# Diurnal Variation of Oceanic Deep Cumulus Convection

## Paper I: Observational Evidence

By
R.W. Jacobson Jr.

## Paper II: Physical Hypothesis

By
W.M. Gray

Department of Atmospheric Science
Colorado State University
Fort Collins, Colorado

Preparation for this report has been financially supported by NOAA National Environmental Satellite Laboratory Grant No. 04-3-158-51.
February 1976

Colorado State University

Department of Atmospheric Science

Paper No. 270

DIURNAL VARIATION OF OCEANIC DEEP CUMULUS CONVECTION

PAPER I: OBSERVATIONAL EVIDENCE

by

Robert W. Jacobson, Jr.

# DIURNAL VARIATION OF OCEANIC DEEP CUMULUS CONVECTION

PAPER I: OBSERVATIONAL EVIDENCE
By Robert W. Jacobson, Jr.

PAPER II: PHYSICAL HYPOTHESIS
By William M. Gray

Preparation of this report

has been financially supported by

NOAA National Environmental Satellite Laboratory

Grant No. 04-3-158-51

Department of Atmospheric Science

Colorado State University

Fort Collins, Colorado

February 1976

Atmospheric Science Paper No. 243

ABSTRACT

This is the first of two papers on the diurnal variability of oceanic deep cumulus convection. This paper presents observational evidence in support of the existence of a significant morning maximum and afternoon-evening minimum of oceanic deep convection. The second paper of this report advances a physical hypothesis for this diurnal variation.

This diurnal variation in deep cumulus convection has been detected from upper air, rainfall, satellite, and radar data over oceanic and many land regions. Diurnal variations in low level, layered, and total cloudiness are very much smaller than variations of deep cumulus.

Composited upper air wind data relative to satellite observed cloud clusters and clear regions in the West Pacific has yielded divergence-convergence patterns which are substantially larger at 00Z (∿ 10 L.T.) than at 12Z (∿ 22 L.T.). Thirteen years of hourly rainfall data for eight tropical Pacific atolls and small islands, plus six-hourly rainfall data for shorter time periods from nine other atoll-island stations in this same region show that heavy rain showers occur much more frequently in the morning hours than in the afternoon or early evening. Two months of DMSP satellite photographs (∿ 1/3 n.m. resolution) also indicated increases in convective activity between 22 L.T. and 10 L.T. with decreases between 10 L.T. and 22 L.T. Other supporting evidence is presented and discussed.

The diurnal variability of deep convection is not readily observable from satellite pictures which cannot resolve individual convective cells, or from surface observations of percent sky coverage which are heavily weighted to the presence of low level and layered clouds.

# 1. BASIS FOR THE PRESENT STUDY

In a study of tropical oceanic cloud clusters (for definition see GARP report, 1968) Ruprecht and Gray (1974) found large diurnal differences in convective rainfall for the five hour periods centering on 00Z ($\sim$ 10 L.T.) and 12Z ($\sim$ 22 L.T.). They reported over twice as much rainfall from cloud clusters in the morning (07-12 L.T.) than in the evening (19-24 L.T.). The diurnal variation of the heaviest rainfall was most pronounced (see Table 1). Such large variations were quite unexpected. Heavy rain from the cloud clusters (defined as being >1.0 cm per hour) was nearly three times as prevalent in the morning. It was shown that as the precipitation intensities became smaller, the dominance of the morning rainfall maximum decreased. The tropospheric divergence profiles for 00Z and 12Z also showed large differences and generally corroborated these daily rainfall changes.

Ruprecht and Gray concluded that such large diurnal convection changes could not be due to heat island effects. They suggested that the variations were induced by changes in the net tropospheric radiation. The present study was undertaken to further document these variations.

TABLE 1

Comparison of Morning vs. Evening Occurrence of Various Rainfall Intensities for Western Pacific Cloud Clusters (from Ruprecht and Gray, 1974)

| Rain Intensity | 5 Hours Centered at 00Z (07-12 L.T.) | 5 Hours Centered at 12Z (19-24 L.T.) |
|---|---|---|
| Cluster Precipitation (% of Total) | | |
| >1.0 cm/hr | ∿75 | ∿25 |
| .25-1.0 cm/hr | ∿60 | ∿40 |
| trace- .1 cm/hr | ∿55 | ∿45 |
| Total | ∿70 | ∿30 |
| All 13 years of Precipitation whether associated with Cluster Precipitation or not (% of Total) | | |
| >2.0 cm/hr | 70 | 30 |
| 1.0-2.0 cm/hr | 60 | 40 |
| .5-1.0 cm/hr | 57 | 43 |
| .1- .5 cm/hr | 55 | 45 |
| trace- .1 cm/hr | 50 | 50 |
| Total | 57 | 43 |

## 2. DATA SOURCES AND ANALYSIS RESULTS

### 2.1 Composite Scheme

Upper air rawinsondes from twelve West Pacific island stations were positioned in relation to individual satellite observed cloud clusters and clear regions as shown in Figure 1. Two years of wind, temperature, and moisture data from these rawinsondes were then composited around the cluster and clear regions separately as previously described by Williams and Gray (1973), Ruprecht and Gray (1974), and Gray et al. (1975). The stations used and their locations are noted in Figure 2. 00Z (∿10 L.T.) and 12Z (∿22 L.T.) composites were

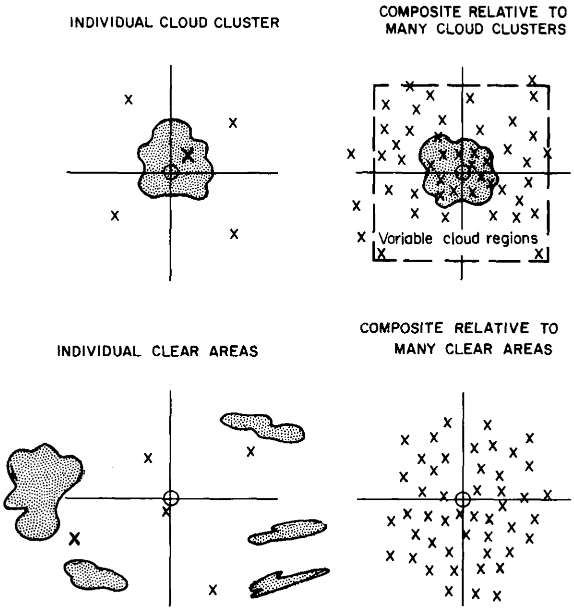



Figure 1. Portrayal of how the rawinsondes are positioned and then composited relative to the cloud cluster and clear regions.

constructed not only for the cloud clusters and clear regions, but also for the areas surrounding the cloud clusters which Gray (1973) called variable cloud regions (see Figure 1). A schematic representation of these three oceanic convective regimes is shown in Figure 3. In this way, the day vs. night thicknesses, wind and humidity patterns, and hourly rainfall data of the convective regimes could be compared.

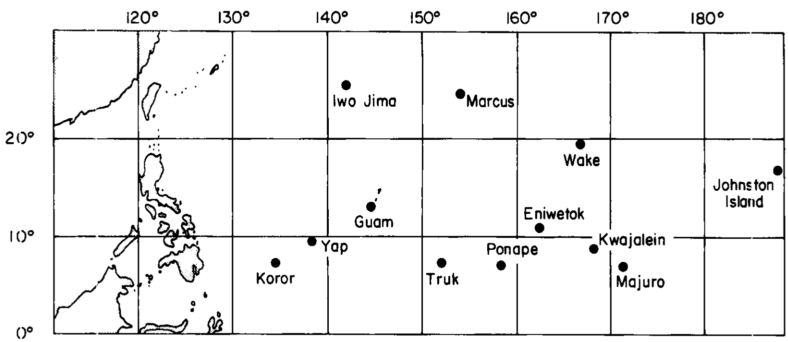



Figure 2. The rawinsonde stations used in forming the data composites around the cloud clusters and clear regions.

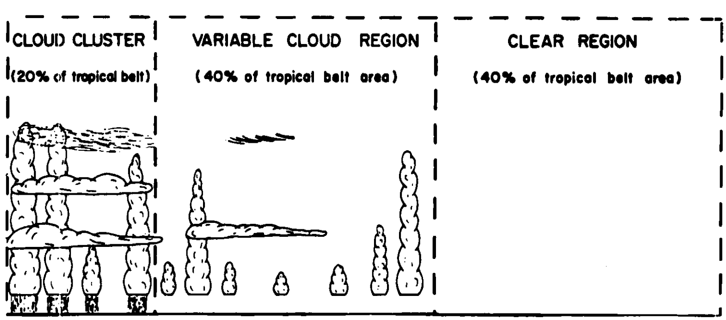



Figure 3. Schematic representation of the three oceanic regions in the tropics.

## 2.2 Wind Data

From the composited wind data divergence profiles were formed. These are shown in Figures 4 and 5. A large 10 L.T. vs. 22 L.T. difference (1½-2 to 1) is apparent for the cloud cluster. Although not as large, a diurnal difference of the opposite sign is observed in the clear region profiles. Thus, the clear regions have more low level divergence and the cloud clusters have more low level convergence at 00Z (∿ 10 L.T.) than at 12Z (∿ 22 L.T.). Divergence profiles for each of the three convective regimes are compared in Figures 6 and 7. It can be seen that the profiles for the variable cloud region and the clear region are generally the same with convergence aloft and divergence below.

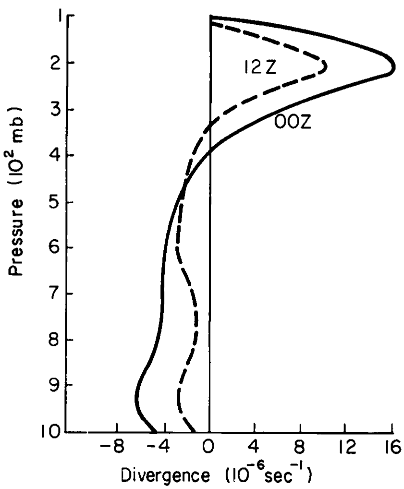



Figure 4. Cloud cluster divergence (00Z-12Z).

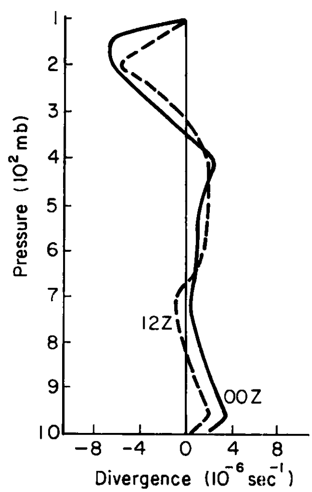



Figure 5. Clear region divergence (00Z-12Z).

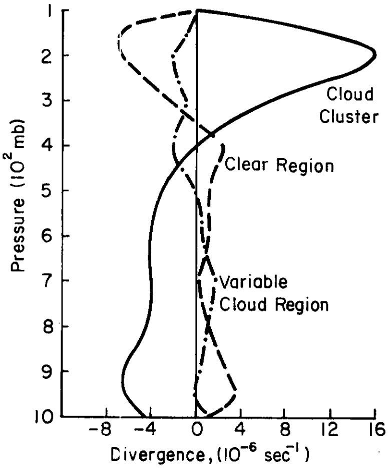



Figure 6. 00Z divergence profiles

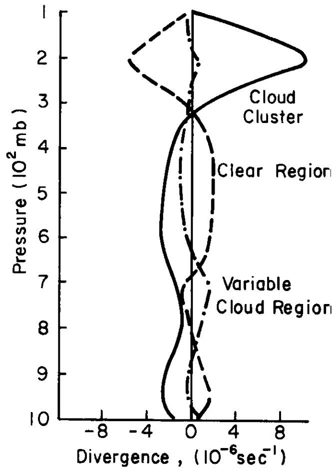



Figure 7. 12Z divergence profiles

From these divergences mean vertical motion profiles for each region were calculated for both time periods. These are shown in Figures 8 and 9. By combining the vertical motions for the clear and variable cloud regions, the total mean sinking motion around the cloud cluster can be derived (Figure 10) which again shows nearly a 2-1 diurnal difference. The depth of these profiles indicates that the large cumulonimbus (Cb) clouds must be primarily responsible for these diurnal variations.

## 2.3 Water Vapor Budget

Using the mean divergence and vertical motion profiles described above and the composited weather system humidity values, water vapor budgets for 00Z and 12Z were constructed to compare with the measured rainfall differences of these periods. It was assumed that the areal

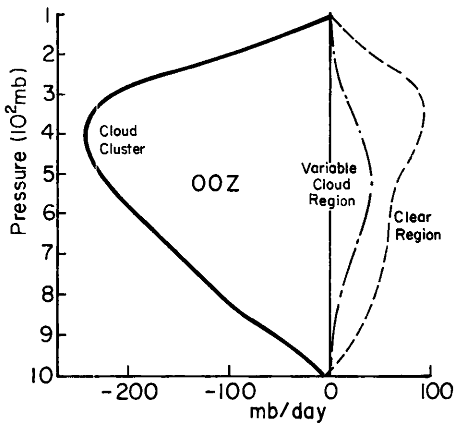



Figure 8. 00Z vertical motion profiles

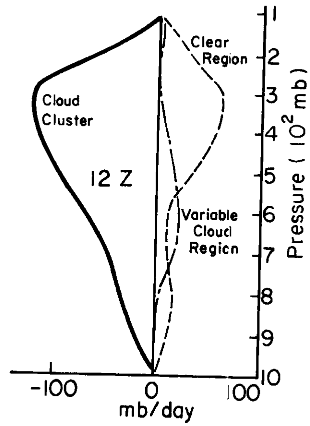



Figure 9. 12Z vertical motion profiles

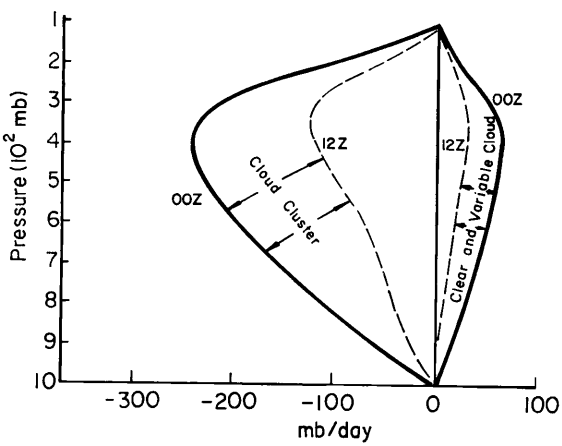



Figure 10. Comparison of vertical motion in and around the cloud cluster.

coverage of the three tropical regions was as shown in Figure 3 with the cloud clusters covering 20% of the area and the clear and variable cloud regions each accounting for 40%. Also, it was assumed that the significant rainfall came only from the cloud clusters, and that the statistical average of the cluster systems was in a steady state. Finally, the evaporation was chosen to have a constant rate of 0.4 gm $cm^{-2}$ $day^{-1}$ for all areas. The vapor budgets are shown in Figure 11.

The large rainfall difference between these time periods is clearly evident. The calculated rainfall values closely agree with the hourly rainfall data discussed in the next section. It is interesting to note that this budget requires a portion of the water vapor be stored in the atmosphere during the afternoon and evening and released as additional rainfall during the next morning. This can be explained by the fact that the upward vapor transport by the cumulus and towering cumulus clouds (which are assumed to produce no significant precipitation) shows little diurnal variation in amount. Thus, the weaker subsidence in the clear regions during the afternoon and early evening results in less moisture loss due to subsidence drying and, with no assumed diurnal variation in evaporation, a net afternoon vapor accumulation should occur. Conversely, the greater early morning subsidence should lead to a depletion of vapor.

## 2.4 Precipitation Data

Hourly precipitation data was obtained from the National Weather Records Center, Asheville, N.C., for the period 1961-1973 for the eight stations shown in Figure 12. Figure 13 gives an example of the form in which the data was received. This rainfall data was analyzed by the month (from March through October) for the thirteen year period. The hourly totals were smoothed and plotted for each station. In the

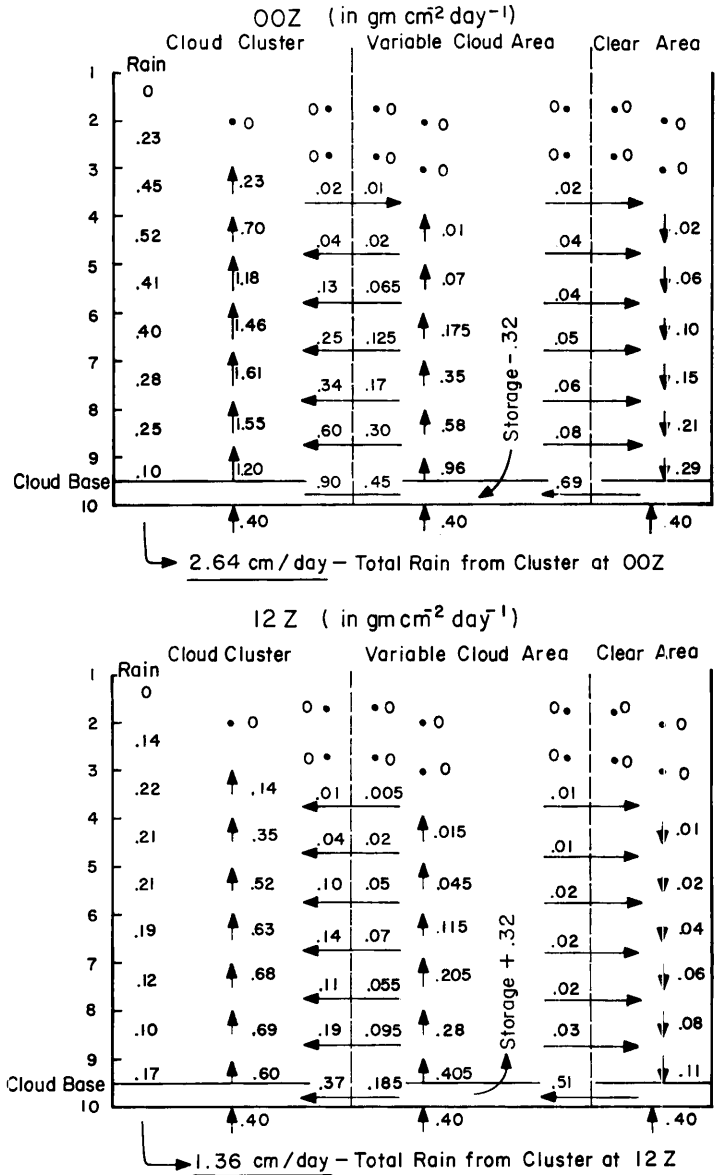



Figure 11. 00Z and 12Z Water vapor budgets for three idealized tropical regions in gm/cm$^2$ per day.

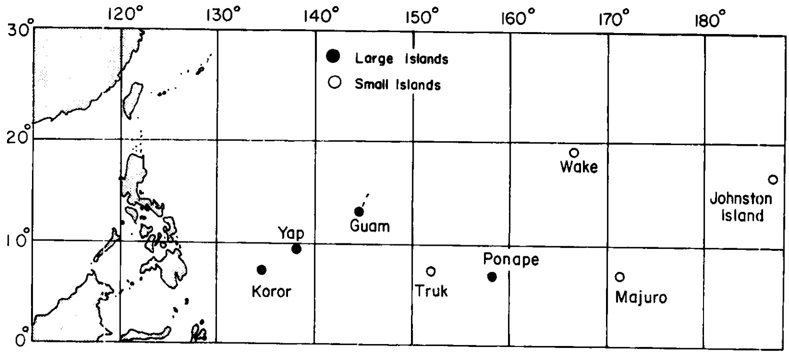



Figure 12. Western Pacific upper air stations for which hourly rainfall data was available (1961-1973).

HOURLY PRECIPITATION (Water equivalent in inches)

| Date | A. M. Hour ending at | | | | | | | | | | | | P. M. Hour ending at | | | | | | | | | | | | Date |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | |
| 1 | .03 | .04 | .12 | T | T | .01 | .24 | .03 | .26 | | | | | | | | | | | | .05 | .03 | .01 | .08 | 1 |
| 2 | | | T | .01 | .31 | .02 | T | T | | | | | | | | | | | | | | | | | 2 |
| 3 | | | | | | | | | | | T | | | | | | | | | | | | | | 3 |
| 4 | T | T | | | | | | | | | | | | | | | | | | | | | | | 4 |
| 5 | | | | | | | | | | | | | | .13 | .01 | .31 | .15 | T | T | | | T | T | | 5 |
| 6 | .47 | .35 | .02 | .01 | .08 | .01 | T | | .01 | T | | | | | T | | | | T | | .05 | T | | | 6 |
| 7 | | T | T | | | | | | T | | | | T | | | | | | | T | | | | | 7 |
| 8 | .01 | | | | | T | | .01 | | | | | | T | | | | | | | | | | | 8 |
| 9 | | | | | | T | | | | | | | | T | T | | T | | | | | T | .01 | | 9 |
| 10 | .02 | .03 | .01 | T | | | | | | | | | | | | T | | T | | | | | | | 10 |
| 11 | | | | | | | | | | | | | | | T | T | .05 | T | T | T | | | | | 11 |
| 12 | T | T | | .05 | T | | | | | | | | .33 | .14 | T | | | T | .12 | .60 | .21 | .03 | .03 | .01 | 12 |
| 13 | T | T | T | T | T | .01 | T | T | | | | | | | | | | | | | | | | | 13 |
| 14 | | | | | | | | | | | | | | | | | | | | | | | | | 14 |
| 15 | | | | | .15 | | | | | | | | | | | | | | .49 | T | | | | | 15 |
| 16 | | | T | .25 | T | .01 | T | T | | | T | T | .03 | T | | | | .33 | .01 | | | .16 | .04 | 1.47 | 16 |
| 17 | .40 | .17 | T | | | | | T | | | | | | | | | | | | | | | | | 17 |
| 18 | | | | | | | T | | | | | | .21 | .11 | .01 | T | T | | | | T | T | T | | 18 |
| 19 | T | T | | | | | | | | | | | | | | | | T | | .01 | | .59 | .12 | .06 | 19 |
| 20 | .05 | .05 | .01 | .01 | T | T | T | T | T | | | | | | T | | | T | T | | .08 | T | | | 20 |
| 21 | | | | | | | | | | | | | | | | | | | | | | | | | 21 |
| 22 | | | .06 | T | T | | .01 | .02 | | | | | | | | | | | | | | | | | 22 |
| 23 | | | | | | T | .21 | .06 | | .01 | T | | | | | | | | | | | | T | T | 23 |
| 24 | .01 | .01 | .01 | .01 | .01 | .01 | T | T | T | | | | | | | | | | | | | | | T | 24 |
| 25 | T | | | | | | | | | | | | | | | | | .03 | | | T | T | T | | 25 |
| 26 | .01 | T | T | | T | T | | | | | | | | | | | .07 | | | | .02 | | | | 26 |
| 27 | .12 | .02 | .07 | .10 | .07 | .09 | .68 | .46 | .09 | T | T | | | | | | | | | | | | | | 27 |
| 28 | | | | | | | T | | .02 | 1.00 | .34 | .01 | .02 | .01 | .02 | T | | .07 | .03 | .35 | .09 | .12 | .16 | .20 | 28 |
| 29 | .17 | .02 | .12 | .01 | T | T | T | T | | | | | | .01 | .08 | .16 | .27 | .01 | .02 | .03 | T | .01 | .01 | T | 29 |
| 30 | T | T | .01 | .02 | T | T | | | | | | | T | T | | | | | | | T | | | T | 30 |
| 31 | .26 | .06 | 1.29 | .01 | | | | .03 | .01 | T | | | | | | | | | | | | | | | 31 |

Figure 13. Typical form of hourly precipitation data (Truk, May 1972).

course of this procedure it was noted that the precipitation curves for the eight stations followed two separate patterns. Four stations (Koror, Yap, Guam, Ponape) showed distinct morning and afternoon peaks throughout the year. The other four stations (Johnston Island, Majuro, Wake, Truk) had only an early morning peak. Because of the similarities in the various rainfall curves, the data from the stations in each set was combined for subsequent analysis.

