Origin and Structure of High-Level Cyclones South of the Maximum Westerlies

By E. PALMÉN, University of Helsingfors

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Abstract

This study deals with a special type of cyclones formed in the belt of the upper westerlies. The development begins as wave-like deformation of the upper flow and this deformation increases in amplitude until the cold air in the southern part of the trough becomes separated from the cold air in the north. The process of separation may be described as a cutting-off of a cold vortex. The process is described in some detail for a special case. High-level cyclones formed in this manner are not rare: on the contrary, such upper air disturbances are of great importance for the weather phenomena in certain regions, especially at latitudes between 30°—50° N. They can also be regarded as an important link in the large-scale quasi-horizontal meridional exchange of air masses.

I. General Results

The westerlies of middle latitudes appear regularly on upper air charts and in meridional cross sections as a rather narrow belt of strong wind embedded in relatively stagnant air masses to the south and to the north. A characteristic example of this type of upper air motion, the case of November 30, 1946, has been discussed in some detail by E. PALMÉN and K. M. NAGLER (5).

In many cases a similarly strong concentration of the westerlies into a relatively narrow "*jet stream*" has been observed over the North American continent, especially during the cold season. The same phenomenon can be observed over the Pacific and the Atlantic Oceans and also in Western Europe. An example of such a jet stream over the British Isles has been given by E. HOVMÖLLER (4).

In most cases the belt of the strongest upper westerlies does not follow a straight course, and considering the whole northern hemisphere we may say that it never happens. *On circumpolar charts for the 500 mb surface the belt of the strongest westerlies has the appearance of a "meandering river", situated just to the south of or in the zone with the strongest horizontal temperature gradient* (8). This zone of crowded isotherms can be regarded as the intersection between the sloping polar front surface (layer) and the 500 mb surface. Therefore, the polar air has its greatest southward displacement in the troughs of the upper disturbances, its greatest northward displacement in the ridges.

The identification of the zone of strongest isotherm concentration at the 500 mb level with a polar front is contradictory to a rather common opinion among meteorologists who regard the polar front as a phenomenon of the lowest troposphere. A careful analysis of a great number of special synoptic situations in North America and also in Europe indicates in many cases the existence of such a frontal layer up to the level of 400 mb or sometimes more. The concentration of the meridional solenoid field is not always so pronounced that the transitional layer between the tropical and the polar atmosphere appears as a frontal discontinuity according to the common de-

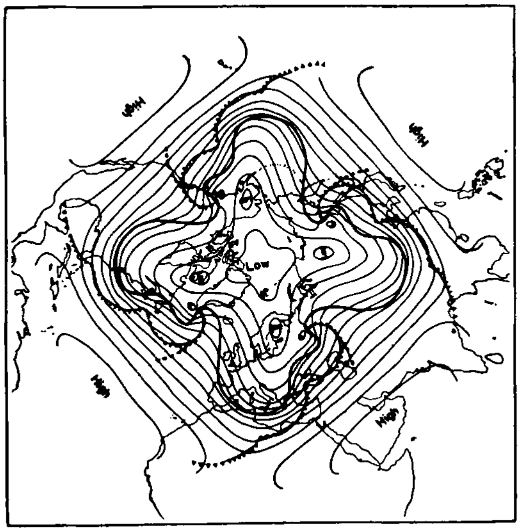

Fig. 1. Schematic circumpolar chart for the 500 mb level showing four long waves. The heavy line indicates the "polar front" and the thin lines contours (topography) of the 500 mb surface; corresponding fronts and frontal perturbations at the earth's surface are marked by usual symbols.

finition of the word. The conception of a front thus needs some modification, if we want to use it in analyses of 500 mb charts.[1]

If the belt of strong isotherm concentration on upper charts can be regarded as the boundary between the polar and the tropical atmosphere, the upper waves, therefore, must be connected with polar front disturbances in the middle troposphere. These upper waves, however, often are of a larger scale than the typical polar front perturbations of the surface layers. Both types of disturbances influence each other and cannot, therefore, be regarded as independent phenomena. Sometimes the long waves in the upper polar front which are very closely connected with the waves in the westerlies, studied by C. G. Rossby (7), correspond to the cyclone families as first described by J. Bjerknes and H. Solberg (2). In such cases the individual polar front cyclones of lower levels appear as minor wave-like disturbances at the 500 mb level superimposed upon these large waves. In fig. 1 a schematic circumpolar chart shows four long waves at the 500 mb level, the corresponding upper polar front at the same level, the minor disturbances and the polar fronts at the earth's surface including their cyclonic disturbances. However, it might be pointed out that in some cases the upper long waves can be directly connected with individual strong surface cyclones. In such cases the upper long waves mostly appear as rather fast moving systems.[1]

The upper waves often show a pronounced instability, indicated by a rapid increase in their amplitude. At the 500 mb level, the band of strongest isotherm concentration, which during the colder season normally is located in the vicinity of latitude 50° N, may then be pushed southward to latitude 40—30° N. In the lower atmospheric layers the process is connected with a pronounced anticyclonic outflow of polar air masses from the polar source region. Thus there must be a pronounced subsidence, combined with horizontal divergence, in the lower layers and a strong convergence, combined with increasing cyclonic vorticity, in the upper layers.

