An Aerological Study of Large-Scale Atmospheric Disturbances

By ALF NYBERG, Swedish Meteorological and Hydrological Institute, Stockholm

(Manuscript received 29 Sept., 1948.)

Abstract.

Pressure change areas in the 500 mb level move with 60 % of the speed of the wind in that level. Advection of cold (warm) air at upper isobaric levels is followed by geopotential falls (rises) at the same levels. The prognostic usefulness of this rule is much higher in the 500 mb level than in the 700 mb level. — There are strong indications of dynamic interaction between flow patterns in distant regions on the hemisphere. A deepening of one upper trough will in special cases lead to a second deepening downstream. — In a deepening trough the cold tropospheric air is shown to be sinking at the same time as the jet stream increases and the front becomes sharper. Models of the distribution of vertical motion and divergence in the frontal zones are given.

Introduction

The study of the formation, movement, deepening and filling of atmospheric disturbances is the principal research problem for a synoptic meteorologist. While considerable progress has been made concerning the movement of disturbances, rather little is known about the formation or intensification of pressure systems. Theories have been developed for the stability of waves as functions of horizontal and vertical wind shear, or of the slope and intensity of fronts. Up to the present time little or nothing of these theories has been directly applicable in prognostic work.

In a study [7] of a cyclogenesis over Scotland and the North Sea April 1939 I found that above 2 km no front was present at the beginning of the deepening though one was formed during the cyclogenesis and I stated that the development evidently must be due to some other factor than the instability of a wave front. At that time, I was not able to give any explanation of the process. After the war I had the opportunity to join a group of meteorologists at the University of Chicago for the year 1946—1947. This paper is a result of research work carried out during that time. I tried to apply and test some of the theories developed by Rossby in forecasting technique. Historical Weather Maps [4] for 500 mb, for the months of October and November 1945, prepared by the USA Headquarter Army Air Forces Weather Service were used. Maps showing the interdiurnal changes in geopotential were drawn, and the behavior of areas of height (geopotential) increase and decrease (pressure rise and pressure fall) was investigated. The actual daily weather maps for various levels were also studied.

Movement of pressure change areas and interdiurnal variation of pressure aloft

It was found that individual centers of 500 mb level change areas can be followed for several days and in some cases for more than a week, and that the 24 hours change in intensity of such a change center was in general much smaller than the intensity of the change area itself.

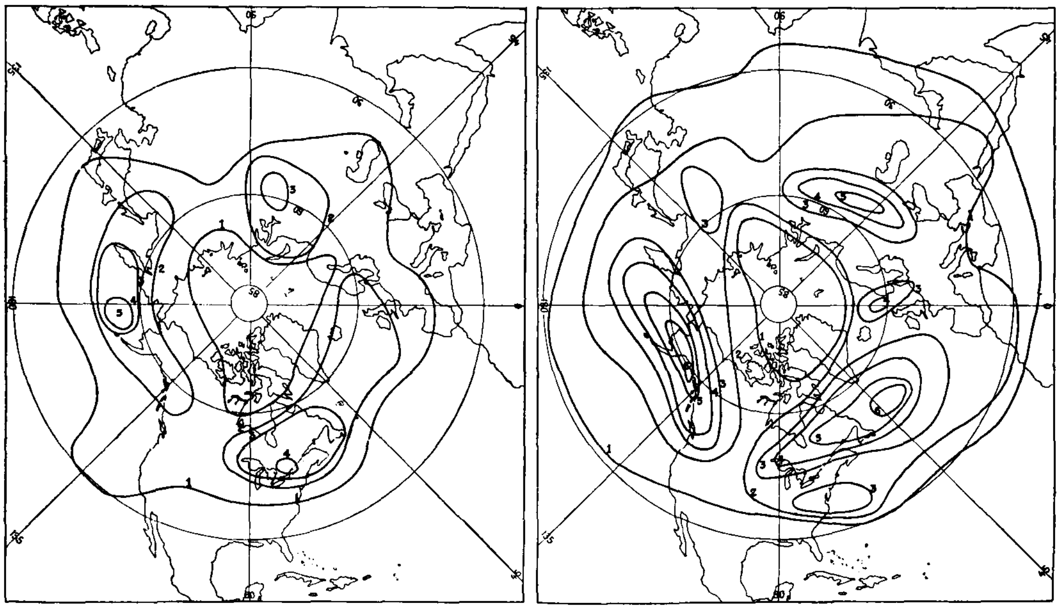
The movement of the change areas is as a rule in the direction of the wind and the speed is in general about 60 % of the wind speed. This rule is of high value in forecasting for

24 and 48 hours and is so obvious that it must be found by anyone working with upper change areas. It will later be shown that the rule can be given a physical interpretation.

