A Comparison of TRMM Microwave Techniques for Detecting the Diurnal Rainfall Cycle

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(Manuscript received 24 February 2005, in final form 18 August 2005)

ABSTRACT

This paper uses rainfall estimates retrieved from active and passive microwave data to investigate how spatially and temporally dependent algorithm biases affect the monitoring of the diurnal rainfall cycle. Microwave estimates used in this study are from the Tropical Rainfall Measuring Mission (TRMM) and include the precipitation radar (PR) near-surface (2A25), Goddard Profiling (GPROF) (2A12), and PR–TRMM Microwave Imager (TMI) (2B31) rain rates from the version 5 (v5) 3G68 product. A rainfall maximum is observed early evening over land, while oceans generally show a minimum in rainfall during the morning. Comparisons of annual and seasonal mean hourly rain rates and harmonics at both global and regional scales show significant differences between the algorithms. Relative and absolute biases over land vary according to the time of day. Clearly, these retrieval biases need accounting for, either in the physics of the algorithm or through the provision of accurate error estimates, to avoid erroneous climatic signals and the discrediting of satellite rainfall estimations.

1. Introduction

Monitoring global rainfall and its diurnal variability is conducted using in situ methods and remotely through the application of retrieval algorithms. Studies of the diurnal cycle from ground radar and rain gauges are limited by regional biases and lack of observations over the ocean. Satellite algorithms using visible and infrared techniques provide excellent temporal and spatial sampling from geostationary orbits, but rainfall is inferred from cloud-top characteristics and estimates are subject to threshold biases and contamination by nonraining cirrus (Todd et al. 1999). Passive microwave retrieval techniques observe the natural radiation from the earth and atmosphere. The main source of microwave radiation atmospheric attenuation is ice and/or liquid hydrometeors, because microwave radiation is relatively insensitive to cloud liquid water and water vapor at the frequencies used for rainfall monitoring. Over the ocean, the absorption/emission signal associated with liquid hydrometeors at frequencies <40 GHz is used to determine rain rates. Over land, high and variable emissivity masks the atmospheric signal in the low-microwave range, and rainfall retrieval has to rely on the scattering of microwave radiation by cloud-ice particles at frequencies >50 GHz. A myriad of passive microwave rainfall retrieval techniques exist today ranging from simple statistical algorithms (Kidd 1998) to the more complex physical algorithms incorporating radiative transfer models (Panegrossi et al. 1998) or highly physical inversion approaches (Kummerow et al. 2001).

Because passive microwave sensors are constrained to low Earth orbiters (LEO), sampling of the diurnal cycle is incomplete. Consequently, several diurnal rain-
fall studies have used a synergy of infrared and microwave techniques, taking advantage of the strengths of each of them (Sorooshian et al. 2002). The Tropical Rainfall Measuring Mission (TRMM) is placed in a non-sun-synchronous, low circular orbit that precesses slowly through the diurnal cycle. TRMM provides passive microwave diurnal estimates with high spatial resolution (30 km × 18 km at 19 GHz and 7 km × 5 km at 85 GHz) and the first spaceborne “active” microwave measurements from the precipitation radar (PR) (~4.3 km). PR assesses the backscattering of an artificially transmitted signal, and provides the most direct means of spaceborne retrievals to date (Iguchi et al. 2000; Anagnostou et al. 2001).

Although technological and scientific advancements have been made in the field of satellite rainfall retrievals, some recent rainfall algorithm intercomparison studies show large discrepancies in absolute rainfall values. Investigations into the forcing of precipitation by El Niño–Southern Oscillation (ENSO) show differences in rainfall trends (Soden 2000; Berg et al. 2002). These biases are attributed to temporal (e.g., seasonal and interannual) and spatial (e.g., East versus West Pacific) variations in cloud microphysical properties, which are currently unaccounted for in the satellite algorithms. These biases lead to the conclusion that current satellite rainfall retrievals are most likely inadequate for quantitatively monitoring short- and long-term rainfall variability for some applications, such as agriculture, hydrology, and climate change (Soden 2000). To our knowledge, no study has shown how algorithm biases vary with the diurnal rainfall cycle at the global and regional scale. Algorithm comparisons are essential for understanding the limitations of the satellite rainfall monitoring, especially if such data are to be used for validating global climate models.

This work supplements a growing number of diurnal cycle studies using data from TRMM (Imaoka and Spencer 2000; Kistawal and Krishnamurti 2001; Shimizu et al. 2001; Nesbitt and Zipser 2003). The goal of this paper is to compare seasonal microwave diurnal cycle estimates at the global and regional scales for the first three years of TRMM (1998–2000). The study evaluates version 5 (v5) rainfall estimates from the PR near-surface algorithm (2A25), the Goddard Profiling (GPROF) TRMM Microwave Imager (TMI) algorithm (2A12) and PR–TMI combined algorithm (2B31). The paper is organized into five further sections. The current understanding of land and ocean diurnal cycles is reviewed in section 2. Section 3 provides a brief description of the datasets and methodology, with a strong emphasis on TRMM sampling considerations. The diurnal cycle and the algorithm differences are described in section 4. The results are discussed and implications are provided in sections 5 and 6.

2. Background on diurnal cycle processes

It is suggested that the tropical diurnal rainfall cycle is highly organized (Lin et al. 2000), as summarized in Table 1. Over both land and ocean, the strongest diurnal cycles are found in regions of intense convective activity, while different precipitation types also peak at different times (Dai 2001; Nesbitt and Zipser 2003). Most studies show a more pronounced diurnal cycle over land compared to over ocean because water bodies have a smaller and slower response to radiational cooling and warming (Meisner and Arkin 1987). Over land, intense solar heating during the day increases lower-tropospheric temperatures and therefore instability, which is conducive to the observed and modeled late-afternoon/early evening maximum (e.g., Yang and Slingo 2000; Sorooshian et al. 2002; Nesbitt and Zipser 2003). Radiational cooling overnight increases stability, resulting in an early morning minimum. Over land, the idealized diurnal rainfall cycle is modulated by regional influences such as sea breezes and mountain winds, which provide regular radiative forcing that enhances rainfall (Garreau and Wallace 1997; Negri et al. 2000; Yang and Slingo 2001).

