

The Diurnal Cycle and Its Depiction in the Community Climate System Model

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ABSTRACT

To evaluate the performance of version 2 of the Community Climate System Model (CCSM2) in simulating the diurnal cycle and to diagnose the deficiencies in underlying model physics, 10 years of 3-hourly data from a CCSM2 control run are analyzed for global and large-scale features of diurnal variations in surface air temperature, surface pressure, upper-air winds, cloud amount, and precipitation. The model-simulated diurnal variations are compared with available observations, most of which were derived from 3-hourly synoptic reports and some new results are reported for surface air temperatures. The CCSM2 reproduces most of the large-scale tidal variations in surface pressure and upper-air winds, although it overestimates the diurnal pressure tide by 20%–50% over low-latitude land and underestimates it over most oceans, the Rockies, and other midlatitude land areas. The CCSM2 captures the diurnal amplitude (1° – 6°C) and phase [peak at 1400–1600 local solar time (LST)] of surface air temperature over land, but over ocean the amplitude is too small ($\leq 0.2^{\circ}\text{C}$). The CCSM2 overestimates the mean total cloud amount by 10%–20% of the sky from $\sim 15^{\circ}\text{S}$ to 15°N during both December–January–February (DJF) and June–July–August (JJA) and over northern mid- and high-latitude land areas in DJF, whereas it underestimates the cloud amount by 10%–30% in the subtropics and parts of the midlatitudes. Over the marine stratocumulus regions west to the continents, the diagnostic cloud scheme in the CCSM2 underestimates the mean stratocumulus amount by 10%–30% and does not simulate the observed large diurnal variations ($\sim 3\%$ – 10%) in the marine stratocumulus clouds even when driven by observational data. In the CCSM2, warm-season daytime moist convection over land starts prematurely around 0800 LST, about 4 hours too early compared with observations, and plateaus from 1100 to 1800 LST, in contrast to a sharp peak around 1600–1700 LST in observations. The premature initiation of convection prevents convective available potential energy (CAPE) from accumulating in the morning and early afternoon and intense convection from occurring in the mid to late afternoon. As a result of the extended duration of daytime convection over land, the CCSM2 rains too frequently at reduced intensity despite the fairly realistic patterns of rainy days with precipitation $> 1\text{ mm day}^{-1}$. Furthermore, the convective versus nonconvective precipitation ratio is too high in the model as deep convection removes atmospheric moisture prematurely. The simulated diurnal cycle of precipitation is too weak over the oceans, especially for convective precipitation. These results suggest that substantial improvements are desirable in the CCSM2 in simulating cloud amount, initiation of warm-season deep convection over land, and in the diurnal cycle in sea surface temperatures.

1. Introduction

Solar heating near the surface and in the atmosphere generates strong diurnal (i.e., 24 h) and subdiurnal (e.g., 12-h semidiurnal) oscillations in surface and atmospheric temperature, pressure, and wind fields. These regular oscillations, which are often referred to as atmospheric tides (Chapman and Lindzen 1970; Dai and Wang 1999), are among the most pronounced signals of earth's climate and weather. Considerable diurnal variations, modulated by synoptic weather events, are also found in moist convection and cloudiness (e.g., Hendon and

Woodberry 1993; Rosendaal et al. 1995; Bergman and Salby 1996; Sui et al. 1997; Garreaud and Wallace 1997; Yang and Slingo 2001), precipitation (e.g., Wallace 1975; Oki and Musiake 1994; Janowiak et al. 1994; Chang et al. 1995; Dai et al. 1999a; Dai 2001b; So-rooshian et al. 2002; Nesbitt and Zipser 2003), and atmospheric water vapor (Dai et al. 2002). These diurnal variations affect surface and atmospheric fluxes of energy (Bergman and Salby 1997), water (Trenberth et al. 2003), and momentum (Dai and Deser 1999). For example, because surface solar heating typically has a sharp midday peak, precipitation that occurs during the day evaporates from the surface more rapidly than that at night. This changes the fraction of rainfall involved in runoff versus evaporation and the partitioning of surface energy into latent and sensible heat fluxes. Climate models without a diurnal cycle cannot simulate these nonlinear processes, resulting in degraded model simulations (Wilson and Mitchell 1986).

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Because of its large amplitude, coherent phase, and short time scale, the diurnal cycle provides an excellent test bed for evaluating model physics in weather forecast and climate models (Lin et al. 2000; Trenberth et al. 2003). Moreover, Trenberth et al. (2003) argue that precipitation intensity is related to atmospheric water vapor amounts through moisture convergence. Accordingly, in a warmer climate, where the amount of moisture in the air is expected to rise faster than the total precipitation amount, large increases in precipitation intensity must be offset by decreases in precipitation frequency or duration. It is therefore important to be able to correctly simulate precipitation characteristics in the models, and the diurnal cycle allows these to be systematically explored.

Most current atmospheric general circulation models (AGCMs) calculate atmospheric radiation at about hourly time intervals [e.g., the version 3 of the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM3), Kiehl et al. 1998], which is sufficient for resolving most of the atmospheric diurnal or subdaily variations. AGCMs are often coupled to a land surface model that also resolves the land surface diurnal cycle. In uncoupled simulations without an ocean general circulation model (OGCM), AGCMs are often forced with specified sea surface temperatures (SSTs) that do not contain diurnal variations. This is true in all AMIP (Atmospheric Model Intercomparison Project)-type simulations (Gates et al. 1999), where monthly mean SSTs are used as the marine lower boundary condition.

In coupled simulations, the AGCM interacts with an OGCM. As has become apparent from this study, no current OGCM can simulate the diurnal cycle in surface oceans. This is because OGCMs distribute the solar heating evenly within the depth (≥ 10 m) of the top model layer. Ocean mixed layer models (Large et al. 1994) with layer thickness less than 1 m have not been incorporated into coupled general circulation models (CGCMs), partly because dynamic instability arises when the top model layer is too thin in a free surface OGCM (W. G. Large 2003, personal communication). Furthermore, the coupling between an AGCM and OGCM is not simultaneous as one of them has to be computed first using finite time steps. This time lag often exceeds one hour, which is significant for the diurnal cycle. The coupling frequency is also too low (e.g., once per day) for diurnal simulations in some CGCMs. Because of these model characteristics, the diurnal cycles in surface air and ocean temperatures (Lukas 1991; Zhang 1995; Webster et al. 1996; Zeng et al. 1999), surface latent and sensible heat fluxes (Zeng and Dickinson 1998), and other near-surface fields over the oceans cannot be simulated well in CGCMs.

Previous analyses of diurnal variability in weather forecast and climate models focus mostly on the tidal variations, in particular the surface pressure tides (Zwiers and Hamilton 1986; Hsu and Hoskins 1989;

Lieberman et al. 1994; Van den Dool et al. 1997), and show that atmospheric models can reproduce many of the large-scale tidal features. Studies of the diurnal cycle in precipitation, cloudiness, and energy fluxes in climate models (Slingo et al. 1987; Randall et al. 1991; Dai et al. 1999a; Lin et al. 2000; Yang and Slingo 2001), however, show that the diurnal cycle of the hydrological processes is still a big challenge for AGCMs and regional climate models. For example, the U.K. Met. Office's AGCM (Yang and Slingo 2001) and ECMWF model (Betts and Jakob 2002a; Trenberth et al. 2003) produce peak precipitation before or near noon over most land areas, in contrast to late afternoon to evening peaks in observations (Dai 2001b). Lin et al. (2000) found that cumulus parameterizations for *deep* and *shallow* convection need to differ from each other in order to capture some aspects of the diurnal cycle of tropical precipitation in the Colorado State University (CSU) AGCM. Many regional features of the diurnal cycle of warm seasonal precipitation, such as the nocturnal precipitation maximum over the central United States (Wallace 1975), are absent in all published climate model simulations. For example, Dai et al. (1999a) showed that none of three well-known moist convection schemes were able to reproduce the broad geographical pattern of the diurnal cycle of summer precipitation over the United States in a regional climate model.

To provide a comprehensive evaluation of the diurnal variability in a state-of-the-art coupled climate model—the Community Climate System Model (CCSM; Blackmon et al. 2001), we have compiled and analyzed a number of observational datasets with high temporal resolution and compared them with the model-simulated variations in various fields. We have also diagnosed the causes of some of the major diurnal deficiencies in the model. The diurnal analysis not only provides an evaluation of model performance (in particular for moist convection, cloud formation, and other hydrologic parameterizations), but also is a necessary step before applying the CCSM to investigate the physical mechanisms underlying the observed diurnal variations (e.g., Randall et al. 1991). Our results should be of interest not only for improving climate models, but also because some of our observation-based diurnal variations (e.g., those of global surface air temperature) have not been reported previously. In this paper we focus on the global and large-scale features of the CCSM-simulated diurnal variations in surface air temperature, pressure, precipitation, clouds, and other atmospheric fields. We plan to report more detailed diagnostic analyses focusing on selected regions in a separate paper.

2. Model, data, and analysis methods

The model examined here is version 2 of the CCSM (CCSM2; Blackmon et al. 2001), which is a fully coupled climate system model. It consists of an AGCM, and OGCM, a comprehensive land surface model (Bon-

TABLE 1. Observational datasets used in this study for evaluating model diurnal variability. All are 3-hourly, global, and averaged over the period, except as otherwise stated. COADS stands for Comprehensive Ocean–Atmosphere Data Set (see <http://www.cdc.noaa.gov/coads/index.shtml>).

Variables	Type	Resolution	Period	Source/reference
Surface pressure	From 7537 stations and COADS	4° lat × 5° lon	1976–97	Dai and Wang (1999)
Surface air temperature	From ~15 000 stations and corrected T_a from COADS	2° lat × 2.5° lon	1976–97	See Dai and Deser (1999) for data source
Precipitation frequency	Reports from ~15 000 stations and from COADS	2° lat × 2° lon	1975–97	Dai (2001a,b)
Low and total cloud cover	Station and marine observations	4° lat × 5° lon	1971–96	Hahn and Warren (1999)
Cloud amounts	Satellite observations	2.5° lat × 2.5° lon	1983–2001	Rossow et al. (1996)
Atmospheric T , q and winds	NCEP–NCAR reanalysis	6-hourly, T63 (~1.875°)	1980–96	Kalnay et al. (1996)

an et al. 2002; Zeng et al. 2002), and a sophisticated sea ice model (Briegleb et al. 2002). The CCSM2 does not use flux adjustments. It has a small trend in surface temperature in a 1000-yr control run that is comparable to those of flux-corrected GCMs (e.g., Stouffer et al. 2000) and the Parallel Climate Model (Dai et al. 2003, manuscript submitted to *J. Climate*). A detailed description of the CCSM2 has yet to be published by the CCSM working groups. Here we briefly outline its main features relevant to the diurnal cycle.

The AGCM in the CCSM2 is the NCAR Community Atmosphere Model (CAM2), which is described in detail by Collins et al. (2003). The CAM2 is an improved version of the CCM3 (Kiehl et al. 1998), but retains the T42 spectral grid (~2.8° grid size). The major modifications (Collins et al. 2003) include prognostic cloud water, longwave radiative transfer improvements, generalized cloud overlap, improved treatment of sea ice, enhanced evaporation of convective precipitation, improved vertical diffusion of dry static energy, and an increase in vertical levels from 18 to 26. The CAM2 retains the Zhang and McFarlane (1995) parameterization for deep moist convection, while shallow dry convection is treated using the Hack scheme (Hack 1994). The Zhang and McFarlane scheme assumes that an ensemble of convective-scale updrafts may occur whenever the lower atmosphere is conditionally unstable. Moist convection occurs only when there is convective available potential energy (CAPE) for which parcel ascent from the subcloud layer acts to release the CAPE at an exponential rate using a specified adjustment time scale (= 2 h). Convective inhibition by a stable boundary layer, descending large-scale motion, and other processes (Fu et al. 1994; Weckwerth 2000) are not considered in this scheme. The earlier version (i.e., CCM3) simulates the geographic, seasonal, and interannual variations of mean air temperature, precipitation, and winds fairly well when observed SSTs are used (Kiehl et al. 1998; Hack et al. 1998; Hurrell et al. 1998). CAM2 simulations are improved in many aspects compared with the CCM3 (see Collins et al. for details).

