

Mass Divergence in Tropical Weather Systems

Paper I Diurnal Variation
Paper II Large Scale Controls on Convection

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ABSTRACT

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 this diurnal variation is very large and has the same basic character in all regions.

Gray and Jacobson (1977) proposed that this diurnal variation in mass convergence results from differences in the net radiative and convective heating profiles of 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 his model output with the observations reveals that a simple radiation-condensation model does simulate most of the observed features. It is hypothesized that radiational forcing is one of the major contributions to the maintenance and modulation of tropical weather systems.

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1. INTRODUCTION

This paper discusses the interaction between cumulus convection and the large scale measurable wind and mass fields. This interaction works in both directions between the two different scales and involves a variety of physical processes.

In this paper the existence of a single diurnal cycle in the mass divergence of summertime oceanic tropical weather systems is documented. The observed diurnal variation is compared with the numerical modelling results of Fingerhut (1978), and the conclusion is made that a simple radiative-convective forcing mechanism is responsible for the observed behavior.

In Paper II an investigation is made of some of the other physical processes controlling tropical convection, and an attempt is made to quantitatively measure the relative role each large scale control has on convection.

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

The above studies were based on less extensive rawinsonde data sources than the current study. These studies concentrated on the variation in precipitation. In the current study the emphasis is placed on the variation of the mass divergence. Use is made of data from the 1974 GATE experiment and from the many tropical data sets composited over the past decade by the tropical storm research

project of William M. Gray (Williams and Gray, 1973; Ruprecht and Gray, 1976a, b; Zehr, 1976; W. Frank, 1977a, b, c; S. Erickson, 1977; McBride, 1977; Frank, 1978a, b; Dewart, 1978 and Grube, 1978).

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

2. WESTERN PACIFIC CONVECTIVE SYSTEMS

Following the technique of Williams and Gray (1973), twice daily rawinsonde observations from the standard observational network in the Northwest Pacific Ocean have been composited relative to the central positions of tropical convective systems. 00Z and 12Z rawinsonde data were used from the summer months of the years 1961-1970. The rawinsonde data network is shown in Fig. 1. The compositing technique has been used in many studies since that of Williams and Gray. A thorough discussion of the technique, its advantages, and limitations has been given by W. Frank (1977a).

Five different types of Western Pacific convective systems have been composited. They are numbered one to five:

1. CLOUD CLUSTER. Summertime cloud clusters were composited for the years 1967 and 1968. Only clusters which did not later develop into tropical cyclones were used. This data set is the "STAGE 00" data set of Zehr (1976).
2. PRETYPHOON CLOUD CLUSTER. This data set consists of cloud clusters which later develop into typhoons. Data for this and the following three data sets are from the full ten year period 1961-1970. This data set is "STAGE 2" of Zehr. It consists of that portion of each storm's track prior to one day before the first reconnaissance aircraft observation. Positions were obtained from satellite pictures and by extrapolation from Joint Typhoon Warning Center, Guam (JTWC) best tracks.
3. TROPICAL STORMS. $P_c > 1000$ mb. This and the following two data sets consist of a stratification of the official best

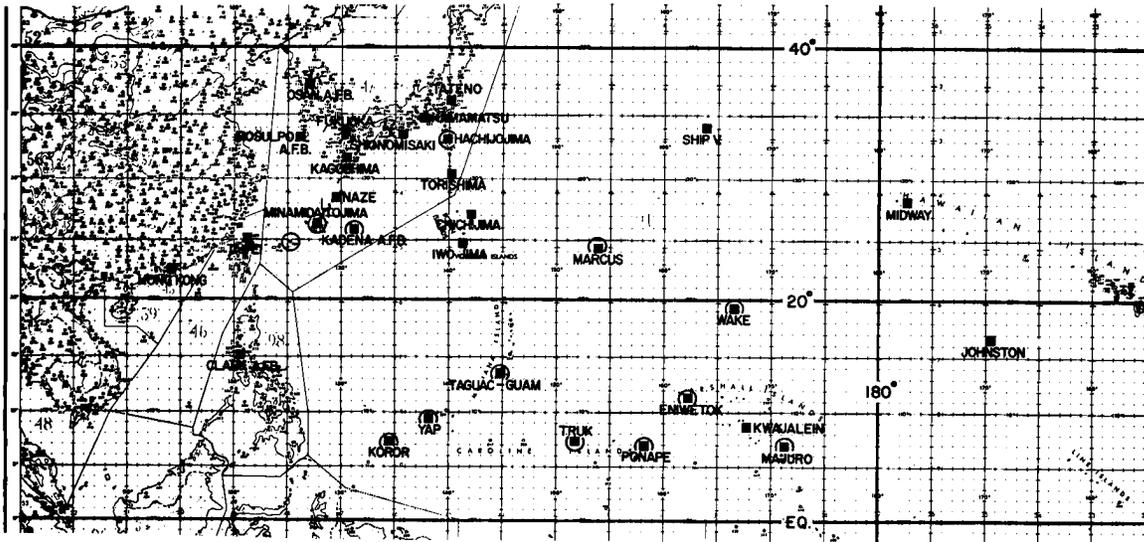


Fig. 1. Northwest Pacific rawinsonde data network.

tracks of the JTWC according to the central Pressure (P_c) of each storm.

4. TROPICAL STORMS: $980 \text{ mb} < P_c \leq 1000 \text{ mb}$.
5. TROPICAL STORMS: $P_c \leq 980 \text{ mb}$.

Positioning for the cloud cluster data sets (data sets 1, 2 and 6, 7 of the following section) was obtained from visual daytime satellite imagery. Interpolation had to be used, therefore, to obtain nighttime positions for these systems. This procedure did not yield any significant day vs. night variation in the composited thermodynamic fields associated with the clusters. Also, as will be seen in Paper II, there is no observed day vs. night variation in the vorticity fields. The observed day vs. night differences in the divergence fields are thus believed to be quite reliable.

Table 1 lists certain properties of the five data sets. In the first two columns are the mean latitude and longitude of the composite weather systems. Column (3) shows the number of rawinsonde observations

TABLE 1

Characteristics of Western Pacific Data Sets

DATA SET	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
	Latitude (Deg N)	Longitude (Deg E)	No. of Sound- ings 2-4 ⁰	Number of Individual Disturb- ances	Estimated Maximum Sustained Winds (m/s)	Mean Tangential Wind at 850 mb 2-4 ⁰ (m/s)	Mass Balance Correc- tion 00Z (m/s)	Mass Balance Correc- tion 12Z (m/s)
1. CLOUD CLUSTER	10	149	190	87	7	1	-.80	.09
2. PRETYPHOON CLUSTER	9	153	222	130	12	5	-.34	-.44
3. STORM $P_c > 1000$ mb	16	143	122	100	15	8	.27	.58
4. STORM $980 < P_c \leq 1000$ mb	22	137	309	200	25	11	-.07	.06
5. STORM $P_c \leq 980$ mb	23	136	362	147	45	17	.04	.00

composited in the annulus between 2° and 4° latitude distance from the center of the system. Column (4) gives the number of individual disturbances making up each composite. Column (5) is a subjective estimate of the maximum sustained wind speed of the system. Column (6) is the composited mean tangential component of the wind at 850 mb in the $2-4^{\circ}$ annulus. This parameter serves as a gauge of the intensity of the system.

To insure that the vertically integrated net radial mass flux between the surface and 100 mb is zero, small mass balance corrections had to be added to the radial component of the wind at each level. These corrections at the two observation times are listed in the last two columns of Table 1. Every feature of the divergence profile described in this study is insensitive to the mass balance correction. The correction had to be made, however, so that vertical velocities would simultaneously be zero at the surface and 100 mb.

Data were composited at 18 vertical levels extending from sea level to 100 mb. The horizontal grid was an annulus, 2° to 4° latitude distance from the center of the convective system. The radial component of the wind was composited in eight octants of 45° azimuthal extent. These eight values of the radial wind, V_R , were considered to be true at 3° latitude distance from the system center. The average horizontal divergence ($\overline{\text{DIV}}$) inside the 3° radius area of each weather system was calculated by an application of the divergence theorem of vector algebra:

$$\begin{aligned}
 \overline{(\text{DIV})} &= \frac{1}{\text{AREA}} \oint V_R \cdot dl \\
 &= \frac{1}{2\pi R} \cdot \sum_{i=1}^8 V_{R_i} \frac{2\pi R}{8} \\
 &= \frac{\overline{2V_R}}{R} \quad .
 \end{aligned} \tag{1}$$

Vertical profiles of divergence averaged over the 0-3° area for the five Pacific data sets are shown in Fig. 2. Profiles are shown for the two standard observation times, 00 and 12 GMT (10 AM and 10 PM Local Time). The corresponding kinematically derived vertical velocities are shown in Fig. 3. The general character of these profiles is convergence or inflow in the lower troposphere compensated for by divergence or outflow near the 200 mb level. The features of interest for the present study are as follows:

- a) The low level convergence in the layer from the surface to 850 mb is greater at 10 AM than at 10 PM.
- b) There is convergence at 10 PM 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.

The corresponding properties of the vertical motion profile are:

- d) The mean tropospheric upward vertical motion is greater in the morning than in the evening.
- e) The level at which the maximum upward vertical velocity occurs is higher in the atmosphere in the evening than in the morning.
- f) The low level upward vertical velocity (at 850 mb) is greater in the morning than in the evening.

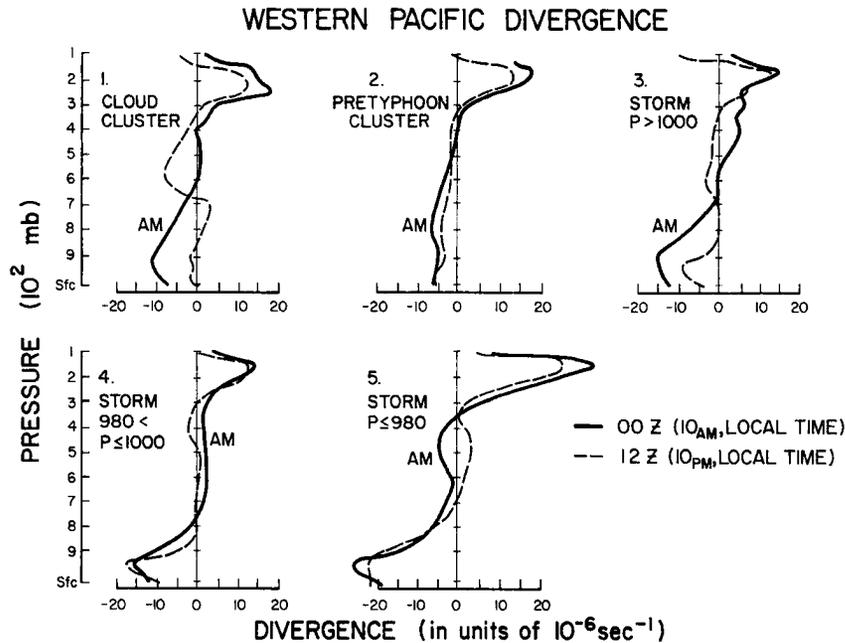


Fig. 2. Mean divergence within the $r = 0-3^{\circ}$ latitude area for the five Western Pacific composite weather systems.

Data sets 1 to 4 exhibit all six of these features. The typhoon data set (data set 5, $P_c \leq 980$ mb) exhibits only properties c) and d).

This diurnal variation can be quantified in various ways as shown in Table 2. The first measure shown is the ratio of the maximum upward vertical velocity found in the troposphere at about 10 AM (00Z) to the maximum velocity at about 10 PM (12Z). 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 2 is the mean tangential wind at 850 mb at 3° radius. This measure of the intensity of each system is included to highlight a general trend which appears in the table: as the intensity of the weather system's cyclonic wind field increases, the magnitude of the diurnal variation of divergence decreases.

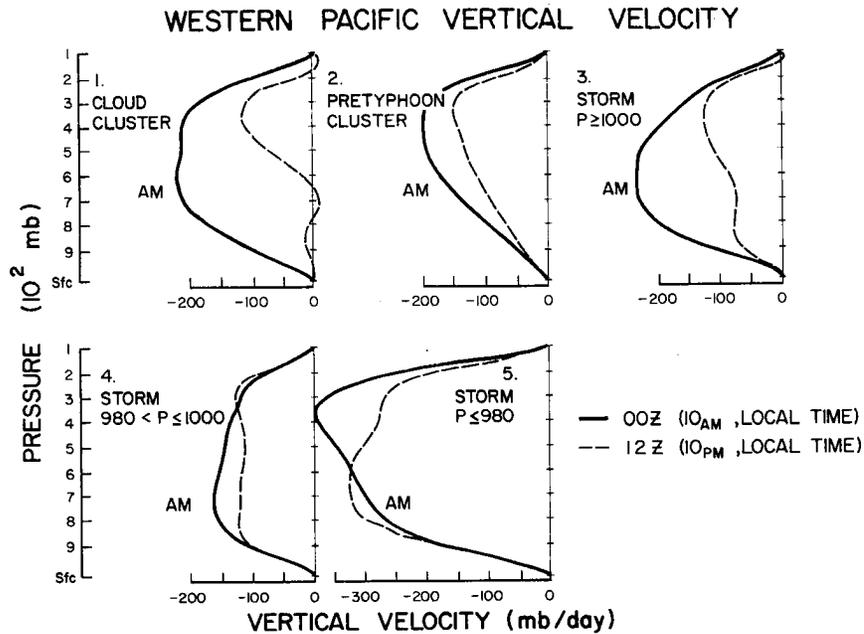


Fig. 3. Mean vertical velocity within the $r = 0-3^\circ$ area for the five Western Pacific composite weather systems.

TABLE 2

Quantitative measures of the intensity of the diurnal variation for the Western Pacific composite data sets.

DATA SET	Mean Tangential Wind at 850 mb (m/s)	Ratio of Maximum Upward Vertical Velocity	Ratio of 850 mb Vertical Velocity	Level of Maximum Upward Vertical Velocity (mb)	
		AM to PM	AM to PM	At AM	At PM
1. CLOUD CLUSTER	1	1.9	9.9	600	350
2. PRETYPHOON CLUSTER	5	1.3	1.2	400	300
3. STORM $P_c > 1000$ mb	8	1.9	2.3	600	350
4. STORM $980 < P_c \leq 1000$ mb	11	1.3	1.2	700	300
5. STORM $P_c \leq 980$ mb	17	1.3	0.9	350	600

3. WESTERN ATLANTIC CONVECTIVE SYSTEMS

Using the same technique as described above for the Pacific, five different types of summertime convective systems were composited in the Western Atlantic - Caribbean area. The data network is shown in Fig. 4. The five data sets are:

6. CLOUD CLUSTER. In collaboration with V. Dvorak of NOAA/NESS Applications Group, positions were obtained from satellite pictures of tropical weather systems which subjectively looked like they had potential for development into tropical storms. If a circulation center for the disturbance was visible, it was defined as the position of the system; otherwise the center of mass of the cloud area was used. Data were compiled from the years 1968-1974.
7. EASTERLY WAVE. N. Frank, Director of the National Hurricane Center, Miami (NHC), has tracked the movement of Atlantic easterly waves since 1968. Using N. Frank's tracks in the Caribbean for the years 1968-1974 a composite was made relative to the centers of these wave disturbances. Only wave systems which had a significant amount of convection associated with them were composited. The center of each system was defined such that the longitude was that of N. Frank's trough axis, and the latitude was the central latitude of convective activity as determined from satellite images.

Composite data sets 6 and 7 have very weak upward vertical velocity. In fact, they have subsidence throughout the troposphere at one of the two observation times. Ruprecht and Gray (1976a) also have composited Western Atlantic cloud clusters and found very little upward vertical

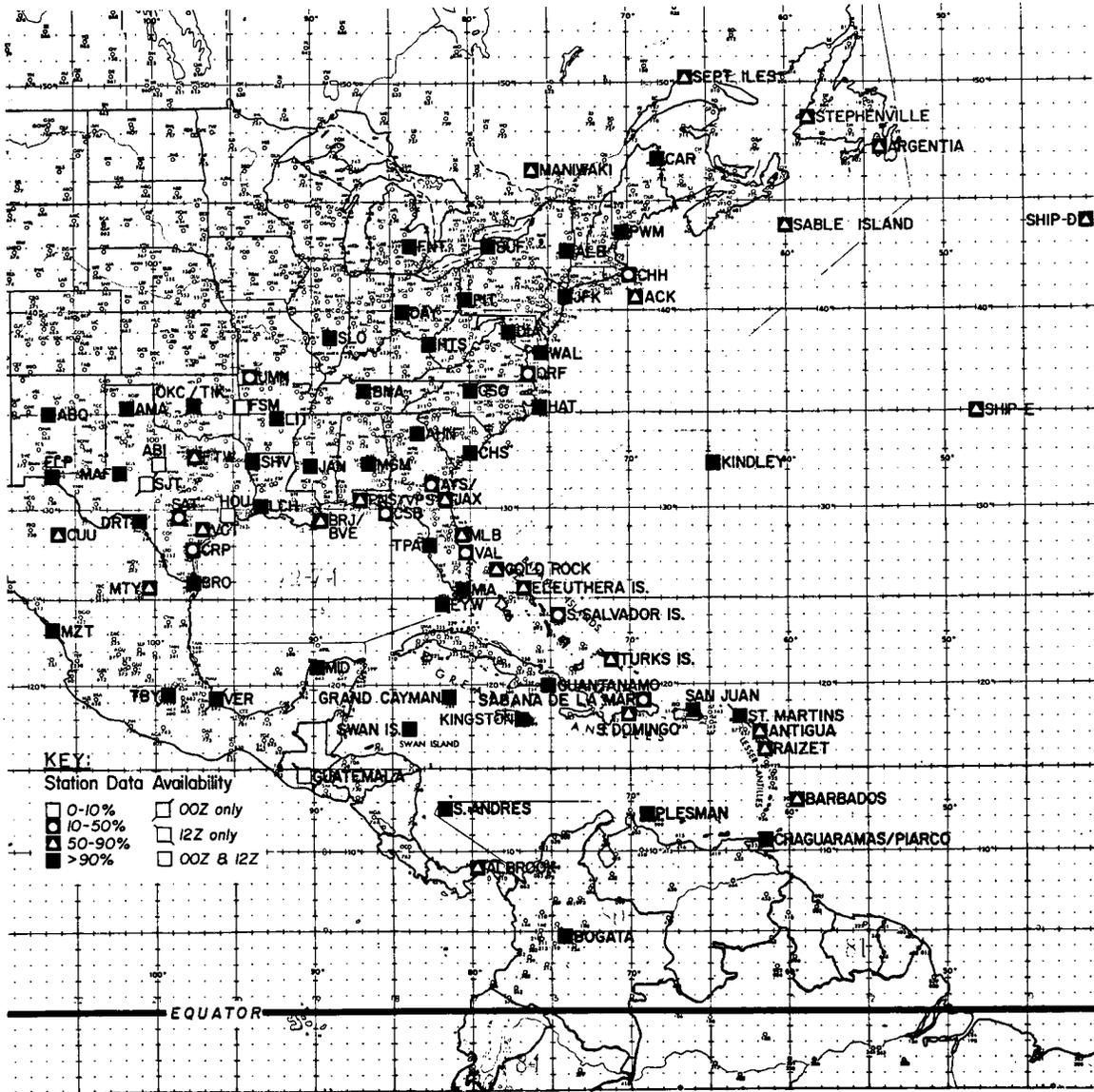


Fig. 4. Northwest Atlantic rawinsonde data network. Only island and coastal stations were used in this study.

motion. These systems actually 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, but to allay fears that data sets 6 and 7 are not centered on the convectively active part of the weather system, east-west vertical cross sections of the meridional wind for these data sets were constructed (Fig. 5). As can be seen, the trough in the meridional wind pattern is very close to the center of the disturbance.

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

8. PRE-TROPICAL STORMS: $V_{\max} \leq 35$ kts.
9. TROPICAL STORMS: $35 \text{ kts} < V_{\max} < 65$ kts.
10. TROPICAL STORMS: $V_{\max} \geq 65$ kts.

Various characteristics of the five Atlantic data sets are shown in Table 3. The observation times 00Z and 12Z for these data sets are within about one hour of 7 PM and 7 AM Local Time respectively. Diurnal divergence profiles and vertical velocity profiles are shown in Figs. 6 and 7.

There is a marked similarity between these figures and the corresponding figures (2 and 3) from the previous section. Data sets 6 through 9 can be seen to have every one of the properties a) to f) listed for the Pacific systems. The hurricane data set (data set 10) has properties a), d) and f).

The quantitative measures of the diurnal variation are shown in Table 4. As in the Western Pacific, the magnitude of the variation decreases as the intensity of the system increases.

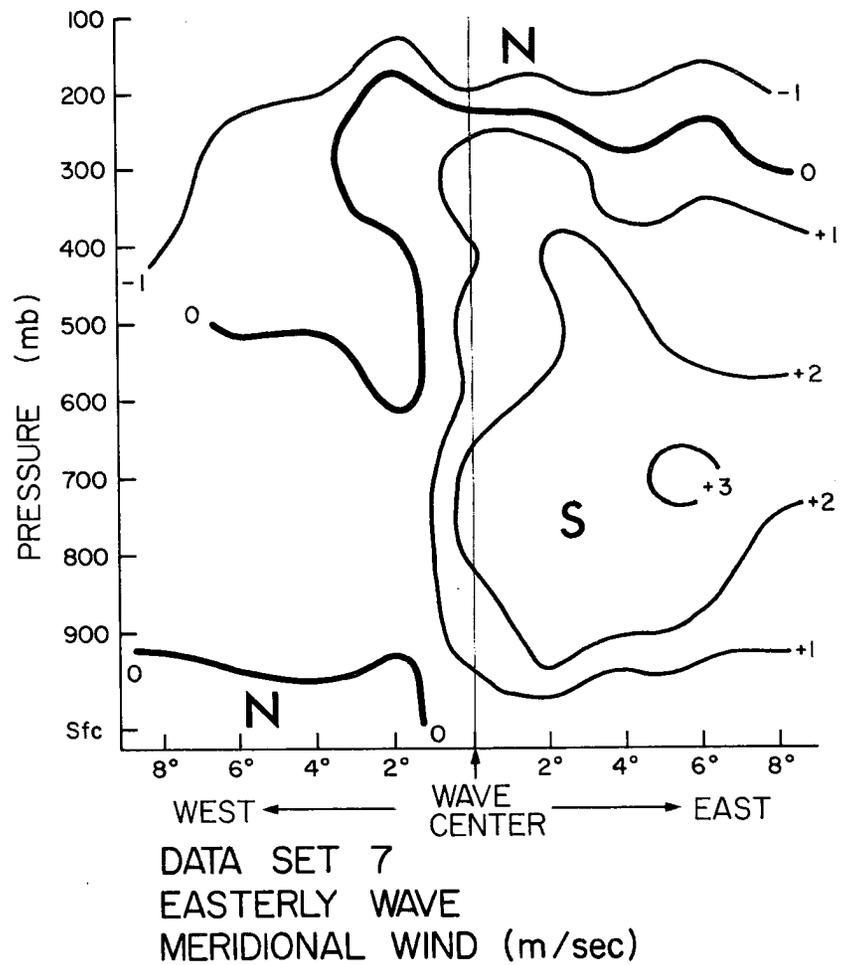
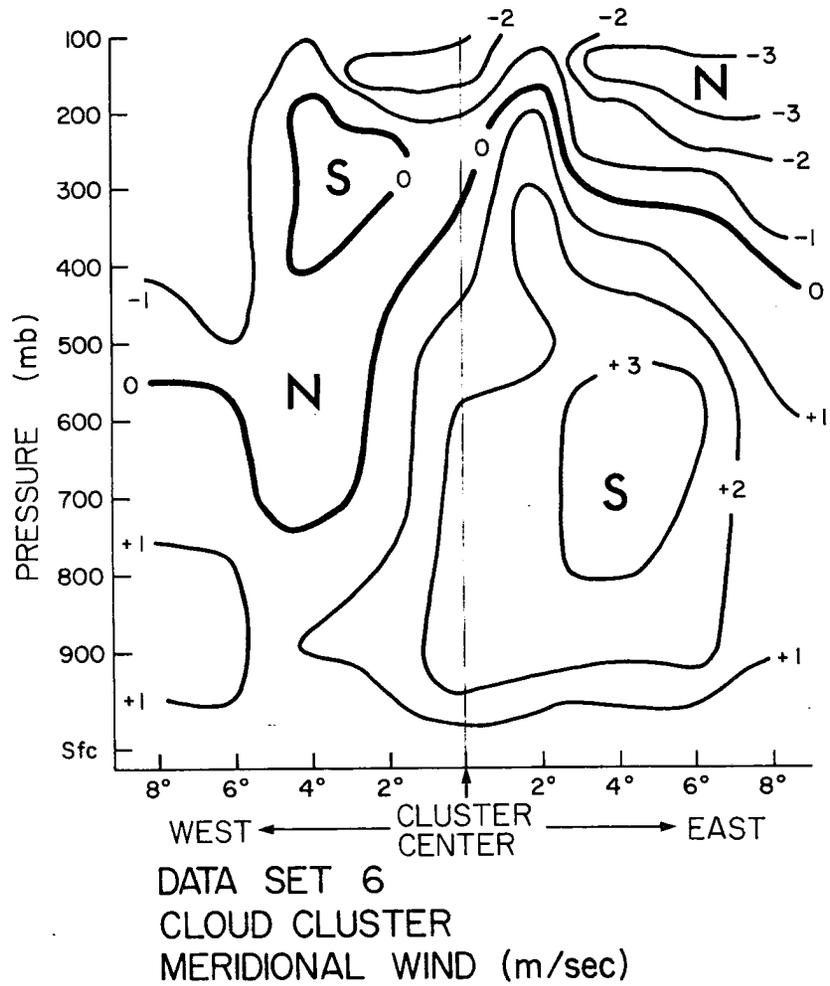


Fig. 5. West-east cross sections of meridional wind for Western Atlantic cloud clusters and easterly waves.

