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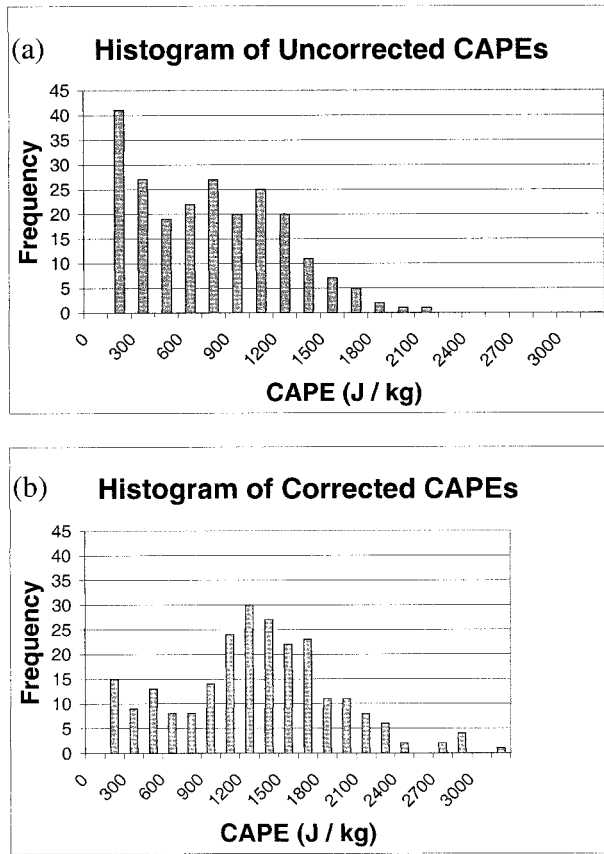


FIG. 4. Histogram of (a) uncorrected and (b) corrected CAPEs.

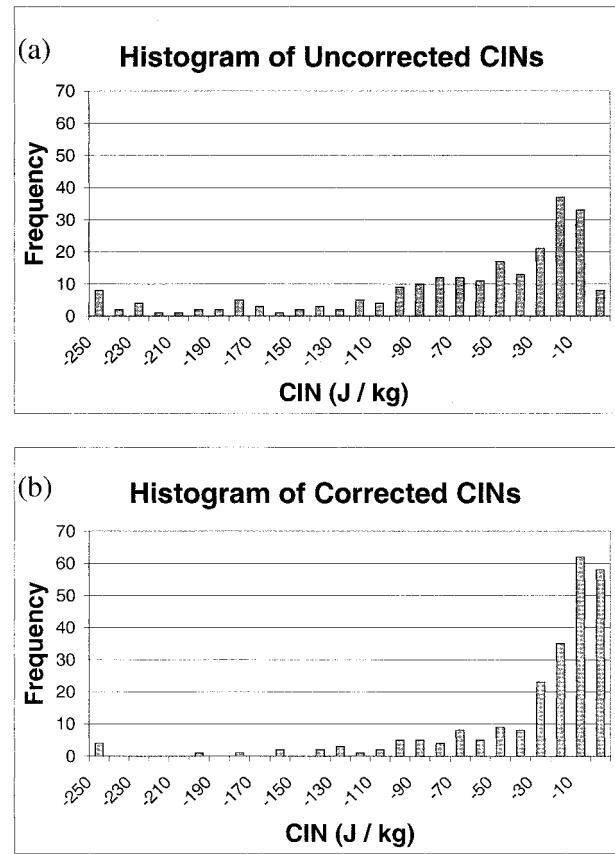


FIG. 5. Same as Fig. 4 but for CINs.

primary difference between the two datasets is that the uncorrected values are far more stable, with more CIN and less CAPE than the corrected counterparts. This general finding is consistent with the results of Zipser and Johnson (1999) and LZ00. The CAPE and CIN values of LZ00 (see their Table 7) lie between our corrected and uncorrected values, but somewhat closer to our corrected values. We suspect that this difference is primarily due to less moisture being added in the boundary layer in LZ00 than with our procedure.