The OGCM in the CCSM2 is based on the Parallel Ocean Program (POP; Smith et al. 1992), with ~1° nominal grid size and 40 vertical layers. The OGCM

uses daily mean values (of the previous day) from the AGCM as surface forcing. While this avoids the effect of diurnal time lag between the AGCM and OCGM, this coupling scheme cannot capture the subdaily variations in air–sea interactions. In the CCSM2, air–sea fluxes (including solar radiation) are distributed evenly within the depth of the top ocean layer (10 m) and the top ocean layer temperature is used as the SST for air–sea coupling. As a result, the SST has almost no diurnal cycle.

Three-hourly instantaneous (i.e., from single time step) fields from a 10-yr period (years 500–509) of a millennial control run using the CCSM2 were averaged to derive the mean diurnal cycle for each season, and then compared with available observations (Table 1). We have compared the diurnal variability in the CCSM2 with that in its earlier version and in AMIP-type simulations (without diurnal variations in the SSTs). As the diurnal variations in these earlier simulations are essentially the same as in the CCSM2, we show only the CCSM2 results.

The comparison with observations was done using both the composite diurnal “anomalies” (i.e., departures from the daily mean) and the estimated diurnal and semidiurnal harmonics [see Dai and Wang (1999) for details of the harmonic analysis]. Our focus is on the mean diurnal cycle in surface air temperature, pressure, cloudiness, and convective, stratiform, and total precipitation because of the availability of observational data with high temporal resolution. Although 3-hourly sampling is sufficient for reproducing a diurnal harmonic and is marginally adequate for capturing a semidiurnal harmonic, substantial errors (probably up to ± 1 h, especially for precipitation) exist in the estimated phases for the diurnal and semidiurnal harmonics due to errors in the harmonic fitting and data averaging (within 1 h for the precipitation data, but close to instantaneous for temperature, pressure, and cloud observations). For the free atmosphere, we compare only a few fields with 6-hourly data from the NCEP (National Centers for Environmental Prediction)–NCAR reanalysis.

Many of the diurnal datasets listed in Table 1 were derived in earlier analyses, which exploited 3-hourly synoptic observations from over 15 000 weather stations

over land and marine weather reports from ships and buoys. These surface observations are the primary source of historical climate records of surface temperature, pressure, winds, clouds, surface humidity, and other variables, even though there are numerous sources of errors related to instrumentation and spatial and temporal sampling (see Dai and Wang 1999; Dai and Deser 1999; and Dai 2001a for details and sampling density maps). The mean diurnal cycles in the surface fields revealed by the synoptic data are reliable over most land and ocean areas, except for the Southern Oceans and the polar regions where sampling is sparse and infrequent.

Surface cloud observations under insufficient moonlight conditions tend to underestimate the cloud cover (by a few percent in seasonal means) and thus distort the diurnal cycle (Hahn et al. 1995). We therefore used only those cloud observations made under sufficient illumination, as defined by Hahn and Warren (1999). We examined the 3-hourly cloud data from the International Satellite Cloud Climatology Project (ISCCP; Rossow et al. 1996) and found similar diurnal variations over the marine stratocumulus and many other regions. The ISCCP mean cloud amount is low (by about 5%–30% of the sky) in the polar regions and within about $\pm 10\%$ in other regions compared with the surface observations (Rossow et al. 1993; Hahn et al. 1995).

To our knowledge, the 3-hourly surface air temperature data have not been used to document the temperature diurnal cycle on a global scale, although Ignatov and Gutman (1999) used air temperature records from four Russian stations to validate the mean diurnal cycle of surface temperature over land derived from the ISCCP dataset. Marine air temperature measurements during sunny days are likely to contain biases due to local solar heating on ship decks. We therefore applied the Kent et al. (1993) correction, which is a function of relative wind speed and incoming solar radiation over the ship based on data collected over the North Atlantic Ocean, although this is likely to be too simple for the global oceans (E. Kent 2002, personal communication). When applied, it shifts the diurnal peak of marine air temperature from noon to midafternoon, consistent with the timing over land and with that in the NCEP–NCAR reanalysis surface air temperature.

3. Results

a. Diurnal cycle of surface air temperature

The diurnal cycle of surface air temperature results primarily from diurnal variations in solar heating near the ground or sea surface, with large modulation by clouds and surface latent and sensible heat fluxes (Dai et al. 1999b). As the stability of the planetary boundary layer (PBL) varies diurnally with the surface solar heating, the temperature diurnal cycle is closely coupled to many processes such as vertical turbulence mixing in

the PBL and thus to the diurnal cycle of atmospheric convection. It is therefore important to simulate the temperature diurnal cycle correctly.

Figure 1 shows the mean diurnal departures from the daily mean of December–January–February (DJF) surface air temperature¹ from surface observations and the CCSM2. In general, the simulated temperature diurnal anomalies are comparable with the observations over most of the continents. For example, the large anomalies (3° – 8°C) over the Tibetan Plateau and the Sahara Desert are captured by the CCSM2. The phase of the westward-propagating wavenumber-1 mode is also close to the observed, despite the weak amplitude over the oceans. This suggests that the diurnal cycle of net surface energy balance is likely to be realistic over most land areas. However, the temperature diurnal cycle over central Africa has incorrect phase and reduced amplitude. The model overestimates cloud cover over central Africa by about 10%–20% (of the sky; cf. Fig. 7), which results in reduced temperature diurnal amplitudes, and the cloud diurnal cycle over the region is also inconsistent with observations (cf. Fig. 8), which likely contributes to the phase error in the surface temperature. However, the excessive clouds over tropical South America and northern Australia (cf. Fig. 7) seem to have smaller effects on the temperature amplitude than over central Africa.

Air temperature diurnal variations over the oceans are too small in the CCSM2 ($\sim 0^{\circ}$ – 0.4°C) compared with observations (0.4° – 1.0°C ; Fig. 1) and the NCEP–NCAR reanalysis (not shown). This is expected because the SSTs in the model lack appreciable diurnal variations (see section 2) and the surface air temperatures are closely coupled to the SSTs.

Figure 2 shows the amplitude and phase [local solar time (LST) of the maximum] of the diurnal harmonic (S_1) of the temperature departures shown in Fig. 1 [also including the June–July–August (JJA) season]. The S_1 dominates the subdaily variation in surface air temperature and accounts for most of its diurnal variance ($>80\%$ over land; cf. Fig. 3). The observed S_1 peaks around 1400–1600 LST and the phase varies little over land and ocean and from winter to summer. The amplitude is much larger over land (1° – 6°C) than over ocean ($\sim 0.5^{\circ}\text{C}$), where evaporative cooling limits daytime surface warming. The amplitude is largest over high terrain during summer resulting partly from radiative cooling at night at high altitudes. The CCSM2 reproduces these broad features. Over the oceans, however, the amplitude in the model ($\leq 0.2^{\circ}\text{C}$) is too small.

Despite the dominance of the diurnal harmonic, the

¹ It refers to the 2-m air temperature in the CCSM2. Station air temperatures are measured around 1.5–2.0 m above the ground. Most buoys measure air temperature around 2-m height, while the height of temperature measurements on boats and ships varies from a few meters to over 20 m above the sea level. No adjustments were made for this height variation.

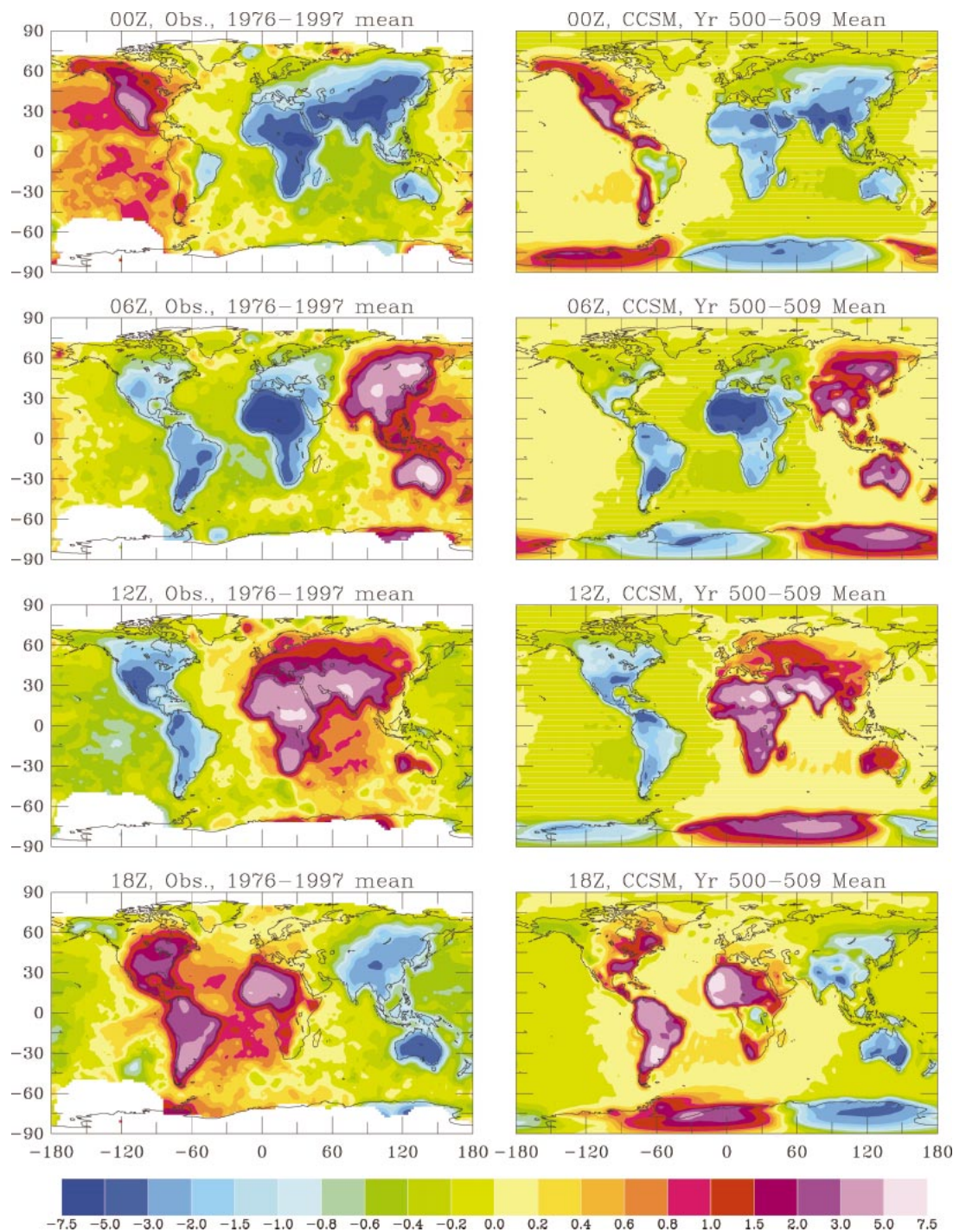


FIG. 1. Mean diurnal anomalies (relative to daily mean) of (left) observed and (right) CCSM-simulated DJF surface air temperature at 0000, 0600, 1200, and 1800 UTC. The color scales are $\pm 0.2, 0.4, 0.6, 0.8, 1.0, 1.5, 2.0, 3.0, 5.0$, and 7.5°C .

semidiurnal harmonic (S_2) of surface air temperature (not shown) has some coherent patterns. The observed S_2 has an amplitude of $0.4^{\circ}\sim 1.5^{\circ}\text{C}$ over land and $0.2^{\circ}\sim 0.4^{\circ}\text{C}$ over ocean, with peaks in early morning (0100–0300 LST) and afternoon (1300–1500 LST) over most land areas; whereas the phase is noisy over the oceans.

The S_2 over land is stronger in winter than in summer, in contrast to S_1 , and is comparable to S_1 over the northern high-latitude land areas and many oceans. The simulated S_2 is slightly weaker than observed during DJF over Eurasia and North America, and is too weak (amplitude $< 0.2^{\circ}\text{C}$) over the oceans, as is the case for S_1 .