The reason for these afternoon differences is apparently due to the size and elevation of the islands. The larger stations experience an afternoon heat island influence which is not present on the smaller atolls. Figures 14 and 15 portray the relative sizes of the islands and the approximate locations of the weather stations.

The monthly results for the two groups showed little variation and were combined into three seasons: spring (March-May), summer (June-August), fall (September-October). Winter precipitation (November-February) was not treated. The hourly values are shown in Figures 16-21.

It is seen that in all cases an early morning maximum occurs between 03 and 06 L.T. Even when the largest precipitation maximum is observed in the afternoon, this early morning peak is apparent. Also, an evening minimum is found in each instance between 18 and 22 L.T. The differences between the maximums and minimums are on the order of 55% for the small islands. If the afternoon rainfall maximums on the large islands are neglected, as in Figure 22, these morning vs. evening differences are about 30%.

It is possible to stratify these results further to show the association between rainfall intensity and the diurnal pattern. Stratifications were made for intensities $\geq 0.3$ in/hr ($\geq 7.6$ mm/hr) and $\leq 0.1$ in/hr ($\leq 2.5$ mm/hr). This coincides with the definitions of heavy and light rain offered in the Federal Meteorological Handbook No. 1 (1970). Comparisons of the heavy, light, and total rainfall amounts are depicted in Figures 23-30. Superimposed on these curves are graphs showing

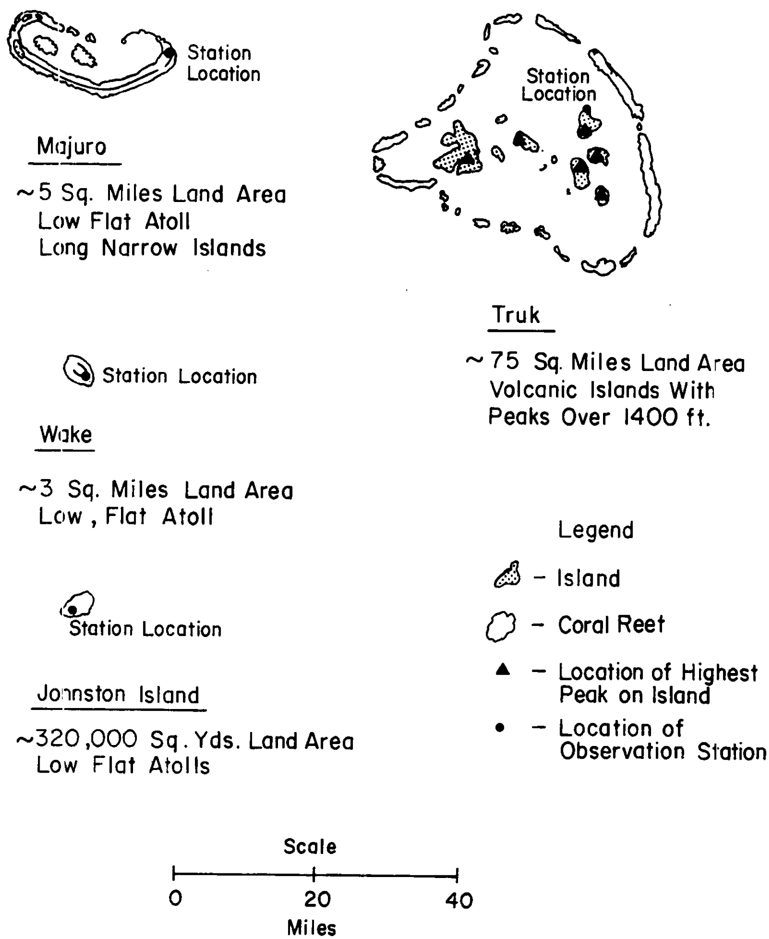



Figure 14. Representations of small islands with no afternoon rainfall maximum.

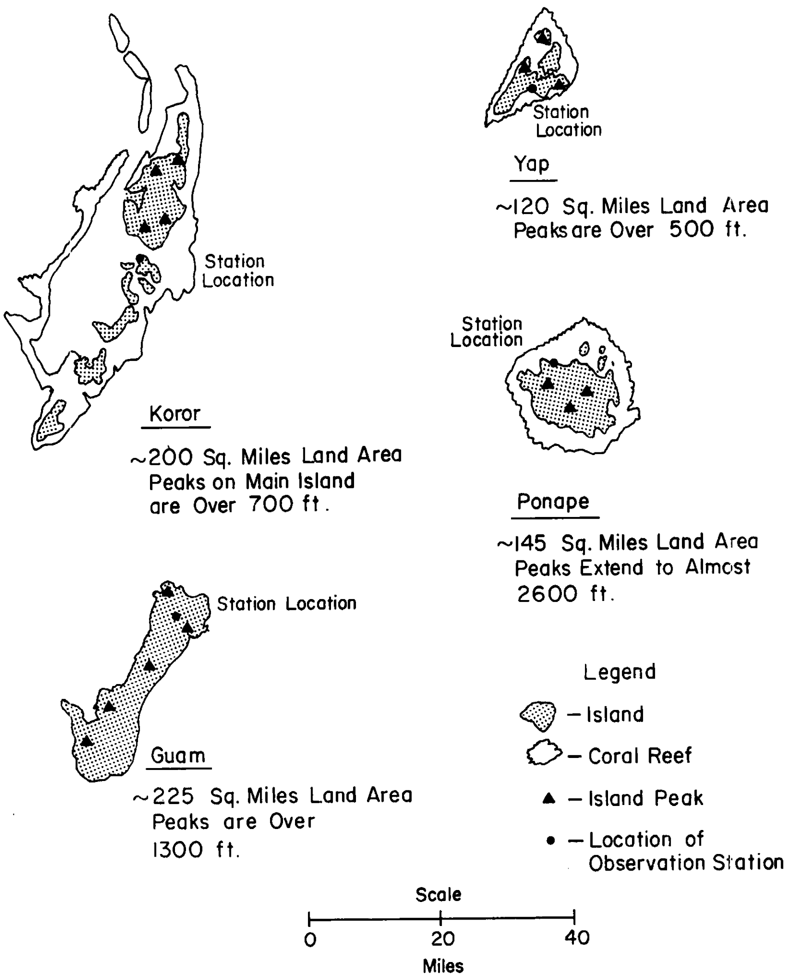



Figure 15. Representations of large islands with a distinct afternoon rainfall maximum.

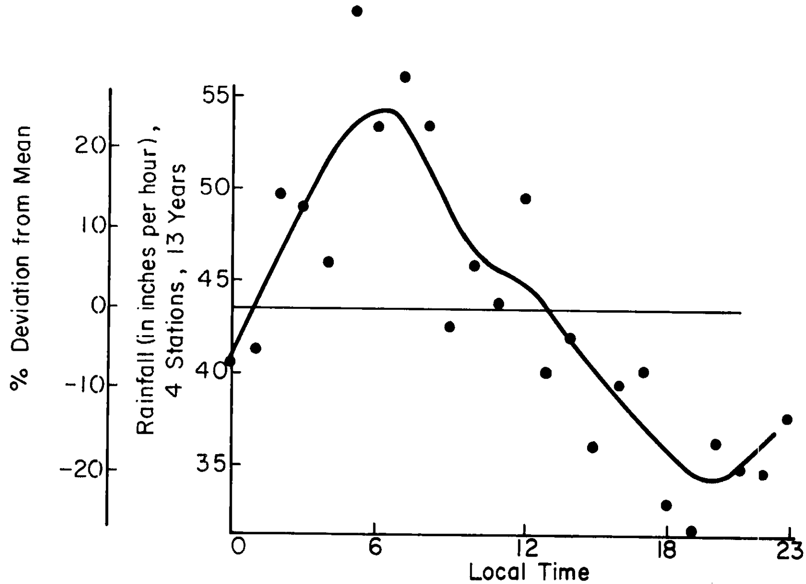



Figure 16. Precipitation curve - small islands - Spring

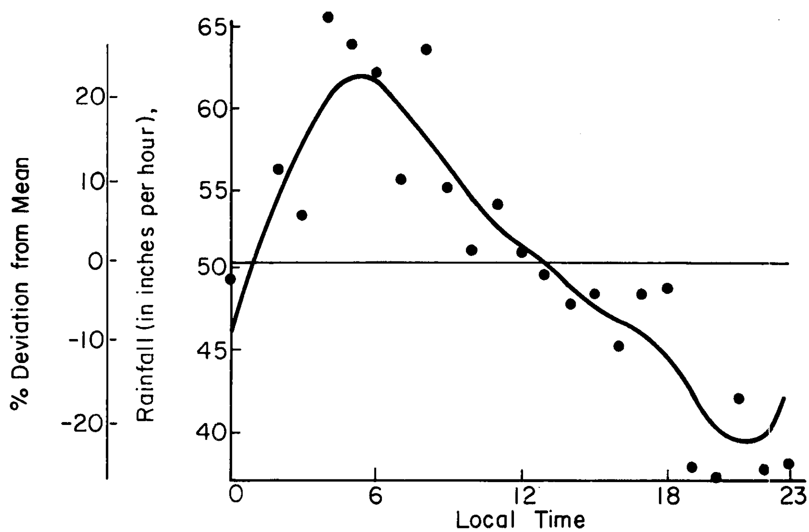



Figure 17. Precipitation curve - small islands - Summer

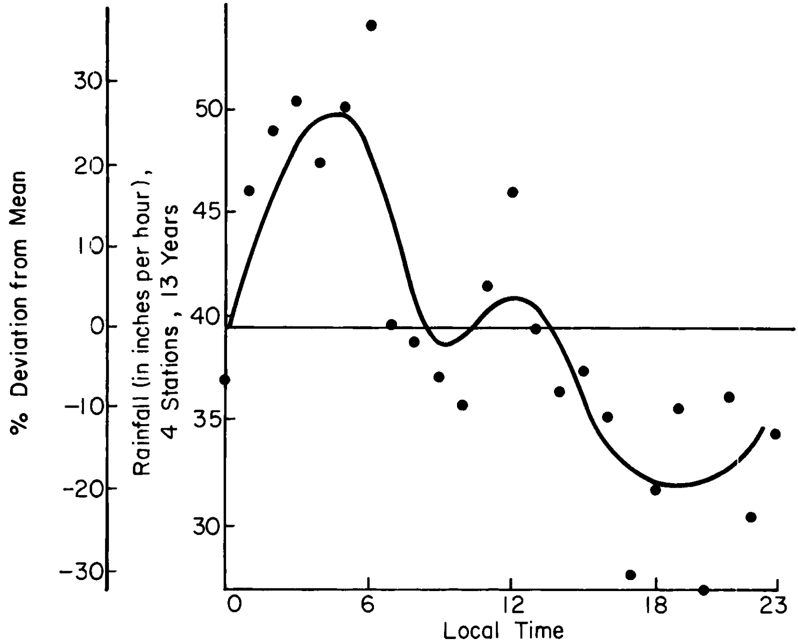



Figure 18. Precipitation curve - small islands - Fall

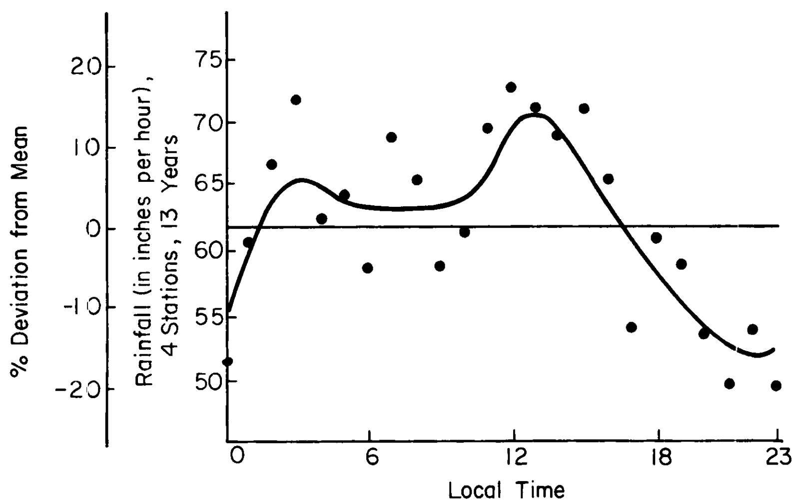



Figure 19. Precipitation curve - large islands - Spring

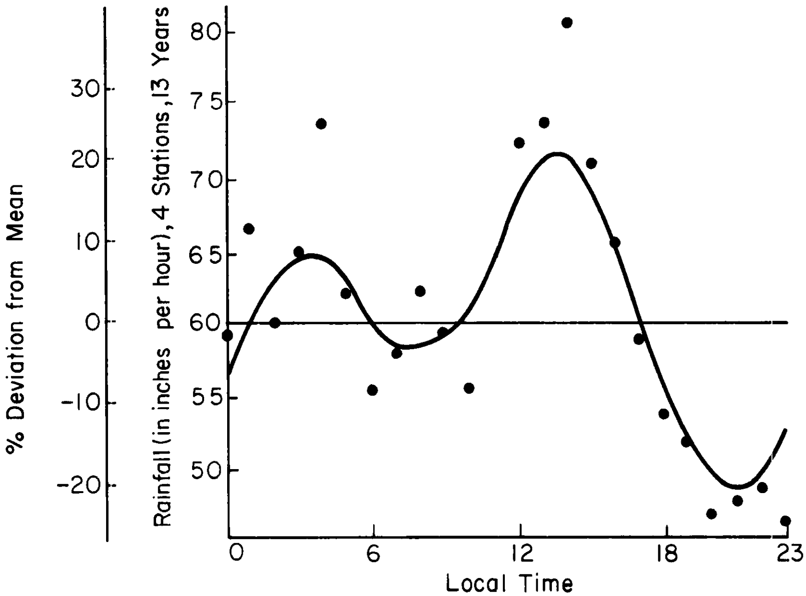



Figure 20. Precipitation curve - large islands - Summer

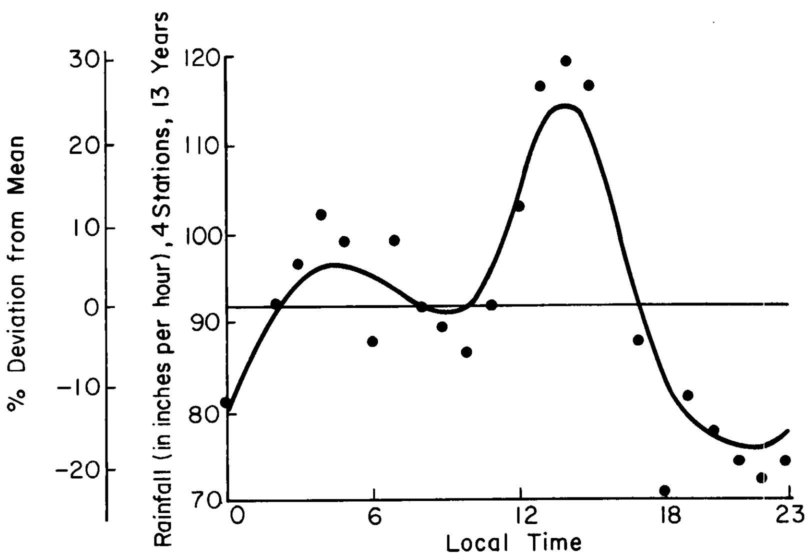



Figure 21. Precipitation curve - large islands - Fall

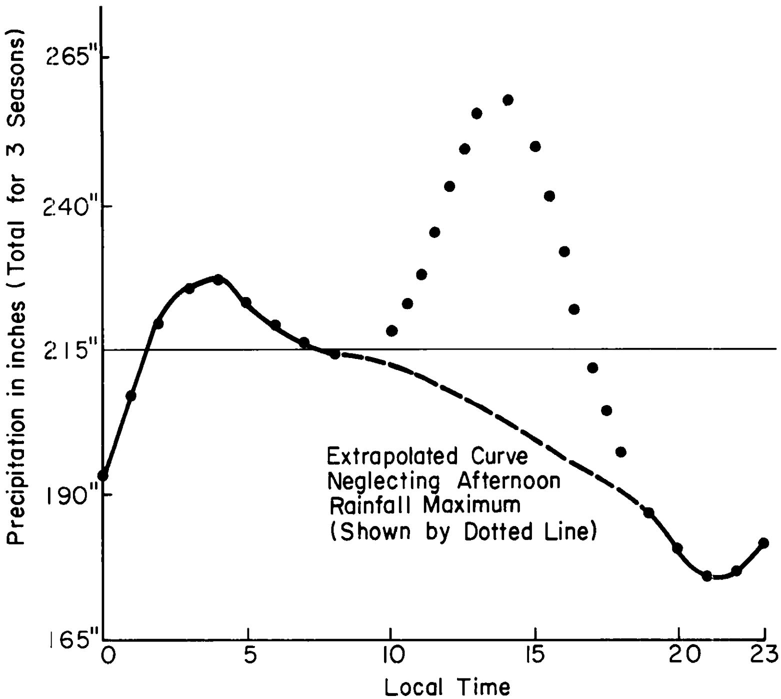



Figure 22. Large island rainfall neglecting the afternoon maximum.

six-hourly precipitation totals caused by rainfall with intensities of ≥1.0 inch per hour (≥25 mm/hr). It was necessary to portray the very heavy rain in this way because the frequency of occurrence was much less than in the other rainfall categories.

In the study by Ruprecht and Gray (1974) it was concluded that the heavy rains were most responsible for the morning rainfall maximums (see Table 1). This conclusion is upheld by Figures 23-30. The heavy rain curves are almost identical in shape with the total rainfall curves. Also, the percentage differences between the morning precipitation maximums and evening minimums are larger than those of the total

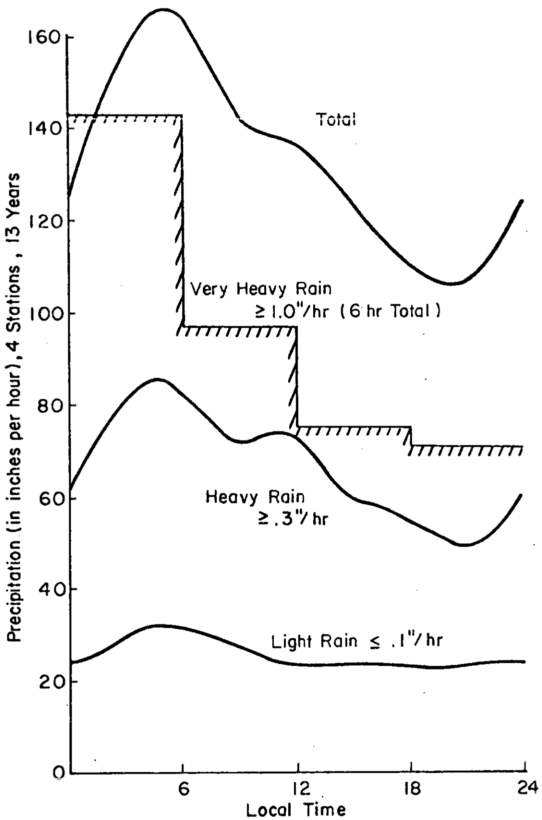



Figure 23. Rainfall components - small islands - 3 seasons.

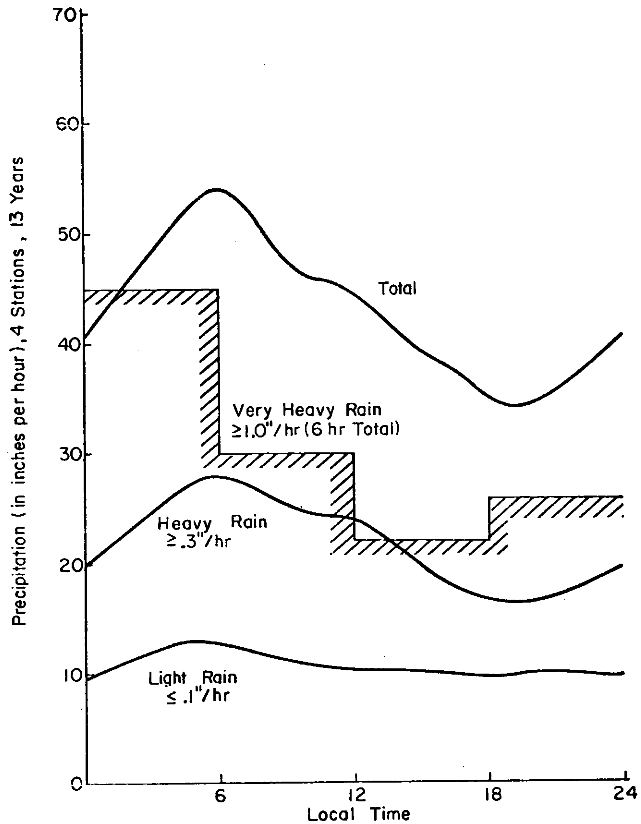



Figure 24. Rainfall components - small islands - Spring.

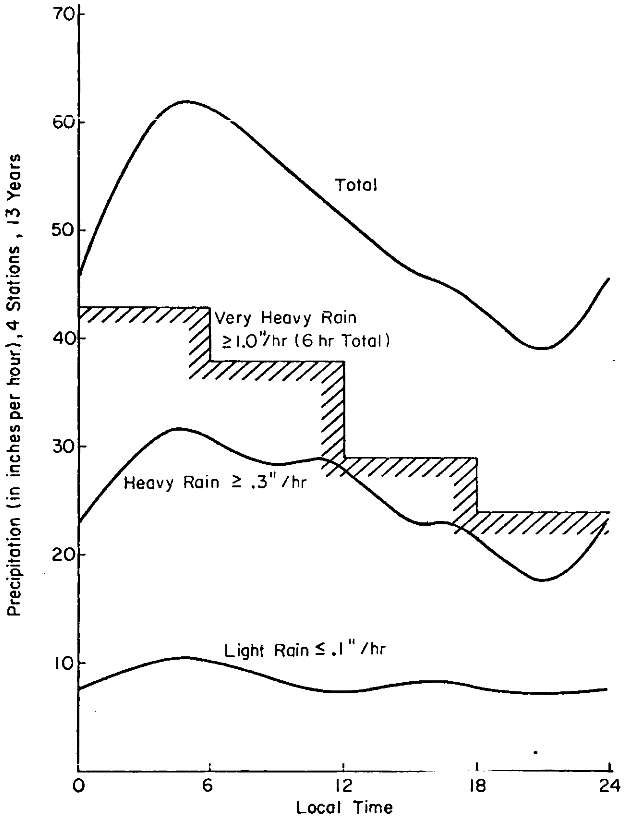



Figure 25. Rainfall components - small islands - Summer.

