The Response of the Tropical Western Pacific Region to a Dry Intrusion from the Perspective of Observations, A Cloud Resolving Model and a Single-Column Model

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Introduction

The appearance of extremely dry air over the Tropical Western Pacific (TWP) has received a great deal of attention in recent years (e.g., Parsons et al. 1994; Numaguti et al. 1995; Yoneyama and Fujitani 1995; Mapes and Zuidema 1996; Johnson et al. 1996; Sheu and Liu 1995; DeMott and Rutledge 1998; Yoneyama and Parsons 1999; Parsons et al. 1999). These dry air events are termed dry intrusions or dry tongues, since the dry air originates aloft at higher latitudes and subsides into the tropics in long filaments, several hundred kilometers in width. Recently, Yoneyama and Parsons (1999) showed that these events are associated with the remnant circulations of middle latitude baroclinic waves.

Current thinking is that these dry intrusions may be relatively common over the TWP. Yoneyama and Parsons found that dry events were present over the region approximately 9% to 17% of the time during the recent Tropical Ocean Global Atmosphere-Coupled Ocean Atmosphere Response Experiment (TOGA-COARE). After a dry intrusion arrives, the typical time scale for the tropical atmosphere to recover to moist conditions is 10 to 20 days. The frequency and magnitude of these extreme lateral drying events, together with the long adjustment times found in recent cumulus modeling 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 dry periods is that the tropical atmosphere spends a great deal of time recovering from this lateral forcing. During this recovery process, convective instability increases, 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 associated with warming sea surface temperatures under relatively clear skies and light winds. In contrast, Emanuel and Bister (1996) had earlier hypothesized that a dry middle troposphere will create increased convective instability through enhanced radiational cooling.

In this study, we will discuss the behavior of the cloud and radiation fields and how these fields interplay to recover the tropics toward moist conditions from three perspectives. We will begin by summarizing recent observational work. We will then discuss very preliminary simulations undertaken with a cloud-resolving model (CRM) and a single-column model (SCM) constructed from a general circulation model (GCM). Both simulations were conducted with forcing derived from observations.

Observations: The Impact of Dry Intrusions on Cloud and Radiative Processes

It is well established that these dry intrusions have a pronounced impact on cloud and radiative processes in the tropics (e.g., Numaguti et al. 1995; Yoneyama and Fujitani 1995; Johnson et al. 1996; Mapes and Zuidema 1996; Brown and Zhang 1997; DeMott and Rutledge 1998; Parsons et al. 1999). In this section, we will summarize the current thinking from an observational perspective on the interplay between dry intrusions, clouds, and radiation in the tropics. Some of the discussion of the observations that follows was taken from Parsons et al. (1999). The reader is referred to that study and the references discussed in that paper for more information regarding the observational studies.

The radiative impact of these extremely dry air masses on convective instability over the tropical warm pool was discussed by Mapes and Zuidema (1996). They noted that the typical vertical profile of longwave cooling was greatly disturbed by the presence of the extremely dry air within the middle and lower troposphere. They further proposed that the vertical profile of longwave cooling contributed to a strong thermal inversion at the base of the dry layer. An example of the vertical profile of longwave cooling calculated from a temperature and humidity profile following the arrival of a dry intrusion is shown in Figure 1. More recently, Parsons et al. (1999) suggested that since shortwave absorption nearly balances the longwave cooling at solar noon (Figure 1), there should be a distinct diurnal variation in the magnitude of these strong inversions. The variation in magnitude of this inversion, in turn, impacts the convective stability, specifically the convective inhibition (for a definition see Bluestein and Jain 1985), making the diurnal variation of the clear air radiative forcing, an important factor in controlling the diurnal cycle of stability and convection over the warm pool during these extremely suppressed conditions. Of course, the contribution of the diurnal changes in the radiation term to the diurnal variation in clouds is likely to be relatively unimportant in the more typical moist regimes over the warm pool where the vertical profiles of radiation are less disturbed (e.g., Raymond 1995).

The changes in the cloud layers following the arrival of a dry intrusion over the tropics correspond in an expected way to the predicted changes in the convective stability. For example, the appearance of dry air has 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. The general suppression of deep convection by the intrusions is due to both the inversion at the base of the dry layer and the negative impact of entrainment of dry air (Mapes and Zuidema 1996; DeMott and Rutledge 1998).

