The Circulation of the High Troposphere over China in the Winter of 1945—46

By TU-CHENG YEH[1], University of Chicago

(Manuscript received 25 April, 1950)

Abstract

The paper studies the structure of the high troposphere over China in the autumn of 1945 and winter of 1945—46. The principal feature is the existence of two belts of maximum westerlies, one flowing around the southern and the other around the northern edge of the Tibetan Plateau. The southern jet stream (the main one) is extremely stationary in position. Its speed increases downstream beyond the edge of the Asiatic continent. The onset of this southern jet stream is abrupt in the middle of October over central and southwestern China and advances downstream at a speed of about 3° longitude per day.

South of this main jet stream is a belt of extremely uniform absolute vorticity which is zero in December and January. In spite of the existence of zero absolute vorticity the circulation above 15,000 feet is strikingly stable. Below 10,000 feet is the regular procession of warm troughs and cold ridges.

Comparison with the conditions along 76° E and the east coast of United States has also been made. The difference in the structure of the basic current along the east coasts of the two continents suggests an explanation of the observed formation of winter typhoons off the Chinese coasts versus the non-existence of such storms in the Caribbean Sea in the cold season.

In the past, our knowledge of the upper-air circulation in the Far East has been limited to the low troposphere. CHU (1934) and TU (1939) first attempted to describe the wind structure in this region. More recently, KAO (1948) studied the upper winds up to 15,000 feet with the aid of pilot-balloon observations of 28 stations in China and adjoining countries. He published monthly streamline charts at 10,000 feet.

In the latter part of World War II, especially during 1945, the Armed Forces of the United States maintained numerous radiosonde and rawin stations in China (fig. 1). The observations from these stations make possible an extension of the previous investigations to include the upper troposphere. This report will describe the structure of the upper troposhere during the autumn of 1945 and during the winter of 1945—46. In addition to the rawin data, use will be made of 300-mb charts prepared at the University of Chicago. Although the observations are limited to one season, it will be shown that the principal features encountered probably are valid in all years.

Onset of winter circulation

Figs. 2 and 3 show the changes in the wind field at Kunming and Chihkiang during October, 1945. The striking feature at both stations is the sudden appearance of strong winds at high levels about the middle of the month. Prior to October 13—14, winds in excess of 50 mph are not observed at all and most of the reported speeds are much less than this (Kunming about 20 mph, Chinkiang

[1] Present address: Geophysical institute, Nanking.

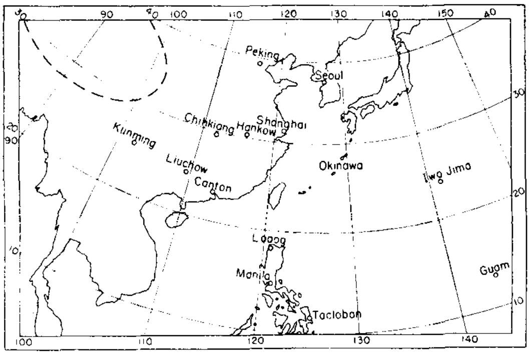

Fig. 1. Distribution of rawin stations. The dotted line at the upper left corner outlines the contour of 3 km height of Tibetan Plateau.

30 mph). After this date wind speeds above 25,000—30,000 feet consistently exceed 50 mph and frequently reach 75 mph and more. In contrast, there is little or no change at low levels. Such increase, as can be observed, lags in time compared to the strengthening of the flow in the high troposphere.

Coincident with this abrupt change, the daily 300-mb charts showed a distinct difference in pattern prior to and following the middle of October. In the earlier part of this month, the contour spacing was wide and the direction of the contours varied irregularly with the passage of troughs and ridges aloft. Later there was a pronounced crowding of the contours south of latitude 30° N in the region where the rawin observations had shown the large increase in wind speed. The shape of the contours was such as to indicate *that suddenly a strong upper flow circling the southern rim of the Himalayan Plateau had set in.*

Fig. 4 is a typical example of the flow pattern at 300-mb after the beginning of the winter circulation. Although the contour configuration over the Himalayas is uncertain as there are no stations, the adjoining observations indicate with reasonable certainty that the westerly current is split into two branches, one situated north and the other south of the mountains. This map pattern is representative for the entire winter season. It is of interest that the width of the southern current decreases downstream (northward displacement of subtropical ridge) and that the maximum wind speed is reached far beyond the boundary of the high plateau. Two lee eddies are shown, one cyclonic and one anticyclonic. Both the data of November 4 and continuity serve to support the analysis as drawn. Eddies that develop in the lee of the mountains and then travel eastward, were observed on numerous occasions. In particular, formation of cyclonic centers near longitudes 100°—105° E appears to be a sufficiently regular event that it is possible to speak of a mean cyclonic circulation in this region.

