Diurnal Heating and Cloudiness in the NCAR Community Climate Model (CCM2)

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ABSTRACT

In this paper, the authors assess the suitability of the heating fields in the latest version of the NCAR Community Climate Model (CCM2) for modeling the thermal forcing of atmospheric tides. Accordingly, diurnal variations of the surface pressure, outgoing longwave radiation, cloudiness, and precipitation are examined in the CCM2. The fields of radiative, sensible, and latent heating are similarly analyzed. These results are subjectively compared with available data.

Equatorial diurnal surface pressure tides are fairly well simulated by CCM2. The model successfully reproduces the semidiurnal surface pressure tides; however, this may result in part from reflection of wave energy at the upper boundary. The CCM2 large-scale diurnal OLR is generally consistent with observations. The moist-convective scheme in the model is able to reproduce the diurnally varying cloudiness and precipitation patterns associated with land–sea contrasts; however, the amplitudes of CCM2 diurnal continental convective cloudiness are weaker than observations. The CCM2 boundary-layer sensible heating is consistent with a very limited set of observations, and with estimates obtained from simple models of diffusive heating. Although the CCM2 tropospheric solar radiative heating is similar in magnitude to previous estimates, there are substantial differences in the vertical structures. A definitive assessment of the validity of the CCM2 diurnal cycle is precluded by the lack of detailed observations and the limitations of our CCM2 sample. Nevertheless, the authors conclude that the global-scale components of the CCM2 diurnal heating are useful proxies for the true diurnal forcing of the tides.

1. Introduction

The latest version of the NCAR Community Climate Model, CCM2, includes a diurnal cycle in the radiative heating algorithm and improved parameterizations of surface exchanges and of boundary-layer and convective processes. In this paper, diurnal variations in the global-scale model surface pressure, outgoing longwave radiation (OLR), and precipitation are analyzed and compared with available observations. The model diurnally varying global radiative, sensible (boundary layer), and latent heating fields are similarly diagnosed.

This study was motivated by a need for realistic tropospheric diurnal radiative, sensible, and latent heating fields as tidal drives. Observations of diurnal tides in the middle atmosphere indicate that vertically propagating nonmigrating components of appreciable amplitude are present along with the main migrating tide (Wallace and Hartranft 1969; Wallace and Tadd 1974; Lieberman 1991). This study follows the usual practice of tidal studies, referring to diurnal fluctuations that are not sun synchronous as nonmigrating. It should be noted that nonmigrating does not imply stationary; the nonmigrating tides do propagate zonally but with a phase speed different from that given by the earth’s rotation.

Nonmigrating tidal components are believed to be driven primarily by longitudinal variations in the diurnal tropospheric heating (Chapman and Lindzen 1970). This premise has already been examined in a number of investigations (McKenzie 1968; Hamilton 1981a; Kato et al. 1982; Forbes and Groves 1987; Tsuda and Kato 1989). These studies employed linear tidal theory with empirical sensible or latent heating functions modulated by simplified land–sea or orographic distributions. It was confirmed that these mechanisms are important sources of tidal excitation; however, realistic three-dimensional representations of these heating fields are needed in order to assess their effects on the tides in a definitive manner. The parameterization of sensible and latent heating mechanisms as tidal drives requires knowledge of the global-scale diurnally varying temperature, cloudiness, and precip-
Fig. 1. The 0600 UTC mapping of the migrating diurnal surface pressure tide. Units are millibars and contour interval is 0.1 mb. Negative contours are dashed. (a) January CCM2. (b) FGGE data, 10 January–10 February 1979. (c) July CCM2. (d) FGGE data 5 May–5 June 1979.

iteration. Unfortunately, the temporal and spatial coverage of the existing observing networks and databases remains insufficient for representing the global, diurnally varying tropospheric heating.

The leading objective of this study is to establish the validity of CCM2 diurnal radiative, convective, and latent heating as proxies for the actual tidal driving. In order to accomplish this objective, we first establish that the principal observed features of atmospheric tides are reproduced in CCM2. Having established some confidence in the CCM2 tides, we proceed to diagnose the forcing terms. Section 2 summarizes the relevant CCM2 attributes and our analysis procedure. In section 3 we examine the model diurnal surface pressure, outgoing longwave radiation (OLR), cloudiness, and precipitation. The
diurnal harmonics of the radiative, sensible, and latent heating are discussed in section 4. Conclusions appear in section 5.

