On the Fields of Wind and Temperature over Japan and Adjacent Waters during Winter of 1950—1951

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Abstract

First, a mean cross section along the meridian 140° E during three months (Dec. 1950—Feb. 1951) is discussed. The main characteristics are:

1) An extremely strong subtropical jet (75 m sec^{-1}) situated at 33° N near 190 mb;

2) A semi-stationary sloping surface which divides almost barotropic warm tropical air in the south from more or less baroclinic temperate air situated in the north;

3) A slight indication of a polar-front jet situated near 230 mb at 41° N; the weak intensity of this jet depends upon the strong variability in position of the polar front.

4) Much warmer temperature in the upper troposphere to the south, and much colder temperature in the lower troposphere north of the subtropical jet, as compared with mean conditions at 76° E and 80° W.

The computed mean winds at 140° E are combined with the mean isotachs given by YEH over China, at the 12-km level. Using the acceleration determined from the isotach pattern, the mean angle of cross-contour flow at 200 mb is determined to be about 8° (between longitudes 110° and 140° E), with a mean cross-contour component of 8 m sec^{-1} toward lower heights, in the center of the mean jet stream.

Second, daily meridional cross sections through Japan are presented for the period 8 to 15 December 1950. These show an extremely strong individual jet stream in which the observed maximum wind speed reached a velocity of more than 110 m sec^{-1}, with near-zero absolute vorticity to the south. An associated nearly-isothermal layer underneath is much deeper (up to 4 and even 6 km) than in other parts of the globe. It is suggested that this unusual depth results from a combination of a stable layer associated with the semi-stationary subtropical jet, and the polar front, when the latter moves southward in east Asia. The behavior of the tropopause in the intermediate atmosphere south of the polar front during this series suggests subsidence of the tropopause and its transformation into the lower boundary of a deep stable layer which is transformed into some kind of front separating the real tropical air from a more temperate air moving in from the regions north of the Himalayas.

Introduction

During the colder season, the high troposphere over east Asia and the western North Pacific is characterized by a remarkably strong and well-defined jet stream. This has been discussed by CHAUDHURY (1950), YEH (1950), and OOI, MATSUMOTO, ITOO and ARAKAWA (1951, 1952). NAMIAS and CLAPP (1949) have published a chart showing the average position and strength of the jet stream for January. According to their figure, the mean subtropical jet stream passes over North Africa and India, increases in speed over

India and south China, and reaches maximum intensity south of Japan, becoming weaker again towards the eastern Pacific.

YEH (1950) made it clear that there exist two jet stream belts over east Asia. The main one (subtropical jet), characterized by very steady winds, comes from south of the Himalayas, the other (polar-front jet) from north of the Tibetan Plateau. According to the experience of Japanese meteorologists, these two belts of strong winds seem to be closely related to two distinct surface frontal systems, the one on the southern sea, the other on the northern sea of the Japanese Islands.

Sometimes during the colder season we observe rather deep nearly-isothermal layers in the upper and middle troposphere over Japanese stations, associated with highly-developed jet streams. These stable layers at times reach a depth of four and even six kilometers, much thicker than frontal layers observed in other parts of the world. So far as is known, this phenomenon (in its extreme form) is limited to the east Asiatic region, suggesting perhaps an orographic effect of the Himalayas and Tibetan Plateau. These considerations suggest that it would be interesting to analyse the wind and temperature fields of the high troposphere over Japan in the mean, and also in individual cases, and to compare the features of the general circulation pattern over east Asia with those observed in other locations.

Data for the following analyses were obtained from "Aerological data of Japan" published by the Japanese Central Meteorological Observatory, Tokyo, supplemented by synoptic data from the "Daily Series, Synoptic Weather Maps" published by the United States Weather Bureau.

Part I. Mean conditions for three winter-months (December 1950—February 1051)

Preparation of mean cross section. — The locations of the radiosonde stations used in constructing the mean cross section are shown in fig. 1; the number of observations used at various levels is shown in Table 1. At least up to 200 mb there were enough observations to give quite homogeneous results, although some error might be introduced by the decreased number of observations at 100 mb.

To avoid the effects of radiation errors (discussed in a later section), 1500 GCT (midnight Japanese Standard Time) observations were used. Data for Tateno, Iwo Jima and Guam were not available for that time, so observations at 0300 GCT (noon local time) were used for those stations, applying at Tateno a mean radiation correction deduced from the neighboring stations. Iwo Jima and

Table. 1. List of stations and number of observations. (Dec. 1950—Feb. 1951 inclusive).

Station	Lat. North	Long. East	No. of observations			
			500 mb	300 mb	200 mb	100 mb
Wakkanai (401)	45°25′	141°41′	86	83	69	46
Sapporo (412)	43°04′	141°20′	79	67	54	28
Akita (582)	39°43′	140°06′	84	83	73	54
Sendai (590)	38°16′	140°54′	89	89	82	56
Tateno (646)	36°03′	140°08′	90	90	90	[1]72
Shionomisaki (778)	33°27′	135°46′	84	77	69	54
Tare (Ship)	29°00′	135°00′	86	81	72	33
Iwo Jima (115)	24°47′	141°20′	67	58	52	[1]33
Guam (218)	13°34′	144°55′	71	69	64	[1]48

[1] 0300 GCT (1200 Japanese Standard Time).

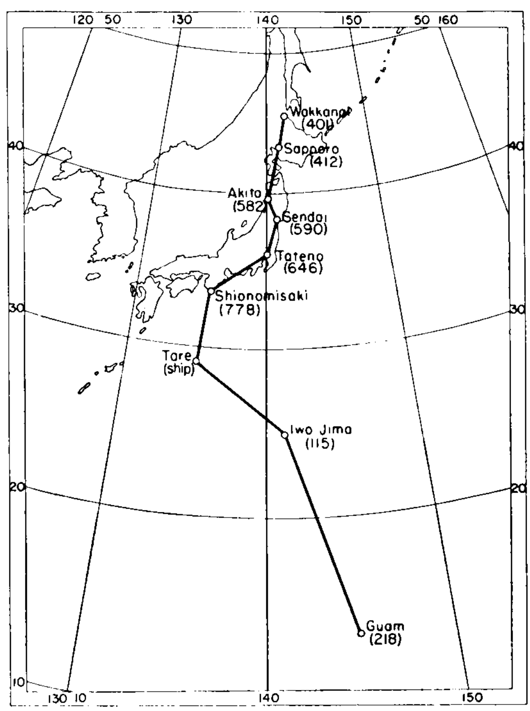

Fig. 1. Map of stations used in cross sections. Index numbers in parentheses.

Guam, which are United States Air Force stations using a different type of radiosonde, were assumed to have been corrected for radiation. A slight mutual adjustment of the analyzed mean isotherms, amounting to about 1° C in upper levels, was made between Iwo Jima and the Japanese ship station TARE, in order to achieve a hydrostatic agreement with the mean observed winds at Iwo Jima.

Mean temperatures were computed at the standard levels 1000, 850, 700, 500, 300, 200, and 100 mb for all stations, and at the additional levels 400, 250, 175, 150, 125, 80, and 60 mb for Japanese stations. The data were projected from the station locations onto the meridian 140° E along the mean contours at the 300-mb level.

The geostrophic zonal wind speed u_g was computed from the slopes of the isobaric surfaces using mean heights at 700, 500, 300, 200, and 100 mb for all stations, and in addition at 250, 150, and 80 mb for Japanese stations. In making such computations in north-south cross sections, it is convenient to plot the heights of individual isobaric surfaces against a horizontal scale proportional to the cosine of the latitude. Then the geostrophic wind u_g is given by

$$u_g = \frac{g}{2\,\omega} \cdot \left[\frac{dz}{d\,(a \cos \varphi)}\right]_p, \qquad (1)$$

where g is the acceleration of gravity, ω the angular velocity of the earth, z the height of an isobaric surface, a the radius of the earth, φ the latitude, and p the pressure. For a given value of u_g, the slope of a height profile plotted in this manner is the same at all latitudes.

The question of how to correct for the effect of curvature in the mean flow is obviously not simple. As a first approximation, it might be assumed that the air flows approximately along latitude circles. In that case the gradient wind, which might be called the *geogradient* wind because of the assumption made, may be computed from the formula (PALMÉN, 1948)

$$u + \frac{u^2}{af} \tan \varphi = u_g, \qquad (2)$$

where u is the geogradient wind and f the Coriolis parameter.

After the geogradient winds were computed using the procedure described above, they were compared with the zonal components of the mean observed winds, wherever the number of direct wind observations was fairly large. The result is shown in Table 2, at 8, 9, and 12 km (cf. fig. 3). Winds at 0300 GCT were used because of the paucity of observations at 1500 GCT. At Iwo Jima and Guam, the mean isotachs were analyzed initially from the wind data, so no comparison is shown.