The rapid change of circulation pattern in the middle of October provides for an interesting comparison with the analysis of the

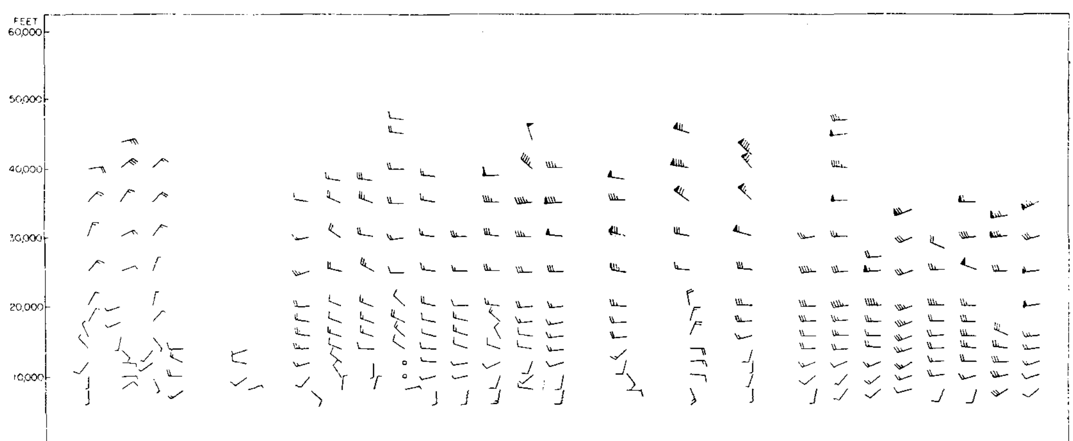

Fig. 2. Vertical time-section at Kunming, October 1945. One barb is 10 mph, one triangular barb is 50 mph and one block is 100 mph.

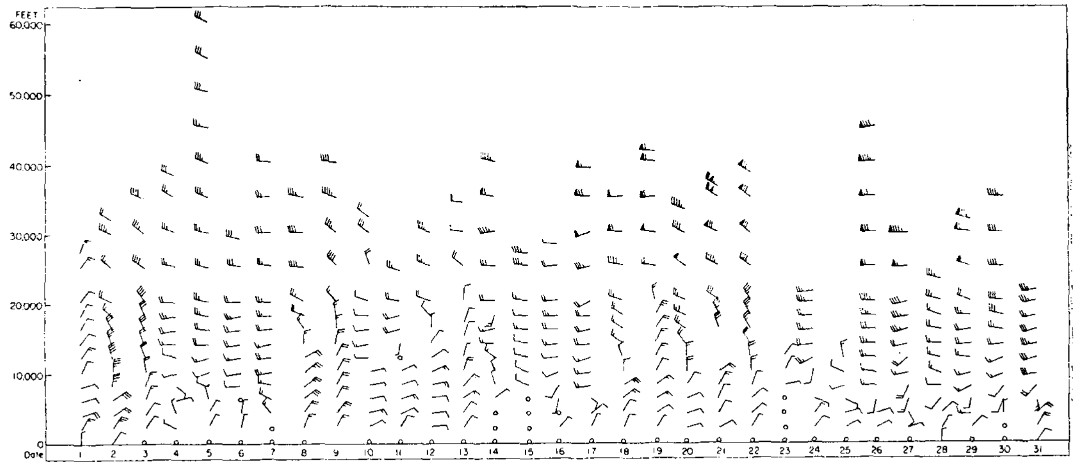

Fig. 3. Vertical time-section at Chihkiang, October 1945. The representation of the wind speed is the same as in fig. 2.

onset of the summer monsoon over India and Burma made by YIN (1949). He indicated how a small northward displacement of the upper westerlies over Eurasia could lead to a rapid transition from winter to summer flow pattern south of the Himalayas as the westerlies suddenly began to circle around the northern rather than the southern periphery of the plateau. The present data suggest an equally abrupt change in the opposite direction at the end of the summer period. In this connection it is of interest to note that Chaudhury[1] observed that the first pronounced invasion of the middle-latitude westerlies into northwest India during October, 1945 took place roughly one week earlier than shown in figs. 2—5.

This eastward progression of the onset of the winter circulation can be traced farther downstream. At Okinawa a rapid increase of the westerly flow above 25,000 feet occured on October 21 and at Iwo Jima as late as October 27. Thus we observe a rather regular downstream progression of the beginning of the winter season aloft over eastern Asia during 1945. The strong westerlies advance from longitude 70°E to 140°E in 20 days, or at a rate of $3\frac{1}{2}°$ long. day^{-1}. This is much less than the upper wind speed itself. The propagation of change from summer to winter conditions differs from what may have been expected, namely a simultaneous southward intrusion of the westerlies at all longitudes. It is plausible to suggest that the downstream propagation is peculiar to eastern Asia and that it is produced by the effect of the Himalayas on the upper circulation.

The zonal wind distribution, December 1945 and January 1946

We shall now turn to the structure of the upper wind field in the middle of the winter season. The discussion will treat the zonal wind component, since the meridional flow is very weak above 15,000 feet. Fig. 5 gives the regional distribution of zonal flow. *A distinct maximum is in evidence in southeastern China. The existence of this maximum is consistent with the presence of a narrow but intense branch of*

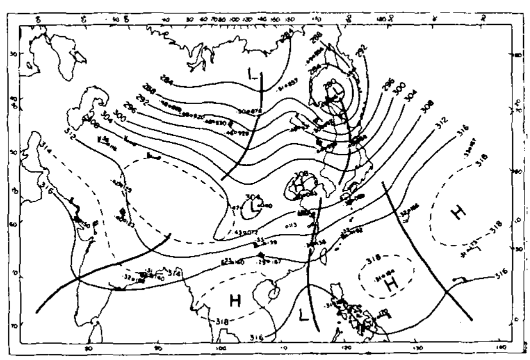

Fig. 4. 300-mb map, 4 November 1945; 0400Z.

