Performance of the New NCAR CAM3.5 in East Asian Summer Monsoon Simulations: Sensitivity to Modifications of the Convection Scheme

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ABSTRACT

The performance of an interim version of the NCAR Community Atmospheric Model (CAM3.5) in simulating the East Asian summer monsoon (EASM) is assessed by comparing model results against observations and reanalyses. Both the climate mean states and seasonal cycle of major EASM components are evaluated. Special attention is paid to the sensitivity of model performance to changes in the convection scheme. This is done by analyzing four CAM3.5 runs with identical dynamical core and physical packages but different modifications to their convection scheme, that is, the original Zhang–McFarlane (ZM) scheme, Neale et al.'s modification (NZM), Wu et al.'s modification (WZM), and Zhang's modification (ZZM). The results show that CAM3.5 can capture the major climate mean states and seasonal features of the EASM circulation system, including reasonable simulations of the Tibetan high in the upper troposphere and the western Pacific subtropical high (WPSH) in the middle and lower troposphere. The main deficiencies are found in monsoon rainfall and the meridional monsoon cell. The weak meridional land–sea thermal contrasts in the model contribute to the weaker monsoon circulation and to insufficient rainfall in both tropical and subtropical regions of EASM. The seasonal migration of rainfall, as well as the northward jump of the WPSH from late spring to summer, is reasonably simulated, except that the northward jump of the monsoon rain belt still needs improvement. Three runs using modified schemes generally improve the model performance in EASM simulation compared to the control run. The monsoon rainfall distribution and its seasonal variation are sensitive to modifications of the ZM convection scheme, which is most likely due to differences in closure assumptions. NZM, which uses a convective available potential energy (CAPE)-based closure assumption,
1. Introduction

The Asian summer monsoon plays an important role in global climate variability (e.g., Ding 1994; Webster et al. 1998; Wang et al. 2001). Numerous studies have documented that the expansive Asian summer monsoon system can be divided into two subsystems: the Indian summer monsoon (ISM) and the East Asian summer monsoon (EASM) systems. These two subsystems are to a large extent independent of each other, but at the same time they interact with each other (Zhu 1934; Yeh et al. 1959; Zhu et al. 1986; Tao and Chen 1987). The EASM is not an eastward extension of the ISM. Lying downstream of the Tibetan Plateau and in between the Eurasian continent and the Pacific Ocean, the EASM has its own unique features. The ISM is a tropical monsoon in which the low-level winds reverse from winter easterlies to summer westerlies, whereas the EASM is a hybrid type of tropical and subtropical monsoon (Zhu et al. 1986; Tao and Chen 1987; Chen et al. 1991; Ding 1994). The generally recognized EASM system consists of three main components, namely, the East Asian mei-yu/baiu/changma front (a major rain-bearing system in the subtropics and midlatitudes), the western Pacific subtropical high (WPSH), and the tropical western Pacific monsoon trough or the western Pacific intertropical convergence zone (ITCZ). These three components are intimately coupled (Nitta 1987; Huang and Wu 1989; Liu et al. 2008).

Simulations of the Asian summer monsoon and its variability are challenging issues (Kang et al. 2002; Wang et al. 2005). Many general circulation model (GCM) results have shown that the basic distribution of monsoon circulation, such as the subtropical high, the equatorial jets, and the monsoon low, etc., can be reproduced reasonably, but their strengths and variations are difficult to simulate (e.g., Hoskins and Rodwell 1995; Zhou and Li 2002). The monsoon rainfall is even more difficult to simulate (Kang et al. 2002). A wide skill range exists among the GCMs in simulating monsoon precipitation, which is largely attributed to the different subgrid-scale parameterization schemes and horizontal resolutions (Sperber et al. 1994). Great efforts have been devoted to the simulations of the Indian monsoon (e.g., Yang et al. 1996; Meehl and Arblaster 1998; Loschnigg et al. 2003; Meehl et al. 2006), but less attention has been paid to the East Asian monsoon. Because of the complex topography and the sharp land–sea thermal contrast, the EASM simulation provides a rigorous criterion to test model physics (Liang and Wang 1998; Zhou and Li 2002).

The Community Atmosphere Model (CAM) has been widely used in climate research (e.g., Ghan et al. 1996; Zhang 2003; Collier and Bowman 2004; Zhou and Yu 2004; Deser and Phillips 2006; Hack et al. 2006). The model has also been used for monsoon studies (e.g., Hoerling et al. 1990; Yu et al. 2000; Liu et al. 2002; Zhou et al. 2008; Wu and Zhou 2008; Zhou et al. 2009a). Meehl and Arblaster (1998) compared the aspects of the Asian–Australian monsoon system associated with El Niño–Southern Oscillation in the global coupled Climate System Model (CSM) and its atmospheric component, the Community Climate Model, version 3 (CCM3). They showed that the CSM captured most major features of the monsoon system in terms of mean climatology, interannual variability, and connections to the tropical Pacific, with the largest discrepancies between the CSM, observations, and the CCM3 simulation in the equatorial eastern Indian Ocean and near the Philippines. Meehl et al. (2006) also examined the simulations of regional monsoon regimes in CAM, version 3 (CAM3) and its coupled Community Climate System Model, version 3 (CCSM3). They noted that the major monsoon features were well represented in all simulations, but the details of the regional simulations were model dependent. Some aspects of the monsoon simulations, particularly in Asia, were improved in the coupled model compared to the SST-forced simulations. Nevertheless, the above analyses concentrated on the tropical monsoon, that is, the South Asian monsoon, and less attention was given to the EASM, especially its subtropical components. A comprehensive evaluation of the performance of CAM3.5 in EASM simulations will potentially reveal strengths and weaknesses of the model and serve as a useful reference for future model improvement.

