.40	17.09	0.85
Global SST	-0.07	10.99	4.00	-3.37	6.22	-2.60	-8.61	5.70	0.78
LULCC	-0.03	6.91	10.87	-1.68	3.61	-9.25	-9.32	-0.23	0.50

Units: surface temperature: K; All other variables: w m^{-2}

Reference Area: 10°N–20°N and 10°W–10°E

Observation and Reanalysis I are based on the differences between the 1980s and the 1950s

Global SST is based on Case SST80 minus Case SST50

LULCC is based on the difference between Case LULCC and Case CTL

CLD total cloud cover, SWD/SWU short wave radiation down/up, LWD/LWU long wave radiation down/up, Net Rad net radiation, LH latent heat flux, SH sensible heat flux

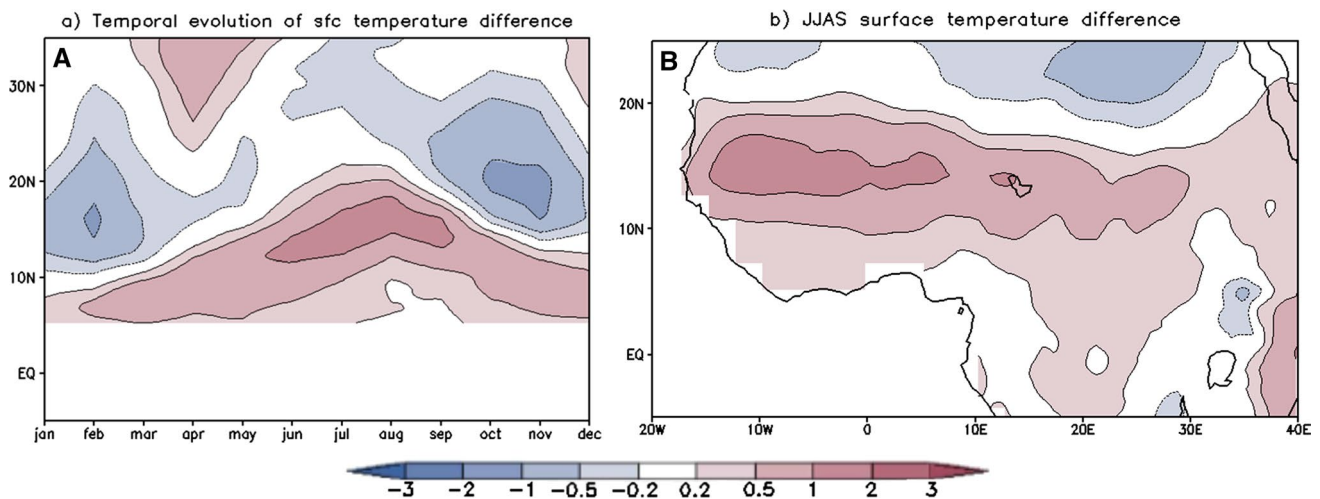


Fig. 10 Simulated temperature difference between Case 80SST and Case 50SST ensemble mean **a** evolution, **b** JJAS difference. Unit: °C

rainfall reduction due to SST effects in the 1980s is also associated with the enhanced Sahel mode and weakened Sahara mode (Fig. 10a, b), in agreement with the WAMME I discoveries. The observed surface temperature in Sahara did not show substantial warming before the 1980s. The effects of the SST and the greenhouse gases on the surface temperature over Sahara seem to have opposite sign before the 1980s. These two modes' changes are the results of the surface energy balance, which are closely related to the circulation/cloud cover change, and also provide a feedback to the changes in WAM precipitation evolution. The present study shows that such dipole changes occur not only during summer (Fig. 10a, b), but also during pre-monsoon development and monsoon retreat. It is a further task to explore how changes in the pre-monsoon season contribute to changes in the circulation and then the summer drought. None of the individual ocean basins seem to contribute dominantly to the surface temperature anomaly patterns (not shown).

3.3 Characteristics of LULCC effect

In the LULCC experiment, we impose a simple LULCC scheme, which represents a maximum feasible scenario. The LULCC data (Hurtt et al. 2006) are prescribed only over West Africa in order to isolate the regional impacts. The areas having conversion from natural vegetation cover are shown in Fig. 11a, which defines the LULCC mask. We then impose a simple set of land class change rules within this area, which is applied to the input land cover datasets for several GCMs. After imposing the LULCC experimental scenario, broadleaf trees decrease by upwards of 30 %, and grasses decrease by approximately half that value; shrublands decrease in the northern Sahel but increase to the south (mostly in place of decreased forest). The bare soil fraction increases upwards of 30 %. Consequently, these land cover changes result in corresponding changes in land surface variables. For instance, the increase of local albedo reaches as high as 0.10 over some areas, but the average

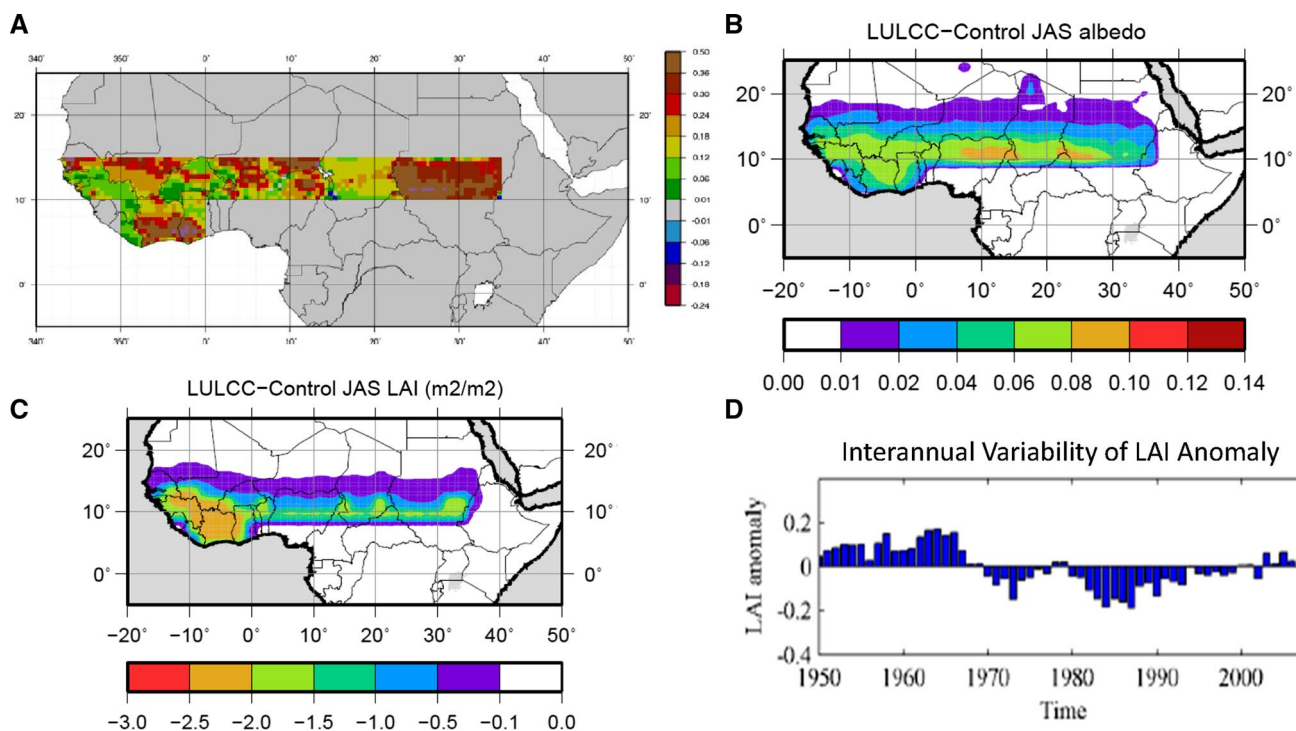


Fig. 11 LULCC experimental design: **a** fraction changes of crop and pasture area between the 1990s and the 1950s (based on Hurtt data (2006)); **b** ensemble mean albedo difference between Case LULCC

and Case CTL; **c** same as **b** but for LAI; **d** dynamic vegetation model-simulated evolution of LAI anomaly over Sahel

increase over the LULCC zone is approximately half of that value (Fig. 11b); the leaf area index (LAI) is also changed accordingly as shown in Fig. 11c. More comprehensive and detailed discussions on the LULCC experiment and its results are presented in Boone et al. (2015, this special issue). In this paper, we show the difference between Case LULCC and Case CTL, which uses a normal or climatology vegetation map, to highlight the LULCC effect.

