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Thermal Modification of the Troposphere due to Convective Interaction

By
William R. Kininmonth

Department of Atmospheric Science
Colorado State University
Fort Collins, Colorado

Colorado State University

Department of Atmospheric Science

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DUE TO CONVECTIVE INTERACTION

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ABSTRACT

The erosion and subsequent mixing to the environment of the updraft of mesoscale convective clouds is examined and modeled as a mechanism for the thermal modification of the cloud environment. Atmospheric soundings made near the cloud edge of mesoscale convective systems at Anaco, Venezuela during project VIMHEX* indicate that these systems are a sink of heat to the prevailing airflow. The convective mixing model is shown to be capable of explaining this heat exchange and thermal modification.

To simulate the effect of an intensifying tropical disturbance the model is applied with higher values of updraft equivalent potential temperature. The model demonstrates that the convective clouds are transformed from a net sink to a net source of heat in the troposphere and remain in phase with the observed thermal changes of tropical disturbances.

* Venezuelan International Meteorological and Hydrological Experiment

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Thanks are due to Professor Herbert Riehl for his support and encouragement to undertake this study. His many helpful suggestions during the preparation of this manuscript are appreciated.

The opportunity to take part in project VIMHEX is acknowledged.

LIST OF SYMBOLS

C_p	Specific heat of air at constant pressure
e_s	Saturation vapour pressure
g	Acceleration of gravity
L	Latent heat of condensation
p	Pressure
q	Mixing ratio
q_1	Updraft mixing ratio
q_2	Environment mixing ratio
q_c	Cloud edge mixing ratio
q_s	Saturation mixing ratio
R	Gas constant for dry air
σ	Cloud hydrometeor mixing ratio
σ_1	Updraft hydrometeor mixing ratio
T	Temperature
T_1	Updraft temperature
T_2	Environment temperature
T_c	Cloud edge temperature
τ_1	Mass flux at a vertical surface from the updraft
τ_2	Mass flux at a vertical surface from the environment
T^*	Virtual temperature
z	Height

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Introduction

One of the major problems of the meteorology of tropical regions is to determine how a cold core disturbance of the lower and middle troposphere of the trade wind easterlies can be transformed into a warm core system, the development of which may lead to the formation of a tropical cyclone. Despite considerable analysis and modeling of these tropical disturbances, and in particular the mature tropical cyclone, fundamental aspects of mesoscale and synoptic scale energy conversions have not been developed. Nowhere is this more apparent than in numerical studies where tropical cyclones are generally the result of weak disturbances, whereas in actuality these weak disturbances infrequently develop to tropical cyclone intensity.

There is no doubt that the energy source of the warm core tropical disturbance is the release of latent energy during condensation within convective clouds. However the processes involved in the conversion from latent energy to kinetic energy are obscure. Most tropical precipitation (with the associated release of latent energy) is produced by synoptic scale disturbances. Predominantly these disturbances are not warmer but are more dense in the lower and middle troposphere than their surroundings (Riehl, 1969).

In the transformation from a weak disturbance, which is more dense in the rain areas, to the more intense warm core vortex there is a requirement for the thermal field to change. The heat sink of the lower and middle troposphere of the weak disturbance must be replaced by a heat source which produces the warm anomaly.

Previous work (Riehl, 1965; Garstang et al, 1967) has indicated that the changes in the tropospheric energy profile associated with tropical disturbances can be attributed to vertical adiabatic exchanges within the cumulonimbi. The middle tropospheric energy is increased by mixing with the higher energy updraft air while the low level energy is decreased by wet adiabatic descent of lower energy middle tropospheric air in the downdrafts.

If the horizontal interaction between the convective updrafts and the cloud environment is such that a noticeable change in the energy occurs then there must be a two way flux between the cloud and its environment. During this horizontal interaction excess enthalpy from the eroding updraft is advected to the environment while enthalpy is consumed during the evaporation of cloud matter.

Objective. It is the purpose of this study to investigate the erosion of the buoyant updraft and the subsequent mixing with the environment. The relative importance of the evaporation of cloud matter as a heat sink and the advection of enthalpy as a heat source during this erosion process will be compared to determine the role of the cloud in the thermal modification of its environment.

Thermodynamics of Cumulonimbus Erosion

Considerable work has been done on the modeling of the structure, dynamics, and thermodynamics of convective clouds. A general discussion of convective theories is given by Simpson et al (1965), and by Newton (1967). Probably the most significant single advance in the evaluation of convective processes was the proposition and enumeration of entrainment by Stommel (1947). Stommel's intent, like that of following workers, was to explain why different clouds grow to different heights and at different rates but under similar environmental conditions. The requirement was to model the growth of convective clouds.

Recognising that different clouds have different growth and size characteristics, the method of this study will be to make use of known characteristics of the cloud forms to explain how the clouds can modify their environment. In this study the classical "top hat" profile introduced by Stommel will be modified to allow for horizontal turbulent exchanges.

In a growing or mature convective cloud the region of the buoyant updraft center is a maximum of total energy (where total energy, or energy, is defined as the sum of the specific enthalpy, the potential energy and the latent energy). The cooler drier environment of the cloud is an energy minimum for a pressure surface. It is assumed that stresses acting between the ascending updraft and the environment results in turbulent exchanges such that in the mean there is a continuous profile of total energy from the updraft to the undisturbed environment.

Away from the updraft energy maximum the energy is decreasing toward that of the undisturbed environment. At any point on a vertical surface surrounding the updraft the actual energy is the result of the relative mass fluxes through the surface from the updraft and from the undisturbed environment.

To evaluate the energy at a point on the vertical surface it is necessary to equate the relative mass fluxes into the surface and the energy transported by these fluxes. A positive flux is defined as being out of the vertical surface (whether toward the updraft or toward the environment) and a negative flux is defined as being into the vertical surface. For conservation of mass the total flux into the surface is equal to the total flux out of the surface.

$$- \tau_1 - \tau_2 + (\tau_1 + \tau_2) = 0 \qquad (1)$$

For conservation of energy the sum of the energy transports of the mass fluxes into the surface from the updraft and the environment is identical with the energy transports due to the mass fluxes out of the surface.

$$- \tau_1(C_pT_1 + Lq_1 + gz) - \tau_2(C_pT_2 + Lq_2 + gz) + (\tau_1 + \tau_2)(C_pT + Lq + gz) = 0 \qquad (2)$$

Similarly the transport of water into the surface must be equated with the transport of water out of the surface.

$$- \tau_1(\sigma_1 + q_1) - \tau_2 q_2 + (\tau_1 + \tau_2)(\sigma + q) = 0 \qquad (3)$$

Isolating the mixing ratio at the vertical surface from equation 3 and substituting this value of q into equation 2 leads to (following rearrangement) an expression for the temperature at the vertical surface.

$$T = T_2 + \left\{\frac{\tau_1}{\tau_1 + \tau_2}\left[(T_1 - T_2) - \frac{L\sigma_1}{C_p}\right] + \frac{L\sigma}{C_p}\right\} \, {}^oK \qquad (4)$$

Also from equation 3 it follows that:

$$\frac{\tau_1}{\tau_1 + \tau_2} = \frac{\sigma + q - q_2}{\sigma_1 + q_1 - q_2} \qquad (5)$$

Hence the temperature of the vertical surface is also given by:

$$T = T_2 + \left\{\frac{\sigma + q - q_2}{\sigma_1 + q_1 - q_2}\left[(T_1 - T_2) - \frac{L\sigma_1}{C_p}\right] + \frac{L\sigma}{C_p}\right\} \;^oK \qquad (6)$$

The vertical surface considered is arbitrary. The temperature at any point of the mixing region about the updraft is consequently specified by the cloud hydrometeor mixing ratio and vapour mixing ratio at that point, together with the boundary parameters of the updraft and the environment at that pressure level.

At the cloud edge the hydrometeor mixing ratio is zero and the vapour mixing ratio (q_c) is near the saturation value for that temperature (T_c). The temperature of the cloud edge is consequently given by:

$$T_c = T_2 + \frac{q_c - q_2}{\sigma_1 + q_1 - q_2}\left[(T_1 - T_2) - \frac{L\sigma_1}{C_p}\right] \;^oK \qquad (7)$$

The temperature excess $(T_1 - T_2)$ and the hydrometeor mixing ratio of the updraft specify the sign of the temperature anomaly $(T_c - T_2)$ of the cloud edge from the environment.

To evaluate the temperature of the cloud edge, environment temperature and mixing ratio profiles are readily available from atmospheric soundings. For large cumulonimbi the updraft temperature and mixing ratio profiles can be approximated as undilute ascent from the sub-cloud layer. Knowing the sub-cloud layer equivalent potential temperature specifies the temperature and mixing ratio of the updraft at any pressure level.

The updraft hydrometeor mixing ratio is an unknown parameter. Adiabatic liquid water values are not realistic as the larger size droplets fall out as rain. With height, as the air density decreases, the terminal velocities of droplets increase and less water would be expected to be held in the updraft. In the lower levels where the updraft velocity is increasing rapidly with height so too the water content would be expected to be increasing sharply with height. Fletcher (1962) gives a summary of observed liquid water in cumulus type clouds. There is a wide variation in the values found by different researchers. These results vary from less than 0.5 g/meter^3 to 10 g/meter^3. Numerical models, for example Squires and Turner (1962), typically give maximum values of cloud water in excess of 6 g/Kg.

At the cloud wall the mixing ratio is near the saturated value for that temperature and pressure. As a first approximation equation 7 can be solved at a pressure level by specifying an updraft profile and using an environment sounding and a reasonable liquid water profile. The mixing ratio of the cloud wall can be approximated at first by the updraft mixing ratio. That is, $q_c = q_1$. Equation 7 is then solved for the cloud edge temperature T_c.

A convergent iterative method can be employed to obtain a better estimate of the cloud wall temperature and mixing ratio. The first estimate of T_c can be used to re-evaluate the cloud edge mixing ratio. The saturation vapour pressure, e_s, is estimated from:

$$e_s = 6.11\exp\left[\frac{0.622L(T_c - 273.2)}{T_c R 273.2}\right] \text{ mb} \qquad (8)$$

which, by substitution into:

$$q_s = \frac{0.622e_s}{p} \ g/g \tag{9}$$

gives a new estimate of the cloud edge saturation mixing ratio. Either the estimate of the cloud edge saturation mixing ratio or a mixing ratio less than the saturated value can be used, through equation 7, to define a new estimate of the cloud edge temperature at that pressure level. This cycle is repeated until the desired level of accuracy is achieved.

