Simulation of the Diurnal Cycle in Tropical Rainfall and Circulation during Boreal Summer with a High-Resolution GCM

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ABSTRACT

The simulation of the diurnal cycle (DC) of precipitation and surface wind pattern by a general circulation model (GCM) with a uniform horizontal resolution of 50 km over the global domain is evaluated. The model output is compared with observational counterparts based on datasets provided by the Tropical Rainfall Measuring Mission and reanalysis products of the European Centre for Medium-Range Weather Forecasts. The summertime diurnal characteristics over tropical regions in Asia, the Americas, and Africa are portrayed using the amplitude and phase of the first harmonic of the 24-h cycle, departures of data fields during selected hours from the daily mean, and differences between extreme phases of the DC.

There is general agreement between the model and observations with respect to the large-scale land–sea contrasts in the DC. Maximum land precipitation, onshore flows, and landward migration of rainfall signals from the coasts occur in the afternoon, whereas peak maritime rainfall and offshore flows prevail in the morning. Seaward migration of precipitation is discernible over the western Bay of Bengal and South China Sea during nocturnal and morning hours. The evolution from low-intensity rainfall in the morning/early afternoon to heavier precipitation several hours later is also evident over selected continental sites. However, the observed incidence of rainfall with very high intensity in midafternoon is not reproduced in the model atmosphere.

Although the model provides an adequate simulation of the daytime upslope and nighttime downslope winds in the vicinity of mountain ranges, valleys, and basins, there are notable discrepancies between model and observations in the DC of precipitation near some of these orographic features. The model does not reproduce the observed seaward migration of precipitation from the western coasts of Myanmar (Burma) and India, and from individual islands of the Indonesian Archipelago at nighttime.

1. Introduction

The diurnal cycle (DC) is one of the most pronounced periodic phenomena in our climate system. The spatiotemporal characteristics of this prominent mode of atmospheric variability have been the object of numerous observational, theoretical and modeling studies. Diurnal signals are discernible in many meteorological variables, among which the precipitation and wind circulation fields are of particular interest. In many tropical and subtropical zones, the amplitude of the DC in precipitation during the wet season is a crucial factor for determining the annual mean rainfall. In these locations, the rainfall maxima generally occur in the middle to late afternoon over land, and at nighttime or early morning hours over open-ocean regions (e.g., Dai 2001). However, many geographical sites exhibit notable departures from this typical scenario, mainly due to local influences of orographic features, coastline geometry, and prevalent circulation regime (e.g., Yang and Slingo 2001; Kikuchi and Wang 2008).

Early observational investigations on the nature of the DC in precipitation were mostly based on in situ measurements taken at station networks on a routine basis (e.g., see summaries provided by Wallace 1975; Gray and Jacobson 1977; Dai et al. 1999; Dai 2001), or at specific regions during targeted field campaigns (e.g., McGarry and Reed 1978; Albright et al. 1981). The spatial data gaps in such monitoring arrays preclude detailed studies of the DC over many maritime regions and remote land areas. More recently, the spatial homogeneity of the data coverage has been much improved by the availability of measurements made by satellites (e.g., Garreaud and Wallace 1997; Yang and Slingo 2001; Yang and Smith 2006; Kikuchi and Wang 2008; Hara et al. 2009).
Many mechanisms have been proposed for explaining the geographical dependence of the DC in precipitation. Various processes contributing to such spatial diversity have been considered by Wallace (1975), Gray and Jacobson (1977), Yang and Smith (2006), and Johnson (2010), among others. The physical causes for various regional details include the following:

- Daytime heating of land surfaces by solar radiation, which alters the thermal structure and stability characteristics of the atmosphere (e.g., Blackadar 1957; Estoque 1961; Simpson et al. 1977).
- Differences in surface heating and cooling at various elevations of sloping terrains, which drive upslope circulations in the afternoon and downslope flows at night (e.g., McNider and Pielke 1984).
- Diurnal contrasts in longwave radiative cooling in cloud systems and solar absorption by water vapor in cloud-free regions, which affect the degree of destabilization and condensation (e.g., Gray and Jacobson 1977).
- Diurnal evolution of the sea surface temperature (SST), which leads to destabilization of the marine boundary layer during the afternoon hours (e.g., Sui et al. 1997).
- Excitation and propagation of gravity waves, which transmit the diurnal signals of convection at the source regions to neighboring sites (e.g., Mapes et al. 2003).

In view of the multitude of dynamical, thermal, and radiative processes that contribute to various phenomena associated with the DC, the realism with which the basic diurnal features can be replicated in general circulation models (GCMs) is a good metric for assessing the skill of these simulation tools. Such diagnostics are particularly suited for evaluating the fidelity of GCMs with high spatial resolution, since those models incorporate the fine details of the underlying orography and land–sea contrasts, as well as subsynoptic-scale dynamics.

Yang and Slingo (2001) have compared the output from a model developed at the Met Office against satellite data for brightness temperature, and have noted considerable discrepancies between model and observations with regard to the phase of the DC in precipitation. Dai and Trenberth (2004) have examined the simulation of the DC by the Community Climate System Model. Lee et al. (2007a) analyzed the simulation by three different global GCMs of the DC of warm-season precipitation over the continental United States and northern Mexico. The sensitivity of these model simulations on horizontal resolution, with grid meshes ranging from approximately 2° to ½° in latitude–longitude, has been investigated in a companion paper (Lee et al. 2007b). Hara et al. (2009) have assessed the performance of a 20 km-grid GCM at the Meteorological Research Institute in reproducing the DC in precipitation over the Maritime Continent, and have attributed model deficiencies over the larger islands in that region to inadequate representation of the coupling between moist convection and local circulations in the model atmosphere.