2. Model and analysis

CCM2 is the most recent version of the NCAR Community Climate Model. CCM2 is described extensively in other papers, some of which are referenced below, and only the features of CCM2 that are the most relevant to this study are discussed here. It should be noted that the changes to the previous version, CCM1, are so extensive that most of the previous documentation has been rendered obsolete. CCM2 still uses the spectral transform method for the dynamical equations but uses the semi-Lagrangian method for transporting water (Rasch and Williamson 1991). The standard resolution of CCM2, which is used here, employs 18 levels in the vertical from ~992 mb to 3 mb and T42 horizontal resolution (~2.8° in latitude and longitude).

Kiehl (1990) has provided an overview of the CCM2 cloud and radiative algorithms. CCM2 has a diurnal
cycle, with solar radiative heating and longwave cooling computed every hour. The solar radiation parameterization employs the δ-Edington approximation, as described by Briegleb (1993). The radiative properties of clouds depend explicitly on droplet properties, which are specified functions of height and latitude. Monthly and zonally averaged ozone mixing ratios are specified from the climatology of Dützsch (1978).

Sensible heating in CCM2 is dominated by the divergence of the turbulent sensible heat flux in the boundary layer. The “nonlocal” boundary-layer parameterization (Holtslag and Boville 1993) diagnoses a boundary-layer depth and determines diffusivity profiles and nonlocal turbulent transports within the boundary layer. The surface exchange formulation is also described in Holtslag and Boville (1993) and the ground temperature is determined by a diffusion equation for the surface layer and three subsurface layers. The land surface wetness is specified as a function of location, varying from 0.05 in the central Sahara desert, to 0.95 in the Amazon basin. Monthly mean sea surface temperatures and sea ice extents are specified from the climatology of Shea et al. (1990) and instantaneous values are obtained at each time step by linear interpolation.

Latent heating in CCM2 is dominated by moist convection in the tropics and also includes large-scale (stable) condensation, which becomes important in middle latitudes. The mass flux convection scheme in CCM2 (Hacker 1994) adjusts the moist static energy over three adjacent layers, allowing for entrainment in the bottom layer, condensation and rainout in the middle layer, and detrainment in the upper layer. The method is applied to each layer sequentially, beginning at the surface, until all of the tropospheric levels have been adjusted. The boundary-layer parameterization diagnoses perturbation temperatures and specific humidities within the rising plumes of convective boundary layers. These perturbations are used for cloud-base quantities to diagnose stability in the convection scheme. The coupling of the convection and boundary-layer parameterizations makes a substantial impact on the diurnal cycle of convection.

The behavior of the individual components of CCM2 is discussed in the papers referenced above and basic climate statistics from a 20-yr control run of CCM2 are discussed by the NCAR Climate Modeling Section (1993). The results presented in this paper are obtained from the seventh year of the CCM2 control run. During the months of January and July, the model output at each grid point was written at 3-hourly intervals for 14 days. A composite day was created by averaging the 14 samples of the 3-hourly grids. A longitudinal Fourier transform was then applied to these composite grids. Subsequently, the zonal wave components were again Fourier transformed in time, to extract the diurnal and semi-diurnal harmonics. All the fields shown were truncated to 12 zonal waves.

3. Comparison of CCM2 fields with observations

a. Surface pressure

The CCM2 migrating diurnal and semi-diurnal surface pressure tides for January and July are compared to the corresponding fields from FGGE analyses below. Following Yagai (1989), we utilize the original (or main) FGGE level IIIb dataset for the special observing periods (10 January–10 February and 5 May–5 June 1979) from the European Centre for Medium-Range Weather Forecasts. Although the second special observing period data is not for the same month as the model data, it is of higher quality than the data available for July. During these time periods, high-resolution data are available four times per day over the globe. We compare these with a CCM2 run centered on a date occurring 6 weeks later, because we employ heating fields from that run in a seasonal tidal study, to be published elsewhere. The FGGE sampling pattern enables the diagnosis of the amplitude and phase of both migrating and nonmigrating diurnal tides. Since the structures of the migrating tides depend only on local time, we display them in map form for a particular universal time.

