The evolution of the tropical western Pacific atmosphere–ocean system following the arrival of a dry intrusion

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SUMMARY

Recent studies using TOGA COARE data have found that extremely dry air from middle-latitude waves frequently intrudes into the equatorial troposphere over the western Pacific. Using sounding data taken during the COARE, the magnitude of the advection of water vapour for one event is calculated, and it is estimated that the time for the atmosphere to recover to moist conditions was $\sim$10–20 days. From the magnitude of the drying and from the frequency of these events, it is proposed that dry intrusions must be a major contributor to the tropospheric moisture budget over the region during the COARE, making it difficult for the atmosphere to reach a radiative–convective equilibrium. Intrusions, instead, can help to recharge the tropical atmosphere by decreasing convective activity and, thus, driving the atmosphere toward unusually large values of convective available potential energy. A variety of atmospheric and oceanic measurements are also used to study the recovery process in detail. A conceptual model is proposed based on this work and previous investigations. As in past studies, the recovery of the atmosphere to moist conditions is accomplished through entrainment from convective clouds that began to form soon after the arrival of the dry air mass and slowly deepen in height as the recovery progresses. Previous investigators concluded that the entrainment of dry air into convective cells is generally the factor that tends to suppress convective activity and limits the height of any convection that does develop under these adverse conditions. The idea that entrainment limits convective activity is consistent with the commonly held perception that the western Pacific is a region where there is little inhibition to deep convection and, when inhibition does occur, it can be removed by surface fluxes within hours. In contrast, it is found that convective inhibition can be large enough to suppress convection following dry intrusions, and that the diurnal variation in rainfall is due partly to modulations in convective inhibition. The modulations in convective inhibition are, in turn, caused by diurnal variations in the vertical profiles of radiation, in surface fluxes, and perhaps in large-scale subsidence, leading to a minimum in convective inhibition during the late afternoon. In contrast, studies of this type of convection have generally emphasized diurnal variations in the surface fluxes, and often ignored convective inhibition and diurnal variations in atmospheric radiative heating.

KEYWORDS: Diurnal cycle Dry intrusions Tropical western Pacific

1. INTRODUCTION

The field phase of the Tropical Ocean and Global Atmosphere (TOGA) Coupled Ocean–Atmosphere Response Experiment (COARE) was carried out in order to improve our understanding of atmospheric and oceanic processes in the tropical western Pacific (e.g. Webster and Lukas 1992). One of the primary concerns of the TOGA COARE was to explore the interactions between the large-scale atmospheric flow, convection, and the warm sea surface temperatures that characterize this region. The importance of understanding cloud and radiative processes over the tropical western Pacific is also noted in the scientific goals and field studies of the Atmospheric Radiation Measurement (ARM) program (e.g. Stokes and Schwartz 1994).

In this study, we employed observations taken during the TOGA COARE to investigate the behaviour of the ocean–atmosphere system following the advection of extremely dry air from middle latitudes into the equatorial troposphere in mid-November. These dry intrusions (or dry tongues) have recently received a great deal of interest among TOGA COARE researchers (e.g. Parsons et al. 1994; Sheu and Liu 1995;
Numaguti et al. 1995; Yoneyama and Fujitani 1995; Mapes and Zuidema 1996; Johnson et al. 1996; DeMott and Rutledge 1998; Yoneyama and Parsons 1999). These events are so named since the dry air originates aloft at higher latitudes and subsides into the tropics in long filaments, several 100 km in width. Recently, Yoneyama and Parsons (1999) showed that these events are associated with the remnant circulations of middle-latitude baroclinic waves.

The radiative implications of extremely dry air on tropospheric stability was discussed by Mapes and Zuidema (1996) as a mechanism to explain the strong thermal inversions observed at the base of these dry layers. These warm, dry layers have been found to be well correlated with a suppression of deep convection, as clearly demonstrated by Brown and Zhang (1997) and DeMott and Rutledge (1998). The suppression of deep convection can be so pronounced that Brown and Zhang (1997) regarded these periods as tropical droughts. While deep convection is suppressed, past studies also show relatively shallow (e.g. 3–8 km in height) convection develops within days after the arrival of this dry air. This shallow convection slowly erodes any dry layers through detrainment (Mapes and Zuidema 1996; Johnson et al. 1996; Brown and Zhang 1997; DeMott and Rutledge 1998). These relatively shallow clouds tend to have a distinct diurnal cycle, with a precipitation maximum in the early evening that has been proposed to be due to diurnal variations in surface fluxes in response to solar heating (e.g. Chen and Houze 1997).

The general suppression of deep convection by the intrusions could be due to either the increased thermal stability at the base of the dry layer or the negative impact of entrainment of dry air (Mapes and Zuidema 1996; DeMott and Rutledge 1998). The Brown and Zhang (1997) and Wei et al. (1998) efforts clearly stress the importance of entrainment. Perhaps the focus on entrainment rather than thermal stability is partly due to the commonly held conceptual model of the western Pacific as a region where there is little inhibition to deep convection, and when inhibition does occur it can be removed by surface fluxes within hours (e.g. Raymond 1995).

We found, in contrast to past studies, that significant convection inhibition exists in these dry environments, that this inhibition also acts to suppress convection, and that the diurnal variations in convective inhibition may partly be responsible for the diurnal cycle of convective rainfall. In particular, we found a minimum of inhibition present during the afternoon due to a combination of diurnal changes in the vertical profile of radiation resulting from short-wave absorption of solar energy and in the surface fluxes. Morning subsidence may also impact the diurnal cycle of convective inhibition.

We also calculated the magnitude of the dry advection from TOGA COARE sounding data and estimated that the time-scale of the recovery to moist conditions should range between 10 and 20 days. Since the magnitude of the lateral advection of dry air in these events is so large and intrusions have been found to occur frequently (every 8–16 days during COARE), lateral advection of dry air must be an important term in the budget of atmospheric water vapour over this region. Recently, Zhang and Chou (1999) noted the importance of lateral drying in reducing water vapour over this region. The frequency and magnitude of these extreme lateral drying events, together with the long adjustment times found in recent cumulus modelling studies (e.g. Tompkins and Craig 1998), suggest that it is difficult for the atmosphere to approach a convective–radiative equilibrium in the presence of these intrusions. Instead the net effect of these tropical drought periods is to increase the convective instability in the tropical atmosphere, which helps to recharge the tropical atmosphere to support subsequent periods of active deep convection. The increase in convective instability results from increases in boundary-layer moisture from warming sea surface temperatures. In contrast, Emanuel and Bister
(1997) hypothesized that a dry middle troposphere will create increased convective instability through enhanced radiational cooling.

These ideas and other new findings about the recovery process are consolidated with results from previous studies to form a conceptual model of the impact of dry intrusions on the equatorial ocean–atmosphere system. We believe that consolidation is necessary, since many past studies have concentrated on different aspects of these events (e.g. origin of the dry air, evolution of the cloud fields, and implications for vertical profiles of radiation).