As an important energy source for atmospheric motion, cumulus convection affects large-scale circulation and wave disturbances through the release of latent heat of condensation and the vertical transport of heat, moisture, and momentum. The large-scale forcing in turn influences and modulates the development and organization of convection and clouds (Wu et al. 2007a,b). The coupling of convective processes with the large-scale dynamics is crucial for modeling the global distribution of precipitation (Zhang 2005). The choice of different cumulus parameterization schemes has a significant influence on...
the simulations of both the climate and synoptic systems (e.g., Hack 1994; Zhang and McFarlane 1995; Maloney and Hartmann 2001; Zhang 2002; Kang and Hong 2008). Zhang (1994) found that changes in the cumulus parameterization in a general circulation model from the Canadian Climate Centre (CCC) have the largest effects in the East Asian and the Indian monsoon areas, because these are regions with the most active convection on the globe. Huang et al. (2001) examined the sensitivity of EASM rainfall and circulation to convection schemes in a five-level spectral AGCM and found that the Kuo scheme (Kuo 1965) generally showed the best performance. However, Singh et al. (2006) indicated that the Arakawa and Schubert (Arakawa and Schubert 1974) and Emanuel (Emanuel 1991) schemes simulate the summer precipitation and its variation better than the Kuo scheme over Korea in a regional climate model.

Recently, with the motivation of improving the performance of CAM, version 3.5 (CAM3.5), the Zhang and McFarlane (1995) convection scheme (hereafter ZM) was modified. There are three modifications: the Neale et al. (2008) convection scheme (hereafter NZM), the Zhang (2002) convection scheme (hereafter ZZM), and the Wu et al. (2003) convection scheme (hereafter WZM). Both the original ZM and NZM schemes employ a convective available potential energy (CAPE)-based closure assumption and emphasize convective instability in determining convection. The NZM modification incorporates the environmental air (e.g., dry versus moist environment) on convection. Differing from the ZM and NZM modifications, instead of CAPE-based closure, a quasi-equilibrium (QE)-based closure is used in ZZM and WZM. The QE-based closure emphasizes the destabilization of the atmosphere through tropospheric forcing (e.g., synoptic-scale disturbances) in driving convection. Because rainfall in the EASM system results from a mix of convection and stratiform clouds, and because there are planetary, synoptic, and mesoscale systems in the EASM rain belts, how to describe the effect of convective activities on rainfall and circulation simulation in the EASM region remains an open question.

The purpose of this study is to evaluate the performance of CAM3.5 in simulating the EASM and to discuss the effects of three revised versions of the ZM convection scheme. We focus on the EASM climatology and its seasonal variation. We want to address the following questions: 1) What are the strengths and weaknesses of the National Center for Atmospheric Research (NCAR) CAM3.5 in EASM simulation? 2) What are the influences of the modifications to the convection scheme on EASM simulations? Our results show that the main characteristics of EASM circulation is reasonably simulated by CAM3.5, but the simulation of the mei-yu/baiu/changma rain belt and the meridional monsoon cell still need to be improved. The precipitation is sensitive to changes in the convection scheme, and the WZM and ZZM schemes, which use quasi-equilibrium closure, show better performance in EASM rainband simulations.

The rest of the paper is organized as follows: Section 2 provides a description of the model and three sets of modifications to the ZM convection scheme, as well as the datasets and analysis methods used in this study. The EASM climate mean state is assessed in section 3, while section 4 addresses seasonal and intraseasonal monsoon variation. Section 5 discusses the possible causes of the model biases. Section 6 follows with summary and concluding remarks.

2. Model, data, and analysis method description
a. CAM3.5

CAM3.5, developed by the NCAR in collaboration with the climate modeling community, is the recently improved version of the state-of-the-art atmospheric general circulation model (AGCM) and serves as an interim version with which to improve the model physics for the next-generation CAM, version 4 (CAM4), which is a global primitive-equation model with 26 vertical levels. The integration uses a finite-volume dynamical core (Lin 2004), and the horizontal resolution is approximately 1.9° latitude × 2.5° longitude. This version is closely related to the previous version (CAM3) and includes changes to convective and cloud processes, the land model, and chemistry modules (Oleson et al. 2008; Stöckli et al. 2008; Neale et al. 2008; Richter and Rasch 2008; Gent et al. 2009). The calculation of cloud fraction is updated, and new hydrology, surface datasets, and canopy integration are introduced in the land model. The other physical processes are the same as those in CAM3, which can be found in Collins et al. (2006).

