Analysis of satellite-observed tropical cloud clusters.

I. Wind and dynamic fields

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ABSTRACT

Composited Northern Hemispheric Data Tabulation rawinsonde data has been gathered relative to three classes of satellite-observed tropical weather systems in the Western Pacific (Eq.–18° N) and in the West Indies ($\sim$15–30° N). These systems include the $\sim$4° wide cloud cluster, the cloud cluster environment ($\sim$2–6° latitude around the cluster center), and the clear region. Information is presented for the years 1967 and 1968. The data gathering philosophy and methodology follows that discussed in a previous report by Williams & Gray (1973).

Composited information is presented on individual weather system winds, divergence, vorticity, etc. Emphasis is put on the regional differences between the Western Pacific and the West Indies weather systems. Significant differences exist. The mean cloud cluster vorticity budget is calculated. The horizontal divergence of the vorticity transport is the predominant term in the budget in both regions. It is necessary to hypothesize a large eddy transport of vorticity from the lower to the upper troposphere. Other observational findings are also presented.

1. Introduction

This is the first of two papers on tropical cloud clusters. This paper discusses the wind and dynamic fields of the cluster. A following paper discusses the cloud cluster thermal and moisture fields.

Satellite meteorology was truly initiated in 1966 with the initial availability of the U.S. National Environmental Satellite Service global daily digitized pictures. The more recent synchronous satellite launches have also greatly expanded our tropical data coverage and research potential. Before these advances in satellite technology can be fully exploited, a few years of picture and rawinsonde compositing, processing, and digesting are required. Intensive statistical stratification of this new information is now possible.

Of particular importance has been the satellite picture observation and documentation of tropical weather systems. Attention has focused on the prominence of the 4° wide cloud cluster (see report on the first session of the JOC Study Group on Tropical Disturbances, Appendix 1, GARP Publ. Series no. 4, also see the paper by Hayden, 1970). These clusters are of growing interest because of their hypothesized role in the general circulation.

This very large increased satellite coverage by itself, however, will not significantly advance our basic understanding of tropical atmospheric dynamics unless the new cloud picture data is properly associated with the flow and thermodynamic patterns in which the cloud fields are embedded. Usually, only small amounts of upper air data are available to go with individual picture information. The authors feel that more involved longterm averaging of the satellite data in terms of the other meteorological parameters would significantly add to our knowledge of these systems.

Most previous tropical weather system investigations have been accomplished without benefit of satellite pictures. The present research represents a new approach: the use of both conventional rawinsonde data and satel-

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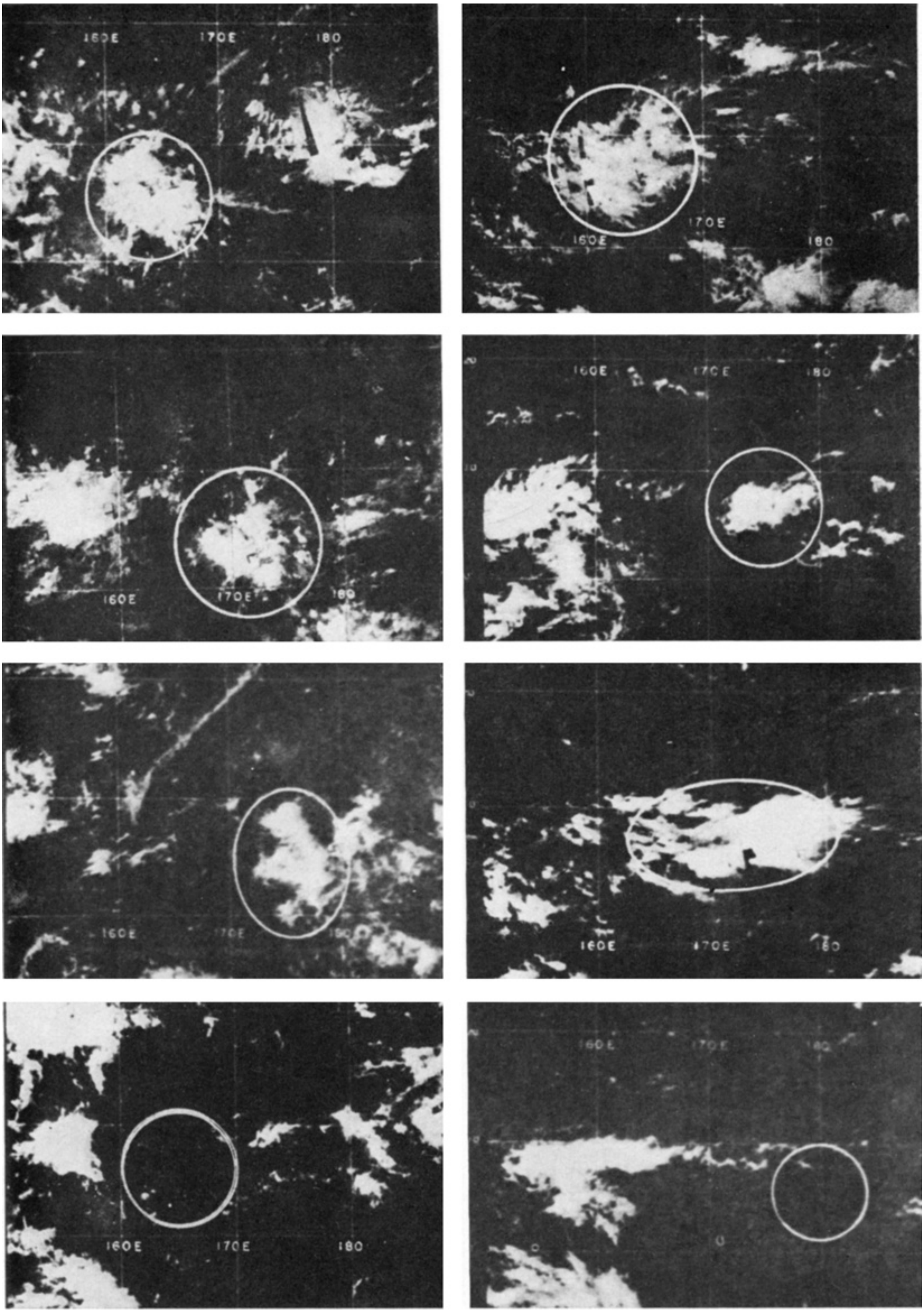


Fig. 1. Typical portrayal of satellite-observed cloud clusters and clear regions.

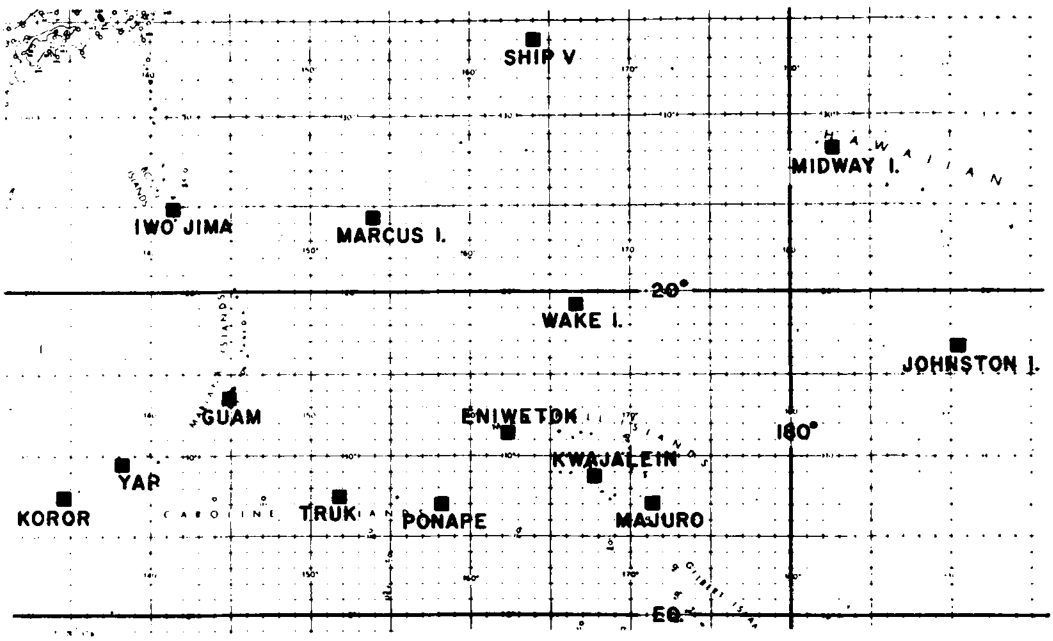


Fig. 2. Radiosonde network in the Western North Pacific Ocean.

lite data simultaneously in a large statistical survey of individual specified cloud systems.

The analysis of tropical weather systems has generally followed one of two methodological paths:

1) Analysis of tropical disturbances as a form of wave phenomena and investigation of wavelength, frequency, dynamic properties, etc. (Chang, 1970; Nitta, 1970; Reed & Recker, 1971; Yanai et al., 1968). A general summary of this research approach has been given by Wallace (1971).

2) Looking at the disturbances as a phenomenon not necessarily associated with a wave and investigating their *in situ* characteristics (Frank & Johnson, 1969; Riehl & Pearce, 1968; Simpson et al., 1967; Williams & Gray, 1973; Zipser, 1969).

This paper follows the latter point of view. It is an extension of the previous tropical analysis research of Williams & Gray (1973), Gray (1973), and Lopez (1973). The purpose of this paper is (1) to present mean data for the cloud cluster; emphasis is placed on the regional differences between Western Pacific and West Indies cloud cluster; (2) to investigate the vorticity budget of the cloud clusters and understand the importance of the up-and-down motions within these tropical disturbances. All available data from routine rawinsonde observations, surface observations (rainfall), and satellite photographs were gathered, combined, and integrated.

2. Data and method of analysis

2.1. *Satellite appearance of tropical cloud clusters*

The tropical cloud cluster appears in satellite photographs as a bright, solid white blob. This is due to the large cirrus canopy typically associated with the cluster. These cirrus canopies are produced by outflow from and remnants of cumulonimbi. Sometimes the cloud clusters are largely composed of middle-layer cloudiness (Sadler, 1962). Flying through the satellite-observed clusters, one sometimes observes very few cumulonimbi. In most clusters, large numbers of Cb's are believed to be present. Fig. 1 shows typical satellite appearances of the tropical clusters we will be discussing. The last two diagrams show examples of clear regions.

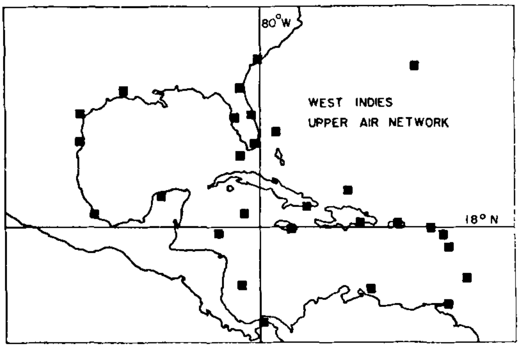


Fig. 3. Radiosonde network in the West Indies.