[1] A. M. CHAUDHURY, manuscript, University of Chicago, 1949.

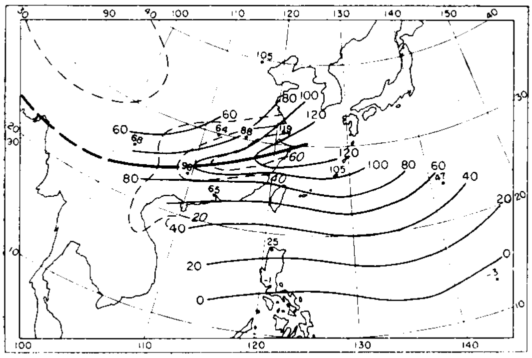

Fig. 5. Distribution of the mean zonal wind speed in mph (solid lines) of December 1945 and January 1946 at 40,000 feet and mean precipitation pattern (dashed lines) in winter. The figures are the mean speeds and the heavy line shows the position of the jet stream.

the westerlies that flows around the southern border of the Himalayas. As seen from the downstream increase of wind speed, this current intensifies even after leaving the vicinity of the mountains. Its center does not closely adjoin the mountains but lies at some distance. Wind speeds in the immediate lee appear to be quite low. This region corresponds to the stagnation point in the potential flow around a smooth cylinder. Since the atmosphere is not an ideal fluid and since its motion is not irrotational, it is very likely that lee eddies of small diameter exist in the region of stagnation as shown in fig. 4. There may even be light easterlies close to the mountains on the average.

The observation at Peking requires special mention. This station lies sufficiently far north to be affected by the westerly current that circles the Himalayas on the north side. Evidence for the existence of this current will be given later. On the other hand, the resultant wind at Peking is obtained from a small number of observations some of which appear very doubtful. For this reason, the analysis on fig. 5 has not been extended to northern China.

Wind steadiness: The preceding discussion has indicated that the winter-time westerlies aloft over central and southern China are derived from a branch of the circumpolar westerlies that flows around the southern edge of the Himalayas. If this is true, we should observe a high degree of steadiness of the upper flow. Such steadiness previously has been noted over Argentina in the upper circulation that crosses the Andes Mountains (BOFFI, 1948[1]; RIEHL, 1949). Wind steadiness (S) in percent, is defined by the formula

$$S = \frac{R}{V} \cdot 100,$$

where R is the speed of the vector resultant wind and V the average speed regardless of direction. Table 1 gives the wind steadiness as a function of height for the rawin stations for all available periods and fig. 6 shows the

[1] J. A. BOFFI, "Effect of the Andes Mountains on the General Circulation over the Southern Part of South America", unpublished part of the M. S. thesis, University of Chicago, 1948.

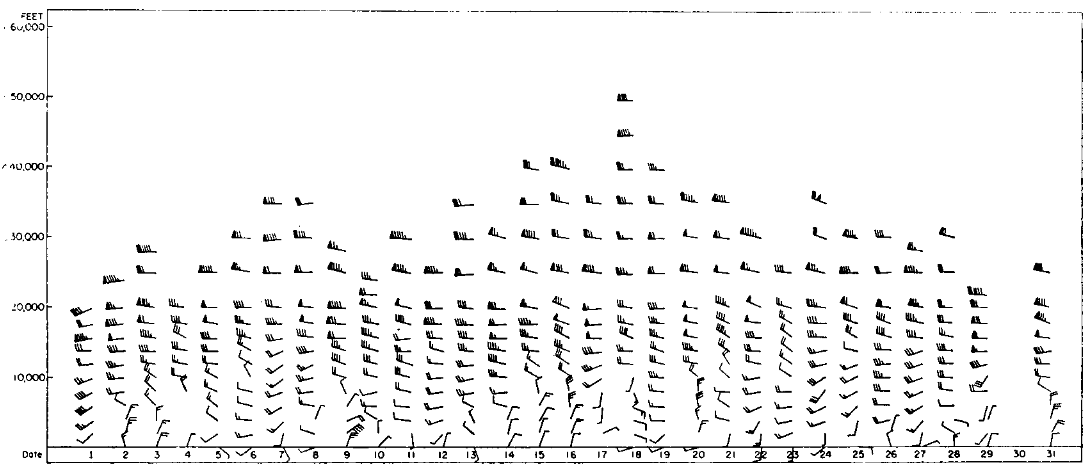

Fig. 6. Vertical time-section at Hankow, January 1946. The representation of wind speed is the same as in fig. 2

vertical time section of upper winds at Hankow, January 1946, as a synoptic example. *Clearly, the wind steadiness at high levels is very great, in support of the suggestion on the origin of the upper current.* At lower heights, the steadiness is considerably less, as can be seen from the table as well as figs. 2, 3, and 6. Near the ground a procession of troughs and ridges moves eastward. These disturbances die out with height and can hardly be observed at 15,000 feet. Therefore the low-pressure areas must be warm, and the high-pressure areas cold, relative to their surroundings. This same procession of thermal disturbances can also be noted farther east at Okinawa and Iwo Jima.