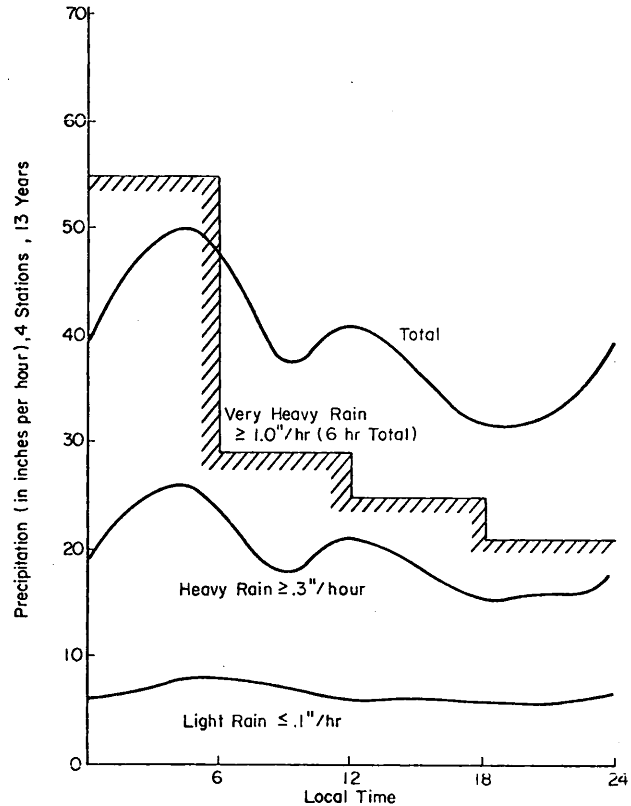
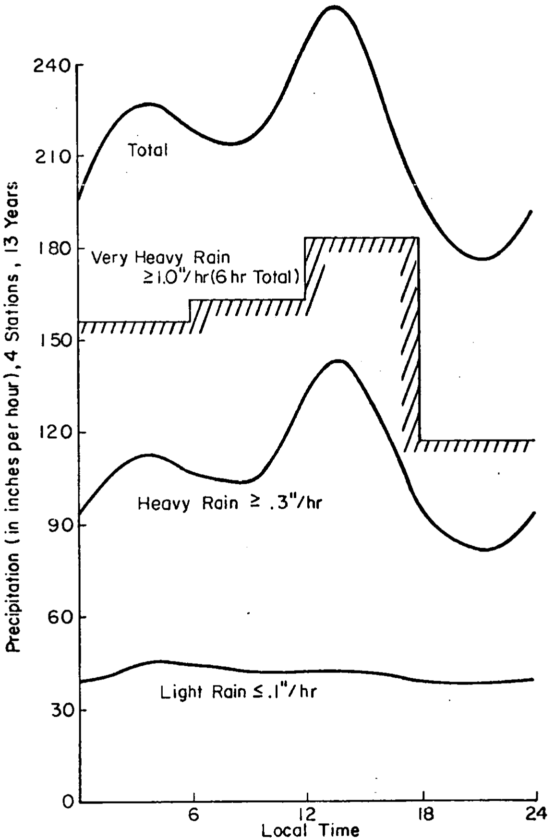



Figure 26. Rainfall components - small islands - Fall.



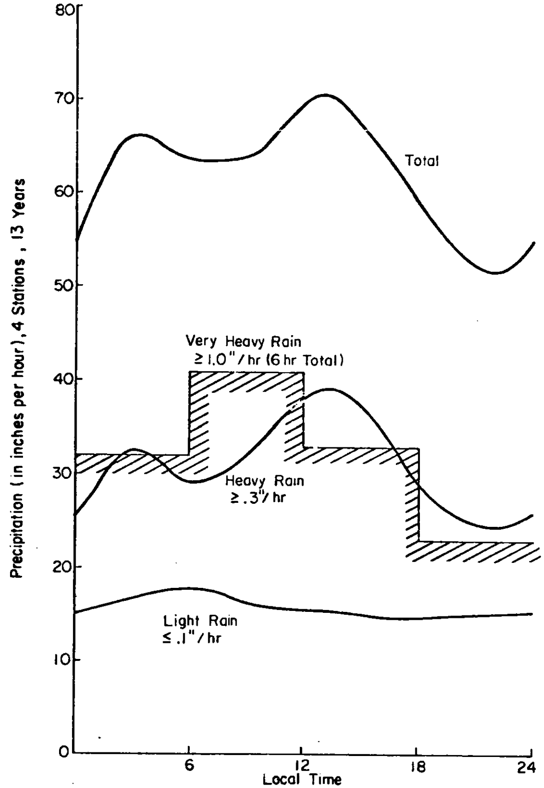




Figure 27. Rainfall components - large islands - 3 seasons.

Figure 28. Rainfall components - large islands - Spring.

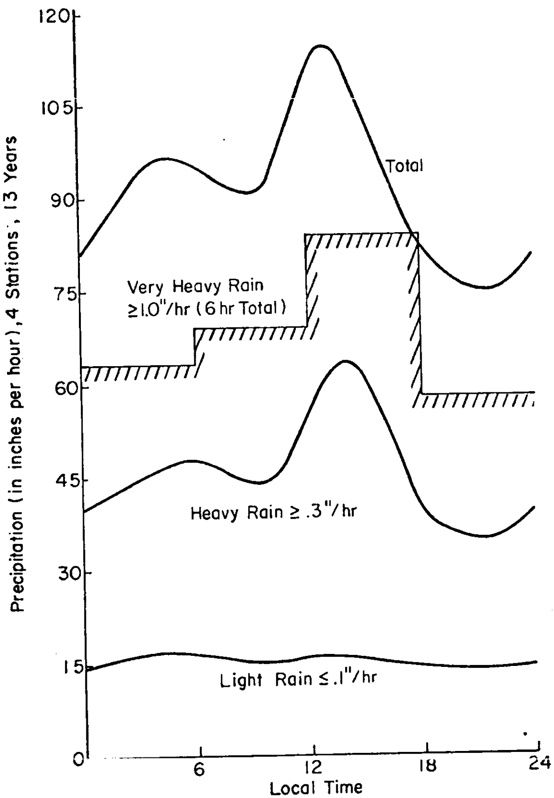



Figure 29. Rainfall components - large islands - Summer.

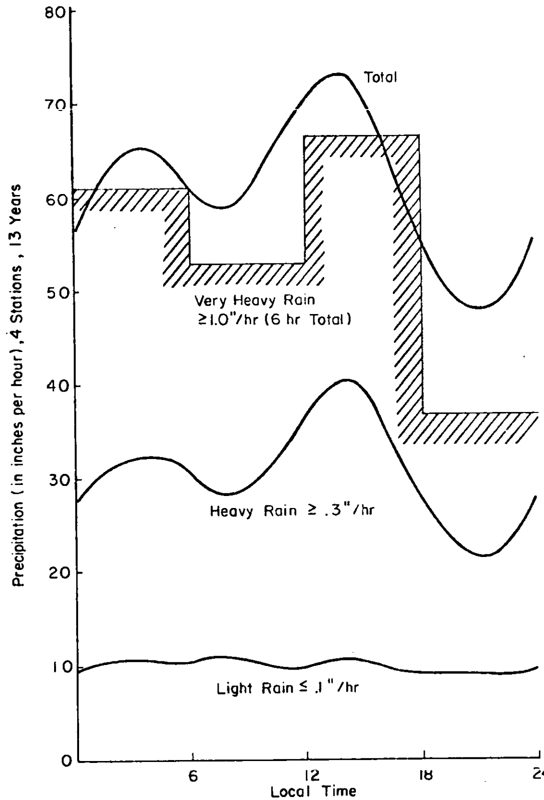



Figure 30. Rainfall components - large islands - Fall.

rainfall curves, being about 70% for the small islands and 35% for the large islands. The light rain curves are almost flat and show a much smaller correlation with the total precipitation curves. The very heavy rain graphs also show the predominance of morning over evening convective activity.

To supplement the hourly data six-hourly precipitation reports for up to eight months (April through November) in 1967 were analyzed for nine other small island stations shown in Figure 31. These stations were also divided into categories based on rainfall distribution.

Figure 32 portrays the six-hour rainfall occurrences of ≥0.5 in. (12.5 mm). Reporting times were adjusted to local times to normalize the data. By subsequently summing the number of occurrences at each hour, an indication of the heavier rain can be found. Again, the highest frequency of the cases is found in the morning.

## 2.5 Supplemental Satellite Photograph Study

To investigate further these diurnal convection variations the Defense Meteorological Satellite Program (DMSP) polar orbiting satellite photographs have been studied for the period of April-May, 1974. These satellites make 3 or 4 passes per day over the Western Pacific area and produce visual and infrared (IR) images having 1/3 nautical mile (n.m.) and two n.m. resolution. For further information on the DMSP, the reader is referred to the DMSP User's Guide (1974).

An area on the photographs between 0° and 20°N latitude and 125°E and 170°E longitude was chosen as the domain for analysis. In this region the 1/3 n.m. resolution IR pictures were available for night hours while the 1/3 n.m. visual pictures were available during the daylight periods. To determine the diurnal convective activity

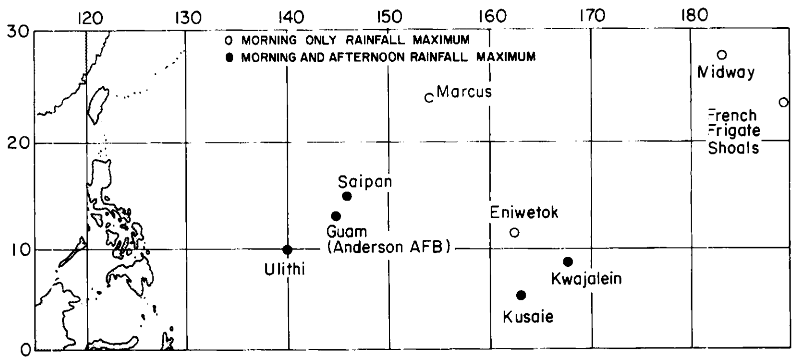



Figure 31. West Pacific upper air stations for which 6-hourly rainfall data was available.

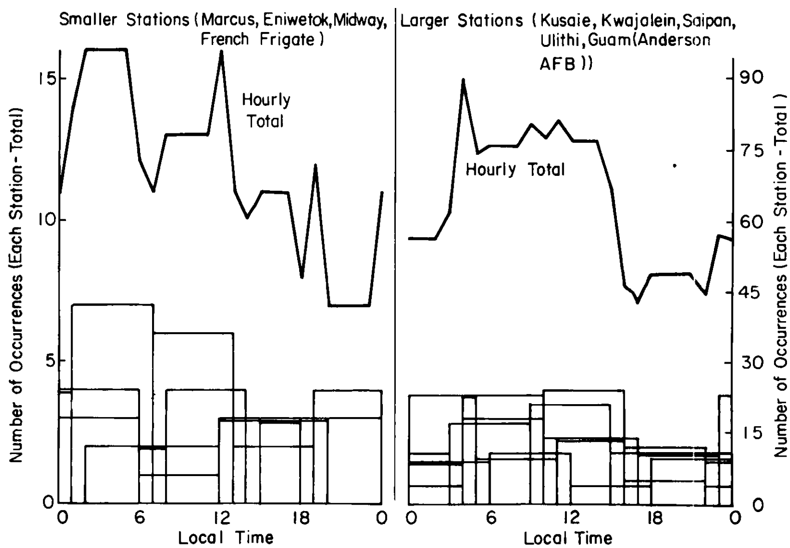



Figure 32. Rainfall occurrences - ≥0.5"/6 hours

changes from these photographs it was necessary to compare cloud temperatures with cloud albedo values. It was thus impossible to use any quantitative scheme for the comparisons. The same situation existed with the 2 n.m. resolution pictures which were used on the few occasions when the 1/3 n.m. photographs were not available.

Picture availability required the comparing of photographs taken at or slightly after 00Z ($\sim$ 10 L.T.) with those pictures taken near 12Z ($\sim$22 L.T.). Visual inspection was used to determine whether the convective elements in the westward propagating cloud systems appeared to increase, remain relatively the same, or decrease in number and area for each 12 hour period. The results of this admittedly subjective procedure are summarized in Table 2. The trend toward a 00Z convective maximum and a 12Z convective minimum is apparent.

TABLE 2

Comparison of Convective Activity Using DMSP Data for April and May, 1974

| Change | 10 L.T. to 22 L.T. | 22 L.T. to 10 L.T. |
|---|---|---|
| Probable Increase | .16 | .71 |
| Probable Decrease | .40 | .12 |
| No Apparent Change | .44 | .17 |

# 3. PREVIOUS INVESTIGATIONS

## 3.1 Diurnal Cloudiness Investigations

The existence of a nocturnal cloudiness maximum over the tropical oceans has been suspected for many years. Because of problems in observing nighttime cloud cover, however, early attempts at quantitatively verifying this existence (e.g. Riehl, 1947) were difficult. Some of the studies which have dealt with these diurnal cloudiness variations are listed in Table 3.

Despite the observational difficulties, a consistent pattern of oceanic early morning and late afternoon or evening cloud cover maximums separated by relative minimums are indicated by these studies. It should be noted that because of observational problems only the lower tropospheric cloudiness was investigated. Variations in deep cumulus convection, therefore, could not be well measured. Except for the study of La Seur and Garstang (1964) based on two short cruises by the WHOI ship Crawford, this oceanic diurnal variability of the low cloudiness was generally found to be small ($\sim$15%). Riehl (1954) estimated this difference to be only about 10% for the tropics in general.

## 3.2 Diurnal Rainfall Variations

This double-peaked cloudiness maximum is not well supported by the diurnal rainfall observations noted in this and other studies which are listed in Table 4. There is a corresponding rainfall maximum in the morning, but a general minimum of rainfall is seen in the evening when the second cloudiness peak occurs. This again indicates that the rainfall patterns are associated mainly with variations of the Cb clouds and not with the total cloud cover.

TABLE 3

Some Studies Concerning Diurnal Variations of Low Level Cloudiness

| Source | Data Source(s) | Time Period | Findings |
|---|---|---|---|
| Riehl (1947) | 3 World War II Atlantic weatherships stationed near 35°N, 40°W; 35°N, 52-55°W; 35°N, 70°W. 3-hourly observations | June-Oct. 1944 April-Oct. 1945 | Low cloud maximum occurred near 06 L.T. with a minimum near 22 L.T. An approximate 15% difference was found. A second maximum at 18 L.T. and a second minimum at 01 L.T. were apparent. |
| LaSeur and Garstang (1964) | 1957 Crawford weathership cruise stationed near 11°N, 52°W. 1963 Crawford weathership cruise stationed near 13°N, 22°W. Hourly observations | 14-20 Aug. 1957 24 Aug.-3 Sept. 1957 12 Aug.-3 Sept. 1963 | Low cloudiness maximum was observed at 07 L.T. with a minimum near 2130 L.T. A 45% difference was found. Secondary minimums and maximums were found near 10 and 14 L.T., respectively. |
| Lavoie (1963) | Hourly reports for Eniwetok | June 1949-Feb. 1959 | Low cloud maximum was observed near 07 L.T. with a minimum at 22 L.T. The difference was about 10%. Small fluctuations were seen in the afternoon. |
| Brier and Simpson (1969) | Hourly surface observations for Wake. | 12 years beginning in Nov. 1949. | Low cloudiness maximum near 08 L.T. with a minimum near 24 L.T. A 20% difference was noted. Secondary minimums and maximums were found near 14 and 19 L.T., respectively. |

TABLE 4

Some Studies Concerning the Diurnal Variations of Oceanic Precipitation

| Source | Data Source | Time Period |
|---|---|---|
| 1. Garstang (1958) | 1957 Crawford Weathership cruise stationed at 11°N, 52°W. | 6-Hourly measurements from 14-20 August 1957 and 24 August-3 Sept. 1957 |

Findings and Discussion: From a small data base it was found that 50% more rain was measured at night (18-06 L.T.) than during the day (06-18 L.T.). This is in general agreement with the findings of the present study. Also, almost all of the precipitation was associated with disturbances.

| | | |
|---|---|---|
| 2. Lavoie (1963) | Eniwetok (supplemented by 1 - 13 years of data from 10 other West Pacific atolls, mainly of the Marshall and Line Islands) | Hourly reports for June 1949 - Feb. 1959 |

Findings and Discussion: A precipitation maximum on Eniwetok occurs near 03 L.T. It is 40% larger than the minimum found near 10 L.T. A small secondary maximum and minimum are observed at 13 L.T. and 17 L.T. Aside from the 10 L.T. minimum this is comparable to the findings of the present study. For the 10 atolls, a 03 L.T. maximum and a 12 L.T. minimum are present with a 25% difference between them. This small difference may be the result of a short data set from some of the stations.

| | | |
|---|---|---|
| 3. Kraus (1963) | 9 Weathership stations in the Atlantic and Pacific Oceans. | 3-Hourly reports of July observations in the period of 1950-1961. |

Findings and Discussion: 50% more rain reports came at night (21-06 L.T.) than during the day (09-18 L.T.). This is in good agreement with the present study even though the data did not come from tropical regions. (See Table 5 for more information)

| | | |
|---|---|---|
| 4. Finkelstein (1964) | 10 Tropical Pacific island stations between 20°N and 30°S lat. and 165°E and 160°W long. | Hourly observations for unspecified lengths of time. |

TABLE 4 (cont'd)

Findings and Discussion: 3 of the stations were on very large islands having dominant afternoon rainfall maximums. For the 7 smaller stations 54% of rainfall was measured between 20-08 L.T. while 46% came between 08-20 L.T. These differences are smaller than those of the present study, but this could be because of the time periods chosen. A general minimum occurs between 20-24 L.T. The percentage differences might be more comparable had the night rain been measured at a later period.

---

| | | |
|---|---|---|
| 5. Holle (1968b) | 1963 WHOI Crawford Weathership at 13°N, 55°W | 12 Aug.-3 Sept. 1963 |

Findings and Discussion: 3 cm radar measurements showed the highest cloud tops occurred near sunrise. Also, radar echoes were viewed twice as often between 02-05 L.T. as at 13-18 L.T. This agrees with the present study.

---

| | | |
|---|---|---|
| 6. Brier and Simpson (1969) | Wake Island | Hourly surface observations for 12 years beginning in Nov. 1949 |

Findings and Discussion: Rainfall frequency maximum occurs at 05 L.T. with a minimum near 15 L.T. The morning rain is almost twice as frequent (.053-.028 occurrences per hour) as the afternoon rain, which is comparable to the findings of the present study.

---

| | | |
|---|---|---|
| 7. Inchauspé (1970) | 4 French Polynesian atolls (Mopelia, Hereheretue, Hao, Takaroa) | Hourly data for 5-6 years between 1965-1970 |

Findings and Discussion: A maximum of rainfall occurrences is observed between 04-08 L.T. and is 20-55% larger than the evening minimum between 22-23 L.T. Smaller maximums and minimums are found near 15 L.T. and 12 L.T.

Maximum amounts of precipitation were recorded between 03-08 L.T. and are 40-60% larger than the measured minimums found either at 12 L.T. or between 21-24 L.T. Smaller afternoon and evening maximums are also seen in the data except at Takaroa.

The findings of Inchauspé generally agree with those of the present study. However, the afternoon and evening maximums were anomalous. It is possible that the location of the Tuamotu Archipelago in relation to the stations used (Figure 33) may be responsible for the afternoon peak. The great number of atolls together with the large areal coverage of shallow lagoons associated with the atolls may act as a heat island. Golden (1974) and Malkus (1957) both found that shoaling in the ocean can cause such results. The evening maximum, which Inchauspé believes is due to atmospheric warming by the ocean, is not generally observed in the data of this study, however.

---

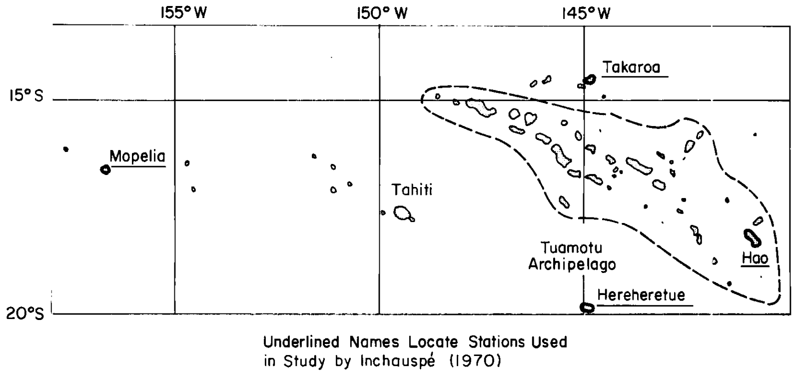



Figure 33. Tuamotu Archipelago

In his weather ship study Kraus (1963) noted the difficulties in distinguishing between convective and non-convective precipitation. This may cause the number of rain reports in each category to be altered somewhat. However, as seen in Table 5, his findings are very comparable to those using the West Pacific data.

TABLE 5

Comparison of from 8-10 Years of July Night and Day Rain Reports
from 9 Ocean Weather Ships
from Kraus, 1963

| Precipitation Category | Number of Night Rain Reports (21-06 L.T.) | Number of Day Rain Reports (09-18 L.T.) |
|---|---|---|
| Convective Precipitation | 283 (60%) | 191 (40%) |
| Non-Convective Precipitation | 1141 (60%) | 747 (40%) |
| All Precipitation | 1424 (60%) | 938 (40%) |
| No Precipitation | 6341 (48%) | 6828 (52%) |

In the study by Inchauspé it is difficult to see how the heating of the atmosphere over the sea can cause an evening increase in precipitation showers. Lavoie (1963) has stated that there is a heat flux from the air to the sea in the afternoon. Therefore, the sea does not begin to warm the lowest layers of the atmosphere until the evening. The rate of the sensible heat flux, according to Warsh et al. (1972) reaches a maximum near sunrise. Since the heating is cumulative, it is believed that the main effects of this heating would be seen near sunrise. Lavoie (op. cit.) has indicated that this sensible heat transfer manifests itself in a slight cloudiness increase, but that this has little or no effect on the deep cumulus convection. Hence, rainfall over the open sea is probably not increased by this type of heating. Thus, we believe the late afternoon and evening peaks in Inchauspé's precipitation are a result of weaker shower activity and are due to atoll lagoon solar heating.