2. OBSERVING NETWORK

The TOGA COARE field phase included a four-month Intensive Observing Period (IOP) extending from 1 November 1992 to 28 February 1993. This study used special observations taken during the IOP, including oceanographic, atmospheric, and air–sea interface measurements. The measurements of primary interest to this study were taken in the vicinity of the inner Intensive Flux Array (IFA) in Fig. 1. The outer edges of the IFA were defined by two ships (Research Vessel (R/V) Kexue 1 and R/V Shiyan 3), the island of Kavieng and the Kapingamarangi Atoll. These four measurement sites included Integrated Sounding Systems (ISSs) (Parsons et al. 1994), each consisting of a 915 MHz wind profiler, a Radio Acoustic Sounding System (RASS), an omega-based NAVAID sounding system, and an enhanced surface measurement system. The surface observing system for each ISS measured horizontal wind speed and direction, air temperature, relative humidity, pressure and various components of the surface
radiation budget. The rawinsonde measurements were used to characterize the large-scale conditions in the vicinity of the IFA in this study, while the surface, RASS and wind profiling measurements were used to monitor the conditions within the atmospheric boundary layer. In addition, the wind profilers also provided information on the nature of precipitation systems over the tropics through providing observations on radar reflectivity and vertical motions (e.g. Gage et al. 1994).

Within the IFA additional integrated measurements were also taken aboard the R/V Moana Wave (Fairall et al. 1997). The measurements included profiling and sounding measurements similar to the ISS, but with significant enhancements that included observations of the surface fluxes of sensible and latent heat by eddy correlation measurements, eddy dissipation techniques and bulk methods (Fairall et al. 1996). Additional flux and rawinsonde measurements within the IFA were also taken aboard the R/V Hakuho-Maru located within the IFA (Suzuki et al. 1995). Some use was also made of oceanic measurements taken by the Atlas buoy and the R/Vs Moana Wave, Hakuho-Maru, and Wecoma. The flux and oceanic measurements allowed us to investigate the atmosphere–ocean interactions that took place in association with the dry intrusions. Finally, we also used other rawinsonde sites within the TOGA COARE sounding array described by Lin and Johnson (1996), and the analysis provided by the European Centre for Medium-Range Weather Forecasts. The convective activity was observed through a Doppler radar located on the R/V Vickers (e.g. DeMott and Rutledge 1998).

3. OVERVIEW OF THE EVENT

Our period of interest coincides with a well-defined drying event in mid-November 1992. The large-scale circulations over the TOGA COARE IFA at this time are documented in previous studies (Kiladis et al. 1994; Gutzler et al. 1994; McBride et al. 1995; Lin and Johnson 1996). According to the Lin and Johnson (1996) analysis, this period coincided primarily with a period of light subsidence and a transition in the zonal winds from westerly to easterly. Specific aspects of this intrusion event have been treated in a number of studies (Parsons et al. 1994; Numaguti et al. 1995; Mapes and Zuidema 1996; Brown and Zhang 1997; DeMott and Rutledge 1998; Yoneyama and Parsons 1999). The origin of the dry air over the IFA in this case was traced back to the middle troposphere over southern Australia by Yoneyama and Parsons (1999) with the dry air arriving over the IFA on ~12 November.

A time–height cross-section of the horizontal advection of water vapour mixing ratio was constructed for the time period of 11 to 22 November from an objective analysis centred on the R/V Moana Wave. This analysis used a Barnes scheme with a preliminary correction for the humidity biases obtained from techniques described by Cole and Miller (1999). Without this correction, the biases, well known to many COARE investigators, affected the magnitude and sometimes even the sign of the advection. The analysis was constructed at vertical intervals of 50 hPa and was designed to capture horizontal gradients observed between the sounding sites at the edges of and within the IFA (i.e. ~200 km) as shown in Fig. 1. On most days a sufficient number of soundings were available over the IFA to produce an analysis approximately every 6 h. The available 6 h estimates were subsequently averaged to produce the daily means shown in Fig. 2. Several points are readily evident in this cross-section. First, the dry air arrives suddenly over a large vertical extent of the troposphere, including the boundary layer, with very sharp horizontal gradients at the leading edge of the dry air mass. While the daily averages of advection at the leading edge are quite large with values
Figure 2. A time–height cross-section of the horizontal advection of water vapour (g kg\(^{-1}\) day\(^{-1}\)) with the convention imposed that the advection of dry air is shown as positive. The advection was calculated from an objective analysis centred on the R/V Moana Wave and plotted as daily averages for the period from 11 to 22 November. The large negative values that appear just following the arrival of the dry air mass are shaded for emphasis.

of over 10 g kg\(^{-1}\) day\(^{-1}\), the 6 h calculations are even larger with values in excess of 15 g kg\(^{-1}\) day\(^{-1}\) for most levels at and below 650 hPa.

The behaviour of the dry intrusion is impulsive, as there is a general absence of strong advection on days following the arrival of the dry air mass. The exceptions to this statement are the relatively strong dry advection on 18–19 November, on 21 and 22 November when the fields are becoming noisy due to increasing convective activity, and the moistening evident at and below 750 hPa just after the arrival of the dry air mass. For this event, the impulsive nature of the intrusion and the relative lack of moist advection afterwards suggests that the general recovery over the IFA is primarily due to local processes, such as surface fluxes, and processes that subsequently mix air vertically, such as turbulence, convection or large-scale vertical motions.

The soundings taken in the vicinity of the IFA are generally consistent with calculated humidity advection, lending some credence to the advection estimates. For example, the deep dry advection is consistent with very dry conditions over the depth of the troposphere at Misima to the south of the IFA (Fig. 3(a)). A sounding taken from the R/V Kexue 1 when the dry air first appeared over the IFA (Fig. 3(b)) reveals extremely dry air in the layer from ~900 to 600 hPa with a stable layer near the base of this dry layer capping a well mixed boundary layer, which is consistent with the magnitude and height of the extreme drying estimated from the objective analysis. Finally, a sounding taken at the R/V Moana Wave near the time of the secondary dry advection event aloft shows very dry conditions developed between ~400 and 300 hPa (Fig. 3(c)). This secondary
drying event aloft is also consistent with the presence of very dry air aloft to the south of the IFA at Misima at this time (not shown). The finding that dry-air advection events can occur at a variety of heights contrasts with the general conceptual model proposed by DeMott and Rutledge (1998) that proposes upper-level drying arises from a decrease in convective activity in response to dry advection taking place in the lower layers.

