Diurnal Variations in Tropical Oceanic Cumulus Convection during TOGA COARE

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ABSTRACT
Diurnal variations in atmospheric convection, dynamic/thermodynamic fields, and heat/moisture budgets over the equatorial Pacific warm pool region are analyzed based on data collected from different observation platforms during the Intensive Observation Period of the Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE). Results reveal that the diurnal variations in rainfall/convection over the TOGA COARE region can be classified into three distinct stages: warm morning cumulus, afternoon convective showers, and nocturnal convective systems. Afternoon rainfall comes mostly from convective cells, but the nocturnal rainfall is derived from deeper convective cells and large areas of stratiform clouds. Results further show that afternoon convective showers are more evident in the large-scale undisturbed periods when the diurnal SST cycle is strong, but the nocturnal convective systems and morning cumulus are more enhanced in the disturbed periods when more moisture is available. The primary cause of the nocturnal rainfall maximum is suggested to be associated with more (less) available precipitable water in the night (day) due to the diurnal radiative cooling/heating cycle and the resultant change in tropospheric relative humidity.

1. Introduction
Diurnal variations in tropical convection have been studied by many investigators. Gray and Jacobson (1977) conducted a comprehensive survey on the existence and cause of the diurnal cycle in deep cumulus convection and heavy rainfall over tropical ocean areas. Starting from the late 1970s, IR measurements from geostationary satellites have provided much-needed information for studying diurnal cycles in convection over different tropical areas. For example, McGarry and Reed (1978) and Reed and Jaffe (1981) analyzed diurnal variations in convection and precipitation over west Africa/tropical east Atlantic. They found that maximum convective activity in the eastern Atlantic occurred in the afternoon and suggested that continental influences may affect the diurnal cycle. Augustine (1984) found the averaged diurnal variations of satellite-infrared rainfall over large tropical areas in the central and eastern Pacific exhibited a maximum in midafternoon and a secondary peak near dawn. Albright et al. (1985) showed an afternoon maximum in the South Pacific convergence zone and morning maxima in the intertropical convergence zone (ITCZ) and the remaining central and eastern tropical Pacific regions. Based on a derived convection index from IR irradiance data, Murakami (1983) found a maximum enhancement of deep convection over the western Pacific around 0600–0900 LT. Mapes and Houze (1993) found that deep convection in organized cloud clusters over the western Pacific warm pool peaks before dawn and decreases through the morning and that the moderately cold cloud areas expand in the afternoon.

By using IR measurements from polar orbiting satellites or by combining measurements from several geostationary satellites, it is possible to study variations in convection and rainfall over the global Tropics. Short and Wallace (1980) studied morning to evening changes in cloudiness. Meisner and Arkin (1987) studied spatial and annual variations in the diurnal cycle in cold cloud and infrared precipitation in the Americas. Fu et al. (1990) used a combined visible–infrared method for detecting deep convective clouds and an adjusted IR method to study diurnal variation in the tropical Pacific. They found maximum deep convective clouds in the early morning and maximum mesoscale cirrus–anvil clouds.
in the late afternoon–early evening throughout the equatorial Pacific.

However, satellite IR sensors provide an indirect estimate of convection and rainfall because they only observe cloud-top parameters that are also affected by extensive stratiform–anvil clouds. On the other hand, microwave sensors measure the thermal emission from liquid water (at low frequencies) or the effects of scattering of upwelling radiation due to precipitation-size ice particles in the tops of convective systems (at higher frequencies), and hence provide a more direct measurement of rainfall than IR observations. Since all present spaceborne microwave radiometers are aboard polar-orbiting, sun-synchronous satellites, the temporal sampling is not as frequent as that available from geostationary IR measurements. Nevertheless, rainfall estimates derived from microwave data can be used as independent information to confirm the results based on IR measurements. Recently, Chang et al. (1995) estimated the annual mean rainfall based on the Special Sensor Microwave/Imager morning and evening passes and examined their differences. The morning rainfall estimates are larger than the evening estimates by about 20% over the oceanic region between 50°S and 50°N, with significant differences mainly located along the ITCZ region. The estimated 24-h harmonic shows a nocturnal or early morning maximum in 35%–40% of the oceanic regions. Janowiak et al. (1994) compared the diurnal variations in tropical cloudiness determined from IR brightness temperature with the estimated precipitation intensity differences between morning and evening observations from microwave satellite data. Both observations indicate maximum convective activity in the predawn hours over the tropical oceans, and an out-of-phase relationship between continental convection and adjacent oceanic convection.

It is generally accepted that the strongest convection over continents occurs in the late afternoon or early evening due to the dominant diurnal surface heating cycle, except in regions where land–sea contrast or orographic forcings are strong. Over the oceanic regions free from continental influence (Silva Dias et al. 1987), however, the coldest clouds and maximum rainfall are frequently observed in the early morning. This requires some mechanisms other than surface heating to maintain strong nocturnal convection over open oceans. The mechanisms that have been suggested include

1) a thermodynamic response of clouds to radiative heating such that the development of cloud (top, cover, and depth) tends to be reduced by solar heating and enhanced by IR cooling (e.g., Kraus 1963); and
2) a synoptic-scale dynamic response to the radiational differences between cloudy regions and clear regions such that low-level convergence into the cloudy region and upper-level rising motion is enhanced during the night and suppressed in the cloudy region during the day (e.g., Gray and Jacobson 1977).