TABLE 3

Characteristics of Western Atlantic Data Sets

DATA SET	Latitude (Deg N)	Longitude (Deg W)	No. of Sound- ings 2-4 ^o	No. of Indivi- dual Distur- bances	Estimated Maximum Sustained Winds (m/s)	Mean Tangen- tial Wind at 850 mb 2-4 ^o (m/s)	Mass Balance Correc- tion 00Z (m/s)	Mass Balance Correc- tion 12Z (m/s)
6. CLOUD CLUSTER	20	82	255	46	7	0.5	-.11	.07
7. EASTERLY WAVE	16	72	265	66	7	1	.03	-.12
8. TROPICAL STORM $V_{\max} < 35$ kts	22	77	351	102	15	5	.27	.38
9. TROPICAL STORM $35 \text{ kts} < V_{\max} < 65$ kts	22	78	189	101	25	10	.10	-.10
10. TROPICAL STORM $V_{\max} > 65$ kts	23	79	326	73	45	12	-.14	-.15

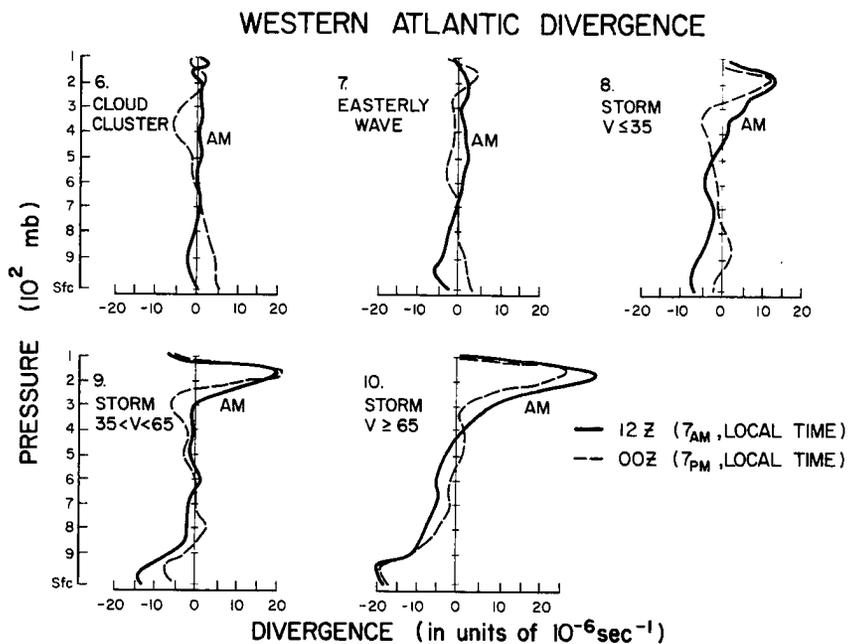


Fig. 6. Mean divergence within the $r = 0-3^{\circ}$ area for the Western Atlantic composite weather systems.

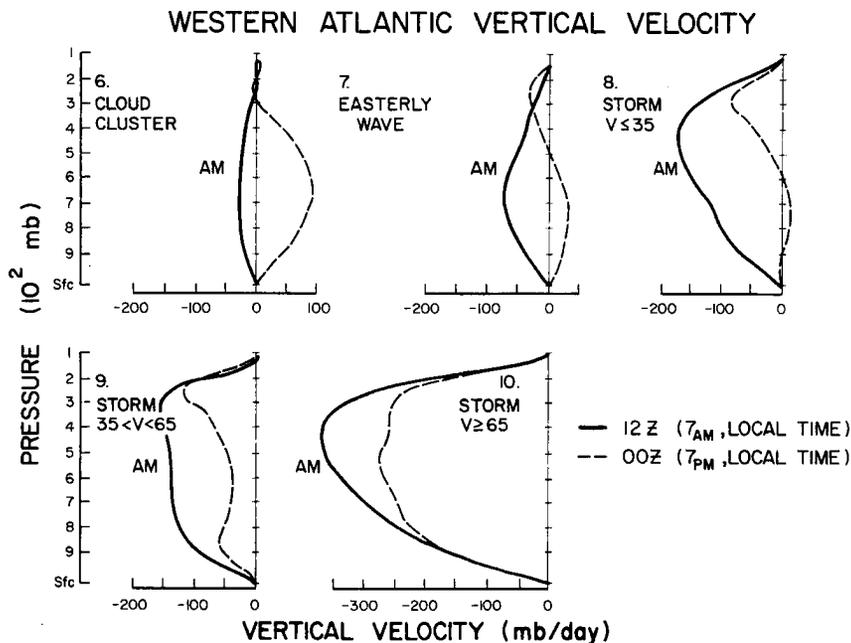


Fig. 7. Mean vertical velocity within the $r = 0-3^{\circ}$ area for the Western Atlantic composite weather systems.

TABLE 4

Quantitative measures of the intensity of the diurnal variation for the Western Atlantic composite data sets.
 ∞ in the third column implies that the PM vertical velocity is downward.

DATA SET	Mean Tangential Wind at 850 mb (m/s)	Ratio of	Ratio of	Level of Maximum Upward	
		Maximum Upward Vertical Velocity	850 mb Upward Vertical Velocity	Vertical Velocity (mb)	
		<u>AM to PM</u>	<u>AM to PM</u>	<u>At AM</u>	<u>At PM</u>
6. CLOUD CLUSTER	0.5	3.2	∞	800	250
7. EASTERLY WAVE	1	2.7	∞	700	250
8. STORM: $V_{\max} \leq 35$ kts	6	2.0	∞	400	250
9. STORM: $35 \text{ kts} < V_{\max} < 65$ kts	10	1.3	1.9	300	250
10. STORM: $V_{\max} \geq 65$ kts	12	1.4	1.1	400	500

The strong similarity between the diurnal divergence and vertical velocity profiles of the two western ocean regions provides convincing evidence that this variation actually does exist. This is not a trivial point, as divergence is a notoriously difficult atmospheric parameter to measure. The appearance of the six common features a) to f) 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.

4. GATE CONVECTIVE SYSTEMS

The GARP Atlantic Tropical Experiment (GATE) was performed in the tropical eastern Atlantic Ocean in June through September 1974. Dewart (1978) has composited data for ten of the most convectively enhanced days of the experiment. The days composited and rainfall amounts as recorded by shipboard rain gauges are listed in Table 5. Rawinsonde data taken by ships stationed in the outer hexagon of the GATE network (the A/B-array) were used in the current study. The A/B-array for Phase I of the experiment is shown in Fig. 8. The rawinsonde data have been processed and validated by the NOAA Center for Experimental Development and Data Analysis (CEDDA).

The GATE array is approximately $3\frac{1}{2}^{\circ}$ latitude radius. The composited divergence and vertical motion can thus be directly compared with the results of the previous section of this study where data are averaged between $2-4^{\circ}$ radius. In GATE, however, observations on the most convectively active days were taken at 3-hourly intervals; so that much greater time resolution is available than in the western oceans.

Vertical profiles of divergence at the eight observation times of these convectively active days are shown in Fig. 9. The corresponding vertical motions are shown in Fig. 10. The character of the divergence profile is different to that in the western oceans. In GATE most of the convergence is below the 800 mb level, whereas the data sets of the western oceans (Figs. 2 and 6), although having a similar total amount of vertical mass exchange, have their inflow spread through a much more deep layer.

Despite this difference the diurnal variation in GATE follows a similar pattern to that in the other systems. This can be seen by

a consideration of the properties a) to f):

Western Oceans

(Figs. 2, 3, 6, 7)

a) The low level convergence in the layer from the surface to 850 mb is greater at AM than at PM.

b) There is convergence at 7 PM and 10 PM 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.

GATE Cluster

(Figs. 9, 10)

a) The GATE low level convergence has a maximum at 7:30 AM (Local Time) and a minimum at 7:30 PM - similar to western ocean profiles.

b) Convergence in the 400-500 mb layer is most apparent at 7:30 PM and 10:30 PM - similar to western ocean profiles.

c) A very deep outflow layer can be seen in Fig. 9 at 7:30 AM and 10:30 AM - similar to western ocean profiles.

d) Again there is general agreement. Vertical velocities are weakest between 7:30 PM and 1:30 AM. The maximum vertical velocities, however, are at 1:30 PM. For reasons discussed in section 6, this appears to be a few hours later than in the western oceans.

Western Oceans (cont'd)

e) The level at which the maximum upward vertical velocity occurs is higher in the atmosphere in the evening than in the morning.

f) The low level upward vertical velocity (at 850 mb) is greater in the morning than in the evening.

GATE Cluster (cont'd)

e) There is some agreement here by virtue of the fact that 7:30 PM is the only time at which the GATE vertical motion maximum is in the upper troposphere.

f) GATE follows the same behavior as the western oceans. The maximum value of 850 mb vertical velocity is at 7:30 AM, the minimum value is at 7:30 PM. The ratio of the morning to evening is 3:1.

The GATE convective systems exist in very different environmental conditions to those in the western oceans. The GATE area has much stronger low level vertical wind shear than the West Pacific and West Atlantic and greater static stability in the lower half of the troposphere (U.S. GATE Central Program Workshop, 1977). There is also the possibility in GATE of modulation of the divergence variation due to downwind effects from Africa and to radiative effects of the Saharan dust.

Despite these differences, the above consideration of properties a) to f) shows that GATE convection follows a very similar diurnal variation in divergence to the other regions. The similarity between the three regions of the world is so striking that it implies there is the same basic forcing mechanism present in every region.

TABLE 5

Rainfall amounts on ten of the most convectively active days of GATE.

Julian Day	Date	A/B + B + C Ship Rain (mm)	A/B, B, C, Ships With > 50 mm Rain/Day (mm)
188	July 7	342	Meteor (73) Oceanographer (147)
189	July 8	333	Oceanographer (61) Vanguard (58) Researcher (122)
195	July 14	305	Poryv (78) Gillis (52)
222	Aug 10	111	Priboy (58)
245	Sept 2	252	Planet (52) Krenkel (64)
248	Sept 5	192	Gillis (47)
255	Sept 12	278	Dallas (98) Fay (83)
256	Sept 13	450	Quadra (107) Okean (52) Vanguard (66) Dallas (71) Fay (85)
257	Sept 14	322	Meteor (68)
259	Sept 16	221	Researcher (65)

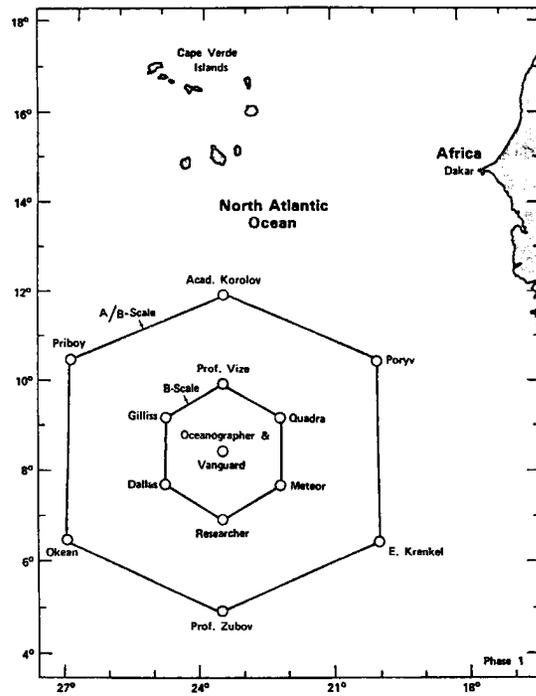


Fig. 8. A/B and B-scale ship arrays for Phase I of GATE.

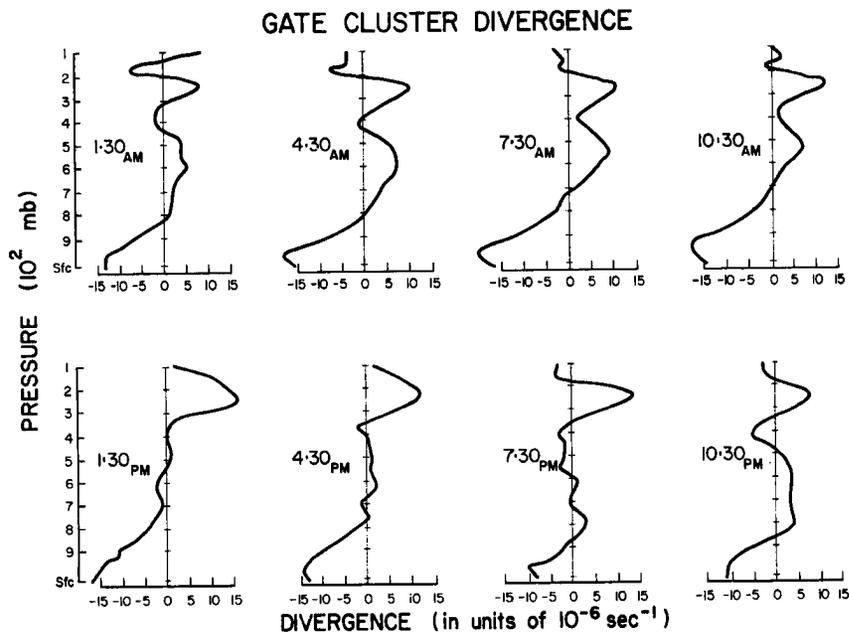


Fig. 9. Mean divergence within the GATE A/B-array for the GATE composite cluster.

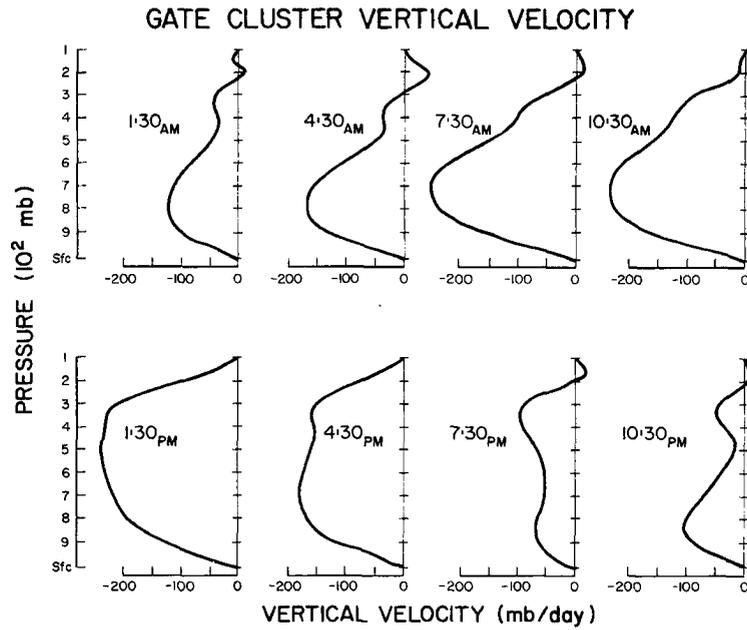


Fig. 10. Mean vertical motion within the GATE A/B-array for the GATE composite cluster.

For a more direct comparison of the data so far presented in sections 2, 3 and 4 of this study, Fig. 11 was constructed. This figure shows the GATE vertical motion profiles at the observation times closest to those in the other locations. The left hand side of the figure can be compared directly with Fig. 3, the right hand side with Fig. 7.

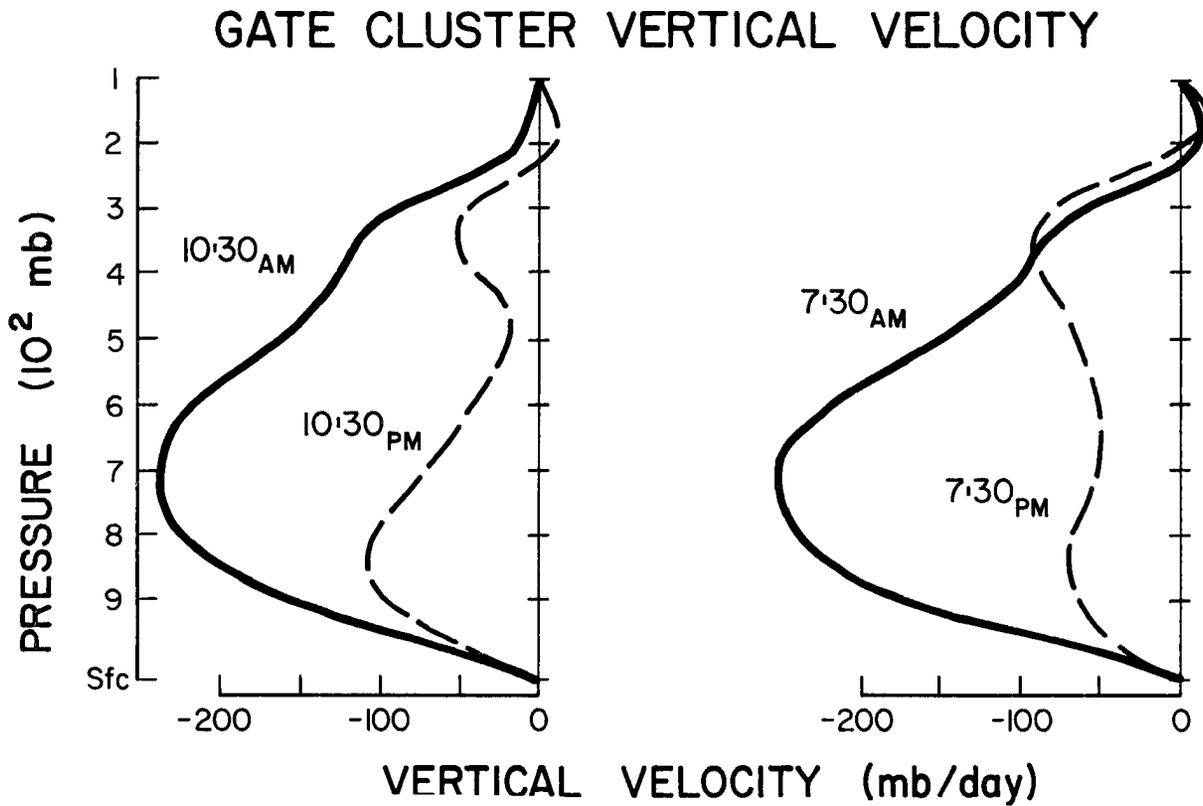


Fig. 11. Mean vertical motion for the GATE cluster at the observation times closest to those in the western oceans. The left side compares directly with Fig. 3, the right side with Fig. 7.

5. PHYSICAL HYPOTHESIS

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

Specifically, the upper layered clouds of organized weather systems are largely opaque to IR energy. They prevent upward IR energy losses from lower layers and prevent a net flux divergence of IR 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, the upper levels of the surrounding cloud-free regions are not able to inhibit IR energy losses from lower layers. Cloud-free areas radiatively cool through IR energy loss at rates significantly greater than that at the same level of the disturbance underneath the cloud shield. The solar absorption of energy is also greatly altered 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. At the same time, the surrounding clear or partly cloudy regions do not undergo significant temperature change from condensation and evaporation.

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

$$\frac{\partial T}{\partial t} + \underset{\sim}{V} \cdot \underset{\sim}{\nabla} T + \omega (\Gamma_a - \Gamma_d) = Q_R \quad (2)$$

Local Change of Tempera- ture	Horizontal Advection	Subsidence Warming	Radiative Cooling
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where ω is the vertical p-velocity and Γ_d , Γ_a are the dry and actual lapse rates.

In the cloud cluster the heat balance is defined as:

$$\frac{\partial T}{\partial t} + \underset{\sim}{V} \cdot \underset{\sim}{\nabla} T = Q_{dis} \quad (3)$$

where

$$Q_{dis} = Q_{Convection} + Q_R \quad (4)$$

Figure 12 portrays our estimate of typical day and night rates of combined radiation and convection temperature change within the tropical weather system. Also shown is the surrounding cloud-free area day and night radiational cooling. This figure was derived from empirical studies of observed temperature change and from discussions with S. Cox and from his group's radiation studies (Cox, 1969a, b, 1971a, b; Fleming and Cox, 1974; Albrecht and Cox, 1975; Cox and Griffith, 1978). The tropical disturbance's surrounding clear or partly cloudy regions radiatively lose about twice as much energy at night as during the day. This radiation (Q_R) is the only diabatic energy source of the surrounding region and is balanced by subsidence warming. In the weather system the situation is more complicated. Besides radiation, diabatic energy sources of condensation (c) and evaporation (e) are also

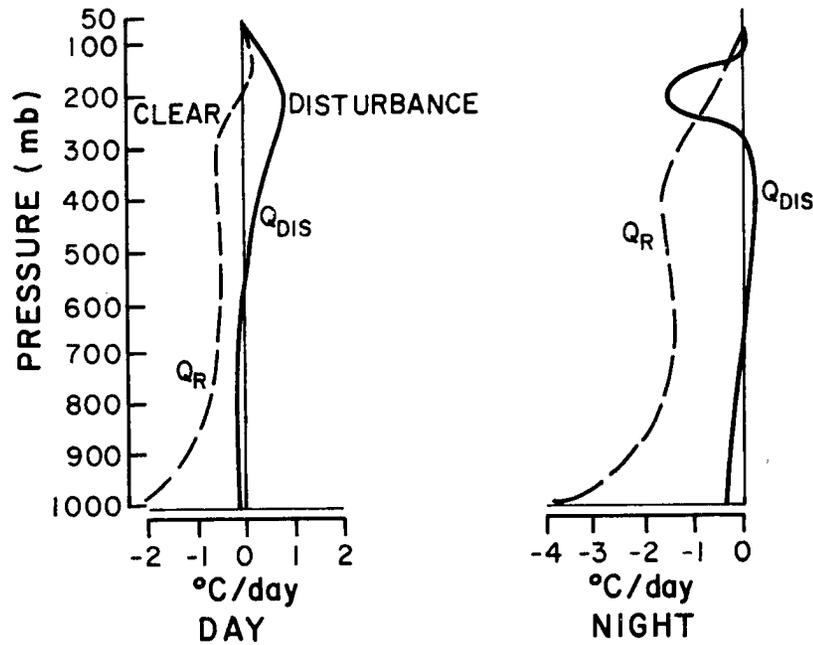


Fig. 12. Estimated typical day and night rates of radiation and condensation temperature change within a tropical disturbance ($r = 0-3^\circ$ radius) and its surroundings. Q_{dis} represent the net radiative-convective heating rate in the cloud cluster (Eq. 3). Q_R is the radiative heating rate in the surrounding clear or mostly clear region.

acting. In conventional notation the convective heating rate,

$Q_{Convection}$ is

$$Q_{Convection} = \bar{\omega} (\Gamma_d - \Gamma_a) - \frac{\partial \overline{\omega' T'}}{\partial p} + (c-e). \quad (5)$$

$\bar{\omega}$ is the vertical p-velocity averaged over the scale at which measurements are taken; and ω', T' are deviations of vertical velocity and temperature from the measurement scale average. In an active tropical weather system the terms on the right of Eq. 5 have no physical meaning since the upward motion is moist adiabatic, taking place in active cumulus clouds. Gray (1973) demonstrated that the actual 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 $\bar{\omega}(\Gamma_d - \Gamma_a)$ actually taking place. For this reason the local heat balance of the cluster has been written as in Eq. 3.