The GCM being examined in this study and the observational datasets for validating the model results are described in section 2. Spatial patterns of the amplitude and phase of simulated and observed DC of summertime precipitation in the overall Asian monsoon region are presented in section 3. Diurnal characteristics of precipitation events in different intensity ranges are examined in section 4. The propagating diurnal signals appearing in several Asian maritime sites are examined in further detail in section 5. The roles of land–sea contrasts and orography in driving the DC in precipitation and surface circulation at selected Asian sites are highlighted in section 6. Corresponding results on the DC over the tropical Americas and Africa are shown in section 7. The major findings are summarized and discussed in section 8.

2. Descriptions of the model and various observational datasets

The GFDL model examined in this study is a global atmospheric GCM, with a dynamical core constructed by Lin (2004). This formulation employs a finite-volume scheme with terrain-following Lagrangian control-volume discretization. This study is primarily focused on the M180 version of the model, with a horizontal resolution of 0.5° latitude by 0.625° longitude. The “180°” indicator in the label for this version refers to the number of grid points between the equator and the pole. Vertical variations are depicted in 24 atmospheric layers. A description of the physical processes that are incorporated in this model has been provided by the GFDL Global Atmospheric Model Development Team (Anderson et al. 2004).
In particular, moist convection is represented using the relaxed Arakawa–Schubert parameterization scheme as formulated by Moorthi and Suarez (1992), surface fluxes are computed using Monin–Obukhov similarity theory, turbulent mixing in boundary layers is treated on the basis of the K-profile scheme of Lock et al. (2000), and ground hydrology is simulated using the land model designed by Milly and Shmakin (2002). Modifications made to these physical schemes in the construction of the M180 model were documented in detail by Anderson et al. (2004). Various facets of the impacts of the model treatments of these processes on the quality of the DC simulation have been noted by Lee et al. (2007a,b). The diurnal cycle is incorporated by performing full radiation calculations at 3-h intervals. The duration of the M180 model integration being analyzed here is 20 yr. Throughout this integration, the climatological seasonal cycle of the observed sea surface temperature (SST) was specified at ocean grid points in all years, so that no interannual variability in the SST conditions was introduced to the model system. Daily mean values of the SST conditions were used in this prescription procedure, so that the oceanic boundary forcing does not contain any diurnal signal. The behavior of synoptic and mesoscale weather systems associated with the East Asian summer monsoon, as simulated in the same experiment, has been examined recently by Lau and Ploshay (2009).

The impact of model resolution on the DC simulation is assessed by comparing the results based on the M180 version with those based on versions with latitude–longitude resolutions of $2^\circ \times 2.5^\circ$ (M45, 20-yr run), $1^\circ \times 1.25^\circ$ (M90, 20-yr run), and $0.25^\circ \times 0.3125^\circ$ (M360, 5-yr run). The parameter changes made to various model versions have been described in Lau and Ploshay (2009). At each resolution, the model data were collected after a statistical equilibrium state was reached. Output for the precipitation, surface wind, and temperature fields are archived at 3-h intervals.

Estimates of the observed precipitation field are based on the 3B42 version 6 of the Tropical Rainfall Measuring Mission (TRMM) dataset (Kummerow et al. 2000), which covers the 1998–2006 period, with a latitudinal and longitudinal resolution of $1/4^\circ$ and a sampling interval of 3 h. This data product is assembled by blending microwave and infrared (IR) measurements taken at various satellite platforms, applying calibration procedures based on the combined TRMM Microwave Imager (TMI) Precipitation Radar (PR) algorithm (Huffman et al. 2007), and rescaling the final results according to rain gauge records. As noted by Kikuchi and Wang (2008), Zhou et al. (2008), and Yamamoto et al. (2008), precipitation estimates based on IR measurements (which are blended into the 3B42 dataset) tend to lag the more reliable in situ or microwave/radar measurements by approximately 3 h. There exist other datasets (such as 3G68) that are based solely on TRMM measurements (i.e., with no blending of IR information applied). Such datasets might depict the DC of precipitation more accurately. However, the size of the data samples in the 3G68 set is much reduced as compared to the 3B42 set, thus yielding rather noisy patterns over the oceans (e.g., see Kikuchi and Wang 2008). After evaluation of the various options, we have chosen to use the 3B42 version in our study mainly to take advantage of the more detailed spatial definition of diurnal phenomena in this dataset by virtue of more abundant data sampling.

We have made an attempt to correct for the systematic bias of the 3B42 dataset by applying a constant 3-h shift to the diurnal phase as inferred from this blended product. This adjustment has been applied to all TRMM results presented throughout this paper and should be borne in mind when comparing between model and TRMM results, as well as between our TRMM results and other findings based on surface observations (e.g., Dai 2001; Dai et al. 2007; Zhou et al. 2008). The simple 3-h phase correction to the 3B42 dataset is an attempt to remove the overall bias that is known to exist due to incorporation of IR measurements. This bias is likely dependent on the specific types of cloud system and synoptic development in question. Hence, our findings need to be checked against future results derived from more elaborate adjustment schemes that take different cloud species and flow environments into account, or from microwave/radar measurements with more adequate space–time sampling.

The simulated surface wind field is compared with the corresponding pattern based on the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analyses (ERA-40; see Uppala et al. 2005) for the 1958–2002 period. This dataset has a latitude–longitude resolution of $1.1^\circ$, and a sampling interval of 6 h (0000, 0600, 1200, and 1800 UTC).

