

Monsoon Regimes and Processes in CCSM4. Part I: The Asian–Australian Monsoon

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ABSTRACT

The simulation characteristics of the Asian-Australian monsoon are documented for the Community Climate System Model, version 4 (CCSM4). This is the first part of a two part series examining monsoon regimes in the global tropics in the CCSM4. Comparisons are made to an Atmospheric Model Intercomparison Project (AMIP) simulation of the atmospheric component in CCSM4 [Community Atmosphere Model, version 4, (CAM4)] to deduce differences in the monsoon simulations run with observed sea surface temperatures (SSTs) and with ocean-atmosphere coupling. These simulations are also compared to a previous version of the model (CCSM3) to evaluate progress. In general, monsoon rainfall is too heavy in the uncoupled AMIP run with CAM4, and monsoon rainfall amounts are generally better simulated with ocean coupling in CCSM4. Most aspects of the Asian-Australian monsoon simulations are improved in CCSM4 compared to CCSM3. There is a reduction of the systematic error of rainfall over the tropical Indian Ocean for the South Asian monsoon, and wellsimulated connections between SSTs in the Bay of Bengal and regional South Asian monsoon precipitation. The pattern of rainfall in the Australian monsoon is closer to observations in part because of contributions from the improvements of the Indonesian Throughflow and diapycnal diffusion in CCSM4. Intraseasonal variability of the Asian-Australian monsoon is much improved in CCSM4 compared to CCSM3 both in terms of eastward and northward propagation characteristics, though it is still somewhat weaker than observed. An improved simulation of El Niño in CCSM4 contributes to more realistic connections between the Asian-Australian monsoon and El Niño-Southern Oscillation (ENSO), though there is considerable decadal and century time scale variability of the strength of the monsoon-ENSO connection.

1. Introduction

Monsoons are central to the seasonal development of tropical rainfall maxima that encompass local, regional, and large-scale atmospheric and ocean circulation patterns. Their year to year variability, and connections between the monsoon regimes themselves as well as to extratropical weather and climate, make monsoon simulations an important part of any climate model.

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This is the first of a series of two papers that document monsoon regimes and associated processes for the Community Climate System Model, version 4 (CCSM4). Here, in Part I the CCSM4 simulation of the Asian-Australian monsoon is examined. General characteristics of the South Asian and Australian monsoons will be discussed, but details of some of the regional monsoon regimes, such as the East Asian Monsoon, will not be specifically addressed. Part II (Cook et al. 2012) documents the CCSM4 simulations of the West African, South American, and North American monsoons. In both papers the fully coupled CCSM4 simulations will be compared to the Atmospheric Model Intercomparison Project (AMIP)-type Community Atmosphere Model, version 4 (CAM4) atmosphere-only runs to show how coupling changes the monsoon simulations.

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Additionally, comparisons will be made where appropriate to the previous generation of this model (CCSM3) character in CCSM

to document any changes or improvements to the monsoon simulations. The monsoons in CCSM3 were previously described by Meehl et al. (2006) and can also be compared to monsoon simulations in a previous version of the model (Meehl and Arblaster 1998).

The purpose of this paper, and why it is included in a CCSM4 Special Collection and not as a regular *Journal* of *Climate* paper, is to provide a basic overview and description of the CCSM4 monsoon simulation characteristics. Thus, it is not a typical *Journal of Climate* paper with analyses that lead to insight into processes and mechanisms. The CCSM4 Special Collection papers are intended to provide the background and description that can then be used as a starting point for more cuttingedge science results papers.

The paper will begin in section 2 with a description of the CCSM4 and the experiments analyzed in this paper. Section 3 will include a description of the South Asian monsoon simulation in CCSM4, Indian Ocean processes focusing in particular on the Bay of Bengal (BoB) in section 4, followed in section 5 with documentation of intraseasonal variability in the Asian-Australian monsoon. Section 6 will address the Australian monsoon, with the Indonesian Throughflow in section 7, and teleconnections involving the Asian-Australian monsoon in section 8. Conclusions will follow in section 9. An extensive review of monsoon simulations in other models compared to CCSM4 is beyond the scope of this paper but is being addressed in Chapters 9 and 14 of the upcoming Intergovernmental Panel on Climate Change (IPCC) Assessment Report 5 (AR5). Previous studies that have compared monsoon simulations in models include Sperber and Palmer (1996) for models run with observed SSTs and Annamalai et al. (2007) for the models that were part of the Coupled Model Intercomparison Project phase 3 (CMIP3). Additionally, the role of land use change is an interesting one for the CCSM4 monsoon simulation since this model includes time-evolving land use change, but this is also beyond the scope of the present paper. It has been addressed in a preliminary way in another of the CCSM4 Special Collection papers by Lawrence et al. (2012) and will be the topic of a future study.

2. Model and observed data descriptions

The standard CCSM3 (e.g., Collins et al. 2006) will be compared to the new CCSM4 (Gent et al. 2011). The CCSM3 had a T85 atmospheric model with 26 levels in the vertical and was coupled to land and sea ice components as well as a nominal 1°-resolution ocean model going down to about ¹/4° in the equatorial tropics. As noted above, characteristics of the Asia–Australia monsoon simulations in CCSM3 were described by Meehl et al. (2006).

CCSM4 includes a finite-volume 1° version of the atmospheric model CAM4, with improved components of ocean, land, and sea ice compared to CCSM3 (Gent et al. 2011). Grid points in the atmosphere are spaced roughly every 1° latitude and longitude, and there are 26 levels in the vertical. The ocean is a version of the Parallel Ocean Program (POP) with a nominal latitude– longitude resolution of 1° (down to ¹/4° in the equatorial tropics) and 60 levels in the vertical. No flux adjustments are used in either CCSM3 or CCSM4. Experiments analyzed will include twentieth-century simulations with a combination of anthropogenic and natural forcings and a multicentury preindustrial control run (Gent et al. 2011). AMIP simulations with CAM4 were run with observed monthly mean SSTs from 1979 to 2005.

For validating the ENSO–monsoon association we use the observed all-India rainfall (AIR) index of Parthasarathy et al. (1994) and Australian land-based monsoon indices from the Australian Water Availability Project (AWAP) rainfall product (Jones et al. 2009). The AIR is constructed for the monsoon season (June– September) based on 306 quality-controlled stations spread over the whole of the Indian subcontinent for the period 1877 to 2000. The observed SST data are from the Hadley Centre Sea Ice and SST dataset (HadISST; Rayner et al. 2003). The Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) dataset with monthly mean values for 1979–2008 is documented by Xie and Arkin (1996).

Additional observed SSTs are used from the National Oceanic and Atmospheric Administration (NOAA) extended reconstructed sea surface temperature (ERSST) v3 dataset (Smith et al. 2008; Xue et al. 2003). The 200-hPa velocity potential results are computed from the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) data (Uppala et al. 2005). The Tropical Rainfall Measuring Mission (TRMM) precipitation data are the 3B42V6 version, and the climatology is calculated from daily values for 1998–2009 (Huffman et al. 2007). The National Centers for Environmental Prediction (NCEP2) climatology was calculated from monthly mean values for 1979-2005. The ERA-Interim analyses are described by Simmons et al. (2006), and the climatology is calculated from daily values for 1990-2005. A longer time series of daily winds at 200 and 850 mb from NCEP-National Center for Atmospheric Research (NCAR) reanalyses (Kalnay et al. 1996) and daily precipitation data from the Global Precipitation Climatology Project (GPCP) (Huffman et al. 2001) are used for the intraseasonal variability analyses.

3. The South Asian monsoon

Figure 1 shows south Asian monsoon summer season average [June–September (JJAS)] precipitation patterns and low-level (850 hPa) vector winds for the CCSM4 compared to the AMIP run with the atmospheric component of CCSM4, called CAM4. The CCSM4 results are an ensemble average of five twentieth-century simulations taken for the period 1979-2005. As with CCSM3 (Meehl et al. 2006, Fig. 1), there is excessive rainfall over the South Asian monsoon region in the AMIP run compared to observations (particularly over the Arabian Sea and East Africa), while the coupled CCSM4 produces rainfall amounts and patterns closer to the observations. However, as was also seen in CCSM3, the rainfall is shifted somewhat too far west in the western Indian Ocean and Arabian Sea. The maximum values in the eastern Arabian Sea and Bay of Bengal are closer to observations, an improvement over both CCSM3 and the CAM4 AMIP run. More details of CCSM4 monsoon precipitation in the Bay of Bengal region will be discussed in section 4, and connections between the Asian-Australian monsoon and ENSO will be shown in section 8.

