A STUDY OF THE EFFECTS OF LARGE BLOCKING HIGHS ON THE GENERAL CIRCULATION IN THE NORTHERN-HEMISPHERE WESTERLIES

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ABSTRACT

The basic problems involved in the investigation of the effects of "blocking action" upon the circulation pattern of the northern-hemisphere westerlies are defined and the mode of attack outlined. By application of turbulence principles seasonal variations in lateral mixing and northward heat transport across the westerlies are related to variations in the extent of blocking action. Evidence is presented indicating that blocking action, lateral mixing, and northward heat transport are all greater in the winters of years when strong north–south thermal gradients exist and *vice versa.* The year-to-year variations in these quantities exhibit a long-period and relatively large amplitude climatic swing with a period of roughly 12–14 years. Considerable year-to-year variation in the radiative properties of the earth's surface and atmosphere, or in incoming radiation, is thus implied.

An analysis of heat convergence through the lateral turbulence process shows in general the heart of the westerlies losing heat, particularly from the east coasts of the continents to the Aleutian and Icelandic Lows. This mechanism is apparently the means by which the large amount of heat picked up by the air from the oceans off the east coasts is dispersed.

A distinction is made between the turbulent eddy sizes identified with unstable, migratory systems and those of blocking action, the latter representing dynamically quasistable systems capable of producing effects upon the general circulation of the westerlies at great distances both upstream and downstream and over a period of many days.

Effects resulting from the intrusion of stable, persistent high-pressure cells into the normal westerly flow of the general circulation of the atmosphere in the characteristic manner of blocking action are investigated by means of empirical, semi-statistical procedures. These processes are found to cause the development of long stable wave patterns in the atmosphere downstream from the point of inception. These stationary patterns may occasionally exist for periods of as long as a month. The wave lengths are found to be longer than those generally considered representative of stationary wave patterns. Presumably this difference is due to the distorting effect of the larger amplitude of the long waves associated with the blocking action process.

A theory of the formation of these blocking high cells is suggested based on the accumulation of heat in low latitudes and the necessity for the readjustment of the general circulation to redistribute this heat. One means of dissipation of the blocking cells is shown to be the formation of a wave pattern in the atmosphere which is out of phase with the stable wave pattern formed by the blocking high.

CONTENTS

INTRODUCTION

1. Definition of blocking action

During the war years the senior author, because of his prewar activity in the relatively new field of extended forecasting, was assigned to the U. S. Navy Long Range Forecasting Unit in the Joint Weather Central, Washington, D. C. During this tour of duty the striking long-period and widespread effects which occasional large-scale quasistationary high-pressure areas impose upon the westerlies created a strong impression. In view of the fundamental nature of the

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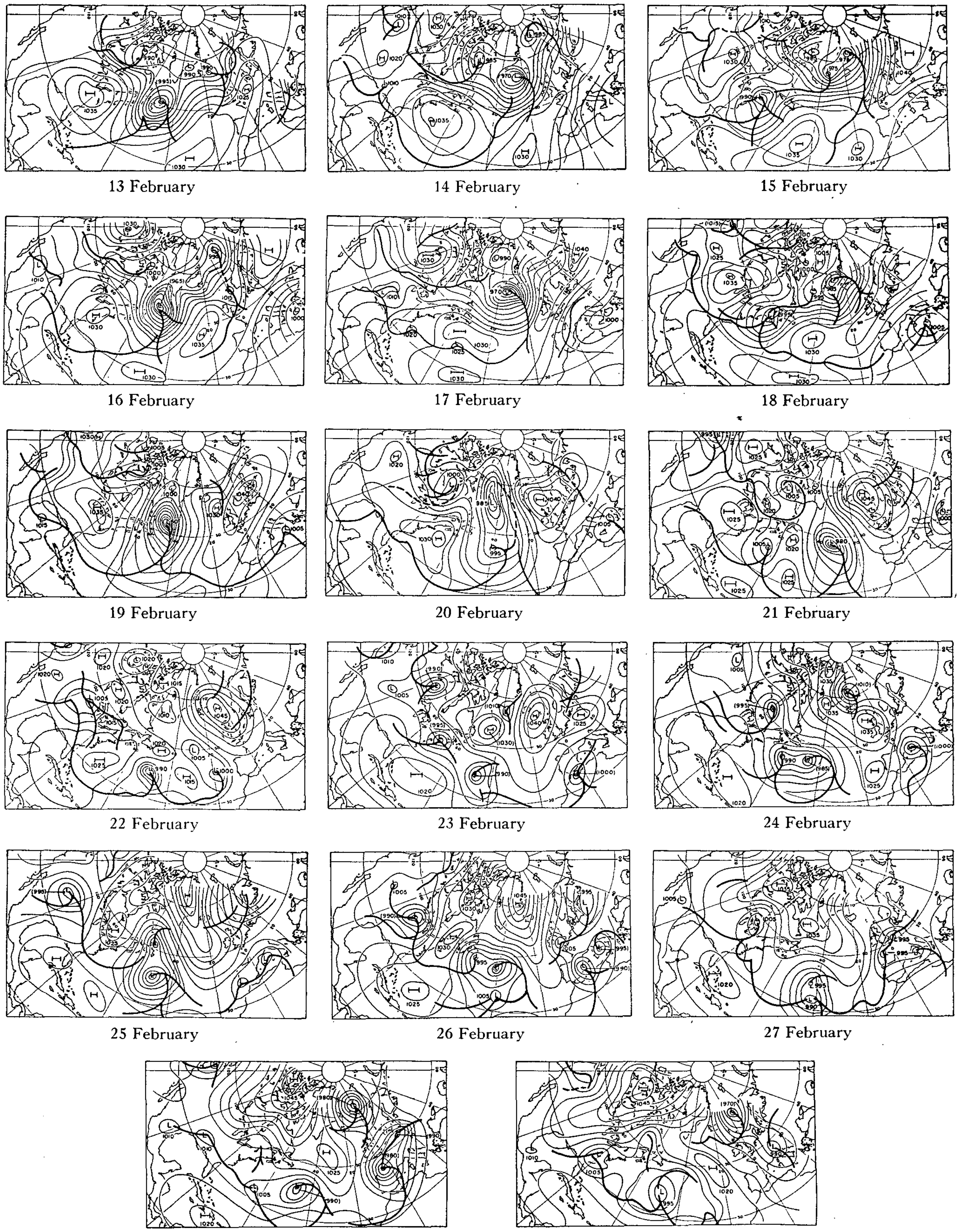

13 February
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FIG. 1. Example of blocking action, 13–29 February 1944.

weather processes associated with such developments and their importance in extended forecasting, the Office of Naval Research has provided funds for research on "Blocking Action" at the Department of Meteorology, California Institute of Technology. The following paper outlines the more important results of this research.

The term "blocking action" may be defined as a state of circulation in which the normal zonal flow is interrupted in a sector or sectors by strong, persistent, meridional-type flow. Synoptically, the condition is characterized by the development of persistent high pressure at high latitudes in such a manner as to obstruct the normal eastward progress of migratory cyclones and anticyclones. In the high-pressure region the pressure and temperature are considerably above normal in the troposphere, and in the stratosphere pressure is above normal and temperature below normal. The high-pressure development is linked, as a rule, to simultaneous, abnormally deep depressions either upstream or downstream or both, which are often trapped at rather low latitudes.

Usually when blocking action is in existence in one region, meridional-type flow will develop subsequently upstream and/or downstream. Some use of the principle of upstream modification has been made in extended forecasting as practiced in this country [4; 12]. This process has also been recognized and used by the Multanovsky school of long-range forecasting [13].

2. Two examples of blocking action

In order to fix in the reader's mind more precisely the meaning of the term "blocking action" and its synoptic implications as far as practical meteorology is concerned, the following two examples are submitted.

The first is that of 13–29 February 1944. Referring to fig. 1, it will be seen that over the Atlantic the three days 13–15 February were characterized by strong zonal-type flow with rapid movement of a deep depression from just east of Newfoundland to the east coast of Greenland. Simultaneously, a high over the eastern United States migrated to the mid-Atlantic. By the 15th a new deep low lies over New England, having moved there from the southwest as a rapidly deepening system. During the three days, 16–18 February, a course of events very similar to that of the first three days occurred, indicating that in this zonal-type flow there exists a three-day period. In accordance with this rhythm one should expect the chart on the 19th to appear analogous to those of the 13th and 16th. A glance at the 19th reveals correspondence with one notable exception; namely, a 1040-mb high has developed over southern Scandinavia. This is the first indication of blocking action. During the three days 19–21 February, this high moves slowly westward and

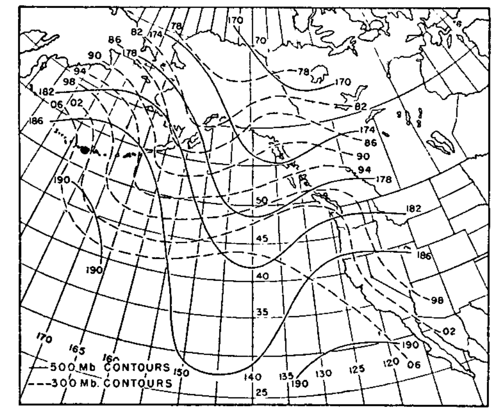


FIG. 2. 500-mb and 300-mb contours, 31 March 1948. The location of Longview, Alaska is indicated by the black dot.

dominates the eastern North Atlantic, forcing depressions to stagnate in the central Atlantic.

During the three days 22–24 February, the blocking action continues with lows being trapped at southerly latitudes in the Atlantic. During the following three days, 25–27 February, the high-pressure area is centered for the most part over Greenland. Note the "plunging type low" which moves southeastward from the east coast of Greenland to Scandinavia on the 24th and 25th. This development is symptomatic of the breakdown of the blocking high and during the last two days, 28–29 February, one finds low pressure in the Norwegian Sea and a 1045-mb high over North America. The retrogression of the blocking high is quite apparent in this series.

The upper-air structure associated with the development of a blocking high can be demonstrated by a second example in the Aleutian Islands area. During the period 31 March–3 April 1948 a large high developed over the central Aleutians and remained more or less stationary in this region during most of April. Fig. 2 shows the 500-mb and 300-mb contour patterns for 31 March and fig. 3 those for 3 April. Note that the trough aloft sloped markedly to the west on 31 March, but that by 3 April anticyclonic contours appear at

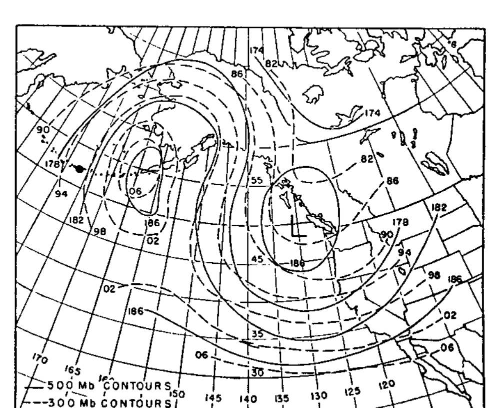


FIG. 3. 500-mb and 300-mb contours, 3 April 1948.

