An Aerological Study of Zonal Motion, its Perturbations and Break-down

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Abstract

This paper contains an aerological analysis of a series of rapid-moving frontal waves associated with a well developed zonal current over central Europe. The gradual destruction of this zonal flow through a retrograde blocking "wave" is described. The relationship between the blocking process and the deepening of series of waves approaching the blocking zone from North America is discussed. The observational data and conclusions from this aerological study are presented for the purpose of rendering some assistance to theoreticians investigating atmospheric wave motions, by providing a few numerical values for certain characteristic parameters of these waves and by calling attention to a factor which appears to play a significant rôle in the deepening of certain types of extra-tropical wave cyclones, viz. the variation with longitude of the basic current pattern.

Introduction.

The objectives of the researches described below are threefold. The well-known investigations conducted by WILLETT (1944), NAMIAS (1947) and others during the last ten or twelve years have shown that during the winter season the general atmospheric circulation in middle latitudes has a tendency to fluctuate between two extreme states, characterized by the predominance of either zonal or meridional motion. The successive stages in the break-down of a well-defined zonal circulation into a complicated pattern of meridional currents has been described on several occasions, but mostly with the aid of consecutive overlapping mean charts in which time averages of the sea level pressure distribution during five or seven day intervals have been plotted. In the present paper a particularly striking case of such a break-down of a fully developed west wind belt into a series of quasi-stationary cyclonic and anticyclonic vortices will be described with the aid of ordinary synoptic charts for the 500 mb level and by means of other synoptic tools. As a by-product the case investigated provides some information concerning the manifestations and westward speed of a typical "blocking wave". Of late, the great importance of this process has been recognized by synoptic and theoretical meteorologists alike but as yet no adequate synoptic description has been published.

The second objective has been to provide a few empirical data for determination of the numerical values of the characteristic parameters of typical fast-moving frontal waves. The total number of investigations of polar front disturbances published since J. BJERKNES' epoch-making paper appeared in 1919 is certainly very great. However, in the light of recent experiences it appears safe to conclude that this collection of case histories is extremely heterogeneous, including typical frontal waves as well as dynamically quite dissimilar major storms associated with the deepening of planetary wave-troughs in the upper west-wind belt. Furthermore, in most cases the systems investigated have been treated as solitary phenomena; with such a starting point one is not likely to arrive at numerical data suitable to serve as an adequate basis for a realistic *wave* theory. The present study of wave cyclones makes no claim to pry as deeply into the structure of the individual wave cyclone and its weather manifestations as do some of the erlier studies by BERGERON, SWOBODA (1924) and others, but a deliberate effort has been made to bring out those factors which from our present point of view appear likely to be of importance in the dynamic interpretation of these waves.

The third objective of our studies has been to shed some light on the nature of the frontal waves and particularly on the occlusion process. Until recently, it has been assumed that the important changes in the flow pattern of the upper troposphere, in particular the break-down of the zonal circulation, are produced by the occlusion of frontal waves. While it is true that a marked deepening and cutting-off of an upper trough in most, if not all cases, is accompanied by a marked deformation of the fronts at the ground in a manner which perhaps best is described as an occlusion process (see Fig. 34 in NYBERG 1945), cases have been observed in which a whole series of disturbances of the short and rapid-moving wave type occlude without bringing about significant modifications of the upper flow pattern. Such a series of occluding waves is described below. It will be shown that the gradual retardation of the occluding waves is associated with a pronounced eastward decrease in the strength of the zonal "steering" current. One is thus led to two conclusions, one definite and one tentative: first, that the process of occlusion by itself is insufficient to bring about significant changes in the upper flow pattern and secondly, that the occlusion of fast-moving frontal waves just possibly may be the consequence of the rapid decrease downstream of the basic current in which they are embedded. If this view point is correct the occlusion process would be comparable to the amplitude increase and breaking of ordinary sea surface waves which occurs whenever the phase velocity of these waves is reduced, by decreasing water depths, below a prescribed critical value. The occlusion would thus represent an instability of form only; the total energy available to bring about these form changes would merely be that of the wave motion itself. This situation differs fundamentally from the one prevailing in those unstable waves which are capable of withdrawing potential and kinetic energy from the basic barocline current system in which the waves are embedded.

In a further attempt to shed some light on the nature of frontal waves a comparison has been made between the family of frontal waves described below and a group of fast-moving waves travelling eastward across the Rocky Mountains in the northwestern part of the United States and studied by HESS and WAGNER (1948). The comparison brings out a striking agreement in the upper-level characteristics of these two series of disturbances, though the surface manifestations of the waves for obvious reasons are quite dissimilar. Further investigations may be required before final and definite conclusions can be drawn but the comparison nevertheless suggests that the short progressive frontal waves observed in the lower tropospheric layers, in spite of their obvious primary importance in as far as weather phenomena go, must be regarded as induced manifestations of relatively short, rapid-moving upper level waves associated in an as yet unknown manner with the jet stream zone and its strongly developed solenoidal field.

1. Zonal Circulation and Frontal Waves.

By R. BERGGREN.

During the second week of February 1948 the weather in western and central Europe and over southern Scandinavia was controlled by a series of fast-moving frontal waves. Large amounts of precipitation and high mean temperatures during the week were observed at most of the stations. The waves moved from the Atlantic over the British Isles and down to the Black Sea with a great regularity. The pressure centre and the occlusion point for each disturbance kept within a belt of about 500—600 km width. The narrowness of this strip (shadowed in Fig. 1) indicates the similarity between the paths of the individual disturbances. Figure 2 demonstrates the pe-

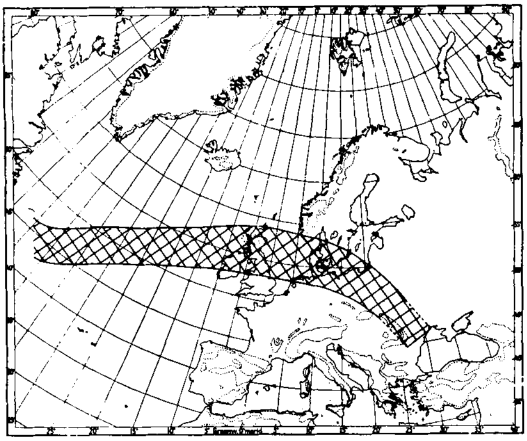

Fig. 1. Cyclone track belt for February 6—12, 1948.

riodicity of the waves. The slanting lines, which represent the consecutive positions of the wave centers, have approximately the same inclination, i.e. the translational velocity from west to east was nearly constant from wave to wave, which could not have been the case unless the upper, steering current remained nearly stationary throughout the period. The positions of the wave centers in this figure were determined from the 0700 GMT positions of the occlusion points. The diagram as reproduced includes five of a total of seven well defined waves observed before the basic current pattern broke down. With the aid of this diagram it is possible to compute an average period for the frontal waves. One finds for this period 31 hours at 30° W, 32 hours at 0°, and 35 hours at 20° E, and for the entire region an average period of 33 hours. The average wave speed over the Atlantic and Western Europe was about 70 km p.h. and the average wave length in the same region 2,500 km. Over Central and Eastern Europe the waves were retarded, as indicated by the break in the lines of constant phase.[1]

In this region the wave length was reduced to about 1 600 km and the speed of propagation to 50 km p.h. If we compute the frequency, or the ratio c/L, where c is the speed of propagation and L the wave length, one obtains, for the western part of the tracks, $2.8 \cdot 10^{-2}$ hours^{-1}, and for the eastern part $3.2 \cdot 10^{-2}$ hours^{-1}. As the values for c and L are smoothed values it appears permissible to conclude that the frequencies (and periods) are approximately constant.

To demonstrate the quasi-stationary character of the upper "steering" current during the period under investigation a mean chart (Fig. 3) has been computed from the seven individual 500 mb morning charts for the time interval February 6—12, inclusive. This chart may be considered as typical of those cases in which the weather over western Europe and Scandinavia is controlled by frontal waves coming in from the west or southwest. The (sea level) frontal system in Fig. 3 is taken from the sea level chart for the morning of February 8. The general orientation of this frontal zone agrees quite well with that of the mean current at 500 mb. Also this chart shows that the distance between consecutive waves decreases eastward. If the mean chart in Fig. 3 is compared with the individual charts for the same period one finds that the changes occurring during this week are of a minor nature. As an example of an individual chart we have reproduced the morning chart for February 9, in the middle of the period (Fig. 4 a). There is good agree-

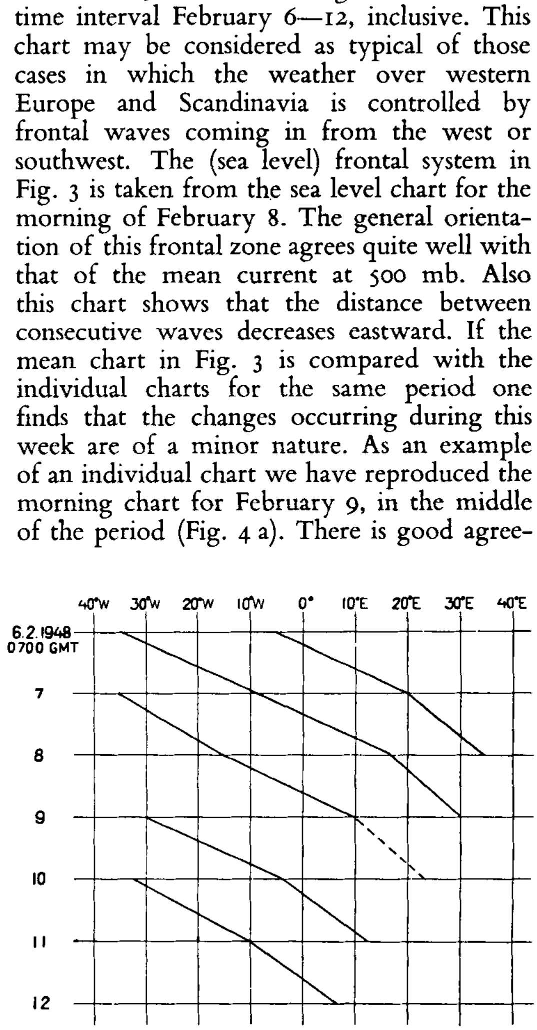

Fig. 2. Phase diagram for February 6—12, 1948, showing successive positions of the occlusion points of five consecutive frontal waves.

[1] The westward motion of the break in the curves of constant phase in Fig. 2 may be taken as an indication of a westward displacement (retrogression) of a decrease in the zonal circulation. This retrogression may possibly be related to the "blocking action" described by NAMIAS and CLAPP (1944). The speed of the westward motion, 7—8 degrees of longitude per day at 50° N, is in good agreement with the results obtained by these authors.

