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# ON THE FORMATION OF TYPHOONS

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ABSTRACT

A typical example of cyclogenesis in the western tropical Pacific Ocean is investigated. In the low troposphere, data do not support the concept of an equatorial front extending across the entire Pacific Ocean. They indicate the presence of several large clockwise-rotating eddies in the equatorial trough. Typhoon development begins in consequence of instability of the northern-hemisphere trades, not as a result of interaction between currents from northern and southern hemispheres. No evidence can be adduced in the low troposphere that suggests a reason for this instability.

Large vortices are found at 200 mb instead of the easterlies characteristic of low levels. Their spacing and speed are studied, and it is found that they are systems of the order of magnitude of long waves in the westerlies. After computations on the structure of the basic zonal current in the trade-wind belt, relations between high-tropospheric vortices and waves in the easterlies are discussed. These two types of perturbations are not propagated at the same speed, but there is relative motion between them. Instability of the trades sets in just as a lower wave trough is bypassed by a broadscale ridge in the high troposphere. The pressure field superimposed on the low levels by the upper vortices appears to furnish an explanation for the instability of the trades.

Finally the question is raised as to why some incipient storms attain typhoon intensity and others do not. It is found that a storm will grow when it becomes located under a ridge aloft, so that upper outflow from the storm is facilitated by the general pressure distribution.

It is well known that there is a low-pressure trough at the surface between the subtropical highs of northern and southern hemispheres that is nearly continuous around the globe. This trough is observed not only in climatological averages but also on most synoptic charts, though with large aperiodic departures from its average seasonal latitude. Its presence, therefore, is a truly normal condition.

A demarcation line is very frequently drawn on synoptic charts through the whole length of the trough —alternately called equatorial front, intertropical front, intertropical convergence zone, or equatorial convergence zone. This line is supposed to represent the boundary between converging trades of both hemispheres. Westerly winds appearing south of this zone in the northern-hemisphere summer are said to be produced by air crossing the equator under conservation of vorticity.

Inspection of maps drawn in the western Pacific during the war shows that some analysts felt that the simple picture just outlined was inadequate to explain the observed circulation on many days. There are also many reports that aircraft flying through the "front" only a few hours apart, encountered widely different conditions, one plane reporting impenetrable lines of cumulonimbus, and the next hardly any clouds at all. This should not happen if a continuous boundary were present and hints at a more complex structure of the equatorial trough. Such complexity has been demonstrated recently by Alpert (1945, 1946), Simpson (1947), and Glenn (1947). Doubt has also been expressed (Deppermann, 1940) that all air moving with a westerly component at low levels over the Indian Ocean is really of southern-hemisphere origin. In fact, flow charts prepared by Solot[2] now prove that northern-hemisphere air constitutes a large portion of this current.

In view of these circumstances, it seemed desirable to reexamine also the structure of the "equatorial front" in the western Pacific. Under contract with the Office of Naval Research several members of the Department of Meteorology, University of Chicago, went to Guam for the period July–November 1947, to study conditions at close range. Maps for the years 1945–46 were also analyzed. This report contains some aspects of the work done under this research program.

## 1. Data utilized and methods of analysis

Although the first-hand impressions obtained in Guam were an invaluable aid, subsequent investigation had to be largely restricted to the year 1945. This is the only year in which a dense coverage at low levels, some coverage at high levels, and high-latitude observations were simultaneously available. Data are presented in map and time-section form. Besides pilot-balloon observations, many pilot reports were available at lower elevations. Pilot reports up to 2000

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[1] Published as a contribution of a research project sponsored by the Office of Naval Research.

[2] S. Solot, "General circulation of the Anglo-Egyptian Sudan and adjacent regions," (to be published).

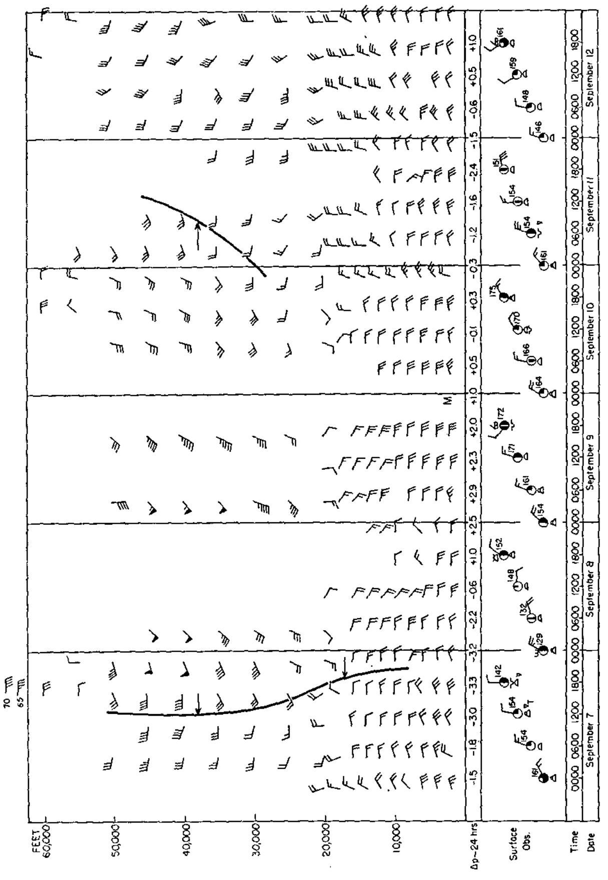

FIG. 2. Vertical time section at Honolulu, 7–12 September 1945.

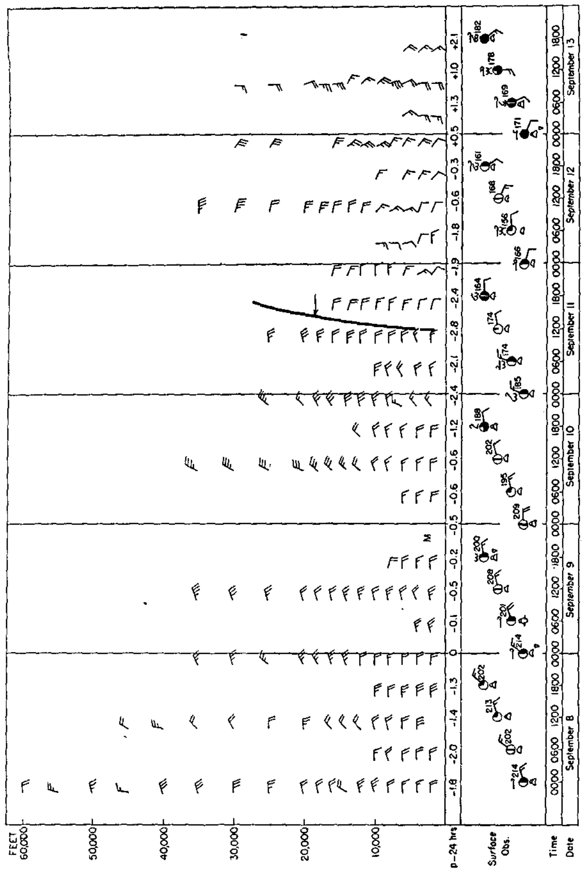

FIG. 3. Vertical time section at Midway Island, 8–13 September 1945.

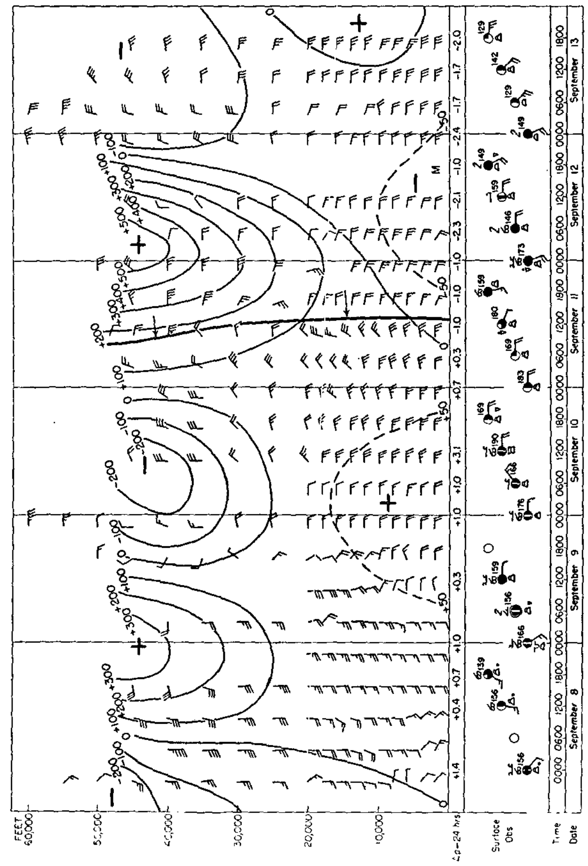

FIG. 4. Vertical time section at Iwo Jima, 8–13 September 1945.

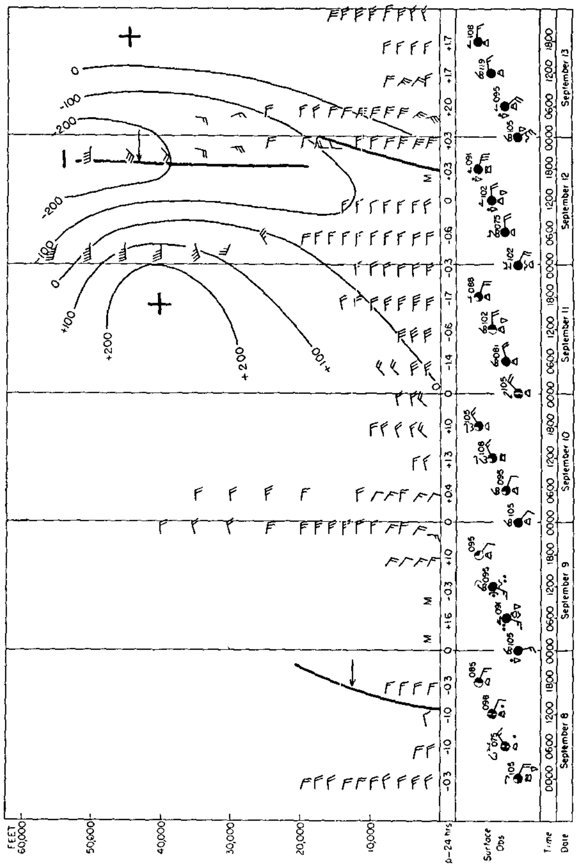

FIG. 5. Vertical time section at Kwajalein, 8–13 September 1945.