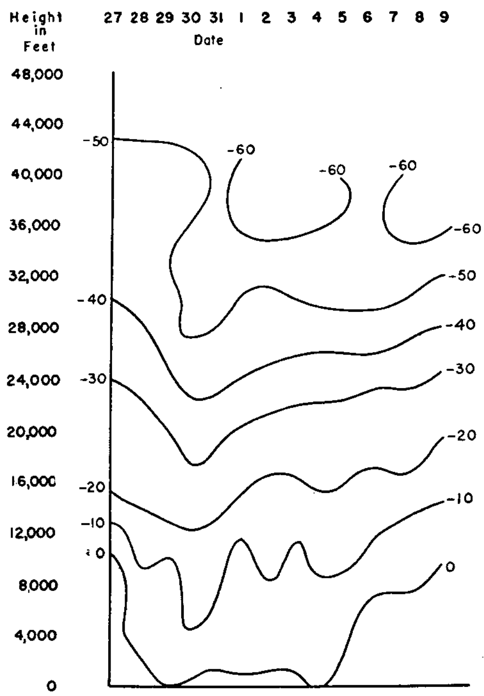


FIG. 4. Time cross section of the temperature at Longview, Alaska, during formative stages of blocking high, 27 March–9 April 1948.

both levels. Fig. 4 is a time cross section at Longview, Alaska, of temperatures during and after formation. Note that there is marked cooling in both the stratosphere and troposphere when the strong sloping trough of 31 March was in existence, indicating that there must have been widespread lifting in the zone of convergence near the trough. Thereafter the troposphere gradually warms, especially the lower part, but the stratosphere remains cold. Since the flow is anticyclonic at all levels, mass advection can no longer be a factor of importance. Therefore, one must attribute the warming in the troposphere to descending motion there, with no appreciable vertical displacement occurring in the stratosphere.

The divergence in the frictional layer must be compensated by convergence at higher levels in order that continuity be preserved and descent continue in the lower troposphere. With a pattern consisting of stationary anticyclonic streamlines at all levels this is difficult to explain. However, an examination of surface charts (not shown) shows that true steady-state conditions do not prevail there, but that as a result of the eastward passage of frontal zones north of the high, a series of small replacement cells move into, or merge with, the high center. It is believed that this process provides the necessary convergence to counteract the frictional layer outflow. Fig. 5 is a schematic diagram of what probably occurs. A mass of air in the south is drawn by frontal action from A to B, and another mass lying in the north is drawn from C to D. If this process is visualized as a continuous process of mixing of air masses across a mean westerly current, then with respect to the mean mass distribution, air with higher absolute momentum is carried north and that with lower momentum to the south. In each case the transferred air mass finds itself out of balance with its surrounding mean pressure field and is in the first case accelerated southward and in the second case northward with resultant convergence. This mixing process will be discussed in more detail below, particularly in connection with the northward transport of heat across the westerlies.

3. Basic data

In the investigation semi-statistical procedures were followed requiring the use of a long series of data in order that the requirement for a reasonable sample be met. For this reason the 40-year Historical Weather Map Series [19] has been used. It is recognized that for this investigation upper-air charts would be more desirable, but the upper-air series is quite short and, therefore, cannot be used. Certain useful inferences can be made concerning the upper-air structure from the surface patterns.

The basic data are daily pressure values at or near 1200 GCT during the January–February season at intersections of 5 degrees of latitude and 10 degrees of longitude in the region 35°N–60°N and 140°E through 180°–40°E. Pressures at 5-degree intersections of longitude for latitudes 50°N–60°N were used in some of the work. This covers the portion of the westerlies for which there exists a fairly reliable reporting net for the entire 40 years of the series.

As a test of significance two independent data samples were used. The even years 1900–1938 were used as the working data and the results were checked against the odd years whenever possible.

4. Blocking-action index

After the preparation of a subjective catalogue of cases of blocking action, suitable objective criteria for the determination of the dates of inception and duration of blocking action in various longitudinal sectors were established. These are based primarily upon the

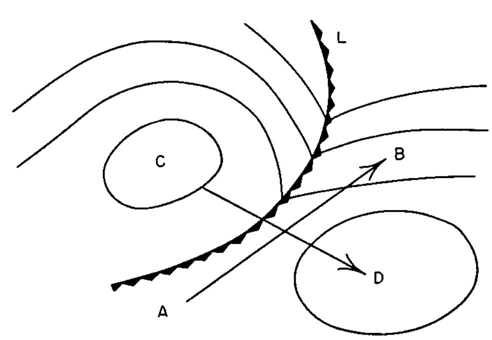


FIG. 5. Schematic diagram of mass transfer into blocking high.

existence of persistent high pressure in latitudes 55°N to 60°N.

A preliminary study of the year-to-year variations in the extent of blocking action for the January–February season was based upon an index N_d, the total number of 5-degree intersections at 55°N and 60°N with pressure departures of +20 mb or more in each of the three sectors:

Pacific:	140°E to 135°W
N. American:	130°W to 60°W
Atlantic:	55°W to 40°E.

Departures from Shaw's [17] normals were used, these normals being considered satisfactory for the purposes of this study.

The more significant results revealed by this study are as follows:

a. There is a marked tendency for blocking activity to occur in the northeastern portions of the Atlantic and of the Pacific.

b. Fluctuations in the year-to-year extent of blocking action in the three different sectors tend to parallel each other.

c. The curve for the sum of all sectors shows a long-term oscillation having a period of about 12–14 years.

The form of the curve is suggestive of the sunspot curve. The two are compared in fig. 6 (curves 1 and 2). These curves indicate some correlation, overlooking the 1918 dip, in the case of the first two peaks but the last peak is considerably out of phase. The possibility that a real relationship exists is within reason in view of Clayton's [1] findings on the relationship between sunspots and pressures at high latitudes.

In order to explore the possibility of a contemporaneous or lag relationship between the day-to-day fluctuations in N_d and in sunspot and magnetic character numbers, curves of daily values of these quantities were prepared. Examination of these curves revealed that although there were some rather striking relationships in some years, other years utterly failed to show the same pattern. The curves suggest that the normally expected interval between peaks in the N_d curve and the other curves was about the same but that the phase relationship was of a rather random nature.

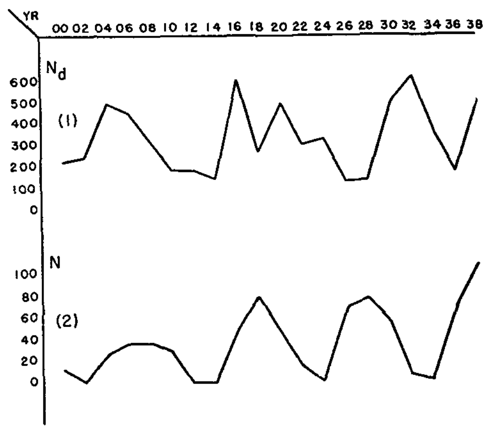


FIG. 6. Long-period variation in index numbers, N_d, and sunspot numbers, N, for the January–February seasons of the 20 even years: 1900–1938.

It seems reasonable to ask the question: If the earth receives more heat than it radiates in the tropics and *vice versa* in the arctic, why do not the tropical regions heat up, and the arctic cool down, until radiation equilibrium is achieved, thus eliminating the need for meridional heat transport? According to Milankovich's [8] figures for theoretical planetary temperature distribution under radiation equilibrium, temperature at the equator would have to exceed by 12F, and at 80°N be 28F lower than, the observed temperature. Under these conditions it is obvious that there would have to exist zonal flow, although this could probably be eliminated at the surface with a properly shaped tropopause. However, it appears that this zonal current with its large vertical wind shear would be unstable, *i.e.*, it would be turbulent, not laminar, in the sense that large horizontal eddies would form, namely, migratory cyclones and anticyclones, which are capable of transporting heat northward by eddy conduction. In the heat budget of the atmosphere it would then become necessary to add to the radiation terms a term for heat diffusion by eddy conduction, and there results a departure from the radiation climate. Likewise, in the mean wind equations shear stress terms must be added in order to achieve a balance of forces.

As early as 1921 Defant [3] suggested this type of treatment of the general circulation problem, considering the migratory cyclones and anticyclones as horizontal eddies in the stream of the westerlies. Later work by Lettau [9; 10] and others has extended the idea with particular attention given to the determination of the *Grossaustausch* or broadscale exchange coefficient.

Jeffreys [7] showed that the loss of momentum suffered through ground friction in the westerlies was not made up by the northward flow of air across the isobars in the friction layer. Surface friction would then destroy the westerlies within several days if a

BROADSCALE LATERAL TURBULENCE

5. Preliminary considerations

Because of the abnormally great meridional flow during blocking action one might infer that the cross-westerly heat transport is excessive during periods of blocking action. Toward the determination of the heat transport and of other dynamically important characteristics it was decided to apply the concepts of broadscale horizontal turbulence.

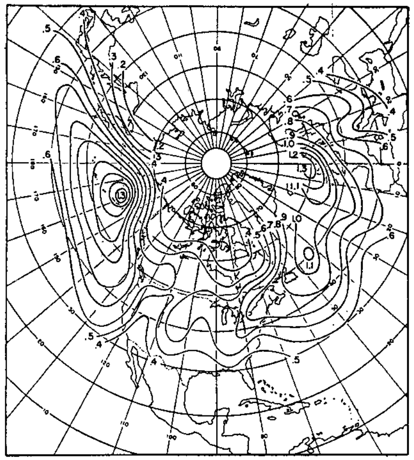


FIG. 7. Mean value of eddy-viscosity coefficient $A_\phi(10^8$ g cm^{-1} sec$^{-1})$ for January–February seasons of the 20 even years: 1900–1938.

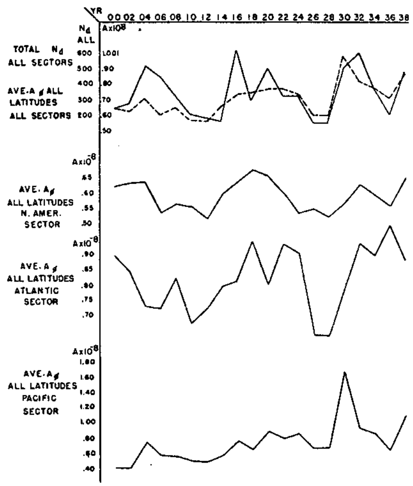


FIG. 8. Variation in eddy-viscosity coefficient $A_\phi(10^8$ g cm^{-1} sec$^{-1})$ by sectors compared to average for all sectors and index numbers, N_d, for January–February seasons of 20 even years: 1900–1938.