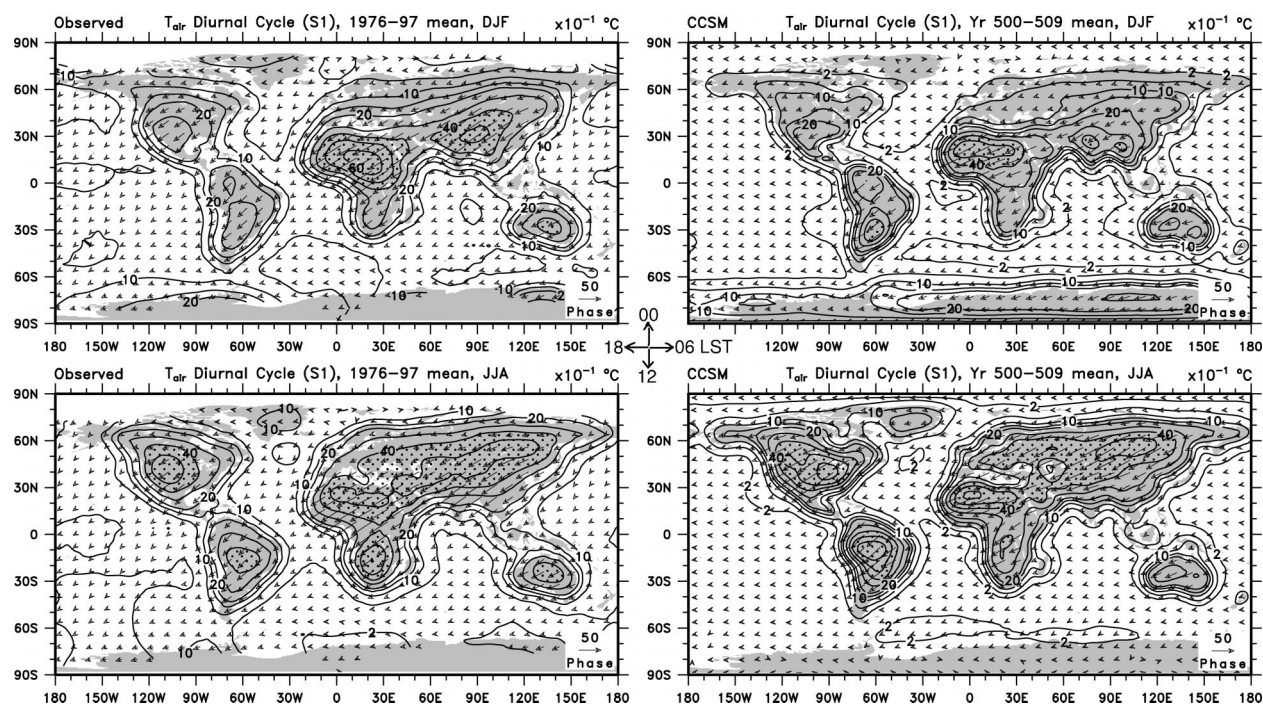


FIG. 2. Amplitude (contours, in 0.1°C) and local solar time at the maximum (arrows, central phase clock) of the diurnal (24 h) harmonic of (top) DJF and (bottom) JJA surface air temperature from (left) observations and (right) the CCSM. Contour levels are 2° , 5° , 10° , 15° , 20° , 30° , 40° , 50° , and 60°C , and values over 4°C are stippled.

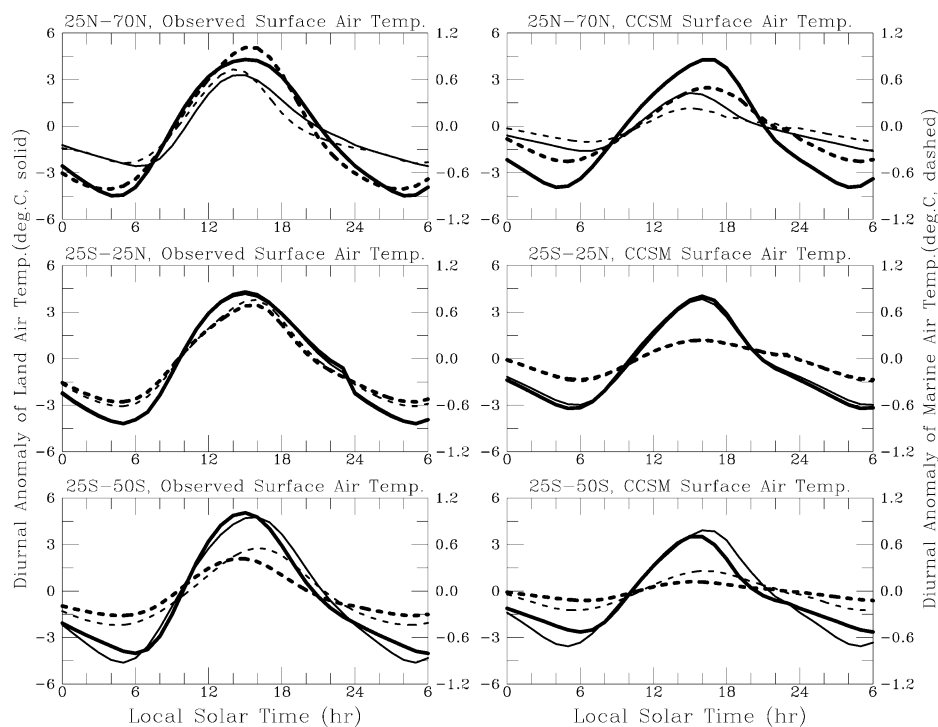


FIG. 3. (left) Observed and (right) CCSM-simulated mean diurnal cycle of surface air temperature (relative to daily mean) area-averaged (at each LST hour) over land (solid lines) and ocean (dashed lines, right ordinate) within three lat zones for DJF (thin lines) and JJA (thick lines).

With coherent diurnal phases, composite diurnal cycles can be derived by averaging (relative to LST) over land and ocean. They (Fig. 3) confirm the dominance of the 24-h harmonic and the peak around 1400–1600 LST. Figure 3 also shows a minimum around 0500 LST in the Tropics and summer extratropics. The minimum occurs later (after 0600) in winter extratropics. The observed diurnal amplitudes over the tropical and northern oceans are only about one-fifth of those over land, whereas this ratio is around one-tenth over 25°–50°S, where landmass is relatively small. The model simulates the phase of the composite diurnal cycles and the amplitude over land reasonably well. However, surface air temperatures over land in the CCSM2 decrease after the midafternoon peak more rapidly than in observations, resulting in slightly skewed diurnal curves. The simulated mean diurnal amplitudes over the oceans are only about half of those observed.

b. Surface pressure tides

Another well-known subdaily variation is the surface pressure tides. The semidiurnal pressure tide is pervasive as it propagates, and results primarily from atmospheric heating due to absorption of solar radiation by ozone and water vapor (Chapman and Lindzen 1970) and, to a lesser extent, latent heat release in tropical convection (Lindzen 1978). In contrast, the diurnal pressure tide is more local and is forced strongly by surface heating associated with terrain.

Figure 4 shows the 6-hourly mean diurnal anomalies of DJF surface pressure from observations and the CCSM2. The model has an atmospheric top at ~2 mb that removes any effects from above the upper stratosphere. The CCSM2 reproduces many regional amplitudes (up to 2 mb) and the westward propagation of wavenumber-2 mode, but it overestimates the amplitudes over tropical land areas (notably, South America; Fig. 4). Figure 5 compares the diurnal pressure tide from observations and the CCSM2. While the overall patterns, such as the larger amplitude over land (0.4–1.4 mb) than over ocean (0.4–0.6 mb), are reproduced, the model overestimates the diurnal amplitude by 20%–50% over low-latitude land areas and underestimates it over most oceans, the Rockies, and other northern midlatitude land areas. The simulated diurnal phase agrees with the observed over most oceans and midlatitude land areas. At low latitudes, the phase is around 0600 LST over both land and ocean in the CCSM2, whereas land lags ocean by ~2 h in the observations. The semidiurnal (S_2) pressure tide (Dai and Wang 1999) is simulated well by the CCSM2 in terms of the amplitude, phase, and seasonal cycle (not shown). For example, both the observations and the CCSM2 show that the semidiurnal pressure tide peaks around 0930–1030 LST (and 12 h later) with an amplitude of 0.8–1.2 mb at low latitudes (30°S–30°N). The model-simulated S_2 is slightly weaker, especially over the eastern Pacific Ocean, and has

less zonal variation than in the observations, and model deficiencies in simulating tropical latent heating associated with the Madden–Julian oscillation (MJO) may be a factor.

c. Diurnal variations in upper-air winds

There exist large diurnal and semidiurnal variations in temperature and wind fields in the free atmosphere (Chapman and Lindzen 1970; Hsu and Hoskins 1989). However, the lack of upper-air data with high temporal resolution prevents detailed analyses of these fields. Figure 6 shows that the CCSM2 reproduces the wavenumber-2 pattern evident in the diurnal anomalies of the 850-mb zonal wind component from the NCEP–NCAR reanalysis for July, even though the magnitude of the anomalies over the oceans is too small. Over land, some large regional anomalies, such as the positive (negative) anomalies over northern Africa at 1200 (0000) UTC and over much of South America at 1800 (0600) UTC, are also reproduced by the CCSM2. Plots at other vertical levels and for the meridional wind component also show similar results; that is, the westward-propagating wavenumber-2 mode is reproduced by the CCSM2 but with too weak diurnal anomalies over the oceans, while many of the regional features over land are captured by the model.

d. Cloud amount

The CCSM2, like other CGCMs, still has substantial biases in seasonal-mean cloud amount over many regions (Fig. 7). Compared with surface and ISCCP cloud amounts, the CCSM2 overestimates the total cloud amount by 10%–20% from ~15°S to 15°N during both DJF and JJA and over northern mid- and high-latitude land areas in DJF, whereas it underestimates the cloud amount by 10%–20% in the subtropics and parts of the midlatitudes (Fig. 7). However, the biggest bias, exceeding 30% in some areas, is over the southern subtropical oceans where the marked contrast between land and ocean in observations is absent in the model. Although cloud amount definitions for the surface observations, ISCCP, and CCSM2 are not entirely consistent with each other (see Rossow et al. 1993; Collins et al. 2003), the resultant differences are much smaller than the model biases shown in Fig. 7.

Maps of low, middle, and high cloud amounts (not shown) reveal that low (below 700 mb) clouds are the main contributor to the diurnal variations in total cloud amount, especially over the marine stratus regions. Both the CCSM2-simulated diurnal variations of total cloud amount and those from surface observations [under sufficient illumination, see Hahn and Warren (1999)] (Fig. 8) suggest a combination of wavenumber 1 and 2 modes moving westward, with large regional variations such as those over the marine stratus regions. However, the model cloud diurnal variations are too large over the equatorial

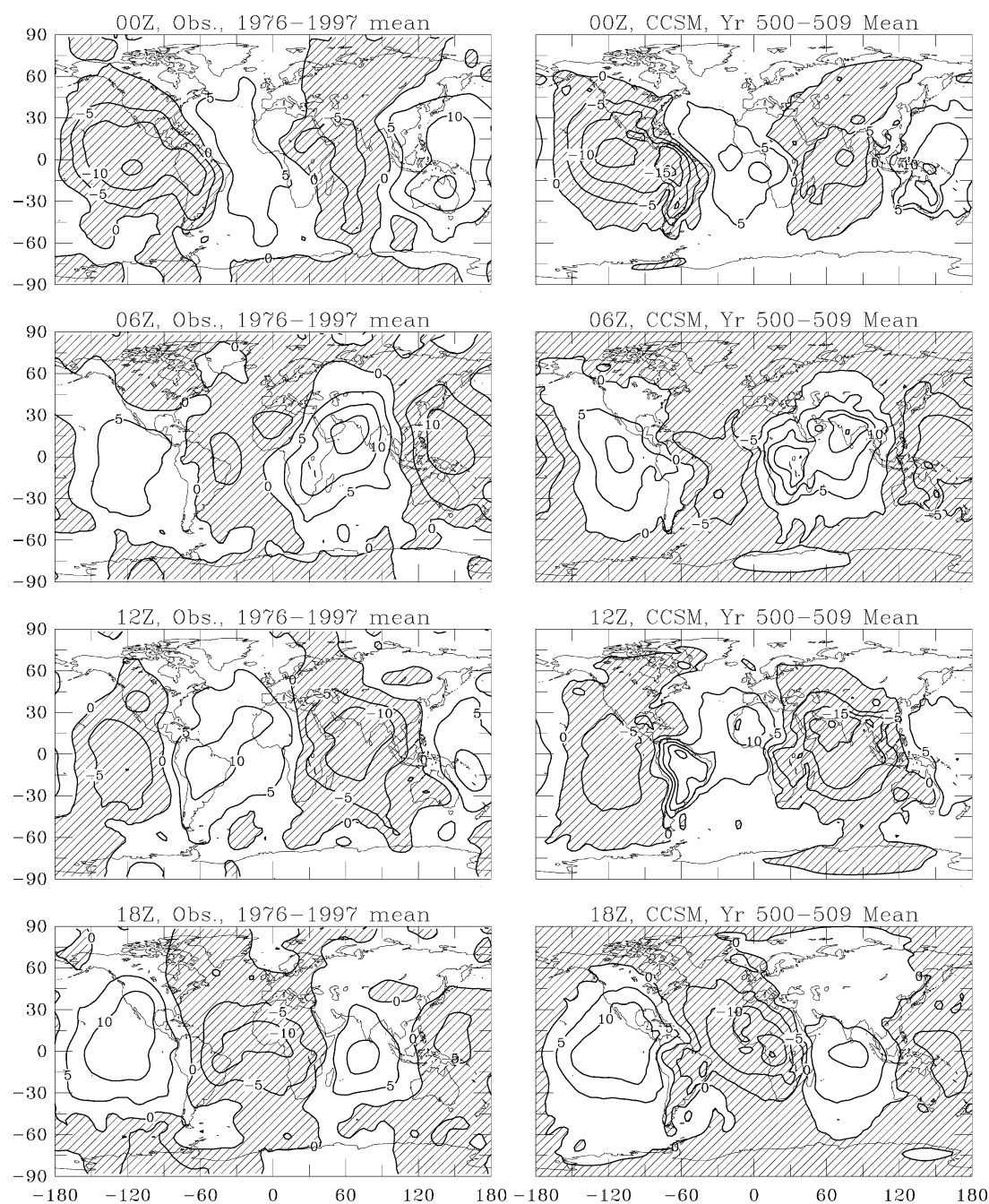


FIG. 4. Mean diurnal anomalies (relative to daily mean) of (left) observed (from Dai and Wang 1999) and (right) CCSM-simulated DJF surface air pressure (in 0.1 mb) at 0000, 0600, 1200, and 1800 UTC.

