Mass divergence in tropical weather systems
Paper I: Diurnal variation

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SUMMARY

The diurnal variation of mass divergence and vertical velocity is documented for tropical summertime oceanic weather systems in the Western Pacific, Western Atlantic and the GATE region. It is shown that it is large and has the same basic character in all regions.

Gray and Jacobson (1977) proposed that the diurnal variation of mass convergence is a result of diurnal differences in the net radiative and convective heating rates in the thick cirrus-shield covered weather systems and their surrounding clear areas. Fingerhut (1978) has developed a numerical model which appears to substantiate this hypothesis. A comparison of observations with results from his model reveals that a simple radiation-condensation model simulates most of the observed diurnal variations of convergence. The hypothesis is that radiational forcing is one of the major contributors to the maintenance and modulation of tropical weather systems.

1. INTRODUCTION

This is the first of two papers discussing the interaction between cumulus convection and the large-scale wind and mass fields. It documents the existence of a single diurnal cycle in the mass divergence of summertime oceanic tropical weather systems. On the basis of a comparison with the numerical modelling results of Fingerhut (1978), it is concluded that radiative-convective forcing is responsible for the observed behaviour. In Paper II some of the other physical processes controlling tropical convection are investigated, and an attempt is made to measure the relative magnitude of each large-scale control.

The existence of a diurnal cycle in the mass divergence of oceanic tropical weather systems was first demonstrated by Ruprecht and Gray (1976a, b), and was further discussed by Gray and Jacobson (1977) and Foltz and Gray (1979). Corroborating evidence on the diurnal variation of heavy precipitation has been presented by Dewart (1978), Grube (1979), McGarry and Reed (1978) and Frank (1979).

The above studies were based on less extensive rawinsonde data sources than the work reported here. They concentrated on the variation in precipitation; in the paper emphasis is placed on the variation of the mass divergence. Use is made of data from the 1974 GATE experiment and from the tropical data sets collected by the tropical storm research project of William M. Gray (Williams and Gray 1973; Ruprecht and Gray 1976a, b; Zehr 1976; Frank 1977a, b, c; Erickson 1977; Frank 1978, 1979; Dewart 1978; Grube 1979 and McBride 1979). A summary of some of this research is contained in the report of Gray (1979).

From these sources, eleven independent composite data sets from three different tropical oceanic regions are assembled. Each represents one type of tropical summertime convective weather system. The diurnal variation of each data set is investigated and intercomparisons are made to determine consistencies in their diurnal behaviour. Preliminary results from this research have been reported by Gray and McBride (1978).

2. COMPOSITE METHOD

Following the technique of Williams and Gray (1973), twice daily rawin soundings from the standard observational networks in the Northwest Pacific and Northwest Atlantic

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oceans have been composited relative to the central positions of tropical convective systems, both cloud clusters and tropical storms. The cloud clusters are positioned from daytime satellite photographs and nighttime positions are obtained by interpolation. This procedure does not yield any significant day v. night variation in the composite thermodynamic fields associated with the clusters. Also, as will be seen in Paper II, there is no observed day v. night variation in the vorticity fields. The observed day v. night differences in the divergence fields are thus believed to be reliable.

Data are composited at 19 levels in the vertical extending from sea-level to 100 mb. The horizontal grid is an annulus, 2 to 4° latitude distance from the centre of the convective system. The radial component of the wind is composited in eight octants of 45° azimuthal extent, and a line integral is used to calculate the horizontal divergence averaged within 3° latitude of the system centre. Mass is balanced in the vertical by the addition of a constant factor to the divergence at each level; but all results presented are insensitive to this correction.

The GARP Atlantic Tropical Experiment took place in the tropical Eastern Atlantic Ocean from June to September 1974. Rawinsonde data from ships stationed in the outer hexagon of the GATE network (the A/B array) have been used. They have been processed and validated by the NOAA Center for Experimental Development and Data Analysis (CEDDA). The A/B array is approximately 3-8° latitude radius, so the line integral evaluations of divergence and vertical motion can be compared directly with the Western Pacific and Western Atlantic results. In GATE, however, there is better time resolution, as observations on the most convectively active days of the experiment were taken at 3-hourly intervals.

