Is the Tropical Atmosphere Conditionally Unstable?

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ABSTRACT

We examine the mean and standard deviation of the bouyancy of lifted boundary layer parcels from a large sample of soundings from the western equatorial Pacific. The inclusion of condensate loading in the definition of buoyancy is emphasized as is the precise originating level of parcels in the subcloud layer. We confirm the observation of Betts that those parts of the tropical atmosphere experiencing deep convection are nearly neutral to adiabatic parcel ascent from the subcloud layer when adiabatic condensate loading is included in the definition of buoyancy. Parcels lifted from the top of the subcloud layer are more nearly neutral than those originating near the surface, and atmospheres with higher levels of free convection (LFCs) are closer to neutral than those with low LFCs. We explore possible reasons for these observations and discuss their implications for tropical meteorology.

1. Introduction

Cumulus convection is among the most visible and perplexing phenomena in the earth's atmosphere, having a very large influence on the behavior of the atmosphere. Progress toward understanding the physics of cumulus convection has accelerated in recent decades due, in part, to large field experiments such as the GATE.[1] The results of GATE raise more questions than answered (Houze and Betts 1981), illustrating the complexity of cumulus convection.

The importance of cumulus convection in the heat and moisture balance of the tropical trough zone was first discussed by Riehl and Malkus (1958) and reexamined by the same authors (Riehl and Simpson 1979). They concluded that the only way to balance the moisture and heat budgets of all layers is to require that all the air, imported by the trade winds, ascend from cloud base level to the upper troposphere, with the virtually undiluted properties of mixed layer air parcels. In order to explain the apparent undiluted ascent of subcloud air, the buoyant cloud towers were hypothesized to be wide enough for their cores to be protected from entrainment.

In general, parcels from the subcloud layer cannot be displaced upward to their levels of free convection (LFC) without external forcing. Over central North America in spring, for example, a relatively large amount of work is necessary to lift near-surface air to its LFC due to the presence of strong capping inversions. For this reason, relatively large amounts of *Convective Available Potential Energy* (CAPE) can build up without being released. In most of the tropics, on the other hand, the LFC is lower and convection is more spontaneous, as evidenced by the much greater population density of cumulus clouds. In view of this, it is surprising that relatively large amounts of CAPE, as defined for undiluted pseudoadiabatic ascent, exist.

Various attempts have been made to explain this apparent paradox. One of these, dating back nearly 25 years, was based on the relatively small thickness of the subcloud layer: without large-scale "moisture convergence," it was contended, convection could not release the stored energy of the tropical atmosphere. This gave rise to the idea of *Conditional Instability of the Second Kind* (CISK) whereby large-scale and convective-scale circulations cooperate in circulations such as tropical storms (Charney and Eliassen 1964; Ooyama 1964, 1969). In retrospect, it is surprising that moisture convergence by separate circulations was called upon since the response of a stably stratified rotating fluid to an isolated heat source has a horizontal scale on the order of the deformation radius; thus a cloud should have no problem supplying its own moisture.

A more reasonable explanation of the apparently finite CAPE of the deep tropics was proposed by Arakawa and Schubert (1974). This explanation is based on the observation that real cumulus clouds are, in the mean, highly diluted mixtures of mixed-layer and environmental air. The "quasi-equilibrium" postulate of Arakawa and Schubert (1974) holds that the observed

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thermodynamic structure of the tropical convective atmosphere is in fact neutral to a spectrum of entraining, *diluted* cumulus clouds which are always colder than their pseudoadiabatic counterparts.

More recently, Betts (1982) revealed a startling property of the tropical atmosphere; namely, that its virtual temperature is very nearly equal to that of an *undiluted* subcloud layer parcel lifted *reversibly* through the depth of the atmosphere, when all the adiabatic condensed water content is included in the definition of virtual temperature.

Recently, Emanuel (1986) and Rotunno and Emanuel (1987) assumed that the average state of the tropical atmosphere in summer is very nearly neutral to real convection. They hypothesized that tropical cyclones are developed through the interactions between vortex and ocean with cumulus convection rapidly redistributing heat acquired at the oceanic source upward and outward to the upper tropospheric sink. They were able to simulate realistic hurricanes starting from conditionally neutral ambient atmospheres.

The Betts (1982) observation of neutrality to reversible displacements of subcloud layer air is surprising in view of the observed properties of cumulus clouds. The main objective of this study is to systematically examine a large number of soundings from several tropical stations in order to determine how general Betts' observation is. We also explore the sensitivity of parcel buoyancy to the precise originating level of the parcel within the subcloud layer.

In section 2 we briefly discuss the methods of analysis and the datasets. Section 3 examines the maximum possible values of buoyancy available in the tropics. The influence of the vertical inhomogeneity of the boundary layer on the buoyancy profile is examined in section 4. In section 5 we discuss and summarize the results.

2. Methodology

In an emagram, the amount of buoyant energy of a pseudoadiabatically displaced parcel is represented by the area between the environmental temperature sounding and the temperature of the parcel. In such a diagram, a logarithmic scale of pressure is used as a vertical coordinate axis, while temperature varies along another axis. In this study, the buoyancy is instead defined as the virtual temperature difference between cloudy and ambient air. All results, presented in sections 3 and 4, are in log$p$ coordinates, so that the convective energy is clearly indicated.

