A FURTHER STUDY ON THE RELATION BETWEEN THE JET STREAM AND CYCLONE FORMATION

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ABSTRACT

Events at sea level and aloft over the United States are related to the approach of a speed maximum in the jet stream. In the left hand portion of the area downstream from the jet maximum where air at the jet level is decelerating, frontogenesis, cyclogenesis, and spread of precipitation occur. Other indications of high level divergence to the left of the advancing jet maximum are given by changes in the structure of a nearby cold dome. Insofar as the changes in the cold dome are precedent, they constitute a means of forecasting cyclogenesis.

INTRODUCTION

In a previous paper (Riehl 1948) the following comment concluded a study on the relations between the jet stream in the high troposphere and cyclone formation:

"It is suggested that the jet stream appearing in connection with a pattern of very long waves in the westerlies provides a mechanism for the initiation of cyclone development and an increase of wave number. But it should be emphasized again that only one species of cyclone formation has been considered here, namely, that associated with initial westerly flow aloft without pronounced streamline curvature. Nor is it suggested that the jet alone can create cyclones. It is evident that the jet is effective only if it is superimposed on a disturbance of the lower atmosphere. Clearly, the amount of cyclonic development to be expected, depends in large measure on this factor. Therefore, when jet stream, long wave pattern, and low tropospheric disturbance coincide in a favorable sense, ensuing cyclone developments will attain the greatest intensity."

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1. Participated under research contracts between the Office of Naval Research and The University of Chicago.

2. Participated while on assignment to the Advanced Forecasters Course sponsored by the Weather Bureau and The University of Chicago.

It is the objective of this report to determine more precisely some of the favorable circumstances mentioned. As in the previous paper, we shall do this with an example whose salient features are typical of a large group of cases, though not of all cases.

In the course of an experiment in forecasting carried out jointly by the University of Chicago and the U.S. Weather Bureau we observed that strong cyclogenesis frequently followed the appearance of elongated and nearly isolated cold domes aloft as shown in figure 2. The observation of these domes as such is not new, and few synoptic meteorologists would dispute that, in the absence of the domes, the ensuing surface events would be quite different. But the laws that determine the course of the surface developments in relation to the thermal and wind structure aloft are far from obvious. We propose to bring out some pertinent facts of the case to be discussed that may provide some suitable hints concerning the routes to be followed in the search for the correct laws.

SURFACE EVENTS

The setting of the period studied, November 12-14, 1951, in the hemispheric picture is at a time when, subsequent to several days of a westerly circulation with weak amplitude (high index or Index Stage NII, Riehl et al 1952), a relative maximum in the westerlies was shifting northward into the higher latitudes. Over the United States two typical high index troughs had passed eastward prior to November 11 when a large low pressure area formed over the western half of the continent and when, with increasing southerly flow, a precipitation area spread rapidly from Texas to the Great Lakes.

On November 12, 1830 GMT (fig. 1a) we observe this large low pressure area east of the Rocky Mountains. The circulation is quite disorganized. There are several weak centers, and the frontal analysis is complex and rather uncertain. Organization of this diffuse pattern begins within 12 hours (fig. 1b) and proceeds rapidly around 1830 GMT on November 13 (fig. 1c). The low pressure center that emerged from Kansas travels slowly toward the Great Lakes on a path with strong counterclockwise curvature. Deepening steadily, it moves toward the NNW late on November 13 and finally becomes nearly stationary south of Lake Superior as a great vortex with central pressure near 976 mb (fig. 1d).

THERMAL STRUCTURE OF THE MIDDLE TROPOSPHERE

A 500 mb, the isotherms over North America initially exhibit the relatively unorganized pattern (fig. 2a), typical of "high index" conditions. Then there travels across the continent a line of maximum spread of the isotherms (minimum temperature gradient), followed by a well marked and nearly isolated elongated cold dome with a great isotherm concentration on its south side (for convenient reference on studies of such concentrations see Palmen 1948, Palmen and Nagler 1948, and Palmen and Newton 1948).

These two features of the isotherm field propagate eastward at a mean rate of about 25 knots. The surface low pressure area lies intermediate between them and initially is closer to the line of minimum temperature gradient. We observe three interesting facts.

(1) The surface low organizes a great distance from the cold dome

(figs. 2b, c) -- as much as 1,000 miles distant. Whatever the role of the dome -- which is pre-existent -- in helping to organize the vortex at the surface, it cannot be simple superposition by means of height falls aloft. Actually, the 12-hour height changes at 300 mb were zero over the area of largest surface pressure falls. A more subtle mechanism is required, capable of downstream transmission over long distances.

(2) The Low deepens not in the region where the Polar front at 500 mb, as seen from the 500-mb isotherm gradients, is strongest, but far downstream in an area where the 500-mb temperature gradient is weak and nearly uniform along an axis drawn through the low center at the time when organization begins (fig. 2b). It is not possible to regard the development as occurring within the zone of strongest baroclinity of the troposphere. Indeed, it has been our observation that only weak and stable frontal waves form when an intense frontal zone at 500 mb extends across North America with nearly uniform strength.

These observations corroborate the work of Ryd 1923, 1927, Pogade 1938, and Scherhag 1948 who point out that deepening preferably occurs in the left hand portion of "delta" regions of the upper wind and thermal field. Sutcliffe 1947 derived an analytical expression suggesting cyclogenesis in regions where there is maximum advection of cyclonic thermal vorticity.

(3) It is possible to track the cold dome over the two-day interval. Figure 2a shows its path. The velocity of the dome is approximately the same as that of the winds in its center. These winds also varied little with height above the mountains so that a vertical time section following the center of the dome (fig. 3a) shows substantially the same air parcels as time progresses. Construction of this section presented some difficulty since the center was never situated precisely over a radiosonde station. Nevertheless, we preferred to draw the section with use of the nearest sounding at each observation time rather than interpolate from analyses. We consider this procedure as most straightforward in this case, since at each sounding period there was at least one alternate station with precisely the same indications.

