The Asian Monsoon in the Superparameterized CCSM and Its Relationship to Tropical Wave Activity

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ABSTRACT

Three general circulation models (GCMs) are used to analyze the impacts of air–sea coupling and superparameterized (SP) convection on the Asian summer monsoon: Community Climate System Model (CCSM) (coupled, conventional convection), SP Community Atmosphere Model (SP-CAM) (uncoupled, SP convection), and SP-CCSM (coupled, SP). In SP-CCSM, coupling improves the basic-state climate relative to SP-CAM and reduces excessive tropical variability in SP-CAM. Adding SP improves tropical variability, the simulation of easterly zonal shear over the Indian and western Pacific Oceans, and increases negative sea surface temperature (SST) biases in that region.

SP-CCSM is the only model to reasonably simulate the eastward-, westward-, and northward-propagating components of the Asian monsoon. CCSM and SP-CCSM mimic the observed phasing of northward-propagating intraseasonal oscillation (NPISO), SST, precipitation, and surface stress anomalies, while SP-CAM is limited in this regard. SP-CCSM produces a variety of tropical waves with spectral characteristics similar to those in observations. Simulated equatorial Rossby (ER) and mixed Rossby–gravity (MRG) waves may lead to different simulations of the NPISO in each model. Each model exhibits some northward propagation for ER waves but only SP-CCSM produces northward-propagating MRG waves, as in observations. The combination of ER and MRG waves over the Indian Ocean influences the spatiotemporal structure of the NPISO and contributes to the differences seen in each model.

The role of ocean coupling must be considered in terms of the time scale of the SST response compared to the time scale of tropical variability. High-frequency disturbances experience coupling via its changes to the basic state, while lower-frequency disturbances may respond directly to SST fluctuations.

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1. Introduction

The Asian summer monsoon is a complex system of wind and precipitation anomalies that spans from the Indian Ocean to the western tropical Pacific Ocean. The summer monsoon season features enhanced rainfall expanding northward over the Indian subcontinent and Southeast Asia, beginning in late April to early May, and retreating equatorward in September–October. In the western Pacific Ocean, onset and retreat occur earlier and later, respectively. Subseasonal eastward-, northward-, and westward-propagating wind and precipitation anomalies produce “active” and “break” cycles with a period of 30–50 days over western India, the Bay of Bengal, and Southeast Asia (Yasunari 1979; Krishnamurti and Subrahmanyan 1982; Lawrence and Webster 2002). The slow (~5 m s\(^{-1}\)) eastward-propagating mode of the Asian monsoon is generally interpreted (Murakami et al. 1984; Madden 1986; Nakazawa 1986) as the boreal summer manifestation of the 30–60-day intraseasonal oscillation (ISO) described by Madden and Julian (1971, 1972, 1994) and Zhang (2005). Unlike the boreal winter ISO, the boreal summer ISO (BSISO) also exhibits northward propagation at a rate of ~1° latitude per day, as the northwest–southeast tilted ISO convective envelope translates eastward (Yasunari 1979; Lau and Chan 1986). Not all northward-propagating monsoon events are linked to eastward propagation, but various studies (Wang and Rui 1990; Lawrence and Webster 2002; Sperber and Annamalai 2008; Klingaman et al. 2008a) estimate the fraction to be around 50%–85%. Like the northern winter ISO, the BSISO is also characterized by 12–24-day westward-propagating equatorial Rossby waves (e.g., Lin et al. 2008 and references therein), which are associated with the NPISO. For the remainder of our discussion, NPISO refers to the northward-propagating component of the BSISO and the Madden–Julian oscillation (MJO) refers to the eastward-propagating component of the BSISO.

Multimodel analyses of the MJO in GCMs (Lin et al. 2006; Zhang et al. 2006) concluded that most models fail to produce significant spectral peaks at MJO spatial and temporal scales and do not propagate intraseasonal convection east of the Maritime Continent. Furthermore, simulated equatorially trapped Kelvin and Rossby waves associated with the MJO are often characterized by excessive equivalent depths, indicative of a lack of convective coupling.

Simulation of the Asian monsoon system is a particularly challenging test for GCMs. Kang et al. (2002) and Waliser et al. (2003) found that a collection of atmosphere-only GCMs mostly failed to produce the eastward-propagating convection and the signature northwest–southeast-tilted rainband that is characteristic of the mature phase of the BSISO. Lin et al. (2008) analyzed the BSISO in 14 ocean–atmosphere coupled GCMs participating in the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC) and found that most models 1) concentrate too much precipitation near the equator and 2) underestimate subseasonal (2–128 days) and 12–24-day westward-propagating variance. They also noted that most models produce at least some form of northward and westward propagation but lack eastward-propagating intraseasonal variability. Sperber and Annamalai (2008) found that most coupled models simulate eastward propagation over the Indian Ocean but fail to extend this mode into the western Pacific Ocean, missing an important part of the MJO life cycle. The missing extension into the western Pacific prevents the models from replicating the observed northwest–southeast tilted rainband structure, which arises from the eastward-moving BSISO and its associated northward-propagating Rossby waves.

MJO-linked NPISO events arise from equatorially trapped Rossby (ER) waves (Matsuno 1966; Wheeler and Kiladis 1999) excited by MJO convection (Wang and Xie 1997; Annamalai and Slingo 2001; Lawrence and Webster 2002), which are modified by the Indian Ocean boreal summer easterly zonal shear profile (Xie and Wang 1996; Lawrence and Webster 2002; Jiang et al. 2004; Drbohlav and Wang 2007) and surface latent and sensible heat fluxes (Kemball-Cook and Wang 2001; Sperber and Annamalai 2008) so that a northward component is introduced to their nominal westward propagation. While the mechanism for northward propagation is not yet fully understood, the combination of eastward-propagating MJO convection and associated northwestward-propagating ER waves produces a northwest–southeast tilted rainband that translates to the east. For an observer at a fixed point, the eastward passage of this tilted rainband appears as a northward-moving convective anomaly. These ideas have been described by Wang and Xie (1997) and Lawrence and Webster (2002) and are illustrated schematically in Fig. 1.