### 3.3 Other Supporting Information for Diurnal Cycle of Deep Cumulus Convection

Oceanic Observations. The following additional data sources also lend much support to the prevalence of an oceanic morning maximum of deep cumulus convection:

1) A new compositing of rawinsonde reports within and surrounding West Pacific cloud clusters independent of the Ruprecht and Gray (1974) cloud cluster rawinsonde composite as reported on page 2 gives additional verifying information in support of large differences in morning and early evening cluster mass convergence-divergence. These composites are shown in Fig.34 . They were obtained from a current 10-year rawinsonde composite study of intensifying and non-intensifying cloud clusters being performed by Zehr (1976). Diagram A of this figure shows 00Z vs. 12Z kinematic determined divergence on a space scale of 4$^{\circ}$ for cloud clusters which do not develop into tropical cyclones. Diagrams B and C portray the same types of 00Z vs. 12Z divergence profile relative to intensifying cloud clusters which later become tropical storms. Diagram B represents the very early stages of the cluster genesis when the maximum sustained cloud cluster

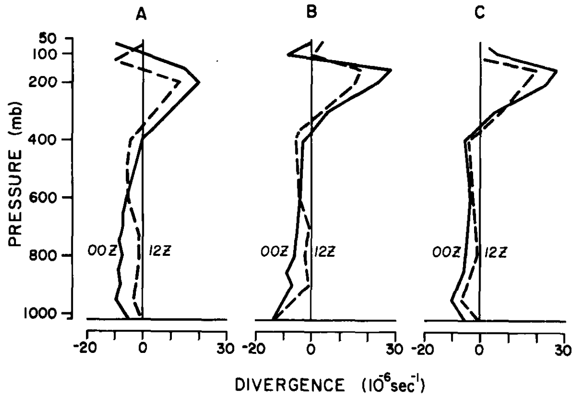



Figure 34. 00Z vs. 12Z 4$^{o}$ average divergence profiles for three groups of West Pacific cloud clusters. Diagram A is for non-intensifying cloud clusters, profiles B and C are cloud clusters which later develop into tropical cyclones. B is for clusters whose maximum sustained winds are less than 10-12 m/sec, C for clusters whose maximum sustained winds are approximately 20 m/sec.

winds are less than 12-15 m/sec. Diagram C represents the time period when the cloud clusters have intensified to the stage where their maximum sustained surface winds are approximately 20 m/sec. Note that in all three cases the strength of the 00Z convergence-divergence pattern through the cluster is approximately one-and-a-half to two times that of the 12Z circulation. These cluster mass flow differences should produce similar magnitude differences in morning vs. evening Cb activity. They add additional verifying evidence to the cluster and clear region divergence profiles of Figs. 4 and 5 on page 5.

2) A kinetic energy study has recently been performed by Kung (1975) using 130 days of rawinsonde data in a 15$^{o}$x15$^{o}$ area in the Marshall Island region which was gathered in conjunction with the 1958 nuclear tests. He found that the integrated 1000-100 mb generation of kinetic energy within the large scale disturbances of this region was nearly three times larger at 00Z (0.33 watt/m$^{2}$) than that at 12Z (0.13 watt/m$^{2}$). Also, the disturbance divergence of kinetic energy flux between 1000 and 100 mb was an order of magnitude larger at 00Z than at 12Z. As his 15$^{o}$x15$^{o}$ mean 00Z vs. 12Z divergence values showed no significant differences, the diurnal kinetic energy variations which he obtained must be largely due to diurnal variability of the local region vertical circulation.

3) Trent and Gathman (1972) have made a massive observational study of thunderstorm occurrence from surface ship observations over the global oceans. Their data sample includes more than 7 million surface vessel reports from 1949 to 1963. Unfortunately, they stratified most of their reports in GMT rather than local time. They did, however, stratify a portion of the data by local time.

As thunderstorms are more easily detectable by lightning discharge at night, their nighttime total far exceeded daytime reports. Thus, one cannot accept their direct comparison of day vs. night frequencies. It is possible, however, to compare their 18-24 L.T. vs. 00-06 L.T. nighttime thunderstorm occurrences and their measured 06-12 L.T. and 12-18 L.T. daytime values. Of 559,956 ocean ship reports which they stratified by local time, thunderstorms were reported on approximately 1% of all observations.

Table 6 lists the local time percentage occurrences in these various six and twelve hour time segments. The much higher late morning vs. afternoon and early morning vs. early evening thunderstorm occurrences are obvious.

TABLE 6

Local Time Percentage Thunderstorm Occurrences

| Nighttime | | Daytime | | All Morning vs. Afternoon-Evening | |
|---|---|---|---|---|---|
| 00-06 vs. | 18-24 | 06-12 vs. | 12-18 | 00-12 vs. | 12-24 |
| 62 | 38 | 64 | 36 | 63 | 37 |

4) Takeuchi and Nagatani (1974) studied sferics observations for 10 days from the Equator to the Fiji Islands, for 32 days between Japan and Guam, and for 7 days between the Philippine Islands and Japan. They concluded:

> "The diurnal variation of the ocean thunderstorm activity shows different patterns as compared with that of the ordinary land storm. The peak activity of the oceanic storm appears in the time between midnight and early in the morning."

5) Using data from the SMS/GOES geosynchronous satellite located over the equator at 45$^{o}$W long., Martin (1974) found that cloud clusters throughout the A-Scale GATE area typically reached their maximum intensity near 06 L.T. This was accomplished by comparing the cluster appearance at various times with its observed rate of growth.

6) Weickmann and Long (1975) also have investigated rapidly expanding cloud clusters in the GATE area with the use of geosynchronous satellite data. These so-called "super novae" storms appeared to expand 1-2 orders of magnitude in area in from 3-6 hours. The vast majority of these expanding cluster cases (14 of 19) occurred between 02-09 L.T.

7) Kidder and Vonder Haar (1976) are using 1.55 cm microwave band data from the Nimbus 5 satellite in attempting to estimate liquid water content of the atmosphere over oceans. Observed radiance temperatures are correlated with liquid water for $5^{\circ}$ latitude-longitude squares near 00 and 12 L.T. daily. Preliminary findings indicate a larger ($\sim$60%) liquid water content at 12 L.T. (result of morning convection maxima) than at 00 L.T. ($\sim$40% - result of evening convection minima).

8) Although for only 5 days, the recently reported diurnal variability of radar echoes for disturbed conditions in the BOMEX area by Hudlow (1975) - see Fig. 35 - also generally supports this type of diurnal convective variability.

9) Holle's (1968a) radar echo statistics over the sea 40-100 n.m. miles east of Miami support an over ocean morning peak of convection. Figure 36 from his report is typical of other months and in contrast with his over land measurements which show strong late afternoon peaks in convection.

10) Nitta and Esbensen (1974) have also obtained significant diurnal variation of tropospheric divergence over 5 days of undisturbed conditions during the BOMEX project. Their variations occurred through a deep layer of the troposphere and were of the magnitude of $\sim 3 \times 10^{-6} sec^{-1}$. Their hours of maximum relative upward and downward velocity were approximately midnight and noon, which is about 6 hours earlier than the near sunrise peaks observed by other investigators. They concluded:

> "The diurnal variations of the large-scale horizontal divergence, large-scale vertical velocity and wind over the Atlantic trade wind region are examined. Maximum amplitude of the diurnal variation of the vertical velocity is 1 mb hour$^{-1}$ which is nearly half of the mean downward velocity under the undisturbed weather situation (22-26 June, 1969). The amplitude of diurnal variations in this study is one order magnitude larger than that of the diurnal tide obtained theoretically by Lindzen (1967). . . .
>
> The large amplitude of the diurnal variations in the large-scale divergence field may affect the activity of cumulus convection . . . . . Hudlow (1970) obtained the average diurnal variation in radar echo occurrences during the BOMEX period. A maximum in echo activity is observed during early morning which is about 6 hours later than the occurrence of the maximum upward motion. The minimum echo activity occurs during midday which is about 4 hours later than the occurrence of the maximum downward motion."

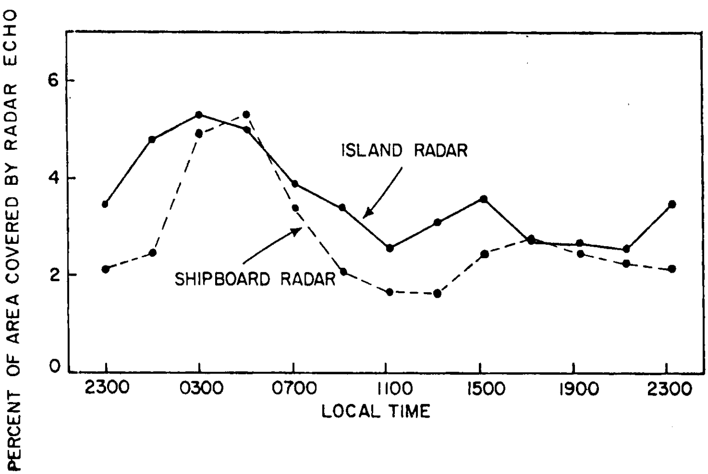



Figure 35. Average diurnal variation in radar echo amounts over the portions of the BOMEX square viewed by each radar during the 5-day period (from Hudlow, 1975).

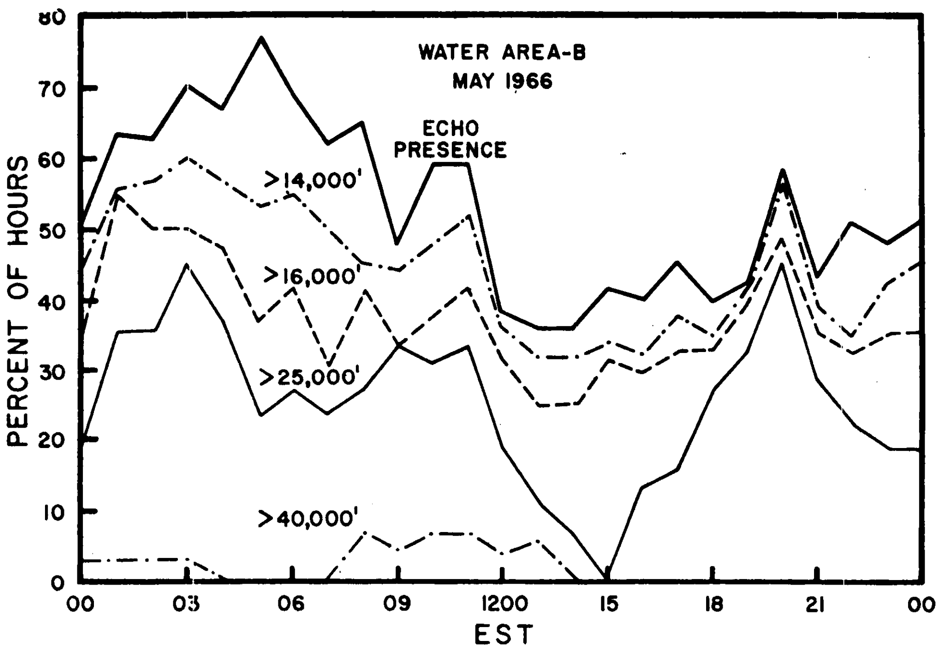



Figure 36. Diurnal variations in percentage of hours with echoes and with clouds above 14,000, 16,000, 25,000 and 40,000 ft during May 1966 over the ocean 40-100 n.m. east of Miami. (from Holle, 1968a)

<u>Tropical Land Observations</u>. Morning thunderstorm and/or rainfall maxima are a frequent phenomenon at a number of tropical and sub-tropical land stations. This lends support to the observed oceanic diurnal cycle. Frequently, as with the large island rainfall reports of Figures 19-22, these morning maxima are present with another afternoon heating peak of rainfall or thunderstorm activity. This has mislead many meteorologists into attempts to associate these double maxima with the semi-diurnal or $S_2$ pressure wave which does not appear to be relevant. The separate morning and afternoon convective peaks appear to be a result of different phenomena (discussed in Paper II).

Recent surveys of tropical meteorology by Atkinson (1971) and monsoon regions by Ramage (1971) have pointed out the often observed nighttime or morning maxima of land station rainshowers and thunderstorms. The following quotations from their reports further document the existence of nighttime and morning rainfall and thunderstorm maxima. From pages 6-26 and 6-27 of the Atkinson report:

> "Contrary to popular opinion many tropical land stations do not show a rainfall maximum during the afternoon period associated with maximum surface heating. Instead, many tropical continental stations show rainfall maximum during the nighttime hours" . . .

> "Ramage (1952) also studied the diurnal variation of rainfall at stations in East Asia during the summer months of May to August. He found a morning maximum between 0700 and 1000 LST to be prominent at tropical and subtropical coastal stations of southeastern China and southern Japan. Some coastal stations and all inland stations also showed a maximum during the afternoon so a pronounced semi-diurnal rainfall variation was evident at many stations" . . .

> "The diurnal variation of monsoon rainfall frequency at 17 stations in Sudan was analyzed by Pedgley (1969). Only one of the stations was located on the coast; nevertheless, a wide variety of diurnal rainfall types are found. An afternoon or early evening maximum is dominant only at stations far removed from the Ethiopian highlands" . . .

"Cocheme and Franquin (1967) presented diurnal rainfall-frequency variations for five inland stations in the semi-arid area of Africa south of the Sahara. These stations also showed a distinct preference for nighttime and early morning maxima . . ."

"Figure 6-21 (Figure 37 of this report) shows the diurnal variation of rainfall frequency at selected stations (10 stations) in Southeast Asia for the month with the maximum rainfall amount. All stations except An-Xuyen on the extreme southern tip of South Vietnam show the maximum frequency occurring sometime during the nighttime hours . . . . Important fact to note from Figure 6-21 is that diurnal variation in rainfall frequencies being at least double the minimum frequency at most stations."

Quoting now from the Ramage (1971) text:

"Except where mentioned below, nocturnal falls [of rainfall] are significant and have been reported over Africa (Hamilton and Archbold, 1945; Jeandidier and Rainteau, 1957), southern Asia (Iyer and Dass, 1946; Iyer and Zafar, 1946; Narasimham and Zafar, 1947; Ramaswamy and Suryanarayana, 1950; Rao and Raman, 1959), eastern Asia (Ramage, 1952), Indonesia (Braak, 1921-1929; Preedy, 1966) and northern Australia (Southern, 1969). It is tempting to try and relate this type of diurnal variation in the summer monsoon circulation (Ramage, 1952), but perhaps the explanation offered by Jeandidier and Rainteau (1957) and earlier suggested by the work of Refsdal (1930) will suffice until detailed studies are made. In a deep, strong convergent monsoon current, considerable cloud persisting through the night and morning inhibits insolation and development of a notable afternoon maximum. At night, radiational cooling from the tops and reduced surface cooling beneath an extensive cloud system presumably increase instability, cloud depth, and rainfall. During the day, radiational warming from the tops and reduced insolation at the surface presumably increase stability and consequently decrease rainfall." (p. 110)

"A strange regime prevails along the southern slopes of Mount Kilimanjaro (3°04'S, 37°22'E) during the southeast monsoon. According to Thompson (1957), thick altostratus or nimbostratus forms at night, giving widespread rains. Sometimes violent thunderstorms develop from embedded cumulomimbus. By afternoon, only scattered convective clouds remain." (p. 133).

Ramos (1975) has shown that the rainfall occurring with organized rain systems around Petrolina in the center of the Northeast Brazil dry zone occurs much more frequently, up to 77% at some locations, in the morning hours of 00-12 L.T. than in the 12-24 L.T. hours. Other

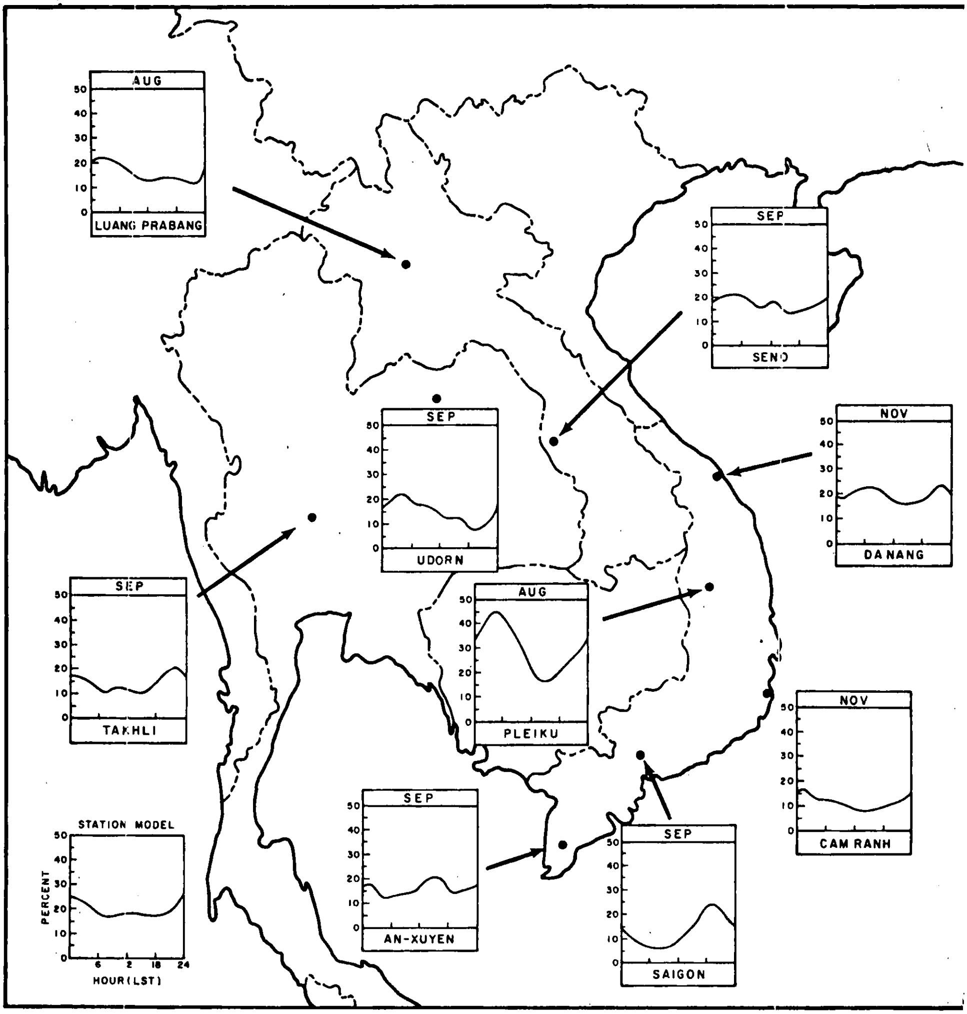



Figure 37. Diurnal variation of rainfall frequency during month of maximum rainfall amount at selected stations in Southeast Asia (from Atkinson, 1971).

researchers also report similar types of morning maxima in land precipitation and thunderstorm activity.

Middle Latitude Nocturnal Convection. This oceanic diurnal cycle should not be confined just to the tropics. It should also be detectable over the middle latitude oceans. Kraus's (1963) data support this. It should also be detectable over selected land areas in situations when the daytime surface energy fluxes to the atmosphere do not dominate. These latter conditions occur more often over land in late autumn and winter. Lower winter solar angles and greater path length lead to a larger ratic of troposphere to surface solar absorption and reduced surface heating. Land nocturnal convection maxima should also be detectable along coastal regions and in humid areas where the diurnal temperature cycle is damped. Snow surface conditions also reduce the surface diurnal temperature cycle.

Quoting from the beginning paragraph of Wallace's (1975) comprehensive study cf diurnal precipitation variation over the U.S.:

> "The documentation of the diurnal variability of rainfall has been the topic of well over a hundred articles, some of which date back to the middle of the 19th century. Ham (1901) attempted to synthesize the results of a large number of earlier investigations in terms of the following interpretation, based upon a simple geographical classification scheme:
>
> (a) In regions with continental climates most of the precipitation falls in convective showers during the afternoons while over the open oceans and in coastal regions with marine climates maximum rainfall occurs at night or during the early morning hours.
>
> (b) In some regions there are pronounced seasonal differences in the character of the diurnal variability. Over much of western Europe precipitation during the winter months exhibits a nocturnal rainfall maximum while during the summer months the maximum occurs during the afternoon. Over parts of the monsoon areas of the tropics there is a shift toward morning maxima during the wet season."

Speaking of wintertime precipitation and thunderstorms, Wallace (1975) makes these additional comments:

> " The 'heavy' precipitation events exhibit a pronounced diurnal cycle with a nocturnal maximum over much of the northern and and eastern part of the country. The maximum falls between 4 and 8 A.M. along most of the eastern seaboard and close to midnight in much of the midwest. The amplitudes are small in comparison to summer but over much of the northeast and the midwest they are large enough to account for about a 2:1 ratio in the frequency of heavy precipitation events between the times of maximum and minimum. "
>
> "Most of the heavy precipitation events during winter are associated with convective cells imbedded within frontal cloud bands of cyclonic storms. A few of these cells are vigorous enough to produce thunder. Thus, it is to be expected that the frequency of winter thunderstorms should display a diurnal variation similar to that of heavy precipitation. . . . . . "
>
> ". . . .From a practical standpoint the most significant finding that emerges from the analysis of the wintertime data is the documentation of a substantial diurnal cycle in convective activity with the highest frequency during the nighttime hours."

Excessive fall and winter IR cooling in the boundary layer over land and the development of temperature inversions apparently acts to cause a shifting of the morning convection maximum backward in time by a number of hours. This may explain the propensity for wintertime thunderstorms and heavy convection over the midwestern states to occur more between the hours of 23 to 03 L.T. (Wallace, _op_. _cit_.) rather than between 04-08 L.T. as happens along the U. S. coastal areas which experience less of a morning temperature decrease. Autumn and winter nocturnal cooling of the boundary layer often develops negative buoyancy which prohibits cumulus convection of any kind at hours around sunrise.

It is very tempting to see this similar type of extra nocturnal vertical mass recycling phenomenon in the following reported events:

1) The higher frequency of winter nighttime and morning thunderstorms over afternoon and evening thunderstorms in the U.S. Gulf Coast states and along the eastern seaboard up to New England revealed in the thunderstorm data compiled by Rasmusson (1971).

This data shows that the Gulf coast and eastern seaboard states experience between 55-65% of their November to March thunderstorm activity between the hours of 21 L.T. and 09 L.T. when the daytime heating cycle is at a minimum. Only 35-45% of thunderstorms occur between 09-21 L.T. when the daytime heating cycle is most pronounced.

2) Grant et al.'s (1974) report of sharp 04 L.T. early morning maxima in wintertime Colorado mountain snowfall, and

3) Brinkmann's (1974) report of wintertime chinook winds at Boulder, Colorado, occurring with much greater frequency during the nighttime hours.

Is it possible that these middle latitude nocturnal convective events are produced by physical processes similar to the processes operating to generate tropical nighttime convection?