3. Amplitude and phase of the DC in precipitation in the Asian monsoon region

At each hour of day for which precipitation data are available (i.e., 0000, 0300, . . . , 1800, and 2100 UTC), the model or observational records were averaged over the months of June, July, and August for all years, thus yielding climatological means of the DC of precipitation. These 3-hourly data were converted from universal time to local time (LT), and then subjected to a harmonic analysis. The shading patterns in Fig. 1 show the amplitude of the first harmonic of the DC, as computed using (Fig. 1a) M180 model and (Fig. 1b) TRMM data.
The orography is depicted by color contours (see caption for details).

In both the observed and model atmospheres, the DC in precipitation exhibits large amplitudes over southeastern China, parts of the Indochina Peninsula and Indian subcontinent, southern edge of the Tibetan Plateau, western portion of Bay of Bengal (BoB), and various islands surrounding the South China Sea (SCS). The amplitudes of the maxima in Fig. 1 are comparable to the local averaged rainfall rates for the summer season.

**FIG. 1.** Distributions of the amplitude of the first daily (24 h) harmonic of climatological precipitation for the June–July–August season, as depicted by color shading. Results are based on (a) M180 and (b) TRMM data. In this figure and Figs. 2–3, topographic heights of 250, 500, and 750 m are indicated using yellow, green, and blue contours, respectively. Elevations of 1000, 2000, 3000, 4000, and 5000 m are depicted using progressively darker purple contours.
(not shown), thus indicating the dominance of the DC in the precipitation climatology in these regions. The observed DC over some maritime sites (specifically, the eastern and northern portions of BoB, off the western coast of Sumatra, and off the northern coast of Borneo), as well as the region on the eastern flank of the Tibetan Plateau (near 28°N, 102°E), is characterized by notably higher amplitudes in the observed pattern (Fig. 1b) than in the model pattern (Fig. 1a).

The phase of the DC of the precipitation is illustrated in Fig. 2 for (Fig. 2a) M180 and (Fig. 2b) TRMM data. Peak phases occurring in daytime and nighttime are depicted using “warm/bright” and “cold/dark” colors, respectively. Thus yellow–orange (red–pink) shadings represent the morning (afternoon) hours; whereas gray–black (blue) shadings correspond to evening (predawn) hours.

In the observed atmosphere (Fig. 2b), the peak diurnal phase occurs during the middle to late afternoon (1500–1800 LT; note red shading) over most land points. The precipitation at many coastal land points (note deep red shades) peaks earlier than that at grid points farther inland (lighter pink shades). This characteristic is discernible over Indochina, southern Borneo, and the western coast of India. Such landward propagation of precipitation signals near land–sea boundaries is also discernible in the observational results presented by Ohsawa et al. (2001), and has been referred to as the “landside coastal regime” by Kikuchi and Wang (2008). The latter authors have attributed this phenomenon to the penetration of weather systems (such as squall lines and fronts) driven by sea-breeze circulations during the afternoon.

Complex phase patterns are observed in the vicinity of high terrain (see color contours). In accord with the observational evidence presented by Basu (2007), the TRMM data (Fig. 2b) show nighttime (afternoon) precipitation peaks near the bottom (top) of the southern slope of the Tibetan Plateau. As will be demonstrated in section 6, this result is consistent with the prevalence of downslope (upslope) winds in nighttime (afternoon) hours over this region. Enhanced precipitation is observed shortly after midnight over the Sichuan Basin in western central China (see minimum in topography centered near 30°N, 105°E, and the site labeled as SB in Fig. 8c). This feature, as well as the nocturnal precipitation maximum over the eastern periphery of the Tibetan Plateau, has been noted by Yu et al. (2007), Zhou et al. (2008), and Li et al. (2008) on the basis of rain gauge data.

The observed patterns over BoB, the southern portion of SCS, and off the western coast of Sumatra are characterized by progressive phase changes from midnight (blue shading) to next morning (yellow and orange shading) at increasing distances from the shores of India, Myanmar, Peninsular Malaysia, and Sumatra. Various aspects of this seaward migration of precipitation signals during nocturnal and morning hours have been documented by Yang and Slingo (2001), Webster et al. (2002), Zuidema (2003), Cesielski and Johnson (2006), Basu (2007), and Aves and Johnson (2008). This phenomenon has been referred to as the “seaside coastal regime” by Kikuchi and Wang (2008). Yang and Slingo (2001) and Kikuchi and Wang (2008) have considered propagation of gravity waves generated by land convection and terrain-induced heating (e.g., Mapes et al. 2003) as a possible mechanism for these features. Other contributory processes include interactions of land breeze circulations with concavity of the coastlines, or with the monsoon mean flow.

The spatial characteristics of Fig. 2b are qualitatively similar to those appearing in the pattern of the time of maximum rainfall in the Asian monsoon region, as presented by Johnson (2010, see his Fig. 8) using the same TRMM dataset. This author has also documented the diurnal characteristics of precipitation over all longitudes within the tropical zone (see his Figs. 2 and 3).

The observed features in Fig. 2b may be compared with those appearing in the model pattern (Fig. 2a). Both the model and observational results show afternoon precipitation peaks over most land points, morning peaks over the oceans, and eastward propagation over western BoB and southern SCS. However, there exist notable discrepancies between the M180 simulation and TRMM dataset near the southern slope of the Tibetan Plateau. Moreover, the observed seaward propagation off the western Myanmar–Sumatra coasts and off the northern Borneo coast is not evident in the model pattern.