In a pseudoadiabatic process, a parcel is lifted adiabatically without sustaining any liquid water content. In a reversible moist adiabatic process a parcel is lifted adiabatically while retaining all condensed water. In the latter process, the inclusion of condensate loading increases the effective density of cloudy air. Hence, the virtual temperature of cloudy air is lower for the reversible moist adiabatic process, as seen from the following expressions:

$$T_{v_{c_1}} \equiv T_p \frac{1 + q^*(T_p)/0.622}{1 + Q_T} \qquad (1)$$

for a reversible moist adiabatic process, and

$$T_{v_{c_2}} \equiv T_p \frac{1 + q^*(T_p)/0.622}{1 + q^*(T_p)} \qquad (2)$$

for a pseudoadiabatic process, where T_p is the temperature of a pseudoadiabatically displaced parcel, q^* the saturation mixing ratio, and Q_T the total water content of the parcel. The temperature of cloudy air (T_p) is calculated from the conservation of equivalent potential temperature (θ_e) defined for a pseudoadiabatic process (reversibly-defined T_p differs only by a few tenths of a degree). The empirical formula of equivalent potential temperature (θ_e) given by Bolton (1980) is used in the calculation. The accuracy of Bolton's formula is better than 0.02°C. The virtual temperature of the ambient air is defined by

$$T_{v_a} \equiv T_a \frac{1 + q/0.622}{1 + q}, \qquad (3)$$

where T_a and q are the temperature and the mixing ratio of the ambient air, respectively. For simplicity, B_1 is referred to as the "buoyancy" including adiabatic condensate loading, ($T_{v_{c_1}} - T_{v_a}$), and B_2 to as the "buoyancy" ignoring condensate loading, ($T_{v_{c_2}} - T_{v_a}$).

The analysis procedure is as follows. The dataset for this study consists of rawinsonde data from three stations from 1965 to 1980 during the period 1 July to 30 September. The three stations are Truk, Koror, and Majuro which are located in the central and western equatorial Pacific. There are more than 2300 soundings at each station in this dataset. Only the soundings with thermodynamic data at every 50 mb level between 1000 mb and 300 mb are used. In order to study different types of convective atmospheres, we divide the whole dataset of an *individual station* into several subsets according to the level of free convection. The LFC can roughly be estimated as the level where the saturation equivalent potential temperature (θ_e^*) of the ambient air is the same as the equivalent potential temperature (θ_e) of cloudbase air parcels. Due to the coarse resolution of the data, we divide it into subsets according to 50 mb intervals of the LFC. There are three subsets having average LFCs at 975 mb, 925 mb, and 875 mb, respectively. The reasons for dividing the dataset according to the LFC will be discussed in section 3.

In section 3, we calculate the maximum parcel buoyancy available in the tropics. The parcels are assumed to be lifted from the maximum θ_e level in the subcloud layer. In section 4 we instead consider air parcels lifted adiabatically from near the top of the subcloud layer. Previous studies (e.g., Johnson and Nicholls 1983) assumed that the mixing ratio and dry static energy are well mixed in the "mixed" layer. We note, however, that θ_e near the top of the mixed layer is generally less than θ_e at 1000 mb.

3. Maximum available buoyancy in the tropics

In this section, we treat parcels lifted adiabatically from the maximum θ_e level in the subcloud layer. These parcels are the warmest of all those lifted adiabatically from the subcloud layer. Therefore, they yield the statistically maximum possible values of buoyancy available in convective atmospheres.

First, we estimate the error in calculating buoyancy due to random instrumental errors. The maximum buoyancy error is obtained by assuming that the temperature error increases linearly with pressure from 0.2°C to 0.6°C between 1000 mb and 300 mb, and that the relative humidity error increases linearly with pressure from 1% to 5% in the same region (WMO 1971). The resulting maximum buoyancy error increases (but not linearly) with pressure from 0.6°C to 1.4°C between 1000 mb and 300 mb (Fig. 1). Thus, buoyancy fluctuations with magnitude less than roughly 1.0°C may be due to instrumental noise.

We next look at the difference between buoyancies including (B_1) and ignoring (B_2) adiabatic condensate loading. From the definition of virtual temperatures of cloudy air [Eqs. (1) and (2)], the buoyancy is reduced when including condensate loading effects, because the liquid water increases the effective density of cloudy air. Figure 2 shows the mean and standard deviation of both buoyancies at Truk. Data at all levels are included in the calculation which uses data from 1 July to 30 September, 1965–80. The mean of B_2 is larger than B_1 by 4.1°C at 300 mb, and by 2.4°C at 600 mb where B_1 has its maximum when including condensate loading (Fig. 2). Also B_2 steadily increases ~~with height~~ even above 600 mb.

The standard deviations of B_1 and B_2 are comparable. The largest standard deviations appear at 300 mb with magnitudes of 2.2°C and 2.4°C for B_1 and B_2, respectively. After eliminating data from conditionally stable atmospheres, we expect to obtain smaller standard deviations of buoyancy, as shown later. Note that the mean B_1 is close to the magnitude of the instrumental buoyancy error shown in Fig. 1.

