Preliminary Analysis of the ARM-TOCS Cruise

S. A. Cohn, W.O.J. Brown, and D. B. Parsons
National Center for Atmospheric Research
Atmospheric Technology Division
Boulder, Colorado

Introduction

The Tropical Ocean Climate Study (TOCS) cruise took place January and February 1997 in the Tropical Western Pacific (TWP) region. Figure 1 shows the research vessel (R/V) Kaiyo used for this study. One goal of the study was to investigate the diurnal cycle of clouds, radiation, and convection. The diurnal cycle of clouds and convection has received a great deal of attention with recent papers by Randall et al. (1991), Hendon and Woodbury (1993), Guichard et al. (1996), Chen and Houze (1997), and Sui et al. (1997). One conclusion from these studies is that there are two distinct modes of convection over the TWP. One with relatively shallow clouds with tops of 3 km to 7 km that have a late afternoon maximum and the other with deep convective clouds having an early morning maximum. Despite the attention focused on these modes, there are few observational studies with measurements taken above the surface with sufficient temporal resolution to resolve the diurnal cycle in atmospheric stability.

Figure 1. The Japanese Marine Science and Technology Center (JAMSTEC) R/V Kaiyo, platform for the TOCS cruise.

Measurements taken during the cruise included rawinsonde ascents at 3-hour intervals, surface radiation, and other basic sea-level meteorological measurements, backscatter from a very high resolution lidar (SABL), detailed oceanographic measurements, and wind and backscatter from a 915-MHz radar wind profiler (RWP) with Radio Acoustic Sounding System (RASS).

We focus on diurnal features of the boundary layer (BL), clouds, and atmospheric stability for suppressed conditions of February 2-9, 1997. This was a suppressed period between strong westerly wind bursts. Our conclusions are specific to this calm period, and must be treated cautiously because they include only 8 days. Measurements from the Nauru99 study conducted in the summer 1999 will provide more data for comparison.

Figure 2 shows sea surface temperature (SST) anomalies and the zonal wind during the first half of 1997 at 156°E longitude. During the suppressed period under study, winds were light and the SST was anomalously high.

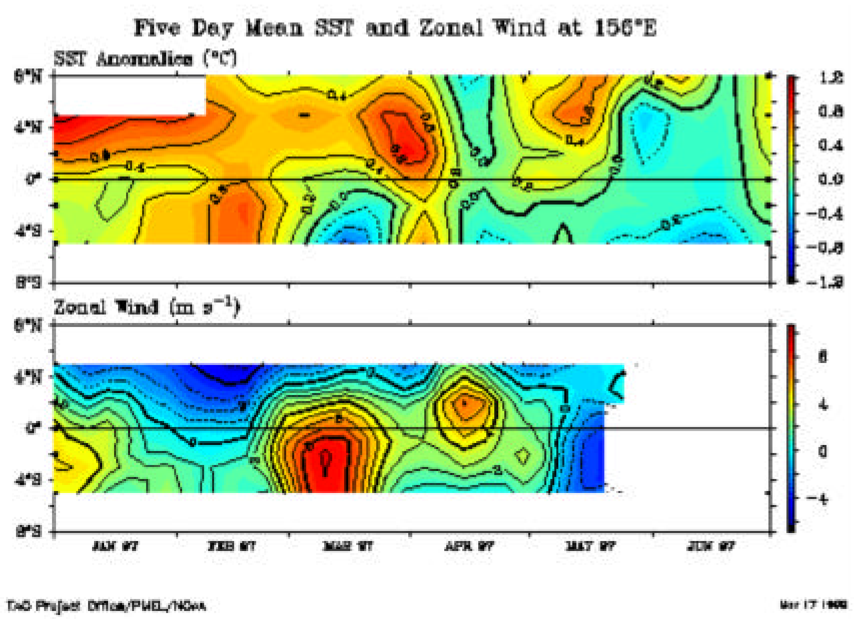

Figure 2. Five-day mean SST anomalies (top) and zonal wind (bottom) at 156 E. longitude during the first half of 1997. The suppressed period under study is early February.

The Tropical Boundary Layer

SABL backscatter and RWP winds and backscatter provide a unique look at the tropical atmosphere. SABL measures aerosol and cloud backscatter profiles with typically 1-second and 3.75-m resolution up to about 5.5 km. Figure 3 shows the complex features present in the BL even during a suppressed day. The BL is characterized by strong aerosol backscatter and its top by strong decreasing backscatter (strong gradient) and small clouds (very large backscatter). However, there is some air with moderate aerosol concentrations above the BL, possibly a "buffer" layer as described by Russell et al. 1998.