Over the ocean, the diurnal cycle is significant but weaker, as illustrated in Table 1. Most studies document an early morning or nocturnal peak (e.g., Gray and Jacobson 1977; Chang et al. 1995; Imaoka and Spencer 2000), although some studies show afternoon maximums (Albright et al. 1985; Sui et al. 1997). A number of processes support the early morning maximum over the ocean, including direct radiation–convection (Kraus 1963; Randall et al. 1991), radiation–dynamic convection (Gray and Jacobson 1977), changes in moisture contents with radiational cooling and heating (Sui et al. 1997), and possibly pressure tides (Dai 2001). The oceanic afternoon maximum is a consequence of surface–cloud–radiation interactions (Chen and Houze 1997) and changes in the life cycle of rainfall systems (Sui et al. 1997). Continental influences modulate the oceanic diurnal cycle through land breezes and gravity waves leading to a more pronounced diurnal cycle (Liberti et al. 2001; Yang and Slingo 2001).

3. Data, sampling, and methodology

a. Data and algorithm description

The main data source for the study is the TRMM satellite, which has provided rainfall estimates between 37.5°N and 37.5°S since mid-December 1997. TRMM's
platform and sensors are comprehensively described in Kummerow et al. (1998, 2000). This study uses data from December–January–February (DJF) 1998 to September–October–November (SON) 2000 from the 3G68 (v5) dataset, available from TRMM’s Science Data and Information System (TSDIS) (online at http://tsdis02.nascom.nasa.gov/). A number of algorithm versions have been released since the launch, with some significant differences (Adler et al. 2000; Ferreira et al. 2001), and reprocessing of the 3G68 data began in April 2004 for the version 6 (v6) release. Three years of data are used to avoid any spatial resolution inconsistencies resulting from TRMM’s orbital boost in August 2001. The dataset provides gridded

<table>
<thead>
<tr>
<th>Data</th>
<th>Reference</th>
<th>Region</th>
<th>Land</th>
<th>Ocean</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gauge</td>
<td>Gray and Jacobson (1977)</td>
<td>Islands in Pacific and Atlantic 15,000 stations, 50°S–70°N</td>
<td>Drizzle maximum 0600 LST; storms/convective maximum late afternoon</td>
<td>Early morning maximum, 2–3 times greater than evening maximum</td>
</tr>
<tr>
<td></td>
<td>Dai (2001)</td>
<td></td>
<td></td>
<td>Drizzle maximum at 2400–0400 LST, storms maximum at 2400 LST, convective rainfall at 0000–0400 LST</td>
</tr>
<tr>
<td>IR</td>
<td>Allbright et al. (1985)</td>
<td>Central Pacific and SPCZ</td>
<td></td>
<td>Within-ITCZ morning maximum, evening minimum; evening maximum for SPCZ</td>
</tr>
<tr>
<td></td>
<td>Shin et al. (1991)</td>
<td>Central Pacific</td>
<td></td>
<td>Maximum early to late evening, semidiurnal exists for ITCZ and SPCZ only</td>
</tr>
<tr>
<td></td>
<td>Mapes and Houze (1993)</td>
<td>Warm pool, 80°E–180°W</td>
<td></td>
<td>Cold clouds peak at dawn (&lt;208 K); moderate cold clouds (&lt;235 K) peak in the evening; small systems have no cycle</td>
</tr>
<tr>
<td>GCM</td>
<td>Janowiak et al. (1994)</td>
<td>Global Tropics 2.5° × 2.5°</td>
<td>Cold cloud greater frequency 1800–2100 LST</td>
<td>Colder clouds peak in early morning, warmer clouds peak in the evening</td>
</tr>
<tr>
<td></td>
<td>Lin et al. (2000)</td>
<td>Tropics</td>
<td>Stronger cycle over land, amplitude of 3–5 mm day⁻¹ during late/early evening</td>
<td>Deep convective zones amplitude 1–2 mm day⁻¹, peaking in the morning; low latitudes and eastern tropical oceans have small amplitudes</td>
</tr>
<tr>
<td></td>
<td>Lim and Suh (2000)</td>
<td>Global</td>
<td>Peaks at 1430 LST, amplitude 2.0 mm day⁻¹; semidiurnal 1400–1500 LST, 1.0 mm day⁻¹</td>
<td>Peaks at 0600 LST, amplitude 0.1 mm day⁻¹, semidiurnal 0.5 mm day⁻¹</td>
</tr>
<tr>
<td></td>
<td>Yang and Slingo (2001)</td>
<td>Tropics</td>
<td>Evening maximum, regional variation (local circulations and MCSs life cycles)</td>
<td>Early morning maximum, continental influences increase amplitude</td>
</tr>
<tr>
<td>SSM/I</td>
<td>Chang et al. (1995)</td>
<td>50°N–50°S, ocean</td>
<td></td>
<td>35%–40% early morning maximum morning greater than evening by 20%</td>
</tr>
<tr>
<td></td>
<td>Sharma et al. (1991)</td>
<td>50°N–50°S, ocean</td>
<td></td>
<td>Regional and seasonal variations but morning-to-evening ratio 1:2</td>
</tr>
<tr>
<td>TMI, SSM/I</td>
<td>Imoaka and Spencer (2000)</td>
<td>Tropical Oceans, 10° × 10°</td>
<td></td>
<td>Peaks at 0400–0700 LST in deep Tropics; amplitude 14% of mean, &gt;10° weaker cycle (except 20°–30°S).</td>
</tr>
<tr>
<td></td>
<td>IR, TMI</td>
<td>Negri et al. (2002)</td>
<td>Amazonia</td>
<td>IR, TMI technique underestimate (overestimate) radar 0200–1100 (1400–1900) LST</td>
</tr>
<tr>
<td>TMI, PR</td>
<td>Nesbitt and Zipser (2003)</td>
<td>40°N–40°S</td>
<td>Maximum evening for precipitation features with ice, and without ice and MCSs; MCSs persist into night; convective rain fraction increases during afternoon</td>
<td>Maximum contribution by MCSs early morning; increase in rainfall because of greater frequency, not changes in area or rain rates; convective/stratiform fraction constant all day</td>
</tr>
</tbody>
</table>
surface rainfall at 0.5° × 0.5° resolution from three retrieval techniques. Rainfall estimates from PR 2A25 (Iguchi et al. 2000), GPROF 2A12 (Kummerow et al. 2001), and PR–TMI 2B31 (Haddad et al. 1997) are available. The PR 2A25 rainfall estimates are calculated from a hybrid of the Hitschfeld–Bordan method (Iguchi et al. 2000) and a surface reference technique (Meneghini et al. 2000) and are sensitive to reflectivities >17 dBZ and rain rates >0.7 mm h⁻¹ (Kummerow et al. 2000). Over the ocean, GPROF 2A12 uses both atmospheric emission and scattering of passive microwave radiation to retrieve rainfall through a physically based inversion profile technique (Kummerow et al. 2001). Over land, only the scattering signal at 85 GHz is used in GPROF 2A12 (McCollum and Ferraro 2003). The PR–TMI 2B31 uses a Bayesian approach to match observed radiances to a likely radar-rain profile and drop size distribution (DSD) (Haddad et al. 1997).