The section shows a rise of the tropopause, in this area defined as the top of the cold dome, by 30-40 mb until November 13, 1500 GMT when it levels off. In the upper troposphere, the isentropes ascend through the period and the lapse rate, initially moist adiabatic (fig. 3b), steepens to become dry adiabatic. Table 1 shows, in millibars, the amount of ascent of several isentropic surfaces during the 36 hours from November 12, 1500 GMT to November 14, 0300 GMT.

Table 1
Ascent of isentropic surfaces in the cold dome.

θ	Δp (mb)
290	-30
292	-60
294	-90
296	-95
298	-95
300	-65
305	-10

The observed cooling particularly in the lowest part of the stratosphere is too large to be explained by non-conservative processes. It follows that the cold dome is rising through the period as may also be confirmed by noting the increase of area covered by the -30°C. isotherm on successive 500-mb charts. In view of the steep lapse rates, the changes in thickness between isentropic surfaces, which can be computed from table 1, may not be sufficiently accurate to warrant a diagram showing divergence and convergence as a function of height. But the table does suggest that in the troposphere there is a deep layer of gradually converging air, that this air ascends at rates with magnitude of about one centimeter per second and that it is evacuated laterally in a narrow layer under the tropopause.

The foregoing observations do not accord with the viewpoint (Margules 1903) that cold domes must sink as a whole during surface cyclogenesis. As just seen, far from subsiding, the cold dome center actually spreads upward during the period of deepening. We are not suggesting, of course, that sinking of cold domes does not take place in the great majority of cases. But we do obtain the impression that such sinking is not uniquely necessary for cyclonic development (Spar 1950) and that the role of the dome in the cyclogenetic mechanism may be other, at least initially, than the simple sinking usually visualized.

The point has come up here that the ascent of the dome takes place while it crosses the mountainous regions of western North America. A possible "mountain effect" enters into almost all detailed synoptic studies that can be made over the continent. We do not see how this can invalidate the inferences just drawn. The point is that the cyclone deepens while the cold dome ascends, irrespective of the reason for such ascent which is not a topic of investigation in this report.

STRUCTURE OF THE JET STREAM

We shall now investigate the structure of the high-tropospheric wind field as a possible connecting link. Figure 4 shows the 300-mb contours for the period and figure 5 the 300-mb isotach analyses, prepared with aid of the observed winds and with computations from the contour field. As is generally the case, this analysis is least certain in the area of strongest winds. We cannot claim that we know the strength of the wind maximum and the gradients around it with precision. However, we believe that the analysis represents as fair an approximation to true conditions as can be secured with the available observations. Many features of the wind field as analyzed accord closely with previous descriptions. The vorticity distribution near the maximum, for instance, is very similar to that computed by Palmen and Newton 1948. To the left of the maximum the absolute vorticity is between 2f and 3f, where f is the Coriolis parameter. To the right, the anticyclonic shear amounts to -1.5f; but this is offset by a curvature term of +0.5 f, so that the total absolute vorticity is very nearly zero.

On November 12, 1500 GMT, a strong jet maximum is situated near the West Coast of the United States, centered about 600 miles south of the cold dome. Farther downstream the organized jet decomposes into the "fingery" structure common in "delta" zones (Riehl et al 1952). This accords with the open and irregular isotherm pattern at 500 mb (figs. 2a-b) over central North America. We also observe that in this area the 500-mb winds across the isotherms at large angles ranging up to 90°; and that winds and isotherms are nearly parallel only in the zone of great isotherm concentration in the west.

On the subsequent maps, the jet maximum propagates eastward parallel to the cold dome and, in the mean, at the same rate. The leading edge of the jet maximum, initially situated a little west of 102°W. (fig. 5a), reaches the states just south of the Great Lakes on November 13, 1500 GMT (fig. 5c). It is at the time of arrival of this leading edge that the surface cyclone organizes on its left hand margin. We can see this clearly from the surface maps and from the three-hourly surface pressure tendency centers entered with dashed lines in figure 5b, c.

In consequence, we can regard the downstream propagation of the leading edge of the western jet stream as a link connecting the cold dome in the west and the cyclone formation in the Central States. There may be other such links. If so, we have failed to notice them. But the corollary evidence which follows suggests that the proposed connecting link is realistic.

ADVECTION OF VORTICITY

As brought out in the literature of recent years (for reference see Palmen 1948 and Riehl et al 1952), it is likely that the surface pressure falls in areas where advection of more cyclonic absolute vorticity takes place in the upper troposphere. This statement is based on mixed dynamic-empirical reasoning that horizontal mass divergence is occurring in regions where higher absolute vorticity is imported from upstream and that this mass divergence exceeds any compensating convergence in the lower levels.

Although vorticity advection charts are not presented, we can deduce readily from figures 4-5 that the relative vorticity decreases downstream at 300 mb to the left of the jet stream axis over the area where the surface pressure is falling November 12-13, and that therewith higher vorticity is brought into this area by the wind. The curvature of the contours changes from cyclonic to anti-cyclonic as we go downstream at a particular time, and the cyclonic shear weakens as the gradient of the isotachs becomes less (cf. fig. 4c, 5c). The downstream variation of the Coriolis parameter is small and may be neglected. We see then that the advection of absolute vorticity has the requisite sign for pressure fall. If the 300-mb surface may be taken as representative for the upper troposphere, as is generally the case, it follows that the law stated initially holds for the present case.

DISTRIBUTION OF VERTICAL MOTION

Although we have just mentioned that a good relationship appears to exist between the signs of surface pressure changes and high level vorticity advection, the correlation between the magnitude of these quantities is poor, except for short period fluctuations. The surface pressure fall is a small residual between low level mass convergence and high level mass divergence. Mass continuity is provided by vertical motions which under the conditions described must be directed upward over areas of surface pressure fall. Such broad scale ascent should be reflected in the precipitation pattern, except perhaps in the lee of mountain ranges. We observe the heaviest precipitation near the three-hourly surface pressure fall centers and this precipitation, in the mean, lies to the left of the 300-mb jet stream axis. Thus the relative geographic positions of jet stream core, region of surface pressure fall, and region of precipitation correspond to that observed by Starrett 1949 and that demanded previously on the basis of dynamic reasoning (University of Chicago 1947 and Riehl, Norquest, and Sugg 1952).

