J1.2 OBSERVED REGIONAL AND TEMPORAL VARIABILITY OF RAINFALL OVER THE TROPICAL PACIFIC AND ATLANTIC OCEANS

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1. INTRODUCTION

Determining the diurnal cycle in rainfall is a fundamental step towards understanding the processes that control convection in the tropics. The dependence of the stage of convection on time of day provides clues as to what constitutes a favorable environment for convective development, and what processes are important at different points in the convective life-cycle. The seasonal and regional modulation of the diurnal cycle provide further clues regarding these processes.

Most studies of the diurnal cycle in tropical oceanic rainfall rely on indirect measurements of precipitation from satellites (e.g. Reed and Jaffe 1981; Albright et al. 1985; Janowiak et al. 1994; Chen and Houze 1997; Yang and Slingo 2001). Sui et al. (1997) use data from surface radars to investigate the diurnal cycle in rainfall in the western tropical Pacific, but this study is limited in both space and time, and also involves indirect measurements of rainfall. Gray and Jacobson (1977) use direct measurements of rainfall from island gauges for investigating the diurnal cycle in rainfall, but these data are subject to island heating effects and are also limited to the western Pacific. Other studies involving surface rainfall or rainfall proxy measurements include Janowiak et al. (1994), who also investigate the diurnal cycle in cloud classification from ship-board data and in rainfall from 5 rain gauges on buoys in the west Pacific warm pool region.

In this study the diurnal cycle in rainfall is investigated for the 1997-2001 time period using measurements from self-siphoning rain gauges on ATLAS buoys within the TAO/TRITON and PIRATA moored buoy arrays in the tropical Pacific and Atlantic Oceans. The TAO and PIRATA buoy arrays offer the first in situ direct measurements of tropical oceanic precipitation over an extended period of time and within critical rain areas throughout the tropical Pacific and Atlantic Ocean basins. Regional and seasonal differences in the diurnal cycles for seven different climatological regimes are discussed and compared with results from satellite and radar studies.

2. RAINFALL MEASUREMENTS

Fig. 1 shows the locations of the buoys within the TAO/TRITON and PIRATA arrays. The seven climatological regimes used in this study are indicated by colored squares. The regimes are defined as regions with similar seasonal cycles in rainfall. In general they include the major rainfall climate zones in the tropics: the western Pacific warm pool (NWPAC, EQWPAC) and SPCZ (SWPAC), the eastern Pacific ITCZ (NEPAC1, NEPAC2), and the Atlantic ITCZ (NWATL, EQWATL).

Standard ATLAS measurements include 3-m air temperature and relative humidity, 4-m wind speed and direction, and ocean temperatures at 11 depths from 1 - 500 m. Next Generation ATLAS buoys, deployed at selected locations since 1997, can additionally measure barometric pressure, shortwave and longwave radiation, precipitation, and salinity at 1-m depth.

The Next Generation ATLAS buoys measure precipitation using RM Young self-siphoning rain gauges mounted 3.5 m above the ocean surface. These measurements are available at 10-min time resolution, which have been smoothed to 1 hour for this study. A complete description of the rain gauges used on ATLAS buoys and their error characteristics is given in Serra et al. (2001). Briefly, this study finds that the random error in hourly rain rates is on the order of 0.15 mm h^{-1}. Therefore, instrument error is not considered a significant source of error in the diurnal cycle estimates for this study.