3. The data sets

The composite data sets are:

(a) Western Pacific convective systems

1. **Cloud cluster**: Summertime cloud clusters from the years 1967 and 1968. Mean latitude is 10°N, longitude 149°E; 87 individual disturbances make up the composite, and 190 rawinsonde observations exist in the 2 to 4° annulus. The estimated maximum sustained wind ($V_{\text{max}}$) of the system is 7 m s$^{-1}$.

2. **Pre-typhoon cloud cluster**: Cloud clusters which later develop into typhoons. Data are from 1961–1970. Latitude is 9°N, longitude 153°E; 130 individual disturbances; 222 observations. $V_{\text{max}} \sim 12$ m s$^{-1}$.

3. **Tropical storms**: $P_{c} > 1000$ mb. This and the following two data sets consist of a stratification of the official best tracks of the Joint Typhoon Warning Center, Guam (JTWOC) for the years 1961–1970, according to the central pressure ($P_{c}$) of each storm. Latitude is 16°N, longitude 143°E; 100 individual disturbances; 122 observations. $V_{\text{max}} \sim 15$ m s$^{-1}$.

4. **Tropical storms**: 980 mb < $P_{c} \leq 1000$ mb. Latitude is 22°N, longitude 137°E; 200 disturbances; 309 observations. $V_{\text{max}} \sim 25$ m s$^{-1}$.

5. **Tropical storms**: $P_{c} \leq 980$ mb. Latitude is 23°N, longitude 136°E; 147 disturbances; 362 observations. $V_{\text{max}} \sim 45$ m s$^{-1}$.

(b) Western Atlantic convective systems

6. **Cloud cluster**: Summertime cloud clusters from the years 1968–1974. Latitude is 20°N, longitude 82°W; 46 individual disturbances; 255 observations. $V_{\text{max}} \sim 7$ m s$^{-1}$.

7. **Easterly wave**: N. Frank, Director of the National Hurricane Center, Miami, has tracked the movement of Atlantic easterly waves since 1968. Using his tracks in the Carib-
bean for the years 1968–1974 a composite is made relative to the centres. Only wave systems which have a significant amount of convection associated with them are composited. The centre of each system is defined from the longitude of the trough axis, and the central latitude of convective activity as determined from satellite images. The mean latitude of the composite is 16°N; longitude is 72°W; 66 individual disturbances with 265 rawinsonde observations make up the composite. $V_{\text{max}}$ is estimated at 7 m s$^{-1}$.

Composite data sets 6 and 7 have weak upward vertical velocity. In fact, they have subsidence throughout the troposphere at one of the two observation times. Ruprecht and Gray (1976a) have also composited Western Atlantic cloud clusters and found similar results. These systems exist in a region of mean environmental subsidence and negative low level relative vorticity. Weather systems are often referred to as being in a ‘coasting’ or ‘weakening’ stage as they move through this subsidence region. This matter will be discussed further in Paper II. To demonstrate that data sets 6 and 7 are centred on the convectively active part of the weather system, east–west vertical cross-sections of their meridional wind values are shown in Fig. 1. As can be seen, the trough in the meridional wind pattern is very close to the centre of the disturbance.

Official best track positions of the National Hurricane Center for the years 1961–1974 are stratified according to the estimated maximum sustained wind ($V_{\text{max}}$) to provide the following three data sets:

8. Pre-tropical storms: $V_{\text{max}} \leq 35$ kt. Mean latitude is 22°N, longitude is 77°W; 102 disturbances; 351 soundings. $V_{\text{max}} \sim 15$ m s$^{-1}$.

9. Tropical storms: $35$ kt < $V_{\text{max}}$ < 65 kt. Latitude is 22°N, longitude is 78°W; 101 disturbances; 189 soundings. $V_{\text{max}} \sim 25$ m s$^{-1}$.

10. Tropical storms: $V_{\text{max}} \geq 65$ kt. Latitude is 23°N, longitude is 79°W; 73 disturbances; 326 soundings. $V_{\text{max}} \sim 45$ m s$^{-1}$.

