Thermodynamic Classification of Tropical Convective Soundings

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ABSTRACT

Soundings from a land tropical experiment (VIMHEX II) are classified according to local area rainfall into four regimes depicting the transition from dry to highly disturbed days. The thermodynamic structure for each of the four regimes is presented, and the implications for convective parameterization theories and the prediction problem are discussed in terms of the potential energy available for parcel convection for different entrainment rates.

1. Introduction

For many years (see, for example, Riehl, 1954) the tropics have been known as a region of small horizontal temperature gradients. Variations of moisture between regions of suppressed or enhanced convection (corresponding to regions of sinking or rising motion on a large scale) have been clearly visible, but in the absence of strong inversions, horizontal variations of temperature have been hard to establish. Small-scale fluctuations, or local effects, or radiosonde calibration errors easily mask any synoptic scale horizontal gradients. However, often the atmosphere on rainy days has appeared colder than on dry days (e.g., Riehl, 1948; Riehl *et al.*, 1973). Ludlam (1966) suggested that atmospheric convection establishes a characteristic stratification and Aspliden (1971) has shown

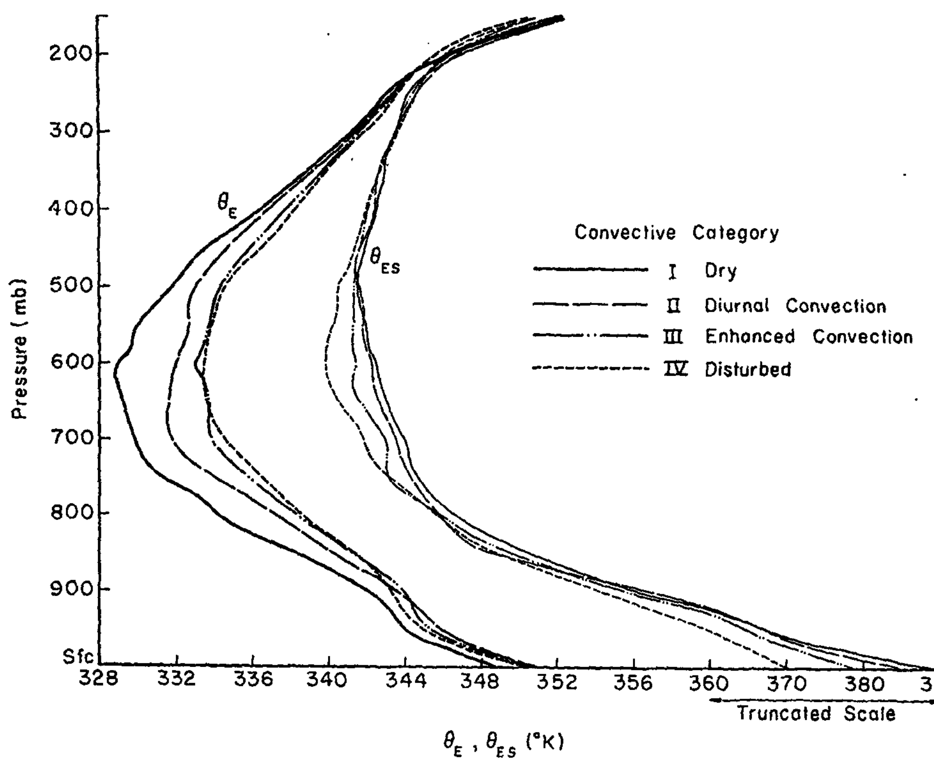

FIG. 1. Vertical profiles of equivalent potential temperature (θ_e) and saturation equivalent potential temperature (θ_{es}) for four convective regimes: dry, diurnal convection, enhanced convection, and disturbed.

how the tropical atmosphere can be classified into modes using the equivalent potential temperature stratification.

Convection parameterization schemes have required as one test for the presence of deep convection that the model atmospheric structure be conditionally unstable, and in some models a structure that is in balance with the convection is maintained (Betts, 1973a; Arakawa and Schubert, 1974). Betts (1973b) has suggested that the thermodynamic atmospheric structure might be used to determine what type of convection might be present. The purpose of this paper is to explore these ideas further using observational data, and to show that, in some average sense, the tropical atmosphere exhibits both a thermal and vapor stratification which is a function of the level of convective activity.