CAM3 uses the standard ZM convection parameterization scheme (Zhang and McFarlane 1995), which is a mass flux scheme inspired by the convective parameterization of Arakawa and Schubert (1974). An updraft ensemble of entraining convective plumes, all having the same mass flux at the cloud base, relaxes the atmosphere toward a threshold value of CAPE. Simulations are not sensitive to the threshold used. In-cloud saturated downdrafts commence at the level of minimum saturated moist static energy. Detrainment of ascending plumes also begins at the level of minimum saturated moist static energy. Therefore, only ascending plumes that can penetrate through the conditionally unstable lower troposphere are present in the ensemble. In the NZM modification,
the calculation of CAPE in the ZM scheme was modified to include the effect of lateral entrainment dilution (Neale et al. 2008). In addition, convective momentum transport (CMT) parameterized by Gregory et al. (1997) was included (Richter and Rasch 2008). In the ZZM version proposed by Zhang (2002) and used in Zhang and Mu (2005), the CAPE-based closure in the original ZM scheme is changed to quasi-equilibrium-based closure, and a relative humidity threshold of 80% at the parcel lifting level was imposed for convection. The WZM version is similar to the ZZZ version, but with the relative humidity threshold replaced by a threshold in large-scale CAPE forcing, based on cloud-resolving model simulation estimates (Wu et al. 2003, 2007a; Zhang and Wu 2003). Furthermore, CMT developed by Zhang and Cho (1991a,b) was also included in the WZM version (Wu et al. 2007b). This CMT parameterization, which is validated and simplified through the study of cloud-resolving model simulations (Zhang and Wu 2003), redistributes the horizontal momentum through the subsidence-compensating convective mass flux, detrainment of in-cloud momentum, and the cloud-scale perturbation pressure gradient.

In this study, four Atmospheric Model Intercomparison Project (AMIP)-type integrations (Phillips 1996), forced by observed global SSts, are performed and compared by using the three revised convection schemes as well as control (CNTL) run using CAM3.5 with the CAM3 convection configuration. Table 1 shows a brief summary of the experiments. All of the experiments are integrated for more than 20 yr and the data from 1980 to 1999 are analyzed in this study.

### b. Data

In this paper, East Asia monsoon region refers to the area between 0° and 45°N and between 90° and 140°E, including the monsoon region over East Asia and the western northern Pacific (Wang and Lin 2002; Ding and Chan 2005). To evaluate the model, the following reanalysis/observation datasets are used: 1) geopotential heights, zonal and meridional winds, air temperature, specific humidity, and surface pressure from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis during 1980–99 on a 2.5° × 2.5° grid (Kalnay et al. 1996); and 2) the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) based on blended satellite and in situ measurements during 1980–99 on a 2.5° × 2.5° grid (Xie and Arkin 1997).

### c. Analysis methods

The main circulation components of the EASM are measured by specific indices. These indices are applied to quantitatively evaluate the model performance. At the upper level (100 hPa), the intensity of the Tibetan high (defined as the averaged geopotential height over 20°–40°N, 60°–120°E), the zonal wind speed of the western jet (defined as the zonal wind speed averaged over 40°–50°N, 60°–120°E) and the tropical easterly jet (TEJ; defined as the zonal wind speed averaged over 10°–20°N, 60°–120°E) are calculated separately. In the middle level (500 hPa), three indices, which respectively represent the intensity of NW, westward extension of NW, and northern edge of the WPSH, are calculated to make an objective comparison between the reanalysis and the model. Here, IS is defined as the regional average of the grids with geopotential height greater than 5860 gpm over the region of 10°–40°N, 100°–140°E. To compare the westward extension of WPSH in reanalysis and model simulations, each of the 500-hPa geopotential height values over the target region are subtracted by IS. Then, IW is defined as the longitude of the westward extension of 0-gpm contours of the subtracted field (Zhou et al. 2009b). The WPSH ridge, defined as u = 0 and du/dy > 0 (Li and Chou 1998), is constructed before calculating IS. Afterward, IN is defined as the latitude of the WPSH ridge position.

To investigate the possible causes of differences between different modifications to ZM, CAPE is estimated and its correlation with rainfall is assessed. The CAPE is defined by $\int_{p_b}^{p_f} R_d (T_v - T_v^d) d \ln p$ (Zhang 2002; Zhang 2003), where $R_d$ is specific gas constant, $p_b$ is the level in the boundary layer at which convective air parcels originate, $p_f$ is the level at which the parcel loses its buoyancy, and $T_v$ is the virtual temperature. Because the NZM uses the dilute CAPE for closure (Neale et al. 2008), the CAPE of NZM is directly derived from model output. Furthermore, to examine the effect of diabatic heating and surface sensible heat flux, the apparent heat source Q1 is calculated following Yanai et al. (1992).
Moreover, a Student’s t test is applied to check the significance of the differences between observations and simulations as well as those between model experiments.