Table 2. Comparison between geogradient wind and observed wind (0300 GCT).

u: geogradient wind component (m sec^{-1})
u_0: observed wind component (m sec^{-1})
n: number of observed winds

Station	8 km		9 km		12 km	
	u	u_0 (n)	u	u_0 (n)	u	u_0 (n)
Shionomisaki (778)	48	47 (38)	55	54 (31)	75	71 (14)[1]
Tateno (646).....	44	44 (89)	50	50 (85)	59	57 (51)
Sendai (590).....	38	36 (23)				
Akita (582)......	36	33 (32)				
Sapporo (412)....	32	29 (32)				
Wakkanai (401)..	23	18 (28)				

[1] Near the main jet center. The value given here was computed on the following basis: the mean vertical shear for the 14 cases where winds were observed at both 9 and 12 km was added to the mean value of wind speed for all 31 cases at the lower level. This was done in order to eliminate the artificially low mean value at the higher level caused by the tendency for balloons to be blown out of range with strong winds.

Fortunately Tateno, with excellent high-level wind data, is located near the latitude of the mean jet stream, where the wind is also very steady. At that station and at stations 778, 590, 582, and 412, the agreement between the independently computed geogradient wind and observed wind components is very good. It is to be noted that, in all cases where there is a discrepancy between geogradient and observed westerly components, the geogradient wind is the larger. This is natural since there are more radiosonde data than wind observations at high levels, and the wind observations tend to be missing when there are strong winds.

At the northernmost station 401, the agreement is not as good. The upper level winds at this station are much more unsteady in direction than at the stations further south, because of frequent passages of troughs and

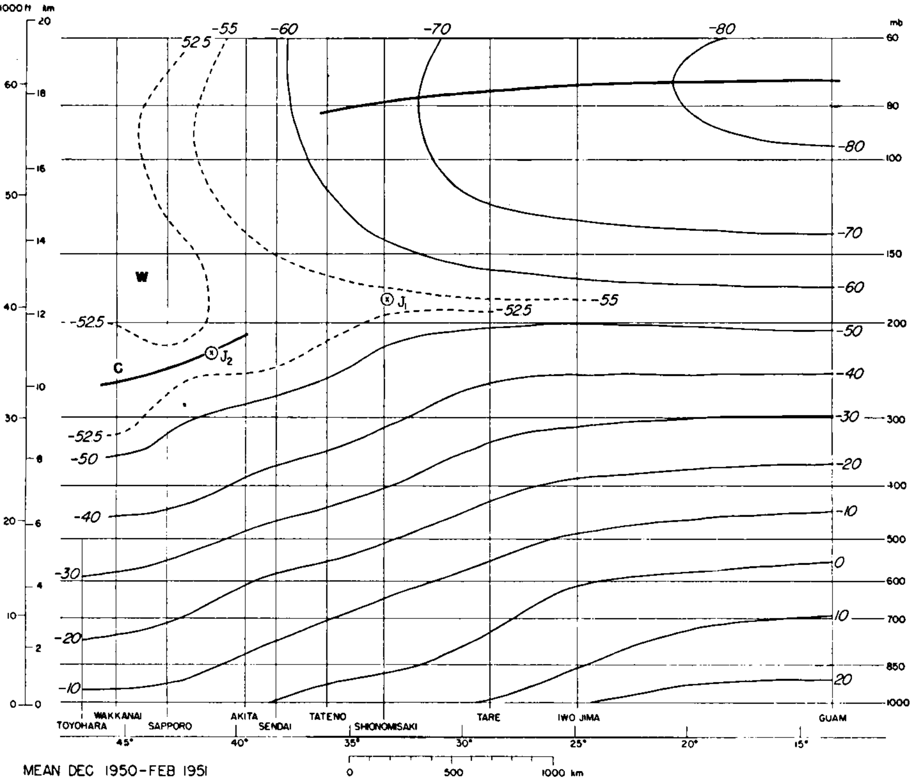

Fig. 2. Distribution of mean temperature (deg C) along meridian 140° E in winter (Dec. 1950—Feb. 1951). Heavy lines indicate level of lowest temperature; J_1 and J_2 locations of principal and secondary mean jet streams.

ridges in the middle latitudes. Obviously no attempt can be made here to consider the effects of curvature on the mean wind computed from the pressure field, since the curvature changes strongly from day to day.

General description. Figure 2 shows the mean temperature field and Fig. 3 the mean wind and potential temperature fields for the three winter months December 1950—February 1951. The lines indicating the tropopauses are located somewhat subjectively according to the normal concept of tropopause, and do not necessarily correspond to the mean locations of the real tropopauses.

The meridian 140° E corresponds to the position of a mean trough in the middle-latitude westerlies. According to the results of NAMIAS and CLAPP (1949) and others, the subtropical jet appears very strong even in mean conditions, indicating great steadiness of the high-tropospheric wind near the latitude of the subtropical highs. Thus the longitude of Japan corresponds to a region where the middle-latitude and subtropical jets are on the average quite close together. This suggests that there should be some indication of two jet streams in the mean cross section. As seen from fig. 3, the southern jet is very clear at 33° N, and some indication of the middle-latitude jet is noticed at 41° N. A similar indication of a double wind maximum can be noticed in the mean conditions at 80° W (HESS, 1949); however, there is no such indication at 150° E (LOEWE and RADOK, 1950), or 170° E (HUTCHINGS, 1950) in the southern hemisphere.

The latitude of the principal jet stream, 33° N, is about 10 degrees north of the position indicated at this longitude by NAMIAS and CLAPP (1949). This difference is not likely to

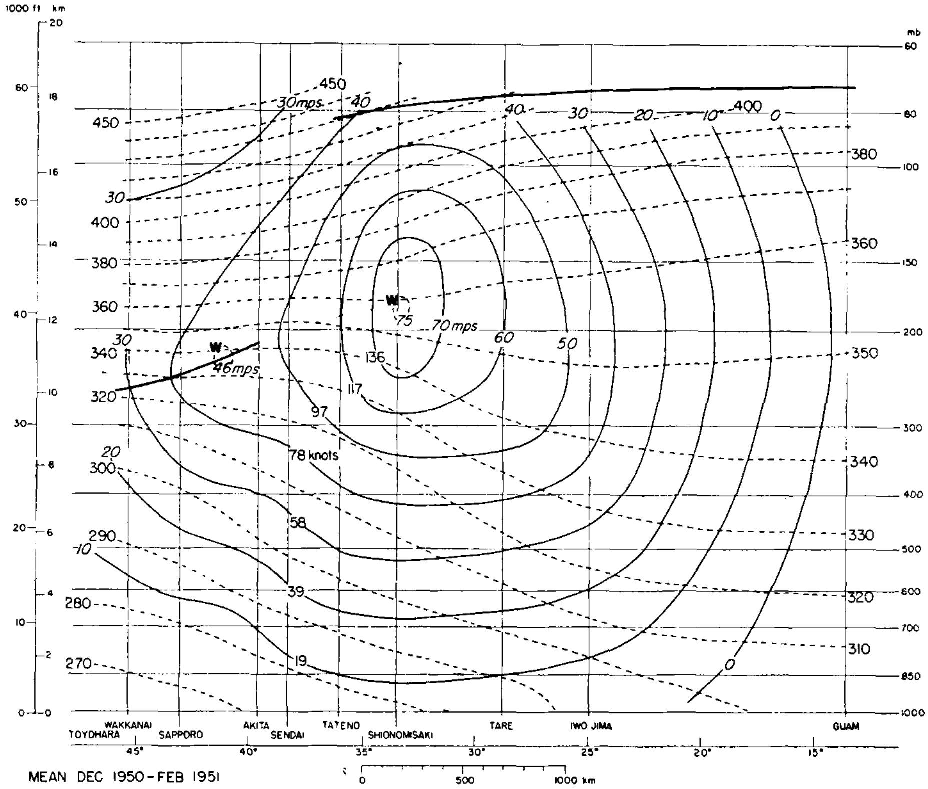

Fig. 3. Mean geogradient west wind (thin solid lines, speeds in m sec^{-1} given by slanting figures, knots by upright figures) and mean potential temperature (deg K, dashed lines) along meridian 140° E *in winter*.

be due to an abnormality in the particular winter studied here, but rather to the fact that more direct observations are available in this case.

The mean zonal wind speed at the jet center reaches the remarkably high value of 75 m sec^{-1}, or 146 knots. This value cannot be directly compared with the results obtained in mean cross sections published for other locations, since these are for different periods; however, such a comparison might be qualitatively useful. YEH (1950) found a winter maximum mean observed wind speed, near 120° E, of about 58 m sec^{-1}. For the cross section shown here, the maximum mean zonal geostrophic speed was 82 m sec^{-1}. At 80° W (in the mean trough in eastern North America), HESS (1948) found a maximum value of 55 m sec^{-1}, while CHAUDHURY (1950) computed, for central India, about 75 m sec^{-1}. In the Southern Hemisphere, LOEWE and RADOK (1950) found a winter maximum geostrophic wind speed of 55 m sec^{-1} at 150° E, and HUTCHINGS (1950) found the same value at 170° E. The geogradient correction would reduce these values by 3 to 7 m sec^{-1}.

The much larger values of wind speed observed in winter over the eastern Asiatic coast are in large part due to the steadiness of the subtropical jet in this location, as pointed out by YEH (1950). Another factor, however, is the fact that the jet is very strong in the individual cases that make up the mean; this will be discussed later.