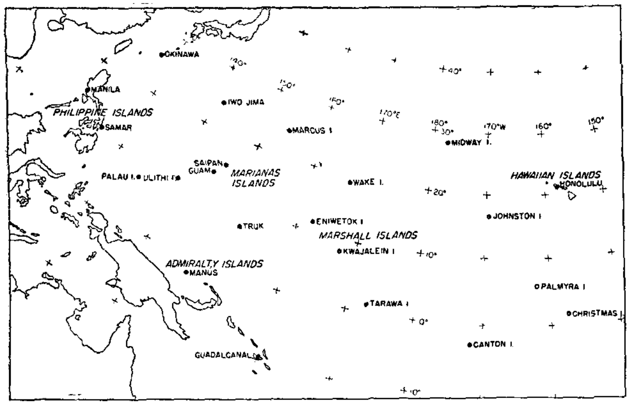

FIG. 1. Name and location of principal island groups and stations used.

ft have been plotted on the 1000-ft charts, and reports between 7000 and 11,000 ft on the 10,000-ft charts. These observations were made at variable times in the course of the day (roughly 2100–0900Z[3]), so that the time difference between reports explains some discrepancies among the winds. Errors in wind calculations account for others. Wind measurements by wind star (triple drift) were infrequent. Commonly, winds were calculated by Loran or single drift, or they were merely estimated. Nevertheless, most reports fit quite well together when coverage is dense, and only a few obviously incorrect reports had to be deleted.

Streamline analysis has been used on the maps. As in earlier reports (Riehl, 1947; 1948) the streamlines are defined by the tangent to the wind; the spacing as drawn is not a function of wind speed. However, some qualitative allowance for convergence and divergence has been made by ending and beginning streamlines in certain areas. On the 200-mb charts, contours are used in the north, and streamlines at 40,000 ft are used in low latitudes. As pointed out previously (Riehl and Schacht, 1946), this is permissible since the height of isobaric surfaces is nearly constant in the tropics. Contour values are given in hundreds of feet. The letter A stands for anticyclonic circulation, and C for cyclonic circulation.

The time section, the standard supplementary tool of analysis in regions with sparse data, was employed extensively. These sections contain surface observations, 24-hr surface pressure changes, upper-wind reports, and isolines of 24-hr changes in the height of isobaric surfaces, as described by López (1948). These height changes are given in feet. The reader will notice many irregularities in the 24-hr surface pressure changes at some of the stations. These irregularities must be ascribed to inexact barometry, a fact which has also necessitated recalculation of all radiosonde data. In the upper-wind reports, a long thin barb represents 10 mi hr$^{-1}$, a short barb 5 mi hr$^{-1}$, and a heavy triangular barb 50 mi hr$^{-1}$.

Fig. 1 contains name and location of the principal stations where upper-air observations were made. The wind data at 200 mb which are plotted below some of the maps, are those of a station located near 22°S, 166°E.

## 2. The weather situation 11–13 September 1945, near the surface

In the subsequent sections, a representative instance of typhoon formation, 11–13 September 1945, will be examined in some detail, beginning near the ground and gradually ascending to upper levels. It should be emphasized that choice of this particular period is arbitrary. Numerous other cases were also available.

On September 9 (fig. 10), a mature typhoon is moving northwestward east of the Philippines, and there is a complex system of shearlines to its north. The southwesterly current flowing from the Dutch East Indies toward the typhoon is partly deflected to the right, in accord with classical concepts. Obviously, however, it is not a "monsoon" current, if the term monsoon is used in its precise sense. A true monsoon is generated by a pressure gradient developed in response to differential heating between two large areas, for instance a continent and an ocean. In our case, the right branch of the southerly current is turning away from the warm Asiatic continent with time, while the left branch is a part of the typhoon circulation east of the Philippines. The name "Southwest Monsoon," so frequently used, should be abandoned for southwesterly currents blowing away from the Asiatic continent in the northern-hemisphere summer. Between longitudes 135°E and 150°E there is a clearly marked frontal zone south of the Marianas, but we must interpolate between longitudes 150°E and 160°E. A wave trough in the easterlies had passed Kwajalein on September 8 (fig. 5) and reaches Eniwetok on the 9th (fig. 7). Southerly wind components and precipitation at these stations therefore are connected with this trough which is advancing westward steadily on the 10th, at the same time that winds return to easterlies in the Marshalls.

Farther southeast, data are very sparse. Easterly winds prevail at Canton, Tarawa, and Guadalcanal; and there are no aperiodic changes of wind direction except introduction of a northerly component near longitude 170°E on the 11th in advance of the next wave trough in the easterlies (fig. 6). Certainly there is no evidence of northward flow across the equator in this area. Some crossing may be taking place between longitudes 160°E and 170°E, but it can hardly be of importance. The vorticity of the air just south of the equator can differ by very little from that to the north,

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[3] Z is used to represent Greenwich mean time.

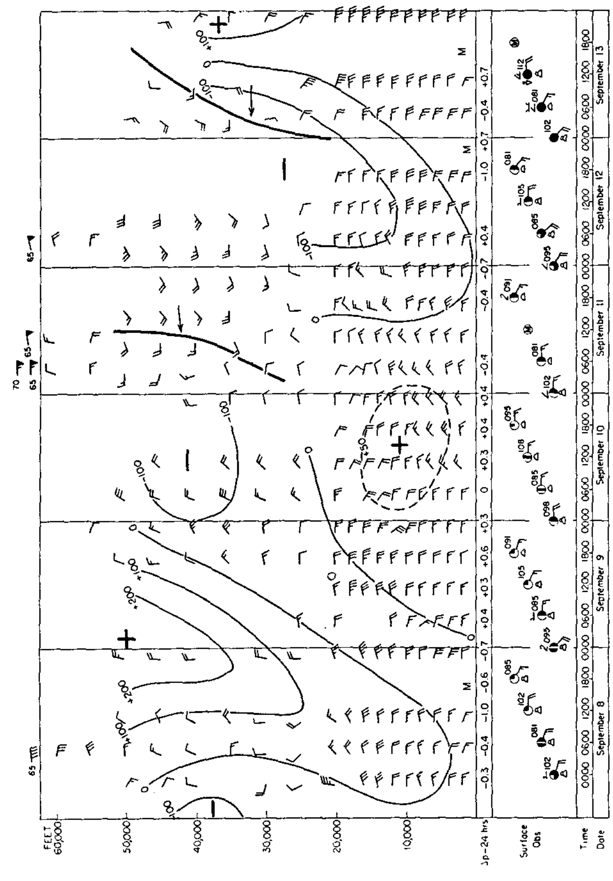

FIG. 6. Vertical time section at Tarawa, 8–13 September 1945.

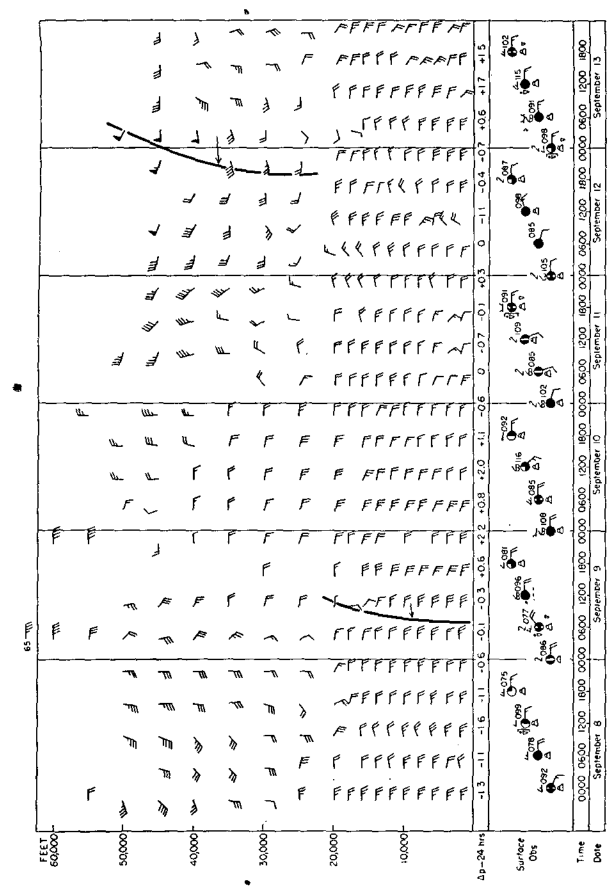

FIG. 7. Vertical time section at Eniwetok, 8–13 September 1945.

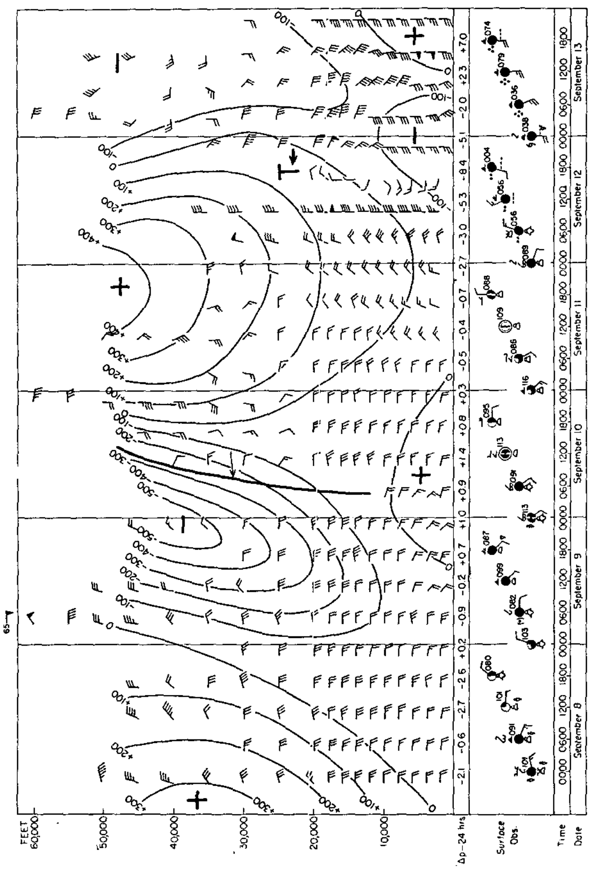

FIG. 8. Vertical time section at Guam, 8–13 September 1945.