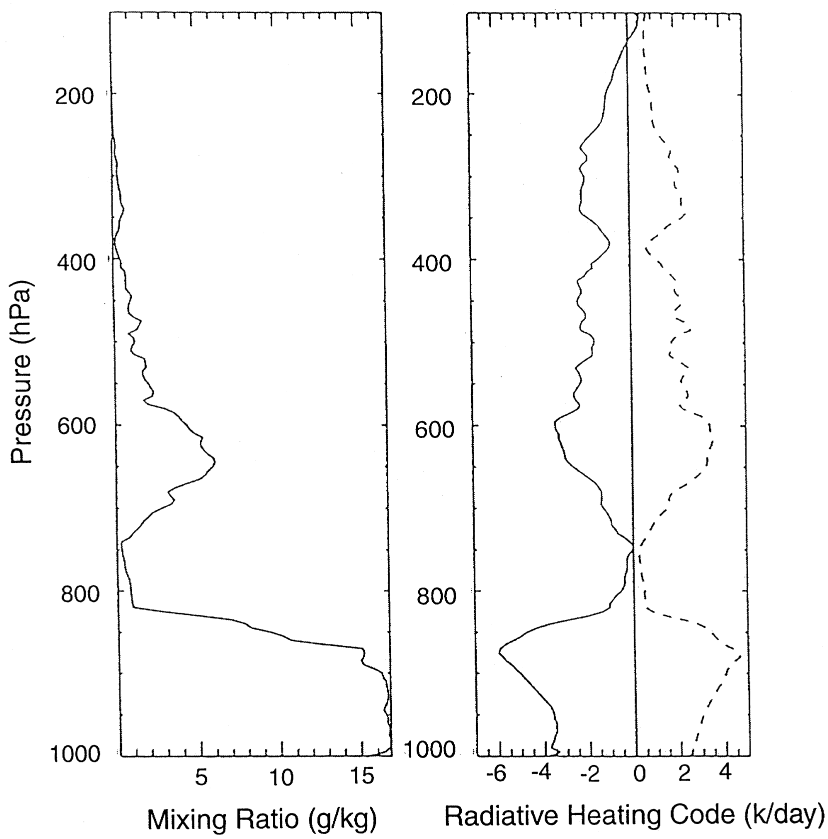


Figure 1. Vertical profile of temperature and specific humidity from a rawinsonde launched at 0440 Universal Time Coordinates (UTC) on November 13 from the Research Vessel (R/V) Shiyan #3. Also shown is the longwave cooling and shortwave heating profiles obtained from the radiation package used by the National Center for Atmospheric Research (NCAR) Community Climate Model Version 3 (CCM3) as used in Mapes and Zuidema (1996). From Parsons et al. (1999).

Following the general suppression of deep convection, past studies also show relatively shallow convection develops within days after the arrival of this dry air. The convection first develops as non-precipitating or lightly precipitating cumulus clouds with echo tops gradually becoming deeper from one day to the next with radar echo tops eventually reaching 3 km to 8 km. An example of this cloud deepening cloud field is shown in Figure 2. This shallow convection slowly erodes any dry layers through detrainment. These relatively shallow clouds tend to have a distinct diurnal cycle with a precipitation maximum in the early evening (Figure 3). This evening maximum has been proposed to be due to diurnal variations in surface fluxes in response to solar heating (i.e., Chen and Houze, Jr. 1997) in combination with the diurnal changes in the vertical profile of radiation discussed earlier. A schematic summary of the recovery process is presented in Figure 4.

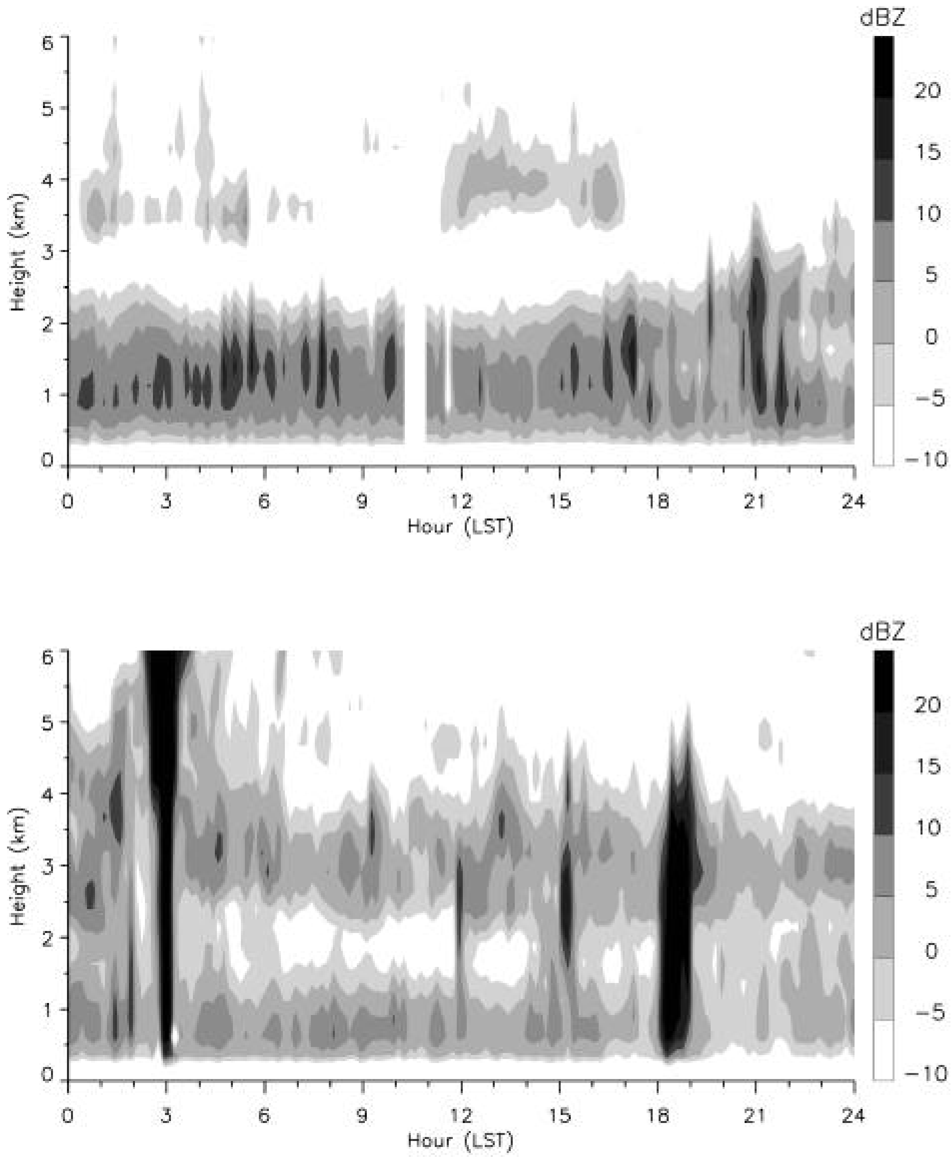


Figure 2. Radar reflectivity from the 915-MHz wind profiler with the Integrated Sounding System (ISS) on board the R/V Kexue #1/ a) November 13 and b) November 15. The time axis is in local time and the highest values are shaded. Note that the data has not been processed to produce a consensus 30-min. average. From Parsons et al. (1999).