Fig. 4 shows a composite mean isotach chart at the 12-km level (near 200 mb) made by combining the analysis given by YEH (1950) over China with 12-km wind speeds taken from fig. 3 at 140° E. Wind and radiosonde

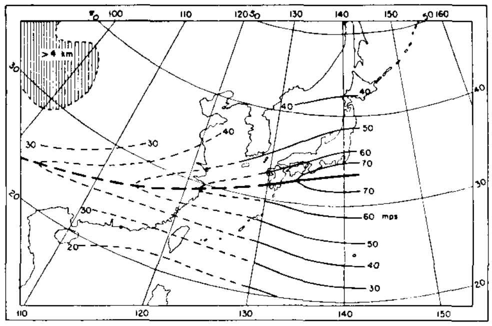

Fig. 4. Distribution of mean west wind (m sec^{-1}) at 12 km in winter, obtained by combining mean isotachs over China and vicinity given by YEH (1950), shown by thin dashed lines, with wind speeds at 140° E from fig. 3 (thin solid lines further east). Heavy line indicates axis of mean jet stream. Hatching in upper left shows approximate land area above 4 km elevation.

observations from China and Japan are not available for identical time periods. However, the data combined in fig. 4 fit very well despite the different periods.

The following discussion does not depend upon whether fig. 4 is correct in all details, since other data are available for substantiation. As shown in fig. 4, the mean speed in the jet center increases downstream from 46 m sec^{-1} at 110° E to 75 m sec^{-1} at 140° E. This indicates that on the average the streamlines must deviate northward across the pressure-contour lines. With certain justifiable assumptions, the magnitude of the mean cross-contour flow may be computed.

If the x-axis is taken along a contour on a constant-pressure surface (fig. 5) the component of acceleration along that axis is given by $du/dt = fv$. Expressing the components u and v in terms of the total velocity v_s along a streamline and the angle α between streamline and contour, and substituting into this expression, we get

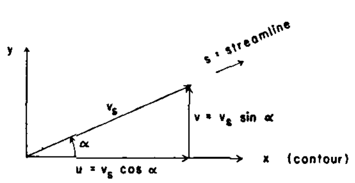

Fig. 5. Sketch of wind components related to direction of pressure contour and streamline.

$$\frac{dv_s}{dt} = v_s \tan \alpha \cdot \left(f + \frac{d\alpha}{dt} \right) \qquad (3)$$

The acceleration may be expressed in the form

$$\frac{dv_s}{dt} = \frac{\partial v_s}{\partial t} + v_s \frac{\partial v_s}{\partial s} + w \frac{\partial v_s}{\partial z} \qquad (4)$$

If the acceleration is computed from mean data, the first term $\partial v_s/\partial t$ is zero. According to the mean cross section in fig. 3 and that given by YEH in eastern China, the 12-km level corresponds nearly to the level of maximum wind. This agrees with the general synoptic experience that the subtropical jet stream varies little from the 200-mb level from day to day. This being the case, the last term can be neglected since $\overline{\partial v_s/\partial z} \approx 0$. Then the mean acceleration can be evaluated from the second, or horizontal-advective term, provided the direction of a mean streamline can be established.

Figure 4 shows that, even in the mean condition, the jet axis separates a region of high vorticity (averaging about 1.3×10^{-4} sec^{-1}) from a region of low vorticity (about 0.3×10^{-4} sec^{-1}). Neglecting the vertical transport of momentum, which as stated above is likely to be negligible at this level, changes of the absolute vorticity are given by

$$\frac{d}{dt}(f+\zeta) = -(f+\zeta)\,\mathrm{div}_H \mathbf{v} = v_s \cdot \frac{\partial}{\partial s}(f+\zeta). \qquad (5)$$

Clearly if a mean streamline near the jet axis is chosen in any direction except parallel to that axis, particles crossing the jet must undergo marked changes in vorticity, which would require large values of mean divergence. It therefore appears justifiable to assume that the mean jet axis corresponds approximately to a mean streamline.

Combining equation (3) with equation (4) retaining only the term $v_s \cdot (\partial v_s/\partial s)$ in the latter,

$$\tan \bar{\alpha} = \frac{\overline{\partial v_s/\partial s}}{f + \overline{d\alpha/dt}} \qquad (6)$$

For steady state, $d\alpha/dt = v_s \cdot (\partial \alpha/\partial s)$. The angle $\bar{\alpha}$, defined as the angle of cross-contour flow, evidently changes little between 110° E and 140° E since, according to fig. 4, the variation of $\overline{\partial v_s/\partial s}$ along the jet axis is small. It is therefore reasonable to neglect $\overline{d\alpha/dt}$ in comparison with f in the denominator of equation (6). Then $\bar{\alpha} = \tan^{-1} \overline{[(\partial v_s/\partial s)/f]}$.

The change in maximum wind speed between 110° and 140° E, a distance of 3,000 km, is 30 m sec^{-1}, giving a value of $\overline{\partial v_s/\partial s} = 0.10 \times 10^{-4}$ sec^{-1}. Using a mean value of $f = 0.73 \times 10^{-4}$ sec^{-1} corresponding to latitude 30°, $\tan \bar{\alpha} = 0.137$ and $\bar{\alpha} = 7.8°$. Since (fig. 5) $\bar{v} = \bar{v}_s \sin \bar{\alpha}$, if we use the average value $\bar{v}_s = 63$ m sec^{-1} in the region considered, the mean cross-contour flow $\bar{v}$ is 8.5 m sec^{-1}, directed toward lower contours. This amount of *mean* cross-contour flow is remarkable, particularly since it represents an average over 30° of longitude or 1/12 of the circumference of the earth. It should be pointed out that fig. 4 gives only west wind speeds; the above values, however, have been computed using the total wind speed.

Although fig. 4 was made by combining data from different winters over China and Japan, the results would not be appreciably changed if only the isotachs over China were used, and these were from a single season. The question whether the particular seasons used in the analysis are representative of mean conditions may of course be raised.

In this regard, the results might be compared with the data given in the British publication "Upper winds over the world" (Brooks *et al.*, 1950). At the 300- and 200-mb levels over east Asia, the charts in that publication show contours running almost straight west-east, in these latitudes. Mean streamlines are given at the 300-mb level only up to latitude 25° N, slightly south of the region we are considering. Near 25° latitude, the mean streamlines between 100° and 140° E are oriented from about 10° south of west, in good agreement with the above computations. On the basis of the available information, admittedly somewhat inadequate, there is no reason to doubt the results.

In fig. 4, the jet axis is oriented from about 15° south of west in the mean. This does not necessarily deny the validity of the computed cross-contour component, since it is entirely possible that in the winter season considered by Yeh (1950) the mean contours were oriented from a few degrees south of west. It must be realized also that the location of the jet axis depends in part upon whether the observations available to Yeh represented true mean values. To what extent this is true cannot determined.

Figure 4 incidentally gives some justification to the assumption, used in obtaining the mean winds at 140° E, that the mean curvature corresponds to that of a latitude circle (geogradient assumption). The mean jet axis in fig. 4 makes almost the same angle with the meridians everywhere, so that the curvature of the jet nearly corresponds to that of a parallel of latitude. It is assumed that the curvature over Japan is the same as that found a short distance to the west.

An interesting thing to be noted from fig. 4 is that the mean subtropical jet deviates northward as it approaches the longitude of a mean trough in the middle-latitude westerlies. Thus the large-scale wave patterns (streamlines) in lower and in middle latitudes are out of phase in this region.

Relation between acceleration pattern and general circulation over east Asia. — It may be worth while to discuss the above results in the light of the findings of other investigators. In particular, the role of the Tibetan Plateau in the circulation of this region has been the subject of much discussion (e.g., Chaudhury, 1950; Yeh, 1950; Bolin, 1950). Both Chaudhury and Yeh point out that the large mean values of wind speed near 30° N are due largely to the steadiness in position of the subtropical jet stream, this being attributed to a tendency for the strong westerlies to flow around the Tibetan Plateau in winter.

Over a very large region, this plateau is more than 4 km above sea level, while peaks rising to 6 km (20,000 ft) or higher cover large enough areas to present an almost solid block to westerly flow at that level. The Himalayas are limited in the north-south direction, and present an obstacle which the bulk of the westerly stream can more easily flow around than over (Bolin, 1950).