At the beginning of the development the depth of the cold air in such an upper through increases northward along the trough line. If strong subsidence occurs the depth of the cold air mass gradually decreases. If the sinking cold air is not replaced by advection the polar front at the level of 500 mb may then sometimes show a displacement to the north. Thus the movement of the polar front at the 500 mb chart, may be in a direction quite different from that of the gradient wind. It is important to observe this difference between the frontal movements on the surface charts, where the vertical wind component vanishes, and on the upper charts.

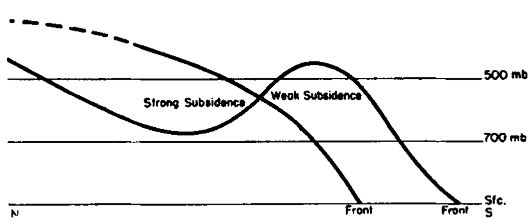

Fig. 2. Schematic meridional cross section through an upper trough showing the profile of the polar air before and after the formation of a "cut-off" low.

[1] The author is preparing a paper about this special subject.

[1] It might be pointed out that the number of upper long waves can vary between about 3 and 7 depending upon the speed of the westerlies.

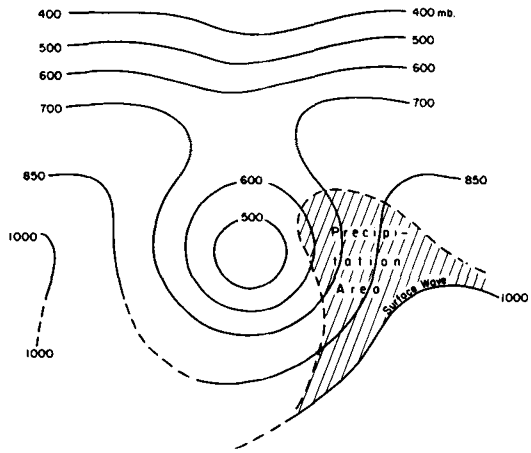

Fig. 3. Topography (mb) of the polar front in the case of a schematic high-level cyclone; the characteristic precipitation area on the earth's surface is also marked.

In many cases, however, the strongest subsidence does not occur in the southernmost parts of the cold upper trough, but in its central parts. As a result the cold air gradually disappears from the 500 mb level in the central parts of the trough, but still remains at the same level farther to the south and to the north. At an upper chart the phenomenon thus appears as a "*cutting-off*" of the cold air from its polar source region. The phenomenon is described in fig. 2 which shows the idealized profile of the polar front surface in a meridional cross section before and after the cutting-off process. In fig. 3 the cut-off cold air is represented with the aid of a schematic frontal contour map. A similar scheme has earlier been presented by A. M. CROCKER, W. L. GODSON and C. M. PENNER (3).

The fact that the shrinking often is much slower in the southern than in the central parts of a trough must depend upon the very strong combined effect of the Coriolis and centrifugal accelerations at the southern boundary of the cold air tongue. This effect tends to produce a cross-stream vertical circulation opposite to the direct solenoidal circulation. In some cases a strengthening of the solenoid field, due to the combined effect of the Coriolis and centrifugal accelerations, has been observed, resulting in a temperature fall of many degrees at the 500 mb level.

On upper air charts the cut-off polar air appears as a cold cyclonic vortex formed from the southern part of the original upper trough. Farther to the north, in the region of the newly formed frontal zone, a belt of very strong west wind can be observed. Thus the whole process may be regarded as a special case of a cyclone formation on the south side of the strongest westerlies.

Upper cold vortices of this type reach their strongest intensity at the tropopause level. Because of the temperature distribution the vortex decreases in intensity downwards in the troposphere, and in many cases it completely disappears in the lowest atmosphere or changes into an anticyclonic vortex. Thus, the frictional loss of momentum is mostly small because of the weak wind at the surface. Sometimes there can be a tendency for spreading of the cyclonic vortex downwards, probably because of a continuous transport of momentum from the upper troposphere. In such cases the original anticyclonic flow in the surface layers gradually changes to a cyclonic one. Many cases of such a development could be observed over the North American continent during the years 1946—1948. I the stratosphere the high-level cyclones appear as vortices with warm core. Thus, their intensity decreases with height above the tropopause.