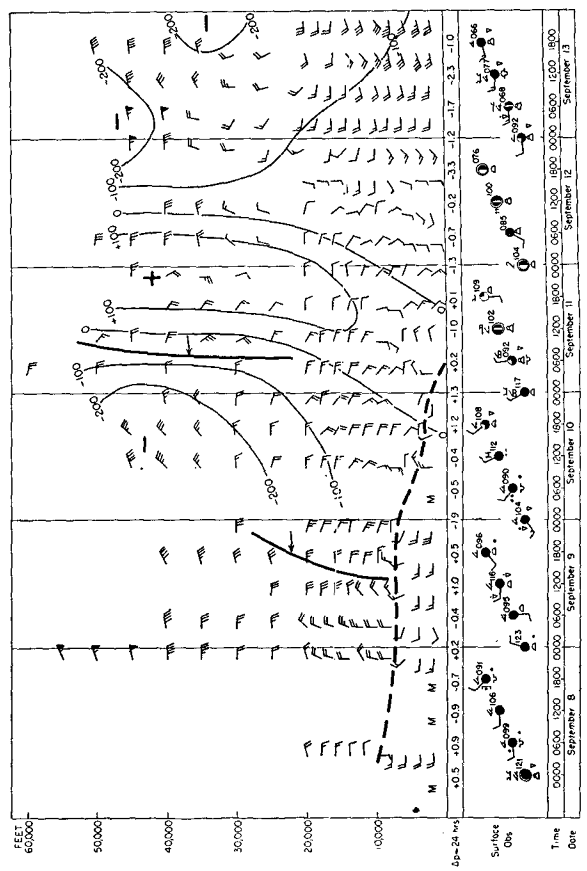

FIG. 9. Vertical time section at Palau, 8–13 September 1945.

since the former current is a portion of a broad, apparently quite uniform easterly stream. The hemispheric origin is of importance only when a current advances against the equator at a relatively large angle, importing vorticity from regions far to the south of the equator, as may be taking place over the East Indies. If a broad easterly current extends along the equator, nothing is gained by referring to hemispheric origin of the air adjacent to the equator. Small meridional displacements of such a current can mean no more than displacements across any latitude circle. Clearly, there is no equatorial front east of longitude 170°E. The excellent streamline charts for the Pacific Ocean published by Werenskiold (1923) confirm that even in the average there is an interruption of the equatorial convergence zone in the Central Pacific.

An interesting possibility remains: the air east of the wave trough at 160°E might be derived from the westerly current south of the Marianas. This is ruled out by the data of the preceding day (not reproduced), by the steady westward progression of the wave trough, the time section at Kwajalein, and events in the Admiralty Islands. On September 9 the wind at Manus turns gradually from southeast to northeast and becomes northwest on September 11. This, taken together with the other wind data, suggests that there is a clockwise-rotating eddy north of New Guinea and that this eddy is contracting with time, or moving westward. In that case, the equatorial front must turn sharply southward east of Truk, forming the boundary between the eddy and the easterly current of the central Pacific Ocean. Instead of an equatorial front extending all the way across the Pacific Ocean, *we find in the equatorial trough a system of eddies*. Such eddies may develop anywhere within the equatorial trough, although perhaps the location north of New Guinea is preferred. Another formation has been indicated south of Johnston on September 10 and 11. The extreme variability of rainfall in the central parts of the equatorial Pacific Ocean, which is well known, also points to an eddy structure within the equatorial trough, a subject first discussed by Brooks and Braby (1921).

Progressive weakening of the eddy north of New Guinea is shown by reversal of winds from westerly to easterly direction between September 9 and 10 west of Truk. At the same time, the depth of the westerly current decreases at Palau (fig. 9) until the westerlies disappear completely on September 11. The trades are advancing everywhere in the western Pacific, and the equatorial front retreats between September 9 and 11 (refer also to displacement arrows in fig. 12). At the same time, a remarkable development is taking place east of the Philippines. As the typhoon initially located east of Luzon continues on its northwestward path, the pressure field that had been drawing air northward from the East Indies disappears. With elimination of the previously strong pressure gradient along the streamlines, the southerly current seeks to continue under conservation of vorticity and turns clockwise. The incipient stages of this turning are indicated by development of a new shearline east of the Philippines on September 10 (fig. 11). One day later (fig. 12) the turning has actually been accomplished, splitting the broad southerly flow that was observed on September 9. A new clockwise-rotating eddy is formed, and very little air can penetrate northward to its east. It is against this background of weakening westerly circulation and an equatorial front south of the Marshalls–Philippines flight route that typhoon formation starts near Truk on September 11.

It is not possible to detect any instability in the wave trough advancing from the Marshalls toward the Marianas on September 10. Even in the noon hours of September 11 general deterioration of weather had not yet begun, although aircraft encountered cumulonimbus and heavy rain near nightfall. But the windfield near map time of the 11th is remarkable. A portion of the air to the rear of the wave trough west of the Marshalls has suddenly ceased to move westward through the troughline; instead, it has turned sharply clockwise, forming a new shearline at the forward edge of the portion of the current that is turning. The sudden clockwise turning evidently takes place entirely within northern-hemisphere air.

Developments of this kind were first noted by the writer during the period 30 July–2 August 1947, under very similar circumstances. On August 1, Guam was situated in the region between old and new troughline, but there were not enough data to substantiate the analysis. A reconnaissance flight, in which the writer participated, established the position of the western principal shearline, while the eastern system moved westward across Guam during the following night. Typhoon formation in the main trough ensued at once. Since that occasion it has been possible to observe the remarkable turning of the trades in numerous instances just prior to typhoon formation. It is not suggested, however, that such turning must always precede deepening.

When a system of waves in the easterlies moves westward in steady state with a speed less than that of the basic easterly current (Riehl, 1945), the air overtaking the trough from the east is subjected to convergence east of the troughline and is therefore able to advance westward through the trough on a trajectory with counterclockwise curvature. In the cases just mentioned, the controls which the wave trough and its associated pressure field normally exercise over the trade stream are radically superseded

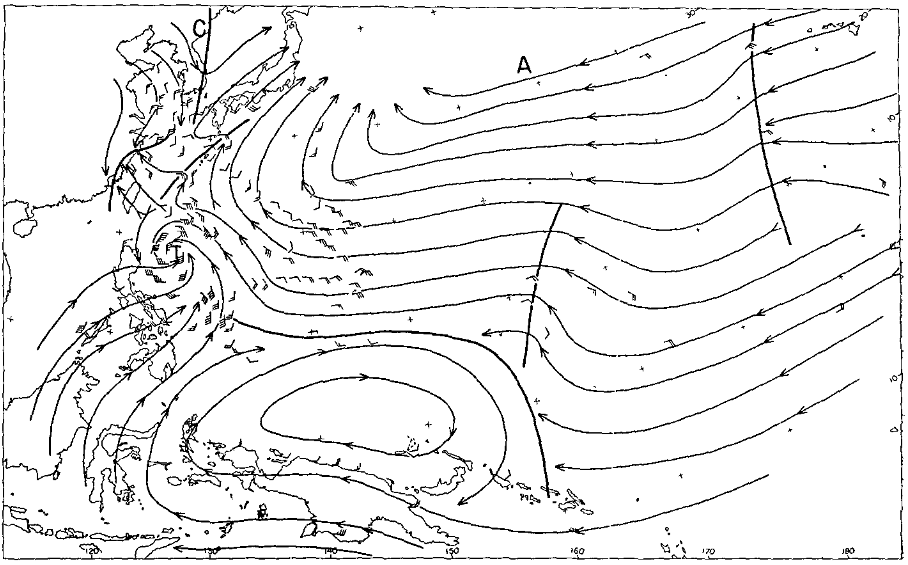

FIG. 10. Streamlines at 1000 ft, 9 September 1945, 0000–0600Z.

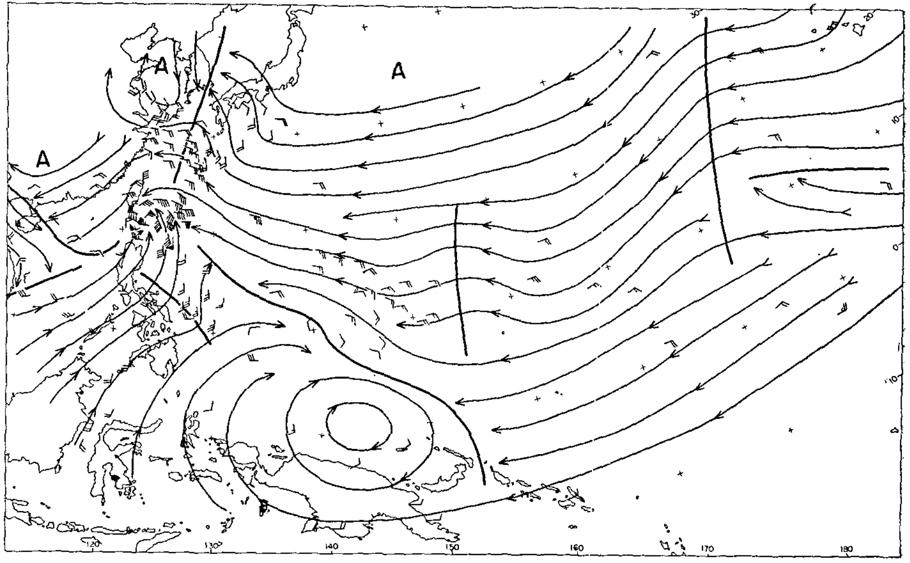

FIG. 11. Streamlines at 1000 ft, 10 September 1945, 0000–0600Z.

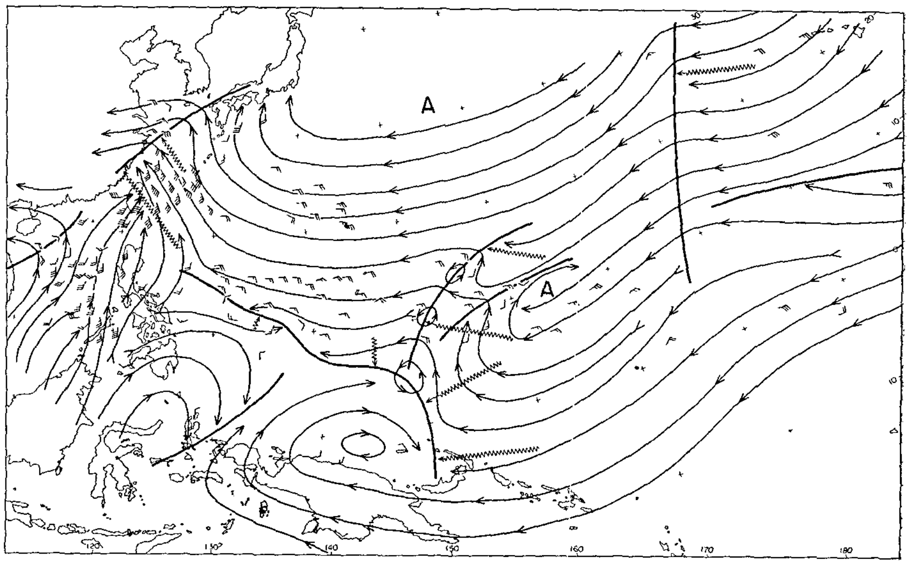

FIG. 12. Streamlines at 1000 ft, 11 September 1945, 0000–0600Z.