In CCSM4 there is a general improvement in the location and magnitude of orographic monsoon rainfall over the southern extent of the Tibetan Plateau and over the Western Ghats compared to CCSM3, though CCSM4 simulates greater than observed rainfall in those regions. There is some evidence of the rain shadow of the Western Ghats over southeastern India in both the CAM4 and CCSM4 that is improved over CCSM3, though there is still more rainfall than observed in that region in CCSM4. The improvement in the location of orographic monsoon rainfall is likely associated with the finite-volume dynamical core in CCSM4 compared to the spectral version of CCSM3, and the higher horizontal resolution (about 1° in the atmosphere in CCSM4) compared to the roughly 1.4° atmospheric resolution in the T85 CCSM3. Additionally, though the rainfall over the western Indian Ocean is still a bit too far west in CCSM4, it is improved compared to CCSM3 as noted above. The southern ITCZ near 5°S is much better represented in CCSM4 compared to CCSM3, where the latter had only a small bit of that feature simulated near Sumatra.

Regarding the time scales of the South Asian monsoon, wavelet spectra shown by Meehl and Arblaster (2011) indicate an improved preferred interannual time scale of the South Asian monsoon near 4 years as in observations, with a somewhat reduced biennial peak associated with a similar change in period of ENSO (Deser et al. 2012). The CCSM3, meanwhile, had a much too dominant biennial peak of interannual monsoon variability as well as ENSO and very little power in the observed range of 4 years (Meehl and Arblaster 2011).

Some of these improvements can be traced to the change in base state SSTs and winds going from CCSM3 to CCSM4 (depicted in Fig. 2 as annual mean differences, CCSM4 minus CCSM3 for surface wind stress and SST) as well as better diffusivity in the ocean (Jochum and Potemra 2008; Jochum 2009). Changes to the convection scheme (Neale et al. 2008) have improved the SST simulation in the tropical Pacific with a less-extensive cold tongue (positive differences in the equatorial Pacific from 140°E to 120°W), an eastern Indian Ocean warm pool that is a bit cooler and closer to observed temperatures (negative differences in the equatorial eastern Indian Ocean), somewhat cooler SSTs near the Somali coast (negative differences), and warmer SSTs south of the equator in the central Indian Ocean, all closer to observations. In general, the equatorial lowlevel winds are weaker in CCSM4 compared to CCSM3, again moving the model simulations closer to observations. These improvements to the base state CCSM4 simulation compared to CCSM3 produce the improvements in precipitation noted above, namely, reduced westward extent of monsoon rainfall in the western Indian Ocean, much better simulation of the southern ITCZ near 5°S in the Indian Ocean, and a reduced precipitation bias north of Papua New Guinea in the Australian monsoon season. These changes are also associated with an improved simulation of ENSO in CCSM4 compared to CCSM3 (Deser et al. 2012).

4. Indian Ocean processes and the Asian–Australian monsoon

The role of the BoB in monsoon variability is well recognized from the northward propagation of the TCZ (Tropical Convergence Zone; Gadgil 1990) that contributes to monsoon onsets and breaks (Gadgil 2000; Bhat et al. 2001). The processes involved with ocean coupling are based on scales of SST variability (Sengupta and Ravichandran 2001) as well as the covariability of intraseasonal rainfall bands and SSTs (Vecchi and Harrison 2002). Using data from three moored buoys over the BoB, Sengupta and Ravichandran (2001) have shown that the temporal scale of air-sea coupling during intraseasonal oscillations over the monsoon region can be substantially less than a week. Yoo et al. (2006) note that Indian Ocean SSTs have a stronger role to play in monsoon variability over South Asia compared to the forcing from the equatorial Indian Ocean dipole mode and central Pacific Ocean SST, and that southern Indian Ocean SSTs are more closely related to Asian summer monsoon variability than northern Indian Ocean SSTs.



FIG. 1. South Asian monsoon precipitation (mm day⁻¹) and 850-hPa wind vectors (m s⁻¹) for JJAS; scaling arrow at lower left (a) observed, (b) CAM4 AMIP, and (c) CCSM4. An average of six ensemble members is shown for the models.

mm day⁻¹



FIG. 2. Change in annual mean base state surface temperature (°C) and wind stress (N m⁻², scaling arrow at lower right) for CCSM4 minus CCSM3.

The role of the SST gradient over the BoB in triggering convection has been documented by Shankar et al. (2007). Furthermore, the impact of freshwater forcing on upper-ocean structure and the feedbacks to the monsoons (Seo et al. 2009), and the interplay between the equatorial Indian Ocean and the BoB via the local Hadley cell during certain years (Slingo and Annamalai 2000), highlight the important role played by the BoB in monsoon convection. Considering the intimate relation between the interannual and intraseasonal variability of the monsoons (Palmer et al. 1992; Goswami and Ajaymohan 2001) and noting the generally deficient simulation of intraseasonal variability in global models (Waliser et al. 1999; Sperber and Annamalai 2008), it is essential to verify some basic relationships between BoB rainfall and regional rainfall variability. Here, we validate CCSM4 simulations against the GPCP precipitation product (used here because it affords a relatively long time series for the correlations) in terms of the correlations between the rainfall over various subregions of the BoB versus rainfall over India and the eastern half of the northern Indian Ocean. This should be instructive for assessing the role of the BoB in regional monsoon variability in the model compared to observations.

Summer precipitation in CCSM4 over the BoB agrees well with observational estimates starting in May with a northeast to southwest orientation of the precipitation pattern (maximum rainfall and its standard deviation of $\sim 3 \text{ mm day}^{-1}$ situated over the BoB, not shown). By June the orographic effect of the Burmese mountains takes control and the rainbands mainly extend southeast to northwest with the maximum rainfall increasing to $\sim 8 \text{ mm day}^{-1}$ in August and then tapering off to $\sim 4 \text{ mm day}^{-1}$ in September (not shown). Since our focus is only on the summer monsoon, we will not consider the rainfall variability over other months related to the northeast monsoon. While reliable prediction of the Indian monsoon at intraseasonal or seasonal time scales remains a daunting task, its response to climate change (e.g., Annamalai et al. 2007) and land use change becomes an additional challenge that cannot be ignored. This is especially true since, while seasonal amounts may remain similar, rainfall may already be characterized by enhanced extreme events (Goswami et al. 2006). This, combined with the importance of the BoB for the Indian monsoon, and the fact that the Indian Ocean has been undergoing the most rapid warming of all tropical oceans (Alory et al. 2007), is motivation to explore the observed relationship between rainfall variability over the BoB and India.

Figure 3 shows the correlations between the mean JJAS rainfall from GPCP (from 1979 to 2008) with the region focused on central India over which enhanced extreme events have been reported by Goswami et al. (2006). We select 4 boxes with $5^{\circ} \times 5^{\circ}$ latitudinal and longitudinal extent to represent the northern, eastern, and western BoB and the tropical eastern Indian Ocean region near the mouth of the BoB. These four boxes were chosen to represent different monsoon regimes over the BoB. The northern box is over the head of the Bay that is a part of the monsoon trough. This part receives one of the highest seasonal mean precipitation totals in the BoB. The western box represents heavy rainfall near the coast of the Burmese mountains. The eastern box, close to the Tamil Nadu coast, encounters the least amount of summer precipitation. The southern box is close to the region where convection is generated and moves north during northern summer. We have tried the same exercise for five more boxes over the BoB. However, the four boxes shown in Fig. 3 best represent the primary relationships. The correlations illustrate possible linkages to processes that connect these BoB areas with rainfall over central India.

Seasonal rainfall over the northern BoB shows significant negative correlations with rainfall over the



FIG. 3. Observed correlations between area-averaged seasonal rainfall over various subdomains (boxes) of the Bay of Bengal and gridded rainfall.

southeastern Peninsular India and Sri Lanka whereas the western BoB (close to the Indian peninsula) shows negative correlations with central India. This is part of the region identified in Goswami et al. (2006) for changes in frequency and amplitude of rainfall events over the last five decades of the twentieth century. The eastern BoB shows no obvious pattern of correlations, but the equatorial Indian Ocean (around 85° –90°E) shows an even stronger negative correlation with central India.

Correlations from CCSM4 for the same boxes (Fig. 4) are broadly consistent with those from the observations in Fig. 3. But the potential flip-flop in the seasonal rainfall between the western BoB and the equatorial region near the mouth of the BoB and the central Indian region are clear and need further detailed analysis for improving process and predictive understanding at intraseasonal to climate change time scales. The seasonality of the ENSO–monsoon and ENSO–BoB correlations, the seasonality of the trends in SSTs, and the intraseasonal to interannual and decadal to multidecadal variability of the BoB SSTs and their role in modulating

Indian monsoon variability can be effectively studied with CCSM4 for the processes involving natural variability and various anthropogenic effects. This is especially so given the improvements in the representation of phenomena that are critical for monsoon simulations such as the Madden Julian oscillation (MJO; see section 5, and Subramanian et al. 2011; Zhou et al. 2012), and ENSO (Neale et al. 2008).

