Structure of Shear Lines near the Tropopause in Summer

C. W. NEWTON, N. A. PHILLIPS, J. E. CARSON, and D. L. BRADBURY
University of Chicago[1]

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Abstract

A detailed study is made of a shear line in the tropopause region. A summer case was purposely selected, in which the lower half of the atmosphere was relatively barotropic, in order to minimize the role that could possibly be played by a subsiding cold air mass in producing a vertical stretching and consequent concentration of vorticity in the higher atmosphere.

During the period studied, the circulation over North America was characterized by the existence of two distinct jet streams, the southernmost of which, though initially having the form of a sinusoidal trough, appears to "buckle", forming a very sharp trough separating two jet streams of opposite direction. A possible mechanism is suggested for the production of strong northerly flow leading up to shear-line formation, which depends upon a mutual interaction between the mass and velocity fields of the jet-stream systems located at different latitudes.

An attempt to determine whether potential absolute vorticity is conserved during formation of the shear line is not wholly conclusive, but indicates a reasonable possibility that vorticity may be generated in the free atmosphere.

Introduction

Although much has been published regarding the importance of such phenomena as cut-off lows and higher-latitude warm highs as cells for the meridional exchange of air masses, certain related features of the circulation have received little attention in the literature. One of these is the shear line frequently found near the tropopause level, with which this paper will be concerned.

For purposes of this discussion, the shear line may be defined as a line of more-or-less abrupt discontinuity of wind direction in the free atmosphere, somewhat analogous to the wind shift found at surface fronts. The free-atmosphere shear line, however, is not generally associated directly with a density discontinuity, but is, in the troposphere, more likely to be associated with a temperature *minimum* relative to the environment.

Two principal questions prompted interest in this subject: (1) what causes the breakdown of the westerlies into a shear line, rather than the more commonly observed type of trough with continuous flow of air through it, and (2) can the high values of vorticity observed in shear lines be accounted for by conservative processes? The latter question appears to be of fundamental importance not only from the general standpoint of fluid dynamics, but in particular because of the increasing interest in numerical weather prediction, present methods of which are based upon the assumption that potential absolute vorticity is conserved (CHARNEY, 1949); (CHARNEY, FJÖRTOFT and VON NEUMANN, 1950).

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Direct evidence that vorticity may not be conserved in limited regions of the free atmosphere has recently been presented by R. FRITH (1950). Photographs of a rift brought about by dry-ice seeding of supercooled Sc clouds over southern England showed a narrow zone with cyclonic shear estimated to be 15 mph in 0.25 mi (1.7×10^{-2} sec^{-1}), two orders of magnitude greater than the shear commonly observed in the higher atmosphere near the jet stream. In this instance, the cloud base was at 2,000 ft and the tops 4,500 ft, with no evidence of a front nearby.

Examinations of time and space variations of upper winds indicate that the strongest wind in the vicinity of a shear line is usually found 200 to 400 km from it, with wind decreasing to calm just in the shear line. The detailed distribution of wind shear is at present indeterminable because of sparsity of observations, but in most cases the mean absolute vorticity over a broad zone spanning the shear line is about three times, and occasionally up to five or six times, the Coriolis parameter. The shear line therefore seems to be a likely place to look for the possible generation of vorticity by nonconservative processes.

Three possibilities exist for the origin of high vorticities: vertical stretching, advection, or generation *in situ*. It is one of the purposes of this study to determine whether the high vorticity observed in shear lines can be accounted for by the first two (conservative) processes, or whether generation of potential absolute vorticity takes place locally.

Some previous investigators (e. g., HSIEH, 1950) have attempted to explain the formation of regions of strong vorticity in shear lines as being the result of vertical stretching above a subsiding cold air mass. Undoubtedly this mechanism is operative in many cases of shear line formation; however, shear lines are also observed to form on occasions when such a cold air mass is lacking. A basic difficulty is also encountered in this explanation in that the subsidence of such a cold dome would lead to a more barotropic distribution of temperature, while actually increased baroclinity is required to achieve quasigeostrophic balance since the vertical wind shear in the bounding currents often increases during the formation of a shear line. It is obvious from the circulation theorem and considerations of continuity, that a purely local process which increases the barocline field cannot at the same time concentrate vorticity in the shear line region.

In order to minimize the possibility that subsidence could play an important rôle in the formation of the shear lines to be studied, summer cases were selected in which frontal activity was weak and confined to the lowest layers. Two cases were studied in detail, of which only part of one is presented in this paper because of the excessive number of diagrams which would otherwise be involved. The conclusions reached were consistent between the two cases, which were very similar in their general features.

Case of 6—10 July 1950

Shear lines are characteristically most pronounced near the tropopause level, which in the present case is most nearly represented by the 200-mb surface over the United States. As the basic charts in the synoptic description a series of 200-mb charts is presented in figs. 1—6, embracing the period 1500 GCT 6 July to 0300 GCT 10 July 1950.

Certain broad-scale features of the circulation pattern are persistent throughout the development considered here: a major trough extending southward from a cold low center near the Canadian west coast, followed downstream by a slowly progressive (until fig. 5) ridge extending from a closed warm high near the southwest border of the United States, and a slow-moving trough in the region south of the Great Lakes.

Near the Arctic Circle, there are two principal quasistationary systems: a cut-off warm tropospheric high over Alaska and a deep cold low north of Hudson Bay. The sparse observations indicate the probable existence of a rather extensive west-wind belt south of latitudes 15—25° N. A series of high centers strung out in an E—W direction between latitudes 25 and 40° N is separated from the lower-latitude westerlies by a series of shear lines or sharp troughs, particularly well developed in the southwest Atlantic and Caribbean.

The major changes during this period are the intensification and eastward movement of the middle-latitude ridge over the western

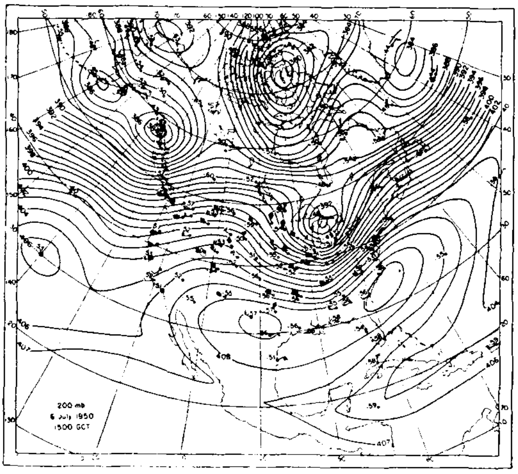


Fig. 1. 200-mb chart at 1500 GCT 6 July 1950. Contours labelled in hundreds of feet, contour interval 100 ft. At stations, temperature in degrees C, wind in knots (full barb 10 knots, solid triangle 50 knots; dashed wind shafts and open triangles represent winds at 35,000 ft).

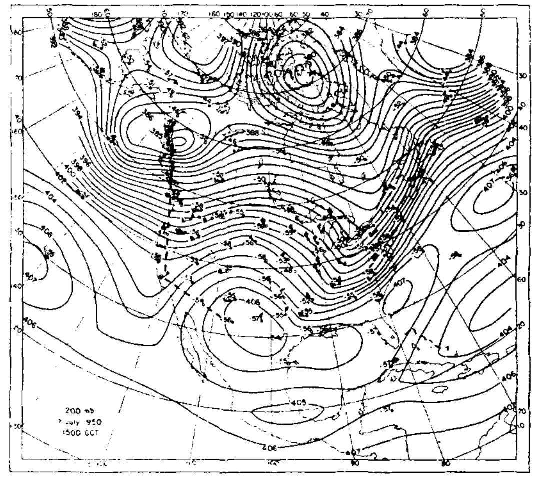


Fig. 2. 200-mb chart, 1500 GCT 7 July 1950.

United States, concurrent with a slow deepening of the west-coast trough. Simultaneously with the intensification of this ridge, the trough over the east central United States underwent a sharpening, at first rather gradual (figs. 1—3, interval between charts 24 hrs), then quite rapidly until a distinct NNE—SSW shear line appeared in the chart for 1500 GCT 9 July (fig. 5). During the first part of the period, there was a considerable flow of air through the trough line at all times, but by 9 July the winds bordering the shear line are almost parallel to it, so the latter time is characterized by only nominal flow through the trough line at 200 mb. The flow pattern in this region may thus be considered as relatively stable and slowly changing up until shortly before 1500 GCT 8 July (fig. 3) and for some time after 1500 GCT 9 July (fig. 5), with a rapid breakdown from one type of flow to the other during the short intervening period.