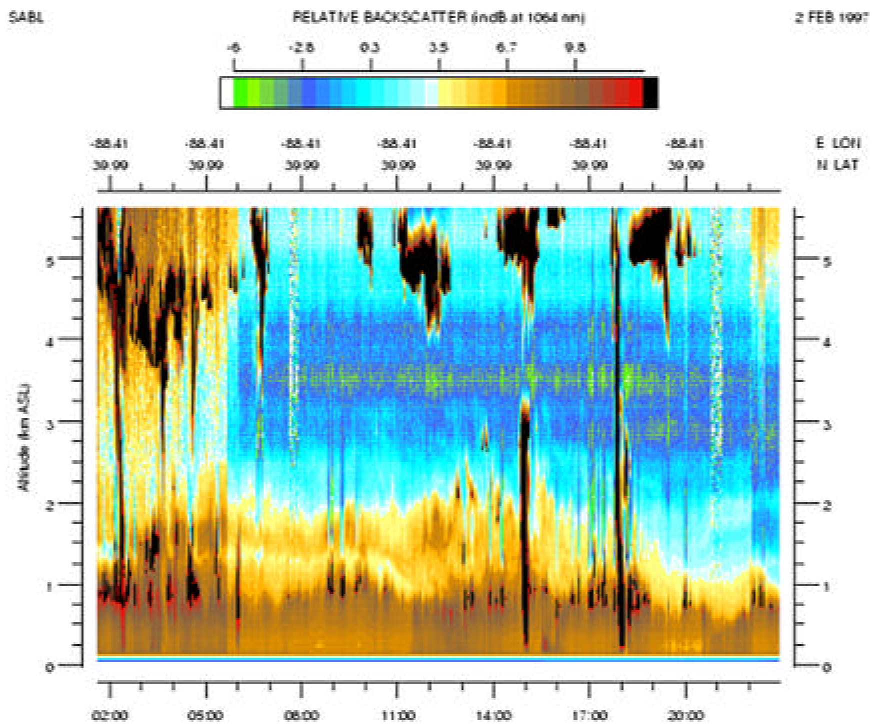

Figure 3. Backscatter intensity from the SABL lidar on February 2, 1997. See text for description of backscatter.

There are also large clouds aloft, as well as clouds beyond the 5.5-km range of SABL. Figure 4 shows the winds available from the RWP with 30-minute and 60-m vertical resolution, and signal-to-noise ratio (SNR) with resolution of several minutes.

Determination of Boundary Layer Height and Cloud Height

As reported at the 1998 Atmospheric Radiation Measurements (ARM) Program Science Team Meeting, Cohn et al (1998) used a wavelet transform technique to find BL height from SABL backscatter. This technique was previously applied to continental convective BL data. The BL height found when this technique is applied to the data in Figure 3 is shown in Figure 5 at 1-minute resolution. The inversion height is also visible from balloon soundings taken every 3 hours (Figure 6), and this corresponds well with the higher resolution lidar result.

We have also calculated cloud hits using the lidar data. This is done using a simple thresholding technique developed for the TOCS data. From the lidar data alone it is unclear if some strong backscatter in the lowest 1 km is from cloud or strong aerosol concentration from an updraft, but photographs during the cruise (Figure 7) show low-level clouds at corresponding times. Further work is needed to refine the cloud detection algorithm and investigate a diurnal cycle in cloud cover.