The JJAS precipitation differences between Case LULCC and Case CTL in Fig. 12a show the land forcing effect. The maximum impact is generally in the land degradation area as shown in Fig. 11a with West Africa having the strongest anomaly. South of this region there are some positive anomalies. Since the LULCC and SST forcings produce very similar spatial patterns, it is sometimes hard to distinguish the contribution of each one to the anomaly patterns shown in Fig. 2b. While the SST effects in most individual models and in the multi-model ensemble mean feature the strongest rainfall deficit signal in the western part of Central Africa, the LULCC effects feature a wet signal there. This wet signal is a feature in the observed precipitation anomalies accompanying the 1980s Sahel drought that cannot result from SST effects in most models. Moreover, the SST impact is apparently larger than that of the LULCC. When we impose the maximum feasible forcing for each case, the SST could produce up to 60 % of the observed precipitation difference

between the 1980s and the 1950s and LULCC produces less but still a substantial amount of observed precipitation difference for the same time period, 43 % (Table 4). In this study we use precipitation anomalies between the 1980s and the 1950s to characterize the Sahel 1980s drought. In the simulations, the SST and LULCC forcings both are prescribed all year long. However, in the SST experiment the precipitation anomaly starts developing from winter near the equator and follows the ITCZ movement afterward (Fig. 3k). In contrast, the precipitation anomaly associated with the LULCC starts developing only when the precipitation band moves to the region where the forcing is applied (Fig. 12b), with the maximum response in the summer monsoon months. These features highlight the role of SST in triggering the Sahel drought and the role of land surface processes in responding to and amplifying the drought.

In contrast with the SST effect, the sea level pressure in Case LULCC shows no noticeable changes over the Sahara but weakens over West Africa (Fig. 13a). The wind from the Bay of Guinea becomes weaker (Fig. 13a). The circulation change due to LULCC is mainly within the 5°N–20°N latitude band with a very similar pattern to what the SST forcing does. A strong sinking or weakened rising motion between 15°N and 20°N is presented (Fig. 13b). The SST effect on precipitation is mostly through change in VIMFC with the ratio of VIMFC and precipitation change being

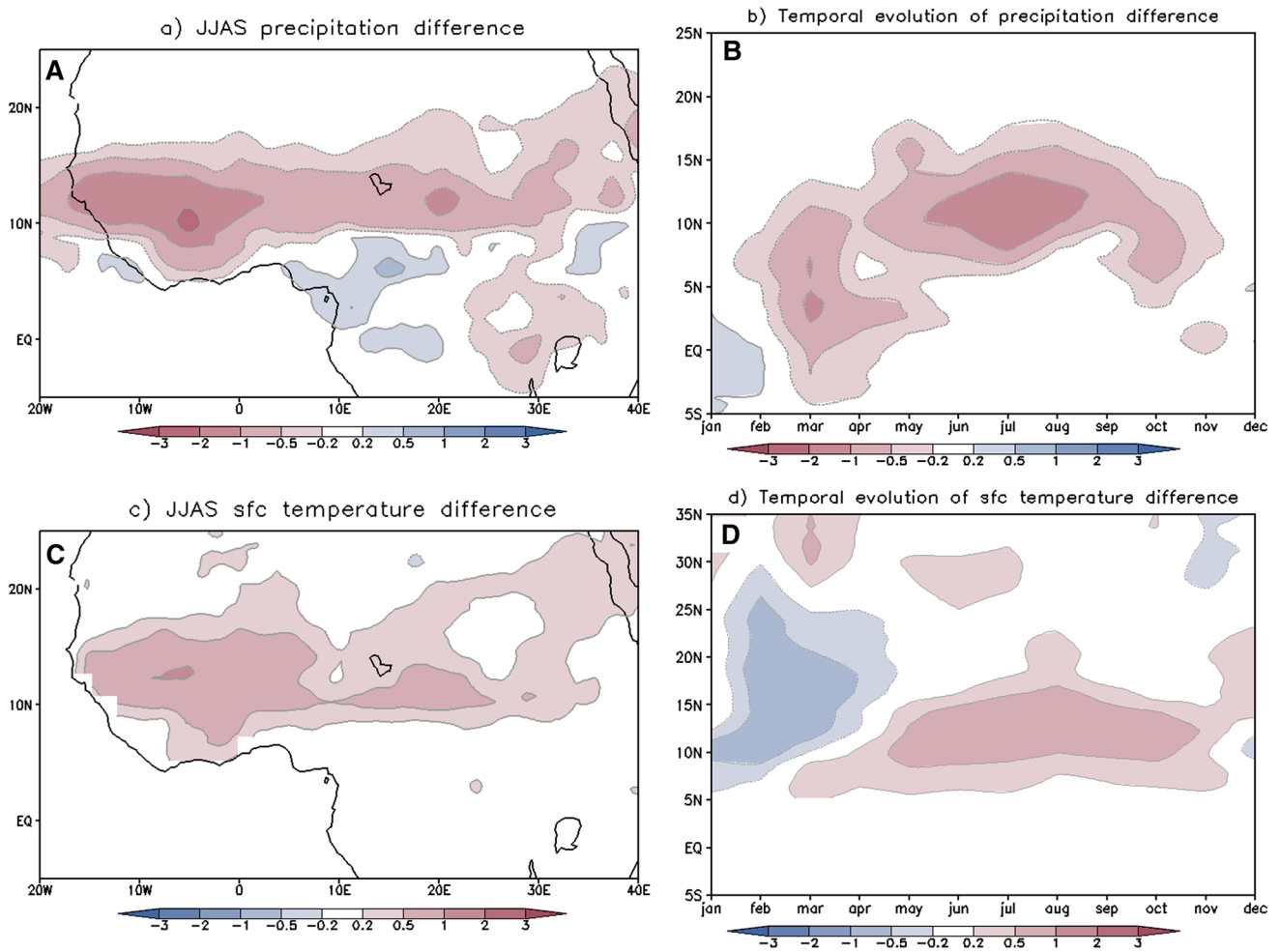


Fig. 12 a JJAS mean precipitation difference between Case LULCC and Case CTL; b temporal evolution of precipitation difference between Case LULCC and Case CTL; c same as a but for surface

temperature; d same as b but for surface temperature. Units: Precipitation: mm day^{-1} ; Temperature: $^{\circ}\text{C}$

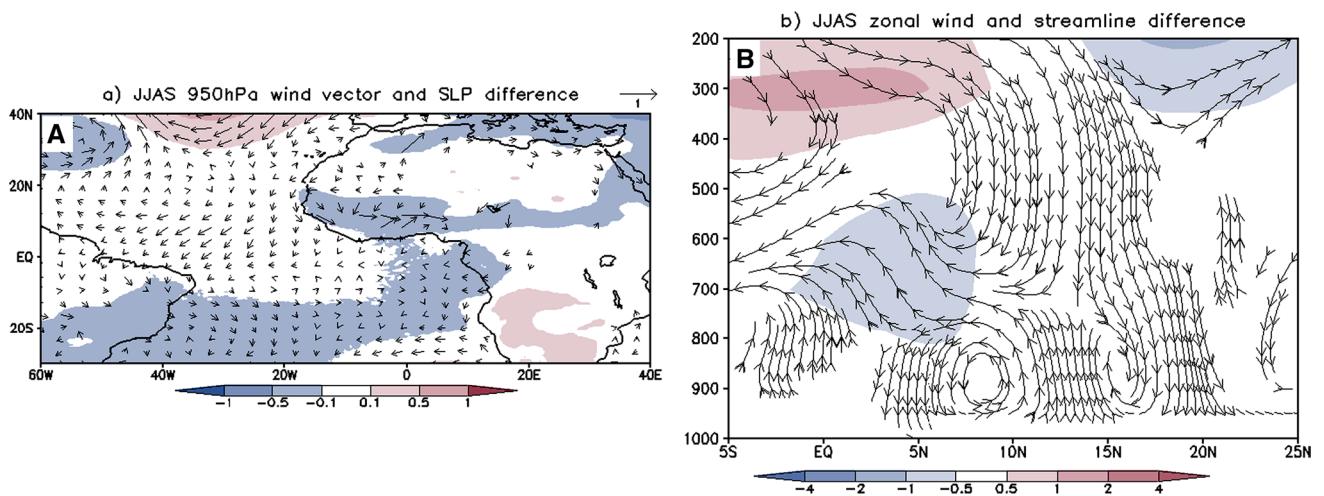


Fig. 13 a Simulated JJAS difference of 950 hPa wind vectors (m s^{-1}) and SLP (hPa) between Case LULCC and Case CTL; b latitude-height cross section of JJAS difference streamline ($v, -100 \times w$) averaged between 10°W and 10°E between Case LULCC and Case CTL