Table 1

No. of weather systems	Western Pacific (equator–18° N, 130° E–165° W) Nov. 1966–Oct. 1968)	West Indies (18°–30° N, 50° W–95° W) Jan. 1967–Dec. 1968
Cloud clusters	1 096	245
Cluster environments	1 096	245
Clear regions	267	182
Total number of rawinsondes	5 690	5 035

2.2. *Data sources*

Figs. 2 and 3 show the rawinsonde data networks which are used in our analysis. From a geographical and climatological viewpoint, these regions have significant differences. The Western Pacific area is far from a continent. The rawinsonde stations are typically located on small islands and atolls. Cloud clusters are often associated with the ITCZ. In the West Indies half the radiosonde stations are located on a continent and 4 on large islands. West Indies clusters are not associated with the ITCZ. The Appendix gives more information on the locations of the individual weather systems relative to these upper-air stations.

2.3. *Compositing philosophy*

Due to the space and time resolution of the upper air observations, the flow patterns associated with *individual* satellite-observed weather systems cannot be adequately described. The surrounding observations that are available at regular time intervals are often not representative, the basic physical processes involved with tropical systems can be understood only from a composite study of many systems with similar satellite-observed cloud patterns.

The rawinsonde data was composited relative to the center of the satellite-observed tropical clusters and clear regions in a way previously described in the paper by Williams & Gray (1973). Table 1 gives information about the numbers of each weather system treated and the number of rawinsondes used. In order to determine the regional differences between the cloud clusters in the deep tropics and those in the subtropics the present investigation is limited in the Western Pacific to weather systems south of 18° N and in the West Indies to systems between 18–30° N.[1] The Appendix shows the locations of all of the cloud clusters and clear regions.

2.4. *Compositing technique*

Following the classification presented by Gray (1973), three different types of typical weather systems are defined and treated separately. These are:

1) The satellite-observed ~4° wide *cloud cluster* (Fig. 1) (as defined by the JOC Study Group on Tropical Disturbances, 1968) in the Western Pacific and in the West Indies.

2) *Cloud Cluster Environment* or variable cloud region defined as the area from 2–6° radius around the cloud cluster center (Fig. 4).

3) *Clear regions* between cloud clusters where no cloudiness is seen on the satellite pictures (bottom diagrams of Fig. 1).

The positions of the centers of the cloud clusters and clear regions were determined from U.S. Weather Bureau ESSA satellite pictures for the two year period of November 1966 – October 1968 for the Western Pacific and January 1967 – December 1968 for the West Indies. All available rawinsonde data are composited with respect to these individual weather systems.

Data have been analyzed from rawinsondes taken at 00Z and 12Z. The daily progression of all observed cloud clusters was westward.

[1] Williams & Gray (1973) included in their Pacific study all clusters and clear areas to 28° latitude. Thus, differences between the results of this study and theirs will occur, especially for the clear regions.

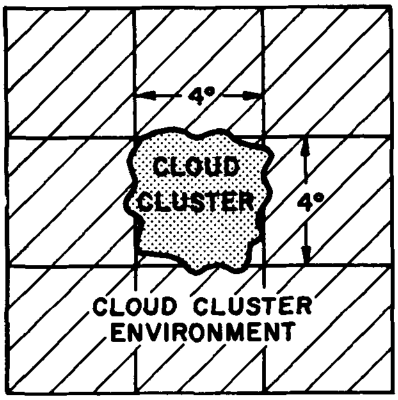


Fig. 4. Portrayal of typical size and relative location of the cloud cluster and the cloud cluster environment. Rawinsonde data has been averaged in each 4° square.

The velocity was assumed to be ~6° longitude per day in accordance with ours and with Wallace (1970) and Chang's observations (1970). Station locations relative to the satellite picture time were adjusted accordingly.

The nearest rawinsonde observations to the satellite pictures (15 local time) were processed. For the Western Pacific stations the 00Z-soundings (~10 local time) were used, for the West Indies the 12Z-soundings (~07 local time) were used. The daytime relative humidities have been corrected for thermistor heating by a scheme given by Ruprecht (1975).

To have as large and as statistically significant a data sample as possible, all information was averaged with respect to the centers of the clusters and clear regions on a 4° latitude-longitude grid. This is similar to the grid averaging previously used by Williams & Gray (*op. cit.*). Thus, our 4° × 4° grid cannot resolve smaller scale phenomena. Fig. 4 portrays this grid resolution.

3. Wind fields associated with Western Pacific and West Indies weather systems

3.1. *Meridional winds*

Riehl (1945) was one of the first to associate concentrated areas of cloudiness and precipitation in the trade winds to wave type disturbances which propagate westward. He developed an easterly wave model with low level convergence and convective activity east of the trough line and subsidence west of it. This model was derived mainly for the West Indies area. A question arises whether the usual ~4° diameter satellite-observed cloud clusters are typically associated with such a wave model and also, how the cluster regions compare with the clear regions, etc. Figs. 5–8 give zonal-height cross-sections of the composited meridional wind for the Western Pacific and West Indies cloud clusters and clear regions.

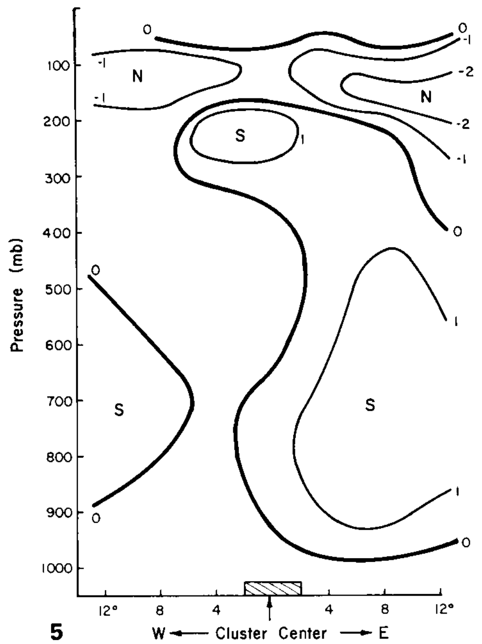
Western Pacific: East of the cloud cluster (Fig. 5) the expected S-winds are observed. However, the corresponding N-wind maximum is hardly in evidence. We might interpret the difference of the maximum of S-winds to the east of the cloud cluster (~1.9 m/sec) and of N-winds to the west of the cluster center as being in general agreement with Riehl's model. Many individual clusters would not fit this mean pattern however. No distinct east–west meridional wind pattern is found in the clear regions (Fig. 6).

West Indies: The distribution of the meridional winds (Fig. 7) shows only that the south winds to the east of the trough are stronger than the south winds to the west of the trough. Only a small west to east shear of meridional wind is observed. Again, this is an average pattern. Many individual clusters may well have fit the Riehl easterly wave model. The West Indies clear regions (Fig. 8) indicate a west to east meridional wind shear in agreement with the ridge region of the Riehl wave model.

The meridional wind distributions do not give a definitive answer to this question. One can, however, conclude that a number of cloud clusters are likely associated with wave disturbances. As discussed by Wallace (1971), spectrum analysis of tropical data indicates a wide range of waves and mutual interactions of tropical disturbances. Our 4° compositing technique could not resolve the smaller scale wave components and our area of coverage precluded resolution of the hemispheric scale wave components.

3.2. *Zonal winds*

The largest differences between the Western Pacific and the West Indies areas occur with the zonal winds. Figs. 9–12 show the large regional differences of the environments in which the cloud clusters are embedded.



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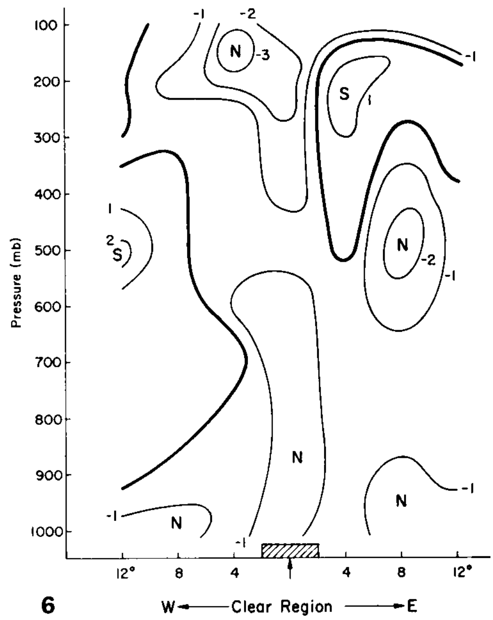
WESTERN PACIFIC CLOUD CLUSTER
MERIDIONAL WIND (m/sec)



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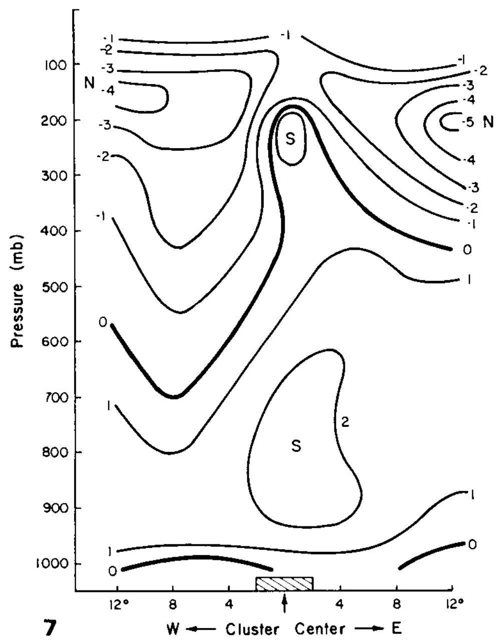
WESTERN PACIFIC CLEAR REGION
MERIDIONAL WIND (m/sec)



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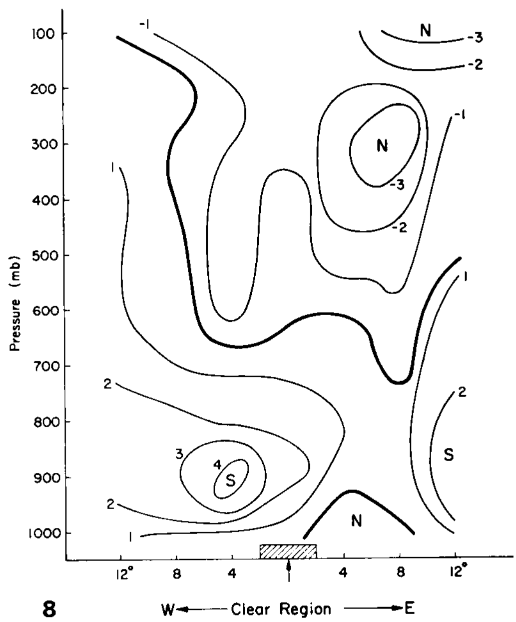
WEST INDIES CLOUD CLUSTER
MERIDIONAL WIND (m/sec)



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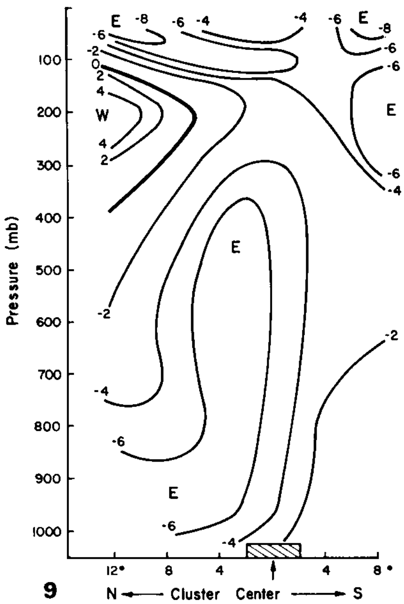
WEST INDIES CLEAR REGION
MERIDIONAL WIND (m/sec)

Figs. 5–8. West–east cross-sections of meridional wind for Western Pacific and West Indies cloud clusters and clear regions. The small hatched rectangular box at the figure bottom defines the 4° cloud cluster or clear area width.



