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TEMPERATURE CHANGES ON A LARGE SCALE IN THE ARCTIC WINTER STRATOSPHERE AND THEIR PROBABLE EFFECTS ON THE TROPOSPHERIC CIRCULATION

BY E. FROGNER

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CONTENTS

69 figures in the text.

Summary. The "Climat-Temp" material has been used to draw monthly mean maps over the Northern Hemisphere. The purpose of this work is to investigate possible influences on the tropospheric circulation from the wintertime behaviour of the stratosphere.

The monthly anomalies from the 5-year mean 1949—53 of the 100/300 mb thickness pattern are used as the indicators of the changes in the stratosphere. The troposphere is represented by the monthly mean departures from the normal values of the 500 mb heights. In this paper the material from the winters 1957—58 and 1958—59 is presented. Both winters are characterized by great changes in the stratosphere. The material from the winters 1959—60 and 1960—61 will be treated in a later paper.

A. It is indicated by the map material in the sections 3b and 4b that changes on a large scale in the arctic winter stratosphere can have an effect on the troposphere in such a way that changes from month to month, and monthly anomalies of the 100/300 mb thickness pattern, are succeeded by similar anomalies of the 500 mb contour heights.

In the same sections the mutual wintertime behaviour of the lower stratosphere and the troposphere is also discussed in connection with (1) the mountain effects, (2) the index cycle of the westerlies, and (3) a yearly cycle of meridional flow in the stratosphere-mesosphere.

B. An attempt to explain the strong stratospheric winter warmings and the breakdown of the circumpolar vortex is made (section 3d), based on combined effects of the baroclinic stratospheric waves and the disturbances of a general annual cycle of vertical velocities in the stratosphere and the lower mesosphere.

C. The annual ozone variation is briefly discussed in section 3e in connection with the general subsidence cycle, and in section 3f the later climatic anomalies in the fall and winter in Norway are judged in relation to the stratospheric temperature changes.

Introduction. Our knowledge of the stratosphere and its behaviour has increased very rapidly during the fifties, thanks to the great expansion of the rawinsonde network and the improved sonde technics.

Among other things, it has been discovered that above 15—20 km height in the Northern Hemisphere in winter, temperature changes may occur which in violence far exceed what can happen in the troposphere. Especially, the warming epochs are very pronounced.

Scherhag (1952) has given the first information about such sudden stratospheric warmings in winter, but the first one recorded is over England and Shetland in February 1951 (Scrase 1953). Later, several investigators have given excellent analyses of stratospheric warmings, Warnecke (1956), Godson and Lee (1958), Teweles (1958), Craig and Hering (1959), and Hare (1960). All these case studies have the objective to gain the best understanding of the physical and hydrodynamical processes involved in the stratospheric warmings and their origin, and the investigations are chiefly directed towards the highest available observations from the stratosphere.

This paper has not directly such an aim, but a more practical one; namely, to investigate whether the great changes in the winter stratosphere in high northern latitudes can have prognostic value for the troposphere.

In this connection, the strong but short-lived fluctuations at 50 mb and higher, only extending weakly downward, usually have little interest. The objective is the great changes covering large areas and dominating the lower stratospheric layers for a long time.

1. The material. The international distribution of "Climat" and "Climat-Temp" started in January 1949. During the first years the "Temp"-material was rather incomplete for drawing hemispherical maps, but has gradually improved. Two steps in this improvement may be mentioned here. Since February 1956 the 100 mb data have been included, and since June 1957 the U.S.S.R. has joined with "Temp" from Siberia. It was hoped that immediately after the I.G.Y., the 50mb-data should have entered in the monthly "Climat-Temp", but it looks as if we have to wait some time yet.

"Climat-Temp" is not quite a reliable material. Some errors are included, originating from different sources. Single computation- or teletype-errors have been corrected by horizontal and vertical checking of the sondes. Possible radiation-errors are only occasionally corrected. The maps drawn are also only occasionally completed by the U.S. Weather Bureau — W.M.O. publication. Great attention has been paid to the 100 mb surface, to make it as complete as possible, by checking and extrapolation of the 100/200 mb thickness. Even so, the 100 mb surface may occcasionally be a little doubtful in some areas, the Pacific, Alaska, the Atlantic and the Polar-Basin, but these possible small errors will not alter the pattern of the great changes.

The 100/300 mb thickness layer is selected as the indicator of the changes in the stratosphere. Primarily, the changes of the thickness pattern from month to month were used, and this is significant in mid-winter. During October—November and February—March, on the other hand, the mean annual trend will be included, and normal values are needed. As a prognostic tool the anomalies are better than the changes of the total heights. Monthly mean maps (1949—53) for the Northern Hemisphere up to 50 mb have been constructed by "Institut für Meteorologie und Geophysik der Freien Universität", Berlin (K. Wege 1957 and 1958). Can 5-year means for winter-time, however, be used as normal values when these levels show such great disturbances? Judging the period (1949—53) in different ways it seems reasonable that the years 1949—53 have been more "normal" in the winter stratosphere than the later years in the fifties, and the (1949—53) mean maps of Wege have been chosen as reference values for the stratosphere. U.S. Weather Bureau normal charts (1952) have been used for the 500 mb surface.

2. Coupling between the stratospheric and the tropospheric circulations.

a. Review of previous papers. In the summer season, the stratosphere in high latitudes is uniformly warm with temperatures near — 40°C. Above a level between 16 and 20 km height (varying with the latitude) anticyclonic easterly circulation is predominant. It is probable that no disturbances of any strength worth mentioning are originating in the summer stratosphere. "The temperature fluctuations in the lower layers of the stratosphere are most likely for the most part induced by the wave systems in the upper troposphere. Above 100 mb the flow is nearly barotropic, and the temperature regime seems almost free of short-period oscillations" (Hare, 1960).

In this manner there should, in the summer season, be only little reason for seeking impulses originating in the stratosphere and influencing the tropospheric circulation.

This has been confirmed by FLOHN, HOLZAPFEL and OECKEL (1959). The authors have made a thorough analysis of the easterly stratospheric flow in summer (May—August), and the results show a very great steadiness of the east wind, up to 95—100% south of the large continents of the Northern Hemisphere. "Above Europe and Eastern Asia practically no relation exists between the easterlies above 20—22 km and the varying winds near the tropopause, and the constancy of the easterlies reaches 80—90%. The stratospheric easterlies in summer seem to be a stationary "climatic" phenomenon, independent of the variable tropospheric systems." (FLOHN and coll. 1959).

In recent years, however, temperature fluctuations of moderate amplitude have been observed at high levels in the summer stratosphere (SCHERHAG 1958, TEWELES 1960).

Looking at the winter season, the facts are quite different, with strong stratospheric fluctuations in flow, temperature and height of the isobaric surfaces in middle and high latitudes. Several papers since 1952 have treated this interesting feature of the winter circulation. HARE (1960, fig. 2) has presented an excellent example on the great difference between the summer and winter behaviour of the stratospheric temperature field.

Only few authors have hitherto dealt with the relations between the stratospheric and tropospheric circulation. AUSTIN and KRAWITZ (1956) have investigated selected periods with great changes in temperature and flow at the 50 mb level over North America during the three winters 1951—54. These authors have made statistics of a great number of cases, including periods of both fall and rise of the 50 mb surface. Among the results of the statistical study, it may be mentioned that, "in low to middle latitudes the 50 mb changes appear merely to be damped tropospheric changes. On the other hand, at high latitudes the 50 mb changes are in general damped downward. Inspection of the data indicated that there are four classes of changes in the northern latitudes, underlying two principal processes; namely, 1) tropospheric developments accompanied by changes at 50 mb which are substantially less intense than those in the upper troposphere (warm stratospheric troughs and cold ridges) corresponding to the mid-latitude cases, and 2) major developments at 50 mb which are accompanied by much smaller changes of like sign in the troposphere (cold troughs and warm ridges)". Supported by case studies from maps and ascents, the authors conclude further that "major activity at 50 mb can have a significant effect upon the pressure distribution in the troposphere, and for this reason should influence the subsequent behaviour of tropospheric systems. The equatorward extent of the 50 mb effect varies with time."

An extensive investigation by PANOFSKY, KRAWITZ and JULIAN (1958) was aiming at (1) a method to specify 100 mb flow, given 500 mb charts, and (2) some understanding of relationships between tropospheric and stratospheric flow. The authors have introduced two definitions; namely, 1) the "polar regime boundary" (PRB), which represents the boundary in the winter stratosphere between the region of cold lows and warm highs to the north and warm lows and cold highs to the south, and 2) the "matching level" at which the wind speed equals the wind speed at 500 mb. Inside the PRB, no matching level exists.

The results of the investigated material (mostly from North America) show that "space-mean 500 mb charts are useful for the estimation of 100 mb flow in all but the summer season. In general, the troughs and ridges on 100 mb and 500 mb space-smoothed charts coincide in geographical position, even inside the PRB.

The wind speed in middle latitudes on 100 mb charts approximates the wind speed on space-smoothed 500 mb charts. However, in regions influenced by the winter-time polar vortex, the 100 mb flow is much faster than space-smoothed 500 mb flow, but well related concerning short-period fluctuations.

In all cases, the 4-day average heights at 500 mb and 100 mb are well correlated. However, the relation is a function of the scale of the oscillation concerned. Whereas the magnitude of long period height changes are similar in middle latitudes, inside the PRB the changes at 100 mb are about twice the changes at 500 mb. The shorter period changes at 500 and 100 mb are about the same at all latitudes."

The authors have also studied the relationships between stratospheric and tropospheric flow inside the PRB in the four winters 1953—57. The stratospheric and tropospheric flow is represented by 8-day running (4-day overlap) averages of the 100 mb and 500 mb height differences, respectively, between Columbia and Resolute. "The shorter-period oscillations at both levels are well correlated, whereas the long period fluctuations vary much, both inter-seasonally and from winter to winter." These variations have been viewed in connection with the warming epochs and the moving of the polar vortex.

WEXLER (1950) had the idea to compute the possible effect on pressure changes at the base of an air-column by warming at the ozone-level. The computations indicate "substantial changes just below the heated level, but using two-layer model, the changes are insignificant near the sea level." Perhaps this result may be interpreted in such a way that isolated warmings in the ozone-layers have no significant influence on the troposphere near the sea level, whereas warmings covering great regions in the stratosphere, and especially the simultaneous subsidence may have real effect.

b. Suggestions of how the stratosphere may influence the troposphere. Theoretical studies of the dynamic coupling between the stratosphere and the troposphere are scanty. In discussing the upper atmospheric circulation, EADY (1950) expressed himself as follows: "At still higher levels (mid- and upper stratosphere) we should expect developments similar to, but not directly related to, those in the troposphere. An account of behaviour in these regions must await further empirical evidence. It seems unlikely that there is sufficient energy, either potential or kinetic, at very high levels for any *direct* major influence on tropospheric behaviour to be possible. It would, however, be erronous to suppose that variations at very high levels (i.e., well above the tropopause) can have no influence on weather. For example, an increase in stratospheric temperature, however slowly developed, would eventually lead to a lower tropopause: The boundary conditions applying to tropospheric developments would be altered, leading to disturbances of smaller size, a reduced rate of heat transfer and consequent modification in zonal characteristics. Most important of all, there would result a change in the patterns associated with topography and the distribution of land and sea."

With our increased empirical information of the arctic stratosphere a little comment may be appropriate. The energy involved in the polar-night jet may have a major effect on the tropospheric circulation. The fluctuations in the stratospheric jet may affect the behaviour of the tropospheric polar front, and especially in periods when the stratospheric jet is coinciding twih the tropopause jet, the influence of the stratospheric flow on the tropospheric developments may be decisive (cf. the changes in January 1959, p. 64).

Otherwise, the writer of this paper is of the opinion that stratospheric influence on tropospheric circulation is mainly to seek in the boundary conditions applying to tropospheric developments. If the convergence and divergence, included in the cyclonic developments, had to be restricted to the troposphere only, not affecting the tropopause layer, a lower tropopause below a critical state, would lead to disturbances of smaller size. Then it is most likely that variations in the tropopause height and the low-stratospheric boundary conditions have the greatest effect on tropospheric developments in areas which usually have a low tropopause. The high latitudes thus are pointed out as the regions where the changes in the stratosphere have the greatest possibilities of influencing the tropospheric circulation, and the great stratospheric changes unveiled in recent years in the Northern Hemisphere draw attention to the winter season.

Limiting the discussion of the stratospheric influence on tropospheric circulation to the mean winter circulation through selected periods at high latitudes, it should be possible to draw some qualitative conclusions.

The map material, presented later in this paper, shows that in most cases the 100/300 mb thickness anomaly pattern is succeeded by a similar 500 mb anomaly in the *next* month. It is probable that the vertical components of the stratospheric flow near the tropopause in this connection are most significant of the low-stratospheric boundary conditions. When, during a period, stratospheric subsidence is uniformly prevailing in an area, the stability in the upper troposphere-lower stratosphere layer will be furthered, that is, the cyclonic developments are damped in that area. In the Arctic and over sub-arctic continents damped cyclonic activity in the winter season will favour anticyclonic departure from the normal tropospheric circulation due to outgoing radiation from the troposphere. In a neighbour area with prevailing rising stratospheric motions, the instability will be higher, cyclonic activity is not damped, but may rather be encouraged. This last effect should also comprise mid-latitudes and the summer season. The great time lag between the low-stratospheric and tropospheric anomaly patterns is mostly due to the persistence of the tropospheric circulation type.

When the stratospheric conditions are uniformly persistent, the writer is considering the stratospheric influences as differentiating boundary conditions in neighbour areas as very important with respect to long-lasting effects on the troposphere, evinced by the succeeding regional distribution of the tropospheric activity.

Craig and Hering (1959) have shown the relationship between simultaneous stratospheric and tropospheric changes in North America and adjacent waters in connection with the great stratospheric warming and circulation changes during January—February 1957.

The reciprocal behaviour of the lower stratosphere and the troposphere in connection with the retarding jet stream due to the mountain effects is disregarded above.

The effect on the circulation at middle to low latitudes of the great stratospheric changes in higher latitudes may be an indirect or secondary one, or it may, according to CRAIG and HERING (1959), be "that the radical stratospheric variation is only one manifestation of a sizable variation in the entire general circulation".

c. *Discussion of the method used.* The possible dynamic coupling between the stratospheric and tropospheric circulation is a very complex phenomenon and can be treated generally only in connection with a complete understanding of the cause of the general circulation and its changes up to great heights in the atmosphere. The empirical investigations and their results must be kept limited to fixed areas and specified concerning the coupling-relationships. This applies to the season and to the latitudes and longitudes on the earth. It refers to the used scale of space and time, that is, the degree of smoothing, and to the time lag between the primary and secondary events. To interpret the results in a right way, importance must also be given to the aggregates used to represent the stratosphere and the troposphere, respectively, and their changes.

In this paper monthly mean charts have been used. It is an advantage to employ mean maps in these investigations, because only stratospheric developments of some duration and continuity can have a significant influence on the troposphere. It is, however, a drawback that these means are restricted to the calendar-month, the possibilities of varying the mean periods consistent with the great revolutions in the stratosphere then being limited. In the two winters 1957—58 and 1958—59 investigated here, the major changes have occurred near the end of the month, and formed a transition to new long-lasting conditions in the stratosphere.

When this work was started, attention was directed only towards the great changes in the winter stratosphere, and their effects on the tropospheric circulation. Later, the investigations developed into inspecting the anomalies and their changes from month to month throughout the winter for the purpose of finding possible influences on the tropospheric circulation from all changes worth mentioning in the stratospheric temperature distribution.

GODSON and LEE (1958) considered the 100 mb level to be representative of the layer above. The writer regards a thickness layer and its changes more appropriate and reliable for this work than the height and temperature of a single level.

It is well known, and discussed by several authors, SAWYER (1951), KOCHANSKI (1954), PANOFSKY, KRAWITZ and JULIAN (1958), that oscillations in the lower stratosphere usually reflect the tropospheric circulation systems and their disturbances. Even so, the 100/300 mb thickness pattern has been chosen in this paper as the indicator of stratospheric changes influencing the tropospheric circulation, and the clue to this choice is the following reasoning.