From the mean uncorrected values of CAPE and CIN one would qualitatively interpret the atmosphere as becoming increasingly stable to deep convection over time as the CAPEs generally decrease and the CINs increase during the measurement period. The stability implied by the average uncorrected CIN is particularly noteworthy, with average values of -60 J kg^{-1} for the first two cruises and of -120 J kg^{-1} for the third cruise. To illustrate the physical meaning of these magnitudes, we will assume that the kinetic energy required for parcels in the mixed layer to overcome the observed CIN is supplied by vertical motions. With this assumption, unrealistically large updrafts of 11 m s^{-1} and nearly 15.5 m s^{-1} would be required to initiate or maintain convection during the two first cruises and third cruise, respectively.

Our interpretation of the analyses of the radar datasets presented in the literature (e.g., Short et al. 1997; DeMott and Rutledge 1998; Rickenbach and Rutledge 1998) is that there is no plausible physical explanation for these relatively large CINs observed during the third cruise in the uncorrected datasets. In contrast, the CINs for the corrected data show that there is generally far less inhibition to convection and that the extreme CINs observed during the third cruise are greatly reduced. Thus, it is likely that the slow stabilization of the tropical atmosphere found in the uncorrected dataset is not physical but in reality results from a change in the dry bias with time, possibly due to an increase in the contamination as a radiosonde batch ages.

It is instructive to compare the uncorrected and corrected datasets with previous studies that use CAPE to study the stability of the atmosphere over the tropical oceans. LeMone et al. (1998) examined the environment in the vicinity of 20 convective systems observed with airborne measurements and found that the COARE convective environment had a range of CAPEs from 812 to 1925 J kg^{-1} with a mean value of 1471 J kg^{-1} . The CAPEs calculated by LeMone et al. (1998) were generally larger than those found during GATE [GARP (Global Atmospheric Research Project) Atlantic Tropical Experiment]. Using data from GATE, Barnes and

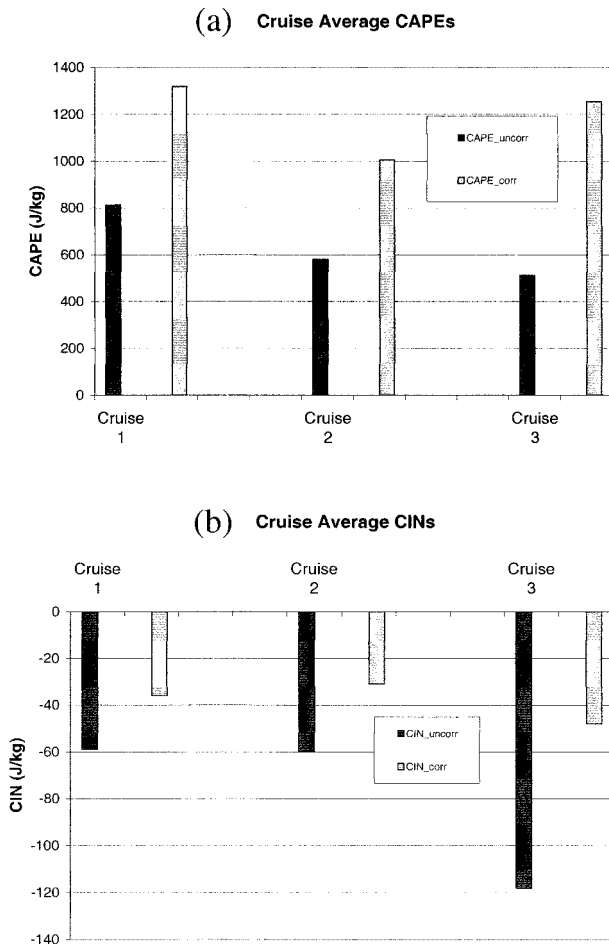


FIG. 6. Cruise average of (a) CAPEs and (b) CINs: uncorrected (black) and corrected (gray).

Sieckman (1984) found the average CAPEs for the so-called fast and slow moving convective lines to be 954 and 1138 J kg^{-1} , respectively. The uncorrected average CAPEs for the R/V *Moana Wave* of between 511 and 812 J kg^{-1} (Fig. 6a) are much smaller than reported by LeMone et al. (1998) and smaller than even the GATE results of Barnes and Sieckman (1984). In contrast, the average corrected CAPEs for the entire R/V *Moana Wave* dataset, which includes convective and suppressed periods, are generally larger than observed for GATE convection. The corrected CAPEs are within the range of the values found by LeMone et al. (1998) but on average smaller. In summary, from the average uncorrected data, one could wrongly infer that the basic state of the atmosphere of the COARE region is an environment where convection would be difficult to initiate and maintain and that the COARE region is more stable than GATE. The corrected dataset is more similar to the aircraft observations.