3. Climatological mean pattern

a. Precipitation

Precipitation is one of the most important variables used to measure monsoon activities. From the observations (Fig. 1a), two major rain belts are evident. One is the tropical monsoon trough located between 0° and 20°N, which reflects the intense rainfall corresponding to the ITCZ in the tropical western Pacific. The other, known as the subtropical mei-yu/baiu/changma front, exhibits a zonally elongated structure extending from East China toward the northwestern Pacific near 30°N, with heavy rainfall along the Yangtze River (100°-120°E) in China, and extending over South Korea and southwestern Japan. These two rain belts are closely related to each other, with a relatively dry region dominated by the WPSH between them. The summer mean precipitation in the CAM3.5 control run has a spatial pattern similar to that of CMAP (Fig. 1b), but some biases are also obvious. The largest deficiency is the significant underestimation of precipitation over both the tropical and subtropical rain belt areas (Fig. 1c). In the tropical region, the model simulates much weaker precipitation over the Bay of Bengal, South China Sea, and the subtropical western Pacific, with the main rainband shifting to the Southern Hemisphere along 5°S. The rainfall centers associated with mei-yu/baiu/changma front over the middle and lower reaches of the Yangtze River valley, South Korea, and southwestern Japan are almost absent. The observed rain belt extends from the eastern...
flank of the Tibetan Plateau to the mid-Pacific, while the simulated rainband is mainly located along the landward side of the eastern coast of the Asian continent. Meanwhile, the model produces excessive precipitation over most land areas of the Asian continent. Overestimation of the rainfall over the Tibetan Plateau and its eastern flank is also evident.

The differences between the three modifications to the convection scheme and the control run are shown in Figs. 1d–f. Overall, all three modifications improve the simulation of EASM precipitation. The rain rate over the two main rainbands is increased. All of these runs show similar biases to those seen in the control run, for example, the underestimated precipitation over the South China Sea and Philippines Sea and the excessive rainfall over the Tibetan Plateau and its eastern periphery. However, regional details are different among the three runs. In the NZM run, the precipitation rate over the subtropical rainband is still weaker than the observations, while in the WZM and ZZM runs, the amplitude is more realistic along the mei-yu/baiu/changma front. The rainfall centers over South Korea and southwestern Japan are only evident in the WZM and ZZM runs, but with a weaker intensity.

To quantitatively evaluate the model’s performance in simulating the geographical distribution of monsoon precipitation, a Taylor diagram (Taylor 2001) is employed (Fig. 2). For better comparison, model results are interpolated to the spatial grid of the observations. The rainfall distributions over the tropical (5°–15°N) and subtropical (25°–35°N) regions are also compared. The standard deviations of the simulations are smaller than those of the observations except for the subtropical rainfall in the ZZM run, suggesting that the spatial variance in the model is smaller than the observation. Compared to the control run, the three revised versions improve the monsoon rainfall simulations. This improvement is evident in both tropical and subtropical rainfall. The NZM run simulates the EASM rainfall more realistically in its tropical rain belt, but not as well in the subtropical rainfall. The subtropical rainfall simulated by WZM is the most reasonable.

b. EASM horizontal circulation

Monsoon rainfall distribution is closely related to monsoon circulation. At the upper level (100 hPa), the most outstanding feature of the EASM is the huge anticyclonic circulation (the so-called Tibetan high) centered over the southern edge of the Tibetan Plateau, with the axis of the anticyclone along 30°N, a westerly jet to the north along 40°N, and the TEJ to the south of 25°N (Fig. 3a). The intensity of the Tibetan high simulated in the control run is stronger than in the reanalysis, and its center shifts westward with a northward-extending ridge (Fig. 3b). As a result, there is an anomalous anticyclone over northeastern China (40°–50°N, 110°–130°E) and strong northerlies west of 90°E in the model (Fig. 3c). The unrealistic anticyclone and northerlies are improved in all of the three revised schemes, but the Tibetan high shows similar patterns in all the runs, suggesting that the upper-level circulation is less sensitive to changes in the convection scheme. The indices related to the Tibetan high are listed in Table 2. In the control run, the intensity of the Tibetan high and the westerly jet are stronger than that in the reanalysis, while the TEJ is weaker. The intensities of the high, westerly jet, and TEJ simulated by the three revised versions of the ZM scheme are weaker than those in the reanalysis. The weak high indicates a weak divergence in the upper troposphere. The weak TEJ in the simulation is consistent with the weak rainfall over the tropical monsoon trough, because the TEJ is closely linked to monsoon rainfall in Asia through the meridional vertical circulation (Zeng and Guo 1982).

In the middle and lower troposphere, the WPSH greatly influences the climate of the EASM. The position, shape, and strength of the WPSH dominate the large-scale quasi-stationary frontal zones and associated rainband in East Asia (Tao and Chen 1987; Ding 1994; Zhou and Yu 2005). Here we use the 500-hPa geopotential height to measure the WPSH, which has been widely used in previous studies (e.g., Yu et al. 2000; Liu et al. 2002; Zhou and Li
In the NCEP–NCAR reanalysis, a strong anticyclone dominates the subtropical western Pacific and a weak trough appears over northeastern China (Fig. 4a).