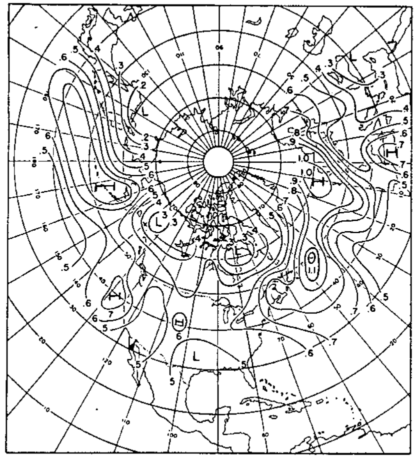


FIG. 9. Mean value of eddy-viscosity coefficient $A_\phi(10^8$ g cm^{-1} sec$^{-1})$ for January–February seasons of years: 1906, 1910, 1912, 1926, and 1928.

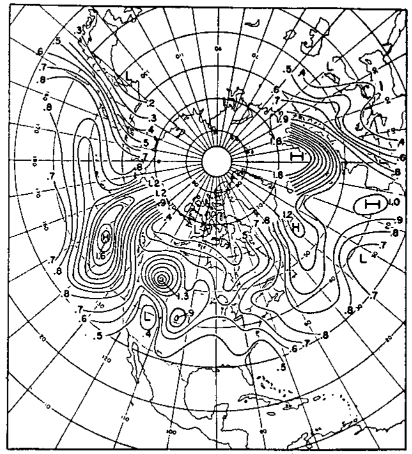


FIG. 10. Mean value of eddy-viscosity coefficient $A_\phi(10^8$ g cm^{-1} sec$^{-1})$ for January–February seasons of years: 1918, 1920, 1922, 1932, and 1938.

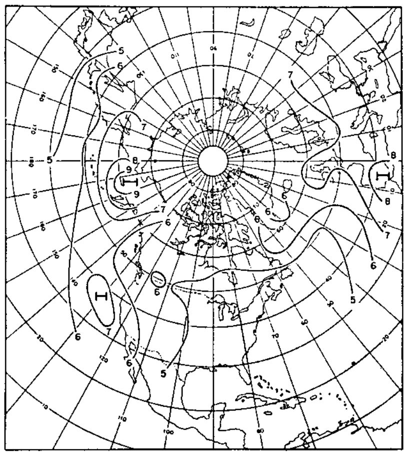


FIG. 11. Mean value of the eddy period in days for the January–February seasons of the 20 even years: 1900–1938.

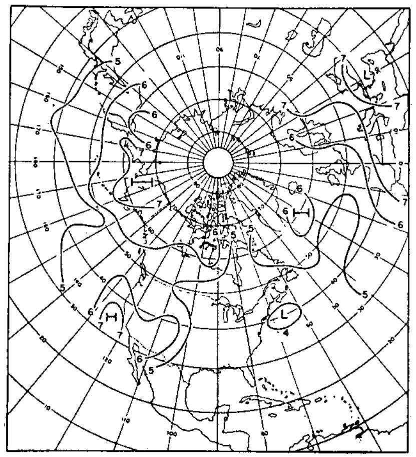


FIG. 12. Mean value of the eddy period in days for the January–February seasons of the years: 1906, 1910, 1912, 1926, and 1928.

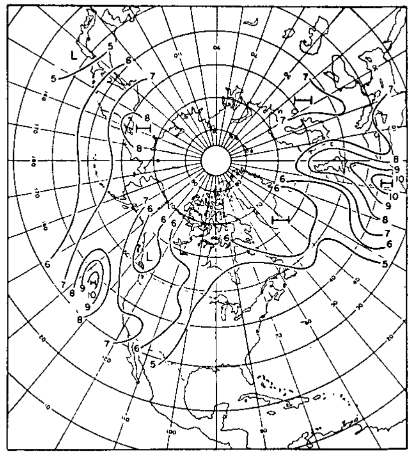


FIG. 13. Mean value of the eddy period in days for the January–February seasons of the years: 1918, 1920, 1922, 1932, and 1938.

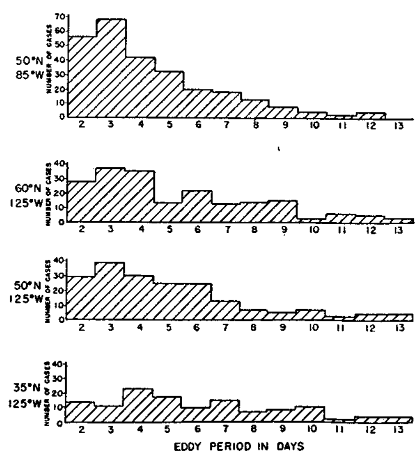


FIG. 14. Frequency of occurrence of eddy periods at selected locations for the January–February seasons of the 20 even years: 1900–1938.

fresh supply of angular momentum were not provided through re-establishment of the westerlies from time to time. Rossby [14] hypothesized that through broad-scale lateral friction the solenoidally produced westerlies near the polar front dragged along air farther south in such a way as to compensate for momentum losses in the ground friction layer. In this manner, he obtains a maximum eddy-viscosity coefficient of 0.49×10^8 gm cm^{-1} sec^{-1} which corresponds well to the values found by Lettau and others by direct measurement of eddies.

In computing the northward heat transport it is necessary to determine the lateral heat exchange coefficient. It has been shown by various investigations that in situations where stability does not appreciably enter into the picture, exchange coefficients for various properties are nearly identical. Therefore, it seems permissible for the present purposes to assume that the exchange coefficients for heat and for momentum are identical.

6. Eddy-viscosity coefficient

A technique used by Lettau [9] to determine a *Grossaustauch* coefficient or lateral eddy-viscosity coefficient is well adapted for use with the type of data forming the basis of this investigation. The basic feature of this technique is the determination of the eddy-viscosity coefficient from the variability in the south to north geostrophic wind flow. Consider the meridional component of the mean geostrophic flow at a point for a two-month period. If $\bar{V}_i'$ is the mean absolute geostrophic departure from this mean flow during one surge of abnormal flow northward or southward as measured by the pressure difference between closely adjacent points at equal longitudinal distances to the left and right of the reference point, then the mixing length for the surge is $\bar{V}_i'n_i$ where n_i is the number of days in the surge. The average over all surges of the product of the mixing length and the mean velocity times the density is the eddy-viscosity coefficient. Thus:

$$A_\phi = (2N)^{-1}\rho \sum^{N} \bar{V}_i'^2 n_i, \qquad (1)$$

where A_ϕ is the eddy-viscosity coefficient, ρ the density, and N the number of surges during the two-month period over which the summation is taken.

The geographical distribution of the mean value of A_ϕ for the 20 even years is shown in fig. 7. There are strong maxima just east of the Aleutian and Icelandic Lows. A band of maximum values extends west-southwestward from the Norwegian Sea to a secondary maximum just east of Newfoundland, thence westward through northeast and north central United States, then northwestward into the southern Prairie Provinces and on into the Bristol Bay maximum. A ridge extends west-southwestward from this maximum to the coast of Asia. These results are comparable to those found by previous investigators for single seasons.

The region of maximum mixing is strikingly similar to the region of maximum frontal activity and to the region of maximum N–S thermal gradient. Minima occur along the southern edge of the westerlies and over the cold continental areas.

As a measure of the overall mixing activity across the westerlies the coefficients have been summed over each of the three sectors for each year of the 20-year series. The results are shown graphically in fig. 8 along with the A_ϕ and N_d curves for all sectors. The graphs of total N_d and total A_ϕ correlate $+0.64 \pm 0.09$. In view of the role played by blocking action in lateral mixing it appears that perhaps A_ϕ should be used as a measure of blocking action rather than N_d. Deleting the unusual year 1930 in the Pacific, the intercorrelations of total A_ϕ between the sectors and their probable errors are:

Atlantic—N. America	$+0.40 \pm 0.13$
Atlantic—Pacific	$+0.28 \pm 0.14$
N. America—Pacific	$+0.18 \pm 0.14$.

Of these positive correlations the one most likely to be of significance is that between the Atlantic and North America.

From these figures one might conclude that there exists a greater tendency for the year-to-year variations in mixing activity to run parallel in the various sectors than for excessive mixing in one sector to be compensated by a deficit in another sector. As shall be shown, the northward heat transport is greater than normal during years of excessive mixing and *vice versa*, so that should the year-to-year variations in mixing run parallel in the various sectors, rather marked year-to-year variations in the radiational properties of the earth's atmosphere or in incoming solar radiation must occur in order that the heat budget of the atmosphere be balanced.

Five years during which the mean A_ϕ was low and five during which it was high for all sectors were selected for special study. Charts were prepared for the mean A_ϕ in each of these two categories. These are shown in figs. 9 and 10. Fig. 9 (low values of A_ϕ) shows a pattern quite similar to the mean condition but with generally reduced values. Greatest reduction in values occurs at about 50°N, 150°W in the eastern Pacific and at about 60°N, 05°E in the Norwegian Sea. Fig. 10 (high values of A_ϕ) shows striking maxima in the eastern Pacific and Norwegian Sea. There is an overall increase over normal with maximum increases at about 45°N, 140°W in the eastern Pacific, 50°N, 115°W in southeastern British Columbia, and 60°N, 05°W in the eastern Atlantic. These charts show clearly that the greatest year-to-year variation in lateral mixing occurs to the west of the continental areas and in the case of the Pacific, south-southeast of the area of normal maximum A_ϕ. One might conclude that in the zone of the westerlies these two areas are the principal controls in determining seasonal characteristics.

7. Eddy period

It is interesting to obtain a measure of the mean size of the eddies responsible for lateral mixing. For the purpose of investigating this, the mean eddy period has been chosen, *i.e.*, the double surge period or the average period from one cyclonic system to the next.

The geographical distribution of the period for the 20-year series is shown plotted in fig. 11. The period ranges from a minimum of between 4 and 5 days in southeastern United States and off the East Coast to a maximum of 9 days in the Bristol Bay area. In the mean frontal zones the mean period is quite close to 6 days.

Figs. 12 and 13 show the geographical distribution of eddy periods for the five years of low A_ϕ and of high A_ϕ. The patterns are essentially the same as for the 20-year mean, with reduced or augmented values corresponding to low A_ϕ or high A_ϕ years. The increased eddy period is most striking for high A_ϕ years in the vicinity of Spain and off the California coast. It appears that in years when blocking action is prevalent, mean eddy sizes are larger than normal; *i.e.*, there are more cases of large and persistent eddies.