2. Data

From June to early September, 1972, over 300 rawinsonde soundings were taken at Carrizal, Venezuela (9°22.8′N, 66°55.0′W) during the second Venezuelan International Meteorological and Hydrological Experiment (VIMHEX II). A network of 40 raingauges were placed within a circle of 60-km radius around the rawinsonde site and the area was scanned by a 10-cm calibrated weather radar. A daily rainfall index (representative of the average precipitation in mm for the area) was determined from the gauges (see Betts and Stevens, 1974). This rainfall index was used to stratify the days according to convective activity (Table I).

The choice of categories by rainfall range was made somewhat arbitrarily so as to give four roughly equal groups of days. However, a subjective assessment based on visual and radar observations would be similar. The number of soundings is biased towards disturbed days since the study of these had the highest priority. The soundings in each category were simply averaged, after interpolation to 25-mb intervals.

The climate of the area is a tropical savannah type, and the period of the experiment was the rainy season. The rainfall is almost entirely convective, with scattered small cumulonimbus on days of low precipitation and larger mesoscale systems or tropical squall lines for the regimes III and IV. There is normally a diurnal variation with a rainfall maximum in late afternoon or evening, but this is modulated by traveling synoptic-scale disturbances. On the type IV days there was usually a low-level trough, and the time period of precipitation was variable. The prevailing low-level wind field was easterly, with a maximum easterly wind speed of about 10 m s^{-1} at 600 mb and westerly winds above 300 mb. Typically, the convective cells moved from east to west at speeds of 5–10 m s^{-1}, and 12–15 m s^{-1} for the squall lines.

Table I. Convective regimes and rainfall.

Convective regime	Rainfall index range (mm)	Number of days	Number of soundings
I. Dry	<0.1	23	44
II. Diurnal convection	0.1–1.0	27	67
III. Enhanced convection	1.0–5.0	28	103
IV. Disturbed	>5.0	21	91

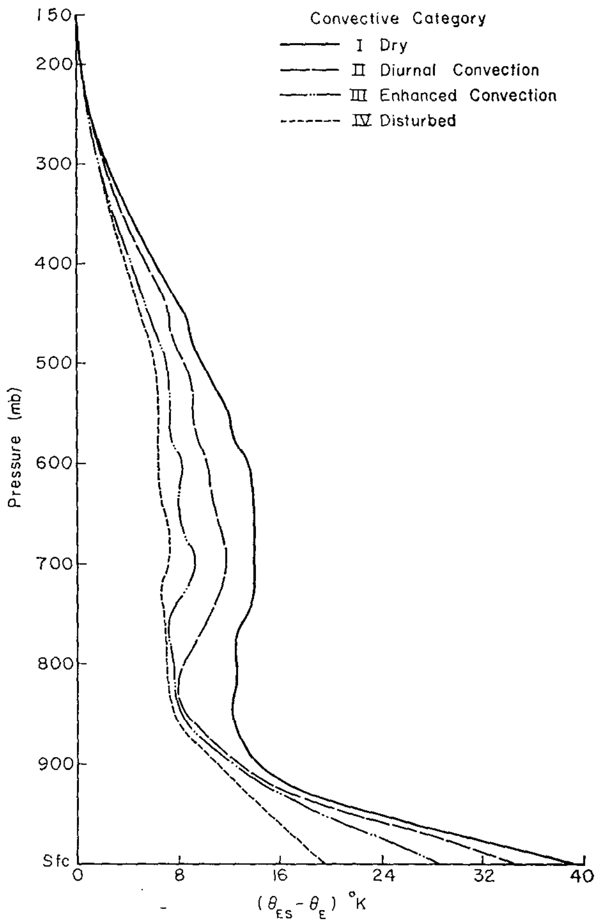

Fig. 2. Vertical profile of $(\theta_{es}-\theta_e)$ for the four regimes.