The temperature anomaly of the mixing region is considered to be dominantly the result of horizontal turbulent exchanges. Even though there are strong vertical motions in the updrafts and downdrafts within the cumulonimbi, these motions are along constant energy surfaces. The updraft air reaching the cloud edge at any pressure surface can be considered the result of ascent or descent along a wet adiabat followed by horizontal mixing.

Outside the cloud the lapse rate followed by air parcels is dry adiabatic. Vertical motions will tend to destroy any anomaly of temperature formed at the cloud wall. The weak vertical motions outside the cloud wall and strong horizontal motions of the environment relative to the cloud ensures that any temperature anomaly formed at the cloud wall will be propogated downwind before dissipation.

A schematic cross-section through a cumulonimbus cloud theorising the effects of horizontal mixing is described in figure 1. Here the cloud is acting as an atmospheric heat sink with cloud edge cooling. Enthalpy of the environment is converted to latent energy through the evaporation of cloud hydrometeors. Due to the mean horizontal motion of the environment around the updraft, the temperature anomaly of the cloud edge is shown to be advected downwind.

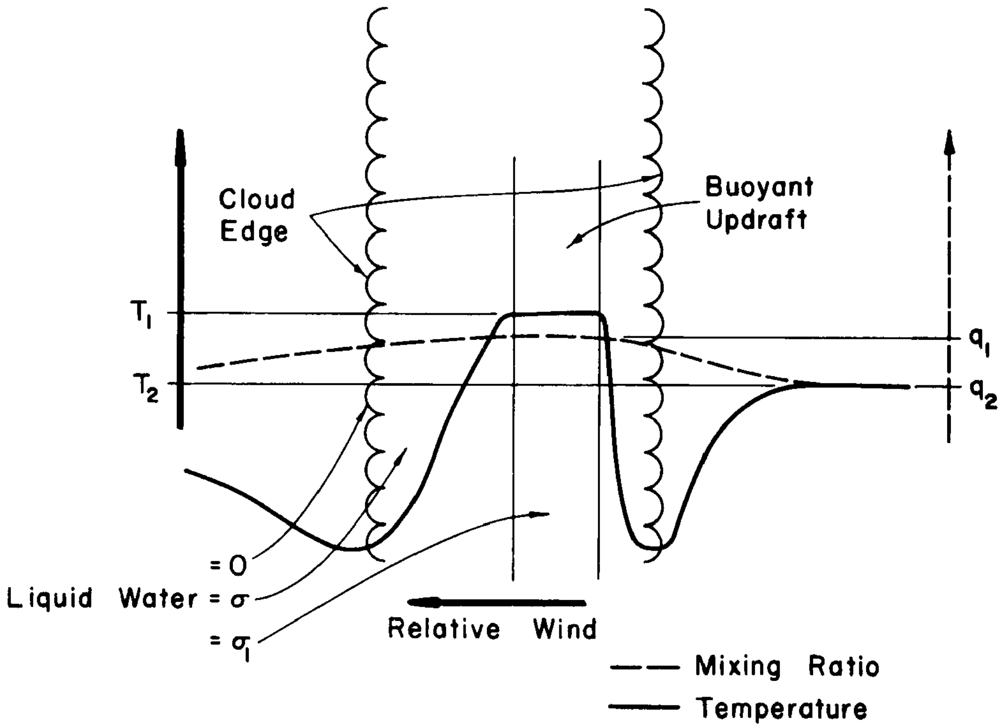


Figure 1. Schematic cross section through the model cumulonimbus showing cloud edge cooling.

Observational Data

Due to the violent nature of cumulonimbus clouds very little is known of the updraft conditions in these mesoscale systems. From radar reflectivities it is known that the hydrometeor content in and around the updraft is high, but a percentage of this will fall out as precipitation. If a cumulonimbus cloud is to act as a heat sink in the atmosphere then the hydrometeor mixing ratio of the updraft evaporated to the environment must be high while the thermal excess is constrained. Within active cumulonimbi the updraft temperature is constrained by the surface equivalent potential temperature, while a vigorous updraft will maximise the mass of cloud hydrometeors supported and detrained during boundary erosion.

1. The VIMHEX Project.

Opportunity to investigate the cumulonimbus as a heat source or sink to the atmosphere was given during the Venezuelan International Meteorological and Hydrological Experiment (VIMHEX) conducted under the direction of Professor Herbert Riehl at Anaco, Venezuela (location, figure 2) during the summer of 1969. Part of the experiment involved the release of radiosonde balloons during periods of disturbed weather

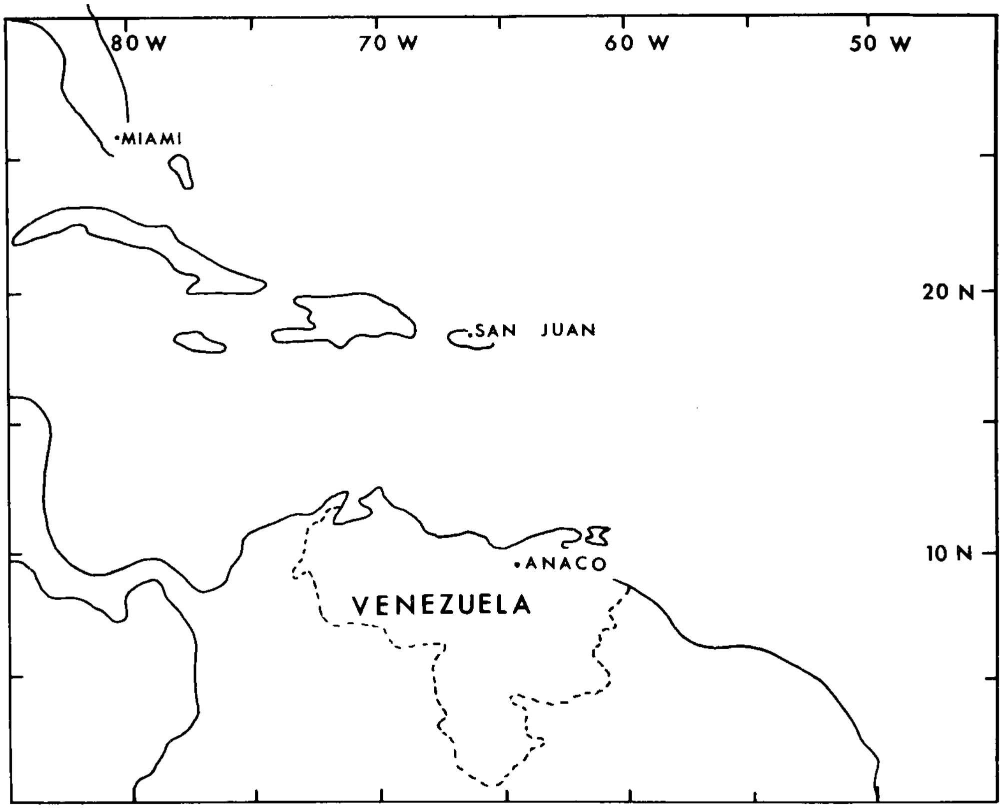


Figure 2. Location diagram - Anaco, Venezuela.

while simultaneously monitoring the location of convective systems by the position of their radar images.

The convective activity at Anaco followed a diurnal trend. The peak of activity was during the middle to late afternoon. Convection rarely started before noon and generally the echoes were dying at sunset.

A sounding was regularly released near 0730 LST each day. As echoes began to appear on the radar scope radiosondes were released at approximately two hour intervals until the cessation of convection during the evening. This plan made for a random scatter of the balloons relative to the echo centers.

The 10 cm wavelength radar used had adjustable antenna tilt and signal attenuation facilities. These allowed for examination of the height of the radar echo and the reflectivity (hydrometeor concentration) of the echo. Following the commencement of convection a sequence of photographs was made of the radar scope at ten minute intervals. This sequence consisted of the four attenuation settings (18 db, 12 db, 6 db, and zero attenuation respectively) at each of the antenna elevation settings (18, 16, 14, 12, 10, 8, 6, 4, 2, and zero degrees elevation respectively). Where there were no visible echoes at the higher elevations no photographs were made.

The primary objective of the experiment was to relate the radiosonde data in a coordinate system relative to the center of, and in the direction of movement of, the mesoscale convective systems in composite form. This is possible by the accurate tracking of the radiosonde balloons and by noting the changing position and intensities of the cloud images as the balloon ascends through the troposphere.

During the three months operation of the experiment a number of balloons were released into and in close proximity to mesoscale convective systems. The following is a series of analyses of data from such radiosondes.

The radar scope photographs have Anaco at the center of the scope. This is also the point of release of the balloons. The elevation of the radar antenna is always at the lowest setting while the signal is unattenuated. North is to the top of the pictures and the persistent echoes in the arc from north to north-east are the mountains near the coast. The inner circle is of 50 Km radius while the outer directional scale is of 130 Km radius.

The atmospheric soundings are plotted on a tropical tephigram, the ordinate being potential temperature and the abscissa being actual temperature. For each sounding the upper plot is the actual temperature against pressure while the lower plot is the dew point temperature against pressure.

It is doubtful that the cloud wall and the edge of the radar echo are coincident. Thus, while the radiosondes were released at the relative position of the scope center to the echo, there is no indication of the relative position of the cloud edge. However, from the proximity of the radar echoes and the deviation of the mixing ratio (or dew point temperature) it is possible to ascertain in-cloud and out-of-cloud conditions.