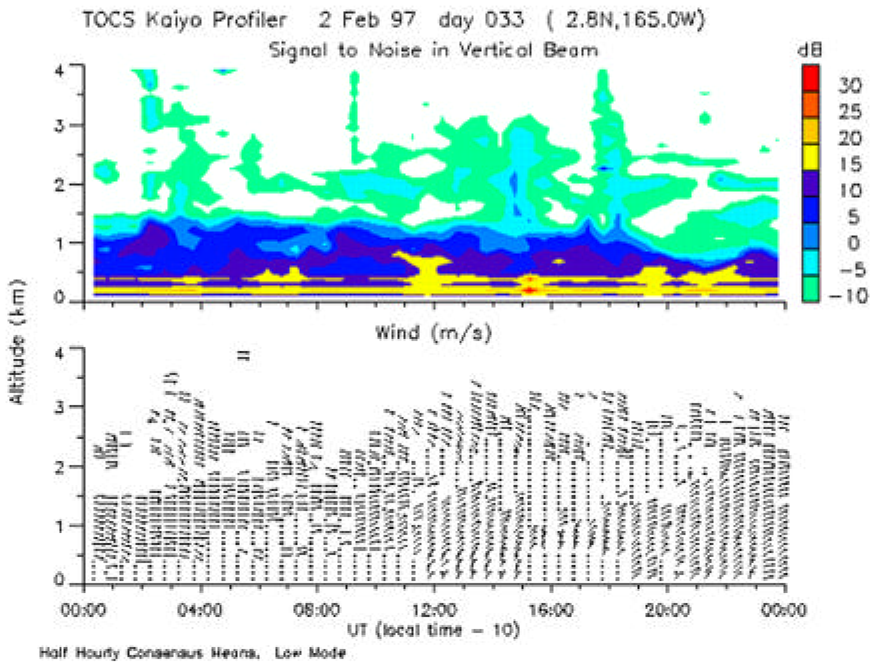

Figure 4. Radar wind profiler signal-to-noise ratio (top) and horizontal wind field (bottom) on February 2, 1997.

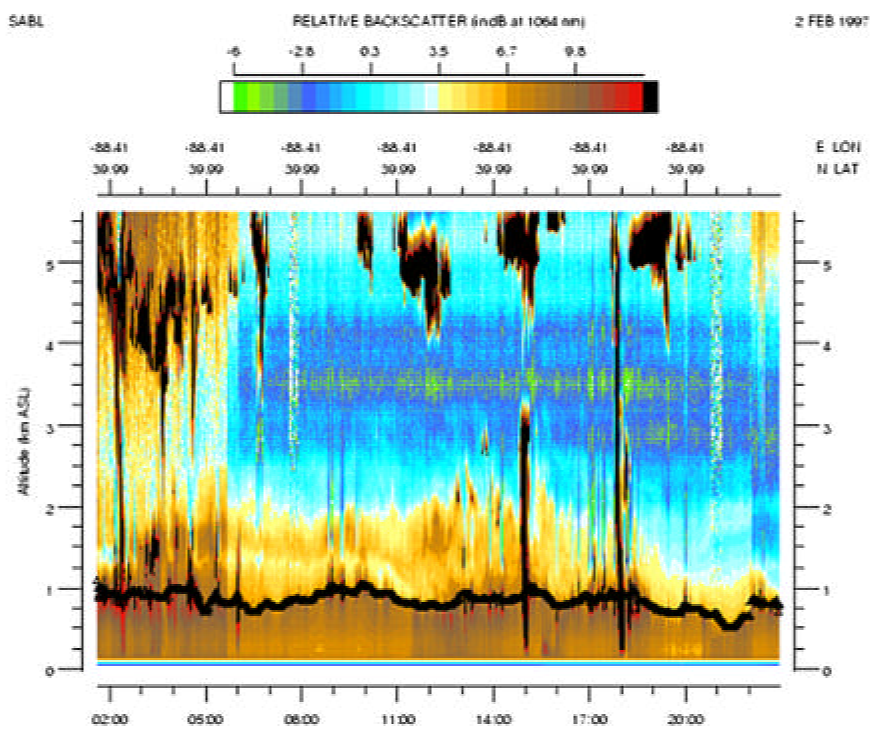

Figure 5. Same as Figure 3, but with boundary layer height determined with the wavelet algorithm overplotted at 1-minute resolution.

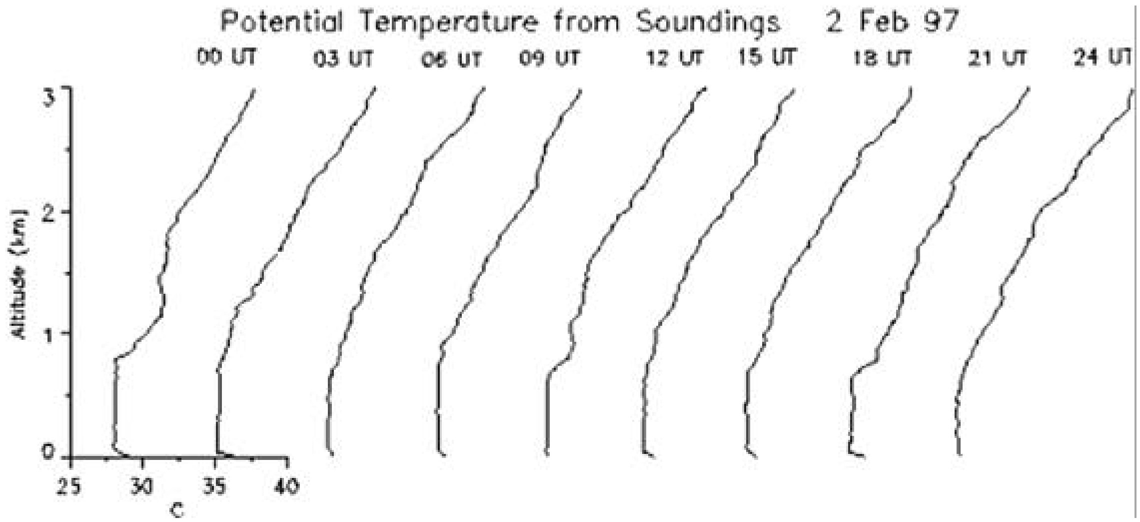

Figure 6. Potential temperature measured with 3-hourly rawinsondes on February 3, 1997. Mixed layer depth corresponds well with the BL height of Figure 5.