The spatial pattern of WPSH in the control run resembles the reanalysis (Fig. 4b). The intensity of WPSH is stronger than that in reanalysis, and it extends westward approximately $5^\circ$ and the ridge shifts northward more than $5^\circ$ (Table 3). The westward extension of the WPSH, along with the downward motion, is consistent with the underestimated precipitation over the subtropical regions of East Asia. Corresponding to the strong WPSH, the westerlies in the higher latitudes and the southerlies in the tropics are both stronger than those in the reanalysis (Fig. 4c). The stronger southwesterly monsoon flows reach relatively higher latitudes in East Asia. The northward shift of the WPSH ridge results in the southerlies deeply penetrating into northern China.

<table>
<thead>
<tr>
<th>Index</th>
<th>Averaged region</th>
<th>Units</th>
<th>NCEP–NCAR</th>
<th>CNTL</th>
<th>NZM</th>
<th>WZM</th>
<th>ZZM</th>
</tr>
</thead>
<tbody>
<tr>
<td>TPI</td>
<td>20°–40°N, 60°–120°E</td>
<td>dagpm</td>
<td>1677.5</td>
<td>1678.7</td>
<td>1670.4</td>
<td>1670.9</td>
<td>1672.2</td>
</tr>
<tr>
<td>$U$ wind</td>
<td>40°–50°N, 60°–120°E</td>
<td>m s$^{-1}$</td>
<td>14</td>
<td>16</td>
<td>14.6</td>
<td>14.4</td>
<td>14.1</td>
</tr>
<tr>
<td>$U$ wind (TEJ)</td>
<td>10°–20°N, 60°–120°E</td>
<td>m s$^{-1}$</td>
<td>−24.7</td>
<td>−23.6</td>
<td>−22.4</td>
<td>−19</td>
<td>−22.8</td>
</tr>
</tbody>
</table>
This corresponds to the excessive precipitation in North China, but deficient mei-yu/baiu/changma rainfall along the Yangtze River (Fig. 1). The midlatitude westerlies in the model seem to be more flat. No trough is evident in the midlatitude westerlies, because an anomalous anticyclone dominates northern China where there is a weak trough in the reanalysis. Although the simulations are improved in the three revised versions of the ZM scheme (Figs. 4d–f), the northward shift of the WPSH still exists, especially in the ZZM run (Table 3). The westward extension is also obvious in all three revised schemes. Generally, the stronger WPSH is associated with a westward extension and northward shift. According to Rodwell and Hoskins (2001), the westward extension can be attributed to the intensified rainfall over the ISM region (Fig. 1) in terms of the Sverdrup vorticity balance. This bias of the WPSH then causes the drier bias of rainfall over the western Pacific, which is dominated by the ridge of WPSH.

The water vapor transport is crucial to monsoon rainfall, and closely resembles the large-scale monsoon circulation in the lower troposphere (Zhou and Yu 2005). From Fig. 5a, three main branches of water vapor transport to East Asia are evident: a strong transport by southwesterlies from the ISM, a moderate transport by southeasterlies from the western Pacific, and a weak branch linked to the cross-equator flow straddling 105°–150°E. The southwesterly transport in all runs is weaker than that in the reanalysis (Figs. 5b–f), which leads to

![Fig. 4. As in Fig. 3, but for 500-hPa wind and geopotential height.](http://journals.ametsoc.org/jcli/article-pdf/23/13/3657/3965857/2010jcli3022_1.pdf)

### Table 3

<table>
<thead>
<tr>
<th>Index</th>
<th>NCEP–NCAR</th>
<th>CNTL</th>
<th>NZM</th>
<th>WZM</th>
<th>ZZM</th>
</tr>
</thead>
<tbody>
<tr>
<td>$I_s$</td>
<td>5868.3</td>
<td>5886.9</td>
<td>5868.5</td>
<td>5868.9</td>
<td>5871.8</td>
</tr>
<tr>
<td>$I_N$</td>
<td>22.5</td>
<td>29.37</td>
<td>27.47</td>
<td>27.47</td>
<td>31.26</td>
</tr>
<tr>
<td>$I_W$</td>
<td>112.5</td>
<td>107.5</td>
<td>112.5</td>
<td>117.5</td>
<td>102.5</td>
</tr>
</tbody>
</table>

The water vapor transport is crucial to monsoon rainfall, and closely resembles the large-scale monsoon circulation in the lower troposphere (Zhou and Yu 2005). From Fig. 5a, three main branches of water vapor transport to East Asia are evident: a strong transport by southwesterlies from the ISM, a moderate transport by southeasterlies from the western Pacific, and a weak branch linked to the cross-equator flow straddling 105°–150°E. The southwesterly transport in all runs is weaker than that in the reanalysis (Figs. 5b–f), which leads to
weak rainfall centers over the Bay of Bengal and South China Sea. The simulated moisture transport turns north-westward to the north of the Bay of Bengal near 20°N and causes a strong northward transport to the mid-latitudes (Fig. 5b). This leads to a weaker contribution of the southeasterly monsoon flow to the water vapor transport over the EASM. The southeasterly transport from the western Pacific is also weak, and shifts westward and extends northward as a result of the biases in WPSH simulation. The southeasterly transport from the ISM and the southeasterly from the western Pacific are strengthened and the northward transport is weakened in all revised model versions, but the spatial distribution of the biases are similar to the control run. Therefore, the deficiencies of the water vapor transport are greatly related to the weak rainfall associated with the subtropical mei-yu/baiu/changma front and the relatively strong rainfall over the midlatitude continent north of 45°N.