For each algorithm, 3G68 provides the total number of pixels, total number of rainy pixels, mean rain rate, convective percentage, and time of the latest overpass for each grid. We discuss both PR collocated (GPROF) and noncollocated (GPROF NC) estimates from the Goddard algorithm. GPROF NC estimates are derived from the entire TMI swath (760 km), while GPROF estimates are derived from the PR collocated swath width (215 km).

b. Sampling considerations

TRMM precesses slowly through the diurnal cycle. Prior to the orbital boost, TRMM took 46 days at high latitudes or 23 days at the equator to fully sample the diurnal cycle, with each progressive orbit being 0.033 h earlier (Negri et al. 2002). Therefore, long periods of time could occur between a sample at a given hour and the next observation at a similar hour. Near the equator, equally spaced samples occurred every 12 h. At 30° there were four to five overpasses every 90 min followed by a 20-h gap (Kishtawal and Krishnamurti 2001). The nonuniform temporal coverage poleward of 30° most likely results in increased noise in the presence of transient regimes. Negri et al. (2002) show that sampling errors at 5° × 5° resolution are similar for PR and TMI despite the differences in swath widths. Moreover, 4-hourly averaging over larger 12° × 12° grid boxes reduces the sampling error for each sensor by a factor of 2 (Negri et al. 2002).

To identify latitudes where unequal sampling occurs, the cumulated number of satellite pixels for each hour is shown for 10° × 10° regions over the Southern Hemisphere’s Pacific Ocean (from 170°E to 180°) (Fig. 1). Results are shown for collocated and noncollocated PR and TMI estimates between DJF 1998 and SON 2000, and all times refer to Local Standard Time (LST). Clearly, the number of samples increases from the equator to 35°S, and the finer spatial resolution of the PR sensor provides a greater number of pixels for each hour compared to the TMI sensor. Note that similar results are found for the Northern Hemisphere, and that variability with longitude is small and results from the time partition. The 3-yr annual mean has equal sampling throughout the day at 10° × 10° resolution. Large climatological averaging reduces the impact of transient regimes to provide stable estimates. Reducing the time period to 3-yr seasonal means (not shown) produces some unequal sampling throughout the day, and therefore less stable diurnal estimates. For this study, spatial resolutions less than 10° × 10° are not considered appropriate. Although comparisons with ground-based data would provide a better assessment of diurnal cycle errors, they are not essential for our scope, which is to compare algorithm retrieval errors.
is found (Fig. 2). Rain rates remain fairly

**c. Methodology**

The average diurnal cycle calculated from $0.5^\circ \times 0.5^\circ$ resolution data is described for various spatial extents, including global ($40^\circ$N–$40^\circ$S, $180^\circ$E–$180^\circ$), regional (Table 2), and $10^\circ \times 10^\circ$ grid boxes. Land, ocean, and coastal areas are determined from a $1/6^\circ$ percentage water database from the National Oceanic and Atmospheric Administration (NOAA) (land = 0%–10%, ocean = 100%, coast = 11%–99%). Land, ocean, and coast constitute 24.8% [number of pixels ($N$) = 28539], 72.6% ($N = 83636$), and 2.6% ($N = 3025$) of the $0.5^\circ \times 0.5^\circ$ global database, respectively. Analysis is not included for coastal areas because of the small sample size and because GPROF has well-documented problems over mixed pixels (Kummerow et al. 2001). Harmonic analysis, which produces a phase and amplitude akin to the timing and magnitude of the peak diurnal rainfall, is used to simplify spatial comparisons at $10^\circ \times 10^\circ$. The harmonics are calculated using the least squares approach (Wilks 1995). The significance of the diurnal and semidiurnal harmonics is calculated from the proportion of the total variance of the mean diurnal cycle and the total variance of the harmonics (Broughs 1994). The significance is expressed as a percentage and shows how well the harmonics represent the observed diurnal cycle. Differences between the phase and amplitude determined from the various algorithms are also shown every $10^\circ \times 10^\circ$.

Significant differences between the PR and GPROF/PR–TMI algorithms are determined using a paired Student’s $t$ test. A paired Student’s $t$ test compares two algorithms at each hour and takes account of the ordering of the observations. The main assumptions for a paired Student’s $t$ test are a normal distribution and independence of observations. Analysis of histograms shows that over large areas the climatological distribution of rainfall is fairly normal (not shown). According to the central limit theorem, the hourly mean values of a given time series possess a Gaussian distribution, regardless of the gamma distribution of the time series (Wilks 1995). A significance level of 0.01 ($p = 0.01$) is used throughout the study.

**4. Results**

**a. Global diurnal rainfall cycles**

1) **LAND**

Figure 2 shows the 3-yr seasonal mean diurnal rainfall cycles for each algorithm (1998–2000) over land and ocean. For all seasons over land, an afternoon rainfall peak (1500–1800 LST) of 3.0–4.0 mm day$^{-1}$ and a morning rainfall minimum (0800–1100 LST) of 1.0–2.0 mm day$^{-1}$ is found (Fig. 2). Rain rates remain fairly large throughout the night and early hours.

There are discrepancies between the algorithms’ detection of the global diurnal cycle. GPROF provides the largest estimates (Fig. 2), while PR’s are the lowest. Moreover, the absolute bias between the algorithms varies throughout the day (Fig. 2). Figure 3 shows the mean annual normalized difference at each hour between PR–TMI/GPROF and PR [(Algorithm-PR)/PR×100], and illustrates that relative biases also vary over land. Product relative and absolute similarities are greatest when ice contents are minimal, namely, during the morning rainfall minima and the initial cloud growth ($\pm 5\%$) (Fig. 3). GPROF exceeds PR to the greatest extent during and directly after the rainfall