The NPISO is accompanied by coherent variations in SST and associated surface fluxes, with most studies concluding that atmospheric anomalies (such as cloudiness and surface winds) drive the surface anomalies (Hendon and Glick 1997; Krishnamurti et al. 1988; Shinoda et al. 1998; Woolnough et al. 2000; Vecchi and Harrison 2002), although modeling studies by Bellon et al. (2008) and Marshall et al. (2008) suggest that surface fluxes may modify the period of the disturbance. Evidence for the role of surface fluxes on the NPISO behavior is found in the improvement of NPISO simulation by GCMs driven with high-frequency specified SSTs (Klingaman et al. 2008a) or coupled to ocean
models (Fu et al. 2003, 2007; Fu and Wang 2004a; Wu and Kirtman 2005; Seo et al. 2007; Weng and Yu 2010). Nevertheless, many coupled models still struggle to produce the observed NPISO magnitude, phase relationship, and local versus nonlocal control of SST, precipitation, and surface flux anomalies (Bollasina and Nigam 2009).

In this study, we analyze the simulated Asian monsoon system in three GCMs using various combinations of conventional cumulus parameterization, “superparameterization,” and ocean–atmosphere coupling. Superparameterization (SP) refers to the replacement of parameterized convective tendencies with those simulated by a cloud-resolving model (CRM) that is run within each GCM grid column (Grabowski 2001). We examine the BSISO in three simulations: 1) the ocean–atmosphere coupled Community Climate System Model (CCSM) version 3 (Collins et al. 2006b,a) with conventional cumulus parameterization, 2) the superparameterized CCSM (SP-CCSM) (Stan et al. 2010), and the superparameterized Community Atmosphere Model (SP-CAM) (Khairoutdinov and Randall 2003), which is the atmosphere-only component of SP-CCSM. Our goals are to determine which monsoon-related processes are improved in SP-CCSM and whether SP and/or coupling are responsible for the improvement.

Section 2 describes the models and provides some background information on their simulated MJOs. Section 3 compares the Indo-Pacific climate and annual cycle of precipitation in models and observations. Monsoon variability and its relation to surface fluxes and equatorially trapped waves are described in section 4. A concluding discussion is given in section 5.

2. Model description and datasets

The implementation of SP into CAM is described by Khairoutdinov and Randall (2001, 2003) and results are presented by Khairoutdinov and Randall (2003), Khairoutdinov et al. (2005, 2008), and DeMott et al. (2007, 2010). The use of SP improves the mean climate
in some ways (diurnal cycle of rainfall, distribution of cirrus clouds) and degrades it in others (reduced marine stratocumulus clouds, high precipitation bias in the northwest Pacific Ocean during boreal summer). The most notable improvement resulting from the addition of SP is an increase in tropical variability, including the MJO.

Benedict and Randall (2009) reported that, compared to the standard CAM, SP-CAM better reproduces the observed space–time structures of the MJO and the vertical transfer of boundary layer heat and moisture to the upper atmosphere but that the MJO is too strong in SP-CAM. Thayer-Calder and Randall (2009) argued that the excessive MJO activity in SP-CAM might be due to overmoistening of the atmosphere due to excessive near-surface winds and latent heat fluxes.

Meehl et al. (2006) found that, while the CCSM improves several aspects of the monsoon simulation compared to the CAM, biases remain, some new ones are introduced, or some are overcorrected. A strong improvement is seen in the precipitation annual cycle.

Stan et al. (2010) performed the first simulation with an ocean–atmosphere coupled superparameterized GCM, adding SP to the CCSM. SP-CCSM exhibited improvements over CCSM, including the mean precipitation pattern, equatorial SST cold tongue structure and associated double intertropical convergence zone, periodicity of the El Niño–Southern Oscillation, the Asian monsoon, and the MJO. Tropical variability was improved without significant degradation of the basic-state atmosphere.

In this study, all simulations are performed using T42 resolution for the atmosphere with a semi-Lagrangian dynamical core (Rasch et al. 2006). The simulation with SP-CAM uses 1986–2003 observed monthly-mean SSTs interpolated to daily mean values as its boundary conditions. SP-CCSM and CCSM utilize a low-resolution 3° version of the Parallel Ocean Program (POP) ocean model (Smith and Gent 2002) initialized at rest with Levitus (1982) temperatures and salinities. Only the final 20 years of the 23-yr coupled simulations are analyzed.

Several observational and reanalysis datasets are used to evaluate the monsoon simulation in each model. All data are obtained at daily resolution (or averaged to daily resolution if finer temporal sampling is available) and regridded to the GCM grid resolution. National Centers for Environmental Prediction reanalysis (Kalnay et al. 1996) is the source of daily mean wind fields. Global-scale precipitation estimates are provided by the Global Precipitation Climatology Project One-Degree Daily (GPCP 1DD) dataset for 1998–2008 (Huffman et al. 2001). Outgoing longwave radiation (OLR) from 1986 to 2003 is obtained from NOAA (Liebmann and Smith 1996). Daily SST and surface wind stress estimates from 1998 to 2008 were taken from the European Centre for Medium-Range Weather Forecasts Interim Re-Analysis (ERA-Interim) dataset (Simmons et al. 2006).
3. Basic state and mean annual cycle

We begin our assessment of monsoon simulations with an evaluation of the Southeast Asian boreal summer climate simulated with the three models. June–September (JJAS) average precipitation and 850-hPa winds for observations and simulations are presented in Fig. 2. Precipitation maxima are observed (Fig. 2a) near the west coast of India and along the eastern shore of the Bay of Bengal (BoB). Western Pacific precipitation maxima are located over the Philippines and just north of the equator in the ITCZ. Maximum mean 850-hPa winds (magnitude \( \sim 15 \text{ m s}^{-1} \)) are observed at 10°–15°N over the Indian Ocean. The low-level westerlies decrease east of the Indo-China Peninsula, shifting to easterlies near 130°E.