3. Results

Figure 1 shows the vertical profiles of saturation equivalent potential temperature (θ_{es}), which is measure of the temperature at a given pressure, and equivalent potential temperature (θ_e), which is a function of both the temperature and the humidity, for averages of the four regimes. The θ_{es}, θ_e curves show opposite trends with increasing precipitation. Figure 2 shows the difference $(\theta_{es}-\theta_e)$: this is a measure of the unsaturation or humidity of the atmosphere. The decrease of the θ_{es} and increase of the $(\theta_{es}-\theta_e)$ curves from dry to disturbed days show that there is a uniform trend towards a cooler, moister atmosphere.

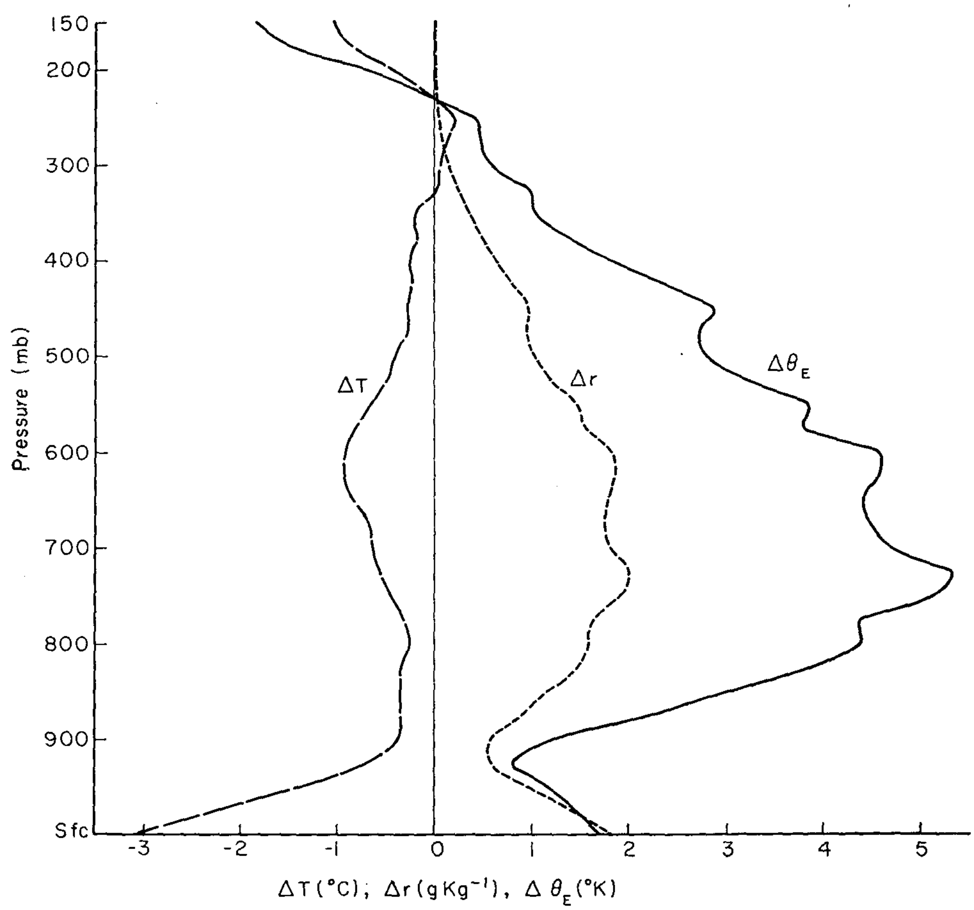

FIG. 3. Difference of temperature (ΔT), mixing ratio (Δr), and equivalent potential temperature ($\Delta\theta_e$) between the disturbed and dry regimes.

Apart from the lowest 100 mb, the resulting θ_e of the atmosphere has an increasing trend with increasing precipitation. These trends are in the direction of the forcing: for example, large-scale mean ascent will tend to cool and moisten the atmosphere in disturbed conditions. The convective transports are a response to this forcing, and in the main produce the opposite effect—a warming and drying. The mean atmospheric state which we observe (Figs. 1 and 2) represents some balance between these opposing processes. However, as one might expect, this "balanced state" shifts in the direction of the forcing. The concept of a balanced state while convection is in progress is closely related to the quasi-equilibrium hypothesis discussed in the next section.