Observed temperature changes in tropical weather systems indicate that 24-h vertically integrated averages of Q_{dis} are about zero. $Q_{Convection}$ closely balances Q_R . In the surrounding clear regions, however, the radiational cooling (Q_R) is always negative. This causes heating rate differences between the disturbance and its surroundings which are about twice as large at night as during the day. These day vs. night diabatic forcing differences are believed responsible for the observed divergence differences.

It is proposed that the diurnally varying radiative-convection heating differences between disturbances and their surroundings cause changes in the inward-outward disturbance pressure gradients. Due to the low value of the Coriolis parameter at tropical latitudes, the divergent and rotational components of the wind field do not change concomitantly. The lack of close wind-pressure balance produces significant ageostrophic flow, which diurnally modulates the observed convergence fields.

It is observed that disturbance temperature varies very little as a function of the amount of cumulus convection. Convection causes small rises in the upper tropospheric temperature and small decreases in the lower tropospheric temperature. Day-night variations of disturbance radiation cause larger temperature variation than do diurnal variations in condensation. This is particularly true in the upper

troposphere where solar absorption causes upper tropospheric warming and enhanced nighttime cooling in comparison with the disturbance surrounding region. This causes day vs. night differences in upper tropospheric Q_{dis} as indicated in Fig. 12 which are only very weakly a function of day-night differences in the disturbance convection. Thus, the disturbance minus surrounding region diabatic energy differences ($Q_{dis} - Q_R$) are largely driven by radiation and have a two to one night vs. day variation. This assessment has been well documented by our project in reports by Jacobson and Gray (1976), Foltz (1976), Frank (1978a), Dewart (1978) and Grube (1978).

Evidence for the large differences in net tropospheric radiative cooling between a tropical disturbance and its environment have been well documented by direct radiometric measurements from aircraft (Griffith and Cox, 1977) and by radiative transfer calculations (Cox and Griffith, 1978). 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 $1^{\circ}\text{C}/\text{d}$ greater in the cluster's environment than it is within the disturbance itself.

There is evidence also for significant day versus night gradients in the radiational forcing. Figure 13 from Cox and Griffith (1978) shows the results of their radiative transfer calculations. A north-south cross-section is shown for an active convective day in GATE. The differences in radiative flux divergence between the disturbance and

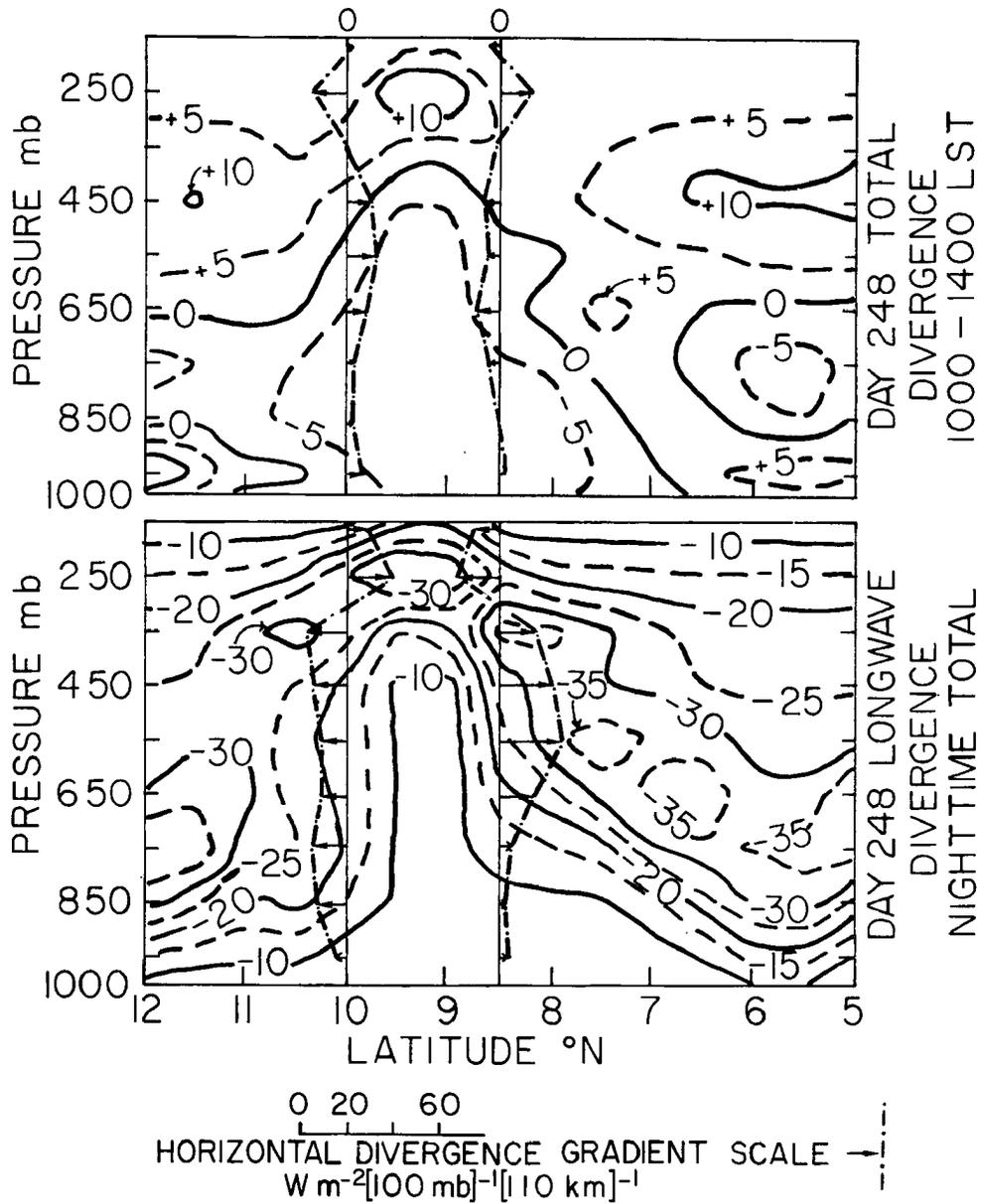


Fig. 13. A pressure vs. latitude (at 23.5°W longitude) cross-sectional view of the GATE A/B-scale array for the 0600-1800 LST period of Julian day 248. The top portion of the figure depicts the 1000-1400 LST total (SW plus LW) radiative divergence ($\text{W m}^{-2}\text{-100 mb}^{-1}$) and the bottom portion depicts the LW component only (nighttime 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 (1978).

its surroundings are much greater at night (lower portion of figure) than they are during the daylight hours (upper portion of figure). Ackerman (1978) also has studied day vs. night active GATE cloud cluster vs. surrounding region tropospheric radiational cooling differences. He finds that at night mean radiation cooling differences between the cloud cluster and its surroundings (averaged from the surface to 100 mb) are 0.63°C greater than during the 12 daytime hours. Such day vs. night cloud cluster and surrounding region radiational differences are hypothesized to be a primary component of the observed diurnal divergence differences.

The atmosphere surrounding the organized tropical disturbance adjusts to its large radiational cooling at night through extra subsidence. This extra nighttime subsidence increases low-level convergence into the adjacent cloud regions. During the day solar heating reduces tropospheric radiation loss. Clear region subsidence warming and cloud region low-level convergence are substantially reduced.

At upper levels the cloud region cirrus shields radiationally cool more at night and less during the day than their 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 3-6h.

Figure 14 shows the hypothesized slope of pressure surfaces from the disturbance to its surroundings resulting from these radiational differences. Note that the daytime solar warming of the upper dis-

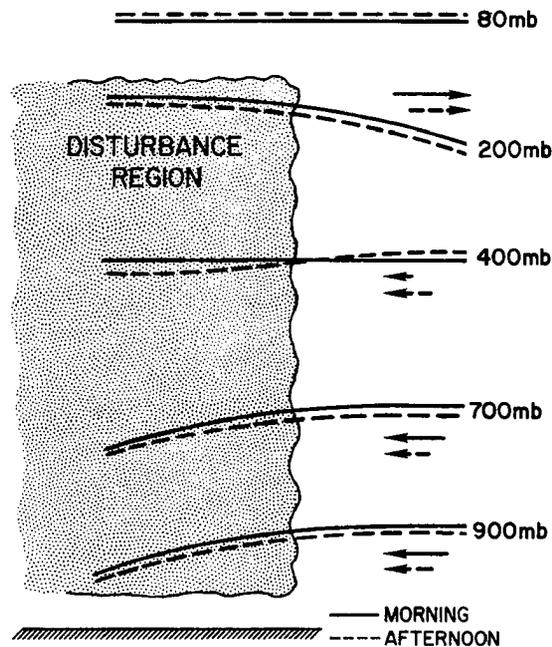


Fig. 14. Idealized slope of the disturbance to surrounding clear region pressure surfaces during the day (dashed curve) and at night (solid curve). Also shown are the corresponding down-the-pressure-gradient radial wind patterns which are established.

turbance cloud layers produces an extra downward bulging of the middle tropospheric disturbance pressure surfaces in comparison with nighttime values. This causes an enhancement of the daytime middle-level convergence and a reduction at night. At lower levels the situation is reversed. Daytime solar warming of the region around the disturbance causes a reduction of the low-level surrounding-disturbance pressure gradients and a consequent reduction of the daytime disturbance inflow as compared to the inflow at night.

The disturbance divergence profiles of Figs. 2, 6, and 9 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 disturbance minus surrounding region diabatic energy differences. Thus, maximum accumulated nighttime radiational cooling effects should occur 1-2 h after sunrise and a maximum in solar

warming effects near sunset. If the wind adjustment were to lag the disturbance's changes of inward and outward height gradients by a few hours time, the lag of the divergence profile would be greater. Maximum and minimum divergence would thus occur in the late morning and early evening. This is observed.

Fingerhut (1978) tested this hypothesis by inserting the above radiation-convective model in as simple a way as possible into a large scale axisymmetric primitive equation model of the tropical cloud cluster and environment. Fingerhut's model was diagnostic, both the convective and radiative heating being specified as in Fig. 12. In the cirrus layer of the cluster energy was absorbed following a sine wave time dependence during the daylight hours (12-h mean absorption = 29 W m^{-2}). An equal amount of energy was 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 were modelled. At tropospheric levels below the high cloud decks net radiative and condensation cooling was close to zero. In the cluster environment diurnally varying climatological net radiative cooling profiles were specified.

The model divergence and vertical motion profiles are presented in Figs. 15 and 16. Once again, features a) to f) are considered.

<u>FEATURE</u>	<u>MODEL RESULTS</u>
a) The low level convergence in the layer from the surface to 850 mb is maximum near 7:30 AM.	a) AGREEMENT: The strongest low level convergence obtained in the model occurs at 8 AM local time.
b) There is convergence between 7 PM and 10 PM in the middle troposphere near 450 mb.	b) GENERAL AGREEMENT: The model has strong middle convergence at 4 PM and at 8 PM.

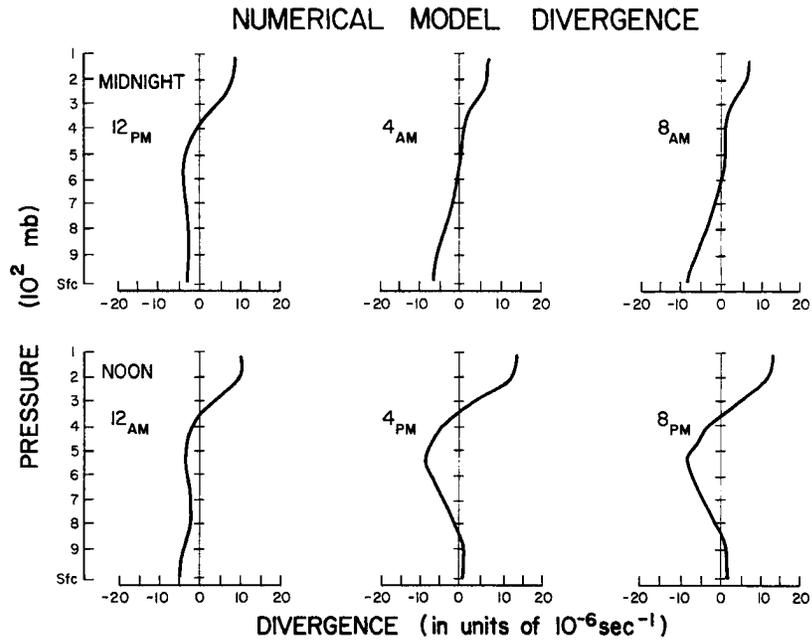


Fig. 15. Mean divergence for a 3° latitude radius cloud cluster as numerically modelled by Fingerhut (1978).

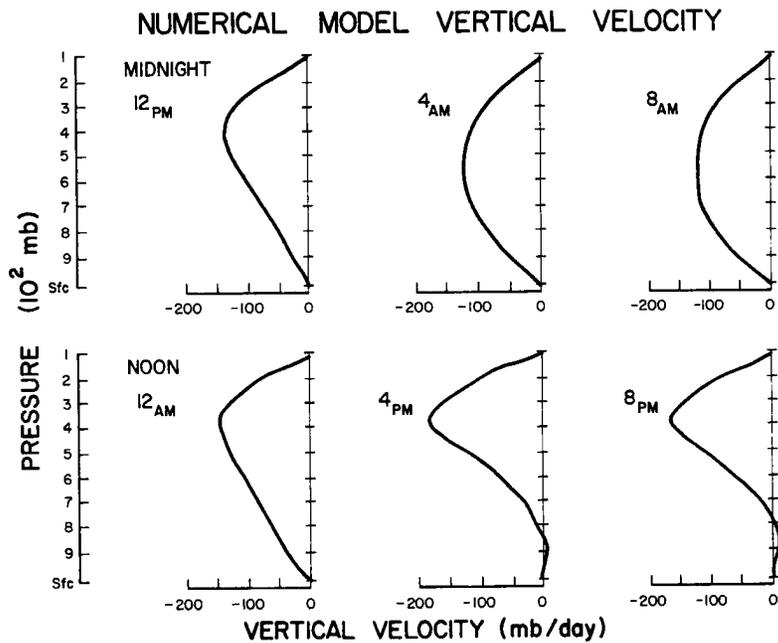


Fig. 16. Mean vertical motion for the 3° radius model cloud cluster of Fingerhut (1978).

<u>FEATURE (cont'd)</u>	<u>MODEL RESULTS (cont'd)</u>
c) In the morning the upper level outflow extends through a deeper layer of the atmosphere than it does in the evening.	c) AGREEMENT: At 8 AM, the level of non-divergence in the model is as low as 580 mb.
d) The mean tropospheric upward vertical motion is greater in the morning than in the evening.	d) DISAGREEMENT: There is little change in the total model vertical mass displacement during the diurnal cycle.
e) The level at which the maximum upward vertical velocity occurs is higher in the atmosphere in the evening than in the morning.	e) AGREEMENT: The primitive equation model has only six vertical levels in the troposphere. At 4 AM and 8 AM the model has maximum vertical velocity at 540 mb. At all other times the maximum is at the next high level, 372 mb.
f) The low level upward vertical velocity (at 850 mb) has a maximum near 7:30 AM and a minimum 12 hours later.	f) AGREEMENT: In the model, both the 708 mb level and the 876 mb level have their maximum vertical velocities at 8 AM.

The agreement shown between the simple radiative model and the observational results is striking. This is even more significant in light of the fact that when Fingerhut performed the model experiment, he had available only data set number one of this study with only twice-a-day time resolution to verify against.

Figure 17 shows diurnal profiles of vertical velocity for each of the three oceanic regions and for the radiation model. The similarity between the four curves is large and obvious. It 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 parameterization 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 was feature d), a 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). Such a feedback was not specified in the numerical model.

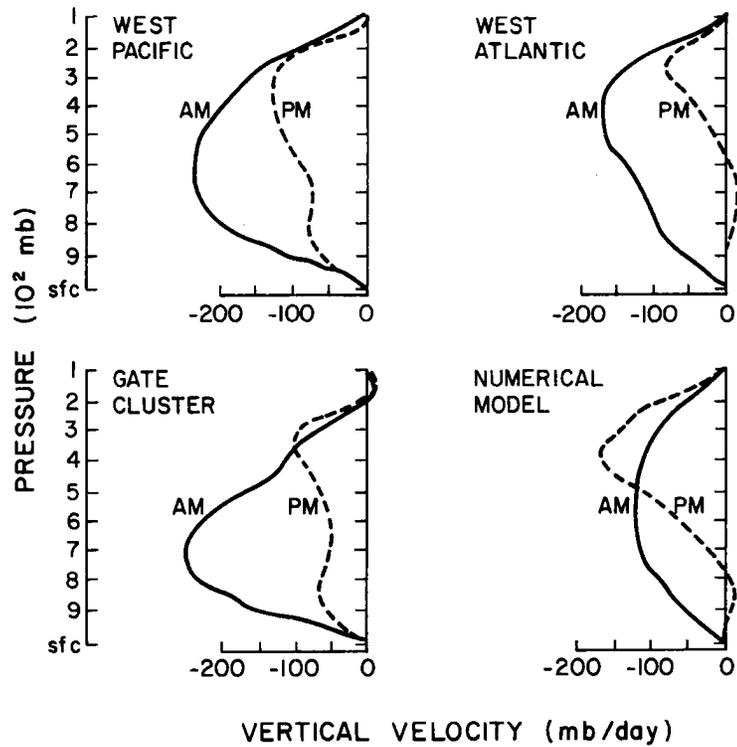


Fig. 17. Diurnal variation of vertical motion for the three oceanic regions and for the numerical radiation model. The actual curves shown are for data set 3 (West Pacific), data set 8 (West Atlantic), the 7:30 AM and 7:30 PM curves for GATE, and the 8 AM, 8 PM curves for the numerical model.

6. DISCUSSION

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

In the current study the diurnal variation of divergence and vertical motion in each region has been documented. Strong similarities have been demonstrated in the diurnal behavior 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 to that of vapor convergence if moisture storage and significant convective feedback occur.

It has been shown above that in the three regions under consideration the mass convergence below the 850 mb level is of maximum magnitude near 7 AM local time. In the western oceans this converging moisture is apparently quickly converted to rain. The diurnal variation of heavy precipitation in Western Pacific weather systems has been well documented by Gray and Jacobson (1977). They showed that rainfall has a maximum near 7 AM and a minimum near 9-11 PM. This is in agreement with the low level convergence being driven by the diurnal variation of diabatic heat sources as here discussed. A preliminary analysis of Western Atlantic

oceanic rainfall by the current authors leads them to the same conclusion for that region. However, rainfall measurements are generally contaminated by the presence of many large islands with their afternoon heating influences and by vapor storage.

In GATE there is also a maximum of moisture convergence near 7 AM local time. This can be seen in the graph of 850 mb level vertical velocity shown in Fig. 18. This vertical velocity is equal to the mass and moisture convergence below that level.

The GATE rainfall maximum, however, is in the early afternoon. This can be seen in Table 6 which shows the actual rainfall for the ten days making up the GATE composite. This time lag in GATE between low level forcing and maximum convection has been observed before, particularly by W. Frank (1978b) and by Ogura *et al.* (1977). It can be seen in Fig. 10 of the current study where the low level vertical velocity has a maximum at 7:30 AM, but the upper level vertical velocity, corresponding to the deep convection, has its maximum value at 1:30 PM.

In comparison with the western oceans, the GATE region is quite stable; in particular in the low levels it is colder (by $\sim 2^{\circ}\text{C}$) and dryer (by ~ 1 gm/kg) than the western oceans. GATE 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.

In consequence of this greater stability and greater wind shear, convection takes some time to develop in GATE. In agreement with the diurnal forcing mechanism, the GATE 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

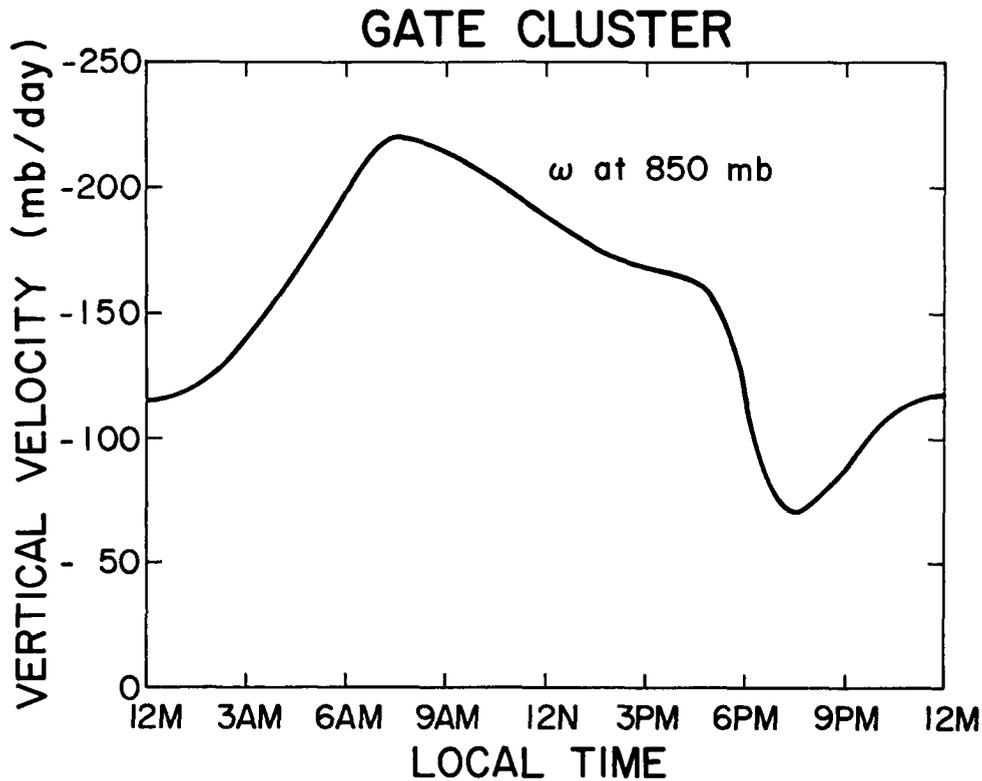


Fig. 18. Diurnal variation of the vertical velocity at the 850 mb level for the GATE composite cloud cluster of section 4.

and squall convection can overwhelm the large-scale forcing and cause rain after the forcing mechanism has subsided. The heaviest convection thus comes later than in the western ocean regions where buoyant instability and low vertical shear permit a faster response to the low level mass convergence.

Summary. It is important that this 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 vs. night and cloud-cloud free radiational and convective heating differences is much required.

TABLE 6

Average rainfall (in mm) in the GATE B-array for the ten convectively active days of GATE investigated in this study. Values are radar rainfall (obtained from Hudlow, 1977¹) averaged over the array and calibrated to agree with ship measurements on the perimeter.