As mentioned in section 2, we divide the dataset of an individual station into three subsets whose soundings exhibit average LFCs at 975 mb, 925 mb, and 875 mb. We do so under the assumption that soundings within each of the subsets will be similar. In the atmosphere represented by each sounding in the subsets so defined, the work needed for lifting a surface parcel

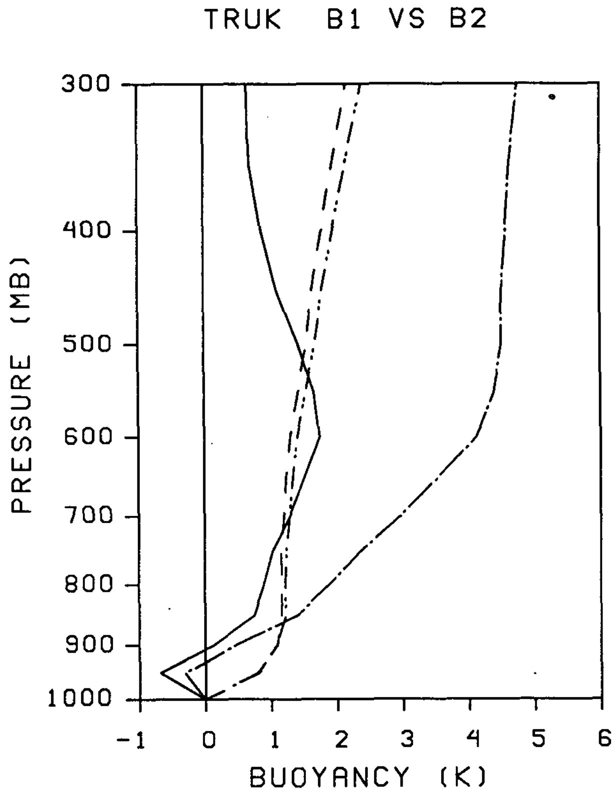

FIG. 2. The mean and standard deviation of buoyancy including (B_1) and ignoring condensate loading (B_2) at Truk. The solid line denotes the mean of B_1, the dashed the standard deviation of B_1. The dot-dashed line denotes the mean of B_2, the dot-dot-dashed the standard deviation of B_2. The cloud parcel is assumed to be lifted adiabatically from 1000 mb.

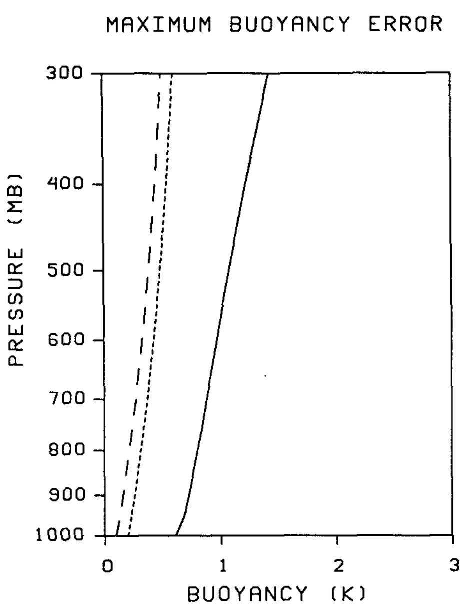

FIG. 1. The maximum buoyancy error (solid) due to instrumental error. The temperature error (short-dashed) increases linearly with pressure from 0.2°C to 0.6°C between 1000 mb and 300 mb, and the relative humidity error (long-dashed) increases linearly with pressure from 1% to 5% in the same region. Note that the values of R.H. error in the plot have to be multiplied by 10%.

to its LFC against stable stratification is approximately the same. Under the assumption that the observed atmosphere is convectively adjusted, we have also calculated the correlation coefficient between the potential and ambient thickness from 1000 mb to 300 mb for each subset. The *potential thickness* is the thickness that the atmosphere would have if it had the same virtual temperature as an undiluted cloud. Using the mixing ratio (q) at 1000 mb, the virtual temperature at 1000 mb (T_{v_a}), the lifting condensation level (LCL), and the LFC as parameters for classifying the dataset, we found that the correlation coefficient between the potential and ambient thickness is the largest when subsets are defined using the LFC (Xu 1987). Thus, we use the LFC to divide the dataset into three subsets. Sets F1, F2, and F3 (pertaining to LFCs at 975, 925 and 875 mb, respectively) include 12%, 35%, and 28% of the whole dataset for Truk station, 16%, 26%, and 25% for Majuro station, and 20%, 41%, and 24% for Koror station, respectively. The remainder are convectively stable soundings. The LFC can be approximately determined by comparing the virtual temperatures of cloudy air and the ambient air.

Because we are interested in the properties of conditionally unstable atmospheres, the soundings representing conditionally *stable* atmospheres are eliminated from the calculations of parcel buoyancy. We do this by assuming that conditionally stable atmospheres correspond to the virtual temperature of a cloud parcel less than the virtual temperature of the environment. In eliminating stable soundings in this way, we bias the results toward positive buoyancy. For example, a *perfectly neutral* atmosphere will generate a mean positive buoyancy roughly equal to the standard deviation caused by instrumental error. This problem is addressed at the end of section 4.

The buoyancy profiles for Sets T (the total set), F1, F2, and F3 at Truk station are shown in Fig. 3 (hereafter, we only show B_1, the buoyancy including adiabatic condensate loading). For brevity, the profiles for the other two stations are not shown here, but they are similar (see Xu 1987).

