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On the General Circulation of the Atmosphere in Middle Latitudes

A Preliminary Summary Report on Certain Investigations Conducted at the University of Chicago during the Academic Year 1946–1947 *

By

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OF THE UNIVERSITY OF CHICAGO

SUMMARY

A synoptic, theoretical and experimental study of the general circulation of the atmosphere in middle latitudes and of the major perturbations superimposed on this circulation pattern was conducted by a group of research workers at the University of Chicago during the academic year 1946–1947. The principal results of these investigations are summarized below in a series of specific statements. Not all of these conclusions are new, and some require further verification, but it is hoped that they may be of some use as a starting point for further discussions and investigations of the general circulation of the atmosphere.

I. Zonal Circulation in the Northern Hemisphere during the Winter Half Year

1. During periods of reasonably straight west wind circulation over the North American continent, there exists normally, at levels between 5 km and 15 km above sea level, a fairly narrow zone of extremely strong west wind circulation (jet stream), reaching its maximum intensity and sharpness at the tropopause level (300 mb–200 mb).
2. South of the jet stream maximum, and at the tropopause level, the zonal wind decreases extremely rapidly with latitude. This decrease may amount to as much as 20 to 25 mph per degree of latitude over a belt of up to or in excess of five degrees of latitude.
3. The intensity of the jet stream maximum decreases with increasing latitude. In its most southerly location (between 30°N and 35°N) it may possibly reach values of in excess of 250 mph.
4. The process of confluence of air currents of different origin, as studied by the extended forecast section of the U. S. Weather Bureau, frequently contributes to the intensification of the jet stream, but well-developed, narrow zonal currents may occur even in cases when no significant confluence is apparent.
5. The jet stream is located in or just south of a zone in which a large fraction of the middle and upper troposphere temperature contrast between polar and equatorial regions is concentrated.
6. The location of the jet stream coincides with a break in the tropopause. Immediately north of the jet stream one observes a low level tropopause, and south thereof frequently a double tropopause, the upper one of which corresponds to an equatorial tropopause.
7. In the region of the jet stream the isentropic layers of the lower portions of the polar stratosphere are in free communication with the upper isentropic layers of the low latitude troposphere.
8. Below the jet stream, it is frequently possible to identify a well-defined frontal zone, intersecting the ground south of the jet stream and reaching the upper troposphere some distance north of the jet stream. The intersection with the ground is often indistinct over large areas of the hemisphere.

II. The Large-Scale Perturbations on the Zonal Circulation

9. The upper west wind belt does not follow a straight course around the hemisphere but meanders north and south. The wave length (distance between consecutive corresponding bends in the jet stream) varies between roughly 50° and 120° of longitude.
10. These upper waves are generally, although not necessarily always, several times longer than the typical, unstable frontal waves.

* Preliminary Report on a Research Project sponsored by the Office of Naval Research.

11. The form and intensity of the upper long waves exercise a controlling influence on the behavior and movement of the shorter, true frontal waves in the surface layers. (Principle of steering.)

12. The occlusion of a well-developed and intense frontal wave may lead to a marked deepening of an upper wave, but the difference in dimensions precludes the possibility of explaining the formation of long upper waves through the instability of the relatively short frontal waves.

13. The individual long waves of the upper west-wind belt behave as typical dispersive waves and cannot be treated as independent phenomena. The effects of displacements, or of amplitude and intensity changes, in one long wave appear to be transmitted downstream to the next long wave at a speed which markedly exceeds the rate of air motion through these waves, suggesting a profound interdependence of the atmospheric circulations in different parts of our hemisphere.

14. With excessive meandering, cyclonically circulating domes of cold, high-latitude air are frequently cut off and completely surrounded by warm air on the south side of the jet stream. Likewise, pools of warm, anticyclonically circulating air are frequently cut off and surrounded by the cold air on the north side of the jet stream. Above the 700-mb level the exchange of air between high and low latitudes appears to take place primarily through this process.

15. Excessive meandering and cutting off of warm anticyclones may, from time to time, lead to the establishment of one (or several) anticyclones covering all or most of the Arctic; when this happens, the polar air masses appear to be displaced southward in such a manner as to form a ring of semi-connected cold cyclonic centers in middle latitudes.

III. Theoretical Results

16. The zonal winds observed at the tropopause level during periods of straight west wind circulation appear to agree fairly well with theoretical wind profiles computed on the assumption that lateral turbulence leads to an equalization of the vertical component of absolute vorticity north of the jet stream maximum.

17. It appears possible to explain the observed sharp breakdown of zonal wind profiles south of the jet stream maximum as a consequence of mechanically generated inertia instability along the southern edge of the zone of intense lateral mixing; this inertia instability leads to the development of meridional circulations which in turn interfere with the lateral exchange of vorticity.

18. The association between the frontal zone and jet stream suggests that the building up of the jet stream is accompanied by a dynamic concentration of solenoids, to build up geostrophic and thermal wind balance.

19. The apparent rapid interaction between long upper waves is at least qualitatively in accord with theoretically established dispersive characteristics of long barotropic waves in the atmosphere.

IV. Experimental Verification

20. It has been possible for us to produce systematic departures from solid rotation in thin hemispherical, rotating fluid shells, by subjecting them to thermally produced large-scale turbulence. As a result of such mixing processes, a belt of excess relative motion (west wind) develops in middle latitudes of the fluid shell; this west-wind maximum increases in intensity and is displaced equatorwards with increasing intensity of the mixing. On a percentage basis, the departures from solid rotation so produced are in reasonable, rough agreement with those observed in the earth's atmosphere.

Introduction

The group participating in various phases of this research project consisted of the following members:

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Dave Fultz, The University of Chicago

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During the academic year 1946–1947 this research group, working in the Department of Meteorology of the University of Chicago, has been engaged in a study of certain phases of the general circulation of the atmosphere and of the principal perturbations associated with this circulation. These investigations were initiated and generously supported by the Office of Naval Research,[3] but in some respects they may be considered as a continuation of investigations conducted with the support of U. S. Department of Agriculture funds during the years 1936–1939, while two of the authors were associated with the Department of Meteorology of the Massachusetts Institute of Technology. It is the purpose of this paper to summarize briefly the principal results of the investigations conducted by the Chicago group during the last year.

There exists at present, among the various meteorological research groups here and abroad, a noticeable divergence of opinion

[1] Not present at time of formulation of the present report.

[2] Project leader.

[3] Certain phases of these investigations have been supported by funds from the U. S. Weather Bureau and from the Pineapple Research Institute of Hawaii.

with regard to the proper interpretation of several of the basic processes in the atmosphere. To some extent, these differences of interpretation may be attributed to the absence of adequate upper-air observations on a world-wide scale. Because of this divergence of opinion, and in order to arrive at a generally acceptable formulation of the basic problems of the general circulation of the atmosphere in our latitudes, an effort was made to bring together research workers representing widely different points of view. Frequent, informal group discussions were arranged to aid in the final formulation of the basic problems to be attacked and solved. As a further link in this program a series of conferences and lectures was held in December 1946, at which time certain basic questions were discussed with a group of specially-invited research meteorologists from other institutions.

The group participating in our studies consisted both of theoreticians and of research workers experienced in the analysis of observational data from the free atmosphere. To insure full contact between these two groups an experimental unit for the analysis of the current weather situation was organized, and daily map discussions were held at which both synoptic and theoretical research workers were present. The experimental analysis unit was used as a testing ground for theoretical deductions and as an aid in the appropriate formulation of the basic problems to be studied, empirically or theoretically.

Progress in our work has been seriously handicapped by the practical impossibilty of securing upper-air data from certain critical regions, even though such data in many cases were collected during the War. Until a genuinely efficient method of data distribution to interested agencies has been set up it is reasonably certain that Government funds invested in research projects at institutions outside Washington will fail to yield maximum returns. The teletype transmission of current weather data is inadequate for research purposes, since the circuits at the present time do not effectively disseminate the many upper-air observations collected at great expense over a large portion of our hemisphere.

The detailed results of our investigations will be presented later in a series of papers by the individual participants in our research program. In this preliminary summary report an attempt will be made to summarize certain general qualitative impressions and conclusions reached by our entire group as a result of synoptic map discussions and related studies; with these impressions as a starting point, elementary theoretical comments on a few of the more striking features of the general circulation in our latitudes will be offered.

I. Summary of Qualitative Impressions and Conclusions

Recently the Air Weather Service published two important volumes [1] containing daily 500-mb charts of the entire Northern Hemisphere, for the months of October and November 1945. Inspection of such charts (Fig. 1) shows that the middle latitude belt of westerlies generally may be traced as a continuous stream around the globe but that the belt itself is surprisingly narrow and may be described as a meandering river winding its way eastward through relatively stagnant air masses to the north and south. Theoretical considerations of the instability of shearing motion [2] suggest that the south side of such a west wind belt would be characterized by the formation of anticyclonic eddies (Fig. 2). Such eddies have, as is well known, been found in the lower layers of the troposphere during the summer season, when the west-wind belt itself is displaced so far to the north that the eddies are brought within the reach of our continental radiosonde network. The eddies may then be identified with the aid of isentropic charts [3].

The 500-mb charts published by the Air Weather Service, as well as similar, daily charts for the winter 1946–47 prepared in our own analysis section, indicate the presence of an entirely different type of instability. They show (Fig. 1) that the west-wind belt relatively seldom follows a

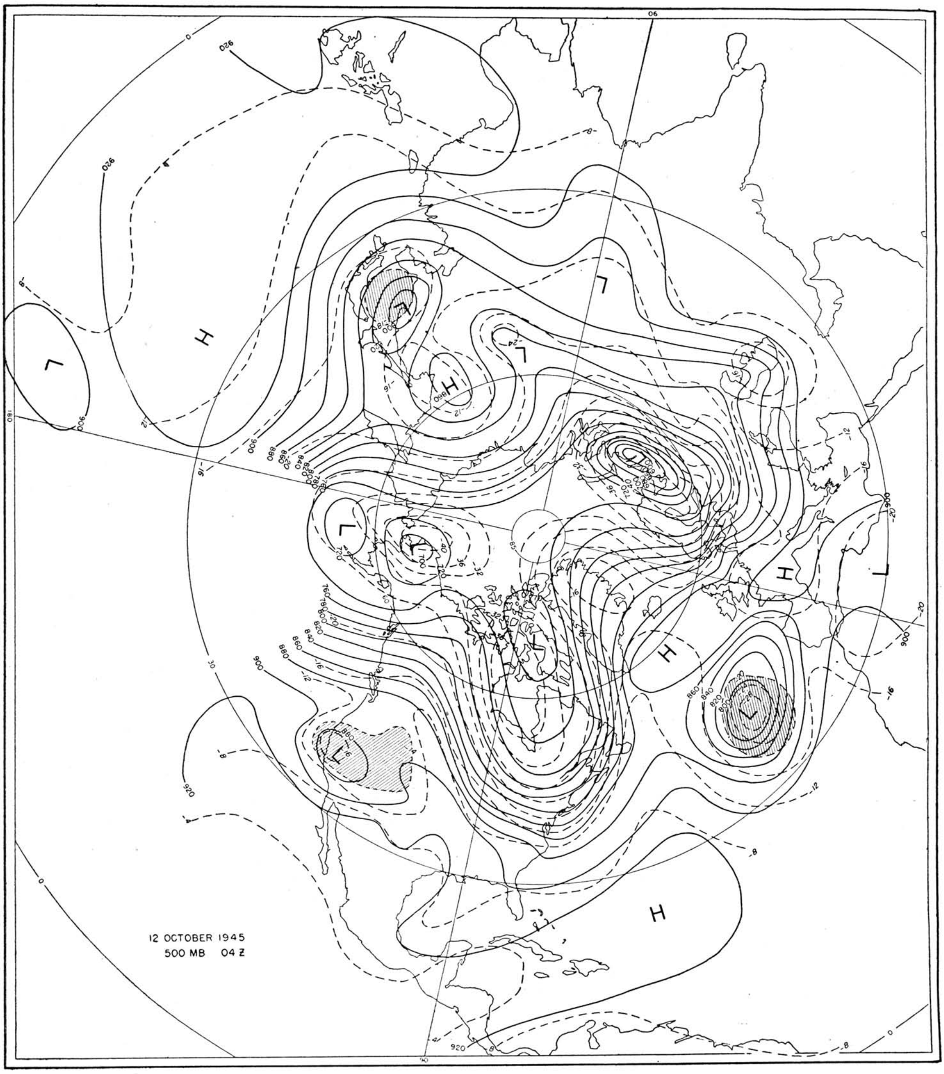

FIG. 1. Contours (full lines) and isotherms (dashed lines) at the 500-mb level on October 12, 1945, 0400Z. Contour lines drawn for intervals of 200 feet. Note cold cyclonic vortices south of zonal current. Note also long-wave pattern with an average wave-length of 120°. Based on data in [1].