In the CCSM (Fig. 2d), precipitation is affected by a double ITCZ bias, with anomalous zonally oriented precipitation bands in the Southern Hemisphere. Northern Hemisphere maxima are located too close to the equator and are too weak, while the western India maximum is largely absent. The CCSM 850-hPa winds are weaker than observed for the Indian Ocean and shift to easterly flow too far west, near 110°E. In SP-CAM (Fig. 2c), the precipitation pattern contrasts sharply with observations with a zonally oriented band of intense precipitation extending 65°–160°E. The Indian Ocean winds are too strong, and the westerly flow extends all the way to 155°E. In contrast, SP-CCSM (Fig. 2b) precipitation shows better agreement with observations, capturing the maxima over western India, BoB, and the Philippines. The Southern Hemisphere secondary maximum is too weak, while the Northern Hemisphere maxima remain somewhat greater than observed. The 850-hPa westerlies are slightly stronger than observed, but their zonal extent is well simulated.

![Image](image.png)

**Fig. 3.** June–September SST bias (K) for (a) SP-CCSM and (b) CCSM, contour interval 0.5 K with positive (negative) biases drawn with solid (dashed) lines. Observed climatology is from the HadISST dataset 1982–2001; rms errors for the region plotted are indicated.

**Fig. 4.** June–September mean zonal shear, defined as the difference in zonal winds at 200 and 850 hPa, for (a) NCEP reanalysis 1998–2008, (b) SP-CCSM years 4–23, (c) SP-CAM 1986–2003, and (d) CCSM years 4–23, contour interval 10 m s\(^{-1}\) with westerly positive (easterly negative) shear indicated with solid (dashed) lines: zero contour is denoted with thick solid line. Model vs observations pattern correlation and rms differences are indicated.
CCSM and SP-CCSM SST biases are shown in Fig. 3. Both models produce a small region of positive SST biases near the African coast. Otherwise, SP-CCSM Indian Ocean SSTs are too cool with a maximum negative bias of ~2.5 K occurring in the northern Arabian Sea. In CCSM, Indian Ocean SST biases are smaller than in SP-CCSM. The western Pacific Ocean is too cold in both models. The cold SST bias in SP-CCSM may be responsible for some of the reduction in mean rainfall compared to SP-CAM.

Figure 4 shows the JJAS zonal shear between 200 and 850 hPa for observations and for each model. Observed easterly shear during JJAS (Fig. 4a) extends zonally from central Africa eastward to near the date line and meridionally from 8°S to ~25°N. SP-CAM (Fig. 4c) simulates easterly shear that is too strong over the Indian Ocean and extending too far east north of the equator, consistent with the zonally extensive band of westerlies in Fig. 2c. SP-CCSM easterly shear (Fig. 4b) is weaker than observed but has realistic areal coverage. CCSM (Fig. 4d) easterly shear is weak and limited in areal coverage. For the region plotted, SP-CCSM easterly shear exhibits the most optimal combination of high spatial correlation and low rms errors compared to observations.

Figure 5 shows the mean annual cycle of precipitation as a function of latitude for the Indian and western Pacific Oceans. The latitudinal extent of the easterly shear is indicated by the black curves highlighting the relationship between the northward excursions of precipitation and shear. In the Indian Ocean, observed rainfall and easterly shear begin their northward advance around 1 May (Fig. 5a). The monsoon precipitation maximum remains at ~15°N through mid-July, while the oceanic convergence zone at ~5°S persists throughout the season (Bollasina and Nigam 2009). Each model (Figs. 5b–d) captures some, but not all, elements of the annual cycle in the Indian Ocean. SP-CCSM and CCSM show a delayed onset of Indian Ocean monsoon rains compared to observations. In SP-CCSM, the southern Indian Ocean precipitation band disappears in mid July (near pentad 43). This coincides with the emergence of the cold SST bias in that model, which may explain the lack of precipitation in this region. In SP-CAM (Fig. 5c), monsoon precipitation is too strong and too persistent. In the western Pacific Ocean (Figs. 5e–h), SP-CAM and SP-CCSM both successfully capture the observed abrupt onset (Wang et al. 2009) of monsoon precipitation, as seen in the near-vertical orientation of the 4 mm day
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contour at approximately pentad 27. In contrast, the northward extension of precipitation in CCSM occurs more gradually. In CCSM, the 850-hPa westerly winds typically do not extend far enough east (Fig. 2d), which may account for the delayed onset of easterly shear and northward-propagating precipitation in this region.

4. Monsoon variability

4.1 Overview

We begin our assessment of monsoon variability by examining the May–October longitudinal distribution of
variance for eastward- and westward-propagating precipitation (Fig. 6), and the latitudinal distribution of northward-propagating precipitation (Fig. 7). The plots are designed to be directly comparable to those shown in Lin et al. (2008), enabling comparison to other CGCMs. Filtering is 24–70 days and positive zonal wavenumbers 1–6 for eastward propagation, 12–24 days and all negative zonal wavenumbers for westward propagation, and 24–70 days and positive meridional wavenumbers for northward propagation [see Fu and Wang (2004a,b) for details of meridional filtering]. The GPCP and simulated zonal distributions of the eastward-propagating intraseasonal modes are shown in Fig. 6a. Observed variance for this mode maximizes in the eastern Indian Ocean and minimizes just east of the date line. CCSM-simulated MJO activity is weak, while SP-CCSM produces a realistic MJO with the magnitude and location of eastward-propagating precipitation in good agreement with observations. When compared to Fig. 9 from Lin et al. (2008), SP-CCSM simulates a more realistic boreal summer MJO than any of the CGCMs analyzed in that study. SP-CAM places too much variance in the western Pacific Ocean, consistent with the high precipitation anomaly in this region (Fig. 2c).