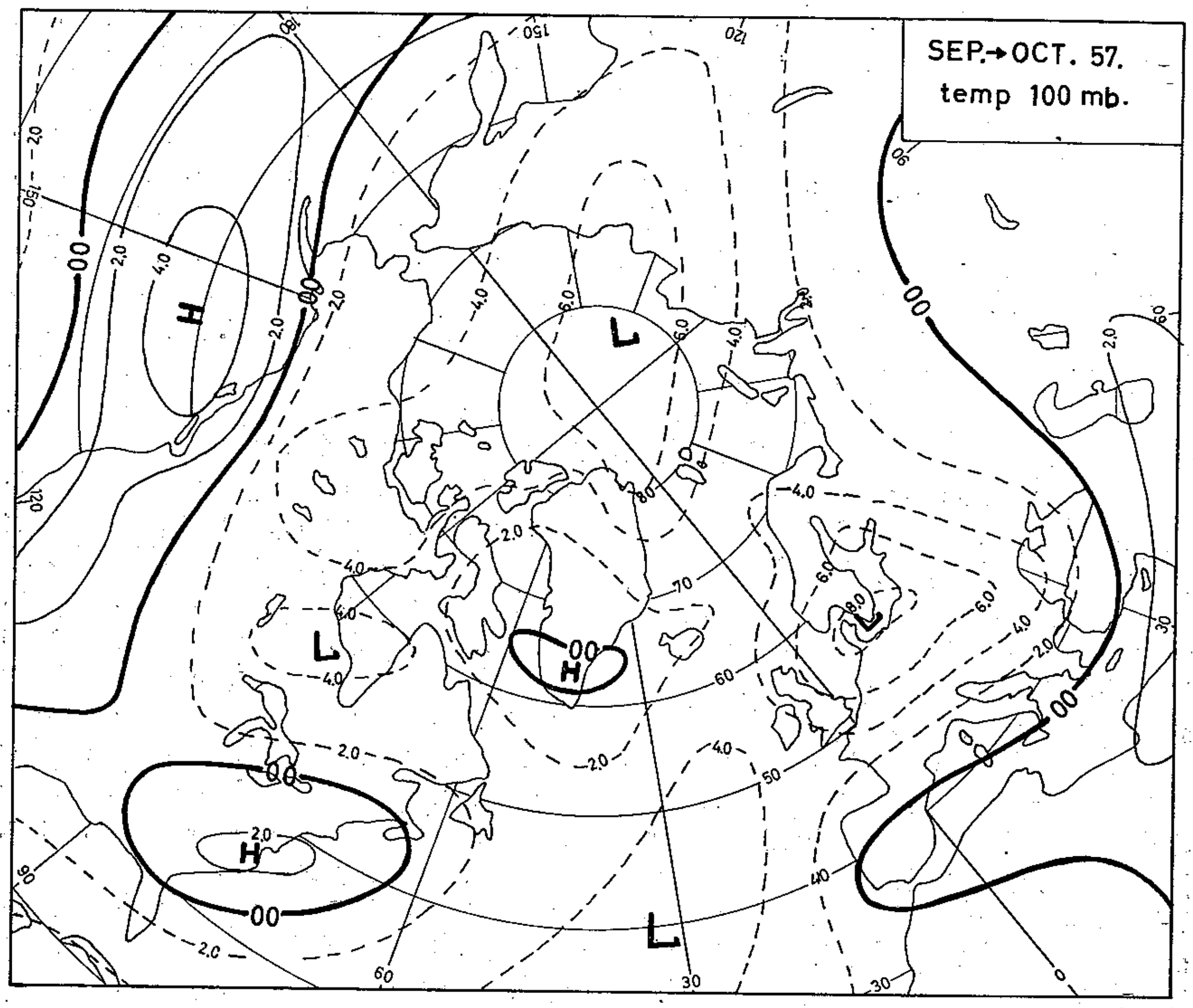


Fig. 1. Change from September to October 1957 of the monthly mean temperatures at 100 mb. Dashed lines negative values. C°.

First, according to Eady, the main stratospheric effect on the troposphere by the dynamic coupling is mostly indirect and founded on the upper boundary conditions applied to the tropospheric developments.

Second, the normal distribution of the 100/300 mb thickness pattern are an interpretation of the normal "adjustments" of the lower stratospheric layer to the normal tropospheric circulation. The "adjustments" may be repressions or favours of the tropospheric disturbances, and may through a longer time be stronger or weaker than normal. These facts will be interpreted in the monthly mean *anomalies* of the 100/300 mb thickness pattern.

Thus, the prognostic use of the monthly mean anomalies of the 100/300 mg thickness pattern as the indicator of the stratospheric effects on the tropospheric circulation may partly be based on the assumption that a weak but long-lasting, in some areas resistance to, in other areas a favouring of the tropospheric cyclonic developments can have a

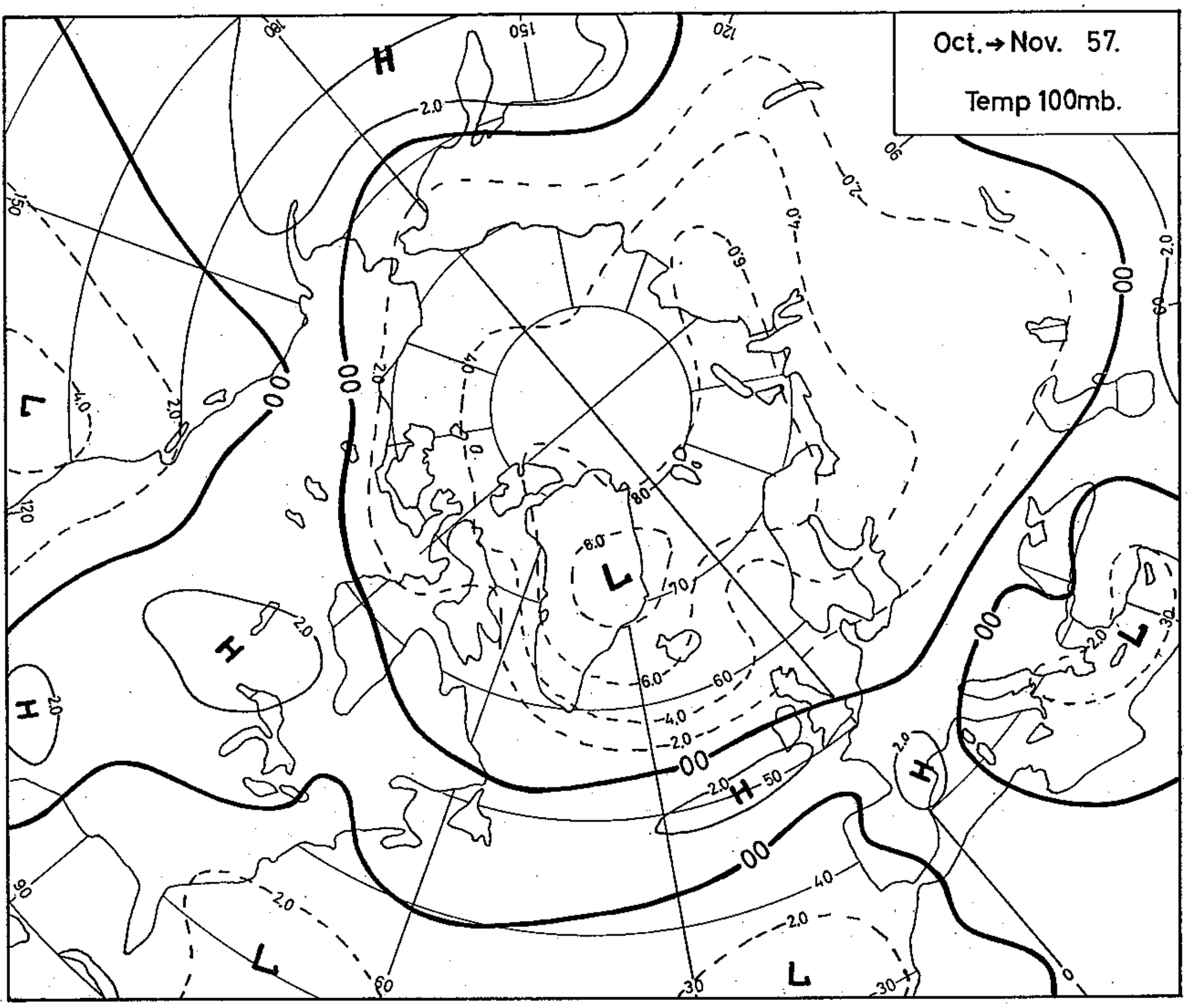


Fig. 2. Change from October to November 1957 of the monthly mean temperatures at 100 mb.

deciding influence on the further regional distribution of the tropospheric circulation pattern.

However, the stratospheric effects can be more active, either by stronger vertical motions or by coincidence of the stratospheric and tropospheric jet. In these cases, the stratospheric influence on the troposphere must act through the lower stratospheric layer, and are, accordingly, also reflected in the 100/300 mb thickness pattern.

Strong but transitory fluctuations in the mid-stratosphere not penetrating the lower stratospheric layer kinematically, are considered only to have disturbing pressure effects and not systematical effects on the tropospheric circulation.

It is the experience of the writer that the changes of the mean monthly 100/300 mb thickness pattern in wintertime are representative of the mean changes in the mid-stratosphere north of, say, 50—60 degrees latitude.

Even so, it is not easy to decide the character of the long stratospheric fluctuations. Extensive investigations are demanded, which the writer has had no resources to make.

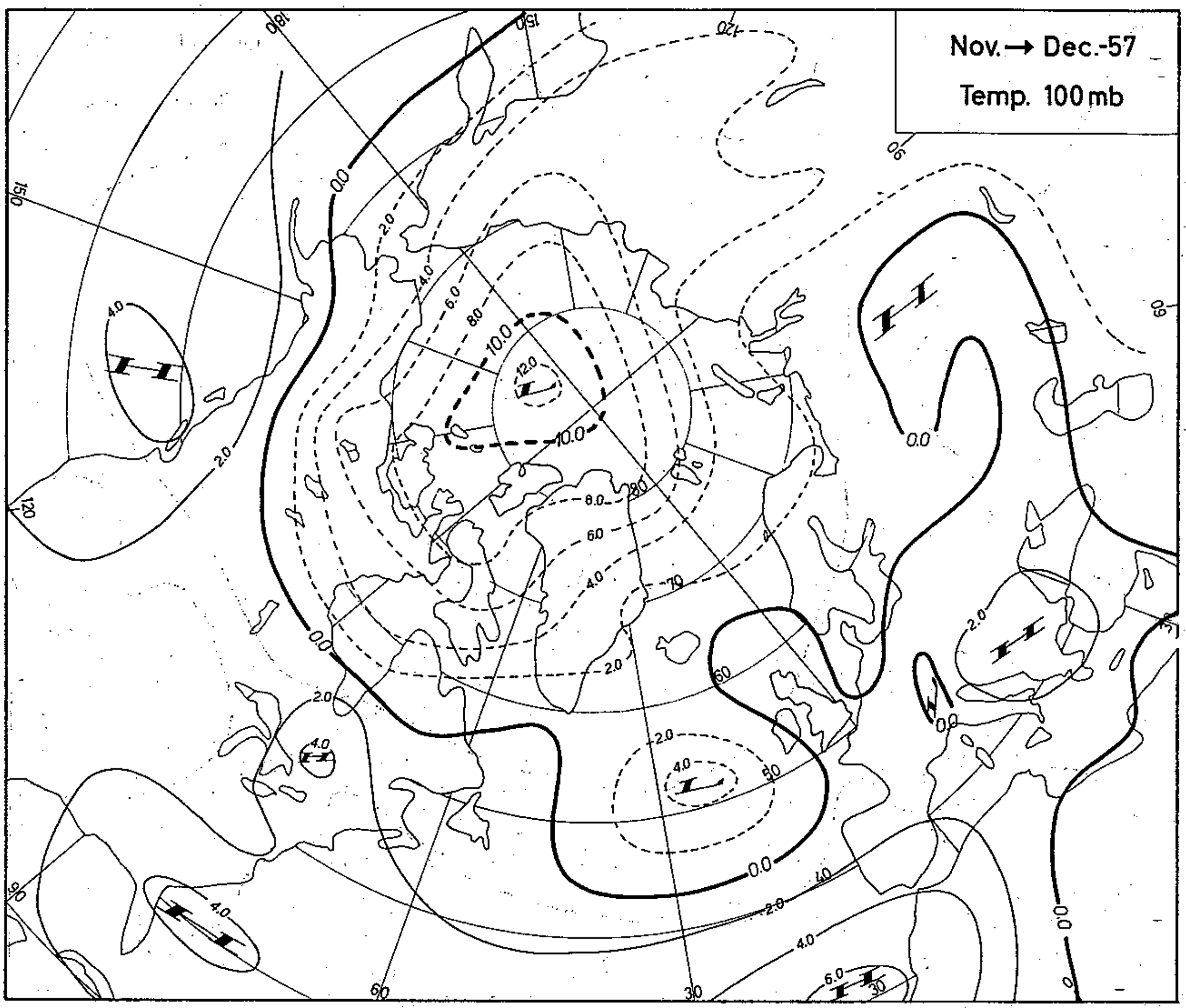


Fig. 3. Change from November to December 1957 of the monthly mean temperatures at 100 mb.

He is relying either on other papers, or he is trying by the mean monthly maps to estimate whether the anomalies of the 100/300 mb thickness pattern are representing tropospheric disturbances, mountain effects included, or stratospheric influences directed downward.

3. The winter 1957—58.

a. Mean temperature changes at 100 mb. Figs. 1—7 give a survey of the changes from month to month of the mean temperature in the 100 mb surface through the period September 1957—March 1958. The change from September to October (Fig. 1) shows an irregular pattern, probably and effect of the tropospheric circulation. Otherwise, the cooling was stronger in the Eurasian than in the American Arctic. This is also the case in Fig. 2, which presents the mean temperature change from October to November. The warming belt south of the cooling area is narrow and weak, but nearly continuous.

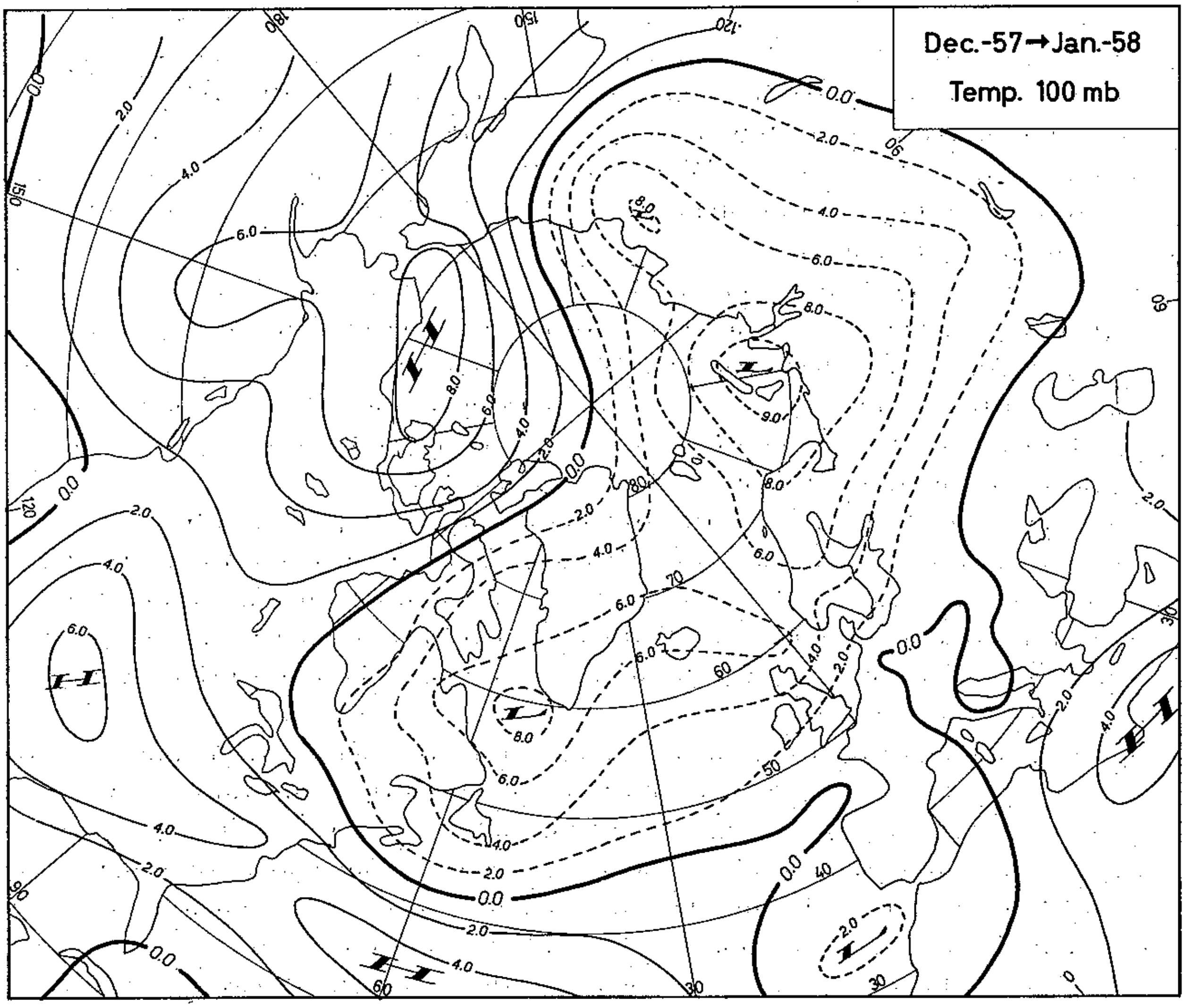


Fig. 4. Change from December 1957 to January 1958 of the monthly mean temperatures at 100 mb.