The mean jet stream maps for winter, given by Namias and Clapp (1949), show that the subtropical jet is well-developed around most of the hemisphere. This indicates that, in contrast with the higher-latitude jets commonly associated with middle-latitude frontal systems, the subtropical jet is quite steady in location at most longitudes. It hardly seems likely that this steadiness at all longitudes can

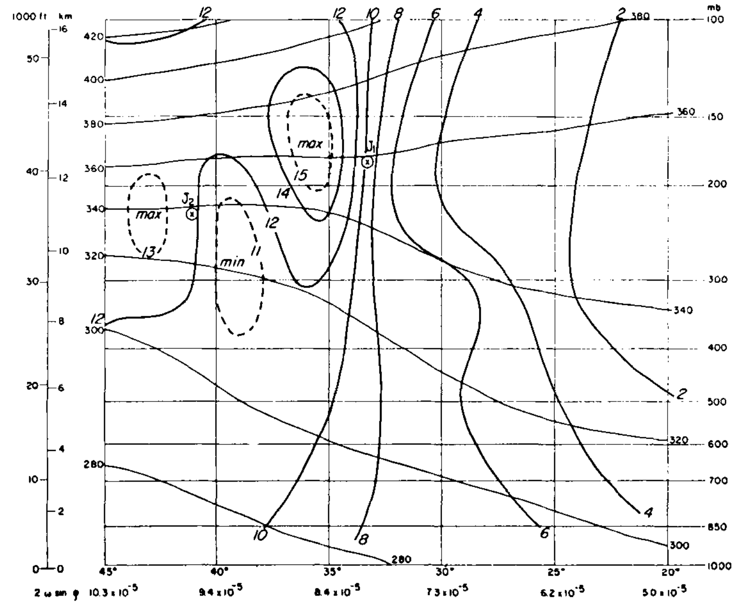

Fig. 6. Mean absolute vorticity at 140°E (heavy solid lines, units 10^{-5} sec^{-1}), Mean isentropes, thin lines (deg K). Values of Coriolis parameter at different latitudes given at bottom. J_1 is principal jet stream.

be attributed to orography alone, but certainly an increased steadiness should be expected in the east Asiatic region where the Himalayas at least present a barrier to northward fluctuations of the jet stream.

BOLIN (1950) has pointed out that the regions east of large mountain barriers are, because of the effect of the mountains in dividing the flow, regions of confluence. Following the reasoning of NAMIAS and CLAPP (1949), this must result in an intensification to the solenoidal field if the two streams undergoing confluence are of different thermal character. BOLIN connects this feature of the flow with the fact that in the mean the most intense jets are found to the east of the major mountain barriers (as shown in the charts given by NAMIAS and CLAPP). The strong cross-contour flow mentioned above is in agreement with the ideas presented by NAMIAS and CLAPP. If indeed a westerly current does divide in order to pass the Tibetan Plateau, it is clear that since the thermal influences are quite different on the north and south sides of that plateau the thermal contrast should be greater where the currents flow together again, than where they separated.

Another extremely important effect of the Tibetan Plateau is that it prevents the southward flow of cold air in the lower levels from the Central Asiatic source region. From near 60° E to near 110° E, the Tibetan and associated plateaus rise everywhere above 2 km. This gives rise to the well-known fact that the southward flow of cold air is stronger in the region from this plateau to Japan in winter than anywhere else around the hemisphere.

The Central Asiatic sea-level high is well-developed from about October to March, with a very strong decrease of pressure southward. Thus there must be a very pronounced flow of air southward in the lower levels in this region, due to frictional cross-isobar flow and other effects. The charts of frequency of anticyclone occurrence and rate of alternation of cyclones and anticyclones in winter, given by PETTERSSEN (1950), bring out in a very striking manner the concentration of southward-moving cold anticyclones between longitudes 110° and 130° E. Synoptic studies (e.g., PALMÉN and NEWTON, 1951) suggest that very strong southward nongeostrophic flow occurs in outbreaks of subsiding cold air in the lower middle latitudes. This takes place through deep layers and adds considerably to the "frictional" cross-isobar flow in the lower levels.

The strong northward nongeostrophic flow at 12 km, discussed earlier, is thus closely connected with a pronounced southward *nongeostrophic* flow in the lower troposphere, which takes place in the same geographic region. The strong acceleration of the mean flow in upper levels agrees with expectations from the strong conversion of potential energy as southward-moving cold domes subside in this region (ROSSBY, 1949 a, 1949 b; PHILLIPS, 1949).

Mean vorticity field. Figure 6 shows the field of mean absolute vorticity (vertical component), computed from the wind field of figure 3. Near the mean subtropical jet stream, there is a sharp gradient of vorticity in a band 4°—5° latitude wide, with more gradual changes elsewhere. This agrees in general with the pattern described by other writers (PALMÉN and NEWTON, 1948; CHAUDHURY, 1950; YEH, 1950). However, in the mean pattern shown here the zone of zero or negative absolute vorticity south of the jet stream, shown in CHAUDHURY's mean section and in some of YEH's mean monthly vorticity profiles, is absent. Here the mean anticyclonic shear reaches half the value of the Coriolis parameter 300 km south, and one third the value 1,000 km south of the mean subtropical jet.

In the vicinity of the weak secondary jet further north, there are similar but much weaker variations of mean vorticity. It should be pointed out that because of the large latitudinal fluctuation of the polar-front jet the mean vorticity field in this region is very much smoothed, while this effect is not so prominent in the case of the more steady subtropical jet.

The maximum value of the mean cyclonic shear 200 km north of the subtropical jet stream is about 6×10^{-5} sec^{-1}. This is slightly larger than the value of the criterion for the development of heavy (horizontal) dynamic turbulence in zonal flow,

$$-\frac{\partial u}{\partial y} = \frac{f}{2} + \frac{2u}{a} \tan \varphi,$$

given by ARAKAWA (1951), which is 5.5×10^{-5} sec^{-1} in the same region.

Mean temperature distribution.—The mean temperature field shown in fig. 2 is in general similar to that observed in other locations. One feature which stands out somewhat stronger than in the mean temperature fields in other locations (HESS, 1948; LOEWE and RADOK, 1950; HUTCHINGS, 1950; CHAUDHURY, 1950) is the rather abrupt *sloping* zone south of the main jet, dividing the almost barotropic air from the baroclinic air farther north. This might be interpreted as an approximate limit to the southward movement of the polar front. The reason for the sharpness of this transition zone is probably connected with the steadiness of the position of the subtropical jet stream, which would probably hinder the southward movement of cold air in the upper levels.

The charts given by PETTERSSEN (1950) indicate that surface anticyclones frequently move almost straight southward over east China and Southern Japan, but turn abruptly eastward on reaching a latitude somewhat south of 30° N. This is the region where the cold anticyclones would come under the strong steering influence of the subtropical jet stream. The thermal contrast through the troposphere near latitude 30° is thus maintained to a considerable extent by the southward movement of cold air masses, but the paths taken by these air masses are in part controlled by the subtropical jet. Thus, as suggested by BOLIN (1950), the actual thermal distribution east of the Tibetan Plateau must be in part controlled by dynamical influences from above (the subtropical jet, whose position is in part controlled by the mountains), and in part from below through thermal influences acting to some extent independently from the higher-level dynamical influences.

Figure 7 shows a numerical comparison of the mean temperature field at 140° E with that over Indo-Pakistan (76° E; CHAUDHURY, 1950) and that in eastern North America (80° W; HESS, 1948). The high troposphere is much warmer south of Japan, and the lower troposphere much colder over and north of Japan, than at corresponding latitudes in the other locations.

In the higher troposphere, e.g., between 500 and 300 mb, the total temperature difference between 30° and 40° N is about one and a half times bigger than that at the other locations. This result is not affected much by the differences in latitude of the strongest barocline zones at the three locations, since this is small (compare jet positions in fig. 7). The vertical gradient of the temperature differences in fig. 7 also shows that, at least above 700 or 800

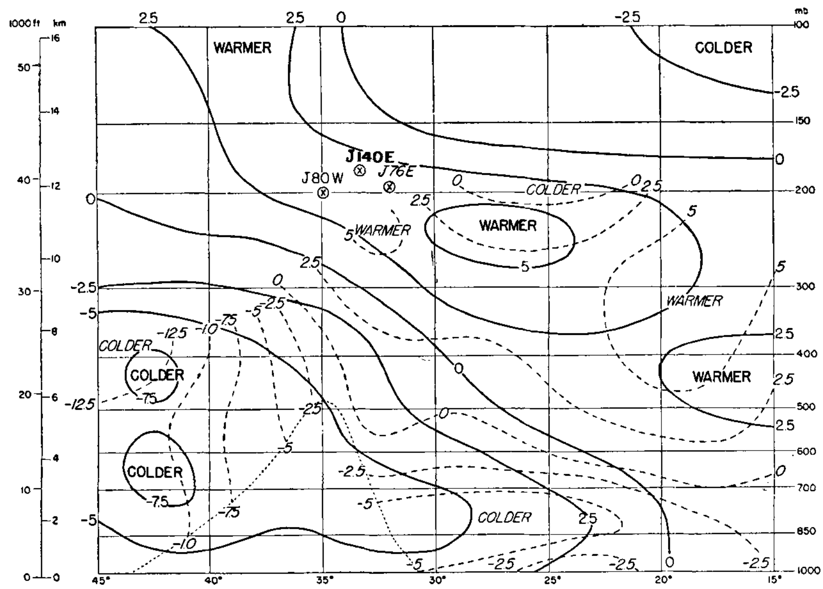

Fig. 7. Comparison of mean temperature at meridian 140° E with that at 80° W (HESS, 1948) and 76° E (CHAUDHURY, 1950). Locations of principal mean jets at the three locations indicated by J with longitude. Solid lines, temperature at 140° E minus temperature at 80° W (C deg, upright numbers); dashed lines, 140 °E minus 76 °E (slanting numbers). Dotted lines indicate profile of mountains at 76° E.

mb, the troposphere is on the average much more stable than in the other locations. This is natural in view of the frequency of cold-air outbreaks over Japan.