If the daily position and intensity of the isallobaric centres at the 500 mb level in October 1945 are plotted on one map a concentration of dots arises as shown in fig. 1 a. There is one marked center south of Bering Strait, another one almost as strong over Labrador and a third and less intense one over northern Russia. The centers south of Bering Strait and over Labrador are close to the areas where cold continental air and warm maritime air meet in lower levels and these two centers could be thought of as a result of extremely intense mixing of air masses. The third center over Russia can not as easily be explained as a result of such an interaction of air masses; it seems to be of a more dynamic origin. It is of interest to note that the difference in longitude between the three centers is approximately 120°. A similar change area chart (fig. 1 b) for November 1945 also shows a rather symmetrical distribution. The variability is greater and there are two additional centers, however; the picture of three main centers is the same as in October and there is only slight westward displacement of the main centers.

ROSSBY [11] made use of the vorticity theorem to study the stationary quasi-horizontal waves that would be set up in a zonal current under the influence of a permanent localized geographic perturbation and showed that 3—5 main troughs would be established around the earth. If two of these troughs are determind by geographical factors a third one is likely to form in such a way as to give a symmetrical picture of troughs and variability.

Dr T. BERGERON and Dr E. BIEHL have brought my attention to the fact that indications of a similar maximum of interdiurnal variability and in roughly the same position has been found on surface maps by VON FICKER, in his studies of Russian air pressure.



a) October 1947 Fig. 1 b) November 1947

Geographical distribution of interdiurnal pressure change a) October 1947, b) November 1947. The lines on the figures represent values of equal mean interdiunal pressure change. They were obtained in the following way. For the whole month each center of the daily pressure (geopotential) change areas was plotted with its value ($\Delta\Phi$ in gpft) on a map. $\Sigma\Delta\Phi$ of all maximal changes in a given area was then plotted and finally the isopleths on the figure drawn. Later the exact interdiurnal variability i. e. changes from day to day on each 5th latitude and longitude intersection has been computed for October. The result in all respects conform the results from fig. 1 a.

Advection and pressure changes

The relationship between temperature advection and pressure change in upper air levels is clear-cut. Statistics by HESS and WAGNER [5] give a correlation coefficient between actual pressure and temperature changes in the 700 mb-level amounting to roughly 0.5. NAMIAS [6] found a correlation coefficient between pressure and temperature varying between 0.3 and 0.8 at the 700 mb level for different places in U. S. and for different seasons. Working with the 500 mb-level I was impressed by the high correlation between temperature change and height change at that level. The correlation between forecast temperature change and actual pressure change over the U.S. in January 1947 was studied. Trajectories were computed from the actual chart assuming steady winds (gradient or measured winds) for 12 hours; then the pressure systems were displaced using extrapolation methods and three hourly pressure changes at the surface, and new trajectories for the following 12 hours were computed from the modified chart. The temperature at the starting point of the trajectory the first day was taken as the forecast temperature at the end point on the second day. This method of computing trajectories and taking the temperature values from the computations is surely less objective than HESS's and WAGNER's method. However, this is a method of the kind that must be used in practical synoptic work. The results are also in accordance with previous results. The correlation coefficient between forecast temperature changes and actual pressure changes for 24 hours at the 700 mb-level amounts to 0.47 while at the 500 mb-level the corresponding value is 0.71. This increase with the height of the value of the correlation coefficient is significant, and it is in accordance with the larger correlation coefficient up to 0.98, between mean temperature in the troposphere and the pressure at 9 km found by DINES [2, 3] and by SCHEDLER [13]. A regression equation computed from my data shows that a temperature increase of 1° C corresponds to an increase in the height of the 500 mb-level by 30 m. Advection of such an intensity that the temperature of the air column is raised by 1° C would lead to an increase of the height of the contours at 500 mb amounting to 20 m. The higher value of 30 m obtained from the regression equation indicates that *the temperature rise in 500 mb also must be statistically combined with a pressure rise at the surface*. The computation of temperature changes in 500 mb is an important aid in constructing upper air prognostic charts; the usefulness of such prognostic charts follows from the fact that the surface change areas follow the upper flow. This method of computing upper air charts is of a rather small usefulness at 700 mb or lower levels, as shown by HESS and WAGNER. However, the usefulness of the method increases with height and is quite considerable at the 500 mb-level. The best application of the correlation will probably be obtained at some higher level, say 400 mb. Advection of warmer air aloft is followed by pressure rise, advection of colder air by pressure fall. The movement of pressure rise and pressure fall areas in the upper levels thus is partly a result of advection.