5. Intraseasonal variability in the Asian–Australian monsoon

Intraseasonal variability plays an important role in the nature of monsoon precipitation (Hoyos and Webster 2007), and thus it is important to investigate the capability of the CCSM4 in simulating such variability. Modes of intraseasonal variability (10–90 days) in the Asian– Australian monsoon system fall within two primary time-scales. The biweekly (10–25 days) variations are associated with westward propagation and are closely related to monsoon active–break conditions (Krishnamurti and Ardunay 1980; Annamalai and Slingo 2001; Wen Correlation of JJAS Precipitation, CCSM4, 1000yr



FIG. 4. As in Fig. 3, but for a 1000-yr control simulation from CCSM4.

et al. 2010). The 30–60-day variability is characterized by poleward propagation and also affects the active– break cycles of the monsoon (i.e., Krishnamurti and Subrahmanyam 1982; Murakami et al. 1984; Webster et al. 1998). Some fraction of the poleward propagation associated with the 30–60-day intraseasonal variability is also associated with the eastward propagation of the Madden–Julian oscillation (i.e., Madden and Julian 1994; Wang and Rui 1990; Annamalai and Slingo 2001; Lawrence and Webster 2001). In this section we compare the atmospheric component simulations of CCSM3 and CCSM4, namely, CAM3 and CAM4, as well as some results from the fully coupled CCSM4, to observed intraseasonal variability.

a. Intraseasonal variance

Figure 5 shows the spectra for Indian rainfall $(15^{\circ}-25^{\circ}N, 70^{\circ}-90^{\circ}E)$ for observations from GPCP (top), for CAM4 (middle), and CAM3 (bottom) for the boreal summer season (May–October). The spectra were

computed using Fourier transform of daily precipitation anomalies for May–October and averaging the squared Fourier amplitude over all seasons. The relative maxima for the observations are in the 20–100-day range and in the 10–20-day range, likely reflecting the abovementioned primary intraseasonal time scales. CAM4 shows a number of relative maxima in the intraseasonal range, whereas CAM3 does not show marked peaks at intraseasonal frequencies beyond 25 days.

Figure 6 shows daily precipitation variance filtered for 30–60-day periods (Fig. 6a) and for 10–25-day periods (Fig. 6b) for the tropics. The precipitation observations from GPCP show regions of strong variance, for both intraseasonal bands, in the central Indian Ocean extending to the Bay of Bengal, over the tropical western north Pacific, and in the ITCZ and SPCZ regions. Daily precipitation variance for CAM4 for 30–60-day (Fig. 6c) and 10–25-day (Fig. 6d) periods again shows that the representation of intraseasonal variance in CAM4 is an improvement over CAM3 (Figs. 6e,f). CAM4 captures a region of high



FIG. 5. Boreal summer (May–October) spectra of daily precipitation anomalies over India ($15^{\circ}-25^{\circ}N$, $70^{\circ}-90^{\circ}E$) for (top) GPCP, (middle) CAM4, and (bottom) CAM3. The *x* axis is in frequency units of cycles day⁻¹ (cpd), such that 0.10 cpd represents a 10-day period, with the first tic mark equal to 0.02 or 50 days, second tic mark 0.04 or 25 days, and so on. The dashed lines are red noise first-order autoregressive (AR1) spectra.

variance over the Bay of Bengal seen in observations but overestimates the variance over the Arabian Sea. Additionally, the variance over the central Indian Ocean that is not seen in CAM3 is represented in CAM4, although it is weaker than observed. CAM4 also somewhat underestimates the variance in the tropical western North Pacific.

b. Spatial correlations and regressions

Following Goswami (2005), we use an intraseasonal variability index for the South Asian monsoon computed

from 10-90-day-filtered precipitation anomalies for 1 June to 30 September for the region 70°–90°E and 15°– 25°N, normalized by its standard deviation. The index is correlated and regressed on 10-90-day-filtered precipitation and 850-hPa zonal wind. The regressed lag-0 precipitation (Fig. 7a) and 850-hPa zonal wind (Fig. 7b) for GPCP and NCEP observations shows a relatively large-scale zonal pattern of precipitation over India associated with westerlies and dry conditions to the north and south associated with easterlies. This agrees well with previous studies (Goswami 2005). For CAM4, there is a large-scale zonal precipitation pattern over India and associated westerlies, although the westerlies are confined to the Indian subcontinent and the negative precipitation anomalies over the Indian Ocean are not present in CAM4 (Figs. 7c,d). There is a more meridional rather than zonal structure to the dipole pattern in the zonal winds for CAM4, which does not match observations. The regressed lag-14 precipitation (Fig. 8a) and 850-hPa zonal wind (Fig. 8b) for observations shows the opposite sign in precipitation and winds between Figs. 7a,b and 8a,b and reflects the active-break cycles of the South Asian monsoon. Figures 8c,d are the same as Figs. 8a,b, but for CAM4 at lag 15 since that time frame best represents the active-break cycle in the model.

We look now at the eastward and poleward propagation patterns for 20-100-day periods using the U.S. Climate Variability and Predictability (CLIVAR) metrics (locations are variable and seasonally dependent, see Table 2 from Waliser et al. 2009). Figure 9 shows the boreal summer (May–October) lag–longitude (Fig. 9a) and lag-latitude (Fig. 9b) cross correlation between 20-100-day-filtered precipitation (colors) and 850-hPa zonal wind (lines) at an Indian Ocean reference point with precipitation over the Indian Ocean. Figure 9a shows eastward propagation at a speed of about 6.4 m s⁻¹ for observations west of the date line. Figure 9b shows laglatitude correlations that represent poleward propagation between the equator and about 20° latitude in both hemispheres. The northward propagation speed is about 1.4 m s⁻¹, and the southward propagation speed is about 3.5 m s^{-1} . Additionally, in observations there is a lag between precipitation and winds documented in previous studies such as Zhang et al. (2006), who found that the precipitation leads the maximum westerlies by about 5 days or an eighth of a cycle. In CAM4 (Figs. 9c,d) there is eastward propagation, however, it appears to be a bit faster than observations. CAM4 also captures poleward propagation, although it is slower than observed with values of about 1 m s⁻¹ northward and 1.6 m s⁻¹ southward.

The lag between winds and precipitation is also captured in CAM4, at approximately 7 days. CAM3 shows eastward propagation in the zonal winds but no noticeable



FIG. 6. Variance of daily GPCP precipitation filtered for (a) 30–60-day periods and for (b) 10–25-day periods; variance of daily CAM4 precipitation filtered for (c) 30–60-day periods and for (d) 10–25-day periods; and variance of daily CAM3 precipitation filtered for (e) 30–60-day periods and for (f) 10–25-day periods.



eastward propagation in the precipitation (Fig. 9e). Furthermore, there is no northward propagation in CAM3 (Fig. 9f) and the winds and precipitation appear to propagate southward at different phase speeds.

In the tropical western North Pacific, we expect northward propagation during boreal summer into the East Asian monsoon region. Figure 10 is the same as in Fig. 9, but for the western Pacific boreal summer (for season and variable dependent reference points, see Table 2 from the Waliser et al. 2009). Although CAM4 captures the northward propagation with phase speeds of order 1.5 m s^{-1} , it shows westward rather than eastward propagation in this region, and thus does not represent much improvement from the propagation shown in CAM3 (Figs. 10e,f).

During boreal winter, we expect southward propagation of precipitation into northern Australia. Figures 11a,b show this, while CAM4 (Figs. 11c,d) does not capture the southward propagation in precipitation. It shows only weak poleward propagation in the zonal wind, and it again shows westward propagation west of the date line, where there should be eastward propagation. There also is equatorward zonal wind propagation at lags 10–20 following the southward propagation at earlier lags in observations (Fig. 11b). This may be related to propagation patterns shown by Wang and Rui (1990), but this behavior is not captured by the model.

The modes of intraseasonal variability in the coupled version of the model (CCSM4) are similar to those in CAM4. This may be due to the changes primarily in CAM4 over CAM3, compared to the benefits of coupling or improvements in the ocean component. CAM4, in addition to having a different dynamical core and higher horizontal resolution, also has two changes in the deep convection scheme. One is an entraining plume approximation in the CAPE calculation and the other is the inclusion of a convective momentum transport (see Raymond and Blyth 1986, 1992; Richter and Rasch 2008; Subramanian et al. 2011). Here, we analyze daily data from one of the twentieth-century simulations of CCSM4 (b40.20th.track1.1deg.008). Figure 12 shows the laglongitude and lag-latitude correlations as in Figs. 9, 10, and 11 but for CCSM4. The boreal summer Indian Ocean eastward propagation pattern (Fig. 12a) and timing closely

FIG. 7. Regressed GPCP (a) precipitation and (b) 850-mb zonal wind at lag 0 relative to the intraseasonal variability index over India (Goswami 2005); regressed CAM4 (c) precipitation and (d) 850-hPa zonal wind at lag 0 relative to the intraseasonal variability index over India (Goswami 2005).