The largest systematic error for siphon gauges is due to wind effects, which may bias the rain

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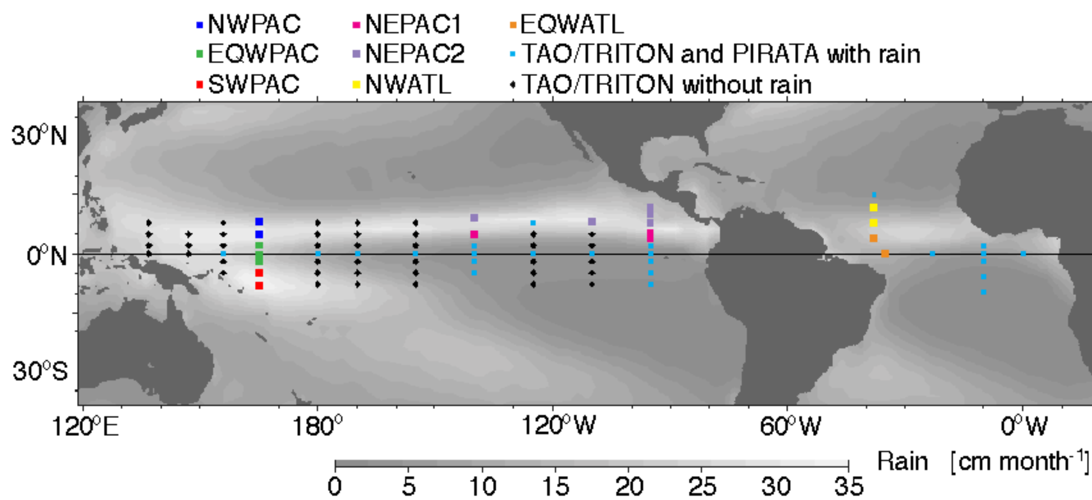

Fig. 1 - Map of the TAO/TRITON and PIRATA buoy arrays. The buoys used for this study are indicated, as are the climatological regimes used to categorize the diurnal cycles in rainfall.

rates low by about 10% - 50% for the wind speeds encountered at ATLAS buoys (Serra et al. 2001). A correction for this effect is provided in Serra et al. (2001) based on Koschmieder (1934). As this correction is not specific to the rain gauges on ATLAS buoys, we have analyzed both the uncorrected and corrected buoy data for this study. In general, we find that the wind speed correction has little affect on the magnitude of the diurnal cycle in rainfall because the phase of the diurnal cycle in wind speed is not necessarily correlated to the phase of the diurnal cycle in rain rate. Thus, there is no bias in the wind speed correction with time of day. We therefore present the results of the uncorrected rainfall data.

Another source of uncertainty in the diurnal cycle is the sampling error associated with measurements from a single point like a buoy. We have attempted to minimize this error in two ways; 1) we combine measurements from at least two buoys to increase the spatial degrees of freedom, and 2) we combine anywhere from 1 - 4 years of data, depending on availability, to increase degrees of freedom in the time domain.

3. CALCULATING THE DIURNAL CYCLE

In order to obtain the diurnal cycle of a buoy variable X, we first separate the data into winter and summer seasons which we define as Dec-May and Jun-Nov, respectively. The data within each season are smoothed with a 2-day Hanning filter, producing a 24-hour smoothed time series. The smoothed data are then subtracted from the original time series to obtain the variability X' about the 24-hour mean. The 24-hour anomalies from the same season for all years are binned according to local hour and combined with similarly treated data from other buoys within the same regime. The mean and standard error are then calculated for every 3-hour block, overlapping by one hour, to provide a diurnal cycle and standard error with 24 points.

Table 1 shows the percent time raining for the seven regimes and for each season, where the threshold for rain are data with $R>0.5$ mm h^{-1}. These percentages are based on over 700 days of data for each regime, with some regimes having over 1600 days of data.

The strong seasonal cycle in rainfall observed in the northeastern Pacific and Atlantic regimes indicates the annual latitudinal migration of the ITCZ. In early spring the ITCZ is close to the equator, producing a maximum in percent time raining for the NEPAC1 and EQWATL regimes in the Dec-May season (Table 1). By late boreal summer or early fall, the ITCZ has reached its maximum northward extension, producing a maximum in percent time raining for the NEPAC2 and NWATL regimes in the Jun-Nov season (Table 1). In contrast, little seasonal variation in rainfall is observed in the western Pacific warm pool (NWPAC, EQWPAC) or SPCZ (SWPAC) regimes.