If there is cold air advection with pressure fall on the left hand of an air current and warm air advection with pressure rise on the right hand of the current, frontogenesis, or confluence, as termed by NAMIAS, is taking place. It means that the speed of the air current increases and that the transversal pressure and temperature gradients aloft increase.

The effect of warm and cold air advection can be expressed in terms of vorticity. The warm air, coming directly from a southern area and keeping its low absolute cyclonic vorticity undergoes anticyclogenesis as it moves towards north, while the cold air moving southwards tends to keep its high absolute cyclonic vorticity by setting up an increased cyclonic circulation relative to the earth. The study of advection, of displacement of pressure change areas and of trajectories give similar results if sufficient data are available. In cases where the circulation is well known the computation of trajectories may give a better result than the rule of advective pressure changes at least for longer range of time. However, in many cases the rule of advective pressure change is more easily applied especially when the analysed area is small and when the curvature of the current, and hence the value of the vorticity, are uncertain. When warm air advection takes place from the north or cold air advection from the south the advective rule still holds and in that case it is almost impossible to apply the trajectory method with success in daily routine work.

Deepening troughs and vertical motion of the cold air in the troughs

In ROSSBY's theory [11] the vertical component of absolute vorticity in general is supposed to be constant and vertical movements are neglected. In a study of ocean currents ROSSBY [10] has shown the mutual adjustment of pressure and velocity distributions. It has been suggested [15] that, in analogy hereto, the cold air on the left hand side of an atmospheric current rises to adjust the pressure distribution to the motion, as the current enters a new area. CRESSMAN [1] states that in one case he has found a considerable lifting of the cold air. This process of lifting is partly based on the fact that the coldest air was found south of a relatively warm ridge. But this fact is normally a result of horizontal advection which varies with latitude and CRESSMAN's reasoning is not convincing. I have studied the vertical motion synoptically during the winter 1946—1947 for several cases. *No upward motion of the cold air has been found at any place.* The deepening in the upper levels is then mainly a result of horizontal advectional effects. If upward motion does exist it must take place only on the left hand side of the cold current and the upward motion must be

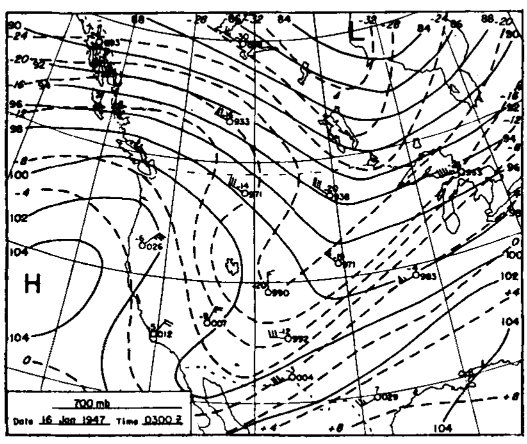


Fig. 2. 700 mb contour chart. North America. Full lines, contours in hundreds of geopotential feet. Dashed lines, isothermes (C). Jan. 16, 0300 G. M. T. 1947.

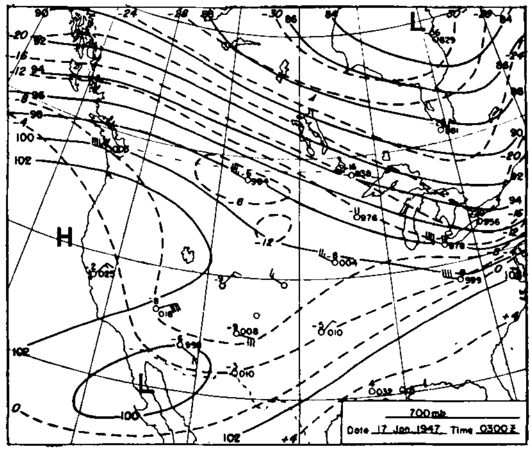


Fig. 3. 700 mb contour chart, North America. Full lines, contours in hundreds of geopotential feet. Dashed lines, isotherms (C). Jan. 17, 0300 G. M. T. 1947.

rather insignificant. In general the movement of isotherms in a constant pressure surface is much slower than the component of the geostrophic wind at right angle to them, a rule which holds both for warm and cold air advection. In special cases the effect of sinking of cold air overcompensates the effect of the cold air advection and the isotherms move against the current. This is particularly the case when an inversion is present close to the isobaric level in question. On the map for 700 mb 16 Jan. 0300 G. M. T. (fig. 2) there is a bend of the —20° C isotherm deep into southwestern U. S. A., 24 hours later the —20° C isotherm in that level and area has completely disappeared (fig. 3). The soundings from Grand Junction (fig. 4) show a strong inversion which is at first, 16 Jan. 0300 G.M.T., slightly above 700 mb and then sinks below that level. Although the actual temperature increase is about 10° C, the sinking motion does not amount to more than about 1 000 m in 36 hours. This sinking occurs on the right hand side of the current in the rear part of a trough and may be regarded as quite normal.