straight course eastward but generally exhibits a wave-like pattern with wave-lengths varying from perhaps 50 to 120 degrees of longitude. Superposition of the surface front patterns on a 500-mb chart reveals, in a very striking fashion, the difference in scale between the individual, short frontal waves and the long waves of the upper west-wind belt (FIG. 3) and suggests that as a rule the two types of wave motion are of entirely different origin. However, this is not necessarily always so, since cases may be observed in which the deepening and occlusion of a wave cyclone are fol-

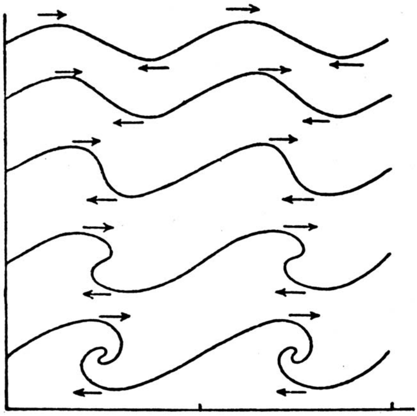

FIG. 2. Vortex formation resulting from shearing instability according to Rosenhead [2].

lowed by the marked deepening of an upper wave or even by the formation of a closed center at the 500-mb level. The general parallelism of the frontal zone, on which the minor, true frontal waves travel, and of the upper wind current does indicate, however, that the development of the frontal zone must be associated with the development of the upper west-wind belt.

The long upper waves often appear unstable in the sense that their amplitudes increase to the point where the troughs or ridges of the waves are cut off from the main stream and form closed or nearly closed vortices to the north and south of the main current. A good illustration is furnished by the cutting off of a cold cyclonic vortex and the elimination of a trough in the Azores region during the period October 9–12, 1945. (FIGS. 4, 5, 6, 1.) *In contrast to the shear-zone eddies, vortices formed in this manner are cyclonic on the south side of the west-wind belt, anticyclonic on the north side thereof.*

It is readily seen that the cutting-off process can lead to a reduction in the wave

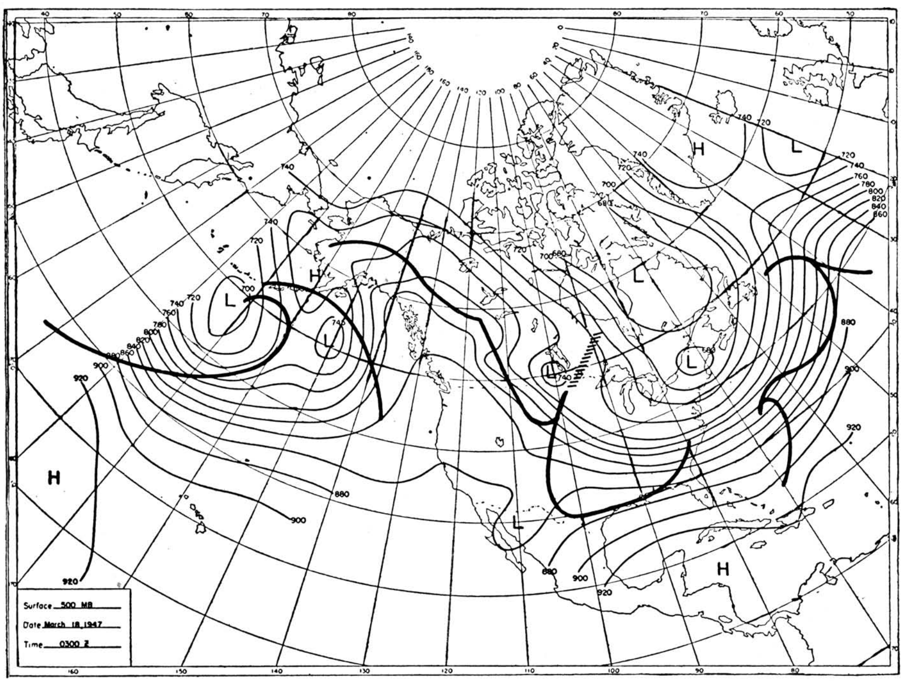

FIG. 3. Surface fronts superimposed on 500-mb chart, for March 18, 1947, 0300Z. Note difference in scale between frontal waves and long-wave pattern in zonal wind current. There are at least six frontal waves as against two major upper waves. Note also general coincidence of zonal wind belt and region of sea-level frontal activity.

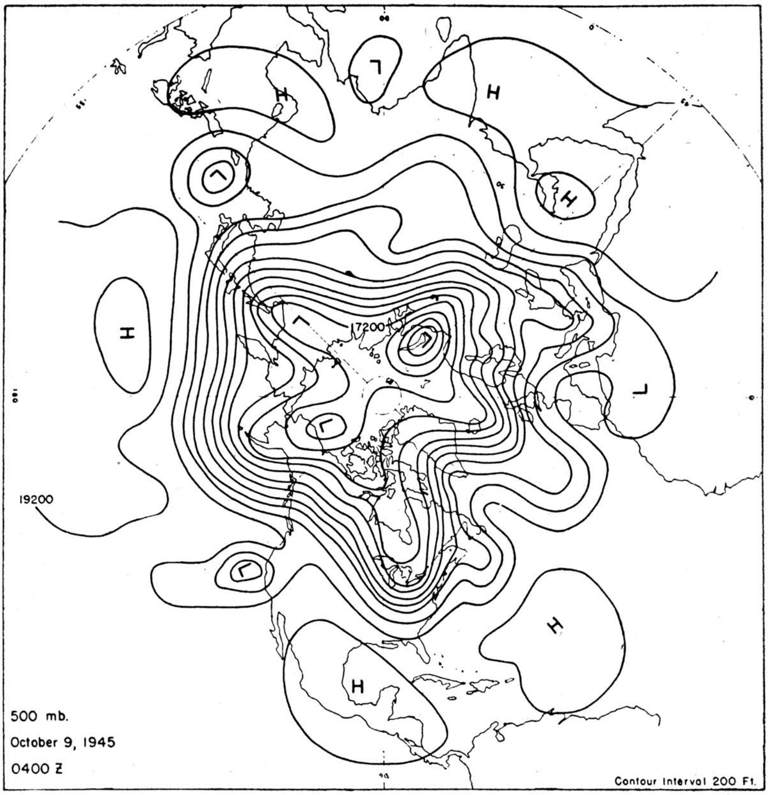

FIG. 4. 500-mb contour chart for October 9, 1945, 0400Z. Note long-wave pattern, with five major waves. Note deep trough in region of Azores.

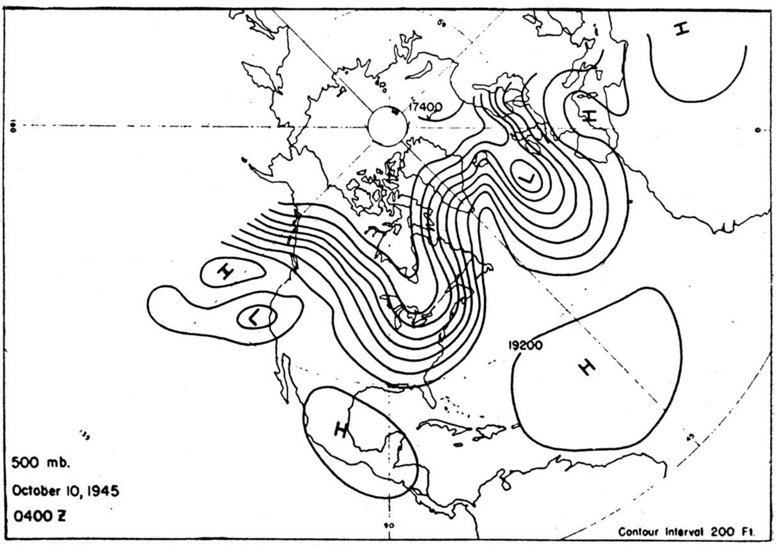

FIG. 5. 500-mb contour chart for October 10, 1945, 0400Z. Note extension northward and eastward of ridge over Atlantic.

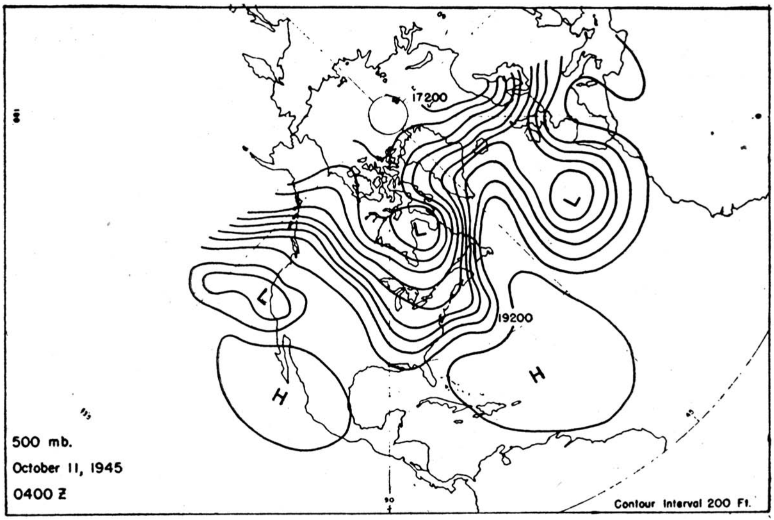

FIG. 6. 500-mb contour chart for October 11, 1945, 0400Z. Note practically completed cut-off of cyclonic vortex in Azores trough.

number. This is apparent from a comparison of FIGURE 4 with FIGURE 1, showinig a reduction in wave number from 5 to 3 between October 9 and October 12, 1945. Inspection of the Northern Hemisphere charts suggests that cutting-off is likely to occur in more than one trough (ridge) at the same time. It should be emphasized, however, that the statistical aspects of the cutting-off process have not been studied numerically, and cannot very well be so studied until several years of complete daily northern hemisphere charts for the 500-mb level are available.