Figure 7. Photograph from the R/V Kaiyo of boundary layer clouds from February 2, 1997.

Diurnal Features

We have calculated the BL height over time for each of 8 days during the suppressed conditions (February 2-9, 1997). Figure 8 shows this data (upper) and the median of all 8 days (lower). The median height for these days is just below 800 m. There is a clear signature visible in this median plot with a maximum around 7 p.m. and the boundary layer height decreasing until nearly 7 a.m. local time. The diurnal changes in the median boundary layer depth are about 120 m. A deeper boundary layer in the late afternoon with boundary layer heights decreasing during the night might be expected from the cycle of diurnal heating. Superimposed on this cycle are other changes during the day with a local maximum near noon and local minimums near 10 a.m. and 4 p.m.

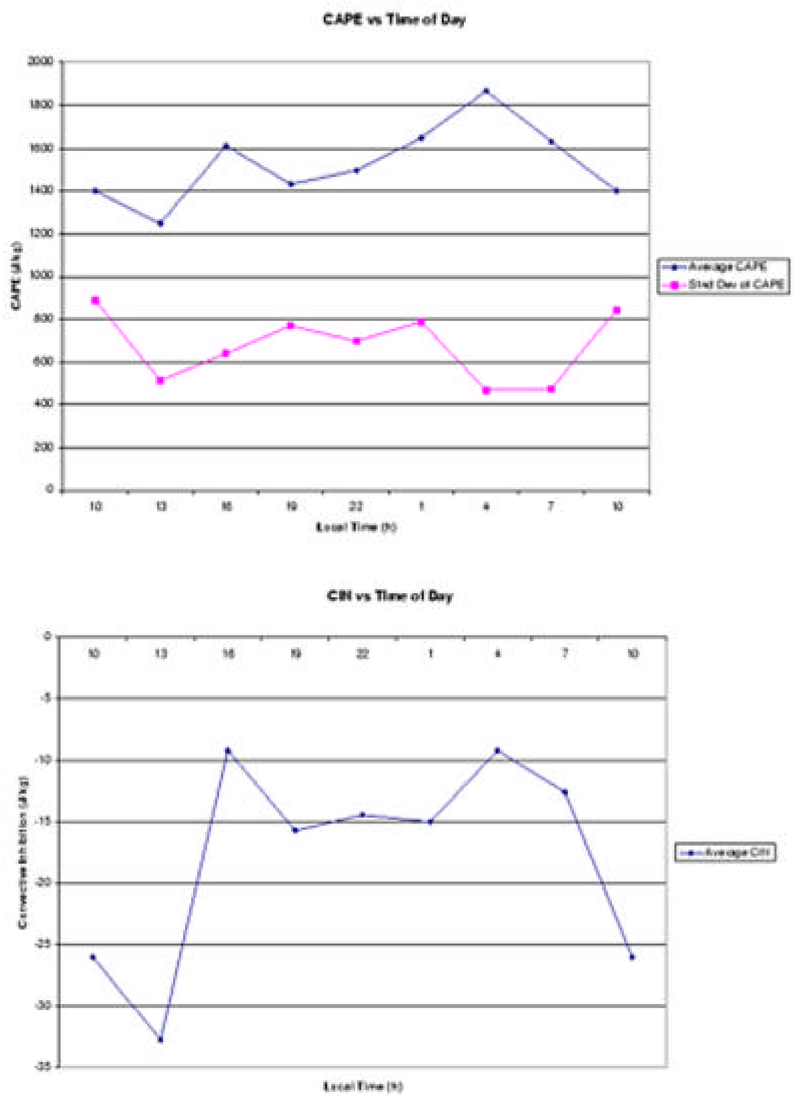

Figure 8. BL height as a function of hour for February 2-9, 1997, (top) and the median of these 8 days (bottom).

