The Diurnal March of Convective Cloudiness over the Americas

RENE D. GARREAUD AND JOHN M. WALLACE

Joint Institute for the Study of the Atmosphere and Ocean, University of Washington, Seattle, Washington

(Manuscript received 16 December 1996, in final form 11 March 1997)

ABSTRACT

Based on nine years (1983–91) of infrared data from geostationary satellites (the B3 ISCCP product), several features of the diurnal march of the frequency of convective cloudiness over the tropical and subtropical Americas are documented with 3-h temporal resolution and 0.5° × 0.5° latitude–longitude spatial resolution. The frequency of convective cloudiness in each grid box is defined in terms of the fraction of temporal samples that exhibit cloud-top temperatures colder than 235 K. The effect of varying the threshold is explored and selected results are compared with rainfall estimates based on microwave (SSM/I) imagery.

Convective cloudiness over most land areas exhibits a coherent diurnal march with relatively clear mornings, a rapid afternoon buildup, and a more gradual nighttime decay. The highest, coldest convective clouds peak a few hours earlier than those with lower tops. Morning to noontime maxima tend to be prevalent over offshore waters that experience significant convection such as the Gulf of Panama, with the highest convective clouds leading the lower ones by 6 h or more. During the austral summer, the strongest diurnal march is observed over the Andes Mountains, in a band just inland of the northeast coast of South America and in two intermediate, parallel bands over Amazonia. Between these bands, weak night and morning maxima are observed in some areas. During the boreal summer the strongest diurnal march is observed over Central America and extending northwest along the continental divide as far as the southwestern United States. In comparison to the microwave imagery, the features in convective cloudiness tend to be somewhat larger in spatial scale, and cloudiness over land tends to be biased toward the late afternoon and evening hours.

Regions with strong late afternoon/early evening maxima in convective cloudiness tend to experience more diurnal-mean convective cloudiness than those with weak (or morning) maxima. Although the evidence is less conclusive, it appears that the same may be true of rainfall. On the basis of these results it is suggested that over regions of relatively flat terrain such as Amazonia, the dynamics of the diurnal march may play a role in determining the spatial distribution of climatological-mean rainfall.

1. Introduction

The diurnal march in convective cloudiness over the tropical and subtropical Americas has been documented, based on both satellite imagery and ground-based observations. During the austral summer (December–January–February, DJF), deep convection reaches its southernmost position and largest areal extent over South America. In this season, convective cloudiness exhibits a quite coherent diurnal cycle in most of the central part of the continent, with a marked maximum around 1800 local time (LST), as reported by Meisner and Arkin (1987) and Minnis et al. (1983) on the basis of infrared satellite imagery with a rather coarse spatial resolution (2.5° lat × 2.5° long). Based on rainfall data, a marked preference for summertime afternoon thundershowers has also been documented over the southern Altiplano (Fuenzalida and Rutllant 1987) and along a band 100–300 km inland of the northeast coast of Brazil (Kousky 1980). Recently, Negri et al. (1994) produced a regional precipitation climatology over the Amazon basin during its wet season (January–May) using 3 years of twice daily Special Sensor Microwave/Imager (SSM/I) data. They found a preference for morning showers over the adjacent tropical Atlantic and alternating bands of morning/evening precipitation maxima over the interior of the continent.

During June–July–August (JJA) most of the convective activity is located over Central America and the adjacent tropical Pacific and extends northwest along the continental divide as far as the southwestern United States, in association with the so-called North American monsoon. Due to the narrowness of Central America, the harmonic analysis of Meisner and Arkin (1987) yielded a rather confused pattern, but a general tendency for a daytime (nighttime) maximum in convective cloudiness over land (water) is still clear. More specific characteristics of the diurnal march of the convection in northern Mexico and the southern United States have been recently documented by Negri et al. (1993, 1994) using high-resolution infrared images and SSM/I data. They found a striking diurnal cycle, with sharply contrasting evening and morning conditions. Convective
precipitation occurs offshore during the early morning hours, with several local maxima in concave-shaped areas of the coastline. During afternoon/evening deep convection reaches its maximum development over land, with a marked maximum along the western slope of the Sierra Madre Occidental.

In the present study, we have used 9 years of equivalent blackbody temperature from geostationary satellites, with 3-h temporal resolution and 0.5° × 0.5° latitude–longitude spatial resolution that enables us to resolve small-scale features in the diurnal march of the deep convective cloudiness over the tropical and subtropical Americas and follow its evolution over the course of the day. Although techniques based on the IR channel alone might overestimate the areal extent and lifetime of the actual convective systems due to the existence of anvil cirrus clouds, satellite-based infrared imagery has been widely used for studying the temporal variability and spatial patterns of tropical and subtropical convection (e.g., Arkin and Ardanuy 1989).