Pacific and the subtropical North Atlantic. The CCSM2 also fails to capture the observed diurnal phase over the marine stratus regions, where the largest cloud diurnal variations exist (amplitude up to 10%, Fig. 8).

Figure 9 shows the mean diurnal evolution of low cloud amount averaged over the central 15° longitudes of four well-known marine stratus regions (selected based on low cloud diurnal anomaly maps): 1) 0°–32°S, 90°–105°W; 2) 0°–32°S, 10°W–5°E; 3) 20°–45°S, 90°–105°E;

and 4) 15°–45°N, 130°–145°W. The cloud amount for the Southern Hemisphere regions is shown for January, and for the region west to California, July is shown. Consistent with earlier analyses (Warren et al. 1988; Rozendaal et al. 1995), the observed marine stratus clouds peak around 0300–0500 LST with an amplitude around 3%–10% of the sky. The early morning maximum and late afternoon minimum of the marine stratus clouds should enhance the daytime surface solar heating and reduce

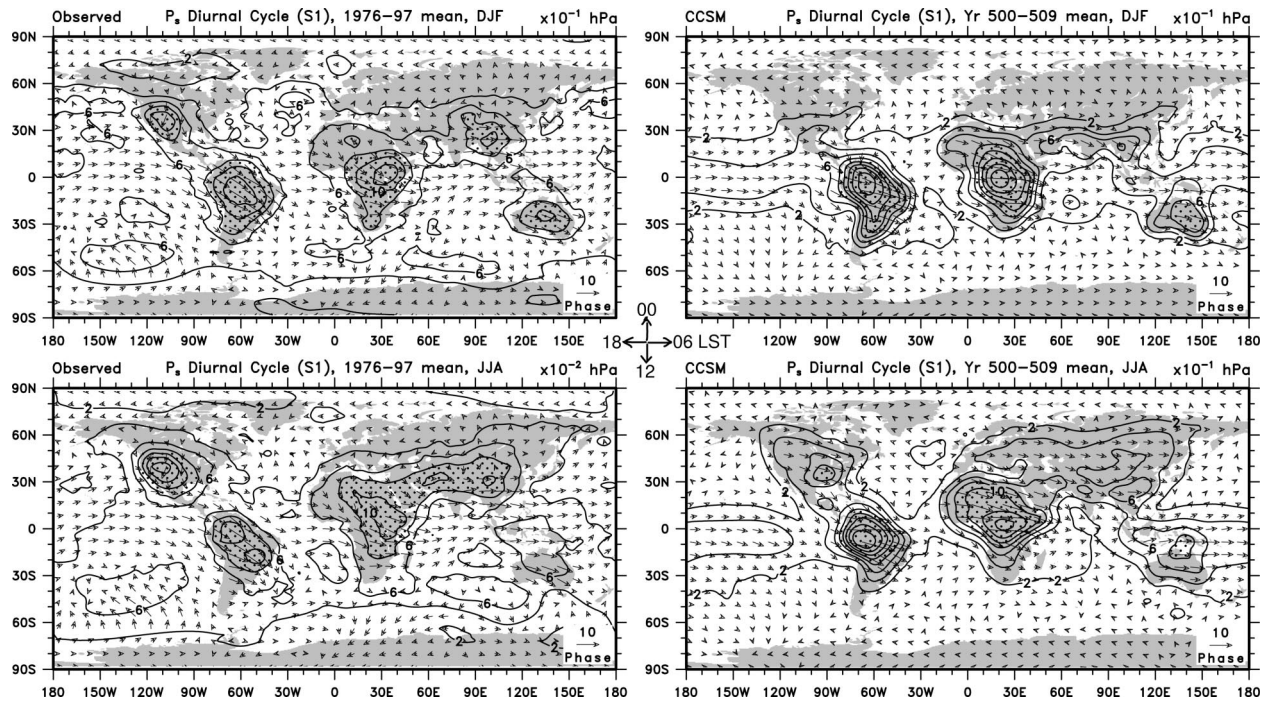


FIG. 5. Amplitude (contours, in 0.1 mb) and local solar time at the maximum (arrows, central phase clock) of the diurnal (24 h) harmonic of (top) DJF and (bottom) JJA surface pressure from (left) observations and (right) the CCSM. The contour interval is 0.2 mb and values over 0.8 mb are stippled.

nighttime surface radiative cooling, resulting in increased surface air and ocean temperatures. Rozendaal et al. (1995) suggest two possible mechanisms for the early morning maximum: 1) absorption of solar radiation within the clouds causes entrainment of dry air from above the PBL that dissipates the stratus clouds during the day and 2) advection of moist boundary layer air from an upwind region favors cloud formation in the morning. Dai and Deser (1999) show that as a diurnal maximum of surface wind convergence (divergence) is in phase with the early morning maximum (late afternoon minimum) of the marine stratus clouds, surface wind convergence plays an important role.

The CCSM2 substantially underestimates (by 10%–30% of the sky) the observed mean low cloud amount over the four marine stratus regions (Fig. 9). Furthermore, the simulated low clouds peak around midnight with relatively weak diurnal amplitudes. As the low clouds dominate the total cloud amount over these regions, these biases in the mean and, to a lesser extent, the diurnal cycle of low clouds² likely distort the energy balances near the ocean surface and within the lower troposphere over the regions.

The CCSM2 uses a diagnostic scheme derived by Klein and Hartmann (1993) using an empirical seasonal relationship found over several regions between marine

stratus cloud fraction (C_{st}) and the potential temperature difference between 700 mb and the surface (Collins et al. 2003): $C_{st} = \min\{1.0, \max[0.0, 0.057(\theta_{700} - \theta_s) - 0.5573]\}$, where θ_{700} and θ_s are the potential temperatures at 700 mb and the surface, respectively. Figure 9 shows that even with observed surface temperature and NCEP–NCAR reanalysis 700-mb temperature (both converted into potential temperature at the 1000-mb reference), this seasonal relationship produced little diurnal variation. Furthermore, substantial negative biases still exist in this calculation, although they are reduced over the regions west of Africa and Australia. This is expected because diurnal variations in the observed potential temperature difference are small over the oceans (amplitude $\leq 1^\circ\text{C}$; cf. Fig. 1) and the relationship does not account for any of the diurnal mechanisms mentioned above. Based on these results, we conclude that this simple seasonal relationship is not capable of simulating the large diurnal cycle in marine stratocumulus clouds, albeit realistic simulations of the stratus diurnal cycle may require improved simulations of the diurnal cycle in local SST and surface winds.

e. Diurnal cycle of precipitation

As argued by Dai (2001a) and Trenberth et al. (2003), precipitation frequency and intensity are important characteristics besides the commonly examined precipitation amount. Figure 10 compares the number of days with precipitation exceeding 1 mm day^{-1} expressed as a per-

² While the peak in clouds around either midnight or 0300–0500 LST has no effect on solar radiation, the daytime minimum allows more sunlight to reach the surface.

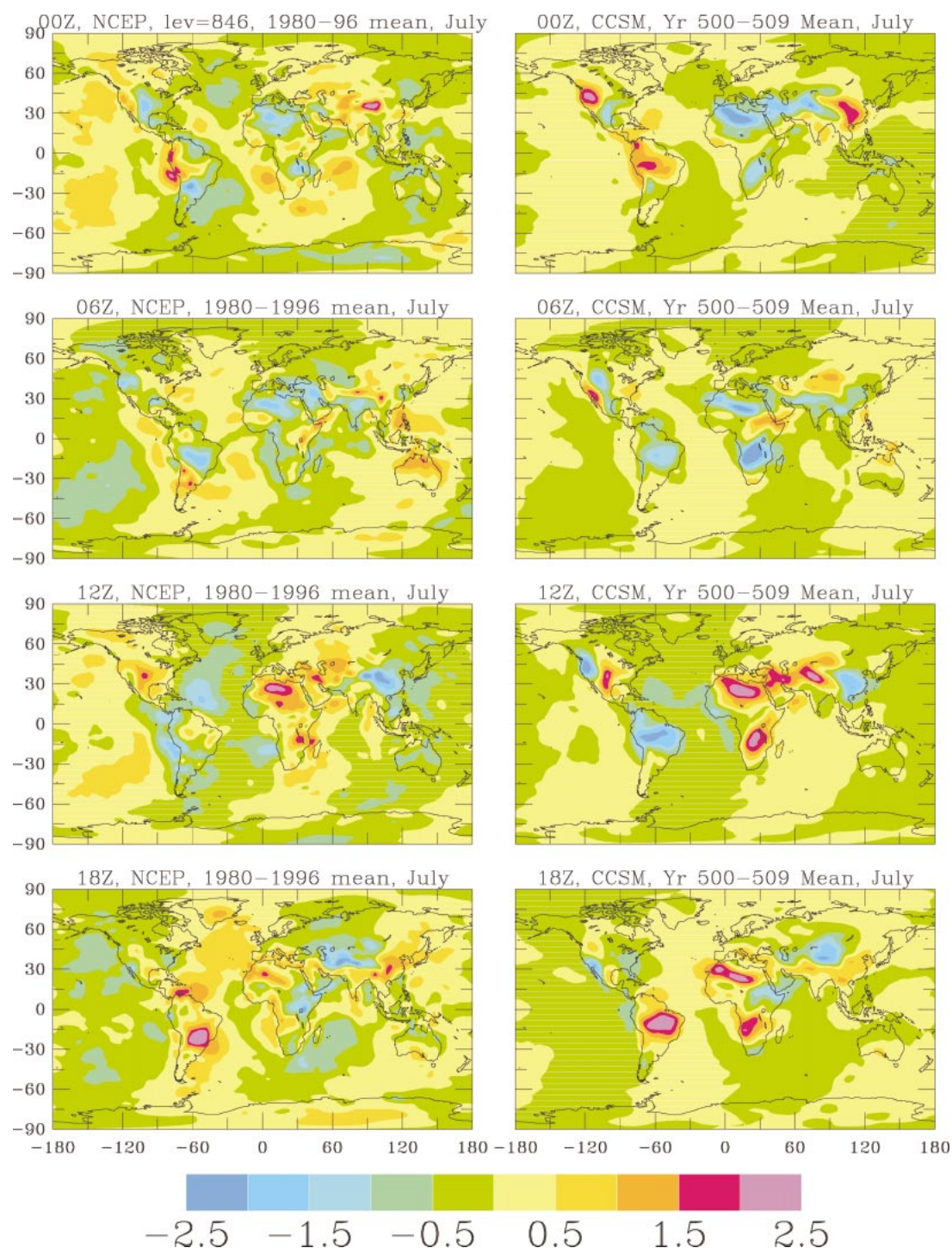


FIG. 6. Mean diurnal anomalies of Jul 850-mb zonal wind component (m s^{-1}) at 0000, 0600, 1200, and 1800 UTC from (left) the NCEP–NCAR reanalysis (1980–96 mean) and (right) the CCSM.