The mean mixing length was computed for 33 points in the region under investigation by use of the relationship

$$l' = (2N)^{-1} \sum^{N} \bar{V}_i n_i, \tag{2}$$

where all symbols have the same definition as that given in section 6. Values for the mean mixing length were more uniform with respect to geographical location than those of A_ϕ or mean period, but showed again, peak values in the Bristol Bay and Norwegian Sea areas. The over-all mean value was 8.2×10^7 cm. This seems consistent with the size of migratory cyclones and anticyclones.

The over-all mean eddy period is seen to be very near to 6 days. However, an examination of the frequency distribution of eddy periods reveals that the most frequent period is far less than 6 days. Fig. 14 shows the frequency distribution of periods for several points. The outstanding feature is the sharp peak at about 3 days with lesser peaks or humps at periods roughly corresponding to multiples of 3 days. This basic eddy period of about 3 days is substantiated by synoptic experience regarding the average interval between the passage of principal troughs, each trough usually containing a cyclone family. From the frequency distribution curves of eddy periods the basic eddy period can be roughly determined within fractions of a day. It appears that the basic period is shorter in the north and longer in the south, ranging from about $2\frac{1}{2}$ days at high latitudes, to 3 days in the heart of the westerlies, and $3\frac{1}{2}$ days in southerly latitudes. This feature conforms to synoptic experience regarding wave motion along the arctic and polar fronts.

From the shape of the distribution curves it appears that the flow pattern at a point is from time to time disturbed in such a manner that one or more single eddies are missing, or masked, so that various multiples of the basic eddy period may occur. A number of explanations present themselves; but most seem to involve blocking action in one way or another. Large polar outbreaks in North America in winter frequently mask at least one principal trough passage by deflecting the migratory surface systems to the north or south of the normal trajectory. At other times complete blocking may occur at the surface although the trough passage may still be detected aloft. This phenomenon is also characteristic of blocking highs over the oceans but the number of troughs masked is usually more than one. In cases where the arctic front is simultaneously active with the polar front, intensification of meridional flow may result from time to time through interference patterns between the eddy motions along the two fronts of slightly different basic periods.

All of these processes having multiple basic period durations apparently involve relatively long-period deformations of the westerly flow which cannot in themselves be considered in the same category as the short-period eddies characterizing unstable baroclinic wave developments in the stream of the westerlies. This point will be discussed in more detail below.

8. Northward heat transport

The northward transport of real heat is

$$- c_p A_\phi R^{-1} \, \partial T/\partial\phi, \tag{3}$$

R being the radius of the earth. At the surface this represents the amount of heat flowing northward across a vertical surface of unit height parallel to the equator. The geographical distribution of the mean northward heat transport for the 20 even years appears in fig. 15. In computing this, the 20-year mean eddy-viscosity coefficients were used in conjunction with mean thermal gradients computed from Shaw's [17] normals. Since regions of maximum thermal gradient correspond roughly to areas of maximum mixing, the heat-transport chart is essentially the same in form as the A_ϕ chart.

It must be remembered that water vapor plays an important role in the northward transport of heat across the westerlies. According to Jacobs [5], large amounts of heat are given to the atmosphere through evaporation just off the east coasts of the continents. This heat is carried some distance northward in latent form before it is realized as sensible heat. When the release occurs, thermal gradients are sharpened and heat transport through lateral turbulence becomes an important factor.

A rough comparison between the heat transport values in the latitude of maximum transport and Simpson's [18] figures for the heat transport required to counteract the radiational unbalance shows that the layer of heat transport need be only about 2 km thick to meet the requirement, or if the mixing coefficient decreases exponentially with height, at 2 km it should be about $1/e$ of its surface value.

It has been impossible to study the year-to-year variation in the northward heat transport in the Atlantic and Pacific sectors due to lack of a satisfactory network of temperature data. In the North American sector monthly surface air temperature departures from 1916 through 1938 were available over an extensive network and the mean seasonal northward heat transport was computed for each even year by using the appropriate January–February mean eddy-viscosity coefficient and the corresponding mean thermal gradient. This was done for each of the even years 1916–1938. This was found to correlate $+ 0.85 \pm 0.05$ with the mean seasonal value of A_ϕ for the North American sector, showing a positive relationship between mixing and northward heat transport. Next, the year-to-year variation in the thermal gradient as represented by the mean seasonal temperature difference between 35°N and 50°N was compared to the variation in mean seasonal values of A_ϕ and northward heat transport. The magnitude of the mean temperature gradient fluctuates considerably more than the other quantities but the trend matches well. In the series of even years, a strong 6-year oscillation was present in

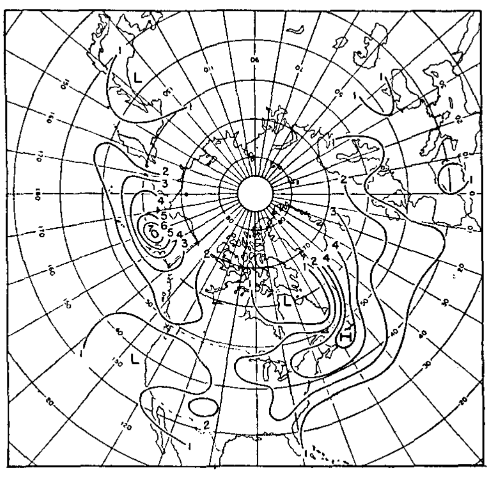


FIG. 15. Mean value of northward heat transport (cal cm^{-2} sec^{-1}) for the January–February seasons of the 20 even years: 1900–1938.

the gradient but when this is smoothed by the use of moving 3-year means (Jan.–Feb. seasons for three consecutive even years), the correlation between the A_ϕ and the temperature gradient curve is $+ 0.76 \pm 0.09$. This correlation shows that, except for the short-period fluctuations, the eddy viscosity is greatest when the magnitude of the thermal gradient is greatest. One would infer from this that when radiation conditions are such as to increase the magnitude of the thermal gradient in a given season, the circulation adjusts itself by becoming more vigorous and increasing the heat exchange across latitude circles so as to partially compensate for the radiation anomaly.

A fact worthy of note is that the thermal gradient anomalies were of the same sign in January as in February in all but two of the years. This suggests that there may have been some peculiarity in the radiational properties of the atmosphere, or the earth's surface, which persisted through the season.

In further support of the hypothesis that mixing is greater when the mean thermal gradient is greater, the following results of a study of Alaskan temperatures are submitted. In the Pacific sector the maximum thermal gradient is normally in the Bering Sea and southern Alaska. Oceanic air temperatures fluctuate less from year to year than do continental air temperatures, therefore, Alaskan mean temperatures during the January–February season should serve as a negative index of the N–S thermal gradient in the Pacific. The curve of the year-to-year variations in mean Alaskan temperature correlated $- 0.86 \pm 0.04$ with the curve of A_ϕ for the Pacific sector.

9. Momentum considerations and the vertical variation of the mixing coefficient

One of the fundamental requirements regarding the nature of the eddies under consideration is that they be unstable, forming and dissipating fairly frequently. Usually, they should appear on the surface synoptic chart as moving closed centers. As one ascends to higher levels one encounters fewer and fewer of these closed centers. This suggests that the lateral mixing must decrease markedly with height. On the other hand, application of equation (1) to 500-mb, 10-, 13- and 16-km surfaces in North America shows values of A_ϕ actually increasing slightly up to the tropopause, then decreasing above that level. Apparently the decrease due to the density term is compensated by the increased strength of the westerly current fluctuations. These fluctuations are not for the most part in the nature of closed centers or eddies, but merely represent a deformation of the upper-level pattern into a sinusoidally shaped wave form incapable of any net momentum transport. Only in the case of blocking action do closed centers appear at very high levels.

Therefore, it is believed that applications of equation (1) to higher levels yield spurious results.

As mentioned above, the equations of mean motion must include shear stress terms. These terms are responsible for the production of a meridional and vertical motion flow pattern. Considering the more important terms one has for straight W–E isobars:

$$-f\rho\bar{V} = \partial\tau_{yx}/\partial y + \partial\tau_{zx}/\partial z,$$

where the customary meteorological coordinate system is used, and $f = 2\omega \sin\phi$.

Integrating first from the surface (0) to gradient level (G), then from gradient level to ∞ one finds:

$$-\int_0^G f\rho\bar{V}\,dz = \int_0^G \frac{\partial\tau_{yx}}{\partial y}\,dz + (\tau_{zx})_G - (\tau_{zx})_0,$$

$$-\int_G^\infty f\rho\bar{V}\,dz = \int_G^\infty \frac{\partial\tau_{yx}}{\partial y}\,dz + (\tau_{zx})_\infty - (\tau_{zx})_G.$$

The integrals represent the momentum lost by the columns. In the mean the total mass crossing a latitude circle must vanish; therefore $\int_0^\infty \rho\bar{V}\,dz = 0$, and also $(\tau_{zx})_\infty = 0$. Adding,

$$0 = \int_0^\infty \frac{\partial\tau_{yx}}{\partial y}\,dz - (\tau_{zx})_0,$$

or

$$\int_0^\infty \frac{\partial}{\partial y}\left(A_y\frac{\partial\bar{U}_a}{\partial y}\right)dz = (\tau_{zx})_0, \qquad (4)$$

where U_a is the absolute velocity in space.

Using typical winter velocity profiles it is easy to show that in order to achieve the balance of momentum gains required by (4), the mixing coefficient must decrease quite markedly aloft, the change being of the same order of magnitude as that arrived at from heat transport considerations above.

The primary purpose of resorting to the use of turbulence principles lies in the need for obtaining a clear understanding of the significance of a mean chart. It has been shown that through the use of mixing coefficients one can set up equations for mean atmospheric motions which express the heat balance and balance of forces involved.

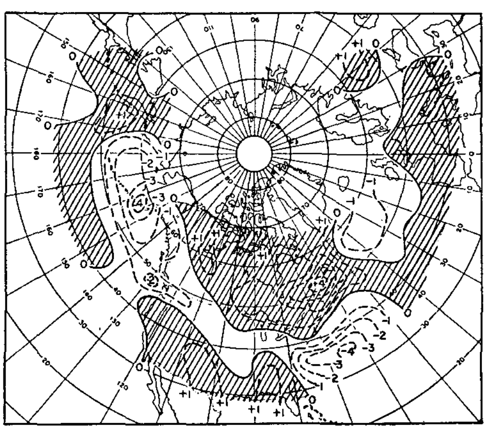


FIG. 16. Mean value of the heat gain at the surface (10^{-8} cal cm^{-3} sec^{-1}) through the action of lateral turbulence during the January–February seasons of the 20 even years: 1900–1938.