The cut-off cold low can be regarded as a cell in a large-scale meridional exchange of air masses. A corresponding transport of warm tropical air masses to the north can often be observed in the upper troposphere and the lower stratosphere. These cells of tropical air appear on the 500 mb charts as tongues or cut-off islands of warm air with the characteristic high tropopause and low stratosphere temperature. The phenomenon has been studied at the University of Chicago, and some results will be published later. Because of the difference in absolute vorticity [see ROSSBY (7)] between the air masses in their original source regions it is obvious that the northward moving tropical air in higher latitudes must be characterized by low absolute vorticity. The tongues or cut-off islands of tropical air thus always show a pronounced anticyclonic relative vorticity and appear therefore as high-level anticyclones formed to the north of the original maximum westerlies.

It is obvious that the formation of cold cyclones to the south and of warm anticyclones to the north of the maximum westerlies

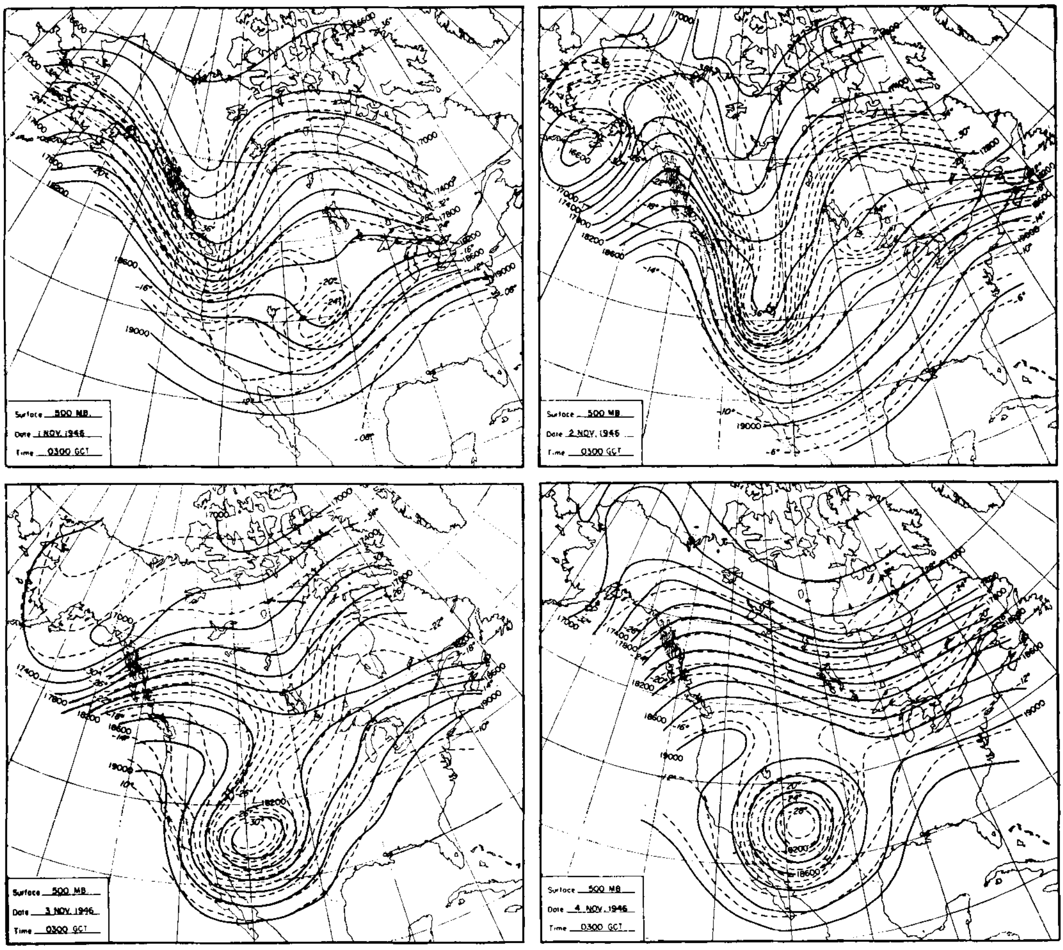

Fig. 4. November 1—4, 1946, 03h G. M. T., 500 mb, with contour lines for every 200 ft and isotherms for every 2° C (dashed lines).

represents a very effective large-scale meridional exchange of air masses, which is necessary for the explanation of the average meridional temperature field. However, this exchange is not possible through a purely horizontal air transport. The vertical component of the air transport, although small compared with the horizontal, is very important from the viewpoint of the maintenance of energy. Connected with the vertical component of air motion during the development of large cells of different air masses are certain types of vertical circulation and horizontal divergence fields. Thus, no complete conservation of absolute vorticity can be expected.

