Variability of Water Vapor, Infrared Radiative Cooling, and Atmospheric Instability for Deep Convection in the Equatorial Western Pacific

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ABSTRACT
In the troposphere of the equatorial western Pacific, the water vapor variability dominates the temperature variability in changing the clear-sky infrared (IR) cooling rate. The large water vapor variability, especially its bimodal distribution at certain levels of the upper troposphere, leads to distinct structures of the clear-sky IR radiative cooling rate. The IR cooling rate, its maximum normally in the upper troposphere (~300 hPa) and minimum in the lower troposphere (~650 hPa), tends to become vertically uniform when the upper troposphere is abnormally dry. A local, maximum IR cooling rate may occur in the boundary layer when the lower troposphere becomes extraordinarily dry. The changes in IR cooling due to the water vapor variability affect the rate of generation of convective available potential energy (CAPE) and the conditional instability for deep convection. Little or no mean rainfall over an area of roughly $3 \times 10^5$ km$^2$ is observed when either the rate of generation of CAPE suffers from a reduction (magnitude of 50 J kg$^{-1}$ day$^{-1}$) or IR cooling decreases with height. The observed variability of water vapor results from both local vertical processes and the large-scale horizontal circulation. Horizontal advection accounts for a large fraction of the drying that is responsible for the changes in the IR cooling profile and in the atmospheric instability for deep convection. These results suggest that interactions among water vapor, radiation, and deep convection must be assessed by fully taking the large-scale circulation into consideration. This study is based on an analysis of upper-air soundings collected during the Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Response Experiment Intensive Observing Period and calculations of a radiative transfer model.

1. Introduction

Water vapor undergoes large variability in the troposphere and exhibits a bimodal probability distribution at certain levels of the upper troposphere in the western Pacific warm pool region (Brown and Zhang 1997). The bimodality is likely to be associated with the tropical intraseasonal oscillation, also known as the Madden–Julian Oscillation (MJO) (Madden and Julian 1971, 1972). The moist mode occurs mainly in convectively active phases of the MJO and the dry mode in suppressed phases. The fluctuations between the two modes can be as large as 50% in relative humidity and 2 g kg$^{-1}$ in specific humidity in the upper troposphere.

Water vapor content in the tropical troposphere above the boundary layer varies with deep convective activity. Feedback from water vapor to tropical deep convection may also exist. For example, dry air entrainment into clouds tends to dilute the buoyancy of convective air parcels and thereby weakens convection (Stommel 1947). Another possible water vapor effect on the large-scale behavior of tropical deep convection is through modifying the vertical structure of atmospheric infrared (IR) radiative cooling rate. Infrared cooling, by destabilizing the atmosphere, promotes a large-scale environment in favor of deep convection not only in the climatological mean, but also for transient behavior of deep convection. Hu and Randall (1994) proposed that interactions among convection, radiation, and surface energy flux may result in a local intraseasonal oscillation in deep convection in the absence of the large-scale circulation. In their proposal, IR cooling plays a critical role in maintaining the intraseasonal oscillation. Fluctuations in tropical deep convection simulated by models of different complexities in the absence of the large-scale circulation all owe their existence in part to IR cooling (Randall et al. 1994). Zhang and Hendon (1997)
questioned whether or not a local convection–radiation–
flux interaction alone acts as the main mechanism for
the dominant signals of the intraseasonal oscillation,
namely, the MJO. Nevertheless, the role of radiative
cooling in the large-scale convective variability needs
further exploration.

The sensitivity of IR cooling to water vapor has been
tested in a number of studies. Artificial changes in
the water vapor profile were made either randomly or uni-
formly to assess the sensitivity of the IR cooling rate
to random or systematic errors in water vapor mea-
surement (e.g., Rogers and Walsh 1965; Doherty and
Newell 1984). Vertically uniform changes in the water
vapor profile alter the amplitude but not the vertical
structure of IR cooling. The structure of the radiative
cooling and heating rate is far more sensitive to changes
in the vertical gradient of water vapor. For example, a
dry layer with a limited vertical extent may have a much
larger effect on IR cooling than drying that is vertically
uniform (Mapes and Zuidema 1996). It is not well
known how the vertical structure of IR cooling varies
with the observed water vapor variability, which is nei-
ther vertically uniform nor random and is not always
confined to a thin layer.

In the present study, we will address the following
questions. Do the observed fluctuations in water vapor
have strong effects on IR cooling? Do the corresponding
changes in IR cooling affect the atmospheric instability
for deep convection? If the answers to these two ques-
tions are positive, then what are the main mechanisms
for the variability of water vapor that cause the changes
in IR cooling and instability? As for the last question,
we are mainly interested in the drying mechanisms,
knowing that the troposphere is moistened primarily by
convective processes.