c. Meridional monsoon circulation

One unique characteristic in the East Asian monsoon region is that the normal Hadley cell is replaced by a meridional circulation of the opposite sense, which is often referred to as the monsoonal meridional cell (Chen et al. 1964; Ye and Yang 1979). The monsoonal meridional cell has been used as an observational metric for evaluating climate models (e.g., Zhou and Li 2002). The averaged meridional circulation in the monsoon region is shown in Fig. 6 (from 90° to 140°E). Strong upward motion controls the region between 10° and 30°N of the Northern Hemisphere, while strong low-level convergence

![Fig. 5. Vertically integrated summer mean water vapor transport (kg m⁻¹ s⁻¹) in (a) NCEP–NCAR reanalysis, (b) control run, (c) the difference between the control run and reanalysis, and (d)–(f) the differences between the three revised versions and the control run, respectively. The gray shading and the black dots denote regions with zonal and meridional components that are significant at the 1% level, respectively.](http://journals.ametsoc.org/jcli/article-pdf/23/13/3657/3965857/2010jcli3022_1.pdf)
dominates the region near 35°N, which is closely related to the mei-yu/baiu/changma rain belt. None of the simulations reasonably reproduces the EASM meridional monsoon circulation (Figs. 6b–f). In the simulations, strong upward motion dominates the tropical area, and weak ascent is seen in the lower troposphere between 20° and 40°N. The middle and upper levels of the subtropical troposphere north of 30°N are controlled by the subsidence, which is the opposite compared to the reanalysis. In the three revised versions, the subsidence flow is improved, because an anomalous upward motion dominates the region north of 45°N, but the intensity of the upward flow is still much weaker than the reanalysis (Figs. 6d–f). This subsidence corresponds to the westward extension of WPSH shown in Fig. 4. Therefore, the biases of rainfall and circulation in subtropical East Asia are closely related to each other, which may be caused by the weak land–sea thermal contrast in the model in the monsoon region (see the discussion below).

4. Seasonal variation

The dominant characteristic of monsoon climate is the seasonal cycle, especially in rainfall (Ding 1994). The seasonal march of the climate mean precipitation averaged over 110°–125°E is shown in Fig. 7. In observations (Fig. 7a), prior to mid-May, southern China experiences a premonsoon rainy season. The monsoon rains extend from southern Asia to the Yangtze River valley in June, and finally penetrate to the northern China in July. The rainy season in northern China lasts for approximately 1 month and ends in August. From August to September, the monsoon rain belt rapidly moves back to southern China. The differences among different model versions are also evident. In the control run (Fig. 7b), both the tropical and subtropical rainfalls are weaker than that observed. These features are separated into two bands, with a dry tongue located between 20° and 25°N from February to June. In midsummer, the strong rainfall

![Fig. 6. Northern Hemisphere summer mean meridional circulations for 90°–140°E in (a) NCEP–NCAR reanalysis and (b) the control run. (c)–(f) The differences between the control run and the reanalysis and between the three revised versions of the ZM scheme and the control run, respectively. The black dots and gray shadings denote regions with meridional and vertical components that are significant at the 1% level, respectively.](http://journals.ametsoc.org/jcli/article-pdf/23/13/3657/3965857/2010jcli3022_1.pdf)
extends northward and the rainband withdraws more rapidly than in the observations. In the three revised versions of the ZM scheme, the northward progress and southward withdrawal of the monsoon rainband are more reasonable. The subtropical rainband simulated by NZM shifts northward, and the rainfall along the mei-yu/baiu/changma front near 30°N decreases dramatically after July, which is approximately 3 months earlier than that observed (Fig. 7c). The simulations of the WZM and the ZZM are more realistic, although the strong raining periods last from June to August over the area north of 40°N, which is approximately 2 months longer than that observed (Figs. 7d,e).

The seasonal progression of the rain belt is closely related to the seasonal change of large-scale circulation. The circulation and related rainfall over Asia undergo abrupt seasonal changes, which are linked to tropospheric warming over the Asian landmass (Murakami and Ding 1982; He et al. 1987; Yanai et al. 1992). The ridge of the WPSH from May to August and 3 mm day$^{-1}$ precipitation rates are shown in Fig. 8. Because the mei-yu/baiu/changma rainfall is more closely related to the WPSH, we show the seasonal march of the rainband in subtropical regions. Two northward jumps of the WPSH ridge in June and July are evident in Fig. 8a. Correspondingly, the rainband over East Asia undergoes two northward jumps (Fig. 8b). In June, the strong rainfall advances from southern China to the Yangtze River valley, and then the mei-yu begins. In July, the strong rainfall further jumps to northern China.