Figure 3 shows the difference in temperature, mixing ratio, and θ_e between disturbed and dry averages. The average temperature difference at 600 mb is still only about $-1°C$ ($\Delta\theta_{es} = -2.7$ K). Δr is smaller near 900 mb than in the middle troposphere, because of the drying effect of downdrafts; this is reflected also in $\Delta\theta_e$. Figure 3 confirms the conclusions of Riehl *et al.* (1973), based on the VIMHEX I experiment, that the disturbed atmosphere over Venezuela is relatively cooler and moister. However, the mean vapor differences are about twice as large from this second experiment. The later values are from a much larger sample, and are more reliable, since the radiosonde used did not suffer from the humidity errors of earlier sondes (Riehl and Betts, 1972; Betts *et al.*, 1974).

These are average results, and the scatter in individual soundings is large. This scatter may well be due to the variations on a smaller scale than that of the disturbances (an average over many soundings represents a larger area average). Thus, it is still often not possible to use the stratification shown by a single sounding as a predictive or diagnostic tool in the tropics. This may be possible on a daily basis given many soundings (see next section), but for predictive purposes there is the additional problem that strong disturbances are often preceded by a strong subsidence zone. It should also be noted that this is a purely thermodynamic classification, and it is probable that other factors such as the wind shear are also important in controlling different regimes of convection (Moncrieff, 1973). The control on the intensity of convection by large-scale forcing represents a different type of control (Betts, 1974), which is also related to the quasi-equilibrium hypothesis discussed in the next section.

4. Implication for parameterization theories and the prediction problem

Figure 2 shows the steady moistening of the atmosphere from dry to disturbed conditions. Clearly this will affect the potential energy available to an entraining cloud parcel which ascends through one of these mean atmospheres. We shall define this potential energy available for parcel convection as a function of entrainment rate (λ) as

$$A_p(\lambda)=\int_{\text{cloud-base}}^{\text{cloud-top}} \frac{g[\theta_p(\lambda)-\bar{\theta}]}{\bar{\theta}}dz, \qquad (1)$$

where the suffix p denotes a parcel property, $\bar{\theta}$ is the mean environmental potential temperature, and λ is the fractional entrainment rate. This is the area on a tephigram between environment and parcel paths. This is closely similar to Arakawa and Schubert's (1974) definition of cloud work function $A(\lambda)$. The difference is that $A_p(\lambda)$ considers only unit mass of cloud parcel, rather than a parcel mass increasing as it entrains. Virtual temperature corrections are also omitted here for simplicity.

Figure 4 shows $A_p(\lambda)$ against λ for the four convective regimes. Parcels were assumed to rise from the subcloud layer with initial $\theta_e=\bar{\theta}_e$ (950 mb), and to entrain at a constant fractional rate λ (%/100 mb). Cloud-top was taken as the thermal equilibrium height as in Arakawa and Schubert (1974).