We consider only clear-sky IR cooling. Radiative
heating and cooling of clouds are usually much larger
than those in clear sky (e.g., Mehta and Smith 1997).
The importance of clear-sky IR cooling, however,
should not be judged simply by its magnitude relative
to the cloud radiative effect. In clouds, radiative cooling
and heating are largely enhanced only in confined layers
near cloud tops and bases (e.g., Webster and Stephens
1980; Ackerman et al. 1988; Guan et al. 1995). Effects
of such vertically confined enhancement in cooling or
heating on the atmospheric instability for deep convec-
tion are unknown. In contrast, water vapor variability
can induce changes in clear-sky IR cooling profiles with-
in a deeper layer and consequently modify the instability
(see section 3b). The importance of clear-sky IR cooling
to deep convection has been demonstrated by model
simulations. Using a cloud model, Dharssi et al. (1997)
showed that IR cooling averaged over a domain in-
cluding both cloudy and clear skies is far more important
to convective development than the cloud radiative ef-
fects alone. Hu and Randall (1994) were able to simulate
30–50-day oscillations in their one-dimensional model
even when all clouds were eliminated in the radiation
calculations. Finally, clear-sky IR cooling is not inde-
dependent from clouds but varies with water vapor, which
is closely related to cloud activity (e.g., Udelenhofen
and Hartmann 1995). Clouds may project their radiative
effects on clear-sky IR cooling by changing clear-sky wa-
ter vapor distribution. For these reasons, effects of the
water vapor variability on clear-sky IR cooling need to
be explored.

The water vapor variability was diagnosed in this
study using soundings collected during the Intensive
Observing Period (IOP; 1 November 1992–28 February
1993) of Tropical Ocean Global Atmosphere Coupled
Ocean–Atmosphere Response Experiment (TOGA
COARE) (Webster and Lukas 1992) and its effect on
the clear-sky IR cooling rate estimated using a radiative
transfer model. The data and the model are described
in section 2. In section 3, we first describe the variability
in water vapor and then demonstrate that the variability
of clear-sky IR cooling depends more sensitively on that
of water vapor than temperature. Next, the variability
of the clear-sky IR cooling rate and the atmospheric
instability for deep convection solely due to the ob-
served fluctuations in water vapor are explored. Inter-
esting changes in the vertical structure of the IR cooling
rate and instability for deep convection motivated us to
further explore mechanisms for the water-vapor fluc-
tuations. We found that horizontal advection by the
large-scale circulation accounts for a large fraction of
the drying in the troposphere. Concluding remarks are
given in section 4. Technical information of the humid-
ity sensor of the TOGA COARE sounding system is
provided in the appendix, where known instrumental
errors are discussed.

2. Data and model

TOGA COARE temperature and humidity soundings
were used for levels between 1000 and 200 hPa. The
soundings, taken four times a day, have an interpolated
vertical resolution of 5 hPa (Parsons et al. 1994).
Soundings with suspicious or missing data between
1000 and 100 hPa were excluded. A total of 1492
soundings from eight sites within and near the Inten-
sive Flux Array (IFA) (Fig. 1) have passed our screen-
ing. Above the 100-hPa level, standard tropical profiles
of temperature and humidity taken from McClatchey
et al. (1972) were used. Between 200 and 100 hPa, the
COARE and standard soundings were merged into a
smooth transition. In addition to the individual COARE
soundings, gridded sounding analyses of Lin and John-
son (1996b) were also used in the diagnoses of the
moisture budget of the IFA.

The IR radiative transfer model of Chou and Suarez
(1994) was used to calculate the clear-sky IR fluxes.
In the model, the IR spectrum is divided into eight
bands, and either the $k$-distribution approximation with
the one-parameter scaling or a table lookup approach
is used to compute transmission functions. In addition
to water vapor, CO$_2$, and O$_3$, the model also includes other trace gases, such as N$_2$O, CH$_4$, and CFCs. Compared to line-by-line calculations, fluxes are computed accurately with an error of $<1\%$.

In the calculations of the clear-sky IR cooling rate, time-independent profiles of concentrations were used for trace gases other than water vapor. A constant surface temperature (303.75 K) was used in all the calculations to exclude its influence on the radiative processes. Included in the calculations are soundings with saturation humidity, which may have gone through clouds. Truly cloud-free situations in the equatorial western Pacific were rare (Chou and Zhao 1997). When soundings with relative humidity exceeding 95% were excluded from the radiation calculations, similar results were obtained. Cloud contamination should be a lesser problem during convectively suppressed phases of the MJO, when observed large-scale rainfall is small, and thick, tall clouds are likely to be isolated.