II. The High-Level Cyclone of November 1—7, 1946

In the following a brief description of the formation and structure of a typical high-level cyclone over the southwestern parts of the United States will be given. Fig. 4 shows the 500 mb charts with contour lines for every 200 ft. and isotherms for every 2° C for the period November 1—4, 1946. On the first day a very pronounced cold trough lay along the American west coast. The strong concentration of isotherms indicates the existence of a real frontal zone separating the polar air in the central parts of the trough

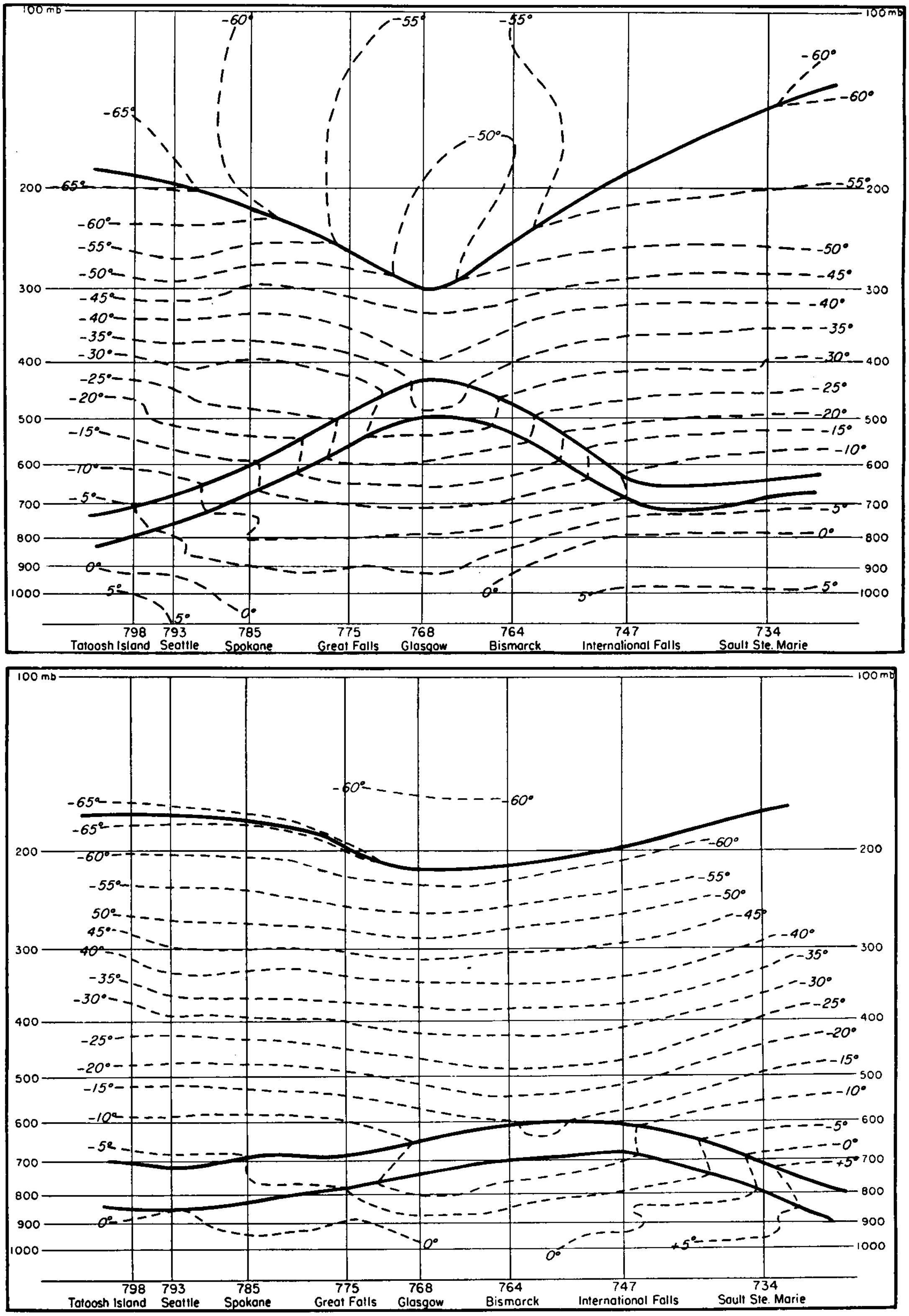

Fig. 5. Vertical cross sections approximately along the parallel 48° N for November 2, 15h and November 3, 03h G. M. T. showing frontal layer, tropopause and isotherms for every 5° C.