In each analysis the modification of the airflow by the cloud is shown. The early sounding prior to convection is representative of the undisturbed airflow and the following sounding is indicative of the modification taking place near the cloud edge. The later

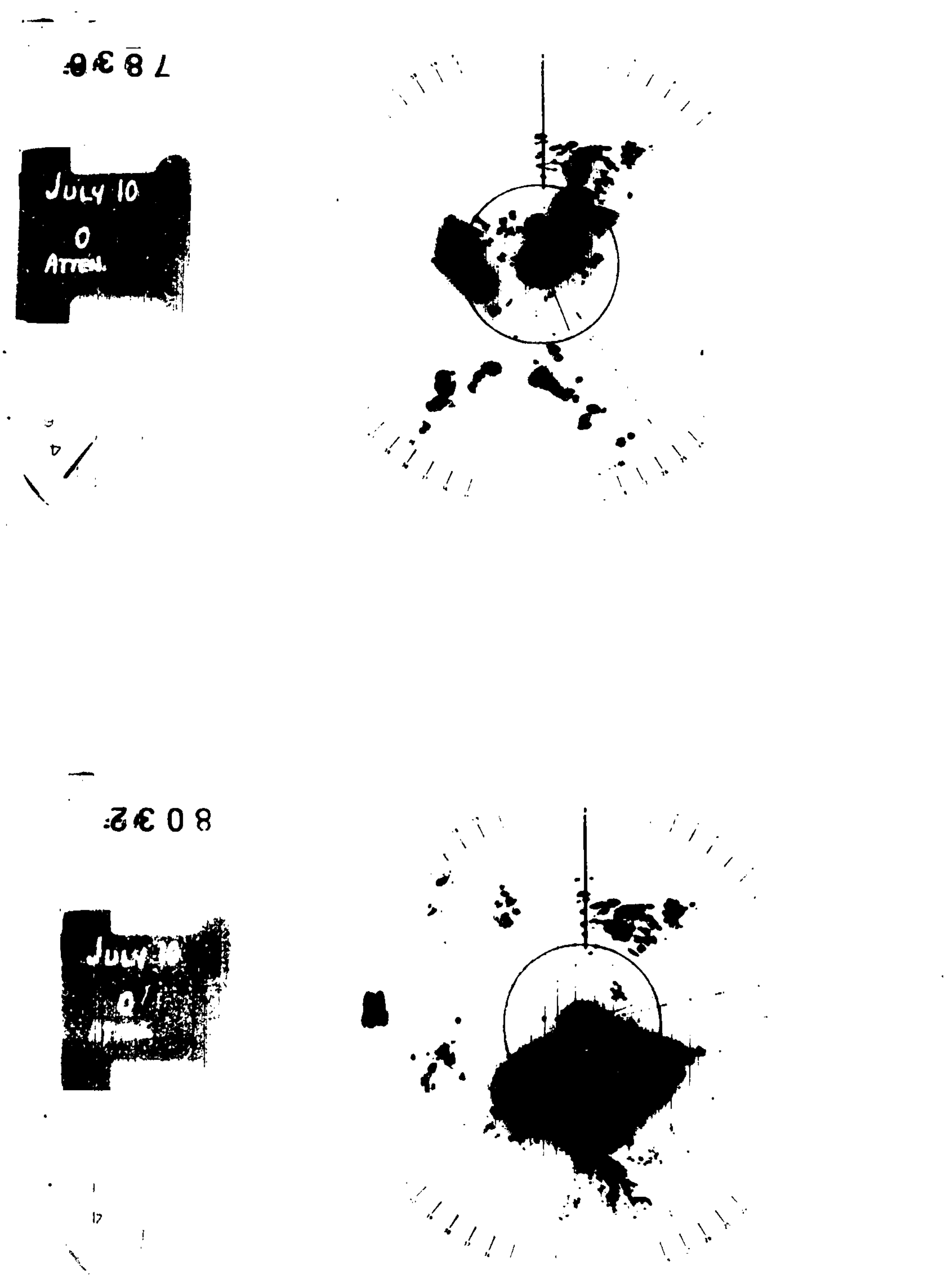


Figure 3. Radar scope photographs at Anaco; a) 1506 LST, b) 1800 LST. 10 July 1969.

soundings demonstrate that the cumulonimbus can act as an atmospheric heat sink as a result of the conversion of enthalpy of the cloud environment to latent energy during the evaporation of cloud hydrometeors.

10 July 1969: Following the passage over Anaco of a surface easterly wave during the evening of 9 July, convective activity was considerably enhanced during the afternoon of 10 July. Frequent radiosondes were released during the afternoon and evening. Figures 3a and 3b are photographs of the radar scope at 1506 LST and 1800 LST respectively as radiosondes were released. An intermediate radiosonde was released at 1645 LST when there were no echoes near Anaco. This balloon prematurely burst at 500 mb. The sounding of 500 mb was similar to the 1500 LST release. The soundings of 1500 LST and 1800 LST are compared in figure 4.

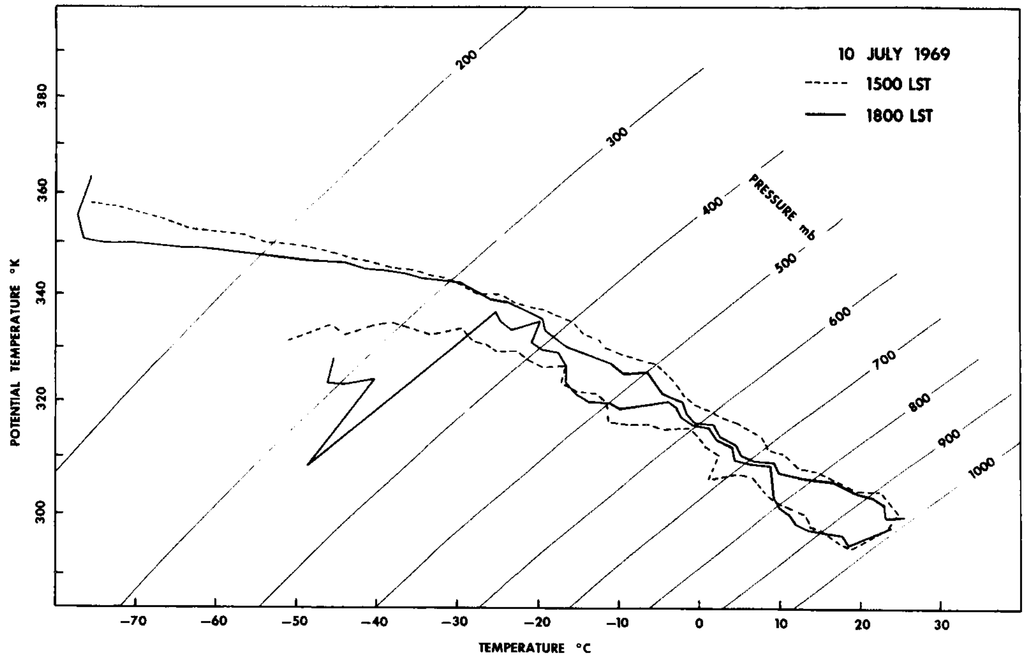


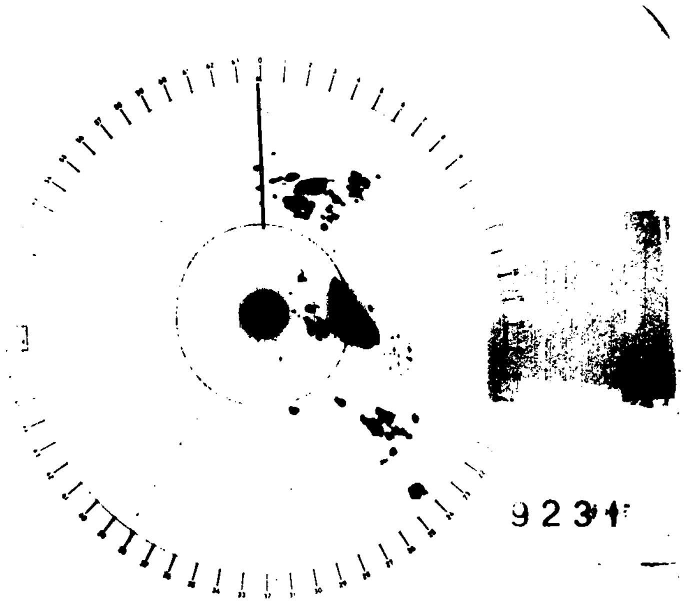
Figure 4. Comparison of atmospheric soundings at Anaco. 10 July 1969.

The radar echoes during the afternoon of 10 July were consistently moving west northwest. The arc of echoes to the south of Anaco moved northwest while developing to the mesoscale system observable at 1800 LST.

The cooling within the large mesoscale system to the south of Anaco which is observed from 800 mb to 400 mb is consistently within the range of from one to two centigrade degrees. The balloon entered the cloud below 750 mb and remained within the cloud to at least 350 mb. It is only near 300 mb that the temperatures are comparable with the 1500 LST sounding. Above 300 mb cooling is again registered, with maximum values near the well-defined tropopause. It is likely that overshoot and mixing of the now relatively cold and dense updraft is responsible for the high level cooling.

12 July 1969: The initial sounding (0810 LST) showed a cool moist layer below 700 mb with near normal temperatures above. The first signs of convective activity were to the south of Anaco where early, but small, echoes began to appear. By 1130 LST a mesoscale system had developed to the southeast of Anaco at a distance of near 60 Km (figure 5a). At this time the radiosounding was similar to the earlier sounding indicating little, if any, downwind influence by the convective system.

The convective system became more active and enlarged during the following two and a half hours as it moved toward Anaco. The convective system was still to the east of Anaco when the subsequent sounding was made at 1400 LST (figure 5b). This and the 1130 LST sounding are compared (figure 6) to show the thermal modification in the cloud air traversed by the balloon.



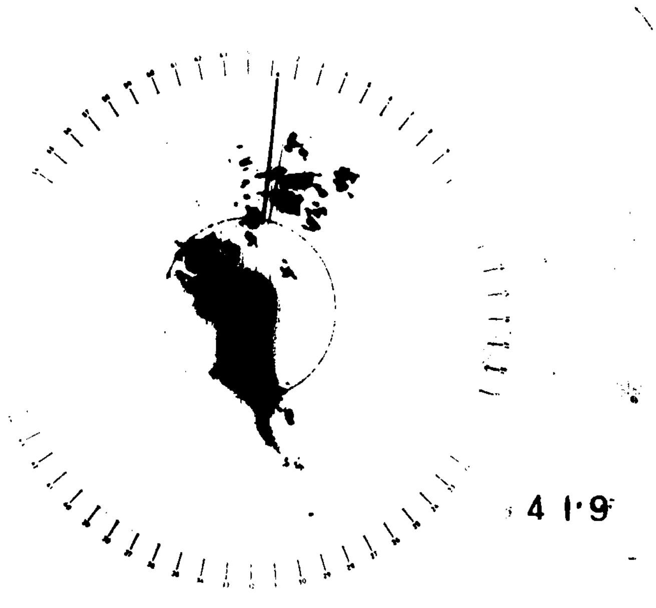


Figure 5. Radar scope photographs at Anaco; a) 1130 LST, b) 1405 LST. 12 July 1969.