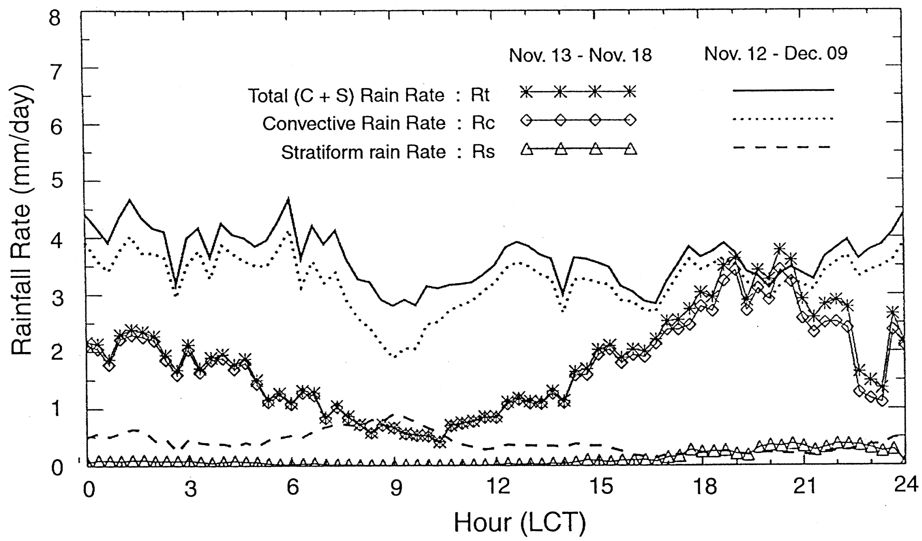


Figure 3. Diurnal variation in rain rates estimated from the Doppler radar aboard the R/V Vickers for the time period of November 12-19 shown in Figure 9. The various curves include total rain rates and rates partitioned into contributes from convective and stratiform rain. The cruise total is also shown in this panel. From Parsons et al. (1999).

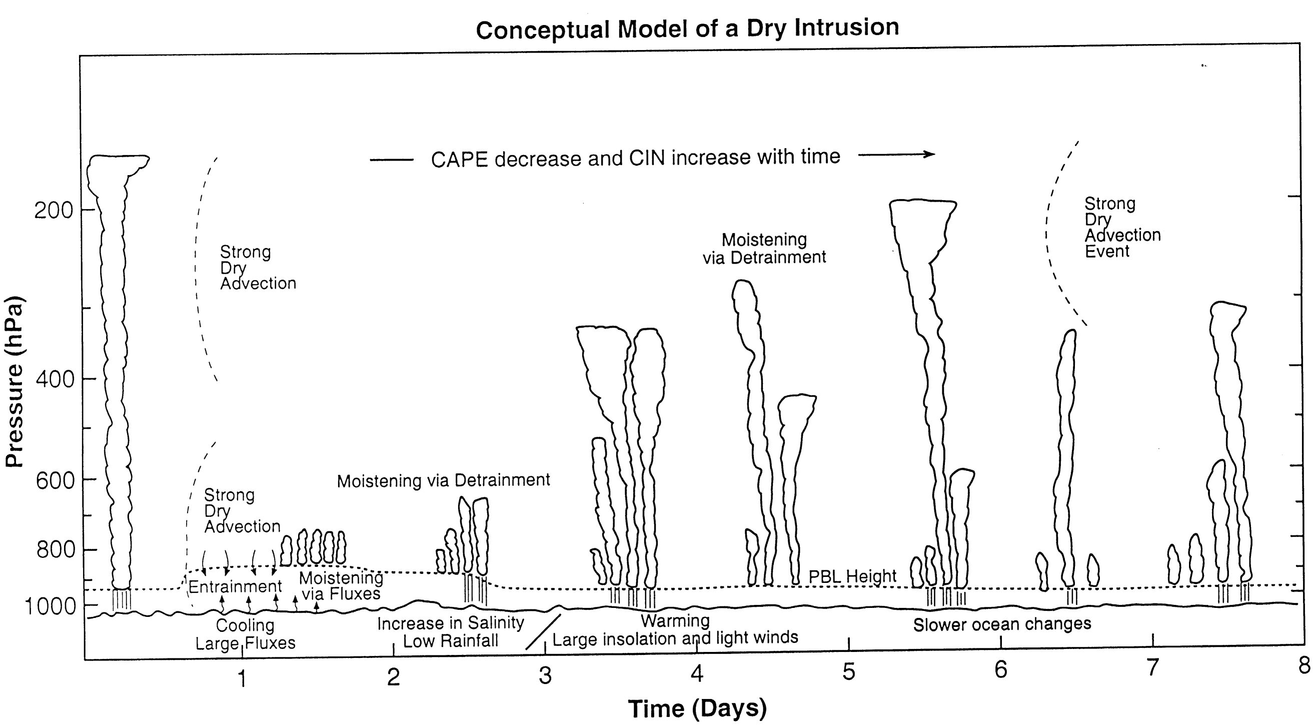


Figure 4. A conceptual model of a dry intrusion moving over the warm pool. From Parsons et al. (1999).