(c) GATE convective systems

11. GATE Composite Cloud Cluster: Soundings at each ship of the A/B array are averaged at each observation time for ten of the most convectively enhanced days of the experiment to provide a composite cloud cluster. The days for which the data are composited are Julian days 188, 189, 195, 222, 245, 248, 255, 256, 257 and 259.

![Figure 1](image-url) East–west cross-section of meridional wind for Western Atlantic cloud clusters and easterly waves.
4. Results

Vertical profiles of divergence for the five Pacific and five Atlantic data sets are shown in Figs. 2 and 4. The corresponding kinematically derived vertical velocities are shown in Figs. 3 and 5. Profiles are shown for the two standard observation times, 00 and 12 GMT (10 a.m. and 10 p.m., respectively, local time in the Pacific, 7 p.m. and 7 a.m. in the Atlantic). The general character of the profiles is convergence or inflow in the lower troposphere compensated for by divergence or outflow near the 200 mb level.

![Figure 2](image)

Figure 2. Mean divergence within the $r = 0-3^\circ$ latitude area for the five Western Pacific composite weather systems.

![Figure 3](image)

Figure 3. Mean vertical velocity within the $r = 0-3^\circ$ area for the five Western Pacific composite weather systems.
Four consistent features can be seen in the figures:

(a) The low level convergence in the layer from the surface to 850 mb is greater in the morning (7–10 a.m.) than in the evening (7–10 p.m.).
(b) There is convergence in the evening in the middle troposphere near 450 mb.
(c) In the morning the upper level outflow extends through a deeper layer of the atmosphere than it does in the evening.
(d) The mean tropospheric upward vertical motion is greater in the morning than in the evening.

Data sets 1 to 4 and 6 through 9 exhibit all four characteristics. The typhoon and

**WESTERN ATLANTIC DIVERGENCE**

![Diagram of divergence](image)

Figure 4. Mean divergence within the $r = 0–3^\circ$ area for the five Western Atlantic composite weather systems.

**WESTERN ATLANTIC VERTICAL VELOCITY**

![Diagram of vertical velocity](image)

Figure 5. Mean vertical velocity within the $r = 0–3^\circ$ area for the five Western Atlantic composite weather systems.
hurricane data sets (data sets 5 and 10) exhibit only properties (c) and (d).

This diurnal variation can be quantified in various ways as shown in Table 1. The first measure is the ratio of the maximum upward vertical velocity found in the troposphere in the morning (7–10 a.m.) to the maximum velocity in the evening (7–10 p.m.). The second measure is the ratio of the values of vertical velocity at 850 mb. The last column of the table shows the levels of maximum upward velocity at the two observation times. Also shown in Table 1 is the mean tangential component of the wind at 850 mb at 3° radius. This measure of the intensity of each system is included to highlight a general trend: as the intensity of the weather system's cyclonic wind field increases, the magnitude of the diurnal variation of divergence decreases.

The strong similarity between the diurnal divergence and vertical velocity profiles of the two western ocean regions provides convincing evidence that this variation exists. As divergence is a notoriously difficult atmospheric parameter to measure, the appearance of the four common features (a) to (d) in these independent data sets shows that they are consistent and realistic features. It also lends confidence to the compositing technique and hence aids in the interpretation of compositing results.

Vertical profiles of divergence and vertical velocity at the eight observation times for the GATE cloud cluster are shown in Figs. 6 and 7. The shape of the profiles is different from that in the western oceans. In the GATE region most of the convergence is below the 800 mb level whereas the western ocean systems have a similar amount of mass inflow spread through a much deeper layer. Despite this difference, the GATE diurnal variation is very similar to that in the other systems. (a) Low level convergence has a maximum at 7.30 a.m. (local time) and a minimum at 7.30 p.m. (b) There is convergence in the middle troposphere near 450 mb at both 7.30 and 10.30 p.m. (c) The outflow layer is deeper in the morning so that the level of maximum vertical velocity is lower in the atmosphere.

Concerning property (d) there is general agreement, vertical velocities being at their weakest between 7.30 p.m. and 1.30 a.m. The maximum upward vertical velocity is at 1.30 p.m. which is probably 3 to 6 hours later than in the western oceans. This difference is discussed further in section 6 and in Paper II.