However, more observational evidence is needed to examine the validity of the above mechanisms. For example, both mechanisms assume the existence of clouds, thus requiring the existence of a correlation between nocturnal convection and extensive cloudiness as supporting evidence. Furthermore, due to the presence of large-scale forcing and extensive clouds in disturbed conditions, the diurnal cycle in large-scale disturbed and undisturbed conditions should be analyzed separately to better understand its causes and to determine if there is a modulation of diurnal variations by the large-scale forcing.

Another issue emerges concerning the role of the boundary layer forcing arising from the diurnal variation of sea surface temperature. Such forcing is expected to be more pronounced under clear skies and weak wind conditions as revealed by recent TOGA COARE observations (Weller and Anderson 1996; Anderson et al. 1996; Sui et al. 1996). A natural question is whether afternoon convection exists over open oceans in response to the boundary layer forcing. An earlier study by Brier and Simpson (1969) suggested a possible forcing of afternoon convection by the semidiurnal tide. But Lindzen (1978) showed that the convergence field due to the tide could not directly account for the rainfall oscillation. Since the atmospheric tides are primarily forced by absorption of solar radiation due to ozone and water vapor, its effect on convection will not be studied in this work.

The purpose of the present paper is to obtain a complete picture of the diurnal variation and its modulation by large-scale environments by analyzing rainfall and cloud statistics, mass and heat budgets, and surface heat fluxes using TOGA COARE observations. Relevant mechanisms responsible for the observed variations are then discussed to address the above issues. Data and analysis methods are described in section 2. To assess the possible role of the convection–radiation interaction and boundary layer processes discussed above, diurnal variations are analyzed in disturbed periods and undisturbed periods separately. The classification of disturbed periods is discussed in section 3. Diurnal variations in both periods are analyzed based on composite results that are discussed in sections 4 and 5. In section 6, the time evolution of the diurnal cycle is further analyzed to identify the features of diurnal cycles obtained from the composite study. A summary and discussion are presented in section 7.

2. Data and methods

This study is based on observations taken during the COARE Intensive Observation Period (IOP) (1 November 1992–28 February 1993) within the domains shown in Fig. 1. The observations include GMS satellite measurements, rain maps from the Massachusetts Institute of Technology (MIT) and TOGA radars, upper-air soundings, and the special mooring data. Detailed de-
FIG. 1. Map of the Intensive Flux Array (IFA) and the Outer Soundings Array (OSA) defined for the COARE IOP. The IFA is the inner quadrangle in the figure, bounded by the two islands and two ships. The OSA is the outer hexagon bounded by the six islands.

Descriptions of these observations and analysis methods are discussed below. These multiplatform measurements provide an unprecedented opportunity to perform a comprehensive analysis of cloud and rainfall statistics, mass and heat budgets, and surface fluxes over the tropical ocean. Such a study of the diurnal variations in convective activity over a tropical oceanic domain would not be possible without COARE observations.

a. GMS IR

Infrared observations from the Geostationary Meteorological Satellite (about 5 km in spatial resolution) were compiled by the Japan Meteorological Agency to yield a temperature histogram along with the mean brightness temperature ($T_{bb}$) and the standard deviation ($\sigma_b$) associated with the spatial variance within each specified mesh. In this study, hourly mean $T_{bb}$ and $\sigma_b$ at 0.1° longitude–latitude resolution were used to analyze convective activity and to compute $T_{bb}$ histograms during the COARE IOP.

b. Radar

Rainfall statistics were derived from rain maps based on radar reflectivity data taken from the MIT 5-cm Doppler radar onboard the vessel R/V John V. Vickers and the TOGA 5-cm Doppler radar onboard the vessel R/V Xiangyanghong #5 for three 30-day periods during the COARE IOP. Reflectivity volume scan data were collected at 10 min intervals. Reflectivity values from the lowest tilts were used by the Tropical Rainfall Measuring Mission office at NASA/Goddard Space Flight Center to produce rain maps (Short et al. 1995, 1996). The relationship $Z = 120r^{1.43}$ (convective) and $Z = 323r^{1.43}$ (stratiform), from raindrop size distribution data observed on Kapingamarangi Atoll, was used to convert radar reflectivity, $Z$ (dBZ), to rainfall rate, $r$ (mm h$^{-1}$), on a 2-km grid. The gridded rain map is within a 5° × 5° latitude–longitude area centered on (2.5°S, 155.5°E), with each gridpoint value representing the instantaneous rain rate averaged over a 2-km × 2-km box. A total of 16 000–16 500 gridpoints are normally covered by each 150-km radar scan. Rain rate at each gridpoint is used to estimate a rain rate histogram $n(r_i)$, rain rate intensity distribution $r_i n(r_i)/N$, and area rain-rate distribution $A_0 r_i n(r_i)/N$, where $r_i$ is the rain rate defining the $i$th bin by $(r_i, r_{i+1})$, $N = \Sigma n(r_i)$ is the total number of counts in all bins, and $A_0$ is the fractional areal coverage for $r > 0$ within the radar scan area. Rain rates are binned logarithmically into 52 bins with the bin sizes chosen to conform to the digitization of rain rates in the dataset.