DAY	Precipitation Occurring 10:30 PM - 4:30 AM <u>Local Time</u>	Precipitation Occurring 4:30 AM - 10:30 AM <u>Local Time</u>	Precipitation Occurring 10:30 AM - 4:30 PM <u>Local Time</u>	Precipitation Occurring 4:30 PM - 10:30 PM <u>Local Time</u>
188	5.268	9.772	21.916	13.581
189	2.373	4.456	10.732	4.041
195	4.031	4.166	6.762	5.650
222	6.327	10.163	1.292	0.290
245	1.825	2.831	11.746	11.713
255	0.886	3.258	9.961	0.685
256	0.939	2.888	8.678	9.310
257	8.949	4.346	5.327	8.145
259	7.060	7.994	6.274	5.615
<hr/>				
TOTAL (mm)	44.55	57.84	90.07	61.77
Average (mm)	4.5	5.8	9.0	6.2
% of Daily Total	17.5	22.8	35.4	24.3

¹Personal communication.

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MASS DIVERGENCE IN TROPICAL WEATHER SYSTEMS
PAPER II: LARGE-SCALE CONTROLS ON CONVECTION

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and
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ABSTRACT

Composited summertime deep cumulus convective weather systems are investigated over three tropical oceans. The occurrence and intensity of the convection are governed by various large scale forcing mechanisms, such as ITCZ convergence, easterly waves, and differences in diabatic heating between the disturbance and its clear or mostly cloud-free surrounding region. Vertical profiles of divergence and vertical velocity for the weather systems are analyzed from the point of view of trying to establish the relative importance of each large scale control on convection.

The dominant forcing in the Western Pacific and the GATE region is ITCZ convergence, with important secondary roles being played by differential radiative-convective heating, easterly waves and convective feedback. In the Western Atlantic trade wind region there is no contribution from ITCZ convergence; the vertical motion of a typical weather system is consequently much weaker than in the other regions.

In all three regions the contribution to vertical velocity from frictional convergence is shown to be quite small.

A large diurnal modulation of convection is observed. In the GATE region a diurnal variation is documented for the low-level convergence into the ITCZ. A physical mechanism is proposed to explain this phenomenon.

It is shown that the actual observed vertical motion of the summertime tropical oceanic weather systems comes from a combination of forcing mechanisms and does not result from any individual large scale control acting by itself. It is important that any idealized picture of tropical weather systems be broad and flexible enough to account for the large regional and multi-scale forcing mechanism differences which exist. A simple 'easterly wave' or 'ITCZ' model of organized tropical convection is quite inadequate.

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1. INTRODUCTION

The tropical atmosphere is characterized by the prevalence of deep cumulus convection. Over the tropical oceans this deep cumulus convection is observed to frequently occur in organized 'cloud clusters' with typical horizontal dimension of about 400-800 km.

Various authors have demonstrated that these organized areas of convection do not occur at random, but rather are well related to various large scale forcing functions such as the ITCZ, easterly waves, upper tropospheric troughs, etc. Each of these large scale forcing functions has a role to play in determining the location and intensity of cumulonimbus convection. Little research has been done, however, into establishing the relative importance of each control or of the interaction between these controls. It is only through such an approach that the cumulus parameterization problem can be better understood.

This paper attempts to estimate the relative importance of these various large-scale controls or forcing mechanisms. The eleven composite data sets of Paper I are analyzed to determine the contribution to the observed convergence fields from each larger scale control. Additional composite runs are made on clear or cloud-free areas in the same vicinity as the clusters, and on the mean regional convergence patterns. A first order estimate is obtained for the magnitude of the contribution from each large-scale control.

2. ITCZ FORCING MECHANISM

The mean low level wind circulation in the tropics during the Northern Hemisphere summer is shown in Fig. 1. The mean position of each of the eleven composite systems of Paper I are marked with crosses and numbers. The weaker Western Pacific systems (1, 2 and 3) and the GATE composite system (11) are in the region of the Inter Tropical Convergence Zone (ITCZ). The Western Atlantic systems (6-10) and the more intense Pacific systems (4, 5) are further poleward in the trade winds. The environmental low-level convergence in which they exist is much less.

The ITCZ is a feature of the global wind circulation and is thus a result of the planetary scale differences in the radiative heat balance of the earth-troposphere system. It was hypothesized by Riehl and Malkus (1958) that the ascending branch of the tropical Hadley Cell takes place in cumulus updrafts, so that the net upward motion is actually the additive result of upward motion in clouds and the smaller magnitude subsident motion around the clouds. This appears to be a valid assessment. The ascending branch of the Hadley Cell takes place largely in the cloud clusters and weather systems composited in Paper I. The observed low level convergence of the cloud clusters, therefore, is to at least some extent forced by planetary scale general-circulation-induced convergence. In this section the magnitude of the vertical motion resulting from the existence of the ITCZ in each of the three tropical oceanic regions of Paper I are investigated.

(a) GATE Cloud Clusters

It can be seen in Fig. 1 that the GATE A/B-array was located in

RESULTANT GRADIENT LEVEL WIND - AUGUST

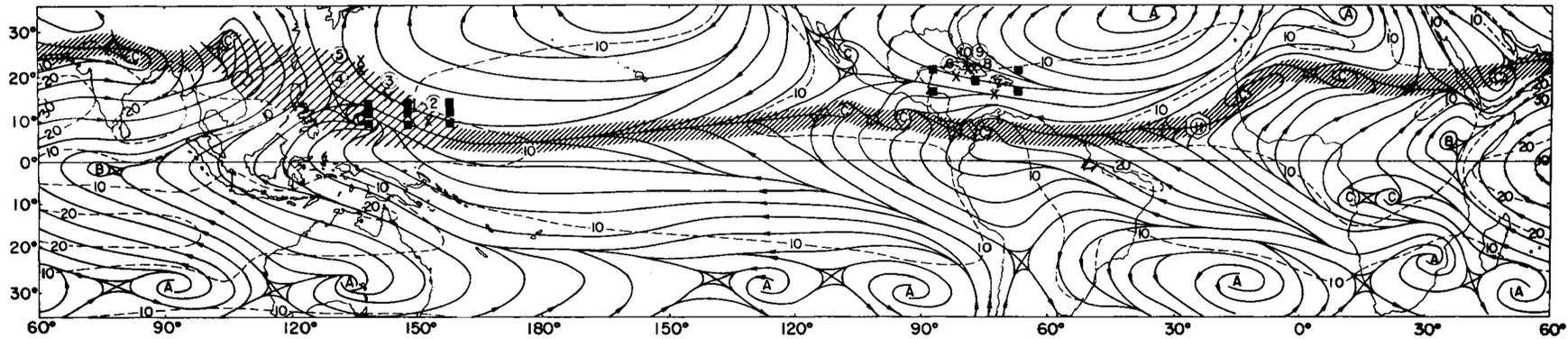


Fig. 1. Resultant gradient level wind for August (from Atkinson and Sadler, 1970). The centroid position of each of the eleven composite weather systems of Paper I is marked with a cross and a circled number. The data points used for the background or 'long term mean' composites are marked with squares. The hatched line represents the mean August position of the InTer Tropical Conversione Zone (ITCZ).

the mean position of the Eastern Atlantic low-level monsoon equatorial trough. In fact, the surface confluence line formed by the converging trade winds from the two hemispheres was located over or very near the A/B-array for most of the experiment. The ITCZ in this region is characterized by a rather narrow east-west belt of intermittent convection and high cloudiness and by very strong low-level convergence. Mean divergence profiles for the GATE convective system of Paper I and for one composite weather system from each of the western oceans is shown in Fig. 2. The low level convergence in the GATE system is two to three times as strong as that for the typical western ocean systems. Inspection of the individual terms of convergence reveals that the GATE convergence consists almost entirely of $\frac{\partial v}{\partial y}$ convergence. This results from the confluence between the northerlies and southerlies from the two hemispheres.

This type of low level ITCZ convergence was present through all of the GATE period and occurred even during days of suppressed convection. It is forced by planetary scale pressure gradients responding to diabatic energy differences between low and sub-tropical latitudes. Dewart (1978) has composited nine of the most convectively suppressed days in GATE (Julian days: 186, 190, 197, 216, 226, 227, 243, 244, and 250). The mean divergence profile for this composite is shown in Fig. 3. The profile shows that low-level ITCZ convergence is occurring even on days with suppressed convection.

It was shown in Paper I that for GATE clusters the low-level ITCZ forcing has a maximum near 7:30 AM and a minimum twelve hours later. The phase and magnitude of this diurnal modulation is such as to indicate a forcing by differences in radiative and convective heating between the cluster and its surroundings. In the cluster's surroundings the local

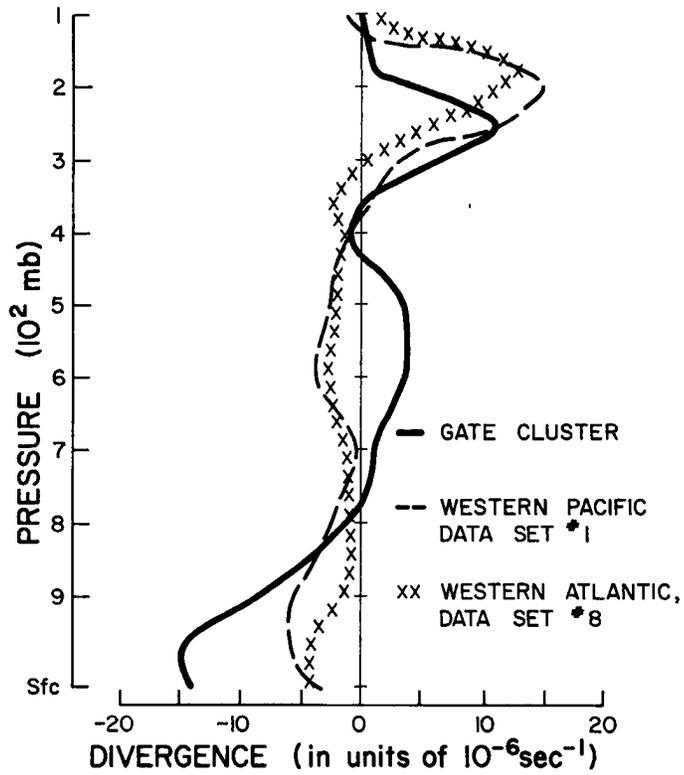


Fig. 2. Mean divergence profiles for the GATE cluster and for one composite system for each western ocean.

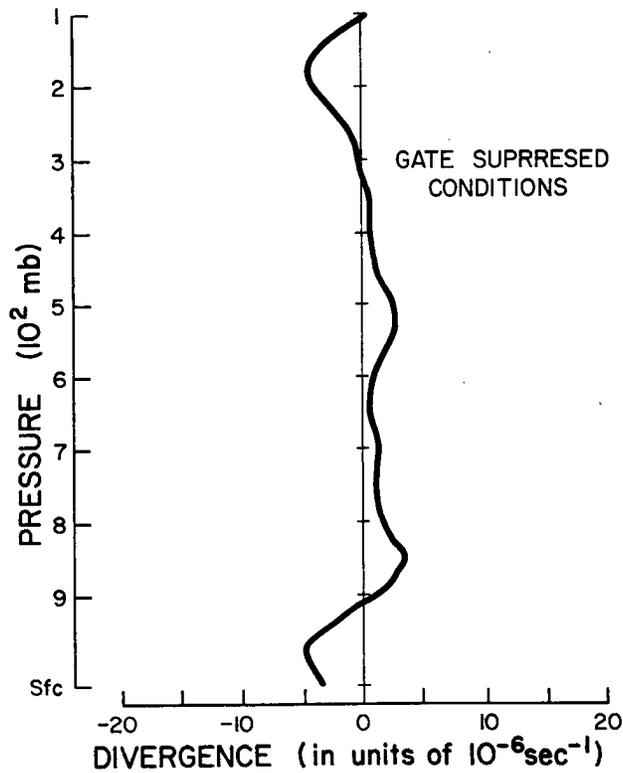


Fig. 3. Mean divergence within the GATE A/B-array for the suppressed days composite.

heat balance is dominated by radiative cooling which must be compensated for by broadscale downward motion. Within the cluster the heat balance is a residual of three very large terms: radiative cooling, local subsidence warming and cooling due to re-evaporation of cloud droplets. The balance of these terms is such that the net radiative-convective heating within the cluster is near zero.

The A/B-scale diurnal divergence profiles for the composite of the nine suppressed GATE days are shown in Fig. 4. Despite the general absence of deep convection the low level convergence is still positive with maximum value at 7:30 AM. The explanation for this is that the whole GATE region, whether enhanced or suppressed is, in a relative sense, an area of convection and local mass recycling compared to the clear and stratocumulus covered trade wind regimes to the north and south. The differential radiative-convective heating mechanism thus operates over a larger scale in GATE than in the western oceans.

In GATE the ITCZ is more distinct and concentrated. It has to the north and south both cold water and subsidence associated with the eastern sides of subtropical high pressure cells. These effects restrict the Eastern Atlantic (or GATE) ITCZ to a quasi-permanent position. By comparison, in the Western Pacific neither of the above effects is operating and the daily latitudinal variation of the ITCZ is quite large.

The GATE ITCZ therefore consists of a narrow east-west extending line of intermittent deep convection and high cloudiness with dry and subsiding air to the north and south. In the subsidence area immediately to the north is a very large region of persistent low level oceanic stratocumulus. Figure 5 shows a typical SMS visual satellite picture taken during GATE. The prominent features in the figure are the narrow

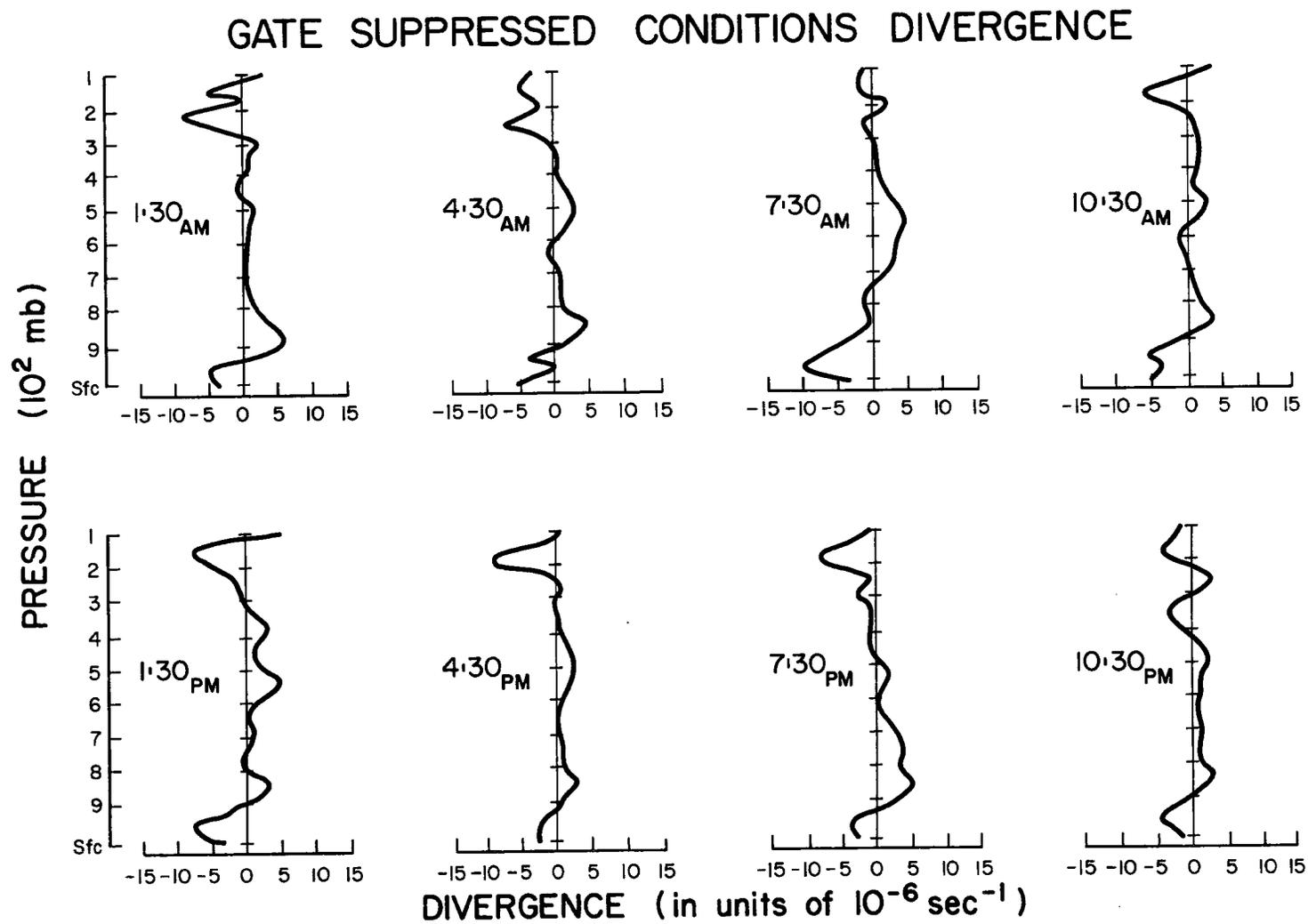


Fig. 4. Diurnal profiles of divergence within the A/B-array for the suppressed days composite.

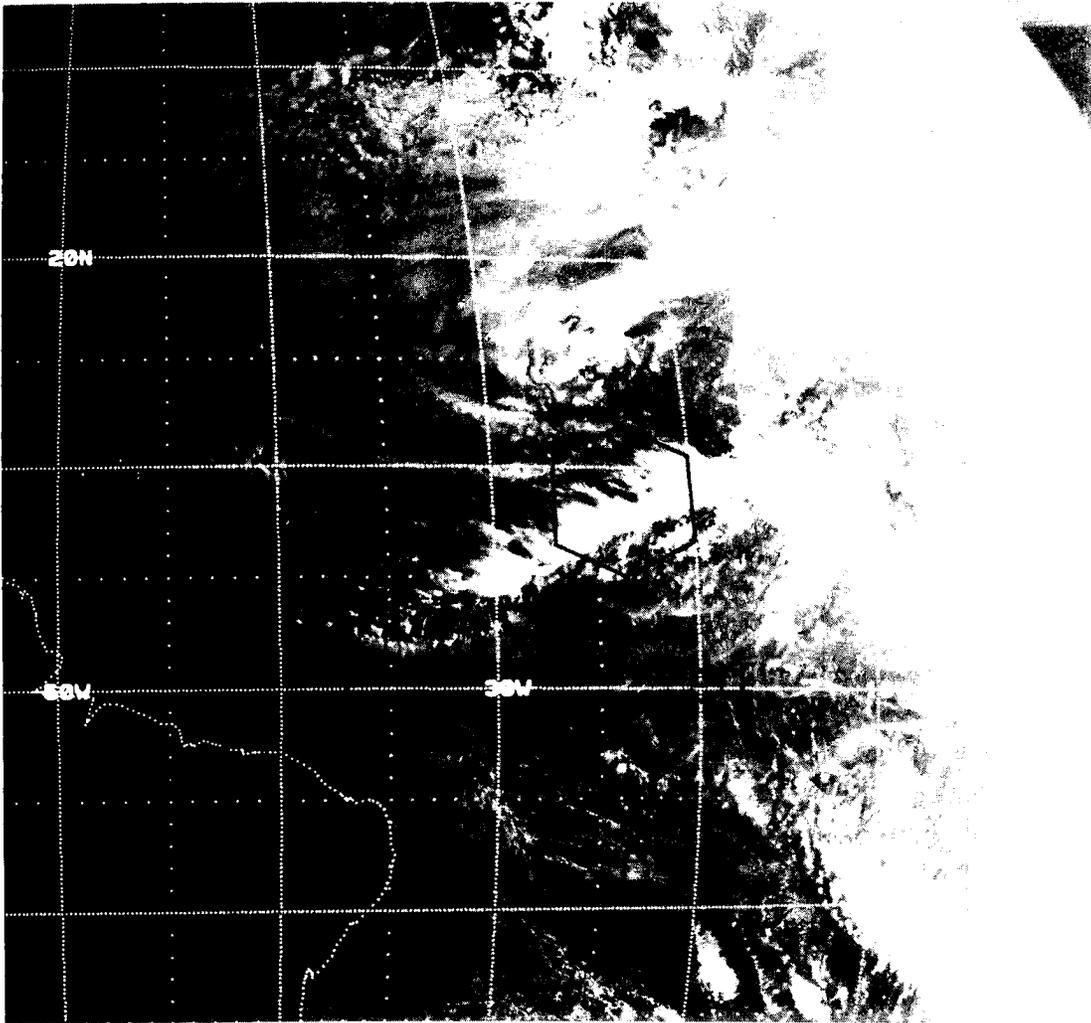


Fig. 5. Typical (0900Z, 12 August 1974) SMS visible image taken during GATE. The image shows the east-west extending line of deep convection constituting the ITCZ, the stratocumulus area to the north, and the northern extent of the stratocumulus area off the coast of southwestern Africa. From Schubert et al. (1977).



Fig. 6. Composite visual satellite imagery for the period August 16-31, 1967. (Obtained from J. Kornfield and A. F. Hasler, Department of Meteorology, University of Wisconsin.)

east-west extending line of convection constituting the ITCZ and the large stratocumulus area to the north. Also visible at the bottom of the picture is the northern extent of a large stratocumulus area off the coast of southwestern Africa, i.e. the source region for the air converging into the GATE ITCZ from the south. The day to day persistence of these features makes them visible on composite average satellite pictures for the northern hemisphere summer (Fig. 6).

The low level stratocumulus greatly enhances net tropospheric radiative cooling. Ackerman (1978) performed radiative transfer calculations for a composite easterly wave in the GATE area. He used the radiative routines of Cox et al. (1976); and he determined cloud top distributions from geostationary satellite $11 \mu\text{m}$ data as described by Cox and Griffith (1978). Figure 7 shows plan views of Ackerman's resultant net radiative flux divergence. Divergences shown are for the tropospheric layer between the surface and 100 mb, and are averaged over 12-h day and nighttime periods. The zero or reference latitude is the latitude of the 700 mb level vorticity maximum. It is just north of the ITCZ. The wave category numbering system follows the convention of Reed et al. (1977), 4 = trough, 8 = ridge. The figure shows that the stratocumulus region immediately to the north of the ITCZ is a region of very large radiative cooling. This cooling must be balanced by compressional warming brought about by large scale adiabatic subsidence. It is this strong radiation induced subsidence which drives the low-level convergence from the north into the GATE ITCZ. There is also net tropospheric radiative cooling at the latitude of the ITCZ. This is not balanced by large scale subsidence, but rather by the local up and down mass recycling typical of tropical oceanic deep convection (Gray, 1973; Yanai et al., 1973).

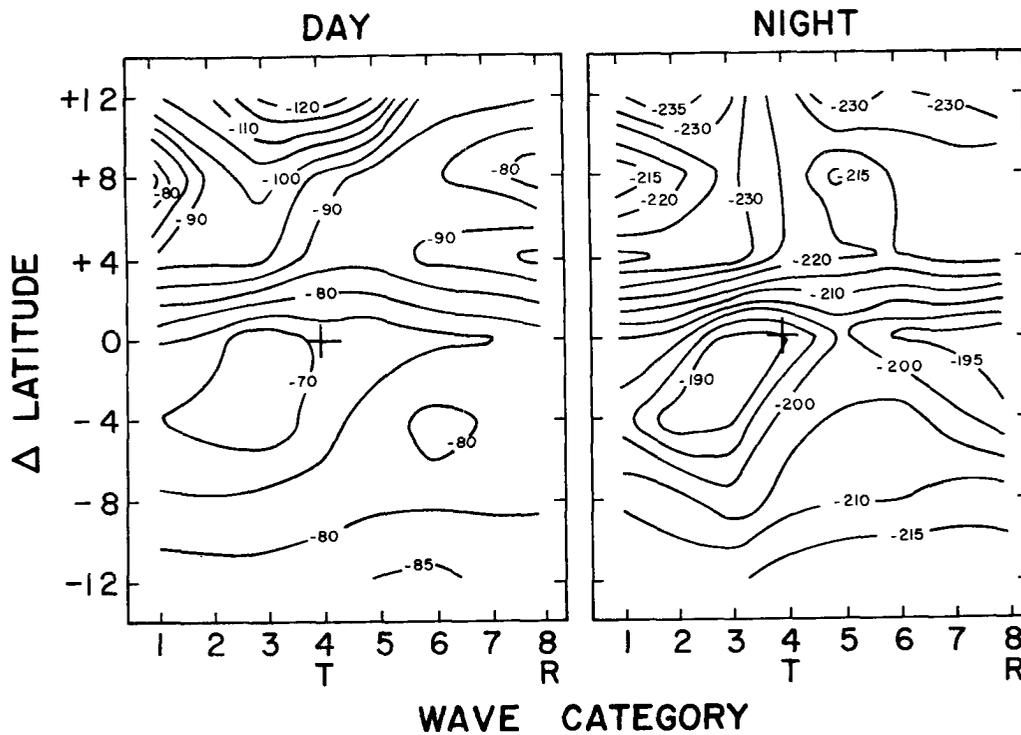


Fig. 7. Total tropospheric radiative divergence in $\text{watts/m}^2 \cdot (912 \text{ mb})^{-1}$ for a GATE Phase III composite easterly wave. The left side represents a 12-h average between 6 PM and 6 AM local time, the right for the period 6 AM to 6 PM (from Ackerman, 1978).