The consideration of properties (a) to (d) has revealed a close similarity in the diurnal

![GATE CLUSTER DIVERGENCE](image)

Figure 6. Mean divergence within the GATE A/B array for the GATE composite cluster.
Table 1. Quantitative measures of the intensity of the diurnal variation for the Western Pacific and Western Atlantic composite data sets. A dash (—) in the third column signifies that the p.m. vertical velocity is downward.

<table>
<thead>
<tr>
<th>Data set</th>
<th>3° radius mean tangential wind at 850 mb (m s⁻¹)</th>
<th>Ratio of maximum upward vertical velocity a.m./p.m.</th>
<th>Ratio of 850 mb upward vertical velocity a.m./p.m.</th>
<th>Level of maximum upward vertical velocity (mb)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pacific weather systems</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Cloud cluster</td>
<td>1</td>
<td>1.9</td>
<td>9.9</td>
<td>600</td>
</tr>
<tr>
<td>2. Precipitation cluster</td>
<td>5</td>
<td>1.3</td>
<td>1.2</td>
<td>400</td>
</tr>
<tr>
<td>3. Storm P &lt; 1000 mb</td>
<td>8</td>
<td>1.9</td>
<td>2.3</td>
<td>600</td>
</tr>
<tr>
<td>4. Storm 980 &lt; P &lt; 1000 mb</td>
<td>11</td>
<td>1.3</td>
<td>1.2</td>
<td>700</td>
</tr>
<tr>
<td>5. Storm P &lt; 980 mb</td>
<td>17</td>
<td>1.3</td>
<td>0.9</td>
<td>350</td>
</tr>
<tr>
<td>Atlantic weather systems</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6. Cloud cluster</td>
<td>0.5</td>
<td>3.2</td>
<td>—</td>
<td>800</td>
</tr>
<tr>
<td>7. Easterly wave</td>
<td>1</td>
<td>2.7</td>
<td>—</td>
<td>700</td>
</tr>
<tr>
<td>8. Storm: ( V_{max} &lt; 35 )</td>
<td>6</td>
<td>2.0</td>
<td>—</td>
<td>400</td>
</tr>
<tr>
<td>9. Storm: 35 ( \leq V_{max} &lt; 65 )</td>
<td>10</td>
<td>1.3</td>
<td>1.9</td>
<td>300</td>
</tr>
<tr>
<td>10. Storm: ( V_{max} &gt; 65 )</td>
<td>12</td>
<td>1.4</td>
<td>1.1</td>
<td>400</td>
</tr>
</tbody>
</table>
variation of divergence in the three tropical oceanic regions. The implication is that the
same diurnal forcing mechanism is operating in each. This conclusion has also been reached
by Foltz and Gray (1979) for the entire tropics from considerations of the troposphere's
diurnal energy budget.

5. PHYSICAL HYPOTHESIS

Gray and Jacobson (1977) proposed that the vertical convergence profile observed in
tropical weather systems is maintained and diurnally modified by differences in the radiation-
condensation heating profiles between thick cirrus-shield covered weather systems and their
surrounding clear areas.

Specifically, the upper layered clouds of organized weather systems are largely opaque
to infra-red radiation. They absorb radiation from lower layers and prevent a net flux of
energy in the layers underneath the cloud tops. In addition, condensation and evaporation resulting from upward vertical motion slightly warm the upper troposphere
and cool the lower troposphere of the typical tropical weather system. By contrast, cloud-
free areas cool radiatively through IR energy loss at rates significantly greater than those at
the same level underneath the cloud shield. The solar absorption of energy is also affected
by the presence or absence of cloud shields. Solar energy acts to increase the temperature of the cloud-free areas throughout the troposphere, but in disturbance regions with thick
layered clouds it acts primarily to raise the temperature within the upper cloud decks.

The heat balance is thus quite different in the two regions. In the cloud free area
surrounding the cluster, the thermodynamic equation in Eulerian coordinates may be
written as:

\[
\frac{\partial T}{\partial t} + \nabla \cdot \mathbf{V} + \omega (\Gamma_d - \Gamma_a) = Q_R
\]

where \( \omega \) is the vertical \( p \)-velocity and \( \Gamma_d, \Gamma_a \) are the dry and actual lapse rates.