In a number of cases it has been possible to anticipate the cutting-off of a cyclonic vortex by considering changes in the constant vorticity trajectory [4] of the air in the next trough upstream. This may be seen from the sketch in FIGURE 7, in which the initial wave pattern, indicated by a single streamline (full line), is assumed to be very nearly stationary. If now as a result of energy conversions in the western trough (A) the wind velocity at the inflection point P increases materially, the resulting trajectory (broken line) will acquire a longer wave-length and the eastern trough (B) will be cut off by a deep warm current to the north. To the south of the trajectory of the invading air current pressure rises will occur in response to the new wind distribution; as a result thereof a closed low pressure system may appear in the southern portion of the trough. The whole process, however, is accompanied by strong subsidence of the cold tropospheric air in the region where the cutting-off happens, probably accompanied by convergence in the invading air current. These vertical motions must produce changes in vorticity, which would further increase the separation of the vortex from the main westerly current.

It is of some interest to point out that the particular type of instability of the

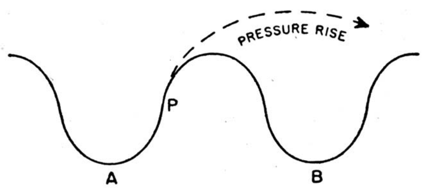

FIG. 7. Sketch of processes involved in formation of cold, cyclonic vortices. Initial long-wave pattern, indicated by single stream line, is assumed to be stationary. Intensification of current on eastern side of trough "A" leads to a trajectory of longer wave-length (dashed line). On the south side of the new trajectory the pressure rises as a result of banking.

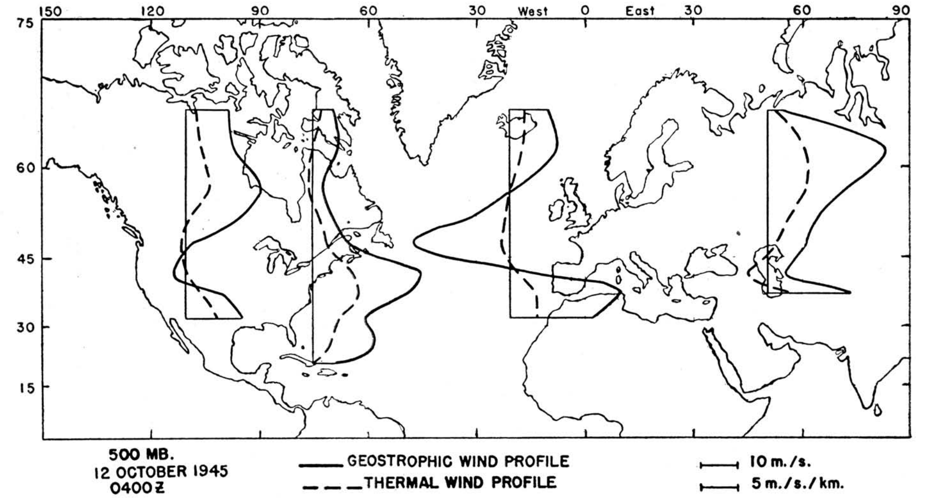

FIG. 8. Profiles of zonal geostrophic and thermal wind components in different longitudes, for the 500-mb level, on October 12, 1945, at 0400Z. Note that strongest zonal motions occur in latitudes of maximum isotherm concentration.

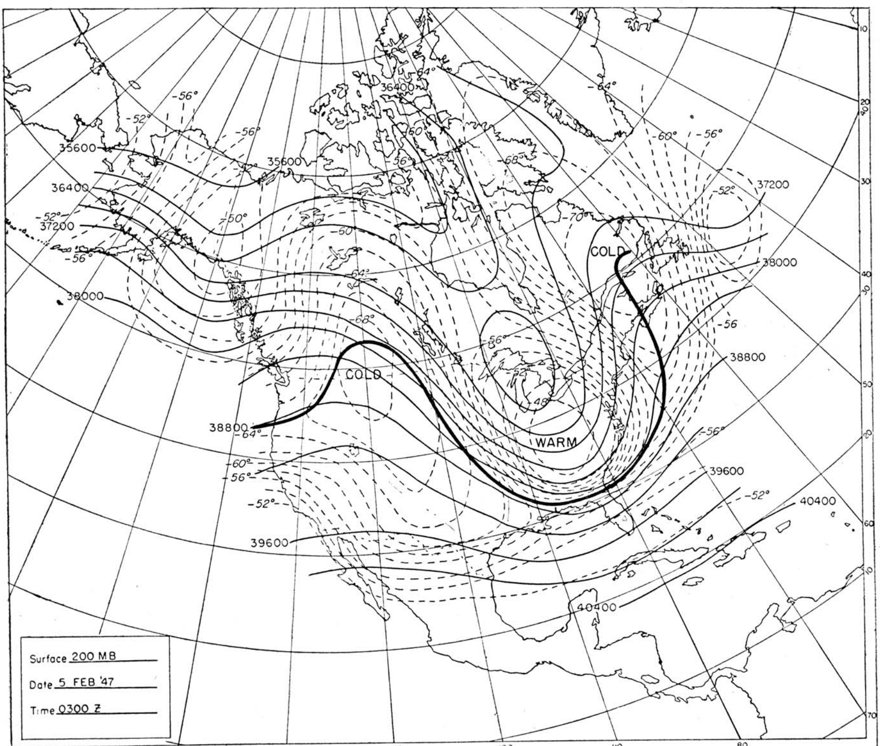

FIG. 9. Height of the tropopause and temperature at the 200-mb surface for February 5, 1947, 0300Z. The figure shows the movement of the air from a cold ridge through a warm trough and to another cold ridge. The temperature changes during this motion show that the air has a strong vertical component (descending motion from ridge to trough, ascending from trough to ridge).

west-wind belt which expresses itself in meandering and cutting-off appears to have a counterpart in the meanderings and vortex formations in certain oceanic current systems, for instance in the Gulf Stream in the region between Cape Hatteras and Nova Scotia [5].

Inspection of the temperature distribution on a large number of 500-mb charts reveals that the temperature contrast between high and low latitudes is not uniformly distributed but rather concentrated to the region of the west-wind belt. This may be brought out in a particularly striking fashion by the device of computing and plotting, for different longitudes, the zonal components of the geostrophic and thermal winds (FIG. 8). It is then seen that the maxima of these two wind profiles generally coincide in latitude. In the upper part of the troposphere, the wind maximum may be slightly south of the thermal wind maximum.

The long waves of the west-wind belt are, in the main, quasi-stationary or slowly progressive. Thus streamlines and isotherms are generally very nearly in phase and the long-wave pattern gives rise to a positive correlation between pressure and temperature in the troposphere. In the lower stratosphere, where the N–S temperature gradient is reversed, the isotherm amplitudes are considerably greater than those of the isobars (contours), suggesting, in most cases, descending motion from ridge to the next downstream trough. Occasionally this vertical motion in the stratosphere reaches so great values that one may observe the formation of stationary or very slowly moving pockets of warm stratospheric air in the troughs, pockets of cold stratospheric air in the ridges, through which the main current appears to move (FIG. 9).

In accordance with the positive tropospheric correlation between pressure and temperature resulting from the long-wave pattern, and in accordance with their manner of formation, one observes that the cyclonic vortices on the south side of the

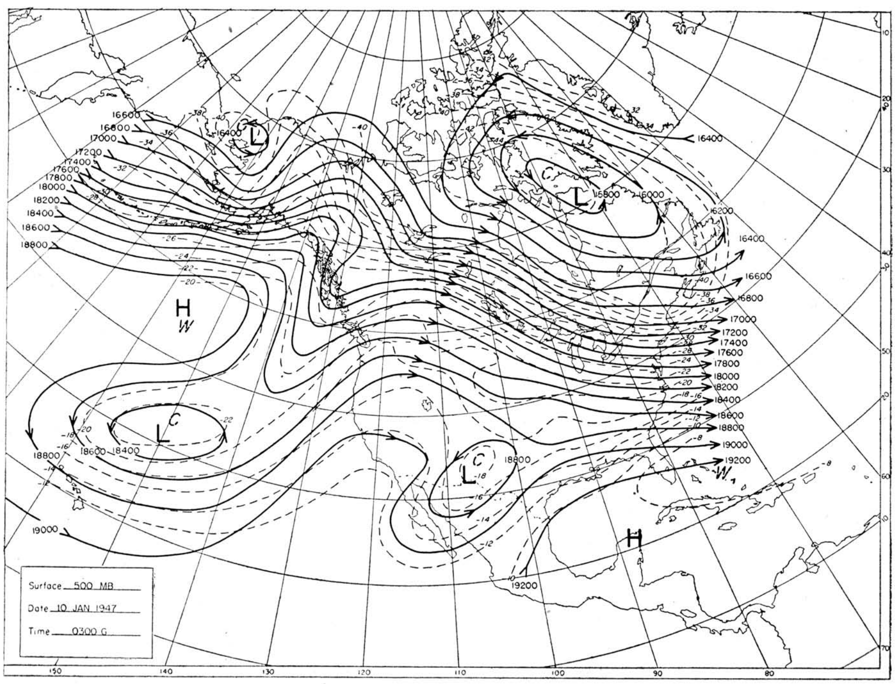

FIG. 10. 500-mb chart for January 10, 1947, 0300Z, illustrating coincidence of isotherm and wind concentration, as well as occurrence of cold cyclonic vortices on south-side of zonal current.

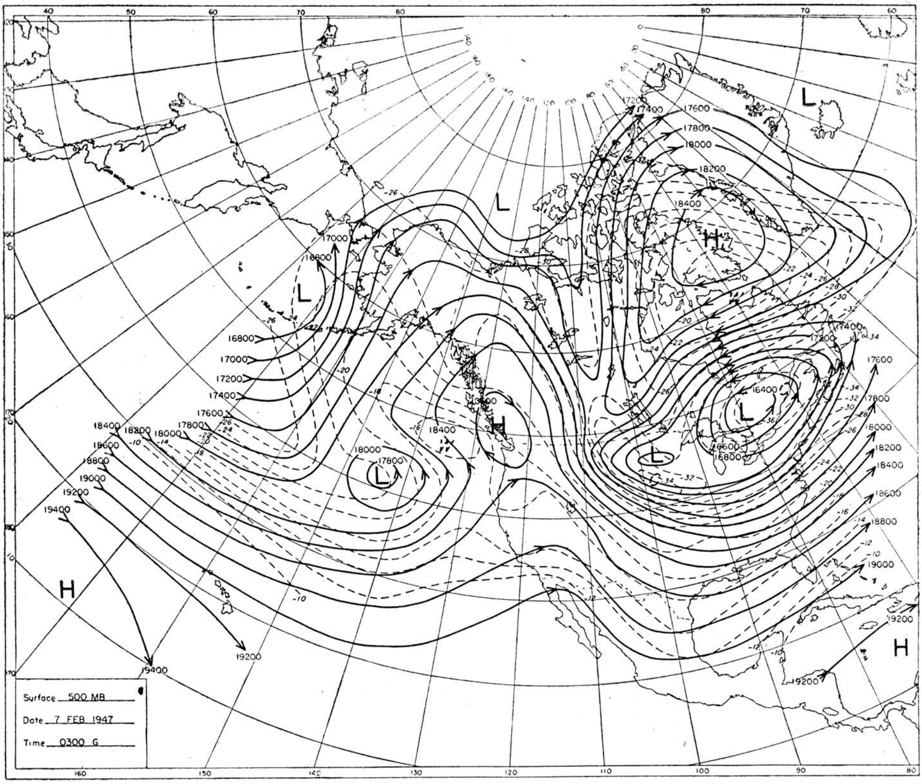

FIG. 11. 500-mb chart for February 7, 1947, 0300Z, illustrating formation of a warm anticyclonic vortex in the Pacific Northwest, through the cutting-off of a warm ridge. Note also warm high east of Greenland.

west-wind belt are cold, the anticylconic vortices on its north side warm (FIGS. 10, 11). Vertical sections through a particularly well developed, nearly symmetric cold cyclonic vortex are given in FIGURE 12 and FIGURE 13.