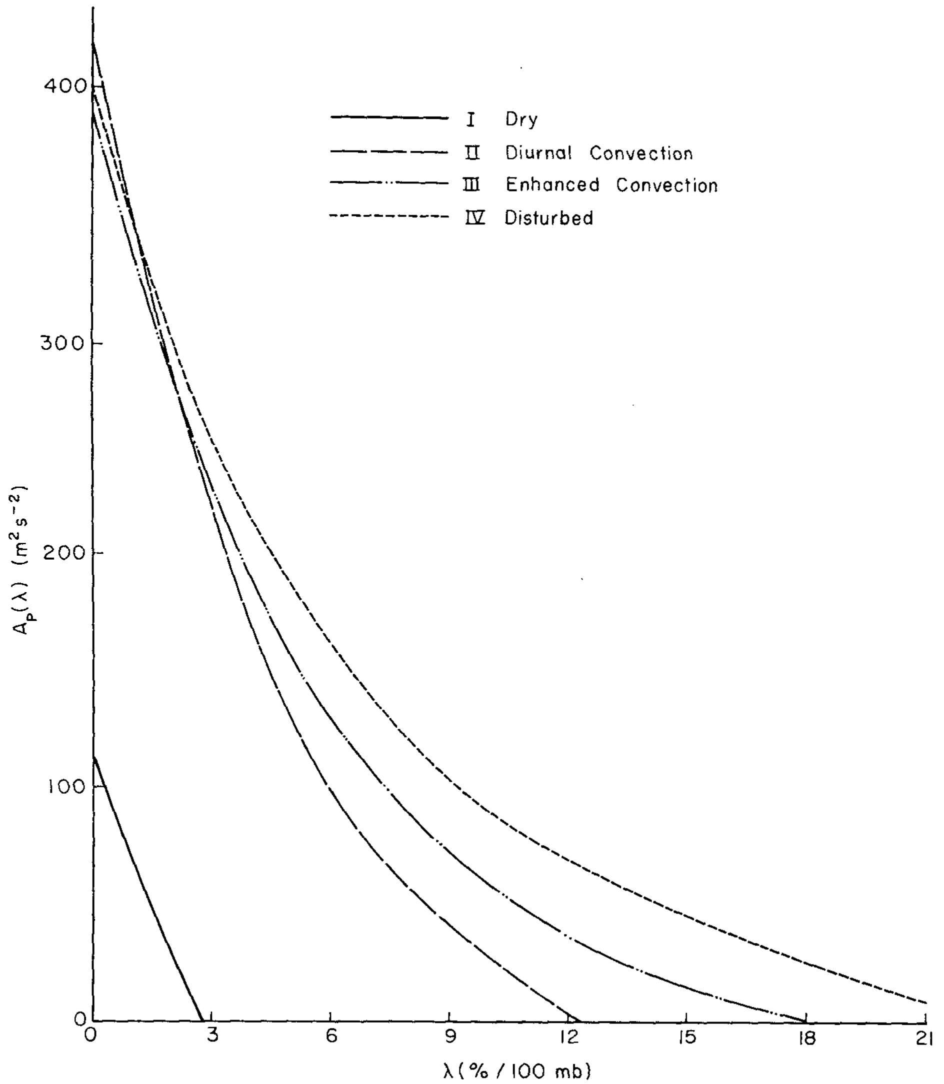

FIG. 4. Variation of potential energy available for parcel convection, $A_p(\lambda)$, as a function of fractional entrainment rate, λ, for the four convective regimes.

These are remarkable curves. First, we note that the dry-day (zero-rainfall) curve is in a different category than the other three, which form a uniform sequence. This is encouraging evidence for the quasi-equilibrium hypothesis (Betts, 1974; Arakawa and Schubert, 1974). It shows that while some deep convection occurs, the whole troposphere can maintain a balanced structure, which changes only slightly in the direction of the forcing. For example, Fig. 4 shows how, for a given λ, $A_p(\lambda)$ increases as conditions change from diurnal to disturbed conditions, when presumably the convection is being strongly forced. Arakawa and Schubert (1974) used a constraint $dA(\lambda)/dt=0$ as a quasi-equilibrium condition: Fig. 4 indicates how much this function can change from one convective regime to another. This change is not very large, however, when compared with the difference in $A_p(\lambda)$ for the dry-day mean. Thus Fig. 4 confirms both that the condition $dA(\lambda)/dt=0$ may be a useful first approximation while deep convection is present, and also, as might be expected, that once significant convection is suppressed, no such constraint on the atmospheric structure exists, and the potential energy available for parcel convection can decrease markedly.

Because of this, one may conclude, from the predictive viewpoint, that while it is possible to predict zero rainfall from the stratification alone, it is difficult to predict the converse: the amount of rainfall. This is determined by the large-scale forcing, and is reflected by only small changes in the nearly balanced state of the convective stratification.

These averages include all soundings for each type of day. Values of $A_p(\lambda=0)$ can typically rise to much higher values (>1000 m^2 s^{-2}) on smaller timescales before the onset of rain. However, precipitation produces downdrafts, and these reduce the low-level θ_e, thus reducing A_p in the average. The variation of $A_p(\lambda=0)$ between categories II, III, and IV is most influenced by the mean low-level θ_e, and should therefore be viewed with caution because the variability of θ_e at low-levels is largest, and the sampling may be less representative.

5. Conclusion

The change of stratification with increase in convective activity from very suppressed convection to a highly disturbed state has been presented. These illustrate the quasi-equilibrium state of the atmosphere when convection is in progress, which makes prediction of the level of convective activity from the stratification alone conceivable but difficult. Further, these changes in this quasi-equilibrium state probably need to be included in parameterization theories if these are to constrain the atmosphere to a structure like that observed.

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