In the cloud cluster the heat balance is:
\[ \frac{\partial T}{\partial t} + V \cdot VT = Q_{\text{dis}} \]  \hspace{1cm} (2)

where

\[ Q_{\text{dis}} = Q_{\text{Convection}} + Q_R. \]  \hspace{1cm} (3)

Figure 8 portrays our estimate of typical day and night rates of combined radiative and convective temperature change within the tropical weather system. Also shown is the surrounding cloud-free day and night radiational cooling. This figure has been derived from observational studies of temperature change aided by personal discussions on radiation problems with S. Cox (Cox 1969a, b, 1971; Fleming and Cox 1974; Albrecht and Cox 1975; Cox and Griffith 1979). The tropical disturbance's surrounding clear or partly cloudy regions lose about twice as much energy radiatively at night as during the day, and this energy sink is balanced by subsidence warming. In the weather system the situation is more complicated. Besides radiation, condensation (c) and evaporation (e) are also present. In conventional notation the convective heating rate, \( Q_{\text{Convection}} \) is

\[ Q_{\text{Convection}} = \bar{\omega} (\Gamma_d - \Gamma_a) - \frac{\partial (\bar{\omega} T')}{\partial p} + (c - e). \]  \hspace{1cm} (4)

\( \bar{\omega} \) is the vertical \( p \)-velocity averaged over the scale at which measurements are taken; and \( \omega' \), \( T' \) are deviations of vertical velocity and temperature from the measurement scale average. Gray (1973) demonstrated that the vertical motion within an active convective disturbance consists of a very large magnitude subsynoptic or local up- and down-circulation, which is not resolved by mean or synoptic scale flow patterns. Thus, there is no synoptic scale adiabatic cooling \( \bar{\omega} (\Gamma_d - \Gamma_a) \). For this reason the local heat balance of the cluster has been written as in Eq. (2).

Observed temperature changes in tropical weather systems indicate that 24-h vertically integrated averages of \( Q_{\text{dis}} \) are about zero. \( Q_{\text{Convection}} \) closely balances \( Q_R \). In the surrounding clear regions, however, the radiational cooling (\( Q_R \)) is always negative but is closely balanced by adiabatic subsidence. This implies subsidence within the surrounding clear or partially cloudy regions which is about twice as large at night as during the day. These day \( \nu \), night surrounding region subsidence differences act to feed more mass into the disturbances in the morning than at night. This is believed to be the primary reason for the large

![Figure 8](image_url)
diurnal divergence differences.

It is proposed that the diurnally varying radiative-convective heating differences between disturbances and their surroundings cause changes in the inward–outward disturbance pressure gradients. Because geostrophic control is weak at tropical latitudes, these lead to significant ageostrophic flow, and large diurnal mass convergence differences.

It is observed that the temperature within disturbances varies very little as a function of the amount of cumulus convection. Convection causes slight warming of the upper troposphere and cooling of the lower troposphere. Day–night radiative variations cause larger temperature variations than do diurnal variations in condensation. This is particularly true in the upper troposphere where solar absorption causes upper tropospheric warming and enhanced night-time cooling relative to the surrounding region. This causes day v. night differences in upper tropospheric $Q_{\text{dis}}$ as indicated in Fig. 8 which are only very weakly a function of day–night differences in the disturbance convection. Thus, the disturbance minus surrounding region diabatic energy differences ($Q_{\text{dis}} - Q_{\text{a}}$) are dominated by radiation and have a two to one night v. day variation. This general assessment of the diabatic influences of convection and radiation has been documented by our project in recent papers by Gray and Jacobson (1977), Foltz and Gray (1979), Frank (1979), Dewart (1978) and Grube (1979).

Evidence for the large differences in net tropospheric radiative cooling between a tropical disturbance and its environment has been documented by direct radiometric measurements from aircraft (Griffith and Cox 1977) and by radiative transfer calculations (Cox and Griffith 1979). Loranger, Smith and Vonder Haar (1978) have obtained net radiative budgets for the GATE-B array under convectively enhanced and suppressed conditions by combining radiative fluxes at the top of the atmosphere (measured from SMS-1 and NOAA-2 satellites) with simultaneous surface radiative flux measurements. They found that the net tropospheric cooling rate averaged for a 24-h period can be more than 1 degC per day greater in the environment than within the disturbance itself.