This paper is organized as follows. Details of the satellite data and its preprocessing to project pixel-map images onto a regular latitude–longitude grid are given in section 2. The general pattern of the diurnal march of the convective cloudiness over the tropical Americas and adjacent waters is presented in section 3. A more detailed description of the diurnal march of the convective cloudiness over three key regions is presented in section 4. The selected regions are the subtropical Andes, the northeast coast of South America, and Central America. Finally, a discussion of the results is presented in section 5.

2. Data and preprocessing

a. ISCCP B3 data

The data used in this work are 9 years (December 1983–February 1991) of reduced resolution radiance images (the B3 International Satellite Cloud Climatology Project [ISCCP] product) from the Geostationary Operational Environmental Satellites (GOES) number 5 (GOES East), 6 (GOES West), and 7. The satellites used in each year and their corresponding sublongitudes are listed in Table 1. Detailed descriptions of the B3 dataset are given elsewhere (e.g., Rossow and Schiffer 1991); here we will comment only on the relevant features of these data in the context of this work. The B3 dataset corresponds to a compressed version of the original images from the geostationary satellites (in particular the GOES series) to a nominal spatial resolution of 30 km, every 3 h (eight images per day beginning at 0000 UTC), normalized with reference to the Advanced Very High Resolution Radiometer mounted on the NOAA polar-orbiting satellites (Schiffer and Rossow 1985; WCRP 1985). The original imaging GOES data come from half-hourly measurements of the Visible Infrared Spin Scan Radiometer (VISSR) mounted on these satellites, with detection wavelength ranges of 0.55–0.75 μm and 10.5–12.5 μm, and nominal resolution of 0.9 km and 6.9 km in the visible and infrared channels, respectively (WCRP 1985). In the IR band, differences down to 1 K in equivalent blackbody temperature can be resolved by the VISSR. In the present study we have used the calibrated values for the infrared channel (nominally centered at 11.6 μm) of the GOES series, expressed as normalized equivalent blackbody temperature $T_b$.

b. Gridding

In an attempt to describe the climatological features of the deep, moist convection over the tropical Americas, a conversion from the original pixel-by-pixel images to a regular latitude/longitude grid was performed. The analysis was limited to the region 40°N–40°S and 120°–30°W. For each image, the approximately 90 000 pixels within this region were sorted, according to their nominal center coordinates, into a uniform 0.5° × 0.5° latitude–longitude grid. Then, a gridded $T_b$ was calculated as the average of the pixel values within those latitude–longitude boxes that contained at least two pixels. The choice of the regular grid size (0.5° × 0.5°, approximately 360 km²) was a compromise between being able to resolve small-scale details of the convective cloudiness and having only a small fraction of boxes with missing data (less than two pixels per box). Figure 1 shows the mean (area and time average) frequency distribution of the number of pixels per box. For a typical map, most boxes (65%) contain between three and five pixels. The 6% of the boxes with missing data are mainly concentrated over the Atlantic Ocean, near the eastern edge of the spatial domain and thus do not introduce a significant problem. Completely missing images over the region of interest comprise less than 8% of the complete 9-yr record.

In the following, most of the analysis is based on the 3-h fractional coverage $F^*$ of clouds colder than 235 K in 0.5° × 0.5° boxes, where $F^*$ is the number of temporal samples with $T_b \leq 235$ K divided by the total number of temporal samples at each grid point. The threshold $T_b = 235$ K, corresponding to a cloud-top height above the 300-hPa level, was chosen because it is cold enough to be unaffected by changes in blackbody temperature related to variations in ground temperature or low-level cloudiness. We did not attempt to apply the GOES precipitation index (Arkin and Meisner 1987) because it is not ideally suited to the high-resolution gridded data used in this study (Richards and Arkin

<table>
<thead>
<tr>
<th>Satellite</th>
<th>Years</th>
<th>Sublongitude</th>
<th>Limb</th>
</tr>
</thead>
<tbody>
<tr>
<td>GOES-5 (East)</td>
<td>1983–84</td>
<td>75°W</td>
<td>0°W</td>
</tr>
<tr>
<td>GOES-6 (West)</td>
<td>1984–87</td>
<td>135°W</td>
<td>60°W</td>
</tr>
<tr>
<td>GOES-7</td>
<td>1987–92</td>
<td>Intermediate</td>
<td>30°W</td>
</tr>
</tbody>
</table>

Table 1. Satellite data used in this study.
1981), and because the lack of an adequate ground-based rain gauge network over a large fraction of the tropical Americas precludes a meaningful calibration. Hence, our results are more indicative of variations in convective cloudiness than variations in rainfall. We also studied the sensitivity of our results to variations in the temperature threshold by calculating the 3-h mean fractional coverage of cloud with thresholds ranging from 190 to 270 K, in 5-K increments over selected regions of active convection.