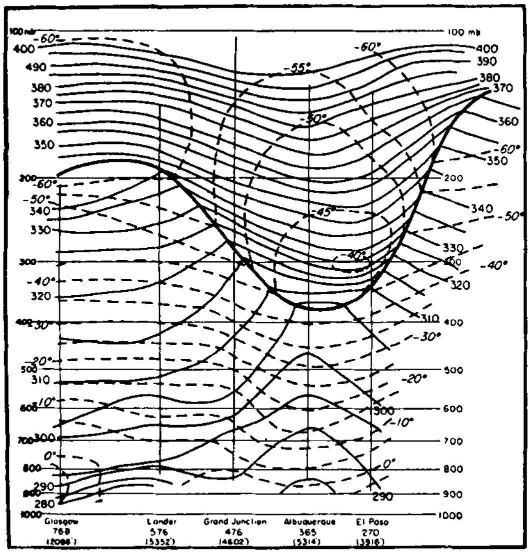

Fig. 6. Vertical north—south cross section through the centre of the high-level cyclone, Nov. 4, 03^h G. M. T. Heavy line indicates the tropopause, thin lines denote isentropes and dashed lines isotherms.

from the surrounding warmer air masses of tropical origin. Instead of moving eastward the amplitude of the trough gradually increased, and the cold air was displaced farther to the south or southeast, as can be seen from the charts. On November 3 a closed cold centre was formed over New Mexico, and on November 4, the upper cold cyclone took definitive shape. The centre over Albuquerque in New Mexico shows a very low temperature (— 29° C) for this latitude, but the temperature has risen about 6° C in the centre, indicating a subsidence in the southern parts of the cold air at the level around 500 mb. Farther to the north, however, the temperature rise was extremely strong, at the latitude 45° N from — 36° C on November 2, to about — 16° C two days later.

It is not easy to estimate quantitatively the vertical velocity component of the moving cold air masses in different parts of the upper trough. However, an idea of the order of magnitude of the subsidence can be obtained from the study of vertical cross sections through the cold air. As an example of such analyses, two west-east cross sections along latitude 48° N are reproduced (fig. 5). The sections are drawn with a time interval of 12 hours, the first for Nov. 2, 1500 GMT, the second for Nov. 3, 0300 GMT. The frontal layer, indicated by two heavy lines, appears in both sections as a distinct feature in the temperature. During these 12 hours the upper boundary of the frontal layer sank from a maximum height of 440 mb to 600 mb which corresponds to a vertical displacement of about 2.4 km. That gives a mean vertical velocity of about — 5.7 cm/sec. During the same time a pronounced increase of the tropopause height (about 2 km) and decrease of the tropopause temperature in the upper trough can be observed.

This type of vertical displacement must be connected with strong upper convergence and lower divergence. In a coordinate system moving eastward with the speed of the trough line, the mean divergence between 5 and 1 km (the approximate surface of the earth) is about $1.4 \cdot 10^{-5}$ sec^{-1} and between 5 and 10 km $-2.1 \cdot 10^{-5}$ sec^{-1}. If ζ is the absolute vorticity of an air parcel the individual change of vorticity $\frac{d\zeta}{dt}$ is determined by the formula

$$\frac{d\zeta}{dt} = -\zeta \ \mathrm{div}_{\cdot H}\mathbf{v} \ldots\ldots\ldots \quad [1]$$

An air parcel moving in the divergence field between 5 and 10 km thus shows an increase

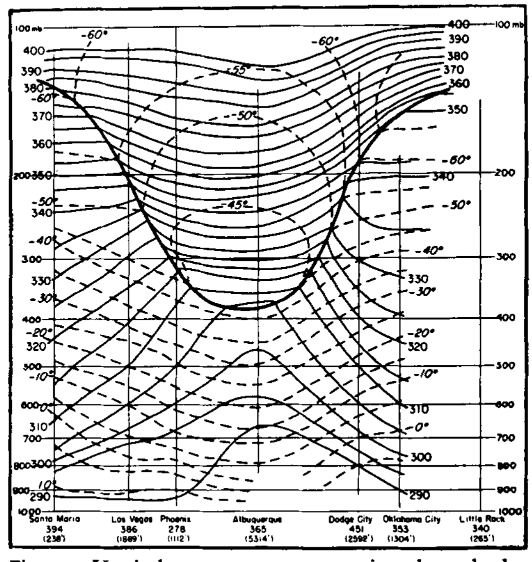

Fig. 7. Vertical west—east cross section through the centre of the high-level cyclone, Nov. 4, 03^h G. M. T, Heavy line indicates the tropopause, thin lines denote isentropes and dashed lines isotherms.

in vorticity of the order of magnitude $2 \cdot 10^{-9}$ sec^{-2}, a very great value, since the absolute vorticity is of the order of magnitude of the Coriolis parameter (10^{-4} sec^{-1}). One can therefore expect, that the upper warm air, which on the chart for Nov. 2 moves anticyclonically along the Canadian west coast, must change its movement in over the continent. That is just what can be seen from the following maps.