There is also evidence for significant day versus night gradients in the radiational forcing. Figure 9 from Cox and Griffith (1979) shows the results of their radiative transfer calculations. In a north–south cross-section for an active convective day in GATE, the differences in radiative flux divergence between the disturbance and its surroundings are much greater at night (lower portion of figure) than during the day (upper portion of figure). Ackerman (1979) using GATE data found that at night mean radiation cooling differences between a cloud cluster and its surroundings (averaged from the surface to 100 mb) were 0.6 degC greater than during the 12 daytime hours.

The atmosphere surrounding an organized tropical disturbance adjusts to its large radiational cooling at night by extra subsidence, which increases low-level convergence into the adjacent cloud regions. During the day solar heating reduces the tropospheric heat loss, and clear region subsidence and cloud region low-level convergence are substantially reduced.

At upper levels radiational cooling of cirrus shields is greater at night and less during the day than surrounding cloud-free regions. This acts in a complementary fashion with conditions at lower levels to alter the cloud region and surrounding area pressure slopes and convergence profiles. This condition results in more convergence occurring in the morning and less in the afternoon-evening. The convergence cycle typically follows the radiational forcing with a time lag of 3–6 h.

Figure 10 shows the expected slope of pressure surfaces from the disturbance to its surroundings resulting from these radiational differences. Note that relative to nighttime values the daytime solar warming of the upper disturbance cloud layers produces an extra downward bulging of the middle tropospheric disturbance pressure surfaces. This causes an
enhancement of the daytime middle-level convergence. At lower levels the situation is reversed. Daytime solar warming of the region around the disturbance reduces the low-level pressure gradients and consequently the daytime inflow to the disturbance is less.

The disturbance divergence profiles of Figs. 2, 4 and 6 indicate a considerable lag in atmospheric response to the proposed diurnal variation of radiational forcing. We believe this to be because the response is to the integrated diabatic heating effects. Thus, maximum accumulated night-time radiational cooling effects should occur 1–2 h after sunrise and a maximum in solar warming effects near sunset. If the wind adjustment lags the disturbance’s changes of inward and outward height gradients by a few hours, the lag of the divergence profile would be greater. Maximum and minimum divergence would thus occur in the late morning and early evening, as observed.

Fingerhut (1978) has tested this hypothesis by inserting the above radiative-convective

![Figure 9](image_url)

Figure 9. A pressure v. latitude (at 23.5°W longitude) cross-sectional view of the GATE A/B scale array for the 0600–1800LST period of Julian day 248. The top portion of the figure depicts the 1000–1400LST total (SW plus LW) radiative divergence (W m⁻² per 100mb) and the bottom portion depicts the LW component only (night-time total). Also shown are the magnitude and direction of the horizontal radiative divergence gradient at two points (arrows point towards regions of greater divergence), from Cox and Griffith (1979).
forcing in as simple a way as possible into a large scale axisymmetric primitive equation model of a tropical cloud cluster and its environment. Fingerhut's model is diagnostic, the convective and radiative heating being specified as in Fig. 8. In the cirrus layer of the cluster energy is absorbed following a sine-wave time dependence during daylight hours (12-h mean absorption = 29 W m⁻²). An equal amount of energy is emitted at a constant rate during the night. In this way both the large solar absorption and the large long-wave emittance of the cloud shield are modelled. At tropospheric levels below the high cloud decks the net temperature change due to radiation and condensation is close to zero. In the cluster environment diurnally varying climatological net radiative cooling profiles are specified.

The model divergence and vertical motion profiles (Figs. 11 and 12) show general agreement with those derived from observations. The strongest low level convergence occurs at 8 a.m., maximum middle level convergence at 4 p.m. and 8 p.m., and the level of maximum upward vertical velocity is higher in the atmosphere in the evening. The model results do not show a large diurnal variation in maximum vertical velocity, however.