Figure 7 shows that the radiative cooling off the stratocumulus is much greater at night than during the day. The atmosphere responds to this cooling with large scale subsidence. The subsidence and therefore the low level mass convergence into the ITCZ consequently must have a diurnal variation. This large scale ITCZ-subtropical (or Hadley Cell) radiative-convective heating difference is analogous to the smaller scale cloud cluster versus clear area radiative-convective heating differences discussed in Paper I and modelled by Fingerhut (1978). The discussion of Paper I and the modelling results indicate that the wind response lags the heating differences by several hours; so that the maximum low level convergence into the area of radiative-convective heating maximum (the ITCZ) occurs between 7 and 10 in the morning local

time, minimum between 19 and 22 local time. This is consistent with the observations.

Figure 8 is an idealized depiction of this diurnal modulation of convergence into the GATE ITCZ. As shown in the schematic the diurnal modulation occurs predominantly in the trade winds from the north, as the stratocumulus area with enhanced radiationally induced subsidence of the Southern Hemisphere (Figs. 5 and 6) is too far to the south to offer a similar influence.

The radiation balance of the trade wind region north of the GATE ITCZ is also strongly influenced by the presence of Saharan dust. During the GATE experiment, there were several large scale outbreaks of airborne dust over the Eastern Atlantic. The dust layer extends from 900 mb to 550 mb in the vertical and in the horizontal often extends from 10°N to 25°N , with the largest concentration in the more northern part of the region.

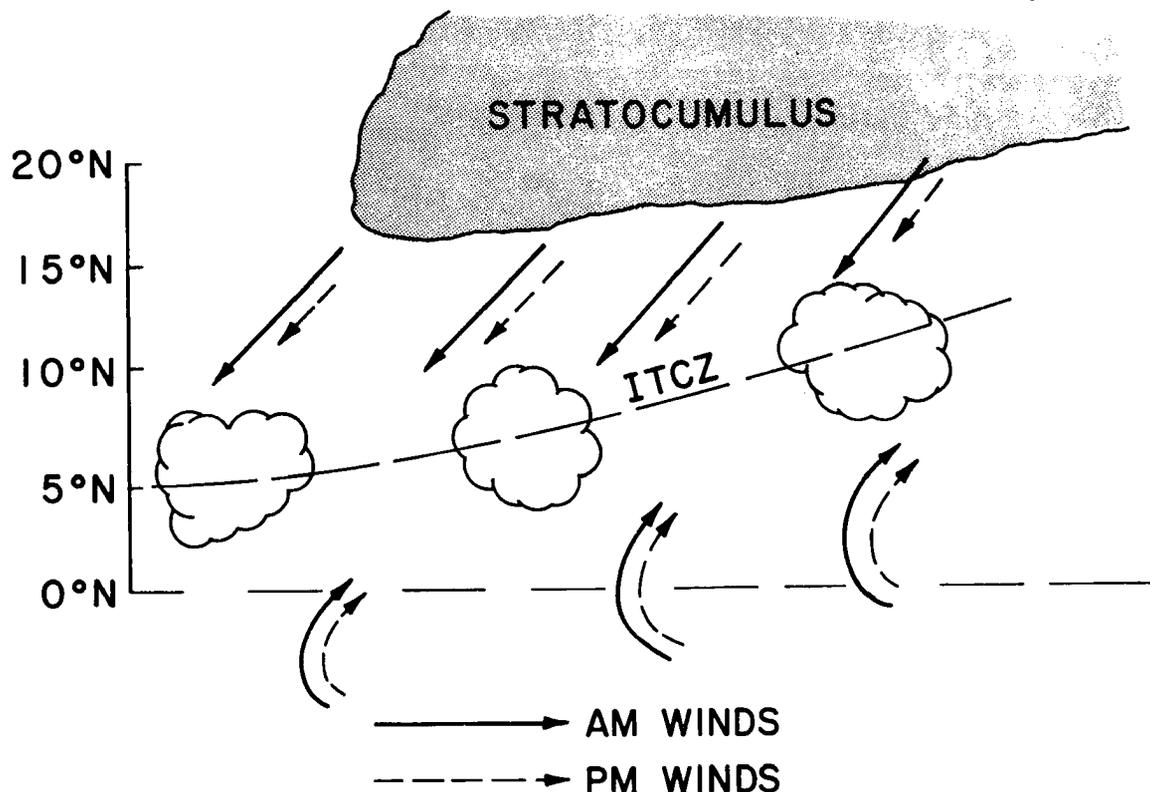


Fig. 8. Idealized model of diurnal oscillation of GATE region ITCZ. Solid streamlines represent morning circulation; broken streamlines represent evening circulation.

The dust layer is one of enhanced absorption of short wave radiation. Estimates by different scientists put the increase of the radiative heating in the low levels at between $\frac{1}{2}^{\circ}\text{C}/\text{d}$ increase (Minnis and Cox, 1978) and $1\frac{1}{2}^{\circ}\text{C}/\text{d}$ increase (Carlson and Prospero, 1977). It is thought that the dust has little effect on the long wave cooling. Thus, the overall effect of the dust is to decrease the net radiational cooling of the region north of the GATE ITCZ in the daytime and to not affect it much at night. It therefore increases the day versus night differences in radiative cooling, and consequently increases the day-night differences in subsidence required to balance this cooling. This in turn must increase the diurnal modulation of the low level flow from the north into the ITCZ. (It should be noted that the effect of the dust has not been incorporated into Ackerman's calculations shown in Fig. 7.)

Figure 9 shows the diurnal variation of the meridional component of the 950 mb level wind as measured by the three USSR ships constituting the northern boundary of the GATE A/B-array. Values shown are averages at that time period for the whole GATE experiment. The 7-10 AM local time maximum in northerly winds is clearly evident.

Figure 10 shows a schematic north-south vertical cross-section through the GATE ITCZ. The top half of the figure shows how the difference in day-night radiative cooling is hypothesized to effect the planetary scale pressure gradients, while the lower half shows the diurnal response of the Hadley circulation winds.

(b) Western Pacific Clusters

Referring again to Fig. 1, the Western Pacific composite cloud clusters (data sets 1, 2 and 3) are seen also to be associated with the

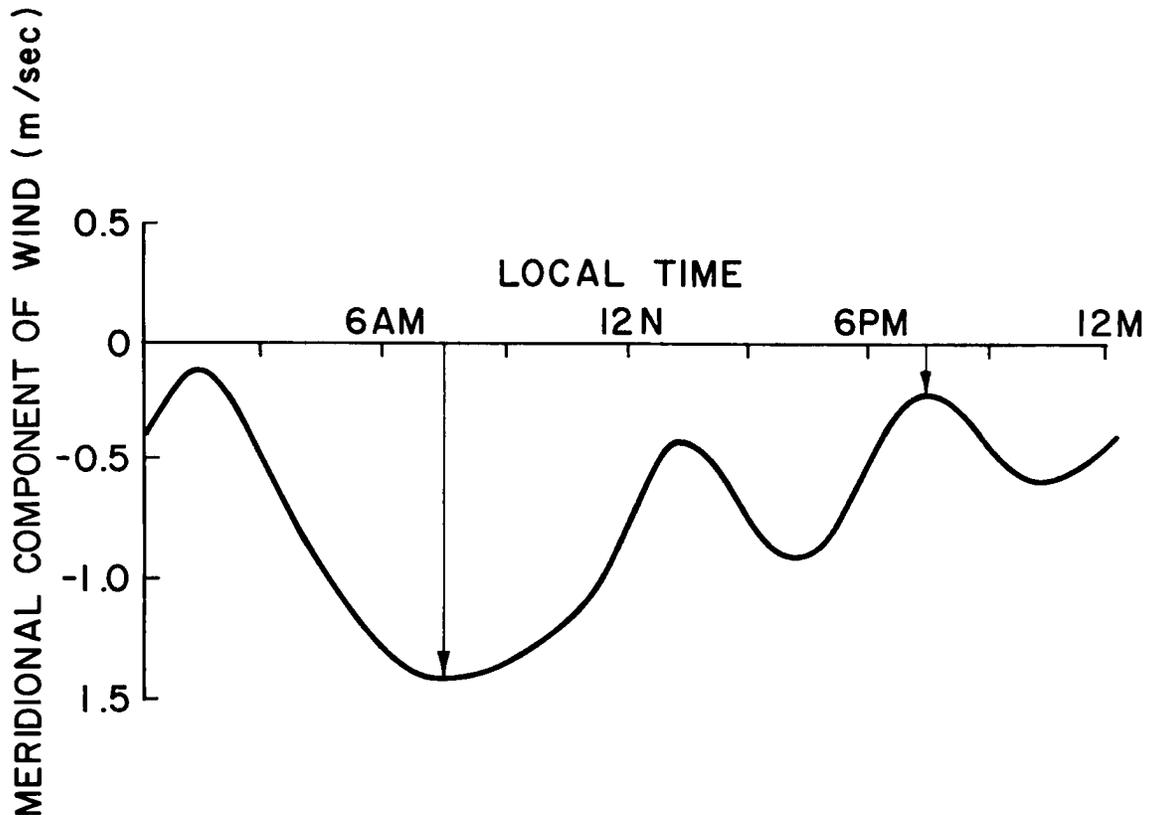


Fig. 9. Diurnal variation of the meridional component of the wind at 950 mb for the three northernmost ships of the GATE A/B-array. Values shown are average of the 3 ships at each time period averaged from the whole experiment. Negative values denote winds from the north.

ITCZ. The ITCZ here, however, has a different character to that in the GATE area. It is not so latitudinally restricted by cold water to the north and south, and its daily position alteration is much larger. It has a general northwest to southeast orientation, and the convergence in the vicinity of the composite clusters is due to the deceleration of the easterly flow as it meets the westerly monsoon flow originating in the other hemisphere. The convergence is thus $\frac{\partial u}{\partial x}$ convergence as distinct from the GATE area $\frac{\partial v}{\partial y}$ convergence.

To evaluate the contribution of background regional convergence to the observed convergences of the weather systems data were composited relative to nine geographically located positions marked with solid closed boxes in Fig. 1. The compositing used all 00 GMT and 12 GMT

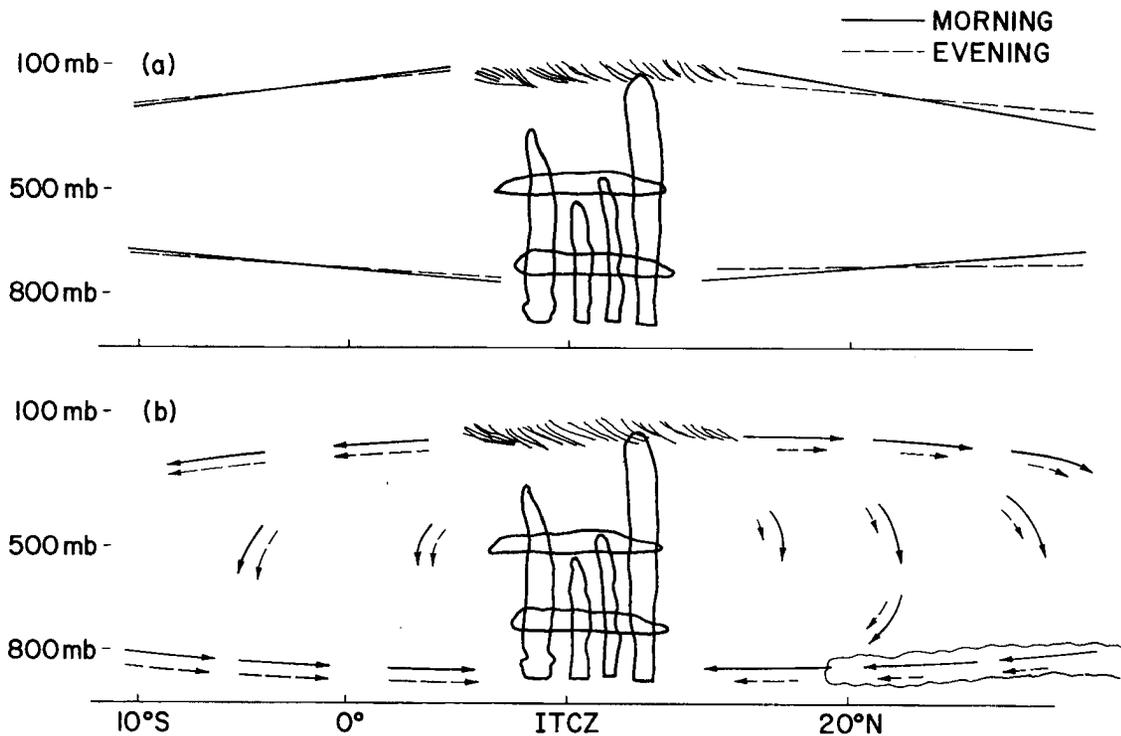


Fig. 10. Idealized model of the diurnal variation of the ITCZ mass circulation with the hypothesized slope of the pressure surfaces (a) and the resulting wind circulation (b). The solid line represents the typical morning circulation, the dashed curve the evening circulation.

rawinsonde data for the period July-October for the two years 1967 and 1968 from the reduced data network shown in Fig. 11.

Divergences were obtained by line integral methods for a 4° latitude radius circle and for an 11° latitude radius circle centered at the background positions indicated. The vertical profiles of divergence and resultant vertical velocity are shown in Figs. 12 and 13. Figure 13 also shows the mean vertical velocities for the weather system data sets 1, 2 and 3 of Paper I.

In Fig. 12 it is seen that the Western Pacific mean summertime Inter Tropical Convergence Zone has a completely different character to that in the GATE area. Rather than being a boundary layer phenomenon

over a very large area at least as large as a 22° latitude diameter circle. This is partly a spacial averaging problem, because the Western Pacific ITCZ moved northward and southward in time much more than the GATE ITCZ. Note in Fig. 14 how the area of 26°C summertime sea surface temperature is much restricted in the latitudinal direction in the eastern as compared with the western oceans. The whole summertime Western Pacific region is an area of general upward vertical motion. This large area of upward motion can also be seen in mean cloudiness (Fig. 6).

Figure 13 indicates that a large portion of the observed cloud cluster vertical motion in the Western Pacific can be attributed to "ITCZ", "general circulation", or "background" convergence. Because the region of cloudiness and upward motion is large and more diffuse, the differential convective-radiative forcing mechanism does not operate on it as a whole as it does in the eastern oceans where the areas of upward vertical motion are much more latitudinally restricted. ITCZ cloud and cloud-free radiative-convective heating differences are thus more diffuse and less geographically fixed than in the eastern oceans. There is much less diurnal variation in the large-scale ITCZ divergence profiles of Fig. 12 than is observed in the A/B-array GATE area of the eastern ocean.

It was shown in Paper I that there exists a large diurnal variation in the upward vertical velocities of Western Pacific cloud clusters; yet Fig. 12 shows no diurnal variation in the total vertical velocity over the whole region. To conserve mass, therefore, there must be a diurnal variation of the opposite sense in the cloud-free or "clear" parts of the region; that is, the clear areas should have maximum low-level subsidence in the morning. Williams and Gray (1973) picked the positions

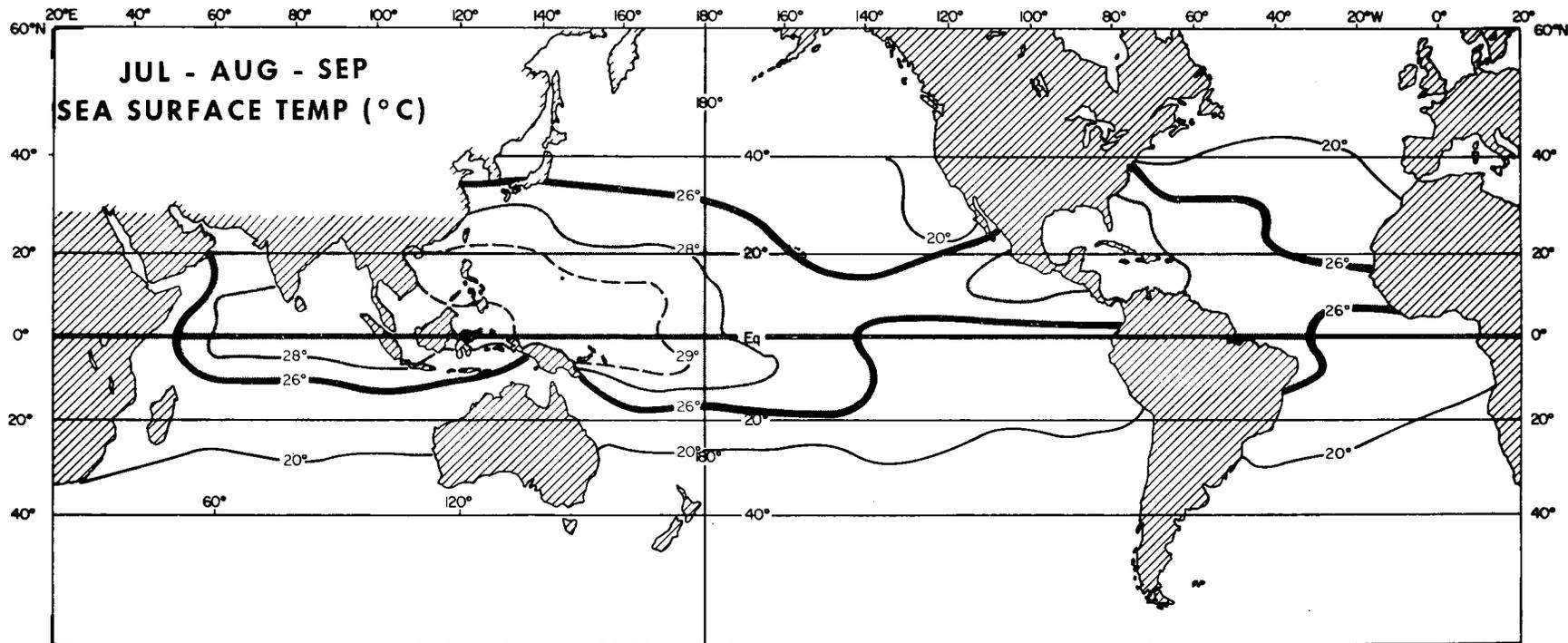


Fig. 14. Illustration of how the area of warm ($> 26^{\circ}\text{C}$) ocean surface temperature occupies a latitudinally restricted region in the Eastern Atlantic and in the Eastern Pacific as compared with conditions in the western oceans.

of Western Pacific clear areas from ESSA satellite pictures for the years 1967 and 1968. Their clear area was defined such that at least a ten degree latitude square area was free of significant cloud cover. Williams and Gray provided positions only for the 00 GMT time period, as this time was close to that of the daily ESSA satellite photographs. The diurnal variation of divergence and vertical velocity was studied by compositing Williams and Gray's clear region positions at 00 GMT. For 12 GMT the same positions were used. Separate composites were made for 12 hours before and 12 hours after each 00 GMT position. Much the same vertical profile of divergence was obtained at each 12Z period, so the average of the two 12 GMT composites was used. The resultant diurnal vertical velocity profiles are shown in Fig. 15. As predicted,

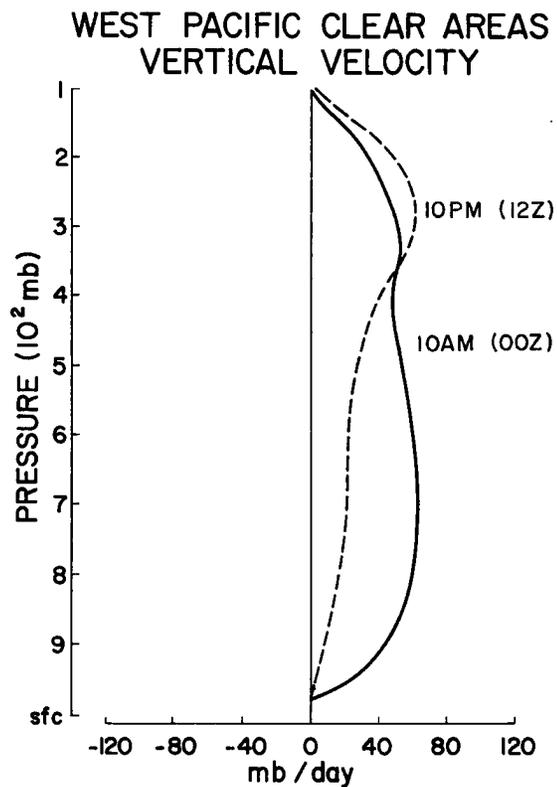


Fig. 15. Mean vertical velocity within the $r = 0-3^{\circ}$ area for Western Pacific clear areas.

there is subsidence or downward vertical velocity throughout the whole troposphere with significantly greater low-level downward motion in the morning than in the evening.

(c) Western Atlantic Weather Systems

In Paper I it was shown that Western Atlantic cloud clusters and easterly wave disturbances are characterized by very small vertical velocities, with slight subsidence even being measured at the early evening period (~ 19 LT or 00Z). This is partly due to the 6° diameter area over which the measurements are made. The area is so large that the radiation induced subsidence within it cancels the upward vertical motion in the cloud area and cloud lines of the weather system. Referring to Fig. 1, it is seen that the Western Atlantic composite systems are not in the vicinity of the ITCZ but are further north in the subsident region of the trade winds. It should thus be expected that the background regional convergence in this area makes a negative contribution to cluster upward vertical motion.

Similar background and clear region composites were made for the Western Atlantic as in the Pacific, using the same 1967-1968 rawinsonde set. Five geographical positions (marked by closed boxes on Fig. 1) were used for the background composite.