3. General patterns of the climatological diurnal march

In this section, we will focus our analysis on the general patterns of the climatological diurnal march of convective cloudiness over the Americas during the rainy seasons. For future reference, climatological mean lower- and upper-troposphere winds, together with terrain elevation, are presented in Fig. 2. For a more comprehensive description of the annual march of the convective activity and the atmospheric circulation over this region the reader is referred to Aceituno (1988), Horel et al. (1989), and Douglas et al. (1993). Figure 3 shows the daily mean fractional coverage of \( T_\beta \) \(< 235 \) K (hereafter denoted as \( F \)) for December–January–February (DJF) and June–July–August (JJA). The standard deviation maps \( (sF) \) of the 3-h fractional cold cloud coverage \( (F^*) \) for each season, also presented in Fig. 3, provide a rough estimate of the amplitude of the mean diurnal march of the convective cloudiness. The seasonal standard deviation of \( F^* \), calculated at each grid point, is the square root of the mean squared anomalies at each hour (0000, 0300, \ldots, 2100 UTC) relative to their seasonal daily mean. Figure 4 shows the seasonal-mean fractional coverage of convective cloudiness during the local evening (2100 and 2400 UTC) and morning (0900 and 1200 UTC) hours, respectively.

a. Spatial distribution

In both seasons, the spatial pattern of \( sF \) over land resembles the daily mean fractional coverage \( (F) \) in most areas with a significant frequency of convective cloudiness. Hence, regions that experience a high frequency of convective cloudiness (primarily aligned with local topographic features) tend to exhibit a strong diurnal march (Fig. 3). Similar results have been reported in other studies (e.g., Meisner and Arkin 1987; Hartmann and Recker 1986).

During DJF, maxima of convective cloudiness over the central and southern part of the Amazonia is organized in two parallel bands, extending from northwest to southeast for more than 2000 km and about 400 km wide (b2 and b3 in Figs. 3, 4). Additionally, there are two bands of maximum convective cloudiness along the subtropical Andes (b1 in Figs. 3, 4) and along the northeast coast of the continent (b4 in Figs. 3, 4). This four-banded structure, particularly evident in late afternoon and early evening (Fig. 4 and 5), differs markedly from the picture of a broad, rather amorphous area of maximum convective cloudiness over the central part of South America, based on coarse resolution imagery. This banded pattern constitutes a robust feature of the summertime convective cloudiness over South America, with three of the four bands of evening high convective cloudiness present in virtually the same locations each year (Fig. 5), whereas the band immediately east of the Andes is more variable in position and less marked but still present in every year of our analysis.

During JJA, the influence of regional factors (concave coastal shapes and mountain ranges) in the North American summer monsoon is evidenced by several local maxima of both evening cloudiness \( (F^* \) at 2100 and 2400 UTC) and diurnal variability \( (sF) \) over the western slope of the Sierra Madre Occidental, the Yucatan peninsula, around the Gulf of Panama, and along the Atlantic coast of northern South America (Figs. 3 and 4). The Florida peninsula and the major Caribbean islands also exhibit a prominent diurnal cycle of convective cloudiness during this season.

Over the ocean it is possible to distinguish between two regimes in the diurnal march. Near coastal regions with a high frequency of convective cloudiness also exhibit strong diurnal variability, particularly in the local concavities in the coastline, although the ratio \( sF/F \) is generally not as high as over land. Such regions include the Gulf of Panama and the Pacific coast of Central America during JJA, and to a lesser extent the northeast coast of South America (most notably between March and May). In contrast, the broad areas of high convective cloud coverage over the adjacent Pacific (JJA) and Atlantic (DIF) intertropical convergence zone (ITCZ) exhibit standard deviations \( (sF) \) of less than 3%, indicative
of a weak diurnal march. Other oceanic areas of extensive convective cloud coverage, including the Gulf of Mexico in JJA and the South Atlantic convergence zone are also characterized by a small standard deviation of $F^*$.

**b. Temporal evolution**

To summarize the mean diurnal march in convective cloudiness, we have applied empirical orthogonal function (EOF) analysis to the seasonal mean 3-h anomalies (relative to the day-mean field, $F$) of the cold ($T_c = 235$ K) cloud coverage. For each season, the EOF analysis was based on the temporal covariance matrix ($8 \times 8$ elements). Although most studies have used harmonic analysis for this purpose, we have used EOF analysis in order to better capture the distinctive shape of the diurnal march in convective cloudiness. In recognition of the pronounced differences between the diurnal march over continental and oceanic regions, we have treated them separately. In all seasons, the first eigen-mode is strongly dominant (see Table 2) and statistically significant at the 99% confidence level using the $N$-rule (Overland and Preisendorfer 1982); therefore, we concentrate our analysis on the leading seasonal principal components (PCs, Fig. 6) and EOFs (Fig. 7) during the austral and boreal summer.