FIG. 8. As in Fig. 7, but for observed lag 14 and model lag 15.

matches CAM4 (Fig. 9c), while the northward propagation in both precipitation and zonal wind is better represented in CAM4 (Fig. 9d) than in CCSM4 (Fig. 12b), but both have slower poleward propagation speeds compared with observations (Fig. 9b).

The propagation patterns for the tropical west Pacific during boreal summer look very similar in CCSM4 (Figs. 12c,d) as in CAM4 (Figs. 10c,d), including the bias of westward rather than eastward propagation compared with observations. The CCSM4 shows a weak southward propagation during boreal winter in the western tropical Pacific (Fig. 12f) and shows a similar westward propagation bias near the date line as was noted in CAM4 (Fig. 11d). Thus, to first order, intraseasonal variability in CCSM4 is similar to CAM4, in spite of different mean states and ocean coupling in CCSM4 that do not significantly affect the simulation of this variability in either the Indian or western Pacific regions. It is possible that this particular twentieth-century simulation shows little improvement between the coupled and uncoupled versions of the 1° model but that other realizations could. It is also possible the above-mentioned improvements from CAM3 to CAM4 represent the main mechanisms responsible for a better intraseasonal representation in CAM4 that does not improve significantly by comparison when coupled. Further aspects of intraseasonal variability in CCSM4 associated with the MJO are discussed by Subramanian et al. (2011) using output from a preindustrial CCSM4 run.

6. The Australian monsoon

The Australasian monsoon region is defined here to include the broad area of northern Australia and islands and seas of the Maritime Continent south of the equator. The rainfall in this region peaks in the DJF season and average rainfall and 850-hPa vector winds are shown in Fig. 13. The CAM4 AMIP simulation simulates excessive rainfall over the region, while CCSM4 reduces the simulated rainfall amounts to produce magnitudes closer to the observations. This was also the case for CCSM3 (Meehl et al. 2006, Fig. 6). However, the CCSM4 still has rainfall totals that are somewhat higher than observations by about 10% to 15%, extends the high rainfall amounts too far south in Australia, and has a strongerthan-observed ITCZ rainfall belt north of Papua New Guinea. This latter feature, which was even larger in CCSM3, has been reduced in CCSM4, which is a notable improvement over CCSM3 in terms of the simulation of precipitation over the west Pacific warm pool.

The annual cycle of rainfall over the Australian "Top End" (land points north of 15°S, between 129° and 137°E) is depicted in Fig. 14. Two observational products are



FIG. 9. Indian Ocean boreal summer (a) lag-longitude correlations and (b) lag-latitude correlations for 20–100-day-filtered GPCP precipitation (colors) and NCEP 850-mb zonal winds (lines) with Indian Ocean GPCP precipitation. (c),(d) As in (a),(b), but for CAM4 precipitation and zonal winds with Indian Ocean precipitation. (e),(f) As in (c),(d), but for CAM3 precipitation and zonal winds. (All following Waliser et al. 2009).

used to compare to the models, AWAP and TRMM. AWAP is a gridded rainfall dataset derived from station observations available over Australian land points from 1900 to the present (Jones et al. 2009). AWAP and TRMM give similar results over the TRMM period of 1998–2009, with the wet season beginning in austral spring and peaking in February, consistent with previous studies (Drosdowsky 1996). Slightly less rainfall is observed over the 1986–2005 period, giving some indication of the range of variability in the observations and the uncertainty in the TRMM climatology that results from having a relatively short record. Nevertheless, the model results lie outside the range of observations, with greater than observed rainfall throughout the year and a somewhat earlier rainfall peak (in January) in CAM4.

It has long been established that rainfall in the Australasian monsoon region has a strong relationship with ENSO (e.g., McBride and Nicholls 1983). During La Niña



FIG. 10. As in Fig. 9, but for the tropical west Pacific.

events, SSTs are typically warmer across the Maritime Continent region with enhanced convection leading to larger rainfall totals. However, for the Australian land points the impact of ENSO is primarily in the austral spring and the early part of the monsoon season (Hendon et al. 2012; Nicholls et al. 1982). Once the monsoon has begun, with monsoon onset typically in late December, the warm SSTs to the north of Australia (during La Niña) are damped because of the increased local wind speeds, and the rainfall becomes less persistent. Consistent with this, Australian summer monsoon rainfall has been shown to be essentially unpredictable on seasonal time scales (Hendon 2003; Hendon et al. 2012).

The influence of ENSO on Australian monsoon rainfall can clearly be seen in Fig. 15, which shows the correlation of an Australian monsoon rainfall index (15°S–equator, 110°–150°E) with SSTs for the September–November (SON) and December–February (DJF) seasons. During austral spring, increased Australian monsoon rainfall is associated with a typical La Niña pattern of anomalously cold SSTs (negative correlations) across the eastern equatorial Pacific and anomalously warm SSTs (positive



FIG. 11. As in Fig. 10, but for the boreal winter.

correlations) throughout the Maritime Continent and the eastern Indian Ocean. In austral summer the correlation has substantially weakened and weak negative correlations are now found to the north of Australia (i.e., enhanced rainfall in the summer monsoon is associated with negative local SST anomalies). This seasonal contrast is also captured by the CAM4 and CCSM4 models, with weaker correlations of Australian–Indonesian monsoon rainfall with SSTs in summer compared to spring. However, in contrast to the observations, both models maintain positive correlations to the north and northwest of Australia in DJF, a bias consistent with previous versions of the model (Meehl and Arblaster 1998). These errors are also associated with an error in the relationship between Australian rainfall and 850-hPa winds in this region, with the models overestimating the connection between monsoon rainfall and local winds in contrast to the very weak relationship in the observations (Fig. 15b). This may be related to errors in the mean winds that are seen in Fig. 13 to be too strong and zonal along the northwest coast of Australia in the models.



FIG. 12. Indian Ocean boreal summer (a) lag–longitude correlations and (b) lag–latitude correlations for 20–100-day-filtered CCSM4 precipitation (colors) and CCSM4 850-hPa zonal winds (lines) with Indian Ocean CCSM4 precipitation; tropical west Pacific boreal summer (c) lag–longitude correlations and (d) lag–latitude correlations for 20–100-day-filtered CCSM4 precipitation (colors) and CCSM4 850-mb zonal winds (lines) with west Pacific precipitation; and tropical west Pacific boreal winter (e) lag–longitude correlations and (f) lag–latitude correlations for 20–100-day filtered CCSM4 precipitation (colors) and CCSM4 850-hPa zonal winds (lines) with west Pacific precipitation; and tropical west Pacific boreal winter (e) lag–longitude correlations and (f) lag–latitude correlations for 20–100-day filtered CCSM4 precipitation (colors) and CCSM4 850-hPa zonal winds (lines) with west Pacific precipitation (all following Waliser et al. 2009).

Recent studies have shown there is some predictability of rainfall in the Australian–Indonesian monsoon region to be gained from knowledge of the phase of the MJO (Wheeler et al. 2009; Wheeler and McBride 2011), with MJO phases 5 and 6 having the largest positive impact (i.e., increased rainfall) on Australasian monsoon rainfall. In addition to the results presented previously in section 5, a detailed discussion of the MJO in the CCSM4 is given in Subramanian et al. (2011). They find a similar progression of the MJO in the CCSM4 as in the observations (their Fig. 14), with maximum convection over the Maritime Continent in phases 5 and 6. However, the



FIG. 13. As in Fig. 1, but for Australian monsoon for DJF averaged over 1986–2005.



FIG. 14. Annual cycle of rainfall over the Australian "Top End" for observations (solid), the CAM4 AMIP (dashed), and CCSM4 (dotted). An average of six ensemble members is shown for the models.

magnitude of the CCSM4 convection and winds in these MJO phases is somewhat weaker than observed.

7. Indonesian Throughflow and the Asian–Australian monsoon

The Maritime Continent at the western edge of the Pacific represents a key region for the Walker circulation,

and the overlying atmosphere is characterized by strong precipitation, convection, and low-level wind convergence. This can be attributed to the high SSTs (Fig. 16) in the Indonesian Seas, which provide a significant source of energy to the atmosphere (e.g., Palmer and Mansfield 1984; Barsugli and Sardeshmukh 2002). Thus, the SST in this region exerts a strong influence on the Walker cell and the Asian–Australian monsoon, and one focus of ocean model development over the last several years has been on improving the fidelity of the water masses in the Indonesian Seas. These developments and their impacts will be documented here.