However, it must be emphasized that this phenomenon cannot be studied as a local process. The upper air masses move very fast, and the warm air in the cross section for Nov. 3 was during the previous day over the ocean. It is difficult to compute the vorticity here because of lack of observations, and therefore it is not possible to determine the individual change of vorticity with satisfactory exactness. C. G. ROSSBY (6) has strongly emphasized the interaction between different atmospheric systems.

The high-level cyclone appears on the 4th of November as an almost symmetric upper vortex. Fig. 6 and 7 show the distribution of temperature and potential temperature in two vertical cross sections drawn in north—south and west—east directions through the cyclonic centre over Albuquerque. In both cross sections the tropopause is marked by a heavy line. No detailed analysis of the tropopause has been made, and the tropopause was therefore regarded as a continuous surface of discontinuity in the temperature gradient separating an upper stratospheric vortex with warm core from a lower tropospheric vortex with cold core. The original frontal surface which during the process became more and more indistinct, has not been marked in the sections.

In the west—east cross section the vortex appears to be almost completely symmetric, but in the north—south cross section a certain asymmetry can be noticed. However, the latter section could not be studied very carefully for lack of observations in the south, especially over Mexico. The distribution of gradient wind, corrected for the approximative curvature of the air trajectories, is given for the west—east cross section in fig. 8. The figure demonstrates the nearly symmetric distribution of wind velocity (N—S component) in the vortex.

In order to give a complete three-dimensio-

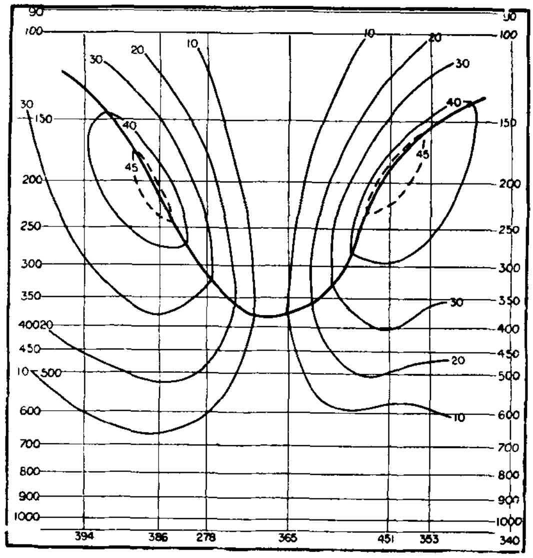

Fig. 8. Distribution of gradient wind corrected for the curvature of the air trajectories in the W—E cross-section through the high-level cyclone of Nov. 4, 03^h G. M. T. Wind velocity is given in m/sec.

nal picture of the vortex the charts for the earth's surface, 700 mb, 300 mb and 200 mb are reproduced in fig. 9. On the surface map no cyclone can be observed in the region where the upper air charts show the deep cyclonic vortex. At the 700 mb level the cyclone is still weak and much better marked in the temperature field than in the pressure field. At the levels of 300 and 200 mb the intersection between the tropopause and the corresponding isobaric surfaces is marked by a heavy curve; the temperature field in the vicinity of these heavy lines indicates the existence of a zone of lowest temperature around the tropopause. The same characteristic temperature distribution can be found in most cases where the tropopause cuts an upper air chart.

The almost complete disappearance of the cyclone at the earth's surface is characteristic of disturbances of this type, which are not formed as the result of any occlusion process. As pointed out above, the weak surface wind reduces the dissipation of kinetic energy to a minimum, and because of the symmetry of the formation there can be no strong tendency for the vortex to move. Actually the centre of the high-level cyclone, discussed here, remained practically over the same place during a period of three days. Cross sections