Summary. These multiple reports of nighttime and/or morning deep convective maxima over land are so pervasive that it is highly unlikely that they can all be related to day vs. night local or regional terrain induced circulations (as might explain the U.S. midwestern summer nocturnal thunderstorm maxima) or by processes associated with land-sea breeze circulations. We believe that most of these over-land nocturnal or early morning peaks of heavy convective activity are primarily due to a physical phenomenon similar to that which produces peaks over the oceans.

### 3.4 Conflicting Evidence

Expectations were raised at the onset of the GATE program that similar large diurnal convective differences would be found. A consensus of many scientists who had worked on the BOMEX was that the convective cells which were to be studied were very active near sunrise and had often weakened by afternoon. However, when some select pieces of GATE data were initially analyzed, such diurnal ranges were not always observed. Marks (1974), for example, found little diurnal variation in

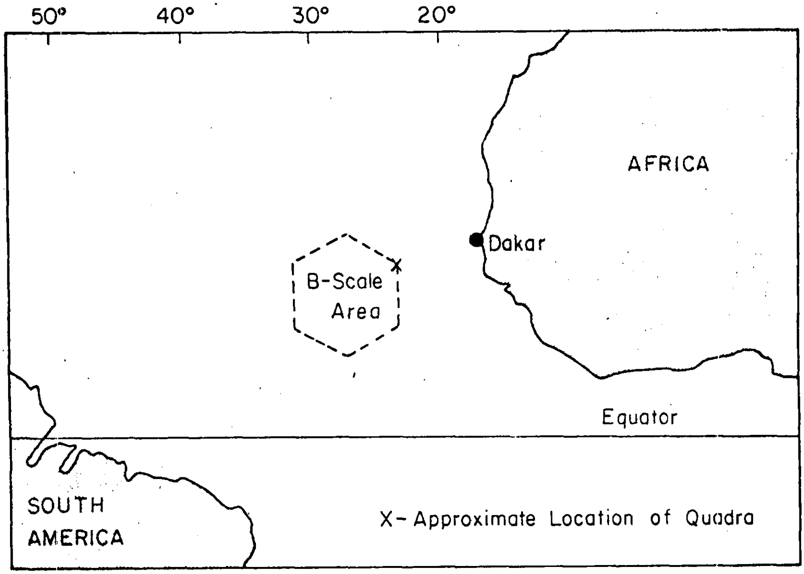



Figure 38. The GATE area.

mean maximum echo heights using radar from the ship Quadra (see Figure 38) and found the cloud coverage to be greatest in the afternoon. It may be that this weathership's proximity to Africa and its downwind location influenced the findings. When all of the GATE data is analyzed, a significant diurnal range similar to that found in the Pacific may yet be uncovered.

Other information recently furnished by J. Winston (1975) also does not support large diurnal cloudiness variations. Using monthly averaged IR information from the NOAA III satellite, little indication of a significant day-night difference in convection over the Pacific was present. Because the diurnal differences are primarily a result of Cb convection, they may have been largely smoothed out in the IR analysis which cannot well resolve individual cumulus elements.

# 4. DISCUSSION

The massive and multiple observational sources discussed in this study (rainfall reports, rawinsonde wind data with its various kinematic and dynamic calculations, satellite pictures, radar information, etc.) singularly and together support the existence of a sizable diurnal variation in deep cumulus convection over most of the tropics and subtropics. Although a morning maximum in oceanic and some land tropical shower rainfall has been qualitatively known to exist at a number of locations for many years, this phenomenon has yet to be well documented as a general global event or thought of as having as large an amplitude as here indicated.

The documentation of this large diurnal cycle in deep convection has been hindered by a lack of a similar diurnal variation in low level, layered, or total cloudiness, or in the frequency of less intensive precipitation. As deep cumulus convection is not a frequent phenomenon, its diurnal variability has been difficult to observe over the oceans. Island stations must have very long records before this diurnal peak can be well established. The recent satellite observations with ability to measure both day and night cloudiness have typically not been able to resolve the individual deep cumulus elements or have had them obscured by upper level cirrus shields. Their often reported statistical observerations of little diurnal variation of total cloudiness (the recent report by Gruger, 1975, for instance) should not mislead one to conclude that deep cumulus convection has a similar small diurnal amplitude. Cb convection and total cloudiness as measured by the satellite or from surface observations are often not well correlated. They should not be directly related on an hour by hour basis.

There appear to be three basic types of diurnal variation of deep cumulus convection in operation within the tropics and middle latitudes.

1. The typical afternoon-early evening maximum of cumulus occurring over land areas in association with the solar heating cycle. This is most pronounced the larger the range of diurnal temperature change. Deserts and semi-arid regions where evapotranspiration rates are low and regions with elevated terrain are most apt to experience this type of diurnal convective cycle. Nighttime and early morning surface cooling and development of low level temperature inversions prevent convection at these times. This regime is most prevalent during the summer. During the winter when the surface heat- is small, nocturnal and morning convection maxima can occur with the passage of well organized weather systems.

2. Oceanic convection regimes which typically have morning maxima and late afternoon-early evening minima. There are little summer and winter differences in this diurnal cycle.

3. Moist land areas where evapotranspiration rates are high and where the diurnal surface temperature cycle is supressed can, depending on the weather conditions, experience either the morning peaks of deep convection occurring with organized weather systems or the afternoon-evening peaks associated with afternoon surface warming. The morning peaks are more frequent in winter, the afternoon peaks in summer.

It is planned that the next phase of this observational research will concentrate on the diurnal variability of deep convection over the Atlantic. The research project of Prof. Gray is presently gathering twenty years of West Indies, Central America and Southeastern United States rawinsonde and precipitation data to study this diurnal cycle at another global location. The GATE data will also be carefully analyzed for similar deep convection diurnal differences. It is time that a thorough global analysis of this phenomenon be accomplished.

Paper II proposes a physical model to explain these diurnal variations in deep cumulus convection.

## ACKNOWLEDGEMENTS

The author is grateful to Prof. William M. Gray for proposing this research topic and for giving much guidance and encouragement. The author also wishes to thank Mr. Fred Stover, Mr. Edwin Buzzell, Prof. Eberhard Ruprecht and Ms. Kathy Shervis (of the Space Science and Engineering Library in Madison, Wisconsin) for their assistance in data gathering and/or analysis. The assistance of Mrs. Barbara Brumit, Ms. Genevra Metcalf, and Mr. Charlie Solomon in preparing the manuscript is greatly appreciated. This research has been financially supported by the United States Environmental Satellite Laboratory of NOAA.

# BIBLIOGRAPHY


Atkinson, G. D., 1971: Forecaster's Guide to Tropical Meteorology. Air Weather Service (MAC), U.S. Air Force.

Braak, C., 1921-1929: Het Climaat van Nederlandsch Indie. Magnet. Meteor. Observ. Batavia, Verhand. No. 8.

Brier, G. W. and J. S. Simpson, 1969: Tropical cloudiness and precipitation related to pressure and tidal variations. Quart. J. Roy. Meteor. Soc., 95, 120-147.

Brinkmann, W. A. R., 1974: Climatological study of strong downslope winds in the Boulder area. INSTAAR Occasional Paper No. 7, 229 pp.

Cocheme, J. and P. Franquin, 1967: An agroclimatology survey of a semi-arid area in Africa south of the Sahara. WMO Tech. Note No. 86, World Meteor. Organization, Geneva.

Defense Meteorological Satellite Program (DMSP) User's Guide, 1974: AWS-TR-74-250, Air Weather Service (MAC).

Federal Meteorological Handbook No. 1. DOC, DOD, DOT, effective January 1, 1970.

Finkelstein, J., 1964: Diurnal variation of rainfall amount on tropical Pacific islands. Proceedings of the Symposium on Tropical Meteor., Rotorua, New Zealand, 5-13 November, 1963, 286-294.

GARP report on the first session of the study group on tropical disturbances (Madison, Wisc., 21 October - 8 November 1968), Joint GARP Organizing Committee, WMO.

Garstang, M., 1958: Some meteorological aspects of the low-latitude tropical western Atlantic, Results of Crawford Cruise, No. 15. Woods Hole Oceanographic Institution, Rept. No. 58-42, 97 pp. (unpublished manuscript).

Golden, J. H., 1974: Scale interaction implications for the waterspout life cycle, II. J. Appl. Meteor., 13, 693-709.

Grant, L. et al., 1974: Weather modification - a pilot project. Final Report to the Bureau of Reclamation No. 14-06-D-6467. Dept. of Atmos. Sci., Colo. State Univ., Ft. Collins, CO.

Gray, W. M., 1973: Cumulus convection and larger-scale circulations, Part I: Broadscale and meso-scale considerations. Mon. Wea. Rev., 101, 839-853.

Gray, W. M., E. Ruprecht and R. Phelps, 1975: Relative humidity in tropical weather systems. Mon. Wea. Rev., 103, 685-690.

BIBLIOGRAPHY (cont'd)


Gruger, A., 1975: The diurnal variation of cloudiness over the GATE B area (outer hexagon). NOAA NESS, MSL Paper, 13 pp.

Ham, J., 1901: Lehrbuch der Meteorologie, 1st ed., Chr. Herm Tachnitz, Leipzig, 338-346.

Hamilton, R. A. and J. W. Archbold, 1945: Meteorology of Nigeria and adjacent territory. Quart. J. Roy. Meteor. Soc., 71, 231-264.

Holle, R. L., 1968a: Frequency and height populations of convective radar echoes over south Florida during May. ESSA Technical Memorandum ERLTM-APCL 4, 27 pp.

Holle, R. L., 1968b: Some aspects of tropical oceanic cloud populations. J. Appl. Meteor., 7, 173-183.

Hudlow, M. D., 1970: Weather radar investigations on the BOMEX. Research and Development Report ECOM-3320, U.S. Army Electronics Command, Fort Monmouth, NJ, 106 pp.

Hudlow, M. D., 1975: Radar and satellite precipitation analysis of a 5-day BOMEX data sample. NOAA Technical Memorandum EDS BOMAP-18, 46 pp.

Inchauspé, J., 1970: Diurnal precipitation variations over atolls of French Polynesia. La Meteorologie, Ser. 5, No. 16, 83-95.

Iyer, V. D. and I. Dass, 1946: Diurnal variation of rainfall at Mahabeleshwar. India Meteor. Dept. Sci. Notes, 9, 37-42.

Iyer, V. D. and M. Zafar, 1946: Diurnal variation of rainfall at Mahabeleshwar. India Meteor. Dept. Sci. Notes, 9, 37-42.

Jeandidier, G. and P. Rainteau, 1957: Prevision du temps sur le bassin du Congo. France Meteor. Nat. Monogr., 9.

Kidder, S. and T. Vonder Haar, 1976: Oceanic precipitation frequency from Nimbus 5 data. Forthcoming CSU Project Report.

Kraus, E. B., 1963: The diurnal precipitation change over the sea. J. Atmos. Sci., 20, 546-551.

Kung, E. C., 1975: Balance of kinetic energy in the tropical circulation over the western Pacific. Quart. J. Roy. Meteor. Soc., 101, 293-312.

LaSeur, N. E. and M. Garstang, 1964: Tropical convective and synoptic scale weather systems and their statistical contribution to tropical meteorology. Final Report on the Observational Field Programme to U. S. Army Electronics Res. and Dev. Lab. Grant No. DA-SIG-360 039-62-623 and U. S. Army Res. Office Grant No. DA-ARD-49-092-63-G23.

BIBLIOGRAPHY (cont'd)

Lavoie, R. L., 1963: Some aspects of the meteorology of the tropical Pacific viewed from an atoll. Univ. of Hawaii, Inst. of Geophysics Report No. 27, 76 pp.

Lindzen, R. S., 1967: Thermally driven diurnal tide in the atmosphere. Quart. J. Roy. Meteor. Soc., 93, 18-42.

Malkus, J. S., 1957: Trade cumulus cloud groups: Some observations suggesting a mechanism of their origin. Tellus, 9, 33-44.

Marks, F., 1974: Study of diurnal variations in convection using Quadra radar. Phase I and II. GATE Report No. 14, 191-204.

Martin, D. W., 1974: Characteristics of West African and Atlantic cloud clusters. GATE Report No. 14, 182-190.

Narasimham, V. L. and M. Zafar, 1947: An analysis of the hourly rainfall records at Poona. India Meteor. Dept. Sci. Notes, 9, 94-110.

Nitta, T. and S. Esbensen, 1974: Diurnal variations in the western Atlantic trades during the BOMEX. J. Meteor. Soc. Japan, 52, 254-257.

Pedgley, D. E., 1969: Diurnal variation of the incidence of monsoon rainfall over the Sudan. Meteor. Mag., 98, 97-107, 129-134.

Preedy, B. H., 1966: Far East Air Force Climatology, Headquarters, Far East Air Force, Singapore.

Ramage, C. S., 1952: Diurnal variation of summer rainfall over East China, Korea, and Japan. J. Meteor., 9, 83-86.

Ramage, C. S. 1971: Monsoon Meteorology. Academic Press, N.Y. 295 pp.

Ramaswamy, C. and N. Suryanarayana, 1950: Rainfall at Peshawar. Mem. India Meteor. Dept., 28, 121-137.

Ramos, R. Paes Leme, 1975: Precipitation characteristics in the Northeast Brazil dry region. J. of Geophysical Research, 80, 1665-1678.

Rao, K. M. and P. K. Raman, 1959: Diurnal variation of rainfall in India. In Meteorological and Hydrological Aspects of Floods and Droughts in India, (S. Basu, ed.), 186-191. Manager of Publications, New Delhi.

Rasmusson, E. M., 1971: Diurnal variation of summertime thunderstorm activity over the U.S. Technical Note 71-4, US ETAC. Scott AFB, IL, plus personal copy of massive printout data.

BIBLIOGRAPHY (cont'd)

Refsdal, A., 1930: Der Feuchtlabile Neiderschlag. Geofys. Publikasjoner 5, No. 12.

Riehl, H. L., 1947: Diurnal variation of cloudiness over the subtropical Atlantic Ocean. Bull. Amer. Meteor. Soc., 28, 37-40.

Riehl, H. L., 1954: Tropical Meteorology, McGraw-Hill Book Co., Inc., N.Y., 394 pp.

Ruprecht, E. and W. M. Gray, 1974: Analysis of satellite-observed cloud clusters. Atmos. Sci. Research Paper No. 219, Colo. State Univ., Ft. Collins, CO, 91 pp.

Southern, R. L., 1969: Personal communication to C. S. Ramage.

Takeuchi, T. and M. Nagatani, 1974: Oceanic thunder storms in the tropical and subtropical Pacific. J. of Meteor. Soc., Japan, 52, 509-511.

Trent, E. M. and S. G. Gathman, 1972: Oceanic thunderstorms. Pure and Appl. Geophys., 100, 60-69.

Thompson, B. W., 1957: The diurnal variation of precipitation in British East Africa. East African Meteor. Dept. Tech. Mem. No. 8

Wallace, J. M., 1975: Diurnal variations in precipitation and thunderstorm frequency over the conterminous United States. Mon. Wea. Rev., 103, 406-419.

Warsh, K. L., W. A. Brower and M. Garstang, 1972: Comparison of sensible and latent heat fluxes computed from measurements by ship and spar buoy in the tropical Atlantic Ocean. Sonderdruck aus der Deutschen Hydrographischen Zeitschrift, Band 25, 62-79.

Weickmann, H. K. and A. B. Long, 1975: The cycle of convective clouds over the tropical ocean - supernova storms. Seminar presented at NCAR, Boulder, CO, February 13, 1975.

Williams, K. T. and W. M. Gray, 1973: Statistical analysis of satellite-observed trade wind cloud clusters in the western North Pacific. Tellus, 25, 313-336.

Winston, J., 1975: Personal communication to William M. Gray

Zehr, R., 1976: Factors favorable to typhoon genesis as revealed by 10-years of composited rawinsonde observations. (Forthcoming CSU report).

# DIURNAL VARIATION OF OCEANIC DEEP CUMULUS CONVECTION

## PAPER II: PHYSICAL HYPOTHESIS

by

William M. Gray

ABSTRACT

This paper accepts as realistic the observational evidence of Paper I on the large observed morning maximum and afternoon-evening minimum of oceanic deep cumulus convection (also frequently observed over land). It proposes that the radiation differences between the cloud regions and the surrounding cloud free regions of tropical weather systems are a fundamental driving mechanism of these systems. It also proposes that the large diurnal range of deep convection is primarily a result of diurnal differences in the radiational cooling profiles of cloud regions vs. surrounding regions.

The atmosphere surrounding cloud regions adjusts to its larger radiational cooling at night by extra subsidence. This extra nighttime subsidence produces extra convergence into the cloud regions at low and middle levels. During the day surrounding region subsidence warming and cloud region convergence is substantially reduced. At upper levels the cloud regions radiationally cool more at night but less during the day than their surrounding cloud-free regions. This acts in a complementary fashion with conditions at lower levels to enhance upper level cloud region nighttime divergence and reduce it during the day.

Reasons are given why this enhanced cloud region nighttime convergence-divergence relieves itself primarily by forming extra deep cumulus clouds and not by changing lower and middle level cloudiness or light precipitation. This day-night cloud region diurnal convergence cycle often follows the radiational forcing with a time lag of 3-6 hours.

A comparison with previously hypothesized explanations for this diurnal cycle is given. A discussion of the likely fundamental role of radiation in tropical weather system maintenance and tropical cyclone formation is advanced.

TABLE OF CONTENTS

## 1. ROLE OF RADIATION IN THE MAINENANCE OF TROPICAL WEATHER SYSTEMS

This paper accepts the observational evidence of Paper I on the existence of a significant diurnal cycle in deep cumulus convection over most of the global oceans and over many of the tropical and middle latitude land areas. The purpose of this paper is to emphasize the likely fundamental role which cloud-cloud free radiational differences play in the structure and maintenance of tropical and middle latitude weather systems and to propose this as the physical explanation for the observed diurnal cycle in oceanic deep cumulus convection.

Radiation vs. Condensation Energy Gain. It may appear surprising to some readers that radiation and not condensation forcing is proposed as the primary explanation of the cloud cluster vertical divergence profile. The magnitudes of the cloud cluster condensation energy releases and the diurnal differences in condensation are approximately three to five times larger than the radiational energy differences between the cluster and its surroundings. Why then does radiation act as the apparent primary driving mechanism for these clusters? The answer to this question rests with the physics of how these two energy sources manifest themselves.

Radiation is a static and direct energy source or sink. It varies with temperature, humidity, the sun angle, and with the presence or absence of cloud layers at other levels, etc. It is not directly related to vertical motion and is always occurring. By contrast, condensation energy release is directly tied to upward vertical motion. It typically does not bring about direct environmental sensible temperature gains but goes primarily into potential energy gains (Gray, 1973). The troposphere experiences warming due to condensation only as a consequence of compensating dry adiabatic sinking motion (Lopez, 1973). Much of this compensating sinking motion occurs at locations removed from the cloud areas. The subsidence warming occurring within the cloud areas is largely expended in balancing the cloud region liquid water reevaporation due to cloud detrainment. Williams and Gray (1973) and Ruprecht and Gray (1974) have demonstrated that the direct influence of the typical $4^{o}$ diameter cluster condensation (equivalent to 1000-1500 calories/$cm^2$ per day) on cluster sensible temperature change is nearly zero. Cluster cumulus clouds typically exert only a very small sensible heat change upon the cluster itself. Most cumulus condensation energy is exported to the surrounding regions to balance the outer subsidence. <u>Cloud cluster minus surrounding cloud-free region radiation induced energy differences can thus be considerably larger than those directly produced by condensation warming</u>. It then appears feasible that cloud cluster condensation may occur more in response to radiational influences than the other way around.

<u>Radiation Produced Vertical Circulations</u>. Albrecht and Cox (1975) have recently discussed the significant role of daily average IR radiation in producing divergence differences between the cloud regions of an assumed easterly-wave trough (so-called perturbed area) and the up- and downstream ridge regions of this wave which are assumed to be largely cloud-free (designated unperturbed areas). Their trough region clouds were taken to be typical of conditions in the Line Island area. They assumed variable cloud tops between 400-250 mb which are significantly lower than the typical West Pacific cloud cluster tops of 150-250 mb. Their perturbed minus unperturbed vertical radiational forcing and resultant wind divergence profiles are shown in Figures 1 and 2. Note the

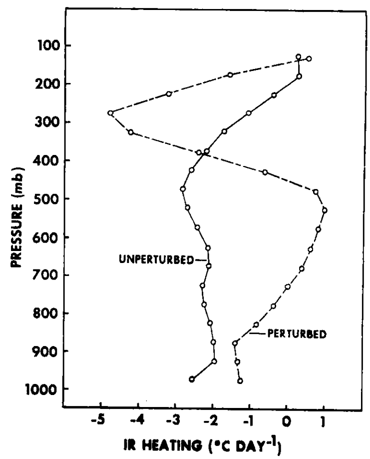



Figure 1. Observed infrared heating profiles for perturbed days (24 March, 2, 10, 16 April) and for unperturbed days (26 March, 4, 6, 7, 8, 12, 14, 28 April). (From Albrecht and Cox, 1975).

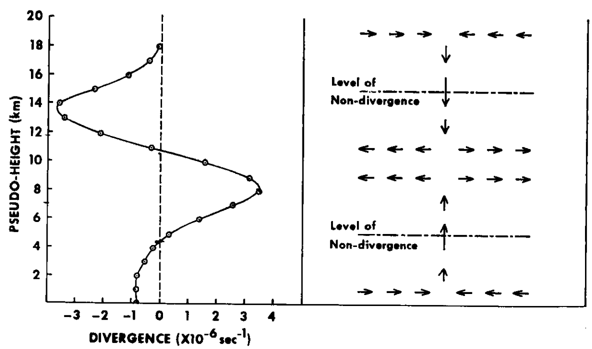



Figure 2. Radiationally forced divergence at latitude and longitude of maximum perturbed radiative heating (a) and sketch of circulation forced by the radiative heating (b). (From Albrecht and Cox, 1975).

magnitude and sign of the radiation induced divergence in their perturbed region. Following Reed and Recker (1971) they assumed that the mean distance between their perturbed and unperturbed regions was $\sim$ 2000 km, or about 2-3 times larger than our observed typical cluster to clear region distance scale of Paper I. Had their assumed wavelength (between perturbed and unperturbed areas) been shorter by a factor of 2-3, their radiation induced divergence values would have consequently been 2-3 times larger. Their analysis of the wind adjustment to radiational infrared (IR) energy forcing indicates that daily average IR radiation differences between cloud and cloud-free regions can exert a powerful influence on the cloud region vertical motion.

Their results indicate that the upper layered clouds of organized weather systems are largely opaque to infrared (IR) energy penetration. They prevent upward IR energy loss from lower layers and produce a net flux convergence of IR energy in the layers underneath the cloud tops. At the cirrus top level the situation is reversed. Downward energy propagation from levels above the clouds is quite small and the IR loss at the cloud tops is large. A large net flux divergence of IR energy results. By contrast, the upper levels of the surrounding cloud-free regions are able to absorb IR energy propagating upward from lower levels. These upper level cloud free areas radiatively cool through IR energy loss at rates significantly less than those of upper cluster cloud shields. At lower levels, however, the cloud-free regions radiatively cool at rates significantly greater than the cluster regions.