The magnitude of dry advection at the impulsive arrival of the air mass can be scaled by the average surface latent-heat flux and surface rainfall amounts to estimate the time-scale for the atmosphere to recover to moist conditions. Realizing that such a calculation ignores subsidence and the advection of clouds and that the uncertainty in the sounding humidity measurements is relatively large, this calculation should be treated with some caution. However, since there is generally weak, dry advection after the arrival of the dry intrusion, there is the possibility that this scaling may actually underestimate the time-scale of the recovery. Using the advection values for the lowest 9 km where the drying is most pronounced (Fig. 2) and using an average latent-heat flux for this event of ~100 W m\(^{-2}\) (the fluxes will be shown later), the time-scale of the recovery is estimated to be ~ 9–10 days. Recalling that some of the moistening by the fluxes will fall out as rain instead of moistening the atmosphere, it is evident that this 9–10 days must be an underestimate. We will later show the rainfall to be of the order of 2 mm day\(^{-1}\), which relative to the input of the fluxes is a precipitation
efficiency of ~57% and slows the recovery time to nearly 20 days. Brown and Zhang (1997) estimated recovery times of ~10 days. Since the frequency of occurrence for dry intrusions during the TOGA COARE is ~8–15 days (e.g. Yoneyama and Parsons 1999), we conclude that the tropical atmosphere over this region during the COARE spends a great proportion of time adjusting to humidity fluctuations arising from higher latitudes. Zhang and Chou (1999) also noted the importance of the lateral advection of dry air during the COARE period. Given this finding it is difficult to image that the atmosphere over the warm pool could be in a state of radiative–convective equilibrium, especially in light of the slow adjustment time-scales found recently in cumulus ensemble modelling (Tompkins and Craig 1998).

(a) Atmospheric surface and boundary-layer changes

A time series of water vapour calculated from rawinsonde data taken aboard the R/V Moana Wave for the time period of 11 to 23 November is shown in Fig. 4. These values have been averaged over various depths and then normalized by the mean value in that layer. The time series within the boundary layer (~1000–950 hPa) shows a sharp drop in humidity near the leading edge of the dry intrusion (Fig. 4(a)), in agreement with the calculated moisture advection (Fig. 2). The non-normalized value of this decrease is ~3.3 g kg⁻¹. The boundary layer recovers from this decrease to values similar to those observed before the intrusion in less than 2 days. In this layer there is the tendency for the highest values of the moisture to be evident toward the end of this period. The transient low values are associated with the local effects of convection.

In agreement with the sounding data, the time series of surface temperature (T) and specific humidity (q) for the R/V Hakuho-Maru (Fig. 5) indicates a near 4 g kg⁻¹ drop in specific humidity on 13 November local time (LST). The specific humidity takes several days to recover from this drop, regaining the higher values late in the day on the 16th. The general pronounced fall in specific humidity at the surface is also evident at other sites within the IFA. For example, a similar sharp decrease occurs at the R/V Moana Wave late on 12 November (Fig. 6(a)). The earlier arrival of the dry air and the earlier recovery is consistent with the southerly location of the site and the trajectory of the dry air mass given by Yoneyama and Parsons (1999). The time-scale of the recovery process at the R/V Moana Wave, however, was slightly more rapid than at the R/V Hakuho-Maru. Also, evident in the R/V Moana Wave time series is a 2 degC fall in temperature on 12 November due to a convective activity in the IFA region present before the arrival of the dry air.

The corresponding surface wind speed measured on the R/V Moana Wave (Fig. 6(b)) indicates a tendency for higher wind speeds to occur earlier in the period. A comparison of the surface wind speed with the time series of specific humidity (Fig. 6(a)) clearly shows a correspondence between the period of high winds and the drop in humidity at the surface. Surface fluxes were obtained according to the techniques of Fairall et al. (1996). The surface flux of latent heat from the R/V Moana Wave (Fig. 6(c)) shows that the early stages of the drying events in the boundary layer correspond to a pronounced peak in the latent-heat flux from the ocean surface, with values of over 200 W m⁻². This magnitude is substantial for the region and is more typical of strong, transient convective outflows (e.g. Parsons et al. (1994) and others) or highly distributed environments such as westerly wind bursts. An elevated, but less strong, peak in the surface flux of latent heat was also observed at the R/V Hakuho-Maru with boundary-layer drying noted by Numaguti et al. (1995). The sensible-heat flux is rather small through the period (Fig. 6(c)). At both sites we believe that the large surface fluxes result
in the relatively rapid recovery (~2 days) of the boundary layer following dry advection. DeMott and Rutledge (1998) noted relatively rapid boundary-layer recoveries following dry intrusions with non-light wind speeds.

Changes were also noted in boundary-layer height as shown in the vertical profiles of the water vapour mixing ratio just following the arrival of the dry air mass at 0500 LST 13 November and several days later at 2200 LST 17 November (Fig. 7). On 13 November there is a deep, well-defined mixed layer extending to nearly 1 km, associated with the arrival of the dry air mass. The depth of the dry mixed layer greatly exceeds the ~500 m typical depth of boundary layers over the tropical ocean (Nicholls et al. 1982). The
vertical gradient of the water vapour mixing ratio is also quite large in this event, with changes of $\sim 13 \text{ g kg}^{-1}$ in only 150 m evident from the mixed layer to the dry air mass. Significant evolution of the vertical profile of water vapour mixing ratio occurs over the next several days as evident by the profile on 17 November. During this time, the boundary layer has moistened in the lowest 600 m by $\sim 2 \text{ g kg}^{-1}$ without significant moistening in the layer from 800 to 1000 m leading to the appearance of a new, shallower boundary layer growing within the older, deeper and drier boundary layer. During this time period, moistening has taken place above 1100 m as values of $\sim 12 \text{ g kg}^{-1}$ are evident where the mixing ratio was $\sim 2$ to $6 \text{ g kg}^{-1}$. The moistening rate for this layer is on the order of $1.5 \text{ g kg}^{-1}\text{day}^{-1}$.

This analysis indicates that the arrival of the dry air corresponds to the entire boundary layer becoming significantly drier and deeper, with increases in wind speeds and highly elevated surface fluxes of latent heat and an extremely sharp drop in humidity at the surface. The very large magnitude of the latent-heat flux and the very deep boundary layer are perhaps initially surprising. Numaguti et al. (1995) proposed that the dry air present just above the boundary layer could be entrained into the boundary layer to account for the observation of pronounced surface drying. From our analysis thus far, we can state that the dry air also has a strong advective component in the boundary layer (Fig. 2). However, a number of factors suggest the entrainment proposed by Numaguti et al. (1995) must also be taking place, bringing dry air with higher momentum into the boundary layer. For example, the large flux of moisture from the surface and the rapid moistening of the air mass points to the need for the dry air mass within the boundary layer to be replenished by some process at the leading edge of the dry intrusion. Further evidence for the entrainment hypothesis is the strong vertical gradient of water vapour above the boundary layer, coupled with a general consensus from past studies (e.g. Numaguti et al. 1995; Yoneyama and Fujitani 1995; Mapes and
Figure 6. Surface measurements from the R/V *Moana Wave* for the period 12–22 November local time: (a) specific humidity and air temperature, (b) surface wind speed, and (c) latent- and sensible-heat flux.