Diurnal variation of observed temperatures. Figure 8 shows the mean temperature field over Japan using only observations at noon local time (0300 GCT), and the differences between day and night mean observed temperatures for the three-month period. Below 850 mb, the variation of 1° to 3° C undoubtedly represents a real effect of surface heating and cooling. From 850 to 200 mb, the differences are rather uniformly of the order of 1° C. Above 200 mb, the day-to-night variation increases rapidly to values of 4° to 6° C at 80 mb. Up to 200 mb, the difference in winds computed from day and night mean temperatures over the Japanese stations is only of the order 0 to 5 m sec^{-1}, however, above 200 mb the differences become much more serious. Between the ship TARE and the U.S. Air Force station at Iwo Jima, the use of daytime values leads to a completely fictitious wind field while by using night temperature values results can be obtained nearly compatible with observed winds at Iwo Jima.

It is not possible to estimate what part of these temperature variations is real, or how much is due to solar heating of the instruments in the daytime soundings. WULF, HODGE, and OBLOY (1946) in a study of the differences between daytime and nighttime soundings over the United States, found differences of 0.5° to 2° C in the troposphere and rapidly increasing differences above the 10- to 12-km level, for winter observations. On the whole, their findings agree with the temperature variations shown in figure 8. From the nature of the changes of the diurnal temperature variations with height, these authors concluded that at least a considerable part of the variations in the first 10—20 km of altitude were real. This is a very important question which could bear further investigation.

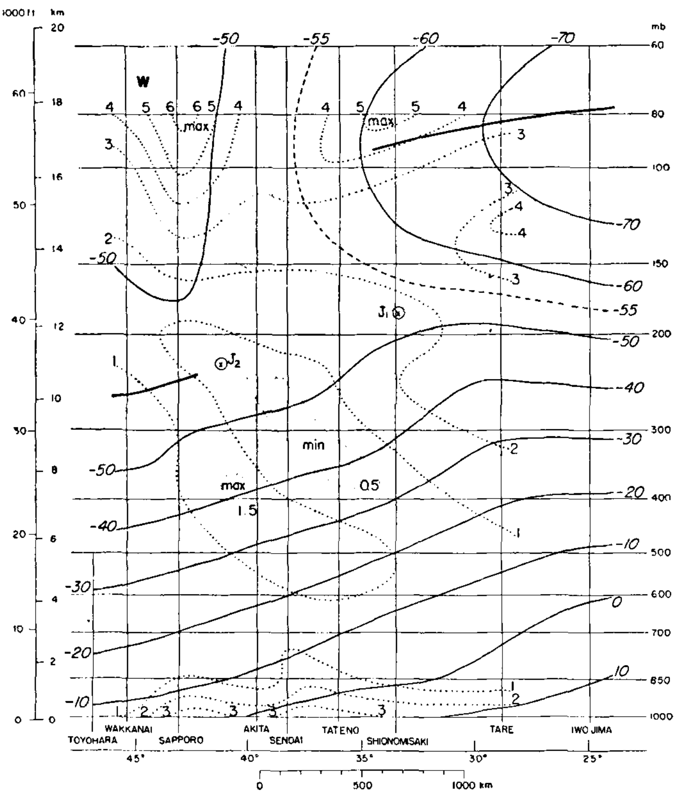

Fig. 8. Mean observed daytime temperatures over Japanese stations, Dec. 1950—Feb. 1951, thin solid lines (deg C). Dotted lines show difference (day minus night, C deg) between daytime and nighttime mean observed temperatures during this period.

Part II. Analysis of the jet stream and the deep stable layer during the period 8—15 December 1950

Purpose of the analysis.—On 10 and 11 December 1950, the Japanese Islands were covered by a moderate-sized anticyclone cell associated with a major cold-air outbreak behind a well-developed frontal cyclone (fig. 9). During these days, a pronounced deep stable layer was observed in the middle troposphere over central and southern Japan, extending west-east at least 20° longitude. At the time of its strongest development, this nearly-isothermal layer was as deep as 4 to 6 km, thicker and more pronounced in total horizontal temperature contrast than the usual frontal zone studied hitherto in many other cases of analysis in other regions. At the same time, above the stable layer a very strongly developed subtropical jet stream was observed in the upper troposphere, with a maximum observed wind of 116 m sec^{-1} (at 13 km, Tateno, 0300 GCT 11 December).

The purpose of this analysis is to describe this nearly-isothermal layer, and to give some suggestions regarding its possible origin. The analysis presented below is based mainly on vertical cross sections along the meridian 140° E twice a day. Cross sections are particularly useful in studying this type of phenomenon in the region of Japan because of the lack of detailed upper-air data to the west, over China, and to the east, over the Pacific Ocean. Unfortunately, the wind observations are insufficient to permit a detailed analysis of the wind field, so they must be supplemented by computations using the thermal field. As

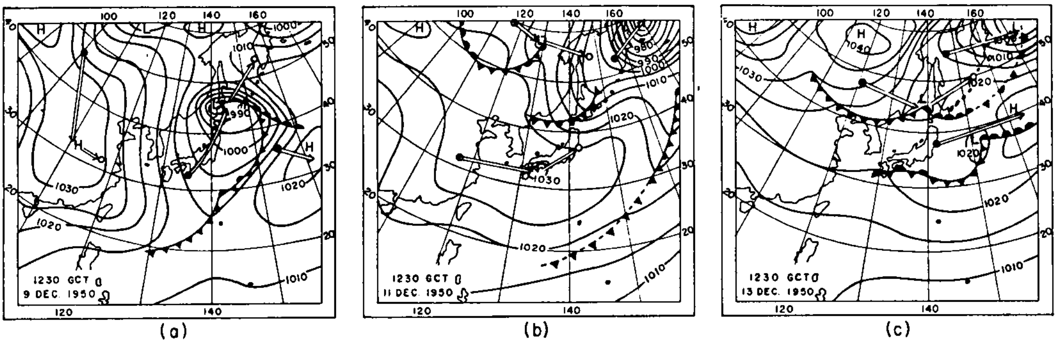

Fig. 9. Surface maps for 1230 GCT (a) 9 December, (b) 11 December and (c) 13 December 1950. Double shafted arrows indicate movements of pressure systems 24 hr before and after each map time.

pointed out in the previous part, the subtropical jet is rather steady, whereas the polar-front jet and associated frontal systems vary strongly in position and intensity; individual north-south cross sections therefore may reveal a good picture of the relative position of both jets.

General description.—The surface synoptic situation is shown in fig. 9. A cyclone formed on the southern coast of Japan on 8 December,

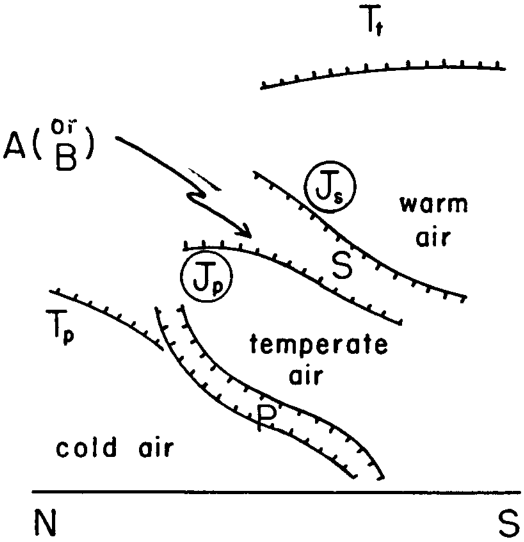

Fig. 10. Schematic diagram showing main discontinuity surfaces observed over Japan and vicinity. Nomenclature used in text and following figures: P, polar front layer; S, subtropical front; A or B lower boundary of S, corresponding at times to "tropopause" of temperate air; T_p, polar-air tropopause; T_t, tropical-air tropopause; J_p and J_s polar-frontal and subtropical jet streams.

Legend for figures 11—19: Meridional cross sections at 140° E, 1500 GCT 8 December to 1500 GCT 12 December 1950. Thin dashed lines, isotherms (deg C); very heavy lines, boundaries of principal stable layers; minor and low-level discontinuities marked with thinner lines. Winds plotted at 3-km intervals, oriented with respect to north at top of figure; full barb 10 knots, flag 50 knots. Potential temperature indicated beside lines for individual soundings (tens of degrees K.) Total sky converage and cloud types given at bottom. On sections for 0300 GCT, isotachs of geogradient west wind (thin solid lines, m sec^{-1}).