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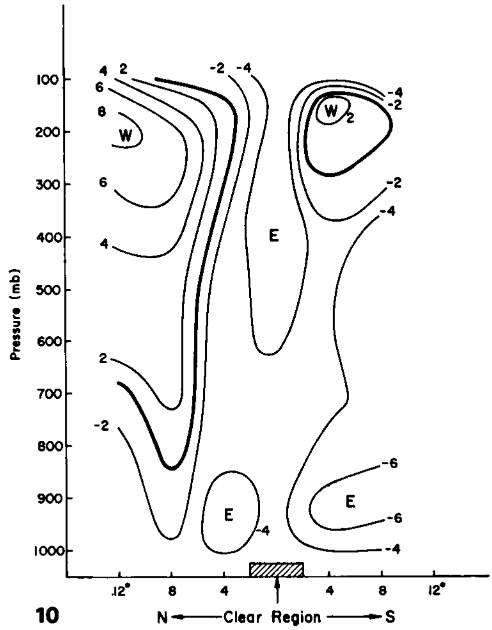
WESTERN PACIFIC CLOUD CLUSTER
ZONAL WIND (m / sec)



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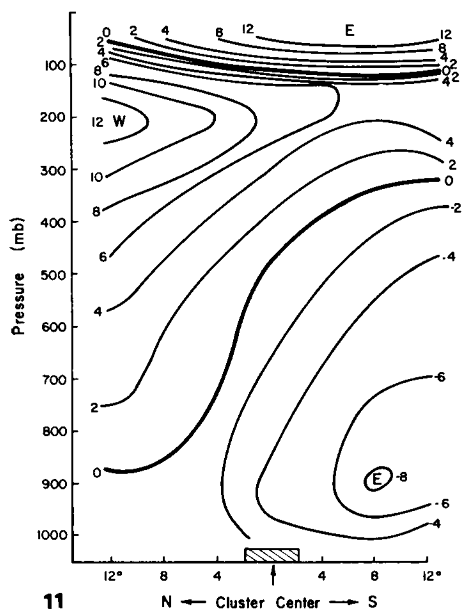
WESTERN PACIFIC CLEAR REGION
ZONAL WIND (m/sec)



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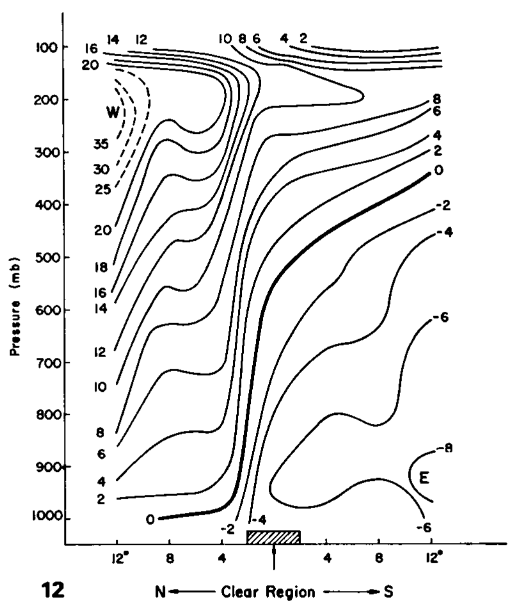
WEST INDIES CLOUD CLUSTER
ZONAL WIND (m / sec)



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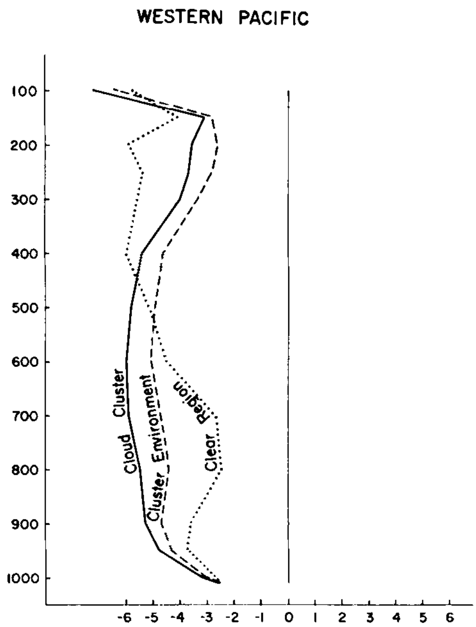
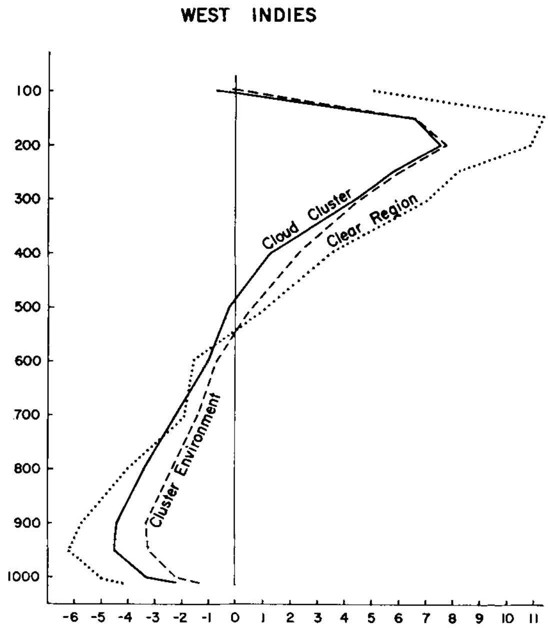
WEST INDIES CLEAR REGION
ZONAL WIND (m/sec)

Figs. 9–12. North-south cross-sections of zonal wind for the Western Pacific and West Indies cloud clusters and clear regions. The small hatched rectangular box at the figure bottom defines the 4° cloud cluster or clear area width.





13 ZONAL WIND (m/sec) 14 ZONAL WIND (m/sec)

Figs. 13–14. Tropospheric distribution of zonal winds by type of weather system.

Western Pacific: The cloud cluster area is entirely within the tropical east-wind regime but the mid-latitude westerlies are present just north of the clear region centers (Figs. 9 and 10).

Although the average positions of the cloud cluster (9° N) and the clear region (11° N) are nearly the same, the trade wind maximum is north of the cloud cluster center (Fig. 9) and slightly south of the clear region center (Fig. 10). These contrasting positive and negative horizontal shearing patterns are believed to be a primary feature in producing the contrasting low level convergence and divergence patterns associated with each system. The changing cloud patterns appear to result from a meridional shift of the trade wind maximum.

West Indies: This region is associated with two different tropospheric wind systems, the tropical easterlies and the mid-latitude westerlies. In contrast to the Western Pacific the trade wind maximum is located well to the south of the cluster center. Both cluster and clear systems exist in a broad-scale anticyclonic north–south shear pattern (Figs. 11 and 12).

Vertical Shear of Zonal Wind. A striking feature in the comparison of the zonal wind distributions is the change of zonal wind with height. The Pacific mean cluster shows zonal wind strengths are nearly constant up to 400 mb. North of the cluster center the vertical gradient of the zonal wind is positive; south of the cluster it is negative. This indicates that the cluster is located at the latitude of warmest surface to 400 mb temperature. The vertical gradient of the zonal wind for the clear areas is always positive except for the center where it is nearly constant. This pattern indicates that the warmest areas are south of the clear region. In the West Indies the zonal vertical shear is much larger and always positive. Warmest tropospheric temperatures are found far to the south.

Figs. 13 and 14 portray the vertical shear of the zonal wind for each weather system in both regions. Significantly more vertical shear is present in the West Indies. As the type and degree of organization of deep cumulus convection is dependent on the magnitude of tropospheric vertical shear (Ludlam, 1966), one would expect more organization of deep convection in the West Indies. Surface reports of cluster squall line activity indicate a greater prevalence in the West Indies. It appears that

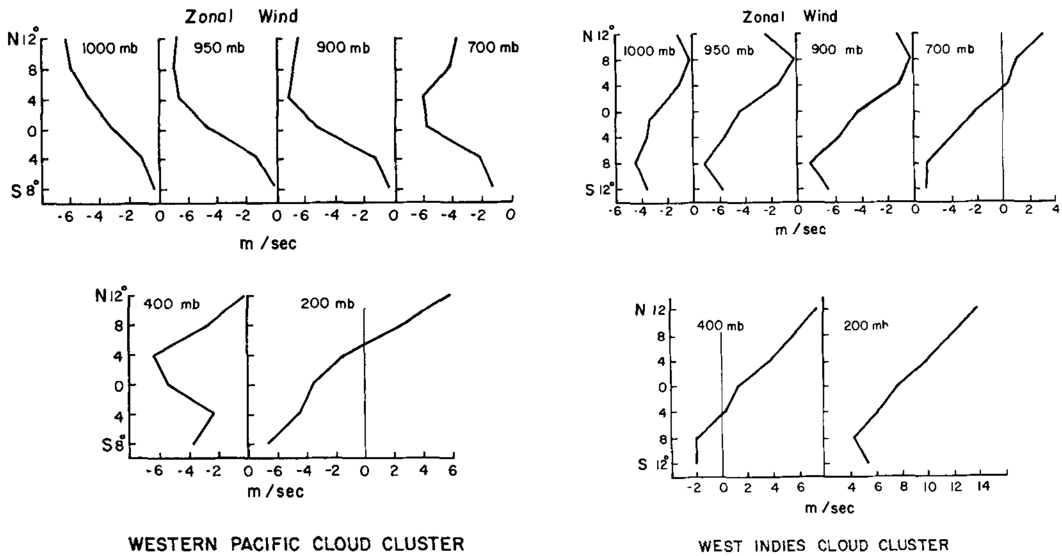


Fig. 15. Portrayal of north–south distribution of zonal wind relative to the Western Pacific and West Indies cloud clusters.

Western Pacific clusters have relatively few squall lines except in the winter and between Johnson and Kwajalein where vertical shear is larger.

3.3. *Barotropic instability of the mean zonal flow*

The north–south gradient of zonal wind in the Pacific gives rise to the possible existence of barotropic instability. Nitta and Yanai (1969) have indicated that at certain times and at certain levels conditions for barotropic instability are satisfied in the Western Pacific. The necessary condition is that the meridional gradient of absolute vorticity, i.e.

$$\frac{\partial}{\partial y}\left(f - \frac{\partial \bar{u}}{\partial y}\right) \quad \text{or} \quad \left(\beta - \frac{\partial^2 \bar{u}}{\partial y^2}\right),$$

changes sign. Here y measures north–south distance, f is the Coriolis parameter, and $\bar{u}$ is the mean zonal wind. Thus a necessary condition for instability is that the zonal flow has an inflection point (i.e. $\partial\bar{u}^2/\partial y^2$ changes sign).

Fig. 15 shows the mean zonal wind profiles at selected levels for the Western Pacific and West Indies cloud clusters. The required inflection point is found at all levels except 200 mb for the Western Pacific cloud cluster. In the West Indies it is only found below 700 mb. The Western Pacific cluster absolute vorticity $(f-(\partial\bar{u}/\partial y))$ shows a local maximum only at 400 mb. At all other levels and in the West Indies the absolute vorticity of the mean cluster does not change its sign.