The leading mode of $F^*$ over land explains more than 80% of the variance associated with the diurnal march. Its spatial pattern (Fig. 7) is reminiscent of the distributions of $F$ and $sF$ (Fig. 3). Its evolution (Fig. 6) is characterized by a rapid increase during afternoon
(1300–1700 LST), a maximum around 1800 LST, and a more gradual decay during the night and morning hours, which may be partially a reflection of long-lived anvil debris. The peak convective cloudiness in the North American monsoon (JJA) occurs around sunset, an hour or two later than in the South American monsoon (DJF). An exception to this diurnal march occurs during DJF along the eastern slope of the subtropical Andes and over the Parana basin (centered at 30°S, 60°W) where cloudiness exhibits a weak predawn maximum. Exceptions to the prevailing daytime maxima of convective cloudiness over land have also been reported in other regions, based on both rainfall records (e.g., Wallace 1975) and satellite imagery (e.g., Short and Wallace 1980; Negri et al. 1994).

EOF analysis of the diurnal march of $F^*$ over ocean also yields significant leading modes in both seasons, with maximum cloudiness around noon in coastal areas where the EOF exhibits large, positive values. The larger amplitude of the leading oceanic PC during JJA reflects the greater prominence of near-coastal convection in the North American monsoon. The highly convective area of the Gulf of Panama (especially during JJA) serves as an outstanding example of this regime, analogous to that observed in the South China Sea (Nitta and Sekine 1994). EOF analysis of $F^*$ restricted to a

Fig. 3. Upper panels: daily fractional cold ($T_c \leq 235$ K) cloud coverage during December–January–February (DJF) and June–July–August (JJA). Lower panels: As in upper panels but for the seasonal standard deviation of the 3-h fractional cold cloud coverage. The 2000- and 4000-m topographic contours are also indicated. Labels b1, b2, b3, and b4 refer to bands of maximum summertime convective cloudiness over South America (see text for details).
coastal band 200 km wide yields similar results to those obtained with the full oceanic domain, although the peak-to-peak amplitude of the leading PC during DJF increases from 9.5% to 13.3% (not shown).

c. Threshold top-temperature dependence

Figure 8 shows the 3-h mean fractional area covered by clouds with brightness temperature $T_b$ within 5-K intervals ranging from 190 to 270 K for Amazonia and the Gulf of Panama. Over Amazonia, the coverage of clouds with $T_b < 200$ K peaks at 1500 LST, about 3 h before the peak in clouds with $T_b$ in the 235–240-K range. Warmer clouds (240–260 K) cover a significantly larger area, and reach their maximum extent during evening hours. Over the Gulf of Panama the very cold clouds ($T_b < 200$ K) reach their maximum areal extent before sunrise, whereas the coverage by increasingly warmer clouds peaks progressively later through the morning and into the early afternoon. In this later case, the time difference between the peak coverage of clouds with tops colder than 200 K and 235 K is about 6 h, and coverage of warm clouds (240–260 K) peaks about 12 h after the onset of the very cold clouds. Qualitatively similar results were obtained for other regions. Table 3 summarizes the dependence of the timing of maximum cloud coverage as a function of cloud-top temperature for several areas of marked convection over land and

![Figure 4](image-url)
FIG. 5. Fractional cold (\(T_b \approx 235 \text{ K}\)) cloud coverage at 2100 UTC (upper panel) and 1200 UTC (lower panel) during the austral summer (DJF) for the years 1984–92 along a transect from the Central Andes to the mouth of the Amazon River (AA'). The small dots indicate the crest of the Andes and the Atlantic coastline. Labels b1, b2, b3, and b4 refer to bands of maximum summertime convective cloudiness over South America (see text for details).

4. Regional features

Within the general framework described in the previous sections, the diurnal march in the frequency of convective clouds exhibits some noteworthy regional characteristics. Rather than exhaustively documenting these features, we will focus on three specific regions during their respective rainy seasons: the subtropical portion of the Andes Mountains (the Altiplano), the northeast coast of South America and Central America.

TABLE 2. Fraction of diurnal variance explained by the leading eigenvectors of the seasonal 3-h fractional cold (\(T_b < 235 \text{ K}\)) cloud coverage.