Figure 13 shows diurnal profiles of vertical velocity for each of the three oceanic regions and for the radiation model. The similarity among the four curves is obvious, and leads to the conclusion that horizontal differences in radiative and convective heating constitute the major forcing mechanism for the observed diurnal variation in the mass divergence of tropical weather systems. This interaction between radiative, dynamic and convective processes is of profound importance for tropical cumulus parametrization studies and could also prove to be significant in research on the early development of tropical cyclones.

The only observational feature not successfully simulated by the model is the diurnal variation in the total vertical mass exchange. This failure is probably a result of the specified constant convective heating. In the actual atmosphere there is a strong cumulus feedback, new cells being forced by the low level convergence associated with the downdrafts of pre-
existing cells (Lopez 1973; Purdom 1976).

6. Discussion

It has been shown that the diurnal behaviour of the divergence profiles of convective systems is very similar in the GATE region, the Western Pacific and in the Western Atlantic. The basic mean vertical velocity profiles on which this diurnal variation is superimposed are quite different in the three regions, however. The observed convergence profiles are the

![Numerical Model Divergence](image)

Figure 11. Mean divergence for a 3° latitude radius cloud cluster as numerically modelled by Fingerhut (1978).

![Numerical Model Vertical Velocity](image)

Figure 12. Mean vertical motion for the 3° radius model cloud cluster of Fingerhut (1978).
result of interactions between many different physical processes acting on various scales. Paper II will tabulate these processes and will present quantitative discussions of their relative roles in the maintenance of convection in each region.

In this paper, the diurnal variation of divergence and vertical motion in each region has been documented. Strong similarities have been demonstrated in the diurnal behaviour of all composite data sets. This study has been concerned only with the diurnal variation of the wind fields. The variation of rainfall can be quite different from that of vapour convergence if there is moisture storage and significant convective feedback.

It has been shown above that in the three regions under consideration the mass convergence below the 850 mb level is a maximum near 7 a.m. local time. In the western oceans the resulting convergence of moisture is, apparently, quickly converted to rain. The diurnal variation of heavy precipitation in Western Pacific weather systems has been documented by Gray and Jacobson (1977). They show that rainfall has a maximum near 7 a.m. and a minimum near 9–11 p.m. This is in agreement with expectation if the low-level convergence is driven by the diurnal variation of diabatic heat sources as discussed here. A preliminary analysis of Western Atlantic oceanic rainfall by the present authors leads to the same conclusion.

During GATE, there was also a maximum of moisture convergence near 7 a.m. local time but the rainfall maximum was in the early afternoon, as has been discussed by a number of authors. This time lag between low-level forcing and maximum convection has been observed before, particularly by Hudlow and Marks (1977), Frank (1978) and Gray and Jacobson (1977). It can be seen in Fig. 7 where the low-level vertical velocity has a maximum at 7.30 a.m., but the upper-level vertical velocity, corresponding to the deep convection, has its maximum value at 1.30 p.m.

In comparison with the western oceans, the GATE region is quite stable; in particular
at low levels the atmosphere is colder (by about 2 degC) and drier (by about 1 gram per kg) than over the western oceans. It is also a region of large low-level vertical wind shear, whereas the other regions, being preferred regions for tropical cyclone genesis, are characterized by quite weak low-level vertical shear.

As a consequence of this greater thermal stability and greater wind shear, rainfall takes some time to develop in GATE. There is an initial storage of the vapour in clouds. In agreement with the diurnal forcing mechanism, convection is initiated in the morning hours (Weickmann et al. 1977) but appears to take 4–8 hours to organize into the observed cloud lines and squall lines. The line and squall convection can overwhelm the large-scale forcing and cause rain after the forcing mechanism has subsided. The heaviest convection thus comes 2–6 hours later than in the western ocean regions where buoyant instability and low vertical shear permit a faster response to low-level mass convergence. It is also possible that the diurnal behaviour of convection in the GATE region is influenced by downwind effects from Africa and by radiative effects involving the Saharan dust.

7. Conclusions

It is important that the large single cycle diurnal variation of mass convergence into tropical weather systems be realized and better understood. The implications for the understanding of tropical convection need to be more fully appreciated. More research into the response of the troposphere to day v. night and cloud–cloud free radiational and convective heating differences is required.

The project report of McBride and Gray (1978) contains more documentation and discussion of this large diurnal cycle in tropical convection.

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