The profiles of mean buoyancy have the following features:

1) The average magnitude of the buoyancy generally increases as the LFC decreases.
2) A local buoyancy maximum always appears near 600 mb for all datasets.
3) The buoyancy is negative for Sets F2 and F3 below their LFCs.
4) The magnitude of the buoyancy depends largely on the mean surface θ_e. The mean surface θ_e of Koror is the highest among the three stations and that of Truk is higher than that of Majuro. As a result, the mean buoyancy at Truk is less than that of Koror, but larger than that of Majuro.

In Fig. 3, the standard deviations are about 1°C at most levels and thus are similar to the instrumental error (Fig. 1). They decrease as LFCs increase, as does the mean buoyancy. Moreover, it is found that the

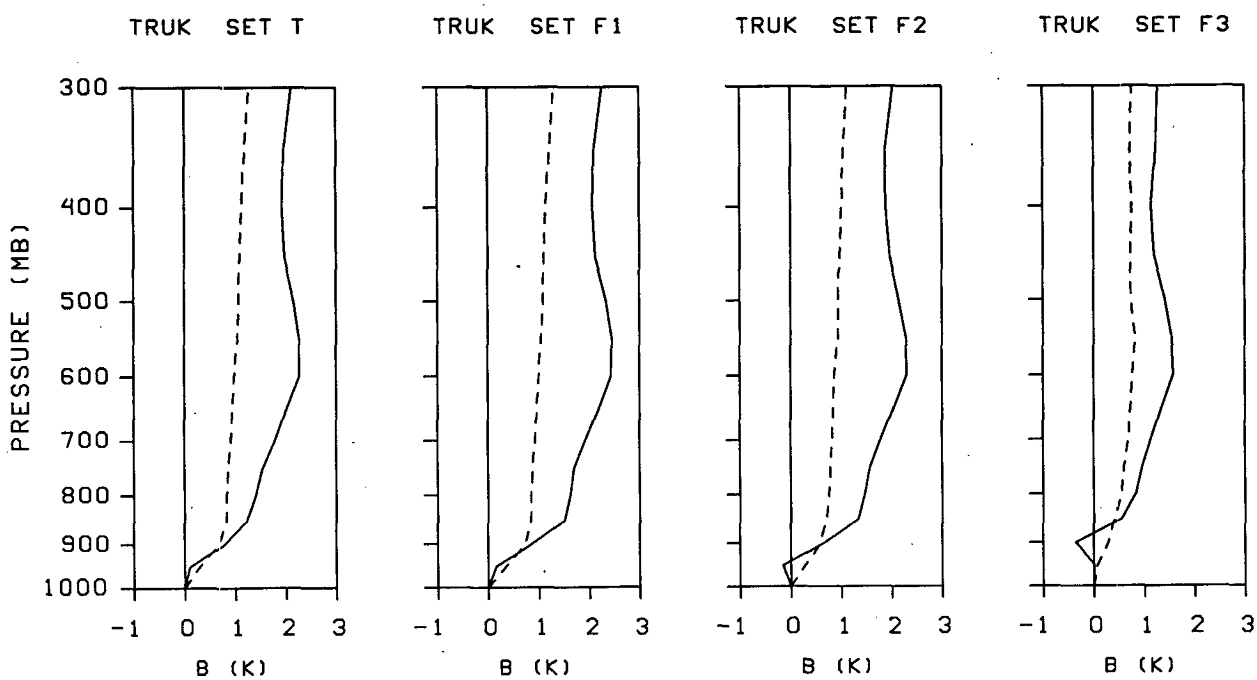

FIG. 3. The mean (solid) and standard deviation (dashed) of buoyancy (B_1) for Sets T, F1, F2, and F3 at Truk. See text for the definition of the datasets. The cloud parcel is assumed to originate at the maximum θ_e level in the mixed layer.

magnitude of the standard deviation is inversely proportional to the correlation coefficient between the potential and ambient thickness from 1000 mb and 300 mb (Xu 1987). This implies that the more convectively adjusted atmospheres have more uniform buoyancies.

4. Influence of parcel originating level on buoyancy

In this section, we attempt to use various mixed layer characteristics to crudely estimate the originating level of air parcels in order to assess its influence on parcel buoyancy. Recall that in section 3, the cloud air parcel is assumed to originate at the maximum θ_e level in the subcloud layer. The buoyancy so obtained is the maximum possible one. Here we seek to determine the buoyancy of parcels lifted from near the top of the subcloud layer.

As mentioned in section 2, the mixed layer is characterized by uniform virtual potential temperature (θ_v), although q and θ_e may individually vary. Because of its coarse resolution, the dataset used in this study cannot give us an accurate estimate of the thickness of the mixed layer. The difference in equivalent potential temperatures between 1000 mb and 950 mb [$\Delta\theta_e = \theta_e$ (1000 mb) $- \theta_e$ (950 mb)], on the average, is about 5°C, which suggests either that the average top of the mixed layer is below 950 mb or that θ_e is not well mixed in the mixed layer.

The average top of the mixed layer is close to 950 mb in the tropics. Thus we may expect that the 950 mb level is just below the mixed layer top in some of the soundings and just above it in others. Since θ_v rapidly increases with height above the mixed layer, soundings in the latter category should exhibit somewhat larger values of θ_v at 950 mb than at 1000 mb. We use the observed difference in θ_v between 950 mb and 1000 mb to estimate the location of the top of the mixed layer.