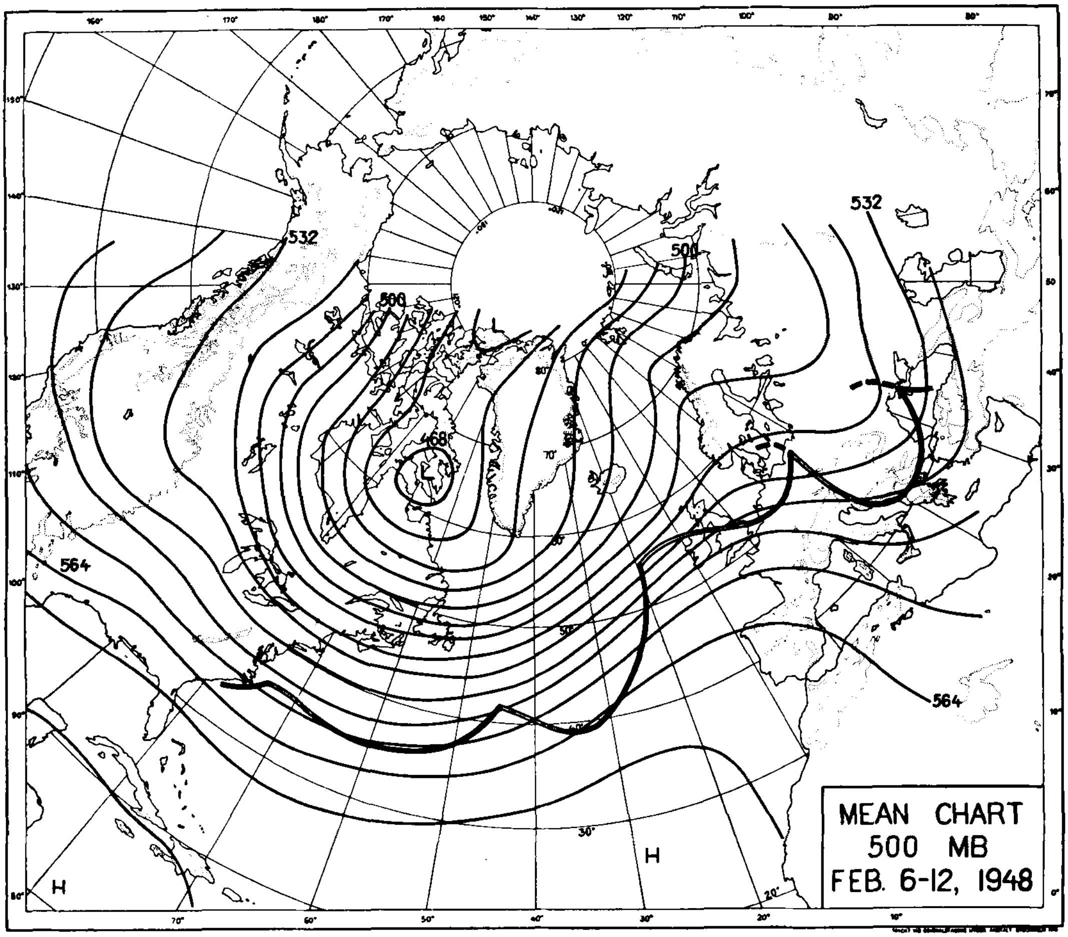

Fig. 3. Mean 500 mb chart for February 6—12, 1948. The frontal system is taken from the sea level chart for February 8, 0600 GMT. Contour heights given in geodynamic decameters.

ment as far as the position of the main ridges and troughs are concerned. Also with respect to the steering current there are only minor differences between the two figures.

Fig. 4 a brings out certain other noteworthy features. A comparison with the sea level chart for the same day (Fig. 4 b) shows that the fairly intense wave cyclone observed that day over the North Sea was reduced to a minor trough over Scotland at the 500 mb level (it should be noted that there is a slight difference in time between the two charts).

Fig. 4 a shows that at the 500 mb level extremely cold air had accumulated in the upper level trough east of Hudson Bay. On this, as well as on several subsequent days, an isotherm for — 50° C can be traced with the aid of the radiosonde data from the north-eastern part of North America. It will be shown in the next section that this cold air plays an important rôle in the subsequent break-down of the zonal circulation.

The sea level chart in Fig. 4 b is typical of the entire period. The well developed high over Russia with a ridge over central and southern Scandinavia is important for later events. It is this system which, a week later, brings down very cold air over Central Europe.

An examination of the variations of the geopotential of the main isobaric surfaces shows that the wave disturbances which

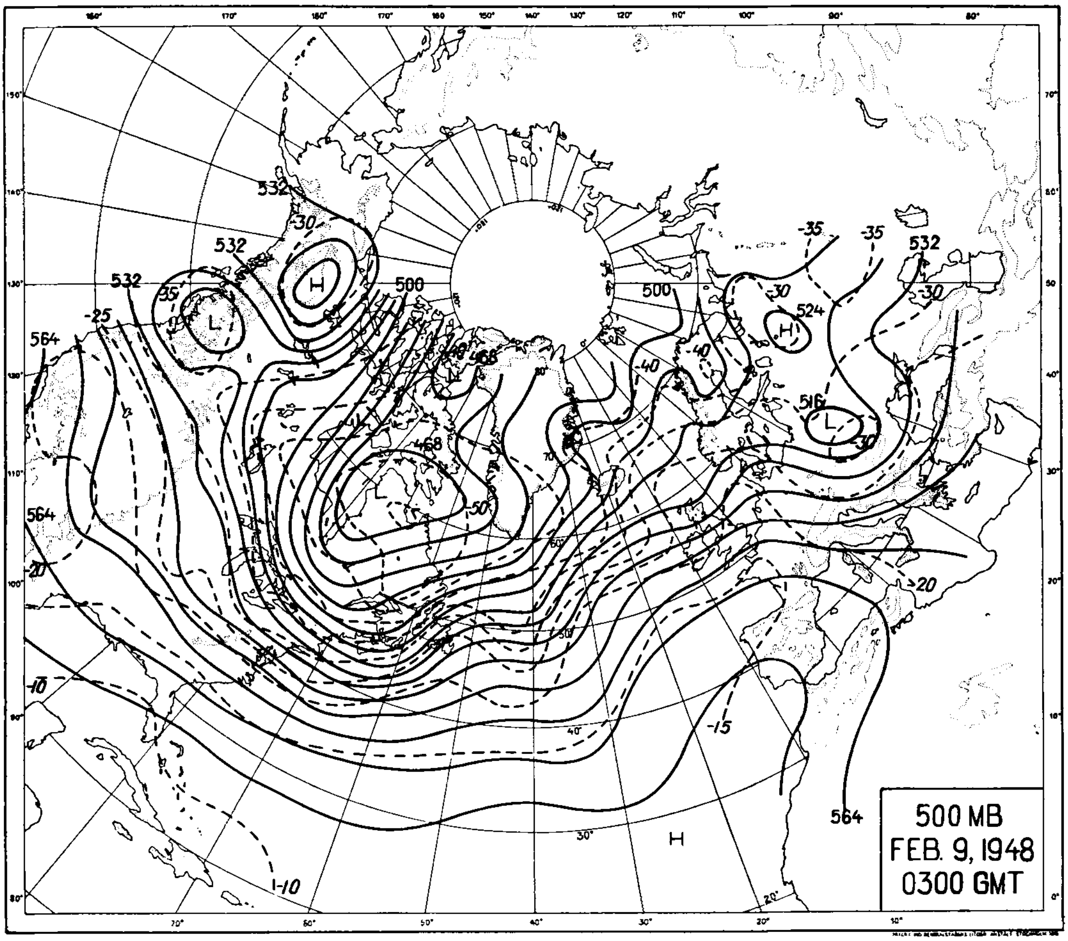

Fig. 4 a. 500 mb chart for February 9, 0300 GMT. Heights in geodynamic decameters.

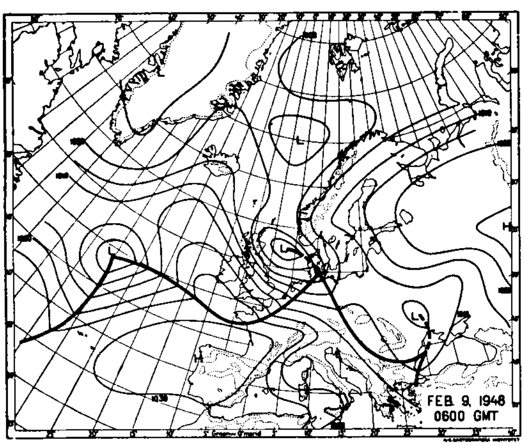

Fig. 4 b. Sea level chart for February 9, 1948, 0600 GMT.

passed Great Britain during this period can not be considered as shallow phenomena. Fig. 5 gives the geopotential variations of the 300, 500 and 700 mb surfaces (and the variations of the sea level pressure) over the radiosonde station Aldergrove in northern Ireland. The curves demonstrate that the individual waves extended through the entire troposphere, though the particular disturbance that passed Aldergrove on February 7, 0900 GMT was considerably weaker than the others and did not acquire full intensity until it reached Germany. *In spite of the great depth of these disturbances and in spite of the fact that all of them occluded they did not bring about any significant major changes in the upper flow pattern.* Aerological experience indicates that this re-

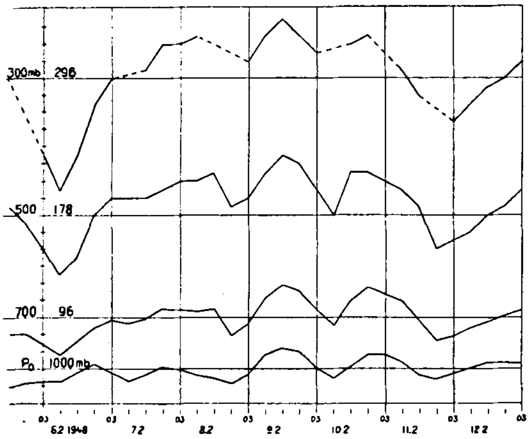

Fig. 5. Diagram showing the height variations of the main isobaric surfaces, 300 mb, 500 mb, and 700 mb. P_0 is the mean sea-level pressure in mb. Ordinate intervals in units of 200 feet, and for the sea level pressure 10 mb.

sult is characteristic of fast-moving frontal waves approaching Europe from the West.

A series of N—S cross sections have been drawn, mainly through the British Isles. Fig. 6 a and 6 b contain a section for February 12 from Cornwall to the Shetland Islands with a characteristic picture of a strong westerly air current. Fig. 6 a shows isentropes (dotted lines) and isopleths (full lines) for the *observed* wind speed perpendicular to the section, i.e. representing a wind direction of 290°. Fig. 6 b shows isentropes (full lines) and isotherms (dotted lines). The wind maximum (jet stream) is situated at the 250 mb level, immediately beneath the tropopause, with the strongest vertical increase of the wind observed between the frontal layer and the 400 mb level. In this particular case the frontal layer is well-defined and seems to extend into the lower stratosphere over the Shetlands. Over southern England there is a weak subsidence inversion. A strong horizontal north-south

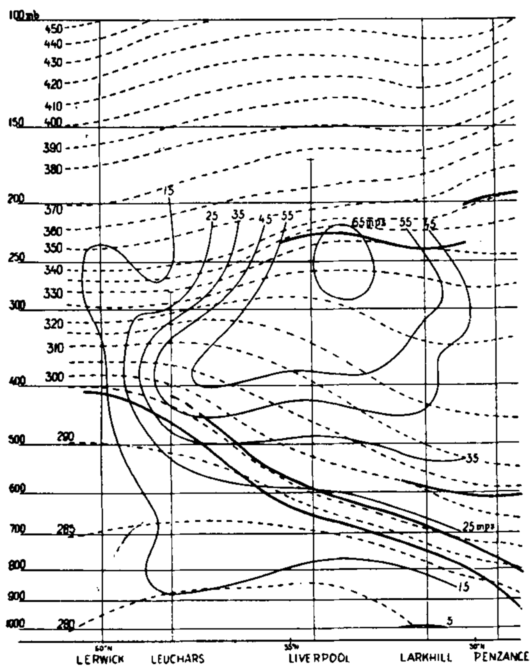

Fig. 6 a. Cross section for Feb. 12, 1500 GMT, 1948, showing isentropes (dotted lines; potential temperature in °K) and isopleths for observed wind speed (full lines; speed in m.p.s.) perpendicular to the section. The heavy line represent the frontal boundaries and the tropopause.

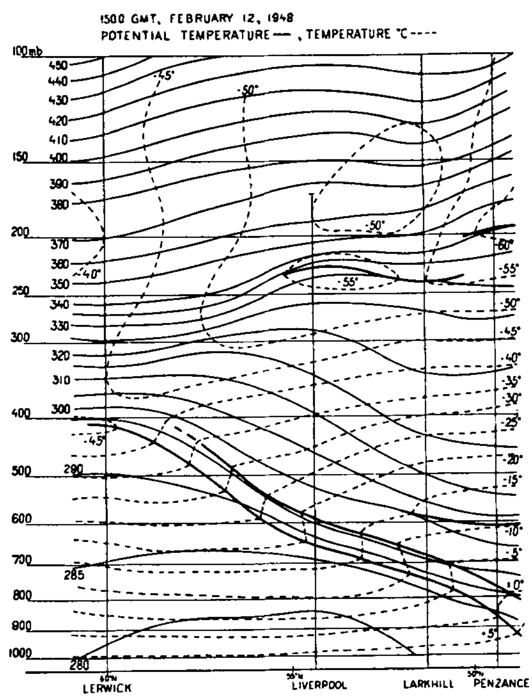

Fig. 6 b. Cross section for Feb. 12, 1948, 1500 GMT, showing isentropes (full lines; potential temperature in °K) and isotherms (dotted lines; temperature in °C). The heavy lines represent the frontal boundaries and the tropopause.