centage of the total number of days in the CCSM2 with the number of days (in %) with one or more reports of nondrizzle precipitation based on synoptic weather reports (Dai 2001a). While these definitions of precipitation frequency are not identical, the weather-report-

based frequency maps are in good agreement with those derived using station rain gauge data using the 1 mm day^{-1} criterion over the United States (Higgins et al. 1996). As long-term records of daily and subdaily precipitation are unavailable over the oceans and not pub-

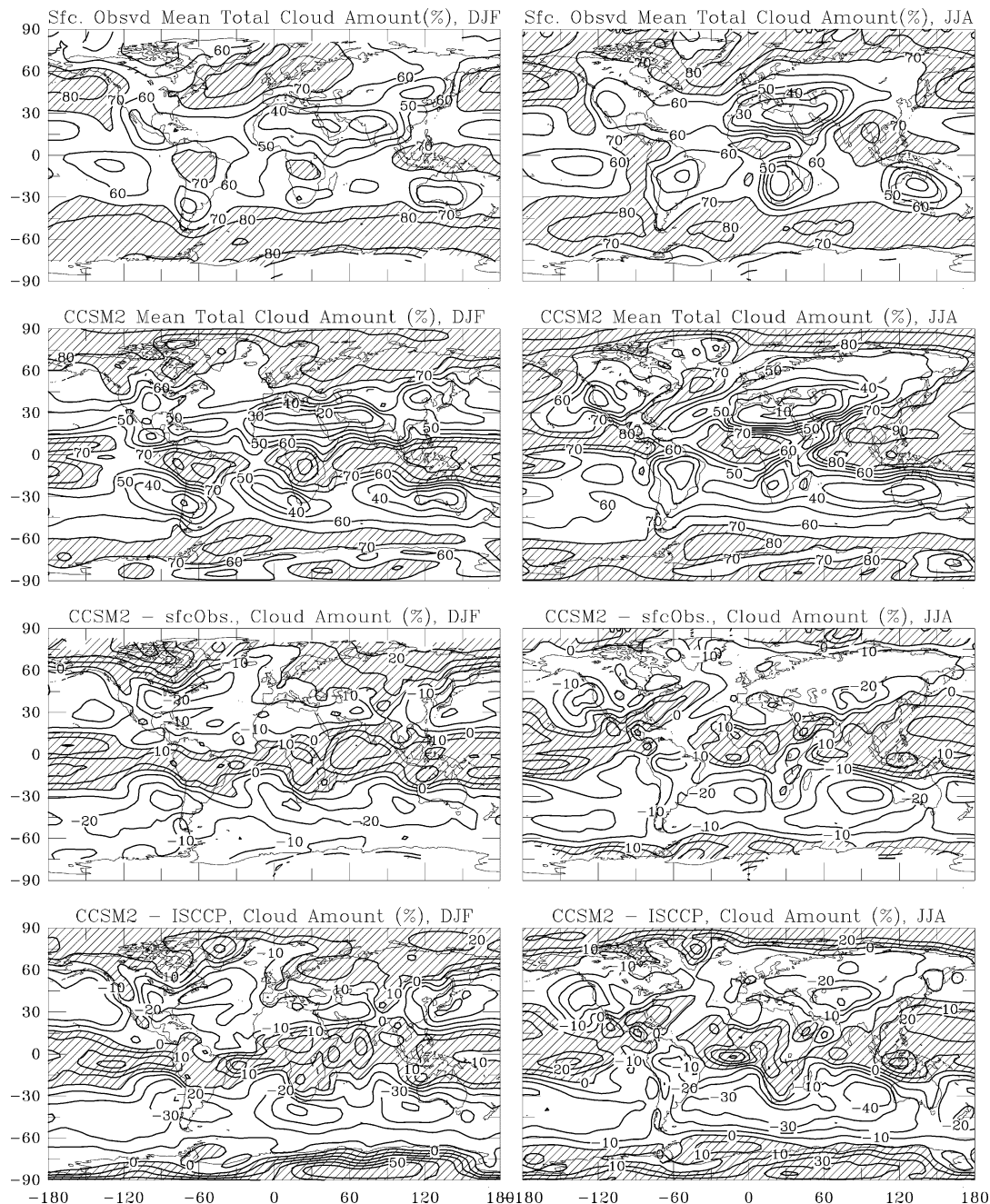


FIG. 7. Mean total cloud amount (%) for (left) DJF and (right) JJA from (first row) surface observations [1971–96 mean, derived from Hahn and Warren (1999)], the (second row) CCSM2, the (third row) CCSM2 minus surface observations difference, and (fourth row) the CCSM2 minus ISCCP (1984–2001 mean) cloud difference. Hatching indicates values over 50% in the top two rows and positive in the bottom two rows.

licly available over most land areas outside the United States, the weather-report-based precipitation frequency and its diurnal variations are reasonable proxies for evaluating the CCSM2.

The CCSM2 reproduces many of the large-scale patterns of the observation-based daily precipitation frequency (Fig. 10). For example, most of the dry areas with daily frequency <20% in the subtropics are cap-

tured by the CCSM2. The land–ocean frequency contrast is also simulated. Substantial differences exist, however, over many regions. For example, the simulated frequency patterns are incorrect over much of the tropical Pacific because of the unrealistic double intertropical convergence zone (ITCZ) and the associated equatorial cold tongue of SSTs in the central Pacific. The model also rains too often in DJF over central Africa

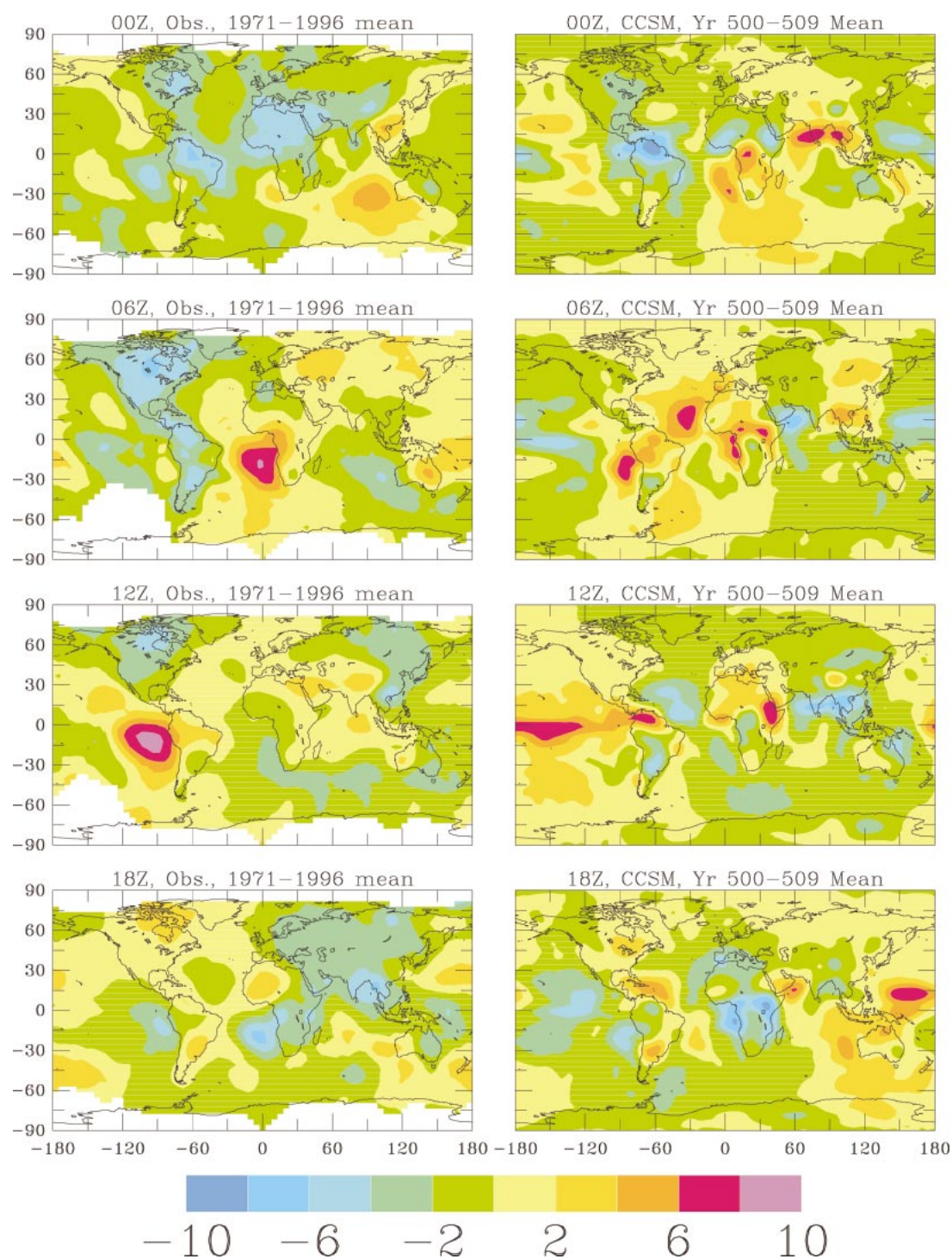


FIG. 8. Mean diurnal anomalies (relative to daily mean) of (left) observed (from Hahn and Warren 1999) and (right) CCSM-simulated DJF total cloud amount (%) at 0000, 0600, 1200, and 1800 UTC.

and South America but too infrequently in Canada, Europe, and northern Asia. A negative bias also exists in JJA over Europe and, to a lesser extent, in northern and central Asia. Interpretation of the observed frequency

over the Southern Oceans is less certain owing to limited sampling (Dai 2001a).

Figure 11 compares JJA diurnal anomalies of convective precipitation amount in the CCSM2 with those

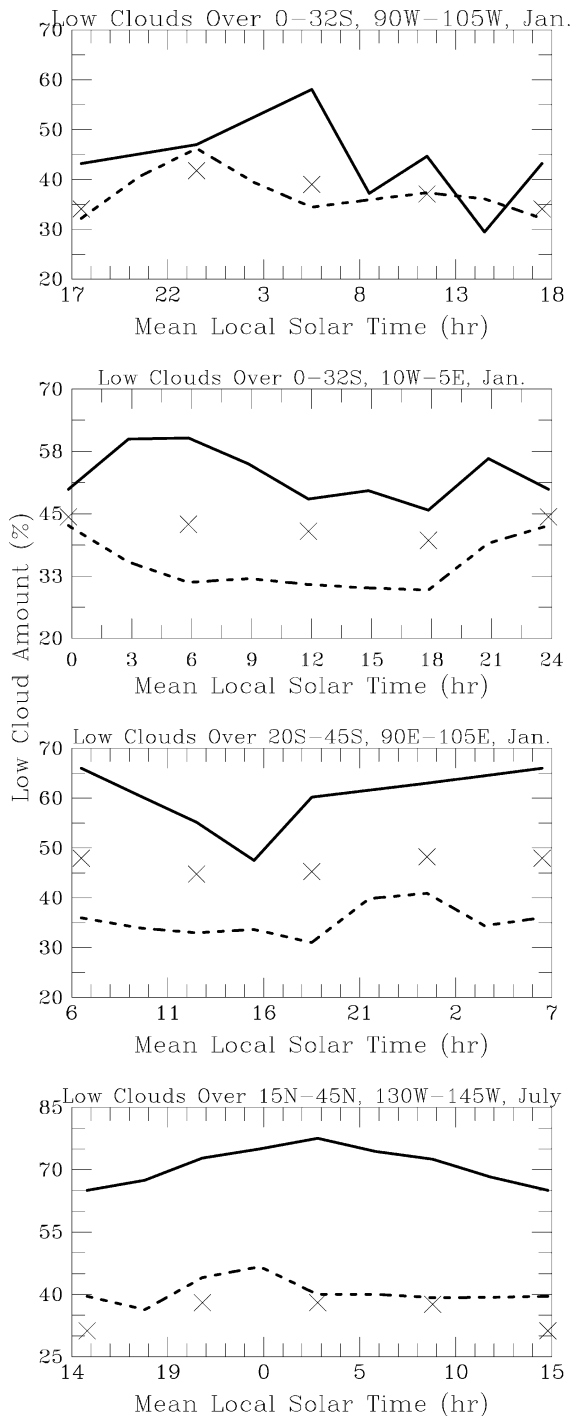


FIG. 9. Mean diurnal cycle of low-level cloud amount (%) from surface observations (solid line) and the CCSM2 (dashed line) averaged over four marine stratocumulus regions. Also shown (crosses) is the marine stratocumulus cloud amount derived using the same diagnostic scheme as in the CCSM2 with observed surface and NCEP–NCAR reanalysis 700-mb 6-hourly potential temperatures (see text for more details).

in the weather-report-based frequency of showery precipitation (from Dai 2001b). This comparison (of patterns) is appropriate since hourly rain gauge data have shown that precipitation frequency contributes most of the diurnal variations in precipitation amount, while mean precipitation intensity varies little diurnally (e.g., Dai et al. 1999a). The CCSM2 captures many of the large diurnal variations over land although regional problems are evident (e.g., over Africa; Fig. 11). For example, at 0000 UTC both the observed frequency and simulated convective precipitation show large positive diurnal anomalies over North America and negative anomalies over Eurasia. At 1200 UTC, these anomalies reverse sign. There are, however, noticeable biases over the oceans. For example, the large diurnal anomalies associated with the ITCZ in the model over the tropical Pacific and Atlantic Ocean are not evident in the observations. However, precipitation diurnal variations over most oceans outside of the ITCZ regions are too small in the model.