Even before the complete breakdown of the trough into a clear-cut shear line, there must evidently have been a pronounced cross-contour flow toward higher contours (deceleration) on the immediate southwest side, and probably a corresponding flow toward lower contours on the southeast side of the trough, since air cannot pass through a sharp trough at a high kinetic-energy level because of the excessive centrifugal force which would be involved. The rapid deepening of the southern part of the shear line trough might be considered as the effect of a tendency to mutual adjustment between the mass and velocity fields as a result of the feeding into the region west of the trough, of a northerly current possessing high momentum. To find an explanation for the generation of these strong northerly winds would therefore seem to be a basic problem in getting at the mechanism of shear line formation.

Twelve hours after the time of fig. 1, the high-latitude trough over Hudson Bay had amalgamated with the trough south of the Great Lakes, forming the extensive trough seen in fig. 2 over the eastern part of the continent. The northern part of this trough moved steadily eastward, eventually shearing off and again (*figs. 5 and 6*) *assuming an identity* entirely separate from the lower-latitude part. The latter is evidently retarded by the persistent lower middle-latitude warm high over the western Atlantic.

As often observed in the case of cut-off lows (University of Chicago, 1947), (Cressman, 1948), the northern part of this shear line is, in the fully-developed stage, bridged across by strong westerly flow, and the shear line trough is out of phase with the long-wave pattern

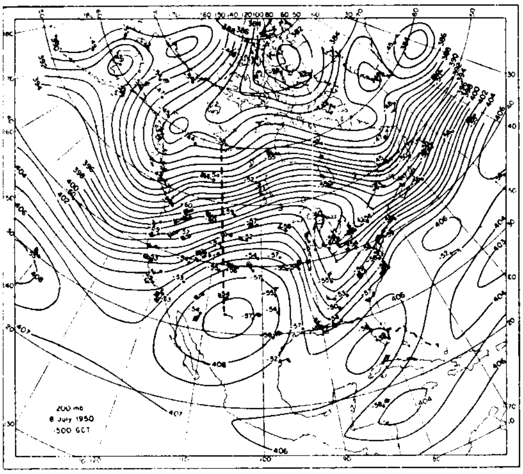


Fig. 3. 200-mb chart, 1500 GCT 8 July 1950.

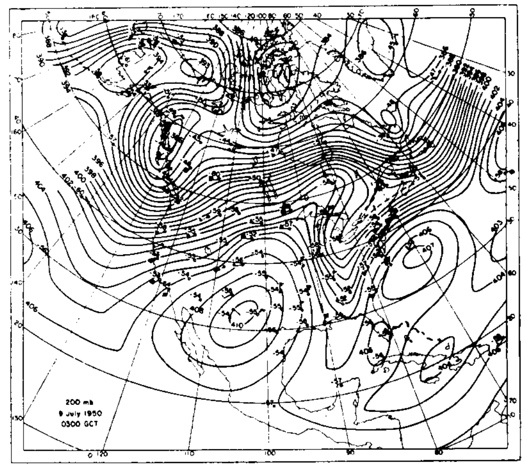


Fig. 4. 200-mb chart, 0300 GCT 9 July 1950.

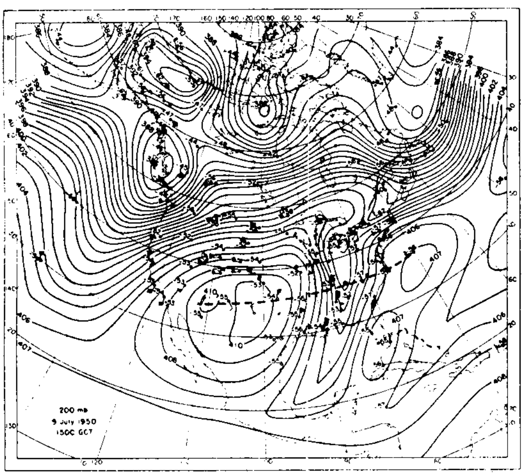


Fig. 5. 200-mb chart, 1500 GCT 9 July 1950.

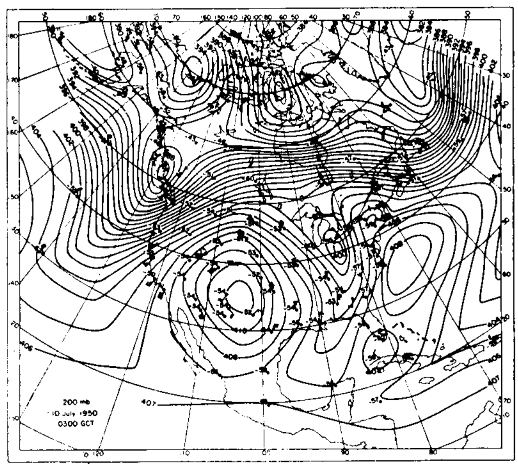


Fig. 6. 200-mb chart, 0300 GCT 10 July 1950.

in the main belt of westerlies. In the early stages (fig. 1), the principal belt of westerlies passes almost entirely south of the low centered over the Great Lakes. This low subsequently fills, but a splitting of the belt of westerlies remains, with successively more of the westerly current passing north of the split in response to an increase in strength of the southwesterly flow over western Canada as the major west-coast trough deepens.

The lower middle latitudes are in the last stages characterized by a series of large-amplitude disturbances bearing some resemblance to "blocking" situations of the higher latitudes described by Rex (1950) and others. In this case, the block starting the train of meandering disturbances is furnished by the strongly developed closed warm high centered over west Texas and New Mexico late in the series. The principal point of dissimilarity between the flow pattern illustrated here, and the typical higher-latitude blocking situation, is that in this case the meandering takes place entirely south of the principal belt of westerlies, which continues eastward without much diminution of the total level of its kinetic energy. In the higher-latitude blocking type, on the other hand, the initial blocking high

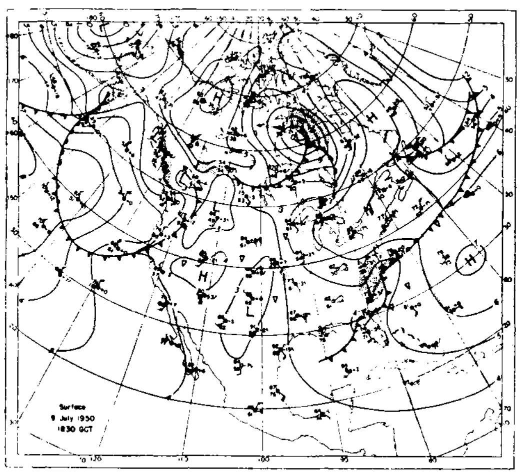


Fig. 7. Surface chart, 1830 GCT 9 July 1950. Contours of the 1000-mb surface in hundreds of feet; fronts indicated by standard symbols.

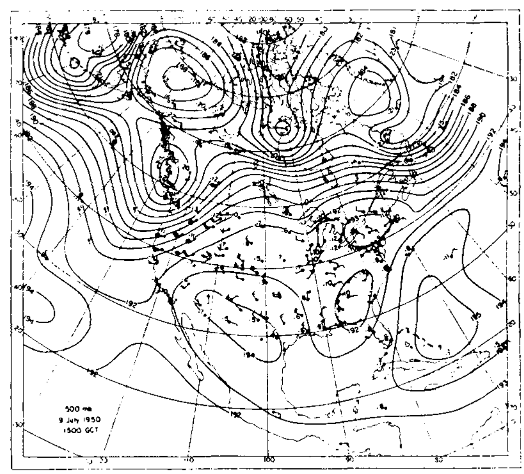


Fig. 8. 500-mb chart, 1500 GCT 9 July 1950.

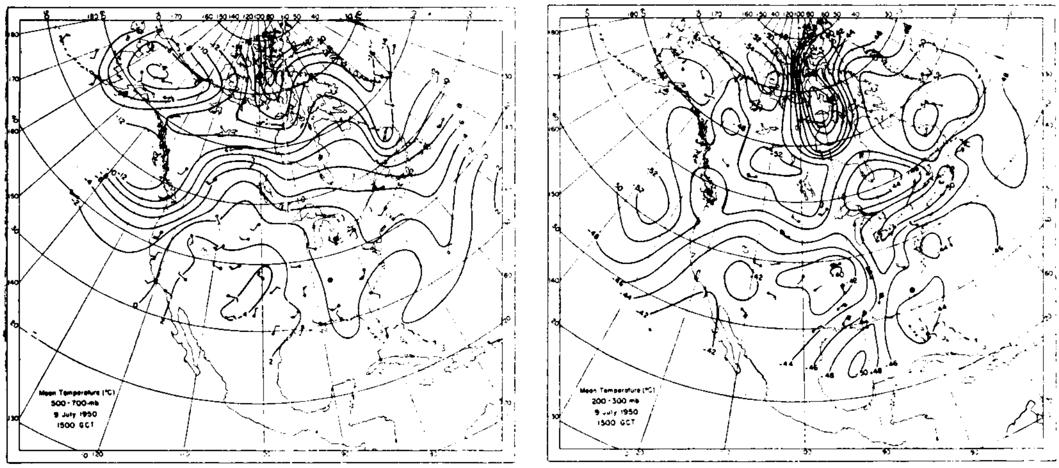
forms north of the latitude of the strong current upstream, and the kinetic energy level is considerably diminished downstream from the block.