Vortices formed by the cutting-off process generally remain fairly stagnant for periods of several days or even a week, until absorbed by the environment. In the particular case illustrated in FIGURE 12 the vortex formed on November 2, 1946 and remained stagnant until November 6, when it slowly began to move eastward and weaken.

It seems reasonable to assume that the cutting-off process described above must contribute in a significant manner to the exchange of air between high and low latitudes. The cut-off anticyclonic vortices bring warm air north at all levels in the troposphere, and the cyclonic vortices bring cold air southward at all elevations within the troposphere. In the original polar-front theory the corresponding exchange was assumed to occur through the northward displacement of occluded wave cyclones in which the warm air had been lifted from the ground, and through the southward displacement of cold (shallow) anticyclones [6].

The low-level exchange mechanism of the polar-front waves obviously results in a net transport of relative cyclonic vorticity northward and of anticyclonic vorticity southward, whereas the exchange mechanism indicated by the 500-mb charts leads to a transport of relative anticyclonic vor-

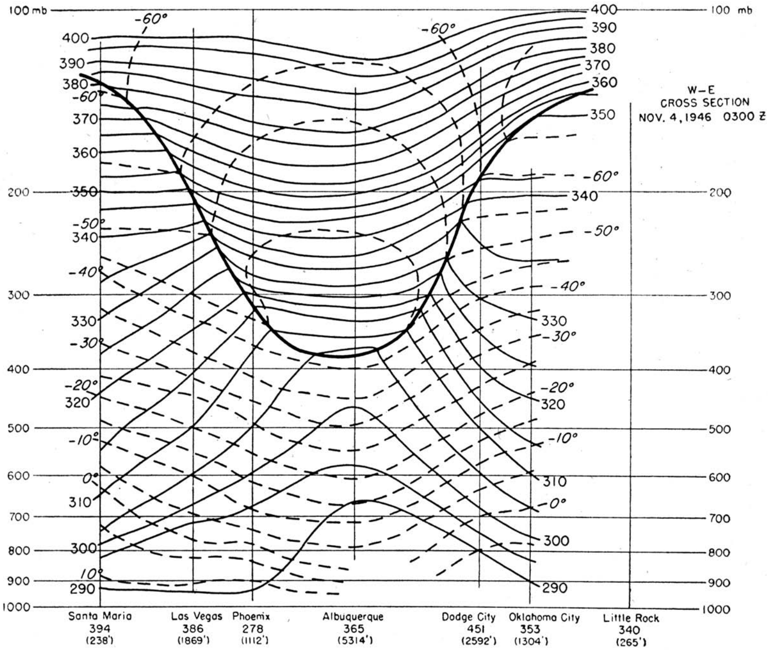

FIG. 12. W–E cross-section through the center of a vortex formed in connection with the cutting-off process. Heavy solid line, the tropopause; thin solid lines isentropes and dashed lines isotherms.

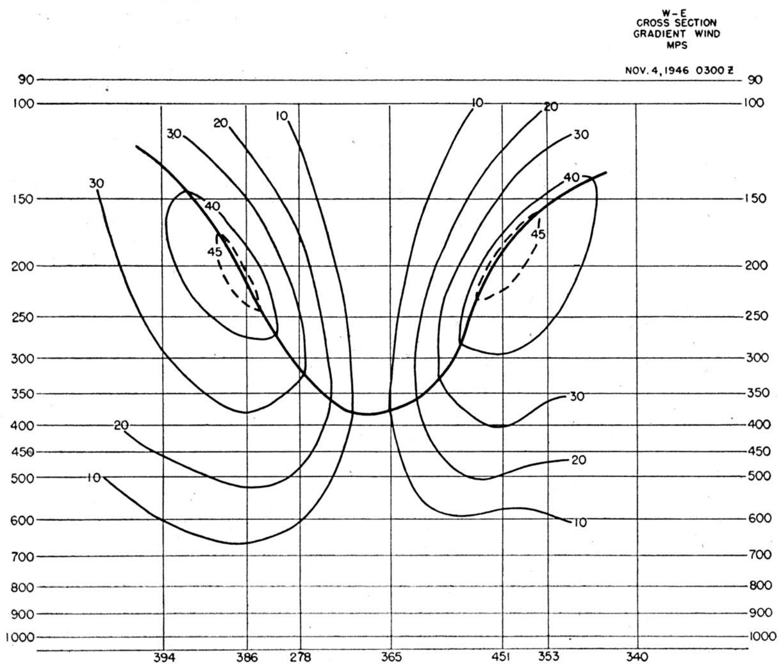

FIG. 13. Wind velocity (gradient wind) in the same cross-section as in FIG. 12. The gradient wind is calculated from the geostrophic wind assuming a radius of curvature corresponding to the distance from the center of the vortex.

ticity northward and relative cyclonic vorticity southward.

It is probable that the exchange mechanism postulated by the polar-front theory and the one observed on the 500-mb charts operate simultaneously in the atmosphere but in the upper half of the troposphere and in the vicinity of the tropopause level the second of these two exchange mechanisms seems to be the only one of major significance.[5]

The exchange mechanism of the cutting-off process may ultimately lead to the accumulation of considerable amounts of warm air at all levels (except near to the ground) in very high latitudes and associated therewith to the development of a high-reaching warm anticyclone over the Arctic. February 1947 appears to furnish a good illustration, with easterly winds and relatively high temperatures prevailing over Greenland and Northern Canada especially at high levels. The vertical cross-section in FIGURE 16 shows, in a striking fashion, the accumulation of warm air in a dynamic high-pressure area in northerly latitudes.

The west-wind belt is not of uniform speed, but varies from trough to ridge. The north-south amplitude of the particular streamline corresponding to the center of the west-wind belt sometimes appears to be greater than the amplitudes of the streamlines farther north and south. As a result, there is a marked crowding of streamlines on the south side of troughs and on the north side of ridges and, because of the quasi-stationary character of the motion, a corresponding crowding of the isotherms, far to the north in the ridges and far to the south in the troughs. The belt of maximum isotherm concentration will therefore tend to possess a somewhat

[5] Oral communications from staff members of the Canadian Meteorological Service indicate that they, through analysis of "frontal contour" charts, have reached similar conclusions regarding the mechanism of exchange between high and low latitudes.

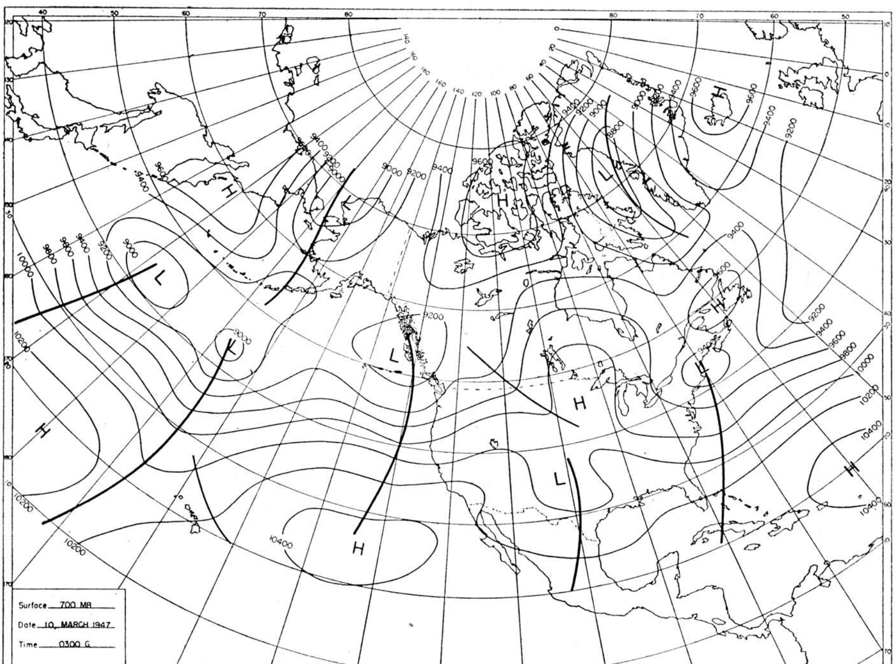

FIG. 14. 700-mb chart for March 10, 1947, 0300Z, illustrating the upper flow pattern in the early stage of a deepening in the Pacific. No well-defined major trough exists over western North America.

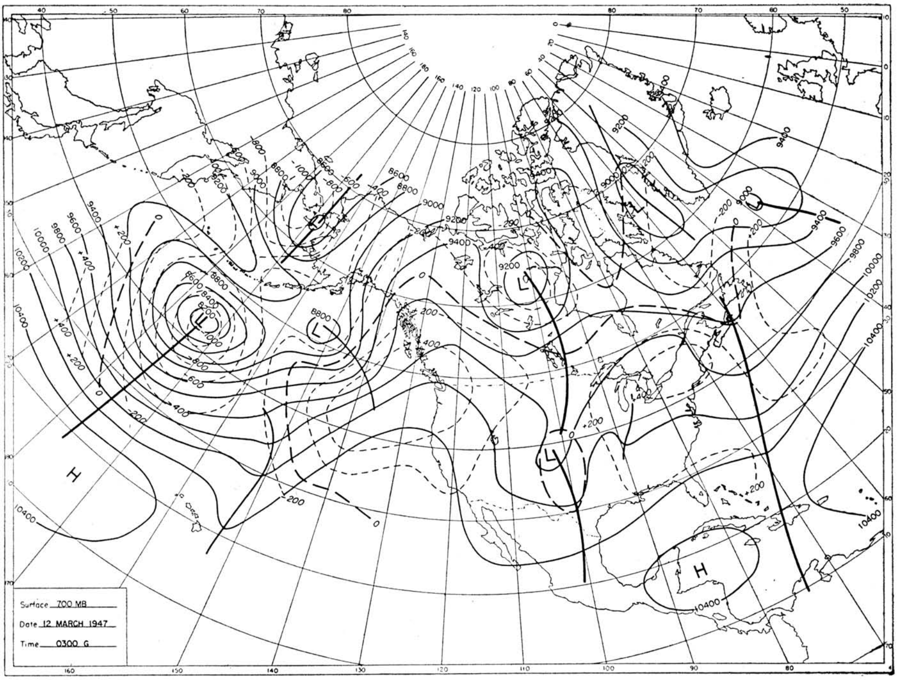

FIG. 15. 700-mb chart for March 12, 1947, 0300Z. Dashed lines give height changes, in feet, at 700 mb during 72-hour period ending at map time. Deepening in Pacific has ended; large rises on Pacific coast and in New England have resulted in trough formation in Middle West.

greater amplitude than the individual streamlines or isotherms of maximum amplitude, but vertical motions may be a contributing factor to this phenomenon.