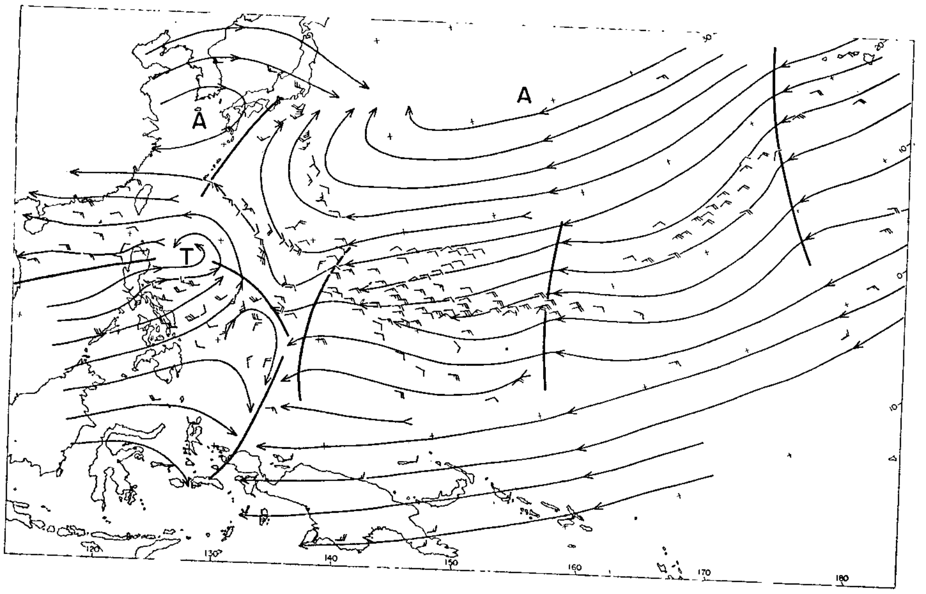

FIG. 13. Streamlines at 10,000 ft, 9 September 1945, 0000–0600Z.

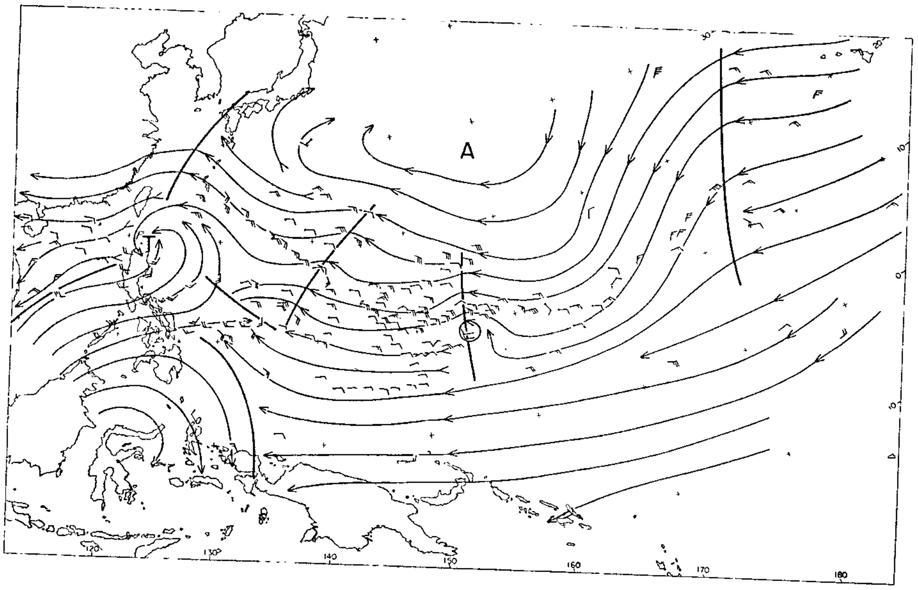

FIG. 14. Streamlines at 10,000 ft, 10 September 1945, 0000–0600Z.

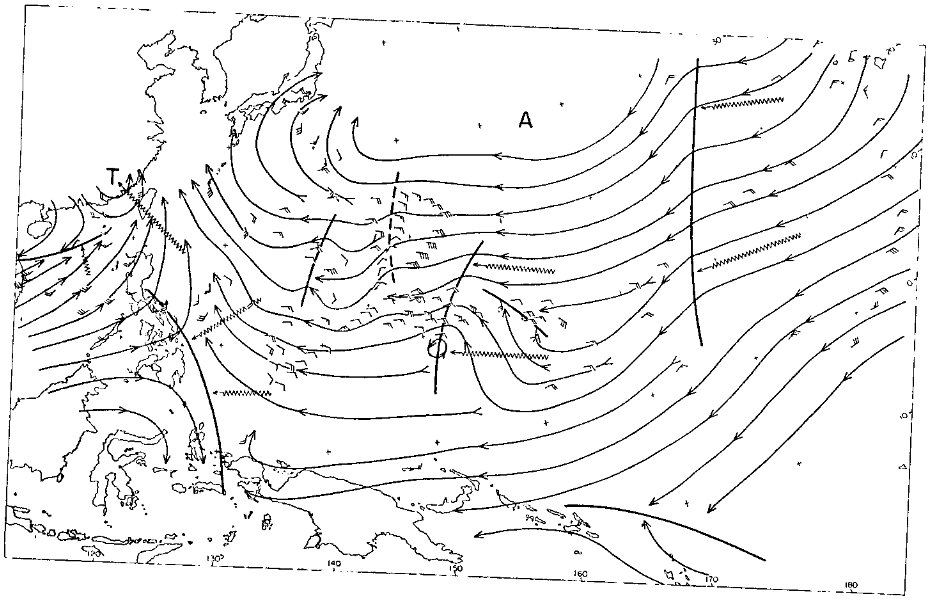

FIG. 15. Streamlines at 10,000 ft, 11 September 1945, 0000–0600Z.

by some agency which is diverting the current back toward the east. As steady state ends within the wave trough, vortices readily begin to form.

The three centers entered along the troughline on September 11 (fig. 12) are of course unsubstantiated. They are meant merely to convey a schematic picture of the circulation frequently observed during the initial transitory phase of instability. The subject of simultaneous generation of several vortices was taken up earlier (Riehl, 1948), and Hubert[4] has a very well-documented example of this type of occurrence. In the case of the cyclogenesis of September 11, reconnaissance planes found a single intensifying cyclone on the following day. Later on September 12 this center passed just north of Guam while deepening rapidly (fig. 8).

In conclusion, we may review the principal points noted: The equatorial trough does not contain a simple equatorial front along its entire length but rather a system of eddies which is alternating with broad easterly flow on both sides of the equator. Typhoon formation started in a wave trough in the easterlies at a time when the low-level equatorial westerlies were collapsing north of New Guinea. The first sign of instability was sudden clockwise turning of one portion of the trade stream. However, with the exception of the development of the new clockwise eddy just east of the Philippines, it has not been possible to explain any of the features observed, especially the veering of the trade and the deepening. In fact, after noting the great weakening of the equatorial westerlies, we must discount one factor frequently called upon to account for instability: the juxtaposition of two air currents with different momentum. Let us now turn our attention to higher layers in the atmosphere.

### 3. The situation at 10,000 ft

The density of reports is greatest at 10,000 ft where, in some parts of the maps, it equals that in the United States. On September 9 (fig. 13) only one major difference exists between flow at 10,000 ft and 1000 ft. The low-level anticyclonic eddy north of New Guinea does not extend to 10,000 ft, as proved by many winds on September 9 and 10. At that level there is no equatorial front in the conventional sense at all. It is not an object of this report to investigate the eddy south of the Marianas in more detail. Exhaustive studies on applicability of the gradient-wind equation and the magnitude of the acceleration terms in the equations of motion at very low latitudes are needed before the wind field associated with such eddies can be taken up in relation to the pressure and temperature fields. Here, we shall be content to note the shallowness of the westerly stream. The time section at Palau (fig. 9) illustrates strikingly how quickly the winds shift at the boundary, even as this boundary lowers toward the surface on September 10.

There are some other differences between 10,000- and 1000-ft charts. At 10,000 ft, the flow over the East Indies and southern Philippines is nearly westerly, whereas it has a larger component from south at 1000 ft. Again, it seems best not to attempt any interpretation of this clockwise turning of wind with height, except to note that presumably the field of acceleration connected with the typhoon decreases in strength between 1000 and 10,000 ft. That would be entirely in accord with what is generally known about the vertical structure and circulation of hurricanes.

Already on September 9 clockwise turning of the equatorial westerlies takes place east of longitude 125°E at 10,000 ft. At their forward boundary a shearline, sharply defined by easterly and westerly winds meeting nearly head-on, is situated between longitudes 130°E and 135°E. As in the surface layer, there is a tendency for the equatorial westerlies to retreat between September 9 and 11 (fig. 15). Westward passage of the shearline at Palau (fig. 9) is clearly marked.

It is necessary to take up briefly the two reports of west wind northeast of Truk on September 10 (fig. 14). These winds would suggest that the wave trough located there already becomes unstable on that day. However, there is no supporting evidence, either in the surface pressure field or the weather reported. All data indicate that deepening did not start until the evening of September 11, but that it was rapid from that time onward. Although the two westerly winds might be incorrect, we should not rule out the possibility that small-scale vortices may exist in stable wave troughs. In fact, the writer has noted several instances of this kind in the Caribbean area.

On September 11, veering of the trades east of the wave trough takes place at 10,000 ft as at 1000 ft. However, the newly formed shearlines do not lie parallel at the two levels, but nearly perpendicular. It is not possible to tell whether this is significant, merely transitory, or due to time difference between reports. It is certain that the amplitude of the principal trough has increased. On the map as a whole the streamline amplitude at 10,000 ft somewhat exceeds that at 1000 ft as should be expected (Riehl, 1945). On September 11 northerly wind components between the Marshalls and Tarawa are pronounced, again emphasizing that no southern-hemisphere air is being transported across the equator east of the deepening wave trough.

In conclusion, we may say that the wealth of observations available at 10,000 ft lends support to the deductions made at 1000 ft. The flow pattern at

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[4] L. Hubert, "High tropospheric westerlies over the equatorial north Pacific Ocean," 1948 (to be published).

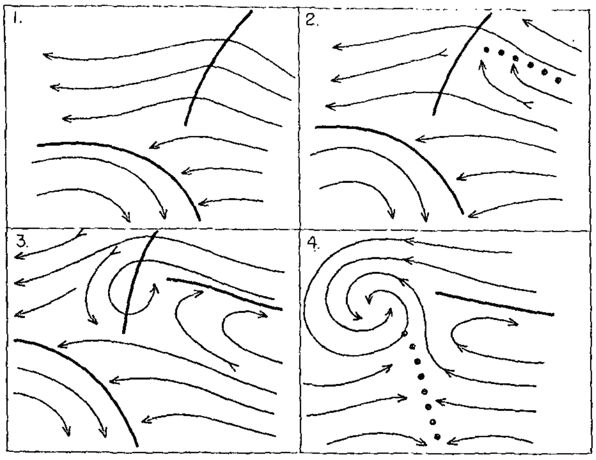

Fig. 16. Model of four stages during typhoon formation in the low troposphere, when the development takes place in the northern-hemisphere trade.