To demonstrate the large-scale variability of water vapor and its radiative effects, we have generated daily mean time series based on individual soundings and then averaged over the six sounding sites in the IFA (Fig. 1). We will refer to such an averaged time series as an “IFA daily mean.” The time average of this IFA daily mean over the IOP is then referred to as the “IFA-IOP mean.” The identification of convectively active and suppressed phases of the MJO in the IFA is based on studies with detailed diagnoses of the dynamic and thermodynamic fields in the IFA (e.g., Lin and Johnson 1996a) and large-scale propagating patterns of convection over a broad region of the eastern Indian and western Pacific Oceans (e.g., Chen et al. 1996) during the TOGA COARE IOP.

3. Results

a. Water vapor variability

During the COARE IOP, tropospheric water vapor in the equatorial western Pacific exhibited a large variability. The middle and upper troposphere were extremely dry during convectively suppressed phases but relatively moist during active phases of the MJO. The variability in moisture is clearly shown in the IFA daily mean time series of $\Delta q$, the deviation of specific humidity $q$ from its IOP mean (Fig. 2a). Positive $\Delta q$ (relatively high $q$) is found during periods of active convection (e.g., IOP days 40–55 and 78–120), which is associated with the passage of MJO events over the IFA. Negative $\Delta q$ (low $q$) is more likely to be found during the suppressed phases of the MJO (e.g., IOP days 10–35 and 60–75). The major dry events are also evident in the time series of relative humidity (Lin and Johnson 1996a). Figure 2a reveals that extremely dry ($\Delta q \leq -1$ g kg$^{-1}$) conditions can occur in the deep troposphere (1000–300 hPa), but also only in the lower troposphere above the boundary layer (850–500 hPa) or only at higher levels (above the 600-hPa level). As will be shown in section 3b, the resulting changes in the vertical gradient in $q$ would lead to a substantial variation in the vertical profile of the IR cooling rate.

Another interesting and intriguing feature of the water vapor variability is a bimodal distribution at certain levels in the upper troposphere. The bimodal distribution in $q$ can be seen in Fig. 3, where the probability distribution function of $\Delta q$ for the COARE IOP is plotted. In the upper troposphere above the 250-hPa level, the specific humidity is very low and the $\Delta q$ distribution is concentrated near zero. But the $\Delta q$ distribution is widely spread below the 450-hPa level. The bimodal structure in the $\Delta q$ distribution is apparent near 300 hPa. The dry and moist branches of the bimodal structure deviate from the mean $q$ at those levels by 50%. The bimodality is not sensitive to sampling procedures and exists outside the IFA and beyond the IOP. The possibility that the bimodality results from instrumental errors is discussed in the appendix.

Brown and Zhang (1997) proposed that the bimodal distribution in humidity is related to the abrupt changes in cloudiness and the circulation associated with the convectively active and suppressed phases of the MJO. From Fig. 2a, three major periods of dry mid- to upper troposphere (IOP days 12–22, 27–40, 59–76) are identified. Hereafter, these periods together will be referred to as “the dry periods” and other days during the IOP will be referred to as “the moist periods.” The dry periods were all within the convectively suppressed phases of the MJO, and most of the moist periods were in the active phases. The separated moist and dry modes of the $\Delta q$ distribution during the moist and dry periods are clearly shown in Figs. 4a and 4c.

b. Clear-sky IR cooling rate

We first discuss the effect of water vapor variability on the clear-sky IR cooling rate versus the effect of the
temperature variability. Four sets of clear-sky IR cooling rates were calculated. The first set was calculated using individual soundings of temperature and humidity ("control" set). In the second set, the temperature profile in each sounding was replaced with the IFA-IOP mean ("mean-T"). Profiles of humidity, instead of temperature, were replaced with the IFA-IOP mean in the third set ("mean-q"). The last set is a single IR cooling profile calculated using the IFA-IOP mean profiles of both temperature and humidity ("mean-T&q"). The total variance of daily mean IR cooling (Fig. 5a, solid line) can be mostly accounted for by the variability of water vapor alone (dotted line). The variance is drastically reduced if the variability of water vapor is removed (dashed). For higher-frequency (6-hourly) IR cooling (Fig. 5b), temperature contributions to the total variance become comparable to the water vapor contribution only in the lowest part of the troposphere. The variability of water vapor alone can still account for most of the variance of the IR cooling rate, especially in the upper troposphere. The mean IR cooling profile can be almost completely duplicated when the temperature variability is neglected, but noticeable biases would exist if the water vapor variability is removed (Fig. 5c). These suggest that the IR cooling rate changes mainly with the distribution of water vapor rather than temperature. In the discussions hereafter, the IR cooling rates were calculated using the IFA-IOP mean temperature and individual humidity profiles to isolate the effect of water vapor.