A crowding of the streamlines implies increased wind velocity. Maximum crowding occurs on the south side of the troughs and this is also the region of maximum wind velocity. The speed of the current increases, sometimes very markedly, along the belt of maximum isotherm concentration from one ridge to the next trough down stream.

The formation of new waves (troughs) in the upper flow is most readily observed when the west-wind belt is located fairly far to the south. The deepening of upper waves may, as previously indicated, sometimes result from the deepening and occlusion of a polar-front wave cyclone. Since the instability of the frontal wave cyclone presumably depends on the presence of strong vertical wind shear, this process would seem to require the presence of a strong horizontal temperature gradient in the lower layers of the troposphere. However, this is by no means the only, or perhaps even the principal, mechanism for the formation of upper waves. New waves appear most likely to form when the wave-length of the existing waves in the westerlies is longer than the stationary wave-length corresponding to the zonal wind at about 600 mb (for a definition of stationary wave-length, see [7]. In our part of the hemisphere it is then frequently observed that an intensification of a trough in the Gulf of Alaska region is followed, first by a strong pressure rise in the Great Basin region and almost simultaneously or shortly thereafter, by a pressure rise in the Atlantic coast region. Between these two regions the new trough aloft develops; at the same time a marked concentration of

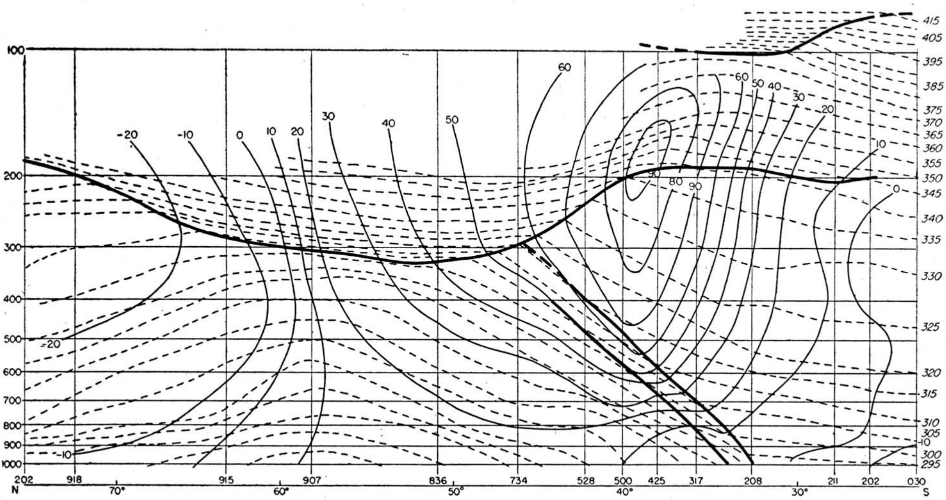

FIG. 16. Meridonal cross-section for January 17, 1947, 0300Z, from Havana in S to Thule (Greenland) in N. Heavy solid lines represent tropopause or boundaries of the frontal layer, thin solid lines constant geostrophic wind (W-wind component in mps); dashed lines represent isentropes.

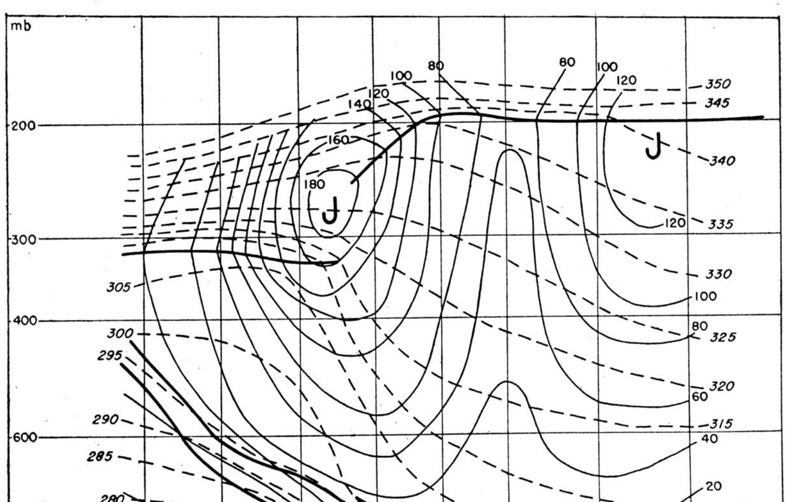

FIG. 17. Vertical cross-section, Bismarck, N. D., to Brownsville, Texas, January 29, 1947, 0300Z. Thin solid lines are lines of equal geostrophic west-wind speed (in mph), broken lines isentropes, heavy solid lines tropopauses or boundaries of frontal layer. The letter "J" (jet) indicates region of west-wind maximum. The section was taken through a current with little curvature of flow.

isotherms occurs. The time scale of the entire process is of the order of magnitude of 3–4 days (Figs. 14, 15).

This particular development, as well as the nature of the cutting-off process, suggests very strongly that *individual long waves cannot be considered as independent physical entities, but that they, on the contrary, interact with each other to produce cut-offs (vortex formations) and to generate new troughs and ridges.*

The impossibility of treating the long waves as individual closed systems is supported by the results from earlier studies. Thus it has been found that in certain cases of cyclone formation and pronounced pressure fall compensation for the mass removed did not occur within the boundaries of the synoptic system under consideration but must have occurred in a great distance [8].

In an attempt to study the west-wind belt in further detail, a number of individual vertical meridional cross-sections for the North American continent were constructed, extending in some cases from Thule in North Greenland to Havana, Cuba (Fig. 16). In most cases these sections cut through an upper level trough but it has been possible to obtain some sections through nearly straight west-wind currents (Fig. 17) and through some flat upper high-pressure ridges in the eastern part of North America. The geostrophic zonal-wind components across these sections were computed from the observed temperature-pressure distributions. These sections show, in a striking manner, the existence of an extremely sharp zonal-wind maximum, at the tropopause, and usually located between 30°N and 50°N. The wind maximum coincides with a narrow zone of extremely steep slope of the tropopause, which further north and south becomes nearly level, thus suggesting that the height of the tropopause is at least in part controlled

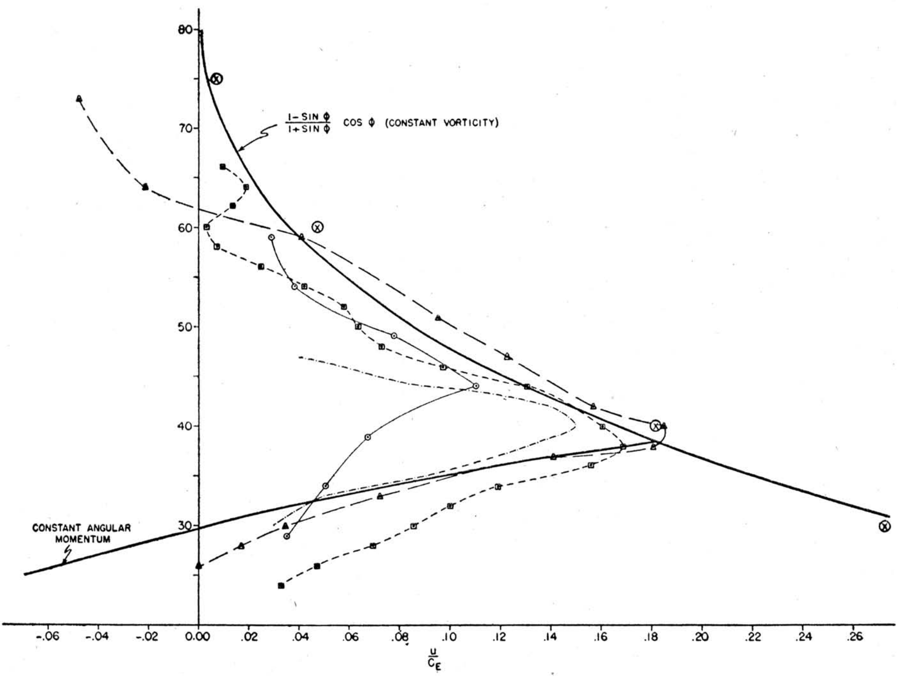

Fig. 18. Comparison between theoretical wind-profiles derived from vorticity transfer and momentum transfer considerations (heavy lines), and a few observed zonal wind profiles from the tropopause level (thin lines). The dashed line, marked by small triangles, was obtained from the 300-mb level in the vertical cross-section in Fig. 16; the full thin line represents an average over North America for nine days with fairly straight westerly flow. The non-dimensional ratio u/C_E is obtained by dividing the actual zonal wind u by the earth's linear equatorial velocity C_E. The large circles represent observed data on solar rotation, reduced to relative, non-dimensional values (see [13]).

dynamically. Below the west-wind maximum, a "frontal" zone of maximum concentration of the horizontal temperature gradient extends from the ground in the south and up to the tropopause north of the region of maximum zonal wind. In some cases the frontal layer is remarkably well developed up to the tropopause, but in other cases it becomes diffuse already in the vicinity of the 500-mb level. The frontal zone itself is very nearly isentropic, but shows mostly a slightly higher potential temperature in its upper parts. In the region of maximum zonal wind the isentropic sheets of the polar stratosphere enter the tropical troposphere, at the same time rapidly increasing in thickness in a manner which suggests centrifugal removal of air from higher to lower latitudes.

In none of the cross-sections analyzed by our unit was the absolute zonal-wind maximum found in the frontal zone itself, but always above and to the south of the upper portion of the front. In some instances a double zonal-wind belt was observed and in those cases the southern maximum appeared to be unrelated to frontal systems.

In the steeply sloping part of the tropopause the isentropic surfaces reach their maximum elevation in a manner suggestive of dynamic lifting; horizontal or isobaric charts intersecting the tropopause in the region of the zonal-wind maximum (about 300 mb) give an indication of this dynamic cooling through the appearance of a narrow ribbon of minimum temperature along the northern boundary of the zonal-wind maximum.

North and south of the zonal-wind maximum the isentropic surfaces at higher levels in the troposphere and in the stratosphere are sufficiently level so that the zonal wind distributions do not vary materially between 500 mb and 200 mb. This observation is important in that it suggests the probable permissibility, in the analysis of large-scale flow patterns, of treating the atmosphere outside the region of the maximum-wind belt as a barotropic fluid sheet. The barotropy is especially well developed north of the zonal-wind maximum.

From the cross sections discussed above, as well as from contour charts, zonal wind profiles at the 200-mb and 300-mb levels were prepared (FIG. 18). *In a number of cases the wind velocities north of the zonal-wind maximum decreased northward at a rate corresponding to a constant vertical component of the absolute vorticity. South of the zonal-wind maximum, it was frequently found that the wind would decrease at a rate corresponding to, or even exceeding, a constant absolute angular momentum per unit mass.* This distribution is of extreme importance to the interpretation of the dynamics of the west-wind belt.