c. Upper-air soundings

Six-hourly upper-air soundings taken at four sites bounding the Intensive Flux Array (IFA), Kapingamarangi, Kavieng, Shiyan #3, and Kexue #1, plus the two ship sites, Xiangyanghong #5 and Moana Wave, were used in this study (Fig. 1). The first four sites were equipped with the Integrated Sounding System (see Parsons et al. 1994). The latter two sites were equipped with National Center for Atmospheric Research (NCAR) Cross-Chain Loran Atmospheric Sounding
Systems. The interpolated data processed at NCAR were further interpolated into 45 levels (1010, 1005, 1000, 995, 990, every 25 mb from 975 to 100, 80, 70, 60, 50 mb). A range-checking program was applied to the time series of $u$, $v$, $T$, and $q_r$ at each level of each site to detect extreme values (three standard deviation levels from the time mean) that were flagged as questionable. The questionable values were either left as undefined or replaced by interpolated values from the two neighboring values if neither of the two were questionable. The fields of $u$, $v$, $T$, and $q_r$ needed for evaluating the mass and heat budgets within the IFA were estimated by least square fitting of linear surfaces to the observed values of the variables at the six sites. The vertical $p$-velocity $\omega$ was obtained from the fitted $u$, $v$ fields by the kinematic method. The divergence field was adjusted to constrain $\omega$ at the surface and 100 mb to be zero.

The heat and moisture budgets were computed as follows:

\[
Q_1 = \frac{\partial s}{\partial t} + \mathbf{V} \cdot \nabla s + \omega \frac{\partial s}{\partial p},
\]

\[
-Q_2 = L \left( \frac{\partial q_r}{\partial t} + \mathbf{V} \cdot \nabla q_r + \omega \frac{\partial q_r}{\partial p} \right),
\]

where $s = c_p T + g \zeta$, and $Q_1$ and $Q_2$ are the apparent heat source and apparent moisture sink, respectively (Yanai et al. 1973).

d. Improved meteorological surface mooring (IMET)

The IMET mooring was deployed at the center of IFA (1.75ºS, 156ºE) with meteorological and oceanographic instruments. The measurements include wind velocity (vector averaged), relative humidity, air temperature, barometric pressure, SST, incoming shortwave radiation, incoming longwave radiation, and oceanographic measurements. The data were analyzed by R. Weller and S. Anderson at Woods Hole Oceanographic Institution. They have adjusted air temperature to remove the daytime solar heating, and calculated the relative wind speed by differencing the absolute vector wind from the shallowest current meter velocity at 5 m. A 4-day gap 9 December 1992 0033:45–13 December 1992 0511:45 UTC was filled with hourly data from the nearby Autonomous Temperature Life Acquisition System buoy (Pacific Marine Environmental Laboratory) that was regridded to the analyzed sample interval (7.5 min). The previous 4 days of incoming shortwave radiation and barometric pressure were duplicated and patched into the gap as a temporary solution for these missing variables.

3. Classification of disturbed periods

The diurnal variations are analyzed in two distinct large-scale environments, disturbed versus undisturbed.
February. The maintenance of the major westerly events may be associated with a quasi-stationary heat source to the east of the IFA near the date line as observed by Sui and Lau (1992) and Weickmann et al. (1990). It may also be related to the establishment of a large-scale east-west pressure dipole between the Maritime Continent and the equatorial central Pacific (Kiladis et al. 1994; Lau et al. 1996).

The most dominant signals in Fig. 2 are the synoptic-scale (2–3 day) oscillations in the convective phase of the MJOs. These oscillations are part of westward propagating 2-day disturbances that are associated with deep divergent circulations as shown in the time–height distribution of $\omega$ within the IFA (Fig. 3a). Such a hierarchy of high-frequency disturbances embedded within the MJOs was analyzed previously by Nakazawa (1988), Lau et al. (1991), and Sui and Lau (1992). A comprehensive analysis of the 2-day wave has been reported in Takayabu (1994). An analysis of the 2-day oscillations during TOGA COARE IOP is reported in a separate paper by Takayabu et al. (1996).

4. Diurnal variations in disturbed periods

In this section, the diurnal variations of various parameters in the large-scale disturbed periods are discussed based on composite results. Parameters considered include cloud and precipitation statistics, thermodynamic stability, divergent circulation, and heat and moisture budgets.

a. Convection

Figure 4 shows the diurnal variations of area rain rate ($R$) and raining area ($A_0$) averaged over the 150-km scan range of the MIT radar, and the areal mean brightness temperature ($T_{bn}$) and the standard deviation of brightness temperature ($\sigma_T$) within the IFA. The variation is dominated by heavier rain during the night from 2200 to 0600 LT, and lighter rain in the remaining part of the day. Other variables are highly correlated with the rain-rate cycle, that is, heavier rain rate is associated with larger $A_0$ and $\sigma_T$ and lower $T_{bn}$. A more careful examination shows a steady increase in $\sigma_T$ and a decrease in $T_{bn}$ after sunset. The minimum $T_{bn}$ is reached at 0100–0200 LT, and the heaviest $R$ and maximum $\sigma_T$ are further developed near 0300 and 0400 LT, respectively. Such a phase relationship reveals the evolution of a nocturnal convective system from maximum cloud development to heaviest rainfall, and to the dissipation of clouds.