<table>
<thead>
<tr>
<th>Season</th>
<th>Land EOF</th>
<th>Sea EOF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer (DJF)</td>
<td>87%</td>
<td>62%</td>
</tr>
<tr>
<td>Autumn (MAM)</td>
<td>83%</td>
<td>69%</td>
</tr>
<tr>
<td>Winter (JJA)</td>
<td>81%</td>
<td>70%</td>
</tr>
<tr>
<td>Spring (SON)</td>
<td>83%</td>
<td>65%</td>
</tr>
</tbody>
</table>

a. Subtropical Andes (DJF)

Frequencies of convective cloudiness in excess of 25% extend from 10° to 22°S slightly eastward of the axis of maximum terrain elevation (Figs. 3, 4), mainly forced by the orographic uplift of moist air over the eastern side of the Andean range (Fuenzalida and Rutllant 1987; Lenter and Cook 1995; see also Fig. 2). This ocean. In all cases, warmer clouds peak later and cover larger areas than cold clouds. Consequently, the cumulative frequency of occurrence of cloud-top temperature less than 235 K (\(F^<\)) is largely determined by the coverage of clouds in the 225–235-K range. The differences between the diurnal march of high cloud over land and ocean, also documented in previous studies (e.g., Hartmann and Recker 1986), suggest differences in the character of the convective systems prevalent over land and ocean.

FIG. 6. Upper panel: amplitude of the first principal components of the EOF analysis of seasonal 3-h mean fractional cold (\(T_b < 235 \text{ K}\)) cloud coverage over land during DJF (solid line) and JJA (dashed line). Lower panel: as in upper panel but for cold cloud coverage over the ocean. Local time referred to 60°W in DJF and 90°W in JJA.
region exhibits a marked diurnal march, illustrated by the evening minus morning cold cloud coverage (Fig. 9) and the time–distance section of $F^*$ that crosses the subtropical Andes at 18°S (Fig. 10a). The Andean crest is virtually free of convection during late night/early morning, with frequencies of cold clouds less than 2% from 0300 to 1100 LST. Coverage increases rapidly from 1200 to 1700 LST and reaches a maximum in excess of 30% (up to 55% over local peaks) around 1800 LST, followed by a more gentle decrease up to midnight. Similar explosive afternoon growth of convection has been documented over the highlands of tropical West Africa by Reed and Jaffe (1981). The results over the subtropical Andes are consistent with the diurnal cycle of precipitation over the southern Altiplano documented by Fuenzalida and Rutllant (1987). The timing of the Altiplanic convection is presumably related to the pronounced diurnal cycle in thermodynamic conditions over this region (Aceituno et al. 1994). For qualitative comparison, Fig. 10 also includes a time–distance section of $F^*$ during JJA that intersects the Sierra Madre Occidental at 23°N. In the North American case, the convection occurs predominantly over the western slope of the range, and convective cloudiness peaks around sunset.

Over the eastern slope of the subtropical Andes and

---

**Fig. 7.** Upper panels: first EOF of the analysis of the seasonal 3-h fractional cold ($T_s \leq 235$ K) cloud coverage over land during December–January–February (DJF) and June–July–August (JJA). Lower panels: as in upper panels but for cold cloud coverage over the ocean. Maximum negative value: $-0.012$. 

---
mean diurnal march of the fractional area covered by cloud with brightness temperature $T_b$ within 5-K intervals ranging from 190 to 270 K as a function of the cloud-top temperature. (a) Amazonia region ($6^\circ\pm12^\circ S, 53^\circ\pm48^\circ W$) during DJF. (b) Gulf of Panama ($1^\circ\pm8^\circ N, 82^\circ\pm77^\circ W$) during JJA. Scale of both panels indicated at the bottom of the figure.

### Table 3. Time of maximum cloud coverage as a function of the cloud-top temperature $T_b$ for different regions of active convection.

<table>
<thead>
<tr>
<th>Region (season)</th>
<th>Reference time</th>
<th>Temperature range (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>230–235</td>
</tr>
<tr>
<td></td>
<td></td>
<td>210–220</td>
</tr>
<tr>
<td></td>
<td></td>
<td>250–260</td>
</tr>
<tr>
<td>Land</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Amazonia (DJF)</td>
<td>17.9</td>
<td>-2.6</td>
</tr>
<tr>
<td>Subtropical Andes (DJF)</td>
<td>17.8</td>
<td>-1.8</td>
</tr>
<tr>
<td>SMO (JJA)</td>
<td>20.2</td>
<td>-1.8</td>
</tr>
<tr>
<td>Southeastern US (JJA)</td>
<td>19.5</td>
<td>-2.2</td>
</tr>
<tr>
<td>Sea</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gulf of Panama (JJA)</td>
<td>13.2</td>
<td>-6.1</td>
</tr>
<tr>
<td>NE coast of Brazil (MAM)</td>
<td>9.8</td>
<td>-5.7</td>
</tr>
<tr>
<td>NE coast of Panama (JJA)</td>
<td>12.2</td>
<td>-6.0</td>
</tr>
</tbody>
</table>