From November to December 1957 (Fig. 3), the cooling region had the centre over the Arctic Basin, between Alaska and the North Pole, with two branches, one towards the Northeastern Atlantic and the other extending far south in Asia. However, the 100 mb surface showed warming in a belt from the Pacific across the Southern U.S., the Atlantic, the Mediterranean and Southern Europe to Southwestern Siberia.

The mean temperature changes at 100 mb from December 1957 to January 1958 (Fig. 4) show a similar pattern of cooling and warming as in Fig. 3, but the cooling region was moving towards South Greenland, Novaja Zemlja and Siberia. South of the cooling area the warming belt was extending round the whole Hemisphere. In Asia and the Western U.S. the warming belt was moving northwards, with the maximum zone north of Alaska. As Fig. 3 and 4 have several similarities, they are combined in Fig. 5.

Large changes and warmings occurred in the stratosphere during the last week of January 1958. The strong revolutions were carried through in the first part of February. The mean result of the great temperature changes in the 100 mb surface is presented in Fig. 6, showing the mean temperature changes from January to February.

During the last part of February and the beginning of March 1958 the stratospheric low at 10 mb moved from the Siberian Arctic across the European Arctic to the American Arctic (BEHR — SCHERHAG — WARNECKE, 1960 c). The mean temperature changes from February to March at 100 mb are presented in Fig. 7, showing warming in Eurasia and cooling in North America.

b. Anomaly maps of the 100/300 mb thickness pattern and of the 500 mb heights. September 1957 (not shown) had positive anomalies of the 100/300 mb. thickness pattern over Canada and Alaska, and from Alaska across the Arctic Basin to Novaja Zemlja. This anomaly distribution may have induced the corresponding 500 mb anomaly pattern in Fig. 8, October 1957, giving low zonal index in temperate latitudes in the western Hemisphere (cf. FRAZIER, October 1957).

Fig. 9 shows the anomalies of the 100/300 mb thickness pattern in October 1957. From Northeastern America across the North Atlantic, Greenland, the Arctic Basin, the Norwegian Sea, Europe to western Siberia the anomalies in Figs. 8 and 9 are in broad features of opposite sign. This is the usual joint action between the troposphere and the lower stratosphere under the influence of tropospheric disturbances (AUSTIN and KRAWITZ, 1956).

The mean anomaly distribution of the temperature through the troposphere and the lower stratosphere shown in Figs. 8 and 9, is similar to the temperature distribution in the daily polar front waves, that is, indicates progressive waves. This fact may be remembered in considering the alterations of the 500 mb anomaly pattern from October to November 1957 (Figs. 10 and 11), with change of anomaly sign and continued meridionally impressed circulation (cf. WOFFINDEN, November 1957).

An active influence of the stratosphere on the tropospheric circulation appears through the stratospheric changes from October to November 1957 (Fig. 12 a), and the tropospheric changes from November to December (Fig. 12 b). The common feature of both figures is an area with negative anomalies from Alaska across the Arctic Basin to Europe, and positive anomalies above the Atlantic. Both figures mark a change towards increasing zonal index in mid-latitudes.

Fig. 12 c shows the change from November to December 1957 of the mean 100/300 mb thickness anomalies. Comparing the Figs. 11, 12a, b, and c with each other, no traces of any systematic influence from the mean tropospheric circulation on the mean 100/300 mb thickness anomaly pattern are observable. On the contrary, as mentioned above, the change towards increasing zonal index appeared first in the 100/300 mb thickness layer. In Fig. 12c is also a tendency towards a long-stretched shape of the stratospheric circumpolar flow-pattern with the long axis from the North Atlantic across the Arctic to East-Siberia (p. 33) to notice.

In December 1957, the 100/300 mb thickness layer was colder than normal in the Arctic, with the maximum zone near the North Pole on the Alaskan side, and warmer than normal in the North Pacific (Fig. 13). This anomaly pattern is in accordance with

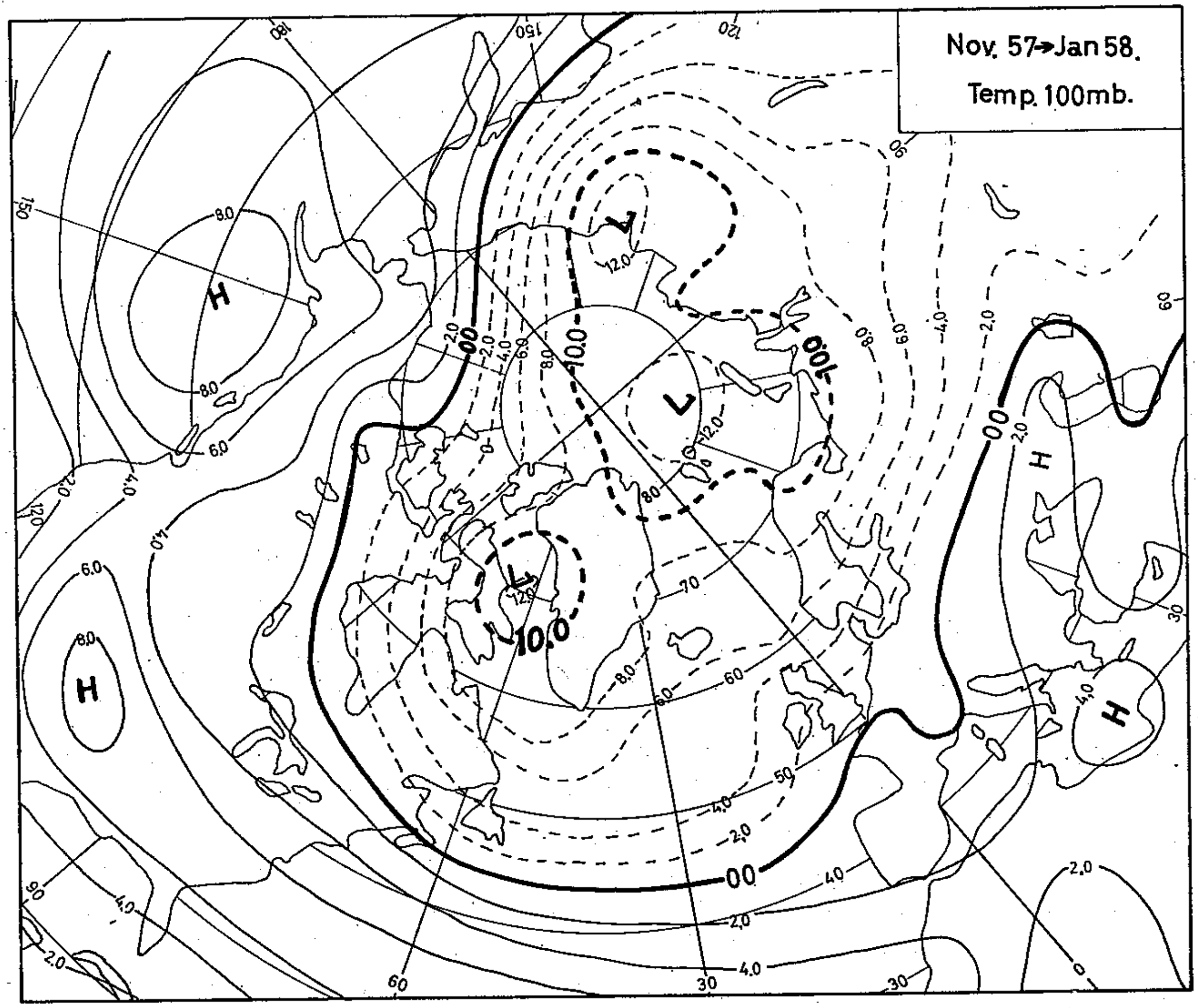


Fig. 5. Change from November 1957 to January 1958 of the monthly mean temperatures at 100 mb.

the high tropospheric zonal index at mid-latitudes in the western Hemisphere in December (cf. DUNN, December 1957).

In January 1958, the warm area moved towards Alaska and Western Canada. The cold dome of the thickness layer intensified and moved towards South Greenland, Novaja Zemlja and Siberia, making three maximum zones (Fig. 14).

Fig. 16 presents the departures from the normal of the 500 mb heights in January 1958. The positive anomalies covered a region from the North Atlantic above Northern America, Greenland and the Arctic Basin to Asia, whereas negative anomalies prevailed in the Aleutian area. This anomaly pattern is nearly the reverse of that in Fig. 14.

It has been previously mentioned (p. 6) that stratospheric warmings in the winter Arctic will favour anticyclonic development, and that cold vortices in the winter stratosphere will favour cyclonic development in the troposphere in the same areas. This hypothesis is founded on the idea that the stratospheric effect on the troposphere is a dynamic one.

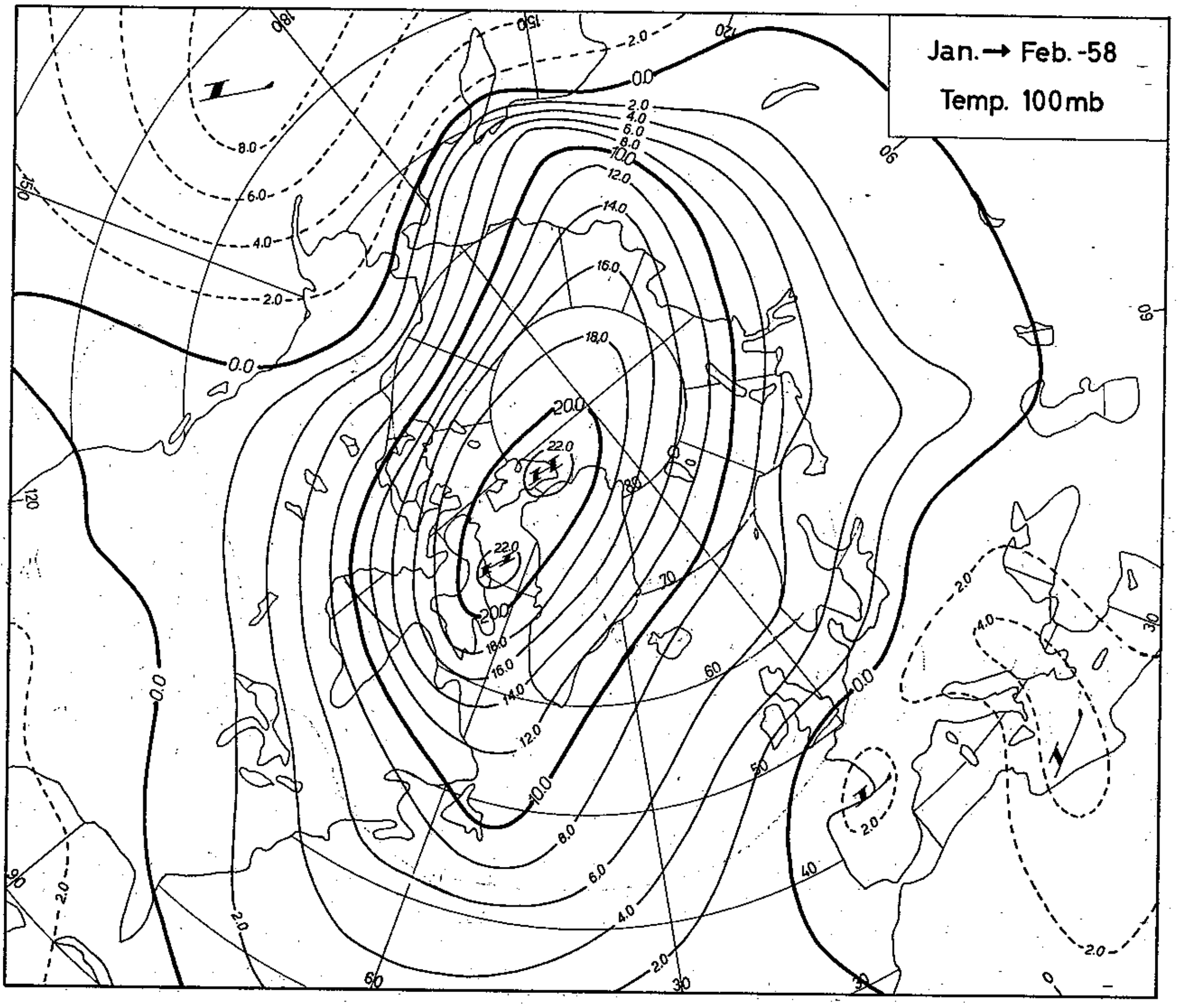


Fig. 6. Change from January to February 1958 of the monthly mean temperatures at 100 mb.

However, the strong, cold vortex in the arctic stratosphere seems, during the greater part of January 1958, not to have influenced the troposphere dynamically, probably with the exception of the Norwegian Sea and North Europe. The anomaly patterns in the Arctic in Figs. 14 and 16 correspond to each other as in situations influenced by tropospheric developments. This explanation is, however, very improbable here. The stratospheric warming over the Aleutians appeared at 10 mb as early as in mid-November 1957 (Teweles and Finger, July 1958), and the anomaly patterns in Figs. 14 and 16 above the Aleutian-Alaskan area are most probably a stationary effect of the Himalaya Mountains on the jet stream.

Teweles (1958) has pointed out this effect from a paper of Yeh (1950): "After passing beyond Japan the jet stream decelerates and in so doing, forces an indirect circulation in the vertical cross-section taken north-south in the central Pacific. In the Aleutian area, the indirect circulation requires rising motion below the level of the jet core and sinking motion above, and in consequence forces the formation of a semi-permanent cold low in the troposphere and a warm ridge in the stratosphere."

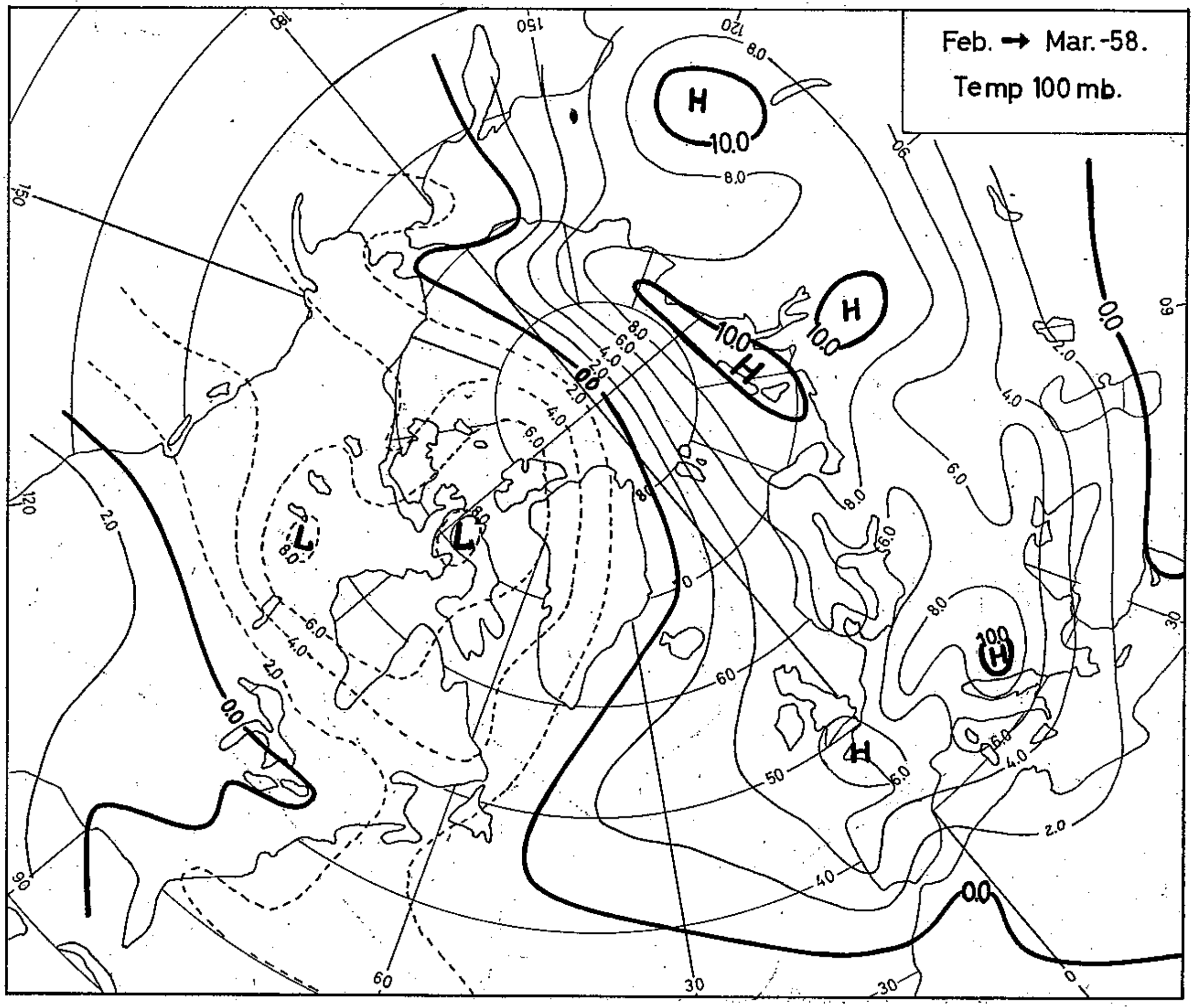


Fig. 7. Change from February to March 1958 of the monthly mean temperatures at 100 mb.