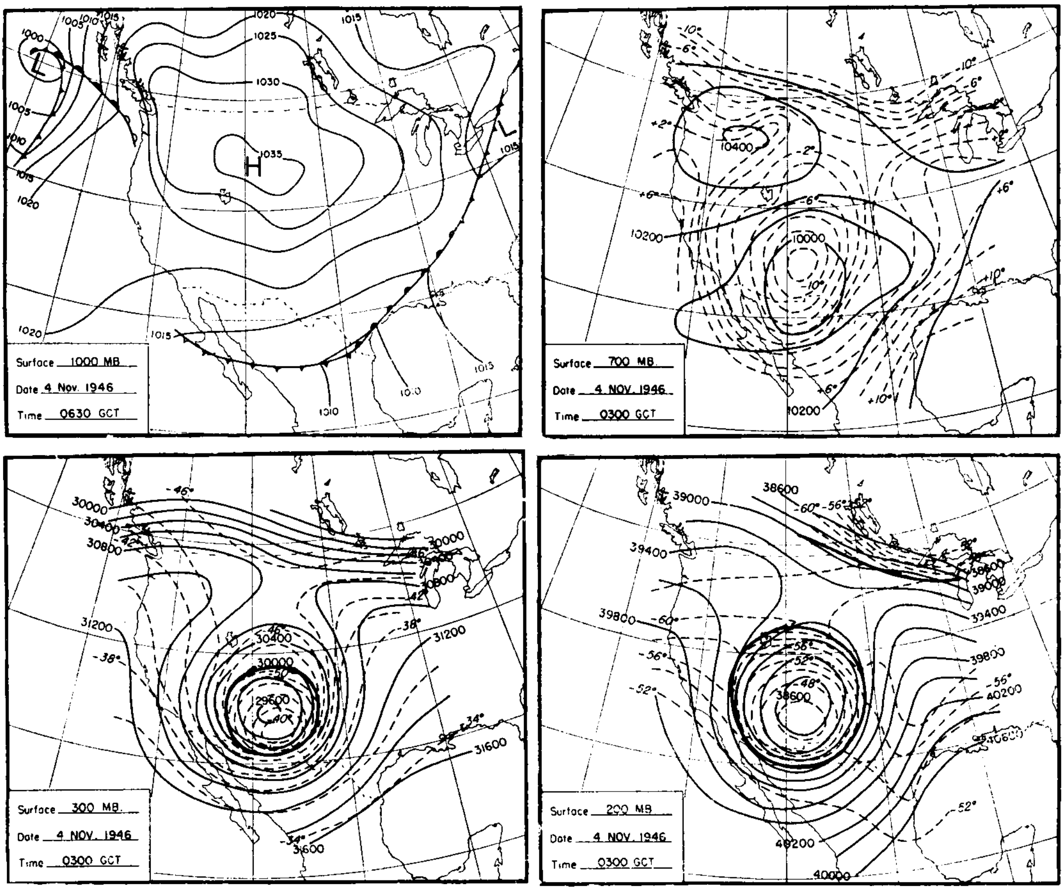

Fig. 9. Charts for the earth's surface and for 700, 300 and 200 mb on Nov. 4, 03h G. M. T. The heavy lines at the charts for 300 and 200 mb indicate the intersection between the tropopause surface and the corresponding isobaric surfaces.

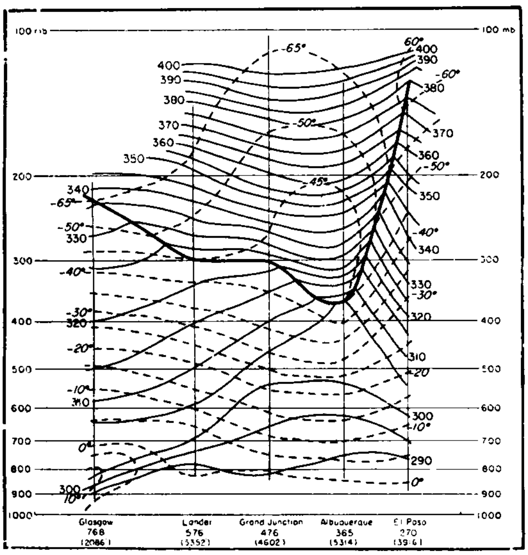

Fig. 10. Vertical N—S cross section through the centre of the high-level cyclone, Nov. 6, 03h G. M. T.

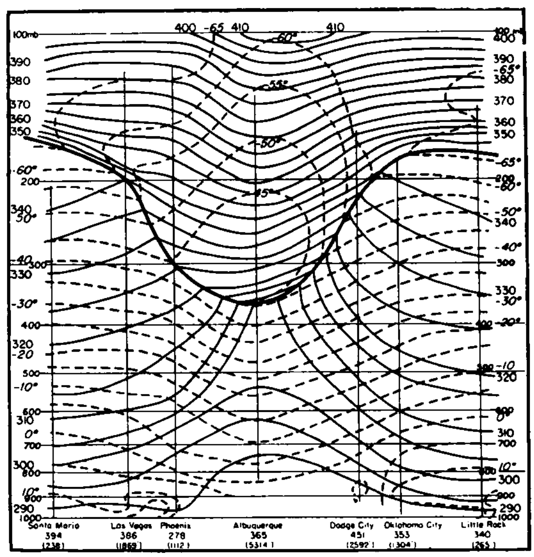

Fig. 11. Vertical W—E cross section trough the centre of the high-level cyclone, Nov. 6, 03h G. M. T.