Rex states that the prime requisites for development of a high-latitude blocking situation are (1) a high kinetic energy level in the basic current and (2) interference from an external perturbation, namely a strong cyclone upstream from the blocking region. There is therefore a certain similarity with the present situation from this point of view, since the development of the block, shear line, and the rest of the meandering disturbances downstream occurs during a period when the trough upstream off the west coast of North America is increasing in amplitude.

The increased amplitude of the shear trough obviously cannot be considered simply as an effect of the change of constant-vorticity paths as a result of the deepening upstream, since the amplitude of the downstream disturbance is considerably greater than that observed upstream. However, it does appear likely that the building-up of the blocking high is at least partly a result of mutual adjustments of the mass and velocity fields as consequence of increased amplitude of the upstream trough.

Certain experiments conducted in the Hydrodynamics Laboratory of the Department of Meteorology of the University of Chicago produce a flow pattern very similar to that observed here (FULTZ and LONG, 1951). One of these experiments consists of placing a circular obstacle of about 20° latitude diameter in the middle-latitude region of a fluid-filled rotating hemispherical shell, and moving the obstacle upstream with respect to the bowl so that a flow is created analogous to a west wind with respect to the obstacle. When this is done, the fluid flows completely around the northern and eastern sides and reaches a stagnation point just east of the south side of the obstacle. The lines of flow observed east of the obstacle show a remarkable similarity to a NE—SW shear line. The obstacle used in this experiment is of a size approximately equivalent to that of the warm high over the southwest United States in the present case (fig. 6).

Associated flow patterns at lower levels. — Figs. 7 and 8 show the surface chart at 1830 GCT and the 500-mb chart at 1500 GCT 9 July, the time of greatest shear line intensity at 200 mb (cf. fig. 5). The surface chart is quite typical of the weather patterns commonly associated with high-level shear lines; a surface ridge beneath the shear line and a series of open waves along a SW—NE oriented cold front to the east. Weak development of surface low-pressure centers is characteristic of the frontal systems associated with shear lines; these ordinarily do not intensify appreciably until they move sufficiently far north to come under the influence of the main belt of westerlies north of the shear-line trough.



9 a 9 b

Fig. 9. Relative topography charts for (a) layer between 500 and 700 mb, and (b) layer between 200 and 300 mb. Isolines labelled as degrees C mean virtual temperature. Wind symbols indicate vectorial wind shear between top and bottom of layer.

At 500 mb (fig. 8), the large-scale perturbations are in general similar to those at 200 mb. However, at the lower level there is little indication of the strong deformation found near the tropopause, the only evidence being a rather weak closed low. The strong winds and concentration of solenoids at 500 mb over southern Canada stand out in striking contrast to conditions over the eastern United States at this level. The concentration of baroclinity in the higher part of the troposphere near the shear line is emphasized by the "relative topography" charts of fig. 9. The gradient of mean temperature in the layer between 500 and 700 mb (fig. 9 a) is very weak in the region south of the Great Lakes. In the same region, however, an extremely strong mean temperature gradient is found in the layer between 300 and 200 mb (fig. 9 b). There is correspondingly a very great difference between the vectorial vertical wind shears (indicated by plotted wind symbols) in these two layers of almost equal thickness.

If the deepening of surface cyclones is in any way dependent upon the presence of a jet stream, as some synoptic studies suggest (Riehl, 1948), the failure of waves to deepen until they move northward beneath the main (northern) belt of westerlies is readily understood from comparison of the structures of the northern and southern jet stream systems. The latter being almost nonexistent in the middle and lower troposphere even though it is quite pronounced at higher levels over the wave systems on the east coast, it seems logical to expect it to have very little effect upon surface cyclogenesis. This might, of course, be interpreted also as a failure to deepen because of the absence of available potential energy in the lower and middle troposphere.

Structure of wind field. — In order to show some variations of the wind field in space and time, the three vertical cross sections of figs. 10, 11, and 12 are presented. These show the isentropes and distribution of the wind component normal to the lines along which the cross sections were drawn (nearly the total wind speed). Isotachs on these sections were constructed from observed winds, supplemented by gradient winds computed from pressure profiles taking trajectory curvature into account[2]. The successive sections are so spaced that an air particle travelling at about 30 m sec^{-1} (60 knots) near the right-hand jet would be found successively at the different cross sections at the appropriate times.

[2] The apparent inconsistency between the strong vertical shear and the weak solenoid concentration between 200 and 300 mb on the cold-air side of the jet near Nashville is balanced by differences of path curvature at these levels. The current is almost straight at 300 mb but anticyclonic at 200 mb, giving a centrifugal circulation acceleration in the same sense as the solenoids.

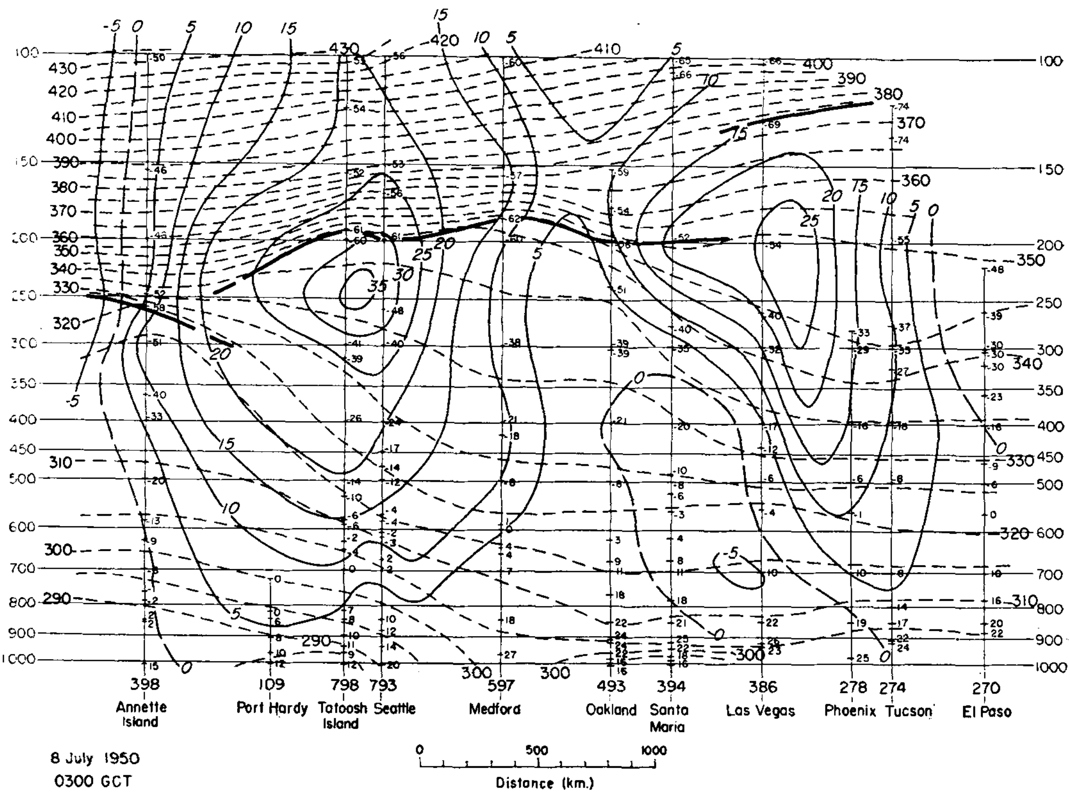


Fig. 10. Vertical cross section from Annette Island to Oakland to El Paso, 0300 GCT 8 July 1950, along heavy dashed line in fig. 2 (note times are 12 hours different). Thin dashed lines, potential temperature (deg. K.); thin solid lines, isotachs of wind component (m sec^{-1}) normal to cross section; heavy lines, tropopauses.

The immediately striking feature of the wind distribution shown here is the pronounced concentration of momentum near the tropopause level in the vicinity of the southern jet stream; particularly in the case of fig. 12, the section through the fully-developed shear line. This is quite different from the vertical wind distribution normally observed in the middle latitudes during winter months, which generally shows rather uniform vertical shear through the troposphere, outside of frontal layers. In the case shown here, no frontal layer was present. There is, however, a marked tendency for the strong vertical shear to be concentrated in a sloping barocline layer entirely within the upper third of the troposphere (upper fourth by weight). While in the typical winter case the strong solenoid concentration and vertical wind shear are found mostly below 400 mb, in the case examined here these properties are found almost entirely above that level. The strongest shear, both vertical and horizontal, is located on the cold-air side of the axis of the jet. The axis of the jet stream at 200 mb west of the shear line (right-hand side in fig. 12) is actually some 500 km distant from the corresponding axis at 500 mb. This type of velocity distribution is not peculiar to this case, but has been noted in a number of summer cases not involving shear lines.