In order to diagnose the structure of convection, we further divide the rain maps into convective and stratiform regions based on the surface rain rate following a method applied by Short et al. (1995) that is similar to Churchill and Houze (1984). The method compares
the rain rate at each grid element with the averaged rain
Fig. 3. Time–height distribution of the computed daily mean $\omega$ (a) and 6-hourly areal mean zonal mean wind (b) computed based on the sounding data from four ISS sites bounding the IFA plus the Xiangyanghong #5 and Moana Wave. Areas of negative $\omega$ and positive zonal wind are shaded. Units are mb h$^{-1}$ and m s$^{-1}$, respectively.

rate of its surrounding 24 grids. If the former is more than twice the latter, the central grid and its surrounding 8 grids are convective. In addition, all grids with rain rate higher than 20 mm h$^{-1}$ are also convective. The remaining rainy regions are identified as stratiform. The composite $R$ and $A_0$ separately for the convective and stratiform types are shown in Fig. 4c. The figure shows that the raining area ($A_0$) and rain rate ($R$) between midnight and 0600 LT are dominated by the stratiform component, confirming that nocturnal convection consists
of extensive stratiform clouds. The figure also shows that the maximum area rain rate of the convective type near 0300 LT makes a significant contribution to the nocturnal rain rate maximum. In addition to the nocturnal signal, the area rain rate of the convective type shows a secondary peak in the late afternoon. This together with the rapid drop of $T_{bb}$ around noon and a secondary peak in $\alpha_s$ around 1300 LT (Fig. 4b) indicates the development of afternoon convection.

To further examine the result described above, diurnal variations of rain rate histograms are analyzed. Because the rain rate histogram is dominated by a quasi-log normal distribution, the diurnal variation shows up more clearly in the anomaly distribution as shown in Fig. 5. The anomaly distribution shows that the evolution of nocturnal rainfall consists of a growing phase from 2200 to 0300 LT and a decaying phase from 0300 to 0800 LT. In the growing phase, a wide range of convection (rain rate $> 0.5$ mm h$^{-1}$) becomes enhanced with most occurrences within 0.5 to 5 mm h$^{-1}$, while in the decaying phase, a systematic shift toward light rain ($< 0.5$ mm h$^{-1}$) is evident. In the daytime, the anomaly distribution shows an overall suppressed distribution except an increase of light rain ($< 0.5$ mm h$^{-1}$) starting around noon till sunset, and a small increase of heavy rain ($> 10$ mm h$^{-1}$) in the afternoon. This indicates that, in the large-scale disturbed condition, total rainfall is suppressed in the daytime, but scattered warm cumuli and showers do develop in the afternoon.

The diurnal variation of the $T_{bb}$ histogram anomalies is shown in Fig. 6. From 2200 to 0400 LT, both high clouds ($T_{bb} < 220$ K) and midlevel clouds ($240$ K $< T_{bb} < 280$ K) are enhanced. The enhanced midlevel nocturnal clouds appear to be originated from the afternoon convection that starts from the rapid increase of cold clouds ($210$ K $< T_{bb} < 250$ K) in the early afternoon and continually evolves into more warmer clouds toward the evening. From early morning to noon, the midlevel clouds dissipate first, then the coldest clouds start to disappear while warm clouds become enhanced.

b. Thermodynamic and dynamic fields

The vertical profiles of time-mean dry static energy, $s$, and moist static energy, $h = s + Lq_v$, are computed based on corrected sounding data from the Shiyan #3, Xiangyanghong #5, and Moana Wave (Fig. 7a). Note that the minimum $h$ is located near 700 mb because of the vertical distribution of moisture, $(h-s)$, that is confined to the lower and middle troposphere. The anomalies from the time-mean profiles $(s' = c_pT', h' = s' + Lq_v')$ composited at four times of the day (0400, 1000, 1600, and 2200 LT) are shown in Fig. 7b. Composite temperature anomalies are generally negative at night and positive in the day, with the magnitude increasing upward. But the moisture anomalies $(h' - s')$ above 700 mb are generally positive at night and negative in the day. Below 700 mb, the moisture anomalies are negative at 0400 and 1000 LT, and positive at 1600 and 2200 LT. Near the surface, the moisture anomaly is negative at 2200 LT, in disagreement with the special buoy measurements shown in Fig. 8 that show the surface layer is wettest at 2200 LT. This reflects uncertainties in moisture measurements near the surface that need further investigation. The vertical distribution of $h$ in Fig. 7 and surface moisture in Fig. 8 together show that positive $h'$ in the lower troposphere and negative $h'$ in the upper troposphere at 2200 LT make it the most unstable for the development of deep convection. On the contrary, the vertical distribution of $h$ at 1000 LT is most stable.
Also shown in Fig. 7 is the horizontal divergence and \( \omega \). The time-mean profiles in Fig. 7a show a pronounced upward motion through most of the troposphere with the maximum near 300–400 mb, the associated convergence below the level of maximum upward motion, and divergence above. The profiles of divergence and \( \omega \) anomalies from the daily mean profiles are shown in Fig. 7c. First note that the amplitude of the diurnal variation in \( \omega \) shown in Fig. 7c is about half of the amplitude of the mean \( \omega \). The \( \omega \) field at 0400 LT shows an anomalous ascending motion in the layer between 500 and 200 mb and an anomalous descending motion below 500 mb, features indicative of organized mesoscale structures. This is consistent with the features of nocturnal convection discussed above. The anomalous \( \omega \) at 1000 LT consists of weak upward motion in the lower troposphere and deep descending motion above. At 1600 LT, the anomalous divergent circulation evolves
Fig. 7. (a) Vertical profiles of time-mean dry static energy \(s\), moist static energy \(h\), divergence, and \(\omega\); (b) vertical profiles of \(s\) and \(h\) anomalies from the time-mean profiles at 0400, 1000, 1600, and 2200 LT; (c) as in (b) except for divergence and \(\omega\) anomalies, for disturbed periods. Here, \(s\) and \(h\) are averaged over soundings from the Shiyan #3, Xiangyanghong #5, and Moana Wave. Units for \(s\) and \(h\) are J kg\(^{-1}\); for divergence, 10\(^{-6}\) s\(^{-1}\), and for \(\omega\), mb h\(^{-1}\).