Figure 3 portrays the estimated values of the daily average IR radiation induced temperature change within the typical West Pacific summertime cloud cluster with a thick cirrus shield and within the surrounding cloud-free regions as derived by the author from discussion

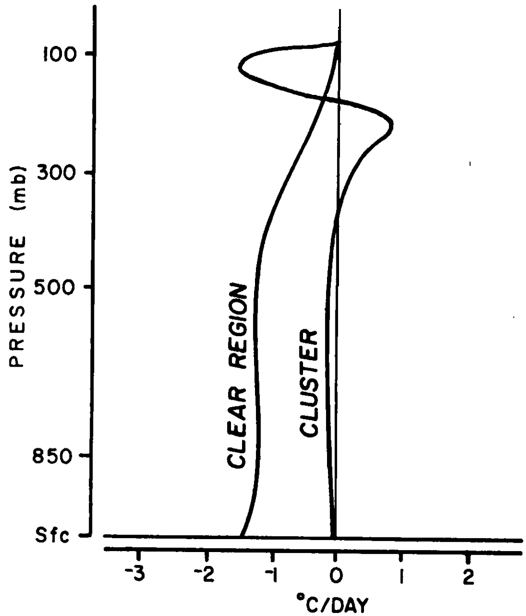



Figure 3. Comparison of typical net radiation induced temperature changes in the cloud cluster and in a tropical clear region from information supplied by Cox (1975) and from the various quoted radiation papers.

with S. Cox (1975) and from information of his quoted radiation papers (Cox, 1969a, 1969b, 1971a, 1971b, 1973). Note the sizable difference in radiation induced IR energy flux at all levels between the cloud cluster with thick cirrus and in the surrounding clear regions. It is hypothesized that these cloud cluster and surrounding region radiational differences are primarily responsible for the cloud cluster or tropical weather systems maintenance. The observation of a rather uniform convergence pattern in the cloud cluster from the surface to 400 mb as found by Williams and Gray (1973), Ruprecht and Gray (1974), Yanai et al. (1973) and Reed and Recker (1971) would argue more for the role of a radiational rather than a boundary layer frictional convergence or CISK[1] mechanism.

Although boundary layer frictional convergence is quite important for maintenance of low level mass and water vapor convergence and in the initiation of cloud shields, it appears that the later cloud cluster structure and maintenance cannot be explained without invoking another physical mechanism to explain the deep layer convergence. Once low level relative vorticity is present and cloud shields are produced, it appears that cloud-cloud free radiational forcing takes over and is the primary mechanism for cluster organization and maintenance. It is obvious that the cloud cluster boundary layer convergence can account for only 20-25% of the total mass convergence into the cluster (see Figures 4 and 34 of Paper I). Multi-day conservatism of organized tropical disturbances is well known. This weather system organization cannot be explained solely by the concentration of cluster low level vorticity. The vorticity within the cloud cluster is not substantially different than that in its surrounding region. By contrast, the radiational energy differences between the cluster and its surroundings are substantial.

It is important that much more attention be given to the likely influence of radiation as a basic dynamic forcing function for weather systems which exist in conditions of weak baroclinicity. In a recent numerical modeling study Pelissier (1975) has also hypothesized cloud and cloud free radiational differences as being the fundamental mechanism to the development and maintenance of sub-tropical upper-tropospheric cold-core cyclones. These cloud-free cyclones form only in the summer

[1]Conditional Instability of the Second Kind as defined by Charney and Eliassen (1964).

in conditions of very weak vertical wind shear. Cloud regions exist at the periphery of these cyclones. When the large scale baroclinicity fields are weak, cloud and cloud-free radiational differences can become a primary source of available potential energy.

Cloud Cluster Divergence Profile and Deep Convection Linkage. Why and how are variations in the 900-400 mb convergence into the cloud cluster related to variations in deep cumulus convection or Cb activity? The author believes that this above boundary layer mass convergence into the cluster can be thought of as going primarily into Cb downdrafts. The more cluster 900-400 mb mass convergence, the more downdraft mass there is to penetrate into the cluster boundary layer, and consequently the more mass to support additional boundary layer Cb growth. The 900-400 mb convergence into the cloud cluster also brings more water vapor into the cluster to support the extra vapor losses from additional Cb rainfall. Williams and Gray (1973) have previously demonstrated that about 60% of the water vapor convergence into the average cloud cluster occurs above the 900 mb level. Other research at CSU on the tropical cloud cluster by Gray (1973), and Ruprecht and Gray (1974) has demonstrated that the up-and-down vertical circulation at the top of the average cloud cluster boundary layer is an order of magnitude larger than the mean upward motion at this level.

To balance simultaneous mass, moisture, and energy budgets for the average cloud cluster it is necessary to hypothesize that most of the mass convergence above 900 mb goes into accelerating downdrafts which penetrate into the boundary layer. The mass entrainment and detrainment into and out of the cumulus updrafts between 900-400 mb is believed to be primarily a self cancelling process. Mass detrainment from shallow cumulus feed the extra entrainment into the deeper cumulus, etc.

The relatively dry and low $\theta_e$ air converging between 400-900 mb does not reduce the cumulus buoyancy if it is not entrained into the cumulus updrafts but instead goes to feed the cluster accelerating downdrafts. In addition, the cluster divergence occurs at the cirrus levels and must be compensated for by upward vertical motion from lower tropospheric levels. The vertical stability of the middle and upper levels dictates that the mass replacement for the cirrus divergence comes not from the middle levels where $\theta_e$ values are quite low but rather from Cb clouds whose roots are mostly below 900 mb.

If these conceptual views of the tropical cloud cluster are approximately correct, then a direct relationship between the intensity of the cloud cluster radiational developed deep inflow convergence from 900 mb to 400 mb and the number of Cbs and intensity of Cb convection rainfall should be expected. The greater the strength of the mass circulation through the cloud cluster, the more frequent and/or intense should be the Cb rainfall.

## 2. DIURNALLY VARYING RADIATIONAL COOLING PROFILES

The differences in rates of radiation energy loss-gain within the cloud cluster and in the surrounding clear or partly cloudy regions undergo considerable day-night variations.

Day-Night Variation in Cirrus Radiation Energy Loss-Gain. The IR radiation energy changes within the cirrus shield do not undergo significant diurnal changes. There are however, significant differences in solar energy gain.

The opacity of Cb produced cirrus clouds to solar energy is very much larger than that of cirrus produced by the large scale vertical motion processes associated with middle-latitude storm systems. The latter are, to a considerable degree, transparent to solar energy while the former can become almost totally opaque to solar energy. Active Cb convection often produces Cb shields which attenuate nearly all of the incoming solar energy.

Assuming that our typical cloud cluster produces cirrus clouds between the levels of 150-250 mb and that a third of the incoming solar energy not reflected or scattered back to space is absorbed within the 150-250 mb layer, a considerable solar energy gain to these upper cloud cluster layers can result. The enhanced absorption of solar energy within the cirrus is produced by the large increase in path length of the incoming radiation due to scattering by the frozen cirrus particles and to some absorption by these frozen particles. Assuming a 70% albedo off of the cluster cirrus shield and penetrating Cb cloud tops, and assuming incoming solar energy of 800 ly/10 hours, then the resulting 150-250 mb energy gain would be 80 ly/10 hours. This would give a layer warming of $\sim 3.2^{o}C/10$ hours. This energy gain is of a magnitude larger than the expected $1.5\text{-}2.0^{o}C/10$ hours IR rate of energy loss in the cirrus shield.[2] A positive cirrus shield radiation energy gain should thus be expected during daylight hours. As the saturation vapor contents of the air with respect to ice at these upper levels are quite low ($\sim$ 0.25 g/kg at 250 mb, $\sim$ 0.08 g/kg at 200 mb, and $\sim$ 0.03 g/kg at 150 mb) any evaporation or phase change cooling must be extremely small, certainly no more than $\sim \frac{1}{2}^{o}C/10$ hours which is insignificant. In addition, the energy loss due to horizontal mass convergence and/or ventilation out of the cirrus layer will be small due to the observed weak temperature gradient between the cluster and its surrounding region. Thus, if thick semi-opaque cirrus is present to the extent that 10% of the incoming solar energy is absorbed in this layer, a cancelling of the IR radiation loss and a slight net radiation energy gain to the cirrus layer should be expected during the daylight hours.

Fleming and Cox (1974) have estimated similar but weaker daytime cirrus warming (See Figure 4) from less absorbent cirrus shields than are to be expected over organized West Pacific cloud clusters.

<u>Day-Night Variations in Clear or Partly Cloudy Regions.</u> The cloud-free troposphere undergoes continuous radiational cooling. At night when no solar heating is available, this radiational cooling is nearly twice as large as in the daytime (see Dopplick, 1974 and Cox's references, also Figure 5). Because the troposphere rapidly responds to

---

[2]The cirrus shield is assumed to be a composite, one with variable tops between 150-250 mb. Cb penetrations of the cirrus shield and layer divergence should produce enough mixing such that the narrow layer of intense radiational cooling just at the top of the cirrus can be expected to mix and cool throughout a deeper layer.

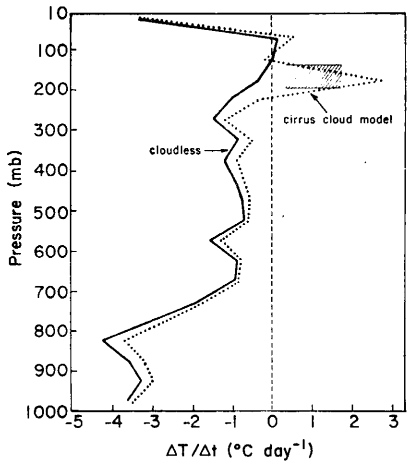



Figure 4. Daily averaged radiational heating rates for a cloudless tropical atmosphere and for the same atmosphere with thin cirrus clouds (from Fleming and Cox, 1974).

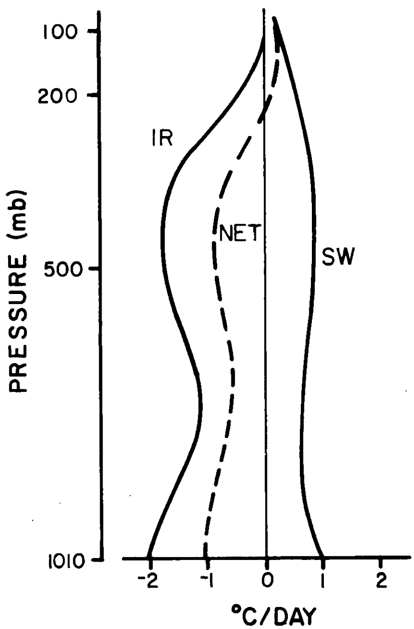



Figure 5. Typical daily amount of tropical clear region infrared (IR) cooling and solar (or short wave) warming (SW) as derived by Cox (1975) and Dopplick (1974). The dashed curve gives the typical net daily radiation induced temperature change.

these day vs. night radiational differences, they only partially manifest themselves in proportional day to night observable temperature and pressure-thickness changes.

## 3. PHYSICAL HYPOTHESIS FOR OBSERVED DIURNAL CYCLE

If the radiational differences between the cloud cluster and its surrounding region are indeed a primary energy source for the cloud cluster vertical divergence profile, then, because of the nearly two to one night vs. day differences in the cloud cluster and the clear region radiational cooling, one should expect substantial day vs. night differences in the cluster vertical divergence profile. Since these day vs. night divergent differences are observed in the composited wind data then additional supporting evidence for the hypothesis of radiational forcing of cloud clusters is available.

The author believes that the diurnal deep convective cycle is a result of the regional vertical recycling response of the troposphere to day vs. night radiational differences between cloud areas and the cloud-free regions which surround them. To the author's knowledge this mechanism has not been previously proposed. He believes that this is quite probably the primary physical mechanism for the large observed morning to evening differences in deep cumulus activity.

The day vs. night radiational differences between the cloud cluster and the surrounding cloud-free areas at cirrus levels are such that a greater outward pressure gradient is developed at night than is present during the day. The reverse situation exists in the middle and lower troposphere where at night there develops a greater inward directed pressure gradient to the cluster than occurs during the day. The cluster-clear region radiational differences of the upper troposphere thus act opposite to the radiational differences of the middle and lower troposphere. Each will now be discussed separately.

Upper Tropospheric Considerations. It is assumed that the area around the cloud cluster has a radiation budget similar to that of our clear regions. The upper levels of the cluster have greater rates of radiational energy gain during the day and loss at night than do the surrounding clear regions at these levels. This day-night radiational energy reversal is portrayed in Figure 6.

It is assumed that near and above 100 mb there are no significant differences in the height gradients between the cluster and surrounding areas although the absolute value of these heights rises during the day and lowers at night. Thus, the extra nighttime IR cooling off the top of the cirrus layers between 150-250 mb results in pressure-thickness decreases and consequent rises in the pressure surfaces relative to the surrounding cloud-free areas. The smaller nighttime cooling rates in the surrounding clear regions yield less 150-250 mb thickness reduction. The surrounding pressure surfaces are thus lowered relative to those of the cluster. A larger nighttime outward cloud cluster height gradient and outward mass divergence is established.

During the day these radiation processes are reversed. Solar warming increases the temperature in the top layers of the cloud cluster more than in the surrounding cloud free regions. The pressure thicknesses are increased more over the cluster than in the surrounding regions. These thickness increases within the upper cirrus levels cause a relative bulging down of the cluster pressure surfaces compared with those of the surrounding regions. This causes a weakening of the daytime outward cluster pressure gradient and mass divergence.

Middle and Lower Tropospheric Variations. At these levels the cluster and surrounding region relative radiational energy losses are

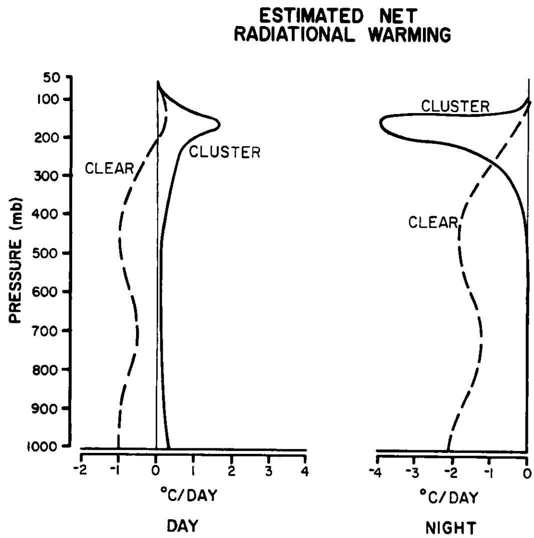



Figure 6. Estimated typical day and night net radiational warming within the cloud cluster and in the surrounding clear or mostly clear regions.

reversed from those occurring at the cirrus levels. The upper cloudiness of the cluster protects its middle and lower levels from significant IR energy loss or solar energy gain. The day and night cluster radiational losses are small and about equal. The surrounding cloud free areas, however, experience nearly two to one day vs. night differences in their radiational losses at middle and lower levels. The surrounding region pressure thicknesses decrease at night and bulge upward relative to the cluster pressure surfaces. This enhances nighttime low level mass convergence into the cluster and compensates for the extra nighttime mass divergence from the cluster at cirrus levels. During the day, the extra solar warming surrounding the cluster increases these pressure-thicknesses more than those of the cluster. The surrounding region pressure surfaces bulge downward relative to the cluster surfaces. This decreases the inward cluster pressure gradient. The daytime inward mass convergence is consequently reduced. This weaker daytime convergence also compensates for the weaker daytime divergence from the cluster at cirrus levels.

Figure 7 portrays estimated cluster minus surrounding clear region daytime (left figure) and nighttime (right figure) radiational energy gain-loss differences. Notice that at middle and lower tropospheric levels the daytime radiational energy differences between the cluster and its surroundings are but half of the nighttime values. At the cirrus levels the cluster and surrounding radiational energy differences reverse themselves.

Figure 8 portrays the author's conceptual view of the typical day and night slope of the pressure surfaces within and surrounding cloud clusters. Note that the lower stratosphere ($\sim$ 80 mb) pressure surface warms and rises during the day (dashed curve) and sinks at night (solid curve), but the cluster-surrounding region pressure gradient is hypothesized not to experience a significant diurnal variation. Below this level, however, the cluster cloudiness induces significant day vs. night differences in the slopes of the pressure surfaces.

Thus, it appears that day vs. night differences in cloud cluster and surrounding region solar absorption can produce a significant diurnal modulation of the cluster inward-outward height gradients and consequently the cluster mass divergence profiles.

<u>Possible Stratospheric Influence</u>. Although not included as a primary factor in this discussion, it is likely that the high percentage of back scattered and reflected solar energy off the tops of the cloud clusters (here assumed to be 70%) produces an extra stratospheric energy gain above the cloud cluster which is not present over the cloud-free

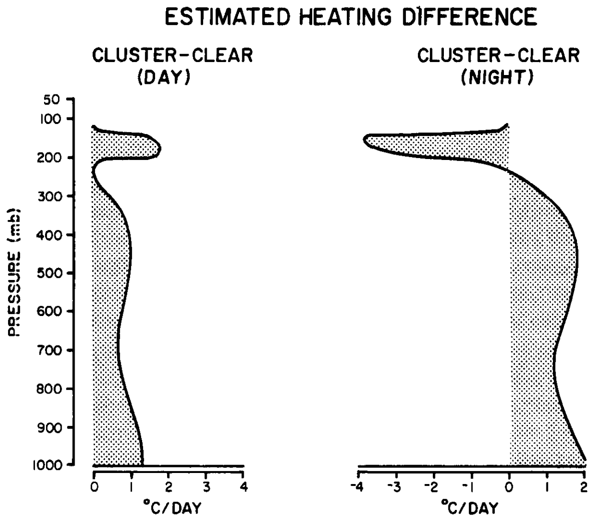



Figure 7. Estimated day vs. night cloud cluster minus surrounding clear region radiation induced temperature changes.

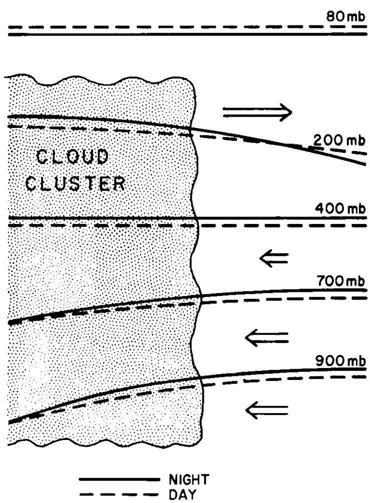



Figure 8. Idealized slope of the cloud cluster to surrounding clear region pressure surfaces during the day (dashed curve) and at night (solid curve).

areas. This would likely lead to a daytime thickness increase of the stratospheric pressure surfaces and reduction of the upper level cluster daytime outward height gradients. This would reduce the daytime upper level cluster divergence. These stratospheric daytime warmings are difficult to detect because of the complicating influence of the Cb overshoot cooling and the time lag between the deep convection and the radiational forcing (to be discussed). Nevertheless, this influence, if significant, should act in the same sense as the tropospheric solar radiational influences to reduce the cluster daytime upper level outward pressure gradient and divergence.

# 4. SUPPORTING EVIDENCE

Although difficult to measure accurately, observed cloud cluster surrounding region 00Z vs. 12Z pressure thickness and height gradients generally support this diurnal radiational difference argument. Only a portion of the radiation-specified cluster minus surrounding region energy differences are observed as temperature differences. This indicates that the atmosphere is responding to most of these radiationally forced temperature differences with compensating vertical circulations. In addition, the 00Z vs. 12Z differences in the height gradients between the cloud cluster and its surroundings support a substantial difference in 00Z vs. 12Z cluster convergence-divergence. 00Z vs. 12Z differences in regional vertical circulation are negligible, indicating that the diurnal variations in cloud cluster and surrounding region vertical circulation are mutually compensating. The diurnally varying divergence profile is of comparable magnitude with the daily average divergence profile.

Diurnal Variations in Clear Region Pressure-Thickness. Diurnal variations of temperature and pressure thickness are difficult to measure accurately. However, the composited rawinsonde data of tropical clear region 00Z vs. 12Z pressure-thickness changes show that these thickness changes are substantially less than those which would be prescribed by radiational cooling alone. This indicates that the clear atmosphere subsidence is larger at night than during the day. This subsidence cancels out most of the daily net radiational energy losses of the clear regions and a significant portion of the day to night differences in radiational cooling.

TABLE 1

Comparison of Theoretical Radiation Specified and Observed Clear Region 12 Hour Thickness Changes in Meters

| Layers | Radiation Forced 12-Hour Thickness Changes | | Observed Difference in 00Z-12Z or 12Z-00Z Thickness | | |
|---|---|---|---|---|---|
| | Day Ave. | Night Ave. | West Pacific Clear Region | West Indies Clear Region | Clear Region Surrounding the Typhoon |
| 300-500 mb | -9 | -17 | ± 7 | ± 4 | ± 5 |
| 500-850 mb | -8 | -14 | ± 4 | ± 4 | ± 1 |
| 300-850 mb | -17 | -31 | ± 11 | ± 8 | ± 5 |

Table 1 compares clear region radiational induced night and day pressure-thickness changes to observed changes. The first two columns show the thickness changes which would result from radiational cooling alone. The last three columns show the observed clear region 00Z and 12Z thickness difference from our composited rawinsonde data (these compositing techniques are discussed in reports of Williams and Gray (1973), Ruprecht and Gray (1974), Zehr (1976) and Frank (1976)). The exact magnitudes of the observed clear region thickness differences are subject to some error since the 00Z and 12Z soundings do not occur at sunrise and sunset. Nevertheless, it can be seen that the observed clear region 12-hour thickness changes are substantially less than the 12-hour radiation specified changes. Often the observed and radiation specified changes are of opposite sign.

If the troposphere did not respond to day vs. night differences in net radiation, the observed 00Z and 12Z thickness differences would be considerably larger than the values shown. Instead, the troposphere appears to selectively adjust itself to most of its night and day radiational energy losses. The larger nocturnal radiational cooling in the cloud-free areas is mostly compensated for by extra subsidence warming and increased lower tropospheric mass divergence. This produces extra nighttime convergence into the adjacent cloud areas. During the day, solar heating is available in the cloud-free areas to compensate for 40-50 percent of the infrared (IR) cooling. The horizontal low level mass convergence into and out of the cloud clusters is consequently lessened.