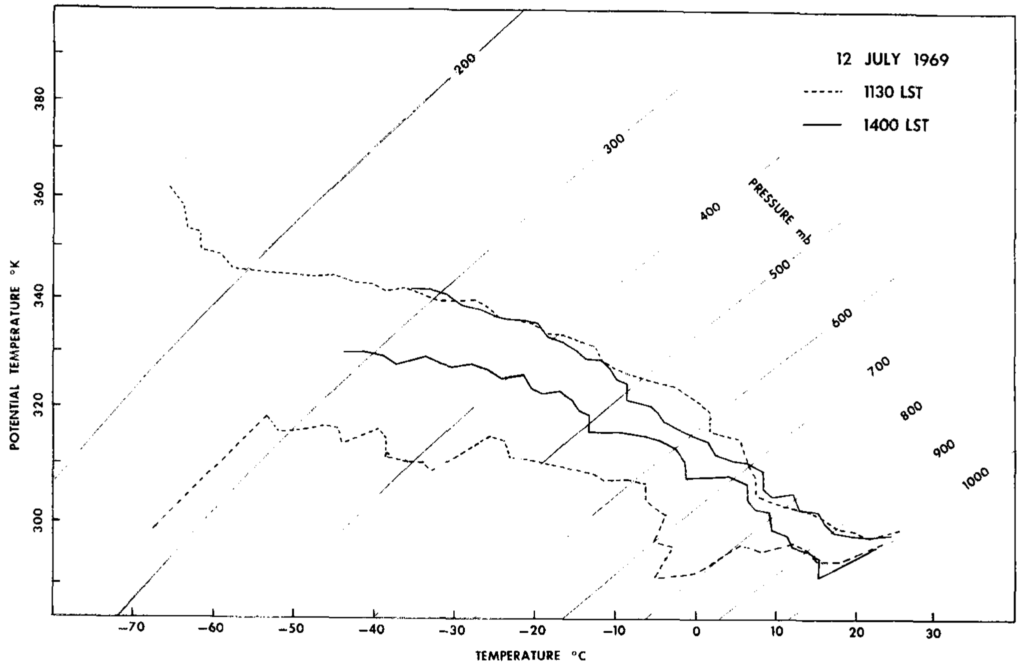


Figure 6. Comparison of atmospheric soundings at Anaco. 12 July 1969.

Between 850 mb and 700 mb the cold anomaly present on the earlier sounding tends to be eliminated at 1400 LST. Between 700 mb and 400 mb, when the relative humidity continues to be high, strong cooling is evident at the later time. Above 450 mb the temperature is near the environment temperature while the dew point depression is small. It is probable that the balloon remains within the cloud.

The weak warming evident in the lower layers at 1400 LST is significant because it shows that, if the temperature excess of the updraft is sufficiently large, then the cumulonimbus can act as a heat source with respect to its environment. In fact a later sounding at 1545 LST shows the cold anomaly of the morning sounding to be completely eliminated as a result of the afternoon convection. At

1400 LST the sounding shows the cloud continuing to modify the environment with maximum cooling at 550 mb and decreasing to no significant value when the balloon burst at above 300 mb.

<u>12 August 1969:</u> Very little convection was being experienced about Anaco on 12 August. During the afternoon a cluster of cells formed to the north and east of Anaco and moved westward. A radiosonde was released as the edge of the main body of radar echoes was a few kilometers to the north of Anaco (figure 7). The near saturation in the lower layers of the 1630 LST sounding (figure 8) indicates that the balloon entered the cloud through the base and emerged into clear air below 650 mb. The wind profile indicated that the balloon emerged from the cloud to the southwest of the system.

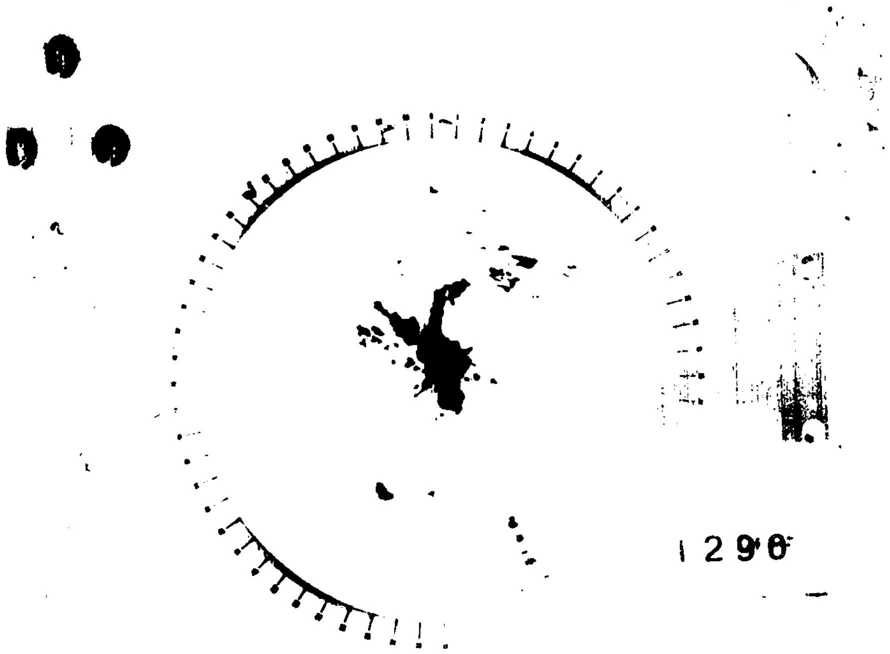


Figure 7. Radar scope photograph at Anaco. 1645 LST, 12 August 1969.

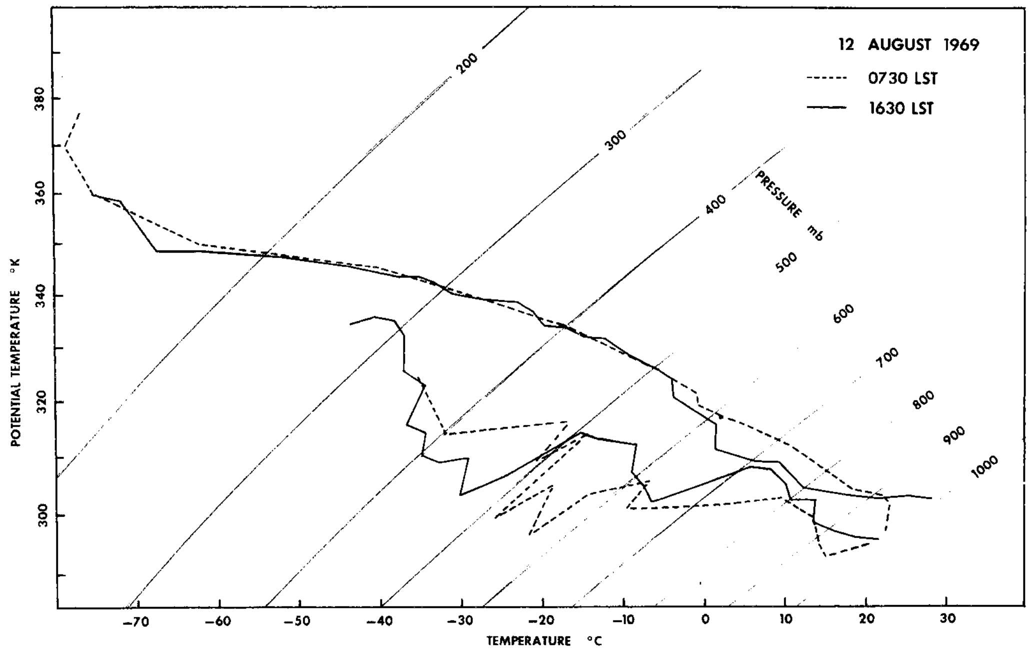


Figure 8. Comparison of atmospheric soundings at Anaco. 12 August 1969.

A comparison of the 1630 LST sounding and the morning sounding given in figure 8 shows cooling in the cloud maintained to 550 mb and after the balloon has apparently emerged from the cloud air. The temperature and dew point profiles above 500 mb are in general agreement between the morning and the afternoon soundings. Cooling above 200 mb and the tendency for the formation of a second tropopause at 150 mb is noticeable on the afternoon sounding. Above 135 mb the 1630 LST sounding resumed the profile of the morning.

<u>13 August 1969:</u> Enhanced convective activity resumed on the afternoon of 13 August. The activity was associated with the approach of an upper level trough from the east. A radiosonde was released at 1200 LST on the right rear flank of a decaying convective system