The diurnal phase differences can be examined more quantitatively through the phase of the diurnal harmonic, as it explains most of the subdaily variance over land and about half of the variance over ocean (Dai 2001b). Convective precipitation peaks around 1400–1600 LST in the afternoon over land and in the early morning (0200–0400 LST) over most oceans in the CCSM2 (Fig. 12); that is, it is out of phase over land and ocean. This phase pattern varies little seasonally in the model. The phase of the observed showery precipitation frequency has much larger spatial and seasonal variations. Showery precipitation peaks from late evening to midnight over low-latitude land areas in all seasons (spring and autumn are not shown). Over northern midlatitude land areas, showery precipitation peaks around 1600–1800 LST in JJA, slightly earlier (1400–1600 LST) in spring and autumn, and around 1200–1400 LST in DJF. These observed seasonal and latitudinal variations suggest that the stronger the solar heating, the later the peak convection. The phase over the oceans is noisy in the data partly because of sampling errors. Nevertheless, it is evident that peak marine showery precipitation occurs around 0600–0800 LST over the North Atlantic, North Pacific, and other large oceanic regions adjacent to the continents. Furthermore, this morning maximum exists in all the seasons, except for the high-latitude North Atlantic and Pacific in boreal winter and autumn where the peak is around 1200–1400 LST.

The diurnal phase of the total precipitation is very similar to that of the convective precipitation in all but the DJF season in the model (Fig. 12) because the CCSM2 convective precipitation accounts for most (>80%) of the total precipitation at low latitudes and over the extratropics during the warm season (not shown). Recent satellite observations (Nesbitt and Zipser 2003) suggest that convective precipitation accounts for just over 50% of total precipitation at low latitudes. The frequencies of showery and nonshowery precipi-

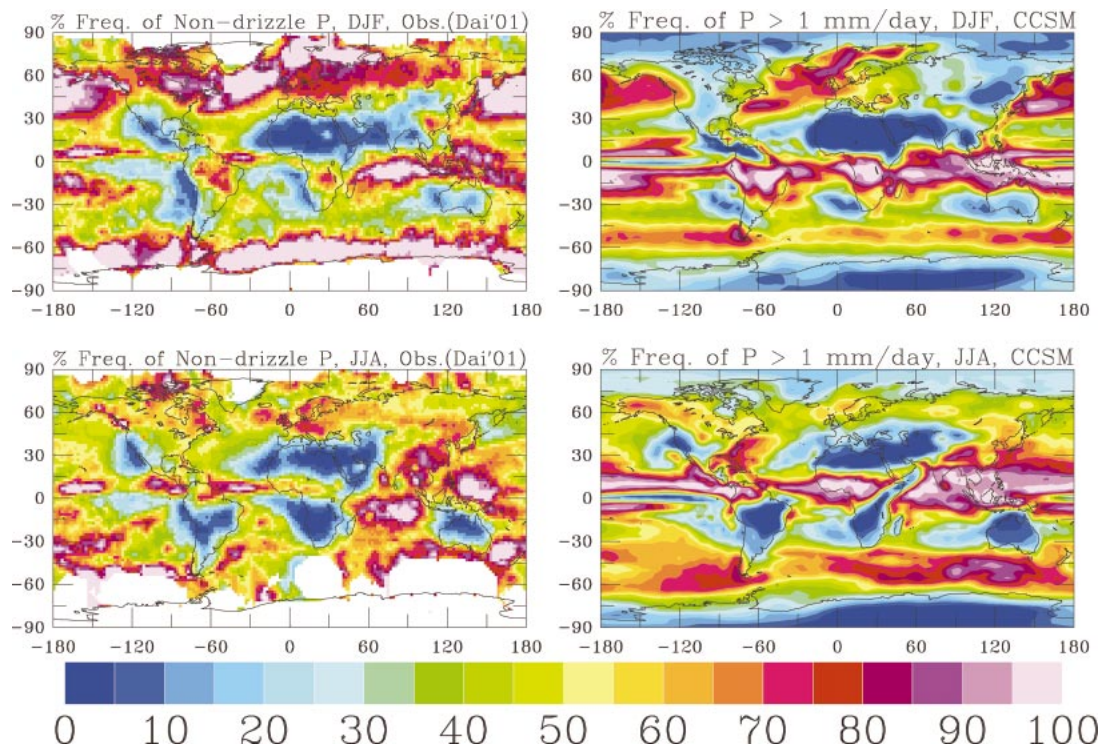


FIG. 10. (left) Weather-report-based (from Dai 2001a) and (right) CCSM-simulated (top) DJF and (bottom) JJA precipitation frequency (%), which is defined as the percentage of the total number of days with one or more reports of nondrizzle precipitation for the left panels and with precipitation exceeding 1 mm day^{-1} in the CCSM for the right panels.

tation derived from synoptic weather reports (Dai 2001a) are comparable over mid and low latitudes, whereas nonshowery precipitation is much more frequent than showery precipitation at high latitudes in winter. These results suggest that the CCSM2 severely overestimates convective and underestimates nonconvective precipitation.

Winter nondrizzle precipitation over northern high latitudes peaks in the morning (Fig. 12), but the CCSM2 has afternoon peaks over most of North America and Eurasia, partly due to a still large ($\sim 50\%$) contribution of convective precipitation over these regions in winter. In the observations, the diurnal timing of the maximum during spring and autumn changes from afternoon at northern low latitudes to morning at higher latitudes, whereas this phase transition does not exist in the CCSM2 (not shown). Some regional features, such as the nocturnal maximum over the central United States, are also absent in the model.

Area-averaged, composite diurnal cycles (relatively to LST) of precipitation (Figs. 13–15) show that warm-season daytime moist convection over land in the CCSM2 starts soon after sunrise (~ 0800 LST), about 4 hours too early compared with observations. Moist convection reaches a plateau around 1100 LST and stays there until around 1800 LST in the model, whereas the observations show a sharp peak around 1600–1700 LST. These diurnal biases result in incorrect diurnal evolution

of CAPE in the atmosphere. According to the NCEP–NCAR reanalysis data, warm-season CAPE over land (Fig. 16) should build up after sunrise and reach a maximum around 1400 LST when strong convection begins to release the energy, resulting in a smooth decline that extends to the evening. In the CCSM2, however, CAPE decreases in the early morning until ~ 1000 LST and then accumulates very slowly until ~ 2100 LST when it reaches its maximum and starts to decline (Fig. 16).

Consistent with Fig. 11, Figs. 13–15 also show relatively weak diurnal cycles over the oceans in the CCSM2. This is particularly true for convective precipitation, consistent with the lack of diurnal variation in SSTs and that atmospheric convection is closely coupled with diurnal variations in surface heating and thus SSTs. However, the diurnal variations in nonconvective precipitation are simulated relatively well, with a small peak in the morning and lower values in the afternoon. Even the relatively weak but nonnegligible semidiurnal component in nonconvective precipitation is evident in the CCSM2.

4. Discussion

The relatively realistic simulations of the tidal variations in surface pressure and upper-air winds by the CCSM2 are consistent with earlier studies cited in introduction that show that large-scale tidal features are

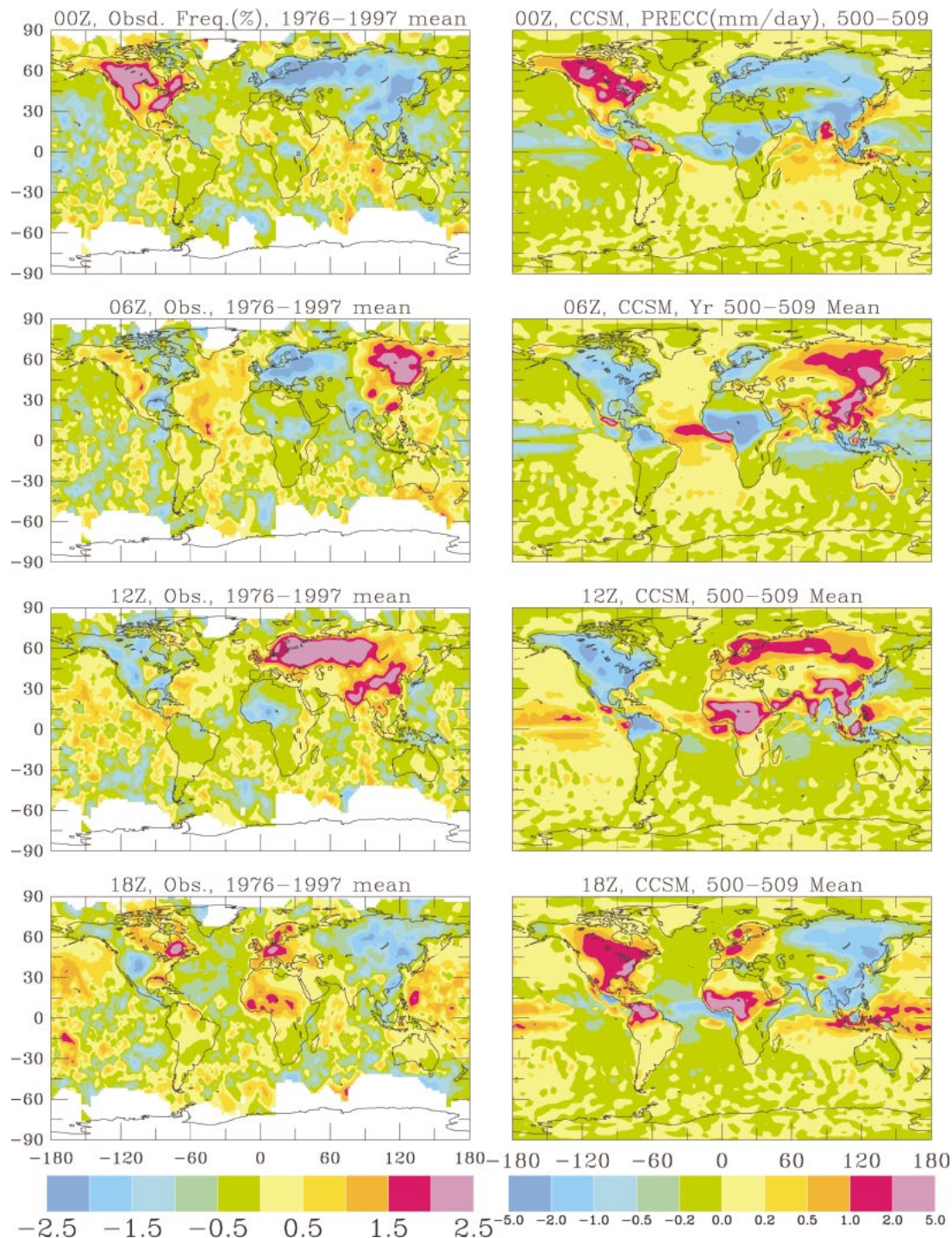


FIG. 11. Mean diurnal anomalies (relative to diurnal mean) of (left) observed JJA mean showery precipitation frequency [%], from Dai et al. (2001b)] and (right) CCSM convective precipitation amount (mm day⁻¹) at 0000, 0600, 1200, and 1800 UTC.

generally reproduced by AGCMs. The tidal results suggest that the treatments of atmospheric absorption of solar radiation by ozone and water vapor and surface sensible heating over land are generally realistic in the CCSM2.