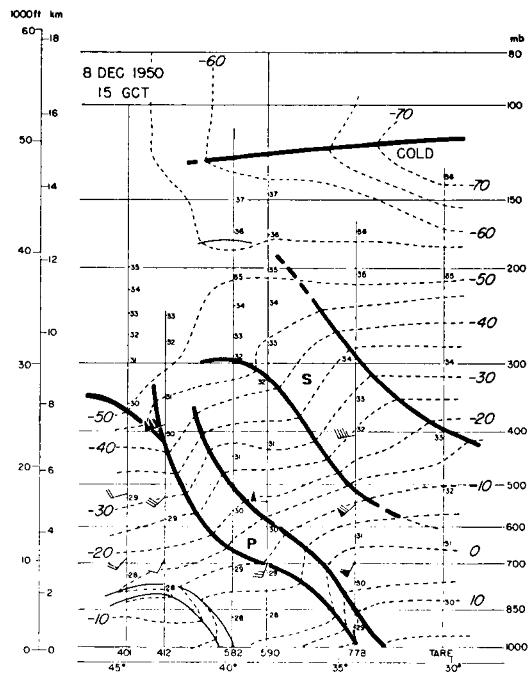

Fig. 11. 1500 GCT 8 December 1950. (Note two frontal layers.)

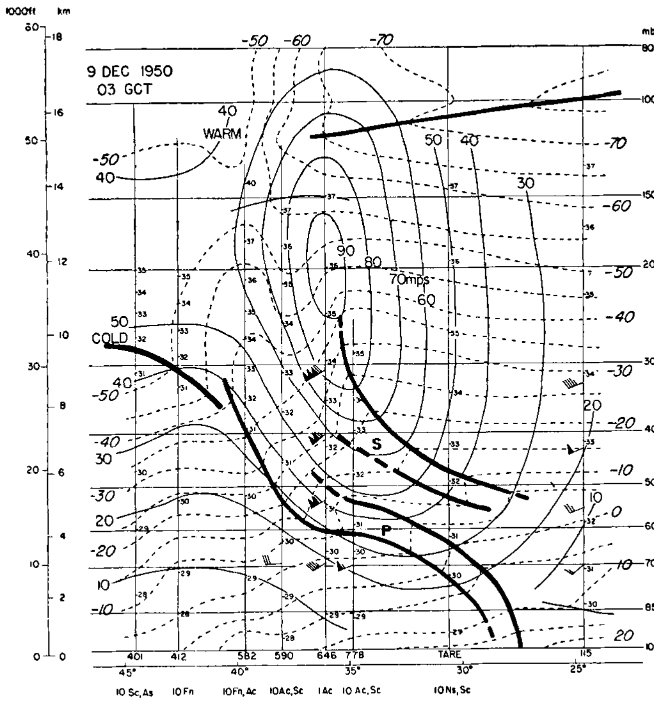

Fig. 12. 0300 GCT 9 December 1950. (Note two frontal layers approaching each other.)

deepening rapidly as it moved northeastward. A cold-air outbreak covered the Japanese Islands from 9 to 13 December with a typical traveling high cell over Japan on the 11th and 12th. A weak cyclone formed on the south coast of Japan on the 14th, and a small high cell covered Japan again in the 15th.

In the cross sections of figs. 11—19, all stations are projected onto the meridian 140° E along the contour lines at 200 mb (locations of stations shown in fig. 1). All available wind data and geogradient winds computed from the slope of the isobaric surfaces were used to obtain the wind field. Discontinuity lines (including frontal zones, tropopauses and inversion layers) were determined from the individual soundings plotted on tephigrams.

In order to describe the sequence of phenomena clearly, the discontinuity lines and jets in the cross sections are presented schematically in fig. 10, which gives the nomenclature of the various features referred to in the text. The discontinuities in fig. 10 agree in general with those shown in the schematic diagram of the winter meridional circulation given by Palmén (1951, p. 353). The lower boundarys of the stable layer "S" in fig. 10 corresponds to the tropopause commonly observed just sonth of the polar-front jet stream in other locations (see, e.g., Palmén and Nagler, 1948), which is found at much lower elevations than the true tropical tropopause.

Ordinarily in other locations the layer of pronounced stability above this tropopause is quite shallow, although it represents a real tropopause in the sense that the air above it does not usually return to the state of instability characteristic of the troposphere below it. In the examples shown in this paper, the layer of pronounced stability is quite deep, and there is a more-or-less distinct upper boundary above which the lapse rate becomes again somewhat more unstable. This stable layer thus may be regarded as some sort of front in the upper troposphere, although in its southern parts it gradually becomes indistinct and evidently does not reach the ground as a surface front. The stable layer is throughout the development most marked near the subtropical jet stream, and by reason of this association will be termed the "*subtropical front*".

The tropospheric air north of the subtropical front has temperatures characteristic of cool tropical air (e.g., at 500 mb, mostly around — 15 to —25° C). Above the stable layer, the air is mostly somewhat warmer than tropical air observed in other locations; e.g., at 200 mb in the series shown here the air is around 5° C or more warmer than in cross sections published for other locations (this was also true for mean conditions; see fig. 7).

The fact that the subtropical front seems to exist in such strength only in this geographical region suggests that its cause may be somehow connected with the influence of the Tibetan Plateau on the westerlies. Prof. E. Palmén suggested to the writer that the subtropical front is formed in this special location through a confluence of real tropical air masses moving in from the region south of the Himalayas, and cooler "temperate" air, originally tropical, which has been modified in passing over the regions north of the plateau of Tibet. Over Europe and North America where there is no pronounced orographic splitting of the westerlies, the modification does not vary so markedly with latitude, and the temperate air south of the real polar front may be regarded as part of the tropical air mass.

Detailed description.—Figure 11 shows a meridional cross section at 1500 GCT 8 December, when a young surface cyclone was developing and shifting along the southeastern coast of Honshu (Japanese main island). There is a definite polar front with potential temperature between 290 and 300 A. The polar front tropopause is also clear.

The upper boundary of the subtropical front is rather clear, connecting 300 mb at Shionomisaki (778) and 400 mb at TARE. In this stage the frontal systems P and S (see fig. 10) are separated by a broad zone of temperate air with weak baroclinity and are quite distinct as two different stable layers. In this stage there exist a polar tropopause and a tropical tropopause, whereas in the intermediate zone the tropopause to some extent can be regarded as the lower boundary of the subtropical frontal layer.

After 1500 GCT on the 8th, the polar front P shifted southward whereas the upper boundary of S was almost stationary with a slight tendency to become lower. Thus at 0300 GCT on the 9th (fig. 12) the polar front and the subtropical front began to form a deep transitional zone between the cold polar air and the warm tropical air. However, still in this stage the subtropical front S and polar front P are both distinct over some stations, such as Shionomisaki (778) and TARE, while the zone of temperate air has become very narrow.

Tateno (646), situated almost under the jet center, has an observed zonal wind component of 86 m sec^{-1} at 9 km, and thus the computed geogradient wind of nearly 100 m sec^{-1} at the jet center is reasonable. At this time, the calculated maximum anticyclonic wind shear (along an isentrope) south of the jet center was nearly 8.5 m sec^{-1} per 100 km, or 8.5×10^{-5} sec^{-1}. This approaches the criterion of dynamic instability for zonal flow (PALMÉN, 1948),

$$\left(\frac{\partial u}{\partial y}\right)_{\Theta} = f + \frac{u}{a} \tan \varphi,$$

since at latitude 34° f is 8.1×10^{-5} sec^{-1} and the correction term for vorticity due to the curvature of latitude circles, $(u/a) \tan \varphi$, is about 10^{-5} sec^{-1} for speeds of this size.

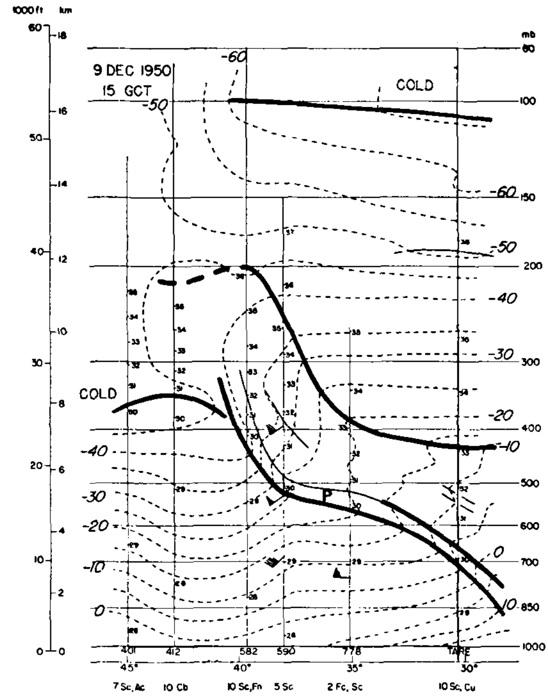

Fig. 13. 1500 GCT 9 December 1950. (Single deep stable layer in upper troposphere.)

A comparison with soundings at earlier times showed that both the subtropical front S and the polar front P became more stable during the past 24 hours, contributing to a sharpening of the horizontal temperature gradient and therefore to an increased wind speed in the main jet. At 1500 GCT on the 9th (fig. 13) the polar front and the subtropical front are being united into one deep stable layer and the temperate air is disappearing. The horizontal temperature gradient at 400—450 mb reached a value of nearly 4.5° C per 100 km, corresponding to a vertical wind shear of 20 m sec^{-1} km^{-1}.