An IFA daily mean time series of IR cooling profile for the IOP is given in Fig. 2b. For most of the IOP,
maximum cooling ($\geq 2.6 \text{ K day}^{-1}$) occurred in the upper troposphere, roughly between 400 and 250 hPa. On certain days, however, the upper-tropospheric cooling was substantially reduced and the level of the maximum cooling descended to 600–400 hPa (e.g., IOP days 30–40, 65–80). Seemingly independent of the upper-level IR cooling, local maximum cooling occurred sporadically within the boundary layer (e.g., IOP days 0–5, 12–17, 61–64, 83–88, 94–98). Maximum cooling appeared at the same level of maximum vertical gradient of $q$. It is clear from Fig. 2 that the two episodes of maximum cooling in the midtroposphere coincided with the extremely dry conditions in the middle and upper troposphere. The episodes of maximum cooling in the boundary layer all occurred when the lower troposphere became extremely dry.

The probability distribution functions of the clear-sky IR cooling rate were calculated respectively for the moist and dry periods defined earlier. The distribution for the moist periods (Fig. 4b) resembles in many respects the distribution for the entire IOP (not shown) and the total cooling rate, including both clear and cloudy skies in the warm pool (Ramsey and Vincent 1995): a maximum near 300 hPa and a minimum near 650 hPa. The cooling distribution for the dry periods (Fig. 4d), however, is quite different. The most striking feature is a lack of vertical structure between 1000 and 300 hPa. The relatively dry middle and upper troposphere reduces the cooling rate in the upper troposphere and enhances it in the lower troposphere (800–600 hPa). This profile of cooling is not very effective in destabilizing the atmosphere.

It has been seen (Fig. 2a) that substantial drying does not always occur simultaneously in both the lower and upper troposphere. There were several episodes (IOP days 1–4, 10–15, 59–63, and 82–100) when drying was more severe in the lower troposphere than aloft. During these episodes, the cooling magnitude decreases quickly with height in the lowest part of the atmosphere, with the clear-sky IR cooling rate being larger (distribution peak is at about 2.5 K day$^{-1}$) than that during the moist periods (2 K day$^{-1}$) at 1000 hPa but remaining the same at 700 hPa. The boundary layer is usually very opaque in the IR spectral region, which reduces radiative cooling. Increases in the surface cooling indicate decreases in the boundary layer IR opacity due to the reduction in moisture. In contrast, when the lower troposphere is relatively moist, the distribution of the IR cooling rate is quite uniform in the vertical from 1000 to 600 hPa and then increases with height up to 300 hPa. These features can be seen from the time series of the cooling rate (Fig. 2b). An IR cooling profile with its maximum in the lowest part of the atmosphere tends to have a stabilizing effect.

The clear-sky IR cooling rate exhibits characteristics very similar to the total IR cooling rate including both clear and cloudy skies estimated by Mehta and Smith (1997). They showed that when deep convective clouds are present during active phases of the MJO, the cooling rate is enhanced, although by a much larger amplitude than in the clear sky, at the cloud top, with the maximum occurring at about 300 hPa. When deep convection is absent during suppressed phases of the MJO, upper-tropospheric cooling rates are reduced, maximum cooling rates descend to the midtroposphere, and the vertical cooling rate profile is nearly uniform. The descended maximum cooling rate in the midtroposphere and the more stable IR cooling profile during convectively suppressed periods are not unique to the intraseasonal timescales but are also found on synoptic scales (e.g., Cox and Griffith 1979).

c. Instability for deep convection

One immediate effect of the changes in the IR cooling rate due to the water vapor variability is modifying the temperature profile, which in turn modifies the atmospheric instability for deep convection. This effect can be illustrated by demonstrating changes in the rate of generation of convective available potential energy (CAPE) and in the criterion of the conditional instability.