These results are consistent with those of Mitchell and Wallace (1992), who find that the western Pacific rainfall variability has more of a semi-annual component, while the eastern Pacific and western Atlantic have more of an annual component. Thus, binning the western Pacific data into 6-month seasons results in nearly equal amounts of rainfall in each season

Table 1: Percent of days with rain

	NWPAC	EQWPAC	SWPAC	NEPAC1	NEPAC2	NWATL	EQWATL
Dec-May	11%	6%	10%	9%	2%	1%	11%
Jun-Nov	10%	5%	9%	3%	10%	9%	4%

4. RESULTS

4.1 Combined oceanic diurnal cycle in rainfall

The buoy rain gauge data indicate that the diurnal cycle in rain rate for the seven ocean regimes combined has a maximum in the early morning hours between about 0200 and 0800 LT (Fig. 2a). A secondary maximum in rain rate from 1200 - 1400 LT and two minima in the late morning and late afternoon are also observed. The rain rates used for this calculation include zero values and have not been corrected for wind induced undercatchment errors. Thus, 700 days or more of data are included in the means and standard errors shown. The magnitude and phase of the maximum varies significantly with both season and regime, producing significant standard errors in the early morning hours. The highly variable nature of the amplitude and phase of the diurnal cycle in rainfall over the ocean is also noted by Yang and Slingo (2001), whose study is based on 3-hourly, 0.5º brightness temperatures and corresponding rain rates.

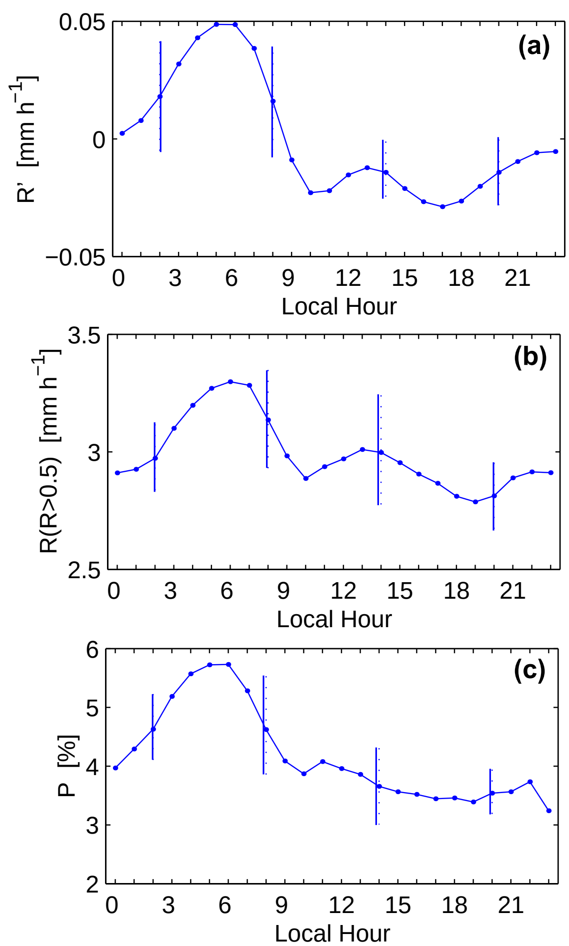

Fig. 2 - Diurnal cycle for all seven ocean regimes and all seasons combined of (a) rain rate R', (b) conditional rain rate R ($R>0.5$ mm h^{-1}), and (c) the percent of conditional rain rate observations P. The vertical lines indicate two standard errors of the mean, or the 95% confidence levels.

An early morning maximum in convection is found in previous studies oceanic rainfall and/or cloudiness (Gray and Jacobson 1977; Reed and Jaffe 1981; Albright et al. 1985; Janowiak et al. 1994; Chen and Houze 1997; Sui et al. 1997; Yang and Slingo 2001; Nesbitt and Zipser 2002). In addition, Nesbitt and Zipser (2002) note that a sharp decrease in convective and stratiform rainfall occurs near sunrise, between 0700 and 0900 LT, which they attribute to the radiative influence on nighttime convection over the oceans. Albright et al. (1985) also find a minimum in convection at sunrise, as well as in the late afternoon.