CROSS-STREAM CIRCULATION

It is one of the assumptions in the derivation of the relation between mass divergence and advection of vorticity, stated earlier, that the flow and vorticity patterns move much more slowly than the wind and that we can neglect local changes compared to advective changes. The validity of this assumption is borne out by figures 5a-c. Although the winds blow at (computed) speeds near 200 knots in the jet stream core, the pattern propagates at little more than 25 knots. The air very rapidly moves through the pattern, and it must suffer extreme deceleration when leaving the area of highest wind. (Earlier case studies, Wobus 1950 and Teweles 1950, have described this phenomenon.) This, following the equations of motion, is accomplished mainly by motion toward higher pressure, i.e., by a clockwise cross-stream circulation looking downstream along the jet core. To the left of the axis this high level cross-stream circulation is likely to be associated with mass divergence, as we have already established for the present case from the vorticity advection pattern. We see the cumulative effects of this circulation in figure 6a, a vertical cross section taken normal to the jet stream axis from Lake Charles, La. (LCH) northward to International Falls, Minn. (INL). The time is 0300 GMT, November 14 -- 12 hours subsequent to the time of figure 5c. The principal wind maximum still is upstream from the section, and the pertinent portion of the isotach pattern has not changed appreciably except for continued gradual downstream displacement of the jet center. Choice of the section shown, rather than an earlier one, was prompted by a very suitable station distribution and availability of data.

On the section, which cuts through the forward edge of the cold dome near Omaha (OMA), we locate the core of the jet at 260 mb, slightly north of Little Rock (LIT), and about 400 miles south of the deepest portion of the cold dome. The section is drawn so that we face upstream. The winds which blow out of the section toward us are decelerating near the jet core as mentioned. According to computations made from the equations of motion, the angle between contours and streamlines in this region must have attained 20° to allow the observed deceleration of the air. From there northward, the rate of deceleration must decrease and eventually become small near the cold dome since, as seen earlier the dome moves with the speed of the winds. Even here, however, some divergence has been taking place as brought out earlier. Computing an approximate value of this divergence from figure 3a and table 1 for the layer between the isentropes 298°A. and 305°A. with the formula

$$\frac{1}{\Delta p}\frac{d\Delta p}{dt} = -\mathrm{div}_2 \mathbf{V}$$

we find that $\mathrm{div}_2 \mathbf{V} \sim 5 \times 10^{-6}\ \mathrm{sec}^{-1}$, a rather small value.

If we combine all the evidence adduced -- the deceleration of air near the jet core, the decrease of this deceleration toward the north, the presence of divergence in the upper troposphere some distance north of the jet core as given by the cold dome computation and the vorticity advection pattern, finally the shower activity under the cold dome -- it becomes plausible to suggest that the configuration of the isentropes in figure 6a indicates in part the accumulated effect of vertical displacements upstream. The sense of the vertical motion and cross-stream circulation pattern (fig. 6b) would be that suggested in an earlier publication (Starrett 1949). We should like to emphasize however, that the arrows of figure 6b are not meant to suggest closed circulation orbits. If air evacuated laterally from the top of the cold dome were to pass to the other side of the jet core with the cross-stream circulation, it would first have to assume the very high vorticities of the zone just north of the jet center, then the very low vorticities to its south. Clearly, this is most unlikely. Besides, we note in the present case that the thick isentropic layers south of the jet lie between 324° and 340°A., those north of the cold dome between 312° and 320°A., whereas the potential temperature of the air evacuated from the cold dome ranges from 298° to 305°A. It is more likely that the part of the jet stream core downstream from a maximum is gradually displaced toward higher contours on any isobaric surface and this indeed is observed in many cases (Sawyer 1950).

VERTICAL VARIATION OF JET STREAM AXIS

In the jet stream publication mentioned initially much emphasis was placed on the marked reversal of temperature gradient across the jet stream axis above the level of strongest wind. It was shown that a band of warm air extends along this axis on the poleward side at 200 mb, and that a narrow band of very cold air parallels its equatorward margin (fig. 7). The axis itself was situated within the zone of strongest 200-mb temperature gradient. Such a position is requisite if the geostrophic component of the wind is to decrease with height. Farther poleward and farther equatorward the temperature gradient again reversed, thereby proving that the 200-mb temperature field found in the jet stream zone could not have advective origin but that vertical motions as shown in figure 6b had to account for its existence. Palmen and Nagler (1948) have reached the same conclusions.

Figure 8 shows the 200-mb isotherms at the time of strongest deepening, November 13, 1500 GMT. In several respects, this chart verifies the description of the 200-mb temperature field given earlier. Relatively warm air is in evidence everywhere at the tropical margin of the chart, and from there the temperature decreases toward the jet axis. We find very warm air on the poleward side of the jet center (cf. fig. 5c) and very cold air on its equatorward side. Indeed there is a suggestion, particularly north of the jet maximum, that the centers of greatest 200-mb temperature anomaly are closely associated with the area of highest wind. In view of the discussion of figures 6a, b this suggestion appears quite reasonable.

We also note some interesting differences between figures 7 and 8. Along the jet axis the 200-mb temperature varies much more in the November 1951 than in the January 1947 case. In fact, the temperature gradient reverses along the axis downstream from the zone of strongest winds. This is also brought out clearly in graphical form in figure 9. It follows that the level of strongest wind must rise downstream above the region of surface deepening. Southeast of Lake Michigan it reaches the 200-mb surface. Over the eastern Great Lakes, it must actually lie above 200 mb.