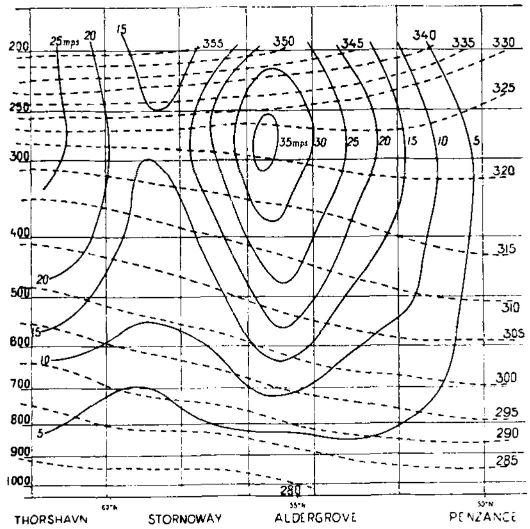

Fig. 7. Mean cross section for Feb. 6—12, 1948, showing isentropes (dotted lines; potential temperature in °K) and isopleths for computed geostrophic wind speed (full lines; speed in m.p.s.) perpendicular to the section.

temperature gradient exists between 52° N and 58° N. The figure shows that the solenoids are not concentrated exclusively to the frontal layer but that they are strongly developed also in the warm air.

The mean cross section from Thorshavn on the Faroes to Cornwall (Fig. 7) gives an interesting picture of the mean distribution during this period of the westerly wind (computed with the aid of the geostrophic assumption.) A rather well-developed wind maximum is formed a little farther to the north than in the individual section in Fig. 6. This mean maximum is situated on the central line of the trajectory belt in Fig. 1.

During the period under investigation the jet stream undergoes an interesting variation which appears as a fluctuation in the direction and strength of the current as well as in the height at which its center occurs. When a trough passes a particular station the wind in the jet shifts from southwest to northwest. At that time its speed reaches values from 50 % to 100 % greater than between troughs. The tropopause sinks to its lowest height as the trough passes and rises rapidly afterwards. At the same time the height of the jet stream center increases. This is indicated by the slanting isopleths at higher levels in Fig. 8 and by Table 1.

Table 1. Variation of the height of the jet stream center with time, as indicated by the observed wind speeds (in knots) over Larkhill (51° N, 2° W).

	February 10			February 12		
Time (GMT)	09	15	21	09	15	21
200 mb	—	69	51	—	70	83
250	70	54	**93**	—	110	**109**
300	82	105	90	100	**112**	96
350	87	**114**	69	**109**	91	89
400	**89**	104	60	102	91	76

The variation in the height of the jet stream center stands out quite clearly in this table. The fact that the table makes use of winds at constant pressure levels does not alter the

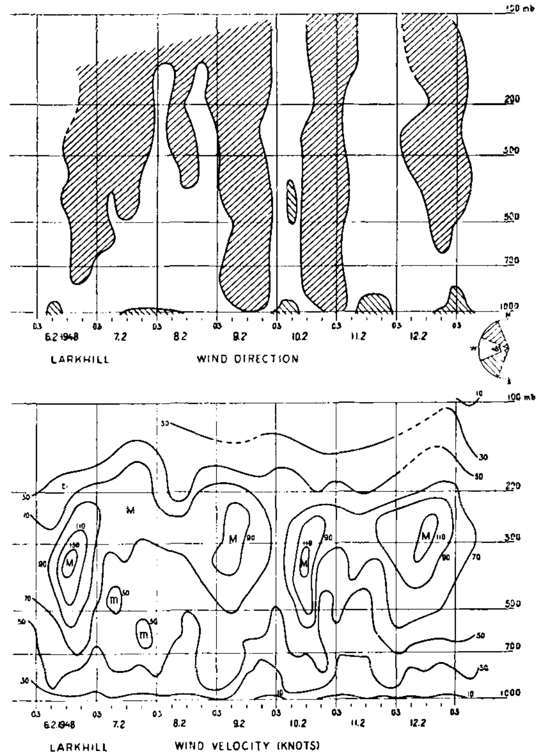

Fig. 8. The variations with time of observed wind speed (in knots) and wind direction over Larkhill (51° N, 2° W). *M* stands for maximum and *m* for minimum wind speed. The meaning of the different shadings in the wind direction diagram is given in the wind rose.

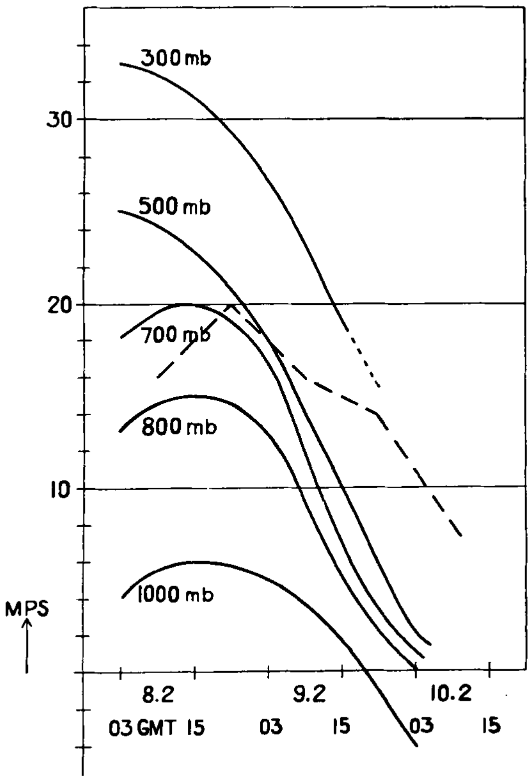

Fig. 9. Diagram showing the zonal wind speed (computed geostrophically) for different isobaric surfaces over an area of the same order of magnitude as a specially selected typical frontal wave. The area moves with the wave so as to occupy the same position relative to the disturbance at all times. The broken line represents the eastward velocity of the wave as determined from the twelve hourly displacements. (On February 10 the material for the 300 mb analysis was sparse, but the indicated drop of the geostrophic wind speed is beyond all doubt real.)

result significantly since the greatest shift of an isobaric surface at Larkhill during these two days is of the order of magnitude 60 m in 6 hours while the shift of the jet stream center is of the order of magnitude 1 km per 6 hours. The secondary maximum on February 10, 1500 GMT at 200 mb is connected with the upper part of a double tropopause. An examination of the jet stream thus gives the result that it undergoes great variations as the fast-moving frontal waves pass by both with regard to strength, direction and height.

In order to investigate the connection between the velocity of the disturbances and the upper current the diagram reproduced in Fig. 9 was constructed. A moving, quasi-rectangular area covering ten degrees of latitude and thirty degrees of longitude was chosen every twelve hours, in such a manner that the particular disturbance which on Feb. 9, 0700 GMT (Fig. 4 b) was situated over the North Sea always remained in the center of the area. The zonal wind component for this strip was then computed geostrophically and plotted in the diagram which gives the zonal wind components for the 1,000, 800, 700, 500 and 300 mb surfaces during the period Feb. 8, 0300 GMT to Feb. 10, 0300 GMT inclusive. The broken line gives the eastward velocity of the disturbance measured as a twelve hour mean and defined as the mean value of the displacement in longitude of the pressure center and of the occlusion point. The diagram brings out, in a striking fashion, two facts: firstly, the disturbance moves faster than the zonal wind up to a certain critical height at which the two speeds become equal (the intersection points between the broken and solid lines), and secondly, this height increases with time (= eastward) from about 2,000 m to more than 5,000 m. This indicates that the height of the steering current for this typical wave disturbance is dependent on time (= eastward longitude).

The objection might be made that the averaging procedure used in determining the lower level zonal current values in Fig. 9 disregards the difference between the air motion on the two sides of the frontal zone. For this reason the geostrophic wind components on the sea level chart in Fig. 4 b were measured. One finds, for the warm sector of the Atlantic disturbance 40—50 km per hour and about the same for the narrow warm sector in the disturbance over Germany. Twenty-four hours earlier the corresponding warm sector values were 40—50 km p.h. over the Atlantic and 30—40 km p.h. over Germany. Thus the basic result, viz. that the disturbances move at a speed considerably in excess of the speed of the basic current near the ground, remains valid.

Near the ground organized vertical motions must be of minor importance in as far as the horizontal temperature distribution is concerned, which therefore, in these lowest layers, must be controlled primarily by advection. It follows that the flat, sinusoidal open warm

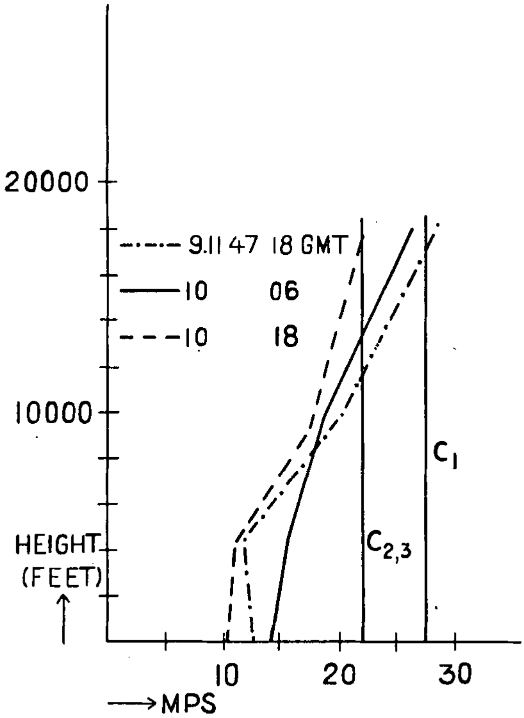

Fig. 10. Diagram showing the zonal wind speed (computed geostrophically) over an area surrounding a moving wave disturbance. C_1, C_2 and C_3 represent the eastward velocities of the disturbance between respectively 03 GMT and 09 GMT on Nov. 10, 15 and 21 on Nov. 9, and 15 and 21 on Nov. 10, 1947.

sectors of these fast-moving disturbances may be considered as a kinematic consequence of the fact that the phase velocity of the waves exceeds the speed of the basic current near the ground (Rossby 1942). In this type of wave motion the center position at the tip of the warm sector will be occupied by ever new segments of the front. However, once a small occluded frontal portion develops the individual fluid elements which make up the occlusion front are unable to move faster than the basic current and they must therefore lag behind the fast-moving wave center. In the formation of this occluded portion of the front it may well be that ground friction plays a significant rôle, as suggested by Scherhag (1938).