Several additional inferences can be drawn from the corrected dataset concerning the nature of the atmosphere over the tropical western Pacific. First, we find

that the CAPEs are within or exceed the range found by LeMone et al. (1998) for over three-quarters of the sampled population (Fig. 4b). Thus, the most common situation for the atmosphere over the tropical western Pacific during TOGA COARE appears to be at least as convectively unstable as the atmosphere was found to be in the vicinity of active deep convection. From the corrected CIN in Fig. 5b, we can conclude that the most common situation over the warm pool is to have CINs less than -10 J kg^{-1} so that convection can be easily triggered and maintained. For these smaller CINs, it seems likely that local variations in the sea surface temperature (SST) or in the temperature at the top of the boundary layer will induce regions where CIN vanishes and convection is initiated. If convection is initiated in these common environments of small inhibition, gust fronts can then force parcels to their level of free convection, and convection can be maintained. LeMone et al. (1998) did not find any environments ahead of convective systems where the CIN was stronger than -25 J kg^{-1} . In our dataset, “stable” values of CIN near or stronger than -30 J kg^{-1} were found for only about 25% of the sampled population, again indicating that the atmosphere during COARE is typically near a threshold where convection can be easily initiated or at least maintained. Since the LeMone et al. (1998) study was restricted to environments in the vicinity of deep convection systems, one can conclude that the “typical” environment, as represented by this one site, has similar CAPEs and CINs, as occurs when organized systems are present.

Our finding of a tropical atmosphere over the COARE region typically being near the convective threshold is similar not only to the LeMone et al. (1998) aircraft study, but also confirms the results of Raymond (1995). While our ground-based approach has the disadvantage of relying on a correction procedure, it does provide a larger sample size than employed in those airborne studies. Raymond (1995), for example, used 44 aircraft soundings to conclude that the typical state of the atmosphere over this region is quite close to the threshold for convection. The previously mentioned radar studies also support the thermodynamic interpretation of a convectively active region, as periods without some form of deep convection are relatively rare over the warm pool.

The uncorrected dataset leads one to a different impression of the tropical atmosphere. The differences in CAPE and CIN are not limited to changes in the mean values and the distribution as the time rate of change of these quantities is also impacted. This time variation is used as a closure assumption in convective parameterizations and in attempting to understand the basic behavior of the tropical atmosphere (e.g., Zhang and Chou 1999). Time series of the CAPE and CIN for the third cruise of the R/V *Moana Wave* are shown in Fig. 7. From this figure, it is evident that the short-term variation between the uncorrected and corrected datasets

This is particularly obvious between days 29 and 32, with a CAPE increase for the corrected dataset only, and between days 32 and 35, with a decrease of the CIN being evident only in the uncorrected dataset. Thus for many purposes, previous studies that employ the TOGA COARE sounding dataset without the bias correction should be treated with a degree of caution that varies with how much the calculations and results depend upon the vertical profile of water vapor.

4. Impact of the correction on radiative fluxes

The impact of the RH correction is not limited to atmospheric thermodynamics. Water vapor also plays a very important role in the earth radiation budget (Spencer and Braswell 1997) and in global climate change estimates (Lindzen 1990; Rind 1998). For instance, global climate change involves a modification of the radiative forcing of the order of a few watts per square meter (mainly related to the atmospheric CO₂ increase), but the actual modification of radiative fluxes might be larger, in particular because of a significant water vapor positive feedback (e.g., Raval and Ramanathan 1989). In this section, we investigate the changes in the radiative fluxes that occur as a result of the RH bias correction.

a. Radiative calculation

The radiative calculations were performed with a stand-alone version of the radiation model used in the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM) (Kiehl et al. 1996, 1998). In this model, the longwave (LW) radiation is treated with a broadband technique. The shortwave (SW) radiation is calculated with a delta-Eddington approximation, with 18 spectral intervals (Briegleb 1992). For cloudy columns calculations, the scheme requires information about the cloud fraction, cloud water mixing ratio, and effective radius at each vertical level. Clouds are treated as gray bodies in the LW, with an emissivity parameterized as a function of cloud liquid and/or ice water paths. Cloud SW optical properties follow the parameterization of Slingo (1989), the coefficients for ice being based on Ebert and Curry (1992). The surface albedo follows Briegleb (1992) for the ocean; surface emissivity equals 0.99, and the SST is fixed to 29.5°C.