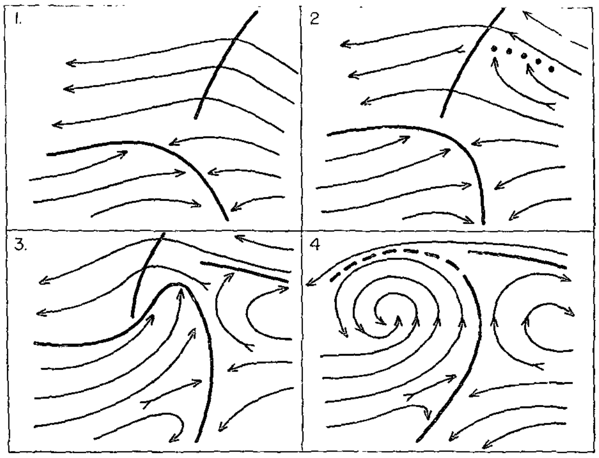

Fig. 17. Model of four stages during typhoon formation in the low troposphere, when the development takes place in the equatorial westerlies.

both levels differs mainly in that the eddy north of New Guinea does not extend to 10,000 ft, demonstrating its shallowness as well as that of the equatorial front. Instability of the trades on September 11 is apparent at both levels simultaneously. The low troposphere as a whole, however, does not appear to supply a clue as to the cause of the instability.

Before proceeding, it is desirable to summarize in model form the sequence of events that takes place when a tropical storm forms after the trades have become unstable. It should be repeated that due to scarcity of data in many situations it is not possible to ascertain how universal such instability is. It is merely depicted here as a frequent event.

The first model (fig. 16) shows four stages during typhoon development when the equatorial westerlies are shallow and do not enter the cyclonic circulation until the storm is well established. This is the course of events observed September 11–13, except that the transitory phase with several small vortices has been omitted. Westerly winds such as appear over Palau on September 12 evidently are produced by the pressure field associated with the storm. They do not precede it. *The developing typhoon forms within one air mass only—the northern-hemisphere trade.*

The second model (fig. 17) depicts a different, yet very frequent sequence. Its occurrence is to be expected when the equatorial sector is much more active and enters the circulation at an early stage. As the trades veer, the equatorial westerlies are drawn at once into the region of deepening. Fig. 17(3) is analogous to the classical picture, except that east of the low-pressure center *the equatorial westerlies do not underrun the trades but move northward side by side with them.* Again, the vortex formation is restricted to one air mass alone, in this case the equatorial westerlies.

Two qualifying statements are necessary. The veering of the trades is not observed near the sea surface. The mixed layer nearest the surface, characterized by almost constant potential temperature and mixing ratio,[5] converges toward the area of deepening from all sides. Veering occurs mainly above the base of the low clouds, and the level at which it begins is variable.

In both models, the clockwise-rotating eddy east of the forming storm has been maintained throughout the period of deepening. There is evidence that this does not always happen. In some situations the eddy disappears as the typhoon strengthens. In other instances, however, the eddy takes on large dimensions and persists throughout the life of the storm, moving northward with it.

## 4. The situation at 200 mb

A glance at any of the 200-mb charts reveals that at this level the flow pattern differs radically from that of the lower troposphere. In place of the broad trade stream we find a system of vortices south of the belt of polar westerlies. The dimensions of these vortices, both cyclonic and anticyclonic, are surprisingly large. Some have a radius of more than ten degrees of latitude. It is apparent that they are phenomena on the scale of the disturbances in the polar westerlies.

It was Bigelow (1904) who noted long ago, from cirrus observations in the Atlantic, that a complex system of anticyclonic eddies overlies the trade stream. An amplification of his observations has been given by the writer (Riehl, 1945; 1948), but high-tropospheric data were too scant even during the recent war years in the Atlantic ocean to obtain an accurate picture of the high-level flow except on rare occasions. In the Pacific, charts at 40,000 ft were first prepared

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[5] J. Wyman and A. H. Woodcock, "Vertical motion and exchange of heat and water between the air and the sea in the region of the trades." Woods Hole Oceanographic Institution, 1946 (mimeographed).

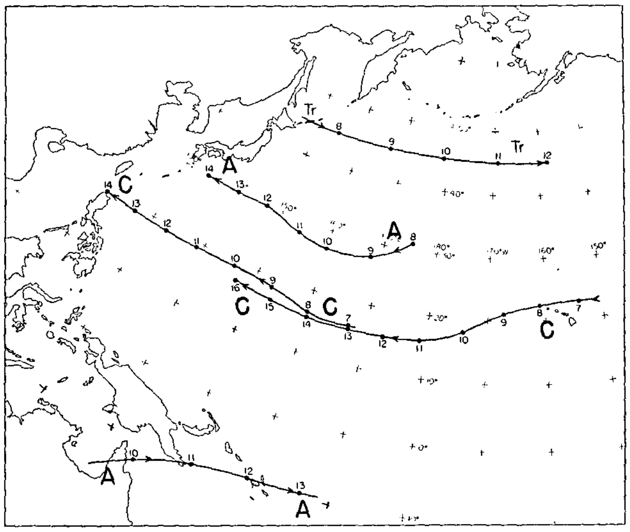

FIG. 18. Continuity chart at 200 mb, 7–16 September 1945.

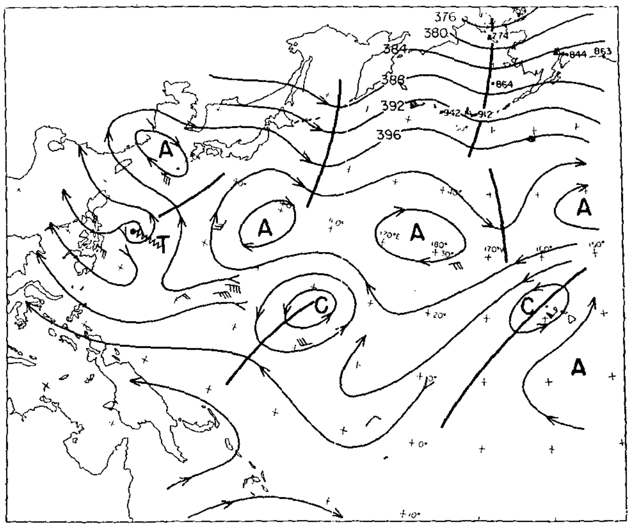

FIG. 19. Streamlines on the 200-mb surface, 8 September 1945, 0000Z.

by Lt. Col. R. J. Shafer of the 20th Air Force Weather Central at Guam and then by D. Fultz,[6] when he was stationed at Guam in 1945. Since it is reasonable to hope that the high-tropospheric vortices may provide answers to some of the questions raised earlier, we will examine them now in more detail.

At first, because of the wide scattering of the reports, it is necessary to give some justification for the analysis as carried out. It has been assumed that the upper streamline patterns are to some degree indicative of the pressure distribution except within three to five degrees of the equator and the immediate vicinity of hurricane centers aloft. This assumption is supported by the observed correlation between variations in pressure and wind field as demonstrated, for example, in fig. 8. Of course, it is not claimed that winds are strictly gradient, but merely that directional departures from gradient flow are far less than 90 degrees. Above latitude 5°N, the deviations probably are less than 30 degrees.

Besides maps and time sections, two additional analysis tools were employed in determining the most probable streamline pattern.[7] One was the analysis at 20,000 ft carried out at the Army Forecast Center at Guam. This analysis helped to obtain some idea concerning the high-level flow pattern over Japan and vicinity. The other aid was the well-known continuity chart (fig. 18). Three major systems were tracked on that chart during the period studied in the tropics and subtropics. To put it more exactly: If three major

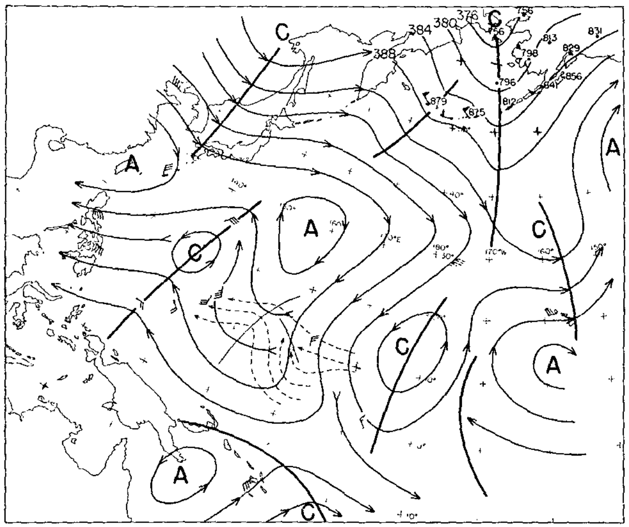

FIG. 22. Streamlines on the 200-mb surface, 11 September 1945, 0000Z.

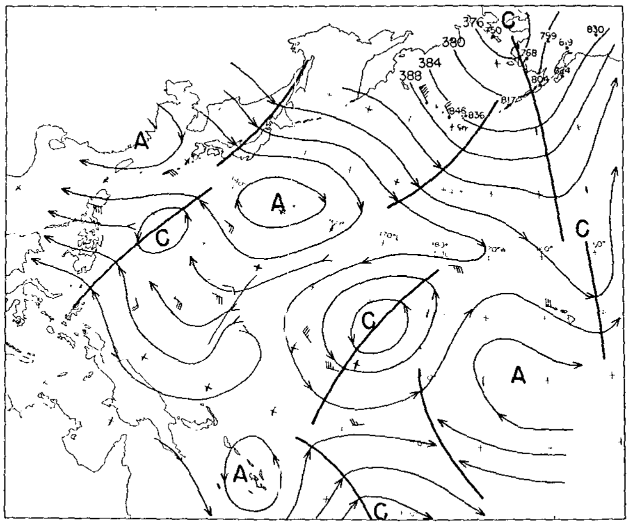

FIG. 23. Streamlines on the 200-mb surface, 12 September 1945, 0000Z.

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[6] Of the Department of Meteorology, University of Chicago.

[7] The 200-mb charts were analyzed by Dr. D. Fultz and the writer separately. Figs. 19–25 show a joint solution.

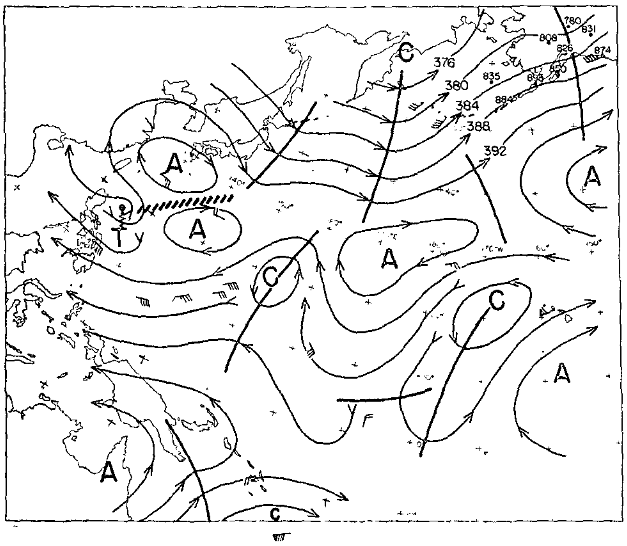

FIG. 20. Streamlines on the 200-mb surface, 9 September 1945, 0000Z.