Figure 1 compares CCM2 migrating diurnal surface pressure tides with FGGE observations. CCM2 peak January values reach 0.57 mb at 7°N, compared to an observed maximum of 0.66 at 11°N. The model tide peaks at 0700 LT, lagging the observed field by about 12 min at the equatorial location. In July, both fields have a 0650 LT equatorial maximum in the Southern Hemisphere. The CCM2 value of 0.55 occurs at 0652 at 7°S, and exceeds the observed peak of 0.44 mb at 11°S. In July, the CCM2 reproduces a secondary maximum in the Northern Hemisphere. However, the model amplitude and phase of 0.4 mb at 0634 LT do not agree as well with the observed amplitude of 0.68 mb occurring at 0556 LT. The CCM2 Northern Hemisphere peak is located farther south, at 32°N, than the observed location at 48°N.

Figures 2 and 3 show the amplitude and phase of the CCM2 and FGGE diurnal surface pressure plotted as harmonic dials. The length of an arrow is proportional to the amplitude, while the orientation represents the hour of maximum in local time (LT). There is a substantial nonmigrating component to the surface pressure tide, which can be seen by the turning of the arrows along a latitude circle. The most prominent nonmigrating modes of the CCM2 diurnal surface pressure are an eastward propagating wavenumber 3 and a westward propagating wavenumber 5 (not shown). These spectral features have amplitude maxima of 0.23 and 0.15 mb, respectively. Tokioka and Yagai (1987) and Yagai (1989) have linked the corresponding waves in the ob-
FIG. 2. Harmonic dials of January diurnal surface pressure, constructed from the zonal mean and 12 eastward and westward traveling waves. Arrows pointing northward indicate hour of maximum at 0000 LT, eastward at 0600 LT, southward at 1200 LT, etc. Length of maximum vector is 1.6 mb. (a) CCM2. (b) FGGE data, 10 January–10 February 1979.
served surface pressure to a standing wavenumber 4 topography pattern at low latitudes.

Both CCM2 and FGGE amplitudes are larger over the continents than over the oceans, in accordance with previous studies (Haurwitz 1965; Haurwitz and Cowley 1973; Hsu and Hoskins 1989). Over landmasses, peak values usually occur between 0600 and 0900 LT. CCM2 amplitudes over land tend to be smaller than FGGE land values. The amplitude of CCM2 pressure over the tropical Pacific Ocean ranges from 0.4 to 0.6 mb in the vicinity of the equator. These values appear to exceed the FGGE observations; however, they compare favorably with those reported by Haurwitz and Cowley (1973). The large-amplitude, small-scale variations in the southern oceans, which are found both in CCM2 and FGGE data, probably arise from sam-
pling errors in the storm track regions. Some small-scale midlatitude oceanic features in the January CCM2 may be attributed to the same cause. However, the features on the larger continent–ocean scale and at lower latitudes are robust in both CCM2 and FGGE.

The migrating component of the semidiurnal tide can be retrieved from the FGGE data by taking the difference between the 0000 UTC and 0600 UTC fields. Figure 4 shows a comparison between the CCM2 migrating semidiurnal tide, and the FGGE 6-h difference field, which indicates extremely good agreement between the two. CCM2 January amplitudes have a maximum of 1.32 mb at 7°N, compared with the observed maximum of 1.25 mb at 6°N. CCM2 tides peak at 1030 LT lagging the observed phase by 30 min.

The upper boundary of CCM2 is placed at 3 mb, a level situated well within the region of the ozone heating, which provides the main driving for the semidi-
urnal tide. At this level a "rigid-lid" ($dp/dt = 0$) condition is imposed. It might well be asked how CCM2 reproduces the semidiurnal tide with such accuracy under these circumstances. A similar question was first posed by Zwiers and Hamilton (1986) in their study of solar tides in a GCM with a rigid-lid upper boundary placed at 10 mb. Using linear tidal theory, Zwiers and Hamilton determined that the omission of the thermal forcing above this level reduces the amplitude of the semidiurnal tide by a factor of nearly 2. However, the presence of a rigid lid at the upper boundary increased these deficient amplitudes by 30% because of reflection of vertically propagating wave energy. It has also been demonstrated that reflection by the upper boundary can cause spurious amplification of vertically propagating waves in a GCM (Boville and Cheng 1988), although tides were not explicitly considered in that study. Therefore, in trying to interpret the CCM2 semidiurnal tides, we must consider the possibility articulated by Zwiers and Hamilton: that part of the success of CCM2 in simulating the semidiurnal tides could result from the reflection of wave energy at 3 mb compensating the absence of the thermal forcing above that level.

b. Outgoing longwave radiation (OLR)