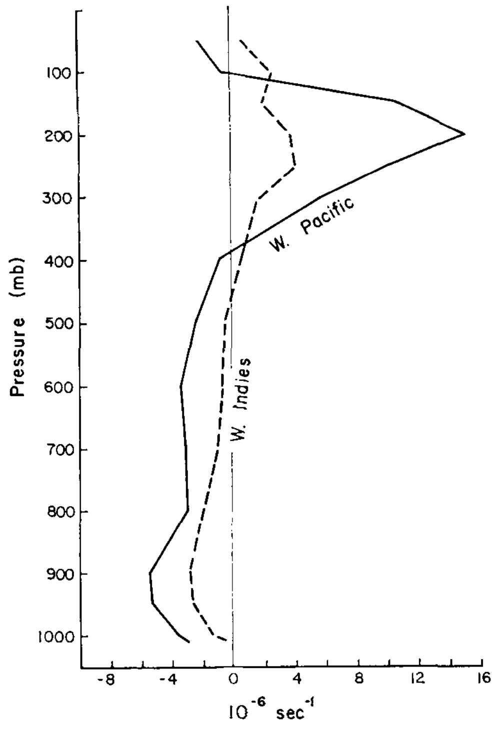
These results indicate that the growth and development of the cloud cluster disturbances cannot be typically explained by the barotropic instability process on this 4° scale of shearing gradient. The decrease of the gradient of absolute vorticity around the Western Pacific cloud clusters, however, leads to the conclusion that the barotropic instability mechanism may, in some situations, be a significant energy source for disturbance growth This can occur in individual situations when:

1) the gradient of north–south shearing fields is much larger than the mean shear shown and/or when

2) our 4° grid resolution was too broad to resolve the real shear which was occurring.

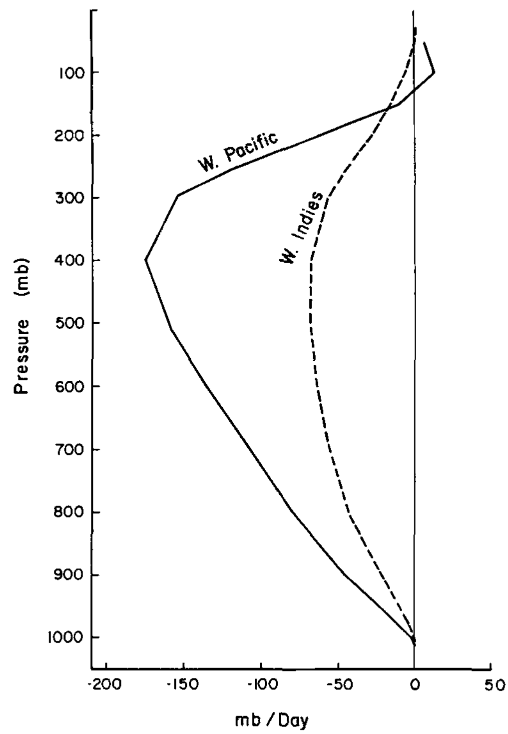
3.4. *Divergence and vertical velocity*

Williams & Gray (1973) have previously shown vertical divergence profiles for Western Pacific cloud clusters and clear regions. The cloud cluster profile showed a nearly constant inflow layer up to 400 mb and an intense outflow layer between 100–300 mb. Reed & Recker



CLOUD CLUSTER DIVERGENCE

Fig. 16. Mean vertical divergence profiles for Western Pacific and West Indies cloud clusters.



CLOUD CLUSTER VERTICAL VELOCITY

Fig. 17. Vertical profiles of the mean vertical velocity for Western Pacific and West Indies cloud clusters.

(1971) have also indicated a similar profile in the trough region of wave disturbances passing through the Marshall Islands. Yanai et al. (1973) have also derived this type of vertical divergence profile in the summer for a large pentagonal region in the Marshall Islands which by favorable circumstance straddled the ITCZ area of convection. A question arises whether this divergence profile is applicable to cloud clusters other than those of the Pacific. The comparison between the vertical divergence and vertical velocity profiles for the Western Pacific and the West Indies cloud cluster are given in Figs. 16 and 17.

We conclude from these results that Western Pacific clusters are stronger and better developed than their counterparts over the West Indies. A two layer structure of divergence can be generally applied to disturbances of both regions. The shape of the profiles are in general agreement. The intensity of the vertical circulation and low-level convergence is significantly less in the West Indies. The maximum vertical velocity at 400 mb is two-and-a-half times higher for the Western Pacific than for the West Indies. Similar conclusions have been advanced by Nitta & Esbensen (1974). The energy supply due to the release of latent heat thus appears to be more than twice as large in the Western Pacific clusters.

3.5. *Boundary layer frictional wind veering*

Table 2 lists observed 1st and 2nd km mean wind veering information for both regions without respect to weather systems. The West Indies data has been divided into two latitude belts (greater and less than 18°). Data north of 18° show larger veerings. This is probably due to the extra terrain influences of large islands. This information indicates that a distinct 10° frictional veering is present in the lowest kilometer layer.

Fig. 18 shows the vertical profile of observed

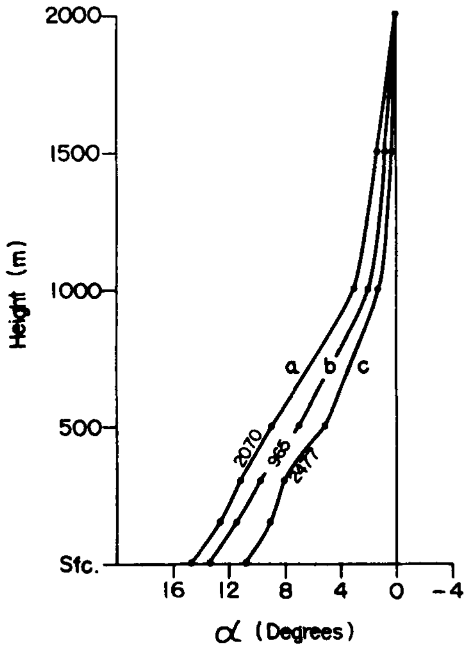


Fig. 18. Observed wind angle veering with height in the lowest two km for island-atoll and ship reports of the tropical Pacific. Wind direction at two km is used as reference. Curve *a* represents veering of wind with height from ships which were located at least 1° latitude from any land. Curve *b* represents the frictional vereing of wind with height as observed from the atoll-island data; curve *c* as observed from ships located within 1° latitude of land. Number of observations in each class is shown on the figures.

wind veering for the Western Pacific atoll and island stations and for selected surface ships surrounding the atolls (see a report by Gray, 1972 for more discussion). This figure shows

Table 2. *1st and 2nd km wind veering from atoll and island stations*

	sfc–1 km	1 km–2 km
Western Pacific (5–18° latitude), 7 210 cases	10°	2°
West Indies (less 18° latitude), 9 205 cases	10°	3°
West Indies (greater 18° latitude), 7 980 cases	14°	3°
Average	11°	3°

Table 3. *Cloud cluster vs. clear area 1st and 2nd km wind veering*

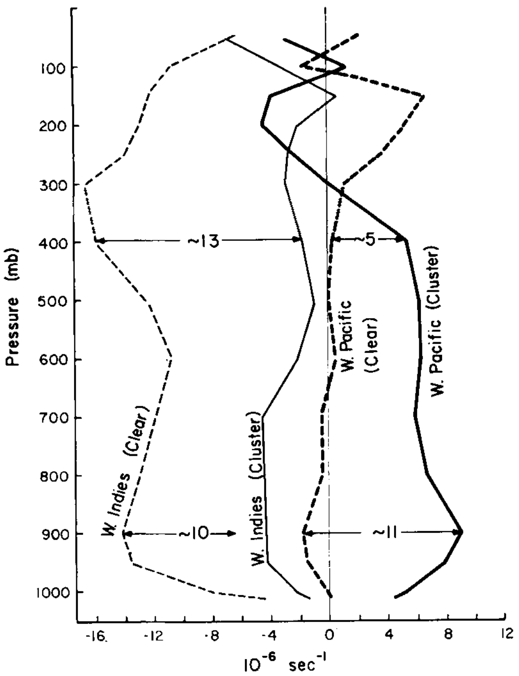
Cloud clusters	(1 km–sfc)	(2 km–1 km)
Western Pacific, 536 cases	10°	4°
West Indies (less 18° latitude), 266 cases	13°	6°
West Indies (greater 18° latitude), 244 cases	27°	9°
Cluster average	15°	6°
Clear areas average	11°	2°

that there is essentially no difference between the statistical averages of the island-atoll and ship veerings.

Table 3 shows the 1st and 2nd km wind veering for stations within 2° latitude of the center of the satellite-observed tropical cloud clusters and the average veering information relative to tropical clear regions. Note that for the clusters the 1st km veering is more (15° vs. 11°), and it extends well into the second km level, (6° veering for the clusters vs. 2° veering for the clear areas). This is to be expected if the cumulus clouds act to carry turbulent sub-cloud layer momentum to higher levels. The clear area veering is very similar to the average of the ship and atoll-island stations. Thus, there does appear to be a significant amount of frictional wind veering associated with the cloud clusters. How closely is this frictional veering related to the observed low level relative vorticity fields?

3.6. *Vorticity*

It is important to understand the vorticity budget of the cloud cluster and to what extent are the tropical weather system's low-level vorticity and convergence related? Figs. 9–12 show that the N–S zonal shear is basically different between the two regions. The location of the maximum trade winds is typically north of the clusters in the Pacific and south of the clusters in the West Indies. North–south wind shears are of opposite sign in each region. If the zonal shear dominates the vorticity, then we would expect vorticity of opposite sign in the two regions. This is verified in the vertical profiles of relative vorticity as seen in Fig. 19. In the Pacific cyclonic corticity is present across the



CLOUD CLUSTER AND CLEAR AREA VORTICITY

Fig. 19. Mean vertical vorticity profiles for cloud clusters and clear regions over the Western Pacific and the West Indies. Arrows give the difference between cloud cluster and clear region vorticity.

cloud clusters up to 300 mb. Above this level there is a change to anticyclonic vorticity. Clear regions show neutral low-level vorticity. Upper tropospheric clear region vorticity is positive. These Pacific vorticity fields are different than those of the West Indies. West Indies cluster vorticity is slightly negative at all levels up to 200 mb. Only small height variations occur. The West Indies clear region vorticity is strongly anticyclonic at all levels.

In both regions the surface to 400 mb difference of vorticity between clear and cluster areas is nearly the same. The reversal of vorticity with height which takes place at 300 mb between the cluster and clear regions in the Pacific does not occur in the West Indies.

These similar region lower tropospheric vorticity differences between cluster and clear regions indicate that the large scale ($\sim$5–15°) mean vorticity fields are less important in explaining the cluster and clear area divergence differences than the smaller scale variations of vorticity which are superimposed upon the mean field.