1) CAPE

CAPE is a necessary, but not sufficient, condition for deep convection and measures its potential strength. According to the “quasi-equilibrium” theory (Arakawa and Schubert 1974), it is the generation of CAPE, rather than CAPE itself, that may be viewed as a large-scale controlling factor of deep convection. CAPE can be generated mainly by surface energy flux, radiative cooling, large-scale temperature advection, and lifting. As-
Assuming that the other processes remain unchanged, the contribution to the generation of CAPE by IR cooling alone can be estimated as (Emanuel 1994)

\[
\frac{\partial \text{CAPE}}{\partial t} \approx R \int_{p_\text{LNB}}^{p_i} \frac{Q}{\rho} \, dp,
\]

where \( Q \) is the IR cooling rate (in unit K day\(^{-1} \)), \( R = 287 \text{ J K}^{-1} \text{ kg}^{-1} \) is the gas constant, \( p_i \) is the pressure level from which an ascending air parcel initially resides, and \( p_\text{LNB} \) is the level of neutral buoyancy for the parcel originating from \( p_i \). Using the IFA daily mean IR cooling (Fig. 2b), we estimated the right-hand side of (1) for selective \( p_i \) and kept \( p_\text{LNB} = 200 \text{ hPa} \) for simplicity. The results are shown in Fig. 6a. Because the other factors affecting the rate of change of CAPE are not considered, the magnitude of the estimated \( \partial \text{CAPE}/\partial t \) due to IR cooling is meaningful only in a relative sense.

Toga COARE observations have shown that air parcels converge into tropical deep convective systems at different levels during different stages of the convective life cycle, ranging from the surface to about the 400-hPa level (Mapes and Houze 1995). Figure 6a shows
Fig. 5. (a) Variance of daily mean clear-sky IR cooling rates (K² day⁻²), (b) variance of 6-hourly clear-sky IR cooling rates (K² day⁻²), and (c) IFA-IOP mean clear-sky IR cooling rates (K day⁻¹), calculated using individual temperature and humidity soundings (control, thin solid lines), the IOP mean temperature profile and individual humidity soundings (mean-q, dotted), the IOP mean humidity profile and individual temperature soundings (mean-T, dashed), and the IOP mean temperature and humidity profiles [mean-T&q, thick solid line in (c)].

that \( \partial \text{CAPE} / \partial t \) for \( p_i = 1000–850 \) hPa are of the same phase and similar amplitude (920–1000 J kg⁻¹ day⁻¹). This suggests that convection whose air parcels ascend initially from anywhere between the surface and 850 hPa may achieve similar strength. Another interesting feature is the sudden reductions in \( \partial \text{CAPE} / \partial t \) of amplitudes 50–100 J kg⁻¹ day⁻¹ found for parcels from all levels. At the lowest levels (\( p_i \geq 850 \) hPa), without these sudden reductions, \( \partial \text{CAPE} / \partial t \) does not change much and stays close to their IOP means (straight solid lines).

The variability of \( \partial \text{CAPE} / \partial t \) is completely due to the water vapor variability, because the latter is the only independent variable in our calculations. Large reductions in \( \partial \text{CAPE} / \partial t \) occur only during periods of severe drying (Fig. 2a). Drying confined to the mid- and upper troposphere is not very effective on \( \partial \text{CAPE} / \partial t \) of parcels from the lower troposphere. If drying extends through the entire troposphere, large reductions in \( \partial \text{CAPE} / \partial t \) occur for parcels from all the levels. Therefore, the rate of generation of CAPE due to IR cooling is more sensitive to low-level water vapor fluctuations than upper-level ones.

2) CONDITIONAL INSTABILITY

The conditional instability criterion for a saturated parcel is

\[
\gamma = \partial \theta_e / \partial p = < 0 \quad \text{stable},
\]

\[
\gamma = \partial \theta_e / \partial p = > 0 \quad \text{unstable}
\]

(2)

where

\[
\theta_e = \theta_e(T,p,q)
\]

(3)

is the saturation equivalent potential temperature. In (3) \( L = 2.5 \times 10^6 \) J kg⁻¹ is the latent heat of condensation at 0°C, \( q_s \) is the saturation mixing ratio, \( c_p = 1004 \) J K⁻¹ kg⁻¹ is the specific heat at constant pressure, and \( T \) is the temperature. The rate of change of \( \gamma \), \( \partial \gamma / \partial t \), indicates whether destabilizing processes are at work. When \( \partial \gamma / \partial t > 0 \), the atmosphere tends toward a state of conditional instability if originally \( \gamma < 0 \), or the instability is enhanced if originally \( \gamma > 0 \). It can be shown that

\[
\frac{\partial \gamma}{\partial t} = \frac{\partial}{\partial p} \left[ P(T) \frac{\partial T}{\partial t} \right].
\]

(4)

where \( P(T) \) is a function of temperature and changes only slightly with pressure. The destabilizing effect of IR cooling, \( Q \), on the conditional instability can then be isolated from others if \( \partial T / \partial t \) is replaced by \( -Q \).

\[
\frac{\partial \gamma}{\partial t} \approx -P(T) \frac{\partial Q}{\partial p}.
\]

(5)

A time series of \( \partial \gamma / \partial t \) due to IR cooling was calculated using the IFA daily mean temperature and IR cooling rate (Fig. 6b).