When an upper trough deepens the warm air on its eastern side is thrown northward and the pressure to the right of that current increases, leading also to a pressure rise area on a 24 hours change chart. Frequently a ridge is thus formed. These are quite normal features and the rule can be used in prognostic work with good success. The pressure rise

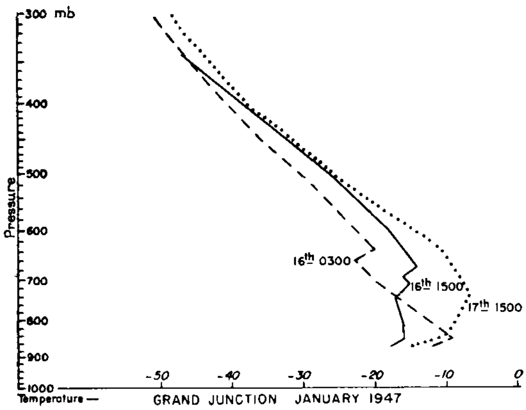


Fig. 4. Soundings from Grand Junction in a T-log p diagram. Temperatures are given in degrees C.

again may lead to an advection of cold air on the eastern side of the ridge so that the first deepening indirectly leads to a second deepening down stream. Cases of that kind may be found, (see for instance CRESSMAN [1]) and in general some prognostic use can be made of this rule although it is not very useful unless it can be combined with some other criteria.

Through the strong subsidence the thermal contrast at the front between the air masses disappears. The subsidence inversion (fig. 4) north of the secluded center (see fig. 3) is not the original front as the air above the inversion originally was much colder than the warm air at corresponding levels further to the west; rather it is believed to be a subsidence inversion formed in the cold air. However, the initially cold air heated by subsidence soon becomes much warmer than the air further to the north which has not subsided as much. We find a broad transitional zone without temperature fronts (see fig. 3). North of that zone a new front is forming concurrently with the development of a new disturbance. Through the subsidence and a simultaneous advection of warm air in the north a core of cold air is cut off in the southern part of U. S. A. forming a cyclonic vortex. A similar cutting off over southern Russia in April 1939 was described by the author in the paper quoted above [7] and this seclusion process which is very frequent over southwestern U. S. has been studied especially by PALMÉN [8], who gives more extreme cases of this cutting off. Other places with frequent seclusions are the Mediterranean area and the Atlantic west of Portugal.

A model of vertical motion and divergence in frontal zones

The two figures 5 and 6 are the 500 mb charts for Jan. 20 and 21, 1947, 0 300 GMT. The mean vertical motion during the time interval between these two charts can be

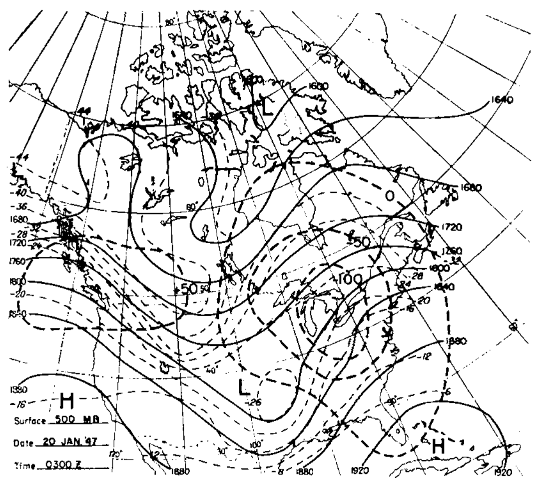


Fig. 5. 500 mb contour chart, North America. Full lines, contours in tens of geopotential feet. Thin dashed lines, isothermes (C). Heavy dashed lines show change in geopotential during following 24 hours.