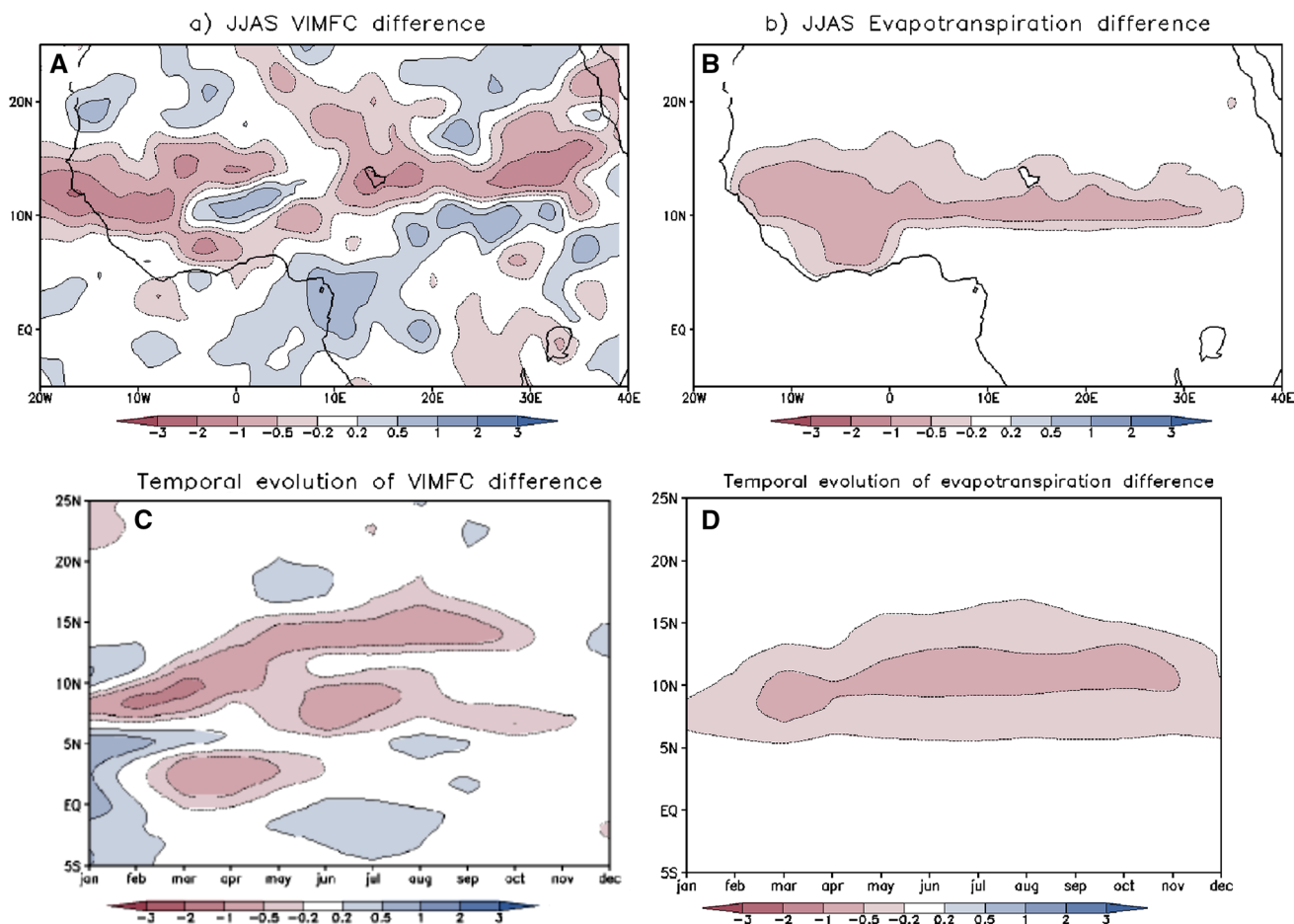


Fig. 14 **a** JJAS ensemble mean difference of vertical integrated moisture flux convergence between Case LULCC and Case CTL; **b** same as **a** but for evaporation; **c** simulated VIMFC difference tem-

poral evolution from Case LULCC–Case CTL; **d** same as **c** but for evapotranspiration. Units: mm day⁻¹

at about 65 % during the monsoon season, while in Case LULCC this ratio is about 54 %, suggesting that local ET plays a more important role in the LULCC Case (Table 4; Fig. 14). The change in ET in Case LULCC is obviously larger than the one with SST effect (Figs. 7b and 14b) but with no latitudinal temporal movement (Fig. 14d). The LULCC affects the surface energy and water balances, which then modify the large-scale circulation. Compared with the SST effect, the upward short wave radiation in Case LULCC is substantially higher due to high surface albedo in the land degradation areas. Thus, the net radiation decreases over the Reference Area (Table 5). The reduction in latent heat flux, however, dominates the change of surface temperature resulting in high surface temperature, which is similar to the SST effect over this semiarid area. In contrast with the SST effect, the changes in the Sahel mode of the surface temperature are dominant (Fig. 12c, d). There is no noticeable change in the Sahara mode in the LULCC experiment.

4 Discussions and conclusions

The present paper presents the major results of the WAMME II SST experiment with an inclusion of the comparison between impacts of SST and LULCC effects. This is the first multi-model experiment specifically designed to simultaneously evaluate the relative contribution of multiple external forcings to the Sahel drought. Furthermore, the impacts of SST anomalies in the world ocean as well as in individual oceans basins are examined. To test the SST effect of each ocean, we subtract its SST anomaly from the global ocean forcing experiment. There is a caveat in our approach in testing SST effect due to specified SST. A full assessment of these effects will require using a coupled ocean-atmosphere-land model. Moreover, the results presented in this study shows that due to the GCM resolution, some intra-seasonal variability, such as WAM jump, as well as the spatial pattern of WAM anomaly, including its heterogeneity, are not well simulated (Figs. 2, 3). Although this paper focuses on the

seasonal and decadal variations, the extent to which the deficiencies in intraseasonal simulation influences understanding of WAM seasonal variability and its link to the decadal variability, and whether the RCMs will help the WAM simulations and understanding in these aspects need further investigation.

The WAMME II experimental design also addresses the aerosol effects. Aerosols have been proposed as a factor contributing to late twentieth century drought in the Sahel via direct and indirect effects (e.g., Yoshioka et al. 2007; Kim et al. 2010; Ackerley et al. 2011; Booth et al. 2012; Zhang et al. 2013). The WAMME II has applied two aerosol data for the WAMME II aerosol experiment: GOCART data (Chin et al. 2014) and the data from the Model of Atmospheric Transport and Chemistry (Mahowald et al. 2003; Luo et al. 2003). In view of the reduced number of ensemble members in the aerosol experiment, however, we have only compared the SST and LULCC effects. Aerosol effects from individual model are presented in the papers by Gu et al. (2015) in this special issue. More multi-model experiments on the aerosol effects are necessary to allow for understand its impact.

In the LULCC experiment, the extent of the perturbation in the model's boundary conditions is based on historical data, but heterogeneity in land degradation (Xue et al. 2004a) is ignored. Furthermore, no different degrees of land degradation are considered. This experimental design, therefore, may exaggerate land effects. However, in addition to anthropogenic land use change, the natural vegetation conditions also present evidence of responding to the Sahel decadal rainfall variability. For instance, it has been shown that an increase in rainfall and the satellite derived Normalized Difference Vegetation Index (NDVI) over the northern edge of the WAM rain belt is associated with the recovery of the Sahel drought since the 1980s, when satellite data first became available (Anyamb and Tucker 2005; Herrmann et al. 2005; Jarlan et al. 2005; Govaerts and Lattanzio 2008; Hiernaux et al. 2009; Dardel et al. 2014). The NDVI-derived leaf area index (LAI) shows similar trends (Song 2013). Although we have no observed large-scale vegetation condition data before 1980, a decreasing LAI trend from the 1950s to the 1980s has been confirmed and identified by a product based on a biophysical/dynamic vegetation model (Fig. 11d, Cox et al. 2001; Song 2013; Zhang et al. 2015). Consistent with the LAI changes, another dynamic vegetation model produces interdecadal variability in surface albedo in the Sahel (Kucharski et al. 2013). Using results from a GCM, Wang et al. (2004) demonstrated that dynamic vegetation-climate interactions enhance the severity of an SST-induced drought in the 2nd half of the twentieth century. Therefore, despite we may overestimate the LULCC situation between the 1980s and the 1950s, which nevertheless is based on the best current available knowledge on this issue, the land degradation due to natural vegetation variability caused by the dry trend

from the 1950s to the 1980s is not included. We believe that the imposed land forcing in the WAMME II should provide a reasonable assessment of land effects.