To obtain a quantitative measure of the correspondence between the two patterns displayed in Fig. 2, the absolute value of the local deviations between the observed and simulated phases was computed at individual grid points. Spatial means of the resulting data were then taken over various ocean and land sites. These domain averages, shown in Table 1, confirm our visual impression that the simulated peak phase of the DC is in best agreement with observations over BoB and northern SCS, where the mean absolute departures are 0.8 and 1.4 h, respectively. The largest discrepancies between model and observations (~ 4 h) occur over the Tibetan Plateau and land sites near the Indonesian Archipelago. The averaged deviations in the remaining regions range from 2.3 to 2.7 h. All computations in Table 1 are based on TRMM data subjected to the 3-h adjustment described in the latter part of section 2. As a test of the sensitivity of the results to this correction procedure, analogous analysis has also been performed using unadjusted TRMM data. The latter calculations yield noticeably larger differences.
FIG. 2. Distributions of the peak phase of the DC of climatological precipitation for the June–July–August season, as depicted by color shading (see scale for the local time of day at the bottom). Results are based on (a) M180 and (b) TRMM data. Phase information is indicated only in grid points where the amplitude of the DC exceeds 1.5 mm day$^{-1}$. Regions A and B, where the DC of precipitation occurrences with various intensities will be examined in Fig. 4, are indicated in (a) using black boundaries. A 3-h adjustment has been applied to the TRMM data (see section 2).
between simulated and observed diurnal phases in nearly all sites considered in Table 1, with the exception of the Tibetan Plateau.

The dependence of the simulated diurnal behavior on model horizontal resolution is examined in Fig. 3, which shows the phase patterns as computed using output from (Fig. 3a) M45, (Fig. 3b) M90, and (Fig. 3c) M360 (see model specifications in section 2). The resolution of the M45 model (Fig. 3a) is evidently too crude to reproduce detailed spatial patterns of the diurnal phase. Phase propagation over the BoB and SCS and land–sea contrasts in many regions start to emerge in the M90 model (Fig. 3b). The phase pattern produced by the M360 model (Fig. 3c) exhibit many of the features seen in the chart based on M180 data (Fig. 2a). The M360 simulation provides a better simulation than the M180 in the following aspects: envelopes of blue shading (i.e., rainfall peaking in predawn hours) over the coastal waters around individual islands of the Indonesian Archipelago; as well as nighttime maximum over the Sichuan Basin. However, differences between the observed and M360-generated phase patterns are still discernible over Indochina, Borneo, and Sumatra, as well as portions of BoB and SCS. The presence of these discrepancies across the full range of model versions examined here suggests that some of the simulation errors may not be attributed solely to insufficient horizontal resolution, and that attention should also be devoted to model treatment of various physical processes related to the DC. In view of the relatively shorter data record (5 summers) available from the M360 simulation, we shall henceforth focus our diagnoses on the 20-yr output from the M180 model.

4. Dependence of DC on precipitation intensity

We proceed to make a more in-depth analysis of the diurnal distribution of precipitation events in various intensity categories. This investigation is performed over two regions: southern China (region A) and Indochina (region B), which are depicted in Fig. 2a using black boundaries. These sites are chosen by virtue of their representativeness of the precipitation characteristics over land areas in the subtropical (region A) and tropical (region B) zones within the Asian monsoon region. The phase of the simulated DC in precipitation is almost uniform within each of these regions (see Fig. 2a). For each of the two regions, the 3-hourly precipitation data at various grid points on individual days are used to compute the joint probability distribution function \( F(I, T) \), which denotes the fraction of the number of occurrences in each intensity \( I \) at each local time \( T \) within the range of \( 0–55 \) mm day\(^{-1} \), with a constant interval of \( 1 \) mm day\(^{-1} \), and \( T \) corresponds to the 8 times of day. To accentuate the diurnal dependence of \( F(I, T) \), the average of this function over all times of day for a given \( I \) is removed from \( F(I, T) \) before the results are presented. Hereafter we shall refer to such deviations from the daily mean as \( F'(I, T) \). The distributions of \( F'(I, T) \) with \( I \) (ordinate) and \( T \) (abscissa) are displayed in Fig. 4, for (top panels) region A and (bottom panels) region B. Results based on M180 and TRMM data are presented in the left and right panels, respectively.

In the model pattern for region A (Fig. 4a), the peak phase for diurnal variations with low values of \( I \) (<4 mm day\(^{-1} \)) occurs near 1200 LT. As diurnal variations with increasing \( I \) are considered, the corresponding \( F'(I, T) \) is seen to attain maxima in progressively later hours of the day. For \( I > 15 \) mm day\(^{-1} \), this fraction peaks at 1500–1800 LT. These results are consistent with the typical diurnal evolution of the boundary layer and convection over land regions, with light rain developing in the morning hours, followed by heavier convective precipitation in the late afternoon [see schematic illustration in Fig. 9.24 of Wallace and Hobbs (2006)]. The observed pattern (Fig. 4c) also indicates an earlier peak of \( F'(I, T) \) for \( I < 15 \) mm day\(^{-1} \) (at about 1200 LT) relative to that for \( I > 15 \) mm day\(^{-1} \) (at about 1500 LT). The frequency of diurnal variations with \( I > 40 \) mm day\(^{-1} \) is much higher in the TRMM dataset than in the model atmosphere.
The results in Figs. 4a,c are consistent with the observational evidence presented by Zhou et al. (2008) on the contributions of the DC of both frequency and intensity to the afternoon maximum in total precipitation amount over southern China.