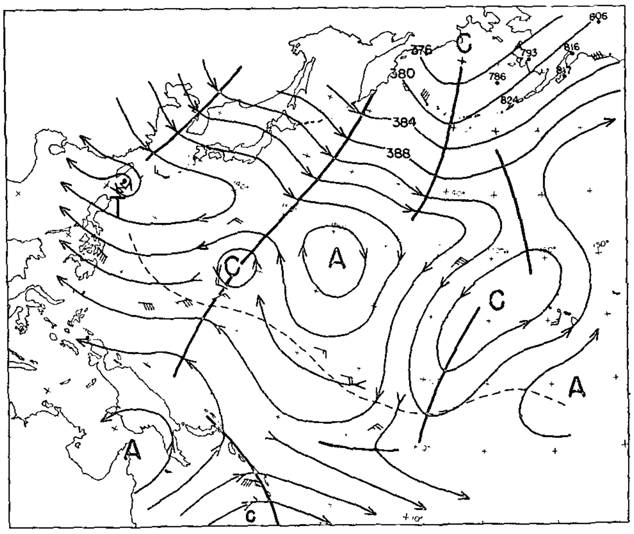

FIG. 21. Streamlines on the 200-mb surface, 10 September 1945, 0000Z.

systems—two cyclones and one anticyclone—are moved westward at constant speed, namely at the speed shown in fig. 18, their motion agrees with the observed windshifts at 200 mb at all stations within the network. It is hardly believable that this could be a mere coincidence. Extension of the tracks outside of the period illustrated provided further continuity. While a variety of streamline patterns can be drawn to fit the data on any particular day, it is not likely that there are several alternate solutions which maintain good continuity throughout the period. The probability is high that the major features of the patterns reproduced closely portray the events which took place.

If this assertion is acceptable, it follows from fig. 18 that upper vortices can move at nearly constant speed across the largest part of the Pacific Ocean; and they can have a life span in excess of one week and possibly much more, for it is impossible to tell how far their trajectories extended east and west of the limits of our map. Such steady motion of course is possible only when the latitudinal boundary between tropical and polar zone does not suffer large displacements. Throughout our series, this boundary remains roughly near 35°N, to judge by fig. 18. If the boundary undergoes large variations in latitude, or if it is located at an appreciably lower latitude, steady westward motion in the tropics is not to be expected because of interference of the westerlies. This is brought out by the maps presented by Hubert.[4]

To proceed with particulars of the analysis: Eastward motion of a principal and a secondary trough

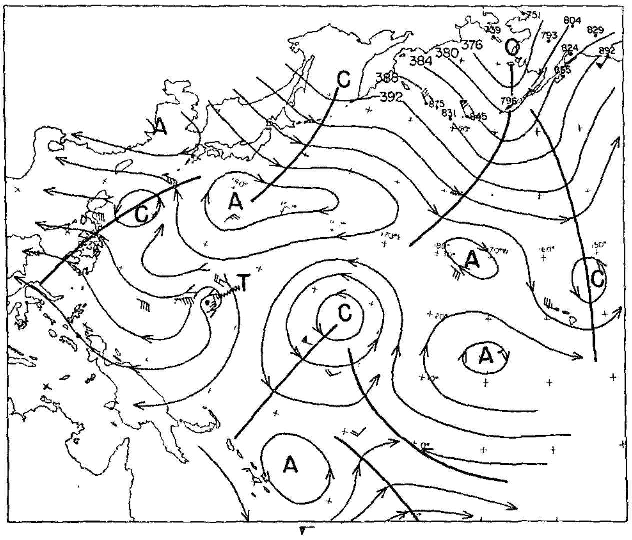

FIG. 24. Streamlines on the 200-mb surface, 13 September 1945, 0000Z.

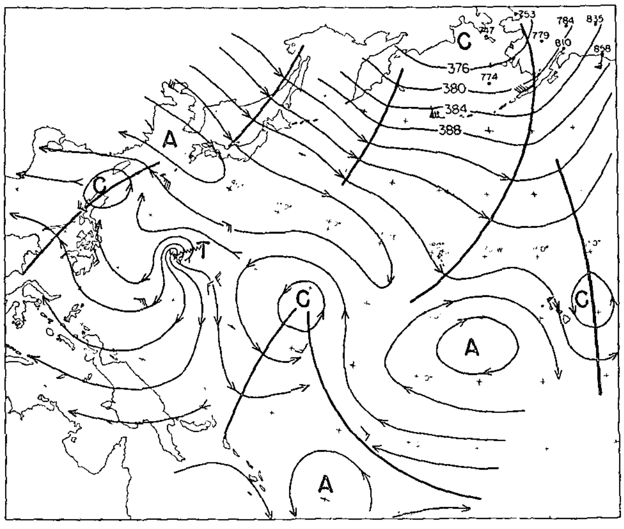

FIG. 25. Streamlines on the 200-mb surface, 14 September 1945, 0000Z.

in the westerlies across the Aleutian chain is easily established from the data. The eastward displacement of this trough at 45°N is shown in fig. 18. Appearance of a new trough over Japan in the last days of the series is suggested by the Guam 20,000-ft charts as well as by what is generally known about spacing of disturbances in the westerlies.

At Midway (fig. 3), deep easterlies prevail September 8–13. Thereafter winds shift quickly through south into northwest, suggesting that the secondary trough in the westerlies is increasing in intensity and extending to lower latitudes as it moves eastward. The turning of the wind at Midway, together with the Aleutian and Alaskan data, hints that the troughline of a long wave in the westerlies (Cressman, 1948) has become more or less stationary between longitudes 160°–170°W in the later part of the map series.

Westward motion of the cyclonic center initially located north of Eniwetok is attested by the sections of Eniwetok, Guam, and Palau on the south side, and by Iwo Jima on the north side. Winds at Okinawa, Manila, and Samar also confirm its westward displacement. This center had passed Honolulu on September 1.

The movement of the second cyclonic circulation is confirmed by Honolulu, Midway, Kwajalein, and Eniwetok data. Passage at Guam and Saipan on September 16 was very pronounced. Both cyclonic centers seemed to attain their greatest circumference when situated north of the Marshall Islands.

At Tarawa (fig. 6), winds shift in accordance with events farther north up to September 13. This is interesting, since Tarawa lies almost on the equator. On September 13 and 14, however, a rapid turning to southeast takes place, and similar turning is noted at Kwajalein and Eniwetok. Reports at Guadalcanal and at a station located at 22°S, 166°E, made it possible to show northeastward advance of a southern-hemisphere shearline during the preceding days, followed by an eastward moving anticyclonic center. The windshift at Tarawa might denote passage of this shearline. Because of the associated turning at Kwajalein and Eniwetok, however, it seemed preferable to adopt the analysis which is shown, though it is impossible to be certain. The same holds for the entire equatorial region east of Tarawa. The analysis there was prepared following Hubert,[4] who had high-level winds at least at Canton. In our period only one wind at 25,000 ft was recorded there, which was westerly. Aperiodic changes of wind direction at high levels over Canton seem to be quite frequent.

It is particularly problematic whether easterly winds really exist at 200 mb over Palmyra and Christmas Island. There is nothing to rely on except the fact that it is necessary to determine a source for the westerly stream constantly passing over Hawaii (fig. 2). Unless the entire current observed there has passed around the low-pressure center west of Hawaii from September 8 to 12, advection from the south must take place. If the transport of air across the equator is small, the current over Honolulu must be derived from an anticyclonic eddy south of the Hawaiian Islands. In any case, the westerlies over Honolulu should be classified as "equatorial westerlies."

Reference should also be made to the dashed line in fig. 21, which represents the equatorial front as copied from the official analysis on September 10. It is clear that even if this line were correct, good forecasting would be difficult, unless the 200-mb chart is available. But we have already seen that there is no equatorial front east of the Marshalls.

### 5. Structure of the trade-wind belt

Although there are just enough winds at 200 mb to gain a fair idea concerning the broadscale pattern over the largest part of our map, winds and radiosonde observations are too scarce to provide a complete description of the high-tropospheric vortices. We can outline only a few of their general characteristics. Figs. 26 and 27 give the average variation of the east–west component of the wind with height, as obtained from the arithmetic mean of rawin observa-

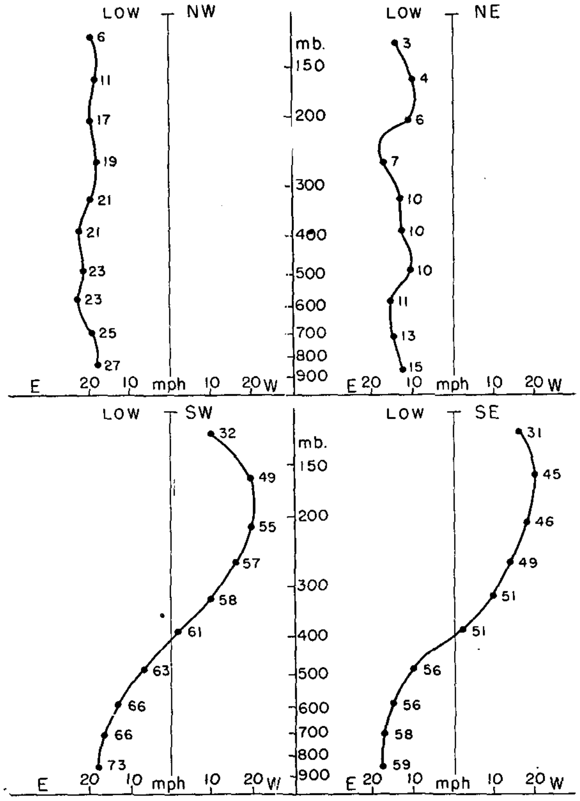

FIG. 26. Vertical profile of the zonal wind in each quadrant of the high-tropospheric cyclones, September 1945. Number of observations at each level is indicated.

tions for each quadrant of several centers noted during September 1945. There are many more observations in some quadrants than in others, due to station distribution. The curve for Low–NE (fig. 26) is particularly irregular because of scarce data. The following principal facts, however, stand out. The low-tropospheric easterlies begin to decrease with height near 500 mb, or even lower, south of high-level cyclones and north of upper anticyclones. The base of the westerlies is at 400 mb below the cyclones and somewhat higher below the anticyclones. The westerlies are strongest between 200 and 150 mb, and there the upper systems attain their greatest intensity. For this reason the 200-mb level has been chosen as the basic level for high-altitude analysis.

North of upper lows and south of upper highs the atmosphere is more barotropic. The easterlies do not increase with height anywhere; in fact, they slowly diminish upward in strength in the anticyclones. This, however, may be accidental. As a first approximation we may say that wind speed around the centers is fairly constant at 200 mb. Since the vortices are centered near 15°N or farther north, winds presumably are nearly gradient except in the southernmost portions. In view of the observed wind speeds and the rate of displacement of the systems, their 'angular amplitude' should vary between 5 and 8 degrees of latitude, according to the calculations of Bjerknes and Holmboe (1944). This is indeed the observed order of magnitude.