The mixed layer depth is used as a means of specifying four (4) subsets of data. When $|\Delta\theta_v| < 0.5$°C, the mixed layer is assumed to extend above 950 mb and the cloud parcel is assumed to originate at 950 mb. This subset is called ML1. The mean buoyancy, obtained by lifting a parcel from 950 mb, is *equal to or larger than* that obtained by lifting a parcel from the actual mixed layer top, since θ_e generally decreases with height in the mixed layer. Set ML2 includes soundings with 0.5°C $< |\Delta\theta_v| <$ 1.0°C, and in this set parcels are assumed to originate at 950 mb as well. But since the mixed layer top may be below 950 mb in this case, we define an additional subset (ML3) in which cloud air is assumed to have T and q values halfway between those of the 1000 mb and 950 mb levels. In Set ML4 it is assumed that parcels originate at 1000 mb for soundings with $|\Delta\theta_v| >$ 1.0°C. The depth of the mixed layer described in the last two datasets is characteristic of squall line wake areas (Johnson and Nicholls 1983).

The mean and standard deviation of buoyancies for subsets ML1–ML4 are shown in Fig. 4. The buoyancies in each of the subsets are very similar to each other and are much less than buoyancies defined by lifting

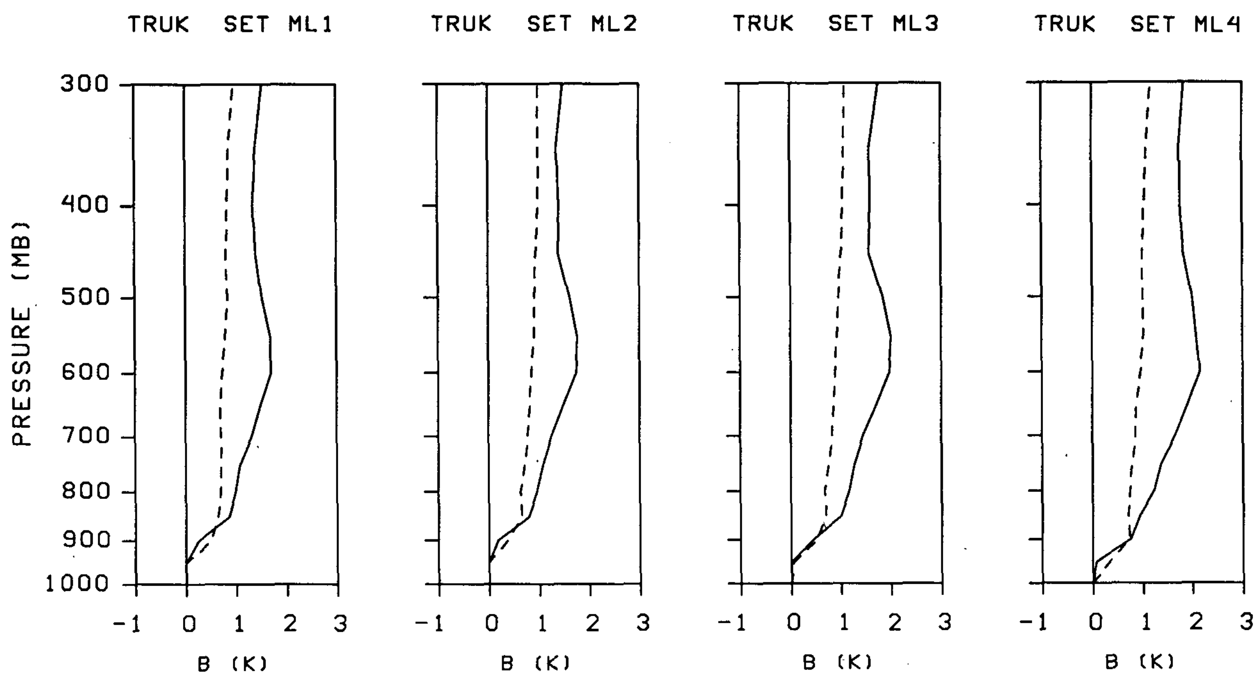

FIG. 4. The mean (solid) and standard deviation (dashed) of buoyancy for Sets ML1–ML4. The cloud parcel is assumed to originate near the top of the mixed layer. See text for the definition of the datasets.

air from the maximum θ_e level in the subcloud layer. Also note that there is no negative buoyancy just above the top of the mixed layer. The standard deviations are also much less than in the previously defined subsets and are about equal to what would be expected from instrumental error alone.

The reduction of buoyancy is most obvious in the upper troposphere: the difference is more than 0.7°C above 650 mb. The mean buoyancy at most levels does not exceed 2°C when the parcel originates from near the top of the subcloud layer. As mentioned earlier, the mean buoyancies would be even *smaller* when obtained by lifting a parcel from the *actual* mixed layer top.

In order to compare these directly with the maximum possible values of buoyancy available in the tropics discussed in section 3, we divide the dataset into the same three subsets as those described in section 3 except that the parcel is assumed to originate from near the top of the mixed layer. These new subsets and the whole dataset are called MLF1, MLF2, MLF3 and MLT, respectively. Figures 5 and 6 show the mean and standard deviation of buoyancy and their difference from those of the maximum available buoyancy described in section 3.

The reduction of the mean buoyancy in Sets T and F1 (compare with Fig. 3) is more than 0.3°C at most levels. The largest reduction of mean buoyancy occurs in Set F1, which, however, has the largest buoyancy among the datasets. The reduction in Set F2 is not as large as that of Set F1, while Set F3 has a relatively small reduction at all levels. Furthermore, note that the standard deviations in Set MLF1 are greatly reduced among the datasets shown in Fig. 6. It is seen that the precise originating level of parcels strongly influences estimates of the stability of tropical atmospheres.