The structure of the southern jet on the west coast and over the Rockies (right-hand sides of figs. 10 and 11) is qualitatively similar to that observed further downstream in the shear line region (fig. 12). All three regions are characterized by a marked tilt of the jet axis and by pronounced concentration of the vertical wind shear in the highest troposphere on the cold-air side of the jet.

Upper wind observations were too sparse to construct a completely reliable picture of the

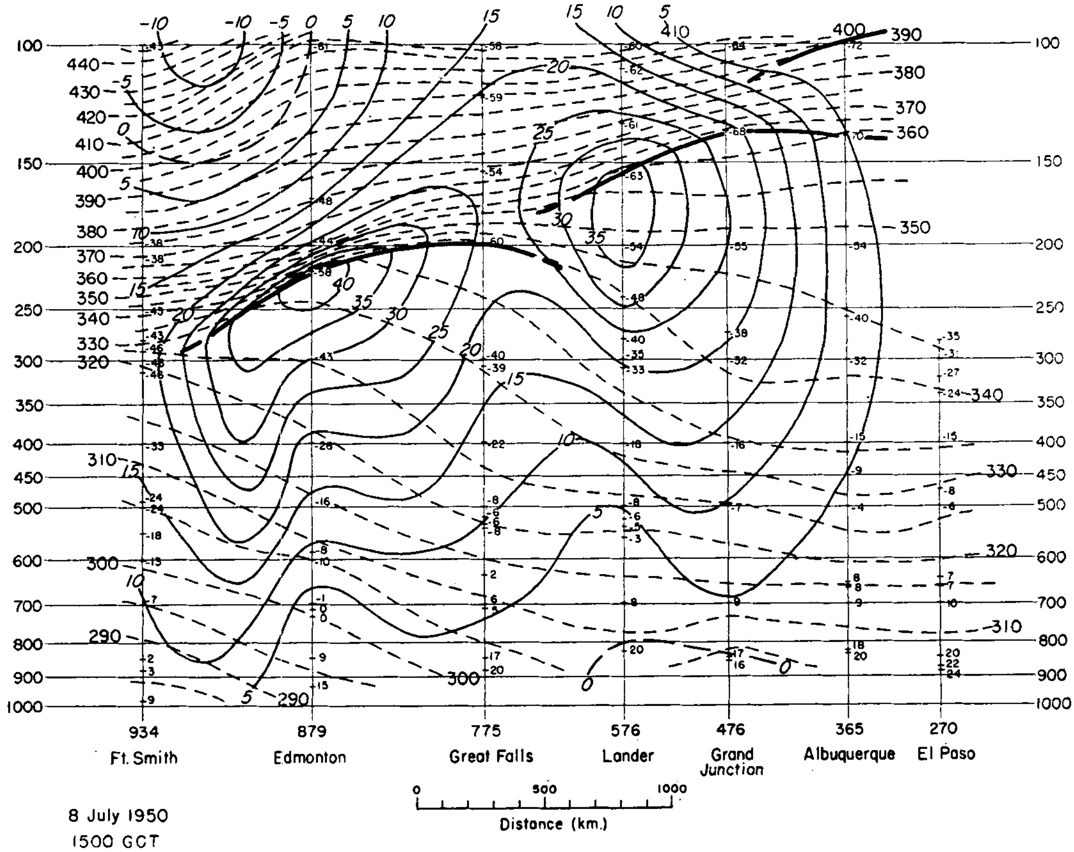


Fig. 11. Vertical cross section from Ft. Smith (Yukon) to El Paso, 1500 GCT 8 July 1950, along heavy dashed line in fig. 3. Legend as in fig. 10.

northernmost jet in figs. 10 and 11, which corresponds to the main belt of westerlies passing north of the shear-line region. Pronounced baroclinity is evident from the earth's surface throughout the troposphere, however, indicating a rather uniform vertical wind shear. Thus the kinematic structure of the northern belt of westerlies is fundamentally different in character from that of the southern wind belt.

Several wind-speed profiles near the axis of the southern 200-mb jet are shown in fig. 13. All of these soundings show the same type of velocity distribution; almost complete absence of vertical shear up to around 30,000 ft and a sharp increase in shear near that level. At Nashville, the 30,000-ft wind is only 6 m sec^{-1}, while at the 200-mb level the wind speed is 58 m sec^{-1}. Theoretical aspects of this type of vertical variation of velocity are dealt with in two papers by Rossby (1951 a, b).

Although the distribution of wind velocity in the southern jet in figs. 10 and 11 is qualitatively similar to that found later on in the shear-line region (fig. 12), quantitatively there is a great difference. In the latter cross section, a very marked increase in velocity has taken place at the levels near the tropopause, with very little change in the velocity at levels below about 300 mb. This is emphasized by the profiles of kinetic energy taken from the cross sections of figs. 10 to 12 and shown in fig. 14 a along the sloping jet axis and in fig. 14 b at the 200-mb level. The total kinetic energy changes relatively little between the first two cross sections, but an extremely large increase takes place between the latter two. This increase is almost entirely confined to a

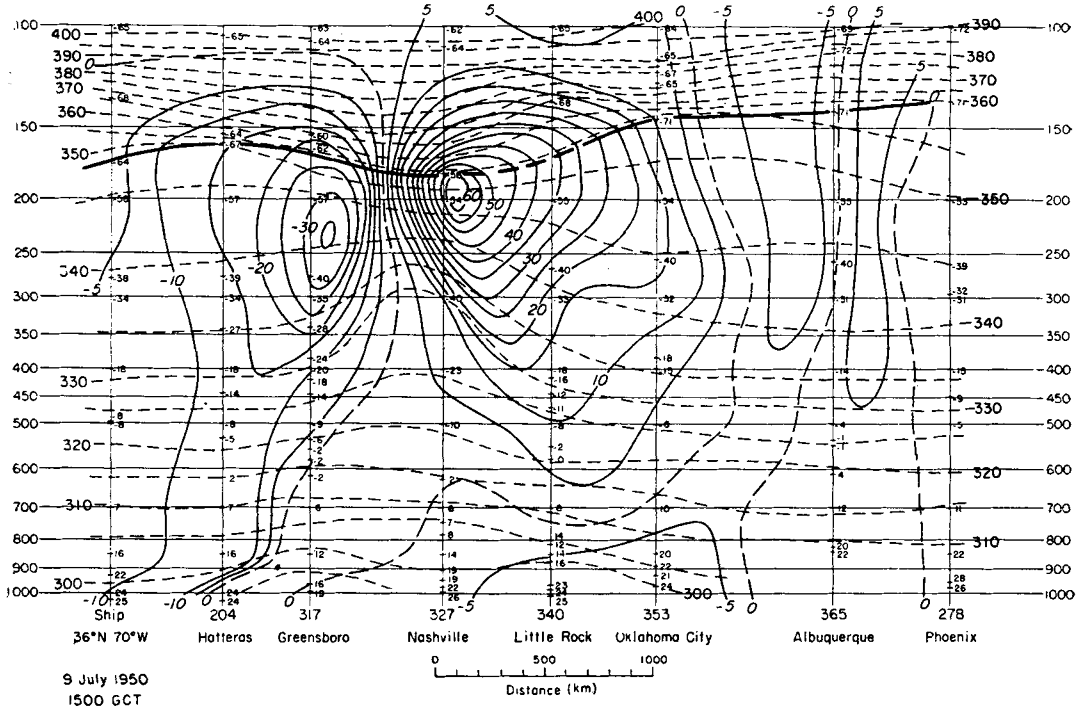


Fig. 12. Vertical cross section from ship located at 36° N 70° W to Phoenix, 1500 GCT 9 July 1950, along heavy dashed line in fig. 5. Legend as in fig. 10.