Taken over both a small region (4° radius circle) and a large region (12° radius circle) the West Indies background composite vertical velocity is slightly downwards and larger at 12Z (07 LT) than at 00Z (19 LT) (Fig. 16). The figure also shows the vertical velocity profiles for the West Indies clear region. Mass continuity arguments dictate that the clear region subsidence have the opposite variation to cloud

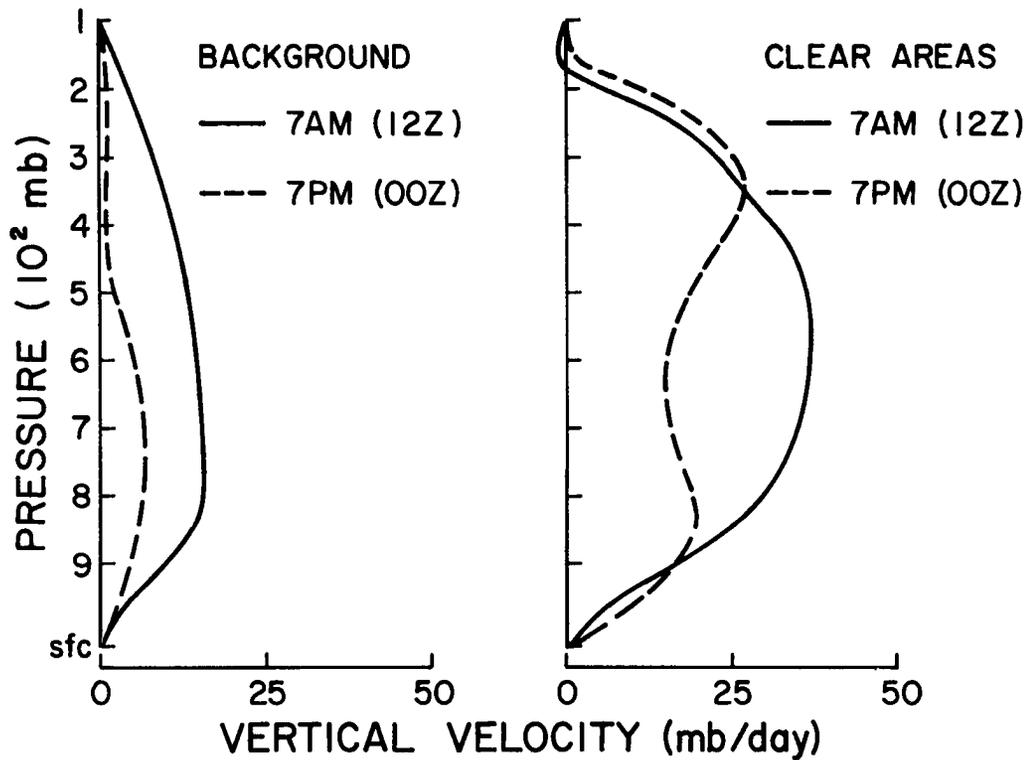


Fig. 16. Vertical velocity for the Western Atlantic background and clear area composites.

clusters, that is maximum low level subsidence in the morning. This is observed.

(d) Summary

The picture that emerges from the above discussion is one of tropical cloud clusters existing in three very different background regimes. The GATE clusters make up the Eastern Atlantic summertime ITCZ. They exist in a region of strong low level convergence resulting from the merging of strong northeast trades with monsoon southwesterlies. A narrow latitudinal and low level concentration of vertical motion exists.

The Western Pacific clusters exist in a much larger latitudinally broad region of mean seasonal upward motion. Convergence occurs over a

deep tropospheric layer. This convergence is brought about by the oceanic trade wind easterlies decelerating as they come across the monsoonal influence of the Asian landmass.

The Western Atlantic composite region is one of general tropospheric subsidence. Weather systems show a general weakening as they move into this region.

3. EASTERLY WAVES

There appear to exist two preferred scales of weather system organization in the tropics, the cloud cluster (400-800 km) and the easterly wave (2000-4000 km).

Empirical synoptic models of the easterly wave relating the wind field to the convection were proposed by Riehl (1945) for the Caribbean and by Palmer (1952) for the Western Pacific. N. Frank (1969) described the movement of easterly waves from Africa westward across the Atlantic. Reed and Recker (1971), Reed et al. (1977) and Burpee and Dugdale (1975) in composite studies have demonstrated that these easterly waves have a pronounced modulating effect on convection and rainfall, i.e. on cluster activity, in the Western Pacific and in the GATE region for Phase III.

Gray (1977a) has determined the time of passage of 700 mb troughs and ridges over the GATE A/B-network for all phases of the experiment by a consideration of horizontal meridional shear over the array. A time series of GATE A/B-values of $\partial v/\partial x$ with troughs and ridges marked is presented in Fig. 17. The time of passage of a trough or ridge was defined as the time of maximum or minimum $\partial v/\partial x$. For comparison, the GATE B-array C-band radar determined rainfall from Hudlow (1977)¹ is plotted on the same diagram.

A preliminary glance at this figure strongly substantiates the findings of Burpee and Dugdale (1975) and Reed et al. (1977). The statistical association of the rainfall events in GATE to the passage of a 700 mb wave trough is quite pronounced, especially during Phase III. This is shown in Fig. 18.

¹Personal communication.

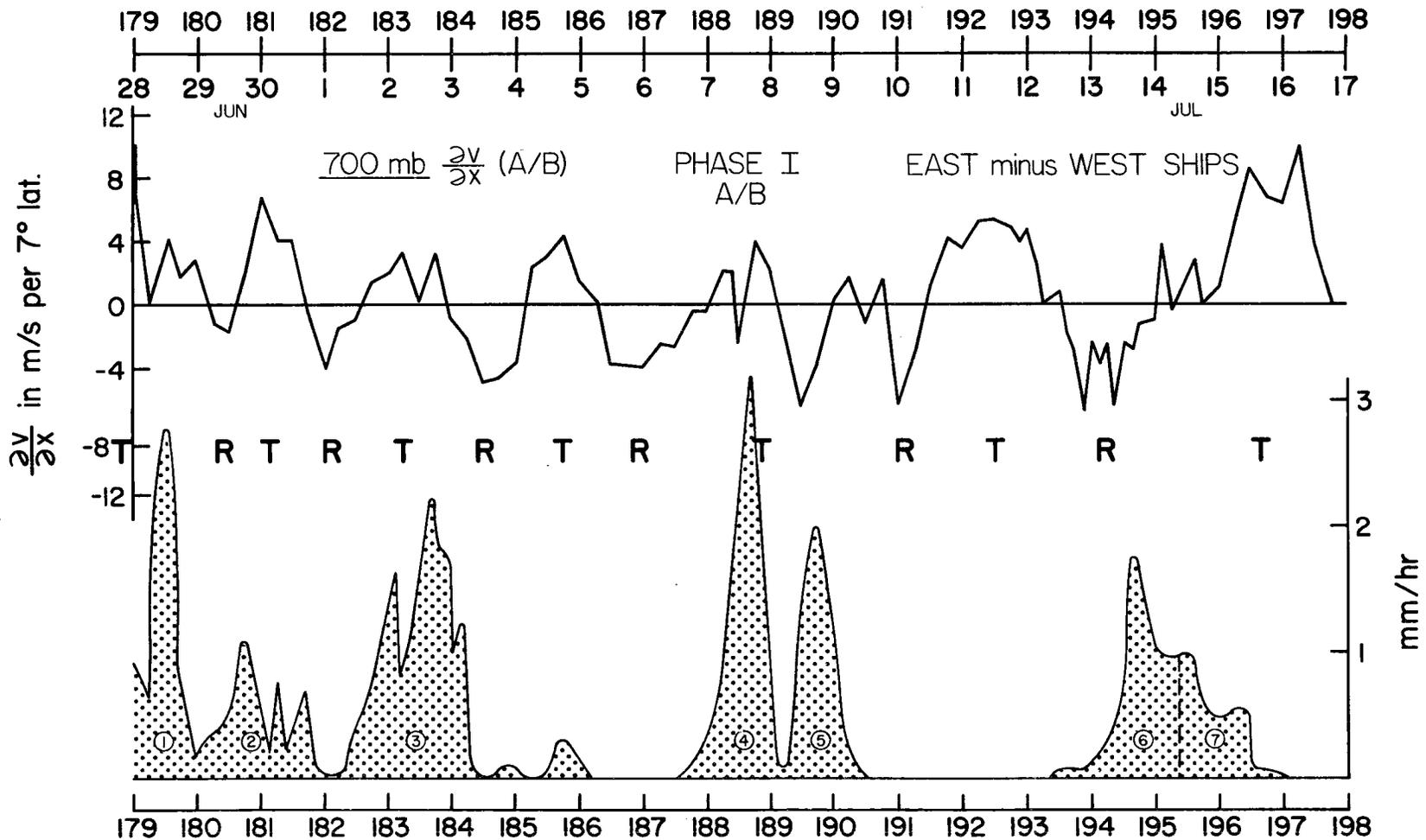


Fig. 17. GATE Phase I, II and III radar determined B-array rainfall in mm/h (bottom) and A/B-array $\partial v/\partial x$ wind changes. R and T denote times of A/B-array passage of 700 mb ridges and troughs.

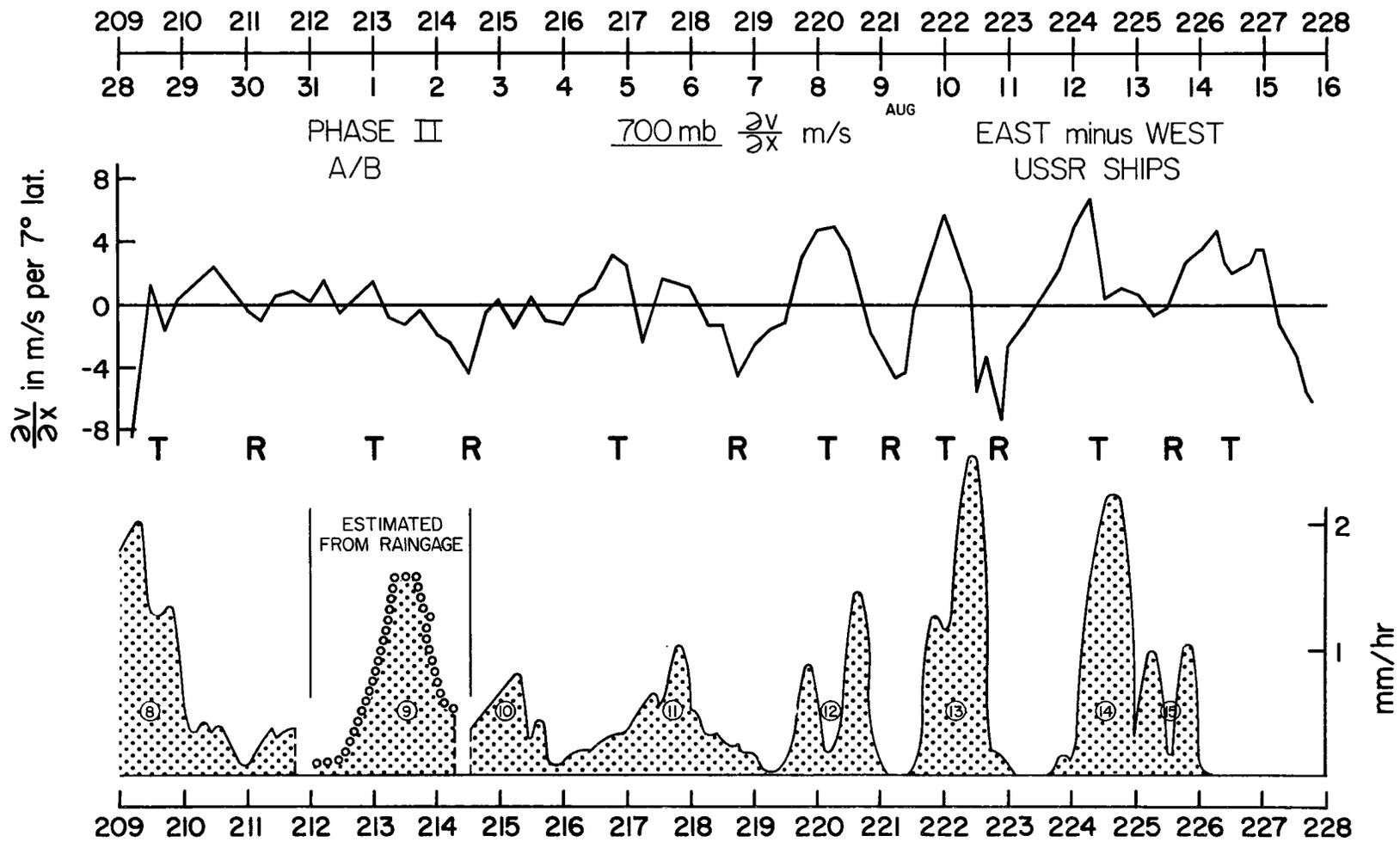


Fig. 17. Continued.

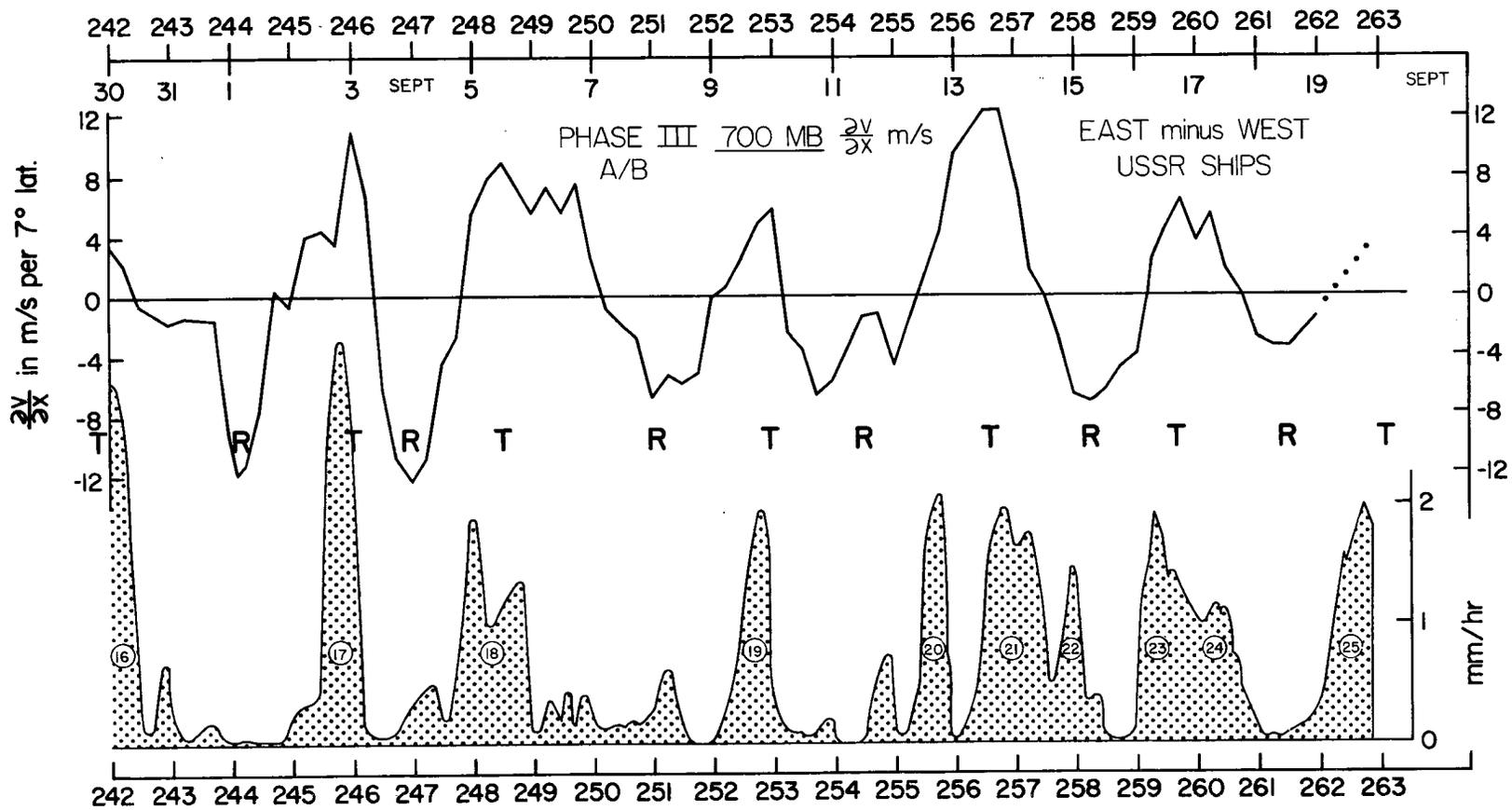


Fig. 17. Continued.

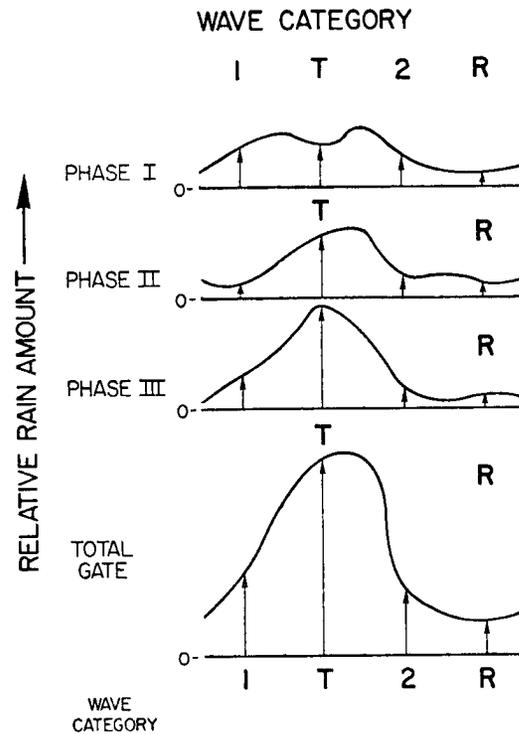


Fig. 18. Phase by phase B-array radar measured relative rainfall amounts (from Hudlow, 1977) as associated with 700 mb determined trough and ridge wave patterns over the GATE A/B-array.

It should be noted in Fig. 17, however, that there is also quite a large variability. Sometimes the heavy rain events appear to the west of the wave trough, and at other times to the east or even in the ridge. There were times during Phase II when significant rainfall occurred without observable wave passage. In turn, there were two pronounced wave passages in Phase I which produced very little rainfall. Nevertheless, an obvious modulation of the rainfall by the passage of the easterly waves is apparent in both figures.

To compare the wave modulation with the diurnal variation documented in Paper I, an A/B-scale composite was made of all 700 mb waves in GATE. Four times daily rawinsonde observations were used. The trough composite consisted of all A/B-scale data from the trough time period plus time periods six hours earlier and six hours later. The composite ridge was

similarly composed of three consecutive time periods centered on the minimum of $\partial v/\partial x$. Wave positions 1 and 2 were defined as halfway positions on the time axis between trough and ridge, as shown in Fig. 19.

The resultant A/B-scale vertical motion profiles for the trough, ridge and two intermediate positions are shown in Fig. 20. Mean upward vertical motion exists in all four wave positions with low level (at 850 mb) deviations of trough and ridge from the wave mean of about ± 30 mb/d. The diurnal ω -profiles for each of the composite wave positions are shown in Fig. 21. Diurnal modulation of the 850 mb ω is the same for each wave category and is about equal to the wave modulation. The curves in Fig. 21 are consistent with those discussed in Paper I for a GATE cluster and in Fig. 4 of this paper for GATE suppressed conditions. In particular, the low level ITCZ forcing is diurnally modulated so that under all conditions the 850 mb level vertical velocity is maximum near 7 AM local time and minimum 12 or more hours later. The diurnal and wave variation modulations to the 850 mb vertical motion in the GATE A/B-array are about the same.

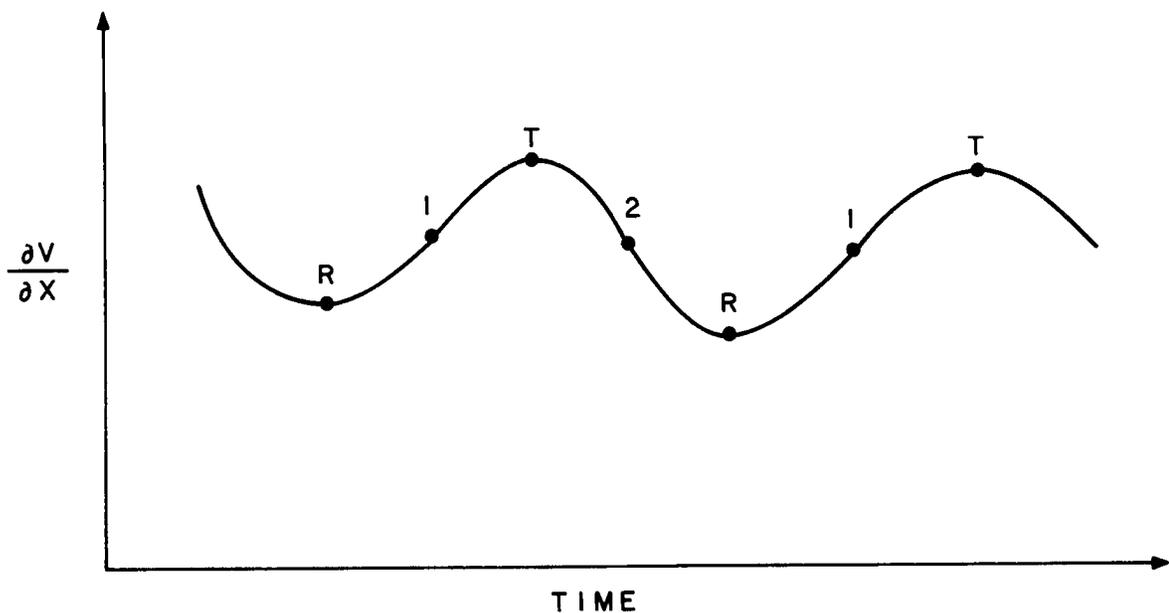


Fig. 19. Relative positions of 700 mb wave trough, ridge and intermediate positions 1 and 2.

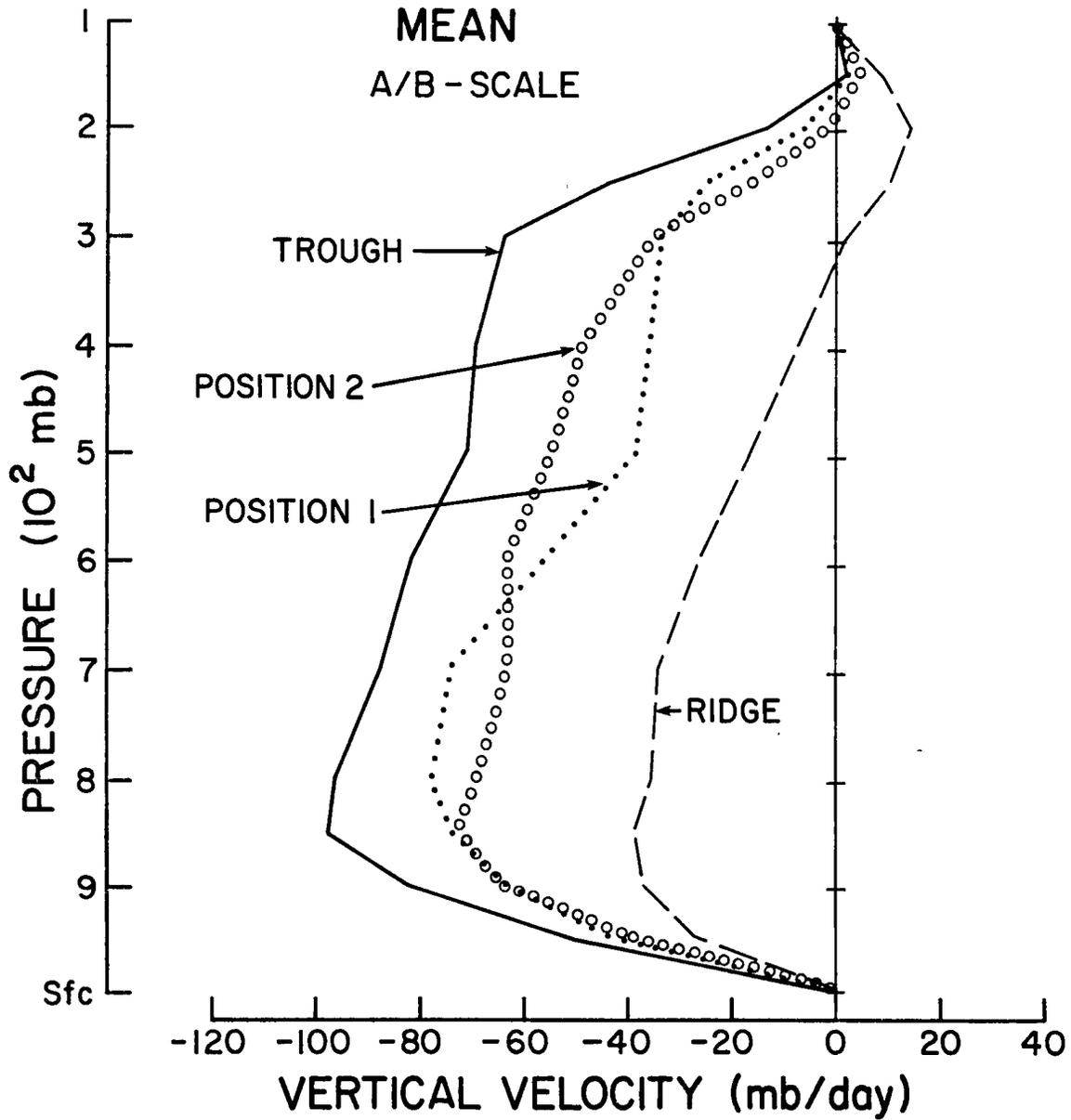


Fig. 20. Mean vertical velocity within the GATE A/B-array for the composited easterly wave.