Zuidema 1996; Yoneyama and Parsons 1997) that subsidence is associated with dry intrusions. In addition, the Lin and Johnson (1996) large-scale analysis also indicates weak subsidence. Using the previously mentioned vertical gradient in mixing ratio of 13 g kg\(^{-1}\) within a vertical distance of only 150 m and an entrainment velocity of only 2 cm s\(^{-1}\) would produce a drying rate of ~1.5 g kg\(^{-1}\) day\(^{-1}\). This drying helps in maintaining the dry conditions at the leading edge of the intrusion against moistening from strong surface fluxes of latent heat as the intrusion moves over the warm ocean waters. A boundary-layer balance that includes significant entrainment of dry air from aloft is different from the perhaps more typical tropical conditions, where the balance is between the destabilization of fluxes from the surface and the stabilizing effect of
convective downdraughts (Raymond 1995). Also, when estimating the recovery time following dry intrusions (e.g. DeMott and Rutledge 1998), it is necessary to consider the wind speed above the boundary layer and how it could impact the boundary-layer winds through entrainment.

(b) Changes in the free troposphere

In the lower free troposphere (900 to 600 hPa), it is evident that the arrival of the dry air is quite sudden, with the atmosphere on average drying out by over 5 g kg\(^{-1}\) in a period of 12 h on 12 November (Fig. 4(b)). The decrease would have been larger, but the air mass had already been modified somewhat by the time it reached this vessel. The recovery of the humidity in this layer is somewhat slower than in the boundary layer, with the atmosphere recovering a substantial portion of this drying in the first four days at a rate in excess of 1 g kg\(^{-1}\) day\(^{-1}\). After this initial recovery period, there are substantial variations in water vapour, but the moisture within this layer approaches, but never recovers to, the values observed ahead of the intrusion of dry air. The upper troposphere in the layer defined between 500 and 250 hPa (Fig. 4(a)) behaves differently, with substantial drying also taking place on 12 November, but with still slower moistening occurring until almost the 19th. After this period, drying again occurs with the water vapour finally returning on 22 November to values consistent with those measured before the intrusion. Our analysis of the different layers is similar to the results obtained by Brown and Zhang (1997). From our layer analysis, we conclude that their estimate of a recovery time of \(~10\) days is reasonable, with the caveat that it may be longer given that the 900 to 600 hPa layer never recovers to pre-intrusion levels and that we did not address humidity changes above 250 hPa.

As mentioned in the introduction, previous studies show that dry intrusions are correlated with a suppression of deep convection and the subsequent recovery of moisture
aloft due to relatively shallow convection. Hence, our discussion of the recovery process will be relatively brief and limited to a few key points. A time series of precipitation was estimated by applying a $Z-R$ relationship to the Doppler radar from the R/V *Vickers* (see Rickenbach and Rutledge (1997) for details) (Fig. 8). In this time series we concentrated on the period from 12 to 19 November to examine the recovery process. The significant precipitation that was evident on the 12th was due to a large convective line (not shown) that existed before the arrival of the dry air and rapidly decayed as it intercepted these unfavourable conditions. The lack of precipitation on the 13th is evidence that no significant amount of rain fell until the boundary layer recovered. Radar data from the R/V *Vickers* (not shown) does indicate that convection began from the south to the north after the boundary layer recovered. The time series shows that precipitation then fell intermittently over the region. Partitioning the rainfall into convective and stratiform components (see DeMott and Rutledge (1998) for details) indicates that the rainfall was generally convective in structure.

During this time there is a tendency for the convection to be weakly organized into areas of active convection rather than be uniformly distributed over the IFA. Thus, we would not term the convection as truly random. An example of this mesoscale organization is presented in the high-resolution satellite image taken at 1330 LST 16 November (Fig. 9). The organization of clouds in this satellite image over the central and southern portions of the IFA has some similarity to open convective cells, with a sharp transition to more linear features in the north. This image also illustrates the general lack of large organized systems with extensive anvils.

The depth of the convection can also be obtained from the wind profilers associated with the Integrated Sounding System (ISS) (Gage et al. 1994), since these clear-air radars are also sensitive to precipitation and some clouds. Precipitating cloud features, in particular, are easily detected in the vertical motions and the radar reflectivities. Time-height cross-sections of radar reflectivity are shown from the R/V *Kexue* 1 for 13 and
15 November (Fig. 10). These data are the ‘raw’ radar reflectivities that have not yet been processed into 30-minute averages. From these data and the time series of vertical velocity (not shown), we hypothesize that there are two types of precipitation systems present during the recovery. Early during the recovery period, the clouds are shallow with the most common echo top ranging between 1.5 and 3.5 km in height as evidenced by the reflectivity plot for 13 November (Fig. 10(a)). In this figure the reflectivity peaks generally extend from ~1.2 to ~3 km with the disruptions in the pattern of reflectivity associated with these peaks extending somewhat further upward. Often these clouds are not associated with significant rainfall, as evidenced by a lack of strong average downward motion in the profiler data that would arise from falling hydrometeors. From a comparison with the sounding data shown earlier, we can conclude that these clouds, which may be non-precipitating, are extending into the base of the dry layer and are likely partly responsible for the recovery in the lower troposphere through detrainment of moisture early in the period. Since these clouds seem to develop shortly after the recovery of the boundary layer, this hypothesis is consistent with the relatively rapid onset of the recovery process in the lower troposphere (Fig. 4(b)) and consistent with earlier studies that show the dry layer moistens from the lower levels upward.

On 15 November the profiler data show heavier rainfall (i.e. stronger reflectivity and strong downward vertical motions) that extends in the vertical through the dry layer. Heights for these events typically extend between 3 and 7 km, with echo tops
for one system reaching $\sim10$ to $12$ km (Fig. 10(b)). These wider variations in the height of convection indicate that, while there is a general deepening of the height of the convection with time, the increase is not a simple linear progression. The deeper convection obviously indicates that some clouds can overcome the negative effects of entrainment and penetrate through the dry layers, although perhaps the environment has been locally moistened by earlier convective events. It is possible that these deeper
precipitating clouds will also detrain moisture, when rising through these stable layers in the lower troposphere, and aid the recovery of the extremely dry lower troposphere.

The presence of the shallow clouds is similar to large areas of the tropics under subsidence. The profiler estimates of convective structure later on the 15th are generally consistent with the analysis of radar data from the R/V Vickers by DeMott and Rutledge (1998). Their study indicates that this dry-intrusion period contained a relatively high percentage of precipitating convection with echo tops mainly in the middle and lower troposphere (i.e. 3–7 km in height). Unfortunately, we feel that the partitioning of the moisture detrainment between the very shallow and perhaps non-precipitating convection, clouds that extend to between 3–7 km, and these less frequent deep systems cannot be accomplished with these data, and is thus best left to cumulus ensemble modelling. However, we do note that since the upper troposphere moistens slowly (see Fig. 4(a)), we propose that there is simply not enough deep convective activity to moisten and maintain moist conditions over the dry upper levels. This finding is consistent with the DeMott and Rutledge (1998) study.