Table 1.
The wind steadiness as a function of height for the stations in China proper

	Canton	Liuchov	Kunming	Chihkiang	Hankow	Shanghai	Peking
2,000 feet . . .	22	8	—	94	32	11	62
10,000 feet . . .	65	89	81	47	57	84	84
20,000 feet . . .	91	95	89	89	95	97	87
30,000 feet . . .	85	96	86	94	97	98	89
40,000 feet . . .	98	98	75	92	96	97	99

It is surprising that the wind steadiness at 2,000 feet, well within the layer of monsoonal flow, is as low as given in the table. Chihkiang is the only station that conforms to the pattern that was expected: high wind steadiness near the ground and in the upper troposphere, with an intermediate region of low steadiness.

Jet stream and rainfall distribution: The foregoing data on wind steadiness aloft, together with the indicated role of the orography in producing the upper current, suggest that the wind distribution noted in the winter of 1945—46 should hold also for other years. It is of interest, therefore, to compare the upper flow pattern with the mean precipitation pattern of winter (Fig. 5, U. S. Weather Bureau[1]). *A maximum of precipitation practically coincides with the center of the upper jet stream. Rainfall decreases northward and southward from there, but much more rapidly on the south side.* This rainfall distribution is exactly like that found by STARRETT (1949) relative to the upper jet stream in the United States. It agrees with the distribution to be expected from the vertical circulation cell postulated for the jet stream region (University of Chicago, 1947). Such remarkable agreement between the precipitation relative to the jet in the United States and the *average* regional rainfall pattern of China relative to the *average* position of the jet over China is possible only if the latitudinal fluctuations of the jet stream over eastern Asia are restricted to very narrow limits, not only from one day to the next, but also from one year to the next.

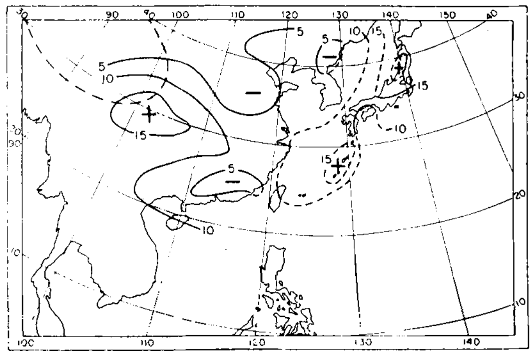

Fig. 7. Mean number of days with rain in October.

It is of interest to add that charts showing the number of days with rain also exhibit a very pronounced maximum that runs east-west along the upper jet. In autumn, at least, there is also a secondary minimum of number of days with rain to the south of the upper jet stream (fig. 7) and, as computed by Liu[2], a region of minimum rainfall variability extends along the jet stream axis.

Meridional cross section: Seven of the rawin stations operative during December 1945 and January 1946 were sufficiently close to the meridian 120° E that they could be projected on this meridian following the mean upper-air flow above 15,000 feet, which was due west. It was possible then to construct a vertical space cross section of the distribution of west wind speed (fig. 8), although the data are not as homogeneous as could be desired. The number of observations at the different

[1] See E. R. BIEL, *Climate of China*, report to U. S. Army Air Service.

[2] see footnote 1.

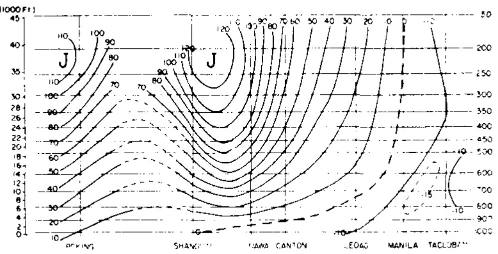

Fig. 8. Meridional cross section of mean zonal wind in mph of December 1945 and January 1946.

stations ranges from 40 to 55 for the two months' period. Moreover, the frequency of observations decreases with height, so that several stations have only 10—15 reports at 40,000 feet. Finally there is a large gap in the important region between Shanghai and Peking where the analysis as inferred has been entered with dashed lines (cf. also figs. 5, 13). Nevertheless, the principal features of the cross section should be representative in view of the high degree of steadiness of the upper wind field stressed before.

Fig. 8 shows two centers of west wind concentrations. The position of the northern maximum is not given by the data presented. It can be inferred from the analyzed daily 300-mb charts mentioned in the introduction which show that the center of the westerly current passing over Siberia, reaches the Pacific coast between Peking and Vladivostok (cf. also fig. 9). The strength of this current at Peking may not be as great as indicated. As mentioned in the discussion of fig. 5, some of the Peking data are quite doubtful. All that can be said with confidence at this time is that there are two westerly maxima.

The position of the southern jet stream center is closely given by the data of fig. 8, together with figs. 5, 9 and 10. Its intensity can be estimated as about 150 mph at the level of strongest wind (highest measured average wind 120 mph at 35,000 feet at Shanghai). The vertical slope of the axis of the southern jet apparently is northward, since the lower and middle-troposphere wind speeds are highest at Okinawa. However, this feature may not be entirely realistic, since Okinawa lies a few degrees east of the main group of stations and the intensity of the jet stream increases downstream (fig. 5).