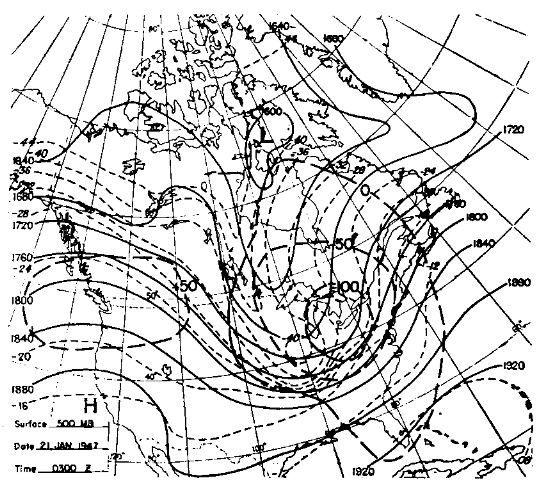


Fig. 6. 500 mb contour chart, North America. Full lines, contours in tens of geopotential feet. Thin dashed lines, isothermes (C). Heavy dashed lines show change in geopotential during last 24 hours.

computed approximately through the construction of trajectories for the geostrophic wind. We find that over the Middle West the isotherms in the 500 mb level have a forward motion below 75 % of the geostrophic wind speed; from the vertical lapse rate we can compute that the mean vertical motion was about 1 500 m in 24 hours. This downward motion in regions of cold air advection (primarily at the cold front) is, as mentioned previously, a normal feature. The corresponding upward motion at the warm front, or in the frontal zone, where warm air advection takes place, is also well established through synoptical experience. Now, if the vertical motion is most pronounced at the point where the advection is strongest we arrive at the distribution of vertical motion indicated in fig. 7.

The divergence described above as concurrent with vertical motion at frontal zones is superimposed on the divergence concurrent with variations of the horizontal wind speed, the path curvature and the latitudinal variation of the Coriolis parameter. These latter effects are not considered here.

It is well-known that there is an almost complete balance between vertical divergence and horizontal convergence, or vice versa, so that in the equation of continuity,

$$\frac{1}{\varrho}\frac{d\varrho}{dt} = \left(\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y}\right) + \frac{\partial v_z}{\partial z}$$

the left hand side is negligible in comparison with either of the two terms on the right hand side. It follows that horizontal convergence,

$$\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} < 0,$$

concurs with

$$\frac{\partial v_z}{\partial z} > 0.$$

Below the cold front the vertical motion is directed downward but increasing in intensity upward; this implies vertical convergence and hence horizontal divergence. Above the cold front we have vertical divergence and thus horizontal convergence. Below the warm-front the upward motion increases with height and there is horizontal convergence, above the front there is decreasing upward motion and horizontal divergence. Through the horizontal convergence over the cold front and the divergence over the warm front circulations are set up aloft in accordance with the figure 8. The air above the cold front zone and in the upper part of that zone acquires increased cyclonic vorticity, while the air above the warm front zone and in the upper part of that zone acquires an increased anticyclonic vorticity. Below the frontal zones and in the lower parts of these zones the change of vorticity is in the opposite direction. Thus the cold air receives an additional component to the South, the warm air a component to the North, which tend to keep the front sharp although the vertical movements act in a frontolyzing way. At the warm front the warmest air is brought northward, the colder air southward and even the front may be sharpened (at the warm front the setting free of latent heat of condensation may be of importance as a frontogenetic factor).

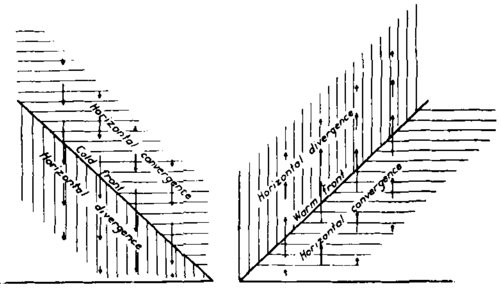


Fig. 7. Normal distribution of large scale vertical velocities and horizontal convergence and divergence at sloping frontal zones. The two parts of the figure i. e. the cold frontal zone and the warm frontal zone should be regarded as independent of one another.

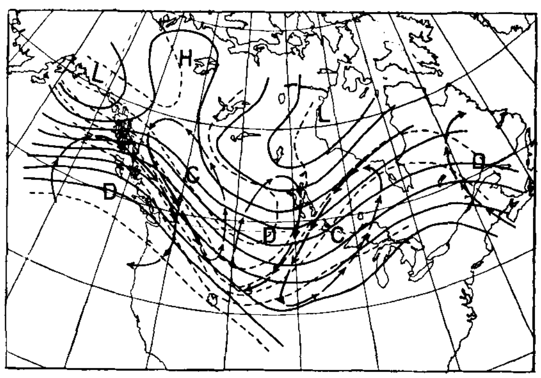


Fig. 8. Typical distribution of contours (full lines) and isothermes (dashed lines) at 500 mb in a deepening trough. Heavy dashed lines with arrows show horizontal circulation as result of vertical motion and convergence and divergence.