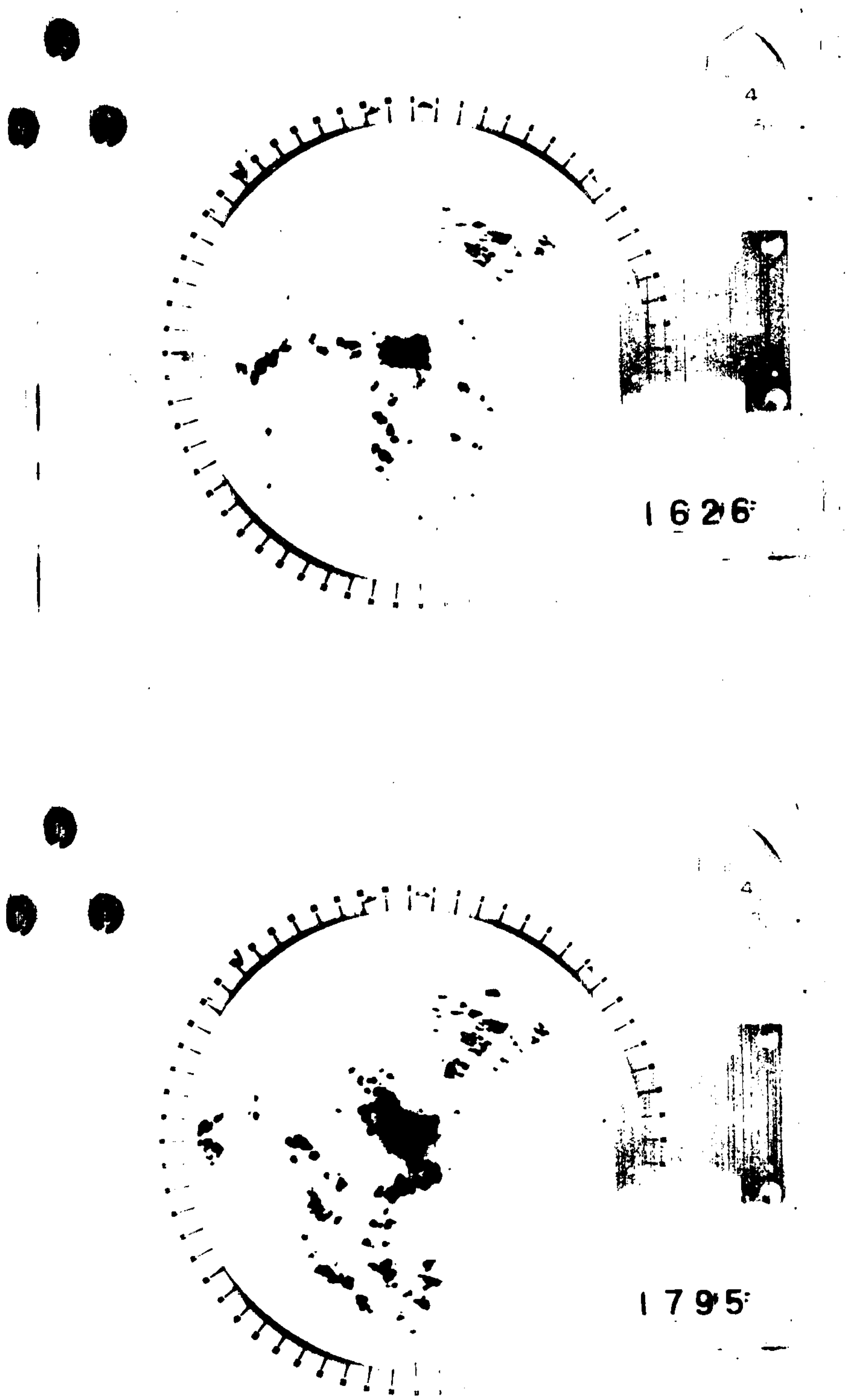


Figure 9. Radar scope photographs at Anaco; a) 1200 LST, b) 1500 LST. 13 August 1969.

(figure 9a). The radiosonde, being upwind of the convective system, did not pass through air which had been thermally modified by the convective system. The following radiosonde at 1500 LST was released in approximately the same relative position to an active system (figure 9b) as the former radiosonde was to the decaying system. This later balloon was, by comparison, entrained into the cloud air and significantly influenced by the mesoscale system.

The comparison of the 1200 LST and the 1500 LST soundings (figure 10) is evidence of the strong modification to the entrained environment on the upwind side of the cloud. Cooling is again indicated to near 420 mb where, prior to the later balloon bursting, the temperature profile corresponds to the earlier sounding. The

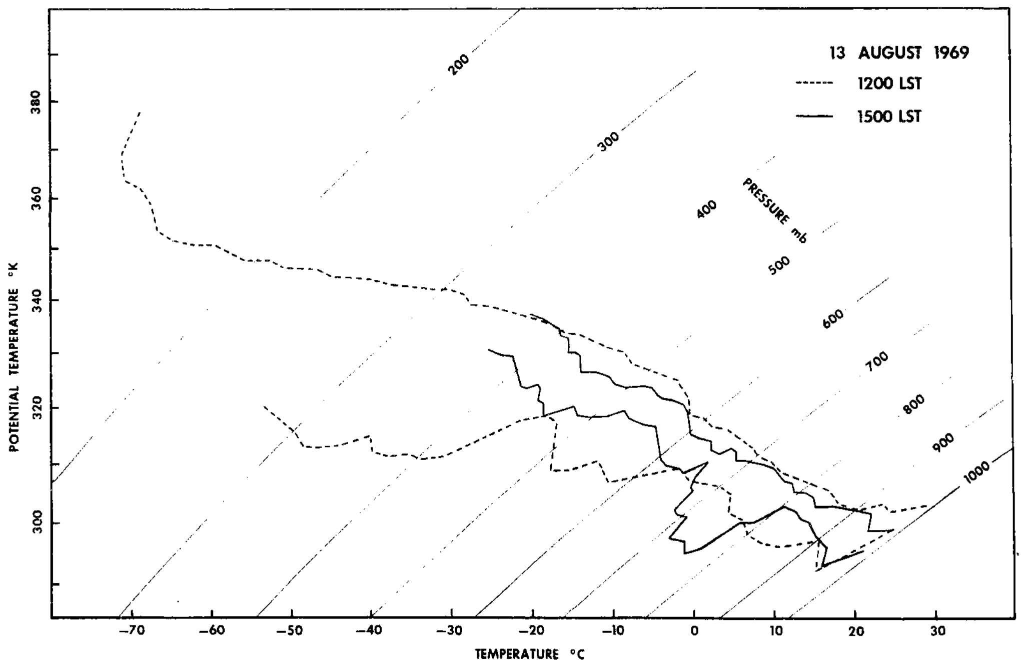


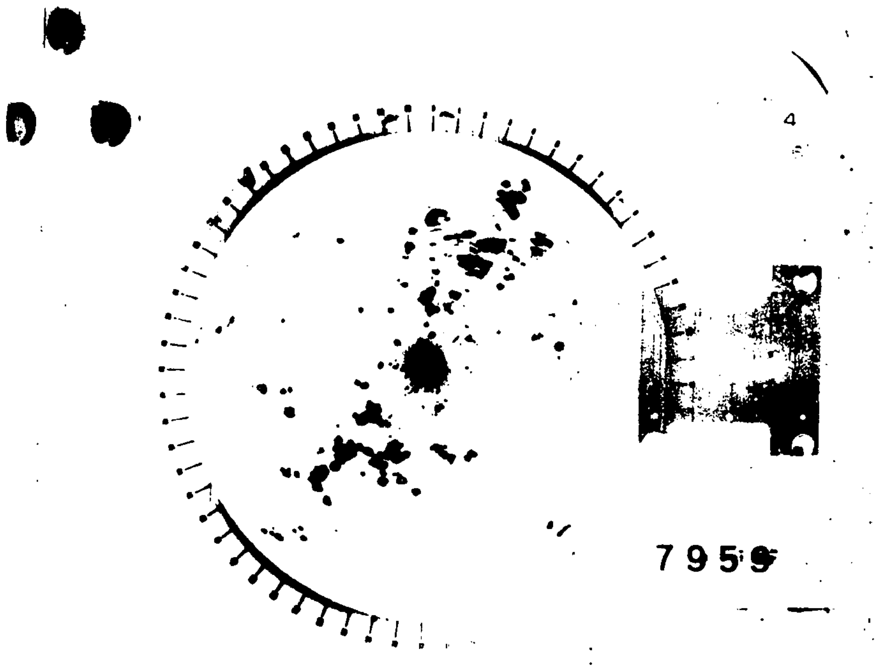
Figure 10. Comparison of atmospheric soundings at Anaco. 13 August 1969.

dew point profile indicates that the balloon entered the cloud near 850 mb and generally remained within the cloud. Although there is a general cooling to 400 mb there are rapid fluctuations in the temperature profile. It is probable that the balloon is oscillating across equivalent potential temperature surfaces and that the horizontal in-cloud temperature gradients result in this profile.

11 September 1969: The convective activity observed on the radar scope was again enhanced on 11 September. Echoes were first appearing on the radar scope before noon and continued into the evening.

A radiosonde was released at 1100 LST and the radar showed no echoes in the near vicinity of Anaco. The following radiosonde release was at 1330 LST as a small convective system passed to the north and east of Anaco (figure 11a). With a prevailing easterly flow the modified air from the convective system was intersected by the balloon during its ascent from 850 mb to 450 mb. The sounding (figure 12) shows cooling (with reference to the 1100 LST sounding) from 850 mb to 450 mb. The tendency for convective overshoot is indicated above 200 mb.

The following radiosonde was released at 1535 LST into a large mesoscale convective system (figure 11b). Even stronger cooling than on the 1330 LST sounding is evidenced at the 600 mb level and between 500 mb and 450 mb. The 1330 LST and 1535 LST soundings are compared in figure 13. The sounding at 1535 LST is nearly saturated throughout indicating that the radiosonde remained within the cloud during ascent. The 1535 LST radiosonde was apparently oscillating in the horizontal between a cold anomaly toward the edge of the cloud and a buoyant updraft region.



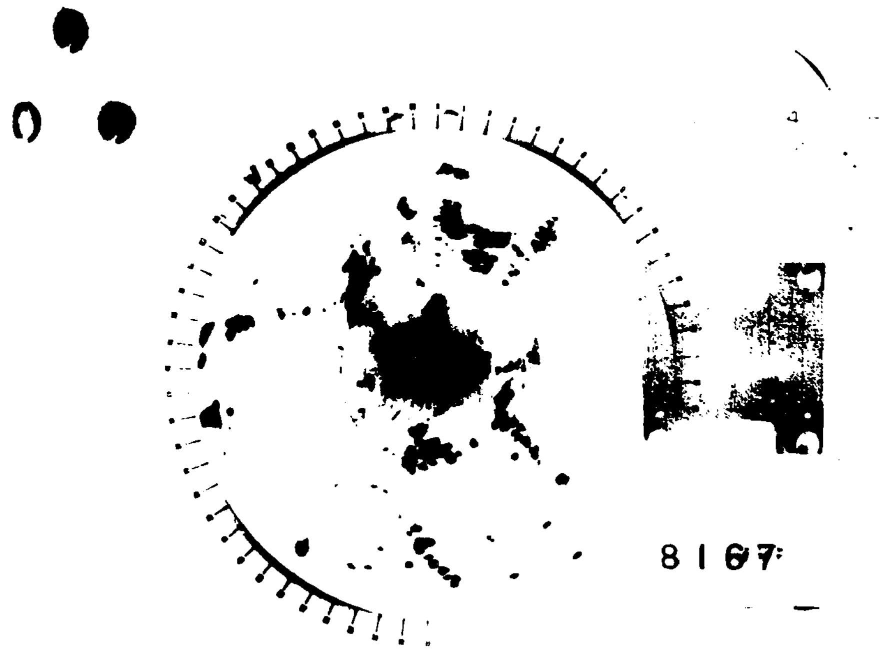


Figure 11. Radar scope photographs at Aanco; a) 1330 LST, b) 1535 LST. 11 September 1969.