Low tropospheric convergence in the West Indies is present despite the negative cluster vorticity. Thus, the association of low level cyclonic vorticity and boundary layer convergence—the assumption of the CISK hypothesis of Ooyama (1964) and Charney & Eliassen (1964)—would seem to be invalid for West Indies cluster convection. If one were to explain the observed West Indies convergence and divergence patterns in terms of deviation of vorticity from an average state, then this direct association can be made. Table 4 portrays individual region mean values of divergence and vorticity and the cloud cluster and clear region deviation of these parameters from their broadscale regional averages. These results leave the question open whether the vorticity beneath the cloud cluster produces the convergence (Ekman pumping) or whether the convergence is responsible for the generation of cyclonic vorticity as described by Ceselski (1974) for the cloud cluster in the West Indies.

Table 4. *Comparison of convergence and vorticity at 900 mb in the Western Pacific and West Indies*

All values in units of 10^{-6} sec^{-1}

	Western Pacific			West Indies		
	Regional mean	Cluster	Clear regions	Regional mean	Cluster	Clear regions
Mean convergence	0	6	−3	0	3	−2
Mean vorticity	3	9	−2	−10	−4	−14
Deviation of convergence from regional mean	—	+6	−3	—	+3	−2
Deviation of vorticity from regional mean	—	+6	−5	—	+6	−4

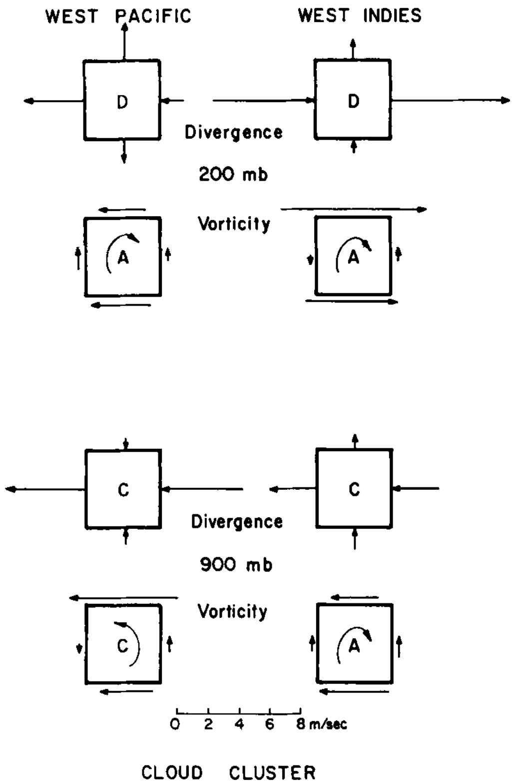


Fig. 20. Schematic diagram of the contribution by the zonal and meridional winds to divergence and vorticity at 900 mb and 200 mb levels for Western Pacific and West Indies cloud clusters.

3.7. *Relative contribution of the zonal and meridional components to divergence and vorticity*

The discussion of the wind cross sections (Figs. 5–12) showed that the horizontal shear of the broadscale zonal and meridional winds was significantly different in the two regions. Williams & Gray (1973) stated that the ratio of zonal to meridional horizontal shear in the Western Pacific is about 3 to 1 at nearly all levels. Reed & Recker (1971), however, have stated that the relative contributions of both components in moving trade wind waves are about equal. This larger contribution of the meridional shearing component in the Reed and Recker study is understandable because of the difference in the compositing schemes. They based their compositing scheme on already determined strong meridional wind changes that are typical of well defined westward moving systems. The typical satellite-observed cloud cluster often appears not to be associated with trade wind waves.

The contribution of the wind components to divergence and vorticity is schematically summarized in Fig. 20. The meridional and zonal wind components are shown with respect to their influence on divergence and vorticity for the 900 and 200 mb levels. These are the levels of maximum inflow and outflow. The striking feature in this figure is the change of the meridional wind north and south of the Western Pacific cloud cluster between 900 and 200 mb. At the lower level the v-components are directed into the cluster, at the upper level both components are directed away from the cluster center. In contrast, the meridional winds blow through the West Indies cloud cluster from south to north at both levels. The zonal winds do not significantly change their speed as they blow through the cluster. This figure also indicates the importance of the N–S zonal wind shear. Despite the change of the zonal wind direction with height in the West Indies the zonal wind provides a large anticyclonic shear at both levels, whereas, over the Western Pacific the E-winds north of the cloud cluster center decrease with height while the E-winds south of the center remain constant.

4. Cluster vorticity budget

Recent diagnostic studies of tropical disturbances (e.g. Gray, 1973; Lopez, 1973; Yanai et al., 1973) indicate a large vertical mass recycling within the cloud cluster. At lower levels the magnitude of this eddy up-and-down mass recycling is an order of magnitude or more larger than the mean upward mass flow. A general consensus on the magnitude of this large vertical recycling is now occurring. The above studies have calculated this vertical recycling circulation from required mass, total energy, and moisture budgets. Little has been accomplished with regard to the cloud cluster vorticity and momentum budgets. This chapter discusses the cloud cluster vorticity budget.

Williams & Gray (1973) have previously calculated the large-scale terms of the vorticity equation and found large lower and upper tropospheric imbalances of opposite sign. They concluded that Cb scale eddy vorticity transports carry an appreciable amount of vorticity from the lower to upper troposphere. A similar conclusion was reached by Riehl & Pearce (1968) for easterly waves in the West Indies area. During the summer monsoon in the Himalaya Mountain region, Holton & Colton (1972) have also indicated that sub-grid scale vertical transport of vorticity must be occurring. They concluded that cumulonimbus convection must play an important role for vorticity balance at 200 mb. The decrease of vorticity due to 200 mb divergence over this region must be largely balanced by a selective Cb importantion of vorticity from lower levels. Reed & Colton (1971) also indicated a sub-grid scale vorticity imbalance in their study of easterly-wave disturbances. In an attempt to determine what these eddy processes might be, a further more detailed analysis of the cloud cluster vorticity equation will now be made.

The vorticity equation for an x, y, p-system will be integrated over the 4° diameter cloud cluster of area A. This includes all the cloud cluster individual cumulus elements and their immediate environment. The vorticity equation for the cloud cluster can be written as:

$$\frac{\partial \bar{\zeta}}{\partial t} = -\frac{1}{A}\oint_s \zeta_a \mathbf{V}_n \, dl - \overline{\omega \frac{\partial \zeta}{\partial p}} - \overline{\left(\frac{\partial \omega}{\partial x}\frac{\partial v}{\partial p} - \frac{\partial \omega}{\partial y}\frac{\partial u}{\partial p}\right)} - \overline{\operatorname{rot} F} \tag{1}$$

where

ζ, ζ_a	= relative and absolute verticity
$\mathbf{V}_n$	= horizontal wind component normal to the cluster, positive outward
dl	= length along cluster boundary s
rot F	= rotation of the friction vector (or rotation of the momentum divergence by small scale turbulent eddies)
$\overline{}$ bar	= denotes an area average

The second and third terms on the right can be expanded into a mean and an eddy contribution. It is assumed that the eddy terms include only the effect of the cumulus and cumulonimbus scale eddies. The smaller scales are combined in the last term, thus

$$\frac{\partial \bar{\zeta}}{\partial t} = -\frac{1}{A}\oint_s \zeta_a \mathbf{V}_n \, dl - \bar{\omega}\frac{\partial \bar{\zeta}}{\partial p} - \overline{\omega' \frac{\partial \zeta'}{\partial p}} - \overline{\left(\frac{\partial \omega'}{\partial x}\frac{\partial v'}{\partial p} - \frac{\partial \omega'}{\partial y}\frac{\partial u'}{\partial p}\right)} - \overline{\operatorname{rot} F} \tag{2}$$

Note that the mean value of the twisting term has been dropped. Yanai (1961) and Yanai & Nitta (1967) have performed case studies for a Western Pacific tropical storm and an easterly wave in the West Indies. They found that the twisting term can at times reach large values at upper levels.

Our analysis of the twisting term indicates that it can be significant at individual locations on the east, west, north, or south side of the cluster but when averages are taken over the 4° cluster, a symmetry in the gradient of vertical motion on the sides of the cluster is present and that the mean values of the twisting term averaged over the cluster becomes small. It is to be noted that the N–S and E–W horizontal gradients of vertical wind shear across the cluster are typically small. For these reasons we have neglected consideration of the mean value of this term.

Equation (2) was used for the evaluation of the cloud cluster vorticity balance in both regions. Two assumptions were made:

1) The horizontal eddy transport at the cloud cluster boundary is negligible and the first term on the right of equation (2) can be calculated by consideration of only mean values.

2) Gust-scale mechanical friction is only effective in the boundary layer. Dissipation of momentum by small-scale (sub-cumulus) turbulence effect is negligible above 900 mb. With the Gauss theorem the friction term can then be written as:

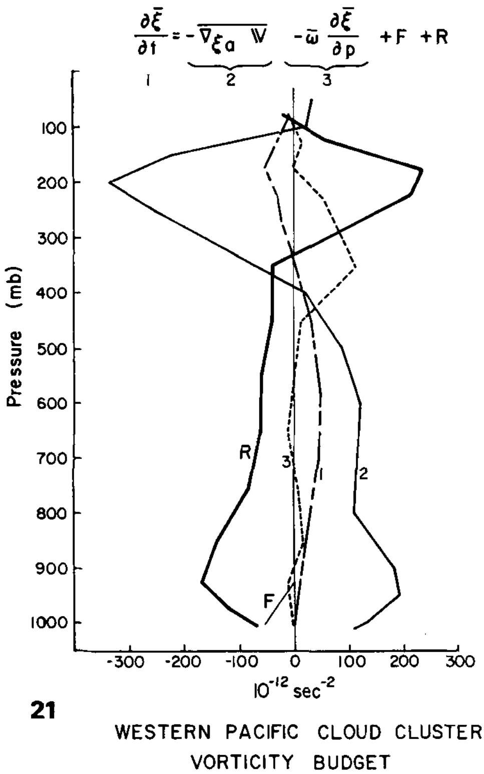
$$\int_A \operatorname{rot} F \, dA = \oint_l F_x \, dx + F_y \, dy$$

The surface friction was estimated using the formula:

$$F = \frac{C_D}{H} |\mathbf{V}_0| \mathbf{V}_0 \tag{3}$$

$C_D = 1.3 \times 10^{-3}$; $H = 750$ m; and $\mathbf{V}_0$ = surface (10 m height) wind.

Figs. 21 and 22 show the results of these calculations. R stands for residual. It is composed of the eddy vertical advection, eddy twisting term and all data and physical hypothesis inadequacies:



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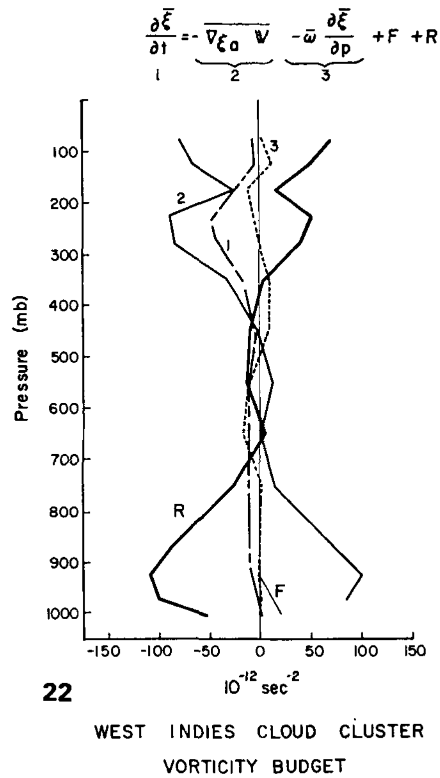

22

Figs. 21–22. Vorticity budgets for the Western Pacific and West Indies cloud clusters. F = friction term, R = residual = eddy plus twisting terms.

$$R = -\overline{\omega' \frac{\partial \zeta'}{\partial p}} - \overline{\left(\frac{\partial \omega'}{\partial x}\frac{\partial v'}{\partial p} - \frac{\partial \omega'}{\partial y}\frac{\partial u'}{\partial p}\right)}$$

$$+ \text{ all data and physical hypothesis inadequacies} \qquad (4)$$

The horizontal divergence of the vorticity transport is the predominant term in the budget in both regions. Clusters gain large amounts of vorticity in the lower half of the troposphere by convergence. They lose large amounts of vorticity through upper tropospheric divergence. Lower convergence and upper level divergence nearly balance. Terms 1, 3, and F are seen to make only small contributions to the vorticity budget. Thus, the residual term is primarily a result of the horizontal transport.