The distinction between the DC in the model atmosphere with low and high values of \( I \) is even stronger over land in the tropics (region B, see Fig. 4b). For \( I < 6 \text{ mm day}^{-1} \), the peak phase is simulated at night and in early morning. On the contrary, the peak phase for categories of \( I > 12 \text{ mm day}^{-1} \) is delayed to the afternoon. Whereas the TRMM pattern (Fig. 4d) does indicate an afternoon peak for high \( I \), the observed diurnal variability for low \( I \) differs noticeably from the model result. Another salient discrepancy between the patterns in Figs. 4b,d is the relatively more frequent occurrence in the model atmosphere of diurnal variations with \( I \) values in the 10–30 \text{ mm day}^{-1} range, and the less frequent variations for \( I > 40 \text{ mm day}^{-1} \).

Yang and Slingo (2001, see their Figs. 5b and 6b) have examined the DC of frequency distributions of brightness temperature \( T_b \) over tropical land regions based on satellite measurements. Their results are in support of our model and observational evidence on preferred timing of heavy precipitation (associated with deep convection and low \( T_b \)) in midafternoon, and lighter rainfall intensity (accompanied by lower cloud tops and higher \( T_b \)) during nighttime and morning hours. Dai (2001, see his Fig. 5) has also documented the diurnal phase of various precipitation types on the basis of station weather reports taken throughout the globe. His analysis also indicates that the frequencies of showery/nondrizzle precipitation as well as thunderstorm events peak in the mid- or late-afternoon hours over regions A and B, whereas drizzle and nonshowery weather mostly occurs in the early morning over these locations.

5. Propagating diurnal features over oceanic sites

Figures 1 and 2 reveal systematic phase evolution of the simulated diurnal precipitation signals over several maritime sites. The spatiotemporal properties of these phenomena are further illustrated in Fig. 5, which shows the development of the simulated precipitation (shading) and surface temperature (contours) with longitude (abscissa) and time (ordinate). These plots are based on data for representative 10-day segments of the model integration, with a sampling interval of 3 h. Results are presented for the longitudinal sectors corresponding to (Fig. 5a) BoB, (Fig. 5b) northern SCS, and (Fig. 5c) southern SCS. The values displayed in each panel are obtained by averaging over latitude bands appropriate for the maritime region of interest. The surface temperature contours represent deviations of the 3-hourly data for a given day from the average of the daily means for the 11-day period centered on that day. For each region considered here, the longitudes of the principal boundaries between the land and oceanic sectors are indicated using bold red lines.
The contour patterns in Fig. 5 indicate that the amplitudes of diurnal temperature variations are highest over the land sector (i.e., India, Indochina, and Malaysia) situated to the west of the respective maritime zones. Precipitation with intensities exceeding 10 mm day$^{-1}$ is initiated over land in afternoon, when or shortly after the maximum surface temperature is reached. The precipitation signals then travel continuously eastward to the neighboring ocean, reaching as far as 10$^\circ$ longitude off the coast in BoB (SCS) in the morning hours of the next day. This scenario is discernable almost every day in the 10-day period in each of the three regions examined in Fig. 5. The zonal component of the seaward propagation speed of these diurnal precipitation signals is approximately 12 and 8 m s$^{-1}$ over BoB and SCS, respectively.

The maps in Fig. 6 show the diurnal changes of precipitation and the attendant wind circulation in the vicinity of BoB in greater detail. Spatial patterns of climatological precipitation (color shading) and surface wind (streamlines) are presented in this figure at a 6-hourly interval. Results based on the M180 and TRMM/ERA-40 data are plotted in the left and right panels, respectively. The daily averages of precipitation and surface wind have been removed from the corresponding data at individual times of day. Hence, the features mentioned in the following discussion are to be interpreted as deviations from the daily mean condition.

In the observed atmosphere, onshore flows are well established in this region in the late afternoon (Fig. 6e), with prevalent convergence (divergence) over the land.
FIG. 5. Time–longitude distributions of precipitation (shading, see scale at bottom) and deviations of surface air temperature from the 11-day running mean of daily averaged data (contours; interval of 1°C; solid and dashed contours indicate positive and negative values, respectively) during selected 10-day periods in the June–July–August season of the M180 simulation. Results are based on averages of 3-hourly data over the latitudinal zones between (a) 10°–15°N in the India–BoB sector, (b) 12.5°–17.5°N in the Indochina–northern SCS sector, and (c) 5°–7.5°N in the Peninsular Malaysia–southern SCS sector. Blue (red) tick marks on the time axis indicate 0300 (1500) UTC, corresponding to approximately 0830 (2030) LT at 82.5°E in the BoB sector, 1030 (2230) LT at 112.5°E in the northern SCS sector, and 1000 (2200) LT at 105°E in the southern SCS sector. Blue and red local time labels are shown only at the beginning and end of the selected 10-day periods. Red lines indicate longitude of land–sea boundaries. A 3-h adjustment has been applied to the TRMM data (see section 2).
FIG. 6. Distributions of the deviations from the daily mean of the climatological precipitation (shading, see scale at bottom) and surface wind circulation (streamlines), as computed using (left) M180 output and (right) TRMM/ERA-40 data for the June–July–August season. Results are shown for (a),(e) 1200 UTC (1700 LT); (b),(f) 1800 UTC (2300 LT); (c),(g) 0000 UTC (0500 LT); and (d),(h) 0600 UTC (1100 LT). Local times in parentheses pertain to sites at 75°E. In this figure and Figs. 7–10, topographic heights of 250, 500, and 750 m are indicated using yellow, red, and blue contours, respectively. Elevations of 1000, 2000, 3000, 4000, and 5000 m are depicted using progressively darker purple contours. A 3-h adjustment has been applied to the TRMM data (see section 2).
(ocean) areas. This circulation pattern is accompanied by wet conditions over land, and dryness over BoB and the eastern Arabian Sea. Near midnight (Fig. 6f), weak precipitation departures are observed over much of India, and offshore flows develop along the western coast of this subcontinent. The only remnants of positive precipitation departures at this time are found off the southeastern coast of India. In early morning (see Fig. 6g), much of BoB is under the influence of an offshore flow regime. The negative precipitation departures over the Indian landmass attain maximum amplitudes. The positive maritime precipitation center originating from the eastern Indian coast expands and intensifies as it migrates east-northeastward by about 300 km during the 6-h period after midnight (Figs. 6f,g). A cyclonic circulation prevails in this wet region in early morning (Fig. 6g). The movement of this positive precipitation feature toward the open waters of BoB is discernible through noontime (Fig. 6h), when the local flow pattern is distinctly convergent. At noontime, the precipitation departures over much of India weaken considerably. However, coherent negative departures are still identifiable over the southeastern Indian coast. The subsequent amplification and eastward displacement of this dry feature and the ambient circulation during the following 12 h (Figs. 6e,f) is analogous to the evolution of the wet feature over the same location during the morning-to-noon period (Figs. 6g,h), as described above.