All simulations reasonably reproduce the poleward jumps of the WPSH ridge (left panel, Fig. 8), but the position of the ridge shifts northward. The ridge is flat in the zonal direction, which is different from the southwest-northeast-tilted pattern in the reanalysis. The simulated northward jumps of the rain belt are not well simulated (right panel, Fig. 8) except for the WZM run (Fig. 8j). The simulated rain belts all shift northward approximately 10°, which is inconsistent with the bias of the WPSH ridge. Meanwhile, it is worth noting that the evolution of the rain fronts with month is much more smooth for ZZM and WZM (Figs. 8h,j). This is because they are more weakly coupled to surface inhomogeneities resulting from the closure and relative humidity minimum closure conditions that are not present in CNTL and NZM.

![Seasonal migration of the monsoon rainband](image-url)
Fig. 8. Seasonal evolution of (left) 500-hPa WPSH ridge and (right) the corresponding 3 mm day\(^{-1}\) precipitation rate contour line averaged for May (black solid line), June (blue dashed line), July (red dotted line), and August (purple dotted–dashed line).
5. Discussion

The different responses to the revised versions of the ZM convection scheme indicate the important role of convective processes in EASM simulation. The disparities between the NZM and WZM runs may be attributed to closure assumptions, because these two modifications both have momentum transport parameterization, even though they are formulated differently (Neale et al. 2008; Wu et al. 2003). The WZM and ZZM runs both use QE closure. The differences between these two may reflect the roles of convective-scale momentum transport as well as threshold of relative humidity. The above results show that the NZM run improves the EASM tropical rainfall distribution, especially in the Indian Ocean, indicating that NZM may have advantages in simulating tropical deep convection. However, over the East Asian subtropical regions, the tropospheric large-scale forcing is important to convection, especially for the mei-yu/baiu/changma front. The mei-yu/baiu/changma front is a rainband thousands of kilometers long, stretching across East Asia and the northwestern Pacific, and is closely related to the planetary-scale circulation. Precipitation along the mei-yu/baiu/changma front is usually caused by convection and clouds organized into eastward-moving mesoscale convective systems, and the large-scale dynamic circulation associated with baroclinic instability is the driver of the system (Chen et al. 1991). In ZM and NZM, the convection is determined by the amount of CAPE in the atmosphere. Because CAPE strongly depends on the boundary layer equivalent potential temperature, which in turn is largely governed by surface heat and moisture fluxes, this closure ties convection closely to boundary layer forcing (Zhang 2003). Therefore, the NZM does not perform well over these regions. In WZM and the ZZM modifications, convection is determined by the CAPE generation rate from the free-tropospheric large-scale forcing, which has explicitly tied convection to the large-scale forcing through the QE assumption. Thus, they perform better in this region.

To further discuss the influences of different closure assumptions, the spatial correlations between CAPE and precipitation in tropical and subtropical regions are shown in Fig. 9. For comparison, the CAPE is calculated based on NCEP–NCAR reanalysis data, while the precipitation is derived from CMAP data. As shown in section 2, the CAPE calculation uses only temperature $T$ and moisture $q$ of reanalysis. Because both $T$ and $q$ are observable quantities, the impact of model parameterizations in the NCEP–NCAR reanalysis system on CAPE calculation should not be very large. Along the tropical rain belt ($0^\circ$–$20^\circ$N, $100^\circ$–$140^\circ$E), the precipitation and CAPE are positively correlated (Fig. 9a), while along the subtropical rainband ($25^\circ$–$35^\circ$N, $100^\circ$–$140^\circ$E) the correlation is negative (Fig. 9b). In ZM and NZM, which use a CAPE-based closure, the CAPE is positively correlated with the precipitation in both the tropical and subtropical regions. In WZM and ZZM, which use a QE-based closure, the correlation is relatively weak in both the tropics and along the mei-yu/baiu/changma rain belt. This may explain why the NZM simulation performs better in the tropical region but not as well in subtropical regions. The QE closure based on the large-scale forcing should be more favorable in simulating the EASM subtropical rainfall.

Although there are differences among the results of the different runs, the model has some common biases in which the weaker-than-observed land–sea thermal contrasts may play an important role. The Asian summer monsoon is closely related to large-scale land–sea thermal contrasts. The key driving force for the summer monsoon is the available potential energy generated by the differential heating between land and sea (Li and Yanai 1996; Zhang et al. 1997). Because tropical latent heat and plateau sensible heating are the main thermal sources in the EASM, the systematic biases in CAM3.5 should be linked to the biases in heating sources. The zonally averaged ($90^\circ$–$140^\circ$E) cross section of $Q_1$ is shown in Fig. 10 to evaluate the diabatic heating and sensible heat flux. In the Northern Hemisphere, $Q_1$ is positive throughout the troposphere, and intense heating ($>3$ K day$^{-1}$) is located at the middle levels (Fig. 10a), consistent with previous studies that show sensible heating from the plateau surface is the major source of heating during this season (Luo and Yanai 1984; Yanai et al. 1992). In the control run, $Q_1$ is greatly underestimated (Fig. 10b). The tropical heating is positive, but only approximately half of the intensity in the reanalysis. The three modifications improve the simulated heat source mainly in the tropics (Figs. 10c–f). Because all of the model experiments share the same surface land model, the land–sea contrast biases are shared and the $Q_1$ profile in the midlatitude show similar pattern. The significant zonal bias pattern strongly decreases the meridional thermal contrast; thus, the monsoon meridional circulation driven by this thermal effect is much weaker than that in the reanalysis (Fig. 6). The weaker monsoon circulation in the model leads to less rainfall over the East Asian tropical and subtropical regions.