The results for clouds and precipitation are, however, not very encouraging. The large systematic biases in the mean cloud amount (Fig. 7) are of concern because incorrect mean clouds affect solar heating, thereby altering the diurnal cycle near the surface and in the at-

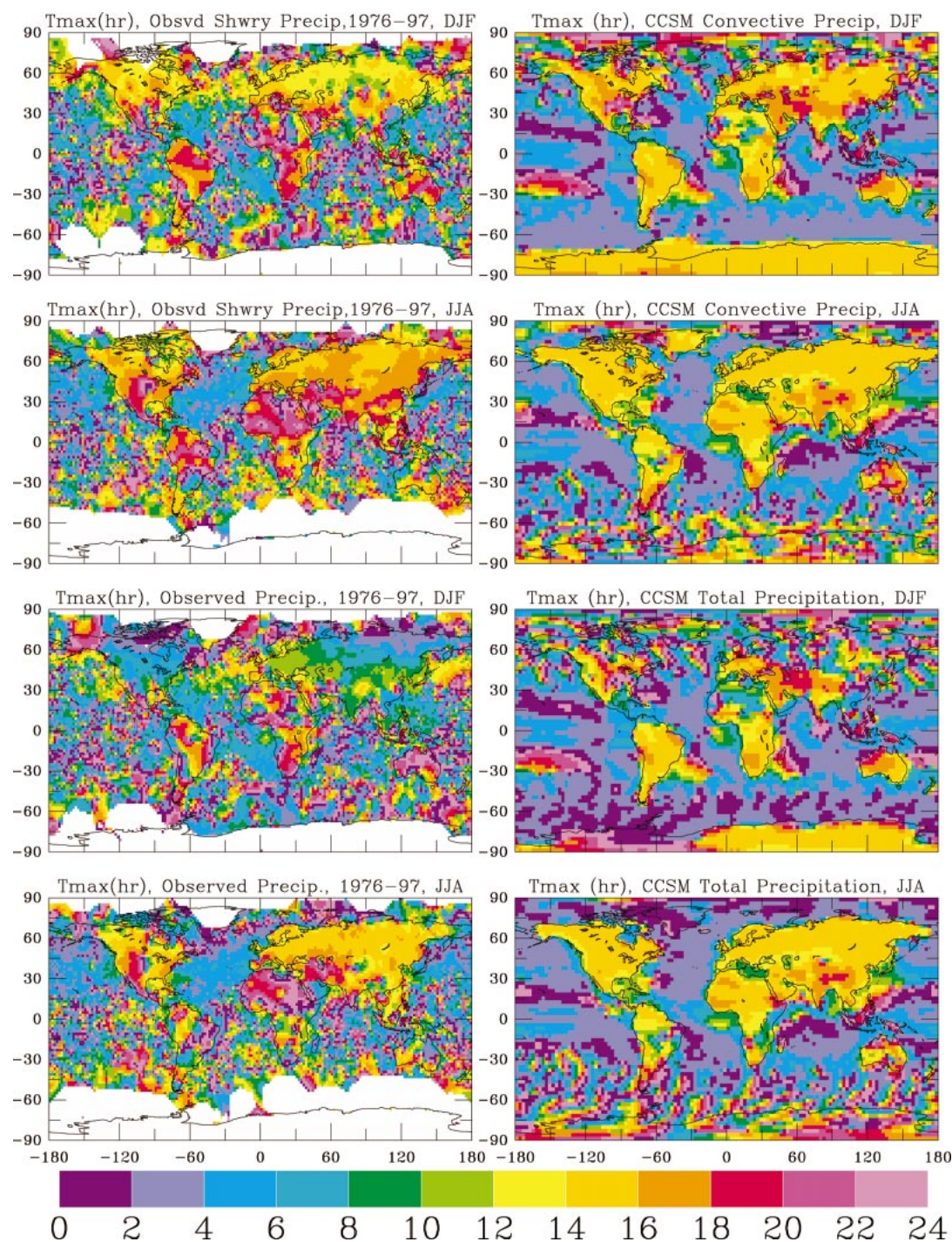


FIG. 12. (left) The LST (h) of the max of the diurnal harmonic of (first and second rows) observed DJF and JJA showery and (third and fourth rows) non-drizzle precipitation frequency (from Dai 2001b). (right) Same as (left) but for (first and second rows) CCSM-simulated convective and (third and fourth rows) total precipitation.

mosphere. A secondary effect arises from the diurnal cycle in the clouds themselves. Thus, it is very important from the diurnal cycle perspective that the mean cloud amount and its optical thickness are simulated correctly. The diagnostic cloud schemes used in the CCSM2 are

apparently too simplistic, as they can neither simulate the mean cloud patterns nor the large diurnal cycle in marine stratocumulus clouds. As shown by Dai et al. (1999a) in a regional model, excessive cloudiness reduces surface solar heating and thus daytime surface

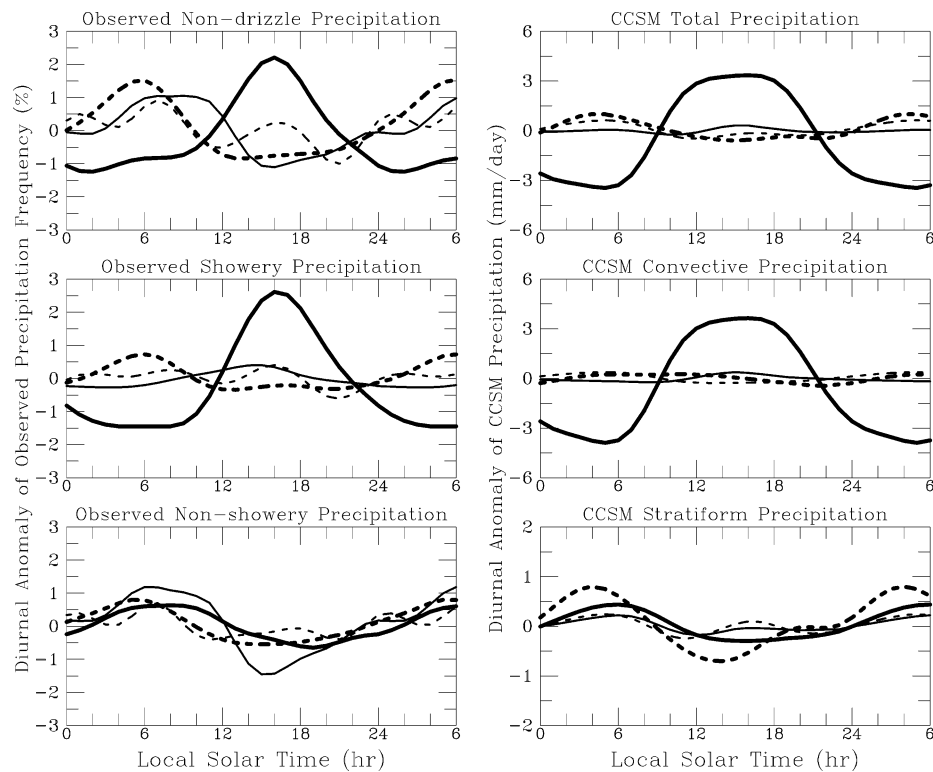


FIG. 13. (left) The 25° – 70° N mean diurnal cycle of observed frequency (%) and (right) CCSM-simulated amount (mm day^{-1}) of (top) total precipitation (excluding drizzles in the observations), (middle) showery or convective precipitation, and (bottom) stratiform precipitation. Area-weighted averaging was done at each LST hour over land (solid lines) and ocean (dashed lines) within the lat zone for DJF (thin lines) and JJA (thick lines).

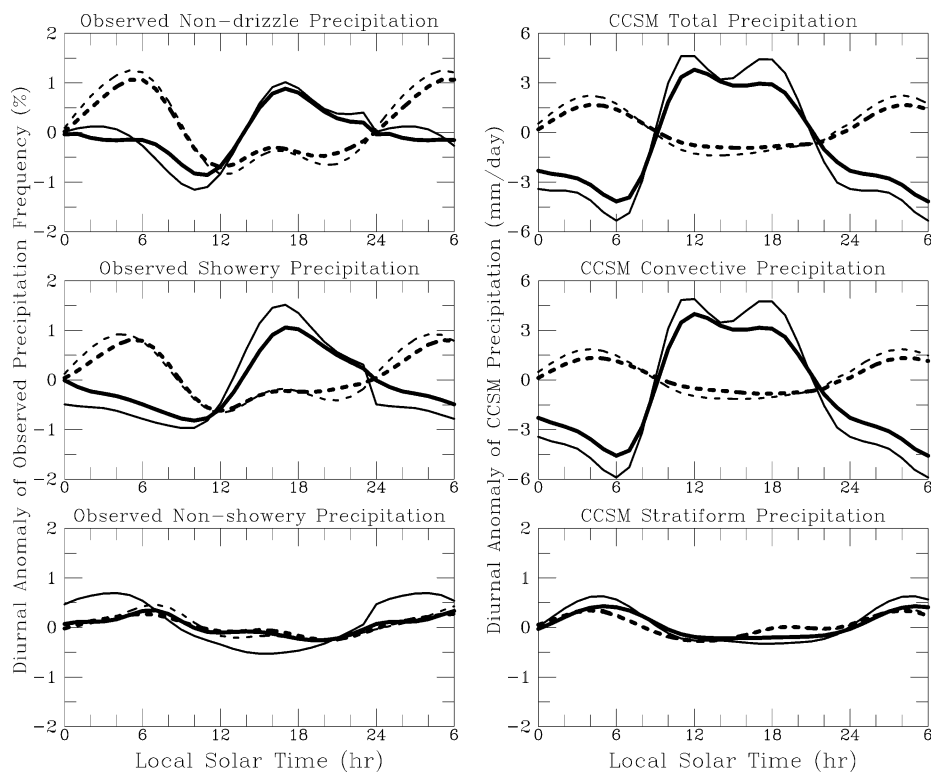


FIG. 14 As in Fig. 13 but for 25° S– 25° N.

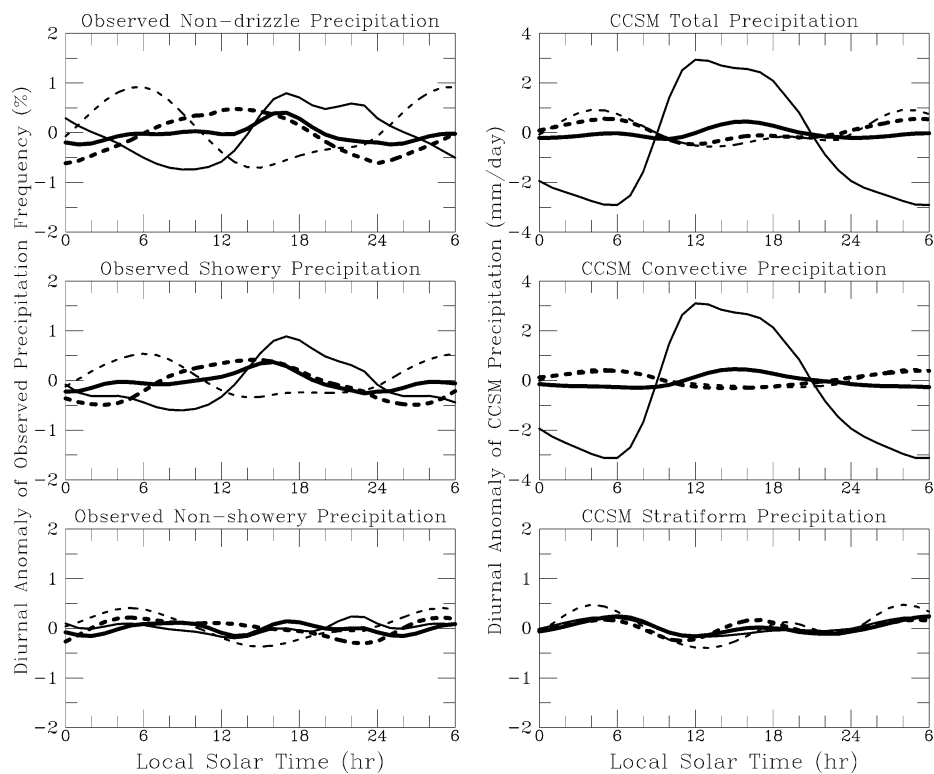


FIG. 15. As in Fig. 13 but for 25°–50°S.