At this point it is perhaps in order to emphasize that the observed concentration of the zonal motion and of the horizontal temperature gradient in a narrow zone

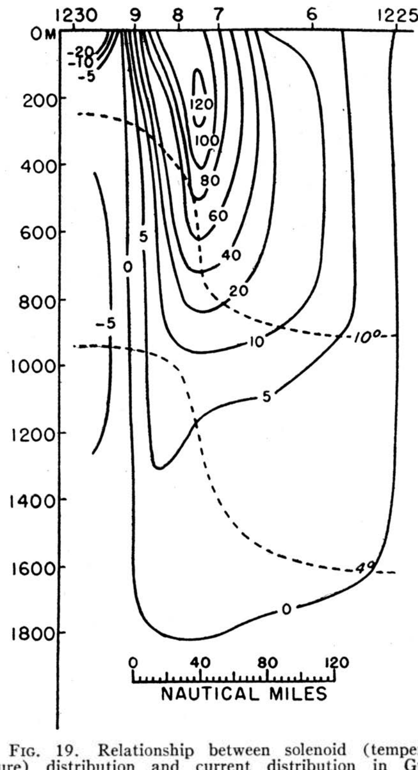

FIG. 19. Relationship between solenoid (temperature) distribution and current distribution in Gulf Stream off Chesapeake Bay. Isotherms are given by broken lines, lines of equal current velocity (in centimeters per second) by full lines. This diagram is a composite of FIG. 5 and FIG. 27 in [5].

does away with a conflict between the terms and models used in describing oceanic and atmospheric current structures. In meteorology it has been customary to work with models based on the assumption that the horizontal temperature gradient is concentrated in a narrow transition zone between two broad currents moving side by side and relative to each other. In oceanography, on the other hand, it is fairly generally accepted that the concentration of the horizontal density gradient in a narrow zone is a dynamic consequence of the motion; the entire region of measurable current velocity must therefore coincide with the region of maximum horizontal density gradient (FIG. 19). Our cross-sections show that in the upper portion of the troposphere a similar coincidence occurs between the region of maximum zonal motion and the region of maximum horizontal temperature gradient.

Our cross-sections indicate very clearly that the strength of the zonal-wind maximum and of the horizontal temperature gradient (solenoid) concentration vary in the same, rather than in the opposite, sense. It would therefore seem that the solenoids contained in vertical sections normal to the zonal wind cannot be directly responsible for the zonal motion, but that on the contrary, *the concentration of solenoids in a narrow vertical section normal to the zonal wind current must be considered as a dynamic consequence of the zonal motion itself*. The necessity for such an adjustment of the mass distribution, and hence of the pressure distribution, to the wind distribution has been stressed previously and the nature of the adjustment process has been studied theoretically [9, 10, 11, 12], but the cross-sections prepared in our analysis unit provide the first tangible synoptic evidence for the existence of such adjustment processes in the atmosphere.

II. Theoretical Comments and Interpretation

The preceding summary of impressions and conclusions gained from the upper-air charts suggests certain basic problems requiring theoretical analysis and interpretation. These problems may perhaps be listed as follows:

1. How are the upper west winds of middle latitudes maintained and why are they concentrated in a relatively narrow belt?
2. Why is it that in spite of the presumably continuous distribution of radiative heat and cold sources in the atmosphere, a large fraction of the horizontal temperature gradients appears to be concentrated in a narrow zone, approximately coincident with the upper west-wind belt?
3. What is the cause of the frontal zone below the upper west-wind belt?
4. How are new major upper-air troughs formed?

a. Zonal Wind Distribution

In a recently published article [13], one of us attempted to show that some of the more striking features of the zonal motion in the atmosphere might be accounted for if one assumes that its statistical distribution with latitude is controlled primarily by large-scale lateral mixing processes. The nature of the redistribution of zonal motion brought about by lateral mixing would depend upon the character of the mixing process itself. Particularly in high latitudes, the inertia stability of the atmosphere is certainly great enough to prevent the development of strong meridional circulations around axes parallel to the latitude circles and accompanying exchange of absolute angular momentum. There is no equivalent resistance against mixing through eddies with vertical axes and, as pointed out earlier, the exchange of air between high and low latitudes appears in fact to occur through such eddies. It was brought out in [13], that lateral exchange of this type would tend to produce an equalization of absolute vorticity, thus transporting warm air and anticyclonic relative vorticity northward, cold air and cyclonic relative vorticity southward. One is thus led to the conclusion that continued lateral mixing would result in the establishment of a zonal wind profile characterized by a constant vertical absolute vorticity.

It was shown in [13] that a polar air cap which as a result of lateral stirring had acquired a constant absolute vertical vorticity, would possess an absolute zonal angular velocity $\dot{\lambda}$, given by

$$\dot{\lambda} = \frac{2\omega_p}{1 + \sin\phi}, \tag{1}$$

where ω_p is the absolute angular velocity of the cap at the pole and ϕ is the latitude. It follows that the relative linear velocity (u) of such a cap would be

$$u = a\Omega \frac{\dfrac{2\omega_p - \Omega}{\Omega} - \sin\phi}{1 + \sin\phi} \cos\phi, \tag{2}$$

a being the radius of the earth and Ω its angular velocity. The earth's linear equatorial velocity (C_E) is given by

$$C_E = a\Omega, \tag{3}$$

and has a numerical value of 1,037 miles per hour. Thus a convenient non-dimensional measure for the zonal wind distribution is provided by the ratio

$$\frac{u}{C_E} = \frac{\dfrac{2\omega_p - \Omega}{\Omega} - \sin\phi}{1 + \sin\phi} \cos\phi. \tag{4}$$

Since the absolute vertical vorticity ζ of the polar cap has the value

$$\zeta = 2\omega_p, \tag{5}$$

it follows that for $\zeta = 2\omega_p < 2\Omega$ a belt of easterly winds must appear in high latitudes. It is not likely that such a polar cap would acquire a higher average absolute vorticity than the maximum vertical vorticity of the earth itself, namely 2Ω. Hence one might perhaps be justified in considering the particular profile for which $\zeta = 2\omega_p = 2\Omega$, namely

$$u/C_E = \frac{1 - \sin\phi}{1 + \sin\phi} \cos\phi \tag{6}$$

as a limiting profile of maximum vorticity.

b. *Tropopause Height as Function of Latitude*

It is clear that (4) and (6) both must lead to increasing westerly winds with decreasing latitude. The zonal winds given by these two expressions must be compatible with the thermal wind equation. We know that near the ground, friction tends to reduce both the actual and the geostrophic winds to values which are low compared to those prevailing at the tropopause level. Because of this restraint on the zonal wind at sea level, it follows that variations in the zonal wind at the tropopause, with latitude or with time, must be associated with changes in the north-south temperature gradient, or with changes in the height of the tropopause, or with both. This interrelationship is easily seen if one writes the zonal thermal-wind equation in the form:

$$\frac{u_H - u_0}{H} = \frac{g\bar{\gamma}}{f\bar{T}}, \tag{7}$$

where u_H and u_0 are the zonal winds at the tropopause height (H) and at sea level, g the acceleration of gravity, f the Coriolis parameter, $\bar{T}$ the absolute mean temperature of the troposphere and $\bar{\gamma}$ the mean tropospheric temperature gradient northward at the particular latitude under consideration. As one proceeds southward towards the maximum west-wind belt u_H increases more and more rapidly and hence $\bar{\gamma}$, or H, or both, must increase sharply towards the south. Inspection of vertical N–S cross sections prepared for selected situations during the winter 1946–1947 shows that both $\bar{\gamma}$ and H increase rapidly as one approaches the west-wind maximum from the polar side; it will be shown below that this fact plays an important role in the sudden break-down of the constant vorticity wind profiles (4) or (6) in lower latitudes.

c. *Break-down of the Constant Vorticity Profile*

In the analysis of the zonal motion presented in [13], it was assumed that the constant vorticity profile under favorable circumstances (intense lateral mixing) might be extended southward as far as latitude 35° or 30°, where the profile necessarily would become unstable because of excessive shear. South of this critical latitude the zonal wind profile would follow a law corresponding to a steady flux of vorticity southward. In the transition zone between those two regions vorticity would have to be replenished by organized ascending motion.

Our synoptic studies, particularly the analysis of N–S vertical cross-sections, have shown that the region immediately below and to the north of the zonal-wind maximum is characterized by such ascending motion, but it has also been established that the previous analysis must be modified to take into account the presence, frequently observed, of a narrow zone of marked inertia instability[6] immediately south of the zonal-wind maximum.

The establishment of such zones of inertia instability may perhaps be understood from the following considerations:

In a straight current system parallel to the latitude circles m, the absolute angular momentum per unit mass, is given by the expression

$$m = a \cos \phi [u + a\Omega \cos \phi]. \tag{8}$$

For isentropic, straight zonal motion, inertia instability is reached when, in an isentropic surface, the angular momentum m is so distributed that it increases towards the axis of the earth, i.e., towards the north, and thus the criterion is, for all practical purposes,

$$\left(\frac{\partial m}{\partial \phi}\right)_\theta > 0, \tag{9}$$

the subscript θ indicating that the differentiation must be carried out in a constant potential-temperature surface. One finds, from a combination of (8) and (9):

$$\left(\frac{\partial u}{a\partial \phi}\right)_\theta > f + \frac{u \tan \phi}{a} = f + \dot{\lambda}_r \sin \phi, \tag{10}$$

$\dot{\lambda}_r$ being the relative angular velocity of the air. In this expression the second term on the right hand side is, even for the high wind velocities encountered in our work, not likely to reach a value much in excess of $0.1f$ and may therefore, in the first approximation, be neglected. Substituting a linear coordinate y, counted positive northward, one then finds as a criterion for inertia instability of straight zonal currents:

$$\left(\frac{\partial u}{\partial y}\right)_\theta > f.^7 \tag{11}$$

Anticyclonic shears of this intensity are readily established as a result of lateral mixing. The process is not unlike the action of small-scale mechanical turbulence on a limited air layer of initially stable vertical temperature gradient near the ground. As a result of the stirring the layer assumes a constant potential temperature corresponding approximately to the arithmetic mean, with respect to mass, of the potential temperatures prevailing in the layer before the stirring. Next to the ground a temperature increase is observed, in the upper part of the mixed layer a temperature drop, and the mixed layer is finally separated from higher strata by a turbulence inversion.

A similar phenomenon must occur in a thin polar cap which is being subjected to strong lateral mixing between the pole and a prescribed latitude ϕ_0. The internal lateral mixing will not affect the total absolute angular momentum (M) of the cap and, hence,

$$M = 2\pi a^4 \int_{\phi_0}^{\pi/2} \cos^3 \phi \dot{\lambda} d\phi \tag{12}$$

must remain constant throughout the mixing process. In this expression it is assumed that the mass of the cap per unit area has the value 1. After mixing, $\dot{\lambda}$ has the value given in (1) and one finds

$$M = 2\pi a^4 \omega_p (1 - \sin \phi_0)^2 \tag{13}$$

from which the value of ω_p in (1) may be determined. The final value of $\dot{\lambda}$ at latitude ϕ_0,

$$\dot{\lambda}_{\phi_0} = \frac{2\omega_p}{1 + \sin \phi_0}, \tag{14}$$

will in all normally prevailing cases be higher

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[6] The development of a criterion for inertia instability applicable to atmospheric phenomena is due to Bjerknes and Solberg [14].