The destabilizing effect of IR cooling was mainly confined to between 700 and 300 hPa. Above the 300-hPa level, IR cooling tended to stabilize the atmosphere locally. Below the 700-hPa level, the effect of IR cooling is either very weak or occasionally stabilizing locally when the water vapor content there was abnormally low (e.g., IOP days 10–15, 60–65, 83–87, 92–100). When the upper troposphere was relatively moist and the IR cooling maximized at its normal level near 300 hPa (Fig. 2b), the destabilization affected a deep layer up to the 300-hPa level. The destabilizing effect, however, appeared to be limited to a shallower layer below the 500-hPa level when the upper troposphere was abnormally dry and the level of maximum IR cooling descended to the midtroposphere (e.g., IOP days 30–40, 66–77, and 85–86). During these periods, the
destabilizing effect of IR cooling is weakened above the 400-hPa level, where the amplitude of the cooling decreases with height. Whether such subtle changes in the distribution of the destabilizing effect of IR cooling really mattered to convective variability is a question that cannot be answered here. Low-level destabilization above the boundary layer is nevertheless in favor of deep convection, which tends to make air parcels positively buoyant when they are lifted by large-scale low-level convergence. Higher-level destabilization may help increase the level of neutral buoyancy and thereby promote deeper convection.

The presumed effects of clear-sky IR cooling due to water vapor on the atmospheric instability for deep convection, although difficult to quantify using observations and unlikely to dictate the variability of deep convection, can be further illustrated by examining the observed IFA mean rainfall rate during the COARE IOP (Fig. 6c). Several episodes of little or no IFA mean rainfall (e.g., IOP days 15–20, 29–34, 61–62, 69–75, 83–88, and 94–99) coincided with at least one of the following phenomena: 1) a sudden reduction in $\partial \text{CAPE}/\partial t$ over 50 J kg$^{-1}$ day$^{-1}$ for parcels originating from the lowest part of the troposphere (below the 800-
hPa level), \(^1\) 2) local stabilization by IR cooling in the lowest troposphere, 3) the lack of a deep layer of destabilization, or 4) penetration of the upper-tropospheric stable layer down to the midtroposphere. It cannot be concluded, however, that the observed variability of precipitation (Fig. 6c) was solely caused by the variability of water vapor. The effects of subtle but potentially critical changes in temperature due to large-scale circulation and other processes have been excluded from the calculations.

d. Mechanisms for water vapor variability

The results presented in the previous three sections motivated us to explore the mechanisms for the water vapor variability responsible for the changes in the vertical structure of the IR cooling rate and consequently in the instability for deep convection. The equation for specific humidity can be written as

\[
\frac{\partial q}{\partial t} = -\mathbf{v} \cdot \nabla q - \frac{\partial q}{\partial p} - S, \quad (6)
\]

where the three terms on the right-hand side represent the drying or moistening effects by, respectively, large-scale horizontal advection, vertical advection, and processes of unresolvable scales. Here, \(S\) is related to the apparent moisture sink \(Q_s\) (Yanai et al. 1973) as \(S = Q_s/L\). The physical processes \(S\) represents include water vapor condensation, evaporation of precipitation, evaporation from the surface, and eddy transport (convective or mesoscale updrafts and downdrafts, cloud detrainment and entrainment). Because \(S\) is calculated as a residual in (6), it also contains errors in data and analysis. Large uncertainties also exist in the term \(\omega \partial q/\partial p\) because \(\omega\) needs to be diagnosed from horizontal wind divergence. In (6), \(\omega \partial q/\partial p\) and \(S\) are the dominant terms, but they are always anticorrelated, with strong moistening by \(\omega \partial q/\partial p\) coinciding with strong drying by \(S\). The combination of the two is comparable in magnitude to the large-scale horizontal advection. Therefore, in discussing the mechanisms for the water vapor variability, (6) can be simplified as

\[
\frac{\partial q}{\partial t} = -\mathbf{v} \cdot \nabla q + R, \quad (7)
\]

where \(R = -\omega \partial q/\partial p - S\) represents all the processes responsible for the observed water vapor variability other than the large-scale horizontal advection. From a large-scale perspective, those processes are essentially local (on unresolvable scales) and one-dimensional in the vertical. For convenience, we will hereafter refer to those processes as “vertical processes.” In one-dimensional single column models and two- or three-dimensional cloud ensemble models, the vertical processes (except large-scale vertical advection) are explicitly simulated or parameterized to different extents, while the large-scale horizontal advection is treated as a given condition.