In this section we briefly summarize the characteristics of the CCM2 OLR fields. The model reproduces many features of the observed diurnally varying OLR documented by Hartmann and Recker (1986) and Gruber and Chen (1988). Both the CCM2 and the satellite observations reveal peak values between 25 and 30 W m$^{-2}$ occurring over
desert regions around 1400 LT. The January and July desert patterns are quite similar. In midlatitude Northern Hemisphere continental regions, the model diurnal cycle shows some seasonal variation, with stronger signals in July than in January. Over the oceans, the diurnal cycle of CCM2 OLR is in general weaker and less organized. There is some organization in the model OLR patterns off the west coasts of North and South America, where the phase shifts to a late afternoon or evening maximum.

This behavior is indicative of the diurnal cycle of model stratiform cloudiness that prevails in these regions in the early morning hours (see Fig. 5).

c. Cloudiness and rainfall

Although diurnal variations in convection and precipitation have been the focus of numerous observational studies, a complete theory of the diurnal tro-
pospheric hydrologic cycle has yet to be formulated. The most fundamental mechanism for diurnal convection is the destabilization of the troposphere due to radiative heating and cooling effects. Over land, the daily heating of the underlying surface usually leads to a late afternoon maximum in convection. Over oceanic regions, the radiative cooling of overlying cloud tops can result in late night or early morning peak convec-
Fig. 9. Harmonic dials of July CCM2 diurnal convective precipitation. Values are smoothed with a five-point running mean in latitude. Maximum amplitude plotted is 3 mm day$^{-1}$. Values exceeding this are enclosed by the 4 and 6 mm day$^{-1}$ contours.

Fig. 10. Harmonic dials of July CCM2 semidiurnal convective precipitation. Arrows pointing northward indicate hour of maximum at 0000 LT, eastward at 0300 LT, southward at 0600 LT, etc. Values are smoothed with a five-point running mean in latitude. Length of maximum vector is 0.86 mm day$^{-1}$. 
tion. The diurnal cycle is conventionally thought to be weaker over water than over land. This is due to the stronger response of the land surface to solar heating, compared to that of the marine atmosphere.

These simple models of diurnal convection can be radically altered by a variety of environmental factors. Gray and Jacobson (1977) demonstrated that the diurnal cycle of oceanic convection and rainfall can be quite strong, especially when associated with the divergence patterns of organized weather systems. Their theory rests on the assumption that the diurnal radiative cycle is more intense in clear regions than in cloudy regions. The ambient atmosphere adjusts to the differential nocturnal cooling with stronger clear-air subsidence, enhancing low-level convergence into the adjacent cloudy regions. Albright et al. (1985) documented considerable variability in the diurnal cycle of deep convection and precipitation throughout the central tropical Pacific, including the presence of afternoon maxima in the southern Pacific convergence zone (SPCZ). Wallace (1975) noted a tendency for late night maxima in summertime precipitation over the central United States. He attributed this to a pattern of nighttime convergence in the boundary layer that results from the diurnal cycle of heating and cooling along the eastern slopes of the Rocky Mountains. In their study of summer convection over West Africa, Reed and Jaffe (1981) noted the tendency for late evening and early morning convection downstream of elevated terrain. They hypothesized that convective cells that develop over the adjacent highlands earlier in the day aggregate into larger mesoscale systems that continue to grow and propagate after the daytime heating has ceased.

Because of the simplified convective parameterizations and the coarse spatial resolution in the model, our validation of the diurnal cycle of CCM2 hydrology has focused only on the broadest spatial scales of the model cloudiness and precipitation. These patterns are interpreted within the context of the simplest conceptual models of diurnal convection, subject to the considerations mentioned above.

The harmonic diurnal cloudiness fraction for both solstice months are shown in Fig. 5. This quantity represents the diurnal variation of the total convective cloud cover in the atmospheric column. Over the oceans, peak diurnal convection occurs between midnight and 0600 LT. These oceanic clouds are mostly confined to the model layers between the surface and 700 mb, so they correspond largely to convection in a stable marine planetary boundary layer. In July, the amplitude of the diurnal harmonic of cloudiness is larger over the continents than the oceans. Over land, peak convection generally occurs during afternoon hours. Exceptions to this are late night maxima occurring over the northwest African coast and the Tibetan Plateau. These clouds originate in the model layers above 700 mb.