Short frontal waves moving faster than the zonal wind in the lower troposphere are frequently observed. We have included one additional example of this, from the second week of November 1947. On the morning of November 10 of that year the state of the circulation over Europe was rather similar to the well-known situation investigated by Bergeron and Swoboda (l. c.). A flat, fast-moving wave of the "Schnelläufer" type passed England during the night between the ninth and the tenth with a speed of about 100 km p.h. For this disturbance a diagram (Fig. 10) similar to the one in Fig. 9 was constructed, showing the zonal wind speed over an area of the same size and shape and with the same position relative to the moving disturbance as in the case discussed above. In Fig. 10 the zonal wind speed is plotted against height; the vertical lines marked c_1, c_2 and c_3 represent the speed of the disturbance during three six-hour intervals centered around consecutive maps. Also in this case the flat, open wave disturbance travelled at a far greater speed than the zonal wind in the lower troposphere. In fact, one must go to the 500 mb level to find the critical level where the zonal wind equals the maximum speed of the wave. The wind observations from Liverpool for November 9 and 10 (Fig. 11) show an interesting distribution of the wind with time and height. The wave passed Liverpool about midnight and at that time a marked increase in the wind at the higher levels was observed. The significant changes occurred above sixteen thousand feet! Thus the fast-moving open

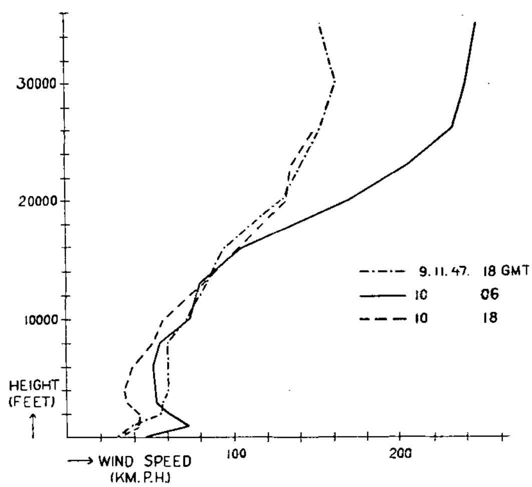

Fig. 11. Diagram showing the distribution with height of the observed zonal wind speed over Liverpool for November 9 and 10, 1947.

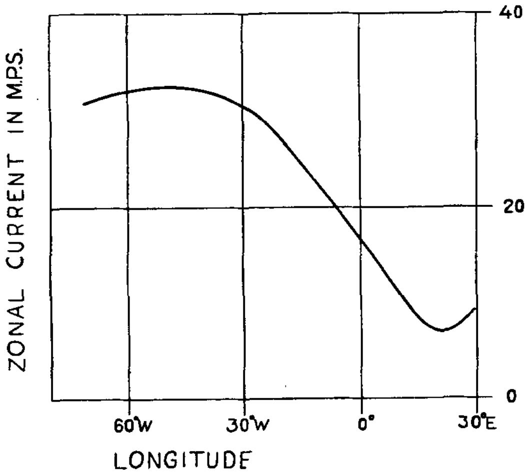

Fig. 12. Diagram showing the zonal geostrophic wind speed in 500 mb for different longitudes during the period Feb. 6—12, 1948. The speed is computed with the aid of the mean chart in Fig. 3 from the height difference between two points, situated 7.5 degrees of latitude on each side of the central line of the trajectory belt on Fig. 1, one computation being made for every ten degrees of longitude.

frontal waves which frequently have been considered as relatively shallow formations, are connected with great changes of the wind in the upper troposphere, while the corresponding changes in the lower troposphere (apart from the surface layers) are somewhat smaller and sometimes hard to detect with the aid of the present wide-meshed aerological station net.

We shall return next to the height variations of different isobaric surfaces over Aldergrove as recorded in Fig. 5. In view of the relatively long time interval between consecutive soundings it is impossible to fix exactly the phase lag between the upper and the lower level waves but it may be stated with certainty that this lag does not exceed 6 hours, or about 60°, i.e. the disturbances are almost vertical. The temperature variations at the various pressure levels have not been entered in the diagram as here reproduced, but the data show, as might be expected, that temperature and height (pressure) vary in phase at all levels except at the ground where inspection reveals that the correlation between the two curves is negative.

It was pointed out in connection with the analysis of Fig. 9 that the critical level at which the zonal speed of the wave disturbance and that of the basic current itself become equal shifts from about 2,000 m to above 5,000 m. In Fig. 12 we have entered the value of the zonal geostrophic wind at 500 mb in different longitudes as computed from the mean chart in Fig. 3. The values calculated refer to a strip with a width of fifteen latitude degrees and centered around the trajectory belt in Fig. 1. This computation indicates quite clearly that the basic zonal current itself, at least at the 500 mb level, decreases markedly eastward. The displacement of the critical level upward must therefore be interpreted as a tendency on the part of individual wave disturbances to continue their movement eastward at a somewhat less decelerated rate than that of the basic current in the lower and middle troposphere. This result lends some support to the impression that the factors controlling the behaviour of the wave disturbances are found in the upper portions of the troposphere.

The discussion presented above has brought out that the period of the waves is fairly independent of longitude and that their phase velocity decreases eastward less rapidly than the basic current in the lower and middle troposphere. With these experimental facts as a starting point it is of some interest to enquire into the stream line amplitude changes that would be observed in a simple transversal horizontal wave, superimposed upon a basic zonal current the speed of which (U) decreases slowly with time. If the transversal wind component is given by

$$v = v_0 \cos \frac{2\pi}{L} (x - ct)$$

the stream lines may be computed from

$$y = \frac{v_0 L}{2\pi U} \sin \frac{2\pi}{L} (x - ct)$$

where v_0 is the transversal wind velocity amplitude, L the wave length, c the phase velocity and y the coordinate northward. Since

$$c = \frac{L}{T},$$

where T is the period of the waves, it follows that

$$y = \frac{v_0 c T}{2\pi U} \sin \frac{2\pi}{L} (x - ct)$$

The experimental data suggest that c and T are relatively constant. If one makes the not unreasonable assumption that the kinetic energy of the wave motion remains fairly constant, the variations in v_0 must be negligible and it follows that the horizontal amplitude of the stream lines increases inversely as U decreases.

The rough calculation presented above should at any rate serve the purpose of suggesting that the amplitude increase of occluding waves might depend upon a down stream decrease in the strength of the basic current and that this factor should be explored in the final formulation of a theory for frontal waves.

We shall finally compare the results obtained from this study of European frontal waves with a somewhat similar study by HESS and WAGNER (1948) of certain atmospheric waves in the northwestern part of the United States. The waves studied by these authors occurred with a strong, anticyclonically curved westerly to northwesterly current in the middle troposphere over the states of Washington, Idaho and Montana, changing into a flat and cyclonically curved current over the Great Lakes region. Station level barograms for a series of points from Seattle eastward to Bismarck (North Dakota) were plotted for nine consecutive days from hourly altimeter settings. The resulting diagrams revealed the existence of a series of pressure waves which moved with great regularity at a speed of about 1 000 miles per day (= 67 km per hour) from west to east; this speed agrees well with the speed of 70 km per hour obtained for the particular case of wave propagation over western Europe studied by us. Aerological data revealed that the Pacific disturbances moved faster than the wind in the lower troposphere, the critical height sometimes reaching values of up to 6 km. Radio sonde data analyzed by HESS and WAGNER indicated that the disturbances west of the Rockies were very nearly in phase at all levels, with high positive correlations between pressure and temperature at all heights except near the ground where the temperature variations were insignificant. East of the mountains the disturbances moved in over low level air masses with a marked north—south temperature gradient; the waves then rapidly acquired the characteristic low level temperature asymmetry of waves on the front between polar air masses of Pacific and Continental origin.

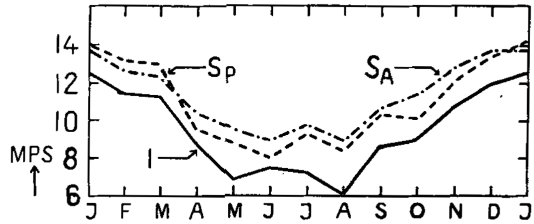

Fig. 13. Monthly average speeds of Alberta (S_A) and North Pacific (S_P) Lows as related to the 10,000 foot zonal index (I). (Reproduced from Namias 1947 a.) The zonal index is computed for a strip extending over North America and limited by 40° and 50°.

In view of the long (12 hours) time interval between consecutive soundings no attempt was made by HESS and WAGNER to determine the period of these waves but an inspection of the published barograms for the 6 km level suggest a period somewhat in excess of 36 hours but definitely much less that 48 hours.

The underlying flow pattern (a strong, anticyclonically curved basic current, increasing rapidly upward) is very similar in the situations investigated by HESS and WAGNER and by us. The speed of propagation and the period of the disturbances appear to be nearly the same.[1] In both cases a critical level well removed from the ground exists and in both cases pressure and temperature variations aloft are positively correlated. It is reasonable to conclude that the perturbations fundamentally are of the same character and origin. With these upper level similarities are associated marked dissimilarities in the frontal histories of the two

[1] The wave periods obtained from the two series studied by HESS and WAGNER and by us should not be considered as fundamental parameters. The phase velocity of this type of wave motion is intimately connected with the speed of the basic current, as may be seen from the extremely illuminating curves in Fig. 13. The advective motion evidently produces a "Doppler" effect which might be eliminated through the introduction of a properly defined orbital frequency.

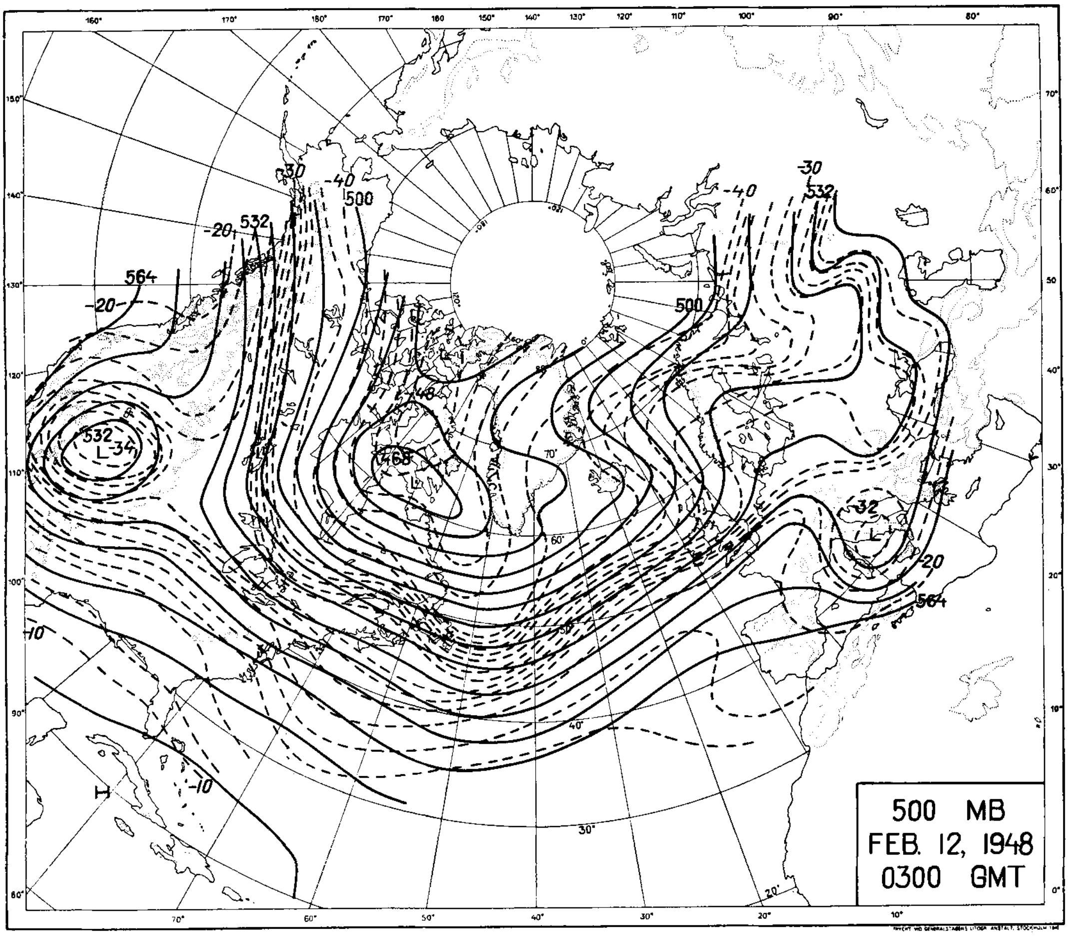

Fig. 14. Contours and isotherms on the 500-mb surface, 0300 GMT, February 12, 1948.

systems. The American disturbances of which the HESS—WAGNER series is a particularly striking example are often traceable back into the Pacific as rapid-moving "occlusions". After crossing the mountains they acquire new fronts and open wave characteristics, while the European disturbances start as open waves but gradually occlude. It is difficult to escape the conclusion that from the dynamic point of view the frontal patterns at sea level must be regarded as induced phenomena, consequences of the fact the that high level jet stream in which the upper waves move, normally, but not always, is associated with a well-developed front at the ground. This conclusion does not exclude the possibility that the upper waves may be intimately connected with the powerful, but somewhat more diffuse solenoidal field associated with the jet. These upper waves would then seem to take over rôle of the primary pressure waves first referred to by VON FICKER (1920). A somewhat similar point of view has been expressed recently also by H. RIEHL (1948).