To estimate tropical heating we use the proxy of total rainfall (Zhou and Li 2002). The observed rainfall has a pronounced center over the northern portion of the Bay of Bengal, with two other centers over the South China Sea and western Pacific (Fig. 1a). This indicates a strong heating associated with upward motion (Fig. 11) and upper-tropospheric divergence (Fig. 12) over these
regions. There are rainfall centers surrounding the Bay of Bengal in all of the simulations; however, a stronger rainfall center is evident in the northern Indian Ocean in both the control run and NZM run (Figs. 1c,d), while strong rainfall centers are located both in the northern Indian Ocean and the west edge of the Philippines Sea in both the WZM and ZZM runs (Figs. 1e,f). Because the mean position and intensity of tropical convection are closely related to the atmospheric general circulation (Branstator 1983), these spurious equatorial rainfall centers (and their associated latent heating) should have negative effects on the large-scale circulation patterns.

FIG. 9. (a)–(b) Scatterplot of the estimated JJA mean CAPE (J kg\(^{-1}\)) from NCEP–NCAR reanalysis vs corresponding CMAP precipitation (mm day\(^{-1}\)) in tropical (0°–20°N, 100°–140°E) and subtropical (25°–35°N, 100°–140°E) regions. (c)–(j) As in (a)–(b), but for each run using different modifications of the ZM scheme. The spatial correlation coefficients are shown above the top-right boundary of each plot.
Both upper-level divergence centers and low-level upward motion over the western Pacific shift westward to the Indian in the control run. The NZM run simulates a similar pattern, because its differences with the control run show an almost reversed pattern compared to that between the control run and reanalysis (Figs. 11c,d and 12c,d). In the WZM and ZZM runs, the convection in the open ocean of the western Pacific (east of 140°E), and the upward motion in the Maritime Continent region are suppressed. Thus, the divergence center shifts eastward to the mid-Pacific (Figs. 12e,f) and the low-level convergence and upward motion are mainly located west of 120°E (Figs. 11e,f). No evident divergence center is seen over the EASM region, and this may partly explain the weak rainfall over the EASM tropical region.

Finally, it should be noted that the EASM is a complex phenomenon including interactions between tropical and subtropical systems. Based on the above analyses, for a better simulation of EASM, the following four factors need to be considered: 1) the influences of the ISM and the moist transport along the south flank of the Tibetan Plateau; 2) the dynamical response to the plateau warming, especially in the downstream region; 3) reasonable simulations of convection related to the mesoscale and synoptic systems in the EASM subtropical region; and 4) a reasonable simulation of the ITCZ and related circulation in the tropical western Pacific. Further efforts are required to improve these factors in current GCMs to obtain more realistic simulation of the EASM.

6. Summary and concluding remarks

In this study, the EASM simulated by CAM3.5 is evaluated in terms of climate mean pattern and seasonal variation. The sensitivity of simulations to ZM convection scheme as well as its three modifications is also examined. Both the strengths and weaknesses of the four versions are documented. The results are potentially helpful to
the climate modeling community in improving the simulation of the East Asian monsoon. The major conclusions are summarized below.

1) The major characteristics of the EASM circulation, including the Tibetan high in the upper troposphere and the WPSH in the middle and lower troposphere, are reasonably simulated by CAM3.5. The model’s main deficiency lies in the simulation of tropical and subtropical monsoon rainfall. All four runs simulate weaker-than-observed rainfall related to the mei-yu/baiu/changma front. This deficiency is closely related to the weak meridional monsoon circulation over East Asia, which is attributed to the weak heating over the subtropical continental model. The tropical rainfall centers over the South China Sea and western Pacific are also weak in the model.

2) The springtime northward advance and late summer southward withdrawal of the monsoon rain belt over East Asia can be reproduced by the model. Two abrupt northward jumps of the WPSH from May to August are also well simulated but the northward jump of the rain belt is poorly simulated, especially for the northward jump in July. The deficiencies in the seasonal cycle are closely related to the climatological pattern, because the northward extension of the WPSH results in longer periods of the rainy monsoon season at higher latitudes near 40°N.

3) Both the distribution and seasonal cycle of monsoon rainfall simulations depend on the convection scheme, while the monsoon circulation shows less sensitivity. The three revised ZM modifications generally improve the simulation of the EASM relative to the control run. The NZM run is better at simulating tropical rainfall partly because it uses a CAPE-based closure assumption. The WZM and ZZM runs are better at simulating the subtropical rainfall and its seasonal variation, implying that the QE-based assumption is more favorable in simulating rain belts associated with large-scale forcing.
By assessing the model performance in EASM simulation, the present study is a useful step toward a better understanding and simulation of the climate system in East Asia. It is our hope that the results shown here can be helpful for future model improvement and development.

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