Regions: Amazonia: $6^\circ\pm12^\circ S, 53^\circ\pm48^\circ W$; subtropical Andes: $17^\circ\pm21^\circ S, 73^\circ\pm69^\circ W$; SMO (Sierra Madre Occidental): $24^\circ\pm27^\circ N, 110^\circ\pm107^\circ W$; southeastern United States: $31^\circ\pm35^\circ N, 92^\circ\pm89^\circ W$; Gulf of Panama: $1^\circ\pm8^\circ N, 82^\circ\pm77^\circ W$; NE coast of Brazil: $2^\circ\pm2^\circ N, 49^\circ\pm46^\circ W$; NE coast of Panama: $10^\circ\pm12^\circ N, 81^\circ\pm79^\circ W$.

Figure 8. Mean diurnal march of the fractional area cover by cloud with brightness temperature $T_b$ within 5-K intervals ranging from 190 to 270 K as a function of the cloud-top temperature. (a) Amazonia region ($6^\circ\pm12^\circ S, 53^\circ\pm48^\circ W$) during DJF. (b) Gulf of Panama ($1^\circ\pm8^\circ N, 82^\circ\pm77^\circ W$) during JJA. Scale of both panels indicated at the bottom of the figure.

an adjacent band of about 300 km that compose the western limit of the Amazon basin, frequencies of cold clouds in excess of 10% persist throughout the day with late afternoon and late night/early morning maxima (Fig. 10a). The late afternoon maximum is almost coincident with the peak cloudiness over the crest of the Andes and over the plains farther to the east. The more marked predawn maximum is a distinctive feature found all along the subtropical Andes. It is presumably related to enhanced low-level convergence during nighttime as a result of the mountain-plain circulation between the Andes and the Amazon basin (Kousky and Bell 1996). Further inspection of Fig. 10a reveals a nighttime displacement of the western edge and axis of the area of strongest convection frequency from the Andean ridge toward the eastern margin of the Amazon basin at a speed on the order of 12 m s$^{-1}$, a feature also documented by Minnis et al. (1983).

### b. Northeastern coast of South America (MAM)

Figure 11a shows the evening (1800 and 2100 LST) minus morning (0900 and 1200 LST) difference of the fractional convective cloud cover over this region during MAM. For comparison, we also included the evening minus morning difference of the estimated precipitation during January–May (Fig. 11b, the same as Fig. 7 of Negri et al. 1994), based on 3 years of SSM/I data over part of this region (Negri et al. 1994). As will be discussed later, the more diffuse features and the weakness of the morning maxima (i.e., areas of negative P.M.-

![Figure 9](image)

**Figure 9.** Evening (1800–2100 local time) minus morning (0900–1200 local time) difference of the fractional cold ($T_b = 235$ K) cloud coverage during the austral summer season (DJF). Topographic contours 2000 and 4000 m are also indicated.
minus-A.M. difference) in Fig. 11a relative to those in Fig. 11b may be an artifact of the fixed temperature threshold \( T_b = 235 \text{ K} \) used in our analysis. The two figures are in close agreement with respect to the banded structure over land. Additional details concerning the convective cloudiness can be inferred from two space–time sections (Fig. 12) of \( F^* \) with axes oriented normal to the Atlantic coastline. The 300-km-wide band just inland from the coast exhibits a marked diurnal march, with nearly convective cloud-free conditions from midnight to noontime and frequencies of cold clouds increasing during the afternoon to a maximum in excess of 30% around 1800 LST. Offshore, the diurnal march is less marked, although weak maxima around 0600 and 1400 LST are evident. The early afternoon peak propagates onshore, perhaps in association with the sea-breeze front (Kousky, 1980).