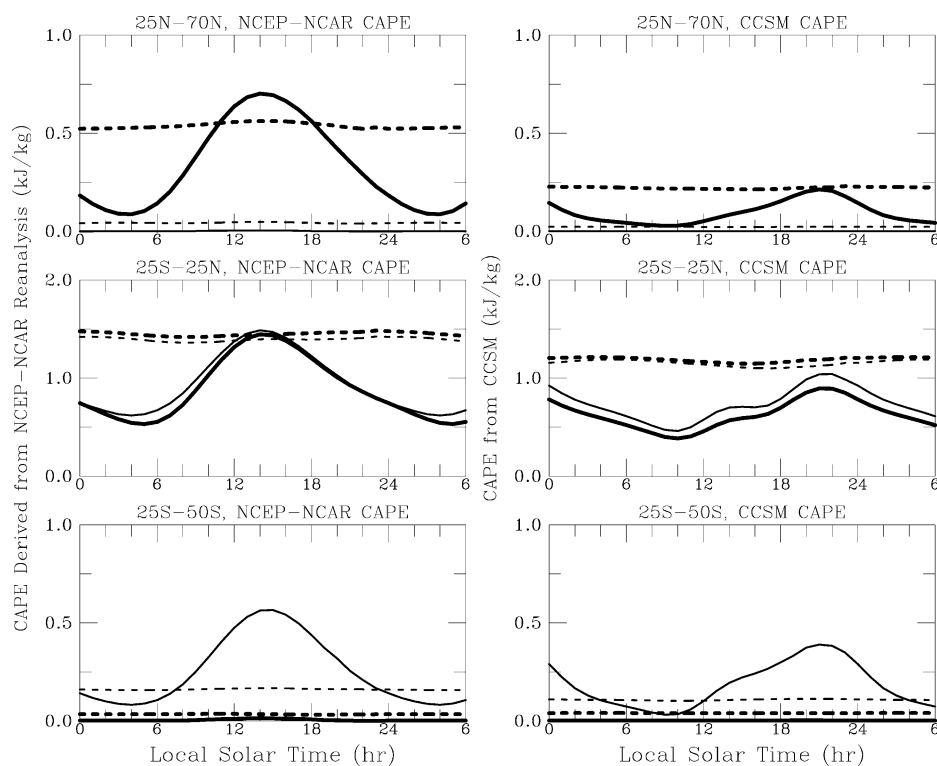


FIG. 16. Mean diurnal cycle of CAPE (kJ kg^{-1}) averaged at each LST hour over the land (solid lines) and ocean (dashed lines) areas within 25°–70°N, 25°S–25°N, and 25°–50°S. CAPE was calculated based on a pseudoadiabatic process using Jan (thin lines) and Jul (thick lines) mean 6-hourly NCEP–NCAR reanalysis data of (left) 1980–96 and (right) 3-hourly CCSM output. Note that one solid line overlays the x axis in the top and bottom panels.

warming. This leads to a generally weakened diurnal cycle at the surface and in CAPE and thus errors in convective precipitation. In the CCSM2, the mean cloud biases (cf. Fig. 7) are associated with a reduced diurnal cycle over central Africa and enhanced diurnal amplitudes over extratropical South America (cf. Figs. 1 and 2). Obviously, more detailed diagnostic analyses focusing on the relationship among the diurnal cycles of various variables over selected regions are needed, and they are under way.

Nevertheless, the global diurnal results presented here point to several other sources of deficiency in the CCSM2. One of these is the lack of appreciable diurnal variations in SSTs in the model. This obviously damps the diurnal cycle in the air above the oceans, leading to weakened diurnal variations in all of the marine fields that we examined. Other related atmospheric processes, such as the afternoon cumulus convection associated with peak ocean skin temperatures (Sui et al. 1997; Chen and Houze 1997), are unlikely to be simulated by the CCSM2. Moreover, this potentially affects the simulated mean climate because of the nonlinear effects. These problems exist in essentially all AGCM simulations as mean SSTs without diurnal variations are often used as the lower boundary condition.

Another deficiency of the model is that deep moist convection starts prematurely (Figs. 13–16), which prevents CAPE from rapid accumulation in the atmosphere and intense convection from occurring in the mid to late afternoon. This is not unexpected since the Zhang and McFarlane (1995) convection scheme used in the CCSM2 does not include convective inhibition induced by, for example, a stable boundary layer or descending large-scale motion. Instead, the scheme allows free convection whenever CAPE is above 75 J kg^{-1} (G. J. Zhang 2003, personal communication).

The extended duration of daytime convection over land (Figs. 13–15) also suggests that the CCSM2 rains too frequently at reduced intensity despite the fairly realistic patterns of rainy days (Fig. 10). This incorrect combination of frequency and intensity was also noticed by Dai et al. (1999a) in a regional climate model and is a fairly common problem in atmospheric models (Trenberth et al. 2003). As moist convection starts too early and occurs too often in the CCSM2, it removes atmospheric moisture prematurely. This reduces the moisture available for nonconvective precipitation and contributes to overestimation of convective precipitation and underestimation of nonconvective precipitation, such as that associated with monsoons at low latitudes and synoptic fronts at mid latitudes.

The fact that the diurnal cycle of CAPE over land is unrealistic in the CCSM2 even though the surface air temperature has a fairly realistic diurnal cycle suggests that CAPE's subdaily variations depend strongly on *atmospheric* variations of temperature and humidity. This appears to be different from the tropical marine atmo-

sphere where mean CAPE is a strong function of mean SSTs (Fu et al. 1994).

The difficulties in simulating the diurnal cycle are caused by several factors. First, the physical processes underlying many of the diurnal variations are not fully understood. Examples of such processes include the nocturnal maximum of warm-season precipitation over the central United States (Dai et al. 1999a) and the early morning maximum of tropical oceanic convection and precipitation (Randall et al. 1991; Liu and Moncrieff 1998; Dai 2001b).

Second, the CCSM2, along with other CGCMs, have substantial deficiencies in their physical parameterizations for land and ocean surface processes (as for SST), the planetary boundary layer, the initiation of moist convection, cloud and precipitation formation, and low-level atmospheric convergence. All of these aspects are closely tied to solar heating near the surface and thus undergo large diurnal variations. As demonstrated by the biases in the mean cloud amount, the CCSM2 still has difficulties in correctly simulating some of the mean patterns. It is then not unexpected that many of the diurnal variations are not realistically captured. Betts and Jakob (2002b) show that the ECMWF model fails to represent the morning development of a growing cumulus boundary layer and that its convection scheme, rather the large-scale vertical motion field, is likely responsible for the morning precipitation peak in the ECMWF model.

Third, comprehensive evaluation of GCM-simulated diurnal variability has been hampered by a lack of observational global datasets with high temporal resolution. This is especially true for upper-atmospheric fields and precipitation over the open oceans, although recent satellite observations have started to fill some of the data gaps (Sorooshian et al. 2002; Nesbitt and Zipser 2003). Finally, most GCM simulations typically do not archive hourly or 3-hourly data that are required for diurnal analyses, which hampers diagnostic analyses of the diurnal cycle in GCMs. Trenberth et al. (2003) argue that it is important for GCMs to save some hourly data so that the characteristics of precipitation and the diurnal cycle can be examined.

5. Summary and concluding remarks

To evaluate the performance of the CCSM2 in simulating the diurnal cycle and diagnose the deficiencies in related model physics, we have analyzed ten years of 3-hourly data from the CCSM2 control run for global and large-scale features of diurnal variations in surface air temperature, surface pressure, upper-air winds, cloudiness, and precipitation. The model-simulated diurnal variations were compared with available observations, most of which were derived from 3-hourly synoptic observations during the 1976–97 period.

The CCSM2 reproduces most of the tidal variations in surface pressure and upper-air winds, including many

regional patterns. For example, the larger amplitude over land (0.4–1.4 mb) than over ocean (0.4–0.6 mb) and the 0800–0900 LST peak over most land areas in the diurnal pressure tide are simulated by the model. The CCSM2 also reproduces peaks around 0930–1030 LST and amplitudes of 0.8–1.2 mb for the semidiurnal pressure tide at low latitudes. The CCSM2, however, overestimates the diurnal pressure tide by 20%–50% over low-latitude land and underestimates it over most oceans, the Rockies, and other northern midlatitude land areas.

The CCSM2 captures the diurnal amplitude (1° – 6° C) and phase (peak at 1400–1600 LST) of surface air temperature over land. Over the oceans, however, the simulated temperature amplitude ($\leq 0.2^{\circ}$ C) is too small. The observed surface air temperature also shows a coherent semidiurnal cycle with amplitudes of 0.4° – 1.5° C over land and 0.2° – 0.4° C over ocean and peaks around 0100–0300 and 1300–1500 LST over most land areas. The model simulates some of the semidiurnal features, but they are too weak over Eurasia and North America during DJF and over the oceans in all seasons. The results suggest that, while the daytime solar heating and nighttime radiative cooling near the ground are generally realistic in the CCSM2, the diurnal cycle over the ocean surface is too weak in the model owing to the absence of diurnal variations in SSTs.

Compared with observations, the CCSM2 overestimates the mean total cloud amount by 10%–20% of the sky within $\sim 15^{\circ}$ S– 15° N during both DJF and JJA and over northern mid- and high-latitude land areas in DJF, whereas it underestimates the cloud amount by 10%–30% in the subtropics and parts of midlatitudes. Because of high albedo of clouds, these biases affect daytime solar heating and thus the diurnal cycle near the surface. Over the marine stratocumulus regions west of the continents where large diurnal variations (amplitude 3%–10%) in low and total cloud amount are observed, the CCSM2 underestimates the mean low cloud amount by 10%–30% and the diurnal variations with incorrect diurnal phase (midnight peak instead of 0300–0500 LST in observations). The diagnostic scheme for marine stratocumulus clouds used in the CCSM2 does not account for factors that contribute to the diurnal cycle in the stratus clouds and fails to simulate the stratus diurnal cycle even when surface observations and the NCEP–NCAR reanalysis data were used.

The CCSM2 simulates the number of rainy days (with precipitation > 1 mm day $^{-1}$) realistically. However, warm-season daytime moist convection over land starts prematurely in the model around ~ 0800 LST, compared with ~ 1200 LST in the observations. In the model moist convection reaches a plateau around 1100 LST and persists until ~ 1800 LST, in contrast to a sharp peak around 1600–1700 LST in the observations. The premature initiation of convection prevents CAPE from accumulating in the morning and early afternoon and intense convection from occurring in the mid to late afternoon. The

extended duration of daytime convection over land results in elevated precipitation frequency and reduced intensity despite the fairly realistic patterns of rainy days. This bias also contributes to overestimation of convective precipitation and underestimation of non-convective precipitation. This convective versus non-convective precipitation bias leads to incorrect diurnal phasing, namely, an afternoon instead of morning peak, in cold-season precipitation over northern high-latitude land. The simulated precipitation diurnal cycle is too weak over the oceans, especially for convective precipitation, consistent with the weak diurnal cycle in marine surface air temperature and pressure that is linked to the lack of a diurnal cycle in SSTs in the model.

These results show that the CCSM2 needs to be improved in simulating the mean cloud amount and the diurnal cycle in marine stratus clouds. The premature initiation of warm-season moist convection over land also needs to be addressed, presumably by properly accounting for convective inhibition and making it harder for an air parcel to reach the level of free convection. Experiments with the ECMWF model (Betts and Jakob 2002b) suggest that this is a rather difficult problem. Improvements in this area will not only increase the probability of a realistic diurnal cycle of precipitation, but also improve the combination of precipitation frequency and intensity and the convective versus non-convective precipitation ratio, albeit the latter also depends on the parameterization of nonconvective precipitation (e.g., from anvil clouds). Work has been under way to improve the treatment of convective inhibition by a stable boundary layer and low-level divergence (Zhang 2002). However, even “perfect” parameterization schemes for moist convection and cloud and precipitation formation may not work correctly in a coupled GCM because of errors in the related fields such as boundary layer processes and low-level wind convergence, as these processes have large effects on initiation of moist convection (Fu et al. 1994). Nevertheless, the uniform biases in the warm-season moist convection over land suggest that changes in the convection scheme alone may substantially improve the diurnal cycle.

The weak diurnal cycle over the oceans in the CCSM2 demonstrates that the lack of a diurnal cycle in SSTs is a significant deficiency. Modeling the complex physical processes (such as mixing due to wave breaking) within the top few meters of the oceans is a difficult task for coupled GCMs. Nevertheless, it is important to include the two-way air–sea interactions on subdaily time scales in coupled models, perhaps through diagnostic equations of ocean skin temperature as a function of surface solar radiation and wind speed (Webster et al. 1996).

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