Many of the observed diurnal characteristics described in the preceding paragraph are reproduced in the GCM (Figs. 6a–d), particularly in regards to timing of the phases of wet/dry conditions and onshore/offshore flow over various land and oceanic sites, and development of the eastward-propagating features over the western BoB. However, the migration of the precipitation center over the waters east of Sri Lanka in the model atmosphere is not supported by the satellite and reanalysis datasets. The observational results also indicate substantial diurnal precipitation changes over the maritime zones along the western coasts of Myanmar and India. These signals are observed to travel in an offshore direction toward the open waters of northeastern BoB, the Andaman Sea, and the eastern Arabian Sea. Such phenomena are not discernible in the model patterns. The observed precipitation features in the northern portion of BoB exhibit a tendency to migrate equatorward with time (see also Webster et al. 2002). This behavior is not as evident in the model atmosphere.

Considering the northeastward orientation of the climatological monsoon flow over South Asia, the western BoB (eastern Arabian Sea) is situated downstream (upstream) of the Indian subcontinent. In the model atmosphere, the presence of seaward migration of precipitation over western BoB, and absence of such features over eastern Arabian Sea (see Figs. 6a–d), suggest that the mean monsoon circulation exerts a strong influence on the subsequent development of convection initiated over land. Specifically, this mean flow pattern promotes (inhibits) the propagation of the precipitation signals downstream (upstream) from their source region. Analogous arguments may also be invoked to interpret the stronger propagation signals simulated over SCS (which lies on the downstream side of Indochina and Peninsular Malaysia), as compared to the weaker signals over the Andaman Sea and the eastern equatorial Indian Ocean (which are located upstream of Indochina and Sumatra). This distinct asymmetry in the strength of the maritime precipitation features downstream and upstream of the mean background flow is much less evident in the corresponding observational results (Figs. 1b, 2b, and 6e–h), which indicate seaward developments both to the east and west of the convective sources over land. The isotropic nature of the propagation as inferred from TRMM data indicates that other mechanisms, such as gravity wave emanation forced by continental latent heating (e.g., Mapes et al. 2003), could also contribute to this phenomenon in the observed atmosphere.

6. Diurnal features over other Asian sites

With the exception of the diurnal variations along relatively narrow coastal and mountainous zones (see section 3) and propagating signals over several maritime regions (section 5), the DC over a large portion of the Asian sector is dominated by maximum precipitation in afternoon over land and by precipitation peaks in morning over ocean (e.g., see Fig. 2). Kikuchi and Wang (2008) have referred to these broadscale characteristics as the “continental” and “oceanic” regimes. The “seesaw” between the wet and dry extremes associated with the DC is illustrated in Fig. 7 in various locations of Southeast Asia. The model patterns for the departures of climatological precipitation and surface wind flow from the daily mean conditions are displayed for (Fig. 7a) 0900 and (Fig. 7b) 2100 UTC, which correspond to 1700 and 0500 LT at 120°E, respectively. The corresponding precipitation charts based on TRMM data are presented in

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1 The hours of the day (i.e., 0900 and 2100 UTC) for constructing the patterns in Fig. 7 are chosen on the basis of two factors: their proximity to the extreme phases of the local DC and the availability of the model and observational precipitation datasets only at 3-h intervals. The selected synoptic hours might not match the extreme phases exactly because of limited temporal resolution of the datasets. Analogous considerations are made in the production of Figs. 8–10.
Figs. 7c,d. Surface streamline patterns are not shown in the latter panels because wind data at 0900 and 2100 UTC are not available from the 6-hourly ERA-40 reanalyses. Both the observed and model patterns in Fig. 7 depict strong land–sea contrasts in the phase of the DC in precipitation. The afternoon maxima and early morning minima over land, as well as the phase reversal of the precipitation cycle over the nearby ocean areas, are noticeable on many spatial scales, ranging from the broad subcontinental regime (e.g., southern China and Indo-China) to the dimensions of smaller islands in the Indonesian Archipelago. This model is also seen to generate onshore (offshore) surface flows in afternoon (early morning) in various coastal zones. The diurnal characteristics
of such circulations are discernible on spatial scales as small as individual islands of the Maritime Continent and the Philippines.