Using the sounding analyses of Lin and Johnson (1996b), we now examine the relative contributions by the drying/moistening processes in (7) to the observed water vapor fluctuations that lead to the previously shown changes in the IR cooling rate and instability. IFA daily mean time series of the vertical processes \((R)\), large-scale horizontal advection \((-\mathbf{v} \cdot \nabla q)\), and the humidity tendency \((\partial q/\partial t)\) are plotted in Fig. 7. The dominant effect of the vertical processes was to moisten the troposphere, especially the mid- and lower troposphere. In contrast, \(-\mathbf{v} \cdot \nabla q\) mainly dried the troposphere. They both show large high-frequency (<5 days) fluctuations. It is interesting that days of strong drying by \(-\mathbf{v} \cdot \nabla q\) coincided with days of strong moistening by \(R\), especially in the lower troposphere (>500 hPa). Almost all periods of strong drying \((\partial q/\partial t < 0)\) were caused by large-scale horizontal advection, and both the mechanisms \((-\mathbf{v} \cdot \nabla q + R)\) were responsible for the moistening \((\partial q/\partial t > 0)\). The upper troposphere (above the 300-hPa level) appears to be dried by both the large-scale subsidence and horizontal advection. The large-scale circulation, therefore, played an indispensable role in the tropospheric water vapor budget of the IFA.

4. Conclusions

Upward motions related to moist convection, carrying water vapor out of the atmospheric boundary layer, is the major moisture source for the rest of the troposphere in the Tropics. The most efficient process of removing moisture is condensation–precipitation. Tropospheric humidity, however, does not decrease because of condensation–precipitation or a lack of convective moistening. Specific drying mechanisms need to be identified. Large-scale subsidence, dynamically forced by remote deep convection or thermodynamically induced by local IR cooling, is commonly considered as a main drying mechanism in the troposphere. Meanwhile, large-scale horizontal advection is known to be very effective in changing the water vapor content. Which one is the dominant large-scale drying process of the tropical troposphere is an issue of scale and location. Our observations have shown that on the intraseasonal or smaller timescales and in the equatorial western Pacific the drying effect of the large-scale subsidence is secondary to that of the large-scale horizontal advection, especially in the mid- and lower troposphere. The magnitude of the large-scale subsidence is too small (<1.0 hPa h\(^{-1}\); see Lin and Johnson 1996a) to account for the observed drying. Unsaturated con-

\(^1\) But notice that episodes of large IFA mean rainfall rates (e.g., IOP days 41–46, 48–56, 76–81, and 88–92) were not always associated with large increases in \(\text{CAPE} / \partial t\) above its means for parcels originating from the lowest levels.
vective and mesoscale downdrafts (Zipser 1969) can efficiently dry the boundary layer and the lower part of the troposphere (e.g., Fitzjarrald and Garstang 1981) but only in confined convective areas. Large-scale horizontal advection appears to be much more efficient in drying the mid- to lower troposphere. Events of dry advection from higher latitudes penetrating into the equatorial region of the western Pacific during the COARE IOP have been clearly observed (Numaguti et al. 1995; Sheu and Liu 1995; Mapes and Zuidema 1996).

The variability of tropospheric water vapor, therefore, cannot be taken only as a consequence of deep convective activity. Deep convection may be affected by
the water vapor variability induced by the large-scale circulation. Entrainment of environmental dry air into clouds as a mechanism controlling deep convection has been discussed by many (e.g., recently, Kain and Fritsch 1990; Brown and Zhang 1997; Lucas 1997). This study has focused on the effect of the water vapor variability on the clear-sky IR cooling, which modifies the atmospheric instability for deep convection.

The results of the present study lead to two concluding statements.

- The water vapor variability in the troposphere may influence the large-scale variability of deep convection in the Tropics through changing the profile of the clear-sky IR cooling rate.
- The roles of water vapor and IR cooling in the large-scale variability of deep convection can be accurately assessed only when effects of the large-scale circulation are fully taken into consideration.

These two statements are based on two assumptions and three observations. The assumptions are that the TOGA COARE sounding analyses and the radiation model calculations are reasonably accurate and that the observations from the limited region of the COARE IFA reflect the physics of large-scale motions in the equatorial western Pacific. The observations are 1) deep convection is lethargic when IR cooling exhibits vertical structures of ineffective destabilization or even stabilization; 2) these structures of the IR cooling rate result from substantial drying at certain levels of the troposphere; and 3) the drying cannot simply be explained by the large-scale substance in the absence of deep convection; it is caused to a large degree by large-scale horizontal advection.