into a deep layer of ascending motion through the troposphere. These appear to support the existence of enhanced warm clouds in the morning and convective clouds in the afternoon. The downward (upward) motion in the lower troposphere at 0400 LT (1000 LT) seems to suggest a link between the nocturnal subsidence and warm morning clouds (the localization of cold boundary layer outflows may initiate the development of shallow convective cells in the morning). But the relative deep ascending motion at 1600 LT and the relative weaker and shallower ascending motion at 1400 LT appear to be inconsistent with the area rain-rate distribution that shows a maximum rain rate near 0400 LT and weak rain rate at 1600 LT. But the diurnal rain-rate distribution is consistent with the net moisture convergence that is maximum at 0400 LT (0.51 mm h\(^{-1}\), see \(Q_2\) budget below). However, causes and effects of the imbalance between rainfall and moisture convergence are difficult to distinguish.

The surface parameters measured by the IMET buoy are analyzed and compared with the sounding data. The diurnal composite of sea surface temperature, surface air temperature, and water vapor mixing ratios are shown in Fig. 8. Both SST and surface air temperature are warmer in the afternoon, but the warmest surface air temperature occurs near noon, about 3 h before the warmest SST. This may arise due to a direct absorption of solar radiation by water vapor. The diurnal cycle of surface water vapor mixing ratio is similar to the variations of surface air temperature such that the highest value of surface moist static energy is found near noon-time. This maximum boundary layer forcing favors the development of afternoon convective clouds as discussed above. But this important feature cannot be resolved by the 6-hourly soundings sampled at 0400, 1000, 1600, and 2200 LT, even when identical measurements are made as those of the IMET (open circles in Fig. 8). Figures 7b and 8 further indicate a noticeable difference between the IMET and sounding measurements of moisture near the surface at 1600 LT other than the difference at 2200 LT discussed above. Because of possible errors known in the sounding measurements, the diurnal cycle obtained by the IMET measurement shown in Fig. 8 is regarded as more reliable.

c. Heat and moisture budgets

The vertical distribution of composite \(Q_1\) and \(Q_2\) at four times of the day is shown in Fig. 9. The source and sink terms that contribute to \(Q_1\) and \(Q_2\) include

\[
Q_1 = Q_{C_n} - \frac{\partial}{\partial p} \omega s' + Q_{\omega n} \quad \text{and} \quad (3)
\]

\[
Q_2 = L_e C_n + L_s \frac{\partial}{\partial p} \omega q', \quad (4)
\]

where \(Q_{C_n} = L_e (c - e) + L_s (d - s) + L_f (f - m)\) is the...
net rate of latent heat release through condensation, deposition, and freezing, among the three phases; \( Q_R \) is the radiative heating rate; \( C_n = (c - e + d - s) \) is the net condensation and deposition; and the other two terms are eddy heat flux convergence and moisture flux convergence by cumulus convection. Because \( L_c(f - m) \) is about an order of magnitude smaller than the other two terms in \( Q_{cn} \), and \( L_c \sim L_v \), therefore, \( Q_{cn} \approx L_c C_n \) is a good approximation.

In the middle and lower troposphere, \( Q_1 \) profiles are characterized by a common heating maximum (positive \( Q_1 \)) and \( Q_2 \) profiles generally show two drying peaks (positive \( Q_2 \)). At 0400 LT, the maximum values of \( Q_1 \) and \( Q_2 \) are centered around 500–400 mb and are quite close to each other, indicating a dominant heating and drying effect of \( L_c C_n \) by stratiform clouds. A similar feature is also evident in the \( Q_1 \) and \( Q_2 \) profiles at 2200 LT, except that the magnitude of \( Q_1 \) and \( Q_2 \) is significantly weaker and the maximum values of \( Q_1 \) and \( Q_2 \) are centered around 500 mb. The \( Q_1 \) and \( Q_2 \) profiles at 1600 LT are similar to those at 2200 LT, but the maximum drying is slightly shifted downward below 500 mb and the maximum heating is distributed in a slightly deeper layer above 500 mb. Such a separation of the maximum \( Q_2 \) values from the maximum \( Q_1 \) values shows some contributions from vertical eddy transports due to cumulus convection as suggested by the following relationship:

\[
Q_2 - Q_1 = \frac{\partial}{\partial p} \omega h - Q_R. \tag{5}
\]

This is a typical feature of the heat and moisture budgets in the convective regime (e.g., Yanai et al. 1973). This feature is also evident at 1000 LT, except that the maximum values of \( Q_2 \) are located at a lower level (800 mb), indicating contributions by shallow cumulus. Note that a drying peak exists in the lower troposphere at all four times of the day. Its location varies from around 800 mb at 0400 and 1000 LT to around 700 mb at 1600 and 2200 LT. Below 800 mb, the moistening (negative \( Q_2 \)) at 1600 and 2200 LT is due to the eddy flux convergence, indicating enhanced surface evaporation.