The zonal distribution of 12–24-day westward-propagating precipitation is shown in Fig. 6b. CCSM variance is again much weaker than observed. SP-CCSM variance exceeds the observed variance at nearly all longitudes, with the largest biases in the Indian Ocean. The three peaks in the SP-CCSM curve correspond to the seasonal precipitation maxima over the Bay of Bengal, the Philippines, and the western Pacific Ocean. SP-CAM westward-propagating precipitation variance maximizes at nearly the same longitudes and is also excessive compared to observations. The standard deviation of precipitation as a percentage of May–October mean precipitation over the same latitude bands is shown in (c) and (d).

Northward-propagating disturbances are extracted using the same method described by Lin et al. (2008); that is, the time–latitude array of precipitation is decomposed into meridional wavenumber and frequency
Fourier coefficients, and the inverse transform is then applied to positive (northward) wavenumber coefficients. In the Indian Ocean (Fig. 7a), observed northward propagation is uniformly distributed from the equator to ~17°N at which point it decreases monotonically to 35°N. In CCSM, northward propagation is weak and peaks at ~5°S. Northern Hemisphere SP-CCSM northward propagation is too strong, similar to the westward propagation results. SP-CAM variability is again excessive with peak variance shifted north of the SP-CCSM variance. Results for the western Pacific Ocean (Fig. 7b) are similar to those for the Indian Ocean, except that SP-CCSM variance is not as excessive and SP-CAM variance is even more excessive. The scaled standard deviations (Figs. 7c and 7d) indicate that SP-CCSM northward variability is slightly low compared to the model’s mean-state precipitation, while SP-CAM variability is excessive, even for its greater mean-state precipitation.

Individual Indian Ocean northward-propagating events in observations and models are shown in Fig. 8. In observations, NPISO events tend to originate near the equator and then propagate northward and, to some extent, southward. Northward propagation is nearly continuous, with few interruptions in the precipitation front. CCSM NPISO events are weak, with northward propagation accomplished by discrete events that propagate poleward for less than ~10° before weakening. SP-CCSM NPISO events are similar to those observed, with meridionally elongated precipitation bands. SP-CAM produces roughly equal numbers of both continuous and discrete NPISO events.

In Fig. 9, the spatiotemporal behavior of “typical” NPISO and eastward-propagating MJO events is illustrated through the use of lag correlation analysis. The figures, based on those of the Climate Variability and Predictability (CLIVAR) Research Program MJO Working Group (CLIVAR Madden–Julian Oscillation Working Group 2009), are constructed by correlating 20–100-day filtered winds or precipitation at each lag and longitude or latitude with a base time series of 20–100-day filtered precipitation averaged over the equatorial Indian Ocean (10°S–5°N, 70°E–100°E). Use of equatorial 20–100-day precipitation as the base time series implicitly selects NPISO events that are linked to the MJO. In observations, the MJO appears first in the western Indian Ocean at lag ~18 (Fig. 9a), propagates eastward across the Maritime Continent and into the
western Pacific Ocean, and decays near the date line at lag 23. Low-level easterly (westerly) winds lead (lag) precipitation by one quarter cycle. The MJO in CCSM is confined to the Indian Ocean and fails to propagate across the Maritime Continent. SP-CCSM produces the most realistic MJO of the three models but with lower correlations east of the Maritime Continent and slightly faster phase speed compared to observations. SP-CAM (Fig. 9d) also propagates convection into the western Pacific Ocean, but it decays more rapidly than in SP-CCSM.

The quadrature relationship between rainfall and low-level winds is also observed in the Indian Ocean NPISO (Fig. 9e), with easterly winds preceding NPISO events. Precipitation starts near the equator and then propagates northward and southward. Southward propagation is faster, weaker, and of shorter duration than northward propagation. SP-CCSM (Fig. 9f) produces lag correlations slightly lower than those in observations but with realistic northward and southward phase speeds. In contrast, CCSM (Fig. 9h) weakly simulates northward, but not southward, propagation. SP-CAM (Fig. 9g) simulates some of the southward-propagating mode and a realistic temporal duration for the northward mode but with lower correlations than observed, suggesting that a larger fraction of variability north of ~20°N in SP-CAM results from disturbances not associated with the NPISO. The visual similarity of lag correlations between SP-CCSM and observations suggests that SP-CCSM captures many facets of the MJO/NPISO that both CCSM and SP-CAM fail to simulate.

Spatial composites of BSISO OLR give further insight into the ability of the models to simulate the BSISO. Figure 10 (also based on the CLIVAR MJO diagnostics package) presents May–October composite OLR anomalies for observations and models based on the combined EOF method of Wheeler and Hendon (2004). In observations (Fig. 10a), convection begins in the Indian Ocean (phases 1 and 2) and expands and/or propagates eastward in phases 3–5, corresponding to the eastward BSISO propagation across the Maritime Continent and the northward propagation of convection onto the Indian subcontinent, resulting in the northwest–southeast tilted rainband (seen in phases 3–7). Eastward translation continues over the South China Sea in phases 6 and 7, leading to the northward shift of precipitation over Southeast Asia, while the western Pacific Ocean is characterized by in situ decay in phases 6–8. SP-CAM and SP-CCSM (Figs. 10c and 10b, respectively) capture many of these elements in phases 1–4. Both models propagate convection across the Maritime Continent. In phases 5–7, however, SP-CAM convection is concentrated north of the equator so that it becomes zonally oriented rather than tilted. SP-CCSM,
on the other hand, reproduces the tilted rainband structure in phases 3–6. CCSM (Fig. 10d) does produce some BSISO activity, especially in the Indian Ocean, but it is weak and does not propagate into the western Pacific Ocean.