No pretension is made toward explanation of the physical processes within an eddy which account for its role in exchanging heat, momentum, *etc.* In attempting such an explanation the classical method of linearized perturbation equations fails because the exchange takes place in the unstable case; namely, when the basic assumption that second order terms in the perturbation equations can be neglected is no longer valid. The treatment of heat transfer in wave patterns with horizontal solenoidal fields by Jaw [6] is a step in the right direction.

Rossby [14] has pointed out that the vertical velocity field resulting from lateral turbulence establishes certain "dynamically prescribed heat sources and sinks." He suggests that a full treatment of the problem might have to await a better understanding of the radiational processes in the earth's atmosphere. From the foregoing it seems likely that lateral turbulence is adequate to disperse the heat accumulated by dynamic heating in the regions of descending motion.

10. Diffusion of heat

The heat gained per cm³ at the surface due to lateral turbulence is

$$\frac{c_p}{R\cos\phi}\left[\frac{\partial}{\partial\phi}\left(A_\phi\cos\phi\frac{\partial T}{\partial\phi}\right) + \frac{\partial}{\partial\lambda}\left(\frac{A_\lambda}{R\cos\phi}\frac{\partial T}{\partial\lambda}\right)\right]. \qquad (5)$$

A computation of A_λ, the mixing coefficient for E–W mixing, showed that the importance of this quantity was equal to that of A_ϕ in N. America during Dec. 1945–Jan. 1946. This is probably due to the fact that in the mean flow over N. America there exists a considerable meridional component. On the other hand, the W–E thermal gradient is in general less than the N–S gradient and, therefore, the computation of the mean heat gain, neglecting the second term, should represent a fairly good approximation to the true value. Fig. 16 shows the distribution of heat gain due to turbulent diffusion as computed from the mean northward heat transport chart.

The advective rate of heat loss, dQ/dt, is given by:

$$dQ/dt = c_pT\nabla\cdot\rho V + c_p\rho V\cdot\nabla T. \qquad (6)$$

The first term on the right is the total mass divergence and must equal zero at all levels under steady-state conditions. The second term may be divided into two parts, one representing the horizontal advection effect, the other the vertical motion effect. Wexler [20] has computed the mean horizontal advective effect between the surface and 10,000 ft using mean isotherms and winds. Fig. 17 is a reproduction of his results and shows a similarity to fig. 16. Wexler has compared his values to the values of heat given from the ocean to the atmosphere in accordance with Jacobs' [5] calculations and finds a general correspondence.

According to Jacobs' data, sensible plus latent heat is given by the ocean to the atmosphere in almost the whole hemisphere in winter, but most markedly so just off the east coasts of North America and Asia where total amounts are over 500 calories day^{-1} cm^{-2}. Wexler explains the fact that his charts show heating before the east coast of the U. S. is reached by the presence of subsidence (the vertical-motion effect) in the polar outbreaks moving southeastward to the coast. The area of cooling over Alaska (and the northeast Atlantic) is likewise explained by vertical motion, *i.e.*, widespread ascent in the area. However, acceptance of this explanation for the differences between Wexler's and Jacobs' charts introduces certain difficulties in that no means is provided for the westerlies to remove the heat picked up off the east coasts and thus the westerlies should, in the mean, gradually warm up. Transfer of heat to the ground by vertical turbulence over continental areas in the region of the westerlies could hardly be of the required magnitude due to the suppression of turbulent mixing by stable thermal stratification. Numerous investigations [11] of the vertical transfer of heat from the atmosphere to the surface and *vice versa* by the action of turbulence have definitely shown that heat is much more readily transferred to than from the atmosphere. Radiational losses from the atmosphere itself do not provide a reasonable explanation as the air particles in the westerly current lie in a given region with a given radiational regime for only a short time.

With these considerations in mind one may compare Wexler's chart (fig. 17) with the heat gain through lateral turbulence (fig. 16), and find a solution to the difficulty. The lateral turbulence acts to remove heat from the westerlies as they move from the east coasts to the Aleutian and Icelandic lows. It is all along this region that heat is being transferred to the atmosphere, primarily through the release of latent heat resulting from the evaporation of water from the ocean areas just off the east coasts. The heat is easily transferred from ocean to atmosphere through evaporation, then convective activity, which may be considered for purposes of this discussion as being a form of vertical turbulence, transforms it to real form and mixes it through a thick layer. The heat removed through the action of lateral turbulence is spread principally in polar regions where a relatively stagnant condition prevails, allowing the warm air to slowly subside to lower levels where it is gradually cooled over a large area by radiational processes, eventually moving south to the westerlies as polar outbreaks. Because the westerlies are warmest downstream from the east coasts and coldest upstream, there exists a thermal pattern which creates the well-known mean trough in the eastern continental areas and crest in the eastern oceanic areas in the Northern Hemisphere.

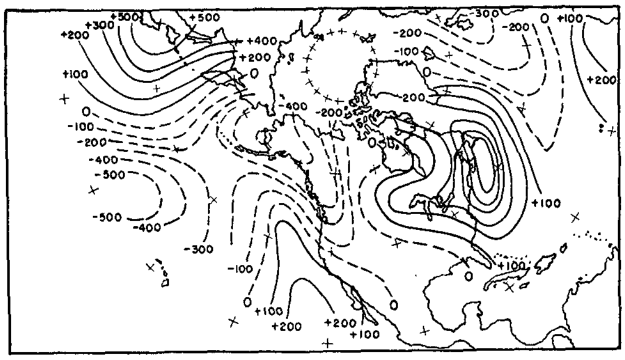


FIG. 17. Regions of normal heat gain (solid lines) and loss (dashed lines) for the layer from sea level to 10,000 ft for February (by permission of H. Wexler). Isolines are labeled in units of cal cm^{-2} day^{-1}.

The existence of a subtropical anticyclone can be explained as follows. If a straight westerly flow pattern is assumed to be the initial state, the divergence in the friction layer would be less over the sea than land due to smaller surface roughness, thus permitting the building up of an anticyclone to the right of the current over the sea. A general trough condition would then prevail over the continents, with a high over the sea. The sea surface would, over a long period of time, respond to this meteorological condition by starting a circulation bringing warmer waters northward off the east coasts. Cold winter continental air crossing the coast line would gain heat rapidly by evaporation and transfer of sensible heat. The lateral mixing is also greater over the oceanic area which may contribute to the building of the subtropical cell.

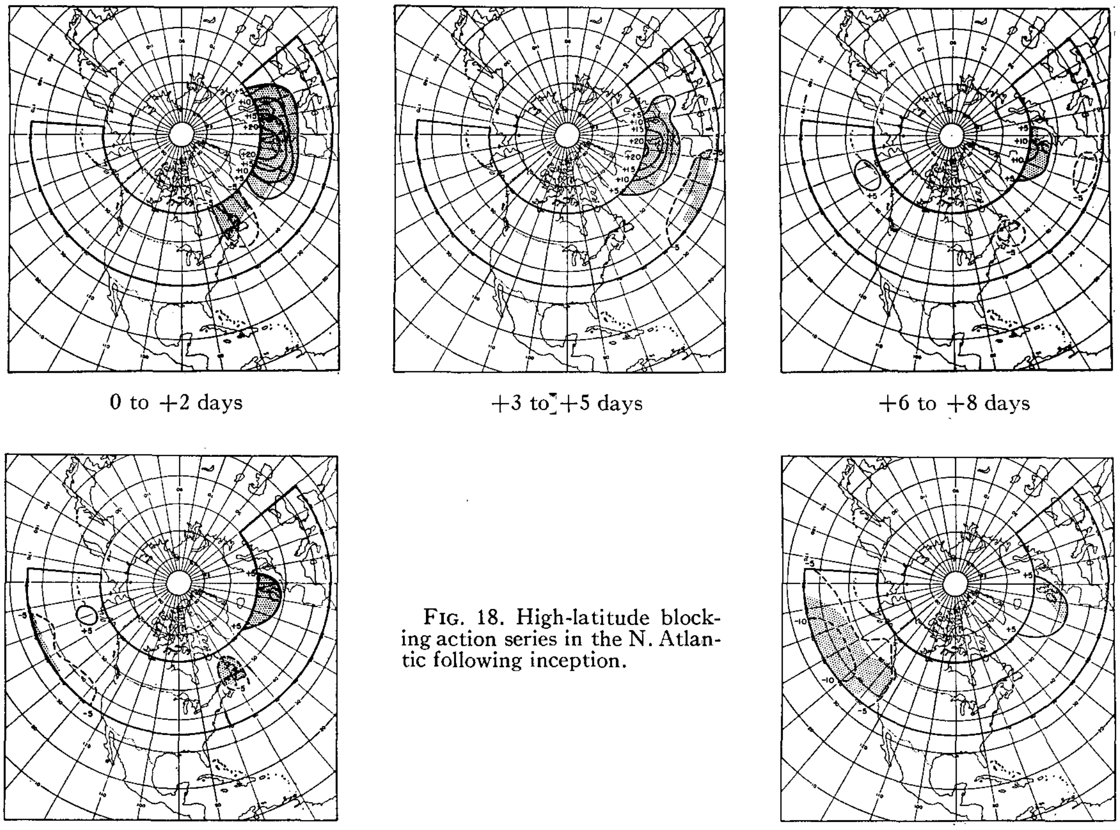
Long-period anomalous behavior in circulation patterns might well be explained on the basis of long-period anomalies in the heat loss from the ocean surface off the east coasts.

BLOCKING HIGHS

11. Blocking highs and heat transport

An acceptable hypothesis is needed to explain why a more or less zonal flow pattern breaks down from time to time into the meridional patterns characteristic of blocking action. The ordinary cyclonic and anticyclonic eddies having a mixing length of about 10^8 cm apparently are not always effective in producing the required heat exchange between equator and pole, especially at higher levels, and from time to time larger scale eddy motion must appear in the form of blocking action eddies.

This scale of eddy is so large that it produces a semipermanent deformation of the westerlies and perhaps



0 to +2 days

+3 to +5 days

+6 to +8 days

+9 to +11 days

+12 to +14 days

FIG. 18. High-latitude blocking action series in the N. Atlantic following inception.

should not be treated as eddy motion at all, as it represents a quasi-stable dynamical system. This latter approach is used in the following sections in which cases of blocking action are discussed in detail, treating each as a broadscale (a large portion of the northern hemisphere) and long-period (15–30 days) weather process. Ordinary migratory and transitory cyclonic and anticyclonic eddies are treated as a turbulence phenomenon superimposed upon the blocking action pattern.