II. Blocking and Breakdown of the Zonal Motion

By B. BOLIN

It was shown in the last section that the strong zonal motion which characterized the middle and upper parts of the troposphere

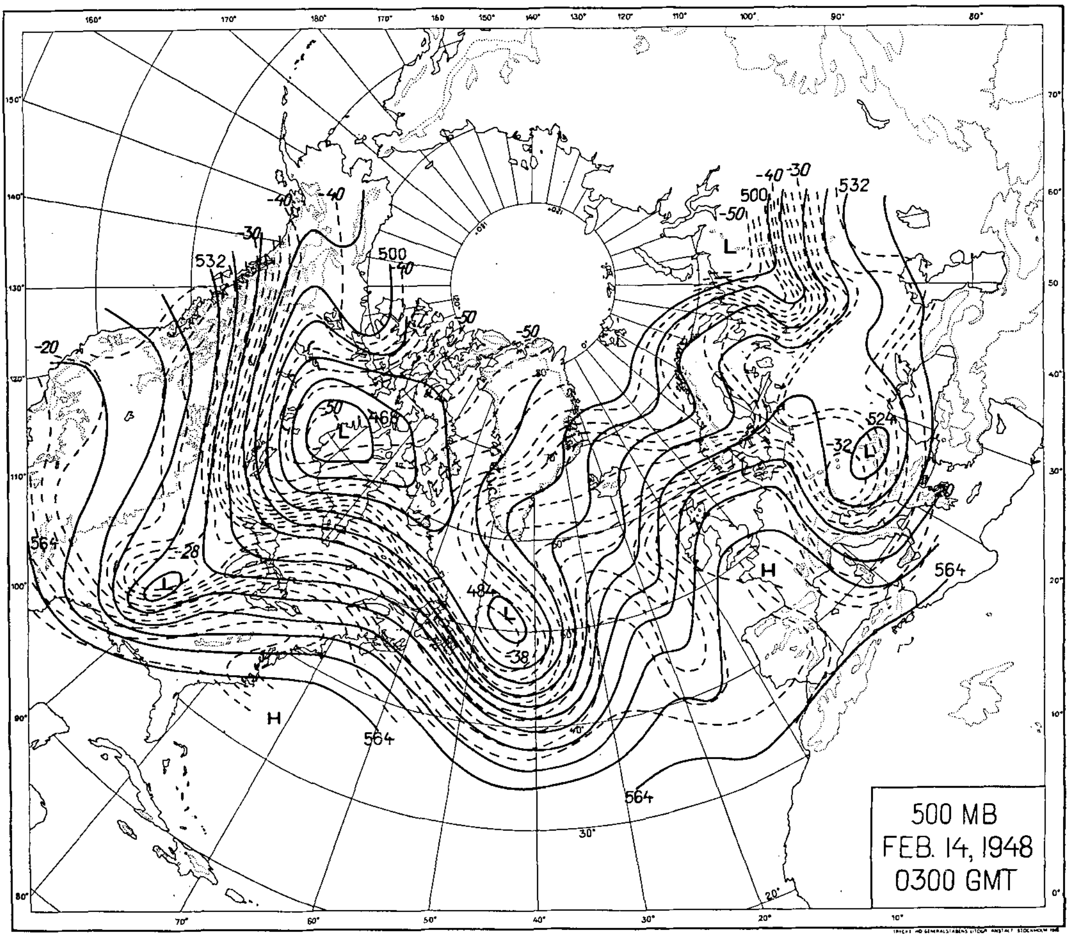

Fig. 15. Contours and isotherms on the 500-mb surface, 0300 GMT, February 14, 1948.

over Northwestern Europe and the Atlantic during the second week of February 1948 remained essentially unchanged in spite of the occlusion of a series of frontal waves. It is the purpose of the present section to describe how this apparently stable flow during the following week, between the twelfth and twentieth of February, broke down into quasi-stationary cyclonic and anticyclonic vortices. The broad features of this development are readily seen from a comparison between the synoptic 500 mb charts for February 9, 12, 14, 16, 18 and 20 (Fig. 4 a and 14—18 inclusive).

On February 9, prior to the breakdown (see Fig. 4 a) a well-defined belt of strong zonal motion was located in about 55° N and extended into Central Europe. Over the extreme northern part of Europe the westerly winds at 500 mb were weak. South of Europe, over the western part of the Mediterranean, there was no air motion from the West. Over the Baltic this central belt of westerlies split in two branches, one extending northward towards the White Sea while the other, stronger branch flowed southeastward towards the Balkans and into Asia Minor. The ultimate destruction of the zonal current appears to be associated with the gradual intensification and slow westward displacement of this split. On February 12 (Fig. 14) it occurred further to the West, in the region of the North Sea, and the process was by then far more striking and abrupt than three days earlier. During the following eight days the region of branching

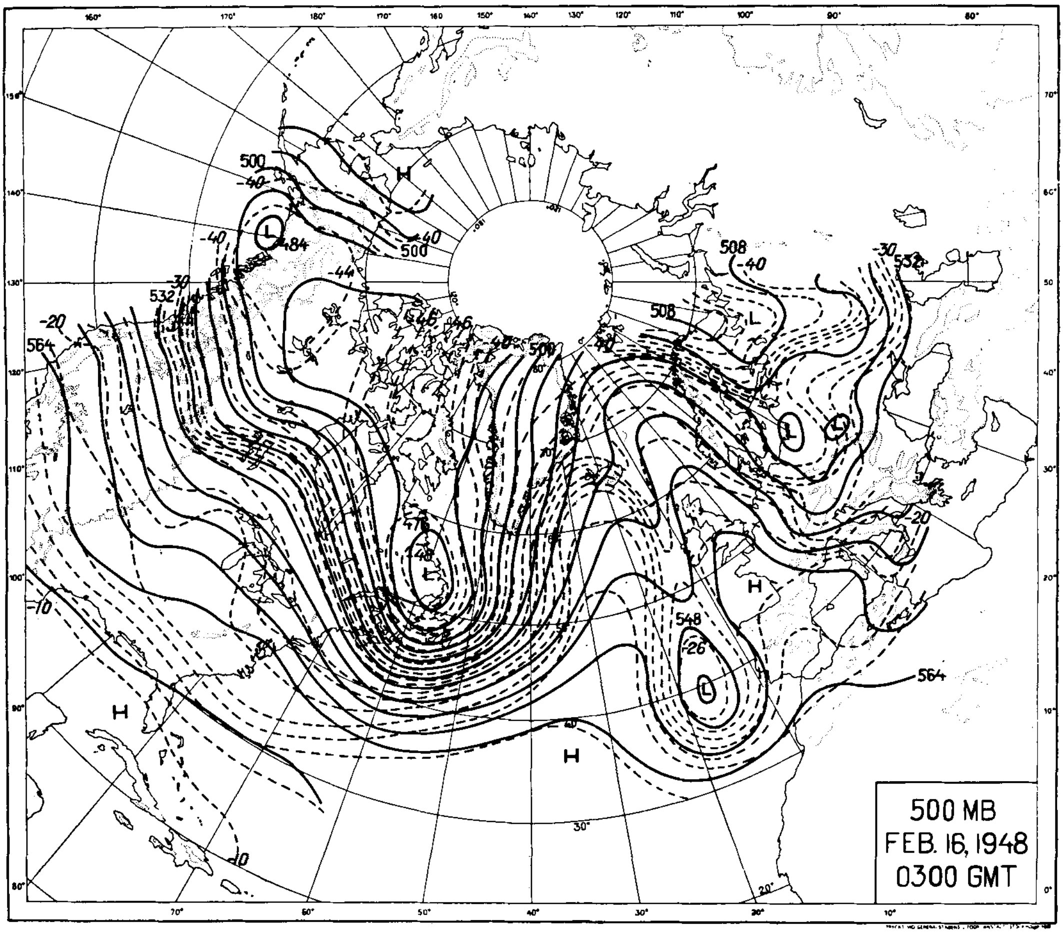

Fig. 16. Contours and isotherms on the 500-mb surface, 0300 GMT, February 16, 1948.

pushed steadily westward, dividing and destroying the zonal motion behind it until on February 20 (Fig. 18) the process had reached Labrador and Newfoundland. By that time the jet stream in lat. 55° N over the Atlantic had disappeared. In its place one observed a series of closed cyclonic and anticyclonic vortices, capped to the North by a belt of westerlies which extended eastward from Greenland at about 70° N, while another, apparently continuous zonal belt could be followed from Newfoundland across the Atlantic through the Mediterranean. It should be mentioned that such a zonal wind belt south of Europe is observed frequently. There are good reasons to believe that this southern jet may be identified with the well-defined zonal wind maximum at the tropopause level over northern India indicated in the normal meridional cross section for the winter season published in Physikalische Hydrodynamik (1933).

A comparison between the synoptic surface charts for February 9 (Fig. 4 b) and February 18 (Fig. 19) shows that this breakdown of the zonal motion aloft was accompanied by a corresponding breakdown of the zonal motion at sea level. On the latter day the prevailing zonal motion over the eastern part of the Atlantic and over Europe between about 35° N and 55° N was in fact from the East. This easterly current was surrounded by two belts of westerly winds, one to the North of Scandinavia and the other over North Africa.

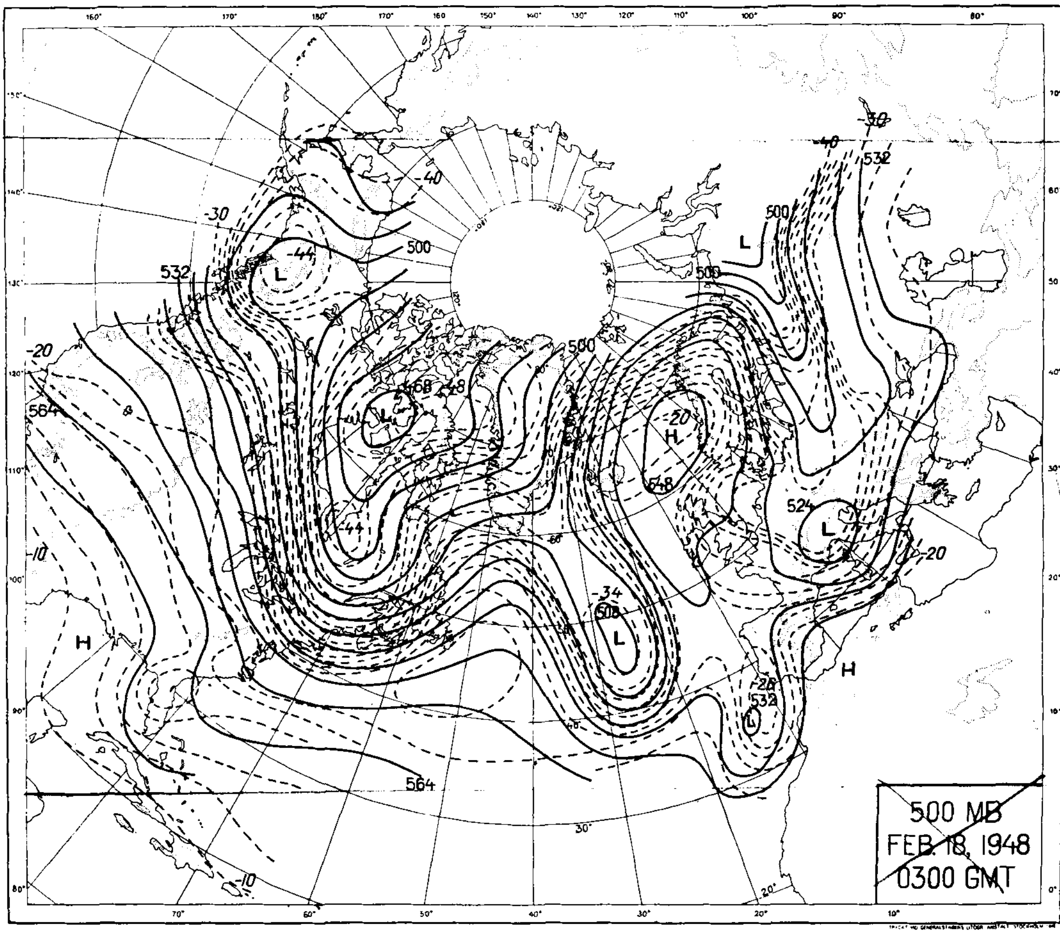

Fig. 17. Contours and isotherms on the 500-mb surface, 0300 GMT, February 18, 1948.