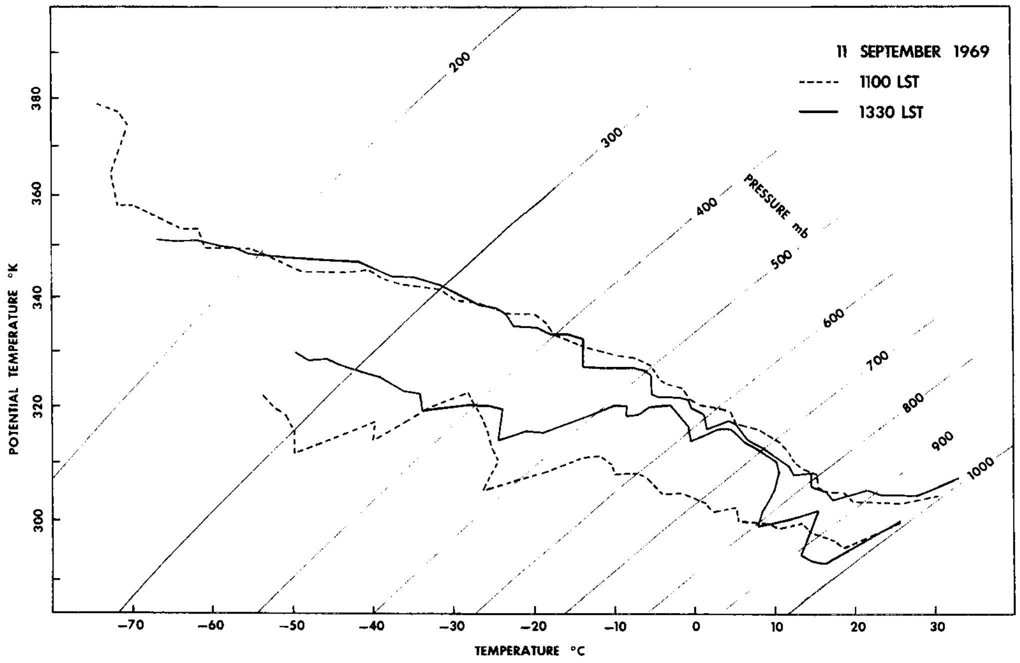


Figure 12. Comparison of atmospheric soundings at Anaco. 11 September 1969.

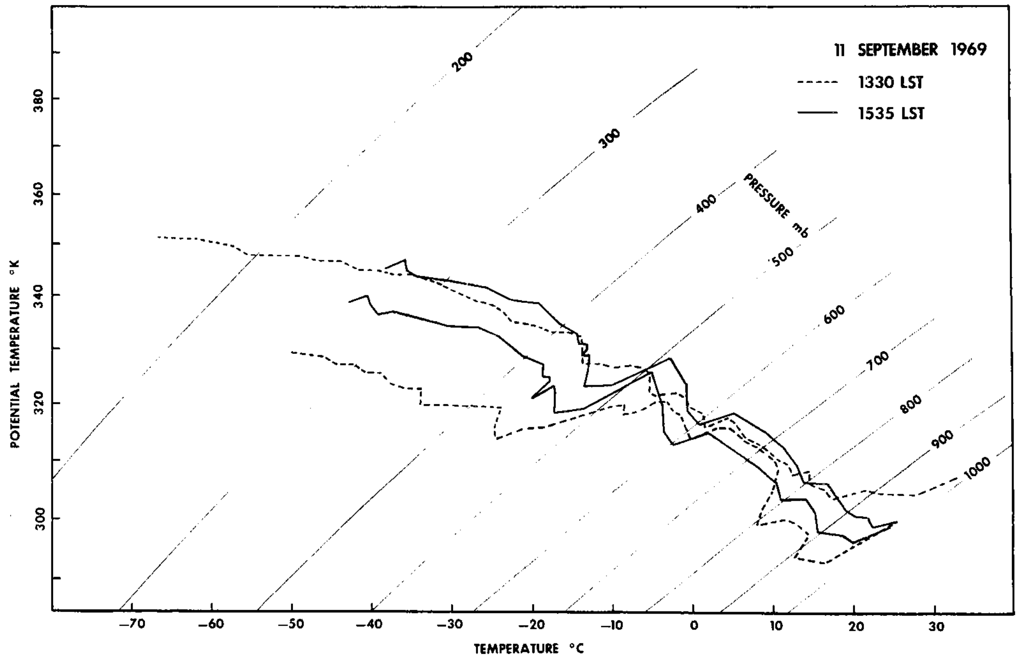


Figure 13. Comparison of atmospheric soundings at Anaco. 11 September 1969.

A common characteristic of the above thermally modified soundings of the atmosphere is the tendency for cooling in the lower and middle troposphere (with a cold anomaly maximum near 500 mb), the resumption of the original profile near 200 mb, and followed by another layer of cooling as apparently the buoyant updrafts overshoot the equilibrium level and tend to mix with the prevailing airflow.

2. The Convective Model.

The soundings examined for five of the disturbed days at Anaco will now be used to test the model of horizontal mixing as a viable mechanism of atmospheric modification. The five soundings prior to the onset of convection (1500 LST, 10 July; 1130 LST, 12 July; 0730 LST, 12 August; 1200 LST, 13 August; and 1100 LST, 11 September) have been meaned at 50 mb intervals to give a representative sounding of the undisturbed environment. The five following soundings when radiosondes were launched into and about mesoscale convective systems (1800 LST, 10 July; 1400 LST, 12 July; 1630 LST, 12 August; 1500 LST, 13 August; and 1330 LST, 11 September) are considered representative of the modified atmosphere near the cloud edge.

Through the use of equation 7 for the determination of the cloud edge temperature, the sounding representing the undisturbed environment, and a suitable updraft profile it is now possible to determine a vertical profile of the modified atmosphere at the cloud edge. For this illustration the cloud edge is assumed to be the vertical surface on which the cloud hydrometeor mixing ratio goes to zero and the relative humidity is 90%. The iteration process previously described was carried out until succeeding estimates of the cloud edge temperature at each 50 mb level differed by less than 0.05 centigrade degrees.

The radar echoes of the particular convective systems were large. It is considered that the updraft centers of these systems can be approximated by undilute ascent from the boundary layer. Noting that the actual modified sounding approximated the undisturbed sounding near 200 mb and on the 347° K equivalent potential temperature surface (figure 14), then 347° K was taken as the pseudo-adiabat of undilute ascent. This pseudo-adiabat is thermally buoyant from below 850 mb to 200 mb.

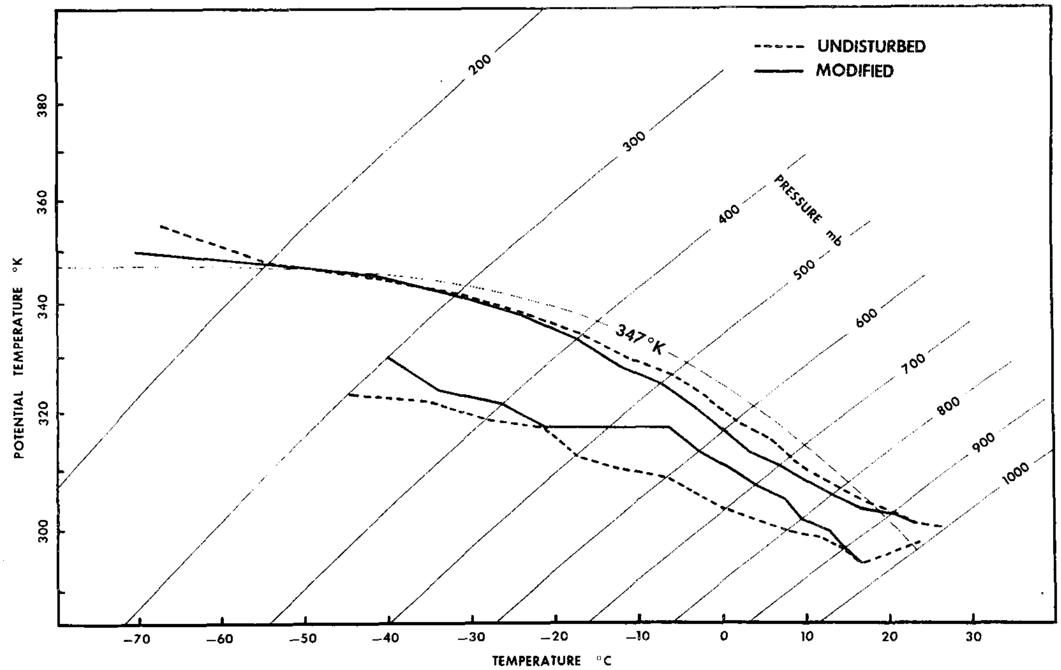


Figure 14. A comparison of the mean of five soundings prior to the onset of convection with the mean of five following soundings made near the cloud edge of mesoscale systems.

Estimating the updraft hydrometeor mixing ratio which is eroded into the environment is difficult. An upper limit would be less than 10 g/Kg, the maximum predicted by numerical models and also observed by research flights. For this model the updraft hydrometeor mixing ratio which is eroded and evaporated into the environment is taken as the profile of figure 15. The effective hydrometeor mixing ratio increases sharply to 600 mb, representing the increased holding capacity of the updraft with increased vertical velocity. The effective hydrometeor mixing ratio then falls from the maximum of 6 g/Kg to insignificant values above 200 mb. The profile of Figure 15 may be regarded as the effective hydrometeor mixing ratio for the purposes of this model. The actual hydrometeor mixing ratio of the updraft is in excess of this, the difference falling out as precipitation.

Using the undisturbed environment profile of figure 14, the effective updraft hydrometeor mixing ratio of figure 15, and the updraft pseudo-adiabat of 347° K equivalent potential temperature, the cloud edge temperature profile was calculated. The difference between the calculated cloud edge temperature profile and the undisturbed environment temperature profile is the temperature deviation, or anomaly, at the cloud wall.

The calculated temperature anomaly at the cloud edge is compared to the actual anomaly of the modified air near the cloud edge in figure 16. It is readily apparent that the model of horizontal mixing from the eroding updraft can reproduce the observed modification brought about on the edges of mesoscale convective systems.