The mostly North Pacific waters are carried through the Indonesian Seas in a multitude of currents, which are summarily referred to as the Indonesian Throughflow (ITF). They bring about 0.5 PW (10¹⁵ watts) of heat from the tropical Pacific to the Indian Ocean (Vranes et al. 2002). The ITF in CCSM3 had excessive transport through the Torres Strait (between Australia and Papua New Guinea), the Lombok and Ombai Straits (8°S, 116°E; and 6°S, 105°E, respectively; between Sumatra, Java, and Timor), and through the Sulu Sea (8°N, 120°E; between the Philippines and Borneo). The total ITF transport is controlled largely by the Southern Ocean and equatorial winds ("Island Rule", Godfrey 1989), but how



FIG. 15. Correlation of Australasian monsoon rainfall with surface temperature and 850-hPa winds for austral (left) spring and (right) summer for (a),(b) observations, (c),(d) CAM4 AMIP, and (e),(f) CCSM4. An average of six ensemble members is shown for the models. Scaling arrows in lower-left corners indicate the magnitude of a 1.0 correlation for the 850-hPa winds.



FIG. 16. SST (colors) and vertically integrated transport (contour interval 2 Sv for transports between -15 and +15 Sv, and 10 Sv beyond that) in CCSM4 for 1996–2005, based on six twentieth-century ensembles.

it is distributed among the different straits is crucially dependent on topography and viscosity (Wajsowicz 1993). Inspection of the ocean topography in CCSM3 showed that Palawan (10°N, 119°E), which blocks the connection between the South China Sea and the Sulu Sea, had been obliterated during the smoothing of the topography in CCSM3. Restoring it reduced the flow through the Sulu Sea from 8 Sv (1 Sv $\equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) in CCSM3 to a more realistic transport of less than 2 Sv (Cai et al. 2009) in CCSM4 (Fig. 16). This also reduces the warm water entering the South China Sea through the Luzon Strait (20°N, 120°E) and leads to reduction of the annual mean precipitation over the South China Sea of approximately 1.5 mm yr^{-1} (not shown). Similarly, the Ombai and Lombok straits have been made shallower in CCSM4, and the Torres Strait has been closed altogether to achieve more realistic transports (compared to the latest estimates by Gordon et al. 2010). These changes did not lead to any significant changes in water mass structure or precipitation. It should be noted, though, that during past climates the Torres Strait could have been quite important (Cane and Molnar 2001) and made a significant contribution to the ITF (Jochum et al. 2009).

Further improvements in the water mass structure and precipitation were achieved by replacing the constant vertical background diffusivity with the observed diapycnal diffusivity. The thermocline diffusivity is enhanced in the Banda Sea near 6°S, 130°E (e.g.; Wyrtki 1961, Ffield and Gordon 1992), and reduced along the equator (Gregg et al. 2003). Accounting for the former in CCSM4 leads to realistic Banda Seawater properties and reduced precipitation over the Banda Sea (Jochum and Potemra 2008); accounting for the latter leads to a shift of the precipitation away from the warm pool toward the central Pacific (Jochum 2009), and together the two improvements account for the approximately 40% reduction of precipitation bias of CCSM4 over the Maritime Continent (compared to CCSM3, Fig. 13 of Gent et al. 2011).

The strength of the ITF in the 1850 control version of CCSM4 is 12.5 \pm 1.3 Sv (Fig. 16), consistent with the observational estimate of 13-15 Sv (Gordon et al. 2010). The CCSM4 value is 87% of the 14.3 \pm 1.6 Sv predicted by the Island Rule (Godfrey 1989), which makes the annual mean surface wind stress a good predictor for long term changes in the ITF. In the six twentieth-century CCSM4 ensembles, the transport weakens steadily to 12.0 ± 0.1 Sv (based on 6 means of the years 1996–2005), only slightly less than the reduction expected from the Island Rule of 0.8 \pm 0.2 Sv. Thus, one can analyze the twentieth-century shifts in the wind stress to understand the reason for this decline. It is found that the weakening of equatorial easterlies and the poleward shift of the Southern Hemisphere westerlies both account for about half of it. At least over the twentieth century this weakening of the ITF has not yet led to a noticeable temperature or salinity change in the source waters for the Indian Ocean.

8. South Asian Monsoon–ENSO connections

Walker (1924) and Walker and Bliss (1932) noted the simultaneous relationship between surface pressure variations associated with the Southern Oscillation and Indian monsoon rainfall fluctuations. Following on this work and consistent with the hypothesis embodied in Charney and Shukla (1981), many observational and modeling studies have suggested that the mean and interannual variability associated with the planetary-scale monsoon are influenced by slowly varying boundary conditions. Of all the boundary forcing elements, ENSOrelated SST anomalies are perhaps the dominant forcing element of monsoon interannual variations (e.g., Sikka 1980; Webster and Yang 1992). However, for atmospheric models (run with observed SSTs) to capture the ENSO-monsoon association, a realistic simulation of mean monsoon precipitation is a necessary ingredient (Meehl and Arblaster 1998; Sperber and Palmer 1996). An additional challenge for global coupled atmosphereocean general circulation models (AOGCMs) is to capture the mean state in the tropical Pacific as well (Turner et al. 2005). In section 6 above (Fig. 15), results were presented that showed linkages between the Australian monsoon and ENSO. Here, we examine such connections for the South Asian monsoon in more detail.

Annamalai et al. (2007) analyzed the ENSO–South Asian monsoon association in the AOGCMs that participated in the CMIP3 that were assessed in the IPCC AR4.



FIG. 17. (a) Seasonal mean (June–September) precipitation (shaded, mm day⁻¹) and SST (contours, °C) climatology from the twentieth-century integrations of CCSM4; (b) monthly temporal evolution of composite SST anomalies (in standard deviations) averaged over the Niño-3.4 region (5°S–5°N, 120°–170°W) from CCSM4 (dashed line) and observations (thick line). In (a) the 28°C isotherm is shown in heavy dotted and 20°C isotherm in heavy lines, respectively. In (b) year (0) represents the developing and year (+1) the decaying, phases of El Niño, respectively.

They concluded that apart from mean monsoon precipitation, models that correctly simulate the timing and location of SST and diabatic heating anomalies over the equatorial Pacific, and the associated changes to the largescale east–west Walker circulation, capture the monsoon– ENSO relationship realistically. Following the conjectures of Annamalai et al. (2007), we examine first the ability of CCSM4 in capturing the basic state over the South Asian summer monsoon and ENSO characteristics. Then, the ENSO–monsoon association is examined in six ensemble members of the twentieth-century integrations.

a. Mean monsoon precipitation

The simulation of mean precipitation climatology has proven to be rather difficult and therefore provides a severe test of the AOGCMs. Figure 17a shows the

seasonal average (June-September) precipitation (shaded) and SST (contours) climatology constructed from the last 30 years (1976-2005) of the simulation. Consistent with observations and model results shown in section 3, the model captures the three regional precipitation centers or heat sources in the south Asian region that include (i) the Indian summer monsoon (ISM: 10°-25°N, 70°-100°E), (ii) the western North Pacific monsoon (WNPM: 10°–20°N, 110°–150°E), and (iii) the equatorial Indian Ocean (EIO: 10°S-0°, 60°-90°E). Compared to the AOGCMs analyzed by Annamalai et al. (2007), CCSM4 has higher horizontal resolution that results in a more realistic representation of orographically induced intense rainfall along the Western Ghats of India, the Burmese coast, and the foot-hills of the Himalayas as noted in Fig. 1. While the rainfall along the monsoon trough extending

from northwest India to the tropical west Pacific is realistic, the local maximum along the EIO is shifted westward as noted in section 3. A realistic representation of these centers is important to adequately investigate the ISM variability because these centers do not respond in unison to ENSO forcing and the convective variability over the EIO and WNPM modulate the ISM at interannual time scales (Annamalai and Liu 2005; Annamalai 2010). Other realistically simulated rainfall features include ITCZs along the Pacific and Atlantic at 10°N, and the West African monsoon (see Part II of this paper, Cook et al. 2012). However, rainfall along the SPCZ is much stronger. As regards to SST basic state, the east-west SST gradient along the equatorial Pacific and the spatial extent of the Indo-Pacific warm pool $(SST > 28^{\circ}C)$ are apply simulated. Given the overall reasonably realistic basic state in CCSM4, next we investigate its ability in capturing the space-time evolution of SST during ENSO, and the associated precipitation (diabatic heating) anomalies along the equatorial Pacific.

b. ENSO characteristics

Figure 17b shows the composite evolution of Niño-3.4 SST anomalies during strong El Niño events [defined as area-averaged Niño-3.4 (5°S-5°N, 120°-170°W) SST exceeding 1.0 standard deviation during the monsoon season]. Though the canonical ENSO event begins and ends in northern spring in both CCSM4 and observations, some events last more than one year so the composite average event lasts about 18 months in Fig. 18. Here, year (0) represents developing and year (+1)decaying phase of El Niño, respectively. In terms of amplitude and temporal characteristics the model El Niño is in general agreement with observations. Spectra of the Niño-3.4 time series also show much improvement in CCSM4 compared to CCSM3 with regards to the time scale of ENSO being less biennial in CCSM4, with the more dominant periods being about 3-5 years as in observations (Meehl and Arblaster 2011; Deser et al. 2012).