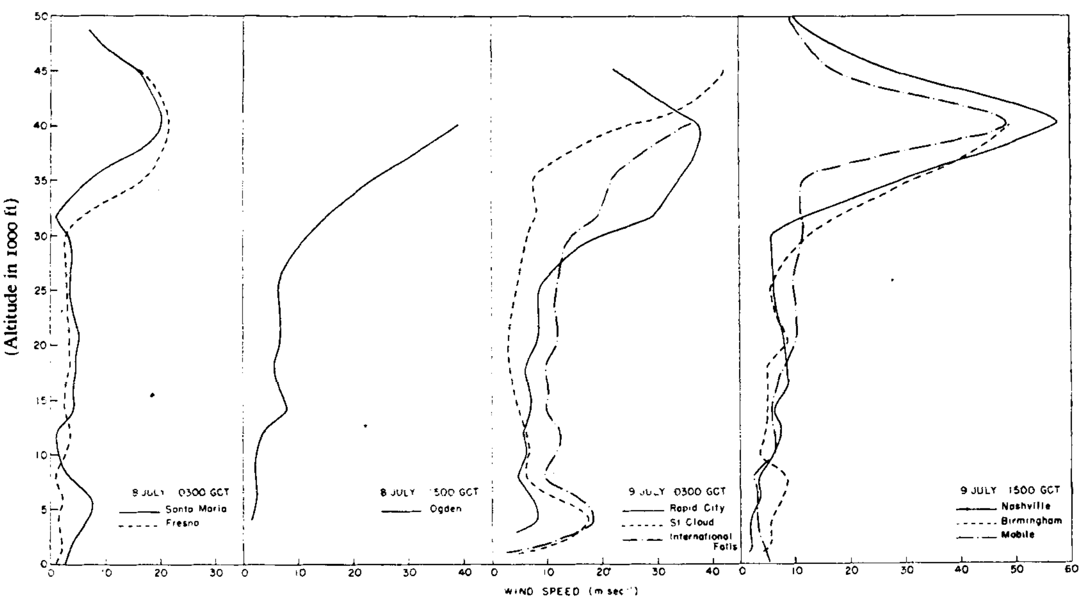
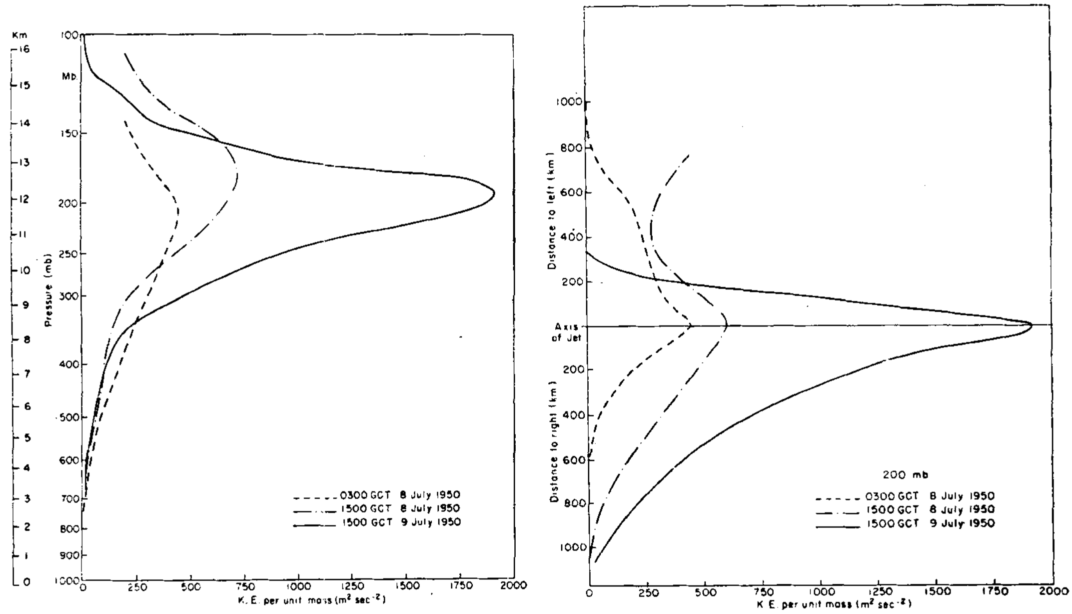


Fig. 13. Profiles of wind speed at various locations (time and location indicated in figure) near axis of 200-mb jet stream, 8—9 July 1950.



14 a 14 b

Fig. 14. (a) Profiles of wind speed along sloping axis of maximum wind, taken from southernmost jet in figs. 10, 11, and 12; (b) profiles of wind speed at 200-mb level through same jet.

layer about 150 mb (4.5 km) thick and 1,000 km wide[3]. Wind observations were too sparse to construct an intermediate cross section along the ridge line at 0300 GCT 9 July (fig. 4), but available observations seem to indicate that the increase in kinetic energy took place almost entirely on the east side of this ridge.

The structure of the 200-mb wind field over North America at the time of maximum shear line development is shown by the isotachs of fig. 15. The coexistence of two distinct jet-stream systems at different latitudes is clearly shown by this figure. The double jet structure was evident from a careful examination of the wind observations throughout the series from fig. 1 to fig. 6. At 1500 GCT 9 July (figs. 12 and 15), the maximum wind speed on the west side of the shear line was approximately twice that observed on the east side. This is not an invariant feature, though it is commonly observed during the earlier stages of shear line formation. Available wind observations and the contour analyses clearly show a marked acceleration of the air on the east side of the ridge just upstream from the shear line; that is, it is plainly evident that the strong concentration of momentum in this region did not originate wholly by advection from points further west[4].

Distribution of vorticity in shear line region. — Fig. 16 shows the distribution of the vertical component of absolute vorticity at various levels, taken from the cross section through the shear line (fig. 12). The high degree of concentration of the strong vorticity, both in a vertical and horizontal sense, is strikingly evident from this figure. At 200 mb, Nashville had a wind of 360° 112 knots, while

3 Since the velocity does vary along the profile, comparing these profiles in fig. 14 does not really indicate changes in velocity of the same particles at the different cross sections. However, the flow in the lower-velocity boundary regions of the jet was sufficiently close to steady-state to allow a comparison to be qualitatively correct.

4 J. Bjerknes, in an article to be published in the Compendium of Meteorology of the American Meteorological Society, has suggested that strong acceleration occurs east of sharp ridges due to the fact that curvature of the pressure-contours in such ridges often exceeds a critical anticyclonic curvature for the velocity of the air passing through the ridge, this necessitating a flow toward lower contours and therefore an acceleration east of the ridge. This appears to be a significant factor in the present case.

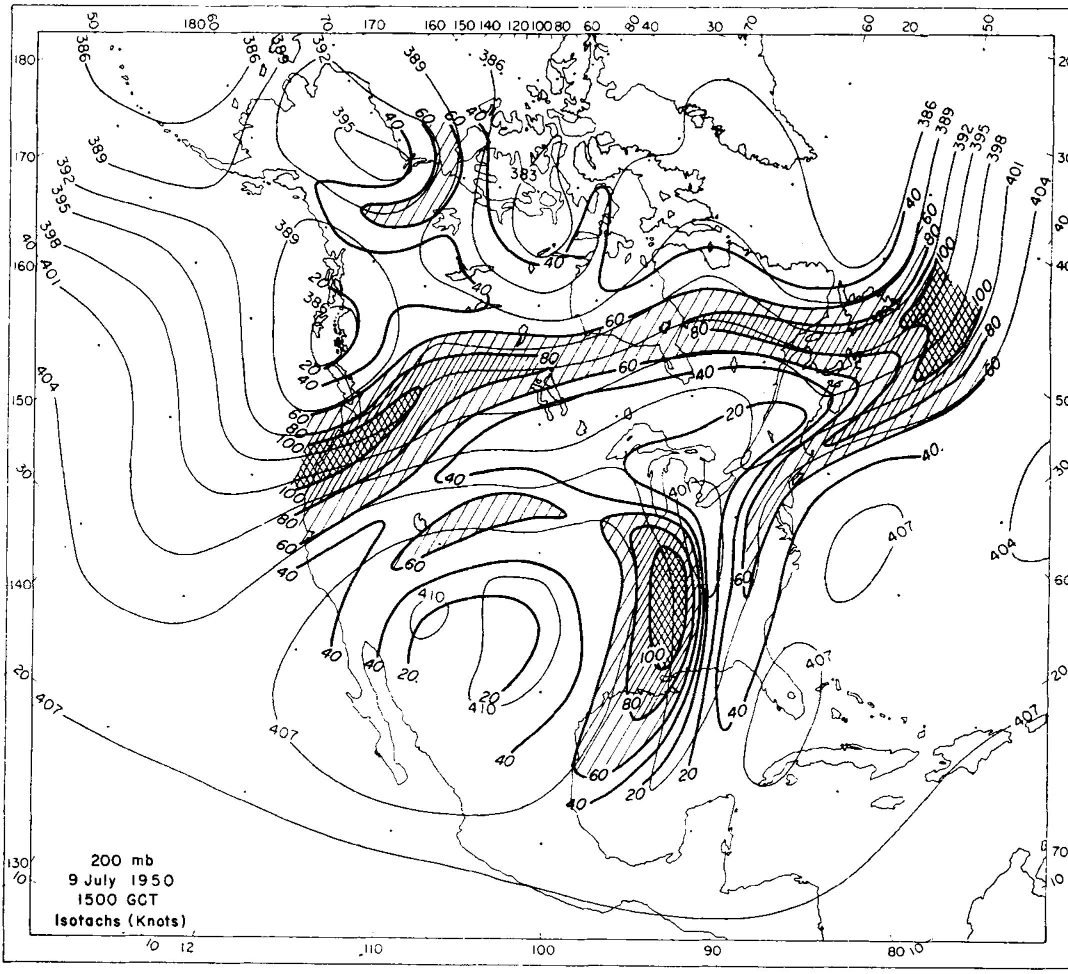


Fig. 15. 200-mb chart, 1500 GCT 9 July 1950. Thin lines, contours of 200-mb surface, contour interval 300 feet. Heavy lines, isotachs (knots). Hatching indicates regions of wind speed greater than 60 knots, cross hatching speed greater than 100 knots.