Lahey, et al. (1960) have shown that between October and November, and again between November and December, there are marked increases in the mean strength of the jet stream over Japan. This portion of the jet stream remains at a peak strength through February, and has lost from one-half to three-quarters of its energy by the time it crosses the North American coastline. The conversion of kinetic energy into potential energy requires the above mentioned arrangement of vertical motions (Teweles, 1960).

It may be that the Himalaya effect on the jet stream can have induced the anomaly pattern of the 100/300 mb thickness layer and the 500 mb heights over Siberia and North America too, these areas getting impulses of opposite phase in relation to the North Pacific.

As to the central Arctic, however, the most probable explanation is that a cold stratospheric core, without causing extra cyclonic development in the troposphere, is connected with relatively high tropospheric pressure.

This idea is also in accord with the inflow (flux) to the Arctic during the fall and early winter in the undisturbed circumpolar vortex, branching in subsidence to the

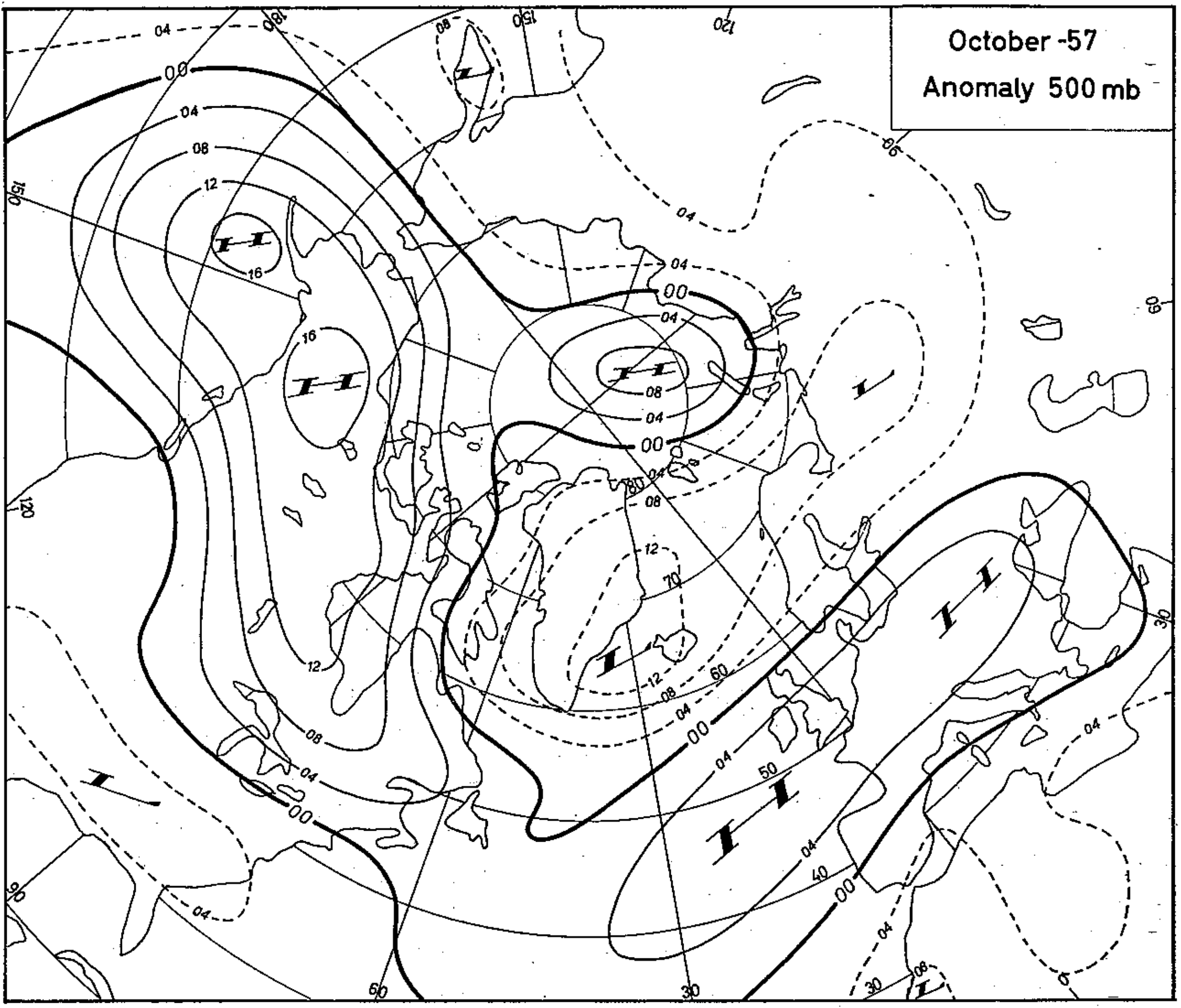


Fig. 8. October 1957. Mean departures from normal of the 500 mb heights. Numbers in decametres.

middle and lower troposphere and in ascendance to the upper stratosphere and the lower mesosphere (see later p. 35).

Schumacher (1958, p. 86—89) has discussed the relations between different thickness layers and heights from the radiosondes at Maudheim. The result was negative correlation coefficients between $\triangle H_{500}$ and $\triangle h_{150/200}$ throughout the year. $\triangle H$ refers to interdiurnal height differences, and the calculation is based on simultaneous readings.

By studying the circumpolar vortex at different levels up to 10 mb in January 1958 (Behr — Scherhag — Warnecke, 1960a, b and c), it will also be seen that the tropospheric circulation in great regions in the Arctic was rather unaffected by the stratospheric vortex during most of January. In the stratosphere, the strong vortex was narrow with a deep center, while the mean circumpolar vortex in the troposphere was more expanded, forming a "flat" center.

An exception formed over the Norwegian Sea and Europe where the stratospheric

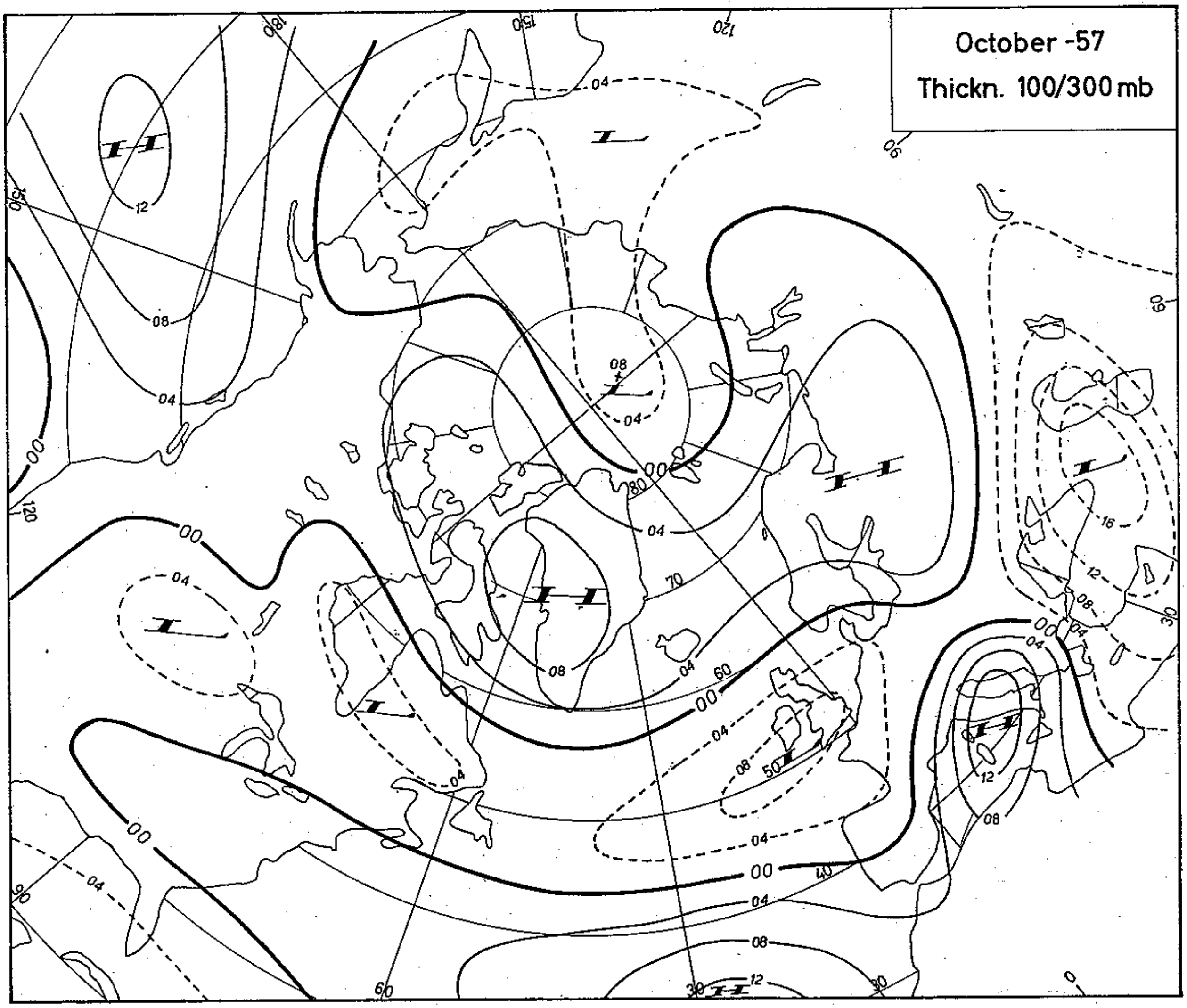


Fig. 9. October 1957. Mean departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer. Numbers in decametres.

disturbances seem to have affected the troposphere. Especially during the last part of January, the great stratospheric overturnings influenced the troposphere immediately.

The mean 100 mb contours and isotherms for December 1957, January and February 1958 are presented in Figs. 17—19, respectively. Fig. 17 shows a nearly circular vortex with the centre and the lowest temperatures near the North Pole. In January 1958, the mean vortex at 100 mb is extended with the long axis from Greenland and the Norwegian Sea to Siberia, and with the coldest air between the North Pole and Novaja Zemlja.

The mean pattern in the 100 mb surface, after the great warmings had been carried through in the first days of February, is presented in Fig. 19. The mean vortex has continued an oval shape with the long axis across Scandinavia and Siberia, and with the centre above Novaja Zemlja. The contours are forming a local through above Hudson Bay and central Canada.

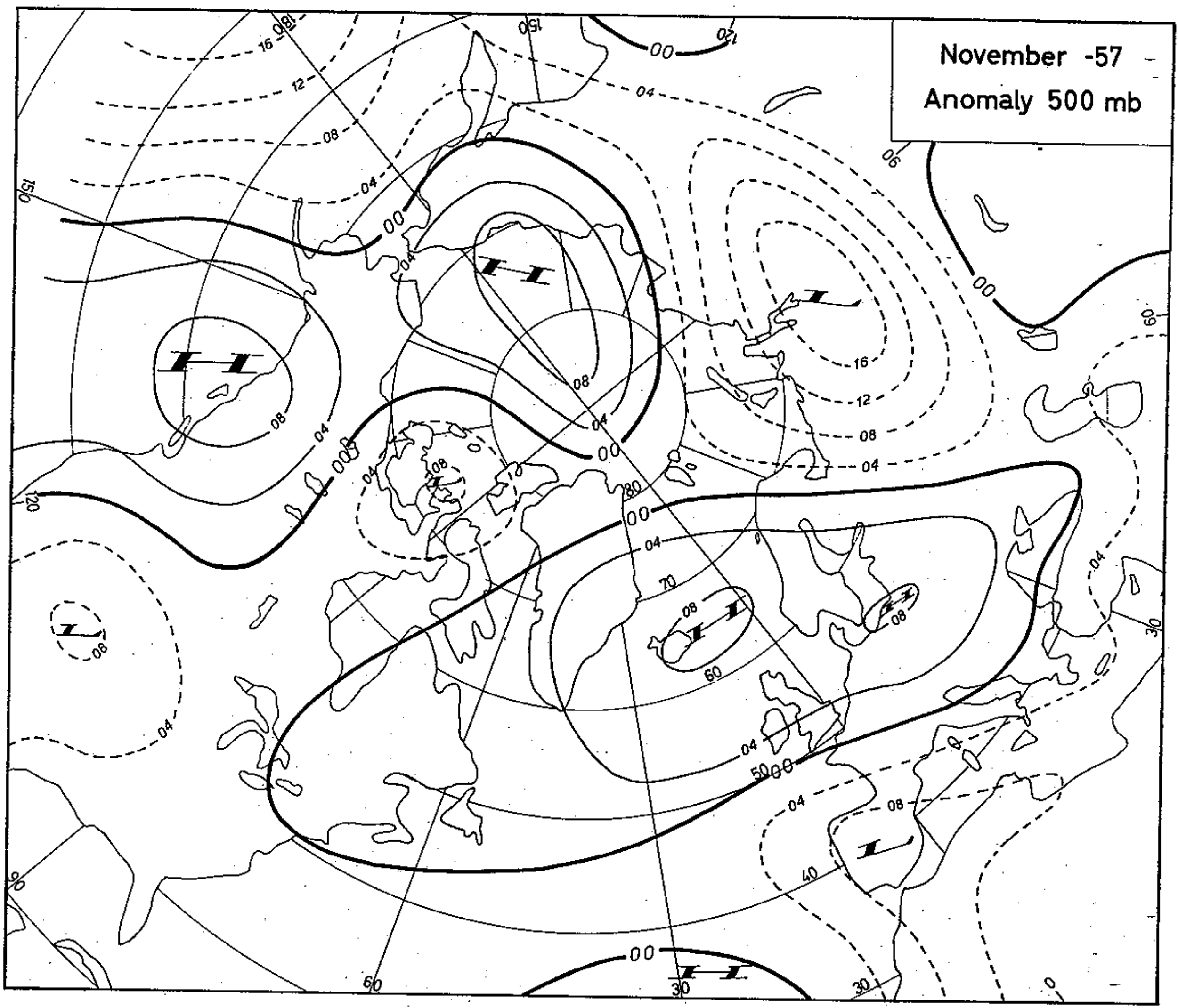


Fig. 10. November 1957. Mean anomalies of the 500 mb heights.

The temperature pattern in the 100 mb surface February 1958 can only be explained by a continual and wide-ranging regime of vertical motions. Assuming stationary conditions, subsiding motions must have been prevailing from the cold core to the warm core, and ascending motions from the warm to the cold core.

The great difference between the mean temperature patterns in the troposphere and in the lower stratosphere in February is shown in Figs. 20a and 20b, presenting the 300/1000 mb and 100/300 mb thickness layer, respectively. The great temperature changes in the stratosphere have only reached the tropopause layer (lowered the tropopause).