Apart from the Niño-3.4 index, the spatial evolution of equatorial Pacific SST and precipitation anomalies that determine the changes in the Walker Circulation are elements that influence the fidelity of the model ENSO. Typical features include the eastward movement of warmest SST (Fig. 18a) that is accompanied by an eastward migration of rainfall in the central Pacific and a reduction in rainfall over the Maritime Continent (Fig. 18c). It is indeed this redistribution of diabatic heating sources and sinks that determines the rising–descending branches of the anomalous Walker Circulation (Meehl et al. 2003; Turner et al. 2005; Annamalai et al. 2007). In CCSM4, the simulated equatorial Pacific SST during El Niño (Fig. 18b) is higher than observed, but the warm SST signal depicts a clear eastward migration and attains a peak toward the end of year (0) and beginning of year (+1). Compared to observations, the model warm SST extends farther east ($\sim 125^{\circ}W$) during the developing phase [year (0)] but retards rather quickly during the decaying phase [year (+1)]. Compared to the earlier simulations with the NCAR Parallel Climate Model (PCM), where the warm SST did not develop in the western-central Pacific during El Niño (see Fig. 3b in Annamalai et al. 2007), the El Niño simulation in CCSM4 is indeed more realistic. It is notable that the model (Fig. 18d) simulates reduced precipitation over the Maritime Continent and increased rainfall over the central-eastern Pacific as observed (Fig. 18c). The simulated decreased rainfall over the Maritime Continent occurs despite the presence of high-mean SST, indicative of subsidence induced by the anomalous atmospheric circulation.

c. ENSO-monsoon association

We analyze the role of ENSO on the South Asian monsoon by plotting correlation coefficients of the SST throughout the tropical Pacific and Indian Ocean region with the AIR index (to be consistent with observations, the AIR index for the model results is constructed using only the land points over the region 7° - 30° N, 65° - 95° E). The diagnostics are calculated in each of the six realizations separately, and then a grand ensemble mean is computed for presentation and discussion. Barring slight differences in the amplitude of the correlations, the teleconnection patterns computed from the individual members (not shown) show features similar to the ensemble mean.

Figure 19 shows the simultaneous correlation coefficient between AIR and SST from observations (Fig. 19a) and the ensemble-mean pattern from CCSM4 (Fig. 19b). While the inverse association between ENSO and monsoon is well captured by CCSM4, the strength of the negative correlations along the equatorial central-eastern Pacific is much weaker (note different shading intervals between Figs. 19a and 19b). However, for the period analyzed here (1850–2005), an absolute value of 0.22 is significant at the 95% level, and therefore the results are notable. Based on this statistic, one can suggest that ENSO explains only about 10%–15% of AIR variability in the model.

To check if the timing in ENSO–monsoon association is faithfully represented, we examine lead–lag correlations between Niño-3.4 and AIR anomalies for a 2-yr period (one year before and after the monsoon season). The choice of Niño-3.4 rather than Niño-3 region is due to the fact that both in observations and the model the strongest anticorrelations between AIR and SST occur over this region (Fig. 19). In observations, the occurrence



FIG. 18. Monthly SST evolution in the equatorial (5°S–5°N) Pacific during El Niño years for (a) observations and (b) CCSM4. Values are not plotted below the 27.5°C threshold for deep convection to occur. (c),(d) As in (a),(b), but for precipitation anomalies. The time coordinate starts with January of year (0) of the developing phase and ends with December of year (+1) of the decaying phase of El Niño. The composites are based on strong El Niño events (>1.0 standard deviation of SST anomalies over the Niño-3.4 region during June–September).

of negative correlations (Fig. 19c, black line) begins in April. The maximum correlation after the monsoon season has prompted investigators to propose the hypothesis that variations in the intensity of monsoon rainfall potentially influence the surface wind stress in the equatorial Pacific and thereby modify the statistical properties of ENSO (Kirtman and Shukla 2000). As before, the negative correlations in the model are weak (Fig. 19c, dashed line) but the timing appears realistic. In particular, the appearance of negative correlations in boreal spring indicates that ENSO can be treated as a potential predictor of seasonal mean AIR anomalies in CCSM4.



FIG. 19. Correlation patterns between seasonal mean (June–September) AIR indices and SST anomalies from (a) observations and (b) CCSM4. (c) Lead–lag correlation between AIR anomalies and Niño-3.4 SST anomalies from observations (thick line) and CCSM4 (dashed line). Lag 12 corresponds to Niño-3.4 SST anomalies one year before the monsoon season.

Why are the model anticorrelations weaker compared to observations? While there may be many reasons, we speculate the following, but a detailed investigation is deferred to a future study. In each of the six twentieth-century ensemble members, of the total 156 years of integrations (1850–2005), ENSO occurs in about 52–56 years, that is, in about one-third of the total period. During boreal summer of El Niño, the simulated warm SSTs extend too



FIG. 20. (a) Correlation between the Niño-3.4 SST time series and all-India rainfall for observations for 13-yr running time periods; (b) as in (a), but for the individual ensemble members from CCSM4, along with the ensemble mean in black; and (c) as in (a), but for the 500-yr period of the CCSM4 control run.

far east in the equatorial Pacific (Fig. 18b) resulting in a local maximum in positive rainfall anomalies east of the date line (\sim 150°W) while observed positive rainfall anomalies are more prevalent to west of the date line (\sim 170°E). Thus, the rising branch of the anomalous Walker Circulation also shifts into the eastern rather than central Pacific. This, in conjunction with the westward shift in the location of EIO precipitation in the basic state (Fig. 17a) can influence the intensity and location of the descending branch of the anomalous Walker Circulation over South Asia (i.e., the entire Walker Circulation could shift, Annamalai et al. 2007). Another possible reason may be differences in the model internal variability over the monsoon region.

However, there are changes in the strength of the monsoon-ENSO connection that have been noted in observations such that there has been a recent apparent weakening of this connection (e.g., Kumar et al. 1999). It was speculated that this change could be affected by external forcing from increasing GHGs. But could this just as easily be due to internally-generated decadal time-scale variability? To test this possibility, we first calculate the correlation between Niño-3.4 and all-India rainfall from observations for 13-yr running time periods in Fig. 20a. As shown by Kumar et al. (1999) for a somewhat shorter period of record, Fig. 20a shows a recent deviation from negative correlations ranging from about -0.3 to over -0.8, to near zero or even slightly positive values since the mid-1990s. Next, we perform this calculation for the five twentieth-century ensemble members from CCSM4 (Fig. 20b), as well as from the 500-yr control run from CCSM4 (Fig. 20c). Results show multidecadal time-scale variability of the monsoon-ENSO relationship occurs naturally in the model, with values in the individual ensemble members and the long control run ranging from somewhat positive to about -0.8, with the ensemble average value of around -0.35for the twentieth-century simulations. Thus, as with most climate variability in the Indo-Pacific region, internally generated SST changes can affect the strength of Indo-Pacific connections without invoking external forcing. However, it is likely that external forcing, which produces an SST pattern similar to that associated with internally generated variability, could play a role as well (e.g., Meehl et al. 2009). Annamalai et al. (2007) arrived at similar conclusions based on an analysis of CMIP3 models.

9. Conclusions

The general simulation characteristics of the Asian– Australian monsoon system are documented for CCSM4. Some comparisons are made to AMIP simulations of the atmospheric component in CCSM4, the CAM4, to suggest changes to the monsoon simulations with and without accurate simulations of SSTs and/or ocean– atmosphere coupling, and to CCSM3, an earlier version of the coupled GCM. In general, monsoon rainfall is too heavy in the uncoupled AMIP runs with CAM4, and monsoon rainfall amounts in some regions are better simulated with ocean coupling in CCSM4.

Many aspects of the Asian–Australian monsoon simulations are improved in CCSM4 compared to CCSM3. For the South Asian monsoon, this includes a reduction of the systematic error of rainfall extending too far west in the western Indian Ocean, and the southern ITCZ near 5°S is much better simulated in CCSM4. Observed connections between SSTs in the Bay of Bengal and regional South Asian monsoon precipitation are represented in CCSM4. Improvements are also seen in the pattern of rainfall in the Australian monsoon, in part due to contributions from the inclusion of the ITF in CCSM4 (it was absent in CCSM3) and through the change to diapycnal diffusion in the ocean. Both of these improve the SST and associated precipitation simulations in the Australian monsoon over the Maritime Continent. A simulated weakening of the ITF during the course of the twentieth century is traced to a reduction in strength of equatorial easterlies and the poleward shift of the Southern Hemisphere westerlies. Intraseasonal variability of the Asian-Australian monsoon is much improved in CCSM4 compared to CCSM3 both in terms of eastward and northward propagation characteristics. However, monsoon intraseasonal variability is still somewhat weaker in CCSM4 compared to observations.