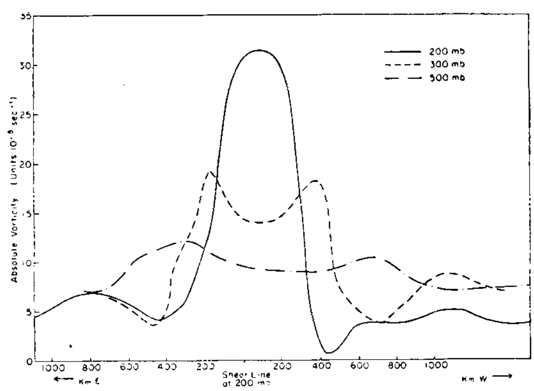
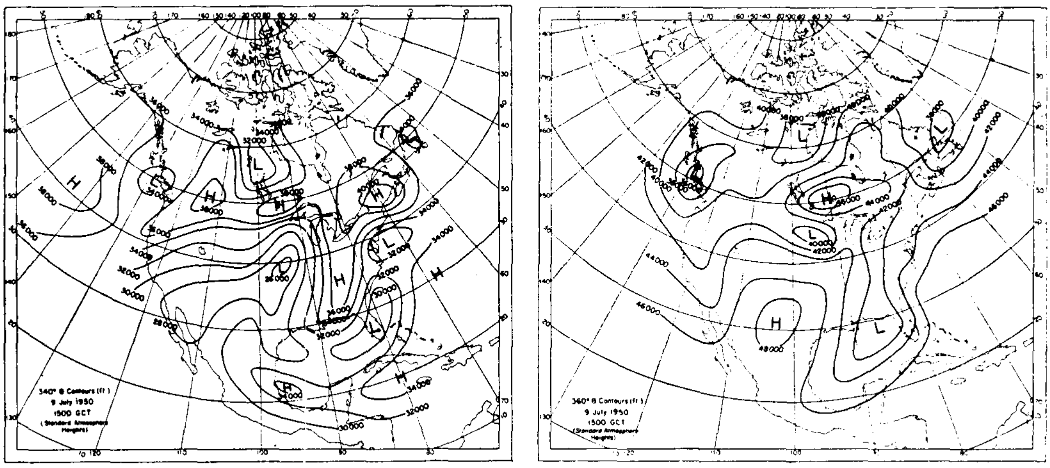


Fig. 16. Profiles of the vertical component of absolute vorticity at 200 mb (solid line), 300 mb (dashed line), and 500 mb (dash-dotted line), across the shear line at 1500 GCT 9 July 1950, taken from fig. 12. Units 10^{-5} sec^{-1}.



17 a 17 b

Fig. 17. Contours (feet) of isentropic surfaces at 1500 GCT 9 July 1950, (a) 340K and (b) 360K potential temperature. H and L indicate regions of high and low elevation.

Knoxville, 260 km farther east, had a wind of 40° 2 knots. Making the too-conservative assumption of a linear variation of wind speed between these two stations, an average absolute vorticity of 30.4×10^{-5} sec^{-1}, or 3.6 times the Coriolis parameter, is obtained.

Comparing the distribution of vorticity with that of stability as shown by the vertical packing of isentropes in fig. 12, we see that the strongest vorticity is located, in the upper troposphere, just in that region where the vertical stability is most pronounced. The weakest vertical stability in the vicinity of the shear line is found at lower levels (300—400 mb) where the horizontal shear is weak.

In order to find out whether the high vorticity was generated locally or transferred advectively from other regions, an attempt was made to estimate paths of particles in relation to the potential absolute vorticity field. Since the 340 and 360K isentropic surfaces in this case straddle the tropopause, envelope a semi-nodal surface and also enclose the layer wherein the greatest vorticity and momentum concentrations are found, this layer was chosen as the most logical to examine for the purposes of this study.

Fig. 17 illustrates the topography (standard atmosphere heights) of the 340 and 360K surfaces at the time of maximum shear line intensity. The outstanding feature is the almost complete out-of-phase relationship of the troughs and ridges of these two surfaces over the United States. Over the major anticyclonic regions, the 360K surface is found at high, and the 340K surface at low elevations, while the opposite is observed in the vicinity of the shear line. This implies, as does the cross section of fig. 12, that regions of high absolute vorticity are, near the tropopause, associated generally with pronounced thermal stability, while the anticyclones, regions of

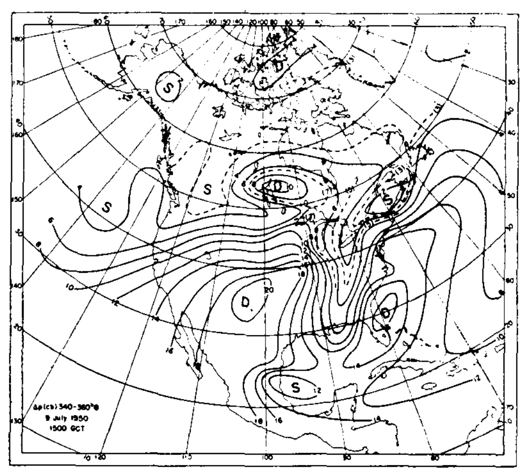


Fig. 18. Isentropic "weight chart" for the layer 340 to 360K at 1500 GCT 9 July 1950. Contours labelled in centibars. S, "shallow" and D, "deep" represent regions of small and large pressure differences respectively.

low vorticity, are relatively unstable in their upper parts. Thus both the horizontal and vertical gradients of potential vorticity are in general in the same direction and of greater magnitude than the gradients of absolute vorticity.

As an aid in the analysis of the potential absolute vorticity field, a series of isentropic weight charts (340 to 360K) was analyzed at 12-hourly intervals. Fig. 18, showing the weight distribution at 1500 GCT 9 July, the time of maximum development of the shear line, is reproduced from this series[5]. The preceding charts showed considerable irregularity of smaller-scale features, but the continuity of the broadscale features was fairly well preserved. During the three days preceding fig. 18, a gradual amplification of the "deep" thickness tongue over the southwest was observed, together with the extension southward of the "shallow" tongue in the shear line region which at first appeared upstream as a much smaller perturbation. South of latitude 45° N, there was close parallelism between the isentropic thickness pattern and the contours of the 200-mb surface, throughout the development (compare figs. 5 and 18).

The conservation of potential absolute vorticity. — The equation expressing the conservation of potential absolute vorticity and potential temperature may be written in the form

$$\frac{d}{dt}\left(\frac{f+\zeta}{\triangle p}\right) = 0, \qquad (1)$$

where f is the Coriolis parameter, ζ is the vertical component of the relative vorticity, and $\triangle p$ is the pressure difference between two isentropic surfaces[6]. This equation is valid only when the non-conservative "frictional" forces are negligible and the motion is adiabatic. Thus in the absence of frictional and non-adiabatic effects, air columns contained between two isentropic surfaces can increase their absolute rate of rotation about the vertical only when the isentropic surfaces move farther apart (so that $\triangle p$ increases).

Application of this theorem to the cross section through the shear line (fig. 12) implies that the appearance of high absolute vorticities in the fully-developed shear line must have previously been associated with even smaller values of isentropic thickness (or greater vertical stability) than they are associated with on the cross section. The isentropic weight charts showed that there were regions to the north of this shear line where the pressure difference $\triangle p$ between the potential temperature surfaces 340 K and 360 K was smaller than the smallest value of $\triangle p$ in the cross section through the fully-developed shear line (fig. 12). It may therefore be possible that the shear line was formed by a process whereby columns of air with great stability ($\triangle p \sim 20$ mb) and normal values of absolute vorticity stretched vertically and increased their absolute vorticity at least approximately in conformance with the vorticity equation (1).