In the upper troposphere, \( Q_R \) is the dominant term in the \( Q_2 \) budget.\(^1\) The heating at 1000 LT and cooling (negative \( Q_2 \)) at 2200 LT above 250 mb in Fig. 9 are consistent with the diurnal cycle of \( Q_R \). But heating at 0400 LT and cooling at 1600 LT need further explanations. It may be contributed by the eddy heat flux convergence, and the cloud–radiation interaction that produces cooling near cloud top and warming below. It also suggests a possible error introduced by the estimate of \( \omega \) in the \( Q_2 \)-budget computation based on the 12-h center-differencing scheme. To examine this possibility, the second estimates of \( Q_1 \) profiles at 0400 and 1600 LT are shown in Fig. 9 (thin solid curves), based on an alternative scheme of \( \omega \) by least-square fitting the composite \( s \) fields to a sinusoidal function. There are significant differences between the two estimates of \( Q_1 \) in the upper troposphere. Contrary to the first estimates, the second estimates show cooling at 0400 LT and heating at 1600 LT and are thus more consistent with the diurnal radiative heating cycle. This indicates that the diurnal temperature change \( \omega / \delta t \) is the dominant term in the upper-tropospheric heat budget, and it cannot be estimated accurately by 6-hourly sounding data.

5. Diurnal variations in undisturbed periods

The same composite fields as those discussed in the previous section but for undisturbed periods are discussed in this section. For comparison purposes, some basic time-mean quantities for the two periods are summarized in Table 1. SST in the undisturbed periods is slightly warmer than SST in disturbed periods, by 0.4°C, and the surface air in the undisturbed periods is 0.7°C warmer and 0.2 g kg\(^{-1}\) less humid compared to the

\(^1\) The conventional Vaisala radiation corrections were not applied to sounding data measured by the Integrated Sounding System. Thus the observed diurnal change in temperature and \( Q_1 \) should be viewed with caution.

<table>
<thead>
<tr>
<th>Fields</th>
<th>Fractional</th>
<th>SST (°C)</th>
<th>Ts (°C)</th>
<th>q (g kg(^{-1}))</th>
<th>R (mm h(^{-1}))</th>
<th>A(_v) (%)</th>
<th>T(_{so}) (K)</th>
<th>(\sigma) (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Disturbed</td>
<td>63/120</td>
<td>29.2</td>
<td>27.6</td>
<td>19.4 (82%)</td>
<td>0.35</td>
<td>21</td>
<td>259.0</td>
<td>21.5</td>
</tr>
<tr>
<td>Undisturbed</td>
<td>42/120</td>
<td>29.6</td>
<td>28.3</td>
<td>19.2 (78%)</td>
<td>0.18</td>
<td>4</td>
<td>284.5</td>
<td>135</td>
</tr>
</tbody>
</table>

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\( C \)
surface air in the disturbed periods. But the difference in $R$, $A_0$, $T_{wb}$, and $\sigma_b$ between the two conditions is much larger: $R$, $A_0$ and $\sigma_b$ in disturbed periods are about two times, five times, and two times higher than those in the undisturbed periods, respectively. The time-mean histograms of rain rate and $T_{wb}$ shown in Fig. 10 illustrate that the cloud clusters in the disturbed periods vary greatly in height and depth, producing a wide spectrum of rain rate, while convection in the undisturbed periods is dominated by warm cumulus. In the disturbed periods, the large population of light raining events (peak around 0.3 mm h$^{-1}$) in the rain-rate histogram and the lack of corresponding warm clouds in the $T_{wb}$ histogram suggest the existence of extensive stratiform clouds that may cover many warm cumuli below it. These basic differences arise due to the large-scale forcing as represented by the time-mean fields of horizontal divergence and $\omega$. A comparison of these fields for disturbed periods (Fig. 7a) and undisturbed periods (Fig. 11) shows that the former is representative of an ascending branch of the large-scale circulation and the latter a descending part of the large-scale circulation. The corresponding thermodynamic state as represented by the vertical profiles of $s$ and $h$ shown in Figs. 7b and 11 further reveal that the mean state in undisturbed periods is slightly warmer near the tropopause and notably drier above the boundary layer, leading to a more unstable stratification than that in disturbed periods.

Despite the differences in the time-mean quantities for the disturbed and undisturbed periods, the diurnal variation of convection in the undisturbed condition is also dominated by nocturnal convection as exemplified by the composite area rain rate shown in Fig. 12. Our analysis indicates that there are many similarities in diurnal variations between the disturbed and undisturbed conditions. However, a further divide of the diurnal rain rate distribution within the two groups of the undisturbed periods (see discussion in section 3), one with scattered rain cells (34 days), and the other with weakly organized rain patterns (8 days), shows that the second group contributes most of the nocturnal rainfall (Fig. 12). On the other hand, the diurnal variation in the first group is dominated by afternoon rainfall (Fig. 12). The anomalous rain rate and $T_{wb}$ histograms shown in Figs. 13 and 14, respectively, indicate a consistent variation of convection from a sporadic development of light-raining cells around 1000 LT to more frequent occurrence of cumulus in the afternoon. The most rapid development is found around 1300–1400 LT when SST
is reaching a maximum (Fig. 8). Note that, like the variation in the disturbed periods shown in Fig. 6, nocturnal cold clouds also exist in the undisturbed periods shown in Fig. 14. This suggests a common mechanism for producing nocturnal clouds.