The influence of the large stratospheric changes on the tropospheric circulation will now be discussed. This effect appears at first not so convincing because, as mentioned above, the tropospheric circulation in great regions in the Arctic seems to have been rather unaffected dynamically by the stratosphere for a time prior to the changes. In great areas the 500 mb surface had an anomaly pattern, too, which was to be ex-

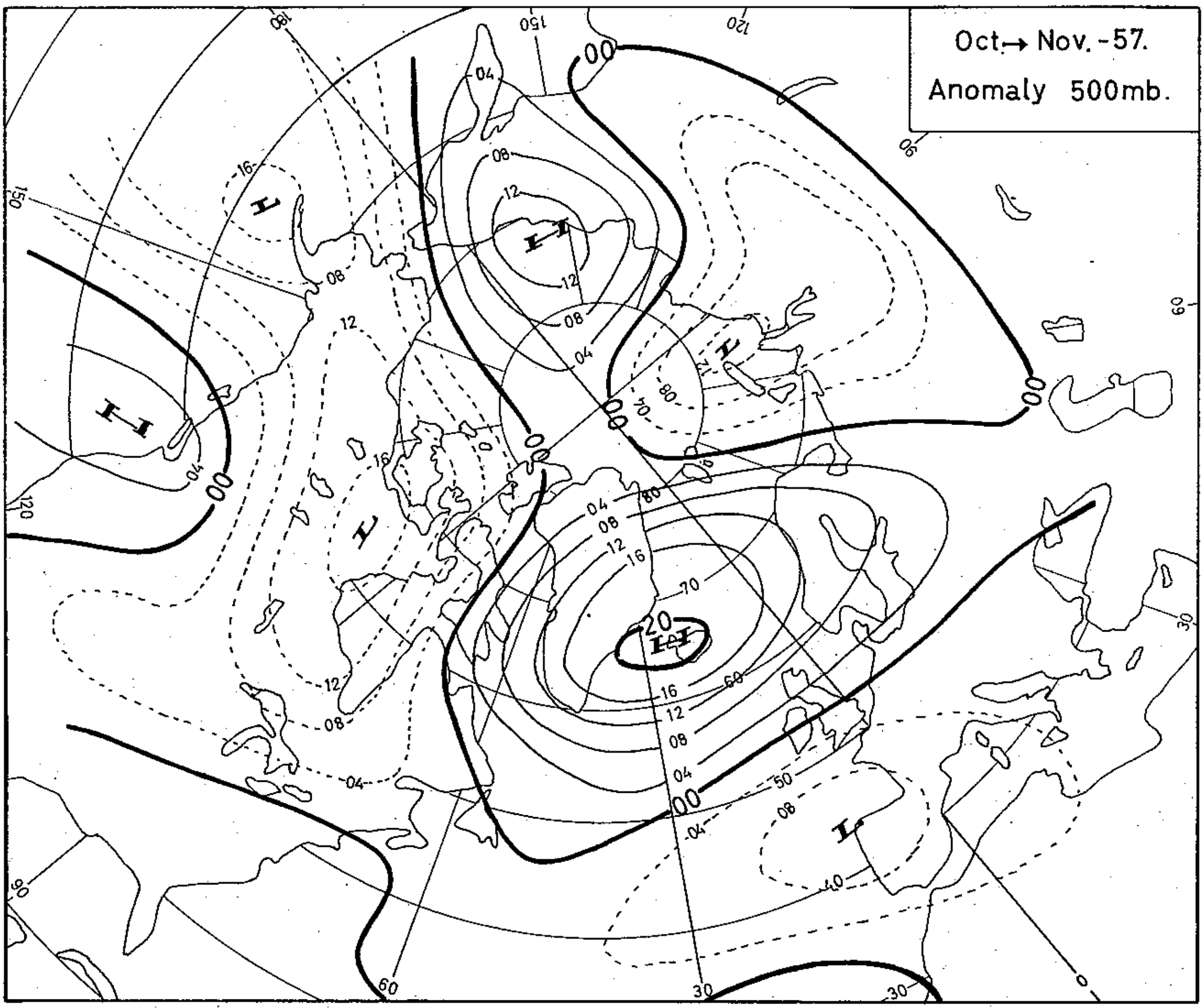


Fig. 11. Change from October to November 1957 of the mean monthly anomalies of the 500 mb heights.

pected after the warming. The long duration of this pattern, however, must be attributed to the stratospheric warming.

On Fig. 19 two arrows *A* and *B* are plotted. *A* is the displacement December 1957 —January-February 1958 of the mean vortex centre at 100 mb, and *B* is the displacement of the 500 mb mean vortex centre from February to March. During the next months February to March and March to April, respectively, both the 100 mb and the 500 mb vortex centre moved eastward, the 500 mb centre a little shorter than the 100 mb centre.

On the 500 mb map February 1958 (Figs. 21 and 22) the trough above North Europe and Siberia, the ridges above Alaska and Greenland and the trough over Hudson Bay are in accordance with the 100 mb map (Fig. 19), but not the trough northwest of Greenland.

As January 1958 is the month of transition, the mean change from December 1957 to February 1958 has been calculated, also. Fig 23a shows such a change of the 100 mb

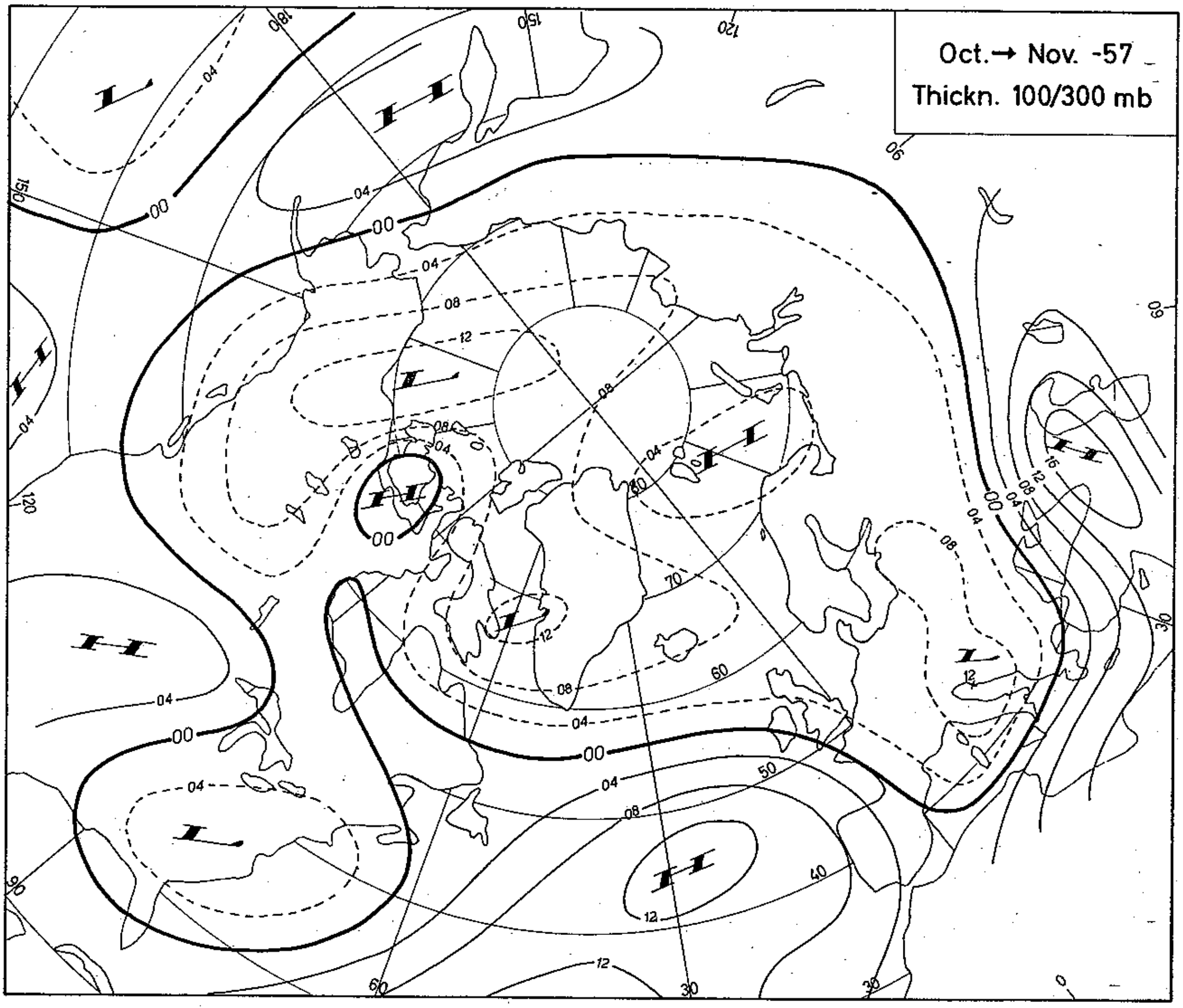


Fig. 12a. Change from October to November 1957 of the mean monthly departures from the 5-year mean (1949—53) of the 100/300 mb thickness pattern.

mean heights. Fig. 23b presents the sum of the 500 mb anomalies in February and March 1958. It is seen that these two maps have positive and negative areas in very good accordance, but there is one particular difference, which is quite peculiar too. The fact is that nearly straight beneath the maximum peak of Fig. 23a is a local low on Fig. 23b. This singularity may probably be attributed to a local tropospheric effect in connection with the topography. As mentioned above, the stratospheric warming penetrated to the troposphere only to a little extent, and the problem here, as in similar situations, is the continuity and the distribution of the cold, arctic airmasses in the troposphere.

During January 1958, warm tropospheric airmasses were transported from the North Pacific across Canada, while the coldest tropospheric airmasses were situated above East Siberia. In connection with the February ridge above the Bering Straits, the cold tropospheric airmasses in East Siberia moved partly westward and partly

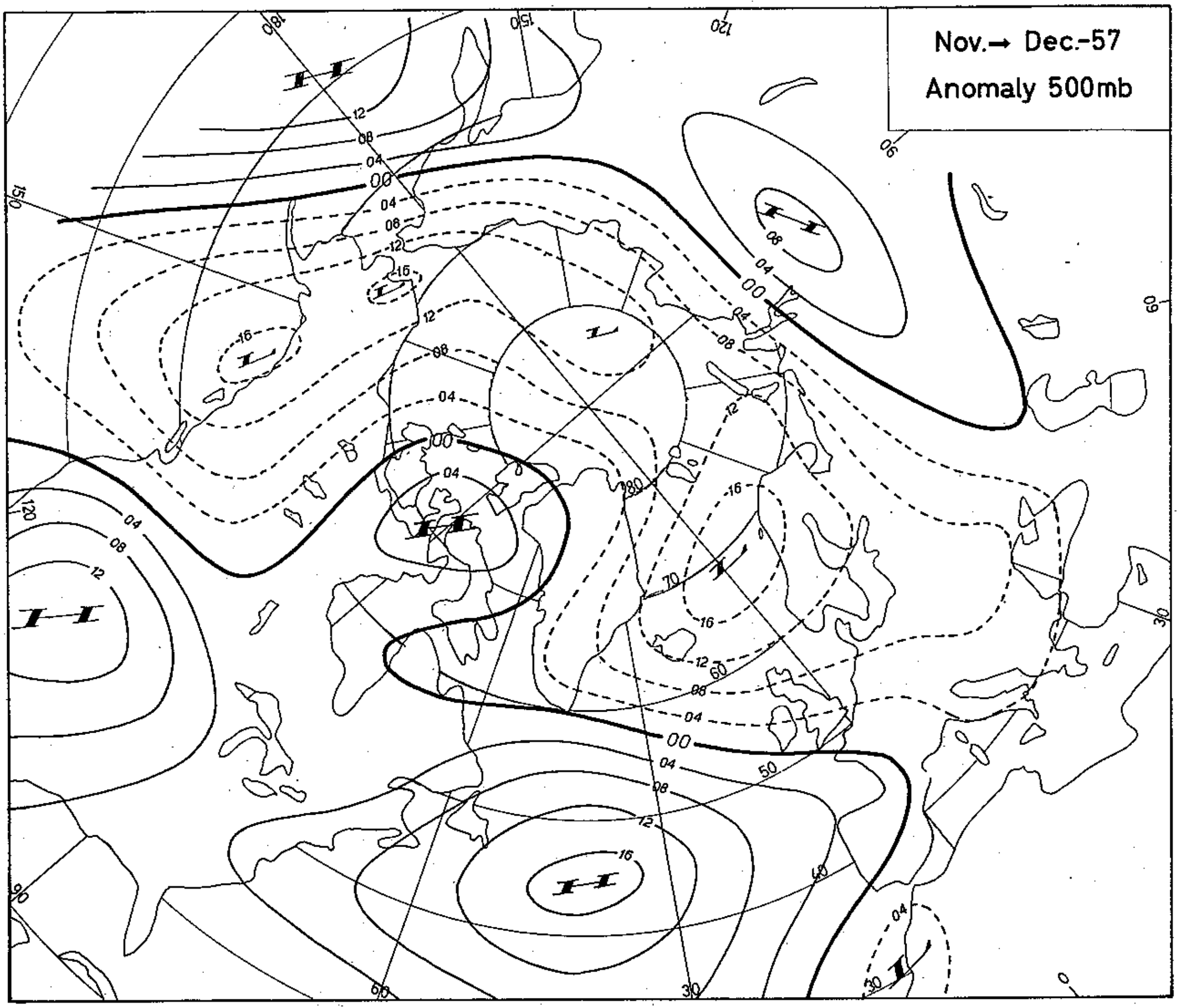


Fig. 12b. Change from November to December 1957 of the mean monthly anomalies of the 500 mb heights.

across the Arctic Basin towards the Canadian Arctic, a natural place for the cold airmass, and formed a trough there in the 500 mb level. At sea level the trough was not so pronounced. This fact, that the coldest airmasses partly moved towards the Canadian Arctic, prevented North Europe from being invaded to such a high degree as the great stratospheric changes seemed to indicate.

The Figs 24a and 24b also have good agreement concerning positive and negative areas, except in the Atlantic and NW of Greenland (the 500 mb singularity).

The mean anomalies of the 500 mb heights after the great warmings in the lower arctic stratosphere during the last week of January 1958 are presented in Figs 22, 23b and 24b. The figures illustrate low zonal index in polar and temperate latitudes, and a high zonal index in subtropical latitudes (i.e., an expanded circumpolar vortex). This is a result which was to be expected, using the ideas of Eady concerning the effect of the stratospheric circulation on the troposphere.

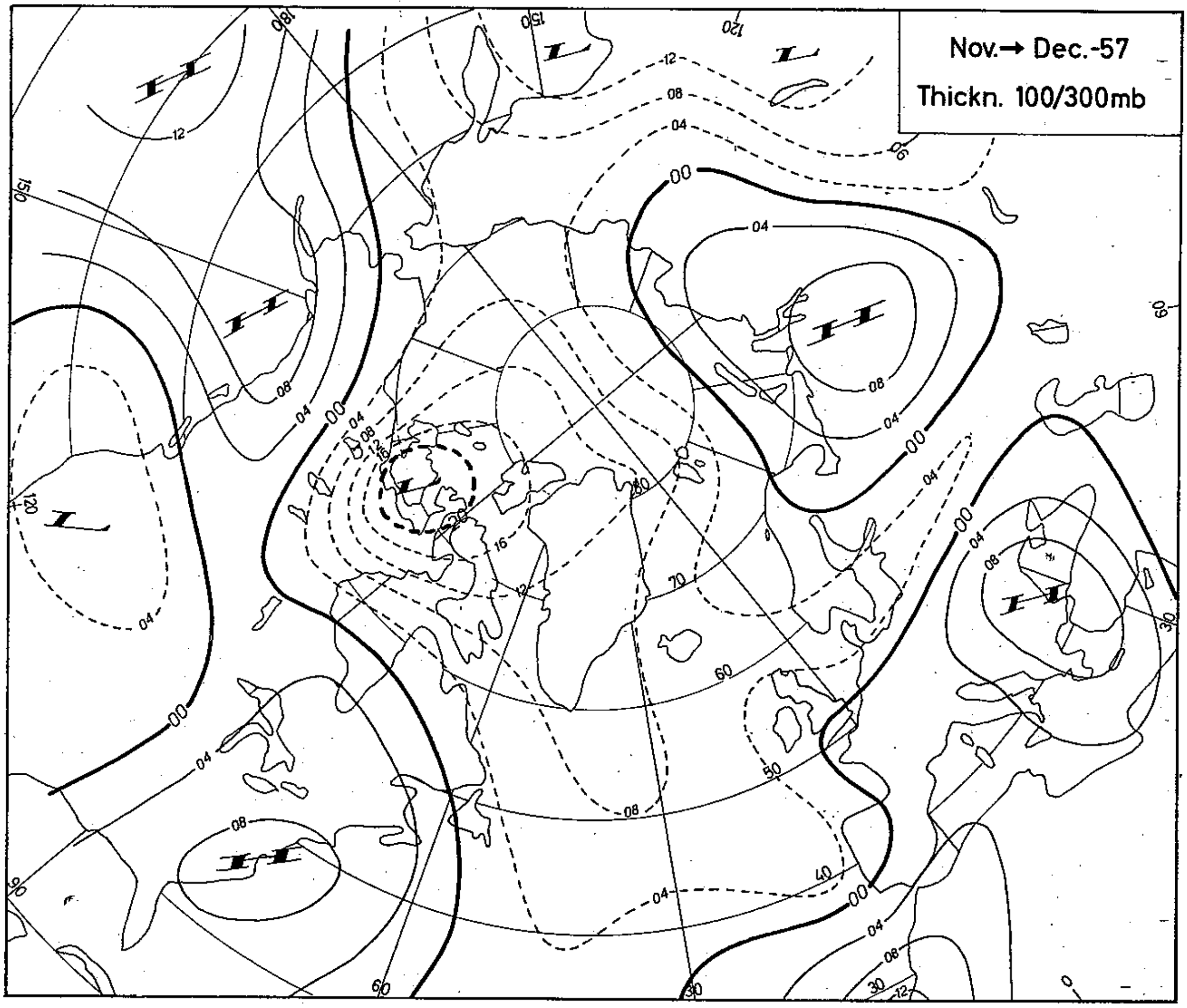


Fig. 12c. Change from November to December 1957 of the mean monthly departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer.