Moreover, land surface processes include many biogeophysical and biogeochemical processes, such as radiative transfer at the canopy, soil moisture, turbulence at the surface layer, photosynthesis, etc. (e.g., Zeng et al. 1999; Xue et al. 2004b, 2010b; Koster et al. 2004; Ma et al. 2013). The SST effects considered in this paper include the land surface biogeophysical and soil moisture effects as presented in our discussions on surface water and energy balances (Figs. 7, 8, 9, 10). These aspects are not comprehensively investigated in the present study. More experiments tackling multiple external forcings with different experiment designs are necessary to explore these issues further.

As presented in this and other papers (e.g., Boone et al. 2015; Wang et al. 2015, in this special issue), the model results submitted to the WAMME II broadly agree on the important role played by SST anomalies and LULCC in WAM. Among the different ocean basins, the results suggest that the Indian Ocean SST anomalies have the largest impact during the precipitation anomaly development during the winter and spring, while the Pacific Ocean SST anomalies greatly contribute to the summer decadal precipitation anomalies. The consensus on the global SST effect on annual evolution of precipitation anomalies over West Africa suggests a potential predictability of WAM development linked to SST variability. More modeling and analyses studies are necessary to explore these issues further.

This study confirms the conjecture that the Sahel dry period in the 1980s resulted primarily from remote effects of SST anomalies amplified by local land surface/atmospheric interactions. The WAMME II results indicate that land surface processes are, similarly to SST, a first order contributor to the Sahel drought of the 1980s and to the West African climate system in general. The results of this study suggest that when each ocean's SST and land influences combine synergistically to favor drought, catastrophic consequences are likely to happen in certain regions. Similar phenomena have also appeared in East Asia and in the U.S. Southern Great Plains (Xue 1996; Xue et al. 2005, 2016). Since the SST and land forcing in the real world are likely smaller than specified in this study, further investigations on the effects of aerosols as well as of other external forcings, such as greenhouse gases, and of atmospheric internal variability, are necessary.

It is also shown that SST and land effects share many common characteristics in affecting the simulated Sahel precipitation decadal anomalies (e.g., Figs. 2k, 3k, , 6b, 7a, b, 11a, 12b, 13b, 14a, b). This raises the difficulty of separating these two effects locally in the real world. However, the outputs of WAMME II models nevertheless reveal some different characteristics of how these two forcings affect the development

of precipitation anomalies and affect the surface energy and water balances as well as other climate features. More comprehensive observational data and analyses with more components plus modeling study should help to identify and improve understanding of the natural and anthropogenic causes and guide models' development and improvement.

Finally, although the WAMME models are broadly consistent on the effects of the global SST and Pacific and Indian SST, large discrepancies exist regarding the effects of Atlantic SST. Similarly, the models are unable to reach agreement on the Mediterranean SST effect. Due to the scope of this paper, we mainly report the results with consensus among the WAMME II models. A large number of issues that require further studies remain open. For example, why does the Indian SST greatly affect the WAM onset? Also, the WAMME II results show that the models have different responses to the 1950s SST and the 1980s SST (not discussed in this paper). We cannot cover all these issues within the scope of this paper. We wish this paper will serve to stimulate further analysis of these simulations, as well as suggest new research on different external forcing's effect and dynamic and physical mechanisms contributing to the Sahel

decadal variability and drought. The WAMME I (Xue et al. 2010a) and this WAMME II experiment have generated large amounts of data that are available for public use. The WAMME I data are available at the NCAR database (openly available to the scientific community (http://data.eol.ucar.edu/master_list/?project=WAMME)), and WAMME II data are available at the AMMA data base (<ftp.bddamma.ipsl.polytechnique.fr>). The work presented only uses a small portion of these data sets. More studies utilizing these data sets are highly encouraged.

Acknowledgement We appreciate AMMA's support for the WAMME project, including use of the AMMA database for the WAMME II experiment. The WAMME activity and analysis are supported by U.S. NSF Grants AGS-1115506 and AGS-1419526. Each WAMME model group's efforts are supported by U.S. NSF and NASA, the European Union, and other funding agencies. A number of WAMME models simulations were conducted with the NCAR Supercomputers.

Appendix

See Figs. 15, 16 and 17.

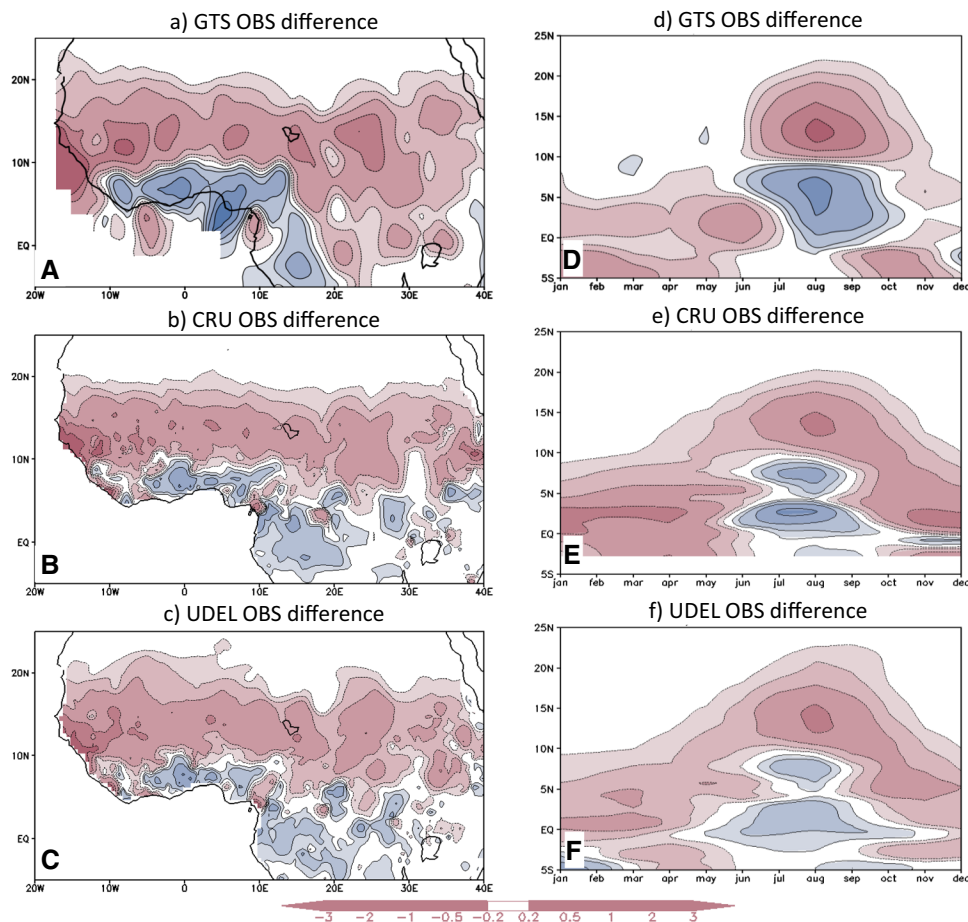


Fig. 15 Comparison of JJAS precipitation difference between the 1980s and the 1950s from different observational data sets (mm day^{-1})