Simulations with a Cloud Resolving Model

The simulations with the CRM were conducted to determine whether this type of model could capture the general evolution of the tropical atmosphere following the arrival of a dry intrusion. One of the dry intrusions for TOGA-COARE, November 13-20, 1982, was selected for this study. This event was focussed on in several of the observational papers discussed in the previous section. The CRM used in this study was the non-hydrostatic CRM of Redelsperger and Sommeria (1986). Ice phase microphysics were treated with the scheme developed by Caniaux et al. (1994). Radiative effects were computed fully interactively with the cloud field using the radiation scheme of the European Centre for Medium-Range Weather Forecasting (ECMWF) (Guichard et al. 1996). The surface fluxes are computed with the COARE bulk algorithm with the sea surface temperature observed on R/V Moana-Wave. Large-scale forcings deduced from COARE observations (Lin and Johnson 1996) are imposed on the model. During this time period, the forcings were generally weak partly due to the absence of large organized systems (see Lin and Johnson 1996). The mean horizontal wind within the model domain was also nudged toward the observed wind. A two-dimensional framework is used with a horizontal resolution of 1 km over a total domain of 100-km width. The vertical resolution is 80 m around in the boundary layer, gradually increasing to greater than 600 m at the top of the troposphere. Numerical tests have shown that the results were not sensitive to both an increase in the horizontal resolution and to the dimensionality of domain (i.e., two-dimensional [2-D] versus three-dimensional [3-D]).

The design of the experiment with the mean wind over the domain being nudged, the sea surface temperature being prescribed by observations, and the large-scale forcing being tied to the observations suggest that one would hope that the simulation reproduced the observations. Even with these high expectations, several aspects of the simulation are noteworthy. For example, Figure 5 shows that the model reproduces the observed fluxes over the period reasonably well. The general intensity and the timing of rainfall (Figure 6) are also reproduced quite well. The model cloud field (Figure 7) reproduces many aspects of the different observed convective stages described earlier (Figure 4): a) non-precipitating shallow boundary layer clouds, b) precipitating shallow convection (more developed than in the boundary layer clouds with cloud heights up to 3 km to 4 km, and c) precipitating deep convection. The model seems to also capture the gradual moistening of the environment due to detrainment from the slowly deepening cloud field.

While the overall evolution of the simulations reproduces many aspects of the observed evolution of the cloud fields, there are, nevertheless, several aspects of the observed conditions that the model is not able to capture. One aspect not reproduced is the first rain event over the first 12 hours on November 14. This rainfall event, however, is difficult to reproduce as it is revealed through satellite and radar to be quite inhomogeneous with more activity in the south of the domain. The explanation for the inhomogeneity is that the dry intrusion arrived with substantial drying in the boundary layer from the south so that the boundary layer recovered to moist conditions first in the southern portion of the domain. The model simulation with uniform conditions was unable to capture this evolution. Another and perhaps more important difference between the observation and the model is that the model appeared to be cool relative to the observations by ~2 K at the end of the simulation. This problem has frequently arisen in simulations over this region driven by observed forcing as found in the GCSS