4—835339

Behaviour of the jet stream and of the frontal surfaces in a deepening trough

The area covered by aerological observations is in general too small to permit a study of the origin of variations of the subtropical high pressure cells. We know, however, that the variations are frequent and that a westward movement or increase of one of the cells will be associated with an increased southwesterly current with warm air advection on the northwestern side of that cell.

Through the advection of warm air from the west and southwest and of cold air from the northwest (i.e. confluence) into the central parts of the ridge following a trough, the westerly current aloft is intensified and a jet stream is formed. The jet stream was found in mean charts over North America by Willet [16]. On daily charts it was studied in Chicago primarily by Palmén [8], who has illustrated its structure in several papers. It consists of a zone with a high maximum velocity, decreasing upward and downward as well as northward and southward. From theoretical considerations Rossby [12] found the wind maximum to be about 40 m/sec at 50° N and about 90 m/sec at 35° N. This theory is developed from the assumption that the jet is the result of a lateral mixing process. On the north side of the jet the horizontal gradient of the vertical component of the absolute vorticity vanishes, on the south side it is constant. Some vertical cross sections have indicated the presence of inertia instability on the south side of the jet, and several writers, for instance Solberg [14] and Riehl [9], have proposed this instability as the cause of frontal wave development.

If the inertia instability south of the jet is of great importance for the formation of cyclones, the appearance of the jet must be the first sign of the coming development. The question then arises, how is the jet created? We shall consider a synoptic case, which is a rather typical one (Jan. 20 and 21, 1947, fig. 5 and 6). On Jan. 20 a high over the eastern Pacific brings warm air up over the North American Pacific Coast, the pressure rises and the speed of the upper current increases in that region. Through this rise cold air over Western Canada starts moving towards southeast. Concurrently a frontal disturbance

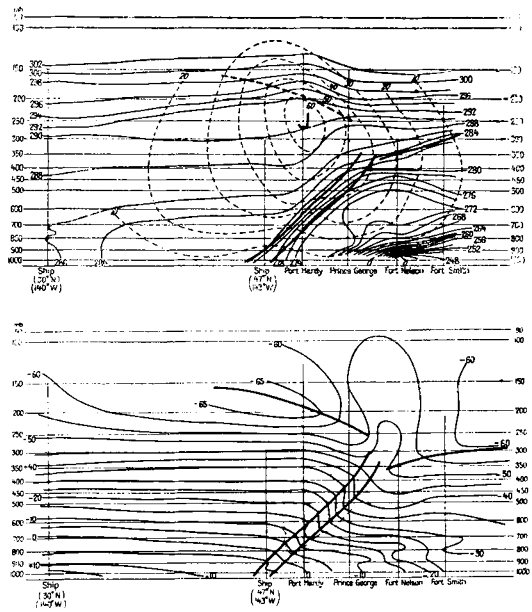


Fig. 9. Cross-sections of temperature, below, and wet bulb potential temperature, above. Dashed lines show distribution of computed gradient wind velocity. The cross-sections are taken normal to the jet stream. Heavy lines show fronts and tropopauses. Jan. 20, 0300 G. M. T. 1947.

appears at the ground moving in a southeasterly direction. On a frontal zone running from New Mexico towards northeast another disturbance appears at the ground. These two disturbances converge upon each other, and a deep cyclone is formed at the surface and a deep trough aloft, over Middle West. Through the confluence over western Canada a jet is intensified and a frontal zone is sharpened in that region.

The lower half of fig. 9 shows a cross section of temperature (°C) normal to the current, from the Pacific to Fort Smith in Canada and, the upper half, the same cross section with wet bulb potential temperatures and lines of equal wind velocities in m/sec. In fig. 10 the corresponding data are given for the following day but the cross sections are now made between Miami, Florida and International Falls in Canada. Since the wind velocity varies considerably with height, there is only one level (about 600 mb) at which the air of the second cross section can be identified with the air in the cross section for the previous day. Nevertheless, the figures