Secondary maxima in tropical oceanic convection around local noon are noted by Janowiak et al. (1994), Chen and Houze (1997), Sui et al. (1997), Albright et al. (1985), and Yang and Slingo (2001). These studies collectively find that the afternoon maximum is related to shallow, less organized convection, while the nighttime maximum tends to be the result of deep, organized systems.

In order to understand what governs the diurnal cycle in convection, we also calculate the diurnal cycle of conditional rain rate ($R>0.5$ mm h^{-1}) (Fig. 2b) and the percent of conditional rain observations (Fig. 2c). The former gives an indication as to the intensity of rain and the latter the frequency of events. Unlike in Table 1, the percentages in Fig. 2c are defined with respect to the total number of rainy days. Thus, the sum of the 24 hourly values shown is 100%. Fig. 2b shows that the diurnal cycle in rainfall intensity follows that of average rain rate, with an early morning maximum, minima at sunrise and sunset, and a

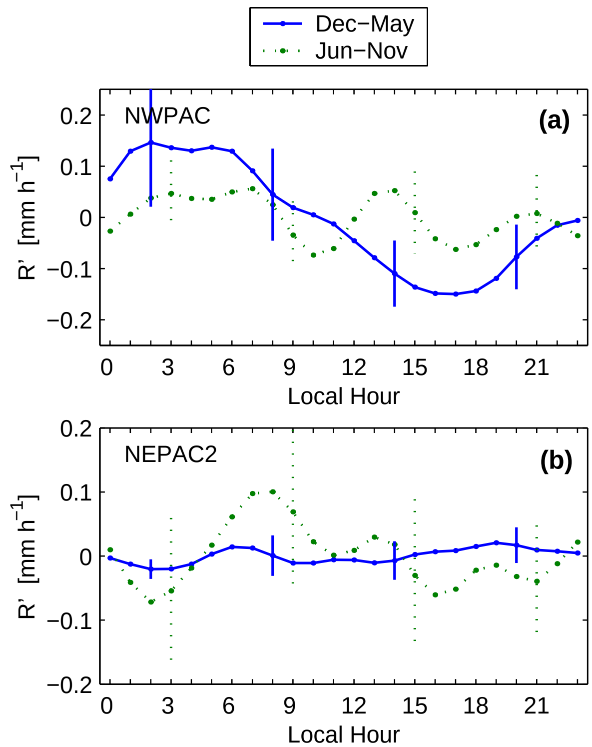

Fig. 3 - Diurnal cycle of rain rate R' for (a) the NWPAC, and (b) the NEPAC2. The vertical lines indicate two standard errors of the mean or the 95% confidence levels.

secondary maximum in the afternoon. From Fig. 2c it is seen that the diurnal cycle in the frequency of events has a single early morning maximum, tapering off to a minimum in the early evening.

These results suggest that both the intensity and frequency of rain are important for determining the phase and amplitude of the diurnal cycle over the ocean. Nesbitt and Zipser (2002) find that the diurnal cycle in rainfall over tropical oceans is determined primarily by the frequency of events, rather than their intensity, using microwave imager (TMI) and radar (PR) data from NASA's Tropical Rainfall Measuring Mission (TRMM) satellite for the 1998-2000 time period. This discrepancy may result from the different definition of intensity used by the two studies. Nesbitt and Zipser (2002) use information about the vertical structure of the precipitation area to define the convective intensity, whereas the present study uses the magnitude of conditional rain at the surface. Resolution of this issue is important for establishing confidence in both the satellite and buoy measurements of precipitation.