Neglecting data and previously assumed hypothesis inadequacies, the eddy processes included in R must transport vorticity out of the lower troposphere and transport it into the upper troposphere. If this does not occur, then the eddy processes must act in some fundamental way to dissipate vorticity at low levels and generate it at high levels. The authors know of no physical basis for the latter conclusion. They, therefore, accept the idea of a large magnitude of eddy transport of vorticity from lower to upper troposphere. It seems reasonable that Cb convection could accomplish this.

Vorticity transports by cumulonimbi. In order to show more details of the likely cloud cluster vertical eddy transports we shall discuss what we think to be a reasonable estimate of the vorticity distribution and mass flux within the cloud cluster associated with cumulonimbus convection. The 4° diameter cloud cluster area will be divided into three regions:

(*a*) region of strong updrafts—horizontal fractional area σ_u,

(*b*) region of strong downdrafts—horizontal fractional area σ_d, and

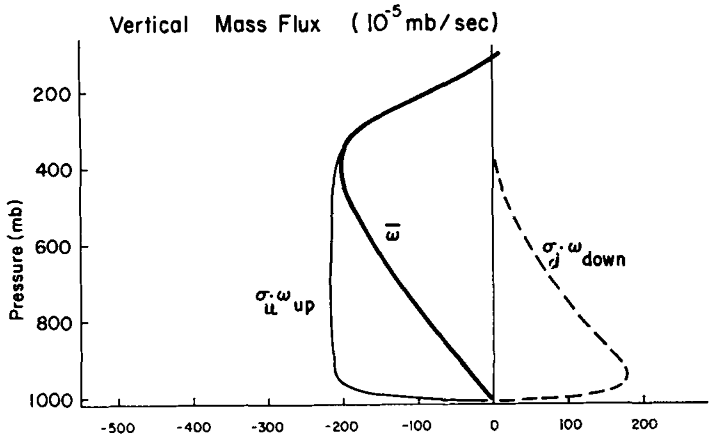

Fig. 23. Mean cloud cluster vertical motion (in 10^{-5} mb/sec) by the mean circulation ($\bar{\omega}$), the upward component by the deep cumulus convection ($\sigma \cdot \omega_{up}$) and the downdraft motion ($\sigma \cdot \omega_{down}$) of the deep cumulus.

(c) region within the cloud cluster between the strong up and downdrafts, where it is assumed that the average of the residual vertical motion is zero—horizontal fractional area $\sigma = 1 - (\sigma_u + \sigma_d)$.

Our up- and downdrafts will be composed only of the vertical motions associated with deep cumulonimbus convection. The mean upward motion within the cloud cluster is assumed to result from a greater mass transport by the strong updrafts compared to the strong downdrafts.

Figs. 23–26 show the results of our hypothesized model. The essential aspects of these figures are not the exact values of the profiles shown but the magnitudes of the differences between the updrafts and downdrafts and the mean values.

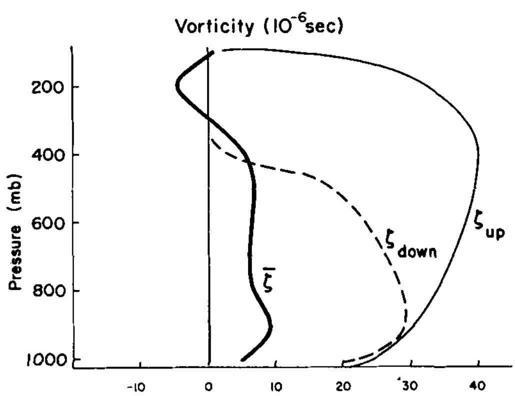

Fig. 24. Cloud cluster mean vorticity ($\bar{\zeta}$) and hypothesized vorticity within deep cumulus updrafts (ζ_{up}) and downdrafts (ζ_{down}).

The vertical mass flux (Fig. 23) is derived from our calculated mean Western Pacific cloud cluster divergence. Assuming a cluster horizontal fractional area $\sigma_u = 1\,\%$ for the intense updrafts at 900 mb we calculated a mean vertical velocity of about 2 m/sec at this level for the Cb updrafts. E. Zipser has indicated (personal communication, 1973) that in squall line systems over the West Indies the downdraft region exceeds the updraft area by a factor of 2 or more. We will assume the horizontal fractional area of the cloud cluster downdrafts (σ_d) to be $\sim 2\,\%$. With this we calculated a mean downdraft velocity of about 0.8 m/sec at 900 mb assuming that the 97 % fractional environmental area (σ_e) between the up- and downdrafts has no vertical motion ($\omega_e = 0$) from the defining vertical mass flux equation

$$\bar{\omega} = \sigma_u \omega_u + \sigma_d \omega_d + \sigma_e \omega_e \tag{5}$$

The hypothesis is made that the middle level (400–900 mb) convergence feeds into the downdrafts. The updraft origins below 900 mb result from the mass accumulation of the downdrafts into the surface to 900 mb layer and the mean cloud cluster upward transport. The upward mass flux in the middle levels (500–900 mb) is nearly constant, and the downward mass flux increases from zero at 400 mb to its maximum at the top of the boundary layer ~ 900 mb.

This very large mass flux in the strong up-

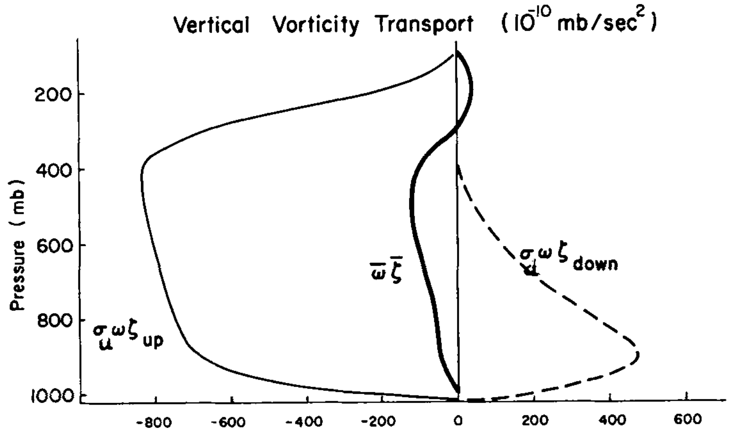

Fig. 25. Comparison of mean cloud cluster vertical vorticity transport by the mean circulation ($\bar{\omega}\bar{\zeta}$) and by updraft and downdraft components of the deep cumulus convection.

and downdrafts (Fig. 23) is in reasonable agreement with the estimated values of Gray (1973). He indicated a larger magnitude of recycling in the lower troposphere than we are showing. We are neglecting the influence of the extra recycling by small cumulus up-and-down motion because we believe they do not contribute in a significant way to the net cluster mass and vorticity budget. By contrast, the deep cumulus are the major contributors. These values are slightly larger than those found by Yanai et al. (1973). He and his group made calculations in a large pentagonal region in the Marshall Islands enclosing both cloud and clear regions on both disturbed and undisturbed days. By contrast, our calculation is made only for the disturbed cloud cluster area.

A scale analysis of vertical motion beneath and within up-and-downdrafts indicates that these draft values of 1–10 m/sec can be two or more orders of magnitude larger than the cloud cluster average vertical motion of 0.1 to 2 cm/sec. Thus, $\omega_{up}/\bar{\omega}$ and $\omega_{down}/\bar{\omega} \sim 10^2$ to 10^3. A scale analysis of vorticity shows that the convergence requirements of up-and-downdrafts dictate that the vorticity spin-up (from the vorticity conservation equation, $A_c \zeta_c = A_e \zeta_e$, where A stands for the area and the subscripts c and e refer to area values of the cumulus and of the environment respectively) by a factor of

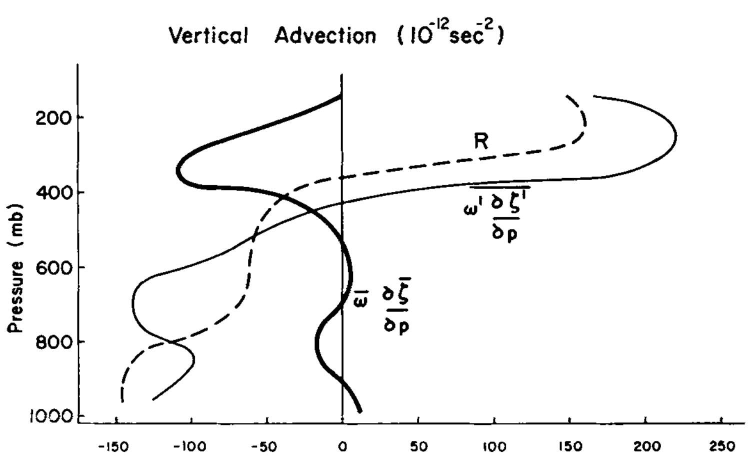

Fig. 26. Comparison of the mean and eddy vertical vorticity advection terms with the vorticity residual calculated in Fig. 25.

3 to 10 beneath and within Cb up- and downdrafts. Thus, ζ_{up}/ζ and $\zeta_{\text{down}}/\zeta \sim 3$–10, where ζ_{up}, ζ_{down}, and ζ are the vorticity within the updrafts, downdrafts, and the mean within the cluster. The vertical cluster vorticity profiles shown in Fig. 24 were derived in accordance with this scale analysis. The relative vorticity in the up- and downdrafts must be much larger than it is in the surrounding environment.

The vertical profiles in Figs. 23 and 24 show our estimates of the typical magnitudes of mass transport in the up- and downdrafts and our assumed vorticity values within the drafts. These values are next combined in Fig. 25 to portray the typical vertical vorticity transport within the up- and downdrafts where the total average cluster vertical vorticity transport $(\overline{\omega\zeta})$ is defined as

$$\overline{\omega\zeta} = \sigma_u \omega_u \zeta_{\text{up}} + \sigma_d \omega_d \zeta_d + \sigma_e \omega_e \zeta_e \tag{6}$$

The vertical vorticity transport of the environment between the up- and downdrafts $(\sigma_e \omega_e \zeta_e)$ is assumed to be negligibly small. The vorticity in the updrafts increases slightly up to 400 mb. It is dissipated by spindown processes of the upper level outflow. Downdraft vorticity reaches its maximum at the top of the boundary layer where it spreads out and reduces its spin. The vertical vorticity transport within the cumulonimbus updrafts is an order of magnitude greater than the vorticity carried by the mean upward circulation. This figure also shows the large magnitude of the vorticity downdraft transport to the boundary layer to balance the large export by the updrafts. These strong downdrafts are a necessary ingredient of the vorticity budget. It becomes more and more clear how important the cumulus scale vertical transport processes are to the dynamics of the cloud cluster.