4. SLOW CHANGES IN ATMOSPHERIC STABILITY AND THE UPPER OCEAN

Thus far we have looked at the general characteristics of the dry intrusion and how the atmosphere recovers toward moister conditions generally thought of as typical of this equatorial region. In this section we will examine changes that occur in the upper ocean, and variations in atmospheric stability that arise as a result of the dry intrusion and the subsequent recovery process.

(a) Oceanic changes

A general characterization of the conditions of the upper ocean can be seen in the CTD (Conductivity, Temperature and Depth) vertical profiles obtained from the R/V Wecoma (Figs. 11(a) to (c)). The general tendencies shown in the R/V Wecoma data were also represented in the CTD data from the other vessels. The time–depth cross-section of density from the R/V Wecoma for 11 to 22 November (Fig. 11(a)) shows a general tendency for a reduction in the depth of the ocean mixed layer from ~70 m on the 12th to less than 40 m at the end of the period. This reduction is typical of the warm pool under light winds and low precipitation. Superimposed on this general trend of a decreasing depth of the mixed layer is a local increase in the depth late in the day on 12 November following a convective event and near the time of the arrival of the dry intrusion. This time series also suggests that the upper ocean is becoming slightly denser.

The time–depth cross-section of temperature indicates that the upper ocean is warming by less than 1 degC in the upper 25–40 m of the ocean (Fig. 11(c)). Below that depth the ocean is actually cooling during this time, so that the upper ocean is also becoming more stratified in temperature with depth. Also, as was noted in the density data, the mixed layer is becoming thinner. The general warming in the upper layers suggests that any increase in density must be due to increases in salinity. This point is confirmed by the salinity data (Fig. 11(b)) that show the upper ocean mixed layer becoming slightly more saline from 11 to 17 November, which is typical of conditions where evaporation dominates over precipitation. The tendency for the mixed layer to become shallower is also clearly evident in the time–depth cross-section of salinity.

A time series showing daily averages of the various components of the surface energy budget, including the net flux of energy into the ocean, the incoming solar short-wave radiation, the net long-wave cooling, and the latent- and sensible-heat flux,
Figure 11. CTD (Conductivity, Temperature and Depth) vertical profiles obtained from the R/V *Wecoma* from 1100 LST 11 November to 1100 LST 21 November 1992 with time marks every 24 h. The depth shown is from the surface to 100 m. (a) Specific density (g kg$^{-1}$). [Note: 21.4 in the Figure means 1.0214 g kg$^{-1}$], (b) salinity (practical salinity units), and (c) temperature (°C).
Figure 12. Time series of daily averages of the components of the surface energy budget, including the short-wave solar incoming radiation ($Q_s$), net long-wave radiation cooling ($Q_l$), the latent ($Q_e$) and sensible ($Q_h$) heat flux and the net change in the radiation budget ($Q_n$). The net long-wave cooling was calculated from the measured downward directed long-wave energy and the loss from the ocean surface estimated from the sea surface temperature. Cooling of the ocean surface due to rain and lateral heat flux due to advection is not shown. An albedo of 0.0 was used for the short-wave energy. The sea surface temperatures (SSTs) are also shown (solid line).

is presented in Fig. 12. These daily averages show that the net effect of these terms following the arrival of the dry air mass is an energy flux into the ocean, assuming an albedo of 0.0, of the order of 80–110 W m$^{-2}$ for most of the period. In contrast, before the arrival of the dry air, cloud cover from convective activity in the region on the 12th greatly reduced the incoming short-wave energy and a net heat loss was observed. A relatively large surface flux of latent heat also contributed to this cooling. There was also a smaller net heat flux into the ocean on the 13th just following the arrival of the dry air mass due to high surface fluxes and a slightly larger long-wave loss despite the relative high values of short-wave flux. The CTD data shown earlier seem to reflect these processes through producing a cooler and deeper mixed layer on these days. With this exception, one can conclude that, under this regime, large values of the incoming short-wave solar energy are responsible for a net flux of energy into the ocean following the arrival of the dry air mass. The observed increase in the sea surface temperature (SST) of 0.6 degC from the 13th to the 19th is also consistent with the general trend in the surface energy balance. The 19th corresponds to a more disturbed period of convection.

The net heating of the upper ocean takes place despite a strong decrease in the long-wave downwelling radiation emitted by the atmosphere to the surface. This point is not readily evident in Fig. 12 where only the net difference between the long-wave loss by the ocean surface and long-wave downwelling from the atmosphere is observed. When
just the long-wave downwelling energy is shown (Fig. 13), it is evident that the dry period corresponds to the dry atmosphere emitting less infrared radiation to the surface. According to Long (1995) the variation in this term is very small. From Fig. 13 one can see that the changes associated with the dry intrusions account for a significant fraction of the variation in the long-wave downwelling energy relative to the short-duration changes induced by convection.

(b) Variations in atmospheric stability

One measure of the changes that occur under this regime is to monitor variations in the convective available potential energy (CAPE). CAPE, defined as the amount of buoyant energy available for a boundary-layer parcel lifted to its level of free convection, is a frequently used measure of the potential intensity of convection. In-depth descriptions of how to calculate CAPE can be found in numerous papers and texts. The definition of CAPE is meant to describe the amount of thermodynamic energy available to a rising parcel. However, it is clearly evident that CAPE will in general not represent the thermodynamics of a rising parcel, as it does not include mixing, vertical pressure effects which can have a strong impact on rising parcels associated with convective systems even over the TOGA COARE region (Trier et al. 1996), difficult-to-predict variations in precipitation loading that can decrease CAPE substantially (e.g. Xu and Emanuel 1989), and the corresponding releases of the latent heat of fusion that add
a significant amount of the available energy (Williams and Renno 1993). In this study we will put the controversy aside through realizing that using CAPE as a predictor of the actual amount of energy an individual parcel will experience in the atmosphere is fraught with difficulties, and instead examine relative changes in CAPE as simply one measure of changes in the amount of thermodynamic energy available to a convection parcel.

The other relevant thermodynamic parameter is the amount of energy that must be realized before a parcel can reach its level of free convection. This energy is termed the convective inhibition or CIN. This quantity is discussed by Colby (1983) and Bluestein and Jain (1985). In our calculations we included both the CIN below the level of free convection and any relatively small additional energy required by stable layers just above the level of free convection. The interpretation of CIN in terms of actual atmospheric convection is far more straightforward than the interpretation of CAPE, since ice processes and water loading are not an issue for CIN. The mean daily values of CAPE and CIN calculated from the R/V Moana Wave and the R/V Shiyan 3 are shown in Fig. 14 for the time period of 12 to 22 November. The values of CAPE and CIN were calculated using the average conditions in a 50 hPa level just above the surface. Neither water loading nor ice phases were included in these CAPE calculations. Soundings corrected by the Cole and Miller (1999) technique were used in this calculation.