4.2 Regional and seasonal modulation of the diurnal cycle

Diurnal cycles in rain rate for the NWPAC and NEPAC2 are shown in Fig. 3 for both boreal summer and winter seasons. The NWPAC exhibits a semi-diurnal component in rain rate variability during the summer season (Jun-Nov), with maxima observed from 0200 - 0800 LT and from 1200 - 1500 LT. During the winter season (Dec-May), the diurnal component is better defined, with an early morning maximum from about 0000 - 0700 LT, followed by an afternoon minimum from 1300 - 2000 LT. In the NEPAC2 the diurnal cycle in rain rate is only present during the boreal summer season (which is the rainy season), with a maximum in rainfall from 0600 - 0900 LT. As in the NWPAC, there is a hint of a secondary afternoon peak from 1200 - 1400 LT, however this peak is less prominent than that in the NWPAC.

Figs. 4 and 5 are similar to Fig. 2b and c, but for the NWPAC and NEPAC2. In Fig. 2 the error estimates are based on the seven regimes. In Figs. 3 and 4 the errors are based on the standard deviation within each regime. As percent of conditional rain has only one value per hour per regime, no error bars are shown in Fig. 5. Because of the small number of rain events observed in the NEPAC2 winter season and the consequent lack of a diurnal cycle in rainfall (Fig. 3b), we do not show the diurnal cycle in conditional rain rate or percent conditional rain in FIgs. 4b and 5b.

In the NWPAC summer season the frequency of events is maximum around 0300 - 0500 LT, with small secondary maxima in the afternoon and evening (Fig. 5a), coincident with the peaks in rain rate for this season shown in Fig. 3a. The intensity of rainfall has the opposite variability, being maximum in the afternoon to late evening, with only a small secondary maximum in the early morning around 0700 LT (Fig. 4a). Thus, the observed semi-diurnal cycle in rainfall for the NWPAC summer results from the out of phase relationship between the intensity and frequency of rain events throughout the day. During the winter season, both the intensity and frequency of rain events are greatest at night, producing the well defined diurnal cycle shown in Fig. 3a for this season.

Nesbitt and Zipser (2002) find a maximum in mesoscale convective systems between 0600 and 1100 LT in the northwestern Pacific. These authors also note that over tropical oceans in general there is a maximum in less organized systems, both with and without ice scattering, from 0100 - 0500 LT, and a secondary peak in those with ice scattering in the afternoon. Similarly, Sui et al. (1997) find that IR-derived rain rate has a maximum between 0300 - 0400 LT, using satellite data collected over the west Pacific warm pool from Nov 1992 - Feb 1993. Their study also uses surface radar data from the same time period to determine the diurnal cycles of convective and stratiform rain. Their results indicate that stratiform and convective rain have a maximum at 0300 LT, and convective rain has an secondary maximum from 1200 - 1800 LT. Using the same IR data as Sui et al. (1997), Chen and Houze (1997) also find that the coldest IR temperatures (T < 208K) in the tropical western Pacific occur between midnight and 0600 LT, while warmer clouds (208K < T < 235K and 235K < T < 260K) peak between 1400 - 1800 LT.

The Sui et al. (1997) and Chen and Houze (1997) results should be compared to the buoy results for the NWPAC boreal winter season given the time period from which their data were collected. However, the afternoon maximum appears only in the summer season for the buoy data. This discrepancy may in part be due to the fact that 1992-1993 was an El Nino year, so that