The westerlies do not maintain a fixed position but swing northward in certain areas and southward in others, the amplitude of the swing depending upon the extent of the blocking action. When the westerlies act in this fashion there is in general a lengthening of the girdle of west winds which presents in effect a greater length for the operation of the processes of heat transport out of tropical regions and into polar regions through the turbulent action of migratory cyclones and anticyclones. In the region where the westerlies bulge northward, the higher tropopause moves north with them and in the troposphere there is an increase in heat in that sector. This heat is transported outward through the turbulent action of the individual migratory cyclones and anticyclones running along the periphery of the bulge so that there is a strong heat loss from the center and eventually the bulge diminishes, approaching a more normal condition. Prior to its disappearance, it apparently produces various involved effects upon the circulation both upstream and downstream from the point of inception. In the trough regions formed on either side of the anticyclonic bulge the turbulent heat transport is directed inward toward the trough line so that there is a heat gain in this region.

12. Blocking highs in the northeastern Atlantic area

As the highest frequency of occurrence of blocking action exists in the band from 0°–25°W in the northeastern Atlantic Ocean, this region was selected as the first region to be investigated.

The objective criterion of blocking action formulated to cover this case is as follows: "A band fifteen degrees of longitude wide and covering 55°N and 60°N latitudes must experience pressure departures of +20 mb or more for at least three consecutive days." The requirement of +20 mb or more eliminates all but the extreme cases of blocking action. The continuation for at least three days is for the purpose of excluding the possibility of the high cell being migratory in nature.

The examples satisfying this criterion have been divided into two classes designated by (*a*) high-latitude blocking and (*b*) low-latitude blocking, where the distinction is obviously the latitudinal position of the center of the blocking high cell. The high-latitude cases are the more extreme disturbances, being characterized by a high-pressure cell far to the north and trapped low-pressure areas along the southern periphery. In terms of the displacement of the west-wind belt they may be considered as cases of displacements far to the north around the northern limb of the blocking high or displacements far to the south around the southern limb of the trapped low-pressure areas. The low-latitude cases are characterized by a northward extension of the subtropical high cell where the connection with this cell is not broken; any trapped low-pressure centers being formed well to the east of the blocking high cell. In terms of the displacement of the west-wind belt these cases represent a less extreme northward displacement of the belt.

The inception date for each example satisfying the above criterion is listed in table 1, together with the duration of the blocking eddy in days.

It is apparent from this table that the high-latitude type is a considerably more stable development, even though it represents the more extreme disturbance. It would therefore be expected to exert more influence on the circulation patterns in other regions of the northern hemisphere than the less extreme, less persistent low-latitude type.

In order to eliminate some of the small-scale fluctuations caused by migratory cyclones and anticyclones from the larger scale fluctuations caused by the blocking action process, three-day mean pressure-departure charts[2] have been prepared for consecutive three-day periods following the inception of each type of Atlantic blocking. Due to space limitations, only the high-latitude series representing the more extreme disturbances is reproduced here. These charts are shown in fig. 18. Solid lines surround areas for which the total departure (sum of the even-year cases and odd-year cases) during the given three-day period exceeded an average of +5 mb per day. Dotted lines surround areas for which the total departure exceeded an average of −5 mb per day. Shaded areas are those areas where an average departure of at least 5 mb per day was observed independently in both odd and even years. These areas are considered particularly significant. Areas showing no large anomalies were omitted for clarity.

A check has been made at the point 60°N–140°W of the statistical significance of an average deviation from the normal pressure such as is indicated by the shaded areas of the figures. This check has indicated that at that particular point an average deviation of about 7 mb per day calculated from a total of ten independent three-day periods is statistically significant at the five per cent level. An additional check at the point 60°N–170°W under similar conditions shows an average anomaly of +5 mb per day to be significant at the five per cent level for that point. Since the standard deviations of the pressure at these points are larger than those of most of the points in the area under consideration it is believed that most of the shaded areas represent statistically significant deviations from the normal pressure.

The second chart in fig. 18 (+3 to +5 days) shows the production of a widespread low-pressure area in the vicinity of the Azores. This is in contrast to the above normal condition in the corresponding area for the low-latitude case. In the low-latitude series (not shown here) a long east–west belt of abnormally high pressure extends westward from the vicinity of the British Isles to the Mississippi Valley 3–5 days following inception of the blocking action. From a consideration of the individual weather maps involved and of some of the surface temperatures included on these maps, it appears that the western portion of this east–west high-pressure belt is made up of a warm high-pressure cell in the initial stages followed by cold polar outbreaks. This corresponds to the picture already formed of the distinction between the high- and low-latitude classes. The low-latitude type being a less intense deformation of the normal pressure field extends its influence farther in a longitudinal direction than the more disturbed high-latitude type.

No further large significant areas are observed in fig. 18 until the last two periods (+9 to +11 and +12 to +14 days). In the high-latitude case only, an extensive low-pressure area begins to develop in the east Pacific, continuing into the last period. From the check of the statistical significance at the point 60°N–140°W mentioned above, the peak average deviation of over 10 mb per day reached on the last chart in fig. 18 is considered highly significant in a statistical sense. It should be noted that there is no evidence on the charts

TABLE 1. Inception dates and durations (in days) of Atlantic blocking actions.

Even years			
High latitude		Low latitude	
Date	Duration	Date	Duration
30 Jan. 1902	12	13 Jan. 1902	4
11 Jan. 1914	5	20 Jan. 1904	3
21 Feb. 1916	3	19 Jan. 1906	4
23 Feb. 1928	7	4 Feb. 1908	6
7 Feb. 1930	6	2 Jan. 1922	3
8 Feb. 1932	23		
12 Feb. 1934	10		

Odd years			
High latitude		Low latitude	
Date	Duration	Date	Duration
7 Feb. 1901	16	26 Jan. 1905	4
13 Jan. 1903	7	13 Feb. 1909	3
30 Jan. 1911	10	16 Jan. 1911	4
16 Feb. 1913	3	18 Jan. 1923	3
20 Jan. 1917	5	11 Feb. 1933	4
8 Feb. 1917	4		
16 Jan. 1935	8		

[2] In a special statistical study of the significance and use of mean pressure charts, the writers found that the 3-day mean chart achieves, on the average, the greatest possible smoothing of daily cyclonic fluctuations and the maximum delineation of the large-scale processes considered here.

to show that this deviation is produced by any effects acting upstream from the blocking high. The possible methods of formation then, are an upstream effect from the blocking high at latitudes north of 60°N or a downstream effect through Siberia. Either method could lead to the development of high pressure in the Alaskan area (outside of the network of available data) and a subsequent southward displacement of the west-wind belt in the Pacific Ocean. The examination of some of the northern hemisphere sea-level charts for which an adequate network of data is available in Siberia has shown that the downstream method of formation is the more probable.

An excellent example of this high-latitude development is afforded in a recent example which may serve as an independent illustration. During mid-January of 1947 a high-pressure cell developed in the eastern Atlantic area reaching the high-latitude blocking action stage over the British Isles about 19 January with the pressure remaining predominantly high in that area into March. An extensive low-pressure area subsequently developed in the eastern Pacific Ocean at low latitudes with the pressure remaining abnormally low in that region from 1–14 February. A more normal pattern was reestablished for a short period in the Pacific but beginning again about 22 February and continuing into the middle of March, pressures were again predominantly low at low latitudes in the eastern Pacific. In general terms, then, once the blocking action eddy is established it continues to exert its controlling influence on the over-all circulation pattern at least until the disturbing eddy is dissipated.

13. Blocking in the mid-Pacific area

The investigation of Atlantic blocking described in the preceding paragraphs has shown the probability of downstream effects originated by the introduction of a disturbance in the form of a blocking high-pressure cell. In order to study this downstream effect a region was selected for investigation for which a large downstream area of data was available. The region selected is bounded by the longitudes 180° and 170°W. The strong west-wind belt normally lies between 50°N and 45°N latitude in this sector of the hemisphere so that these latitudes form the other boundaries of the region. In the Atlantic criterion the restrictive requirements regarding high pressure departures have probably excluded some cases which should be included in the blocking action category. These restrictions are necessary in order to produce disturbances of sufficient strength to carry through the blank area of data in Siberia and be observable in the Pacific region where data are available. When the immediate downstream effects of the disturbance are observable the objective criterion may be weakened considerably and the disturbance may still be expected to produce meaningful pressure deviations at least for a short distance downstream.

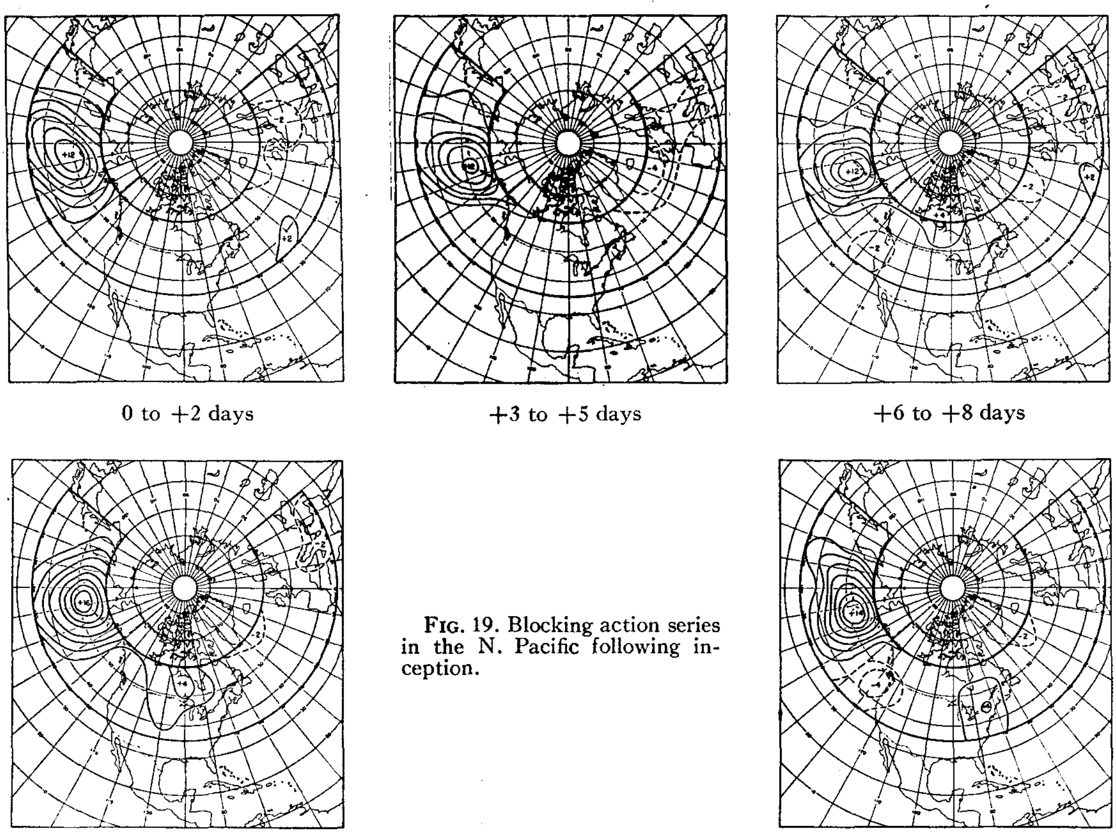
The following objective criterion was therefore formulated for the Pacific blocking: "The total pressure departure of the four intersections formed by 180° and 170°W with 50°N and 45°N must average 5 mb above normal for a two-day period. This average must be maintained for at least three days." This is obviously a weak definition. It could allow negative departures on any given day providing the positive departures on the preceding and following days overbalance the negative values to the extent of an average of +5 mb for the two-day period. This is for the purpose of allowing a minor trough passage after which the surface pressure is re-established. In terms of the upper-level flow, the controlling wave pattern may still exert its influence on the circulation even though a minor low-pressure trough has passed.