The results presented in this subsection demonstrate that SP-CCSM simulates a more realistic Asian monsoon than either SP-CAM or CCSM. The addition of SP improves the monsoon in coupled simulations, and coupling improves the monsoon in SP simulations. We now turn our attention to the relative contributions of SP and coupling to the improved monsoon simulation.

b. NPISO and surface fluxes

Figure 11 illustrates the essential phase relationships (as depicted by the +0.2 correlation contour) of NPISO precipitation, SST, and surface stress anomalies in observations and the three simulations. Both land and ocean points are used for the precipitation correlation, but only ocean points are used for the surface temperature and surface stress contours. Surface wind stress is highly correlated with latent heat flux, but is more directly observable than the latter and thus provides a more direct comparison to simulations. As with Fig. 9, all variables were correlated with equatorial Indian Ocean precipitation anomalies. In observations (Fig. 11a), a coherent relationship is seen between precipitation, SST, and Northern Hemisphere surface stress. Warm SST anomalies lead northward-propagating convection by \( \sim 10 \) days. Maximum surface stress is associated with westerly winds (Fig. 9a) and maximizes near the equator and \( \sim 15^\circ \)N, lagging precipitation by approximately five days.

Not surprisingly, the surface flux relationships in SP-CAM (Fig. 11c) bear only limited resemblance to the observations. The specified SSTs in this model are based on monthly-mean data interpolated to daily resolution. Northern Hemisphere surface stress correlations are weak, but exhibit positive correlations over a broad region ahead of southward-propagating precipitation in the Southern Hemisphere. Ocean coupling in SP-CCSM (Fig. 11b) improves the phasing and northward extent of

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**FIG. 9.** (a)–(d) Longitude vs lag correlation and (e)–(h) latitude vs lag correlation of 20–100-day filtered Indian Ocean precipitation (10°S–5°N, 75°–100°E) with 20–100-day filtered precipitation (shaded) and 850-hPa zonal wind (contour) at each longitude–lag or latitude–lag pair. Data are averaged from 10°S to 10°N for (a)–(d) and from 80° to 100°E for (e)–(h). Positive (negative) correlations are indicated by red (blue) shading or solid (dashed) contours. Contour interval is 0.1.
NPISO precipitation, SST, and surface stress anomalies. As observed in the Northern Hemisphere, SP-CCSM warm SSTs lead precipitation by about 7 days, which leads positive surface stress anomalies by 7–10 days. Simulated correlations are greater in the Southern Hemisphere than in observations, with surface stress leading southward-propagating convection by a few days. The relationship of precipitation, SST, and surface stress anomalies in CCSM (Fig. 11d) is similar to that simulated by SP-CCSM (including the Southern Hemisphere extension of the correlation curves), indicating that ocean coupling, rather than SP, controls the temporal relationships among these variables. SP clearly improves the realism of the NPISO simulation (represented by the precipitation contour in these plots), but it appears to have little influence on the interplay of SST and surface stress anomalies as they relate to the NPISO. Because the phasing of the surface fluxes are similar in CCSM and SP-CCSM, improvements to the NPISO in SP-CCSM may be more directly related to internal atmospheric dynamics, rather than surface fluxes, on these time scales.

c. NPISO and internal atmospheric dynamics

Here we assess the relationship between the NPISO and equatorial waves. The westward-propagating $n = 1$ equatorial Rossby wave is frequently studied in connection with the NPISO. We also examine the westward-propagating mixed Rossby–gravity (MRG) wave because it is of similar spatial scale to the ER wave and has been linked to NPISO activity by Straub and Kiladis (2003), who found that 1) MRG group velocity accounted for a large fraction of MJO OLR variance and 2) individual MRG waves exhibited northwestward propagation, similar to ER waves. Takayabu and Nitta (1993), Dunkerton and Baldwin (1995), and Yang et al. (2007b, c) described the tendency of westward-propagating waves to transition from MRG to ER waves, or to exhibit hybrid MRG–ER structures over the western Pacific.

Figure 12 shows the distribution of equatorially symmetric OLR signal-to-noise ratio in wavenumber–frequency space for observations and models. In observations (Fig. 12a) significant peaks correspond to the MJO (wavenumber $+1$ and periods ~50 days), the Kelvin wave, the $n = 1$ ER wave, and the higher-frequency $n = 1$ inertia–gravity (IG) wave. The peak observed near wavenumber 14 and frequency 0.1 is an artifact of combining OLR observations from multiple geostationary satellites and is not physical (Wheeler and Kiladis 1999). The simulation of equatorial waves improves as one moves from CCSM (Fig. 12d) to SP-CAM (Fig. 12c) to SP-CCSM (Fig. 12b). In CCSM, the MJO peak is poorly defined and appears at a lower frequency than in observations. Typical of many coupled models (Lin et al. 2006),
Kelvin wave activity lies to the left of the $h = 50$ m equivalent depth dispersion curve, scaling to a deeper equivalent depth ($h \sim 90$ m) compared to observations. The $n = 1$ ER wave is reasonably well simulated by CCSM. SP-CAM (Fig. 12c) is an improvement over CCSM in its ability to produce a spectral peak in the ISO region, although the peak is not clearly separated from the Kelvin wave. The SP-CAM Kelvin wave convection is aligned with the $h = 25$ m dispersion curve, as in observations. SP-CCSM (Fig. 12b) produces spectral peaks similar to those observed. The MJO peak is well separated from the Kelvin wave, the Kelvin wave peak is comparable to that seen in observations, and the $n = 1$ ER wave peak is present.