One way in which this possibility can be tested is to analyze synoptic charts of the potential absolute vorticity, $(f+\zeta)/\triangle p$. Charts of $(f+\zeta)/\triangle p$ at the 200-mb level were constructed for the six observation times preceding and including the time of maximum development of the shear line at 1500 GCT 9 July. For these charts, the relative vorticity was computed geostrophically, using the relation

$$\zeta = \text{scale factor} \times \nabla^2 z_{200},$$

where z_{200} is the height of the 200-mb surface. $\nabla^2 z_{200}$ was evaluated by finite differences from the analyzed 200-mb contours, using an overlapping rectangular grid with intervals of approximately 500 km (except over the eastern part of the United States where a non-overlapping grid interval of 250 km was used). It must be pointed out that this method will tend to minimize extreme vorticity values if they exist over a small region — as is the case with the shear line. After adding the relative vorticity to the Coriolis parameter, the potential absolute vorticity $(f+\zeta)/\triangle p$ was then obtained by dividing the absolute vorticity $f+\zeta$ (in units of 10^{-5} sec^{-1}) by $\triangle p$ (in cb) for the isentropic layer 340—360 K.

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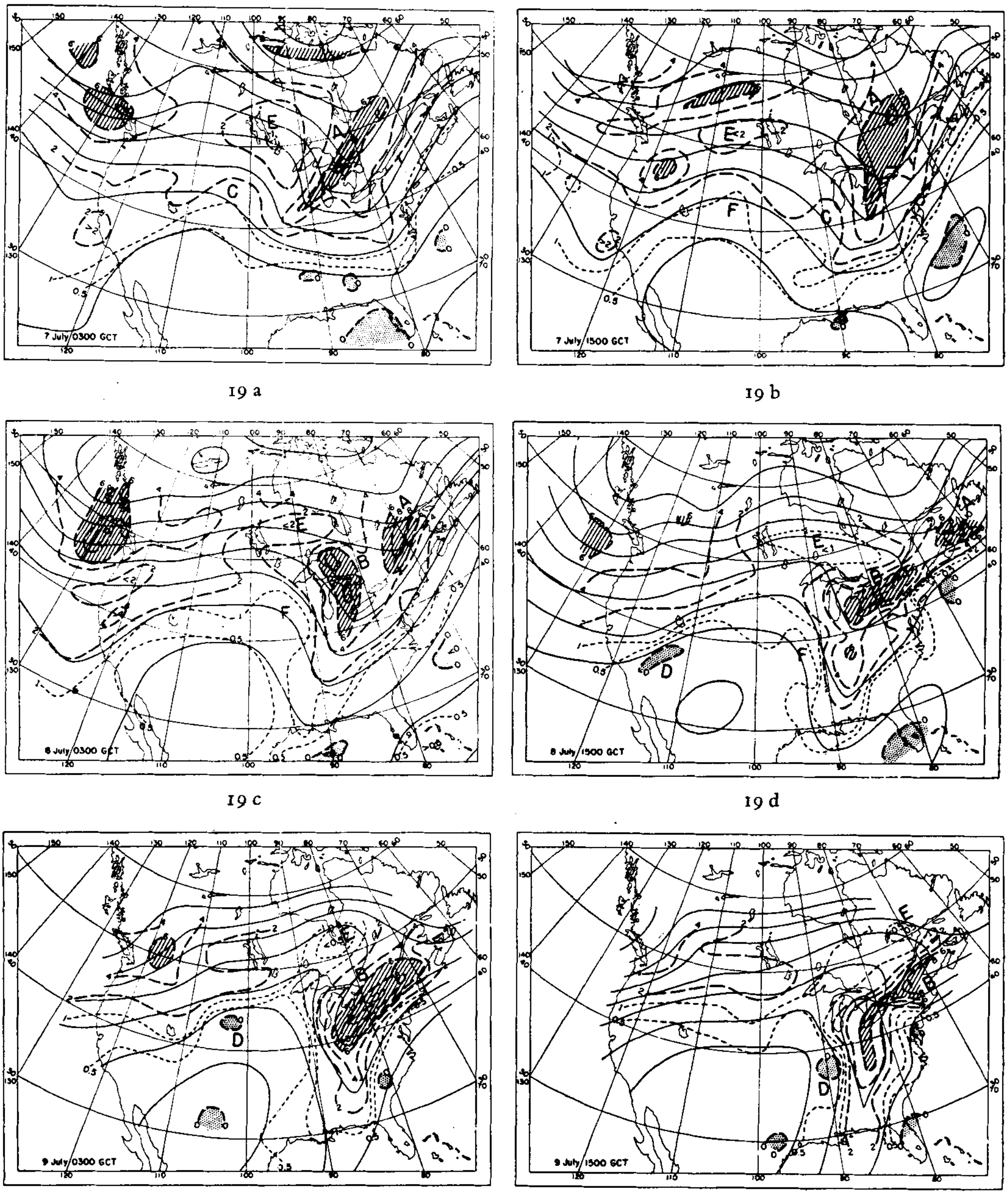
[5] As a qualitative guide in the analysis of these charts, $\partial^2 \mathbf{v}/\partial z^2$ (parallel to the weight lines) was computed from wind observations at the 35-, 40-, and 45-thousand foot levels.

[6] In general, the conservation theorem must also include the horizontal components of the absolute vorticity vector and potential temperature gradient, as shown by Ertel (1942), but these may ordinarily be disregarded when the isentropic surfaces are not excessively steep.



19 a 19 b 19 c 19 d 19 e 19 f

Fig. 19. Charts of potential absolute vorticity, $[(f + \zeta)/\Delta p]$, 0300 GCT 7 July to 1500 GCT 9 July 1950. Dashed lines are isolines of potential vorticity (units of 10^{-5} sec^{-1} cb^{-1}) and thin continuous lines are contours of 200-mb surface at intervals of 300 feet. Regions of maximum (> 6) and minimum (< 0) potential vorticity are indicated by hatched and stippled areas respectively.

The charts prepared in this manner are reproduced in fig. 19.

An examination of these charts shows that in general the continuity of the isolines of $(f + \zeta)/\Delta p$ was quite good in the central part of the region. For example, the features indicated by the letters A, C, D and F show very good continuity on successive charts. On the other hand the center of minimum potential vorticity (indicated by E) that is present near the Canadian-United States border throughout the series, exhibits little tendency to move with the flow pattern. Similarly, the center of maximum potential vorticity indicated by B appears suddenly on the chart for 0300 GCT 8 July (fig. 19c) with no previous history. There are two possibilities of explaining these two discrepancies: (*a*) The analysis of potential vorticity in southern Canada is essentially correct and the anomalous behavior of systems B and E is to be explained by the non-conservative processes neglected in (1); (*b*) the more scattered aerological observations over southern Canada make a sufficiently accurate analysis of the potential absolute vorticity impossible at this level. It is difficult to conceive of non-conservative processes sufficiently intense to explain completely the sudden appearance of center B, but it is possible that a combination of the two possibilities (*a*) and (*b*) can account for its appearance.

The marked intensification of the shear line occurs between 1500 GCT on the 8th and 9th, corresponding to the last three charts in fig. 19. The simultaneous formation of the strong but shallow meridional jet on the west side of the shear line is associated with the increase of the zonal gradient of potential vorticity in the same region. At least a part of this increase (and thus in a sense the formation of the jet itself) can be explained by advection of potential vorticity since the center of minimum $(f+\zeta)/\Delta p$ (letter D) shows excellent continuity from 1500 GCT 8 July to 1500 GCT 9 July. During this 24-hr period there is also a local increase of potential vorticity in the region of the shear line itself. It is quite possible that this also can be explained qualitatively by advection since the charts for 1500 GCT 8 July and 0300 GCT 9 July show that the "ridge" in potential vorticity lagged slightly behind the trough in the contour lines. A more detailed analysis of this last point is not possible, however, because of the relatively long time- and space-intervals between observations.

Especially important with respect to the conservation of potential vorticity is an exact determination of the velocity field in the vicinity of the shear line. As pointed out previously, the most conservative assumption (that of a quasi-linear field) gives the smallest possible value of the maximum absolute potential vorticity in the shear zone. If the velocity field is not this simple, the maximum values of potential absolute vorticity in the shear zone may be considerably larger than the values indicated on the charts. It would then be very difficult, if not impossible, to explain the final formation of the shear line by a process compatible with the individual conservation of potential absolute vorticity.

Interpretation of mechanism of shear-line formation. — As pointed out earlier, the most significant event connected with the shear-line formation appears to be the initial generation of a strong northerly current on its west side. Once such a current has been established, the wave trough in which it is initiated must become unstable, since a large amount of momentum is then injected into lower latitudes where the pressure gradient is insufficient to balance the Coriolis force acting upon the current. Deepening of the pressure trough comes about through a tendency to mutual adjustment of the mass and velocity fields, and the generation of an oppositely-directed current east of the shear line follows from intensification of the zonal pressure gradient as a result of this deepening[7].