Since the mean buoyancy in these subsets is comparable to the standard deviation to be expected from instrumental error, elimination of the "stable" soundings will result in the deletion of some soundings that in reality are unstable but which the instrumental error makes appear stable. This will bias the averaged soundings toward positive buoyancy. We illustrate this effect in Fig. 7, which shows at each pressure level the frequency distribution of soundings classified according to their buoyancy for Truk Set ML1. Inspection of Fig. 7 shows that the maximum medium buoyancy is less than about 1.5°C and that except for the long tail of negative buoyancies, the data are roughly normally distributed about the mean with a standard deviation comparable to the instrumental error. The long negative tail, especially evident at higher levels, is to be expected as there is no physical constraint on the magnitude of negative buoyancies. Figure 7 also suggests that the mean buoyancy (accounting for instrumental errors) reaches a maximum in the middle troposphere and decreases to small values in the upper troposphere.

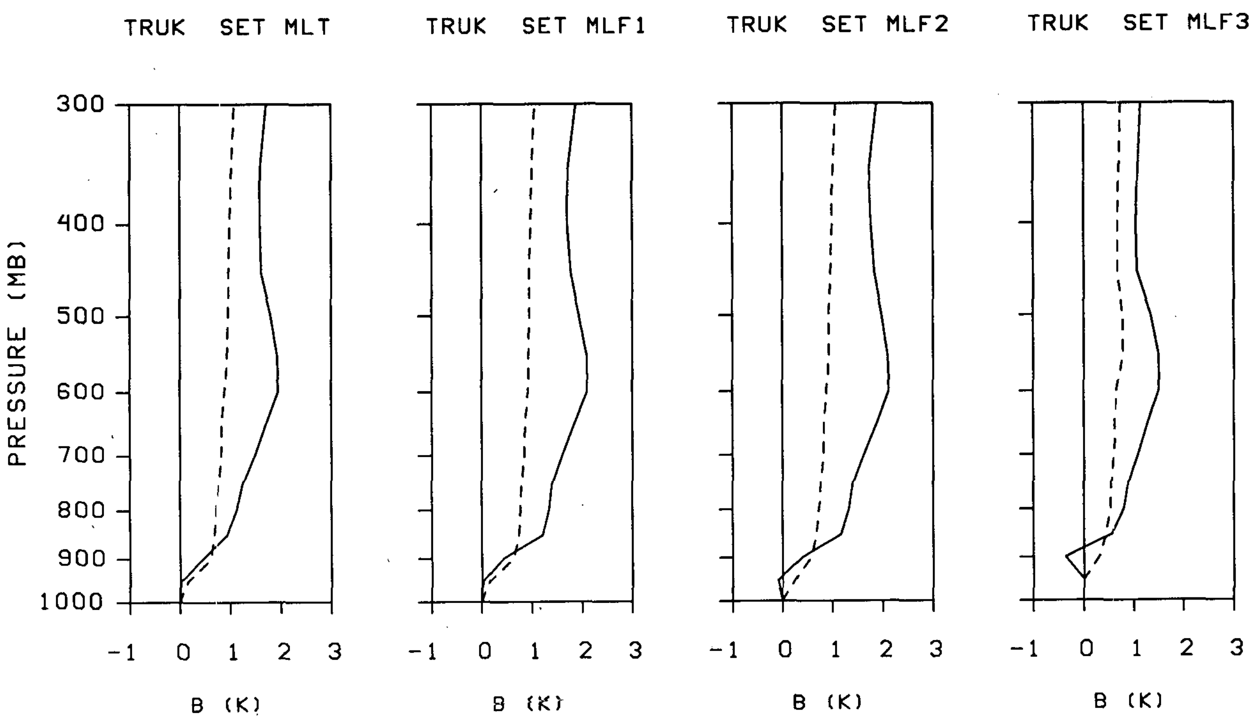

FIG. 5. As in Fig. 3 except assuming that the cloud parcel originates near the top of the mixed layer. See text for the definition of the datasets.