The distribution of equatorially antisymmetric OLR spectral peaks is shown in Fig. 13. In observations (Fig. 13a) a spectral peak corresponds to the $h = 25$ m equivalent depth dispersion curve for the MRG and eastward inertia–gravity (EIG) waves. A weaker peak is also observed near wavenumber $-5$ and frequency $\sim 0.25$ cycles per day, associated with tropical depressions (Wheeler and Kiladis 1999; Straub and Kiladis 2003). The artificial “data merging” peak near wavenumber 14 is also present. The simulation of antisymmetric waves again improves as one moves from CCSM (Fig. 13d) to SP-CAM (Fig. 13c) to SP-CCSM (Fig. 13b). In CCSM, few signals rise above the noise threshold (the signal-to-noise ratio is less than 1). In SP-CAM, spectral peaks begin to emerge near the dispersion curves, but they are weak and disorganized. SP-CCSM simulates MRG/EIG waves that stand out above the background. Compared to the 14 coupled models analyzed by Lin et al. 2006, the SP-CCSM is unique in its ability to simulate MRG waves with the same dispersion characteristics seen in observations. The clear presence of MRG waves in SP-CCSM, but not in CCSM or SP-CAM, suggests that coupling may modify the basic-state environment so that SP-generated variability produces MRG waves. The role of coupling in improving tropical variability is discussed in section 5.

We examined the spatial distributions and magnitudes of ER and MRG wave power for observations and models, following the extraction method described in Wheeler and Kiladis (1999) but filtering the full (symmetric + antisymmetric) time series so as to capture the effects of landmasses and seasonality. ER waves were extracted
from the full time series by applying the reverse Fourier transform to coefficients contained in a wavenumber–frequency region bounded by the 8-m and 90-m equivalent depth curves, westward (negative) wavenumbers 1–10, and periods 10–35 days. MRG waves were extracted using the same equivalent depth and wavenumber boundaries but for periods 3–8 days.

Observed and simulated May–October ER wave OLR variances are plotted in Fig. 14. In the observations (Fig. 14a), ER variance is greatest in the Northern Hemisphere, consistent with the northward shift of the ITCZ during this season. ER variance maximizes in the western Pacific Ocean, just east of the Philippines, with two ridges of enhanced variability at 5°N and 15°N, extending toward the Indian Ocean. ER variance patterns are similar in SP-CCSM and SP-CAM (Figs. 14b and 14d) but with excessive magnitudes. CCSM ER variance (Fig. 14c) differs from observations and the two SP models, with maximum variance located near the date line and no variance ridge in the Indian Ocean. The increase of Indian Ocean ER variance with latitude in CCSM may indicate a midlatitude source of tropical ER wave activity in that model.

Figure 15 shows observed and simulated May–October MRG wave OLR variance. For all models, the spatial distribution of simulated MRG variance

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**FIG. 12.** Wavenumber vs frequency distribution of spectral-power-divided estimate background spectra for equatorially symmetric OLR anomalies. Signal-to-noise ratios (SNRs) greater than (less than) one are shaded in yellows and reds (blues). Contour interval is 0.1. Shallow water dispersion relationships for equivalent depths of $h = 12, 25,$ and $50$ m are shown for the $n = 1$ equatorial Rossby, Kelvin, and $n = 1$ inertio–gravity waves.
agrees well with observations although, as mentioned earlier, only SP-CCSM produces statistically significant MRG spectral peaks. Observed Northern Hemisphere MRG variance (Fig. 15a) is centered at −10°N with maximum variance spanning the Indian and Pacific Oceans. The longitudinal extent of MRG variance in SP-CCSM (Fig. 15b) is similar to observations, although the maximum is located near the Maritime Continent rather than the date line. Both CCSM and SP-CAM (Figs. 15d and 15c, respectively) appear to concentrate MRG variance over a smaller longitudinal range.

Because so many coupled models fail to simulate realistic NPISO events (Sperber and Annamalai 2008), while SP-CCSM succeeds, we ask whether the northward propagation in SP-CCSM results from the same mechanisms as in nature, particularly with regard to ER and MRG wave activity. The distribution of ER and MRG variance with latitude over the Indian Ocean (60°–100°E) for observations and models is shown in Fig. 16. The excessive Northern Hemisphere ER variance in SP-CCSM and SP-CAM is apparent, although south of −10°N the biases are smaller in SP-CCSM than in SP-CAM. SP-CCSM MRG variance agrees well with observations, while SP-CAM and CCSM MRG variances do not. In all cases, however, Northern Hemisphere MRG wave power ranges from 25% to 80% of

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**Fig. 13.** As in Fig. 12 but for equatorially antisymmetric OLR anomalies. Shallow water dispersion relationships for equivalent depths of $h = 12$, 25, and 50 m are shown for the mixed Rossby–gravity (MRG) and $n = 0$ eastward inertio–gravity (EIG) waves.
ER power, with an average ratio of ~50%, so MRG wave activity is nonnegligible compared to ER wave activity.

Next we address how simulated ER and MRG waves contribute to the Indian Ocean NPISO. Lag-correlation plots of the observed and simulated latitudinal and longitudinal propagation of ER waves at 5°N and 15°N and MRG waves at 10°N are shown in Figs. 17 and 18, respectively. We also examine the propagation of the MRG wave envelope, described by the 9-day running variance of MRG OLR anomalies. Unlike the results shown in Fig. 9, there is no implied link to the MJO in Fig. 17 because each wave type defines its own base time series. Zonal propagation of ER and MRG waves is primarily determined by filtering constraints, whereas meridional propagation of ER and MRG waves is not. Therefore, differences in lag–latitude correlations are attributable to model differences.