The idea that water vapor variability above the boundary layer may influence the large-scale behavior of deep convection, through either changing the IR cooling profile or dry-air entrainment into clouds, is complementary to the conventional wisdom that convection is controlled by conditions in the boundary layer. Quantitatively how much the water vapor–IR cooling process modulates the large-scale variability of tropical deep convection, such as that associated with the MJO, has yet to be further explored.

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<table>
<thead>
<tr>
<th>Measurement range</th>
<th>0%–100%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Accuracy at 20°C</td>
<td>±1%</td>
</tr>
<tr>
<td></td>
<td>±2% (0%–90%)</td>
</tr>
<tr>
<td></td>
<td>±3% (90%–100%)</td>
</tr>
<tr>
<td>Temperature dependence</td>
<td>±0.04 °C</td>
</tr>
<tr>
<td>Response time at 20°C and 90% RH</td>
<td>1 s</td>
</tr>
<tr>
<td>Operating temperature</td>
<td>−40°C–60°C</td>
</tr>
</tbody>
</table>

APPENDIX

Errors in Upper-Tropospheric Humidity Measurements of TOGA COARE

The reliability of the upper-tropospheric humidity measurements of the TOGA COARE soundings is a critical issue because some of the most interesting features in the water vapor variability discussed in this study (e.g., its bimodal distribution) are found above the 400-hPa level. Traditionally, U.S. weather stations do not report humidity measurements if temperature (T) is less than −40°C, and measurements of relative humidity (RH) less than 20% are regarded as unreliable and discarded. Gutzler (1993) has shown that uncertainty in calculated radiance due to such uncertainty in humidity measurement is comparable in magnitude to radiative fluctuations due to water vapor response of the tropical atmosphere to doubling CO₂. To what degree the water vapor variability in the upper troposphere discussed in this study is contaminated by instrumental errors is a serious and difficult problem. The bimodality of water vapor distribution is independent of the sampling procedure (Brown and Zhang 1997) and exists in a subset of soundings collected from sites that deployed the Integrated Sounding System (ISS). We now provide technical information of the ISS humidity sensor to facilitate readers’ own judgments on the possibility of contamination in our analyses by ISS instrumental errors.

The ISS radiosonde system uses the Vaisala RS80-15N radiosonde consisting of a Vaisala HMP35C temperature and humidity probe. The humidity sensor is a Vaisala “humicap” capacity RH sensor. Relevant technical specifications of this sensor are listed in Table A1. The traditional practice of the U.S. Weather Service to discard measurements of RH < 20% should probably not be applied to data collected using the Vaisala radiosonde. Two potential error sources, however, exist. The first is the slower response time of the humidity sensor at low temperatures. Vaisala estimates

2 Technical information provided here is from technical reports of NCAR (Cole 1993) and Vaisala Inc. (Vaisala 1990), Internet ISS documentation from NCAR ATD (http://atd.ucar.edu/ssf/facilities/ssf_facility_descrip/ISS_site.html/), and discussion with Vaisala engineers (W. G. Richards and K. Goss 1996, personal communication).
that at $T < -40^\circ$C (normally above the 250-hPa level in the equatorial western Pacific) the response can be as slow as 30–60 s. This would lead to overestimates or underestimates of RH depending on whether RH decreases or increases with height. A false constant RH reading may result from an extremely slow response above the 200-hPa level (Sherwood 1995). But this is unlikely to be a main problem here because the large water vapor variability and its bimodality exist at altitudes where $T > -40^\circ$C (e.g., 400–300 hPa). There is high correlation between specific humidity above and below the $-40^\circ$C level. Another potential error source is deposition of liquid water on the sensor, which leads to RH readings $>100\%$. This tends to occur when a radiosonde passes through clouds. When supercooled water droplets exist in the cloud, ice can form on the surface of the sensor. Icing on the sensor insulates the sensor from the environment and results in a constant RH reading with height. Ice may fail to melt even after the radiosonde comes out of the cloud because of the low temperatures. Soundings with such errors are easy to detect by their relatively constant RH values in the upper troposphere and their impossibly high RH values in the lower stratosphere. Lin and Johnson (1996a) have reported the icing problem in the TOGA COARE soundings. In the present study, the icing problem cannot explain the relatively low probability distribution of humidity that separates the moist and dry modes and makes the humidity distribution bimodal.

Other errors in the TOGA COARE soundings are those associated with solar radiative heating of the instrument, humidity response lag because of the protective cap, and radiosondes launched from balloon launchers that had been stored inside air-conditioned shelters. These errors existed mainly at the lowest levels of the troposphere and were partially corrected by NCAR ATD ers that had been stored inside air-conditioned shelters.

Finally, should the observed bimodality in water vapor be indeed a result of instrumental errors, it would indicate an identifiable error source.

REFERENCES