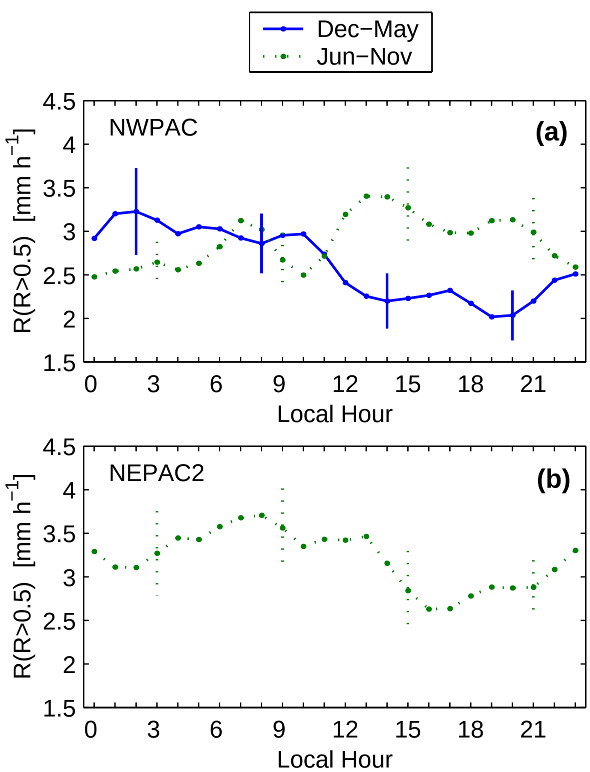

Fig. 4 - As in Fig. 3 but for conditional rain rate R (R>0.5 mm h^{-1}). There are too few rain events to permit an accurate calculation of the diurnal cycle in conditional rain for the NEPAC2 winter season.

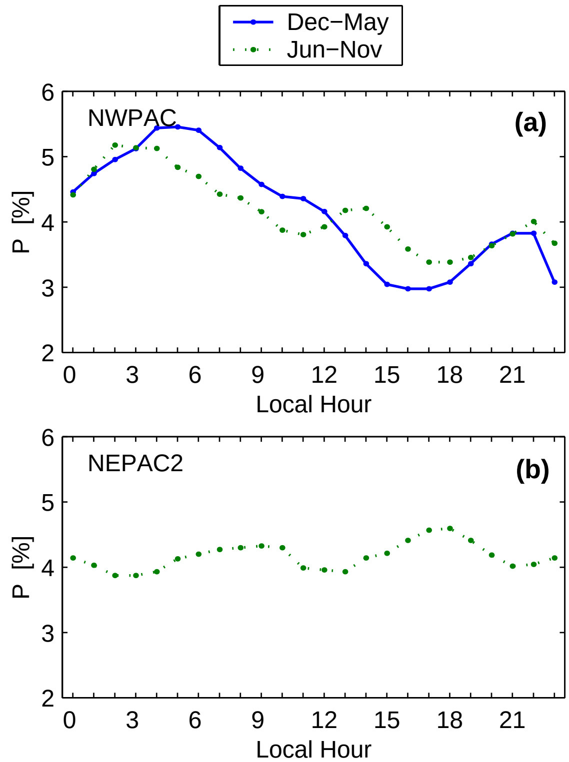

Fig. 5 - As in Fig. 3 but for percent time raining P. Also as for conditional rain, there are too few rain events to permit an accurate calculation of the diurnal cycle in the frequency of rain events for the NEPAC2 winter season.

the wintertime surface temperatures were warmer than usual and possibly supported afternoon convection normally only present during the summer season. The buoy data are biased towards La Nina conditions, which persisted from mid-1998 through 2000. The Nesbitt and Zipser (2002) results, which are from nearly the same time period as the buoy data, are not sorted by season, so it is unknown whether these data would resolve this discrepancy or not.

The boreal summer season in the NEPAC2, like that in the NWPAC, indicates an out of phase relationship between the intensity and frequency of rain events (Figs. 4b and 5b). In this case, however, the maximum intensity occurs between roughly 0400 - 1300 LT, while the frequency of events is greatest around 1800 LT. A small secondary maximum in the frequency of events is also observed in the morning.

In general, the amplitude of the diurnal cycle in the frequency of events is less than one percent for the summer season in the NEPAC2, in contrast to the NWPAC, where the amplitude for both seasons is greater than two percent. On the other hand, the amplitude of the intensity in the NEPAC2 summer season is similar in magnitude to the NWPAC, suggesting that the intensity of systems is more variable than their frequency in the NEPAC2 regime.