As the intense 200-mb temperature gradient north of the jet axis over the north central Plains is directed nearly parallel to the contours (not reproduced but similar to those of fig. 4c), and as the air even at some distance from the axis moves at a rate several times greater than the speed of the system, we can safely conclude that ascending motion is taking place through the 200-mb surface. Here, the schematic vertical motion picture of figure 6b cannot hold entirely. The ascent in the rain area of figure 5c appears to extend to very high levels, a conclusion similar to that reached by Fleagle 1947 in his studies of upper troughs and ridges. South of the jet axis, th temperature gradients are much weaker. We can only state that the region w... subsidence, compensating for the upward mass transport through the 200-mb surface takes place, is not completely delineated by the charts presented here.

We can verify some of the conclusions drawn from figures 6-9 by constructing isotach cross-sections along the axis of the jet stream, a representation which we have not yet encountered in the literature. This is done in figure 10a-c. The sections shown in these figures follow the axis of the jet stream, and tick marks refer to points along the jet at 300 mb as given in figure 5a-c. It is to be noted that the sections do not portray conditions along the vertical but that they pass through the axes of a strongest wind at all levels. For the construction we first analyzed isotach charts at 700, 500, 300, and 200 mb, and in part also at 250 mb. We then drew lines connecting the points of highest speed on each surface and projected these lines to coincide with the 300-mb axis. Finally we plotted the wind values so oriented on cross-section paper and drew isotachs. The distance projected did not exceed 200 miles and from 400-500 mb upward the axes nearly coincided, i.e. the jet stream was almost vertical as is commonly the case.

The first impression that one gets from figures 10a-c is that they resemble sections taken normal to the upper westerlies. Variations of wind structure along the current have the same order of magnitude in the present case as variations normal to the current, although the regions of intense baroclinity seen in figure 6a of course are not present. Figure 10a-c verifies the upward displacement of the jet axis downstream from the wind center (marked J) as inferred above. It will be an interesting problem for the future to draw corresponding cross-sections of isentropes and attempt to determine the actual rate of deceleration of the air particles. This deceleration would be less than indicated by constant pressure charts if the air ascends substantially in the region where the isotachs trend upward.

Figure 10a-c reveals another curious feature. We have already noted on figure 5a-c that a secondary jet maximum formed at 300 mb downstream from the main maximum on November 13 and that it was this maximum which was most directly connected with the cyclone formation. The longitudinal sections indicate how this center began to form early on November 13 (fig. 10b) and then became a separat entity later on that day (fig. 10c). Qualitatively, one gets the impression that an "impulse" becomes detached from the main maximum and propagates forward at a more rapid rate than the parent center. If future synoptic studies should establish a general connection between such "impulses" and deepening, a new approach to the problem of the dynamics of cyclogenesis would indeed be provided (cf. Riehl and Jenista 1952). At this time, further speculation on this topic is not warranted.

CONCLUSION

In an attempt to enlarge on previous description of the relations between cyclogenesis and the structure of the upper atmosphere, the following has been noted.

(1) The developing surface pressure fall in the case studied could not be explained by simple superposition of an upper pressure fall.

(2) The cyclone did not develop on an intense frontal zone but far downstream from this zone.

(3) While the deepening progressed, the center of the cold dome upstream ascended. Therefore simple sinking of the cold air could not account for the deepening.

(4) An intense jet stream maximum located to the right of the cold dome, looking downstream, elongated rapidly from the central Rocky Mountain area toward the Great Lakes by means of sending forward an "impulse" which could be observed forming a new wind maximum on November 13.

(5) Downstream from the main jet center the axis of strongest wind ascended to reach levels above 200 mb. over the zone of deepening. The ascent of the axis is coupled with upward motion of the high tropospheric air to its left.

The scheme of cross-stream circulation in the jet stream zone proposed earlier (University of Chicago 1947) is supported by the data in the vicinity of the main maximum. In the region where the jet axis ascends, some modification is required. Nevertheless, the major part of the precipitation area is observed to lie to the left of the axis, in accordance with the earlier findings.

(6) At the arrival of the forward edge of the "impulse" above a pre-existing weak surface Low, the latter began to deepen strongly. Dynamically, this deepening could be related to the observed pattern of vorticity advection aloft.

ACKNOWLEDGMENT

The authors are indebted to Mr. Gordon E. Dunn, Meteorologist-in-charge of the Chicago District Forecast Center of the United States Weather Bureau who generously placed the facilities of his office at the disposal of the joint Weather Bureau-University of Chicago course in forecasting during the autumn of 1951. They also wish to thank Dr. Harry Wexler and Mr. Roger A. Allen who facilitated completion of this paper at the Central Office of the United States Weather Bureau.

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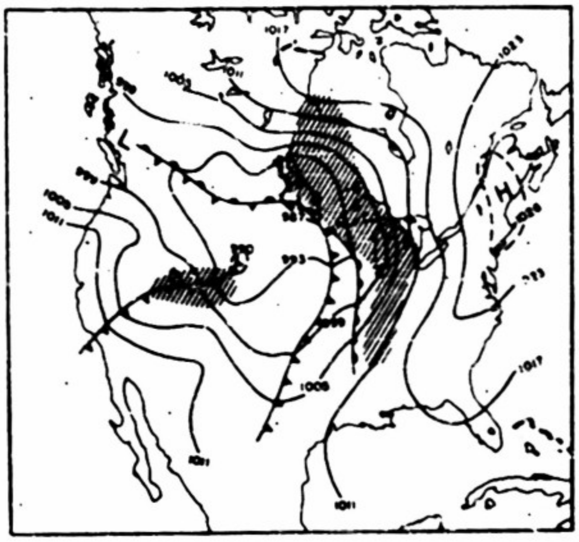

Fig. 1a. Surface isobars (mb) and fronts, Nov. 12, 1951, 1830 GMT. Area of steady precipitation is shaded.