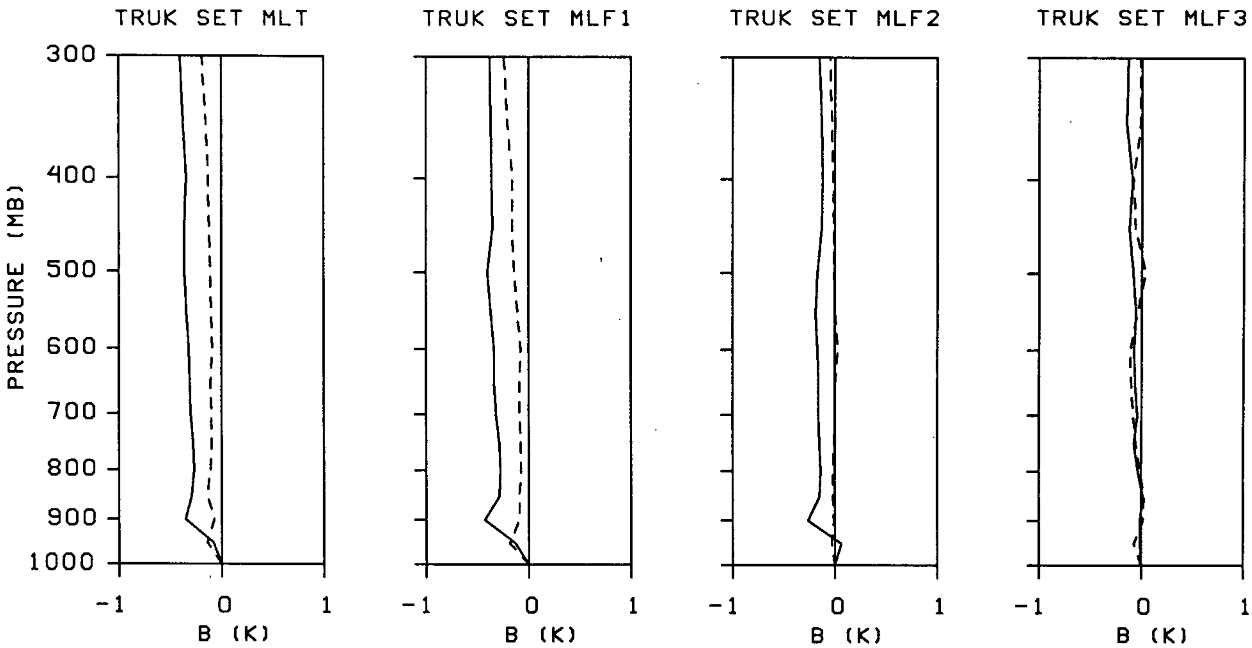

FIG. 6. The difference between the mean (solid) and standard deviation (dashed) of buoyancy of Fig. 5 and the maximum possible ones in Fig. 3.

5. Discussion and summary

We have systematically examined the stability of the tropical western Pacific atmosphere to reversibly displaced parcels from the subcloud layer. The results tend to confirm the observation by Betts (1982) that deep convective atmospheres are nearly neutral to such parcels when the adiabatic condensed water is included in the definition of buoyancy.

As pointed out by Betts (1982), the "mixed layer" is not actually well mixed in any conservative scalar except the virtual potential temperature (θ_v), which measures buoyancy in unsaturated air. Rather, the mixed layer is really a convectively adjusted layer. In particular, θ_e generally decreases with height in the mixed layer so that the buoyancy of reversibly created cloud parcels depends on precisely where in the subcloud layer they originate. Within the limitations imposed by the dataset we used, we made a crude attempt to estimate the buoyancy of air lifted from near the top of the mixed layer. These parcels have buoyancies that are remarkably close to zero through a deep layer; moreover, the standard deviation of buoyancy is close to what would be expected from instrumental noise alone. This points to a need to use more accurate data with better vertical resolution in the subcloud layer.

Our main conclusion, then, is that the tropical atmosphere, when and where it is not conditionally stable, is almost neutral to parcels displaced reversibly from the top of the subcloud layer. This is so in such a systematic way that it begs a physical explanation.

The numerous investigations of cumulus cloud properties (e.g., Malkus 1954; Paluch 1979) have all shown that such clouds are extremely inhomogeneous and, on the average, highly diluted. Careful measurements of deep, nonprecipitating cumuli (Paluch 1979) reveal, however, that these clouds contain samples of air that do represent nearly undiluted ascent from the top of the subcloud layer. Despite the fact that these parcels have the highest condensed water contents, it is easy to show that they have the largest buoyancies, since any evaporation leads to cooling that outweighs the reduction of condensed water content insofar as buoyancy is concerned. Thus we may conclude that the *highest* virtual temperature a *conditionally neutral* atmosphere could have is that of an undiluted subcloud layer parcel lifted through its depth, provided the convection is nonprecipitating.

Much of the deep convection occurring in the region which our soundings were taken in is precipitating, though. Naturally, if undiluted parcels lose condensed water by precipitating, their buoyancies will be larger. Yet precipitation may also fall into parcels from above, decreasing their buoyancy. In general, parcels near cloud base may be expected to *lose* buoyancy due to precipitation falling from above, while those closer to cloud top will have increased buoyancy due to fallout of water. Heat released by freezing in the upper portions of deep clouds will also lead to increased buoyancy. With these considerations, it is not difficult to understand why the calculated parcel buoyancies decrease above 600 mb in nearly all our estimates. The relevance of a reversibly lifted parcel to the virtual temperature of the upper troposphere is thus questionable.