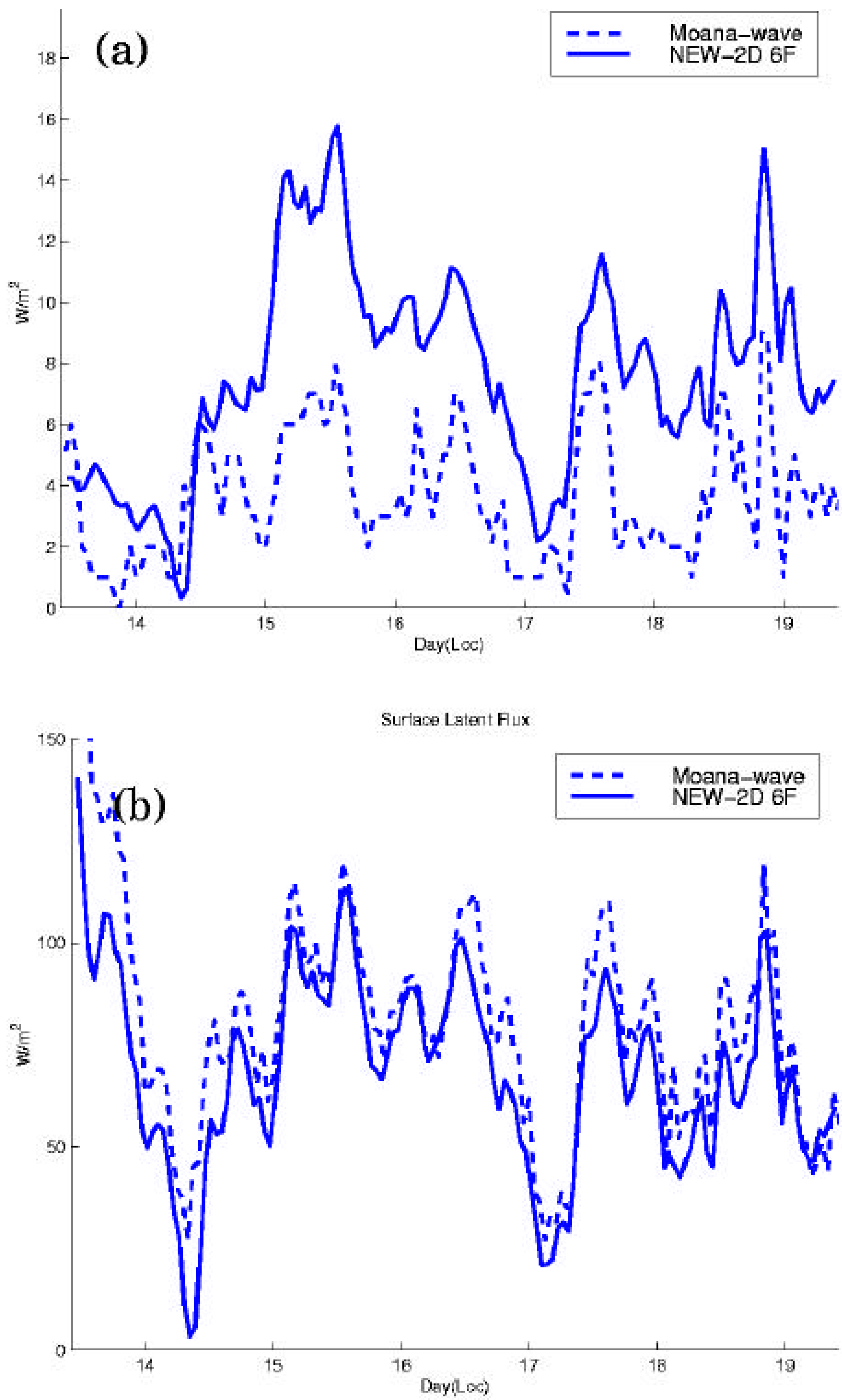


Figure 5. Surface fluxes as observed on board the R/V Moana Wave (dashed lines) and simulated (solid lines): a) sensible heat flux and b) latent heat flux (in W m^{-2}).

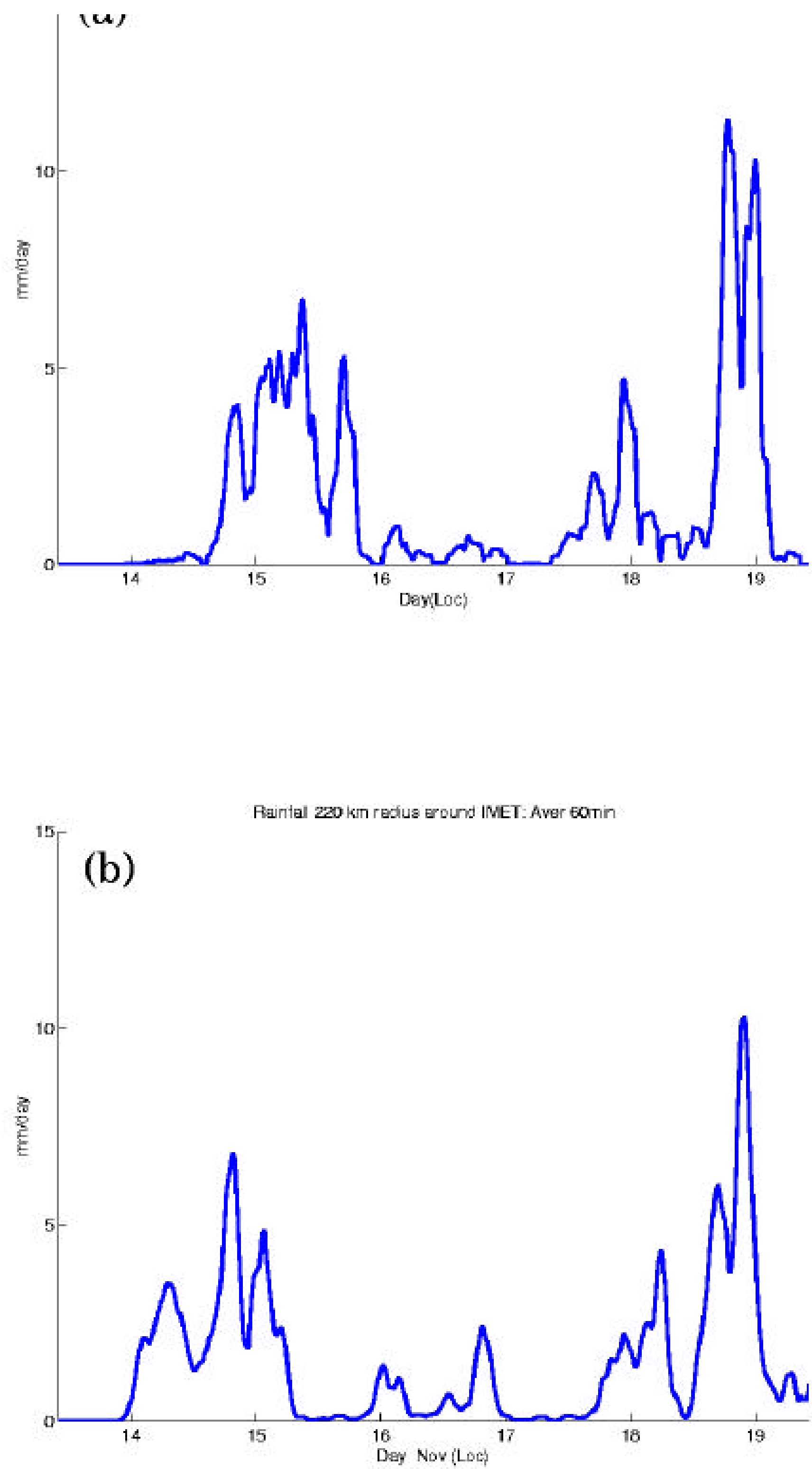


Figure 6. Hourly averaged rainfall as simulated a) and b) estimated from MIT radar over a circle of radius of 220 km around IMET (in mm day^{-1}).