The development outlined above exerted great influence on weather conditions at the ground. The synoptic chart for February 18 (Fig. 19) provides an excellent illustration of the typical weather conditions over Europe and the eastern Atlantic following the breakdown. The most striking feature is the intense anticyclone built up over northern Europe. To the north of this anticyclone disturbances were travelling in the strong westerly current and caused rather mild and moist weather in Iceland and northern Scandinavia. The temperature was between 3 and 6 degrees above normal. Most of the precipitation in Scandinavia fell on the Western side of the mountain range. To the south of the anticyclone cold air was brought into Europe from the Arctic Sea and Russia. Thus the temperature decreased rapidly between February 14 and 18, and there was a negative temperature anomaly of 5 to 10 degrees during the rest of the month in southern Scandinavia, central Europe, France and southern England. Because of the very low temperature at the upper levels the air was fairly unstable and snow showers were frequently observed in this easterly current. However, in general the amounts of precipitation were not large since the moisture content of the continental air was low.

It is evident that the development occuring during this period provides a striking illustration of blocking as described in recent years by NAMIAS (1947 a) and by ELLIOT and SMITH (1948). NAMIAS has presented a particularly

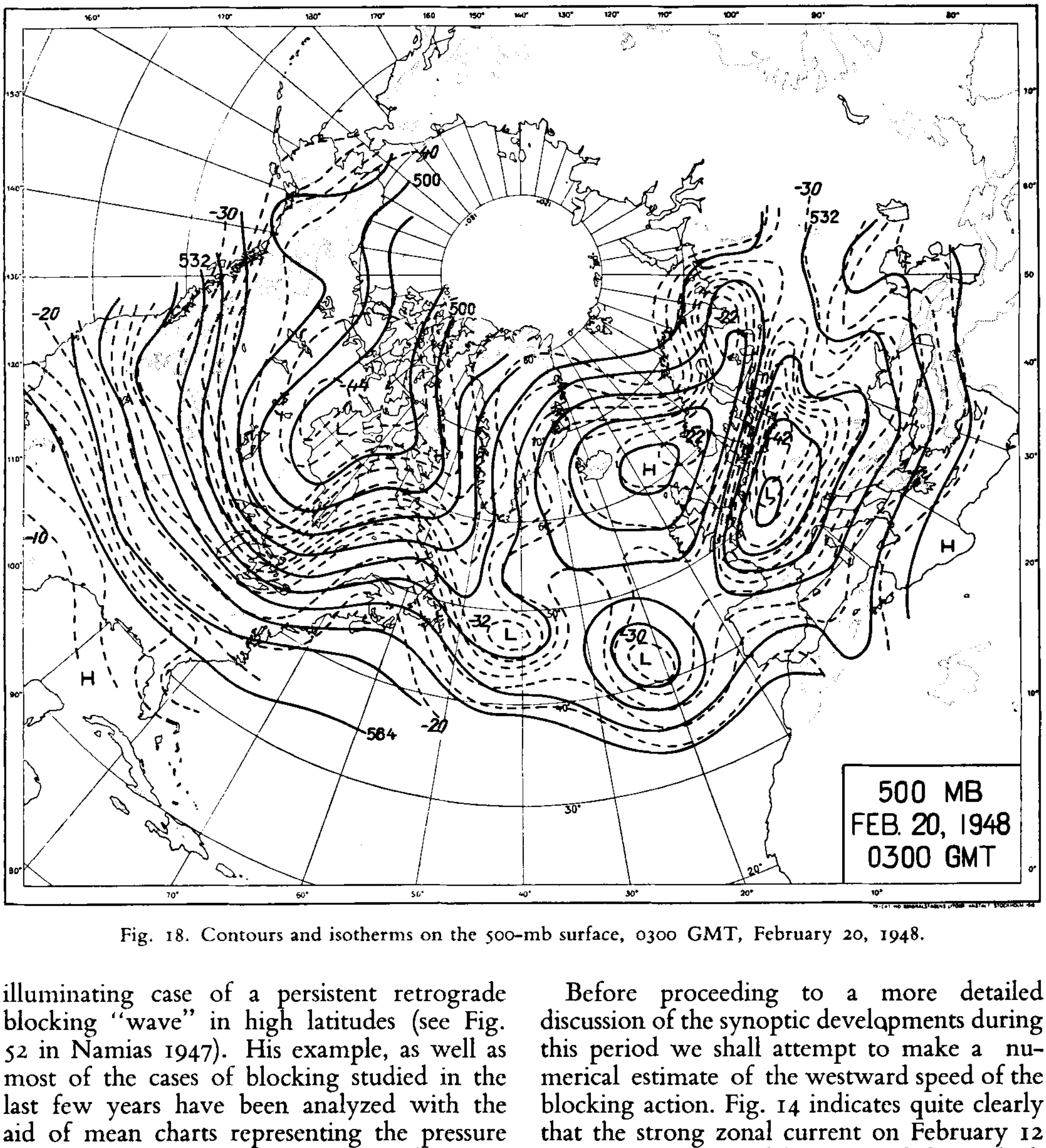

Fig. 18. Contours and isotherms on the 500-mb surface, 0300 GMT, February 20, 1948.

illuminating case of a persistent retrograde blocking "wave" in high latitudes (see Fig. 52 in Namias 1947). His example, as well as most of the cases of blocking studied in the last few years have been analyzed with the aid of mean charts representing the pressure distribution at sea level or at 10 000′ for overlapping or consecutive five or seven day periods. Even though this method is of obvious advantage in bringing out slow changes in the large-scale flow pattern it also tends to obscure details which may be of basic importance. For this reason we shall describe below the strongly developed blocking wave through its manifestations on the daily synoptic charts. It should be remembered, however, that individual blocking waves may vary considerably in their characteristics.

Before proceeding to a more detailed discussion of the synoptic developments during this period we shall attempt to make a numerical estimate of the westward speed of the blocking action. Fig. 14 indicates quite clearly that the strong zonal current on February 12 was limited eastward by a trough line which extended from Iceland towards Scotland and the North Sea. Similarly, on February 20 (Fig. 18) the zonal current was limited by another trough extending NW—SE and located just off the Labrador Coast. From the intersections between these two trough lines and the latitude circle for 60° N one finds a total displacement westward of about 55° of longitude in eight days, or an average speed of seven longitude degrees per day, in 60° N. This rate was not uniform. The maps for the

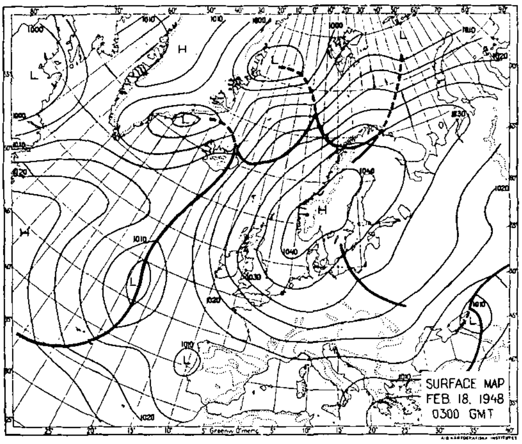

Fig. 19. Surface map for 0300 GMT, February 18, 1948.

intermediate days show that the blocking in its early stages developed quite rapidly westward but later on slowed down considerably.

An accurate estimate from the daily maps of the westward speed of the blocking process is rendered extremely difficult by the marked distortions in the basic 500 mb flow pattern brought about by a series of rapidly deepening waves which during this period moved out over the Atlantic from North America. The signifiance of these disturbances will be discussed later. To eliminate their effect and make possible a determination of the rate of retrogression of the blocking process a series of overlapping three-day mean contour charts for the 500 mb level were constructed. Fig. 20 gives the consecutive positions of a particular contour line taken from these charts. One finds from this diagram a westward speed of four longitude degrees per day in 60° N, or about 10 km. p.h. Since the anticyclone which developed in the rear of the blocking wave changed shape and steadily grew in extent, the speed obtained in this manner cannot be expected to agree with the first estimate more than to the order of magnitude. For comparison it should be mentioned that NAMIAS (1947) gives a displacement of about sixty longitude degrees per week as a characteristic value for high latitude blocking waves.

It was pointed out above that the advance westward of the blocking process did not take place at a steady rate but that the entire evolution was associated with the rapid deepening (and gradual stagnation) over the Atlantic of a series of waves which appear to have originated in the western part of North America. Figure 21 shows the vertical structure and approximate period of these waves. The data for this diagram were taken from the aerological soundings at Caribou, Me (47 N°, 68° W). The full lines represent, from top to bottom, the contour values at 500, 700 and 850 mb as well as the pressure at sea level, all as functions of time. The broken lines give the temperature data for the same levels. The figure proves that three well-defined waves passed the station between February 12 and February 20 with an average period of slightly less that three days. *These rapidly deepening waves were thus characterized by a period almost twice as long as that of the relatively stable waves described in section I of this paper.* Temperature and pressure variations were in phase at the upper levels (i.e. the troughs were cold, the ridges warm) but at the 850 mb level there was a phase lag between the two curves, particularly well-marked in the second and deepest wave. In this trough the axis of the temperature wave was very nearly vertical while the pressure wave at the ground was ahead of the upper pressure wave by about 24 hours.

The horizontal amplitude increase of these deepening waves is of particular interest in connection with the breakdown of the zonal circulation. An approximate idea of the movements of individual air particles in the

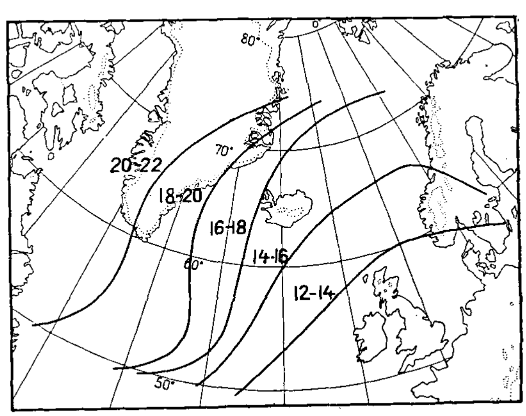

Fig. 20. Successive positions of the contour line for 5300 geodynamic meters as taken from overlapping three-day mean charts for the 500 mb level in February 1948. The numbers indicate the periods for which these mean charts were constructed.