In observations (Figs. 17a and 17b) and SP-CAM (Figs. 17i and 17j), ER waves propagate smoothly from the equator to 20°N. In SP-CCSM and CCSM, ER waves (Figs. 17e and 17f and 17m and 17n, respectively) exhibit less northward propagation. In observations, high-frequency MRG waves appear in groups with individual waves propagating northward (Fig. 17c). This behavior is well simulated in SP-CCSM (Fig. 17g), but not in SP-CAM or CCSM (Figs. 17k and 17o, respectively), where MRG waves propagate due west or slightly southward. The observed MRG wave envelope (Fig. 17d) is nearly fixed at 10°N, a feature also seen in SP-CAM and CCSM (Figs. 17l and 17p). In contrast, the MRG wave envelope in SP-CCSM exhibits some northward drift. ER and MRG waves also propagate zonally (Fig. 18), as dictated by their wave-number filtering limits. Zonal propagation of ER waves is especially well maintained in the SP models (i.e., the two left-hand columns of Fig. 18). All three models also reflect the eastward MRG group velocity (right-hand column of Fig. 18), given by \( c_g = \delta \omega / \delta k \), where \( c_g \) is group velocity, \( \omega \) is frequency, and \( k \) is zonal wave-number.

To understand how propagation characteristics of ER and MRG waves relate to the NPISO, composite wave trajectories were computed, with the aid of Figs. 17 and 18, by plotting the latitude–longitude pair of maximum correlation for each wave and lag (Fig. 19). For each wave type, lag = 0 is the day at which OLR minimizes in the 70°–80°E longitude band. In observations (Fig. 19a), both ER and individual MRG waves propagate to the northwest. ER waves that propagate across the eastern Indian Ocean originate near the equator, presumably in association with eastward-moving MJO convection and then propagate to the northwest. The northern ER wave originates farther east than the southern ER wave and
takes longer to reach western India than the southern ER wave, consistent with the idea of the MJO acting as the ER source. MRG waves, whose convective signal maximizes at \(-10^\circ\)N, also propagate to the NW, reinforcing the slow northward propagation over western India associated with the northwestward-propagating ER wave train.

Each of the models (Figs. 19b–d) simulates ER and MRG wave trajectories that differ somewhat from those seen in observations. In both SP-CAM and SP-CCSM, ER waves originate near the equator, as in observations, and propagate to the northwest, especially in SP-CAM. In CCSM, ER waves exhibit more westward than northward propagation with the southern ER wave propagating to the southwest. For MRG waves, only SP-CCSM simulates the observed northwestward propagation of individual waves; SP-CAM and CCSM produce westward- or southwestward-propagating MRG waves.

The relationship of ER and MRG waves to the slow northward drift of NPISO convection over western India may be interpreted (with reference to Fig. 8) as follows: In observations, northwestward-propagating MRG waves reinforce the northward propagation arising from ER waves. In SP-CCSM, the reduced northward propagation associated with ER waves is apparently compensated for by MRG waves, both from individual waves and their northeastward group velocity (Fig. 17h). In SP-CAM, realistic northwestward propagation of ER waves is somewhat disrupted by westward- or southwestward-propagating MRG waves. In CCSM, the unrealistic southward-propagating ER waves at low latitudes and northward-propagating ER waves at higher latitudes may be sufficient to explain the poor simulation of the NPISO over western India (Fig. 8d). However, westward- or southwestward-propagating MRG waves may also account for some of CCSM’s errors in the NPISO.

Several recent studies have focused on mechanisms responsible for the northward propagation of ER waves. Lawrence and Webster (2002) analyzed the NPISO in NOAA OLR and NCEP reanalysis winds. They concluded that surface frictional convergence into the low pressure center Rossby waves accounted for the northward movement of monsoon precipitation, with oceanic surface fluxes playing a supporting role. Several modeling studies (Jiang et al. 2004; Wu et al. 2006; Drbohlav and Wang 2007; Seo et al. 2007; Boos and Kuang 2010) support the hypothesis that internal atmospheric dynamics—particularly the interactions of waves with easterly shear—are primarily responsible for the northward-propagating monsoonal rainfall, with surface fluxes playing a secondary role. Other modeling studies, however, suggest that high-frequency SST variability and their associated surface fluxes play a crucial, rather than secondary, role in the NPISO (Klingaman et al. 2008a; Wang et al. 2009; Weng and Yu 2010).
5. Discussion and conclusions

We have attempted to understand the improved simulation of the Asian monsoon in SP-CCSM in terms of how cloud representation and ocean coupling influence three factors important to the monsoon: the atmospheric mean state, atmospheric variability and internal dynamics, and ocean–atmosphere interactions. In terms of the seasonal-mean climatology, SP-CCSM produces a more realistic distribution of Northern Hemisphere summer rainfall and easterly shear than CCSM, with Indian Ocean (western Pacific Ocean) SSTs slightly cooler (warmer) than CCSM. Realistic simulation of zonal shear is especially critical for the monsoon because it has been hypothesized to provide a mechanism for the northward propagation of precipitating systems. However, successful simulation of this basic-state variable is not sufficient to produce a realistic monsoon. This is seen in CCSM, which, despite its realistic Indian Ocean mean shear profile, fails to produce the NPISO.

Subseasonal tropical variability is critical to the Asian monsoon, which is frequently linked to the eastward-propagating 20–100-day MJO and the associated northwest-propagating equatorial Rossby waves. A model that struggles to produce the MJO may also struggle to produce the Rossby waves associated with the NPISO. A model that produces some semblance of the MJO, but fails to propagate the MJO across the Maritime Continent into the western Pacific Ocean, may struggle to propagate the NPISO as far north as it does in the real world because the Pacific source region of the northernmost ER waves would not be simulated.