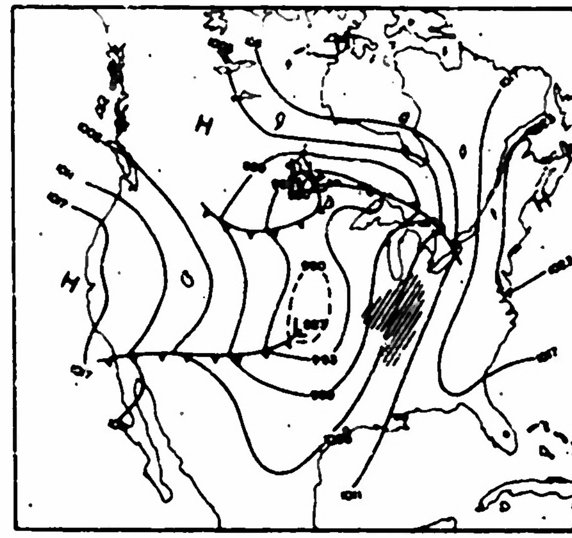

Fig. 1b. Surface isobars and fronts, Nov. 13, 1951, 0630 GMT.

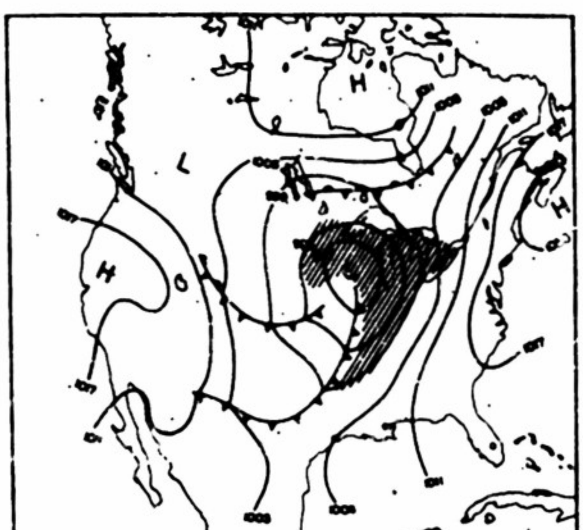

Fig. 1c. Surface isobars and fronts, Nov. 13, 1951, 1830 GMT.

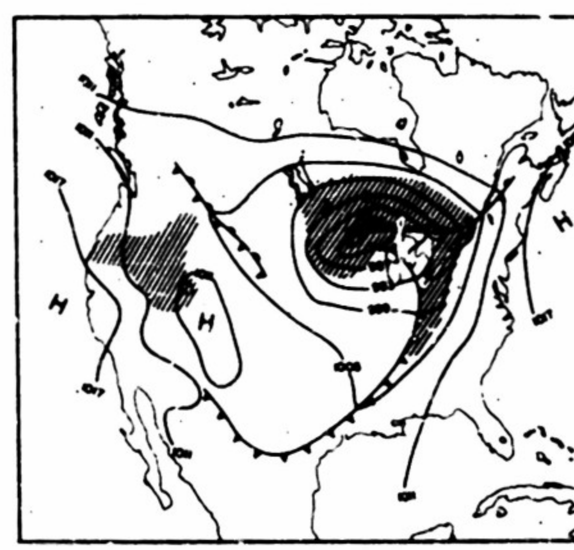

Fig. 1d. Surface isobars and fronts, Nov. 14, 1951, 0630 GMT.

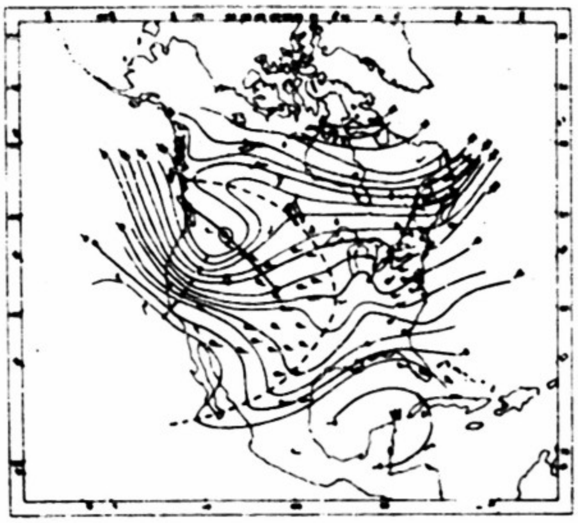

Fig. 2a. Winds and isotherms (°C) at 500 mb, Nov. 12, 1951, 1500 GMT. Progression of cold dome in 12-hourly steps, Nov. 12, 0300 GMT to Nov. 14, 0300 GMT, is marked by heavy line. Axes of minimum and maximum temperature gradient are marked by dashed lines. "W" denotes warm center and "C" denotes cold center. On wind vectors a long barb denotes 10 knots, a short barb 5 knots, and a heavy triangular barb 50 knots.

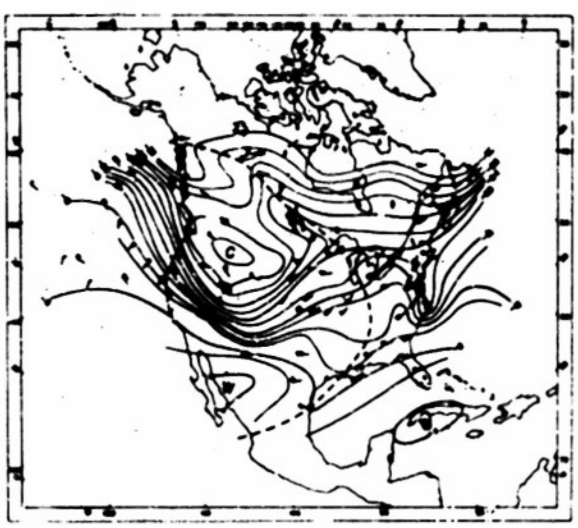

Fig. 2b. Winds and isotherms at 500 mb, Nov. 13, 1951, 0300 GMT. Heavy dot marks position of surface low pressure center.