The observed diurnal rainfall patterns over parts of Sumatra and Borneo, as well as the nearby waters, are not well reproduced by M180. Specifically, the observed nighttime precipitation peaks at the equator over Borneo, as well as the seaward migration of diurnal signals off the western Sumatra and northern Borneo coasts, are not evident in the model atmosphere (see also the phase distribution in this region in Fig. 2). Hara et al. (2009) reported that the simulation by a GCM at the Meteorological Research Institute exhibits similar discrepancies from the TRMM data. These authors concluded that the fidelity of the simulation of such detailed diurnal features depends crucially on the treatment of cumulus convection in the model. Differences between the model and observational results are also notable along the southeastern edge of the Tibetan Plateau (near 25° N, 90°–100°E) and the northern portion of Indochina.

We next turn our attention to the Tibetan Plateau region, where departures from the daily mean of precipitation and surface flow patterns in late afternoon are generally opposite to those in early morning (not shown). This phase reversal is accentuated by the difference charts in Fig. 8, which are obtained by subtracting the climatology at 0000 UTC from those at 1200 UTC for (Fig. 8a) M180 simulation and (Fig. 8b) TRMM/ERA-40 data. These synoptic times correspond to 0600 and 1800 LT at 90°E, respectively. Hence, the difference charts in Fig. 8 highlight the prevalent features in late afternoon. The principal terrain features and river systems in this region are depicted in Fig. 8c.

In late afternoon, the observed surface flow (Fig. 8b) is directed up the slope of the local terrain. The diurnal wind pattern in the vicinity of the Tibetan Plateau is similar to that presented by Luo and Yanai (1983, see their Fig. 4b). These anabatic flows (and the accompanying “valley winds”), which are primarily driven by differential solar heating at the mountain slopes [e.g., see Figs. 3–4 of Nitta (1983) and Fig. 9.31 of Wallace and Hobbs (2006)], lead to convergence and enhanced precipitation near the apex of the major mountain ranges, as well as divergence and reduced precipitation over valleys and basins. Such orographic influences on the DC of precipitation are exemplified by the observed wet afternoon conditions near the Himalayan, Heng–Duan (HD), and Indo–Burma (IB) ranges, and northern portion of the Indian Plateau (NIP).²

On the contrary, dry afternoon conditions prevail over the relatively low-lying areas in this region. The latter sites include the valleys and flood plains of the major river systems in South Asia (e.g., Indus, Ganges, Brahmaputra, and Irrawaddy), as well as large-scale depressions [e.g., Sichuan Basin (SB)]. The observed circulation and precipitation patterns at 0600 LT, as may be inferred by reversing the polarity of the changes depicted in Fig. 8b, are dominated by katabatic winds down the mountain slopes, convergence and wetness over valleys and basins, as well as divergence and dryness near mountain peaks. Similar impacts of the terrain in the Himalayan region on the DC of the local precipitation have been noted in the observational studies of Barros and Lang (2003) and Fujinami et al. (2005).

Some of the observed diurnal changes noted above are reproduced in the model atmosphere (Fig. 8a). Particularly noteworthy is the simulation of the contrasts between flow patterns in afternoon and early morning in the vicinity of the major orographic features. The model mimics the DC in precipitation over the elevated plateaus and ridges. The negative precipitation departures in late afternoon in the upper reaches of the Indus, Brahmaputra, and Irrawaddy Rivers, as well as the Sichuan Basin, are also evident in the model results. However, the corresponding rainfall signals over much of the Ganges Valley are weaker in the M180 pattern than in the TRMM pattern.

7. Diurnal features over the tropical Americas and Africa

To complete our survey of the DC in various tropical sites of the globe during northern summer, we proceed to examine afternoon-minus-morning difference charts similar to those in Fig. 8 for two selected regions outside of Asia. The model and observational patterns for a portion of the Americas, as constructed using data for 0000 and 1200 UTC, are presented in Fig. 9. These synoptic times correspond to 1800 and 0600 LT at 90°W, respectively. The results based on TRMM/ERA-40 data (Fig. 9b) indicate that the onshore flows in late afternoon over most coastal regions are accompanied by divergence over the northern Gulf of Mexico, subtropical western North Atlantic, and the Pacific waters off the west coast of Central America. Convergence prevails over the southeastern United States, Cuba, and western Hispaniola, as well as near the continental divide extending through the length of Central America. This flow pattern is associated with mostly positive (negative) precipitation departures over land (ocean). The precipitation pattern in Fig. 9b is in general agreement with that documented by Garreaud and Wallace (1997) using

² The locations of various topographic features are indicated in Fig. 8c using the abbreviated labels given in parentheses in this discussion.
Fig. 8. Distributions of the difference between 1200 UTC (1800 LT) and 0000 UTC (0600 LT) of climatological precipitation [shading, see scale below (b)] and surface wind circulation (streamlines), as computed using (a) M180 output and (b) TRMM/ERA-40 data for the June–July–August season. Local times in parentheses pertain to sites at 90°E. A 3-h adjustment has been applied to the TRMM data (see section 2). (c) Relief map indicating locations of orographic features (see abbreviated black labels) and river systems (see blue labels) mentioned in the text. Scale bar for topographic heights in various ranges is shown on the right side of (c).
FIG. 9. Distributions of the difference between 0000 UTC (1800 LT) and 1200 UTC (0600 LT) of climatological precipitation (shading, see scale at bottom) and surface wind circulation (streamlines) in the tropical Americas, as computed using (a) M180 output and (b) TRMM/ERA-40 data for the June–July–August season. Local times in parentheses pertain to sites at 90°W. A 3-h adjustment has been applied to the TRMM data (see section 2).
satellite-based data for convective cloudiness and microwave imageries. The late-afternoon/early-evening precipitation maximum in the TRMM pattern over the ascending slope of the Sierra Madre Occidental (25°–30°N, 108°W) is consistent with the satellite observations reported by Tian et al. (2005). These authors also noted that the strong DC observed along the coastal zone of the Gulf of Mexico, Cuba, as well as the Floridian and Yucatan Peninsulas are the result of sea-breeze fronts.