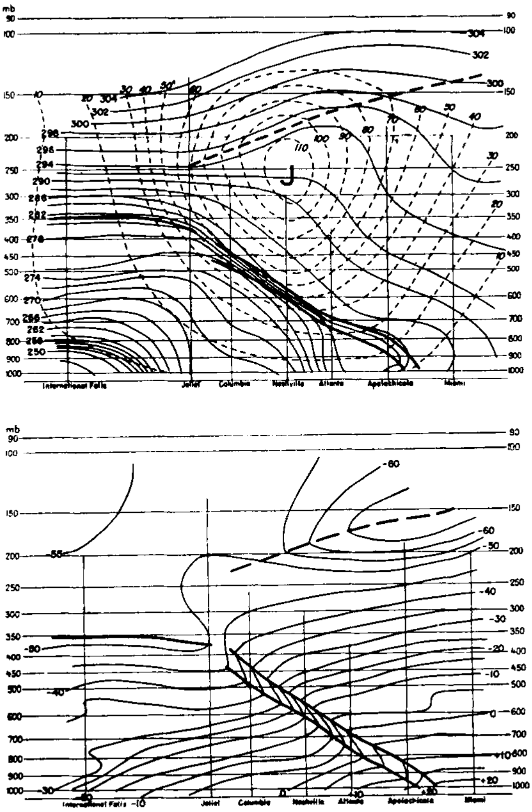


Fig. 10. Cross-sections of temperature, below, and wet bulb potential temperature, above. Dashed lines show distribution of computed gradient wind velocity. The cross-sections are taken normal to the jet stream. Heavy lines show fronts and tropopauses. Jan. 20, 0300 G. M. T. 1947.

show certain significant differences. In the first place, the front has intensified considerably which is seen from the temperature cross section, and also from the wet bulb potential cross section. The same fact appears even more marked in fig. 11 which shows the soundings in Prince George and Port Hardy on the 20th and in Nashville and Columbia on the 21st. While there are three isotherms (in degrees C) in the front on the 20th, there are five or six on the 21st and the isothermal frontal sheet has developed into an inversion. The same feature can be seen at all levels. At the same time the wind velocity in the jet has increased from about 60 m/sec on the 20th to about 110 m/sec on the 21st, when the jet is in a more southerly position. The tendencies in 500 mb in tens of feet from Jan. 20 to Jan. 21 are given in the charts in fig. 5 and 6 showing the deepening and movement of the trough. The sharpening of the front, the deepening of the trough and the increase of the jet stream thus seem to be simultaneous.

We find that the jet approximately has the normal wind profile given by ROSSBY. The winds are gradient winds computed from the geopotential differences on the cross section and corresponding level charts. There are temperature errors in the soundings and uncertainties in the proper choice of path curvature; additional errors are introduced through the projection of the sounding stations on the cross section plane. Though the given values of the wind velocities may deviate 20 % from the real values it seems nevertheless probable that the wind velocities are larger than those obtained from ROSSBY'S theory under the assumption of constant vertical absolute vorticity north of the jet and a vorticity at the pole equal to twice the angular velocity of the earth. This discrepancy may be due to the vertical motion of the air. However, no definite answer can be given to that question at the present time and the problem of vertical motion has to be studied further. It is obvious that the constant

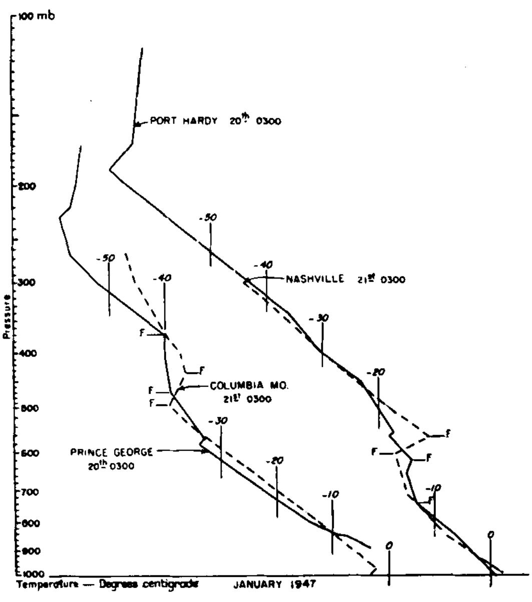


Fig. 11. Soundings plotted in a T-log p diagram. Temperatures are given in degrees C.

absolute vorticity profile is obtained in regions where cold air through large scale turbulent mixing has been brought far to the south carrying its high absolute cyclonic vorticity with it. Similar profiles but with weaker winds will probably be found in sharp ridges where warm air is brought far to the north.