During late afternoon/early evening, there is a secondary band of maximum convective cloudiness over central Amazonia, centered 1500 km inland from the coast and about 300 km wide (Fig. 12). The configuration of this feature in the space–time sections is suggestive of an afternoon reactivation of the landward propagating coastal convection from the previous day, which moves with a phase speed on the order of 15 m s\(^{-1}\). A similar situation is observed during DJF. The nocturnal migration is more clearly defined over the northwestern transect (Fig. 12a). The phase speed and typical dimensions of this climatological nighttime landward displacement of the area of maximum convective cloudiness suggest that it might be the signature of the Amazon coastal squall lines (ACSLs) documented by Cohen et al. (1995). These mesoscale to synoptic-scale systems are particularly frequent during the austral autumn, and during their mature stage they appear as discontinuous lines of discrete cells, with mean longitudinal and transverse dimensions of 700 km and 200 km, respectively. The ACSLs move inland at mean speeds.
of 13 m s$^{-1}$, and the presence of onshore low-level flow is thought to be instrumental in producing their westward migration (Cohen et al. 1995). The more pronounced landward propagation of the convection in Fig. 12a might be a reflection of the more prevalent onshore low-level flow northwest of the mouth of the Amazon (Fig. 2b).

c. Central America and Gulf of Panama (JJA)

The evening minus morning difference of the frequency of convective clouds over Central America during JJA (Fig. 11c) reveals the typical land–sea contrast in the timing and maximum development of the convection. A similar pattern emerges in the P.M.-minus-A.M. difference of the estimated precipitation during July–September as shown in Fig. 11d, based on the SSM/I data (Negri et al. 1994). As in the northeast coast of South America, the features tend to be more concentrated in the SSM/I imagery, particularly the offshore morning maxima. The infrared imagery in Fig. 11c also shows some suggestion of an evening maximum over the tropical Pacific about 700 km off the coast of Central America, which is not evident in SSM/I imagery.

The time-latitude section of $P^*$ centered along 79°W (Fig. 13a) indicates that convective cloudiness reaches its maximum frequency and areal extent over the Gulf of Panama (see also Fig. 3) with values in excess of
30% from 0600 to 1700 LST and a maximum just after noontime. Convective cloudiness is also more frequent over the Caribbean side of the isthmus during this part of the day, although the values of $F^*$ are not as high. Convection over Panama exhibits an equally strong but more concentrated maximum centered around 1800 LST. Over the eastern boundary of the Gulf of Panama (not shown), convective cloudiness over land exhibits a maximum between 2200 and 0600 LST, which moves seaward during the predawn hours, where it spreads out and intensifies around noontime. This feature is better defined in the equinoctial seasons (centered in April and October) when the seasonal migration of the ITCZ brings on the wet season over the northwestern part of South America.

A diurnal migration of the area of maximum frequency of convective cloudiness from land southward or southwestward into the Pacific is evident all along Central America. There is a suggestion of this behavior in Fig. 11c and a clearer indication in Fig. 13b, which shows a transect normal to Central America that intersects the Pacific coastline at 14°N. As in the previous section, convection over land is largely restricted to the hours between sunset and midnight. During late night and early morning the convection over land subsides and the axis of maximum frequency of convective clouds migrates offshore in both directions. The continued southwestward migration of this feature appears to be responsible for the band of enhanced afternoon convective clouds about 700 km off the Central American coast, pointed out in the discussion of Fig. 11c. This propagation and spreading of the high, cold clouds over the tropical Pacific could reflect the behavior of the convection itself, but it could also be at least partially the reflection of anvil clouds emerging from the coastal thunderstorms and being advected by the mean northeasterly flow at upper levels over this region (Fig. 2c).

5. Discussion

This paper documents the climatological diurnal march of convective cloudiness over the tropical and subtropical Americas, based on the occurrence of cloud-top temperatures lower than 235 K. Our results agree with the general patterns documented in previous studies based on 2.5° × 2.5° latitude–longitude spatial resolution (e.g., Meisner and Arkin 1987), but they also resolve a number of small-scale signatures that have been only recently documented in regional studies based on...
twice daily high-resolution microwave imagery (e.g., Negri et al. 1994).

Over land during the rainy season, the evolution of the clouds with top temperatures in the 190–235-K range is consistent with our current understanding of the life cycle of deep cumulus convection. The buildup phase during the afternoon is more rapid than the decay phase, which extends through most of the night. The coldest clouds, which are presumably connected with the strongest cumulus updrafts, reach their peak amplitude by 1500 LST. Moderately cold (~235 K) clouds peak within the next few hours and warmer (~250 K) clouds do not reach their peak frequency until around midnight. Continental areas that exhibit the highest frequency of convective cloudiness tend to be primarily aligned parallel to certain coastlines, such as the north-east coast of South America, and the main topographic barriers (i.e., the Andes Mountains and the Sierra Madre Occidental). Within these regions, local thermally driven circulations (land–sea breezes and mountain–valley winds) provide a regular day-to-day dynamical forcing (boundary layer convergence) that favors convection at the time of day that the boundary layer stratification is most unstable. Over the central part of South America, a conspicuous banded pattern of areas of maximum afternoon/evening convective cloudiness is evident during the rainy season even though the local terrain is rather flat. A similar banded pattern is apparent in the afternoon mean precipitation inferred from the SSM/I measurements (Fig. 5 of Negri et al. 1994). We speculate that this feature might be a remote response to the strong, locally forced diurnal cycle in convection over the subtropical Andes and the northeast coast of South America, via gravity currents (e.g., Rottuno et al. 1988) or gravity waves (e.g., Nicholls et al. 1991). Clearly, further modeling studies are required to understand this point.