<table>
<thead>
<tr>
<th>Region</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Surface</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tropical Africa</td>
<td>015°N–015°S</td>
<td>010°W–050°E</td>
<td>Land</td>
</tr>
<tr>
<td>Subtropical Africa</td>
<td>015°N–015°S</td>
<td>010°W–050°E</td>
<td>Land</td>
</tr>
<tr>
<td>Tropical South America</td>
<td>015°N–015°S</td>
<td>080°–020°W</td>
<td>Land</td>
</tr>
<tr>
<td>Subtropical South America</td>
<td>015°N–015°S</td>
<td>080°–030°W</td>
<td>Land</td>
</tr>
<tr>
<td>Australia</td>
<td>015°N–015°S</td>
<td>120°–170°E</td>
<td>Land</td>
</tr>
<tr>
<td>India–Southeast Asia</td>
<td>035°–010°N</td>
<td>060°–140°E</td>
<td>Land</td>
</tr>
<tr>
<td>Southeast United States–Mexico Gulf</td>
<td>025°–015°N</td>
<td>120°–060°W</td>
<td>Land</td>
</tr>
<tr>
<td>East Pacific</td>
<td>015°N–005°S</td>
<td>130°–080°W</td>
<td>Ocean</td>
</tr>
<tr>
<td>Central Pacific</td>
<td>015°N–005°S</td>
<td>180°–130°W</td>
<td>Ocean</td>
</tr>
<tr>
<td>Northwest Pacific</td>
<td>015°–005°N</td>
<td>130°E–180°</td>
<td>Ocean</td>
</tr>
<tr>
<td>Southwest Pacific</td>
<td>005°N–015°S</td>
<td>150°E–180°</td>
<td>Ocean</td>
</tr>
<tr>
<td>South Pacific convergence zone</td>
<td>005°–025°S</td>
<td>180°–130°W</td>
<td>Ocean</td>
</tr>
<tr>
<td>Tropical Atlantic</td>
<td>015°N–015°S</td>
<td>040°W–010°E</td>
<td>Ocean</td>
</tr>
<tr>
<td>Tropical Indian</td>
<td>010°N–015°S</td>
<td>040°–100°E</td>
<td>Ocean</td>
</tr>
<tr>
<td>Maritime Continent</td>
<td>015°N–015°S</td>
<td>100°–150°E</td>
<td>Mixed</td>
</tr>
</tbody>
</table>
peak when larger and greater ice contents are found aloft (>25%). Relative differences between PR–TMI and PR are smaller, with the least (greatest) differences during the afternoon–evening rainfall peak (late-morning minima) at 7.5% (16%) (Fig. 3).

Table 3 shows the mean bias between the algorithms when detecting the global annual and seasonal diurnal cycles during the 3-yr time period over 24 h (number of hours N = 24). The results of the paired Student’s t tests indicate that significant differences (p = 0.01) between GPROF and PR occur for all seasons. Absolute differences are smallest for June–August (JJA) (0.32 mm day\(^{-1}\)) and are greatest for March–May (MAM) (0.49 mm day\(^{-1}\)) and DJF (0.47 mm day\(^{-1}\)). GPROF and GPROF NC differences are small (0.0–0.08 mm day\(^{-1}\)) and insignificant for the annual and seasonal diurnal cycles, and so are not included. The PR–TMI rain rates mirror PR estimates, but are slightly larger (0.25–0.31 mm day\(^{-1}\)) (Fig. 2 and Table 3). The paired Student’s t tests revealed that differences between the PR and PR–TMI are significant for all seasons (Table 3). Differences between PR–TMI and GPROF are significant for
MAM (0.23 mm day$^{-1}$) and SON (0.19 mm day$^{-1}$) (Table 3).

2) OCEAN

The oceanic diurnal cycle is more subdued, with rain rates consistently between 2.0 and 3.5 mm day$^{-1}$, depending on the algorithm (Fig. 2). Early morning (0300-0700 LST) rainfall peaks from ~0.5 to 1.0 mm day$^{-1}$ above the daily average, depending on season. GPROF significantly overestimates PR for all seasons ($p = 0.01$), ranging from 0.49 mm day$^{-1}$ during DJF to 0.33 mm day$^{-1}$ during JJA (Table 3). Over the ocean, PR and GPROF absolute and relative differences throughout the day are more consistent than over land (Fig. 2 and Fig. 3). The PR and PR–TMI global oceanic diurnal cycles agree extremely well and show negligible differences (<0.05 mm day$^{-1}$) (Fig. 2). Note that for MAM and JJA, PR–TMI is less than PR during the global early morning peak. Figure 3 shows that PR–TMI (GPROF) and PR relative differences oscillate around ±5% (14%–22%). The differences between GPROF and GPROF NC are also small and insignificant over the ocean, with a mean bias from −0.02 to −0.07 mm day$^{-1}$ (not shown).

b. Regional diurnal cycles

Specific geographical regions, similar to that in Nesbitt and Zipser (2003), are used to evaluate regional algorithm variations. Table 2 provides the 15 geographical areas assessed in this study with their geolocation and surface type, as determined from the 0.5° × 0.5° resolution. Table 4 provides the mean bias between the algorithms and PR when detecting the mean annual diurnal cycle. Figure 4 shows that regional differences in algorithm biases clearly exist. Because of similarities with GPROF, GPROF NC results are not shown.

The Maritime Continent is the only mixed surface region in the study, and has large rain rates throughout the day (5.0–8.5 mm day$^{-1}$) (Fig. 4h). Differences between PR and GPROF (1.75 mm day$^{-1}$) are greatest over the Maritime Continent (Table 4). The PR and PR–TMI differences are also statistically significant (0.63 mm day$^{-1}$) over the Maritime Continent (Table

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**Table 3.** Mean bias (mm day$^{-1}$) between algorithms when detecting the global diurnal cycle over land and ocean. The mean bias is the average difference between the algorithms over the entire diurnal cycle (column minus row). Paired Student’s $t$ tests were performed and significant differences at the 0.01 level are shown by *.

<table>
<thead>
<tr>
<th>Season</th>
<th>Algorithm</th>
<th>PR–TMI</th>
<th>GPROF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Land</td>
<td>Annual</td>
<td>PR</td>
<td>0.28*</td>
</tr>
<tr>
<td></td>
<td></td>
<td>PR–TMI</td>
<td>0.15</td>
</tr>
<tr>
<td></td>
<td>DJF</td>
<td>PR</td>
<td>0.31*</td>
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<td></td>
<td>PR–TMI</td>
<td>0.16</td>
</tr>
<tr>
<td></td>
<td>MAM</td>
<td>PR</td>
<td>0.27*</td>
</tr>
<tr>
<td></td>
<td></td>
<td>PR–TMI</td>
<td>0.23*</td>
</tr>
<tr>
<td></td>
<td>JJA</td>
<td>PR</td>
<td>0.30*</td>
</tr>
<tr>
<td></td>
<td></td>
<td>PR–TMI</td>
<td>0.02</td>
</tr>
<tr>
<td></td>
<td>SON</td>
<td>PR</td>
<td>0.25*</td>
</tr>
<tr>
<td></td>
<td></td>
<td>PR–TMI</td>
<td>0.19*</td>
</tr>
<tr>
<td>Ocean</td>
<td>Annual</td>
<td>PR</td>
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<tr>
<td></td>
<td></td>
<td>PR–TMI</td>
<td>0.45*</td>
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<tr>
<td></td>
<td>DJF</td>
<td>PR</td>
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<td>PR–TMI</td>
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<td>PR</td>
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</tr>
<tr>
<td></td>
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<tr>
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<td>JJA</td>
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<tr>
<td></td>
<td></td>
<td>PR–TMI</td>
<td>0.47*</td>
</tr>
</tbody>
</table>