The CAPE values in Fig. 14(a) during this period show dramatically increasing values of CAPE, indicating a build-up in the potential of the atmosphere to support more intense convection. The magnitude of the convective build-up deserves some mention also. According to LeMone et al. (1998), CAPEs found over the warm pool can be slightly higher than the 500 to 1500 J kg\(^{-1}\) typically found over the tropical oceans, partly due to slightly deeper convection. In this event, after the arrival of the dry air mass, the CAPE is below 500 J kg\(^{-1}\), while at the end of the period the values exceed 2000 J kg\(^{-1}\). The values at the end of the period are larger than the CAPEs observed in all of the convective cases studied by LeMone et al. (1998). There is also some evidence for a 2-day variation in the CAPE time series, which may correspond to the well-known 2-day variation in convection. The lack of widespread deep convection provides some evidence for a dynamic mechanism for this disturbance, such as internal waves (Liebman et al. 1997; Haertel and Johnson 1998), rather than a local cloud–ocean feedback, such as the “diurnal dancing” mechanism of Chen and Houze (1997).

The CIN values in Fig. 14(b) indicate a general tendency for the CIN to approach zero with time and hence an increasing probability that convection will occur. Extreme values occur on 13 November just following the arrival of the dry intrusion, which correspond to a general suppression of convection on that day. These high values of CIN are expected from the previously shown drying in the boundary layer and the tendency for the dry intrusion to be associated with a strong temperature inversion. The mean CIN on subsequent days approached \(-20\) J kg\(^{-1}\). The maximum vertical motion, \(w\), that would be needed to overcome this magnitude of CIN can be estimated by

\[
    w = (2/CIN)^{0.5}.
\]

For example using a value of CIN of 25 J kg\(^{-1}\), one would need vertical motions of \(\sim 7\, \text{m s}^{-1}\) to overcome the strong inversions at the top of the boundary layer. Although spatial variations in surface thermodynamic, local gravity waves and the observed mesoscale patterning could easily account for several m s\(^{-1}\) of vertical motion, it seems that the 7 m s\(^{-1}\) threshold would be difficult to overcome. Hence, given the magnitude of the daily mean value of CIN it would seem unlikely that much convection would
be initiated or maintained during this period despite increasing CAPE. We will later show that there is a diurnal variation in the CIN, which allows convection to develop preferentially.

A corresponding plot of the water vapour mixing ratio in the lowest 50 hPa is shown in Fig. 14(c). In this figure, there is a gradual increase in the mixing ratio with a strong minimum associated with the arrival of the dry air mass. Large values are observed at the end of the period. The increases in the mixing ratio are expected, given the lack of convective systems and the general warming of the ocean surface. The increase in low-level mixing ratio accounts for much of the increase in CAPE that took place during this time. In contrast, Emanuel and Bister (1997) proposed that dry middle-level air over
the tropics would likely lead to an increase in CAPE due to the relative difference in long-wave cooling between dry and moist air producing more cooling aloft. Our study shows that the atmosphere–ocean system is coupled, and that warm and dry air in the troposphere reduces cloud cover and, thereby, modifies the surface energy budget, which leads to warming of the sea surface, particularly under light winds where vertical mixing of the upper ocean is decreased.

5. DIURNAL VARIATIONS IN THE ATMOSPHERE AND OCEAN

Atmospheric vertical profiles of humidity and temperature with large CINs that occur with dry intrusions, evident in the soundings shown in Fig. 3, would not seem to be conducive to the development of convection. Hence, we are faced with the conundrum that conditions do not support the likelihood of convection, especially early in the period, but shallow convection does develop shortly after the dry air arrives over the IFA. In this section we will show that a diurnal cycle in CIN may partly be responsible for the generation of convection through producing conditions more favourable for convective activity during a portion of the day. We will begin by discussing the diurnal nature of the rainfall.

(a) Diurnal variations in precipitation

The diurnal variation of precipitating convection over tropical oceans has long received a great deal of attention as discussed by Randall et al. (1991), Hendon and Woodbury (1993), and Janowiak et al. (1994). Recent discussions of the diurnal cycle of convection during the TOGA COARE can be found by Guichard et al. (1996), Chen and Houze (1997), and Sui et al. (1998). One finding that can be summarized from these studies is that deep convection has an early morning maximum in precipitation but that shallower convection (i.e. echo tops less than 5–7 km in depth), such as was observed in this case, has an evening maximum. Since this period is suppressed in terms of deep convection, the time period we selected allowed us to study the physical processes responsible for this evening maximum without the interference of deep convection. Chen and Houze (1997) also used a portion of this period to illustrate that convection extending to middle levels had an early evening maximum in contrast to deep convection with an early morning maximum.

The diurnal cycle of rainfall was also obtained by using Z–R estimates from the Doppler radar aboard the R/V Vickers and averaging the results over the seven-day period. The diurnally varying rainfall distribution (Fig. 15) indicates a pronounced minimum at approximately 1000 LST followed by an abrupt increase in convective activity with a broad maximum in rainfall between 1800 and 2100 LST. This rainfall variation is generally consistent with the previously discussed Chen and Houze (1997) study. The broad maximum in Fig. 15 arrives in part due to the peak occurring at different times on different days during the period. Again, the overwhelming contribution to the total precipitation is from convective rainfall. The lack of stratiform precipitation is consistent with the dry layers aloft which would inhibit the formation of anvils through evaporation.

The magnitude of the precipitation and the diurnal tendency for this time period can be compared with the cruise average rainfall statistics also shown in Fig. 15. Surprisingly, the two rainfall rates are similar during the late afternoon and evening hours, but with significantly less precipitation during the night and early morning hours for the intrusion event. There is also a tendency for the precipitation during the dry intrusions to have a smaller precipitation contribution from stratiform precipitation.
Stratiform precipitation contributes to cruise average during the morning hours. The lack of convective anvils is consistent with the evening maximum in convection as various authors have proposed, that the early morning maximum is associated with anvil precipitation.