Table 3. Stable layer at 0300 GCT 10 December.

Station	Depth (km)	Height (km)	Lapse-rate °C km^{-1}
TARE	3.5	2.0—5.5	1.4
646	6.5	3.5—10.0	2.3
590	5.5	5.5—11.0	1.1

By 0300 GCT 10 December (fig. 14), the nearly-isothermal layer in the middle tropo-

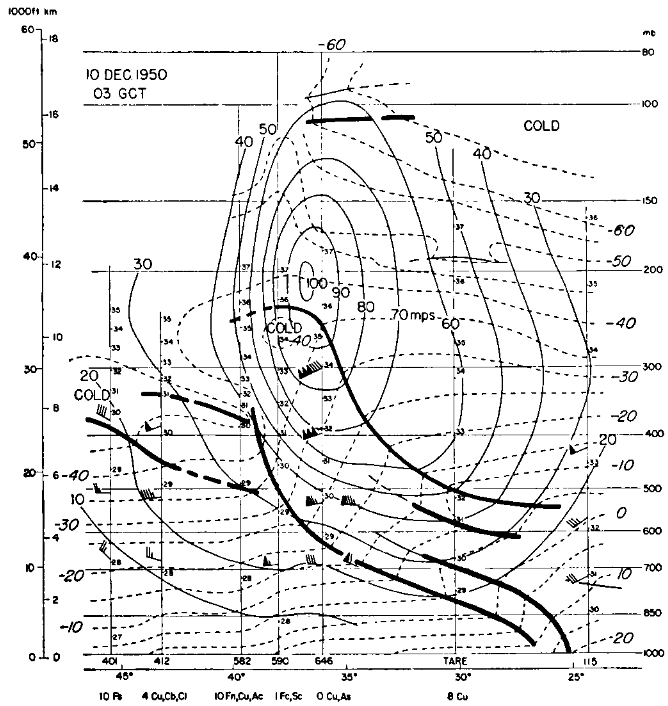

Fig. 14. 0300 GCT 10 December 1950.

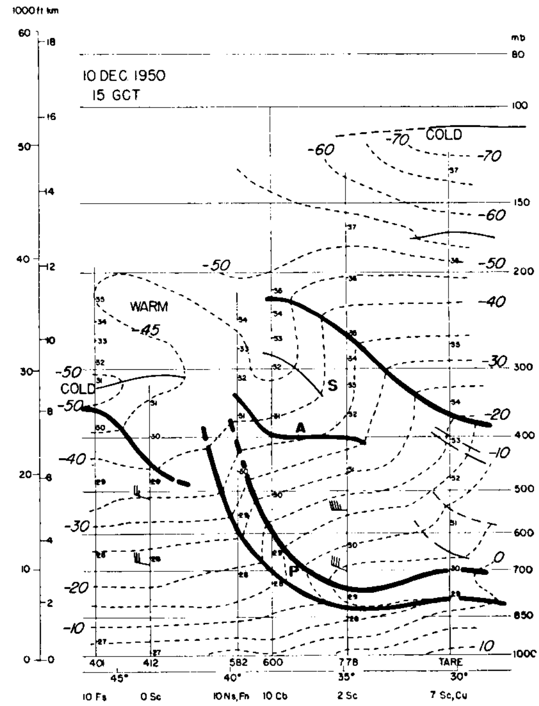

Fig. 15. 1500 GCT 10 December 1950. (Note subsidence of polar front.)

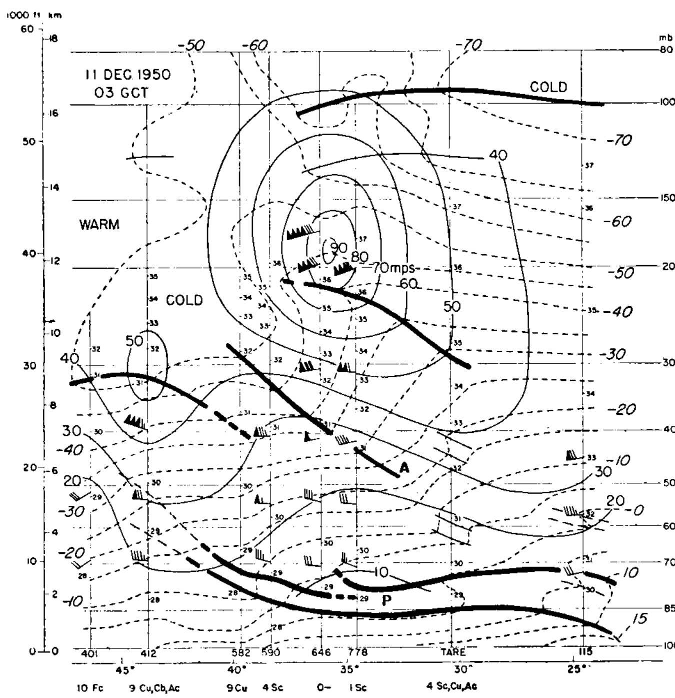

Fig. 16. 0300 GCT 11 December 1950. (Note boundary A subsiding, subtropical front deeper.)

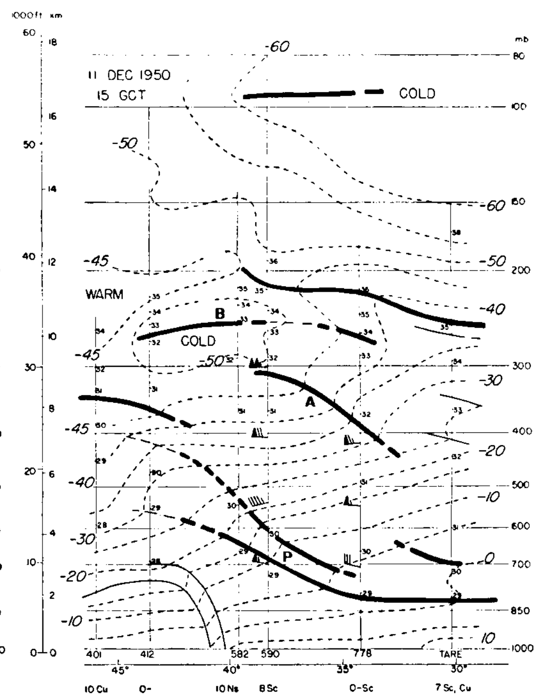

Fig. 17. 1500 GCT 11 December 1950. Note new tropopause B.)

sphere reached a remarkably pronounced development. The depths of the combined stable layer over some stations are shown in Table 3. At this time, the mean horizontal temperature gradient within the stable layer was approximately 3.4° C per 100 km at the the 7-km level, corresponding to a vertical wind shear of 15 m sec^{-1} km^{-1}. Actually, over Tateno (646) the zonal component of the observed vertical wind shear was 22 m sec^{-1} km^{-1} between 6 and 7 km, and 17 m sec^{-1} km^{-1} between 5 and 8 km. The computed zonal geogradient wind speed, 105 m sec^{-1} at the jet center at 0300 GCT on the 10th might be regarded as reasonable since it is verified by the observed wind at 9 km over Tateno extrapolated according to the horizontal temperature gradient. The calculated maximum anticyclonic shear south of the jet near 210 mb is 8×10^{-5} sec^{-1}, whereas $u/a \tan \varphi \approx$ $\approx 1.0 \times 10^{-5} sec^{-1}$; thus the vertical component of absolute zonal vorticity is

$$\zeta_a = -\left(\frac{\partial u}{\partial y}\right)_\Theta + f + \frac{u}{a}\tan\varphi \approx$$
$$\approx 1.5 \times 10^{-5}\ sec^{-1}.$$

This is very small, but it does not reach zero vorticity.

In the analyzed wind field of this and several other sections, there appeared to be branch of the jet extending downward toward the south, with strongest computed winds near 300—400 mb and decreasing winds above that level. It is not known to what extent this is real, since wind observations were lacking in this region. Undoubtedly part of the increase in temperature northward in the upper levels between Iwo Jima and TARE may be due to the different types of instruments used at these stations. However, a reversal of the horizontal temperature gradient at relatively low levels is also observed at some times north of TARE.