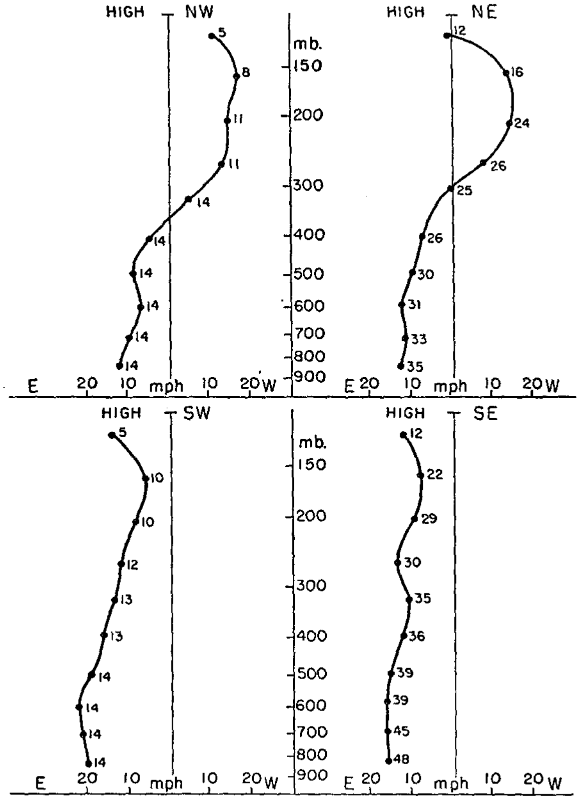

Fig. 27. Vertical profile of the zonal wind in each quadrant of the high-tropospheric anticyclones, September 1945. Number of observations at each level is indicated.

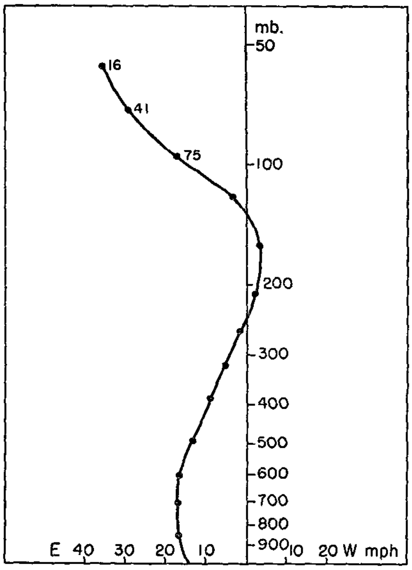

Fig. 28. Vertical profile of the basic zonal current, September 1945. Computations include stations between 8°N and 25°N, 160°W and 120°E. Number of observations in the stratosphere is indicated.

Fig. 28 shows a composite wind curve obtained by computing the arithmetic mean of the eight sectional curves of figs. 26 and 27. Combining the upper-wind data in such a form should lead to cancellation of all winds associated with perturbations, so that only the basic zonal current remains. The curve also has been extended as far as possible into the stratosphere, using all available data for September 1945. The number of observations used has been indicated above 50,000 ft.

The basic easterly current is entirely barotropic only up to 600 mb. Above that level it decreases slowly, but steadily, with height. Between 240 and 140 mb there is a light westerly wind component. This component, however, is so small, that it is probably best to say that the basic current is zero between 300 mb and the tropopause. Extreme caution is necessary in putting forth any explanation for the upward decrease of the easterly current. It should merely be noted that, if the zonal mean motion can be considered as geostrophic, an associated northward-directed temperature gradient of about 0.5C per 5 degrees of latitude must exist between 600 and 200 mb.

This vertical structure of the basic current does not apply everywhere within the tropics and probably does not even persist without interruption throughout the summer over the trade-wind belt. In our series we note that at Palau (fig. 9) the easterlies increase with height through the troposphere. Hubert[4] has observed high-level westerly winds near the equator that are not directly connected with closed circulations.

Observations near the tropopause level (50,000–55,000 ft) and in the stratosphere are scant but significant. Nearly all rawins that enter the stratosphere show resumption of general easterly flow there, irrespective of the lower flow pattern. Moreover, deviations from the easterly direction are small. For this reason it is permissible to combine the stratospheric observations with the mean curve for the troposphere as determined here.

There was only one station where stratospheric winds were variable and sometimes even westerly. This is Canton Island, located near 3°S. Comparison with Batavia (Shaw, 1930, p. 272) shows that at that station westerly winds near 20 km have also been observed in July, while the easterlies attain their greatest speed near 15 km. This is the inverse of fig. 28 and intimates at least the possibility that within the inner equatorial zone a zonal wind field exists, which is very different from that of the trade-wind belt. Such a different structure, however, was not reported by the "Meteor" expedition (Kuhlbrodt and Reger, 1933). The upper-air circulation observed during that voyage is in very good agreement with the Pacific data presented here.

In the latitude belt dominated by the upper vortices, we may distinguish three principal layers: the low-level trades, which are barotropic between the surface and roughly 500 mb, and disappear near 300 mb; the high tropospheric layer between 400–300 mb and 100 mb, in which there is no basic current and which is characterized by large cyclonic and anticyclonic circulations; and the stratosphere, in which easterly winds of relatively high speed prevail. Aperiodic changes of wind direction are largest and occur most frequently in the high troposphere, in marked contrast to the stable easterly flow of the trades and the stratosphere.

## 6. Relation between high- and low-level flow

Our 200-mb charts pose many interesting questions concerning origin, kinematics, and dynamics of the high-level circulations. Their effect on the polar westerlies must be profound. In this report, however, we are concerned mainly with interaction between upper and lower parts of the atmosphere within the trade-wind belt and will confine our discussion to that topic.

It is unlikely that the stratosphere exercises much influence on circulations nearer the ground. As just noted, there is a strong tendency for undisturbed easterly flow to persist in the stratosphere. Rapid upward damping of the vortices observed at 200 mb therefore must take place through the tropopause, a conclusion which is indicated by the rather abrupt changes of wind with height in the tropopause layer. As in the case of cold lows and warm highs of middle latitudes, such damping is readily accomplished by variations in the height of the tropopause (Palmén, 1932). Over low-pressure centers, the tropopause must dip downward, and over highs it must be raised. At the low latitudes which we are considering, only small temperature changes are necessary to eliminate the closed circulations, and the undulations of the tropopause should be correspondingly small. This is especially true, since the tropopause usually is well marked, and since temperature begins to increase with height immediately above it. It is an observed fact that despite large vertical wind shear aperiodic changes in the height of the tropopause comparable to middle-latitude changes are never observed. In the following discussion, we will consider only relations between the high and low troposphere.

It is a remarkable fact that the high-tropospheric cyclones observed here not only move westward over several thousands of miles, but also move at the same speed, and the anticyclonic center between them is displaced at the same rate. The distance between the cyclones therefore remains nearly constant, and when a new counterclockwise circulation appears east of Hawaii on September 13 and 14, the distance between it and the vortex located in the Central Pacific is again roughly the same. This observation points to systematic relations between these centers. One is reminded of steady-state wave motion, except that in our case the perturbations are very large and no basic current is present in which the perturbations are embedded. The height of the 200-mb surface above any given station, of course, executes wave-like oscillations (fig. 8). If we can speak of wave length in connection with our disturbances, this wave length is 45 degrees longitude at 15°N, corresponding to a hemispheric wave or vortex number of 8. There is, of course, no evidence that a train of vortices girdles the entire globe. A hemispheric wave number of 8 at 15°N corresponds to a wave number of 5 at 50°N and 4 at 60°N. The tropical systems therefore have roughly the linear magnitude in absolute units of long waves in the polar westerlies.

A distinct difference will be noted when a comparison is made between high-level and low-level wave length within the trade-wind belt. The number of wave troughs in the lower easterlies on our 10,000-ft charts is far greater than the number of cyclones at 200 mb. The westward displacement of the upper vortices at 15°–20°N is about $6\frac{1}{2}$ degrees longitude per day, an amount very nearly equal to the speed of the lower easterlies. A given station, therefore, experiences passage of one upper 'wave' per week, in view of the wave length. In the lower troposphere, as mentioned elsewhere (Riehl, 1945), roughly two wave passages may be expected in a week. Our charts and time sections support this statement. These observations lend support to characterization of the low-tropospheric disturbances as true waves. While initial perturbations no doubt are derived very frequently from the upper troughs (Cressman, 1948a), a set of disturbances is found in the final state that has a frequency entirely different from that apparent at high levels. Corresponding to different properties of the atmosphere below 500 mb, a different type of low-level perturbation is excited by the upper disturbances.

By comparing high and low latitudes, we find systems of long and short waves in both regions. The short waves in the lower easterlies correspond to systems of the wave length of wave cyclones in higher latitudes. It is well known that these cyclones are closely connected with warm troughs and cold ridges, therefore with the strong meridional temperature gradient existing in higher latitudes near the ground. In contrast, waves in the easterlies exist within a nearly barotropic current, and the temperature field accompanying these waves is largely produced by the field of vertical motion.

The high-tropospheric vortices of low latitudes are comparable to the long waves in the polar westerlies, and also to the extensive system of vortices that sometimes takes the place of open wave trains in extratropical regions. The long waves have been connected with inertia oscillations caused by latitude changes of air particles (Rossby, 1940), but little is known about the dynamics of systems of vortices. Here, we have been content merely to note certain observable characteristics of vortices and basic zonal currents within low latitudes, all of which differ greatly from the structure of high-latitude currents and long waves. It is interesting, however, to observe that it is much easier to distinguish between long and short waves in low than in high latitudes. The streamline configurations of our low tropospheric maps show only the short waves, and those of our upper maps reveal only the long waves. In contrast, in the belt of westerlies it is frequently very difficult to decide which is a major and which a minor trough. Methods for dealing with this problem on daily maps have been developed by Cressman (1948b). Most applications of Rossby's wave formula, however, have been made with use of 5-day mean charts (Namias, 1947), which have as one of their principal objects the elimination of the short waves. The tropical forecaster clearly enjoys considerable advantage, since such averaging processes are not necessary in order to recognize the low-latitude long-wave patterns.

It is also interesting to observe that the troposphere in both low and high latitudes is divided into two portions. One of these is characterized by extensive zonal currents with open waves, the other by cyclonic and anticyclonic vortices. In extratropical regions we find vortices near the ground that are overlain by a westerly current. In low latitudes, on the contrary, there is a broad zonal current near the ground, and above it we meet trains of vortices. In order to produce adequate predictions, a forecaster evidently must have maps, and therefore of course data, from both of the tropospheric layers in all latitudes.