Yang and Slingo (2001) find that the amplitude of the diurnal cycle in brightness temperature is maximum in the June-August season in the northeast Pacific ITCZ. Their results also indicate that the minimum brightness temperatures occur in the early morning in this region, in general agreement with the buoy data in Fig. 3b. Nesbitt and Zipser (2002) find a maximum in mesoscale convective activity between 0300 - 0600 LT, somewhat earlier than the maximum observed with the buoy data, and find no diurnal variability in convective intensity for this region.

5. CONCLUSIONS

Results indicate that the seven tropical rain areas defined in this study have a rainfall maximum between midnight and roughly 0800 LT. An additional afternoon maximum is observed at several locations, the most prominent being the northwest Pacific warm pool, the South Pacific Convergence Zone (SPCZ), and the Atlantic ITCZ.

The morning maximum in rainfall is found to result from a maximum in both the frequency and intensity of rain events, while the afternoon maximum results mainly from a maximum in the intensity of systems. However, these results are not generally applicable to all regimes or seasons. For instance, in the northwest Pacific (NWPAC) during boreal summer, the diurnal cycle in rainfall intensity is maximum during the day, in contrast to the regime average, which has a maximum in intensity in the early morning. The northeast Pacific ITCZ (NEPAC2) summer (rainy) season also contrasts the regime average results. In this regime the amplitude of the diurnal cycle in intensity is comparable to that of the average, but the frequency of events has a much smaller diurnal amplitude, and a small maximum in the late afternoon to early evening, rather than in the early morning.

If we can understand the reasons for the diurnal cycles in both the intensity and frequency of rain events, we will have made progress in understanding the processes which control convection in the tropics. In addition, understanding the discrepancies between satellite and in situ measures of the diurnal cycle in these quantities will further our progress towards accurate measurements of tropical rainfall. This study highlights the level of agreement between the satellite and buoy observations of the diurnal cycle. Further study is needed to understand any discrepancies and their implications.

6. REFERENCES

Albright, M. D., E. E. Recker, and R. J. Reed, 1985: The diurnal variation of deep convection and inferred precipitation in the central tropical Pacific during January-February 1979. *Mon. Wea. Rev.*, **113**, 1663-1680.

Chen, S. S., and R. A. Houze, 1997: Diurnal variation and life-cycle of deep convective systems over the Pacific warm pool. *Quart. J. Roy. Meteor. Soc.*, **123**, 357-388.

Janowiak, J. E., P. A. Arkin, and M. Morrissey, 1994: An estimation of the diurnal cycle in oceanic tropical rainfall using satellite and in situ data. *Mon. Wea. Rev.*, **122**, 2296-2311.

Koschmieder, H., 1934: Methods and results of definite rain measurements. *Mon. Wea. Rev.*, **62**, 5-7.

Mitchell, T. P., and J. M. Wallace, 1992: The annual cycle in equatorial convection and sea surface temperature. *J. Climate*, **5**, 1140-1156.

Nesbitt, S. W., and E. J. Zipser, 2002: The diurnal cycle of rainfall and convective intensity according to three years of TRMM measurements. *J. Climate*, submitted.

Reed, R. J., and K. D. Jaffe, 1981: Diurnal variation of summer convection over West Africa and the tropical eastern Atlantic during 1974 and 1978. *Mon. Wea. Rev.*, **109**, 2527-2534.

Serra, Y. L., P. A'Hearn, P. Freitag, and M. J. McPhaden, 2001: ATLAS self-siphoning rain gauge error estimates. *J. Atmos. Oceanic Technol.*, **18**, 1980-2001.

Sui, C.-H., K.-M. Lau, Y. N. Takayabu, and D. A. Short, 1997: Diurnal variations in tropical oceanic cumulus convection during TOGA COARE. *J. Atmos. Sci.*, **54**, 639-655.

Yang, G-Y., and J. Slingo, 2001: The diurnal cycle in the Tropics. *Mon. Wea. Rev.*, **129**, 784-801.