(b) Diurnal variations in the upper ocean and atmospheric boundary layer

Time series of the components of the surface energy budget are shown in Fig. 16(a) in contrast to the daily averages shown earlier. From this time series it is evident that there is a strong net flux of energy into the ocean during the day (again assuming no reflection) due to incoming solar short-wave energy. A smaller net energy loss is observed at night. The time series of SST and the average daily variations (Figs. 16(b) and (c)) clearly show a diurnal cycle, reflecting this variation in the energy flux with the warmest temperatures observed near solar noon. This diurnal cycle is well known to occur in this region under conditions of light wind and clear sky (e.g. Ostapoff and Worthem 1974; Lukas 1991). With the diurnal cycle in the SST there is also a diurnal cycle in the fluxes of latent and sensible heat from the ocean surface as obtained from the R/V Moana Wave (Figs. 16(d) and (e)) with peaks near noon local time. In addition to these peaks, a secondary maximum is evident near 0300 LST, possibly in conjunction with a semi-diurnal peak in the wind speed. A similar diurnal cycle in virtual temperature measured by the Radio Acoustic Sounding System (RASS) is noted through the depth of the boundary layer in Fig. 17. The magnitude of this variation is ~0.5 to 1.5 degC. These variations extend through and somewhat above the mixed layer, but with smaller amplitudes aloft. In these data there is a time delay of a few hours between maximum virtual temperature in the boundary layer and the elevated heat fluxes, which take place closer to local noon. There also seems to be some suggestion of a 2-day cycle in these measurements.
Figure 16. (a) Time series of the components of the surface energy budget including the short-wave solar incoming radiation ($Q_s$), net long-wave radiation cooling ($Q_l$) and the latent ($Q_e$) and sensible ($Q_h$) heat flux and the net change in the radiation budget ($Q_n$). The net long-wave cooling was calculated from the measured downward directed long-wave energy and the loss from the ocean surface estimated from the sea surface temperature. Cooling of the ocean surface due to rain and lateral heat flux due to advection are not shown. An albedo of 0.0 was used for the short-wave energy. (b) Time series and (c) average diurnal cycle of sea surface temperature. Average diurnal cycle of (d) sensible- and (e) latent-heat flux obtained from the methods previously discussed.
Given the diurnal cycles in the fluxes, it might be expected that diurnal cycles will be evident in CAPE and CIN. Unfortunately, the detection of a diurnal cycle is difficult for several reasons. One reason is that when convection is present it will change the convective parameters. Fortunately, in this study, the cloud fractions are small so that convective processes do not dominate the fields. Another difficulty is that the diurnal cycle is barely resolved with soundings every 6 h. In addition, the previously mentioned correction procedure employed for soundings modifies the diurnal signal, particularly for CAPE which is more sensitive to small changes in humidity. With these points in mind we show the diurnal changes in CAPE and CIN (Fig. 18). The diurnal change in CAPE is very small (Fig. 18(a)) and well within instrumental error. There is a detectable diurnal cycle in CIN with a minimum in the late afternoon (Fig. 18(b)) just preceding the maximum in convective activity. After this time the CIN increases at night as might be expected from radiation cooling. However, the strongest inhibition is in the morning, when precipitation is also at a minimum. Foltz and Gray (1979) found that at 10 LST there is lower tropospheric subsidence over this region. We propose that this change in CIN is potentially important in modulating when convection is initiated and maintained in this regime with dry, warm air above the boundary layer. During moist conditions, in contrast, Raymond (1995) found that CIN is relatively small and thus unimportant in the life cycle of convection.

(c) A possible explanation for the evening maximum in convection

It is tempting to postulate that the diurnal cycle in SST and the resulting changes in the fluxes are primarily responsible for the diurnal cycle in CIN and rainfall. However, we believe that the SST variation explains only a portion of the variation in CIN and rainfall. In our analysis the magnitude of the diurnal variation in the sensible-heat flux is
Figure 18. Bar graph of the diurnal variation in (a) convective available potential energy (CAPE) and (b) convection inhibition (CIN) calculated from a composite of rawinsondes launched from research vessels in the intensive flux array during the time period from 12 to 22 November 1992.

\[ \sim 4 \text{ W m}^{-2} \] Converting this diurnal variation into a net heating over the 600 m depth of the mixed layer (by dividing by the density of air and the specific heat of air at constant pressure) equates into a maximum variation of the heating rate of only 0.5 degC day\(^{-1}\). It is readily evident that this variation alone cannot sustain the 0.5 to 1.5 degC day\(^{-1}\) diurnal cycle in the virtual potential temperature observed with the RASS. In contrast, the magnitude of the heating rate from the absorption of solar short-wave radiation (primarily by water vapour) at solar noon (Fig. 19) ranges between 2 and 4.5 degC day\(^{-1}\) within the boundary layer for a sounding taken at the start of the dry intrusion. This heating rate was calculated assuming clear-air transfers of radiation. Hence, we can conclude that this radiation term is at least as important in the diurnal variation of temperature within the boundary layer as the variation in surface sensible heat (and likely to be more important!), but that both terms act together to drive the diurnal cycle in temperature and contribute to the diurnal variations in CIN. While such a diurnal temperature variation due to absorption of solar energy by water vapour would take place over land, it will often be relatively less important as the diurnal variations in the sensible-heat flux are far larger and conditions are generally not so moist.

The combination of the vertical profile of short-wave absorption and long-wave cooling (Fig. 19) should also produce strong diurnal changes in the strength of the inversion and thereby influence the CIN and CAPE in the soundings as follows. First, we note that the long-wave cooling is very strong at the inversion layer near the sharp gradient in moisture. The cooling in this case reaches nearly 6 degC day\(^{-1}\). It is also relatively large within the boundary layer, while being negligible within the
extremely dry layer above. Hence, from the long-wave cooling alone there is a tendency to destabilize the mixed layer by cooling the top, while stabilizing the atmosphere to deep convection through developing or strengthening the thermal inversion above the mixed layer. The importance of the vertical variation in the long-wave cooling to the inversion strength above dry layers was noted by Mapes and Zuidema (1996). The diurnal variation in the vertical profile of temperature comes about since there is a nearly matching pattern of short-wave absorption that makes this process of strengthening the inversion far less important during the day. Thus during the day (night) there is not only a relatively warmer (cooler) boundary layer, but the capping inversion is also growing less (more) slowly. From these thermal considerations alone it is not surprising that there is less inhibition during the afternoon with strong heating.

The diurnal cycle in moisture can in turn be estimated from the diurnal variation in the surface flux of latent heat, which has a diurnal variation of $\sim 35 \text{ W m}^{-2}$. The diurnal cycle of this flux can be obtained by dividing by the latent heat of vaporization and the density of air to obtain a diurnal maximum enhancement to the moistening rate of $1.75 \text{ g kg}^{-1}\text{day}^{-1}$. This large change will greatly modify the CAPE and to a lesser degree the CIN to again favour convection in the afternoon hours consistent with the behaviour of the soundings. From our analysis we propose that a number of factors, absorption of solar short-wave radiation within the boundary layer, diurnal changes in
the radiation balance near the inversion level, and the diurnal cycle of surface fluxes, all point to times between solar noon and sunset being favourable for the initiation of convection.