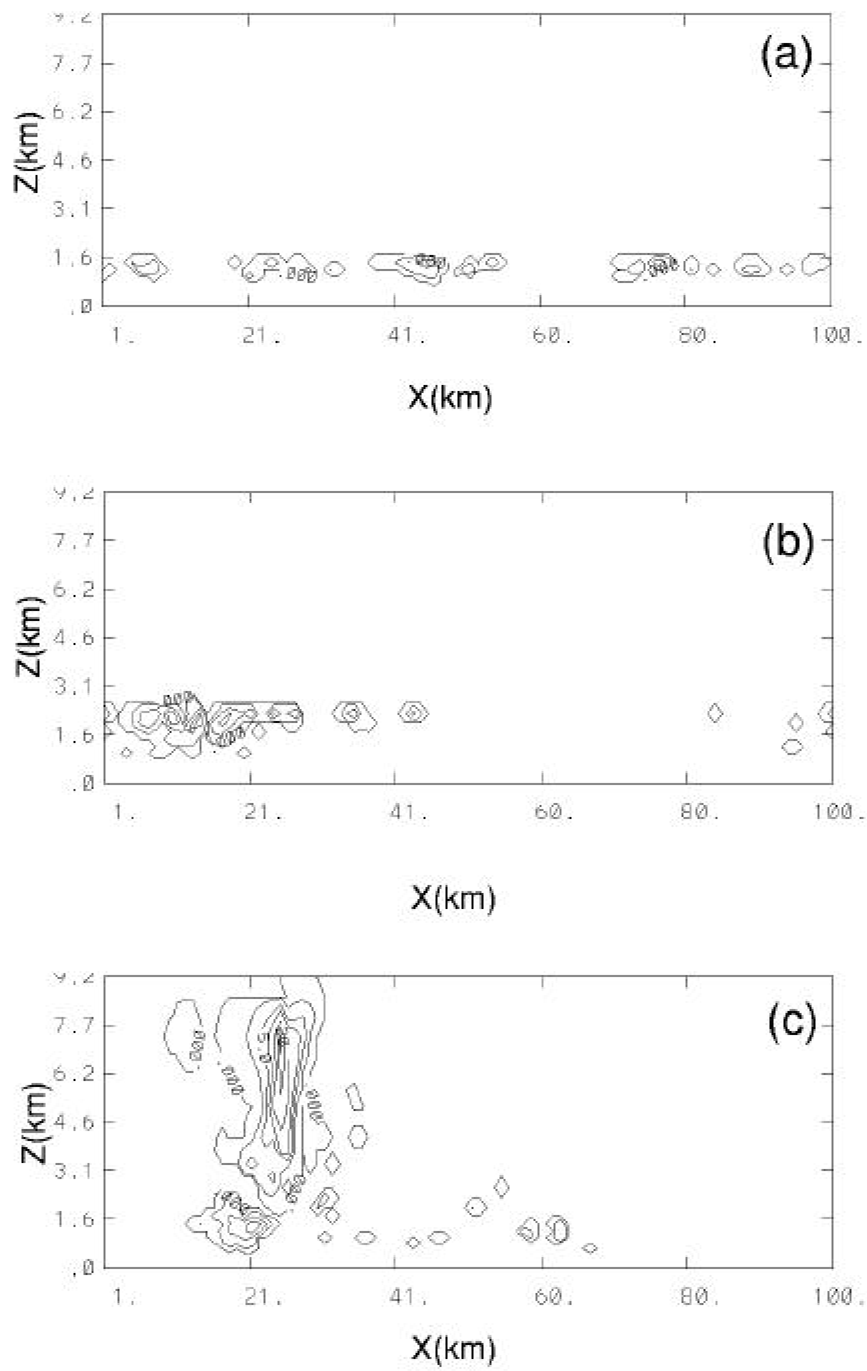


Figure 7. Vertical cross-section of simulated total water content (liquid + solid) at a) November 14 00 h Loc, b) November 15 00 h Loc, and c) November 18, 18 h Loc. The isolines correspond to values of 0.00001, 0.1, 0.5, 1, 2, and 3 g kg^{-1}.

intercomparison as recently described in Su et al. (1999). The cause of this problem is currently under (GEWEX [Global Energy and Water Experiment] Cloud System Study) investigation. In addition to the model bias in temperature at the end of the simulations, there was some tendency for the model to be dry in the lower troposphere. The moisture differences between the radiosonde measurement and the model varied considerably from sounding to sounding, but were generally less than 1 g kg^{-1} to 2 g kg^{-1}. The bias in the model predictions lead to an overestimation of the sensible surface heat flux and an underestimation of the latent heat flux (Figure 6).