## 7. Relation between high-tropospheric flow and instability of the trades

The displacement of the high-level eddies (fig. 18) is about 17–18 mi $\mathrm{hr}^{-1}$, which is the same as the speed of the lower trades. If waves in the easterlies are displaced at a lesser rate, as is commonly the case, then relative motion between upper and lower disturbances must take place. This is strikingly borne out in our series. We shall concentrate on the wave trough moving from the Marshalls toward Guam between September 9 and 11. As this trough passes Eniwetok, it lies west of a high-level anticyclone and east of a high-level cyclone. Several days previously it had been situated immediately below this upper cyclone. On September 11 (fig. 22) this is no longer the case. The upper ridge has moved forward relative to the trough and is bypassing it on its way westward. This is clearly seen from the Guam time section (fig. 8) on September 12. There the heights of isobaric surfaces rise strongly on September 11, but on the following day this rise ceases as winds back from southeast to northeast. The wave trough in the easterlies, now being transformed into a typhoon, passes Guam 18 hours later. Northerly winds aloft continue to prevail after typhoon passage, reflecting the approach of the next high-level cyclone. On September 12 the typhoon is located between a high tropospheric anticyclone to its west and a cyclone to its east. This is a typical position observed in many instances of typhoon formation.

In noting the relative motion between waves at low and high altitudes in the trade-wind belt, we have encountered a clue that may explain the instability of the trades and the beginning of typhoon formation on September 11. *Deepening starts at the surface just as the upper high center bypasses the low-level trough.* Near the ground the air suddenly is accelerated eastward. Therefore the low-level pressure field must change so that a pressure gradient along the streamlines appears that produces the desired acceleration. Our data suggest that the necessary pressure changes are produced by the isallobaric minimum that is approaching from the east. *As soon as the wave trough in the lower easterlies passes from the region of pressure rises ahead of the upper ridge to the area of pressure falls to its rear, instability develops.*[8]

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[8] Note also, conversely, that many young, deepening wave cyclones of middle latitudes are situated just west of eastward moving ridges in the upper westerlies.

The question arises as to why instability develops at the particular site at which we observe it. Since a disturbance in pressure and windfield is already present at the wave trough, it is apparent that superposition of an additional disturbance will more readily set off deepening there than it will in the undisturbed easterly current. But we may ask why only a small slice of the easterlies veers rather than the entire body of air east of the troughline.

The effect produced by superposition of high-level pressure falls on the field of motion near the ground mainly depends on three factors: latitude, pressure gradient along the troughline, and vertical stability of the air. The higher the latitude and the greater the meridional pressure gradient, the weaker must be the influence of the particular type of superimposed disturbance that is in question here. The greater the vertical stability, the more readily will small vertical motions compensate for the imposed pressure changes so that their intensity diminishes toward the ground. In the easterlies, both stability and pressure gradient increase north of the equatorial low-pressure trough. It is therefore in the flat pressure field of the area where the wave trough in the easterlies joins the equatorial trough that the trades are most easily diverted. As shown by many radiosonde observations, the air there has a lapse rate slightly in excess of the moist adiabatic. There is no trade-wind inversion and the humidity is high, at least throughout the lower troposphere. Effective compensation of impressed pressure changes through vertical motion is therefore not possible

It is necessary now to compare briefly the sequence of events just described with circumstances preceding cyclogenesis that have been noted in the western Atlantic (Riehl, 1948). In the latter cases, superposition of the low-latitude portions of polar troughs on waves in the easterlies and equatorial shearlines played a principal role. If the polar westerlies aloft overlie the entire trade-wind belt, tropical cyclogenesis does not occur at all. The latter statement also applies to the Pacific, but our map series demonstrates that no polar trough was involved in the cyclone development beginning on September 11. We observe here a type of situation not previously noted in the Atlantic, and this is one of the principal reasons for presenting this particular set of maps. This is not to say that typhoon developments produced through the effect of extratropical troughs do not occur in the Pacific. On the contrary, they happen quite frequently. The storm located east of the Philippines at the beginning of our series was derived from such superposition, and the same holds for the development described by Hubert.[4] Conversely, it is probable that some Atlantic hurricane formations follow the pattern of typhoon formation indicated by our maps. This is particularly likely with regard to the "Cape Verde" type of hurricane. It is, however, impossible to make a comprehensive analysis of hurricane formations in the central and eastern parts of the Atlantic Ocean because of the complete lack of data.

Whether a tropical storm forms because of some type of interaction between high and low latitudes, or whether it develops as shown in our series, its generation still takes place in response to a common event, namely, *superposition in consequence of relative motion*. In one set of cases, relative motion between extratropical and tropical perturbations is involved. In the other set, perturbations within the tropics move relative to each other. One sequence of events favors generation of tropical storms at relatively high latitudes—north of the region of strongest easterlies; the other favors formation at much lower latitudes—south of the strongest easterlies.

It is advantageous now to make a distinction between the three classes of broadscale circulation patterns which appear to be the most common types encountered by the forecaster during the hurricane season:

*1.* The polar westerlies aloft and the anticyclonic shear zone to their south overlie the trade-wind belt. No tropical storms form.

*2.* The polar westerlies are of moderate intensity only. Their influence ceases about 10 degrees of latitude north of equatorial shearlines, the most intense portions of wave troughs in the easterlies or the northern boundary of the equatorial trough, whichever is farthest north. Tropical storms form as a result of interaction between disturbances of high and low latitudes.

*3.* The polar westerlies are weak and confined to relatively high latitudes at all elevations. Middle-latitude control of events in the tropics is then shut off. Tropical storms form as a consequence of interaction between disturbances at low and high altitudes within the tropics.

It is to be noted that the equatorward boundary of the zone of influence of the polar westerlies is determined by the direction of motion of the anticyclonic eddies at 200 mb in the subtropical ridgeline rather than by the lowest latitude to which the westerlies penetrate at that level. The latitude of the boundary is, of course, a function of longitude.

## 8. Relation between high-tropospheric flow and development of storms of full hurricane intensity

Up to this point we have investigated circumstances attending the initial formation of tropical storms. Now we shall briefly consider what happens to these vortices once they are created. This is a serious problem. Every hurricane forecaster knows that many tropical depressions remain weak and do not "wind up." Once a closed counterclockwise circulation has been established in the tropical air mass, convection usually is intense and sometimes excessive to an amazing degree. If condensation energy were all that the circulation needed for its growth, every such vortex should become an intense tropical storm. But we know from experience that this is not so. Condensation energy therefore not only cannot be used to give a complete explanation of initial formation, but the growth of storms also cannot be entirely attributed to it. Liberation of condensation energy is a necessary, but not sufficient, condition for both inception and growth of storms.

In the Atlantic studies (Riehl, 1948), several factors were suggested as tending to prevent hurricane growth. But the discussion had to remain incomplete because of lack of data. It is possible on the basis of the Pacific reports to suggest an additional criterion. We are certain today that tropical storms are warm low-pressure centers and that outflow takes place in the middle and upper troposphere. Streamlines in both typhoons of our series have been drawn accordingly. The pressure gradient between the interior and the immediate vicinity of tropical storms should reverse in the upper troposphere, as actually computed by Schacht (1946). At upper levels, tropical storms are high-pressure centers, though presumably with cyclonic circulation. It should assist the growth of the storms if they were located in a region where high-pressure exists in a broadscale sense, so that the general pressure field is added to the pressure distribution produced by the storm itself. Such an addition would provide optimum circumstances for upper outflow.

It is indeed a fact that tropical storms of great intensity, such as those illustrated here, display an unquestionable affinity for the 200-mb high-pressure cells. The center that begins to form on September 11 is located midway between the high-tropospheric cyclones on September 13 and 14 (figs. 24 and 25). This is a curious and remarkable phenomenon. The same holds true for the typhoon of September 8–11, for the storm shown by Hubert,[4] and many similar cases. Frequently the strength of the high-level ridges increases after typhoon formation, which indicates that apparently the storm helps to build up the large-scale highs.

If we now adopt the view that proximity of an upper ridge to an incipient storm leads to deepening to full hurricane strength, we may well ask what happens when storms are generated through interaction between a tropical perturbation and a cold trough of the polar westerlies. In those instances, the broadscale high-tropospheric pressure field opposes the perturbation pressure field generated by the storm and impedes the outflow aloft. We should expect that in such situations storms would develop far more infrequently than when associated with a broadscale flow pattern of the September 11–13 type. We may have here an indication as to why Atlantic storms are rare compared with Pacific typhoons. If an incipient center is to attain great strength, the relative motion which has produced the cyclogenetic situation must continue after cyclogenesis. The polar trough and the forming center must be separated, or the polar trough must weaken.

If we look for observational data to substantiate these suggestions, we are handicapped because small and weak cyclones usually escape the notice of the analysts in regions with sparse data. In the formative stage the diameter of a storm often is only 50 miles or even less. We have, however, the case of the typhoon 8–11 September 1945, a storm formed by superposition of tropical and extratropical troughs, as previously noted. First formation of this center took place September 5–6, but it remained weak until late on September 7 when it suddenly began to "wind up."[9] It attained great strength on September 9. Inspection of 200-mb charts and Iwo Jima and Okinawa time sections shows that on the morning of September 7 the original polar trough was still situated between Iwo Jima and Okinawa. The high-latitude portion of the trough broke off on that day and began to move eastward across the Pacific. Simultaneously, the tropical portion began to weaken rapidly. Although it was still present on the morning of September 8, the two high-pressure centers situated on both sides of it moved toward each other on that day, as clearly seen from the time sections, and merged on the 9th. Just at that time activity in the surface depression picked up rapidly.

The University of Chicago group on Guam witnessed another disturbance of this kind. On 28 August 1947, it was suspected that a depression had formed south of Guam and a reconnaissance flight was sent there in which Mr. L. Hubert participated. The aircraft encountered a weak depression and only moderately bad weather. On the following day, the writer accompanied the naval reconnaissance which located the disturbance somewhat farther west. Surface pressure had fallen to about 1000 mb and winds had increased to 30–35 mi hr$^{-1}$. The rain was torrential, and the naval aerologist, veteran of many typhoon flights, classified the rain as fully of the intensity observed in mature storms. The writer expected that the depression would shortly become a mature typhoon, but nothing happened. The depression traveled in steady state all the way into the China Sea, a clear proof that condensation energy alone cannot create intense tropical storms.

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[9] Seventh Amphibious Airforce, United States Navy, "Typhoons of the western Pacific, August–October 1945." Produced by the Photographic Unit of U.S.S. Catoctin.

High-level winds in 1947 were so scarce that no 200-mb charts could be drawn. But inspection of the Guam time section showed that winds there shifted from northeast to southeast at 30,000 and 40,000 ft as the surface depression formed, indicative of westward passage of a high-level trough. Subsequently the Guam winds remained southeast, and we may presume that the upper trough continued to move westward with the surface cyclone.

Considering positive and negative evidence together, there appears to be a fair amount of proof that an incipient vortex, in order to grow to full typhoon strength, must be associated with a ridge in the 200-mb broadscale flow pattern or at least become separated from a polar trough, if formed through interaction with such a trough. Extension of the present work to many more storms will show to what extent the present description of the formative and immature stages of hurricanes and typhoons is adequate, and to what extent further complicating factors must be introduced.

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