We can now compare our hypothesized model of vertical vorticity advection with the observed cluster residual vorticity budgets of Figs. 21 and 22. These mean vs. eddy vertical transports are compared in Fig. 26. This figure gives a direct comparison between our hypothesized model and our calculated mean vorticity budget. The eddy advection term $\omega'(\partial\zeta'/\partial p)$ was calculated with the values from Figs. 23–24. The dashed curve shows that the Western Pacific determined residual of vorticity is in good agreement with the eddy advection term. We thus feel that this explanation of the mean vorticity residual as being primarily due to selective transport by the deep cumulus is, indeed, a reasonable one. We believe the eddy twisting term, therefore, is of only minor importance. Reed & Johnson (1974) have also recently calculated the mean vorticity budget in the trough of composited W-Pacific wave disturbances. Their results on the role of selective transport of vorticity by deep cumulus agree well with our model.

These results indicate the vorticity equation can be treated as a transport equation where the sink of vorticity is the divergence term. A large amount of vorticity is horizontally transported into the lowest layers of the cloud cluster, and is selectively transported upwards by the Cb updrafts. This eddy vertical transport is necessary for the maintenance of the vorticity loss due to upper tropospheric divergence.

For a number of years the second author has believed in the fundamentally important role played by deep cumulus convection in selective vertical transport of momentum and vorticity.

5. Summary

The main characteristics of the cloud clusters and some of their important regional differences are listed below:

	Western Pacific	West Indies	Remarks
Size	$4° \times 4°$	$4° \times 4°$	Same
Movement	$\sim 6°$ long./day towards W	$\sim 6°$ long./day towards W	Same despite different ambient wind fields
4° Scale divergence	Convergence below 400 mb divergence above	Convergence below 450 mb divergence above	Shape of the profiles similar, magnitude 2–3 times higher in Pacific, less middle level convergence in West Indies
4° Scale vorticity	Cyclonic below 300 mb; anticyclonic above	Anticyclonic at all levels	Large differences because of opposite sign of north–south zonal shear
4° Scale vorticity differences between clear and cluster regions	Much larger cluster vorticity $\sim 10 \times 10^{-6}$ sec^{-1}	Much larger cluster vorticity $\sim 10 \times 10^{-6}$ sec^{-1}	Pattern similar for both regions
Vorticity budget	Large required lower to upper tropospheric transport by Cb's	Large lower to upper tropospheric transport by Cb's is required	Upward vorticity transport by mean circulation is grossly insufficient to accomplish vertical vorticity balance
4° Scale vertical shear of zonal wind	Nearly constant with height	Large and positive sign of zonal wind changes	Squall lines more probable over the West Indies, more unorganized cluster Cb's in W. Pacific
Wave disturbances	S-winds E of center, weak N-winds W, rough agreement with wave model	Less agreement with wave model, vorticity of the v-component positive	Some clusters are associated with wave disturbances, some are not.
Barotropic instability	Inflection point in the u-profile at all levels below 400 mb. Change of sign in the meridional gradient of absolute vorticity only at 400 mb. Decrease of the gradient around cloud cluster center at all levels	Inflection point in the u-profile at all levels below 700 mb. No change of sign in meridional gradient of the absolute vorticity. Decrease of gradient of absolute vorticity at low levels around cloud cluster center	Not primary energy source for cloud clusters, but possible contributing mechanism for smaller scale growth in some cases
Ekman pumping	Low level cluster convergence and vorticity well associated, clear region divergence not well associated with anticyclonic vorticity	Low level divergence and clear region anticyclonic vorticity well associated, cluster vorticity and divergence not in agreement	Concept of vorticity and convergence association is valid only if vorticity is interpreted as a deviation from the region mean value

5.1. *Regional similarities*

From the viewpoint of the digitized satellite pictures, the majority of cloud clusters look much the same. Still, a variety of different types of disturbances are likely to be embedded within similar $\sim 4°$ clusters. This generally agrees with the conclusions of the NOAA Hurricane Center in Miami as discussed in their yearly surveys of Atlantic tropical disturbances (see Simpson et al., 1968, 1969; Frank, 1970).

It was unexpected that the cluster movement towards the west would be similar in both regions despite significant differences in the ambient wind fields. As discussed by Wallace (1970) cluster propagation may not be well related to the velocity of the embedded flow field.

The two layer divergence structure of these disturbances as observed by Williams & Gray (1973) is supported by both divergence profiles. Although the basic flow fields in both regions are different, the individual region cluster minus clear region differences of vorticity and divergence are nearly the same. Weather system deviations of vorticity from a mean state environmental flow field are very similar. Cluster convergence below 400 mb is primarily due to the v-component gradient, the vorticity is due mainly to the zonal wind shears.

5.2. *Regional differences*

The major differences between the two regions appear to be due to location. Clusters averaged $\sim 9°$ N in the Western Pacific and $\sim 24°$ N in the West Indies. The Western Pacific area is located equatorward of the trade wind maximum, the West Indies area is located on the poleward side of the trades. Therefore, the mean zonal flow provides cyclonic shear over the Western Pacific and anticyclonic shear over the West Indies. Cluster and clear region wind deviations in the Western Pacific were superimposed on a basic cyclonic shearing pattern while West Indies systems were superimposed on an anticyclonic shearing field. This seems to indicate that the cluster systems can travel and maintain themselves in an unfavorable or anticyclonic flow field.

The West Indies region is not a favorable area for cluster genesis. Frank (1970) and Simpson et al. (1968, 1969) have shown that West Atlantic weather systems typically tend to be in a "coasting" or "running down" mode as they cross the West Indies. They found that 90% of the disturbances observed in the West Indies during 1967 and 1968 were generated far to the east, 50% coming out of Africa. The much higher values of the cluster divergence profile for the Western Pacific and the typical lack of cluster genesis in the West Indies support this assessment. A following paper (Part II) discusses the cloud cluster thermal and moisture fields.

A more expanded discussion of this material is available in a Colorado State University Project Report by Ruprecht & Gray (1974).

REFERENCES

Ceselski, B. F. 1974. Cumulus convection in weak and strong tropical disturbances. *J. Atmos. Sci. 31*, 1241–1255.

Chang, C. P. 1970. Westward propagating cloud patterns in the tropical Pacific as seen from time-compositing satellite photographs. *J. Atmos. Sci. 27*, 133–138.

Charney, J. & Eliassen, A. 1964. On the growth of the hurricane depression. *J. Atmos. Sci. 21*, 68–75.

Frank, N. 1970. Atlantic tropical systems of 1969. *Mon. Wea. Rev. 98*, 307–314.

Frank, N. & Johnson, T. M. 1969. Vortical cloud systems over the tropical Atlantic during the 1967 hurricane season. *Mon. Wea. Rev. 97*, 124–129.

Gray, W. M. 1972. A diagnostic study of the planetary boundary layer over the oceans. Colorado State University Department of Atmospheric Science Research, Report No. 179, 95 pp.

Gray, W. M. 1973. Cumulus convection and large-scale circulations, Part I: Broadscale and mesoscale inversions. *Mon. Wea. Rev. 101*, 839–855.

Hayden, C. M. 1970. An objective analysis of cloud cluster dimensions and spacing in the tropical North Pacific. *Mon. Wea. Rev. 98*, 534–540.

Holton, J. R. 1972. *An introduction to dynamic meteorology*. Academic Press, New York. 319 pp.

Holton, J. R. & Colton, D. E. 1972. A diagnostic study of the vorticity balance at 200 mb in the tropics during the northern summer. *J. Atmos. Sci. 29*, 1124–1128.

JOC Study Group on Tropical Disturbances, 1968: Report on the first session of the JOC Study Group on Tropical Disturbances, Appendix I, GARP Publ. Series No. 4.

Lopez, R. E. 1973. Cumulus convection and larger scale circulations. Part II: Cumulus and mesoscale interactions. *Mon. Wea. Rev. 101*, 856–870.

Ludlam, F. H. 1966. Cumulus and cumulonimbus convection. *Tellus 18*, 687–698.

Nitta, T. 1970. Statistical study of tropospheric wave disturbances in the tropical Pacific region. *J. Meteor. Soc. Japan*, Ser. II, *48*, 47–60.

Nitta, T. & Esbensen, S. 1974. Heat and moisture budget analyses using BOMEX data. *Mon. Wea. Rev. 102*, 17–28.

Nitta, T. & Yanai, M. 1969. A note on barotropic instability of the tropical easterly current. *J. Meteor. Ser. Japan*, Ser. II, *47*, 127–130.

Ooyama, K. 1964. A dynamical model for the study of tropical cyclone development. *Geofisica Intern. 4*, 187–198.

Reed, R. J. & Colton, D. E. 1971. Properties of synoptic scale wave disturbances in the equatorial western Pacific: vorticity budget. Paper presented at 7th Technical Conf. on Hurricanes and Tropical Meteor., Barbados, West Indies, 6–9 December 1971.

Reed, R. J. & Recker, E. E. 1971. Structure and properties of synoptic-scale wave disturbances in the equatorial western Pacific. *J. Atmos. Sci. 28*, 1117–1133.

Reed, R. J. & Johnson, R. J. 1974. The vorticity budget of synopticscale wave disturbances in the tropical western disturbances. *J. Atmos. Sci. 37*, 1784–1790.

Riehl, H. 1945. Waves in the esterlies and the polar front in the tropics. Misc. Report No. 17, Dept. Meteor., Univ of Chicago.

Riehl, H. & Pearce, R. P. 1968. Studies on interaction between synoptic and mesoscale weather elements in the tropics. Atmos. Sci. Paper 126, Dept of Atmospheric Science, Colorado State University.

Ruprecht, E. 1975. Diurnal temperature corrections for rawinsonde humidity sensors. *Mon. Wea. Rev.*, 103, 352–355.

Ruprecht, E. & Gray, W. M. 1974. Analysis of satellite-observed tropical cloud clusters. Colorado State University Department of Atmospheric Science Research Report No. 219, 91 pp.

Sadler, J. C. 1962. Utilization of meteorological satellite cloud data in tropical meteorology. Procedings of the First International symposium on Rocket and Satellite Meteorology, Washington, D.C.

Simpson, J., Garstang, M., Zipser, E. J. & Dean, G. A. 1967. A study of a non-deepening tropical disturbance. *J. Appl. Meteor. 6*, 237–254.

Simpson, J., Frank, N., Shideler, D. & Johnson, H. M. 1968. Atlantic tropical disturbances, 1967. *Mon. Wea. Rev. 96*, 251–259.

Simpson, J., Frank, N., Shideler, D. & Johnson, H. M. 1969. Atlantic tropical disturbances of 1968. *Mon. Wea. Rev. 97*, 240–255.

Wallace, J. M. 1970. Time-longitude sections of tropical cloudiness (Dec. 1966–Nov. 1967). ESSA Technical Report, NESC 56 (available from G.P.O., Washington, D.C.), 37 pp.

Wallace, J. M. 1971. Spectral studies of tropospheric wave disturbances in the tropical western Pacific. *Rev. Geophys. Space Phys. 9*, 557–612.