Another illustration of the weakness of the definition lies in the small area it encompasses. The center of the blocking cell may lie anywhere outside of the given area and the definition may still be satisfied. An obvious advantage of the definition is that it allows the inclusion of a larger number of cases so that some of the disadvantages mentioned above tend to be smoothed out. The inception date and duration of the cases which satisfy this criterion are listed in table 2.

The downstream effects of this disturbance are shown in fig. 19 in the form of three-day mean maps representing a composite picture of the cases listed in table 2. Solid lines enclose areas where the average daily pressure during the three-day interval is 2 mb,

TABLE 2. Inception dates and durations (in days) of mid-Pacific blocking actions.

Even years Date	Duration	Odd years Date	Duration
21 Jan. 1900	6	18 Jan. 1899	15
29 Jan. 1904	31	7 Feb. 1899	4
5 Jan. 1906	5	30 Dec. 1900	6
12 Jan. 1906	12	3 Feb. 1901	14
23 Feb. 1908	7	2 Jan. 1903	7
28 Dec. 1909	36	7 Feb. 1903	9
1 Jan. 1912	9	6 Jan. 1907	12
4 Jan. 1914	5	25 Jan. 1907	12
19 Jan. 1914	10	5 Jan. 1909	32
26 Dec. 1915	37	3 Jan. 1911	37
11 Feb. 1918	7	1 Jan. 1913	16
8 Jan. 1920	22	9 Jan. 1915	6
4 Feb. 1922	25	18 Jan. 1915	8
30 Jan. 1924	19	29 Jan. 1915	22
31 Dec. 1929	6	14 Jan. 1917	6
25 Jan. 1930	20	2 Jan. 1921	24
9 Jan. 1932	7	10 Feb. 1921	5
9 Jan. 1934	8	9 Jan. 1923	19
1 Feb. 1936	7	12 Feb. 1923	12
21 Jan. 1938	25	9 Jan. 1925	13
		7 Feb. 1925	13
		21 Feb. 1929	5
		11 Jan. 1933	20
		22 Feb. 1933	24
		6 Jan. 1935	15
		24 Feb. 1935	10
		6 Jan. 1937	13
		23 Jan. 1937	26



0 to +2 days

+3 to +5 days

+6 to +8 days

+9 to +11 days

+12 to +14 days

FIG. 19. Blocking action series in the N. Pacific following inception.

4 mb, *etc.* above normal and which are common to both the odd year and even year samples. Dashed lines enclose similar areas of below normal pressure. In order that no extraneous influences accompanying the breakdown of this process would be introduced, individual examples were eliminated from consideration as soon as the criterion was no longer satisfied, although it is probable that the effect of this breakdown is propagated downstream rather slowly and is not immediately apparent at distances far downstream. This is in contrast to the procedure used for the Atlantic blocking action cases. In this way the number of cases considered is gradually reduced from about twenty for the period 0 to +2 days to eight for the last period (+12 to +14 days) for the even year examples and a similar reduction for the odd year examples.

Fig. 19 shows the gradual establishment of a consistent wave pattern downstream becoming well established in the period 12–14 days following inception. The pressure ridge observed in eastern Canada in this period has been found to exist for periods longer than 12–14 days and represents the first downstream crest of the stable wave pattern. From this surface pattern the upper-level flow may be inferred. The wave length for this pattern is about 100°. While the pressure abnormalities observed are not large, account should be taken of the weakness of the objective criterion. Larger deviations could be produced with a more rigid definition. Because of the decreased number of cases this series was terminated at +14 days.

An independent illustration of this process occurred in March–April 1948, as mentioned in section 2. Beginning about 8 March pressures in the critical area ranged generally above normal and the wave pattern downstream resembled the pressure departure pattern shown with the principal differences occurring following marked pressure fluctuations in the controlling area. Beginning again at the end of March and continuing through most of April, pressures in the critical area were well and consistently above normal. During this period the above pressure departure pattern was followed closely, breaking down only after the breakdown of the Pacific blocking high.

Stable wave patterns of the form shown in fig. 19 have been studied by numerous investigators. Rossby [15] and co-workers have, through theoretical treatment and under the assumptions of no horizontal divergence and no vertical motion, arrived at an expression for the stationary wave length of stable long waves in the atmosphere:

$$L^2 = 4\pi^2 U/\beta, \tag{7}$$

where L is the wave length, U the zonal wind speed and β the northward rate of change of the Coriolis parameter. In a recent paper Cressman [2] has in-

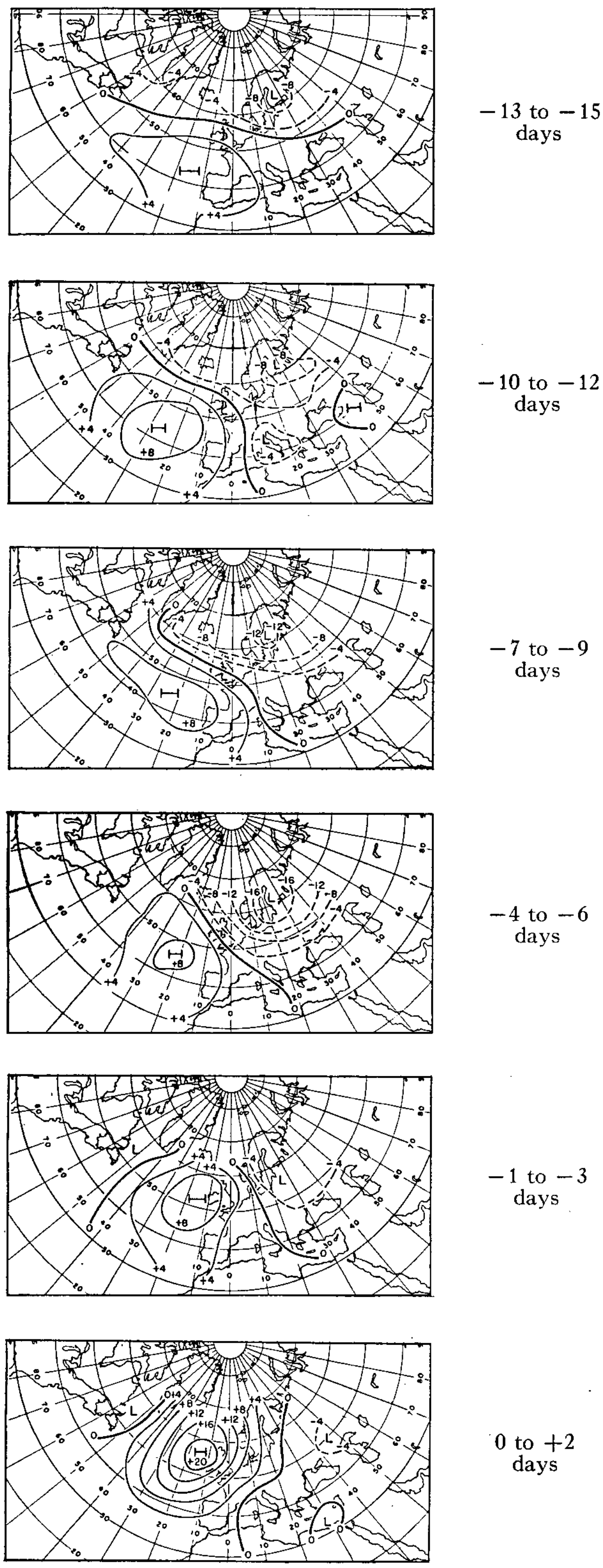


FIG. 20. High-latitude blocking action series in the Atlantic preceding inception.

vestigated empirically a number of properties of these long waves. He has found an average wave length of about 78 degrees of longitude for the stationary waves on the basis of upper-level flow patterns. The stationary wave lengths found in the study of blocking action processes are considerably longer, being of the order of 90–100 degrees of longitude. These blocking action processes are normally associated with closed cyclonic and anticyclonic centers in the upper-level flow pattern and, as Cressman has pointed out, this characteristic makes the use of the above formula questionable due to the difficulty in the determination of the mean zonal wind speed U. Cressman himself has eliminated these upper-level flow patterns from his study so that no conflict exists between the two results. The larger amplitude, more extreme stable waves in the atmosphere are apparently characterized by longer wave lengths than the less disturbed stationary stable waves.

14. Energy propagation in the atmosphere

The problem of the rate of propagation of energy in atmospheric waves has recently come into considerable prominence due to the work of Rossby [16] and others. Some information on this subject can be gained from the preceding discussions of blocking action processes by considering the intrusion of a blocking high into the normal westerly circulation as an energy source. The use of three-day mean maps and the consequent elimination of the smaller scale pressure fluctuations serves also to eliminate most of the effects of propagation of energy downstream as hypothesized by Rossby, at a group velocity exceeding the simple wave velocity. The three-day mean maps then show the propagation of energy associated with the downstream development of a stationary wave pattern.

This is most clearly shown in fig. 19 for the mid-Pacific blocking case where the disturbance reaches the northeastern Canadian area twelve to fourteen days after the inception of the disturbance in the Pacific. This is a distance of about 100 degrees of longitude traveled at the latitude of about 60°N so that its speed is apparently less than the mean zonal flow. The high-latitude Atlantic blocking case is less clearly defined since the low pressure forming in the Pacific 12–14 days following inception is assumed to be a secondary effect of the production of high pressure in Alaska at latitudes outside of the network of data. However, the rate of downstream propagation is comparable to the Pacific example when consideration is given to the more northerly latitude traversed through Siberia and Alaska.