Diurnal Variation of Cluster Minus Surrounding Region Pressure-Thickness. Table 2 (left column) lists the night and day 12-hour thickness change differences between various pressure levels which would result from radiational cooling differences between the cloud cluster and the surrounding clear regions as estimated from Figure 6. These cluster minus surrounding radiation induced thickness change differences are considerably larger than the observed cluster minus surrounding region thickness changes between 00Z and 12Z for four classes of weather systems in the West Pacific and in the West Indies (right columns of this table). Although these measured 12-hour thickness differences are subject to instrument inadequacies and are not taken at the hours which would show maximum radiational influences (sunrise and sunset), they do indicate that the observed cluster minus surrounding region thickness gradients are considerably weaker than the cluster minus surrounding thickness gradients which would occur if the full day and night radiational forcing differences were to manifest themselves. It is suggested that these radiational forcing differences establish horizontal pressure accelerations and vertical circulation changes which feed-back and cancel out a large portion of the radiation specified gradients.

TABLE 2

Comparison of Expected and Observed Magnitude of Cloud Cluster Minus Surrounding Region 12-Hour Thickness Changes Between 850-100 mb in Meters Per 12 Hours

| Layers | Expected magnitude of 12-hour cluster minus surrounding region thickness changes predicted from radiation data of Figure 6. | | Magnitude of observed 12-hour difference in cloud cluster and surrounding region thickness change (00Z-12Z, or 12Z-00Z) | | | |
|---|---|---|---|---|---|---|
| | Day Average | Night Average | Non-developing West Pacific clusters | Non-developing West Indies clusters | Intensifying cloud clusters | Typhoons |
| 100-200 mb | +16 | -22 | ± 8 | ± 6 | ± 5 | ± 8 |
| 300-500 mb | +7 | +12 | ± 1 | ± 3 | ± 3 | ± 3 |
| 500-850 mb | +6 | +11 | ± 3 | ± 4 | 0 | ± 1 |

This response to the radiational forcing does not balance all of the cloud and cloud-free radiation induced thickness differences, however. Some diurnal variations in the cluster-clear region height gradients do occur, and these height gradient differences produce the day-night differences in divergence-convergence.

00Z vs. 12Z Height Gradients. Observed 00Z vs. 12Z West Pacific cloud cluster outward height gradients are shown in Figure 9. Gradients are shown from the cluster center out to $8^{\circ}$ radius. Note the large diurnal difference in the outward 150-250 mb height gradient from cloud cluster center to $4^{\circ}$ radius. Also note the slightly larger inward pressure gradients at middle and low levels at 00Z.

Note the 00Z vs. 12Z reversal of pressure gradient at 100 mb compared with those of the 150-300 mb levels. The cross-over level of no 00Z vs. 12Z slope difference must be somewhere between 100 and 150 mb. It is to be expected that some convergence occurs near the 100 mb level or above to balance the overshooting Cb tops which entrain mass and then collapse back into the cirrus shields.

Although the accuracy of the upper level heights is often questioned due to the possible errors resulting from addition of systematic instrument inaccuracies, there may not be significant errors in the day vs. night differences in the gradients of these pressure-heights. They are believed to be generally correct. Similar 00Z vs. 12Z height gradients have been obtained from other independent cloud cluster data sets which we are currently processing.

Table 3 lists observed 00Z (10 L.T.) minus 12Z (22 L.T.) height gradient differences between the cluster center and $4^{\circ}$ radius at several pressure levels (column A). This table shows the resulting 00Z minus

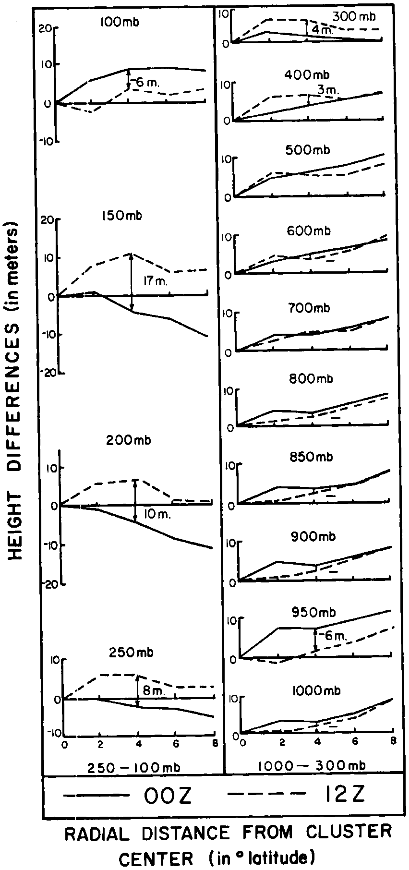



Figure 9. Comparison of observed 00Z and 12Z outward height gradients of pressure surfaces from the cloud cluster center to the 8° radius surrounding region.

12Z outward pressure accelerations (column B). The resulting outward radial wind changes per 6 hours (column C) and divergence changes per 6 hours (column D) which would occur if these 00Z and 12Z height gradient differences were acting are also shown. Note that these divergence differences qualitatively agree with the observed 00Z vs. 12Z differences in divergence (column E) as indicated in Figure 4 of Paper I, but their magnitudes are larger at upper levels. Undoubtedly other influences, such as the feed-back of the Cb convection, are acting to

TABLE 3

Observed 00Z Minus 12Z Differences in Mean Outward Height Gradients between the Cluster Center and its 4° Radius and Resulting Outward Accelerations and 6-Hourly Wind and Divergence Changes

| Pressure Level (mb) | A Cluster Center Minus 4° Radius Height (m) | B Outward Acceleration Differences $(-g\frac{\partial Z}{\partial r})$ (in $10^{-3}$cm/sec$^2$) | C Outward Wind Change per 6-hr from this acceler. (m/sec) | D Div. Change per 6-hrs. from outward acceleration ($10^{-6}$sec$^{-1}$) | | E 00Z-12Z Observed Divergence Differences ($10^{-6}$sec$^{-1}$) |
|---|---|---|---|---|---|---|
| 100 | -6 | -11 | -2.4 | -11 | Inflow | -3 |
| 150 | 17 | 39 | 8.4 | 38 | | 7 |
| 200 | 10 | 23 | 5.0 | 23 | | 7 |
| 250 | 8 | 18 | 3.9 | 18 | Outflow | 7 |
| 300 | 4 | 9 | 1.9 | 9 | | 6 |
| 400 | 3 | 7 | 1.5 | 7 | | 2 |
| 500 | -2 | -5 | -1.1 | -5 | | -1 |
| 600 | -2 | -5 | -1.1 | -5 | | -2 |
| 700 | 0 | 0 | 0 | 0 | | -3 |
| 800 | -2 | -5 | -1.1 | -5 | | -4 |
| 850 | -3 | -7 | -1.5 | -7 | Inflow | -4 |
| 900 | -2 | -5 | -1.1 | -5 | | -4 |
| 950 | -6 | -14 | | | | |
| 1000 | -2 | -5 | -1.1 | -5 | | -4 |

reduce the direct influence of these day-night differences in outward height gradient. Nevertheless, the level by level signs and relative magnitudes of the 00Z vs. 12Z height gradient differences appear to agree well with observed diurnal convergence differences.

Because this diurnal modulation of the height patterns occurs through the depth of the troposphere, it manifests itself primarily in an alteration in the number and/or intensity of cumulonimbi and heavy rain showers. The more readily observable shallower cumulus and middle and lower level layered clouds show much less of a diurnal cycle. This is why the diurnal deep convection cycle has yet to be well established; i.e., it has been difficult to measure properly from long record oceanic surface reports and low resolution satellite pictures.

Regional Mass Compensation. Another important question to be considered is whether the upward mass transport in the cloud clusters is regionally compensated for by subsidence in the surrounding cloud-free areas or rather results from broader scale Continental-Oceanic mass adjustments. It has been implied previously (e.g. Nitta and Esbensen, 1974; Wallace and Hartranft, 1969) that continental-oceanic heating differences may be a primary cause for the observed oceanic diurnal cycle in convection. Thus, the day vs. night vertical motion patterns of the continents and the oceans ought to be about 12 hours out of phase with maximum daytime continental convection. This would imply that a large portion of the mass compensation for the oceanic vertical motion might be due to a diurnal macro-scale continent and ocean oscillation. In the West Pacific this was not found to be the case.

Figure 10 gives an idealized view of the typical arrangement of tropical cloud clusters ($\sim 4^{\circ}$ in diameter), clear regions, and the variable

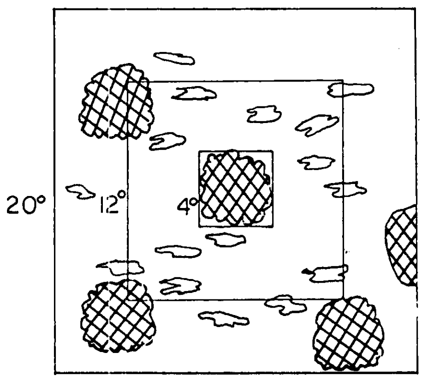



Figure 10. An idealized cloud cluster with its environment plan view.

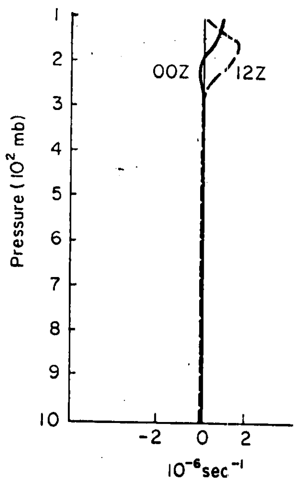



Figure 11. Net divergence in $20^{\circ}$ square.

cloud regions between them. Two rectangular areas around the cluster with sides of $12^{\circ}$ and $20^{\circ}$ of latitude are shown. The composited rawinsonde data from stations around each cluster and within each shell were combined and analyzed for over 1000 cloud clusters and 500 clear regions. Paper I discusses the type of rawinsonde compositing procedures which were used.

The $20^{\circ}$ square contains cloud clusters, clear regions, and variable cloud regions. By combining all rawinsonde data around the cloud clusters and clear regions the mean divergence pattern for the areas $20^{\circ}$ on a side can be found. This is shown in Figure 11. It indicates that mass circulations within a $20^{\circ}$ square are almost entirely locally compensated for at each time period. No detectable 00Z vs. 12Z upward or downward net mass circulation is observed. Mass compensation for the

00Z and 12Z differences in cluster divergence does not thus come from great horizontal distances, but is, to a large extent, regionally compensated for. Macro-scale continental-oceanic influences do not appear to be the primary cause of these observed 00Z vs. 12Z vertical motion differences.

Kung (1975) has also indicated that no significant differences exist between 00Z and 12Z mean vertical motions averaged over a $15^{o}$ by $15^{o}$ area in the Marshall Island region from like composites of rawinsonde data for 130 days. This occurred despite his large calculated 3 to 1 excess of 00Z over 12Z kinetic energy generation and export within the individual large scale disturbances. The mean upward and downward circulations of the Hadley, Walker and continental-oceanic circulations are quite weak and steady in comparison with the diurnally varying up-and-down cloud-cloud free local circulations. A large diurnal variation of the local circulation does not mean there is a similar percentage diurnal variation in the regionally averaged circulations.

Hourly Changes in Clear Region Sinking Motion. Figure 12 shows the 850-500 mb clear region IR cooling profile as calculated by Cox (1973) for summer conditions at the island of Guam. The average daily solar warming as calculated by Dopplick is also shown. By adding the solar and IR curves together, a daytime average net tropospheric cooling profile is obtained. This is significantly less than the nighttime cooling profile when the solar warming is not included. If Reynolds et al. (1975) are correct in their findings that Dopplick's solar heating rates for the tropics may be too small, this difference between day and night cooling would be even larger. However, for the present study these curves are assumed to give the approximate values for tropical clear regions and for the typical regions surrounding the cloud clusters.

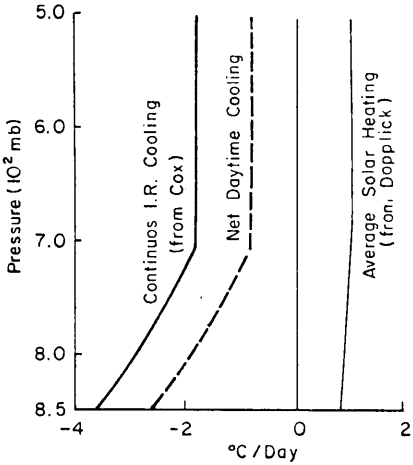



Figure 12. Radiative cooling (850-500 mb) at Guam during the summer.

The hourly solar heating rate was calculated from tropical temperature, moisture, and ozone values which were inserted into a numerical model supplied by S. Cox. Solar heating was calculated at 50 mb intervals for every hour from sunrise (assumed to be 06 L.T.) to local noon. This heating rate was assumed to be symmetrical about local noon with sunset occurring at 18 L.T. After averaging these solar heating rates between 850 and 500 mb, they were plotted as shown in diagram a of Figure 13. The daily average of this solar curve is the same as that found from Dopplick's data. By adding the IR cooling and solar heating rates, the total hourly net radiational cooling curve is found as portrayed in this same diagram.

This hourly net cooling curve indicates how the atmospheric temperature would vary if radiation was the only influence involved. To find the actual temperature variations 3, 6, and 12 hour thickness changes between 850 and 500 mb for various stations in the West Pacific and West Indies were calculated. We also used the special 6-hour rawinsonde

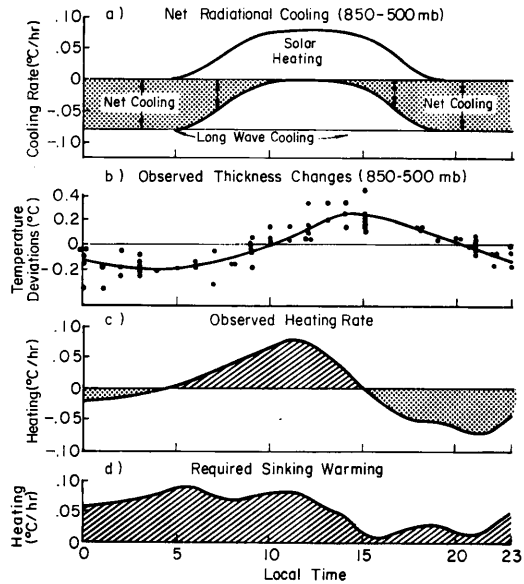



Figure 13. Calculated 850-500 mb typical hourly changes occurring in a clear region surrounding the usual cloud cluster of:

Diagram a) Solar and IR heating rates,

Diagram b) Hourly layer temperature deviations from the daily layer average as derived from observed thickness changes,

Diagram c) Observed layer heating rate, and

Diagram d) Required clear region sinking warming to simultaneously account for the observed layer temperature changes and the radiaticn induced temperature changes.

data taken during the nuclear testing operations in the Marshall Islands of May, June, and July 1956 and 1958, supplied to us by Prof. E. Kung. Extra thickness values could also be determined from a utilization of the change in rawinsonde release time in 1957 from 03Z and 15Z to 00Z and 12Z. 850 mb and 500 mb height data were chosen because they were nearly always available and because it was thought that any land heating influences exerted by the larger islands would not extend to these levels.

These pressure-thickness values were calculated from the height differences for each station for each time period. At least two months of data were taken from each station. The mean thickness for each station was found along with the average hourly deviations at each reporting time. These inter-daily thickness deviations were then coverted into temperatures and were plotted as hourly deviations from the daily mean of all stations. Diagram b of this figure depicts the smooth curve drawn through these thickness determined 850-500 mb hourly temperature changes. From the time rate of thickness change, the actual or observed diurnal heating rates within the 500-850 mb layer can be determined. These are shown in diagram c of this figure. Note that between 05-15 L.T. the layer is warming, while between 15-05 it is cooling.

By subtracting the net radiational cooling rate from the actual or observed heating rate, the required hourly clear region compressional sinking warming can be determined. This is portrayed in diagram d. Notice the larger sinking warming required during the morning hours in contrast with the smaller amounts of sinking required in the late afternoon and early evening. Although these calculations probably overestimate the real diurnal variation in clear region sinking, they do indicate that significant morning vs. evening differences in clear region or partly cloudy region sinking motion are to be expected. These diurnal sinking patterns agree qualitatively with observed heavy and very heavy rain patterns.

Following our findings on the local region mass compensation as discussed in the previous section, we believe this enhanced morning sinking motion in the clear regions around the cloud cluster areas relieves itself in extra convergence into the cloud areas and in enhanced deep cumulus convection.

Comparison of Diurnal vs. Daily Average Radiational Forcing. If a diurnally varying short wave radiational profile is superimposed on the Albrecht and Cox daily average IR profile with the proper corrections for their lower cloud heights and larger horizontal scale, a diurnally varying cloud cluster divergence profile would likely result.

Figure 14 gives a level by level comparison of our inferred cloud and cloud free radiational forcing with the IR daily average radiational forcing as determined by Albrecht and Cox. These can be interpreted as the cloud region or perturbed area radiation forcing. Curve 1 gives daily average values of cluster minus surrounding clear region rate of radiational induced temperature change as derived from Figure 3. Here the day vs. night net radiational cooling in the cirrus tops largely cancels itself. Curve 2 portrays Albrecht and Cox's (1975) trough (i.e. perturbed) minus ridge (non-perturbed) daily average IR radiational differences. As previously discussed, the tops of their assumed disturbance cloudiness are 100-200 mb lower than our assumed cluster cirrus. Their IR-only radiational induced perturbed minus unperturbed temperature changes are very similar to our daily average of cluster minus surrounding region radiational differences. Curve 3 portrays the vertical distribution of this paper's 00Z minus 12Z difference in cluster minus surrounding region net radiational forcing. This latter vertical profile is not so different from the other radiational profiles when proper adjustments are made. Note that the diurnal variations in radiational forcing are very similar to the daily mean values at levels below 300 mb but substantially larger from 150-250 mb.

Figure 15 compares the observed 00Z vs. 12Z cluster divergence differences (curve 3), with the daily average measured cloud cluster

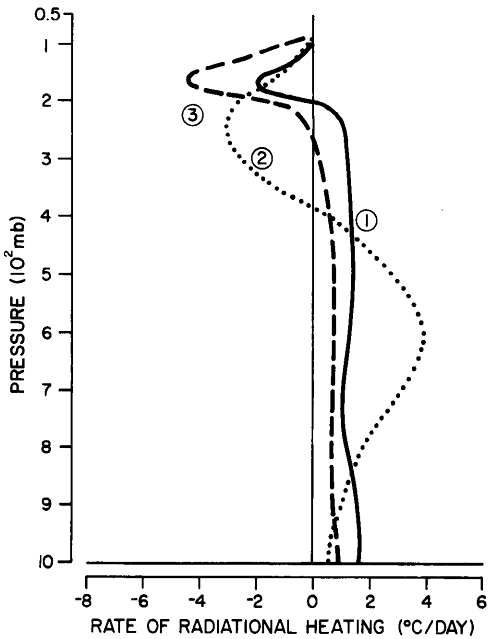



Figure 14. Cloud cluster minus surrounding region rate of radiation induced temperature change of:

Curve 1) Daily average of cluster minus surrounding region values from Figure 7.

Curve 2) Albrecht and Cox's (1975) trough (perturbed) minus ridge (unperturbed) IR only daily average net radiation, and

Curve 3) This paper's 00Z minus 12Z differences in assumed net radiation induced temperature change.

divergence from Paper I (curve 1) and the inferred daily average IR only radiation induced divergence profile of Albrecht and Cox (1975) multiplied by two (curve 2) because of their 2-3 times larger distance scale between trough and ridge as discussed on page 55. Note that the daily mean curves and the 00Z minus 12Z values are of similar magnitude. Remember that the Albrecht and Cox (1975) data is based on Line Island disturbances whose cloud tops are substantially lower than our summertime

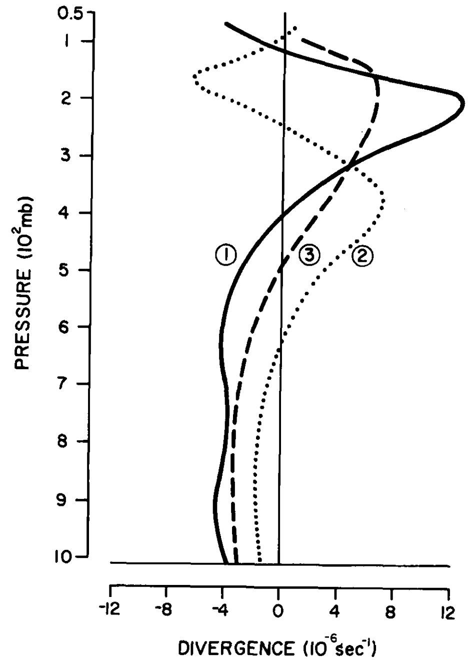



Figure 15. Vertical cloud cluster divergence profile resulting from cluster minus surrounding region radiation induced energy differences:

Curve 1) The daily average observed cluster vertical divergence profile from Figure 4 of Paper I,

Curve 2) Albrecht's and Cox's (1975) trough (perturbed) minus ridge (unperturbed) IR only daily average radiation induced divergence differences multiplied by two, and

Curve 3) Observed 00Z minus 12Z divergence differences from Paper I.

West Pacific cluster tops. Note the similarity of these three profiles and how the radiation induced divergence profile of Albrecht and Cox fits rather well our observed daily average and 00Z minus 12Z cluster divergence.

<u>If the daily average IR radiational differences between cloud and cloud-free areas can require vertical divergence profiles which are</u>

<u>similar to the observed vertical divergence profiles, then it is likely that radiation is the primary determinant of the observed profile. Also, if these radiational influences are fundamental, then it is likely that two to one differences in night vs. day cluster minus surrounding radiation is also fundamental to the large diurnal modulation of the cluster.</u>

## 5. TIME LAG OF CONVECTIVE RESPONSE TO RADIATIONAL FORCING

The rainfall data and the cloud cluster divergence profiles discussed in Paper I indicate a considerable lag in atmospheric response to the hypothesized day-night radiational forcing. We believe this to be a natural consequence of the hourly accumulation of the cloud minus surrounding region radiational energy differences. Thus, if the response of the troposphere to the radiationally specified thickness is only partial, as the author believes, then a near sunset/sunrise accumulated cluster minus surrounding radiational energy excess/deficit should occur. Thus, the maximum accumulated nighttime radiational cooling should occur around sunrise, and maximum accumulated solar warming near sunset.