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**Fig. 3.** Normalized difference between the algorithms and PR rain rates [i.e., (algorithm – PR)/PR ×100] for the global mean annual diurnal rainfall cycle (1998–2000) over (a) ocean and (b) land. Land and ocean is determined from a percentage water database at 0.5° × 0.5° resolution.
For pure land regions, the largest diurnal rainfall peaks and algorithm differences occur over tropical South America, with peak rain rates of 7.5–9.0 mm day\(^{-1}\) and algorithm biases of 0.86 (PR–TMI) and 1.33 mm day\(^{-1}\) (GPROF) when compared to PR (Fig. 4c and Table 4). Note that a smaller secondary peak is detected during the night/early morning over tropical South America (4.0–6.5 mm day\(^{-1}\)) (Fig. 4c). Over subtropical South America, rain rates are consistently between 3.0 and 4.5 mm day\(^{-1}\) throughout the day. Tropical Africa has a similar diurnal cycle to that of tropical South America, although the rainfall peaks are smaller and occur slightly later (Fig. 4a). In addition, algorithm differences are smaller over tropical Africa, especially for PR–TMI (0.23 mm day\(^{-1}\)) (Table 4). The differences between PR and PR–TMI are insignificant over subtropical Africa, which has a distinct afternoon rainfall peak (~4 mm day\(^{-1}\)). Excellent agreement between PR and GPROF (0.02 mm day\(^{-1}\)) occurs over the southeast United States and the Mexico Gulf region, where a fairly large afternoon rainfall peak (5 mm day\(^{-1}\)) is found (Fig. 4g and Table 4). Small diurnal cycles are noted over Australia (Fig. 4e) and India–Southeast Asia (Fig. 4f) because annual averaging reduces the distinct rainy season cycles (not shown). For both these regions, differences between PR and PR–TMI (0.19 and 0.36 mm day\(^{-1}\)) are greater than that between PR and GPROF (0.00 and 0.20 mm day\(^{-1}\)) (Table 4).

All of the oceanic regions show an early morning maximum of various amplitudes (Figs. 4i to 4o). The largest rain rates are found over the southwest and northwest Pacific Ocean, ranging from approximately 4.0 to 7.0 and from 3.5 to 6.5 mm day\(^{-1}\), respectively (Figs. 4k and 4l). The largest algorithm differences occur over the southwest Pacific, East Pacific, and central Pacific Oceans. For example, over the East Pacific, GPROF exceeds PR by 1.4 mm day\(^{-1}\) (Table 4). Similarities between the algorithms are greater where rain rates are the smallest, namely, in the tropical Atlantic and Indian Oceans. For example, over the tropical Atlantic PR and PR–TMI differences are insignificant and small (0.04 mm day\(^{-1}\)), while PR and GPROF biases are slightly larger (0.24 mm day\(^{-1}\)) (Table 4).

### c. 10° × 10° diurnal cycles

At 10° × 10° resolution, annual and seasonal diurnal cycles are assessed over the globe using two approaches. Primarily, regions where there are significant differences between the algorithms and PR mean annual and seasonal diurnal cycles are determined using paired Student’s t tests and are plotted in Figs. 5–6.

#### 1) Paired Student’s t Test results

Figure 5 shows that some differences between PR and PR–TMI are significant when detecting the annual and seasonal diurnal rainfall cycle over the smaller 10° × 10° areas. The PR–TMI significantly underestimates PR in some regions of the northern and southern oceans poleward of 30°, mainly during periods of less intense convection in the winter hemisphere. In contrast, PR–TMI significantly exceeds PR in some regions of strong convection, such as the intertropical convergence zone (ITCZ) over Africa, South America, and the tropical oceans.

For the annual and seasonal diurnal cycles, GPROF significantly exceeds PR in regions of intense convection over land and ocean (e.g., the Maritime Continent, Africa, South America, and the central, eastern, and western Pacific Ocean). In contrast, during the dry season over East Africa/Madagascar, Australia, the southeast United States, and parts of northern Africa, GPROF significantly underestimates PR. GPROF is significantly less than PR in the drier subtropical high pressure regions over the ocean. A seasonal dependence on the sign of the bias is noted in some regions of the oceans beyond 30°. GPROF overestimates (underestimates) PR during the summer (winter) when con-
Fig. 4. The annual diurnal cycle of regional rainfall (DJF 1998–SON 2000) for TRMM microwave algorithms over (a) tropical Africa, (b) subtropical Africa, (c) tropical South America, (d) subtropical South America, (e) Australia, (f) India–Southeast Asia, (g) southeast United States–Mexico Gulf, (h) Maritime Continent, (i) East Pacific, (j) central Pacific, (k) northwest Pacific, (l) southwest Pacific, (m) SPCZ, (n) tropical Atlantic, and (o) tropical Indian Ocean. For regional definitions see Table 2.
Convection is stronger (weaker), as shown in the southern Indian Ocean and northern Pacific Ocean. GPROF and GPROF NC both produce similar results when compared to PR, despite the differences in sampling (not shown).

2) Harmonic Results

Table 5 provides a summary of the 3-yr mean annual amplitude and phase over 42 land (<10% water) and 150 ocean (100% water) $10^\circ \times 10^\circ$ grid boxes. Note that because of the coarse resolution, many mixed grids are removed from the analysis (e.g., Madagascar). Over land, the mean amplitude of the annual diurnal harmonic is highest for GPROF (1.03 mm day$^{-1}$) and lowest for PR (0.75 mm day$^{-1}$), with a mean peak phase between 1700 and 1800 LST. GPROF peaks later than the other algorithms. Over the ocean, the mean annual diurnal harmonic peaks between 0600 and 0700 LST with a small amplitude (0.31–0.45 mm day$^{-1}$).