The CCM2 diurnal cloud fractions are compared with observed diurnal cloud cover taken from the climatology compiled by Warren et al. (1986). These patterns are shown in Figs. 6, 7, and 8 for seasons centered around January and July. Separate maps have been prepared for continental cumuliform cloudiness and oceanic stratiform cloudiness, since the latter correspond most closely to the CCM2 convective cloudiness confined to the layer below 700 mb.

The pattern of CCM2 July diurnal continental cloudiness is qualitatively consistent with the observations shown Fig. 6. Over land, cumulus cloud cover maximizes shortly after noon, while the deeper cumulonimbus clouds generally peak close to 1800 LT. The regions of highest continental cloudiness in the CCM2 correspond to the locations of the observed maxima. These are found over the Northern Hemisphere landmasses, equatorial belts of Africa and South America, and the Himalayan highlands. The values of the climatological maxima are higher than the CCM2 cloud fractions by about a factor of 2.

The pattern of CCM2 oceanic convection is generally consistent with the climatological diurnal variation of oceanic stratus shown in Figs. 7 and 8b. The most notable features, captured by both the model and climatology, are the cloudiness maxima off the western coasts of North America, South America, and Africa. In January, CCM2 diurnal convection of appreciable magnitude is largely confined to the Southern Hemisphere. An exception is noted over East Asia. From Fig. 8, this localized convection in the model has no observational counterpart. There is little indication in the model or the climatology of the enhanced diurnal
convection documented in the west and central Pacific (Albright et al. 1985). It is likely that the coarse model resolution, the truncation to zonal wavenumber 12, and the large grid boxes used in the Warren et al. climatology smooth out the small-scale features. Daily mean convective cloudiness maxima occur from the Philippines southwestward through southeast Asia and off the west coast of Mexico in July, and also near New Guinea and in the central equatorial Pacific in January. These correspond to total precipitation maxima in the same regions.

Figure 9 displays the harmonic dials of CCM2 July diurnal convective precipitation. Over the largest landmasses, the amplitude of the diurnal harmonic varies between 2 and 3 mm day$^{-1}$, and peaks in the middle to late afternoon hours. These values are consistent with the findings of Hamilton (1981b) for coastal and inland regions. The amplitude of the diurnal harmonic is weaker in equatorial Africa and South America, where the hour of maximum occurs closer to noon. There is no evidence in the model of late night summertime precipitation over the central United States observed by Wallace (1975). Over the tropical oceans, the model precipitation maximum generally occurs between 0000 LT and 0600 LT. While this behavior is consistent with the overall findings of Gray and Jacobson (1977), it should be noted that a number of studies report the occurrence of regional afternoon rainfall maxima over tropical oceans (Gray and Jacobson 1977; Reed 1983; Augustine 1984; Albright et al. 1985; Meisner and Arkin 1987). CCM2 produces early afternoon precipitation over the Bay of Bengal, the Arabian Sea, and northeast of the Philippines. A prominent region of 0600 LT precipitation is seen off the northeast coast of South America. This pattern may reflect the convergence of the large-scale easterly wind with the offshore flow of the nighttime sea breeze, as suggested by Kousky (1980). The CCM2 semidiurnal convective precipitation is shown in Fig. 10. At low latitudes, peak values of close to 1.0 mm day$^{-1}$ generally occur between 0100 and 0300 LT. Poleward of 30°, the hour of maximum occurs between 0400 and 0600 LT. These results agree well with the findings of Lindzen (1978) and Hamilton (1981b).

Zwiers and Hamilton (1986) analyzed the diurnal and semidiurnal rainfall in the Canadian Climate Centre model, and isolated very weak amplitudes (0.5 mm day$^{-1}$). They attributed these results in part to a fixed cloud distribution in the model ra-

![Fig. 12. Vertical distribution of the Hough components of CCM2 diurnal solar heating, in kelvins. Solid curve: Θ$^1_1$; dashed curve: Θ$^2_1$; x's: Θ$^4_2$; filled circles: Θ$^4_4$; broken curve: Θ$^4_6$. (a) July; (b) January.](image)

![Fig. 13. Vertical distribution of the diurnal Hough components of water vapor heating, in units of mW kg$^{-1}$. Multiply abscissa values by 0.0137 to convert to Fig. 12 heating units. (a) Negative of Hough mode 1 (−J$^1_1$) plotted together with Hough mode 3 (J$^3_3$). (b) Hough mode −1. (c) Negative of Hough modes −2 and −4 (−J$^2_{−2}$ and −J$^4_{−4}$) from Groves 1982.](image)
4. CCM2 diurnal heating
   a. Tropospheric shortwave radiative heating