SP-CCSM simulates the eastward-, westward-, and northward-propagating components of the Asian monsoon, whereas they are either missing or very weak in CCSM. We ask How does the explicit representation of convection help simulate the observed modes of variability seen in the Asian monsoon? Thayer-Calder and Randall (2009) and Benedict and D. A. Randall (2011) demonstrated the ability of SP-CAM and SP-CAM coupled to a slab ocean model, respectively, to replicate the observed gradual deepening with height of low-level humidity leading up to MJO convective events. DeMott et al. (2007) described the differences in convective behavior in SP-CAM and the conventionally parameterized CAM, noting: 1) frequent and light rain rates in CAM versus a more realistic broad rain rate distribution in SP-CAM and 2) SP-CAM’s gradual deepening of

![Figure 16](https://example.com/figure16.png)

**Fig. 16.** Indian Ocean (60°–100°E) ER (thin solid) and MRG (dashed) OLR variance and the MRG/ER ratio (thick solid) as a function of latitude for (a) observations, (b) SP-CCSM, (c) SP-CAM, and (d) CCSM.
low-level moisture in advance of major precipitation events versus the CAM tendency to simultaneously moisten the upper and lower troposphere coincident with maximum precipitation. In the presence of positive CAPE, CAM with parameterized convection generates convective plumes that are required to detrain above the level of minimum moist static energy (Zhang and McFarlane 1995), thereby eliminating “preconditioning effects” of detrainment and moistening at lower levels (Thayer-Calder and Randall 2009). On the other hand, detrainment of SP convection occurs whenever convective elements reach their level of neutral buoyancy. Entraining convection in a dry environment quickly loses buoyancy, resulting in detrainment and reevaporation of cloud liquid water at low levels. This moistening by reevaporation is favorable for later deeper convection. Kuang and Bretherton (2006) and Waite and Khouider (2010) have demonstrated the ability of two different CRMs to simulate this sequence of events.

Fig. 17. Latitude vs lag autocorrelation of eastern Indian Ocean (70°–80°E) ER and MRG OLR anomalies, and 9-day running variance of MRG OLR anomalies. Base time series are computed as (first column) ER OLR 2.5°–7.5°N, (second column) ER OLR 12.5°–17.5°N, (third column) MRG OLR 7.5°–12.5°N, and (fourth column) MRG OLR variance 7.5°–12.5°N. Positive (negative) correlations are dark (light) shaded with solid (dashed) contours. Contour interval is 0.1, zero contour omitted. Averaging range is indicated with heavy horizontal line at lag = 0. Latitude of Northern Hemisphere maximum positive correlation for each lag is indicated with small diamond.

This conceptual model of SP convection is hypothesized to explain the role of SP in the simulation of convectively coupled MRG waves and other tropical modes not explicitly discussed here. Wheeler and Kiladis (1999) discuss the reduction of the effective static stability by convective heating as a mechanism favoring slower propagation. As discussed by Kiladis et al. (2009), with the exception of ERs, heating and moistening profiles in convectively coupled equatorial waves (CCEWs) of all scales (easterly and westerly inertia–gravity waves, Kelvin waves, MRG waves) are observed to evolve in a manner...
consistent with the notion of shallow convection preconditioning the tropical atmosphere for subsequent deeper convection (Johnson et al. 1999; Straub and Kiladis 2003, Khouider and Majda 2008), a sequence that is simulated by SP-CAM and SP-CCSM but not by CCSM. Further study is needed to determine if SP improves the convection–wave coupling via proposed mechanisms such as moisture–stratiform instability (Mapes 2000; Kuang 2008) or interactions with gross moist stability (Raymond and Fuchs 2007; Frierson 2007).

The role of SP in improving ER waves is more difficult to identify. Over most of the tropics, ER waves exhibit a deep unimodal structure in their temperature and wind profiles (Kiladis and Wheeler 1995), but in the convectively active warm pool region they acquire a highly baroclinic structure (Wheeler et al. 2000; Yang et al. 2007a). Such changes may arise from convective coupling but also from a combination of the slow propagation speed of these waves and the presence of strong easterly shear (e.g., Wang and Xie 1996; Sperber and Annamalai 2008).

Tropospheric moistening in ERs differs from other CCEWs. Rather than exhibiting a gradually deepening layer of moisture from the surface, ER moistening occurs throughout the entire depth of the troposphere. Deep tropospheric moistening begins about four days ahead of the minimum ER OLR (Fig. 18d of Kiladis et al. 2009). The unusual CCEW moistening sequence resembles that simulated by SP-CAM in the monsoon region. This may help to explain why ER variance is so high in SP-CAM compared to observations and the coupled models. It has been hypothesized (Luo and Stephens 2006; Benedict and Randall 2009) that the cyclic boundary condition on the CRM in SP-CAM can lead to a positive feedback.
between convection and column moisture in regions of warm SSTs. SP-CCSM may be less vulnerable to this problem because convectively driven surface winds and reduced surface shortwave fluxes cool the SST, potentially reducing the vigor of subsequent convection (Wu and Kirtman 2005). A more detailed study of moistening associated with simulated ER waves, as well as their interaction with environmental shear and surface fluxes, is needed to better understand their interactions with the simulated monsoon.

The role of ocean–atmosphere feedbacks on the monsoon simulation must be considered over a range of temporal scales. On one hand, we have demonstrated that the atmosphere-only model (SP-CAM) produces a different basic state from the coupled model (SP-CCSM). Seasonally, the atmosphere and ocean surface temperature evolve together via seasonal solar flux variations and ocean dynamics (Webster et al. 1998) so that the ocean directly affects the seasonal mean state of the atmosphere. On the other hand, for a high-frequency wave such as the MRG (period ~5 days), even intra-seasonal SST fluctuations are “slow” and unlikely to directly impact wave development. However, high-frequency waves may respond to SST variations indirectly via their interactions with the basic-state climate, which is itself affected by coupling with the ocean. We hypothesize that ocean coupling in SP-CCSM leads to a basic-state climate that is favorable to MRG waves and their northward propagation but that the presence of those waves is attributed to the enhanced convective variability of SP. ER waves and the MJO represent an intermediate time scale on which ocean–atmosphere feedbacks may be experienced both directly as the disturbances evolve and indirectly as they interact with the basic-state atmosphere.

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