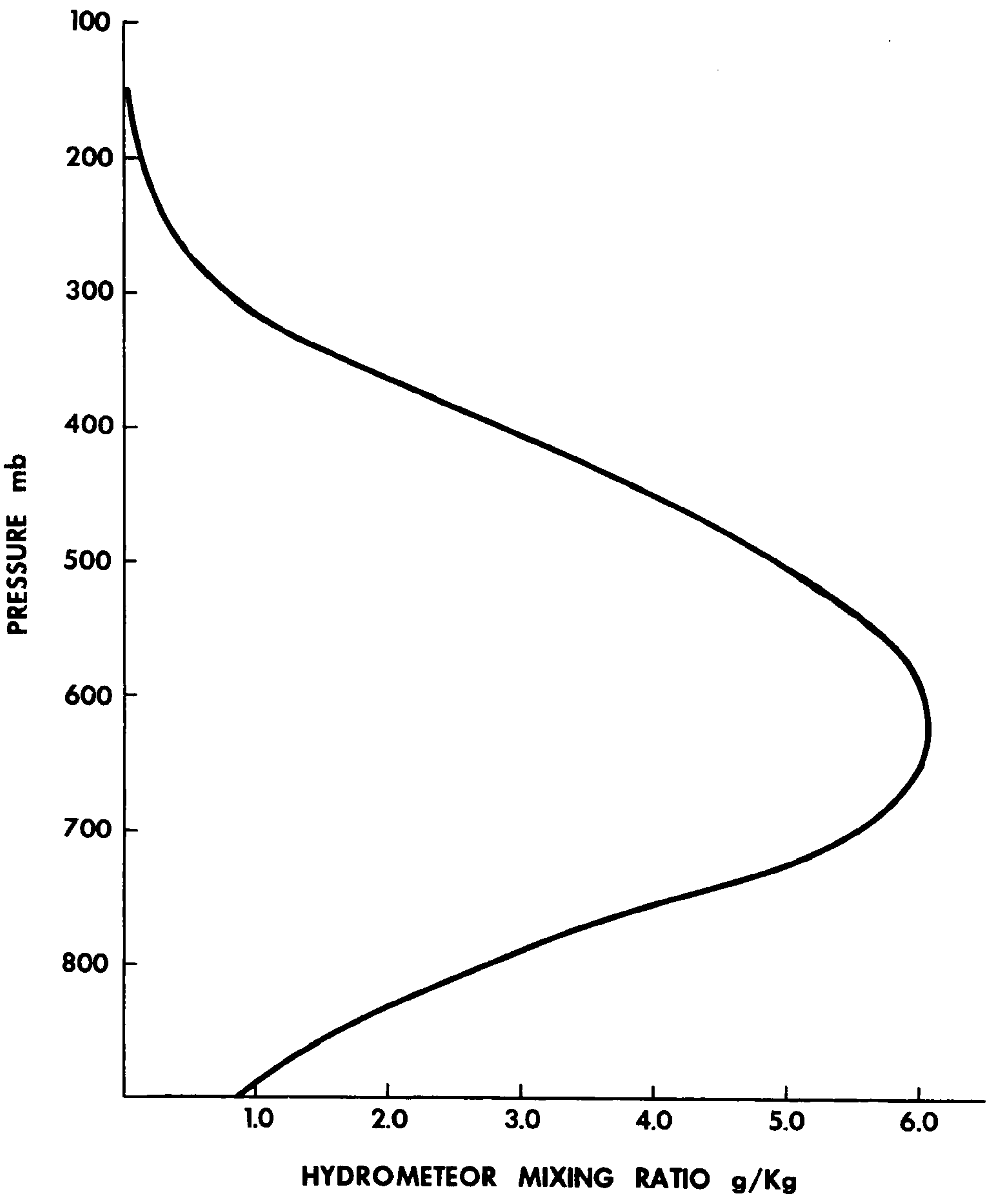


Figure 15. The effective updraft hydrometeor mixing ratio of the mesoscale convective clouds.

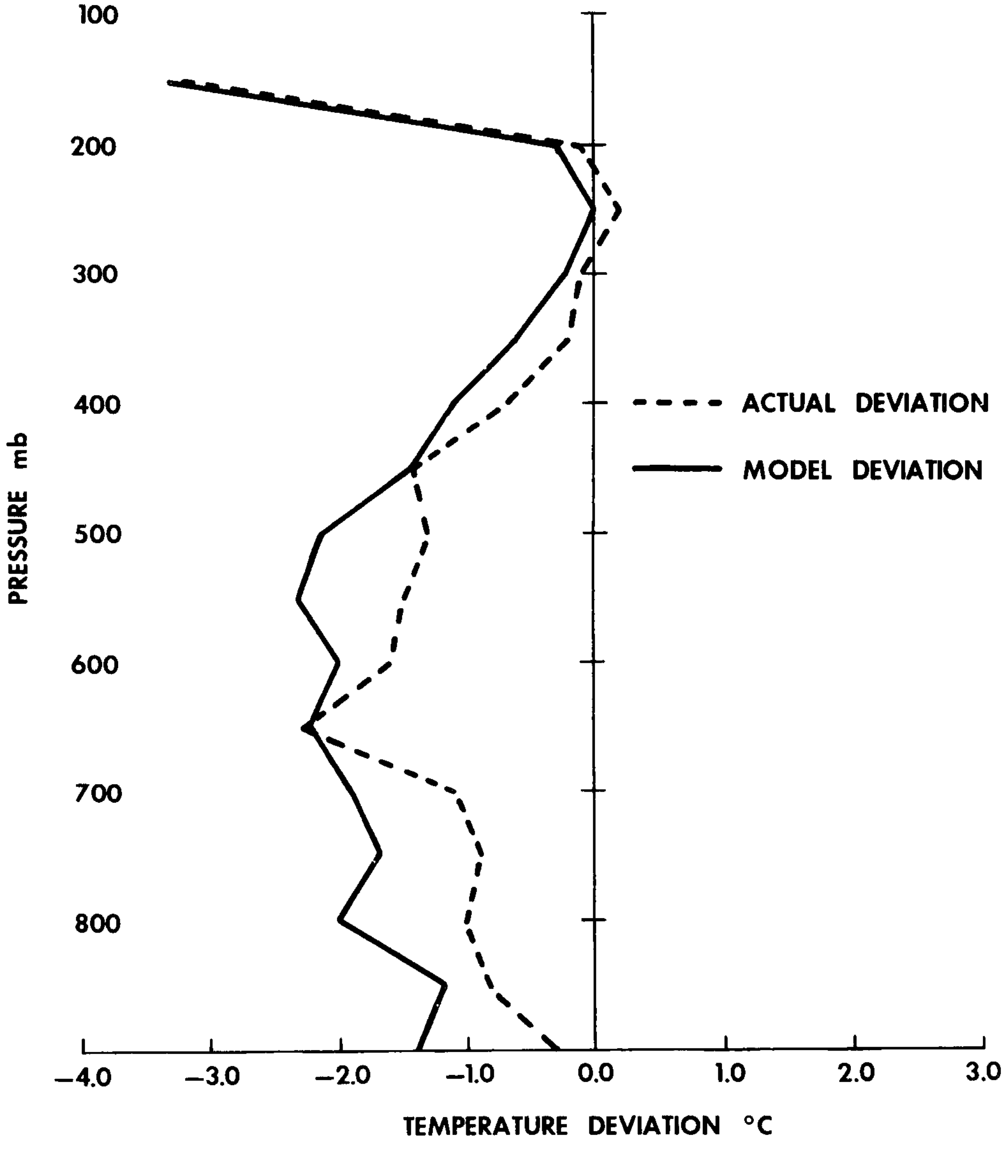


Figure 16. A comparison of the mean deviation of the temperature observed near the cloud edge and the temperature deviation predicted using the horizontal mixing model.

Meso and Synoptic Scale Interactions

Typically the surface equivalent potential temperature at Anaco, Venezuela was in excess of 360° K during the afternoon and preceding convection. The vertical gradient of moisture and temperature was such that at 950 mb a typical value of the equivalent potential temperature was less than 335° K. Under these boundary layer conditions, except for a very narrow and weak layer near 250 mb, the model and the actual data show the convective clouds to be a sink of sensible heat to the atmosphere.

Over tropical oceans the summertime equivalent potential temperature at the surface is typically near 355° K (Gray, 1967). With turbulent mixing and an abundant heat and moisture supply from the ocean's surface, the mean sub-cloud layer equivalent potential temperature will be near 350° K. In regions of synoptic scale low level convergence the mean sub-cloud layer equivalent potential temperature will approach 355° K as the high energy surface air is lifted. Typically, the mean equivalent potential temperature of the sub-cloud layer will rise with the increase of surface convergence during the intensification of a tropical disturbance.

It is of interest to extend the model of horizontal mixing from the convective updraft to determine the cloud edge temperature modification resulting from changes in the sub-cloud layer equivalent potential temperature. This changing sub-cloud layer equivalent potential temperature will be noticed within the cloud as a change in the equivalent potential temperature of the updraft pseudo-adiabat.

The horizontal mixing model has been applied using the same undisturbed environment (figure 14) and effective updraft hydrometeor mixing ratio (figure 15) as used previously. Updraft pseudo-adiabats of 347° K, 351° K and 355° K reflect the changing mean equivalent potential temperature of the sub-cloud layer with disturbance intensification. Equation 7 was used to calculate the cloud edge temperature (as previously) at 50 mb intervals between 900 mb and 150 mb for the three sub-cloud layer conditions.

The vertical expansion or contraction of an atmospheric layer between pressure surfaces is dependent on the layer mean mixing ratio as well as the layer mean temperature. The effect of the moisture is accounted for if the virtual temperature, T*, is used in calculating the layer thickness.

$$T^* = (1 + 0.61q)T \ {}^{o}K \qquad (10)$$

From the cloud edge temperature profiles calculated using the 347° K, 351° K and 355° K updraft pseudo-adiabats the cloud edge virtual temperature profiles were evaluated. The deviation of these virtual temperature profiles from that of the undisturbed environment is shown in figure 17.