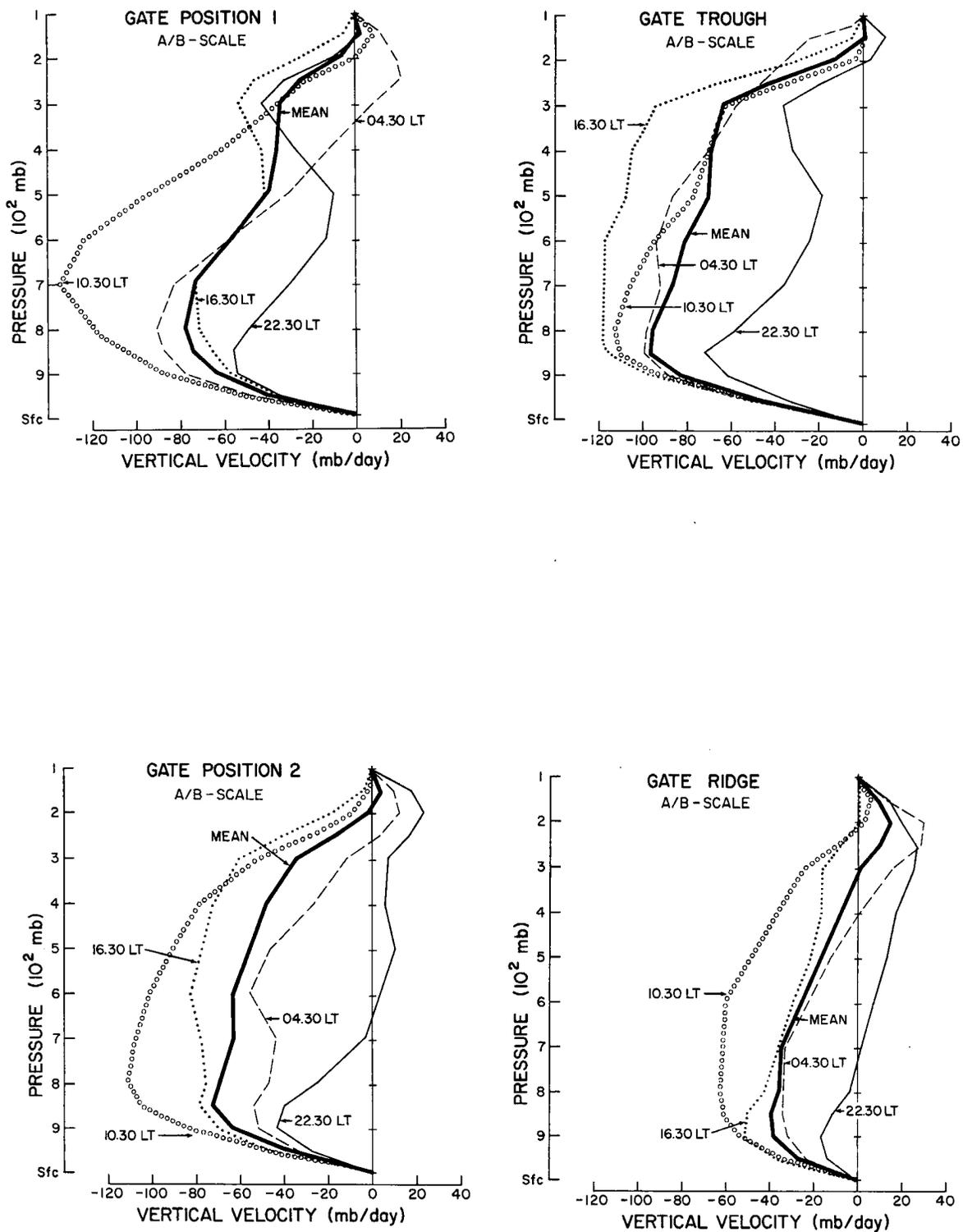


Fig. 21. Diurnal variation of vertical velocity for the four composite easterly wave positions as described in Fig. 19.

4. BOUNDARY LAYER FRICTIONAL CONVERGENCE

In regions of low-level cyclonic vorticity, quasi-geostrophic-Ekman dynamics (as described by Charney and Eliassen, 1949) determine that low level frictional convergence takes place. In regions of anti-cyclonic vorticity there is boundary layer suction carrying down air from aloft.

This frictional mechanism is no doubt operating in tropical disturbances; but it has been shown in recent years (Gray, 1975; Gray and McBride, 1976; Gray, 1977b) that it plays only a small role in cloud cluster convergence and in the early stages of development of tropical storms. The basis for this comment is that the observed convergence of mass into these systems is much greater in magnitude and extends through a much deeper layer of the atmosphere than can be explained through boundary layer frictional influences. The following discussion describes the portion of weather system vertical motion which can be attributed to boundary layer frictional convergence.

Gray (1972) and Gray and Mendenhall (1974) statistically analyzed pibal and rawinsonde wind data over the tropical oceans. They showed that, after the thermal wind effect had been eliminated from the data, the typical veering of the surface wind relative to the wind at the top of the boundary layer is about 12° . These results have been verified with the GATE data. It is assumed that this veering is due to mechanical frictional processes. It is emphasized that the veering angle referred to is that between the surface wind and the wind at the top of the layer of frictional influence. It is not the angle between the surface wind and the surface isobars. In the tropics, the latter angle

may be much larger. Gray (1972) observed also that over the tropical oceans the decrease of wind speed from the top of the boundary layer to the surface (from ship data) is small. This is demonstrated in Fig. 22.

In Fig. 23 a schematic is shown of the wind at the top of the boundary layer at a distance R from the center of the disturbance. From the figure it is seen that the convergence at the top of the boundary layer is equal to $\tan \beta$ times the vorticity at this level (or $\overline{\zeta_T}$). If it is assumed that the tangential wind maintains its strength through the boundary layer, and that the boundary layer frictional processes bring about a further 12° turning of the wind, then the convergence at the surface is equal to $\tan(\beta + 12^\circ) \times \overline{\zeta_T}$. Thus, if the veering varies linearly with height, the average convergence in the boundary layer is $\frac{1}{2} \overline{\zeta_T} \times (\tan\beta + \tan(\beta + 12^\circ))$. Defining C_F as the additional convergence due to the frictional turning of the wind:

$$C_F = \left[\begin{array}{c} \text{calculated boundary layer} \\ \text{convergence} \end{array} \right] - \left[\begin{array}{c} \text{convergence at the top of the} \\ \text{boundary layer} \end{array} \right]$$

$$C_F = \frac{1}{2} \overline{\zeta_T} (\tan\beta + \tan(\beta + 12^\circ)) - \overline{\zeta_T} \tan\beta$$

$$C_F \approx \frac{1}{2} \overline{\zeta_T} \tan 12^\circ$$

$$C_F \approx 1/10 \overline{\zeta_T} \quad . \quad (1)$$

As documented in Paper I, the low level convergence of tropical weather systems (except for special conditions of GATE clusters) extends

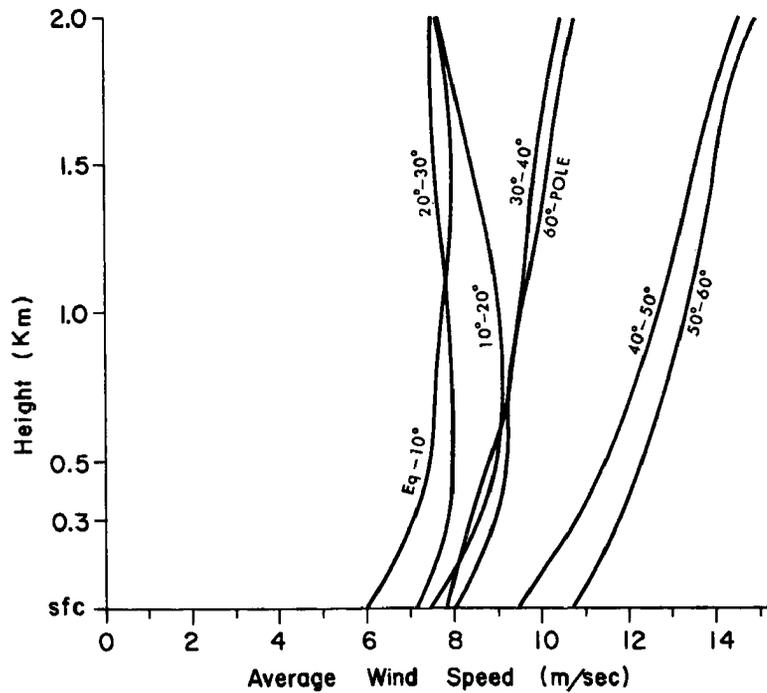
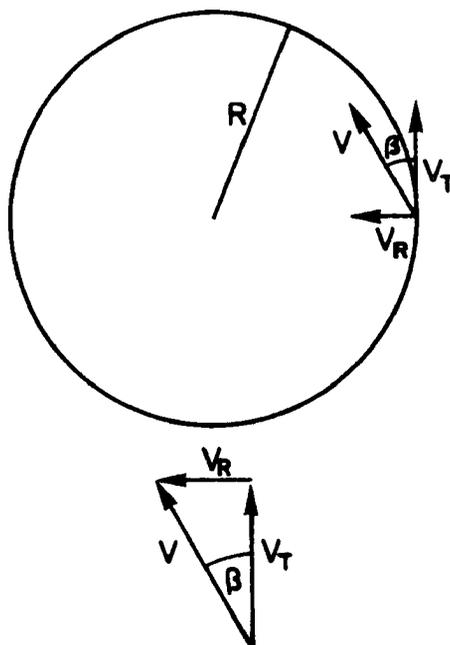


Fig. 22. Variation of wind speed with height in the lowest 2 km. The data shown are from a composite of over 100,000 pibal and rawin soundings over the Northern Hemisphere oceans, performed by Gray (1972).



V = Wind at top of Planetary Boundary Layer (P.B.L.)

β = Inflow angle at top of P.B.L.

V_T = Tangential wind = $V \cos \beta$

V_R = Radial wind = $V \sin \beta$

Fig. 23. Schematic depiction of the wind field at the top of the boundary layer at the radius R . By Stokes' theorem, the vorticity averaged over the circle = $2 V_T/R = \zeta_T$. By Gauss' theorem, the divergence averaged over the circle = $2 V_R/R = \tan \beta$. $2 V_T/R = \tan \beta \cdot \zeta_T$.

through a deep tropospheric layer. The convergence in the planetary boundary layer can only account for 10-30% of the total mass convergence in these systems. In addition, the divergence profile of the typical weather system undergoes a large diurnal variation. Referring to Table 2 and Table 4 of Paper I, it is seen that the vertical velocity at 850 mb is at least twice as large in the morning as in the evening. If boundary layer frictional convergence played a dominant role in the maintenance of this vertical velocity, then the low level vorticity would also be expected to vary diurnally. In Table 1 the ratios of the relative vorticities at the two observation times (AM/PM) for the Western Pacific and Western Atlantic data sets are listed. Despite the large diurnal variations in divergence, no consistent diurnal variation is observed in vorticity.

In Fig. 24 the low level and upper level vertical motions for the GATE cluster (data set 11) are plotted as a function of time of day. Also plotted is the 850 mb level relative vorticity. The GATE vorticity does exhibit a diurnal variation but it is out-of-phase with the low level vertical motion. This shows that the changes in vertical motion are not a balanced flow response to changes in vorticity. In fact, the observations indicate that the opposite is true.

To further demonstrate the independence between low level vorticity and divergence in actual weather systems, Table 2 is presented. This table shows the ratio of convergence to vorticity at 900 mb for all observation times for the eleven basic data sets. The lack of uniformity in the table and the large differences between the AM and PM values for each data set clearly show that a direct relationship between boundary layer convergence and vorticity is not present and is not a dominant

TABLE 1

Ratio of morning to evening vorticity in various tropical weather systems. Vorticity is averaged over the $0-3^{\circ}$ area at 900 mb and 700 mb levels. AM times are about 07-10 LT, PM times are about 19-22 LT.

WESTERN PACIFIC	900 mb	700 mb
<u>DATA SETS</u>	<u>ζ AM/ζ PM</u>	<u>ζ AM/ζ PM</u>
1. Cloud Cluster	.8	.6
2. Pretyphoon Cluster	1.2	1.1
3. Storm $P_c > 1000$ mb	1.0	1.1
4. Storm $980 < P_c \leq 1000$ mb	1.1	1.0
5. Storm $P_c \leq 980$ mb	1.0	1.0
WESTERN ATLANTIC		
<u>DATA SETS</u>		
6. Cloud Cluster	.4	.5
7. Easterly Wave	1.4	1.1
8. Storm $V_{\max} \leq 35$ kts	1.0	.9
9. Storm $35 \text{ kts} < V_{\max} < 65$ kts	1.0	1.0
10. Storm $V_{\max} \geq 65$ kts	1.0	1.0

controlling factor to the magnitude of low level convergence in tropical weather systems.

Figure 25 shows the vertical profiles of vorticity and divergence over the GATE A/B-array averaged over the whole GATE experiment. There is large scale cyclonic vorticity across the array, but it is too high in the troposphere to drive the boundary layer convergence. Indeed, in the boundary layer the convergence is an order of magnitude greater than the vorticity, showing that in this region at least the

GATE CLUSTER

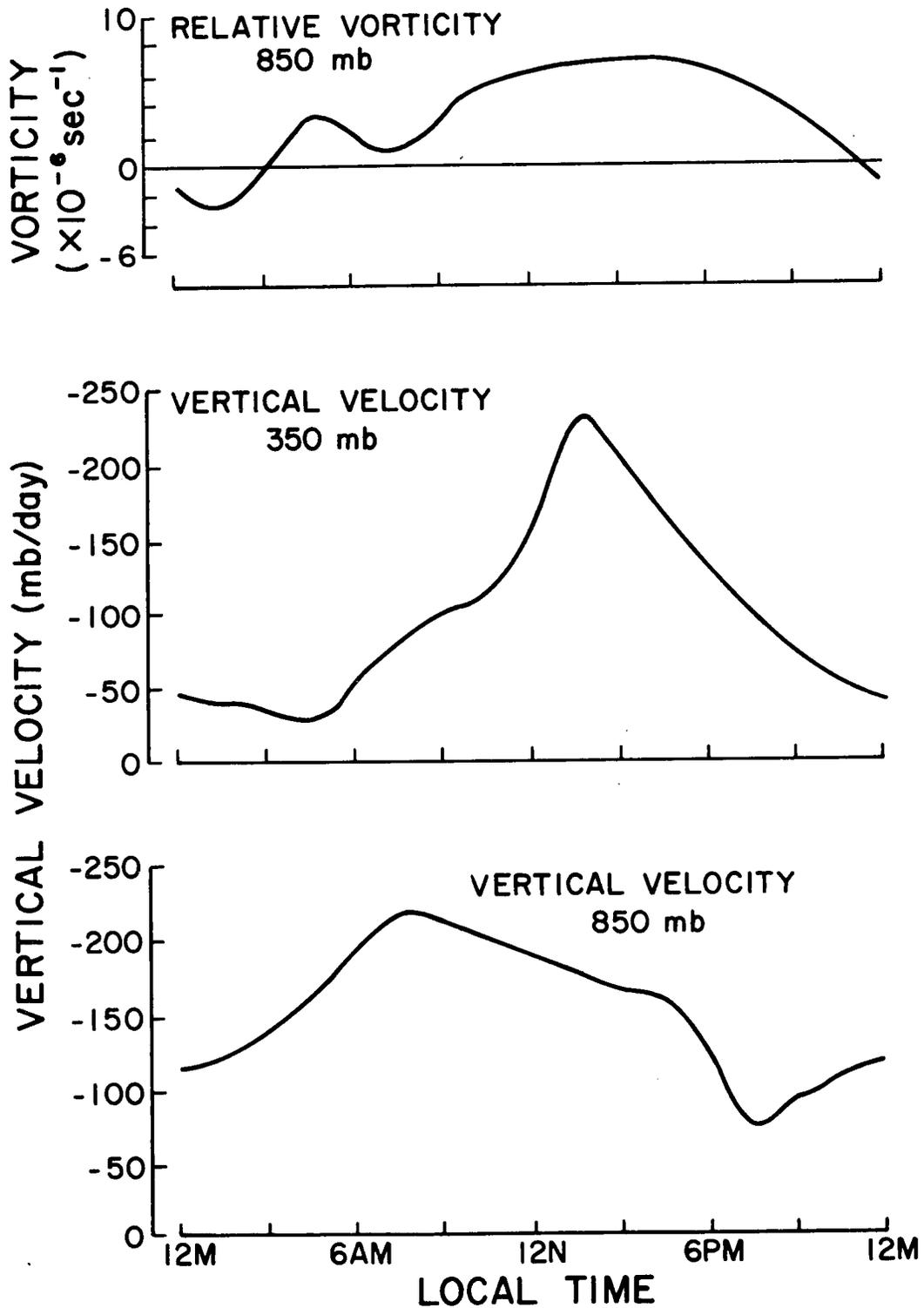


Fig. 24. Diurnal variation of A/B-scale vorticity at 850 mb and vertical motion at 850 mb and 350 mb for the GATE composite cluster.

TABLE 2

Ratio of 900 mb level convergence to vorticity for all of the composite weather systems. AM times are about 07-10 LT and PM times are about 19-22 LT.

PACIFIC

	<u>AM</u>	<u>PM</u>
1. Cloud Cluster	0.9	0.3
2. Pretyphoon Cluster	0.1	-.002
3. Storm $P_c > 1000$ mb	0.4	0.2
4. Storm $980 < P_c \leq 1000$ mb	0.1	0.1
5. Storm $P_c \leq 980$ mb	0.2	0.2

ATLANTIC

6. Cloud Cluster	6.5	-5.8
7. Easterly Wave	0.5	-0.5
8. Storm $V \leq 35$ kts	0.1	-0.1
9. Storm $35 < V < 65$ kts	0.1	.03
10. Storm $V \geq 65$ kts	0.2	0.2

GATE

<u>CLUSTER</u>	<u>1:30AM</u>	<u>4:30AM</u>	<u>7:30AM</u>	<u>10:30AM</u>	<u>1:30PM</u>	<u>4:30PM</u>	<u>7:30PM</u>	<u>10:30PM</u>
	-0.9	6.9	12.4	0.9	1.9	0.6	0.2	-5.6

quasi-geostrophic assumption and a frictionally driven ITCZ model are invalid.

Background or long term low level vorticity was also calculated for the three regions and compared with the mean low level background divergence. Results are shown in Table 3. Comparing the two ITCZ

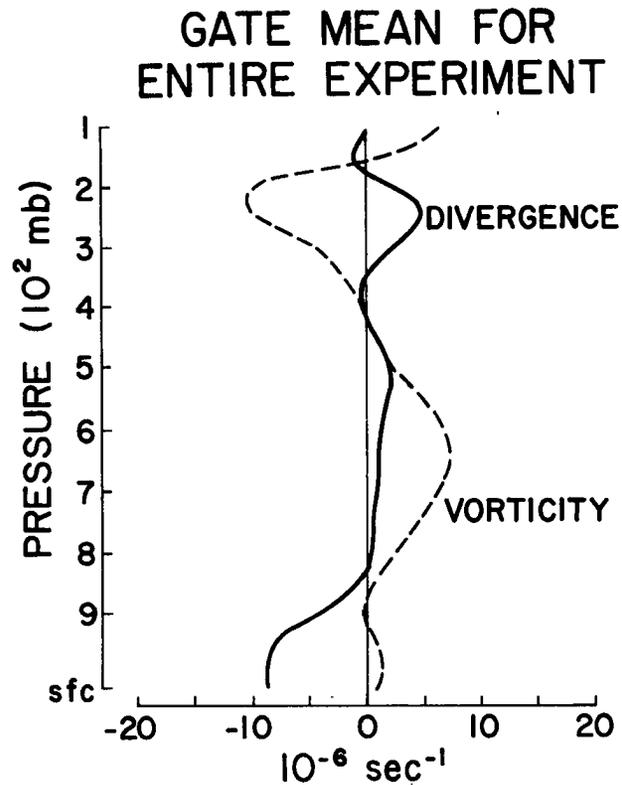


Fig. 25. Mean divergence and vorticity over the A/B-scale for the whole 60 days of the GATE experiment.

TABLE 3

Background or 'long term mean' vorticity and vertical motion at 850 mb.

	ζ (850 mb)	ω (850 mb)	$-\zeta/\omega$
	$(10^{-6} \text{ sec}^{-1})$	(mb/day)	(Arbitrary units)
Western Pacific: Background	+3	-30	1.0
Western Atlantic: Background	-1.5	+10	1.5
GATE Mean for Whole Experiment	+0.5	-80	0.1

regions it is seen that the West Pacific has much stronger vorticity than the GATE region but less low-level vertical motion. This shows that the strength of the ITCZ mean mass convergence and vertical motion bears little or no relation to the mean vorticity. This is further brought out in the last column of Table 3, which shows the ratio of background low level vorticity to vertical motion for each region. No direct relationship between mean low level vorticity and divergence is apparent.

5. CONVECTIVE FEEDBACK

Four large scale controls of convection have been discussed: the cluster scale differential radiation mechanism as described in Paper I, ITCZ convergence, easterly waves and boundary layer frictional convergence. At any given time, each of these controls is contributing to the observed divergence of the tropical weather system or cloud cluster. The cumulus convective elements themselves have a much smaller horizontal scale than any of the controls so far discussed but once released, they further accenuate the convective processes through the release of their condensation heat. These convective feedbacks can act as an additional source of mass, energy and momentum to the large scale. Thus, a portion of the large scale vertical motion field in weather systems is likely not a direct result of the large scale forcing mechanism alone but rather the indirect result of the cumulus convective feedback processes.

The classical convective feedback is release of latent heating in the cumulus clouds. This is the major feedback mechanism and is probably responsible for most of the observed cluster vertical velocity above that attributable to the large scale forcing mechanisms. It should be pointed out, however, that only a very small fraction of the released latent heat is realized within the cluster. The profiles of divergence presented in Paper I for all systems show import of mass in the lower troposphere where the dry static energy is significantly less than in the upper troposphere where the export of mass takes place. This leads to a net export of static energy which almost exactly balances the latent heat energy released as rainfall.

The actual mesoscale mechanics of the feedback are of interest. Modelling studies (Lopez, 1973) and observational studies (Matsumoto *et al.*, 1967; Zipser and Gautier, 1978) have shown that to overcome the low level gravitational stability of the tropical atmosphere, the individual cumulus elements require under cloud base a convergence of the order of $5 \times 10^{-3} \text{ s}^{-1}$, three orders of magnitude larger than the cluster scale convergence. It has been demonstrated by Lopez (1973) and Purdom (1976) that this large required convergence is sustained by the downdraft air from other cumulus cells penetrating the boundary layer from upper levels. The convection thus self-enhances, new cells growing from the downdrafts of old cells.

Besides acting as a source of heat for the large scale, the cumulonimbus clouds also vertically transport momentum and kinetic energy. This momentum injection into the large scale at the outflow and downdraft levels of detrainment represents another mode of convective feedback.

A third mechanism is relative radiational heating. As discussed extensively in Paper I the cloud covered areas with thick cirrus shields radiatively cool significantly less than the surrounding clear atmosphere. In a relative sense then, the convective areas are being warmed in comparison with the other regions. This brings about additional large scale convergence.