The deepening mentioned above was, as far as one can see, an effect of the confluence of warm and cold air masses over western Canada, a process whereby cold air could be brought southward on the eastern side of the area of rising pressure. The process was going on for a rather long time and ceased when no more warm or warmer air was brought into the ridge and no more cold or colder air was available to be brought into the system. We thus may conclude that the deepening was influenced by previous processes over the Pacific. The warm air coming from a southerly position with a low absolute cyclonic vorticity acquired a relative anticyclonic vorticity and created a ridge over the eastern Pacific. Following an approximately constant absolute vorticity trajectory it run southeastward over the U. S. creating a pressure fall. But through the pressure fall the cold air was being drawn into the system. We obtained a direct circulation with sinking cold air and rising warm air, as shown by the temperature observations; the potential energy which was set free accelerated the jet and lead to a pressure fall over the Middle West which was stronger than the initial pressure rise over the Rockies, as seen from fig. 5 or fig. 6.

A a result of its sinking the cold air has a tendency to deviate from a constant vorticity trajectory; In the lower layers the cyclonic curvature is reduced, in the upper layers increased; In the warm air the modifications are in the opposite sense.

Every situation is unstable to some extent. A weak disturbance acting during a short time interval may start a development (deepening) where potential energy is converted into kinetic energy, but if the disturbance ceases to act as destabilizing force and turns to act as a stabilizing one, the process will stop. On the other hand if the disturbing force is acting in the same direction for longer periods the development may be considerable. The whole problem, as is well known, is very complex. It seems to me that it might be attacked with more hope for success by studying the positions, movements, intensity and amount of cold and warm air masses together with the possibility of bringing these air masses together, than by restricting oneself to a study of the situation in small areas. The latter study can only give the sign of a development in the first moment.

REFERENCES

1. CRESSMAN, G. P. 1948: On the forecasting of long waves in the upper westerlies. *Journ. of Meteor.* **5** No 2, pp. 44—57.
2. DINES, W. H. 1912: Total and partial correlation coefficients between sundry variables of the upper air. *Geoph. Mem.* **1** No 2, pp. 31—47.
3. DINES, W. H. 1919: The caracteristics of the free atmosphere. *Geoph. Mem.* **2** No 13, pp. 47—76.
4. HEADQUARTERS, AIR WEATHER SERVICE: Historical Weather maps, for the northern hemisphere (sea level and 500 mb), October and November 1945, Washington D. C.
5. HESS, S. L. and WAGNER, H. 1948: On the prognostic usefulness of the correlation between pressure and temperature changes at 3 km. *Bull. Amer. Meteor. Soc.* **29** No 5, pp. 219—226.
6. NAMIAS, J. 1947: Extended forecasting by mean circulation methods. Washington, U. S. Department of Commerce, Weather Bureau. 89 pp.
7. NYBERG, A. 1945: Synoptical-aerological investigation of weather conditions in Europe, 17—24 April 1939, *Statens meteorologisk-hydrografiska anstalt, Communications, Series of papers*, No 48, 122 pp.
8. PALMÉN, E. 1948: On the distribution of temperature and wind in the upper westerlies. *Journ. of Meteor.* **5** No 1, pp. 20—27.
9. RIEHL, H. 1948: Jet stream in upper troposphere and cyclone formation. *Trans. of the Amer. Geophys. Union*, **29** No 2, pp. 175—186.
10. ROSSBY, C. G. 1938: On the mutual adjustment of pressure and velocity distributions in certain simple current systems, II. *Journal of Marine Research*, **1**, No 3, pp. 239—263.
11. ROSSBY, C. G. 1939: Relation between variations in the intensity of the zonal circulation of the atmosphere and the displacements of the semipermanent centers of action. *Journal of Marine Research*, **2**, No 1, pp. 38—55.

12. ROSSBY, C. G. 1947: On the distribution of angular velocity in gaseous envelops under the influence of large-scale horizontal mixing processes. *Bull. Amer. Meteor. Soc.*, **28**, No 2, pp. 53—68.
13. SCHEDLER, A. 1921: Über den Einfluss der Lufttemperatur in verschiedenen Höhen auf die Luftdrucksschwankungen am Erdboden. *Beitr. Physik fr. Atmos.*, **7**, pp. 88—101.
14. SOLBERG, H. 1936: Le mouvement d'inertie de l'atmosphère stable et son rôle dans la théorie des cyclones. *Procès-verbaux de l'Assoc. de Météor., U. G. G. I.*, Edinbourg, pp. 66—82.
15. UNIVERSITY OF CHICAGO 1947: On the general circulation of the atmosphere in middle latitudes. *Bull. Amer. Meteor. Soc.*, **28**, No 6, pp. 255—280.
16. WILLETT, H. C. 1944: Descriptive meteorology. New York, Academic Press. 310 pp.