On the south side of the southern jet the isolines of equal wind speed are almost vertical above the lower levels; a high degree of barotropy must prevail in this region. The subtropical ridge line intercepts the ground near 30° N. At first, it slopes strongly toward the equator with height until it reaches latitude 16° N at about 15,000 feet. From there on, the ridge line (base of the polar westerlies) is almost vertical, and it becomes entirely vertical above 18,000 feet at Manila.

Comparison with conditions at 80° W and 76° E: It is of interest to compare the upper wind field at 120° E as given by fig. 8 with conditions at 76° E (Chaudhury, 1950) and 80° W (Hess, 1948). In so doing, we again make use of the earlier contention that our data are representative of mean winter conditions. It is to be noted that the cross sections of Chaudhury and Hess are both based on computations from pressure data with use of the geostrophic assumption, whereas fig. 8 is constructed from actual wind observations.

Outside the tropics fig. 8 is in good agreement with the section of Chaudhury. At 76° E one principal west wind maximum appears just south of the Himalayas and a second center is indicated north of the mountains. These jets correspond to those shown in fig. 8. Chaudhury has a third maximum in the far south, just where the subtropical line is indicated at 120° E. An explanation of this difference is not offered.

Comparison of the two east-coast sections shows that the heights of the principal jet-stream centers are roughly equal. The speed of the North American jet is somewhat smaller than that of the Asiatic jet, presumably because the latitude of the strongest westerlies undergoes considerable changes over the United States whereas it is steady over China. Both sections show a secondary maximum in the north. But the separation of the two centers over North America is not as clearly marked as over eastern Asia, probably again on account of the variability of the jet stream position. This variability can also be seen from the fact that the isolines of equal speed have a lesser slope with height in the lower latitudes at 80° W compared to 120° E and that the intensity of the anticyclonic shear is much smaller (cf. also fig. 13). Finally, the subtropical ridge lies 4—5° latitude farther south over North and Central America and its slope does not become vertical in the tropics as over the western Pacific.

Figs. 5 and 8 and Chaudhury's section, taken together, finally show that over Asia the air in the troposphere mainly flows *around* the mountains. In contrast, it largely moves *over* the North American barrier. This suggests that the mountains exert different dynamical effects on the upper-air currents of the two regions which must be allowed for in theoretical analyses and in estimates of downstream influences such as positions of troughs and ridges.

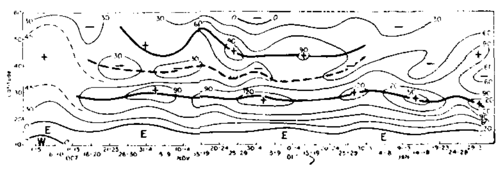

Fig. 9. Time-section of the 300-mb geostrophic zonal wind along 115° E from October 1945 to January 1946.

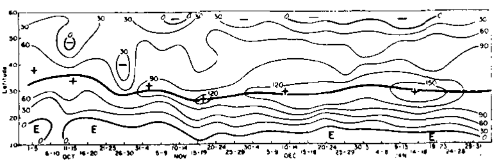

Fig. 10. Time-section of the 300-mb geostrophic zonal wind along 135° E from October 1945 to January 1946.

Fluctuations of position and intensity of jet stream and subtropical ridge

In order to demonstrate further the high degree of steadiness of the upper-air flow pattern over the Far East in winter, figs. 9—10 show the zonal wind distribution at 300-mb from October 1945 to January 1946 in synoptic sequence. The west wind speeds given in these figures were obtained with use of the geostrophic formula from the daily 300-mb charts mentioned in the introduction. In the longitude intervals 110—120° E and 130—140° E the 300-mb height gradient was averaged on each day and the geostrophic wind profiles were then combined in successive 5-day periods and plotted and analyzed in time-section form. Fig. 9 is representative of conditions in the sectors 110—120° E (center 115° E) and fig. 10 for 130—140° E (center 135° E). *The most remarkable feature seen on both figures is the stationary position of the principal jet stream center* beginning about October 15 at 115° E and somewhat later at 135° E. Oscillations of the center are within very narrow limits. The mean deviation from the mean position is less than 1° latitude at 115° E and less than 2° latitude at 135° E. The west wind speed in the central zone is more variable and also increases generally from October to the middle of the cold season.

Farther north, the secondary belt of westerlies evident in fig. 8 is present continuously over China (fig. 9) in November and December 1945, when its mean latitude is 45° N. It is less distinct during January 1946, although still indicated in the data. This second jet cannot be found in fig. 10, and this is due to considerable variability of position and intensity as seen from inspection of the daily charts. The intensity of the northern maximum at all times is less than that of the southern one: highest speeds in the north are near 100 mph whereas they reach 175 mph in the south. This difference holds not only for the 5-day but also for the daily values.

Coincident with the steadiness of the southern jet stream center we observe that the subtropical ridge undergoes only minor fluctuations in latitude over the tropical part of the western North Pacific Ocean. This may be seen also by inspection of time sections of daily rawin observations, especially at Manila and Guam. At both stations the resultant wind is nearly zero in the middle of winter (cf. also figs. 5 and 8). This small resultant is not produced by alternating high westerlies and easterlies which in the average cancel each other. Wind speeds in the upper troposphere are light on all days. This contrasts with summer conditions when a train of large upper vortices keeps passing over the region (Riehl, 1948 a).