The diurnal variation of temperature in the undisturbed periods is similar to the variation in the disturbed periods. However, the variations of moisture in the two periods are different. As shown by the IMET measurements (Fig. 8), in the undisturbed periods, the boundary layer moisture does not follow the diurnal variation of temperature; instead it closely reflects the diurnal variation of convection, that is, the driest boundary layer occurs around 1000 and 1600 LT when convection is most active. Variations in lower-tropospheric moisture observed by soundings (not shown) are consistent with IMET measurements in the undisturbed periods.

The composite $\omega$ fields for the two groups of undisturbed periods show a strong contrast (not shown). Group I shows that descending motion persists through the day, except at 1600 LT when a weak upward motion is developed up to 500 mb. This is consistent with the
development of scattered afternoon cumulus discussed above. However, variation in group 2 shows that upward motion persists through the day.

6. Modulation of diurnal cycles by the large-scale environment

The results presented in the previous sections are based on composite analyses with a priori classification into disturbed/undisturbed conditions. There is some ambiguity about the interpretation of the composite diurnal variation in the time evolution sense. There is also a need to seek a more direct relationship between the diurnal variations and the relevant large-scale conditions to help clarify the causes of the diurnal variations. Due to the existence of multiscale, propagating disturbances in the large-scale environment, a definite relationship between the diurnal variation and the large-scale disturbances is difficult to obtain from observations covering a limited spatial domain. In this section, an EOF analysis is adopted to seek the dominant time-evolving diurnal modes and their statistical phase relationship with indices of large-scale processes within dry and wet phases of MJOs.

The EOF analysis is applied to the hourly rain-rate histogram, distribution of rain-rate intensity, and distribution of area rain rate. The data are divided into a sequence of daily matrices such that the hourly data in each day are treated as a two-dimensional field of 24 h × 52 rainbands. The EOF analysis produces a sequence of eigenfunctions and a principal component in time corresponding to each eigenfunction.

The first eigenfunction (EF1) of rain rate histogram $n(r)$ (Fig. 15) shows three peaks: a nocturnal peak (0.5–2.0 mm h$^{-1}$ around 0200–0300 LT), a morning peak (0.2–0.4 mm h$^{-1}$ around 1000 LT), and an afternoon peak (0.4–0.6 mm h$^{-1}$ around 1600 LT). The analysis shows that a significant portion of the diurnal variation in rain rate frequency distribution (53% of the total variance) occurs in the lower end of the rain rate spectrum. Further EOF analyses of the area rain rate distribution, $A_r(r)/N$, and the rain-rate intensity distribution, $r_r(r)/N$, provide different aspects of the diurnal rain-rate variability, because the former is weighted by rain rate and the latter is further normalized by total number of rain counts, $N$.

The first eigenfunction of area rain-rate distribution (Fig. 16) shows a dominant nocturnal peak at 2–15 mm h$^{-1}$ between 2300–0600 LT. It accounts for 40% of the variance of diurnal area rain rate. On the other hand, the first eigenfunction of rain-rate intensity distribution (Fig. 17) shows a dominant afternoon peak between noon and 1800 LT. It accounts for 31% of the variance of diurnal rain rate intensity. The second and third eigenfunctions in both analyses form a degenerate pair as a random mixture of the true eigenfunctions (North et al. 1982). Other higher modes show more complicated patterns that cannot be identified as diurnal modes, therefore they are not discussed here.

To examine the relationship between the nocturnal mode and the large-scale cloudiness, the principal components corresponding to the EF1 of area rain-rate distribution (PC1a) are plotted as a function of the daily mean $T_{bs}$ (Fig. 18a) averaged over the IFA. Here the
daily mean $T_{bb}$ is used as an index of cloudiness to represent the combined cloud cover and cloud amount information. The plot is made separately for the dry phase (12 November–7 December, 30 December–16 January) and the wet phase (7–30 December, 16 January–1 February) characterizing the inactive and active phase of Madden–Julian oscillations (see discussion in section 3, and Sui et al. 1996). Figure 18a shows that values of PC1a are almost all negative when $T_{bb}$ is higher than 280 K (mostly in the dry phase). For $T_{bb}$ lower than 280 K, PC1a does not show a clear relationship with $T_{bb}$ but tends to be more positive in the wet phase than in the dry phase. There are two important implications here. First, the lack of correlation between ex-
tensive clouds (low $T_{bb}$) and nocturnal rainfall/convective does not support cloud-radiative forcing as a dominant mechanism for the formation of nocturnal convection. Second, nocturnal convection appears to be modulated by low-frequency disturbances. The evolution of PC1a is found to be closely related to the evolution of the daily-mean vertically integrated water vapor (or total precipitable water) shown in Fig. 19. This indicates a clear tendency for nocturnal convection to be suppressed in the large-scale undisturbed periods, and to occur in the large-scale disturbed periods when there is a more abundant moisture supply by the scale circulation. A more detailed study about the interaction between diurnal cycle and synoptic-scale disturbances in the TOGA COARE is reported by Chen and Houze (1996).