We have calculated convective available potential energy (CAPE) and convective inhibition (CIN) from the 3-hour soundings. The median of data on these 8 days (Figure 9) shows that the two times where the atmosphere is most favorable for deep convection are late afternoon (4 p.m.) and early morning (4 a.m.) as there is high CAPE and diminishing CIN. There are also hints of shorter-term variations during the day with the morning and early afternoon as less favorable times. The growth in CIN during the morning suggests the presence of subsidence, which has been predicted to occur based on thermodynamic considerations.

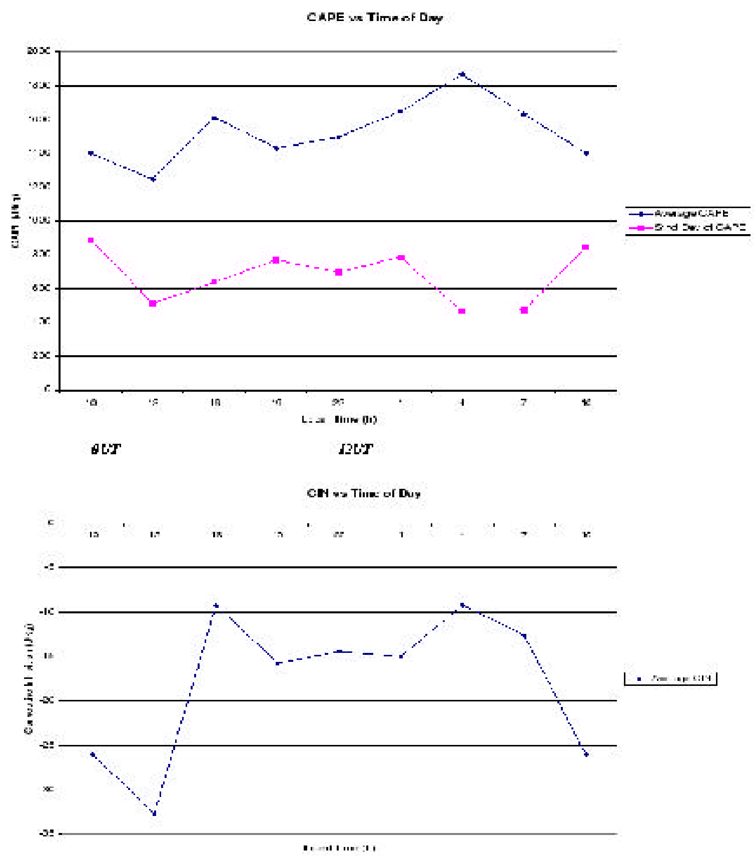

Figure 9. Diurnal variation of CAPE (top) and CIN (bottom) for the same 8-day period.

Conclusions and Future Work

We are currently working to extract a diurnal cycle in low clouds and the radiation cycle. We are optimistic that we can use this data together with other periods from TOCS and Nauru99 to diagnose the mechanisms for the diurnal cycle in clouds over this region. Some aspects of our results are surprising in that we did not expect to see such a strong peak in CAPE at 4 a.m. in the sounding data during these suppressed conditions. This raises the possibility that the increase in precipitation is not due to such factors as enhancement of anvil rainfall by cloud versus cloud-free radiation differences but due to real increases in atmospheric instability.

Acknowledgments

This work was supported by the U.S. Department of Energy's ARM Program. NCAR is operated by the University Corporation for Atmospheric Research under sponsorship of the National Science Foundation.

References

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

Cohn, S. A., C. L. Holloway, and S. D. Mayor, 1998: Boundary layer height determination with the LIFT dataset. In *Proceedings of the Eighth Atmospheric Radiation Measurement (ARM) Science Team Meeting*, DOE/ER-0738, pp. 167-171. U.S. Department of Energy, Washington, D.C.

Guichard, F., J.-L. Redelsperger, and J.-P. Lafore, 1996: The behaviour of a cloud ensemble in response to external forcings. *Quart. J. Royal Meteor. Soc*., **122**, 1043-1073.

Hendon, H. H., and K. Woodbury, 1993: The diurnal cycle of tropical convection. *J. Geophys. Res*., **98**, 16,623-16,637.

Randall, D. A., Harshvardhan, and D. A. Dazlich, 1991: Diurnal variability of the hydrologic cycle in a general circulation model. *J. Atmos. Sci*., **48**, 40-62.

Russell, L. M., D. H. Lenschow, K. K. Laursen, P. B. Krummel, S. T. Siems, A. R. Bandy, D. C. Thornton, and T. S. Bates, 1998: Bidirectional mixing in an ACE 1 boundary layer overlain by a second turbulent layer. *J. Geophys. Res*., **103**, 16,411-16,432.

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