The inverse holds over the eastern Pacific Ocean, especially near the Hawaiian Islands. There the upper winds are most steady in summer and most variable in winter when large troughs and ridges move over the island chain. It is of interest that the rainy season occurs in winter in the eastern tropical Pacific and in summer in the western tropical Pacific, *therefore in the season when the high-level flow is most unsteady*. The dry seasons are associated with a relatively steady regime aloft.

There is also a great difference between the upper-air circulation of the tropical West

Pacific and the Caribbean Sea. The subtropical ridge drops below 10° N east of the United States whereas it holds at 15° N in the western Pacific (cf. fig. 13). Moving troughs and ridges in the upper westerlies dominate the wintertime weather of the Caribbean. The westerlies have a north component and travel down the Antilles chain from the Gulf of Mexico. Wind speeds often are very high. Over Puerto Rico (18° N) WNW winds of 80 mph are observed frequently and even the resultant winds aloft reach 45 mph (STONE, 1942).

At this time it is not possible to explain the striking difference in the low-latitude flow patterns east of the two major continents. It is clear, however, that the existence of polar westerlies over the Caribbean prevents any possibility for tropical storms to develop in that region in winter (RIEHL, 1948b). *However, because of the persistence of the subtropical ridge near 15° N in the western Pacific, dynamical conditions aloft remain basically favorable for storm generation in that area.* When the other synoptic conditions necessary for tropical cyclogenesis are met, a typhoon can develop. Actually, several such cyclones form in most winters, as is well known.

Monthly profiles of west wind speed and vorticity at 300-mb

Figs. 11 a—b summarize the data of figs. 9—10 in form of monthly west wind profiles. Here, the averaging is done following the position of the southern jet stream in order to show the wind structure relative to the west wind maximum. It turns out that the mean latitude of the jet is 28° N at 115° E and 30° N at 135° E. At 115° E the highest speed rises from 75 mph in October to 135 mph in January. A secondary maximum is evident in the north in spite of the method of averaging. On the south side of the jet an east wind of 10 mph is found in all four months 15° latitude south of the center. At 135° E the speed of the jet is about 10—15 mph higher than at 15° E in all months except October.

Figs. 12 a—b illustrate the distribution of absolute vortivity (ϱ_a) with latitude computed from figs. 11 a—b with the formula

$$\varrho_a = f - \frac{\partial n}{r \partial \Phi} + \frac{u}{r} \tan \Phi \ldots \qquad (1)$$

Here f is the Coriolis parameter, u the west wind speed, r the radius of the earth and Φ the latitude. Use of this equation is justified, since the wind nearly always is due west.

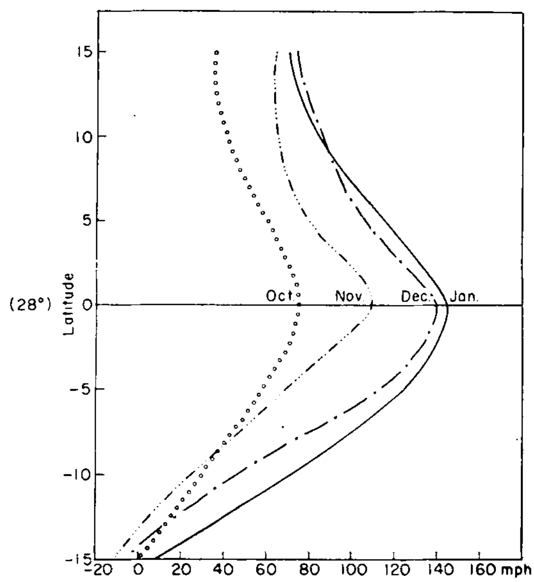

Fig. 11 a.

Fig. 11. Monthly mean profiles of 300-mb zonal wind along 115° E (a) and 135° E (b) averaged with respect to the center of the southern jet stream.

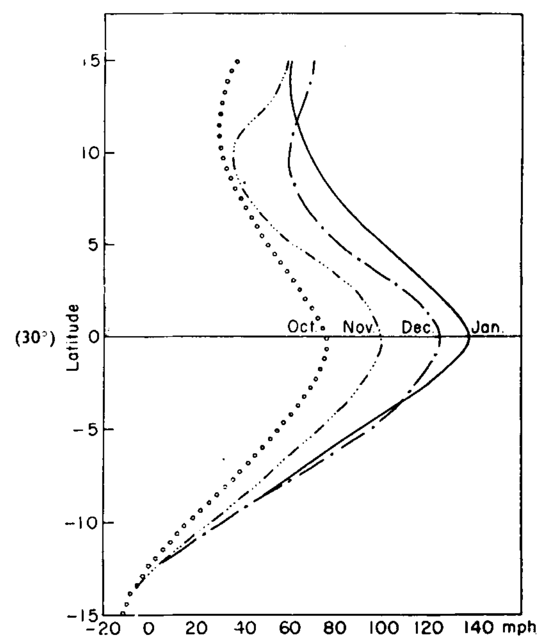

Fig. 11 b.