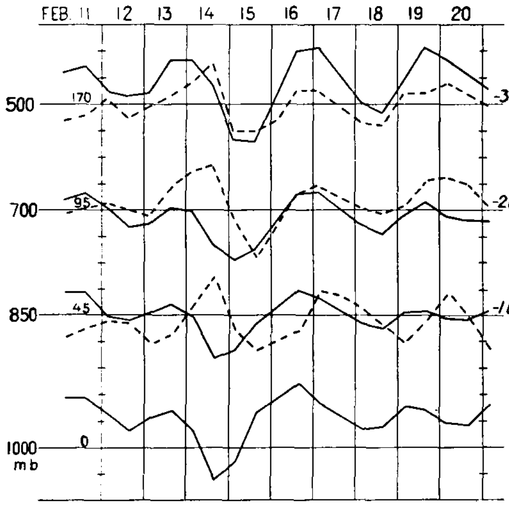

Fig. 21. Contour and temperature (broken lines) variations with time at different isobaric levels above Caribou, Me. The full lines represent, from top to bottom, the contour variations at 500 mb, 700 mb and 850 mb, and finally the pressure at sea level. The broken lines represent the corresponding temperature variations.

north—south direction may be obtained by studying the changes of the temperature field in the 500-mb level. In doing so we assume that the particles move along an isobaric surface and that the processes are adiabatic. In reality these conditions are not fulfilled. However, the non-adiabatic processes normally produce a cooling of the warm air masses as they are carried northward and a warming of cold air which is transported southward. Furthermore, by and large experience indicates that the cold air masses in the lower troposphere are subsiding while warm air is rising. Hence the effect of vertical movements on the temperature field at 500 mb is in the same direction as that of the non-adiabatic processes and both counteract the effect of horizontal (or isobaric) advection. It follows that the meridional displacements must be at least as large as those indicated by the isotherm patterns in the 500-mb level.

The 500-mb chart for February 12 (Fig. 14) shows a narrow, nearly zonal belt around 50° N, in which the isotherms are very closely packed. Superimposed on this pattern one observes a series of fairly evenly spaced isotherm troughs, located (at 50° N) in about 30° E, 5° W, 50° W and 85° W. The easternmost of these became important in the final stage of the blocking process. For the time being we shall consider the development of the two western systems. Between February 12 and February 14 (Fig. 15) the isotherm amplitudes in these troughs increased markedly, particularly in the trough located in longitude 85° W on February 12, which during these two days moved easward about forty degrees into the Atlantic just off the east coast of Newfoundland. As a result of differential motion in the northern and southern parts of the trough, the horizontal axis of the trough took on a marked slant NW—SE. During these two days a marked deepening (pressure fall) occurred in it; this deepening in turn gave rise to intense southerly winds and warm air advection on the eastern side of the trough. Through this warm current the stagnant southern portion of the next cold trough downstream was cut off, to form a cold air pool off the Bay of Biscay (Fig. 15). The same intense warm current produced pressure rises along its eastern boundary which resulted in a broadening towards the West of the relatively narrow ridge of high pressure which on February 12 had reached from Central Europe and Scandinavia towards the northeastern coast of Greenland. On February 14 another deep trough, with cold air advection, was located over the Mississippi Valley. Two days later (Fig. 16) it had reached Newfoundland and Labrador where it set off a new and intense air current northward. The warm air advection and pressure rises associated with this current led to a complete cutting off of the southern portion of the preceeding trough which on February 16 had been reduced to an isolated cold cyclonic vortex off Spain.

The processes described above were repeated a total of four times during the eight day period under investigation. Each trough, as it deepened and slowed down over the central part of the Atlantic, gave birth to a strong southerly current on its eastern side through which the preceeding trough was cut off and partially filled. These intermittent warm currents gradually broadened the initially quite narrow anticyclonic ridge until finally, on February 20, a large warm anticyclone extended along 60° N from Scandinavia to Greenland while the cut-off troughs remained as a series of cold cyclonic vortices on the south side of this anticyclone.

PALMÉN (1949) has stressed the fundamental importance of subsidence in the development of cut-off cyclonic vortices in the upper troposphere. According to him this subsidence occurs primarily in the northern part of the cold air tongues, though there is no particular reason why such a distribution of the sinking motion should be expected a priori. In the series of cut-offs described above the required subsidence would be a dynamic consequence of the strong, anticyclonically curved warm current arriving from the next trough upstream and would occur mainly along the right edge of this current. In a large measure, however, the initial cutting off seems to result simply from horizontal advection. [Compare Staff Members, University of Chicago (1947)].

The development discussed above is very similar to the seclusion process first described by BJERKNES and SOLBERG (1922), though the changes in the upper flow pattern studied by us are of greater dimensions than the seclusion pattern of ordinary short frontal waves.

During this period every disturbance travelling across the American continent developed in conjunction with a wave-like deformation of the frontal pattern at the ground. These waves were unstable and their amplitudes grew rapidly throughout the entire troposphere. Rapidly occluding frontal disturbances of short wavelength are often observed in which no such large-scale changes in the upper flow pattern or temperature field are found. Section I of this paper contains a description of a series of such short waves from the preceeding week, just before the blocking of the westerlies set in. These short waves generated closed cyclonic vortices in the lower troposphere below about 600 mb. Above this level, however, only minor waves were observed and the individual air particles performed an oscillating motion with an almost constant amplitude while travelling eastward. No closed vortices were formed by the cutting-off of cold air in the manner described above. *Consequently these shorter waves were not unstable in the sense of producing a complete breakdown of the wavelike flow pattern at the upper levels.* It is then necessary to observe the difference between these two kinds of waves in the atmosphere, both of which appear as occluding frontal waves at the ground but represent quite different processes aloft: 1) short frontal waves (wave length 2 500 km or less) which may occlude but nevertheless do not change the upper flow pattern and 2) somewhat longer, apparently unstable waves which develop in conjunction with a complete breakdown of the zonal flow.

The reservation implied in the term "apparently unstable" is made advisedly. A generally accepted theory for the instability of these somewhat longer waves has not been developed yet. Assuming however, that such a theory could be formulated it would presumably lead to the conclusion that a certain range of wavelengths is unstable; it might even be possible to show that there exists a particular wave length of maximum growth rate, as was done by EADY (1948) for the particular atmospheric model studied by him. By dropping the restriction to small amplitudes it might be possible to show that the speed of propagation of the individual troughs would slow down as the amplitudes grow, and that adjacent waves would influence each other. However, such a theory could not by itself account for the characteristically uneven spacing of the troughs that developed during the period studied by us and that expressed itself as a tendency of the troughs to pile up off the coasts of Europe. Any theoretical treatment of the growth and breakdown of a series of small, sinusoidal perturbations would inevitably lead to identical behaviour of the individual troughs and would not by itself be able to account for the marked shortening in wave length observed over the eastern part of the Atlantic. One is forced to conclude that this observed decrease in the spacing between consecutive troughs as they come in from the Atlantic, inexplicable on the basis of instability in the commonly accepted sense, must be looked upon as a manifestation of the blocking process, whatever its cause, and must be attributed to the successive (rather than simultaneous!) retardation of each consecutive trough in a given region. Under those circumstances it is impossible to exclude completely the possibility that the observed amplitude increase of the horizontal isotherm pattern in each trough (i.e. the instability) somehow may be the consequence of the retardation of the troughs, rather than the cause. One is once more led to the same conclusion as in the previous section, viz. that the theory for these different types of perturbations should take into consideration the variation with longitude of the underlying basic current pattern itself.

In spite of the complications pointed out above it may be of some interest to explore the applicability of existing theories for unstable waves to the deepening disturbances analyzed in this section. The first studies of the dynamics of long waves were carried through for an incompressible and barotropic atmosphere to clarify the rôle played by the variation of the Coriolis' parameter with latitude in determining the scale of atmospheric disturbances. In view of the restrictive assumptions on which these studies were based they could not provide any information concerning the stability of atmospheric waves. Subsequent investigators attempted to solve the stability problem by taking the baroclinc state of the atmosphere into account. JAW (1946) studied the problem in two dimensions by introducing the effect of the horizontal solenoidal field and found a critical wave length beyond which the waves become unstable. A more general attack was made by CHARNEY (1947) who treated the three-dimensional problem and took the vertical shear into consideration. He derived an instability criterion which indicated that the instability increases with vertical shear, vertical temperature lapse rate and latitude but decreases with wavelength.

When applying these theories to the upper air charts it is necessary to keep in mind that they were derived for a rather simplified atmosphere and by assuming small pertubations, in order to permit linearization of the hydrodynamic differential equations. Thus, while the theories may give the conditions necessary for the initiation of instability they cannot describe in what manner the subsequent development takes place. In spite of this, it is of interest to compare the initial conditions in February 1948 with those prescribed by the theory. The mean value of the vertical shear has been computed for each day in a 1 500 km broad strip along the jet between longitude 130° W and 60° W, the region where the waves grow most rapidly. The figures thus obtained (table II) show an increase of the shear as the development of the blocking starts, indicating a qualitative confirmation that the waves have become unstable.

Table II. Mean vertical shear in a 1 500 km broad band along the jet between longitude 130° W and 60° W, computed from the isotherms in the daily 500-mb charts for 0300 GMT

Date	Mean vertical shear m/sec/km
Febr. 11	1.7
12	1.9
13	2.2
14	2.3
15	2.4

The magnitude of the shear, however, is considerably greater than is necessary in order to get instability according to CHARNEY. The computations represent a mean value for a rather broad band. Nevertheless, the actual shear so determined is almost twice as great as the critical value demanded by Charney's theory: 1.0 m/sec/km. In the frontal zone, where the strongest concentration of isotherms exists, the shear is still greater: 4—6 m/sec/km. Charney himself reached the conclusion that the westerlies are the seat of a nearly permanent instability. Since a complete breakdown of the waves in the manner described by us occurs fairly infrequently it appears impossible to explain this development as the consequence of a nearly permanent instability of the zonal flow. Once more the impression is gained that the major change in the flow pattern associated with the blocking process may be the primary factor and that the amplitude of the individual waves grow as a result of the blocking and of the retardation in phase velocity produced by it. In this connection it is instructive to consider the changes in the longitudinal distribution in zonal motion at 500 mb which are observed during the period (Fig. 22). In this figure we have plotted the zonal motion at 500 mb between 45° N and 65° N as a function of longitude for three slightly overlapping three-day periods. The westward progress of the blocking is brought out by the diagram as is the fact that the zonal motion throughout the entire period decreases in strength towards the east.

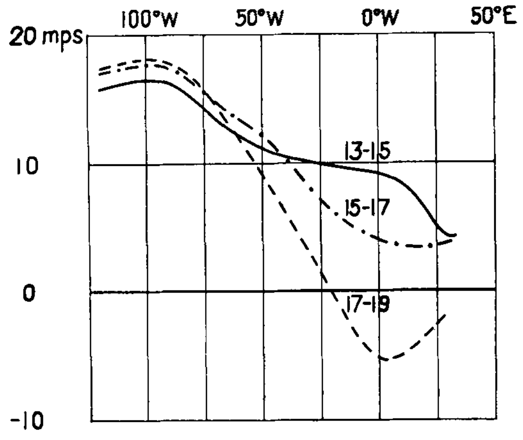

Fig. 22. Distribution at 500 mb of the zonal geostrophic motion between 45 and 65° N as a function of longitude, at different stages of the blocking process. The numbers indicate the partly overlapping three-day periods for which the computations were made.

At least superficially the sharp break in the zonal current at the western boundary of the particular blocking wave described in this paper resembles the standing wave or hydraulic jump which develops in open channel flow when the depth of the water current drops below a certain critical depth. The existence of a critical relationship between speed and depth may be derived from the laws of conservation of mass and momentum but can not be established by means of linearized equations. Fig. 23, reproduced from ADDISON (1944) gives an idealized representation of the flow through such a standing wave. One finds that the total energy is less at section 2 than at section 1; the surplus is transformed into violent turbulent motion in the standing wave. If this analogy is accepted the large-scale turbulence which appears in the present case as the waves break down would be the consequence of the formation of a blocking wave. The principal problem is then to determine the conditions under which such a change in the basic current takes place. In an investigation of this process it may well be necessary to take into consideration the three-dimensional structure of the jet stream even though it has been shown (YEH 1949) that a considerable amount of information concerning the blocking process may be obtained from the analysis of barotropic models.