   Figure 11 indicates the latitude–height distribution of the migrating component of the January CCM2 tropospheric shortwave solar heating. This field has a broad distribution with latitude, and extends to the 200-mb level. Quantitatively, the CCM2 diurnal heating is expressed in terms of the amplitude of the corresponding diurnal temperature perturbation. A maximum of 0.34 K is observed at about 800 mb at 30°S. In order to facilitate comparison with previous studies, this heating field is projected onto the meridional eigenfunctions of the diurnal tide, or Hough functions [see Chapman and Lindzen (1970) for further discus-
Fig. 15. Harmonic dials of diurnal CCM2 diffusive heating, at 993 mb. Length of maximum vector is 5.0°. (a) January. (b) July.

The vertical distributions of the most prominent of these modes, $\Theta^1$, $\Theta^2_1$, $\Theta^2_2$, and $\Theta^2_4$, are shown in Fig. 12, together with $\Theta^1$. In Fig. 13 the diurnal water vapor heating components calculated parametrically by Groves (1982) are shown for comparison. The CCM2 solar heating magnitudes generally agree with those of Groves, but the vertical distribution peaks well below 5 km in CCM2, while Groves' values peak between 7 and 9 km. It is likely that this disagreement results from the differences between CCM2 and Groves' model in the parameterizations of cloud absorption and scattering.

In Fig. 14 the amplitude of the 787-mb January and July CCM2 shortwave heating is mapped. The hour of maximum heating (not shown) is almost uniformly 1200 LT. The strongest heating is concentrated over
tropical and subtropical oceanic regions to the west of the continents. These locations are characterized by stratus cloud cover. These clouds can enhance the multiple scattering and increase the absorption of solar radiation incident on the clouds from below, resulting in stronger net heating rates. The July Northern Hemisphere maximum (0.63 K) is lower than the January Southern Hemisphere maximum of 0.78 K. This difference may reflect the influence of the diurnal cycle of cloud amount on the shortwave heating rate. As indicated in Fig. 5, the amount of July daytime cloudiness over land is greater than over the oceans. This results in higher daytime cloud albedos over land, which lower the solar heating rates. The afternoon continental precipitation may reduce the amount of water vapor available to absorb shortwave radiation.

b. Sensible and moist convective latent heating

Figure 15 shows the January and July CCM2 sensible heating at the lowest level above the surface. This quantity is calculated as the vertical derivative of the boundary-layer turbulent sensible heat flux. As expected, this heating is much more intense over land than oceans. For these locations, the hour of maximum generally occurs no later than 1200 LT. The vertical extent of this heating is largely confined between the surface and the 850-mb level.
The boundary-layer depths, diffusivity profiles, and nonlocal transports can vary greatly through the diurnal cycle over continental regions (Holtslag and Boville 1993). However, the diurnal behavior of the CCM2 sensible heating is qualitatively consistent with the simple conceptual models of boundary-layer heating employed in the previous works of Siebert (1961) and McKenzie (1968). This heating was defined as that which would produce a diurnal temperature oscillation satisfying the vertical diffusion equation

\[ \frac{\partial T}{\partial t} = K \frac{\partial^2 T}{\partial z^2}, \]  

with constant eddy diffusivity \( K \). The resulting heating function then becomes a function of the diurnal surface temperature perturbation. The vertical structure is a damped oscillation, with a spatial scale given by \( H_s = (2K/\sigma)^{1/2} \approx 400 \) m for \( K = 5 \times 10^4 \) cm\(^2\) s\(^{-1}\). The observations of Wallace and Patton (1970) reveal that 12-h surface temperature differences of 10 K are typical for many inland stations in the western United States, although this value can be exceeded locally by almost 100%. These surface temperature extrema tended to drop off rapidly within a few hundred meters. The amplitudes of the CCM2 continental diurnal heating are typically about 5 K at the lowest levels. These amplitudes are consistent with the temperature ranges observed by Wallace and Patton. We note, however, that the highly localized observations of Wallace and Patton are insufficient to enable a global comparison with CCM2.
Before proceeding to a discussion of the CCM2 diurnal latent heating, we discuss briefly the "time mean" (14 day) of the CCM2 moist-convective heating. Figure 16 shows latitude–height sections of this field averaged 30° in longitude about 135°E and 170°E. These are longitudes where observational studies of in situ convective heating rates are presented. The deepest convection extends to nearly 100 mb, with maximum heating occurring in the vicinity of 600 mb. Over Australian latitudes, the peak model heating reaches 15° day⁻¹. The magnitude and the vertical structure are consistent with heating estimates of Frank and McBride.
heating does not reveal the upper-tropospheric warming documented by Johnson and Young (1983) over the South China Sea.