[7] It is perhaps worth emphasizing that inertia instability can appear in straight currents of arbitrary direction, and not only in straight west-wind belts. (See also a discussion of this topic in [10].) To see this, one may compute the absolute angular momentum per unit mass of a symmetric atmospheric vortex with a vertical axis. One finds:

$$m = Vr = r\left(v + \frac{fr}{2}\right),$$

where v is the linear relative velocity (positive for cyclonic motion), and r the horizontal radial distance from the axis of the vortex. Inertia instability requires that

$$\frac{1}{r}\frac{\partial Vr}{\partial r} < 0,$$

or

$$-\frac{\partial V}{\partial r} > \frac{v}{r} + f = \frac{v_g}{v} f,$$

where v_g is the geostrophic wind. For large values of r, the criterion reduces to

$$\frac{\partial v}{\partial n} > f,$$

where n now is a horizontal normal pointing to the left of the downwind direction, which is arbitrary.

than the undisturbed value of $\dot{\lambda}$ in the zone south of the cap. Thus a sharp drop in the zonal wind will be observed southward across the boundary of the polar cap. In the particular case in which the initial zonal motion was characterized by a constant absolute angular velocity $\dot{\lambda}_i$; one finds, from an evaluation of (12) for the initial state, that M has the value

$$(15) \quad M = 2\pi a^4 \dot{\lambda}_i (1 - \sin \phi_i)^2 \frac{2 + \sin \phi_0}{3}.$$

Through substitution of (14) in (13) one obtains

$$(16) \quad M = \pi a^4 \dot{\lambda}_{\phi_0} (1 + \sin \phi_0)(1 - \sin \phi_0)^2.$$

It follows, from a comparison of (15) and (16), that the drop in angular velocity $\Delta\dot{\lambda}$ at latitude ϕ_0 may be computed from

$$(17) \quad \Delta\dot{\lambda} = \dot{\lambda}_{\phi_0} - \dot{\lambda}_i = \tfrac{1}{3}\dot{\lambda}_i \frac{1 - \sin \phi_0}{1 + \sin \phi_0}.$$

Since $\dot{\lambda}_i$ in the first approximation corresponds to the earth's own angular velocity Ω, it is evident that the discontinuity in angular velocity has an approximate value of

$$(18) \qquad \Delta\dot{\lambda} = \frac{\Omega}{3} \cdot \frac{1 - \sin \phi_0}{1 + \sin \phi_0}.$$

The corresponding discontinuity in $\Delta \dfrac{u}{C_E}$ has the value

$$(19) \qquad \Delta \frac{u}{C_E} = \frac{1}{3} \cdot \frac{1 - \sin \phi_0}{1 + \sin \phi_0} \cos \phi_0,$$

corresponding to a value of almost 100 mph at latitude 30°N, with smaller values for mixing zones terminating farther north. A graphical, theoretical comparison of zonal wind profiles before and after mixing is given in FIGURE 20.

As a result of inertia instability, the wind discontinuity at the southern boundary of the polar cap must spread out over a zone of finite width and with a shear corresponding to constant absolute angular momentum per unit mass. The width of this zone is obtained by dividing Δu by the critical shear f. One obtains, for latitude 30°N, a width of about 615 km or about $5\frac{1}{2}$° of latitude, in fair agreement with observations (see FIG. 18).

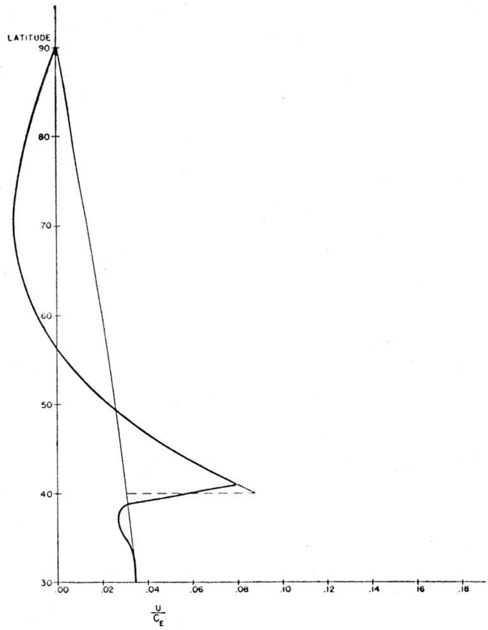

FIG. 20. Theoretical diagram, illustrating the effect of lateral mixing within a limited polar cap on the zonal wind distribution. The initial wind distribution, corresponding to constant angular velocity, is given by the thin line; the final wind distribution by the full line. The discontinuity at latitude 40° is presumably spread out over a zone of finite width as a result of meridional circulation and momentum transfer.

In the layers well below the tropopause wind maximum, the horizontal wind shears are weaker than at higher levels, but in these regions the isentropic surfaces may at times acquire so steep a slope that the isentropic anticyclonic shear reaches, or even exceeds, the critical value given in (11). For the case of vanishing horizontal shear, this happens when the slope of the isentropic surfaces reaches the value $f/\sqrt{gs}$, in which expression s represents the vertical stability. For latitude 43°N ($f = 10^{-4}$ sec^{-1}) and a typical value of $s = 10^{-7}$ cm^{-1}, the critical slope has a value of 10^{-2}, a value not infrequently attained in the region below a well-marked zonal-wind maximum.

The preceding interpretation of the inertia instability observed to the south of and below the zonal-wind maximum, was based on the assumption that the lateral mixing process is restricted to a polar cap, sharply limited equatorwards. It is evident, however, that even without such a restriction, the gradual extension southward, through mixing, of a polar cap of constant vorticity must lead to the development of an increasingly strong tropopause west-wind belt in middle latitudes and to increased tilting of the underlying isentropic surfaces along the southern boundary of the zone of maximum mixing. When the critical value of inertia instability has been reached, meridional circulations in vertical planes must develop and interrupt the lateral exchange of vorticity. *Thus it would appear that the lateral mixing process necessarily must create its own sharp forward boundary, at which the vorticity transfer is interrupted*, much in the same way as mechanical turbulence in a vertically stable atmosphere, by creating a temperature inversion, creates a sharp upper boundary limiting further penetration of turbulence upward.

d. Meridional Circulation Pattern

The inertia instability observed on the south side of the west-wind belt is, according to our analysis, a dynamic consequence of two separate processes, first, the establishment, through lateral mixing, of a constant vorticity profile at the tropopause level, and secondly, the deformation of the isentropic surfaces well within the troposphere to meet the thermal-wind requirement. As the west-wind belt is being built up, the inertia instability will presumably first arise at the tropopause level; as a result, at least a part of the air in the maximum west-wind belt will be flung southward. Compensating northward displacements will occur in the lower isentropic surfaces, in which vertical tilting associated with the adjustment process may have led to the establishment of isentropic instability. One obtains in that manner a reverse solenoidal circulation with sinking, southward motion in the upper troposphere to the south of the west-wind belt and ascending northward motion below the west-wind belt in the region of maximum concentration of the isentropic surfaces. The existence of such a reverse circulation is made more plausible by the extreme dryness of the tropospheric air south of the west-wind belt and by the frequent occurrence of heavy precipitation in the frontal zone below the west-wind maxi-

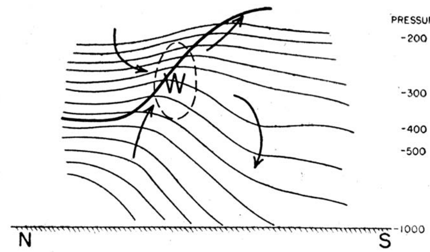

FIG. 21. Probable meridional displacements and adjustments associated with intensification of zonal wind maximum. Isentropes given by thin lines, tropopause by heavy line.

mum. These reverse circulations are illustrated in the sketch in FIGURE 21.

The reverse circulation accompanying the early stages of the building up of a new zonal-wind maximum is most likely in the nature of a continuous adjustment process. The general nature of such adjustments in terrestrial current systems was discussed by one of us in a paper published several years ago [11] and may be seen from the sketch in FIGURE 22.

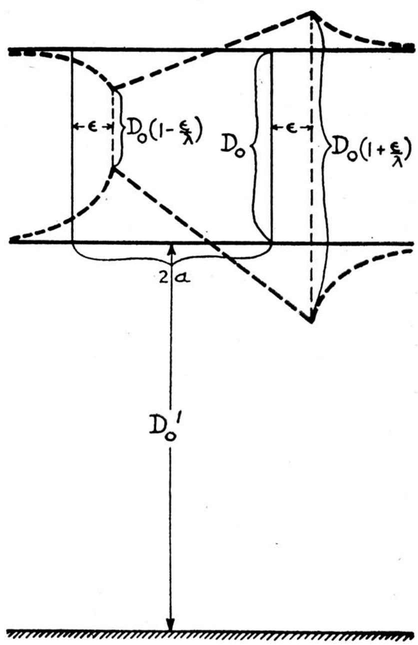

FIG. 22. Adjustment of mass distribution to a mechanically established limited current in a baroclinc system. Note in particular deformation of internal boundary (lower broken line) and compare with isentropes in FIGS. 16 and 21. From [11].

There is an obvious similarity between the deformations indicated in that sketch and the observed deformations of the isentropic surfaces in FIGURES 16 and 17.

These adjustment processes are likely to change their character in an abrupt manner as soon as inertia instability is established. In that moment, violent southward displacements aloft and up-slope northward displacements near the ground become possible and might lead to a sudden activation of a frontal zone in lower levels.

In a gravitationally stable atmosphere the reverse circulations described above cannot result in closed orbits in the meridional planes, but merely in readjustments of the mass distribution. In a gravitationally unstable atmosphere, on the other hand, closed orbits would result and, as a consequence, true vortex rings would be formed along the equatorial boundary of the polar cap of constant absolute vertical vorticity. In this connection it may be of some interest to call attention to the fact that V. Bjerknes [15] many years ago postulated the existence of such vortex rings, to account for certain characteristic properties of solar activity.

e. Frontogenesis

The processes described and analyzed above suggest an explanation for the fact that it frequently appears possible to construct, from the now available upper air data over the continent of North America, a practically continuous "frontal" boundary between "polar" and "tropical" air masses.[8] The lateral exchange processes within the polar cap must reduce the horizontal temperature gradients at the same time as they reduce the horizontal gradient of absolute vertical vorticity, and hence they must gradually build up a steep horizontal temperature gradient along the southern boundary of the well-stirred polar cap. Furthermore, coincident with the establishment of a strong zonal-wind maximum along the southern boundary of the polar cap, transverse indirect solenoidal circulations are impressed upon the atmosphere, to bring about the necessary adjustment of the pressure and mass distribution to the wind field; these circulations must lead to further solenoid concentrations. This two-fold process would seem to account for the frequent occurrence of a well-marked frontal zone, the product of intense lateral mixing to the north, and of a strong solenoidal field in the warm air south of the front, resulting from the transverse indirect vertical circulation. This latter circulation may occasionally bring about an increased tilt of the frontal zone itself.

Our studies have shown that not infrequently a marked zonal-wind maximum and a sharp frontal zone extending practically throughout the entire troposphere above the 700-mb level may appear in our latitudes, as early in the winter half-year as November, or even in October. This fact, plus the further fact that the upper portion of the frontal zone appears to be steep and well developed even when it is located far to the south, suggest that at higher levels the boundary between cold and warm air at least partly is created *in situ*, through concentration of available temperature contrasts, rather than through southward advection of cold air from very high latitudes.