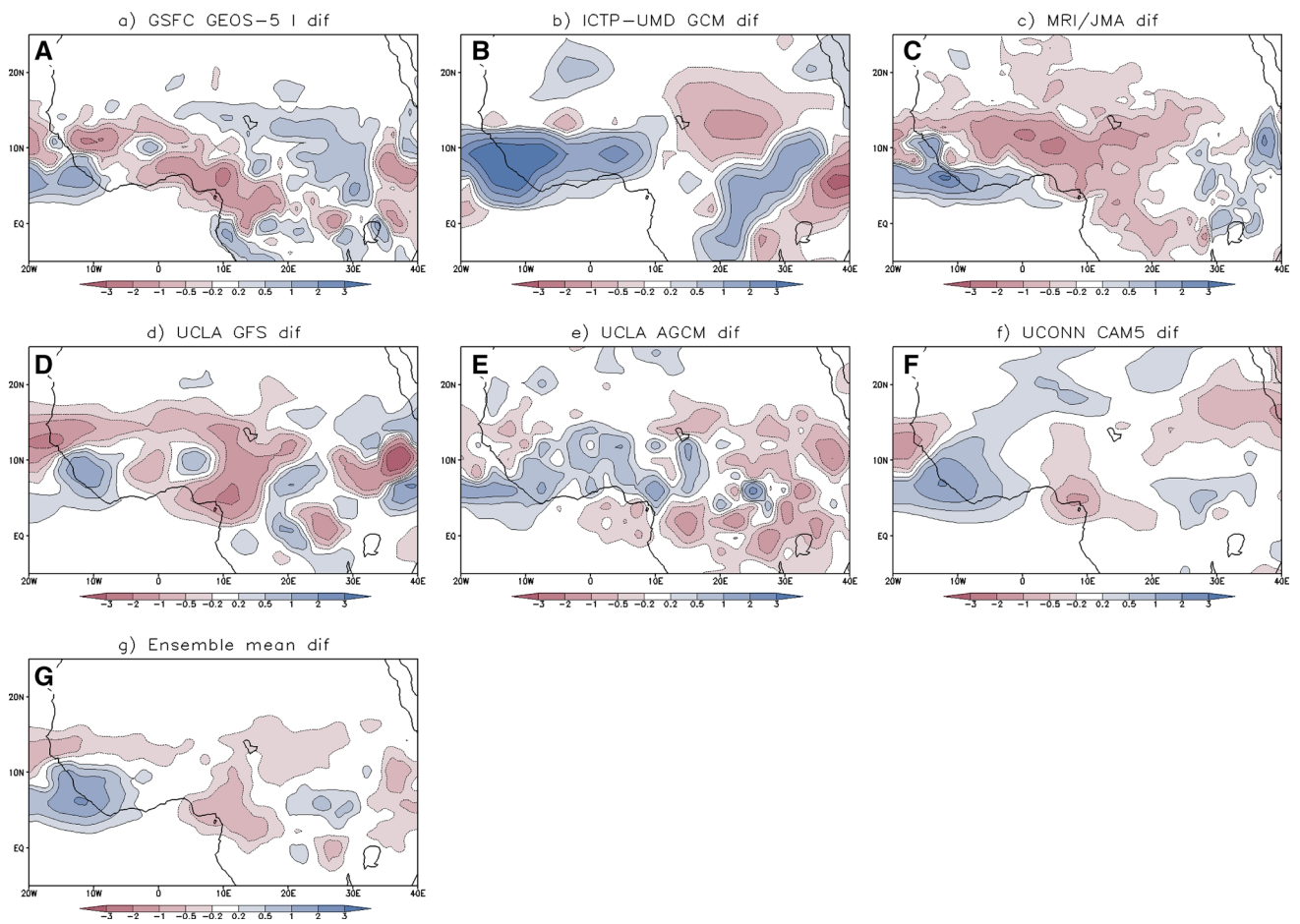


Fig. 16 JJAS anomaly precipitation (mm day⁻¹) due to Atlantic SST anomaly

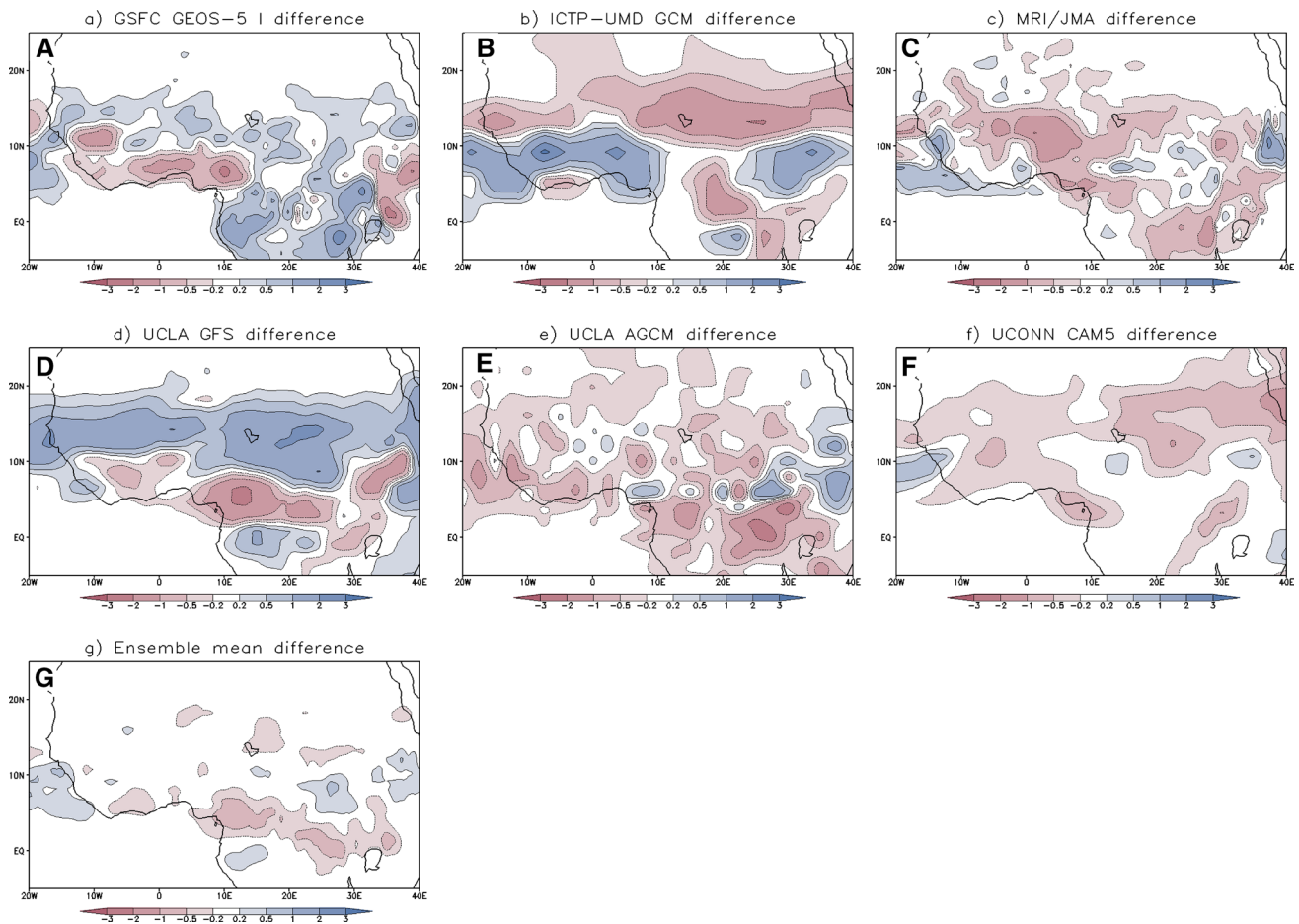


Fig. 17 JJAS anomaly precipitation (mm day^{-1}) due to Mediterranean SST anomaly