Daily mean convective cloudiness and late afternoon–early evening convective cloudiness exhibit remarkably similar spatial signatures over the continents. Results of Negri et al. (1994; their Figs. 1, 5, 8 and cover) suggest that the same is true of rainfall, provided that the sum of their twice-daily rainfall estimates is reasonably representative of the true diurnal mean. Hence, it appears that on average, regions with late afternoon–early evening maxima tend to receive more (diurnally averaged) rainfall than regions with maxima at other times of day. The former are presumably regions in which the dynamical forcing discussed in the previous paragraph occurs in phase with the thermodynamic forcing of the
convection and the latter to regions in which the two types of forcing occur out of phase, resulting in a smaller net forcing. Since deep convection can occur only when the forcing exceeds some critical threshold, it seems plausible that regions that are subject to a strong diurnal forcing should experience a higher frequency of convection averaged over the course of the day even though the hours during which it occurs might tend to be more restricted.

Over the ocean it is possible to distinguish two regimes of the diurnal march. Near-coastal regions with a high frequency of convective activity exhibit a marked diurnal variability, strongly influenced by land–sea breezes, in contrast to the much weaker diurnal march observed over the open ocean in the intertropical convergence zones. During the boreal summer (JJA) the near-coastal convection in the North American monsoon exhibits a diurnal march as marked as over land but characterized by a near noontime maxima and midnight minima. A similar diurnal march has been documented by Nitta and Sekine (1994) over the South China Sea. Very cold clouds ($T_b < 200$ K) develop, reach their maximum horizontal extent and disappear during late night/early morning, followed by the predominance of successively warmer (200–240 K) clouds throughout the morning and early afternoon. This behavior seems consistent with the model of convection over the sea near Borneo proposed by Houze et al. (1981), where the maximum frequency and spatial extension of the cold cloud coverage around midday is related to the horizontal expansion of the anvil shield and the decay of convective cells (with very cold-top temperatures), which actually peak around sunrise.

We have refrained from making any quantitative inferences concerning the diurnal march of rainfall because it is not possible to distinguish the precipitating area of convective systems. Additionally, a substantial fraction of tropical and subtropical rainfall is associated with stratiform cloudiness, whose presence is not fully captured in our IR-based analysis. Both of these problems are evident in the comparison between the climatological evening minus morning difference in fractional cold cloudiness and precipitation inferred from SSM/I measurements over northeast Brazil and the southern United States (Fig. 11). First, the features in SSM/I tend to be more concentrated than their counterparts in convective cloudiness (as defined by $T_b < 235$ K), indicating that 235 K may be too warm as a threshold for outlining areas of convective rainfall, in agreement with results of Negri et al. (1993). It is also noted that climatological P.M. minus A.M. estimated precipitation over the northeastern part of tropical South America (Fig. 11b, Fig. 7 of Negri et al. 1994) exhibits alternating areas of positive and negative values of comparable magnitude, indicating that some regions experience substantially more morning than afternoon precipitation. In contrast, the climatological evening-minus-morning difference in the frequency of clouds with $T_b < 235$ K (Fig. 11a) is strongly biased toward positive values (P.M. > A.M.). Similar discrepancies are evident during JJA over some areas in the southern United States (Figs. 11c and 11d). Hence, it appears that over part of the domain, the frequency of cold (<235 K) clouds may be biased toward late afternoon/early evening, relative to the associated precipitation. In recognition of these problems, an algorithm based on both IR data (with their 3-h sampling) and SSM/I data (with their more faithful representation of hydrometeors) appears to be the most reliable strategy for producing precipitation estimates.

**Acknowledgments.** The B3 ISCCP data were obtained from NASA Langley Research Center EODIS Distributed Active Archive Center. Andrew J. Negri kindly provided climatological precipitation fields inferred from SSM/I measurements. Sarah Doherty provided a crucial code for data processing. Helpful comments and discussion from Brad Smull, Chris Bretherton, and Dennis L. Hartmann are greatly appreciated. One of the authors (RG) is partially supported by the National Science Foundation under Grant 9215512 and by the Department of Geophysics of the Universidad de Chile.

**REFERENCES**


Lente, J. D., and K. H. Cook, 1995: Simulation and diagnosis of the