In discussing the circulation patterns of January and February 1958, O'CONNOR (1958) and KLEIN (1958) point out a marked index cycle in the Western Hemisphere westerlies at 700 mb (NAMIAS, 1950). Figs 25 and 26 illustrate this. It is, however, interesting to note a peculiarity here. The change of the zonal index in the Western Hemisphere had begun already on 21. December 1957, and the most rapid declining of the temperate westerlies occurred between 11. and 18. January 1958 (Fig. 26). From Fig. 28 it is seen that the warming at 100 mb began 13. January 1958 at Aklavik, 20. January at Alert and Thule, and 21. January at Clyde, Frobisher and Keflavik; That is, the strong declining of the temperate westerlies at 700 mb began previous to the warming at 100 mb in the Canadian Arctic. The strong anticyclonic ridge above the Davis Straits was built up during the last half of January 1958 (O'CONNOR 1958), but chiefly at a time when the same area in the stratosphere was influenced by the cold

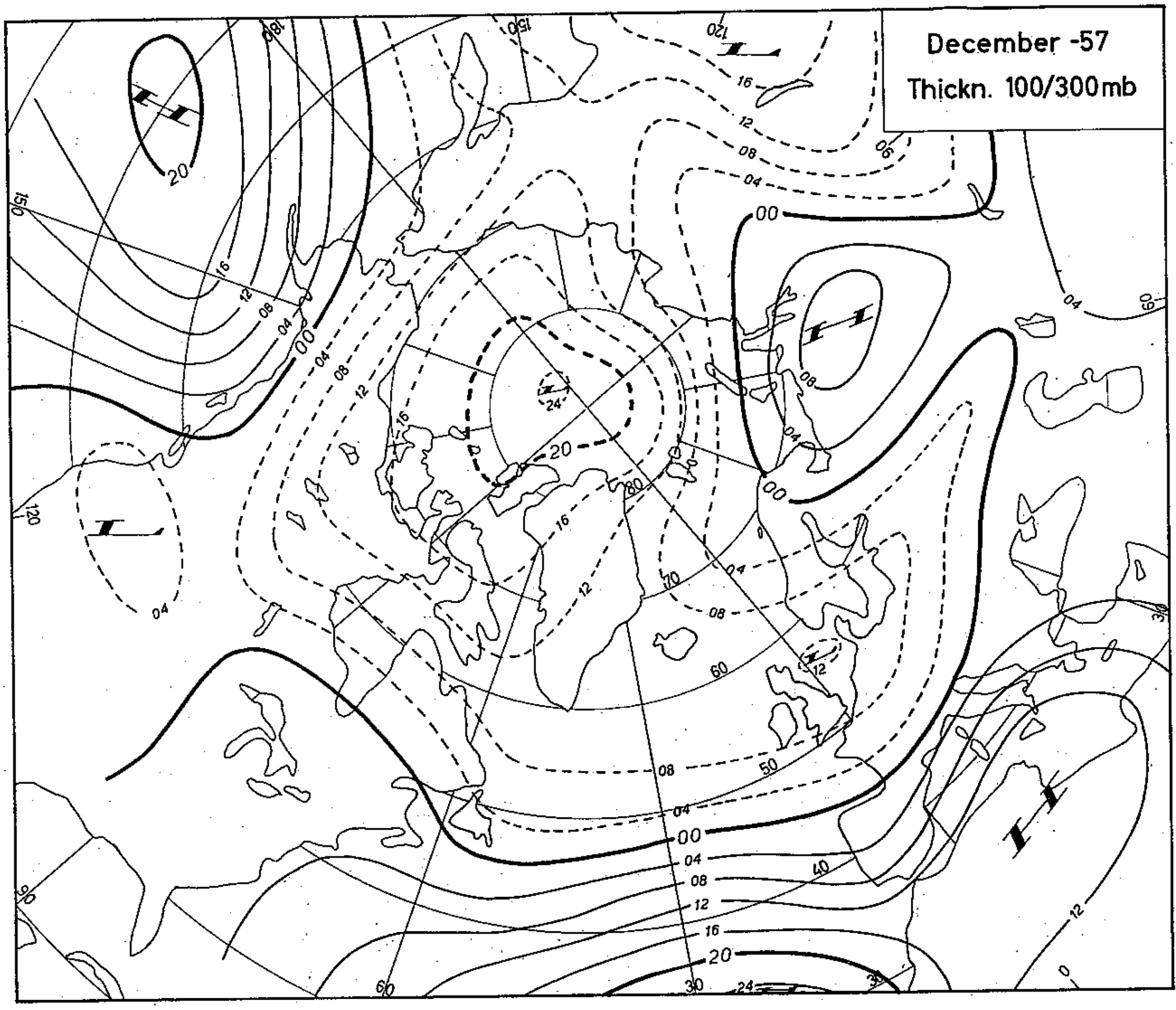


Fig. 13. December 1957. Mean departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer.

vortex. As mentioned above (p. 13) this is in contradiction with the usual idea of a circulation effect of the stratosphere on the troposphere.

Craig (1959) has, after index-analyses of Stark and Woffinden, pointed out that two index minima of the temperate westerlies at 700 mb occurred during the great stratospheric warmings of January—February 1957, the first one in mid-January and the second one in mid-February, at the beginning and at the end of the period of the large stratospheric changes, respectively, but the circulation patterns accompanying the two mid-latitude index minima were strikingly dissimilar. It is perhaps worth-while to note that the declining to the first minimum started previous to the stratospheric warming above North America.

Most probably, this feature must be taken into account by estimating the origin and nature of the atmospheric changes of this kind. Taking into consideration the

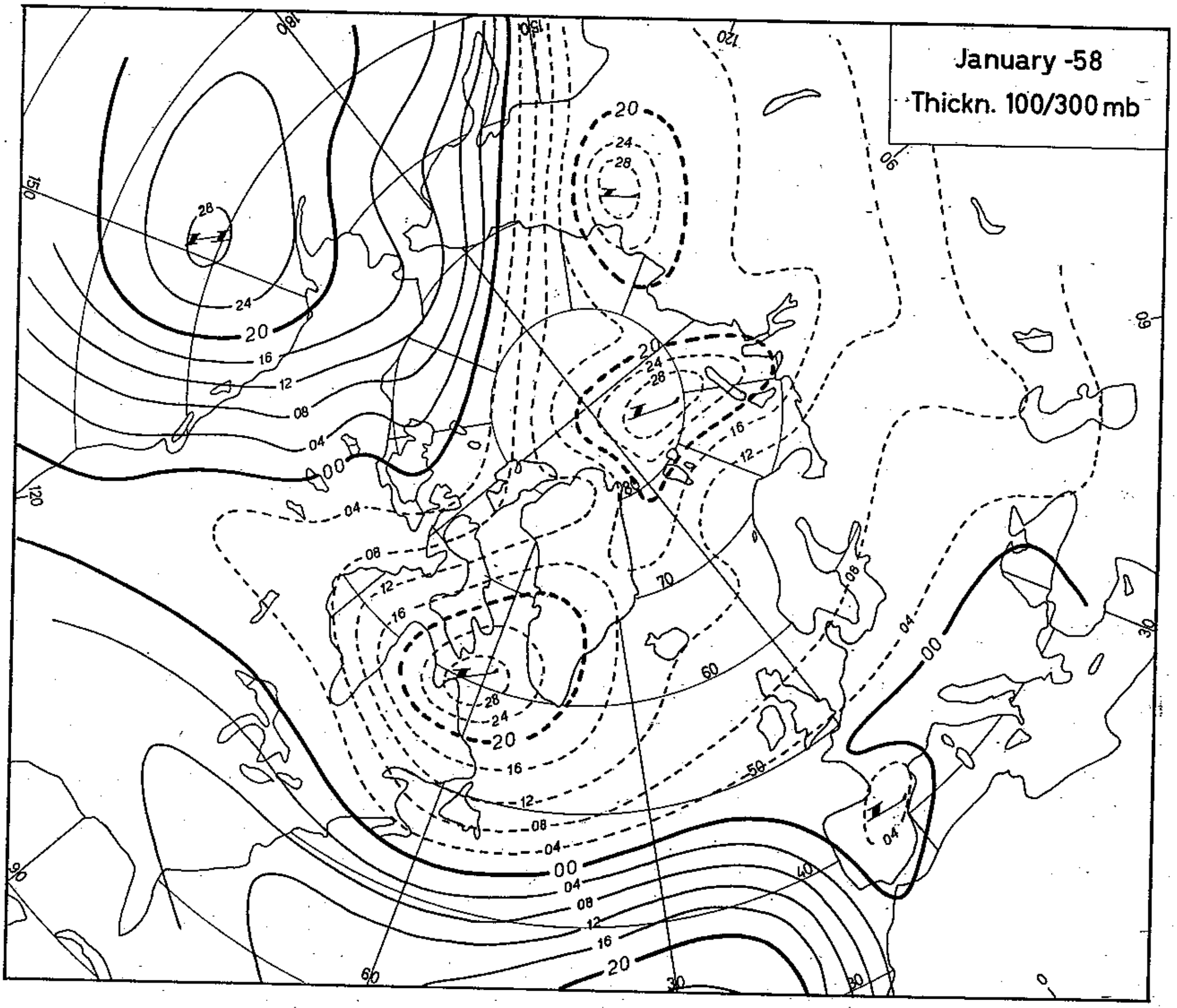


Fig. 14. January 1958. Mean departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer.

mountain effect on the jet stream prior to the great stratospheric warming, we may explain the occurrences as follows:

"The decelerating jet stream from the Himalaya Mountains is enforcing a vertical circulation in the meridional plane over the North Pacific, creating cyclonic development in the troposphere and subsidence in the lower stratosphere. The mountain waves have short wave-length, and over Canada and the Northern U.S. (probably modified and displaced westward by the Rocky Mountains) a meridionally vertical circulation of opposite sign in relation to the North Pacific is induced. When the jet stream is increasing over Himalaya, the vertical circulation effect is also increasing, and the Himalaya effect is increasing in relation to the Rocky Mountain effect, that is, anticyclonic development is increasing in the troposphere over Canada and the Northern U.S. and moving a little eastward. The index cycle is started.

If now strong stratospheric subsidence and warming from higher levels follows the

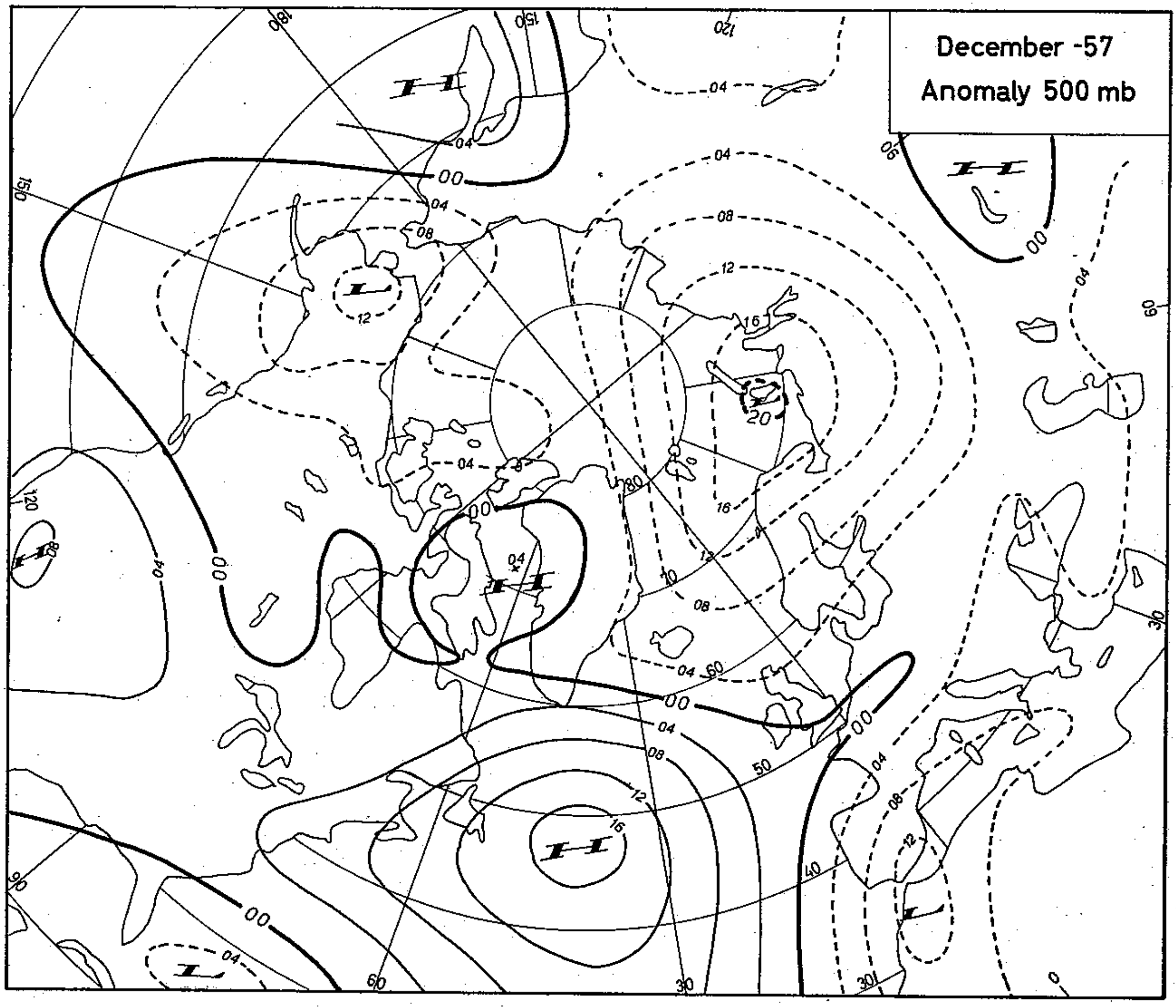


Fig. 15. December 1957. Mean anomalies of the 500 mb heights.

jet-increase over Himalaya, waves of much longer wave-length are created in the stratosphere. Over Canada and the Northern U.S., the general stratospheric subsidence is favoured downward by the pre-existing anticyclonic anomaly pattern in the troposphere. The anticyclonic circulation in the troposphere over Canada and the Northern U.S. will increase further, and the index cycle will be pronounced and of long duration."

In March 1958, the stratospheric circumpolar vortex increased a little again, chiefly in the American sector. The mean change from February to March 1958 of the 100/300 mb anomaly thickness layer is shown in Fig. 25a, and Fig. 25b presents the change of the 500 mb anomaly heights from March to April (the 500 mb anomalies in April were favoured by a continued cold 100/300 mb thickness layer over North America).

c. Discussion of the stratospheric temperature changes. As 1957 and 1958 constitute the IGY, great efforts have been made to analyse the upper-air material of this period to as high a level as it is possible today. There are at present several 10 mb maps from this period under consideration (Teweles and Finger, January 1958 and July 1958, U.S.