A question arises as to why the tropical atmosphere

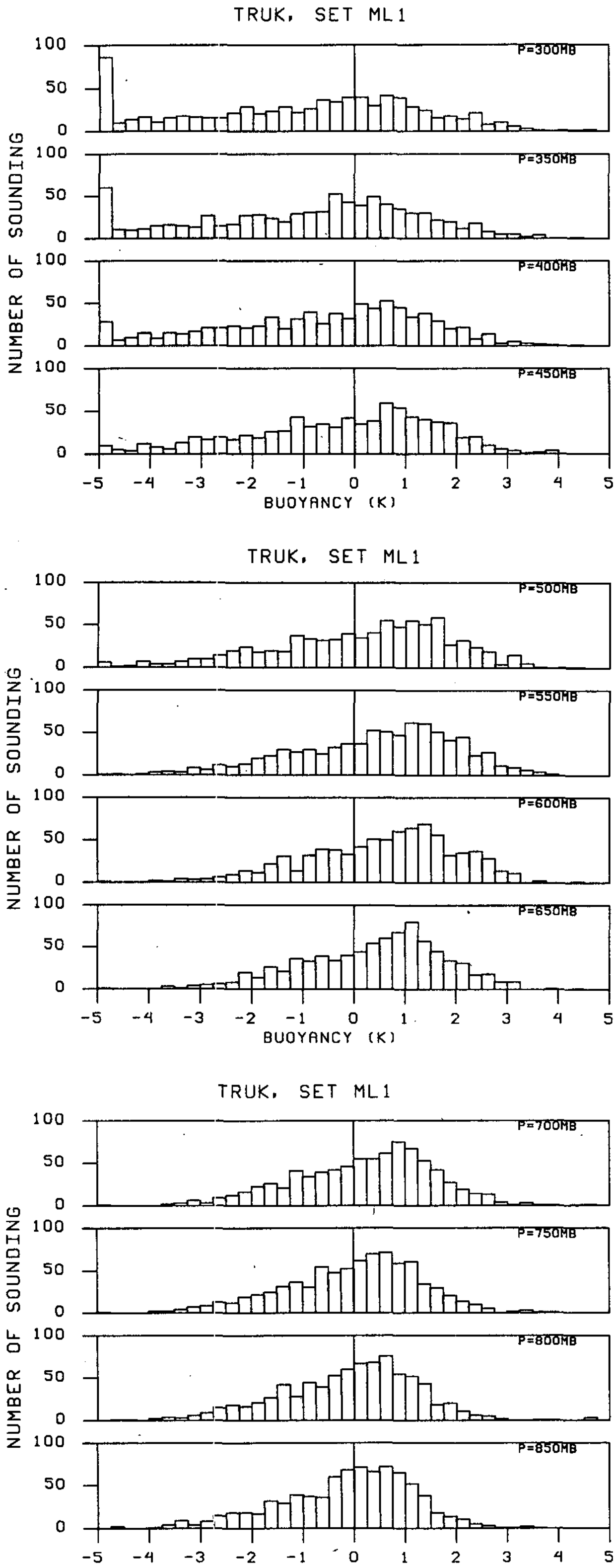

FIG. 7. The number of soundings at Truk in each buoyancy range of 0.25°C for each pressure level. The leftmost bar contains all soundings with buoyancies less than −4.75°C. The buoyancies were calculated from air lifted reversibly from 950 mb in those cases for which the θ_v at 950 mb was within 0.5°C of the value at 1000 mb.

appears to be more systematically neutral to parcels lifted from near the *top* of the mixed layer. Again we refer to the results of Paluch (1979) who showed that the undiluted samples in middle and high levels within Colorado cumuli originated near cloud base rather than near the surface.

We are not in a position to answer conclusively the question posed by the title. We have shown, however, that the tropical atmosphere is nearly neutral to nonprecipitating clouds originating in air lifted from near the top of the subcloud layer. It is difficult to assess the degree of instability to nonprecipitating clouds from individual soundings since the measured buoyancies are likely to be comparable to those resulting from instrumental error. More substantial buoyancies may be realized by lifting air from near the bottom of the subcloud layer *or* by relieving the water loading by precipitation. We note, however, that the latter process does not necessarily reduce the average condensate loading of clouds, as is often assumed. Indeed, the water loading at cloud base is always superadiabatic, and may be so aloft if the terminal velocity of precipitation particles is not too much greater than the updraft velocity, as may be the case in the upper portion of clouds where the precipitation is often in the form of snow. We emphasize that the observation of subadiabatic water content does not contradict the foregoing argument since subadiabaticity can result from mixing as well as from precipitation. As mentioned previously, mixing always reduces buoyancy in spite of reducing the water loading. Finally, we note that the measured vertical velocities in tropical clouds are consistent with the small buoyancies shown by the soundings.

To the extent that the tropical atmosphere is very nearly neutral to reversible parcel ascent from the subcloud layer, the actual density of the free troposphere is strongly tied to the moist entropy of the mixed layer. This means that generation of potential energy in such an atmosphere must be due mostly to processes that change the mixed layer entropy and not to latent heating per se. That is, vertical motion that results in latent heat release unaccompanied by boundary layer entropy increases cannot lead to kinetic energy generation since no virtual temperature perturbations can result. Put yet another way, no generation of kinetic energy can occur in a conditionally neutral barotropic atmosphere with fixed subcloud layer moist entropy. The oft-used theoretical device whereby latent heating tied to moisture convergence drives growing disturbances is as problematic as a theory that used mass convergence to drive unstable modes in a dry atmosphere.

The observations presented here point to possible problems with the concept of CISK (*Conditional Instability of the Second Kind*), since no amount of moisture convergence can lead to energy production in a conditionally neutral atmosphere.[2] We believe that

[2] Small instability likewise implies small energy production, assuming that it is not released strictly by cumulus clouds themselves.

it is more likely that large-scale disturbances result either from redistribution of preexisting kinetic energy (as in barotropic instability) or feedback mechanisms that increase the entropy of the subcloud layer, as is the case with tropical cyclones (Emanuel 1986; Rotunno and Emanuel 1987), and possibly the tropical intraseasonal oscillations (Emanuel 1987; Neelin et al. 1987). We believe that future investigations of large-scale tropical circulations should focus on processes affecting subcloud layer entropy content rather than on local release of small amounts of conditional instability.

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