Among the ensemble of soundings collected on the R/V *Moana Wave*, only those reaching 100 hPa are retained, resulting in 219 soundings being used in this analysis. The soundings were interpolated on a 25 hPa vertical grid between 1000 and 100 hPa for the radiation calculations. Above the 100 hPa level, 6 additional layers were added up to 1 hPa for radiation computations but without any modification of moisture at those levels. The modification of radiative fluxes induced by the correction of water vapor profiles was not very sensitive

to the vertical resolution or to the relative humidity assumptions above 100 hPa.

Calculations were performed assuming either clear sky or cloudy conditions. Temperature and moisture data can be used directly for the computation of radiative fluxes assuming clear sky conditions. However, because of the very large modulation of radiation by clouds over the COARE area, a more realistic picture of the impact of the water vapor correction is obtained when clouds are included in the calculations. Indeed, these calculations show the greatest sensitivity to assumptions related to the cloud field. The main objective of this study, however, is not to get the true cloud fields and radiative properties for each sounding, but rather, a rough estimate of the probable cloud field to assess the likely impact of the error and its correction. Thus, some relatively crude assumptions were used to define a cloud field.

For each sounding, a vertical profile of the cloud field, including vertical profiles of cloud fraction, cloud water content, and effective radius, is diagnosed. Then, this two-dimensional cloud field, a function of time and height, is used for the radiative calculations, with uncorrected and corrected moisture profiles. This framework allows us to isolate the radiative impact of only the changes in the water vapor field for cloudy conditions.

In our calculation with clouds, the cloud fraction was diagnosed as a function of the corrected relative humidity for each layer. The RH with respect to liquid water (to ice) was considered for temperatures greater (less) than 0°C (−23°C). For intermediate or “mixed phase” temperatures, the RH was interpolated in the same way as in the European Centre for Medium-Range Weather Forecasts (ECMWF) model (Simmons et al. 1999). When the relative humidity is larger than a given threshold value, the level is assumed to be cloudy. A threshold value of 90% was used above 800 hPa. Below that height, a slightly smaller value (87%) was adopted. (A comparison between simulated and observed radiative fluxes suggests that a value of the threshold that is constant in the vertical leads to a relative underestimation of low clouds.) The cloud water path (CWP) was expressed as a function of height, as in the cloud parameterization of the NCAR CCM version 2 (Kiehl et al. 1994). For each vertical level it is given by $CWP = \int \rho_0 e^{-z/h_c} dz$, where ρ is the cloud water density and z the height above the surface, ρ_0 is fixed to 0.18 g m^{−3} and h_c is a cloud water scale height expressed as $h_c = A + B \cos^2 \alpha$, with $A = 1080$ m, $B = 2000$ m, and α is the latitude. Finally, the effective radius is fixed up to 10 microns for liquid droplets and varying between 10 and 30 microns for ice crystals (Kiehl et al. 1994).

In the following, we focus on radiative fluxes (noted ϕ) at the TOA and at the surface (SRF) in the LW and the SW range. Upward and downward fluxes will be noted ϕ^\uparrow and ϕ^\downarrow , respectively. Mean values of these fluxes, derived using previous assumptions about the cloud field, are summarized in Table 1. For each vertical

Ciesielski (2000) show that rainfall biases of NCEP and ECMWF reanalysis over the COARE region can be partly explained in terms of the moisture bias of the measurements by radiosondes manufactured by different vendors. In addition, cloud parameterizations in general circulation models usually relate the cloud fraction to the relative humidity. This moisture correction will act to enhance the cloud cover, especially at high altitude, because of the large increase of relative humidity in the upper troposphere. These additional high clouds, in turn, may have a strong impact on the model radiative budget.

The discovery of an error in the measurement of humidity with Vaisala radiosondes led to a significant improvement of the physics of the measurement technique. Similarly, it can be expected that this moisture correction will also contribute to an enhancement of our current understanding of the atmosphere.

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