15. The formation of the blocking-action eddy

The general characteristics of the pressure field leading to the development of a blocking high in the northeastern Atlantic area are shown in fig. 20. These charts represent the average pressure departures for the even-year cases listed in section 12. The figures show clearly that the principal origin of the blocking-action high in this region is the Atlantic subtropical high-pressure cell and that the principal dynamic development is an abnormally high zonal flow directly to the north of the subtropical cell. This high zonal flow increases through the early stages of the development and is deflected gradually to a more northwesterly flow thus representing a gradual change from a strong zonal flow to a strong meridional flow. The anticyclonic deflection and the subsequent formation of the low-pressure area in western Europe is explainable in a qualitative sense by the zonal flow acquiring anticyclonic vorticity in the area of horizontal divergence in extreme western Europe. This is not the only means by which the blocking high may form. Another method will be discussed briefly in a later paragraph. One important fact should be pointed out in connection with this case. The start of the development takes place at least 12–15 days prior to the actual formation of the high cell at high latitudes. If one attempts an explanation of the formation on the basis of extra-terrestrial radiation, the change in that radiation should occur over a comparable period. This would appear to rule out such short-period phenomena as solar flares, daily fluctuations in magnetic numbers, *etc.*

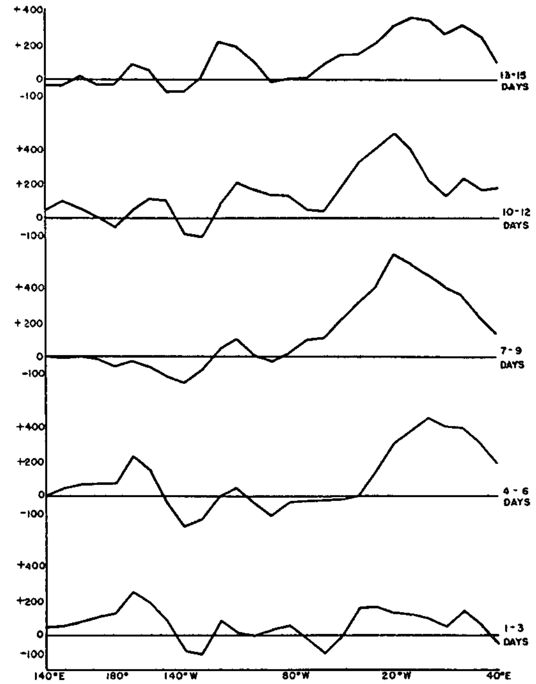


Fig. 21. Pressure-gradient anomalies (50°N minus 60°N) preceding blocking action for the Atlantic blocking action series.

The mean pressure gradient between 50°N and 60°N for the cases mentioned above is investigated in more detail in fig. 21. The large magnitude of this gradient corresponding to the strong zonal flow occurring in the Atlantic area is clearly shown. The only suggestion of a consistent, meaningful pattern upstream from this strong flow is that occurring between fifteen to seven days prior in the neighborhood of longitude 110°W. This suggests the existence of a stable wave of length 90–100 degrees upstream from the subtropical high cell which may aid in its development into the blocking stage.

An explanation for the dynamical development consisting of the abnormal extension of the subtropical high cell into the region normally occupied by a westerly flow involves the accumulation of heat in the region of the subtropical cell. If the characteristics of the general circulation are such as to lead to an accumulation of heat in a given area, the blocking action process is postulated as the mechanism by which the circulation adjusts itself in order to redistribute this heat. This corresponds to the results of the seasonal studies of section 6. The process may be visualized as follows. The accumulated heat in the subtropical high cell causes an increase in the latitudinal temperature gradient and consequently an increase in the zonal flow. This increased zonal flow is brought about by the transformation into kinetic energy of a small portion of the potential energy which exists in the form of the abnormal latitudinal temperature gradient. The mechanism by which this excess heat can be transported northward and redistributed is assumed to be the pressure disturbance superimposed on the zonal flow. The additional meridional velocity component brought about by this disturbance is apparently sufficient to allow the flow of heat northward and to lead to the formation of the high pressure at the higher latitudes. At the high latitudes the heat is then dissipated by lateral mixing around the periphery of the blocking high as discussed in section 11. The maintenance of the high pressure at these high latitudes is considered in section 2.

A study has been made of the sea-level air temperatures in the area bounded by latitudes 50°N and 40°N and by longitudes 30°W and 10°W. This corresponds to the region occupied by the subtropical high cell in fig. 20 (-10 to -12 days). The temperatures were taken from ship reports in the area and are necessarily subject to some inaccuracy. The results of the study show the temperatures averaging 1–2F above normal throughout the area during the period 21–6 days prior to inception, being gradually dissipated during the following six-day period. This represents a considerable amount of excess heat when summed over the entire area and lends some weight to the above explanation.

The pressure departures for three-day intervals prior to the inception of the twenty even-year cases of mid-Pacific blocking have been investigated in a manner analogous to fig. 20. The characteristic pressure pattern leading to this type of blocking is a strongly disturbed state of the circulation with large below normal areas in both the Atlantic and Pacific

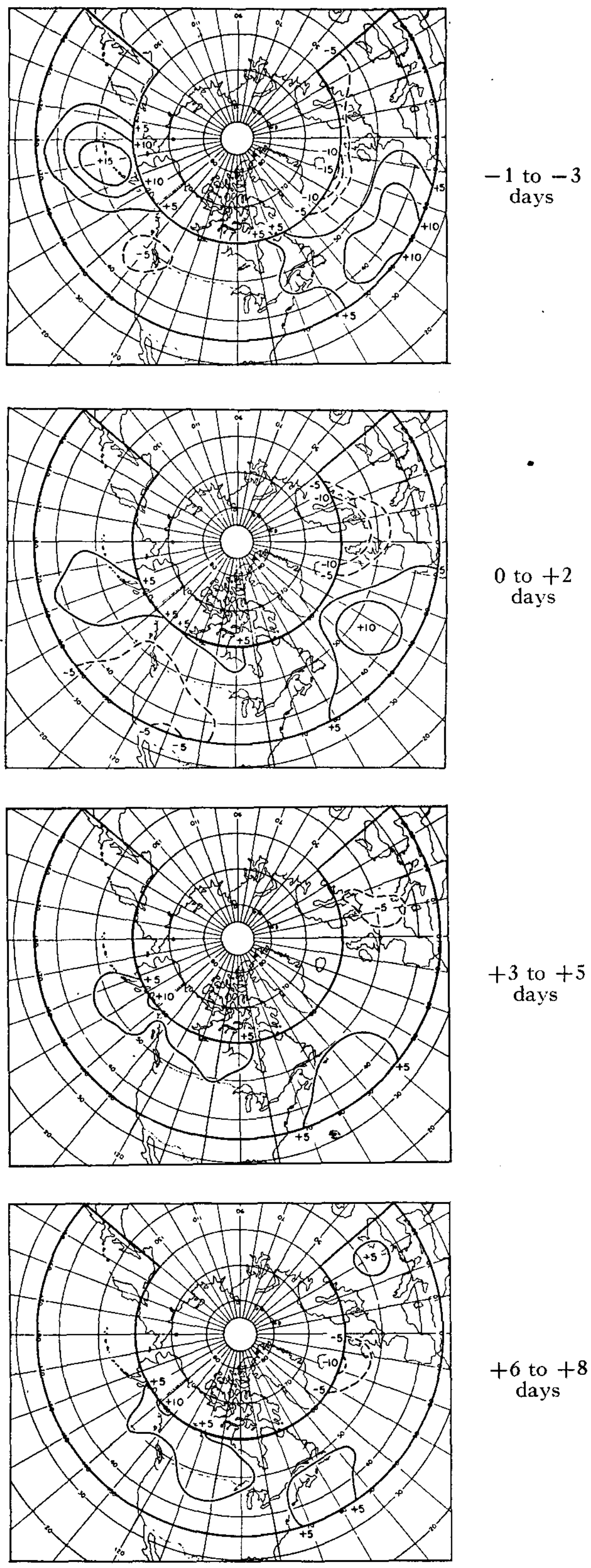


FIG. 22. Dissipation of blocking action in the N. Pacific.

Oceans. During the consecutive three-day intervals, these areas drift slowly toward the northeast. The origin of the high cell which subsequently becomes the blocking high is in Siberia, moving out to the blocking area in a general west to northwest flow. There is no evidence for the gradual northward extension of the belt of the subtropical high as was the case in the Atlantic. These results suggest that the circulation is already in a highly disturbed state prior to the inception of this type of blocking and that a slight readjustment of the long-wave patterns such as is discussed in section 16 results in the destruction of the blocking in one area and its reformation in the new area. This type of formation would then be considered as a secondary effect rather than as the introduction of a primary disturbance. It should be mentioned here that a few of the high-latitude Atlantic cases are apparently formed in this way, the high cell coming down from the Greenland Sea area. They show little relation to the Atlantic subtropical high cell and should be considered as secondary in origin.

16. The dissipation of the blocking-action eddy

Only a small amount of information has been obtained as yet on the mechanism of the dissipation of the blocking high cells. Aside from the natural degradation to a zonal type flow through the dissipative action of migratory systems as discussed in section 2, interaction between blocking highs in two different regions might lead to the destruction of the weaker if it were not located at the correct wave length from the stronger. This process is shown in fig. 22 and represents the deterioration of about one half of the cases of mid-Pacific blocking for the even years. Although it must be considered as only one breakdown process, it must represent one of the most common types. The pressure departures are again presented in groups of three days. The first chart in fig. 22 shows the average daily departures for the final three days of blocking in the mid-Pacific. In the Atlantic the subtropical high cell is considerably above normal in pressure with an abnormally high zonal flow to the north, a condition strikingly analogous to that leading to the formation of the blocking high in the northeastern Atlantic, although displaced slightly to the west. The Atlantic subtropical cell continues its analogous development in the second chart while the mid-Pacific blocking cell deteriorates considerably in intensity. The Atlantic high cell does not remain as an important blocking cell in the last two charts of fig. 22 but a new, important high cell apparently develops in northwest Canada.

The mechanism visualized in this transition is as follows: As shown in section 14 the first stable wave crest downstream from the mid-Pacific blocking high is located in northeastern Canada. Due to the mechanism suggested in section 15 a blocking cell develops in the mid-Atlantic region for reasons independent of the action of the Pacific blocking. The position is out of phase with the stable wave pattern formed downstream from the Pacific blocking cell. The mid-Atlantic cell tends to set up its own stable wave pattern independent of that formed by the Pacific cell. If there is greater energy associated with the Atlantic disturbance than with the Pacific, then the Atlantic pattern will dominate in the general circulation and the original wave pattern will be destroyed.

17. Conclusion

Various empirically observed facts in regard to the blocking-action process have been presented in this study and some attempt has been made to offer tentative explanations for these facts. An adequate theory is not yet available but the observations presented here must be taken into account in the formation of any future theory. It is hoped that enough information has been presented to demonstrate the extreme importance of these processes in the anticipation of weather trends covering periods of a few days to three to four weeks and that this information will serve to stimulate further research in this field.

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