After 0300 CGT 10 December (figs. 14—16) the polar front subsides and moves southward in the lowest layers. During this process new temperate air masses gradually occupy the space between the subsiding polar air and the sloping subtropical front. Again after 1500 GCT 11 December (figs. 17—19), the subtropical front appeared to split into two layers, the layer marked "A" subsiding in very much the same manner as the polar frontal

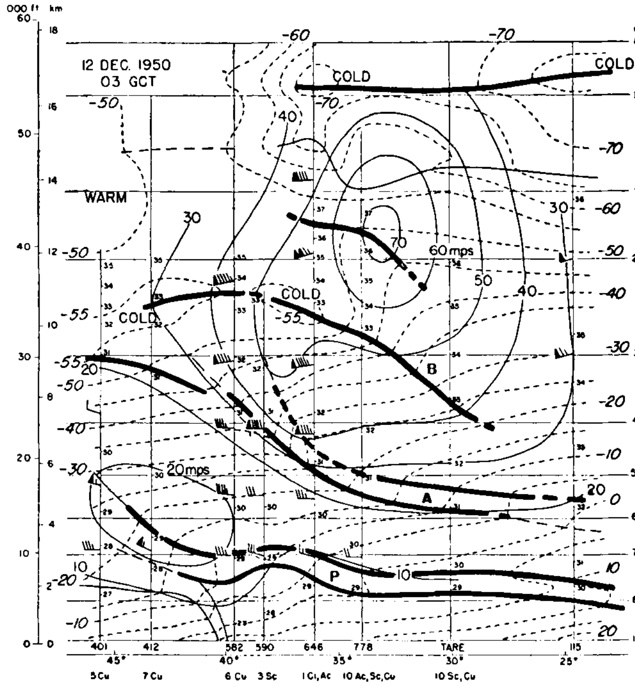

Fig. 18. 0300 GCT 12 December 1950. (Note boundary A now subsiding after second splitting of subtropical front).

layer did previously. This is difficult to understand but might become more clear if previous events upstream from Japan were known. At the end of the series (fig. 19) conditions again appear very much as they did at the beginning.

At first sight, the stage in figs. 14—16 appears to be a reversal of the process involved in the formation of the deep stable layer. However, a characteristic difference is that in the formative stage the polar front P shifted southward and did not change its inclination so much, thus a shifting of relatively parallel frontal layers was observed in the cross sections, whereas in the weakening stage of the isothermal layer, the polar front P reduced its inclination strikingly and this front became a rather flat inversion layer below the 700 or 800 mb level. During the formative stage the main jet center kept almost the same position and also its strong speed, but in the weakening stage it shifted southward and decreased in intensity (see fig. 20).

The stable layer above tropopause A near the end of the series is deepest at 0300 GCT 11

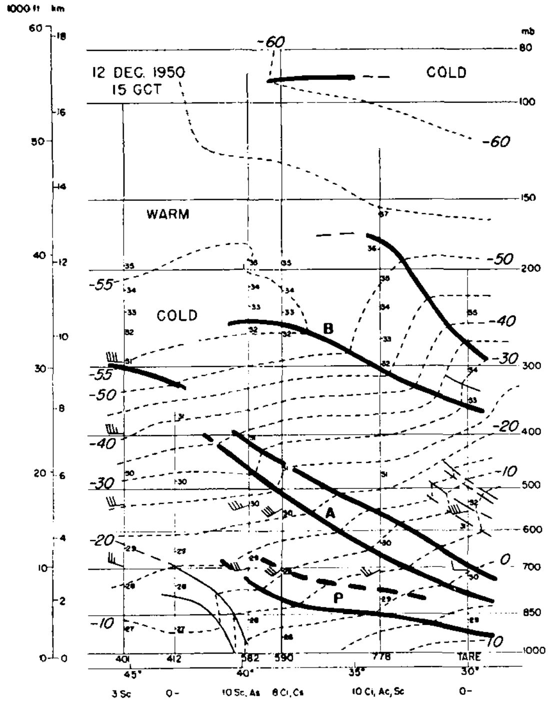

Fig. 19. 1500 GCT 12 December 1950. (Note subtropical front again well developed above discontinuity B.)

December, reaching a depth of 3.5 km over Tateno (646) and 4 km over Wazima 300 km northwest, the lapse rate being 1° C km^{-1} in both cases. This tropopause is recognized over Kagoshima (131° E, 32° N in western Japan) and over the fixed ship X-Ray (153° E, 39° N in the eastern sea of Japan) and thus is not a purely local phenomenon.

After 10 December (figs. 15—19), a region of cold air (< — 50° C) between 200 and 300 mb was observed to move southward. Associated with this southward shift of the cold region, a new jet appeared at 300 mb over Sapporo (412) at 0300 GCT 11 December (fig. 16). This jet shifted very rapidly southward, as seen in figs. 16 to 18. As shown in fig. 20 this jet moved to higher elevation and higher potential temperature with time, (cf. PHILLIPS, 1950) while the subtropical jet moved much less. Thus the main jet itself is of a rather stationary character, whereas the polar-front jet varies much in its position and merges into the main jet. The vertical wind shear above tropopause A is remarkable; over Tateno a wind of 116 m sec^{-1} was observed at 13 km at 0300 GCT on the 11th.

It might also be interesting to note that on the 10th a trough at 500 mb passed with a slight wind shift. It can be said that the deep isothermal layer was observed on the east side of the trough, and the layer split after its passage. However, the detailed relationship is not clear at present.

After 1500 GCT 11 December (fig. 17), tropopause A subsided and transformed into a lower boundary of some kind of front in the middle troposphere, and a new tropopause B appeared. The statement that a tropopause transformed into some kind of front might be subject to some discussion. It is difficult to define what is a tropopause and what is a front and the phenomenon observed here further complicates the matter. However, we know that in cases with multiple tropopauses the potential temperature at each one of the tropopauses is usually relatively constant at a given time, even though the tropopause height varies (BJERKNES and PALMÉN 1939). Therefore the successive positions of the tropopause on the cross sections might be determined with the aid of potential temperature. Table 4 shows the potential temperatures of tropopauses A and B.

Table 4. Tropopause A.

Station Date/Time	778	646	590	582
10 15 GCT	315	—	314	311
11 03	308	310	313	320
15	314	305	313	(312)
12 03	306	(304)	306	—
15	(297)	—	303	301
13 03	(298)	—	—	—
Tropopause B.				
11 15	336	—	—	330
12 03	328	322	326	333
15	318	322	321	—
13 03	328	—	—	—

Summary.—To understand the phenomena as analyzed above more clearly, it might be useful if we formulate the four stages into a series of models. As already stated at the beginning of Part II, the main point is the formation and dissolution of the deep nearly-isothermal layer. The changes observed in figs. 11 to 19 are illustrated schematically in fig. 21, and may be described in four stages as follows: Stage I — The polar front and the subtropical front exist as completely different layers separated by a broad zone of temperate air with weak baroclinity. In this stage there exist a polar tropopause and a tropical tropopause, whereas in the intermediate zone the tropopause to some extent can be regarded as the lower boundary of the subtropical frontal layer.

Stage II — The two frontal layers are approaching each other so that the zone with temperate air becomes very narrow and gradually disappears.

Stage III — Only one very deep almost isothermal zone separates the real tropical air in the middle troposphere from the polar air.

Stage IV — The polar air subsides and moves southward in the lowest layers. During this process new temperate air masses occupy gradually the space between the subsiding polar air and the sloping subtropical front. At the end of this stage the type I gradually forms.

There were not enough direct wind observations to study the changes of the wind field exactly. However, from the observed temperature fields it appears likely that at first separate jets (marked J_1 and J_2 in fig. 21) are associated with the subtropical and polar fronts, merging as these fronts combine into a single layer (stage III). Fig. 21 is only intended to give a simplified picture of the larger-scale process. As pointed out above, the subtropical front split a second time, repeating the process in stages III and IV. There also were several weaker stable layers and secondary fronts in

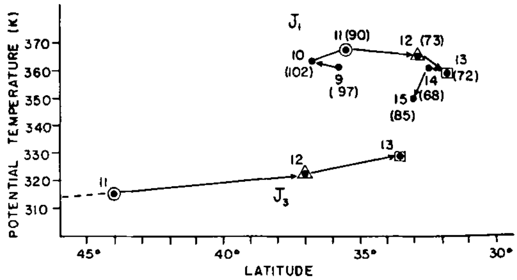

Fig. 20. Changes in latitude and potential temperature at centers of jet streams, 9 to 15 December 1950 (all 0300 GCT), dates given beside dots showing daily positions. J_1, main subtropical jet (maximum wind speed in parenthesis, m sec^{-1}); J_3, polarfront jet (note large latitudinal movement compared with subtropical jet on same days).

lower levels during the above series, which have not been discussed in detail.

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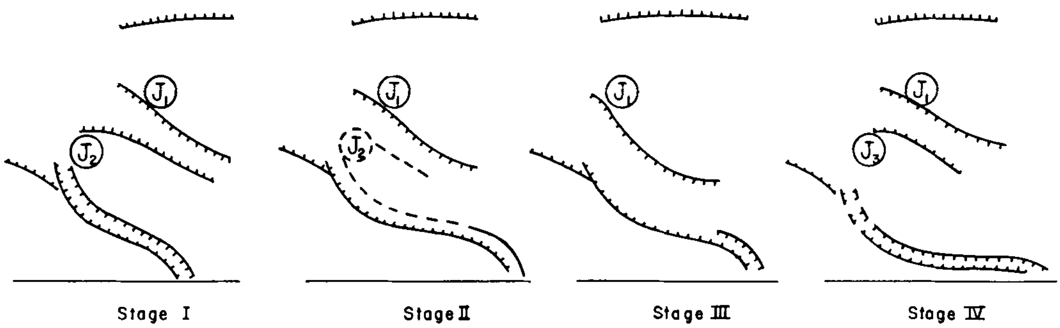

Fig. 21. Schematic diagrams showing changes of structure of main discontinuity layers over Japan, during period 8 to 12 December 1950.

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