Cross correlations between the algorithms’ annual diurnal harmonic phases and amplitudes identify which algorithms produce similar regional variations. Table 6 shows the results for land, ocean, and all $10^\circ \times 10^\circ$ grids within the study area. Geographical variations in the amplitude of the diurnal harmonic are similar for each algorithm, with correlations between the algorithms ranging from 0.73 to 0.98 (Table 6). The relationships between the algorithm diurnal phases are weaker, ranging from 0.61 to 0.80 over land and from 0.44 to 0.8 over ocean (Table 6). In particular, the correlations between GPROF NC and PR/PR–TMI diurnal phases are weak, especially over ocean (0.44–0.57). The improved sampling of the larger TMI swath width may contribute to the different phases determined by GPROF NC when compared to the other algorithms.

Figure 7 shows the global distribution of the PR’s mean annual diurnal and semidiurnal harmonics, along with their significance. The phases of the diurnal harmonic are represented by the direction of the arrow (Fig. 7a). The semidiurnal harmonic phase represents a dual peak and therefore has no direction plotted on the arrow (Fig. 7c). The shading in Fig. 7a and Fig. 7c represents the amplitudes of the diurnal and semidiurnal harmonics, respectively. The significance of each harmonic shows how well the phase and amplitude represent the observed variance in the diurnal cycle. It is not the purpose of this paper to describe in detail the geographical/seasonal variations in the diurnal harmonics. As a result, only the annual harmonic results are presented. However, it is worth noting that strong and significant diurnal peaks occur during the summer hemisphere, when insulation is maximized (not shown). For more detailed descriptions of the diurnal cycle,
readers are referred to Sorooshian et al. (2002) and Dai (2001).

The annual diurnal harmonic (Fig. 7a) generally explains a large proportion (>50%) of the observed diurnal variance over land and ocean (Fig. 7b). The largest annual diurnal amplitudes (1.0–2.9 mm day\(^{-1}\)) are found over regions of intense convection over land, such as within the ITCZ (South America, Africa, Maritime Continent) and during the rainy season or summer monsoons (e.g., India, East Asia, Central America, Southern United States). The annually averaged diurnal harmonic phase maximum occurs predominantly in the afternoon/early evening maximum (1500–1800 LST) (Fig. 7a). The largest oceanic diurnal amplitudes
(<1.0 mm day\(^{-1}\)) occur over the ITCZ (Pacific Ocean, Atlantic Ocean, and surrounding the Maritime Continent) and the northern and southern ocean convergence zones during summer. However, the oceanic diurnal amplitudes are generally less than 0.5 mm day\(^{-1}\), and there is a clear tendency for early morning peaks over most regions (0300–0600 LST) (Fig. 7a).

Generally, the semidiurnal cycle does not capture the annually averaged diurnal variance well, explaining less than 50% of the diurnal variance (Fig. 7d). The main exception occurs over northern South America and central Africa. For both regions, the phase is approximately 1800 or 0600 LST, while the amplitude is ap-
approximately 1.0–2.9 mm day\(^{-1}\) (Fig. 7c). Other regions that show a significant (>50%) annual semidiurnal harmonic include parts of the Maritime Continent, Asia, Japan, and central Australia. Over the ocean, the southern Pacific convergence zone (SPCZ) also shows a significant semidiurnal harmonic, which peaks at approximately 0600–1800 LST (Fig. 7c and Fig. 7d). Examination of the seasonal results indicates that the semidiurnal harmonic has the largest amplitudes and greatest significance during intense summer convection (not shown).

Regions where the percentage differences between the PR and algorithm annual diurnal harmonics are greater than 50% or 100% are shown in Fig. 8a and Fig. 8b, respectively. As expected, PR and PR–TMI diurnal amplitudes are similar, with differences mostly <50% (Fig. 8a). The main exceptions occur over parts of the southern and northern Atlantic/Pacific Oceans, where PR–TMI underestimates PR amplitudes by >100% (Fig. 8b). GPROF and GPROF NC exceed PR diurnal amplitudes by 50%–100% over Central America/northern South America, the Maritime Continent, and parts of the Pacific Ocean (East, West, and central). GPROF underestimates PR diurnal amplitudes by more than 100% over parts of Northern Africa, while over the subtropical high pressure cells GPROF’s underestimation is between 50% and 100% (Fig. 8).

Figure 9 shows 10° × 10° grids where the algorithm annual diurnal phase maximums differ from PR by more than 1 or 2 h. The PR–TMI and PR diurnal amplitudes are mainly within an hour of each other, except for parts of the Maritime Continent and the northern and southern oceans. GPROF and PR differences are found over parts of Amazonia and northern Africa, where GPROF lags PR by 2 h or more (Fig. 9b). GPROF also peaks later than PR over many areas of the ocean, and in particular the Atlantic and Pacific Oceans. Of all of the products, GPROF NC phase maximums are the farthest in time to PR’s, and this is attributed to the sampling differences of the swath widths at the lower spatial resolution.

### 5. Discussion

**a. Global algorithm differences**

Over land, the global annual and seasonal results show that a large 3.5–4.0 mm day\(^{-1}\) afternoon rainfall peak exists between 1500 and 1800 LST. These results are clearly consistent with the research listed in Table 1 and supports that solar heating destabilizes the lower troposphere. The large nocturnal rain rates coincide with an increase in stratiform rain fraction, as observed by PR data in the Nesbitt and Zipser (2003) study.

Comparisons of the algorithms’ global means revealed discrepancies over land. Generally, GPROF detects the highest and PR detects the lowest global mean rain rates. Absolute differences between GPROF/ GPROF NC and PR are greater than the PR–TMI and PR differences. Moreover, GPROF and PR biases vary throughout the day, which is consistent with changes in cloud microphysics during precipitating system life.

### Table 6. Cross correlations between the mean annual diurnal amplitudes and phase detected by the algorithms every 10° × 10° for land, ocean, and total (DJF 1998–SON 2000). Note the correlations are all significant at the 0.01 level.