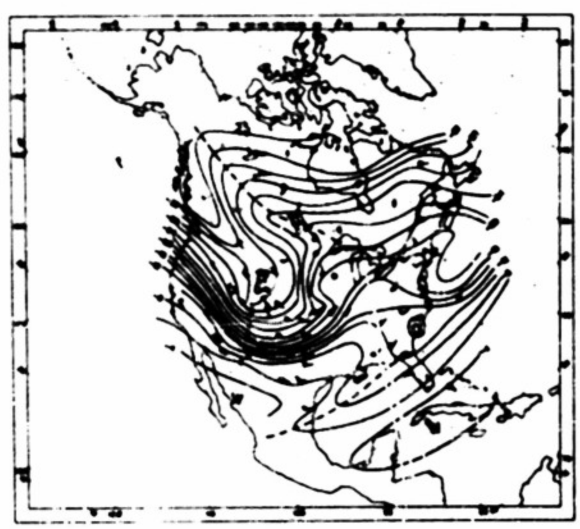

Fig. 2c. Winds and isotherms at 500 mb, Nov. 13, 1951, 1500 GMT. Heavy dot marks position of surface low pressure center.

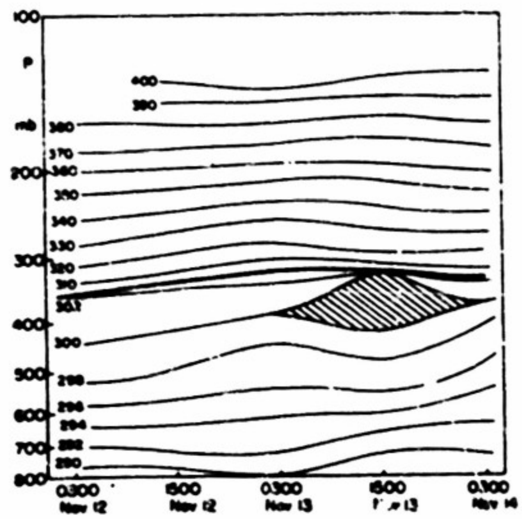

Fig. 3a. Vertical cross-section of potential temperature (°A) following cold dome on track given in figure 2a. Heavy line denotes tropopause and shaded area indicates adiabatic layer. Note that isentropes in the stratosphere are drawn for intervals of 10°A.

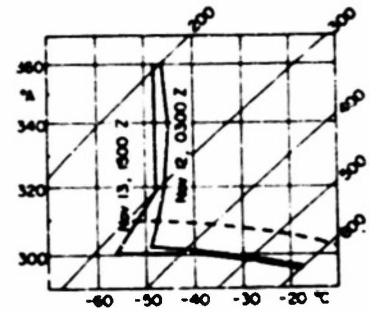

Fig. 3b. Tephigram showing soundings near center of cold dome. Horizontal lines are isentropes, vertical lines isotherms, sloping lines isobars (mb) and dashed line shows the moist adiabatic lapse rate.

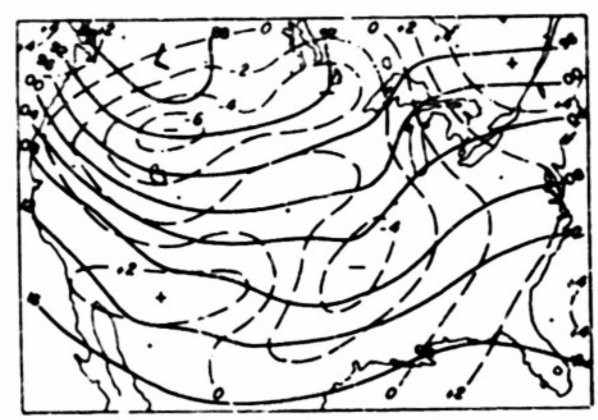

Fig. 4a. 300-mb contours (100's feet, first digit omitted) and 12-hour height changes, (100's feet), Nov. 12, 1951, 1500 GMT.

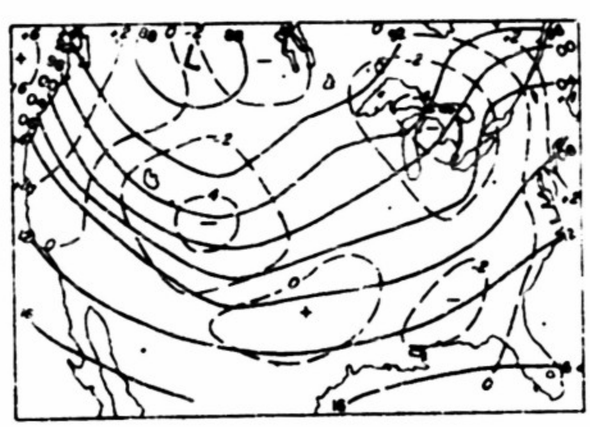

Fig. 4b. 300-mb contours and 12-hour height changes, Nov. 13, 1951, 0300 GMT.

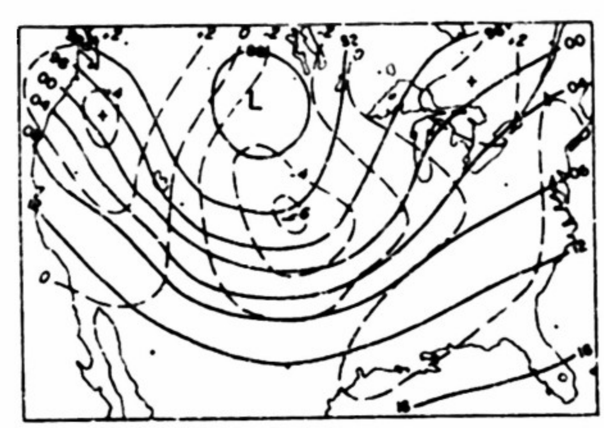

Fig. 4c. 300-mb contours and 12-hour height changes, Nov. 13, 1951, 1500 GMT.