The layer thickness, Δz, between 50 mb increment pressure surfaces is given by:

$$\Delta z = \frac{R}{g}\bar{T}^* \mathrm{Ln}\left(\frac{p + 25}{p - 25}\right) \ \mathrm{cm} \qquad (11)$$

where $\bar{T}^*$ is the layer mean virtual temperature and p is the mid-layer pressure surface. The virtual temperatures previously calculated at 50 mb intervals for the 347° K, 351° K and 355° K updraft pseudo-adiabats were used as the mid-layer (approximately layer mean) virtual temperature in the calculation of the cloud edge thickenss of each

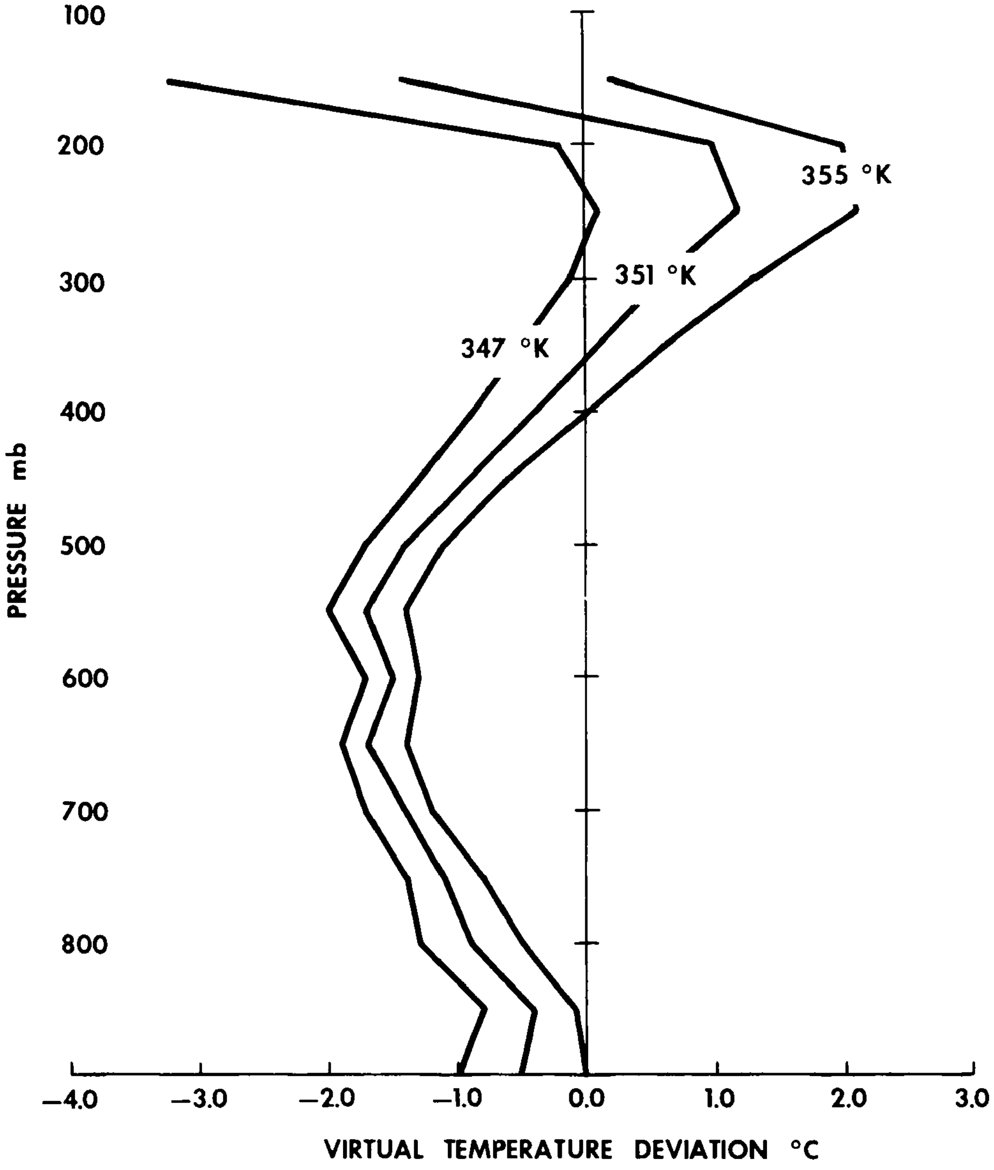


Figure 17. The virtual temperature deviation at the cloud edge predicted for varying updraft pseudo-adiabats.

layer for the three updraft conditions. The profiles of the deviations of the thickness of these 50 mb layers from the thickness of the corresponding layer of the undisturbed environment are shown in figure 18.

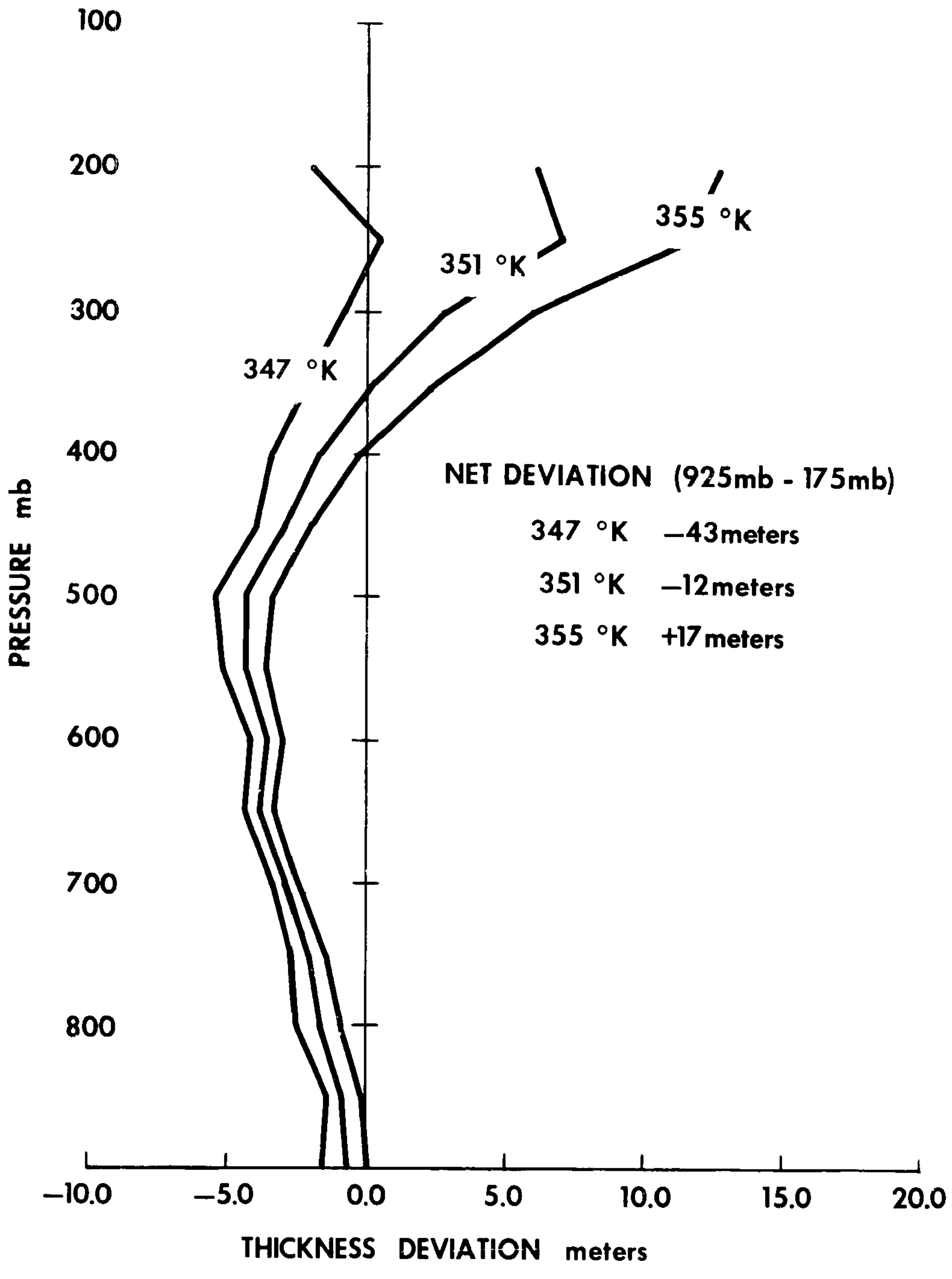


Figure 18. The predicted thickness deviation of 50 mb layers at the cloud edge for varying updraft pseudo-adiabats.

The main variation of virtual temperature, and consequently thickness, occurred near 250 mb. Where cooling was at all levels except 250 mb with a 347° K equivalent potential temperature updraft, as the updraft equivalent potential temperature increased warming was apparent from 350 mb to 200 mb for the 351° K pseudo-adiabat and from 400 mb to 150 mb for the 355° K updraft pseudo-adiabat.

Consequently the net thickness change between 925 mb and 175 mb varied from -43 meters for the 347° K updraft, through -12 meters for the 351° K updraft, to 17 meters for the 355° K updraft. As the sub-cloud layer, and hence updraft, equivalent potential temperature increases beyond 355° K the horizontal mixing model shows the meso-scale convective clouds to increase in effectiveness as a net heat source to the atmosphere.

Whereas the convective clouds of the rain areas of the disturbances over eastern Venezuela are a sink of heat throughout the troposphere, the rain areas of the disturbances of the oceanic areas of the tropics are colder in the lower and middle troposphere and warmer above. The model of convective mixing can reproduce the cold anomaly in the vicinity of the clouds of eastern Venezuela and predicts the formation of a warm anomaly in the high troposphere about the clouds of weak disturbances.

Yanai (1963), in a study of tropical disturbances of the Pacific, found that those disturbances which subsequently develop to Typhoon intensity have an upper level warm anomaly. During the transformation process from a cold core disturbance to a warm core disturbance this warm anomaly increases in intensity and propogates downward (Yanai, 1964).

The model of convective mixing shows the clouds to be a heat source in the upper troposphere when they located over a mean sub-cloud layer equivalent potential temperature typical of a weak disturbance. As the disturbance intensifies and the surface pressure lowers, the mean sub-cloud layer equivalent potential temperature will increase due to isothermal expansion of the low level airflow. Continuing the trend shown by the model, the upper levels of the convective clouds will increase in intensity as a heat source to their environment and this heat source will propogate downward with the intensification of the disturbance.

Conclusion

The principle of erosion and mixing of updraft air into the environment has been shown to be capable of explaining the observed temperature anomalies near the cloud wall of mesoscale convective clouds at Anaco, Venezuela.

The model of horizontal mixing between the convective updraft and its environment can readily be extended for application over tropical oceanic areas. Here the mean sub-cloud layer equivalent potential temperature is higher than that of Anaco, Venezuela. The model shows the clouds to act as a sink of heat in the lower and middle troposphere, and a source of heat above. The heat exchange is acting in exactly the same sense as is observed in weak tropical disturbances. With intensification of the disturbance the model indicates that the heat source in the upper regions of the convective clouds intensifies and propogates downward.

It has been shown by the model that the heat sources and heat sinks of the convective clouds are at least acting in sympathy with the thermal field of the tropical disturbances in which they are imbedded.

While it is suggested that the heat sources and heat sinks of the convective clouds as described by the model are the controlling factors of the thermal fields of the disturbances in which the clouds are imbedded, further studies are required to verify this hypothesis. A full analysis of the data obtained during project VIMHEX will give the extent of the thermal modification about the clouds of eastern Venezuela. Synoptic studies will be required to incorporate the model in the framework of the dynamics and thermodynamics of synoptic tropical disturbances.

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