On the other hand, afternoon convection is found to be evident in the undisturbed periods when surface forcing is strong. Thus the principal components corresponding to the EF1 of the rain-rate intensity distribution (PC1r) as a function of diurnal amplitude of SST ($\Delta SST$) are plotted in Fig. 18b. There is a general positive correlation between the two quantities; that is, PC1r is positive (negative) when $\Delta SST$ is large (small), except near the higher end of the diurnal amplitude of SST ($\Delta SST > 1.2^\circ C$). The strongest $\Delta SST$s occur near the end of the two dry phases (2–5 December, 12–15 January) when tropospheric downward motion is strong (Fig. 3a). This indicates that both surface heating and the large-scale circulation control afternoon convection.

7. Summary and discussion

The diurnal variations in convective activities, stability distributions, divergence and $\omega$ fields, and heat and moisture budgets are analyzed using observations collected during the Intensive Observation Period of TOGA COARE. Our major goal is to identify statistically significant modes of diurnal variations that have large day-by-day variations because of modulations by large-scale disturbances. The diurnal composite and EOF analyses reveal a consistent picture of the diurnal cycle over the tropical open ocean that is characterized by warm cumulus in the morning, afternoon convective showers, and organized nocturnal convection. Nocturnal convection and morning cumulus tend to be dominant in the large-scale disturbed periods, and afternoon convection is evident in the large-scale undisturbed periods. Cloud and rain-rate statistical distributions, thermodynamic stability, mass, and heat and moisture budgets indicate that cumulus convection in the morning and afternoon consists of shallower convective elements, and the nocturnal convective system consists of deeper convective cells and larger areas of stratiform clouds.

The evolution and strength of afternoon convection is found to be correlated to the diurnal SST cycle. Diurnal variation in the ocean surface layer is particularly strong in the undisturbed periods due to weak winds.
and strong solar heating at the surface that maintain a shallow and warm mixed layer in the daytime (Weller and Anderson 1996; Anderson et al. 1996; Sui et al. 1996). However, the strongest diurnal SST cycles appear to occur in the presence of strong large-scale subsidence that acts to suppress the development of afternoon convection.

The lack of a clear correlation between the nocturnal rain-rate maximum and cloudiness, and the appearance of cold nocturnal cloud in both disturbed and undisturbed periods, indicate a possible existence of a more fundamental mechanism responsible for the development of nocturnal convection, related to the availability of moisture. We suggest that the nocturnal rainfall maximum is related to the destabilization by radiative heating/cooling cycle and the resultant change of available precipitable water (APW) that can be expressed as RH ($W^*$), that is, the difference between the vertically integrated saturation water vapor amount $W^*$ and a reference value of $W^*$, with respect to a reference columnal relative humidity. Assuming the diurnal variation of RH is small (the diurnal composite values of RH in TOGA COARE are near 69%), the diurnal variations of rain rate due to radiative destabilization can be approximated by $-\Delta RH \frac{\partial W^*}{\partial t}$. The estimates of $W^*$ and $-\Delta W^* / \partial t$ based on the composite TOGA COARE temperature are listed in Table 2. Considering $-\Delta W^* / \partial t$ as a theoretical limit of the nocturnal rain-rate distribution, the nocturnal rain-rate maximum appears to be related to the diurnal temperature cycle. This simple thermodynamic relationship is suggested to be the fundamental destabilizing mechanism responsible for the nocturnal rainfall maximum. Further modeling studies based on cumulus ensemble models are needed to verify this explanation.

The cloud-radiation interaction mechanisms may play some role in destabilizing clouds in the disturbed periods that contribute to the enhanced nocturnal rainfall. But the more humid troposphere, due to the large-scale moisture supply by low-frequency disturbances, appears to play a more important role in providing a favorable condition for realizing the potential of increased APW at night, enhancing the development of nocturnal convective systems.

The prevalence of warm cumulus in the morning may be initiated by the nocturnal subsidence driven by radiative cooling. Warm cumulus interacts with large-scale dynamics by providing moisture and heating in the lower troposphere, tending to excite low-frequency oscillations. Warm clouds also cause a significant cooling in cloud–SST interactions. The frequency of occurrence of warm cumulus over the warm ocean and its role in the climate system need to be more thoroughly investigated.

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**Table 2.**

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**REFERENCES**


Fu, R., A. D. Del Genio, and W. B. R состоять в том, что они поддерживают развитие конвекции в области субседентации.

Однако присутствие теплых облаков в течение дня может быть инициировано ночным субседентационным процессом, управляемым радиационным охлаждением. Теплые облака взаимодействуют с крупномасштабными процессами, обеспечивая подпитку влагой и теплоносителем в нижнем тропосфере, что приводит к возбуждению низкочастотных колебаний. Теплые облака также могут вызывать значительное охлаждение в облачно–ССТ взаимодействиях. Частота появления теплых облаков над теплым океаном и ее роль в климатической системе требуют более тщательного исследования.

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**Таблица 2.**

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<td>$-\Delta W^* / \partial t$ (мм·ч$^{-1}$)</td>
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**Список литературы**


Ганюй, Й., П. А. Аркн и М. Моррисей, 1994: Исследование...