References

- Ackerley D, Booth B, Knight SHE, Highwood EJ, Frame DJ, Allen MR, Rowell DP (2011) Sensitivity of twentieth-century Sahel rainfall to sulfate aerosol and CO_2 forcing. *J Clim* 24:4999–5014
- Anyamb A, Tucker CJ (2005) Analysis of Sahelian vegetation dynamics using NOAA-AVHRR NDVI data from 1981–2003. *J Arid Environ* 63:596–614
- Bader J, Latif M (2011) The 1983 drought in the West Sahel: a case study. *Clim Dyn* 36:463–472
- Biasutti M, Held IM, Sobel AH, Giannini A (2008) SST forcings and Sahel rainfall variability in simulations of the twentieth and twenty-first centuries. *J Clim* 21:3471–3486
- Biasutti M, Sobel AH, Camargo SJ (2009) Role of the Sahara low in summertime Sahel rainfall variability in the CMIP3 models. *J Clim* 22:5755–5771
- Boone A, Pocard-Leclercq I, Xue Y, Feng J, De Rosnay P (2010) Evaluation of the WAMME model surface fluxes using results from the AMMA land-surface model intercomparison project. In: Special issue “West African Monsoon and its Modeling.” *Clim Dyn* 35:127–142. doi:10.1007/s00382-009-0653-1
- Boone A, Xue Y, De Sales F, Comer R, Hagos SM, Mahanama S, Schiro K, Song G, Wang G, Li S, Mechoso CR (2015) The regional impact of land-use land-cover change (LULCC) over West Africa from an ensemble of global climate models under the auspices of the WAMME2 project. *Clim Dyn*. In this Special Issue, under revision
- Booth BB, Dunstone NJ, Halloran PR, Andrews T, Bellouin N (2012) Aerosols implicated as a prime driver of twentieth-century North Atlantic climate variability. *Nature* 484:228–232
- Caminade C, Terray L (2010) Twentieth century Sahel rainfall variability as simulated by the ARPEGE AGCM, and future changes. *Clim Dyn* 35(1):75–94. doi:10.1007/s00382-009-0545-4
- Carton JA, Cao X, Giese BS, Da Silva AM (1996) Decadal and inter-annual SST variability in the tropical Atlantic. *Ocean J Phys Oceanogr* 26:1165–1175
- Chen M, Xie P, Janowiak JE, Arkin PA (2002) Global land precipitation: a 50-yr monthly analysis based on gauge observations. *J Hydrom* 3:249–266
- Chin M, Rood RB, Lin SJ, Muller JF, Thompson AM (2000) Atmospheric sulfur cycle simulated in the global model GOCART: model description and global properties. *J Geophys Res* 105:24671–24687
- Chin M, Ginoux P, Kinne S, Torres O, Holben BN, Duncan BN, Martin RV, Logan JA, Higurashi A, Nakajima T (2002) Tropospheric aerosol optical thickness from the GOCART model and comparisons with satellite and sun photometer measurements. *J Atmos Sci* 59:461–483
- Chin M, Diehl T, Tan Q, Prospero JM, Kahn RA, Remer LA, Yu H, Sayer AM, Bian H, Geogdzhayev IV, Holben BN, Howell SG,

- Huebert BJ, Hsu NC, Kim D, Kucsera TL, Levy RC, Mishchenko MI, Pan X, Quinn PK, Schuster GL, Streets DG, Strode SA, Torres O, Zhao XP (2014) Multi-decadal variations of atmospheric aerosols from 1980 to 2009: a perspective from observations and a global model. *Atmos Chem Phys* 14:3657–3690
- Cook KH (1999) Generation of the African Easterly Jet and its role in determining West African precipitation. *J Clim* 12:1165–1184
- Cox PM, Betts RA, Jones CD, Spall SA, Totterdell IJ (2001) Modeling vegetation and the carbon cycle as interactive elements of the climate system. In: Pearce R (ed) *Meteorology at the Millennium*
- Dardel C, Kergoat L, Hiernaux P, Mougouin E, Grippa M, Tucker CJ (2014) Re-greening Sahel: 30 years of remote sensing data and field observations (Mali, Niger). *Remote Sens Environ* 140:350–364
- De Sales, F. Y. Xue, G. Okin (2015) Impact of burned area on the northern African seasonal climate from the perspective of regional modelling. *Clim Dyn*, this special issue. doi:10.1007/s00382-015-2522-4. <http://link.springer.com/article>
- Dieppois B, Diedhiou A, Durand A, Fournier M, Massei N, Sebagn D, Xue Y, Fontaine B (2013) Quasi-decadal signals of Sahel rainfall and West African monsoon since the mid-twentieth century. *J Geophys Res Atmos*. doi:10.1002/2013JD019681
- Druryan LM, Feng J, Cook KH, Xue Y, Fulakeza M, Hagos SM, Konare A, Moufouma-Okia W, Rowell DP, Vizu E, Ibrah SS (2010) The WAMME regional model intercomparison study. In: Special issue “West African Monsoon and its Modeling”, *Clim Dyn* 35:175–192. doi:10.1007/s00382-009-0676-7
- Evan AT, Flament C, Lavaysse C, Kosha C, Saci A (2015) Water vapor-forced greenhouse warming over the Sahara Desert and the recent recovery from the Sahelian drought. *J Clim* 28:108–123
- Folland C, Owen J, Ward MN, Colman A (1991) Prediction of seasonal rainfall in the Sahel region using empirical and dynamical methods. *J Forecast* 1:21–56
- Fontaine B, Janicot S (1996) Sea surface temperature fields associated with West African rainfall anomaly types. *J Clim* 9:2935–2940
- Fontaine B, Roucou P, Sivarajan S, Gervois S, Chauvin F, Rodriguez de Fonseca MB, Ruti P, Janicot S (2010) Impacts of Warm and Cold situations in the Mediterranean Basins on the West African monsoon: observed connection patterns (1979–2006) and climate simulations. *Clim Dyn* 35:95–114. doi:10.1007/s00382-009-0599-3
- Giannini A, Saravanan R, Chang P (2003) Oceanic forcing of Sahel rainfall on interannual to interdecadal time scales. *Science* 302:1027–1030
- Goldewijk KK (2001) Estimating global land use change over the past 300 years: the HYDE database. *Global Biogeochem Cycles* 15:417–433
- Govaerts YM, Lattanzio A (2008) Estimation of surface albedo increase during the eighties Sahel drought from Meteosat observations. *Global Planet Change* 64:139–145. doi:10.1016/j.gloplacha.2008.04.004
- Gu Y, Xue Y, De Sales F, Liou KN (2015) A gcm investigation of dust aerosol impact on regional climate of North Africa and South-East Asia. *Clim Dyn*, in this Special Issue. <http://link.springer.com/article/10.1007/s00382-015-2706-y>
- Hagos SM, Cook KH (2008) Ocean warming and late-twentieth-century Sahel drought and recovery. *J Clim* 21:3797–3814
- Hagos S, Leung LR, Xue Y, Boone A, De Sales F, Neupane N, Huang M, Yoon JH (2014) Assessment of uncertainties in the response of the African Monsoon precipitation to land use change simulated by a regional model. *Clim Dyn* 43:2765–2775. doi:10.1007/s00382-014-2092-x
- Hastenrath S, Polzin D (2010) Long-term variations of circulation in the tropical Atlantic sector and Sahel rainfall. *Int J Climatol* 31(5):649–655. doi:10.1002/joc.2116
- Herrmann SM, Anyambab A, Tucker CJ (2005) Recent trends in vegetation dynamics in the African Sahel and their relationship to climate. *Glob Environ Change* 15:394–404
- Hiernaux P, Diarra L, Trichon V, Mougouin E, Soumaguel N, Baup F (2009) Woody plant population dynamics in response to climate changes from 1984 to 2006 in Sahel (Gourma, Mali). *J Hydrol* 375(1–2):103–113
- Hoerling M, Hurrell J, Eischeid J, Phillips A (2006) Detection and attribution of twentieth-century northern and southern African rainfall change. *J Clim* 19:3989–4008
- Hoerling M, Eischeid J, Perlwitz J (2010) Regional precipitation trends: distinguishing natural variability from anthropogenic forcing. *J Clim* 23:2131–2145
- Horel JD, Kousky VE, Kagano MT (1986) Atmospheric conditions in the Atlantic sector during 1983 and 1984. *Nature* 322(6076):248–251
- Hourdin F, Musat I, Grandpeix J-Y, Polcher J, Guichard F, Favot F, Marquet P, Boone A, Lafore J-P, Redelsperger J-L, Ruti PM, Dell’Aquila A, Filiberti MA, Pham M, Doval TL, Traore AK, Gallée H (2010) AMMA-model intercomparison project. *Bull Am Meteorol Soc* 91:95–104. doi:10.1175/2009BAMS2791.1 <http://journals.ametsoc.org/doi/pdf/10.1175/JCLI-D-12-00819.1>
- Hurtt GC, Frolking S, Fearon MG, Moore B, Shevliakova E, Malyshv S, Pacala SW, Houghton RA (2006) The underpinnings of land-use history: three centuries of global gridded land-use transitions, wood harvest activity, and resulting secondary lands. *Global Change Biol* 12:1208–1229
- Hurtt GC, Chini LP, Frolking S, Betts R, Feddema J, Fischer G, Fisk JP, Hibbard K, Houghton RA, Janetos A, Jones C, Kindermann G, Kinoshita T, Klein Goldewijk K, Riahi K, Shevliakova E, Smith S, Stehfest E, Thomson A, Thornton P, van Vuuren DP, Wang Y (2011) Harmonization of land-use scenarios for the period 1500–2100: 600 years of global gridded annual land-use transitions, wood harvest, and resulting secondary lands. *Clim Change* 109:117–161. doi:10.1007/s10584-011-0153-2
- Janicot S, Harzallah A, Fontaine B, Moron V (1998) West African Monsoon dynamics and eastern equatorial Atlantic and Pacific SST anomalies (1970–88). *J Clim* 11:184–1882
- Jarlan L, Tourre YM, Mougouin E, Philippon N, Masega P (2005) Dominant patterns of AVHRR NDVI interannual variability over the Sahel and linkages with key climate signals (1982–2003). *Geophys Res Lett* 32:L04701. doi:10.1029/2004GL021841
- Joly M, Voltaire A, Douville H, Terray P, Royer JF (2007) African monsoon teleconnections with tropical SSTs: validation and evolution in a set of IPCC4 simulations. *Clim Dyn*. doi:10.1007/s00382-006-0215-8
- Kanamitsu M, Ebisuzaki W, Woollen J, Yang SK, Hnilo JJ, Fiorino M, Potter GL (2002a) NCEP–DOE AMIP-II Reanalysis (R-2). *Bull Am Meteorol Soc* 83:1631–1643
- Kanamitsu M, Kumar A, Juang HMH, Wang W, Yang F, Schemm J, Hong SY, Peng P, Chen W, Ji M (2002b) NCEP dynamical seasonal forecast system 2000. *Bull Am Meteorol Soc* 83:1019–1037
- Kim KM, Lai WKM, Sud YC, Walker GK (2010) Influence of aerosol-radiative forcings on the diurnal and seasonal cycles of rainfall over West Africa and Eastern Atlantic Ocean using GCM simulations. *Clim Dyn* 35:115–126
- Knight JR, Folland CK, Scaife AA (2006) Climate impacts of the Atlantic multidecadal oscillation. *Geophys Res Lett*. doi:10.1029/2006GL026242