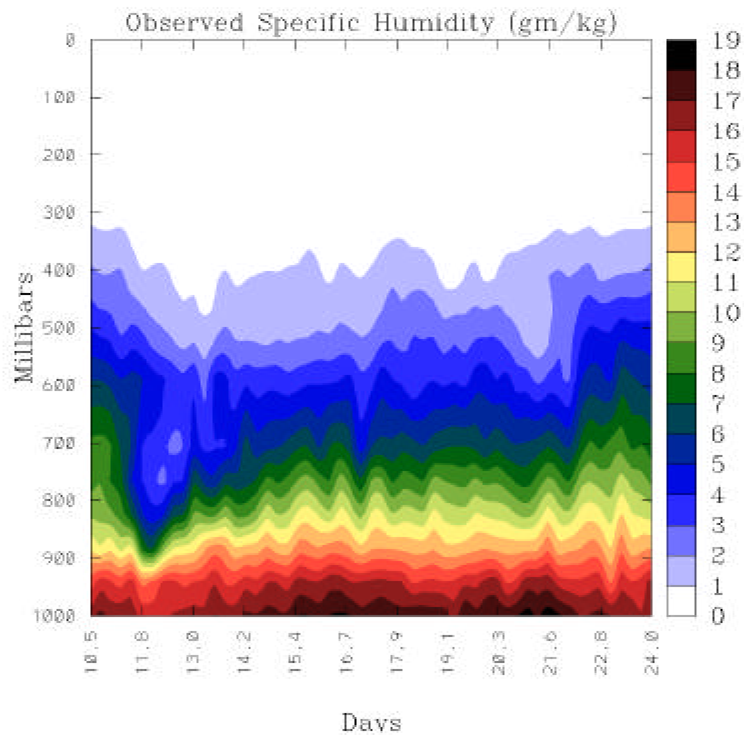
Some Very Preliminary Simulations with a Single-Column Model

In order to test the ability of the physics in a GCM to replicate how the tropical atmosphere recovers to moist conditions following the arrival of an extremely dry air mass from middle latitudes. For this test we used the SCM approach, where a single-column of a GCM is driven with observed forcing. The large-scale forcing of Lin and Johnson (1996) was again used to drive the simulation. The observed sea surface temperatures were also employed. The SCM used is based on the NCAR community climate model (Hack et al. 1998; Hack and Pedretti 1999). A number of different simulations were undertaken using the NCAR SCM. First, we conducted a number of simulations of the dry intrusion simulated with the CRM, but with a number of different starting periods on November 10-13. We used a variety of starting times due to the strong sensitivity of SCMs to the initial conditions (Hack and Pedretti 1999).

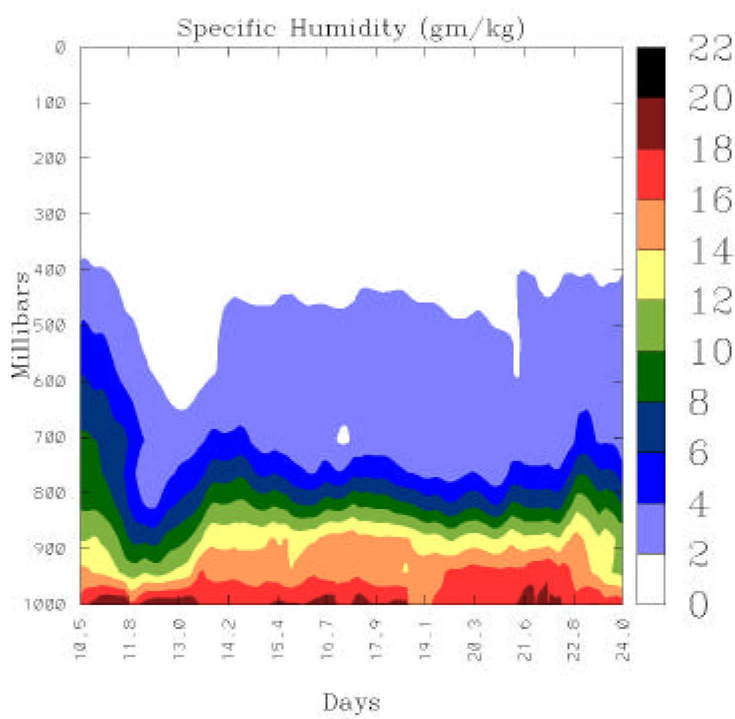
Our SCM results are very preliminary and must be considered with some caution. However, from our results thus far we note several aspects of the SCM solution. The time series of water vapor from November 10-24, 1982, measured and predicted by the SCM are shown in Figure 8a and b. The measured water vapor field shows a strong decrease in the water vapor over most of the troposphere on November 11 followed by a general increase in the water vapor over the next several days in the lower troposphere. At the end of this simulation period, the water vapor content over the bulk of the troposphere has returned to the values observed before the arrival of the dry air mass. The observations also show evidence of a strong diurnal cycle. The SCM simulation does not reveal strong evidence for a diurnal cycle in water vapor (Figure 7b). This lack of a diurnal cycle might be due to shortcomings in the forcing/experimental design or in the model physics. Another shortcoming of the SCM simulation is the lack of a sustained recovery in the moisture field. The recovery, in contrast, in the observations and the cloud model simulation was driven by a slowly deepening field of convective clouds. A preliminary examination of the SCM cloud field reveals that this behavior was not replicated. The result of the difference in the recovery is large errors in the SCM prediction of water vapor (Figure 7c). A large error was also incurred in the SCM prediction of temperature with a layer structure (Figure 8) most likely driven by the interaction of the radiation code with the incorrect water vapor field.