The fourth major form of convective feedback is that whereby the cumulonimbus elements become so well organized that the cloud-scale dynamics take over and become rather independent of the larger scale forcing which initialized the convection. An example of this process is found in the GATE cluster composite. GATE is a region of strong

low-level vertical shear. As the convection develops in the presence of shear, it often becomes organized into squall lines. The low level vertical velocity driving the convection is maximum at 7:30 in the morning (Fig. 24). Once the convection has developed, the internal dynamics of the squall lines take over and the actual maximum of convection and of mid-tropospheric vertical motion is not until early afternoon. This squall convection can continue into the early evening when large-scale forcing has become quite weak.

6. RELATIVE MAGNITUDE OF VARIOUS FORCING FUNCTIONS

A summary of lower and upper tropospheric vertical motion for all composited weather systems is given in Tables 4 and 5. Table 4 shows the mean vertical motion at 850 mb and 350 mb for the ten composite weather systems in the Western Pacific and in the Western Atlantic. Also shown are the AM (\sim 07-10 LT) and PM (\sim 19-22 LT) vertical motion values. Table 5 gives the same information for the GATE composites. Large variations in vertical motion occur at both levels between the various weather systems. Of the same magnitude as these variations and apparent in every data set are the large diurnal differences.

The total observed vertical velocities in Tables 4 and 5 are the result of the interaction between many different large scale forcing functions and convective feedbacks. To gain an appreciation of the relative effects of each forcing function, the typical magnitude of the contribution of each mechanism to the lower tropospheric vertical motion in each region will now be estimated. Weather system total vertical motion might be thought of as equal to the sum of the

- (A) Contribution of ITCZ forcing, as described in section 2.
- + (B) Contribution due to the diurnal modulation of the ITCZ convergence, as described for GATE in section 2a.
- + (C) Modulation of the vertical motion due to the passage of middle tropospheric easterly waves, as described in section 3.
- + (D) Contribution due to frictional convergence, as described in section 4.
- + (E) Diurnal modulation of vertical motion forced by cluster-scale differences in radiative-convective heating between deep convective cloud clusters and surrounding subsidence areas.

TABLE 4

Mean measured vertical motion inside $r = 3^\circ$ radius and AM (07-10 LT) and PM (19-22 LT) values at two tropospheric levels for the composite western ocean weather systems. Values in mb/d.

	<u>At 850 mb</u>			<u>At 350 mb</u>		
	<u>Mean ω</u>	<u>AM</u>	<u>PM</u>	<u>Mean ω</u>	<u>AM</u>	<u>PM</u>
<u>WESTERN PACIFIC</u>						
1. Cloud Cluster	-76	-138	-14	-164	-210	-118
2. Pretyphoon Cluster	-62	-68	-56	-177	-201	-153
3. Storm $P_c > 1000$ mb	-124	-172	-76	-182	-237	-127
4. Storm $980 < P_c < 1000$ mb	-135	-148	-122	-146	-163	-129
5. Storm $P_c \leq 980$ mb	-251	-242	-260	-358	-389	-327
<u>WESTERN ATLANTIC</u>						
6. Cloud Cluster	+17	-26	+60	+29	-28	+86
7. Easterly Wave	-14	-56	+28	-54	-75	-33
8. Storm $V \leq 35$ kts	-40	-84	+4	-128	-170	-86
9. Storm $35 \text{ kts} < V < 65$ kts	-88	-115	-61	-140	-160	-120
10. Storm $V \leq 65$ kts	-201	-206	-196	-324	-375	-273

+ (F) Contribution due to cumulus convective feedback, or

$$\omega(\text{low level}) = A + B + C + D + E + F \quad (2)$$

From the magnitudes of vertical motion presented in Tables 4 and 5 and other knowledge as discussed in this paper, one is able to estimate the contribution of each term in the above equation for the typical convective system (i.e. for a cloud cluster) in each of the three regions, Western Pacific, Western Atlantic and GATE.

Term (A) - ITCZ Forcing. For the Western Pacific and Western Atlantic this term is assumed to be equal to the 850 mb level vertical velocity obtained from the long period summertime average or background composites described in section 2. This is slightly negative in the Western Atlantic (+10 mb/d), but upwards at about -30 mb/d in the Pacific. In GATE it is equal to the 850 mb level vertical velocity averaged over the A/B-array for the entire experiment, and is large, approximately 80 mb/d.

Term (B) - Diurnal Modulation of ITCZ Forcing. In the background composites for the western oceans this influence appeared to be small. It was measured to be about zero in the Western Pacific (Fig. 12) and only about ± 5 mb/d in the Western Atlantic (Fig. 16). For the GATE A/B-scale it is quite significant and is considered to be equal to the observed diurnal variation for the suppressed conditions composite, about ± 45 mb/d.

Term (C) - Wave Forcing. This is taken to be equal to the difference between the observed 850 mb level vertical velocity between the A/B-scale GATE trough and ridge as tabulated in Table 5, i.e. ± 30 mb/d. The GATE region is immediately downstream of the area where African waves are generated; thus the wave forcing is stronger here than in the

TABLE 5

Mean measured vertical motion ($\bar{\omega}$) and diurnal values for the GATE A/B-array composite systems at 850 and 350 mb. Values in mb/d.

	Mean $\bar{\omega}$	<u>At 850 mb</u>							
		<u>Value at diurnal time periods</u>							
		1:30 AM	4:30 AM	7:30 AM	10:30 AM	1:30 PM	4:30 PM	7:30 PM	10:30 PM
Cluster	-154	-121	-165	-220	-203	-176	-163	-72	-112
Suppressed	-16	+16	+3	-75	-36	-22	-4	+6	-16
Wave Position 1	-78		-87		-104		-66		-55
Wave Trough	-98		-99		-110		-115		-67
Wave Position 2	-69		-53		-104		-78		-40
Wave Ridge	-39		-34		-61		-50		-12
		<u>At 350 mb</u>							
Cluster	-99	-41	-31	-86	-114	-199	-164	-101	-53
Suppressed	+42	+48	+68	+8	+10	+55	+52	+75	+20
Wave Position 1	-26		+5		-34		-48		-27
Wave Trough	-67		-65		-68		-100		-34
Wave Position 2	-36		-19		-64		-68		+7
Wave Ridge	-4		+8		-30		-16		+21

western oceans. Reed, Recker and Norquist (1977) composited easterly waves in GATE and found a perturbation in the 700 mb level meridional wind field of ± 5.5 m/s between trough and ridge. Reed and Recker (1971) composited Western Pacific waves and found a perturbation of ± 3.5 m/s; the wavelength also was larger. On the right hand side of Fig. 5 of Paper I of the present study it is seen that in the Western Atlantic the perturbation is only $\pm 1.5-2.0$ m/s. Scaling the wave vertical velocity forcing in a linear relationship to the meridional wind perturbation the following magnitudes are obtained: GATE, ± 30 mb/d; Western Pacific, ± 20 mb/d; Western Atlantic, ± 15 mb/d.

Term (D) - Frictional Convergence. This is calculated using Eq.1 of section 4. Inserting mean vorticities from the cloud cluster data sets (data sets 1, 6, 7, 11) into that equation it is proportional to the low-level relative vorticity in these systems and is seen to be small, about -7, -4, -3 mb/d in the Western Pacific, Western Atlantic, and GATE regions respectively. Thus, for a typical oceanic weather system, it seems that boundary layer frictional convergence plays very little role in the forcing and maintenance of the system.

Term (E) - Cluster Scale Diurnal Modulation. This is the difference between the observed diurnal variation in cloud cluster systems as tabulated in Tables 4 and 5 and the contribution of the ITCZ diurnal variation or term (B). In magnitude it is observed to be about ± 45 mb/d.

Term (F) - Convective Feedback. For cloud clusters of the western oceans and GATE A/B-scale region the addition of terms (A) to (E) is insufficient to account for all of the measured vertical velocity. The difference from these measurements will be considered to be the

contribution due to the cloud cluster convective feedback.

The feedback is enhanced by squall lines. The Western Pacific has few squalls except with tropical cyclones, the Western Atlantic has more but not as many as occurred in the GATE region. This convective feedback will thus be taken to be 20, 25 and 40 mb/d in the three regions respectively.

The results of the above calculations for each region are tabulated in Table 6. Adding together the contributions from each term the total magnitude of the lower tropospheric vertical velocity for enhanced and suppressed conditions in each ocean can be approximated as in Table 7. Tables 6 and 7 both show the large role of the diurnal modulation of vertical motion and that wave induced vertical motion is not the dominant mode of large-scale forcing for the tropical weather system. Figure 26 schematically represents the results of Tables 6 and 7. It shows how the actual observed vertical motions of the various region weather systems are a combination of forcing mechanisms and do not result from any individual large-scale control acting by itself.

There are a number of important findings embodied in the results of Tables 6 and 7 and Fig. 26. The first is the major contribution to the net convergence of a tropical weather system in the GATE and Western Pacific area that comes from the ITCZ forcing. Secondly, the overwhelming role played by the diurnal variation must be considered. As discussed in Paper I and in section 2 of this paper, this variation appears to be forced by the differences in radiative-convective heating between an area of deep convection and its surrounding areas of tropospheric subsidence. This contribution to the dynamics of oceanic tropical weather systems has largely been ignored in the past. Its full ramifications

TABLE 6

Estimated typical magnitude of various large-scale 850 mb vertical motion forcing components in mb/d.

		(A)	(B)	(C)	(D)	(E)	(F)
$\left(\begin{array}{l} \text{Magnitude of Lower} \\ \text{Tropospheric Vertical} \\ \text{(\sim 850 mb) Motion in} \\ \text{a Convective Weather} \\ \text{System} \end{array} \right)$		$= \left(\begin{array}{l} \text{ITCZ} \\ \text{Forcing} \end{array} \right) +$	$\left(\begin{array}{l} \text{Diurnal} \\ \text{Modulation} \\ \text{of ITCZ} \\ \text{Forcing} \end{array} \right) +$	$\left(\begin{array}{l} \text{Easterly} \\ \text{Wave} \\ \text{Forcing} \end{array} \right) +$	$\left(\begin{array}{l} \text{Frictional} \\ \text{Convergence} \end{array} \right) +$	$\left(\begin{array}{l} \text{Cluster} \\ \text{Scale} \\ \text{Diurnal} \\ \text{Modulation} \end{array} \right) +$	$\left(\begin{array}{l} \text{Convective} \\ \text{Feedback} \\ \text{in Cluster} \\ \text{Region} \end{array} \right)$
Region	Term	(1)	(2)	(3)	(4)	(5)	(6)
			AM/PM		Trough/Ridge	AM/PM	
Western Pacific		-30 mb/d	0 mb/d	$\bar{+}20$ mb/d	-7 mb/d	$\bar{+}60$ mb/d	-20 mb/d
Western Atlantic		+10 mb/d	$\bar{+}5$ mb/d	$\bar{+}15$ mb/d	-4 mb/d	$\bar{+}40$ mb/d	-25 mb/d
GATE		-80 mb/d	$\bar{+}45$ mb/d	$\bar{+}30$ mb/d	-3 mb/d	$\bar{+}30$ mb/d	-40 mb/d

TABLE 7

Relative magnitudes of lower tropospheric (850 mb) upward vertical motion at AM (07-10 LT) and PM (19-22 LT) times for various weather systems. The values were obtained by summing the contributions of terms (A) to (F) of Table 5 in mb/d.

a) Enhanced conditions of wave trough at AM and PM.

	<u>AM</u>	<u>PM</u>	<u>Daily Mean</u>
Western Pacific	-137	-17	77 mb/d upwards
Western Atlantic	-69	+1	34 mb/d upwards
GATE A/B-array	-228	-78	153 mb/d upwards

b) Suppressed conditions of wave ridge. The contributions from the cluster diurnal (E) and feedback modulation (F) are obviously not included.

	<u>AM</u>	<u>PM</u>	<u>Daily Mean</u>
Western Pacific	-10	-10	10 mb/d upwards
Western Atlantic	+30	+20	25 mb/d downwards
GATE A/B-array	-95	-5	50 mb/d upwards

in various areas of tropical research such as cumulus parameterization studies and tropical storm development have yet to be explored.

Another question that need be resolved is the role of midtroposphere easterly waves as a large scale forcing mechanism. Figure 18 shows the strong modulating of the GATE rainfall which these waves can make. GATE is the region with strongest waves and the strongest wave rainfall modulation. However, as seen in Figs. 17a, b and c GATE rainfall can occur without significant wave passage. Conversely, when wave passage occurs

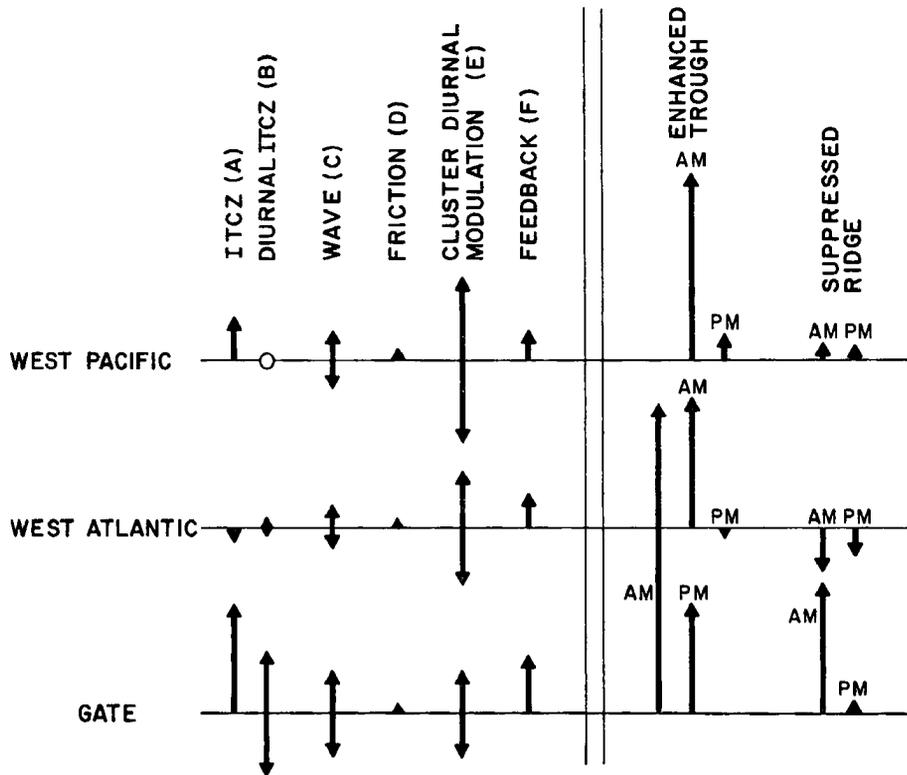


Fig. 26. Schematic representation of the relative magnitudes of the large scale forcing mechanisms on low level vertical motion in the three regions. Arrow lengths are drawn in proportion to the relative magnitudes. The right hand side depicts the magnitude of low level vertical velocity for various weather systems obtained by summing together the contributions from the separate forcing mechanisms.

outside the ITCZ region little rainfall results. Observational studies in other oceanic regions where wave forcing is weaker than in GATE have indicated a weaker wave modulation to convection. Very heavy rainfall occurs in some low latitude oceanic regions without apparent wave passage. The role of easterly waves, therefore appears to be one more of modulation of convection rather than one of cause of convection. Thus, it is observed by N. Frank (1975) that in the Western Atlantic there is a remarkable stability from year to year in the number of easterly waves present in that ocean, yet there can be a significant

variation in West Indies summer rainfall and very large interannual variation in the amount of tropical cyclone activity.

The fourth item worthy of comment in Fig. 26 is the very minor role played by frictional convergence.

The interaction between the various large scale forcing functions can be idealized as shown in Fig. 27. The ITCZ forcing with its diurnal modulation is shown in the top curve. The wave forcing by itself may be visualized by the second curve of this figure. If all three of these mechanisms were to act together then significant low level convergence would be established and cloud cluster growth and internal convective feedback could take place. This feedback is denoted by the dotted area. The dashed line denotes the magnitude of the upward vertical motion necessary to activate a meso-scale radiation and condensation cloud feedback.

Once this cloud cluster growth takes place the cluster scale differential radiative-convective heating forcing mechanism and the condensation forcing mechanism further act to feedback upon the cluster system and significantly enhance the vertical motion. This feedback acts to more strongly concentrate the vertical motion within the cluster region and cause a smaller scale area of vertical motion in comparison with the scale of the wave induced motion. The bottom curve of Fig. 27 combines all four mechanisms. This is an idealized picture. The relative magnitude of each of these forcing mechanisms can be significantly different in the various tropical regions. For example, in the trade wind region of the Western Atlantic the ITCZ forcing is observed to be negative. In the other two regions it is positive.

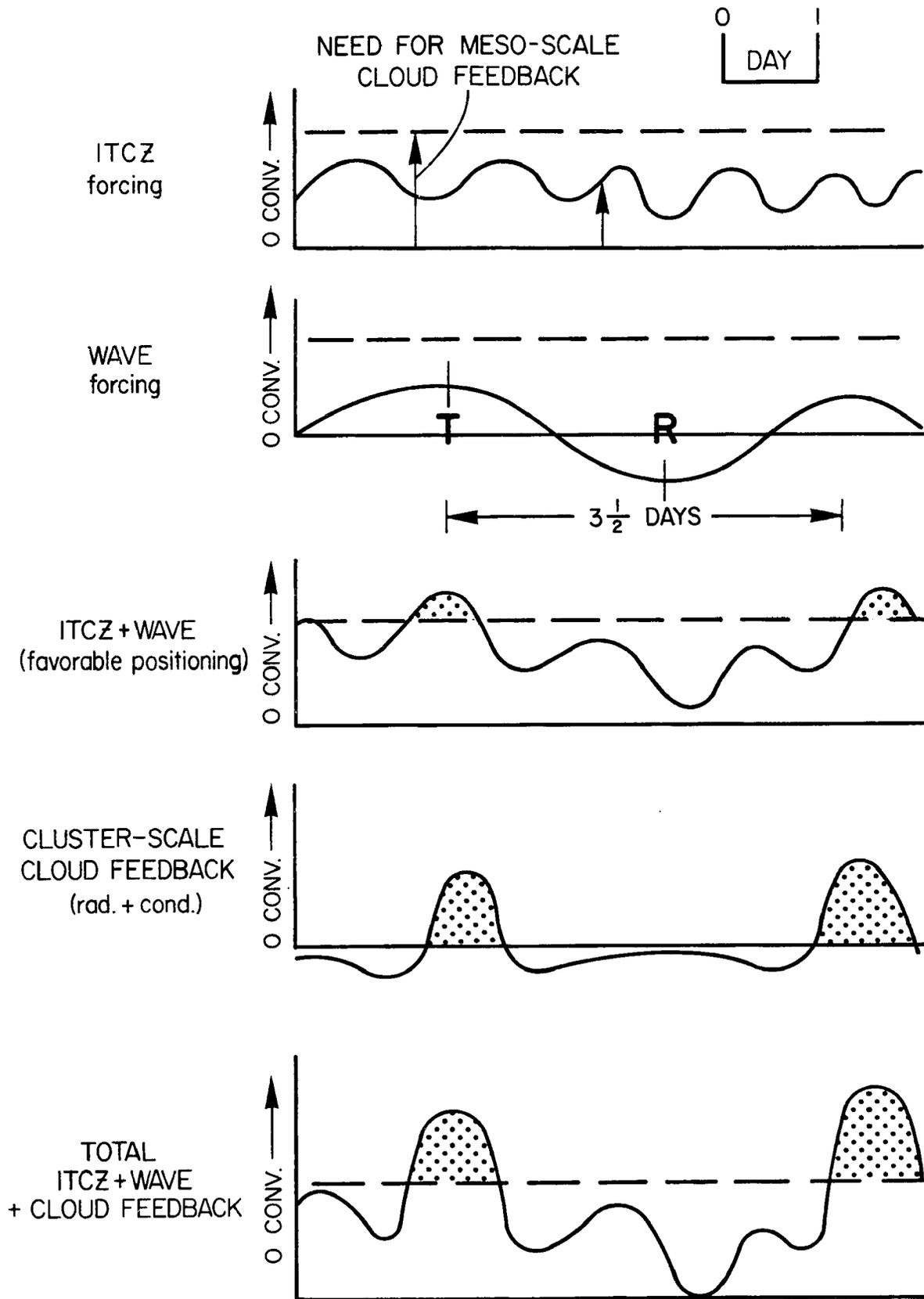


Fig. 27. Idealized association of ITCZ vertical motion forcing with diurnal cycle (top curve) and easterly wave modulation (second curve). Third curve is sum of top two curves. Dotted region denotes required upward vertical motion such that cloud region feedback will occur. Second from bottom curve denotes cloud region feedback. Bottom curve denotes the sum of ITCZ, diurnal, wave and feedback forcing.

It is important that any idealized picture of tropical weather systems be broad and flexible enough to account for the large regional and forcing mechanism differences which exist. A simple 'easterly wave' or 'ITCZ' model of organized oceanic tropical convection is quite inadequate.

7. DISCUSSION

Various authors in recent years (e.g. Stevens et al., 1977; Chang, 1976; Stark, 1976) have formulated mathematical models designed to simulate tropical oceanic convection. To the author's knowledge none of these models realistically include the large scale ITCZ forcing. This should be done. Sadler (1964, 1975), Ramage (1971) and others have stressed the relationship of cloud clusters to the ITCZ and the likely dominant role of the ITCZ in explaining a high percent of tropical rainfall and tropical cyclones. It is time the large role of the ITCZ be quantitatively stressed.

The differences between the heat balance of areas of deep convection and areas where radiative cooling is balanced by broad scale subsident warming appear to play a major role in maintaining and diurnally modulating convection on the cloud cluster and Hadley cell scales of circulation. This has been documented in the past, for the cloud cluster scale, by Gray and Jacobson (1977), Fingerhut (1978) and in Paper I of this study. The current study and Dewart (1978) have demonstrated a large diurnal variation of the low level convergence into the whole ITCZ area in the GATE region.

Riehl (1945), Palmer (1952) and Reed and Recker (1971) have demonstrated that tropical oceanic convection is influenced by the passage of easterly waves in the middle level flow fields. A major result of this paper has been to show that the easterly wave forcing is only one of a number of primary large-scale forcing mechanisms. Tropical convection cannot likely be well understood until all of the large-scale forcing mechanisms here discussed are treated in unison.

The term Conditional Instability of the Second Kind (CISK) has been proposed by Charney and Eliassen (1964) to describe a mechanism by which boundary layer friction convergence acts to force latent heat released in cumulus convection, and feedback to intensify large scale circulations. In the original CISK models the large scale convergence of moisture was brought about by frictional effects in the boundary layer. More recently Yamasaki (1969) proposed a second type of CISK instability, called Wave-CISK by Lindzen (1974). Wave-CISK describes the growth or maintenance of a tropical easterly wave. In this model, the low level convergence of moisture rather than being frictionally driven is assumed to be part of the convergence field associated with the wave dynamics. Other non-frictional modes of CISK have been proposed by Bates (1973) and by Yamasaki (1975, 1977a, b). In these models the low level convergence is associated with gravity waves propagating away from the point of release of latent heat. There is also a frictional aspect to the ITCZ forcing. Charney (1968) and Bates (1970) have proposed that the Hadley Circulation can be thought of as being a balanced flow response to the planetary scale radiative imbalance with low level friction producing the equatorward flow. It is noteworthy, though, that Fig. 12 reveals that in the Western Pacific the ITCZ convergence extends from the surface up to the 400 mb level, well above the level of frictional influence. In addition, mean boundary layer convergence into the GATE A/B region is 3-5 times larger than the mean vorticity. Figure 25 implies a large wind pressure imbalance and down-gradient flow convergence which cannot be explained by any type of boundary layer frictional veering or by easterly wave induced convergence.

The observations of this paper indicate that boundary layer frictional convergence plays only a minor role in the large-scale forcing of organized convection. In addition, the boundary layer convergence associated with easterly waves appears also to be quite small. Thus, the important roles previously ascribed to the physical process associated with the so called "CISK" and "Wave-CISK" hypotheses appear to overly exaggerate the role which surface and wave-induced low level convergence play in tropical weather system maintenance.

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