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

Yanai, M. 1961. A detailed analysis of typhoon formation. *J. Meteor. Soc. Japan*, Ser. II, *39*, 187–213.

Yanai, M. & Nitta, T. 1967. Computation of vertical motion and vorticity budget in a Caribbean easterly wave. *J. Meteor. Soc. Japan*, Ser. II, *45*, 444–466.

Yanai, M., Esbensen, S. & Chu, J. 1973. Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. *J. Atmos. Sci. 30*, 611–627.

Yanai, M., Maruyama, T., Nitta, T. & Hayashi, Y. 1968. Power spectra of large-scale disturbances over the tropical Pacific. *J. Meteor. Soc. Japan*, Ser. II, *46*, 308–323.

Zipser, E. J. 1969. The role of organized unsaturated convective downdrafts in the structure and rapid decay of an equatorial disturbance. *J. Appl. Meteor. 8*, 799–814.

АНАЛИЗ СКОПЛЕНИЙ ТРОПИЧЕСКИХ ОБЛАКОВ, НАБЛЮДАЕМЫХ СО СПУТНИКОВ. ЧАСТЬ I. ВЕТЕР И ПОЛЕ ДИНАМИЧЕСКИХ ЭЛЕМЕНТОВ

Были собраны составные данные радиозондирования по северному полушарию соответственно трем классам тропических погодных систем, наблюдаемых со спутников в западной части Тихого океана (экватор $-18°$ с. ш.) и в Вест-Индии ($\sim$15–30° с. ш.). Эти системы включают облачные скопления шириной $\sim 4°$, их окружение ($\sim$2–6° широты относительно центра скопления) и безоблачные районы. Информация представлена за 1967 и 1968 гг. Общие концепции и методология сбора данных обсуждались в предыдущем сообщении Вильямса и Грея (1973). Составная информация содержит данные по ветру, дивергенции, завихренности и т. д. для индивидуальных погодных систем. Особо подчеркиваются региональные различия между погодными системами западной части Тихого океана и Вест-Индии. Существует значительная разница между ними. Подсчитывается средний бюджет завихренности облачного скопления. Главной составляющей бюджета в обоих районах является горизонтальная дивергенция переноса вихря. Необходимо предположить наличие большого переноса завихренности из нижней тропосферы в верхнюю. Представлены также другие наблюдательные факты.

Appendix. Locations of cloud clusters and clear regions

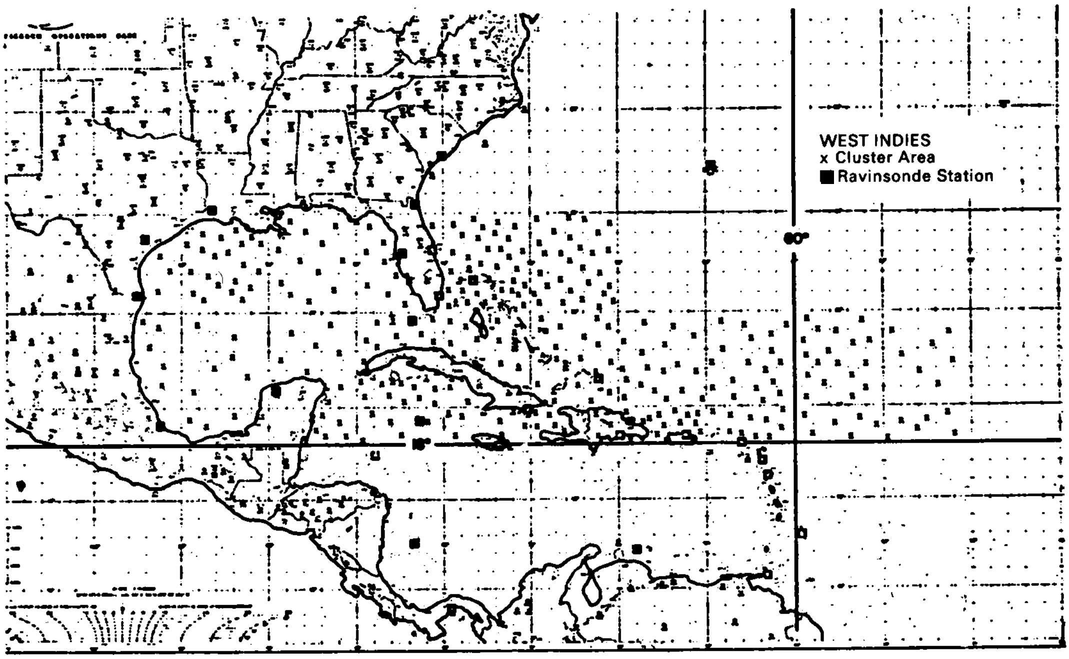

Fig. A 1. Location of cluster regions in the West Indies.

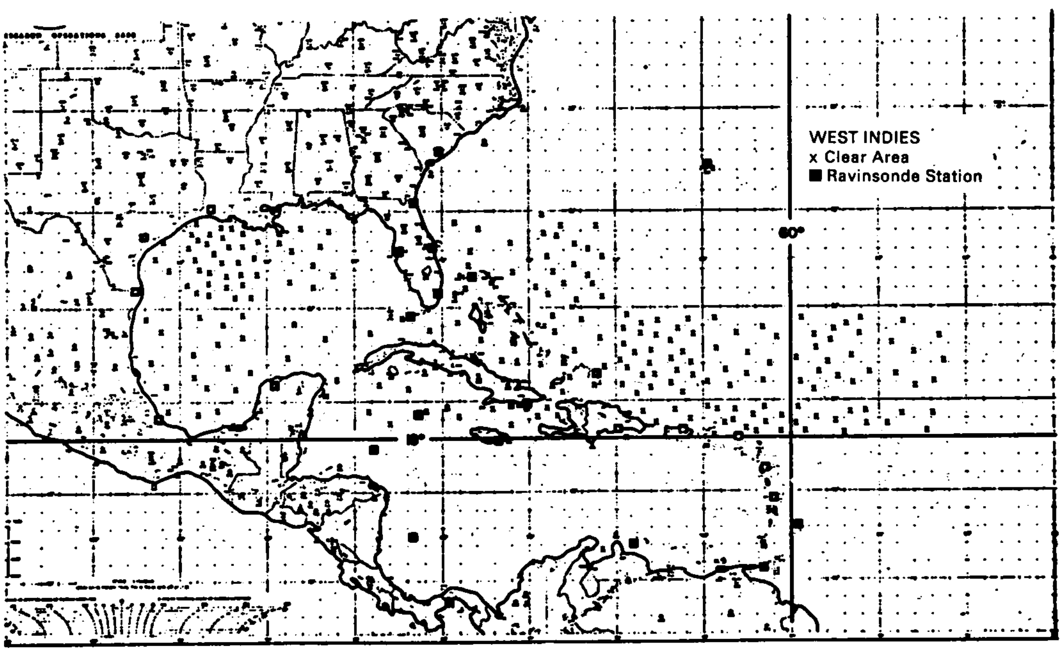

Fig. A 2. Location of clear regions in the West Indies.

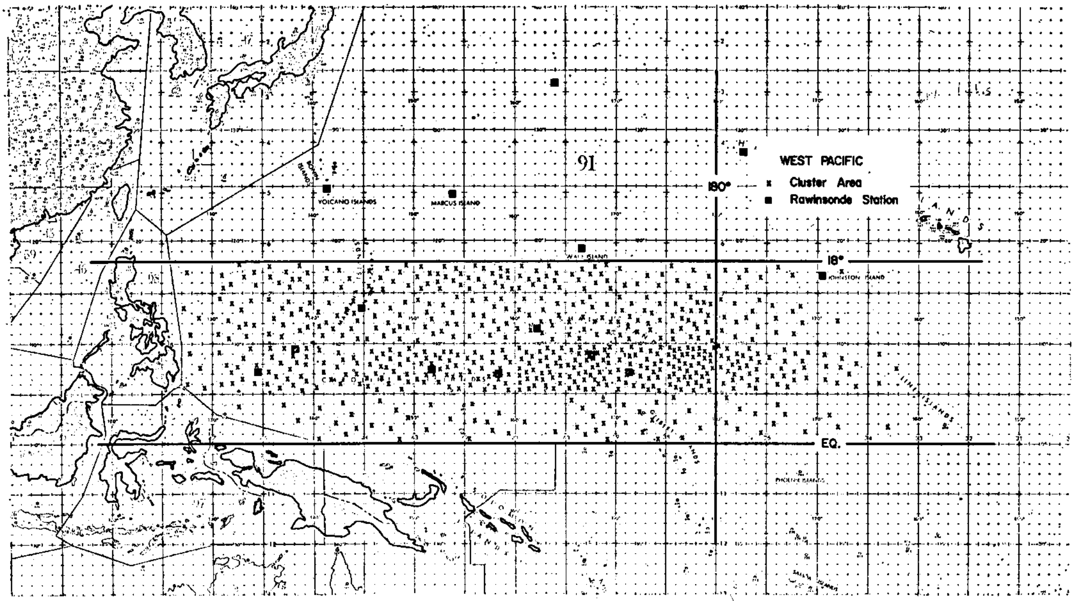

Fig. A 3. Location of cluster regions in the Western Pacific.

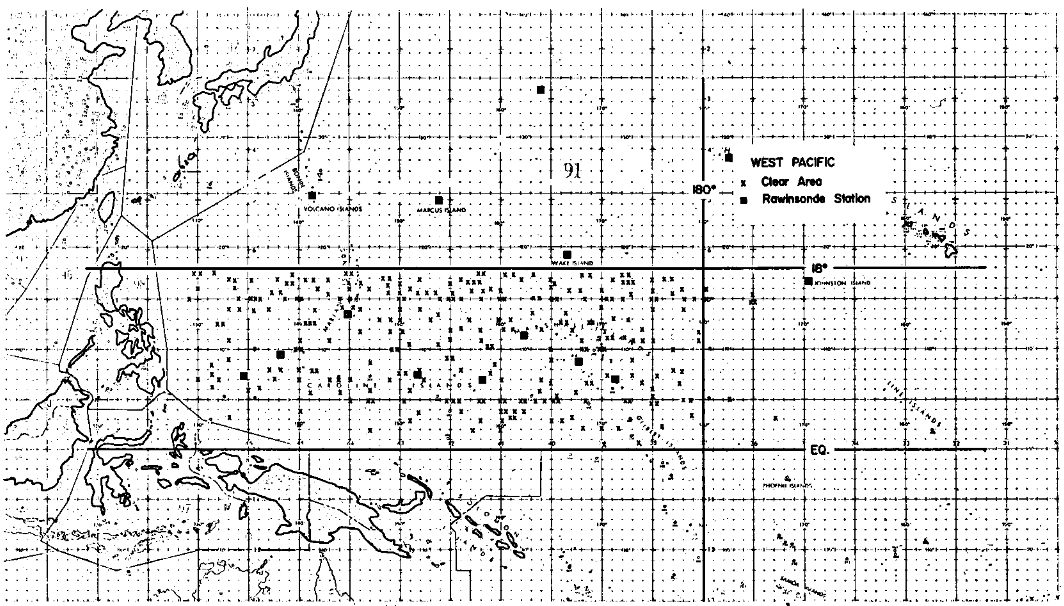

Fig. A 4. Location of clear regions in the Western Pacific.