Although an increase in amplitude took place in the trough off the west coast at about the time of most pronounced development of the shear-line trough (figs. 1 to 6), the dispersion of energy from such a deepening trough could not in itself cause the formation of a trough of even greater amplitude downstream (see Yeh, 1949). The deepening likewise did not seem to take place through local barocline processes of the type ordinarily associated with concentration of vorticity; baroclinity in the lower troposphere was very weak, while baroclinity in the upper troposphere increased with time in the shear-line region.

It was concluded from the analyses presented earlier (e. g., fig. 15) that the pronounced local

[7] It should be pointed out that the discussion here refers only to the type of shear line that develops initially in a single westerly current. The type that forms, e. g., between a strong W or SW current of considerable extent upstream, and an initially separate E or NE current flowing around an anticyclone located at a higher latitude, appears to be of a fundamentally different character.

increase of northerly winds in the region west of the shear line did not evolve through advection from other regions. Since none of the above processes appears to furnish a complete explanation for the local acceleration, it is clearly necessary to look for a possible mechanism due to some other external influence. Since such an influence is not apparent in the southern jet-stream system alone, it is logical to consider what possible interactions could take place between the momentum and pressure fields in the vicinity of the jet stream further north, and the wind field of the southern jet-stream system.

To determine what this interaction might be, an attempt was made to study the isallohyptic and wind-velocity fields. For the case described above, this attempt was unsuccessful because of lacunae or inaccuracies of data in certain critical regions. A second case was therefore studied (31 July to 5 August 1949), in which it was possible to arrive at a satisfactory analysis of these fields. The flow patterns in the two cases were quite similar, and qualitatively the velocity and height-change fields in the July 1950 case were similar to those of the August 1949 case, insofar as the points discussed below were concerned. In order to conserve space, the charts already presented for the former case will be referred to in this discussion, together with a schematic diagram prepared with the help of more adequate data from the latter case.

Examination of the 200-mb charts between 1500 GCT 7 July and 0300 GCT 9 July shows that during this period the southwesterly flow over western and central Canada, in the northern jet-stream system, increased strongly as the major ridge in the northern belt of westerlies moved eastward to a position north of the shear line. This suggests that the height rise SW of James Bay (figs. 3 and 4) may be considered as the result of the amplification and eastward movement of the major ridge in response to deepening of the west-coast trough, in combination with a "banking" or mutual adjustment of pressure and velocity fields as suggested by Rossby (1937), following the injection of a current possessing high momentum into a region initially possessing a pressure-gradient field insufficient to balance the Coriolis force acting upon the injected stream.

The field of height changes imposed in this manner is illustrated by the dashed lines in fig. 20, in relation to the two jet-stream systems. The greatest rises are found just to the right of the northern jet, and the greatest total rises over a longer period of time (but not necessarily in a given 12-hr period) take place at approximatively the half-wave length $\frac{1}{2}\lambda_s$ of a stationary wave from the upstream trough, given (Rossby, 1940) by

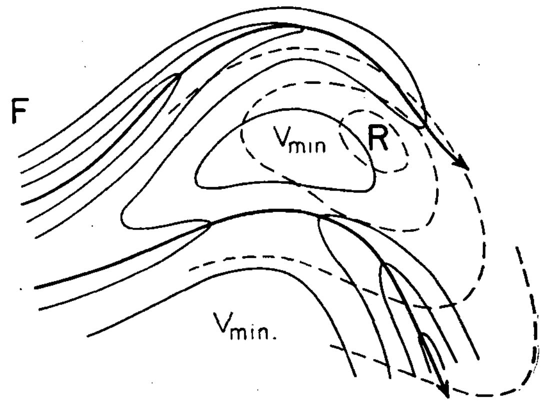


Fig. 20. Schematic diagram of height-change field accompanying increase in a amplitude of northern part of complex ridge due to processes acting upstream in northern jet-stream system. Solid heavy lines, axes of jet streams; light solid lines, isotachs; light dashed lines, lines of equal height rise decreasing in magnitude away from rise center R. V_{min} denotes regions of slow wind speeds.

$$\frac{1}{2}\lambda_s = 90\sqrt{\frac{2\,U}{\omega a\cos^3\varphi}},$$

where U is the speed of the basic westerly current, a is the radius of the earth and ω its angular rotation, and λ_s is measured in degrees longitude.

The rise-center R thus imposed in the northern part of the ridge relaxes the height gradient across the southern jet near the ridge line. In the case of a ridge moving at a speed much less than that of the wind in the center of the jet, the radius of trajectory curvature R_T is approximately given by the gradient wind equation

$$R_T = \frac{v^2}{f\,(v - v_g)}, \qquad (2)$$

where R_T is taken as positive for anticyclonic curvature, v is the actual and v_g the geostrophic wind speed. The relaxation of the height gradient due to the rises centered at R in fig. 20 must, by decreasing v_g, also decrease R_T. In the limiting case ($v_g \to 0$), the trajectory curvature approaches that of an inertia circle of radius v/f. As seen from figs. 2 to 4, the relaxation of the height gradient in the southern part of the ridge was very pronounced, so the progressive clockwise turning of the southern jet east of the ridge is in accord with the conclusion from equation (2). This conclusion assumes that v does not decrease with time in the vicinity of the ridge, an assumption that appears to be borne out by the observations and would be expected from the nearly constant input of momentum into the ridge region from farther upstream.

In both of the cases studied, it was clear that an axial acceleration of the winds developed east of the ridge upstream from the shear line, as indicated schematically by the isotachs in the lower right-hand portion of fig. 20. This implies that a pronounced cross-contour flow developed where previously it was negligible. The magnitude of the movement across contours may be estimated from the energy equation for adiabatic motion on an isobaric surface,

$$\frac{c^2}{2} + gz = \text{const.}$$

For the air to accelerate from 60 to 100 knots, a total height change of 80 m toward lower heights is necessary. This is approximately in agreement with the amount determined by inspection from fig. 15 in the region SW of Lake Michigan, along the axis of maximum wind speed. Since fig. 15 does not represent a steady-state condition, however, the angle of flow across contours as estimated from this figure alone is somewhat exaggerated. This angle may be computed from the equation

$$\frac{dv}{dt} = -g\left(\frac{\partial z}{\partial s}\right)_p = f v_g \tan \alpha,$$

where v is the total wind speed along the direction s of a streamline and α is the angle measured counterclockwise from contour to streamline. This angle was computed to be about 15° in the region of strong acceleration SW of Lake Michigan in fig. 15.

No satisfactory explanation for the development of this pronounced axial acceleration has been found. It is recognized, of course, that the imposition of a localized region of height rises such as indicated in fig. 20 contributes not only to increasing the anticyclonic curvature of the streamlines according to equation (2), but also to increasing the curvature of the contours in the southern part of the ridge. If the increase in contour curvature through this process exceeds the increase in streamline curvature, the required axial acceleration can be produced east of the ridge in the southern jet. It is by no means obvious how this condition can be produced. Thus the simple mechanism proposed above for the interaction between jet streams at different latitudes appears to partially answer the question of how the strong northerly winds can be produced in the region west of the incipient shear-line formation, insofar as a turning of the winds is involved, but does not provide an explanation for the simultaneous increase in total momentum of the northerly current.

Conclusion. — No claim can be made that a completely satisfactory answer has been given to either of the questions stated in the opening paragraphs. Through the expedient of choosing cases at the outset in which no pronounced lower-tropospheric cold dome was present, it is believed that the process of vertical stretching of the warm air above such a subsiding cold dome has been eliminated as a significant factor in the formation of the shear lines or the production of the strong vorticity observed therein. The study of changes in the potential absolute vorticity field, while not wholly conclusive because of limitations in the data, at least suggests that potential vorticity may have been generated through nonconservative processes in the free atmosphere.

A possible mechanism has been suggested for the production of a strong meridional current leading to instability of the shear line trough; this mechanism may have more general application in understanding the interference patterns of waves in westerly currents at different latitudes. The question as to why the instability should take on the form of a shear line rather than simply an increase in amplitude of a continuous current flowing around a trough, needs further investigation. The answer appears to be connected with the initially weak baroclinity throughout the troposphere in the lower latitudes into which the strong meridional current is injected; the rapidity with which changes in the kinematic field take place may make it impossible to build up a solenoid concentration rapidly enough and of sufficient intensity to deflect the current eastward.

The extreme concentration of momentum in a shallow layer near the tropopause observed in the case discussed above, appears to be of great synoptic as well as theoretical interest. Particularly in view of the interest shown by synoptic meteorologists in the connection between the jet stream and cyclogenesis, it would appear profitable to study the differences in behaviour of jets characterized by this concentration of momentum, and those characterized by more even distribution of baroclinity throughout the troposphere. Lastly, studies of this nature emphasize the need for more adequate observational data in the higher troposphere and lower stratosphere; particularly wind data, since it is difficult if not impossible to construct a completely satisfactory picture of the kinematic field from the sparse data now available in the higher levels over North America.

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