Once established, such a boundary must, for short periods of time, behave as a material surface, particularly in the lower layers, but over longer time intervals a changing zonal motion may build up a new frontal zone or solenoid concentration in a different portion of the atmosphere.[9] The mechanism of frontogenesis presented above would seem to provide a necessary supplement to the kinematic mechanism suggested by Bergeron [6] and studied by Petterssen [16].

f. Long-Wave Pattern and Isotherm Concentration

The concentration of solenoids below the maximum west-wind belt is often intensified by the long-wave pattern. The upper-level charts have shown that in these wave patterns the streamline corresponding to the maximum wind sometimes has a much greater amplitude than the surrounding streamlines corresponding to the weaker winds further north and

[8] This is very clear at the 500-mb level, where there still is a great temperature contrast between the polar and the tropical atmosphere. Higher up, e.g., at the 300-mb level, the pre-existing meridional temperature distribution is such that the frontal boundary cannot be studied by means of the temperature field alone.

[9] The vertical cross-section for January 29, 1947 (FIG. 17) contains indications of such transfer of the maximum solenoid concentration to a level above the original frontal zone.

south. This result might have been anticipated from the fact that the wave-length of a constant vorticity trajectory increases with wind speed unless at the same time the amplitude is increased in a compensating manner. Thus if all the streamlines of a limited west-wind belt are to have the same wave-length, the inner streamlines must possess a larger amplitude than the surrounding streamlines.

The difference in amplitude described above gives rise to an intense temperature and streamline concentration on the south side of major troughs, leading to a further sharpening of the maximum west-wind belt, increased inertia instability and increased tilting of the isentropic surfaces. A similar, although perhaps somewhat weaker, concentration appears to occur in the northern part of major ridges.

It would thus seem that the energy concentration takes place in two stages, first through the building up of a principal west-wind belt with its associated solenoid concentration, secondly through the long-wave pattern which results in further energy concentration in the southern portion of troughs and in the northern portion of ridges. *In both instances vertical motions are impressed upon the atmosphere by the field of horizontal motion.*

g. Energy Dispersion and Formation of New Troughs

The long-wave pattern does not merely result in a concentration of energy but also in energy dispersion [17], and thus in the establishment of new long waves in regions previously characterized by straight motion. FIGURE 23 contains the results of a theoretical analysis of the effect of the establishment of a new trough, on a previously straight west-wind current downwind from the trough. The calculation is based on the assumption of barotropic, non-divergent motion and can thus serve merely as a crude guide to the exceedingly complex character of this process. It is assumed that beginning at a time $t = 0$, cyclonic relative vorticity is injected, at a constant rate, into a uniform west-wind belt of the speed u, at a prescribed longitude ($x = 0$). Downstream from this point a transient develops, in the region $x - ut > 0$, but for values $x - ut < 0$ a steady-state wave pattern prevails with an amplitude determined by the value of the relative vorticity which is steadily being injected at $x = 0$. This theoretical development should be compared with the charts in FIGURE 14 and FIGURE 15. The computed sequence of events, namely the es-

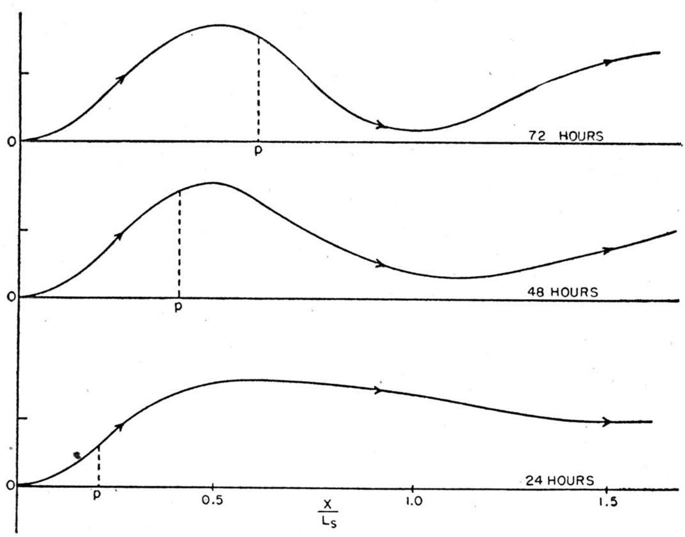

FIG. 23. Theoretical diagram illustrating the establishment of a new wave in a straight zonal current as a result of the injection of cyclonic vorticity at a prescribed longitude. Streamlines are drawn for three consecutive days, following the initial injection of vorticity.

tablishment of a ridge immediately east of the vorticity injection point, followed or accompanied by the development of a second, somewhat weaker ridge further east, between which two ridges a new trough develops, has been observed repeatedly.

III. Experimental Studies

In view of the importance attributed in our analysis to the processes of lateral mixing associated with the heat exchange between high and low latitudes, it seems desirable to study experimentally the departures from solid rotation that would be produced in a rotating hemispherical fluid shell, subjected to thermally produced lateral mixing. Accordingly, an apparatus was constructed which consisted of two concentric glass hemispheres, mounted in a rigid framework which could be rotated at speeds varying between 5 and 60 revolutions per minute. The space between the two shells was filled with water and thermal lateral turbulence was introduced through heating of the fluid from below (at the pole). The thickness of the fluid shell was 1.6 cm and its mean radius 10 cm. The ratio between these two numbers is obvi-

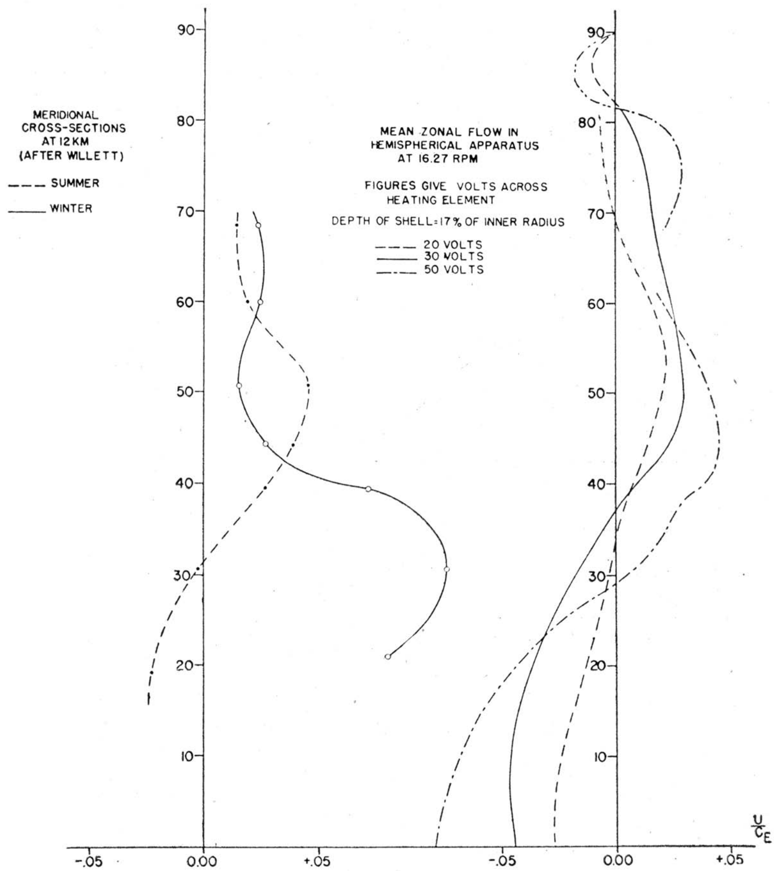

Fig. 24. Graphical representation of zonal relative motion produced by thermal turbulence in a rotating hemispherical fluid shell for different rates of heating. Left hand side represents relative zonal motion at the tropopause level in the earth's atmosphere.

ously much larger than the corresponding ratio in the earth's atmosphere (20/6400); it may therefore be stated that our experimental set-up in some measure tended to favor the development of axially symmetric meridional circulations with momentum transfer rather than vorticity transfer. The experimental work is continuing, but the following preliminary results may be stated with some degree of assurance:

(*a*) Thermal lateral turbulence produces systematic departures from solid rotation.

(*b*) The order of magnitude of these departures (expressed as fractions of the linear equatorial motion of the solid framework) agrees with the values given by Willett [18] for the zonal wind profiles at the tropopause level, and the general character of the distribution of these departures with latitude is not unlike the distribution with latitude of the zonal motion in the atmosphere.

(*c*) Increased heating seems to produce a displacement of the belt of maximum relative zonal motion equatorwards, but in our experiments it has not yet been possible to force this maximum belt as far equatorwards as in the earth's atmosphere. A simple graphical representation of these results and a comparison with Willett's data are given in FIGURE 24.

The conclusions and analyses presented above concern themselves with the structure of the zonal wind distribution and with the processes involved in the concentration and dispersion of energy in the atmosphere. Little progress has been made with the basic problem of ascertaining the manner in which the ultimate source of atmospheric energy, solar heat, is converted into motion.

The lack of evidence for the existence of strongly developed direct meridional circulations suggests that effective conversion of solar energy into motion through solenoidal action does not take place except in flow patterns containing well-developed northerly and southerly current branches and hence also solenoids in vertical W–E planes. If this suggestion should be borne out by further studies, it would in some measure be justifiable to state that the zonal motion may be considered as the end product of the large-scale lateral motions, rather than as a basic current, maintained through direct meridional circulations and serving as the energy source for the large-scale north and south currents. This viewpoint appears to be supported by our experimental work.

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The authors of this report have profited greatly from a study of the bold attack on astrophysical hydrodynamics contained in a recent monumental study by Wasiutynski [19]. Our emphasis on the importance of large-scale lateral mixing processes has received considerable support from the recent estimates by Tuominen [20] of the value of the lateral mixing length in the surface layers of the sun, which would seem to point to a marked anisotropy of solar turbulence.

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European Relief Needs Still Urgent

An appeal has been received from the Committee of Personnel of the Central Bureau for Meteorology and Geodynamics, Hohe Warte 38, Vienna XIX, Austria, on behalf of the 46 members of the staff of the Bureau. They and their families are virtually starving. Thanks to a gift of $20 just received, 2 *CARE* packages were immediately ordered sent. But large though these are, they can provide supplementary rations of only 300 calories a day for 46 people for less than a week. It is time to order two more, and, soon after, two more still. Our relief fund needs for the Austrians alone some $20 per week. But we must continue to help the still larger staff of the Hungarian weather bureau, and to send an occasional package to individuals.

Following my report at the Washington meeting that the total collected since the appearance of my appeal in the Dec. BULLETIN had been only about $100 ($115), Professor Rossby made a moving appeal to the meeting for further contributions. Several members of the Society are making substantial contributions direct, and several of the local branches have contributed food or clothing. So our Society's contribution to the relief of meteorologists abroad is not to be measured by what has come into the relief fund. Nevertheless, the contributions direct or through the relief fund cannot be too generous in the face of the enormous need for clothing and food that is still to be met.

Of the $135 spent so far, about 75% went for food, 17% for clothing, including postage, and 8% for books. There is $5 on hand waiting for another $5 to cover a CARE package.

The contributions received since December have averaged a nickel per member. Let's try to raise this. How about $1, or what have you, from each member who has not yet contributed to the relief fund or in some other way? Send your contributions to the Am. Meteorological Society Relief Fund, 5 Joy St., Boston 8, Mass.—*C. F. B.*

International Geographic Congress, Lisbon, Sept., 1948

At the next International Geographical Congress, to be held at Lisbon the last half of Sept., 1948, among the questions to be discussed will be: "The seasons of the year in extra-tropical climates: their definition, their limits and their characteristic elements," and "Study of climatic variations." The Secretariat of the Congress is c/o Centro de Estudos Geograficos, Praça do Rio de Janeiro, 14, Lisbon, Portugal.