The improved simulation of El Niño in CCSM4 contributes to more realistic connections between the Asian– Australian monsoon and ENSO, though the pattern of the correlation is somewhat weaker than observed. However, results from the long CCSM4 control run reveal considerable decadal and century time-scale variability of the strength of the monsoon–ENSO connection.

In general, there is an improvement in going from CCSM3 to CCSM4 for the Asian–Australian monsoon. Some of these improvements are related to the betterresolved representation of regional topography in CCSM4, and some are associated with improvements in the simulated SSTs in CCSM4. CAM4 provides insights into the role of ocean–atmosphere coupling and the correct simulation of SSTs. The CAM4 produces excessive monsoon precipitation in the Asian–Australian monsoon that is reduced in closer agreement to observations in the CCSM4. However, intraseasonal variability in CCSM4 is mostly similar to CAM4, in spite of different mean states and ocean coupling in CCSM4 that do not significantly affect the simulation of this variability in either the Indian or western Pacific regions.

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REFERENCES

- Alory, G. S., and Coauthors, 2007: Observed temperature trends in the Indian Ocean over 1960–1999 and associated mechanisms. *Geophys. Res. Lett.*, 34, L02606, doi:10.1029/2006GL028044.
- Annamalai, H., 2010: Moist dynamical linkage between the equatorial Indian Ocean and the South Asian monsoon trough. J. Atmos. Sci., 67, 589–610.
- —, and J. M. Slingo, 2001: Active/break cycles: Diagnosis of the intraseasonal variability over the Asian summer monsoon. *Climate Dyn.*, 18, 85–102.
- —, and P. Liu, 2005: Response of the Asian summer monsoon to changes in El Niño properties. *Quart. J. Roy. Meteor. Soc.*, 131, 805–831.
- —, K. Hamilton, and K. R. Sperber, 2007: The South Asian monsoon and its relationship with ENSO in the IPCC AR4 simulations. J. Climate, 20, 1071–1092.
- Barsugli, J. J., and P. Sardeshmukh, 2002: Global atmosphere sensitivity to tropical SST anomalies throughout the Indo-Pacific basin. J. Climate, 15, 3427–3441.
- Bhat, G. S., and Coauthors, 2001: BOBMEX: The Bay of Bengal Monsoon Experiment. Bull. Amer. Meteor. Soc., 82, 2217– 2243.
- Cai, S., Y. He, S. Wang, and X. Long, 2009: Seasonal upper circulation in the Sulu Sea from satellite altimetry data and a numerical model. J. Geophys. Res., 114, C03026, doi:10.1029/ 2008JC005109.
- Cane, M. A., and P. Molnar, 2001: Closing of the Indonesian seaway as a precursor to East African aridification around 3–4 million years ago. *Nature*, **411**, 157–162.
- Charney, J. G., and J. Shukla, 1981: Predictability of monsoons. Monsoon Dynamics, J. Lighthill and R. P. Pearce, Eds., Cambridge University Press, 99–109.
- Collins, W. D., and Coauthors, 2006: The Community Climate System Model, version 3 (CCSM3). J. Climate, 19, 2122– 2143.
- Cook, K. H., G. A. Meehl, and J. M. Arblaster, 2012: Monsoon regimes and processes in CCSM4. Part II: African and American monsoon systems. J. Climate, 25, 2609–2621.
- Deser, C., and Coauthors, 2012: ENSO and Pacific decadal variability in Community Climate System Model version 4. J. Climate, 25, 2622–2651.
- Drosdowsky, W., 1996: Variability of the Australian summer monsoon at Darwin: 1957–92. J. Climate, 9, 85–96.
- Ffield, A., and A. L. Gordon, 1992: Vertical mixing in the Indonesian thermocline. J. Phys. Oceanogr., 22, 184–195.
- Gadgil, S., 1990: Poleward propagation of the ITCZ: Observations and theory. *Mausam (New Delhi)*, **41**, 285–290.
 - ----, 2000: Monsoon-ocean coupling. Curr. Sci., 78, 309-323.
- Gent, P., and Coauthors, 2011: The Community Climate System Model, version 4. J. Climate, 24, 4973–4991.
- Godfrey, J. S., 1989: A Sverdrup model of the depth-integrated flow for the World Ocean allowing for island circulations. *Geophys. Astrophys. Fluid Dyn.*, 45, 89–112.
- Gordon, A. L., and Coauthors, 2010: The Indonesian Throughflow during 2004–2006 as observed by the INSTANT program. *Dyn. Atmos. Oceans*, **50**, 115–128.

- Goswami, B. N., 2005: South Asian monsoon. Intraseasonal Variability in the Atmosphere-Ocean Climate System, W. K.-M. Lau and D. E. Waliser, Eds., Praxis, 19–61.
- —, and R. S. Ajaymohan, 2001: Intraseasonal oscillations and interannual variability of the Indian summer monsoon. *J. Climate*, **14**, 1180–1198.
- —, and Coauthors, 2006: Increasing trend of extreme rain events over India in a warming environment. Science, 314, 1442–1445.
- Gregg, M. C., T. B. Sanford, and D. P. Winkel, 2003: Reduced mixing from the breaking of internal waves in equatorial waters. *Nature*, 422, 513–515.
- Hendon, H. H., 2003: Indonesian rainfall variability: Impacts of ENSO and local air-sea interaction. J. Climate, 16, 1775–1790.
- —, E.-P. Lim, and G. Liu, 2012: The role of air-sea interaction for prediction of Australian summer monsoon rainfall. *J. Climate*, **25**, 1278–1290.
- Hoyos, C. D., and P. J. Webster, 2007: The role of intraseasonal variability in the nature of Asian monsoon precipitation. *J. Climate*, **20**, 4402–4424.
- Huffman, G. J., R. F. Adler, M. M. Morrissey, D. T. Bolvin, S. Curtis, R. Joyce, B. McGavock, and J. Susskind, 2001: Global precipitation at one-degree daily resolution from multisatellite observations. J. Hydrometeor., 2, 36–50.
- —, and Coauthors, 2007: The TRMM Multisatellite Precipitation Analysis (TMPA): Quasi-global, multiyear, combined-sensor precipitation estimates at fine scales. J. Hydrometeor., 8, 38–55.
- Jochum, M., 2009: Impact of latitudinal variations in vertical diffusivity on climate simulations. J. Geophys. Res., 114, C01010, doi:10.1029/2008JC005030.
- —, and J. Potemra, 2008: Sensitivity of tropical rainfall to Banda Sea diffusivity in the Community Climate System Model. J. Climate, 21, 6445–6454.
- —, B. Fox-Kemper, P. H. Molnar, and C. Shields, 2009: Differences in the Indonesian Seaway in a coupled climate model and their relevance for Pliocene climate and El Niño. *Paleoceanography*, 24, PA1212, doi:10.1029/2008PA001678.
- Jones, D. A., W. Wang, and R. Fawcett, 2009: High-quality spatial climate data-sets for Australia. *Aust. Meteor. Oceanogr. J.*, 58, 233–248.
- Kalnay E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. Bull. Amer. Meteor. Soc., 77, 437–471.
- Kirtman, B. P., and J. Shukla, 2000: Influence of the Indian summer monsoon on ENSO. *Quart. J. Roy. Meteor. Soc.*, **126**, 1623– 1646.
- Krishnamurti, T. N., and P. Ardunay, 1980: The 10 to 20 day westward propagating mode and "breaks" in the monsoons. *Tellus*, **32**, 15–26.
- —, and D. Subrahmanyam, 1982: The 30–50-day mode at 850 mb during MONEX. J. Atmos. Sci., 39, 2088–2095.
- Kumar, K., and Coauthors, 1999: On the weakening relationship between the Indian monsoon and ENSO. *Science*, 284, 2156– 2159.
- Lawrence, D. M., and P. J. Webster, 2001: Interannual variations of the intraseasonal oscillation in the south Asian summer monsoon region. J. Climate, 14, 2910–2922.
- Lawrence, P. J., and Coauthors, 2012: Simulating the biogeochemical and biogeophysical impacts of transient land cover change and wood harvest in the Community Climate System Model (CCSM4) from 1850 to 2100. J. Climate, in press.
- Madden, R. A., and P. R. Julian, 1994: Detection of a 40-50-day oscillation in the zonal wind in the tropical Pacific. *Mon. Wea. Rev.*, **122**, 813–837.