- Koster R, Suarez MJ, Ducharne A, Kumar P, Stieglitz M (2000) A catchment based approach to modeling land surface processes in a GCM—Part 1: model structure. *J Geophys Res* 105:809–822
- Koster RD, Dirmeyer PA, Guo Z, Bonan G, Chan E, Cox P, Gordon CT, Kanae S, Kowalczyk E, Lawrence D, Liu P, Lu CH, Malyshev S, McAvaney B, Mitchell K, Mocko D, Oki T, Oleson K, Pitman A, Sud YC, Taylor CM, Verseghy D, Vasic R, Xue Y, Yamada T (2004) Regions of strong coupling between soil moisture and precipitation. *Science* 305:1138–1140
- Kucharski F, Zeng N, Kalnay E (2013) A further assessment of vegetation feedback on decadal Sahel rainfall variability. *Clim Dyn* 40:1453–1466. doi:10.1007/s00382-012-1397-x
- Lamb PJ (1978) Large-scale tropical Atlantic surface circulation patterns associated with sub-Saharan weather anomalies. *Tellus* 30:240–251
- Lau KM, Shen SSP, Kim KM, Wang H (2006) A multimodel study of the twentieth-century simulations of Sahel drought from the 1970s to 1990s. *J Geophys Res* 111:D07111. doi:10.1029/2005JD006281
- Lavaysse C, Flamant C, Janicot S, Parker DJ, Lafore JP, Sultan B, Pelon J (2009) Seasonal evolution of the West African heat low: a climatological perspective. *Clim Dyn* 33:313–330
- Le Barbé L, Lebel T, Tapsoa D (2002) Rainfall variability in West Africa during the years 1950–90. *J Clim* 15:187–202
- Lebel T, Ali A (2009) Recent trends in the Central and Western Sahel rainfall regime 1990–2007. *J Hydrol* 375:52–64
- Losada T, Rodríguez-Fonseca S, Janicot S, Gervois S, Chauvin F, Ruti P (2010) A multi-model approach to the Atlantic equatorial mode: impact on the West African monsoon. *Clim Dyn*. 35:29–43. doi:10.1007/s00382-009-0625-5
- Lu J, Delworth TL (2005) Oceanic forcing of the late 20th century Sahel drought. *Geophys Res Lett* 32:L22706. doi:10.1029/2005GL023316
- Luo C, Mahowald N, Del Corral J (2003) Sensitivity study of meteorological parameters on mineral aerosol mobilization, transport and distribution. *J Geophys Res* 108(D15):4447. doi:10.1029/2003JD0003483
- Ma HY, Mechoso CR, Xue Y, Xiao H, Neelin JD, Ji X (2013) On the connection between continental-scale land surface processes and the tropical climate in a coupled ocean–atmosphere–land system. *J Clim* 26:9006–9025
- Mahowald N, Luo C, Del Corral J, Zender C (2003) Interannual variability in atmospheric mineral aerosols from a 22-year model simulation and observational data. *JGR*. doi:10.1029/2002JD002821
- Martin ER, Thorncroft C (2014) Sahel rainfall in multimodel CMIP5 decadal hindcasts. *Geophys Res Lett* 41:2169–2175. doi:10.1002/2014GL059338
- Mathon V, Laurent H, Lebel T (2002) Mesoscale convective system rainfall in the Sahel. *J Appl Meteorol* 41:1081–1092
- Mechoso CR, Yu JY, Arakawa A (2000) A coupled GCM pilgrimage: From climate catastrophe to ENSO simulations. General circulation model development: Past, present and future. In: Proceedings of a symposium in honor of professor Akio Arakawa, D. A. Randall. Academic Press, USA, pp 539–575
- Mizuta R, Oouchi K, Yoshimura H, Noda A, Katayama K, Yukimoto S, Hosaka M, Kusunoki S, Kawai H, Nakagawa M (2006) 20-km-mech global climate simulations using JMA-GSM model, Mean Climate States. *J Meteorol Soc Japan* 84:165–185
- Mohino E, Janicot S, Bader J (2011) Sahel rainfall and decadal to multi-decadal sea surface e temperature variability. *Clim Dyn* 37:419–440. doi:10.1007/s00382-010-0867-2
- Molod A, Takacs LL, Suarez MJ, Bacmeister JT, Song IS, Eichmann A (2012) The GEOS-5 atmospheric general circulation model: mean climate and development from MERRA to Fortuna. NASA Tech. Memo. 104606, vol 28, technical report series on global modeling and data assimilation, edited by: Suarez, M. J., 117 pp
- Neale RB, Chen CC, Gettelman A et al (2012) Description of the NCAR Community Atmosphere Model (CAM 5.0), Technical Note NCAR/TN-486 + STR. National Center for Atmospheric Research, Boulder, p 274
- New M, Hulme M, Jones PD (1999) Representing twentieth century space-time climate variability. Part 1: development of a 1961–90 mean monthly terrestrial climatology. *J Clim* 12:829–856
- Nicholson SE, Dezfuli AK, Klotter D (2012) A two-century precipitation dataset for the continent of Africa. *Bull Am Meteorol Soc* 93:1219–1231. doi:10.1175/BAMS-D-11-00212.1
- Pope VD, Gallani ML, Rowntree PR, Stratton RA (2000) The impact of new physical parametrizations in the Hadley Centre climate model: HadAM3. *Clim Dyn* 16:123–146
- Ramankutty N, Foley JA (1999) Estimating historical changes in global land cover: croplands from 1700 to 1992. *Global Biogeochem Cycles* 13:997–1027
- Rayner NA, Parker DE, Horton EB, Folland CK, Alexander LV, Rowell DP, Kent EC, Kaplan A (2003) Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J Geophys Res* 108(D14):4407. doi:10.1029/2002JD002670
- Redelsperger JL, Thorncroft CD, Diedhiou A, Lebel T, Parker DJ, Polcher J (2007) American monsoon multidisciplinary analysis. *Bull Am Meteorol Soc* 88:1740–1746
- Rodriguez-Fonseca B, Mohino E, Mechoso CR, Caminade C, Biasutti M, Gaetani M, García-Serrano J, Vizy EK, Cook K, Xue Y, Polo I, Losada T, Druyen L, Fontaine B, Bader J, Doblus-Reyes FJ, Goddard L, Janicot S, Arribas A, Lau W, Colman A, Rowell DP, Kucharski F, Voltaire A (2015) Variability and predictability of West African droughts. *J Clim* 28:4034–4060
- Roehrig R, Bouniol D, Guichard F, Hourdin F, Redelsperger JL (2013) The present and future of the West African monsoon: a process-oriented assessment of CMIP5 simulations along the AMMA transect. *J Clim* 26:6471–6505. doi:10.1175/JCLI-D-12-00505.1
- Rowell DO (2001) Teleconnections between the Tropical Pacific and the Sahel. *Q J R Meteorol Soc* 127:1683–1706
- Rowell DO (2003) The impact of Mediterranean SSTs on the Sahelian rainfall season. *J Clim* 16:849–8610
- Scaife A, Kucharski F, Folland C, Kinter J III, Fereday D, Fischer A, Grainger S, Jin E, Kang I, Knight J, Kusunoki S, Lau N, Nath M, Nakaegawa T, Pegion P, Schubert S, Sporyshev P, Syktus J, Yoon J, Zeng N, Zhou T (2009) The CLIVAR C20C project: selected 20th century climate events. *Clim Dyn* 33:603–614. doi:10.1007/s00382-008-0451-1
- Schrage JM, Fink AH (2012) Nocturnal continental low-level stratus over tropical West Africa: observations and possible mechanisms controlling its onset. *Mon Weather Rev*. doi:10.1175/MWR-D-11-00172.1
- Song G (2013) Global and Sahel regional biophysical processes, vegetation dynamics, and climate interactions. Ph.D. Dissertation. University of California, Los Angeles, pp 183
- Stein THM, Parker DJ, Delanoë J, Dixon NS, Hogan RJ, Knippertz P, Maiment RI, Marsham JH (2011) The vertical cloud structure of the West African monsoon: a 4 year climatology using CloudSat and CALIPSO. *J Geophys Res* 116:D22205. doi:10.1029/2011JD016029
- Sultan B, Janicot S (2000) The West African monsoon dynamics. Part II: the “preonset” and “onset” of the summer monsoon. *J Clim* 16:3407–3427
- Taylor KE (2001) Summarizing multiple aspects of model performance in a single diagram. *J Geophys Res* 106:7183–7192
- Ting M, Kushnir Y, Seager R, Cuihua L (2011) Robust features of Atlantic multi-decadal variability and its climate impacts. *Geophys Res Lett* 38:L17705. doi:10.1029/2011GL048712