We shall now consider the final stage in the development of the high latitude blocking anticyclone. It has been shown above that the gradual westward growth of this system was associated with a series of intermittent warm southerly currents emanating from successively deepening Atlantic troughs. Towards the end of the period (February 17—20) this anticyclone was cut off completely from the warmer air in southerly latitudes by a major change in the flow pattern over Finland and Russia. As a result of the strong pressure rises over the northernmost part of the Atlantic the contours to the east of the ridge took on an almost straight meridional orientation. The cold air over the Arctic Ocean then had the opportunity to move southward. In the eastern part of this current the cold air advection gave rise to rapid cyclogenesis, but a portion of the cold air was deflected westward and appeared over northern Germany and southern England as a very intense *easterly* jet. With this development the breakdown of the westerly flow had been completed. On February 20 (Fig. 18) the zonal index at the 500 mb level, computed for the region between 45° and 65° N and between 60° W and 40° E, had dropped to — 3.5 mps. To get an idea of the structure of the warm anticyclone and of the cold vortex on its south side a cross section has been constructed extending from Jan Mayen in 70° N to Maison Blanche (Algiers) in North Africa (Fig. 24). From this section, as well as from the individual 500-mb charts one gets a strong impression of the large-scale horizontal mixing associated with the breakdown process, warm tropical air having been brought far to the north while cold arctic air has been drawn into the temperate and subtropical zones.

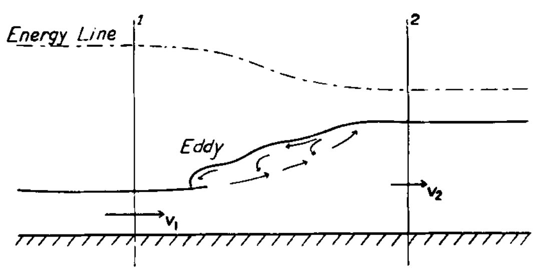

Fig. 23. Section through a standing wave (hydraulic jump) in a water current. (Compare e.g. ADDISON, 1944.)

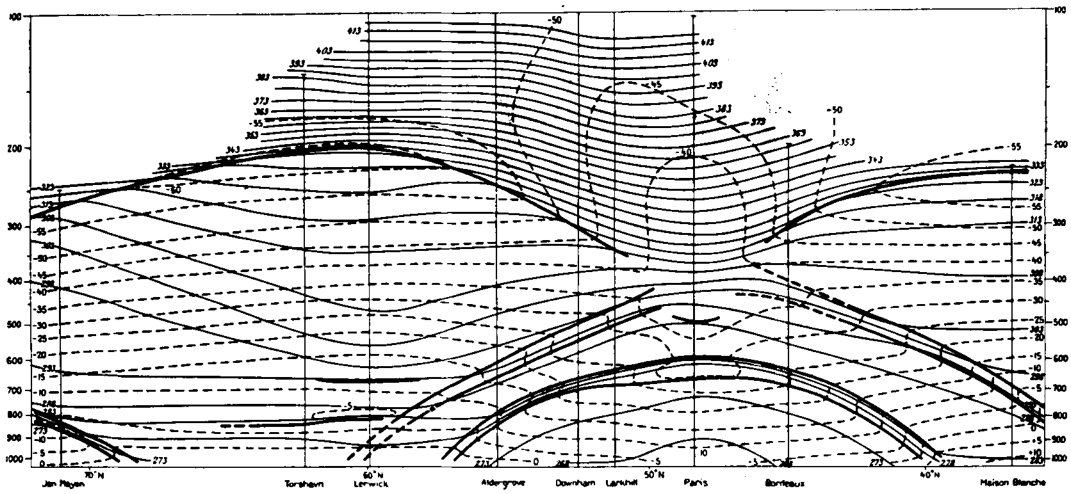

Fig. 24. Vertical cross section from Jan Mayen to Algiers (Maison Blanche) for February 20, 1948, 0300 GMT. Frontal boundaries, inversions and tropopause are indicated by thick solid lines when distinct, and by thick broken lines when indistinct. Thin solid lines represent isentropes and thin broken lines isotherms.

Some interesting features of the easterly jet may be pointed out. The relative vorticity of an individual air particle in the current (at 500 mb) *decreases* in spite of the southward movement. The complete vorticity equation

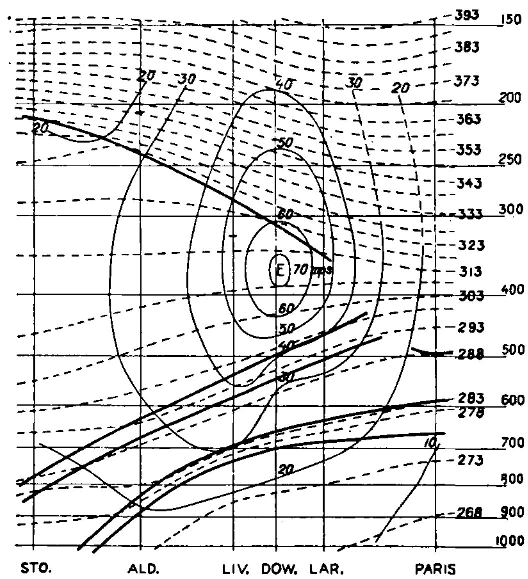

Fig. 25. Distribution of wind speed in the easterly jet over southern England on February 20, 1948, 0300 GMT. Thin solid lines indicate the wind speed normal to the section and thin broken lines the isentropes for dry air. Frontal boundaries, inversions and tropopauses are given by the thick lines.

shows that this decrease may be explained as an effect of horizontal divergence within the current, of a horizontal solenoidal field or of certain vertical movements. In the present case the solenoids act in the opposite direction to the one required, i.e. they tend to produce cyclonic vorticity, as may be seen from the indicated cold air advection. There are two observations which suggest that the decrease in vorticity was brought about by divergence. In the first place, the height of the tropopause over the cold current dropped to lower and lower levels, reaching a minimum of 4 500 m over Paris on February 20 (see Fig. 24). In the second place, from an approximate analysis of the vertical motions it can be shown that there was no general descending motion in the lower troposphere, below about 2 000 m. The last statement is borne out also by low-level convection in the cold air referred to earlier. One is thus led to the conclusion that the upper portions of this cold air current must have spread out laterally during the movement southward.

The wind distribution in the jet is shown in Fig. 25 which is based on observed winds extrapolated by means of the temperature field and the geostrophic assumption. On February 20 the jet was located over southern England with a pronounced maximum of about 70 mps in the 400 mb level. The maxi-

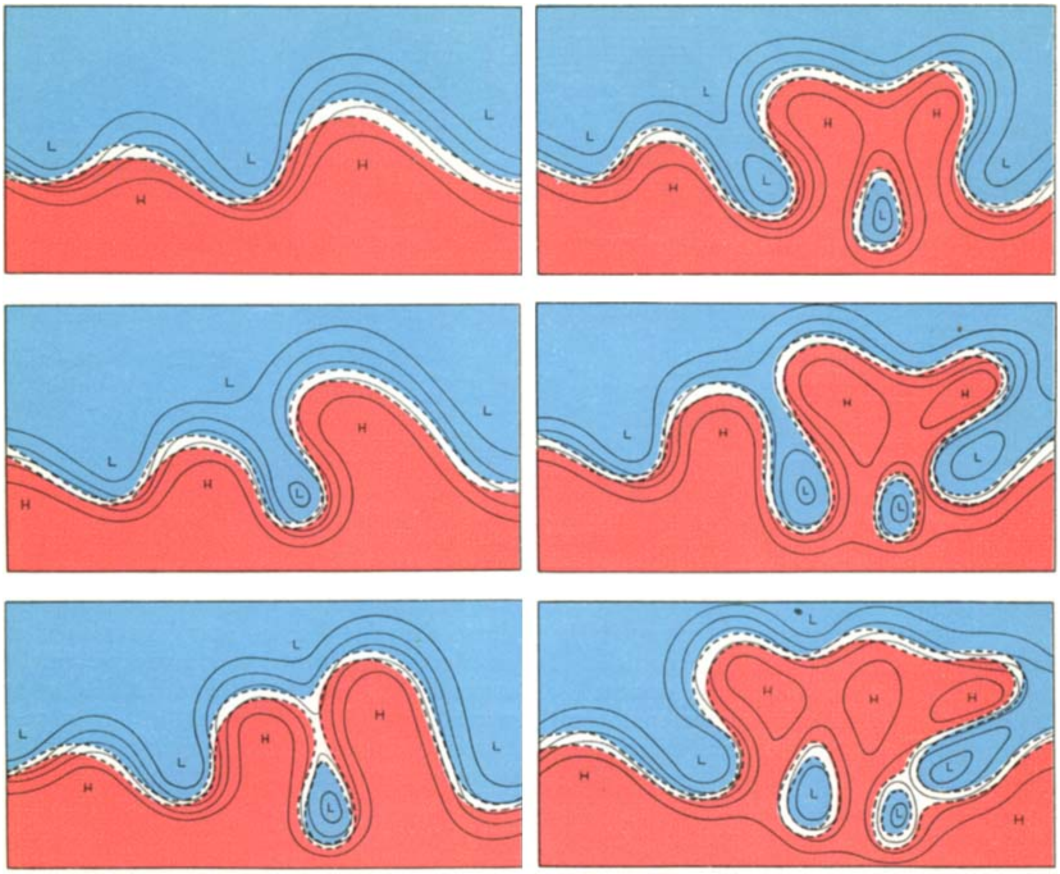

Fig. 26. Idealized sketches of the development of unstable waves at the 500 mb level, in association with the establishment of a blocking anticyclone in high latitudes. Cold air in blue colour, warm air in red. Solid lines are stream lines and broken lines the frontal boundaries.

mum anticyclonic shear which is likely to be observed in a zonal current, except as a transient phenomenon, should correspond to constant absolute angular momentum. In this case both observed and computed winds seem to indicate that this critical value was reached on the northern side of the easterly jet. In this connection it should be mentioned that professor J. Bjerknes in an as yet unpublished investigation has attributed the anticyclonic sweepback of deep upper troughs of the type here studied to the presence of inertia instability on the south side of a westerly jet as it crosses the upstream ridge.

There are two additional factors which appear to be of importance in connection with the establishment of the easterly jet. As a result of the blocking process one branch of the zonal westerlies had been shifted to high latitudes. The stationary wave length of planetary waves, measured in longitude degrees, is given by

$$\lambda = 180 \sqrt{\frac{2\,U}{a\,\Omega\,\cos^3\varphi}},$$

where U is the velocity of the zonal current, a the radius, Ω the angular velocity of the earth (in c.g.s. units) and φ the latitude. λ is the wave length measured in degrees longitude. In the derivation of this formula the curvature of the earth was not taken into account and the expression therefore can not give more than an approximate value for the wavelength. Nevertheless, it is evident that the stationary wavelength is greater in high than in low latitudes for a given value of the zonal current. This difference in wave length between high and low latitudes shows up rather clearly on the 500 mb charts for the latter part of the period. It appears safe to conclude that the ultimate establishment of a quasi-stationary cold trough over the White Sea region may be regarded as the consequence of the long quasi-stationary wave length in high latitudes and of the development of a strong southerly current in the region of Greenland. Through this quasi-stationary White Sea trough cold arctic air was brought far to the south. It was shown earlier that the westward deflection of this cold current was associated with marked horizontal divergence. However, it is not unlikely that the marked anticyclogenesis on the south side of the developing northern zonal current may have been a significant contributing factor, establishing air motion from the east over central Europe prior to the arrival of the cold arctic air. This appears particularly likely from the character of the flow pattern at the 500 mb level on February 18.

NAMIAS (1947 b) has described the establishment of a very intense blocking anticyclone in February 1947. This period seems also to have been characterized by a development similar to the one described in this paper.

In order to sum up the various essential features of the blocking process described above we have prepared the idealized sketch maps reproduced in Figure 26. In these sketches stream lines are solid and the dashed lines represent the frontal zone. Cold air is indicated by blue colour, warm air by red. The width of the frontal zone on the sketches is inversely proportional to the intensity of the front. In reality the front is not as sharp as shown in these figures but this overemphasis was made to give a clear picture of the principles involved. For the same reason the anticyclones to the north have been drawn as separate vortices though they actually appear to merge into a single system. The retrogression, the cutting off of cold cyclonic vortices and of warm anticyclones as well as the final injection of a cold current from the northeast have been incorporated in these figures.

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