<table>
<thead>
<tr>
<th>Number</th>
<th>PR–TMI</th>
<th>GPROF</th>
<th>GPROF NC</th>
<th>PR–TMI</th>
<th>GPROF</th>
<th>GPROF NC</th>
</tr>
</thead>
<tbody>
<tr>
<td>Land</td>
<td>0.98</td>
<td>0.86</td>
<td>0.77</td>
<td>0.99</td>
<td>0.76</td>
<td>0.80</td>
</tr>
<tr>
<td>Ocean</td>
<td>0.95</td>
<td>0.89</td>
<td>0.73</td>
<td>0.71</td>
<td>0.75</td>
<td>0.44</td>
</tr>
<tr>
<td>Total</td>
<td>0.97</td>
<td>0.87</td>
<td>0.78</td>
<td>0.82</td>
<td>0.82</td>
<td>0.69</td>
</tr>
</tbody>
</table>

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5. Discussion

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cycles. GPROF exceeds PR most during and directly after the rainfall peak when larger and greater ice contents are found aloft. At the rainfall peak, there are stronger vertical updrafts and greater buoyancy because of the lower-tropospheric instabilities caused by solar heating. The stronger updrafts transport greater quantities of supercooled water to the mixed zone, where ice and liquid particles coexist, to allow ice crystals to grow in size (riming) and produce a stronger ice scattering signature (Toracinta and Zipser 2001; Toracinta et al. 2002). It is hypothesized that the stronger ice scattering signature causes GPROF to overestimate rainfall as the scatter-based algorithm does not account for variation in the cloud ice characteristics in the rain-rate conversion (McCollum and Ferraro 2003). Beamfilling errors are also reduced for the larger afternoon rainfall systems, which simultaneously acts to reduce any underestimation in the passive microwave retrieval. Following the rainfall peak, the anvil cloud top spreads out over a large area and rain rates are less intense. GPROF overestimates the rainfall at this time because high ice contents remain in the anvil. The high ice contents in the cloud anvil also lead to an anomalous delayed peak rainfall in the GPROF algorithm during the boreal summer. Previous work has also shown a delayed 85-GHz scattering signature when compared to radar reflectivity measures (Nesbitt and Zipser 2003). At the decaying stages of convective storms, the frequency of downdrafts increases, depriving the cloud of supersaturated air. Ice formation is reduced and stratiform rainfall processes become important (Houze 1997; Nesbitt and Zipser 2003). The PR’s insensitivity to light rainfall (<0.7 mm day\(^{-1}\)), incorrect drop size distribution (DSD) assumptions, and attenuation effects most likely contribute to the differences between products (Iguchi et al. 2000). During the morning minimum, between 0800 and 1100 LST, relative and absolute agreement between the GPROF and PR is greatest.

The global oceanic diurnal cycle has a small early morning peak (0300 and 0700 LST) that increases rain rates by 0.5–1.0 mm day\(^{-1}\). The ocean results therefore agree with previous findings (Table 1). Early morning peaks support the direct radiation (Kraus 1963) or radiation–dynamic convection mechanism (Gray and Jacobson 1977) and coincide with the maximum drizzle and thunderstorm frequency (Dai 2001). Nesbitt and Zipser (2003) also found a small early morning peak and attribute it to an increased number of mesoscale convective systems (MCSs) over the ocean, rather than an increase in convective rainfall or conditional rain rates.

Over the ocean, absolute mean differences between PR and GPROF were slightly larger than over land and were significant for all seasons. However, the algorithm relative and absolute differences are more consistent throughout the day compared to land, which suggests that the nature and microphysical properties of the oceanic systems do not vary greatly throughout the day. According to Nesbitt and Zipser (2003), the oceanic convective/stratiform rainfall fractions and conditional rain rates remain constant throughout the day, which may explain the consistent biases. The PR and PR–TMI absolute differences are also small over the ocean. Also, PR–TMI underestimates PR’s early morning rainfall peak during JJA and MAM, which may be explained by regional biases in the southern oceans during winter.

b. Regional algorithm differences

Regional comparisons are important because climatological biases have been shown to vary both temporally and spatially (Berg et al. 2002). In this study, PR has been taken as reference because it is generally considered to be the most accurate rainfall monitoring system to date (Anagnostou et al. 2001). Nevertheless, the PR algorithm has known biases, with recent comparisons over the oceans revealing 30% underestimation when compared to oceanic buoy measurements in the heavy rainfall area of the tropical Pacific (Bowman et al. 2003; Serra and McPhaden 2003). The PR underestimates rainfall because it is insensitive to light rain rates (<0.7 mm h\(^{-1}\)), and where this rainfall dominates, for example, the subtropical high pressure cells, it can contribute up to 20% of the total rainfall and up to 46% of the rainfall area (Bolen and Chandrasekar 2000; Schumacher and Houze 2000). At higher rain rates, especially over land, underestimation is attributed to attenuation errors (Iguchi et al. 2000). Although an attenuation correction factor exists, it is subject to considerable uncertainties and can affect rain rates by a factor of 10 or more (Iguchi et al. 2000). Moreover, uncertainties exist in the globally derived DSD parameter and reflectivity (Z)–rain rate (R) relationships. These most likely vary regionally, and a more accurate convective stratiform discrimination technique may be required (Masunaga et al. 2002). Furthermore, ground and sidelobe clutter decreases the reliability of the PR estimates at the edge of the swath (Durden et al. 2001). Moreover, the near-surface rain rates do not account for evaporation below the cloud base, although improvements to version 6 estimates may account for this. Another reason for differences between the instrument estimates is that they have significantly different resolutions.

Over land, biases in the GPROF algorithm are linked
to the scatter-based National Environmental Satellite, Data and Information Service (NESDIS) empirical algorithm. The NESDIS algorithm uses a single scatter index–rain rate relationship, and therefore does not take into account regional variations in cloud ice microphysics (McCollum and Ferraro 2003). A new version of the NESDIS algorithm incorporates a convective–stratiform discrimination technique in an attempt to overcome these microphysical biases, although preliminary results show regional biases remain (McCollum and Ferraro 2003). The 10° × 10° paired Student’s t tests showed that GPROF is significantly greater than PR in regions of strong convective uplift, such as tropical Africa, tropical South America, and the Maritime Continent. Percentage differences in the diurnal amplitudes also showed that GPROF exceeds PR by over 50% in these regions. GPROF overestimates PR because larger and greater quantities of ice particles aloft provide a larger scatter signature for the same rain rate. Previous work suggests that increases in cloud condensational nuclei (CCN) from desert and biomass-burning aerosols, as shown over equatorial Africa, increase the ice content of the clouds and produce an artificially high scatter index (McCollum et al. 2000). The greater concentrations of CCN produce smaller drops, which are quick to freeze and enhance the scatter signal. GPROF also exceeds PR in regions of lower precipitation efficiency (e.g., because of evaporation over northern Africa), although the failure of arid/semiarid surface screens may also contribute to this overestimation (McCollum et al. 2000). Conversely, for some regions of light or warm rainfall where scattering signatures are weak, GPROF is less than PR (e.g., parts of eastern and northern Africa, and the southeast United States). In some regions, surface screening removes light rainfall (e.g., northern Africa). Consequently, passive microwave estimates do not provide complete rainfall climatologies. Much work is required to improve surface screening techniques with regionally and temporally varying thresholds. Coastal artifacts also occur because of sharp gradients in emissivity and temperature, and the change in the retrieval algorithm at the land–ocean interface, and may result in errors for the GPROF algorithm.