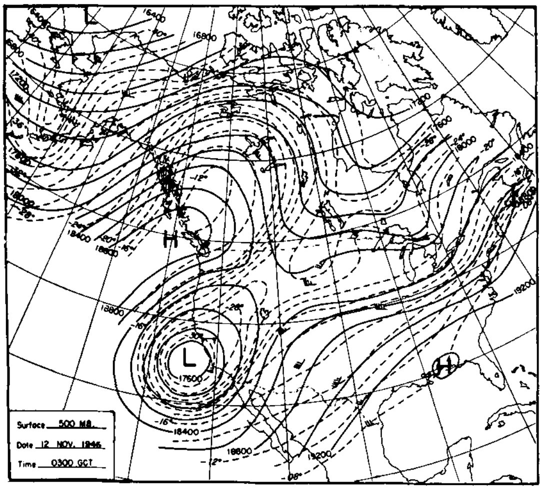

Fig. 12. November 12, 1946, 03^h G. M. T., 500 mb with contour lines for every 200 ft, isotherms for every 2° C and winds.

corresponding to those in fig. 6 and 7 indicate that the centre of the vortex on November 6 was still over Albuquerque (fig. 10—11). In the west—east cross section the disturbance appears as an almost symmetric vortex, but in the north—south cross section the asymmetry is now quite pronounced. At the same time the disturbance begins to move rather rapidly in the ENE direction, and the whole synoptic situation changes quickly.

The symmetric picture of wind and temperature distribution which can be found from the charts and cross sections for Nov. 4, does not carry with it a similar symmetry in the weather phenomena. On the contrary, the high-level cyclone was accompanied by heavy precipitation on its east side and clear skies on its west side, indicating the existence of a strong upper divergence field compensated by convergence in the lower levels on the east side, and vice versa on the west side. The combined effect of the horizontal divergence and the vertical motion must have been such that the local pressure changes practically vanished. Thus, in an arbitrary point of the vortex the vertical mass transport from below must be equal to the total mass divergence above the point, or

$$\int^{\infty} \mathrm{div}_H (\varrho \mathbf{v})\, dz = (\varrho w)_z \ldots\ldots \quad [2]$$

In this expression ϱ denotes the density, $\mathbf{v}$ the horisontal wind vector, and $(\varrho w)_z$ the vertical mass transport through the level z. Because of the asymmetry of the temperature field in N—S direction (fig. 4) the divergence term in [2] increases with height in the

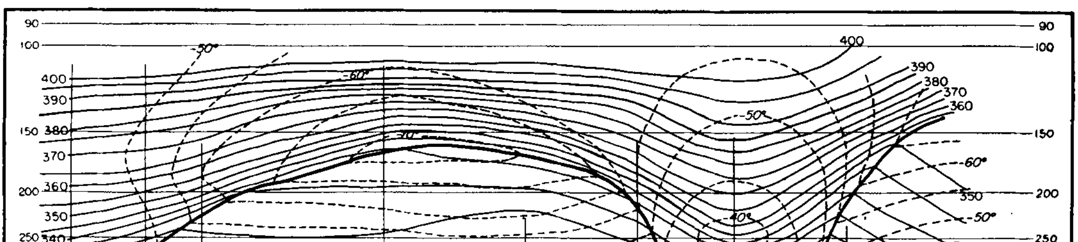

Fig. 13. Vertical cross section approximately along the West Coast of America from Northern Alaska to Southern California. Heavy lines indicate fronts or tropopause, dashed lines isotherms.

eastern part and decreases in the western up to the level of the tropopause.[1] Thus, there must be a level of maximum vertical ascending or descending air movement approximately at the level of non-divergence. The thermal structure of the vortex is therefore in good agreement with the actually observed distribution of cloudiness and precipitation.[2] Because of lack of observations, especially in the southern part of the disturbance, it is not possible to get a quantitative computation of the divergence field and the corresponding vertical displacement of the air in different parts of the vortex.

III. **The Cyclone of November 10—14, 1946**

A similar development of a high-level cyclone occurred some days later over the coastal regions of California. Without giving any detailed description of this case we reproduce only two figures showing the atmospheric conditions on November 12, 03^h G.M.T. at the level of 500 mb and in a vertical cross section almost along the American west coast from San Diego in Southern California to Aklavik in Northern Canada. The centre of the cyclone was on that day situated just south of Oakland (Calif.). The cross section gives a similar picture of the thermal structure of the disturbance as the cross sections in fig. 4 and 5.[1]

Acknowledgement

The writer wishes to thank Miss Chi Kung Chen for her valuable help in analysing the upper air charts and cross sections used in this paper.

[1] See the discussion of this problem given by J. Bjerknes and J. Holmboe (1).

[2] Compare with the area of precipitation in the schematic figure 3, where the curves showing the topography of the frontal surface qualitatively can be regarded as isobars of an upper air chart.

[1] After finishing this paper the author got knowledge of the following paper by W. Schwerdtfeger: »Untersuchungen über den Aufbau von Fronten und Kaltlufttropfen», Bericht des Deutschen Wetterdienstes in den U. S.-Zone. No. 3., which partly deals with the same problem.

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