Essential aspects of the observed diurnal flow patterns in Fig. 9b are well reproduced in the M180 simulation (Fig. 9a). The model resolution is sufficiently high to generate such details as the late afternoon divergence (convergence) over the Gulf of California (Baja California). The simulated negative precipitation departures in late afternoon over most coastal maritime regions have lower amplitudes and less spatial extent when compared with the TRMM results. Differences between model and observations are also discernible along the western slope of Sierra Madre Occidental, and at several isolated sites along the Pacific coast of Central America.

The diurnal changes in precipitation and surface circulation over tropical Africa, as obtained by subtracting the data for 0000 UTC from 1200 UTC, are shown in Fig. 10 for (Fig. 10a) M180 output and (Fig. 10b) TRMM/ERA-40 analyses. These synoptic times correspond to 0100 and 1300 LT at 15°E, respectively. Near noontime, the observed prevalence of onshore winds from the
Atlantic and outflows from the Sahara Desert leads to notable landside convergences along the coasts of western Africa and the Gulf of Guinea (Fig. 10b). Much of eastern Africa is under the influence of onshore flows. Enhanced daytime precipitation is observed over land, with local maxima along the western portion of the African landmass between 5° and 12°N, as well as several interior continental sites near elevated orographic features (e.g., the Ethiopian Plateau at 5°–15°N, 35°–40°E). Increased rainfall is observed during night hours over the coastal waters of tropical western Africa and Gulf of Guinea, as well as Lake Victoria (0°–3°S, 32°–34°E).

The large-scale characteristics of the observed diurnal wind patterns are discernible in the model result (Fig. 10a). The simulated DC of precipitation has relatively lower amplitudes over ocean points in the Atlantic.

8. Conclusions and discussion

The performance of a global GCM with 50-km horizontal resolution in simulating the DC of precipitation and surface circulation is assessed. The model patterns are verified against corresponding results based on satellite observations and reanalysis products with comparable resolutions. Attention is devoted to diurnal features occurring in northern summer over the Asian monsoon region, as well as the American and African sectors within the tropical zone. It is seen that the M180 model is capable of reproducing the large-scale characteristics of the land–sea contrasts in the peak phase of the diurnal precipitation changes, and in the orientation of the surface flow (Figs. 2 and 6–10). Systematic offshore propagation of precipitation signals during evening and morning hours over BoB and western SCS is discernible in both M180 and TRMM datasets (Figs. 2 and 5–6). Model statistics indicate the initiation of low-intensity precipitation over tropical and subtropical land regions in the early hours of the day, followed by heavier precipitation in the middle and late afternoon (Figs. 4a,b). Some aspects of this evolution are supported by the observations (Figs. 4c,d). However, the occurrence of precipitation with very high intensity in midafternoon, as inferred from the TRMM data, is not evident in the model simulation.

The M180 model is less successful in reproducing the fine regional details of the DC. Notable discrepancies between model and observations exist with regards to the timing and amplitude of the peak phase in precipitation over various orographic features, such as the Ganges Valley at the foothills of the Himalayan Range (Figs. 2 and 8) and the western edge of Sierra Madre Occidental (Fig. 9). The observed tendency for maritime precipitation signals to migrate offshore from the coasts of western Myanmar, western India, and various islands of the Indonesian Archipelago is not evident in the model atmosphere (Figs. 2 and 6). Whereas some of the above discrepancies are reduced by increased model resolution (Fig. 3c), improvements in the representation of physical processes in the model atmosphere are also needed to enhance model performance in mimicking the DC. The findings presented in this and other model diagnostic studies (e.g., Yang and Slingo 2001; Lee et al. 2007a,b; Hara et al. 2009) suggest that the following processes play a crucial role in simulations of the DC: heating/cooling mechanisms and their interactions with the flow field in the vicinity of sloping terrain, diurnal evolution of the atmospheric boundary layer and flux exchanges with the underlying land–ocean surfaces, and couplings of various local circulation features (such as land–sea-breeze and gravity waves) with moist convection. The absence of diurnal variability in the prescribed SST forcing in the M180 integration could also affect the simulation of DC in various maritime sites (Dai and Trenberth 2004).

In recent years, there is growing societal demand for information on fine regional details of climate variability and change. High-resolution GCMs are an important tool for understanding and projecting the behavior of the climate system on such small spatial scales. In this study, the adequacy of these models for such applications is critically examined by evaluating their ability to reproduce the characteristics of a quintessential component of regional climates—the DC of precipitation and near-surface atmospheric circulation. We have focused on this particular facet of model performance in view of the strong impacts of the diurnal phenomena on the mean climates in many localities, and the large diversity of processes that contribute to the diurnal features. Diagnosis of the DC in model atmospheres is an efficient way to test the fidelity of the parameterization schemes for these processes. The model discrepancies identified in the present study will facilitate future efforts to incorporate the pertinent processes in the model framework with a higher degree of accuracy.

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