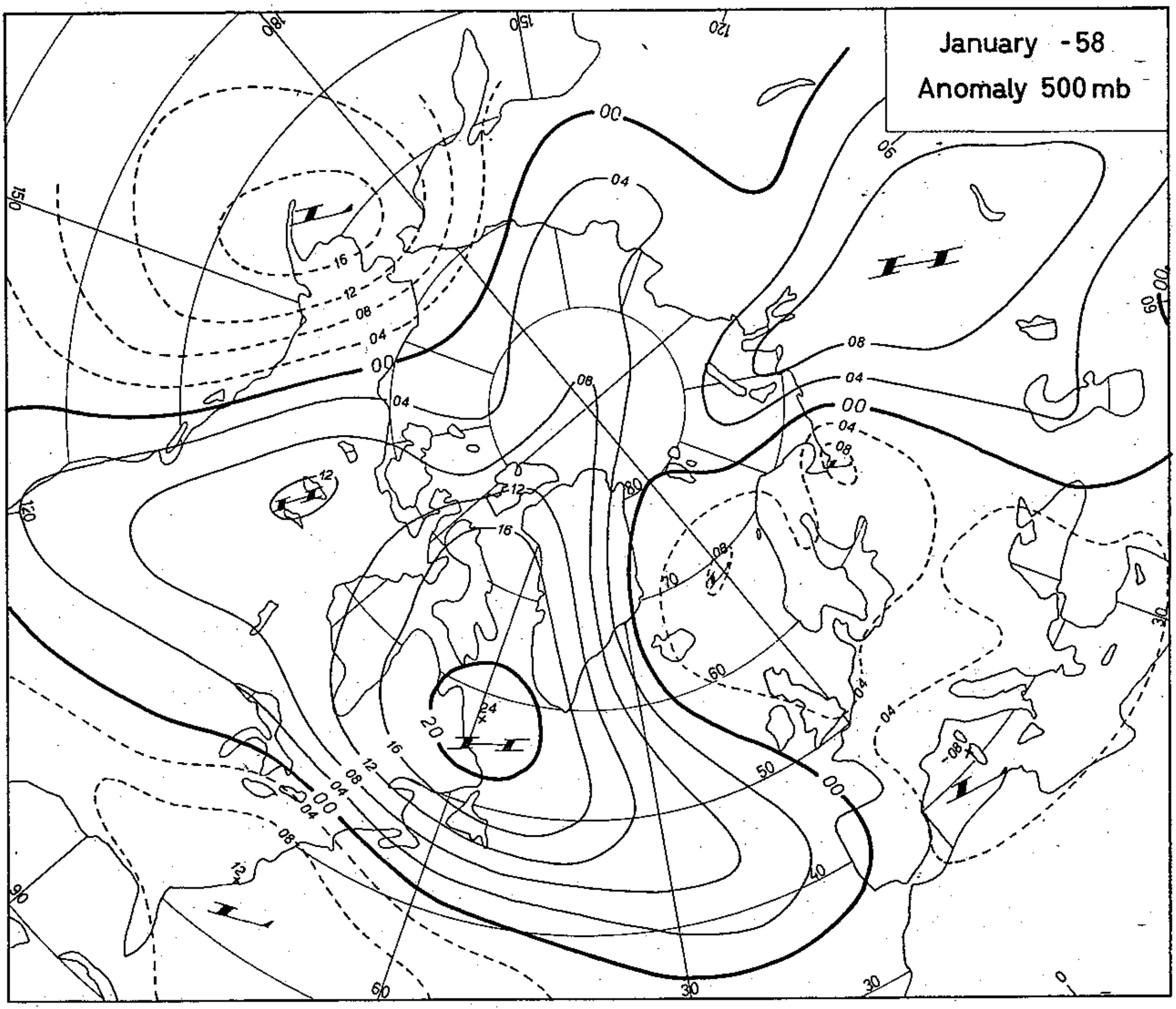


Fig. 16. January 1958. Mean anomalies of the 500 mb heights.

Departmemt of Commerce, December 1959, Behr — Scherhag — Warnecke, 1960c). The mean temperature changes at 100 mb during the winter 1957—58 will now be discussed in connection with the 10 mb maps.

The most marked feature of the mean temperature changes at 100 mb from the fall to the mid-winter is the warming band at temperate to sub-tropical latitudes (Figs. 3—5). The band has its most northern position over the Gulf of Alaska, and appears to be nearly vanishing over India (but is pronounced over Japan). The warming band is most probably an effect of subsidence in the stratosphere, primarily, and secondarily an effect of heat absorption in the ozone layer. During the shift from summer to winter circulation in the fall, a subsidence is created in the stratosphere at temperate to sub-tropical latitudes. In these latitudes, the mean ozone content has its minimum in October (Arosa). This point of time and the increasing ozone content in the atmosphere during the last months of the year at temperate to sub-tropical latitudes is probably influenced by increasing subsidence in the stratosphere.

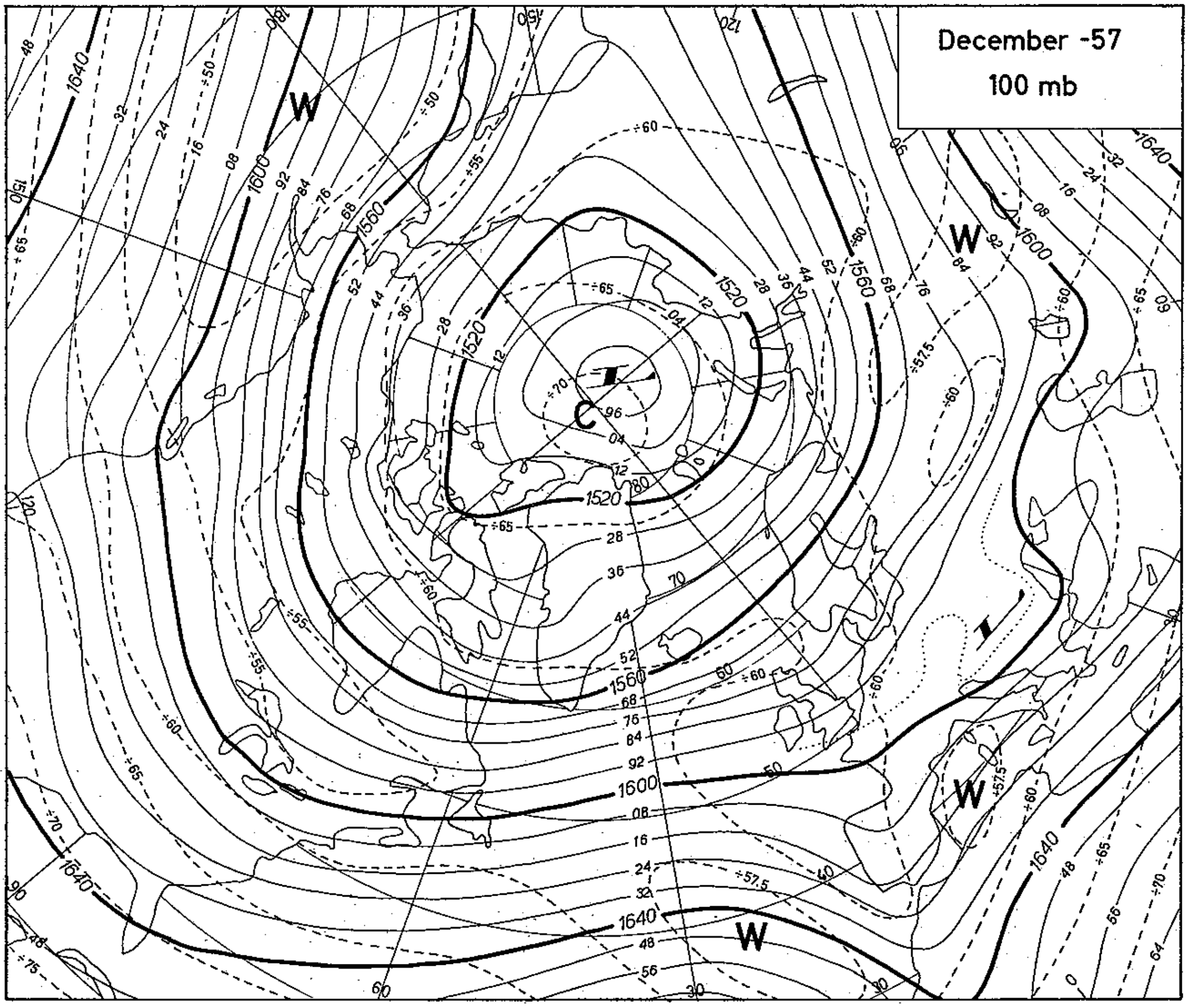


Fig. 17. December 1957. Mean contours (decametres) and isotherms (C°) of the 100 mb surface.

Looking at the latitudinal non-symmetry of the mean cooling at 100 mb in the fall at polar and temperate latitudes, and at the fact that this non-symmetry is changing regionally from month to month, it is reasonable to assume that the radiational cooling in the lower stratosphere in the fall to mid-winter to a high degree is influenced by a long-period regime of vertical motions.

Several authors have pointed out temperature waves with different amplitude and period-length in the middle and lower winter stratosphere, including vertical motions, primarily Godson and Lee (1958), Hare (1960), but also Teweles and Finger (1958), and Craig and Herring (1959).

The predominance of baroclinic waves is one of the most marked features of the stratospheric winter circulation in the Northern Hemisphere. The study of the 10 mb maps from the IGY affirms the concept given above about the lower stratosphere. During the shift from summer to winter circulation a ridge zone was formed at 45°—50°N in the Western Hemisphere in the beginning of September 1957 (U.S. Depart-

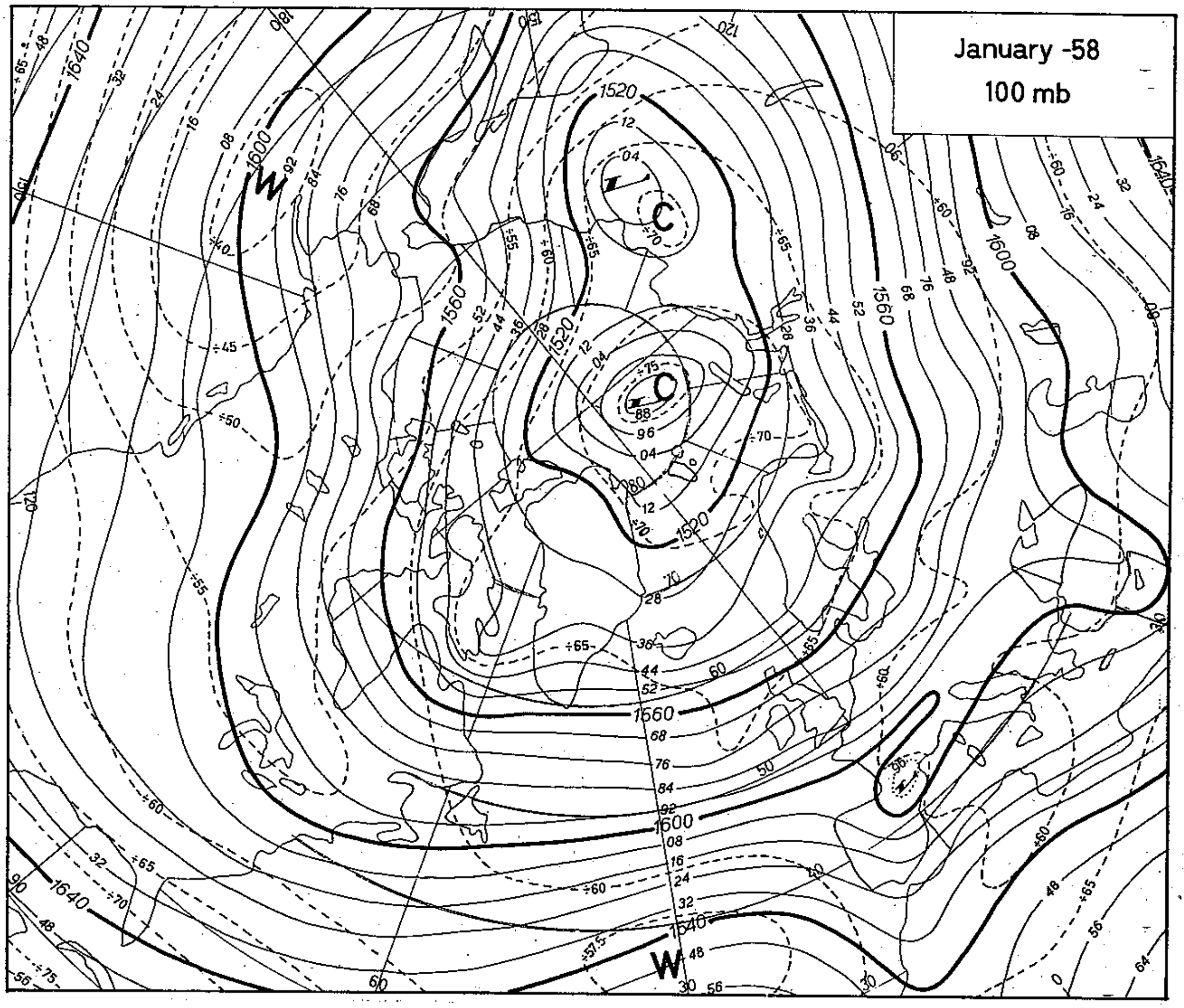


Fig. 18. January 1958. Mean contours and isotherms of the 100 mb surface.

ment of Commerce, December 1959). During September the ridge zone was moving southward to about 35°N and during October into the sub-Tropics (25°—30°N).

In late fall and early winter, the ridge zone at 10 mb is in good agreement with the warming band at 100 mb in Figs. 3—5. Both are split into cells, corresponding to wave motions, and the circumpolar vortex at 10 mb was disturbed by great waves.

About mid-November 1957 a warm subsidence ridge was formed at 10 mb in the Aleutian region. This ridge was soon displaced towards the southwest by colder arctic air in the 10 mb surface.

At New Year 1958 a second, warmer and stronger ridge appeared at 10 mb near the Aleutians. This ridge moved towards Alaska and later into the Arctic. On the Arctic Ice-Floe station "Alpha" (84°N, ca. 160°W) the warming at 50 mb started about 5. January 1958. At 150 mb, the warming appeared about 12. January, and the warm peak was reached on 1. February at all levels up to 10 mb (Wexler, 1959, Teweles 1960). This temperature peak was probably connected with the warming

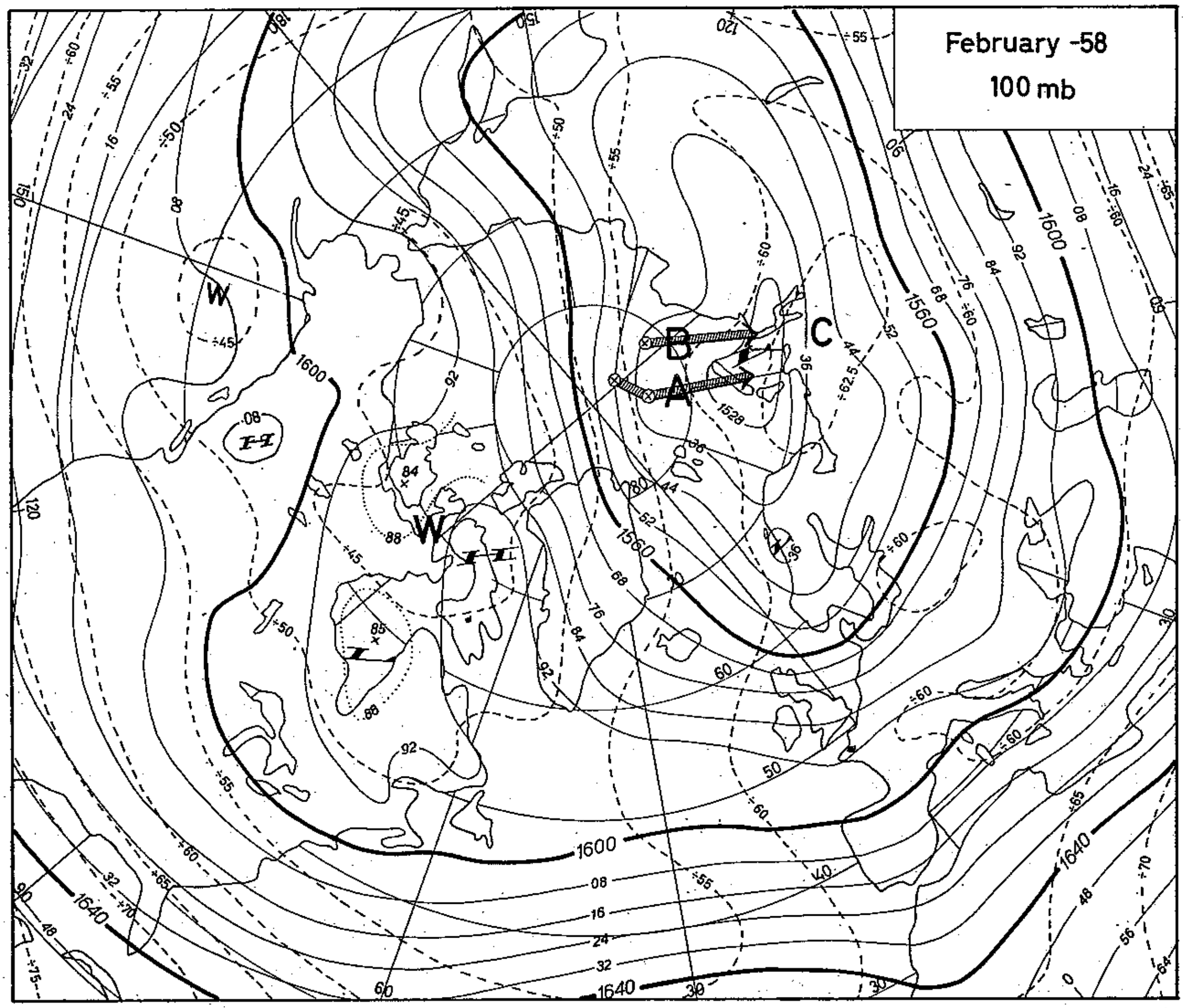


Fig. 19. February 1958. Mean contours and isotherms of the 100 mb surface. The arrow *A* presents the movement of the 100 mb mean vortex center from December 1957 through January to February 1958, and the arrow *B* shows the displacement of the 500 mb mean vortex center from February to March 1958.

wave running northward from the North Atlantic across Greenland. After a pendulating motion, the 10 mb high moved back to the Aleutian area and the North Pacific in the last half of February.