Superimposed upon the CCM2 time mean heating distributions are diurnally varying patterns with complex three-dimensional structure. Fig. 17 indicates the harmonic dials for the CCM2 July convective heating at the 866- and 598-mb levels. At lower levels, convection over land peaks at midday, and a weaker regime of early morning marine convection is apparent. At 598 mb, the model produces most of the convective heating over the Northern Hemisphere elevated landmasses. Peak values range from 0.5 K to 0.9 K between 1800 to 0000 LT. In January, most of the convective heating occurs in the Southern Hemisphere (not shown). The amplitudes of the heating are only 60% of the July values; also, at 598 mb the hour of maximum convective heating over land occurs much closer to 0000 LT. Figure 18 depicts January latitude–height cross sections of the diurnal harmonic of convective heating at 180°E for 0000 and 0600 LT. Peak negative values are interpreted as convective heating maximizing 12 h later than the indicated local time. The heating in the lowest model levels at 0000 and 0600 LT reflects the midnight and early morning marine cloudiness. At 0000 LT there is a narrow band of deeper convective heating between the equator and 25°S. This feature bears some relation to the findings of Albright et al. (1985), who reported a maximum in deep convection around 2100 LT in the southern Pacific convergence zone. Farther south, the deep convective activity shifts to the daytime and early evening hours.

5. Summary

We have analyzed the diurnal variations in CCM2 as seen in two composite solstitial days. Our discussion has focused on certain well-documented tropospheric variations: surface pressure, OLR, convective cloud cover, and precipitation. These fields are reviewed and subjectively compared with available data. Low-latitude diurnal surface pressure tides are well simulated, as is the large-scale model OLR. The model convective scheme generates diurnally varying patterns of oceanic and continental cloudiness that are broadly consistent with ground-based observations. The major exceptions are the phase of July CCM2 convection over the Himalayas and the West African coast and the January CCM2 convection over East Asia. The global CCM2 July diurnal and semidiurnal convective precipitation patterns are generally consistent with the findings of Hamilton for tropical and extratropical regions.

The CCM2 diurnal convective heating is similar to previous estimates gleaned from far simpler models. While the magnitudes of CCM2 tropospheric radiative heating are similar to previous estimates, substantial differences in the vertical structure of this heating have

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**Figure 18.** Latitude–height distribution of the January CCM2 diurnal moist convective heating at 180°. Fields constructed from the zonal mean and 12 eastward and westward migrating waves. Values smoothed with a seven-point running mean in latitude. Contour interval is 0.02°. Dashed contours indicate negative values. (a) 0000 LT. Labels scaled by 1000. (b) 0600 LT. Labels scaled by 10 000.
been noted between CCM2 and the results of Groves (1982). The diurnal latent heating distributions have complicated structures, which are associated with multiple convection and precipitation regimes in latitude, height, and local time.

This study may not be considered a definitive assessment of CCM2 diurnal climate. The 14 samples used to generate the composite CCM2 days are sufficient to enable resolution of the diurnal and semidiurnal harmonics; however, these analyses are prone to aliasing in regions of high synoptic activity. Observations of diurnally varying cloudiness and precipitation reveal patterns of spatial complexity and day-to-day variability that are not reflected in the CCM2 climate presented here. Nor do our samples permit the examination of interannual variability of diurnal forcing and tides. The lack of detailed information on atmospheric global diurnal variations precludes a more comprehensive comparison of this abbreviated CCM2 climate with observations. Nevertheless, our study suggests that diurnally varying CCM2 features associated with ocean–continent differences and large-scale topography are qualitatively consistent with observed climatology. We conclude that, on the global scale, the model distributions of radiative, convective, and latent heating are useful proxies of the true thermotidal forcing. These conclusions are supported by recent calculations that indicate that many characteristics of the surface pressure and middle atmosphere diurnal tides are successfully modeled by linear tidal theory with CCM2 heating as input (Lieberman 1992).

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