Discussion

From the difference between the CRM and SCM results, we hypothesize that the model physics for convective clouds of shallow and medium depth may need to be improved in the SCM. This thinking is motivated by far better performance of the CRM in reproducing the cloud fields and the general recovery process with similar (but not identical) forcing. Based on these preliminary investigations, we feel that the examination of SCM model performance in the context of a specific synoptic feature allows a framework to be developed for the improved physical interpretation of model performance and an improved understanding of model biases.

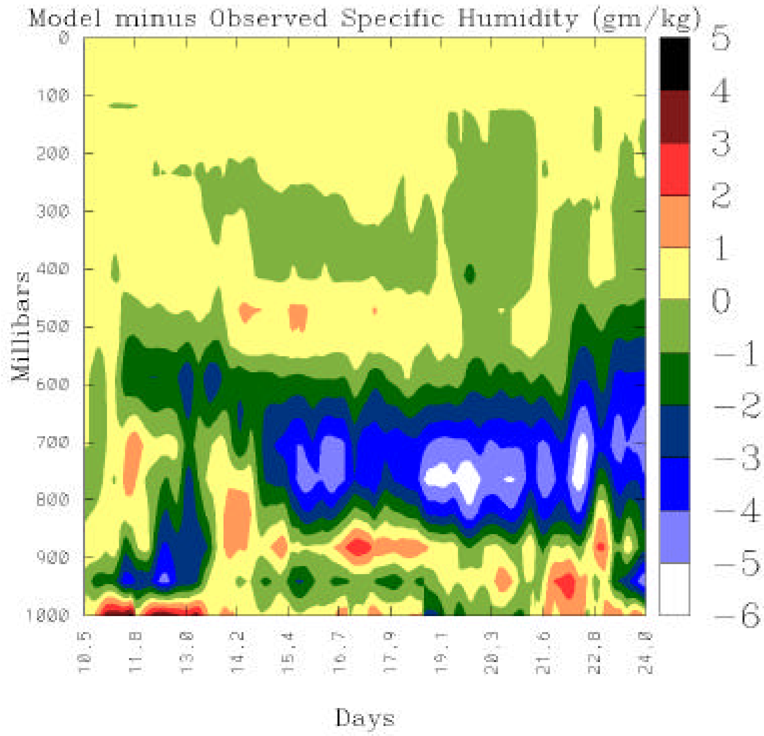


(a)



(b)

Figure 8. a) Observed specific humidity over the center of the TOGA-COARE Intensive Flux Array. The values are color coded and the x-axis is the date in November 1992. b) SCM estimate of the specific humidity. c) The difference field between the observed and model estimates of the specific humidity.



(c)

Figure 8. (contd)

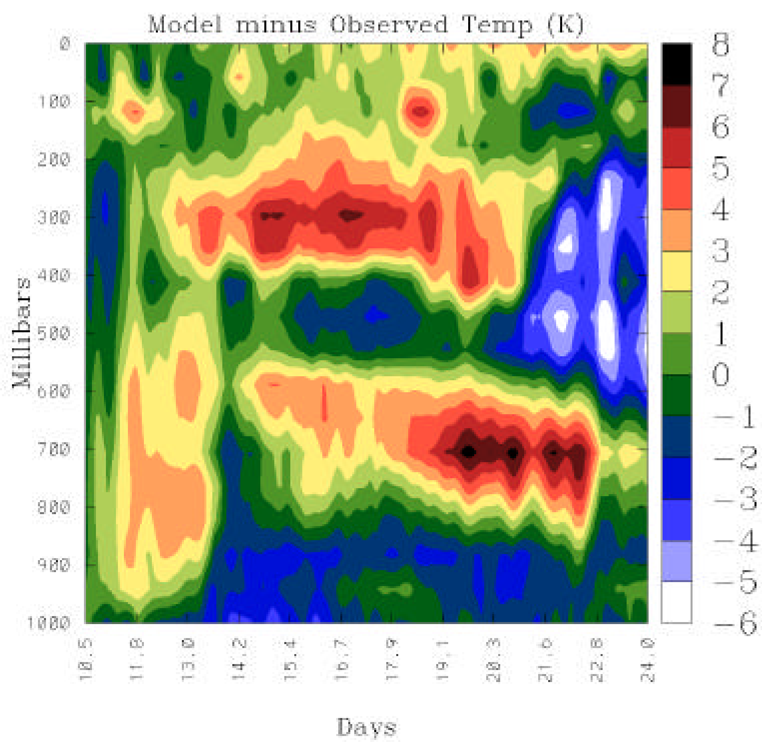


Figure 9. As in Figure 8c, but for the air temperature.

For the CRM, we will focus our future investigations on improving the understanding of the relatively small thermodynamic biases before we undertake a more detailed analysis of the simulations. One point of interest for the CRM simulations is whether the diurnal cycle of the observed cloud field is replicated and whether the vertical gradient of clear air radiative forcing plays a role in controlling atmospheric stability and shallow convection as proposed in Parsons et al. (1999).

The SCM simulations are far more preliminary. As mentioned earlier, the SCM has a strong sensitivity to initial conditions. Although the SCM simulation presented was characteristic of the ensemble of simulations, a more careful examination of the ensemble is clearly necessary. In order to generalize these results, we have also begun to conduct SCM simulations for several of the other dry intrusion periods observed during TOGA-COARE. Our preliminary investigations of other intrusions suggest that the dry bias in the lower troposphere during the recovery process is a repeatable feature. We will also repeat the SCM simulations using the newest version of the SCM and with forcing that has been derived using rawinsonde data that have been corrected for a humidity bias (e.g., Guichard et al 1999a, b; Miller et al. 1999).

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