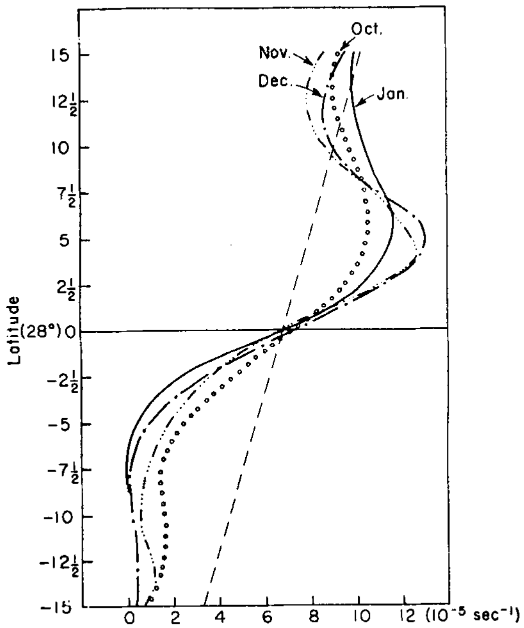

Fig. 12 a.

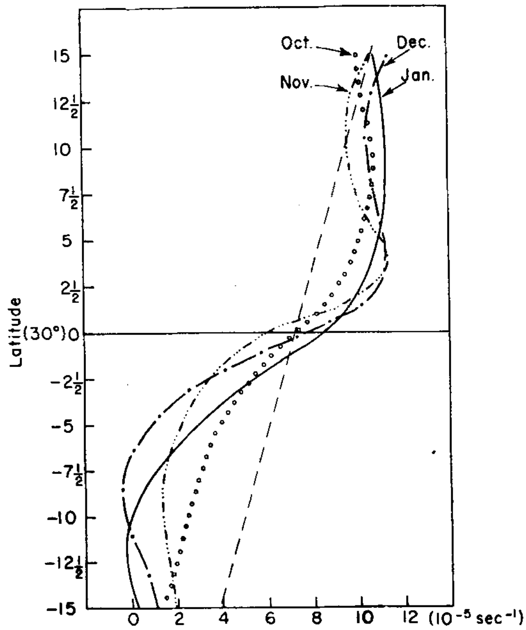

Fig. 12 b.

Fig. 12. Monthly mean profiles of absolute vorticity, computed from fig. 11a—b, along 115° E (a) and 135° E (b).

The general shapes of all vorticity curves are very similar. There is no abrupt change from month to month or from 115° E to to 135° E. A region of maximum vorticity appears 2—5° latitude north of the jet center. In January, this maximum attains a value as high as 1.3×10^{-4} sec^{-1}, almost the vorticity of the earth at the pole. South of the jet there is a broad belt in which the absolute vorticity is low and rather uniform. In December and January, its value is zero. A sharp gradient of vorticity that amounts almost to a discontinuity, extends across the region of west-wind maximum. These features are in accord with the description given by other writers (cf. PALMEN and NEWTON, 1948). The small gradients of vorticity north of the jet stream are in good agreement with the vorticity profile postulated by ROSSBY (1947, 1949).

Comparison of the four profiles along each meridian reveals an interesting change of the vorticity distribution with time. As we pass from summer to winter, there is a general increase of vorticity north of the jet and a decrease to its south. This feature probably holds on all meridians since the speed of the jet increases everywhere from summer to winter. Formally, the opposite change of vorticity in middle and low latitudes agrees with the requirements of the circulation theorem, as changes of wind speed in the equatorial zone are relatively small. Dynamically, however, it is a difficult problem to explain this change. A mechanism is required that is capable of transferring vorticity from low toward high latitudes against the gradient.

Fig. 13 further illustrates the vorticity distribution with latitude and also compares the wind speeds as given by rawin data and geostrophic computation. In this diagram the latitude is plotted according to the cosine scale on the abscissa. According to SOLBERG (1936) "dynamic instability" exists in a zonally symmetric current, provided that

$$f = -\frac{\partial u}{r \partial \Phi} \quad \dots\dots\dots\dots \quad (2)$$

We neglect here the effect due to curvature of the parallels because we are concerned primarily with low-latitude regions. Upon integration with respect to latitude,

$$u = -2\,\omega r \cos \Phi + \text{const.} \dots \quad (3)$$

where ω is the angular velocity of the earth. Equation (3) is represented by a straight line

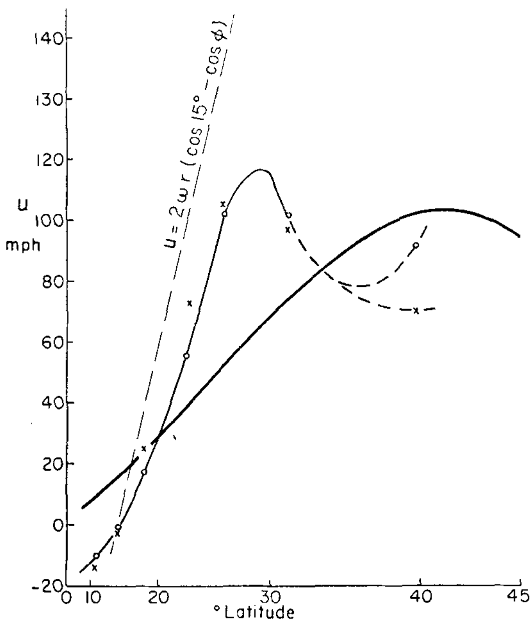

Fig. 13. Comparison of mean 300-mb zonal wind of December 1945 and January 1946 along 120° E (thin solid line) and 65° W (heavy solid line). The abscissa is in the scale of cosine of the latitude. The straight dashed line is the line of zero absolute vorticity with $u = 0$ at 15° N. The circles are the observed winds at 30,000 feet and the crosses indicate the 300-mb geostrophic winds.

on a diagram with the coordinates of fig. 13. In our case, the constant of integration was determined by setting $u = 0$ at $\Phi = 15°$ N.