- Tippett MK, Giannini A (2006) Potentially predictable components of African summer rainfall in an SST-forced GCM simulation. *J Clim* 19:3133–3144
- Wagner RG, Da Silva AM (1994) Surface conditions associated with anomalous rainfall in the Guinea coastal region. *Int J Climatol* 14:179–199
- Wallace JM, Smith C, Bretherton CS (1992) Singular value decomposition of wintertime sea surface temperature and 500-mb height anomalies. *J Clim* 5:561–576
- Wang GL, Eltahir EAB, Foley JA, Pollard D, Levis S (2004) Decadal variability of rainfall in the Sahel: results from the coupled GENESIS-IBIS atmosphere–biosphere model. *Clim Dyn* 22(6–7):625–637. doi:10.1007/s00382-004-0411-3
- Wang GL, Yu M, Xue Y (2015) Modeling the potential contribution of land use changes to the 20th century Sahel drought using a regional climate model: impact of lateral boundary conditions. *Clim Dyn*. doi:10.1007/s00382-015-2812-x
- Ward MN (1998) Diagnosis and short-lead time prediction of summer rainfall in tropical North Africa at interannual and multidecadal timescales. *J Clim* 11:3167–3191
- Weng HY, Lau KM, Xue YK (1999) Long term variations of summer rainfall over China and its possible link to global sea-surface temperature variability. *J Meteorol Soc Jpn* 77:845–857
- Willmott CJ, Matsuura K (1995) Smart interpolation of annually averaged air temperature in the United States. *J Appl Meteorol* 34:2577–2586
- Xue Y (1996) The Impact of desertification in the Mongolian and the Inner Mongolian grassland on the regional climate. *J Clim* 9:2173–2189
- Xue Y, Shukla J (1993) The influence of land surface properties on Sahel climate. Part I: desertification. *J Clim* 6:2232–2245
- Xue Y, Hutjes RWA, Harding RJ, Claussen M, Prince S, Lambin EF, Allen SJ, Dirmeyer P, Oki T (2004a) The Sahelian climate (chapter A5). In: Kabat P, Claussen M, Dirmeyer PA et al (eds) *Vegetation, water, humans and the climate*. Springer, Berlin, pp 59–77
- Xue Y, Juang HM, Li W, Prince S, DeFries R, Jiao Y, Vasic R (2004b) Role of land surface processes in monsoon development: East Asia and West Africa. *J Geophys Res* 109:D03105. doi:10.1029/2003JD003556
- Xue Y, Sun S, Lau JM, Ji J, Pocard I, Kang HS, Zhang R, Wu G, Zhang J, Schaake J, Jiao Y (2005) Multiscale variability of the river runoff system in China and its link to precipitation and sea surface temperature. *J Hydrometeorol* 6:550–570
- Xue Y, De Sales F, Lau KMW, Boone A, Feng J, Dirmeyer P, Guo Z, Kim KM, Kitoh A, Kumar V, Pocard-Leclercq I, Mahowald N, Moufouma-Okia W, Pegion P, Rowell D, Schubert SD, Sealy A, Thiaw WM, Vintzileos A, Williams S, Wu MLC (2010a) Inter-comparison and analyses of the climatology of the West African Monsoon in the West African Monsoon Modeling and Evaluation Project (WAMME) first model intercomparison experiment. In: Special issue “West African Monsoon and its Modeling.” *Clim Dyn* 35:3–27. doi:10.1007/s00382-010-0778-2
- Xue Y, De Sales F, Vasic R, Mechoso CR, Prince SD, Arakawa A (2010b) Global and temporal characteristics of seasonal climate/vegetation biophysical process (VBP) interactions. *J Clim* 23:1411–1433
- Xue Y, Boone A, Taylor CM (2012) Review of recent developments and the future prospective in West African Atmosphere/Land Interaction Studies in the Special issue, *Advances in climate processes, feedbacks, variability, and change for the West African Climate System*. *Int J Geophys*. doi:10.1155/2012/748921
- Xue Y, Oaida CM, Diallo I, Neelin JD, Li S, De Sales F, Gu Y, Robinson D, Vasic R, Yi L (2016) Spring land temperature anomalies in northwestern US and the summer drought over Southern Plains and adjacent areas. *Environ Res Lett* 11:044018. doi:10.1088/1748-9326/11/4/044018
- Yang S, Lau KM, Kim KM (2002) Variations of the East Asian jet stream and Asian–Pacific–American winter climate anomalies. *J Clim* 15:306–325
- Yoshioka M, Mahowald NM, Conley AJ, Collins WD, Fillmore DW, Zender CS, Coleman DB (2007) Impact of desert dust radiative forcing on Sahel precipitation: relative importance of dust compared to sea surface temperature variations, vegetation changes and greenhouse gas warming. *J Clim* 20:1445–1467
- Zeng N, Neelin JD, Lau KM, Tucker CJ (1999) Enhancement of interdecadal climate variability in the Sahel by vegetation interaction. *Science* 286:1537–1540
- Zhang R, Delworth TL (2006) Impact of Atlantic multidecadal oscillations on India/Sahel rainfall and Atlantic hurricanes. *Geophys Res Lett* 33:L17712. doi:10.1029/2006GL026267
- Zhang C, Nolan DS, Thorncroft CD (2008) Shallow meridional circulations in the tropics represented by global reanalyses. *J Clim* 21:3453–3470
- Zhang R, Delworth TL, Sutton R, Hodson DLR, Dixon KW, Held IM, Kushnir Y, Marshall J, Ming Y, Msadek R, Robson J, Rosati AJ, Ting M, Vecchi GA (2013) Have aerosols caused the observed Atlantic multidecadal variability? *J Atmos Sci* 70:1135–1144. doi:10.1175/JAS-D-12-0331
- Zhang Z, Xue Y, MacDonald G, Cox PM, Collatz GJ (2015) Investigation of North American vegetation variability under recent climate—a study using the SSiB4/TRIFFID biophysical/dynamic vegetation model. *J Geophys Res* 120:1300–1321. doi:10.1002/2014JD021963