Upper Tropospheric Response. Assuming that the 150-250 mb levels of the cluster gain radiational energy relative to their surroundings during the day and lose radiational energy relative to their surroundings at night, we can estimate hourly changes in the upper level relative thickness gradients and outward pressure accelerations. As indicated in diagram a at the top of Figure 16, it is likely that the daytime solar energy gains are larger than the IR energy losses. An equilibrium between solar and IR energy should occur a few hours after sunrise and a few hours before sunset, or at about 08 L.T. and 16 L.T. Between 08-16 L.T. solar warming is larger than IR cooling, at other times IR cooling is larger. Diagram b portrays the hourly rates of radiational warming of the cluster cirrus shield and of the region surrounding the cluster. Diagram c gives the estimated cloud cluster minus surrounding region differences in rate of radiational energy gain. If no other processes besides radiation were operating to bring about sensible temperature change, this would require that the maximum and minimum upper

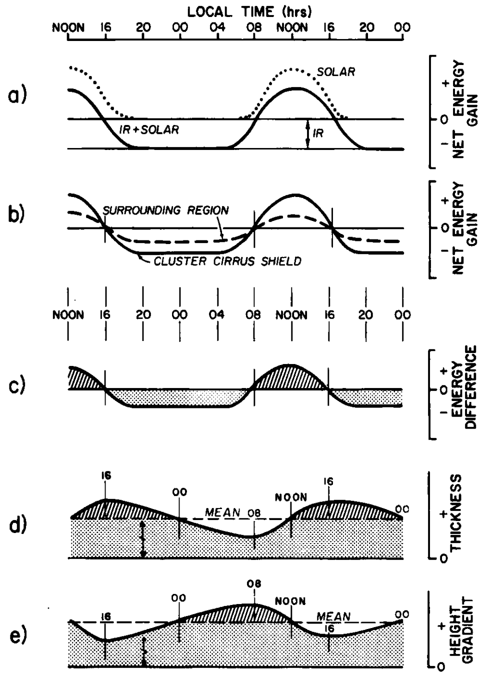



Figure 16. Hypothesized typical hourly changes occurring in the cloud cluster cirrus shield and in the regions surrounding the cluster at these levels:

Diagram a) The cluster cirrus shield rates of IR and solar radiational energy gain and loss,

Diagram b) Comparison of the hourly cirrus shield rate of radiational energy gain-loss with that of the surrounding cluster region,

Diagram c) Cloud cluster minus surrounding region rate of radiational energy gain-loss,

Diagram d) Cirrus shield thickness changes that would result from radiational warming, and

Diagram e) Resulting outward height gradient of pressure surfaces from the cirrus shield to its surroundings.

level radiation induced temperature changes (and pressure thickness changes) occur around 16 L.T. and 08 L.T. as seen in diagram d of this figure. The relative increase of the cirrus pressure thickness between noon and midnight causes a relative downward bulge of the 150-250 mb cluster cirrus height fields compared with those of the surrounding heights and a decrease in the outward directed pressure-height gradient (diagram e). The 150-250 mb cluster outflow would thus be expected to be reduced between noon and midnight.

IR cooling accumulates within the cluster cirrus shield relative to the surroundings after 16 L.T. This causes relative upward bulging of the 150-250 mb cluster pressure heights compared with the surroundings. Minimum cluster outward pressure gradients should occur around $\sim$ 16-18 L.T. The outward pressure gradient and divergence from the cluster are consequently greater between midnight and noon.

It should be noted that these estimated maximum and minimum outward height gradients are without sharp peaks. The multi-station rainfall data of Paper I does not indicate any particular hourly peaks in Cb convection for all the stations. West Pacific data indicates a later morning peak than did Atlantic data or U.S. winter observations.

Middle and Lower Tropospheric Response. The hourly changes of conditions at levels below the cirrus shield within the cloud cluster and in the surroundings are shown in Figure 17. At these levels the surrounding region radiational cooling is always larger than the cluster cooling (diagram a). The night cluster minus surrounding region cooling is greater than the day cooling by a ratio of about two to one (diagram b). The surrounding region resulting thickness changes (diagram c) and the consequent hourly changes in the inward directed height gradient

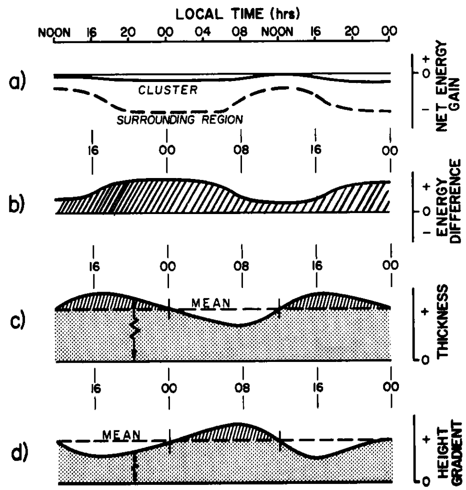



Figure 17. Hypothesized typical hourly changes occurring in the middle and lower troposphere of the cloud cluster and its surroundings:

Diagram a) Hourly variation in rate of radiational cooling within the cluster and within its surrounding region,

Diagram b) cluster minus surrounding region rate of radiation energy gain,

Diagram c) surrounding region hourly thickness changes, and

Diagram d) resulting hourly variation of the typical inward height gradient of the pressure surfaces from the surrounding area to the cluster.

(diagram d) are also shown in this figure. The inward directed height gradients to the cluster from the surrounding areas are larger than the daily average between midnight and noon.

If the wind alterations and consequent deep convection were to lag the cloud cluster's changes of inward and outward height gradients by a few hours, then the day vs. night radiation induced cloud region divergence profile would be shifted further backward in time. The high frequency of 06-12 L.T. deep convection when solar heating influences are becoming large, is the result of the accumulated previous night's IR cooling difference between the cloud and cloud-free areas. Similarly, the early evening minimum in deep cumulus activity, even though occurring after sunset, is a result of the accumulated solar heating differences between the cloud and the cloud-free areas during the previous 3-15 hours. The troposphere's responses to day vs. night radiational forcing are thus seen to follow with a considerable lag in hours. At some locations the time lag in the enhanced morning convection may occasionally carry over into the early afternoon. At other locations, the response of the atmosphere may be very much more rapid and the peak of deep cumulus convection may occur considerably earlier. The sizes, shapes, and relative positions of the cloud areas and the surrounding cloud-free regions; the nature of the flow patterns in which the clouds are embedded; the season and latitude; the thermodynamic characteristics of the atmosphere; etc. all must act to influence the character of the wind response to the cloud and cloud-free radiational forcing functions. In some situations there are likely to be earlier responses, in other situations a later response. In most situations, however, a considerable lag in response is to be expected.

# 6. PREVIOUS EXPLANATIONS

A morning maximum of oceanic shower activity has been known or suspected to exist for a number of decades. Wallace (1975) has summarized the many previous explanations which have been advanced to explain diurnal variations in convection over middle latitude land regions. These are also applicable to the tropics. He groups these into two basic categories.

> "1) those based on thermodynamic processes that affect the static stability, and
>
> 2) those based on thermodynamic processes that influence the mass convergence within the planetary boundary layer."

To these can be added a third phenomenon:

> 3) those processes based on the semi-diurnal pressure wave of $S_2$ oscillation.

The first of these explanations has been a very popular one over the years. This explanation rests primarily with the assumption of enhanced IR cooling off the tops of clouds at night and a consequent increase of the static lapse rate. A number of authors have touched on this explanation, particularly Kraus (1963) and Lavoie (1963). Kraus' (1963) physical explanation might be summarized as follows:

> The tops of the clouds cool more at night than the cloud bases. The resulting instability produces more vertical overturning. The process continues throughout the night. This results in the raising of the quasi-permanent inversion height which allows deeper convection and more rainfall. During the day solar radiation warms the cloud tops and stabilizes them. It also reduces the liquid water content of the clouds. This acts to reduce precipitation.

Lavoie also discusses the effect of extra nighttime radiational cooling of the cloud tops. He, however, relates this to maximum growth of individual elements in the cloud layer and does not mention the inversion. Lavoie also speculates on the possible influence of sensible heat transport from sea to air. This is a cumulative process, he proposes, and adds heat to the lowest layers of the troposphere during the night, thus destabilizing these layers and allowing a small increase in cloudiness in the early morning hours. In his discussion of the air-sea sensible heat transfer, Lavoie states:

> ". . . during the night upward transfer of sensible heat is increasing from the lower boundary with the result that clouds are more easily formed and somewhat more numerous near sunrise. This latter effect probably has little direct influence on the precipitation potential of each cloud, however."

Although these explanations may be generally satisfactory for the small observed diurnal cycle on shallow convection, there are serious problems to the acceptance of these arguments as satisfactory explanations for the large observed diurnal cycle in deep convection. This is because

1) Significant differences between day and night oceanic lapse rates are not observed in the West Pacific and West Indies soundings we have analyzed.

2) Conditional instability is nearly always present in the Tropics. Deep cumulus convection is typically negatively correlated with small positive changes of lapse rate instability (Atkinson, 1971). Increase of the low level instability typically leads to an enhancement of shallow cumulus, not Cb's. Also, the troposphere may actually become more stable as a result of the deep cumulus layer overturning. Given the presence of a conditionally unstable lapse rate, Cb convection is not significantly related to small buoyancy differences.

The second of the proposed basic processes for this diurnal convection cycle has to do with day and night radiation differences within the planetary boundary layer over land. Wallace (op. cit.) lists these as

"(i) the familiar land and seabreeze circulation in coastal areas.

(ii) a uniform diurnal heating cycle in regions of sloping terrain. (Holton, 1968; Lettau, 1967).

(iii) changes in frictional drag associated with the diurnal variation in static stability within the planetary boundary layer (Blackadar, 1957)."

It is difficult to relate very many of the morning deep-convection occurrences over land and the adjacent ocean regions to the land breeze circulations although some coastal areas, particularly mountainous coastlines might be so influenced.

The low level jet phenomenon of summer over a large portion of the Great Plain and Midwest can establish nocturnal thunderstorms due to a combination of processes (ii) and (iii). It is difficult to invoke these latter two processes to explain winter season morning convection, however, and, of course, neither of these mechanisms can be used to explain oceanic morning convection maximums. Wallace (op. cit.) has recently advanced another hypothesis for the winter morning deep convection maximums over land. Quoting again from his paper:

> "The prevalence of a nocturnal maximum in wintertime convective activity over much of the central and eastern U.S. requires a somewhat different explanation, but we believe it also may be interpretible in terms of dynamical processes. Most wintertime convection is associated with the passage of developing cyclonic storms. In the warm sector of these disturbances nighttime inversions are common, and therefore, in a statistical sense at least, the winds near the top of the planetary boundary layer should be stronger during the the late night hours, in line with (iii) above. The increased wind speeds within the warm sector should contribute to the overall amount of lifting above the warm frontal surface where wintertime convection often occurs. In this respect it is interesting to note that Goldie (1936) found evidence of a nocturnal maximum in warm frontal precipitation over Scotland during winter at a time when cold frontal precipitation exhibited a maximum frequency of occurrence during the afternoon. Further evidence of the enhancement of warm frontal precipitation during the nighttime hours was given by Dexter (1944)."

Although this physical hypothesis may be applicable in some winter storm situations, the land nocturnal convection maximum appears too pervasive over the global land areas for this to be a generally satisfactory argument, especially in light of the observed diurnal variations over the oceans. Wallace acknowledges this:

> "A crucial test of this hypothesis will come when satellite derived statistics on the diurnal cycle in convective activity over the middle latitude oceans becomes available. If the above hypothesis is correct, then we should expect to find little or no diurnal oscillation in wintertime convective activity over the sea."

The evidence of Paper I indicates that a significant diurnal variation of deep cumulus convection does exist over the oceans.

The third most discussed physical argument for the diurnal cycle in oceanic convection is the semi-diurnal pressure wave or $S_2$ oscillation. This has been appealing to some meteorologists because of the often observed early morning and late afternoon maximum of convection (see Fig. 18-21 of Paper I for large island stations). This argument has been proposed perhaps most seriously by Brier and Simpson (1969). Their physical explanation goes something like this:

> During each day there occur two successive patterns of divergence and convergence caused by the sun's heating on one side of the globe and lack of heating on the other. This solar tide produces maximum pressure peaks near 10 L.T. and 22 L.T. with minimums near 04 L.T. and 16 L.T. They hypothesize that around 07 L.T. and 19 L.T. there occur periods of maximum pressure rise and convergence. This, they believe, acts to enhance convection. Near 13 L.T. and 01 L.T. there are general pressure falls and hypothesized divergence which should inhibit convection. They believe that the increased large scale convergence around 07 L.T. and 19 L.T. funnels extra mass upwards at selected cumulus locations. This significantly enhances the cumulus cloud activity.

There are basic difficulties to the acceptance of this explanation:

1) The observations indicate that when afternoon surface warming or a heat island effect is not present there are only one maximum and one minimum of deep cumulus convection, not two as predicted by solar tidal theory. In particular, a rainfall minimum occurs near 19 L.T. when a maximum should be found according to their theory. Also, the observations often indicate an increase of Cb activity beginning at 01 L.T. when their theory specifies a minimum.

2) The magnitude of the divergence caused by the $S_2$ tides is very small, of the order $10^{-7}sec^{-1}$. This is substantially less than the 3-5 x $10^{-6}sec^{-1}$ 00Z vs. 12Z divergence differences observed with our $4^{o}$ sized cloud cluster compositing schemes (see Figure 4 of Paper I).

It must be further emphasized that:

1) Double peaks in convection are not observed on small islands and at many land stations.

2) Because of their absence over the oceans, the late afternoon and evening peaks of convection over land must be primarily a direct solar heating influence and not a tidal effect.

Thus, the semi-diurnal solar tide theory does not appear to offer a satisfactory explanation for the observed morning-evening deep convective differences. Likewise, the other previous explanations for this diurnal cycle in deep convection do not appear to be generally valid. Some other physical mechanism must be invoked.

## 7. IMPLICATION OF THESE RESULTS FOR INTENSIFYING WEATHER SYSTEMS

One can view the diurnal variations in cluster-surrounding region deep cumulus mass recycling as being superimposed upon the daily average cluster recycling. The morning recycling maximum is thus about 30% larger than the daily average, and the evening minimum about 30% less. The morning maximum to evening minimum deep cumulus activity is thus $1 \pm 30\%$ or 1.3/0.7. This is almost a two to one diurnal range.

The cloud areas of growing and developed tropical cyclones have cirrus shields similar to those of non-developing cloud clusters but they exhibit a smaller percentage diurnal range of deep convection. This is to be expected. The rainfall of the inner $2^{\circ}$ radius of the average typhoon is about 4 times greater than the rainfall intensity of the typical cloud cluster (Frank, 1976). The typhoon cloud cluster and surrounding mean mass recycling should thus be about 4 times greater than that of the average cloud cluster or $4 \pm$ (30% of 1), giving a 4.3/3.7 diurnal range. This results in a tropical cyclone diurnal convective cycle amplitude of only about 15%. Our observational evidence indicates an approximately similar 15% variation in the typhoon's 00Z vs. 12Z divergence and rainfall patterns. It is thus seen that the magnitude of this diurnal deep convection cycle is similar for most cluster systems but is far more detectable in the weaker ones.

Tropical Cyclone Genesis. For the few cloud clusters which grow into tropical storms, there appears to be a gradual reduction in the ratio of radiation to boundary layer frictional forcing. It is only when the intensifying cloud cluster winds reach velocities of 20-25 m/sec that the boundary layer convergence becomes equal to that of the radiational convergence and the so-called CISK process takes over as the primary cluster convergence mechanism. For the developed typhoon, of course, the CISK process is quite dominant, and the radiational differences between the storm and its surroundings play only a secondary role. Forthcoming CSU reports by Zehr (1976) and Frank (1976) will discuss in more detail the relative magnitudes of the frictional convergence to radiational forcing mechanisms for typhoon genesis and structure.

Figure 18 gives an idealized view of the relative magnitudes of these two processes. At first, conditionally unstable lapse-rates and boundary layer frictional convergence (dashed curve) are necessary to produce cloud layers such that the radiation processes can be started. Once deep cloud layers are formed, the cloud-cloud free radiational differences (solid curve) force deep layer convergence and are the main mechanism for the cloud system maintenance and early intensification. If conditions are ripe for further cloud region intensification into a tropical cyclone, the frictional convergence mechanism gradually becomes more important. The stronger the cyclone becomes, the larger the relative importance of frictional convergence to radiational convergence because the cluster layered clouds, unlike the frictional convergence mechanism, do not significantly change their radiational characteristics as the cluster intensifies.

The author has previously (1975) commented on the fact that the 12-13 numerical models of hurricane intensification which have been developed in the last decade, all have assumed initial vorticity fields which are typically an order of magnitude larger than the usual vorticity fields of cloud clusters from which hurricanes develop. Despite these assumptions of unrealistically high initial vorticity, cyclone intensification (for those using the most realistic cumulus condensation heating schemes) often required 3-5 days, which, considering their initial asssumptions, is too slow. If the results of this research are correct, this difficulty in obtaining sufficiently rapid early cyclone growth and the need to assume a moderately intense cyclone to start might be overcome by proper incorporation of cloud and cloud-free radiational influences. It is likely that future hurricane genesis models which take these cloud and cloud-free radiational influences into account will obtain more rapid and probably more realistic cyclone growth rates. Previously, Lopez (1968) has also speculated on the importance of radiation in tropical cyclone formation and intensification:

> "This general failure of the cumulus convection to provide the necessary initial tropospheric heating for early disturbance intensification indicates that there must be another source of heating. For instance, a significant contribution of early

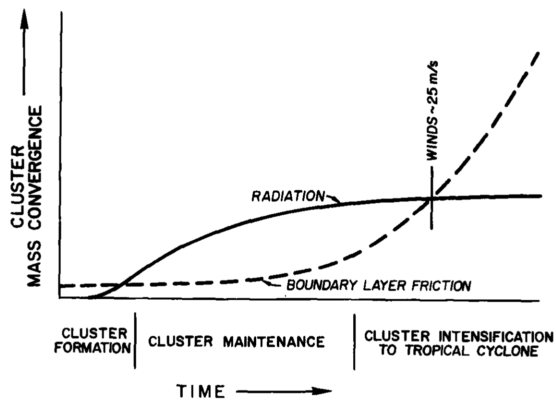



Figure 18. Comparison of the relative magnitudes of radiation induced convergence to boundary layer friction induced convergence of the typical tropical cloud cluster during its usual formation, maintenance, and intensification stages.

disturbance development may be accomplished by the trapping of ultraviolet and infrared radiation by the cloud canopies developed above the incipient disturbance. If this does, in fact, prove to be a realistic contribution, then the primary role of cumulus convection in early disturbance growth will be the indirect one of generating this radiation shield, rather than the direct one of providing the condensation."

# 8. SUMMARY

The large observed radiational differences between cloud and cloud-free regions are probably a significant source of available potential energy for tropical weather systems. These differences are also likely to act as a significant convergence mechanism for middle-latitude cyclone systems. These arguments appear especially relevant when one considers that condensation energy does not manifest itself in direct sensible temperature alteration. If radiation is a fundamental mechanism of tropical weather systems, then one should expect significant day-night differences in weather system mass convergence and deep cumulus convection. These are observed. The recent GATE observations should be carefully studied for additional vertification of the likely fundamental role of cloud and cloud-free radiational differences.

It must be emphasized that we have discussed the diurnal variation of deep cumulus convection and not that of layer or shallow cumulus convection. There is little diurnal variation in these latter types of cloudiness. This diurnal variation of deep cumulus convection may not be readily observable by satellite pictures which cannot resolve the individual Cb elements or by surface cloudiness observations where the percent sky coverage is primarily determined by layer type clouds. It is well known that cluster cirrus and middle level layer cloudiness can persist for many hours in the absence of deep cumulus convection. Thus, reports indicating only very small diurnal variations in total cloudiness as observed by the satellite should not be interpreted as meaning that a significant diurnal variations in deep convection does not exist.

The author would like to encourage numerical modelers to start including cloud and cloud-free and day-night radiation differences in their models. We are currently developing a primitive equation numerical model to test this radiation hypothesis on cloud cluster genesis. The large efforts currently going on by numerical models in attempted simulations of cumulus-broader scale interaction and boundary layer processes may prove to be of no more or perhaps even of less importance than properly accounting for the cloud-environment radiational processes. The author has only reluctantly come to this conclusion in the last few years. Previously, he believed that boundary layer processes and cumulus condensation were dominant to the maintenance of the individual tropical weather system.

## ACKNOWLEDGEMENTS

The author wishes to acknowledge the many discussions and substantial assistance he has received from Stephen Cox on the nature of cloud and clear region radiation processes. Robert W. Jacobson prepared a number of the figures of this report and participated in many discussions. The author would also like to acknowledge helpful discussion on this subject with Messrs. William Frank, William Fingerhut, Raymond Zehr and Gary Foltz of Colorado State University, and also with Ron Holle, of NOAA, Miami. The author is grateful to Prof. Ernest Kung of the University of Missouri for supplying the author with a tape copy of his special Marshall Island atomic test rawinsonde data.

Charles Solomon, Barbara Brumit, Edwin Buzzell, and Genevra Metcalf assisted in the manuscript preparation. This research was supported by a grant from the NOAA National Environmental Satellite Service.

BIBLIOGRAPHY (cont'd)

Ruprecht, E. and W. M. Gray, 1974: Analysis of satellite-observed cloud clusters. Colo. State Univ. Atmos. Sci. Paper No. 219, 91 pp.

Wallace, J. M., 1975: Diurnal variations in precipitation and thunderstorm frequency over the conterminous United States. Mon. Wea. Rev., 103, 406-419.

Wallace, J. M. and F. R. Hartranft, 1969: Diurnal wind variations, surface to 30 km. Mon. Wea. Rev., 97, 446-419.

Williams, K. T. and W. M. Gray, 1973: Statistical analysis of satellite-observed trade wind cloud clusters in the western North Pacific. Tellus, 25, 313-336.

Yanai, M. et al., 1973: Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. J. Atmos. Sci., 30, 611-627.

Zehr, R., 1976: Factors favorable to typhoon genesis as revealed by 10-years of composited rawinsonde observations. (Forthcoming CSU report).

**BIBLIOGRAPHIC DATA SHEET**

**1.** Report No. CSU-ATSP- 243

**2.**

**3.** Recipient's Accession No.

**4.** Title and Subtitle

DIURNAL VARIATION OF OCEANIC DEEP CUMULUS CONVECTION
Paper I: Observational Evidence
Paper II: Physical Hypothesis

**5.** Report Date: February, 1976

**6.**

**7.** Author(s)

Robert W. Jacobson, Jr., and William M. Gray

**8.** Performing Organization Rept. No. CSU-ATSP-243

**9.** Performing Organization Name and Address

Atmospheric Science Department
Colorado State University
Fort Collins, Colorado 80523

**10.** Project/Task/Work Unit No.

**11.** Contract/Grant No.

NOAA 04-3-158-51

**12.** Sponsoring Organization Name and Address

NOAA
National Environmental Satellite Laboratory
Suitland, Maryland

**13.** Type of Report & Period Covered

Project Report.

**14.**

**15.** Supplementary Notes

**16.** Abstracts

This report consists of two papers on the diurnal variability of oceanic deep cumulus convection. The first paper presents observational evidence in support of the existence of a significant morning maximum and afternoon-evening minimum of oceanic deep convection. The second paper advances a physical hypothesis for this diurnal variation. This hypothesis is based on day vs. night differences in cloud region and surrounding cloud free region radiational cooling.

**17.** Key Words and Document Analysis. **17a.** Descriptors

Diurnal Rainfall Variations
Tropical Weather Systems Precipitation
Tropospheric Radiation

**17b.** Identifiers/Open-Ended Terms

**17c.** COSATI Field/Group

**18.** Availability Statement

**19.** Security Class (This Report) UNCLASSIFIED

**20.** Security Class (This Page) UNCLASSIFIED

**21.** No. of Pages 106

**22.** Price