The solid line of fig. 13 gives the wind distribution at 30,000 feet as taken from fig. 8. Crosses mark the corresponding values of geostrophic wind at 120° E as given by the 300-mb charts. It is seen that actual and geostrophic wind values are almost identical south of latitude 30° N. The wind profile between latitudes 18° and 29° closely approximates a straight line, indicating dynamic instability, in agreement with figs. 12 a, b. *It is of interest that this is the latitude belt characterized by extreme steadiness of the winds.*

Disagreement between actual and computed wind is again evident at Peking. Both possible solutions of the wind profile have been drawn as it is not possible to decide which is the correct one.

For comparison, the corresponding velocity distribution east of the United States (65° W) has also been entered. It is seen that the anticyclonic shear in this region is much weaker than in the western Pacific, that the jet stream center lies farther north and that the subtropical ridge lies farther south as discussed before.

Summary of the results

One of the most remarkable features of the winter circulation in the Far East is its abrupt onset manifested by the sudden appearance of strong winds. In 1945 it started abruptly in the middle of October over central and southwestern China. The onset advanced regularly downstream at a speed of about 3½° longitude per day.

After establishment of the winter circulation there appear two belts of maximum westerlies, one flowing around the southern and the other around the northern edge of the Tibetan Plateau. The southern jet stream is strikingly stationary in position while the northern one varies in latitude from day to day. The intensity of the principal jet increases from autumn to midwinter. Speeds also increase downstream beyond the border of the mountains and even beyond the edge of the continent. The maximum intensity is about 160 mph in January at 135° E.

The absolute vorticity south of this main belt of maximum westerlies is remarkably uniform. In December and January this uniform absolute vorticity is practically zero.

In spite of the existence of zero absolute vorticity the circulation above 15,000 feet in this part of the world is extremely stable. The wind direction deviates very little from the mean, which is due west. In contrast to this extremely stable upper-air current are the fluctuations of wind direction and speed associated with passage of small disturbances below 10,000 feet. Since these disturbances rapidly decrease upward in intensity the troughs must be warm and ridges cold relative to the surroundings.

Finally we shall mention a difference between the upper-air circulation of the tropical west Pacific and the Caribbean Sea. The subtropical ridge drops below 10° N east of United States whereas it holds at 15° N in the western Pacific. This difference in the structure of the basic current in these two regions helps to explain the observed formation of winter typhoons off the Chinese coast while such storms do not exist in the Caribbean Sea in the colder season.

Acknowledgement: This study was suggested by Prof. H. Riehl whose constant suggestions and assistance brought this paper to completion.

REFERENCES

CHAUDHURY, A. M., 1950: On the vertical distribution of wind and temperature over Indo-Pakistan along meridian 76° E in winter. *Tellus*, **2**, 1, 56—62.

CHU, C. C., 1934: Circulation of atmosphere over China. *Mem. of Inst. of Met.*, China, No. 4.

HESS, S. L., 1948: Some new mean meridional cross sections through the atmosphere. *J. Meteor.*, **5**, 293—300.

KAO, Y. S., 1948: General circulation of the lower atmosphere over the Far East. *Mem. of Inst. of Met.*, China, **16**, 1.

RIEHL, H., 1948 a: On the formation of typhoons. *J. Meteor.*, **5**, 247—264.

— 1948 b: On the formation of west Atlantic hurricanes. *Dep. Meteor. Univ. Chicago, Misc. Rep.*, No. 24, 1—64.

— 1949: Effects of Andes Mountains. *Bull. Amer. Meteor. Soc.*, **30**, 330—331.

ROSSBY, C.-G., 1947: On the distribution of angular velocity in gaseous envelopes under the influence of large-scale horizontal mixing process. *Bull. Amer. Meteor. Soc.*, **28**, 255—280.

ROSSBY, C.-G., 1949: On the nature of the general circulation of the lower atmosphere. In *The Atmospheres of the Earth and Planets*, G. P. Kuiper, Editor, Univ. of Chicago Press, Chap. 2, 16—48.

SOLBERG, H., 1936: Le mouvement d'inertie de l'atmosphère stable et son rôle dans la théorie des cyclones. *Procès-verbaux de l'assoc. de météor., Un. géod. géophys. Int., Edinbourg*, 66—82.

STARRETT, L. G., 1949: The relation of precipitation patterns in North America to certain types of jet streams at the 300-millibar level. *J. Meteor.*, **6**, 347—352.

STONE, R. G., 1942: On the mean circulation of the atmosphere over the Caribbean. *Bull. Amer. Meteor. Soc.*, **23**, 4—16.

UNIV. OF CHICAGO 1947: On the general circulation of the atmosphere in middle latitudes. *Bull. Amer. Meteor. Soc.*, **28**, 255—280.

YIN, M. T., 1949: A synoptic-aerologic study of the onset of the summer monsoon over India and Burma. *J. Meteor.*, **6**, 393—400.