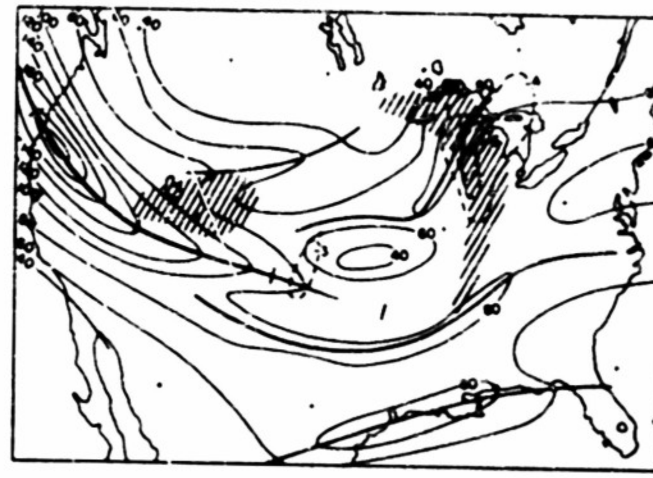

Fig. 5a. Isotachs at 300 mb (knots), Nov. 12, 1951, 1500 GMT. Heavy lines mark jet stream axes. Tick marks along axis of principal current serve to identify the horizontal axis of figure 10a. Precipitation areas are shaded. Dashed lines give 3-hour surface isallobars (mb).

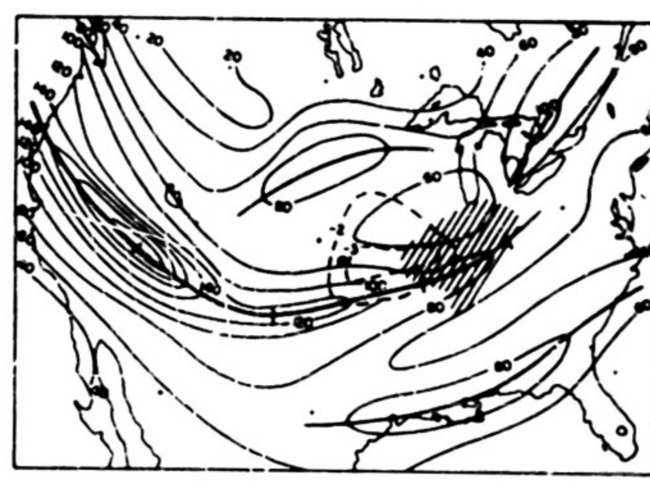

Fig. 5b. Isotachs at 300 mb, Nov. 13, 1951, 0300 GMT. Tick marks correspond to those along horizontal axis of figure 10b. Heavy dot marks position of surface low pressure center.

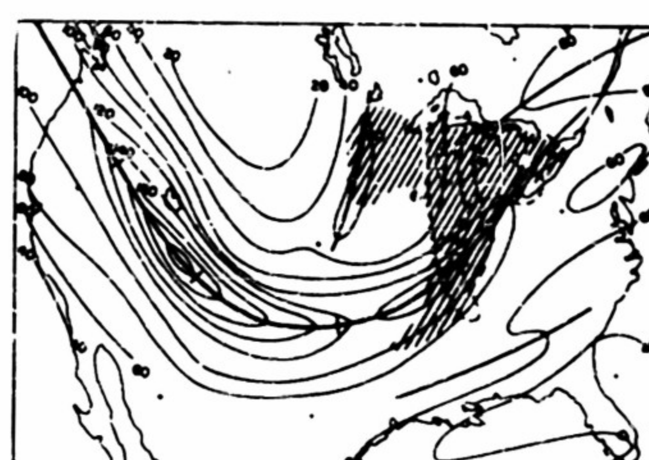

Fig. 5c. 300-mb isotachs, Nov. 13, 1951, 1500 GMT. Tick marks correspond to those along the horizontal axis of figure 10c. Heavy dot marks position of surface low pressure center.

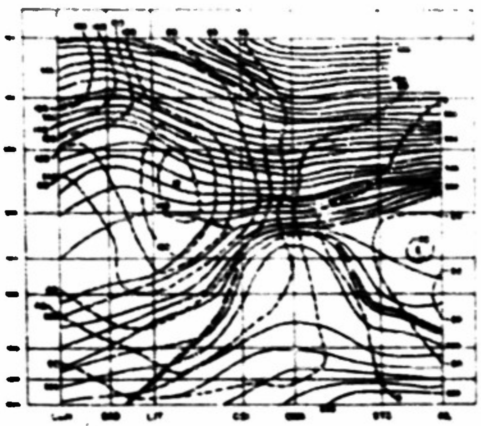

Fig. 6a. Vertical cross-section of isentropes (solid lines, °A) and isotachs (dashed lines, knots), from Little Rock, Ark. to International Falls, Minn., Nov. 14, 1951, 0300 GMT. Heavy lines denote fronts and tropopauses. "W" stands for center of westerly current, and "E" for center of easterly current.

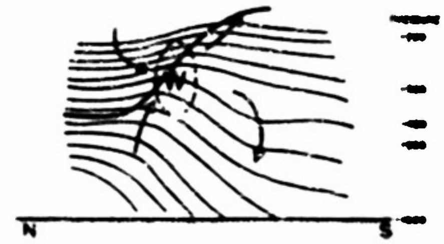

Fig. 6b. "Probable meridional and vertical displacements associated with intensification of zonal wind maximum. Arrows indicate direction of displacement of isentropes. Isentropes given by thin lines, tropopause by heavy line." (From University of Chicago 1947.)

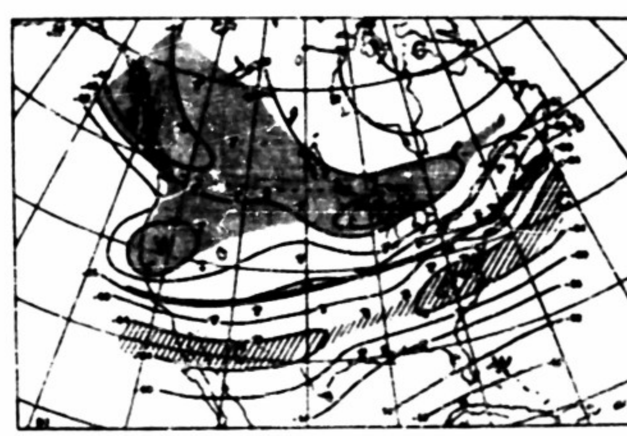

Fig. 7. "Isotherms (°C) at 200 mb, Jan. 28, 1947, 0300 GMT. Jet stream center at 300 mb marked by heavy line." (From Riehl 1948).