The results of the comparison of diurnal harmonic phases and the regional analysis found that GPROF lags PR over parts of tropical South America. Lin et al. (2000) detected this phase difference over Amazonia and attributed it to sampling differences. However, because coincident observations are compared in this analysis, it seems algorithm biases are important. Over Amazonia, the frequency of precipitating systems with no ice scattering signature is greater compared to other continental locations (Petersen et al. 2002). Because the scatter-based NESDIS algorithm is not regionally calibrated to account for weaker ice scatter signatures, the differences are greater than for more continental systems with greater or larger ice particles.

Over the ocean, GPROF uses a physical inversion profile algorithm and so the physical mechanisms contributing to the algorithm biases are different to those over land. Over the ocean, GPROF significantly exceeds PR in regions of heavy convection. This agrees with previous work that has compared algorithm rain rates (Shin et al. 2001; Masunaga et al. 2002). Specifically, GPROF is known to overestimate stratiform rainfall areas in regions of heavy convection (Kummerow et al. 2001). Masunaga et al. (2002) show that in the Tropics, TMI detects larger precipitable water contents than PR, and that this difference is exaggerated further in the conversion to rain rates. Greatest discrepancies were found over the East Pacific region, where climatological rainfall comparisons attribute larger algorithm biases to higher cloud liquid water contents, lower ice contents, and higher stratiform rainfall when compared to the West Pacific (Janowiak et al. 1995; Berg et al. 2002). Moreover, stratiform rainfall increases further in the East Pacific during El Niño, which occurred during DJF 1998 (Schumacher and Houze 2003).

GPROF shows a seasonal bias in the southern and northern oceans, which may occur as a result of changes in the freezing-level altitude. Specifically, GPROF overestimates (underestimates) PR in the summer (winter) hemisphere. There is a lack of midlatitude simulations in the GPROF database, which means that lower TMI precipitable water contents and ice contents may not be accounted for (Masunaga et al. 2002). The finite nature of the profile database is a possible source of GPROF error because an appropriate profile may not exist. In the GPROF algorithm, it is necessary to improve cloud and radiative transfer model physics, such as to account for variability in melting layers and ice microphysics. This may also reduce the significant underestimation of PR by GPROF observed over the subtropical high pressure cells over the ocean.

Overall, the regional comparisons reveal that differences between the PR and PR–TMI detection of the diurnal rainfall cycle are smaller than the PR and GPROF differences. Differences in the phase of the diurnal harmonic are generally less than 1 h. This is expected because PR is an integral part of the PR–TMI algorithm (Haddad et al. 1997). The main exception occurs during winter in the southern oceans poleward of 30° and, to a lesser extent, in the northern oceans.
The PR–TMI underestimates PR because it may be unable to account for lower ice and precipitable water contents found in these winter systems (Masunaga et al. 2002). For intense convective regions, such as the Maritime Continent, absolute differences between the algorithms are maximized and PR–TMI significantly overestimates PR. These biases are consistent with previous work, which has examined climatological biases (Shin et al. 2001; Masunaga et al. 2002).

6. Conclusions

This study documents the mean 3-yr diurnal rainfall climatology from coincident TRMM microwave rainfall estimates (PR 2A25, GPROF 2A12, and PR–TMI 2B31). Algorithm comparisons of both the mean and the diurnal harmonic are conducted at the global (40°N–40°S) and regional level, and in a 10° × 10° grid to identify how diurnal rainfall patterns and algorithm biases impact diurnal estimates. The results show a stronger (subdued) diurnal cycle over land (ocean), which is consistent with previous work (Lim and Suh 2002; Yang and Slingo 2001; Sorooshian et al. 2002; Nesbitt and Zipser 2003). The analysis also shows that some significant algorithm discrepancies exist in the depiction of the diurnal cycle, which is consistent with known climatological biases (Adler et al. 2001; Shin et al. 2001; Berg et al. 2002). Furthermore, GPROF shows a temporal dependence on the magnitude of the biases, exceeding PR to a greater extent at the rainfall peak and directly after, when a stronger ice scattering signal occurs. Geographically, the largest differences occur over tropical South America and Africa (land) and the Pacific Ocean (ocean). Clearly, regional variations in the ice scatter signature need accounting for in the empirical algorithm to which GPROF is constrained over land.

As this study analyses 3-yr seasonal means to reduce noise, it encompasses the tail end of the intense 1997/98 El Niño and the persistent La Niña (1998/2000). How much the observed regional variations and product discrepancies are linked to ENSO forcing is therefore unknown and requires further investigation using a synergy of data sources. At the shorter seasonal time scale the diurnal cycle is also influenced by intraseasonal variations such as the Madden–Julian oscillation and synoptic forcing. These effects have not been removed from the analysis, and therefore additional errors and noise may be introduced into the analysis. Furthermore, there has been no attempt to validate the diurnal estimates with high temporal resolution “ground truth” data that are becoming increasingly available from TRMM ground validation sites, such as the Large-Scale Biosphere–Atmosphere (LBA) experiment in Amazonia. The results of this diurnal cycle study are also limited because of coarse spatial and temporal averaging, used in an attempt to avoid noise caused by the unequal diurnal sampling of the TRMM satellite.

Future work also aims to merge TRMM, Special Sensor Microwave Imager (SSMI), and Advanced Microwave Sounding Radiometer (AMSR) estimates to reduce sampling errors and provide greater confidence in the passive microwave diurnal cycle regional patterns. Such combinations are also important for simulating the diurnal sampling of the Global Precipitation Measurement (GPM), and identifying whether the proposed 3-hourly sampling rate adequately captures the diurnal rainfall cycle in the higher latitudes. In addition, the effect the TRMM orbital boost and the version 6 algorithm changes have had on the diurnal rainfall estimates should be investigated, so that the entire 7-yr dataset can be exploited. In the near future, spatial and temporal retrieval biases need accounting for, either in the physics of the algorithm or through the provision of accurate error estimates, to avoid erroneous climatic signals and the discrediting of satellite rainfall estimates. In summary, care needs to be taken when considering satellite algorithms for validation of global climate models, especially when using data from one satellite.

Acknowledgments. The authors thank the three anonymous reviewers for their detailed comments on this manuscript.

REFERENCES


Liberti, G., F. Chéruy, and M. Desbois, 2001: Land effect on the diurnal cycle of clouds over the TOGA COARE area, as observed from GMS IR data. *Mon. Wea. Rev.*, 129, 1500–1517.


Shimizu, S., R. Oki, and T. Igarashi, 2001: Ground validation of