- McBride, J. L., and N. Nicholls, 1983: Seasonal relationships between Australian rainfall and the Southern Oscillation. *Mon. Wea. Rev.*, **111**, 1998–2004.
- Meehl, G. A., and J. M. Arblaster, 1998: The Asian–Australian monsoon and El Niño–Southern Oscillation in the NCAR Climate System Model. J. Climate, 11, 1356–1385.
- —, and —, 2011: Decadal variability of Asian–Australian monsoon–ENSO–TBO relationships. J. Climate, 24, 4925– 4940.
- —, —, and J. Loschnigg, 2003: Coupled ocean-atmosphere dynamical processes in the tropical Indian and Pacific Ocean regions and the TBO. J. Climate, 16, 2138–2158.
- —, and Coauthors, 2006: Monsoon regimes in the CCSM3. J. Climate, 19, 2482–2495.
- —, A. Hu, and B. D. Santer, 2009: The mid-1970s climate shift in the Pacific and the relative roles of forced versus inherent decadal variability. J. Climate, 22, 780–792.
- Murakami, T., T. Nakazawa, and J. He, 1984: On the 40–50-day oscillation during 1979 Northern Hemisphere summer. Part I: Phase propagation. J. Meteor. Soc. Japan, 62, 440–468.
- Neale, R. B., J. H. Richter, and M. Jochum, 2008: The impact of convection on ENSO: From a delayed oscillator to a series of events. J. Climate, 21, 5904–5924.
- Nicholls, N., J. L. McBride, and R. J. Ormerod, 1982: On predicting the onset of the Australian wet season at Darwin. *Mon. Wea. Rev.*, **110**, 14–17.
- Palmer, T. N., and D. A. Mansfield, 1984: Response of two atmospheric general circulation models to SST anomalies in the tropical east and west Pacific. *Nature*, **310**, 483–488.
- —, C. Brankovic, P. Viterbo, and M. J. Miller, 1992: Modeling interannual variability of summer monsoons. J. Climate, 5, 399–417.
- Parthasarathy, B., K. Rupakumar, and A. A. Munot, 1994: All-India monthly and seasonal rainfall series: 1871-1993. *Theor. Appl. Climatol.*, **49**, 217–224.
- Raymond, D., and A. M. Blyth, 1986: A stochastic mixing model for nonprecipitating cumulus clouds. *J. Atmos. Sci.*, **43**, 2708–2718.
 —, and —, 1992: Extension of the stochastic mixing model to cumulonimbus clouds. *J. Atmos. Sci.*, **49**, 1968–1983.
- Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan, 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. J. Geophys. Res., 108, 4407, doi:10.1029/2002JD002670.
- Richter, J., and P. J. Rasch, 2008: Effects of convective momentum transport on the atmospheric circulation in the Community Atmosphere Model, version 3. J. Climate, 21, 1487–1499.
- Sengupta, D., and M. Ravichandran, 2001: Oscillations of Bay of Bengal sea surface temperature during the 1998 summer monsoon. *Geophys. Res. Lett.*, 28, 2033–2036.
- Seo, H., S.-P. Xie, R. Murtugudde, M. Jochum, and A. J. Miller, 2009: Seasonal effects of Indian Ocean freshwater forcing in a regional coupled model. *J. Climate*, **22**, 6577–6596.
- Shankar, D., S. R. Shetye, and P. V. Joseph, 2007: Link between convection and meridional gradient of sea surface temperature in the Bay of Bengal. J. Earth Syst. Sci., 116, 385–406.
- Sikka, D. R., 1980: Some aspects of the large-scale fluctuations of summer monsoon rainfall over India in relation to fluctuations in the planetary and regional scale circulation parameters. *Proc. Indian Natl. Sci. Acad.*, **89**, 179–195.
- Simmons, A., S. Uppla, D. De, and S. Kobayashi, 2006: ERA-Interim: New ECMWF reanalysis products from 1989 onwards. *ECMWF Newsletter*, No. 110, ECMWF, Reading, United Kingdom, 25–35.

- Slingo, J. M., and H. Annamalai, 2000: The El Niño of the century and the response of the Indian summer monsoon. *Mon. Wea. Rev.*, **128**, 1778–1797.
- Smith, T. M., R. W. Reynolds, T. C. Peterson, and J. Lawrimore, 2008: Improvements to NOAA's historical merged land-ocean surface temperature analysis (1880–2006). J. Climate, 21, 2283–2296.
- Sperber, K. R., and T. N. Palmer, 1996: Interannual tropical rainfall variability in general circulation model simulations associated with the Atmospheric Model Intercomparison Project. J. Climate, 9, 2727–2750.
- —, and H. Annamalai, 2008: Coupled model simulations of boreal summer intraseasonal (30–50 day) variability. Part I: Systematic errors and caution on use of metrics. *Climate Dyn.*, **31**, 345–372, doi:10.1007/s00382-008-0367-9.
- Subramanian, A. C., M. Jochum, M. Miller, A. J. Murtugudde, R. B. Neale, and D. E. Waliser, 2011: The Madden–Julian oscillation in CCSM4. J. Climate, 24, 6261–6282.
- Turner, A. G., P. M. Inness, and J. M. Slingo, 2005: The role of the basic state in the ENSO–monsoon relationship and implications for predictability. *Quart. J. Roy. Meteor. Soc.*, 131, 781–804.
- Uppala, S. M., and Coauthors, 2005: The ERA-40 Re-Analysis. Quart. J. Roy. Meteor. Soc., 131, 2961–3012.
- Vecchi, G. A., and D. E. Harrison, 2002: Monsoon breaks and subseasonal sea surface temperature variability in the Bay of Bengal. J. Climate, 15, 1485–1493.
- Vranes, K., A. L. Gordon, and A. Ffield, 2002: The heat transport of the ITF and implications for the Indian Ocean heat budget. *Deep-Sea Res. II*, 49, 1391–1410.
- Wajsowicz, R. C., 1993: The circulation of the depth-integrated flow around an island with application to the Indonesian Throughflow. J. Phys. Oceanogr., 23, 1470–1484.
- Waliser, D. E., K. M. Lau, and J.-H. Kim, 1999: The influence of coupled sea surface temperatures on the Madden–Julian oscillation: A model perturbation experiment. *J. Atmos. Sci.*, 56, 333–358.

—, and Coauthors, 2009: MJO simulation diagnostics. J. Climate, 22, 3006–3030.

Walker, G. T., 1924: Correlations in seasonal variations of weather—A further study of world weather. World Weather II. *Mem. Indian Meteor. Dep.*, 24, 275–332.

- —, and E. W. Bliss, 1932: World weather V. Mem. Roy. Meteor. Soc., 4, 53–84.
- Wang, B., and H. Rui, 1990: Synoptic climatology of transient tropical intraseasonal convection anomalies: 1975-1985. *Meteor. Atmos. Phys.*, 44, 43–61.
- Webster, P. J., and S. Yang, 1992: Monsoon and ENSO: Selectively interactive systems. *Quart. J. Roy. Meteor. Soc.*, 118, 877–926.
- —, V. O. Magana, T. N. Palmer, J. Shukla, R. T. Tomas, M. Yanai, and T. Yasunari, 1998: Monsoons: Processes, predictability, and the prospects of prediction. *J. Geophys. Res.*, **103** (C7), 14 451–14 510.
- Wen, M., T. Li, R. Zhang, and Y. Qi, 2010: Structure and origin of the quasi-biweekly oscillation over the tropical Indian Ocean in boreal spring. J. Atmos. Sci., 67, 1965–1982.
- Wheeler, M. C., and J. L. McBride, 2011: Australasian monsoon. Intraseasonal Variability of the Atmosphere-Ocean Climate System, 2nd ed, W. K. M. Lau and D. E. Waliser, Eds, Springer, 147–198.
- —, H. H. Hendon, S. Cleland, H. Meinke, and A. Donald, 2009: Impacts of the Madden–Julian Oscillation on Australian rainfall and circulation. J. Climate, 22, 1482–1498.
- Wyrtki, K., 1961: Physical oceanography of the Southeast Asian waters. Scripps Institute of Oceanography NAGA Rep. 2, 195 pp.
- Xie, P., and P. Arkin, 1996: Analyses of global monthly precipitation using gauge observations, satellite estimates, and numerical model predictions. J. Climate, 9, 840–858.
- Xue, Y., T. M. Smith, and R. W. Reynolds, 2003: Interdecadal changes of 30-yr SST normals during 1871–2000. J. Climate, 16, 1601–1612.
- Yoo, S.-H., S. Yang, and C.-H. Ho, 2006: Variability of the Indian Ocean sea surface temperature and its impacts on Asian– Australian monsoon climate. J. Geophys. Res., 111, D03108, doi:10.1029/2005JD006001.
- Zhang, C., M. Dong, S. Gualdi, H. H. Hendon, E. D. Maloney, A. Marshall, K. R. Sperber, and W. Q. Wang, 2006: Simulations of the Madden–Julian oscillation in four pairs of coupled and uncoupled global models. *Climate Dyn.*, 27, 573–592.
- Zhou, L., R. Neale, M. Jochem, and R. Murtugudde, 2012: Improved Madden–Julian oscillations with improved physics: The impact of modified convection parameterizations. *J. Climate*, 25, 1116–1136.