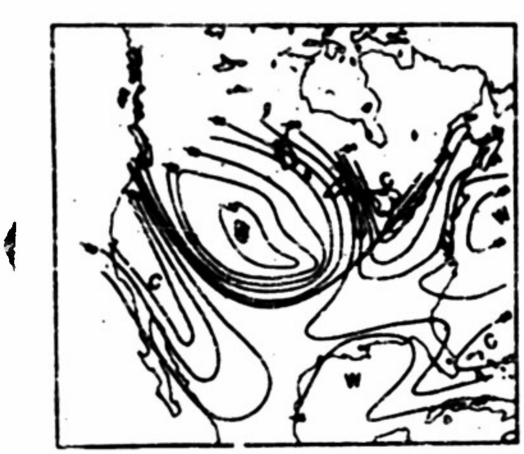

Fig. 8. Isotherms (°C) at 200 mb, Nov. 13, 1951, 1500 GMT. Jet stream center at 300 mb marked by heavy line. Tick marks along jet axis serve to identify the horizontal axis of figure 9.

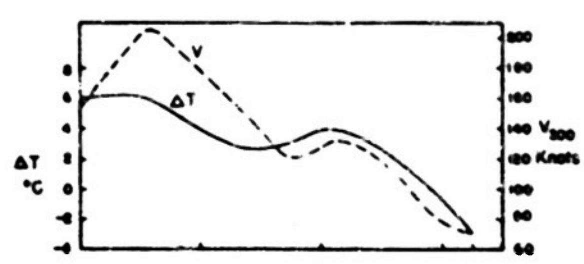

Fig. 9. Wind speed at 300 mb (knots), and temperature gradient (°C) at 200 mb taken over distance of 250 km normal to jet axis (heavy line) of figure 8. Marks at bottom correspond to tick marks along jet axis of figure 8.

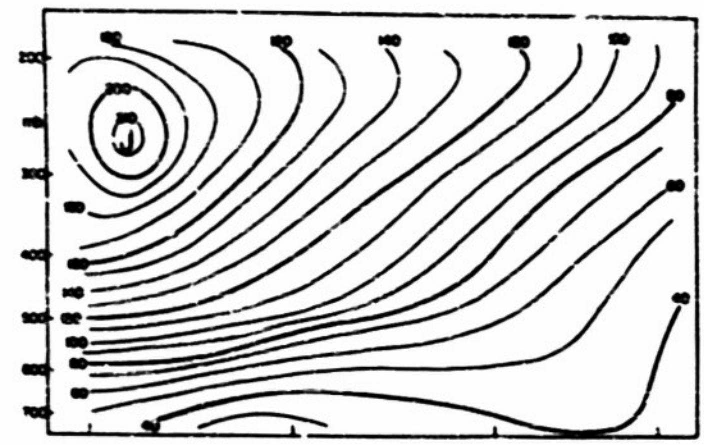

Fig. 10a. Cross-section of isotachs (knots) along axis following jet stream core, Nov. 12, 1951, 1500 GMT. Marks at bottom correspond to tick marks of figure 5a. For details of construction of figure see text. "J" denotes jet stream center.

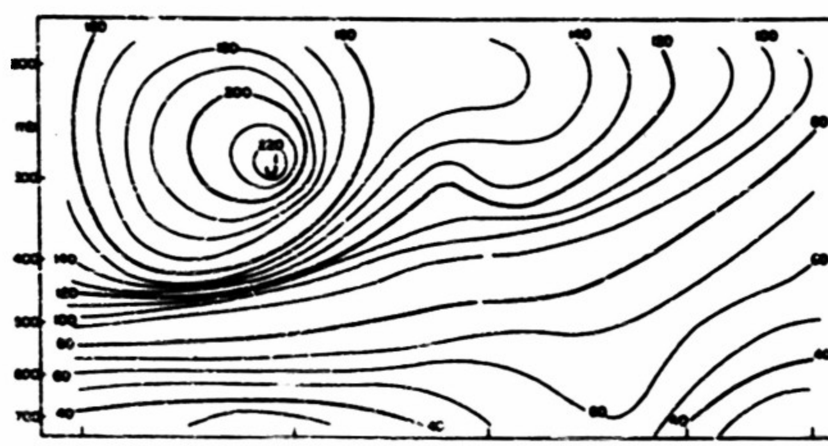

Fig. 10b. Cross-section of isotachs, Nov. 13, 1951, 0300 GMT. Marks at bottom correspond to tick marks of figure 5b.

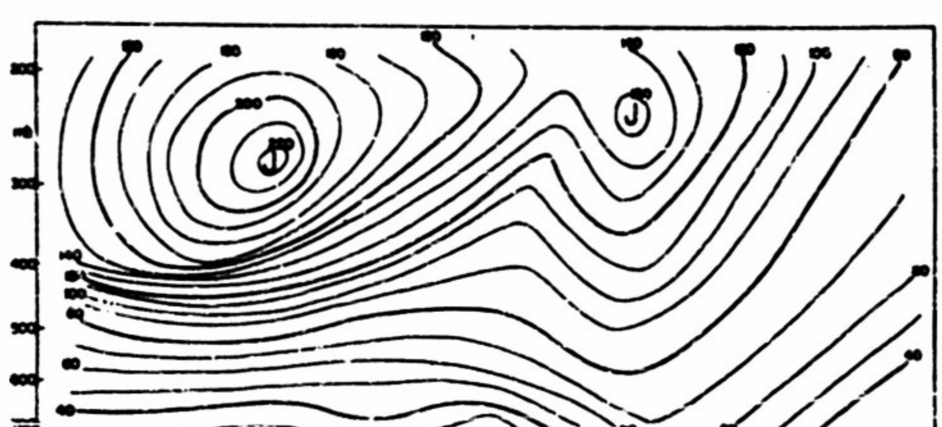

Fig. 10c. Cross-section of isotachs, Nov. 13, 1951, 1500 GMT. Marks at bottom correspond to tick marks of figure 5c.