The relative importance of the various processes can be quantified by following the analysis of Crook (1996), where subject to a number of reasonable assumptions there are equations for calculating the changes in CAPE and CIN that can occur from modifying the temperature or moisture fields in the boundary layer. For the CIN we start with a modified form of the expression from Crook (1996) as

$$\frac{dCIN}{dt} = R \frac{dT}{dt} \left( \frac{p_{avg}}{p_0} \right)^k \ln \left( \frac{p_{lcl}}{p_i} \right)$$

(2)

with $R$ defined as the ideal gas constant, $k$ a standard thermodynamic constant, and $p$ the pressure with the subscript 'i' indicating the pressure at the inversion level, 0 a reference surface pressure, 'lcl' the pressure at the lifting condensation level, and 'avg' the average pressure between the pressure at the lifting condensation level and the inversion layer. To solve Eq. (2) we write

$$\frac{dT}{dt} = L \frac{dq}{c_p + L^2 q_s + R_v T^2}$$

(3)

where $c_p$ is the specific heat at constant pressure, $T$ is the temperature, $L$ the latent heat of vaporization, $q$ the mixing ratio of water vapour, and the subscripts 's' and 'v' denote saturation and vapour, respectively. Note that moisture changes the CIN through primarily changing the condensation level, which results in a different moist adiabat being used for convective ascent. Substituting the maximum diurnal variation in the rate of heating from the fluxes, we obtain a change due to the latent-heat flux of 3.4 J kg$^{-1}$h$^{-1}$ and the sensible-heat flux of 1.7 J kg$^{-1}$h$^{-1}$. The change due to short-wave absorption can be estimated from noting the differences between the mean boundary-layer warming and the warming near the inversion level. Since there is little heating in the dry layer this term is quite large, and depending upon where the inversion occurs we estimate the modification to the CIN from the short-wave absorption to be between 3.5 and 5.2 J kg$^{-1}$h$^{-1}$. The radiative term is the largest, which is surprising considering the previous focus on variations in surface fluxes. Consistent with Crook (1996), we note that the temperature changes are relatively more important for CIN. These simple calculations suggest that the diurnal enhancement in the fluxes, together with the short-wave absorption, act together to remove CIN at a maximum rate of $\sim$10 J kg$^{-1}$h$^{-1}$ of CIN. This rate is larger than the diurnal cycle in CIN shown earlier, but reasonable, since the changes are maximum enhancements at solar noon without any compensating effects. More importantly, these changes in CIN again suggest that the fluxes and short-wave changes will work to destabilize a clear atmosphere with a dry layer above a very moist boundary layer. If the dry layer was not present above the boundary layer, the diurnal changes in the fluxes would still favour a similar diurnal cycle in the rainfall. However, this analysis shows that the radiation terms greatly enhance this cycle.

One must, however, treat these results with some caution, since we used an extreme example at the start of the dry intrusion to illustrate the radiational effects. We also assumed clear air for the radiation transfers, and ignored the impact of convective feedbacks, vertical motions and entrainment into the boundary layer, which are also likely to have diurnal variations. However, even with this large number of caveats, the importance of radiative processes in clear air on atmospheric stability in this unique regime and the combined impact of radiation and surface fluxes on the diurnal cycle of CIN are still relatively solid conclusions. We believe that the importance of radiative processes in the diurnal variation in the inversion and the impact of absorption of short-wave incoming solar radiation on CIN are relative new findings for this region. Chen and
Houze (1997) do mention the possible role of the latter process, but without presenting this type of supporting calculation.

The time rate-of-change of CAPE, which for boundary-layer changes is simply the change in moist static energy, \( \dot{h} \), is given as

\[
\dot{h} = g z + c_p T + L_q
\]

(4)

where \( g \) is the acceleration due to gravity and \( z \) is height. The changes in CAPE due to latent and sensible heat are straightforward to estimate. Substituting in the maximum rate of change in the mixing ratio and temperature respectively, one can estimate that the latent-heat flux changes the CAPE by a rate of 182.3 J kg\(^{-1}\) h\(^{-1}\), and the sensible-heat flux by 20.9 J kg\(^{-1}\) h\(^{-1}\), as the maximum change due to the diurnal cycle. To obtain the contribution from short-wave absorption, the vertical profile of heating must be considered, since the changes will extend through the depth of the column. The greatest heating is in the boundary layer with the general difference between the boundary-layer heating and the free atmosphere being \(~1.5\) degC day\(^{-1}\). This heating difference equates to a change in CAPE of \(~62.8\) J kg\(^{-1}\) h\(^{-1}\). This change is smaller than the contribution from the latent-heat flux, but it is nearly three times larger than the contribution from the sensible-heat flux. The relative importance of these two terms is consistent with the moisture changes being far more important than the temperature changes in modifying the CAPE as noted by Crook (1996). Since the predicted changes in CAPE are not realized, there must be a ‘sink’ of CAPE that balances the increases from these sources. The most likely sinks are convection, which must therefore dry and cool the boundary layer, and long-wave cooling, which reduces CAPE through the relatively ineffective route of cooling the temperature.

6. CONCLUSIONS

In this study we found that dry air associated with a middle-latitude wave was able to reach equatorial waters and produce a noticeable impact on the atmospheric thermodynamics, cloud and radiative processes and the ocean surface. The event greatly reduced the likelihood for deep convective activity, which resulted in a large net flux of energy into the ocean, an increase in SST and a build-up in CAPE until unusually large values were observed. Hence, under favourable (i.e. light wind) large-scale conditions, the intrusions can aid in recharging the tropical atmosphere toward a much more unstable state until the next disturbed period and/or westerly burst arises. Although widespread deep convection was generally absent, the diurnal cycle of radiation and surface fluxes was able to drive shallow precipitating and non-precipitating convection even under relatively adverse conditions with large daily average values of CIN. These shallow clouds moistened the lower free troposphere. On average a reduction in CIN of \(~30\) J kg\(^{-1}\) was observed from 1000 to 1600 LST as the diurnal heating cycle progressed, decreasing the vertical motions needed to trigger convection by \(~7.5\) m s\(^{-1}\).

The concept of a CIN regulated tropical atmosphere is at first an unusual thought, but such a state might be more common in the tropics than one might expect if the frequency of the intrusion events during COARE can be extrapolated to other times and other locations. During the COARE IOP, these events were observed with an average period of 8–16 days. We propose a recovery rate due to local processes of 10–20 days for this strong event. The magnitude of the advection suggests that these intrusions must exert significant influence over the region’s water vapour budget, with the atmosphere spending a great deal of time trying to adjust to the dry conditions
imposed by lateral advection. Clearly, it would be difficult for the atmosphere to reach a radiative-convective equilibrium under such circumstances. Several other new findings are evident from this study including: the concept of the intrusions corresponding to deep boundary layers, to pronounced decreases in downwelling long-wave radiation, to extreme surface fluxes of latent heat, and the illustration of the possible impact of convection and of the arrival of the dry air on the depth of the ocean mixed layer.

Our conclusions are summarized in the conceptual model of a dry intrusion moving over equatorial waters (Fig. 20). The generality of the model will be tested by investigations of other events and through cumulus ensemble modelling. The conceptual model is based on this work and the previous studies discussed in the introduction, especially the work of Numaguti et al. (1995). The boundary-layer portion of this model is, of course, only relevant when the dry-air intrusion exists at a low enough level that boundary-layer entrainment can occur.

We propose that studies of these middle-latitude/tropical exchanges must be investigated for other locations and time periods. It is also evident that soundings with greater frequency and accuracy in the humidity sensors are needed to clarify variations in the observed diurnal cycle. A limitation of this study and other COARE investigations is that the soundings were in need of a humidity correction, and this correction can impact calculation of the diurnal cycle of CAPE and CIN. These problems exist in other research and operational soundings as noted by Cole and Miller (1999).

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