In the meantime, however, several other stronger revolutions had occurred in the mid-stratosphere, unveiled by Institut für Meteorologie und Geophysik, der Freien Universität, Berlin (BEHR, JACOBS, PETZOLDT, SCHERHAG, WARNECKE, 1960, a, b, c). In the first half of January 1958, attempts at warm ridge rises occurred at 10 mb from the Atlantic to Southern Asia, but they dropped off again due to sweeping cold troughs.

On 22. January, however, a warm subsidence ridge was growing over the Orient. It seems that two ridge attempts, one from the Atlantic and the other from Southern Asia, were meeting over Arabia and hence moved northward across Europe towards the Arctic. On 25. January the ridge development met the corresponding one from

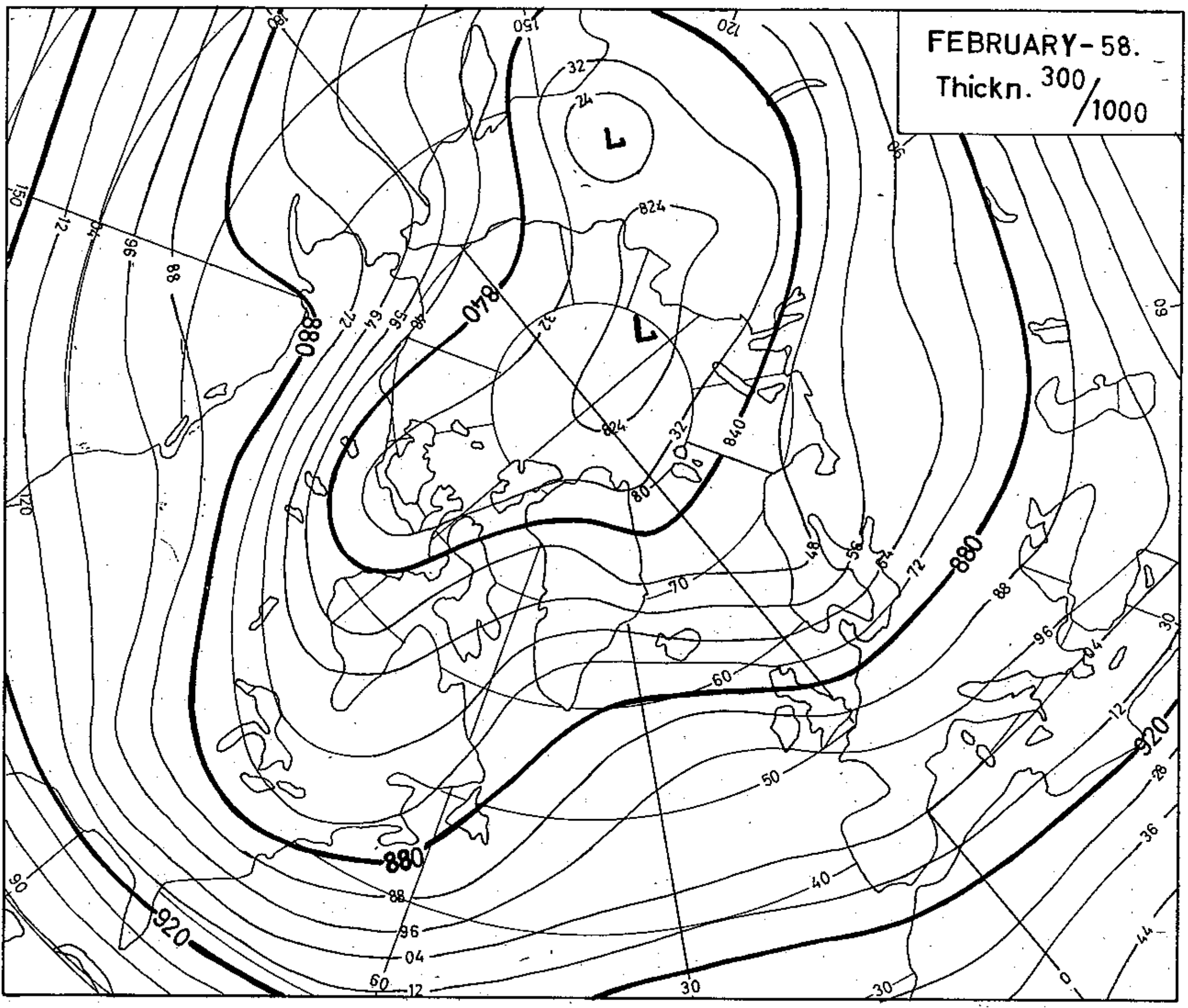


Fig. 20a. February 1958. Mean heights of the 300/1000 mb thickness layer. Numbers in decametres.

the Aleutian side over the Siberian sector of the Arctic Basin, and the result of the amalgamation of the two ridges was a high on 1. February over the American and East Siberian Arctic.

The stratospheric circumpolar vortex at 10 mb was broken down, splitting into two vortices. The western vortex moved across North America towards the Aleutians, steadily weakening, and disappeared later. The eastern vortex strengthened and moved westward across Siberia to North-western Europe (3. February). A new ridge formation started over the Orient, moved northward, broke down and parted the Siberian-European vortex in two, both parts weakening. The Orient high center later moved westward.

On 10. February, the 10 mb map was dominated by a strong high in the Arctic, with only weak local vortices in the Northern Hemisphere.

During the last part of February and the first half of March the Siberian vortex (at 10 mb) strengthened somewhat again and moved across the European Arctic to the American Arctic.

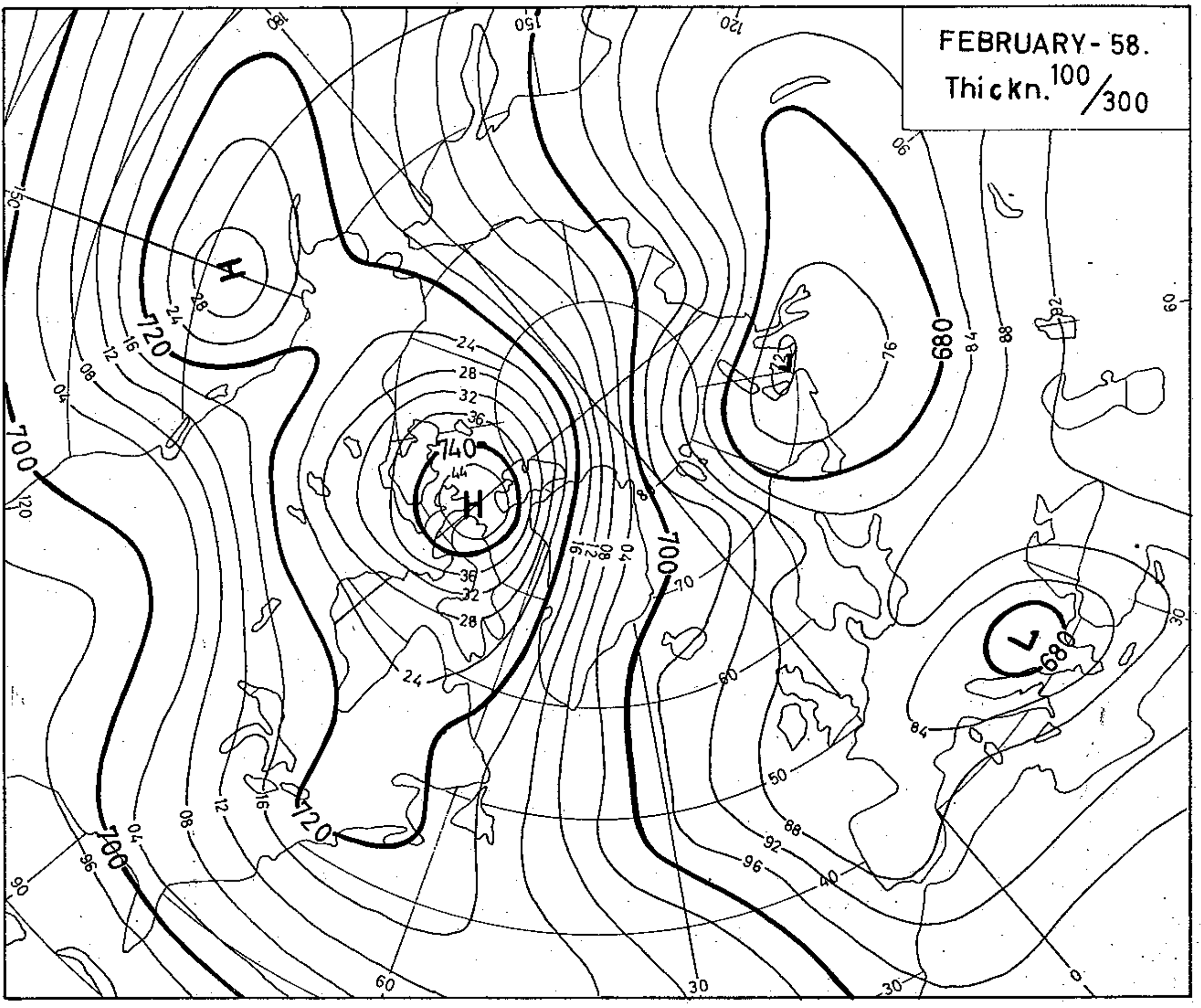


Fig. 20b. February 1958. Mean heights of the 100/300 mb thickness layer.

d. Causes of the stratospheric winter warmings and of the breakdown of the circumpolar vortex. While the changes at 10 mb during the period 20. January to 10. February 1958 were very complicated and involved difficulties in estimating the importance of the different occurrences, the picture was clearer at 100 mb.

Fig. 26 presents 5-day running means of the 100 mb temperature along three cross-sections during the period December 1957 through February 1958. From Aklavik to Alert the point in time of the warming is delayed, and the temperature amplitude is increasing. The cross-section Alert—Frobisher Bay shows only little variation of the warming epochs and the temperature amplitude. The cross-section Aklavik—Frobisher—Keflavik—Trappes shows delayed warming, increasing amplitude to Frobisher, but decreasing amplitude from Frobisher to Valentia and no warming at Trappes.

The geographical distribution of the temperature amplitude of the great warmings, shown by the cross-sections, is in good agreement with Fig. 4. The cross-sections show that at 100 mb the warming moved from Alaska eastward, and had probably started near the coast of North-eastern Asia (Japan) in December 1957.

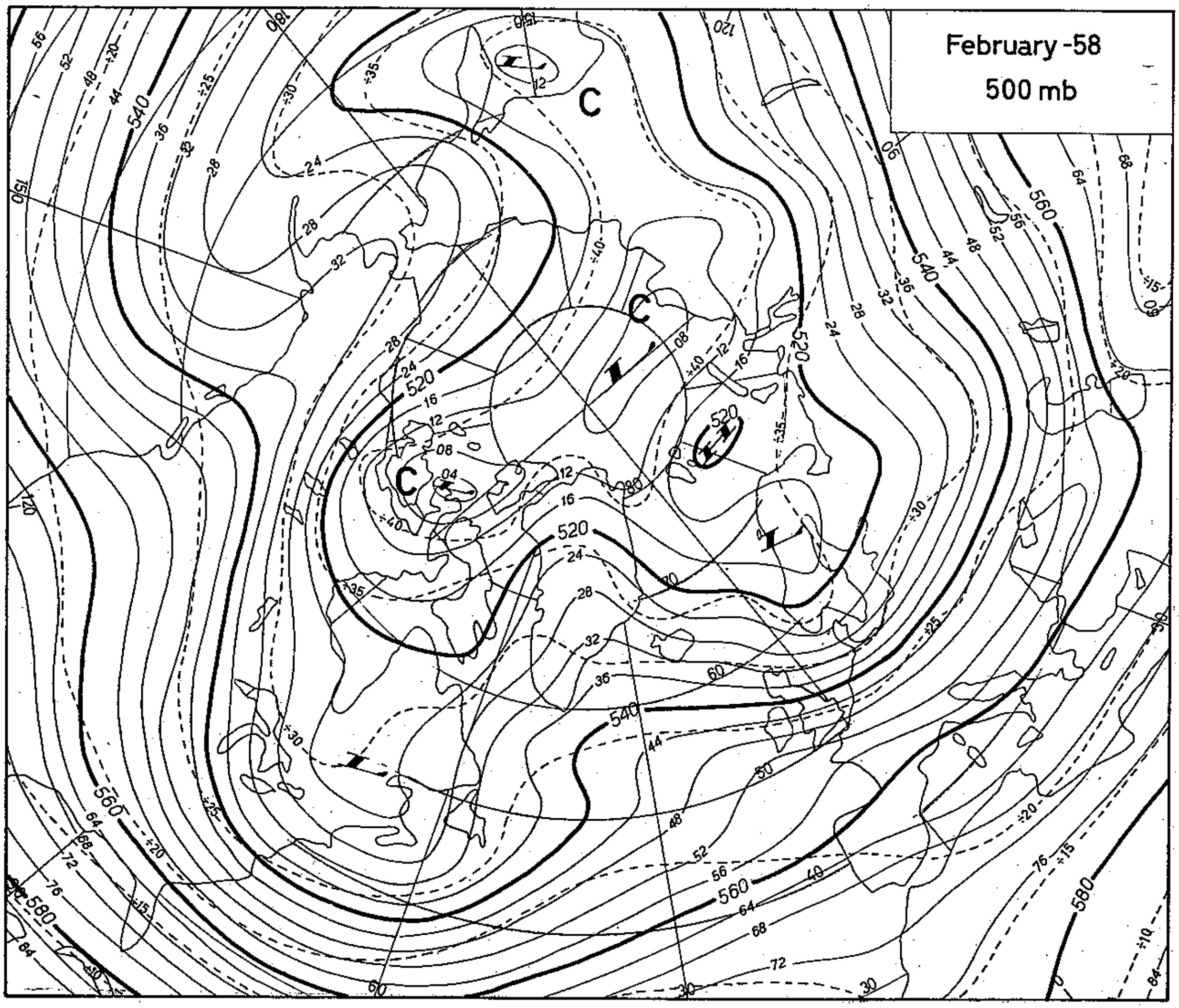


Fig. 21. February 1958. Mean contours and isotherms of the 500 mb surface.

The advection of the warming is not a mere temperature transport, but is chiefly an advection of a disposition to subsidence.

Karin Behr (1960) has summarized the papers on the great winter warmings in the northern-hemispheric stratosphere. The writer would like to give a contribution to this discussion, based on the available papers and the 10 mb maps from the IGY.

As the definition of the upper stratosphere and the lower mesosphere are not quite clear, the writer will explain the terminology used in the following:

tropopause to 20 km — lower stratosphere
20 to 30 km — middle stratosphere
30 km to stratopause (35—40 km) — upper stratosphere
stratopause (35—40 km) to mesopeak (50 km) — lower mesosphere
mesopeak (50 km) to mesopause (80 km) — upper mesosphere

The definition above of the upper stratosphere is affixed to the undisturbed atmosphere, and covers a transitional layer with increasing temperature gradient upward.

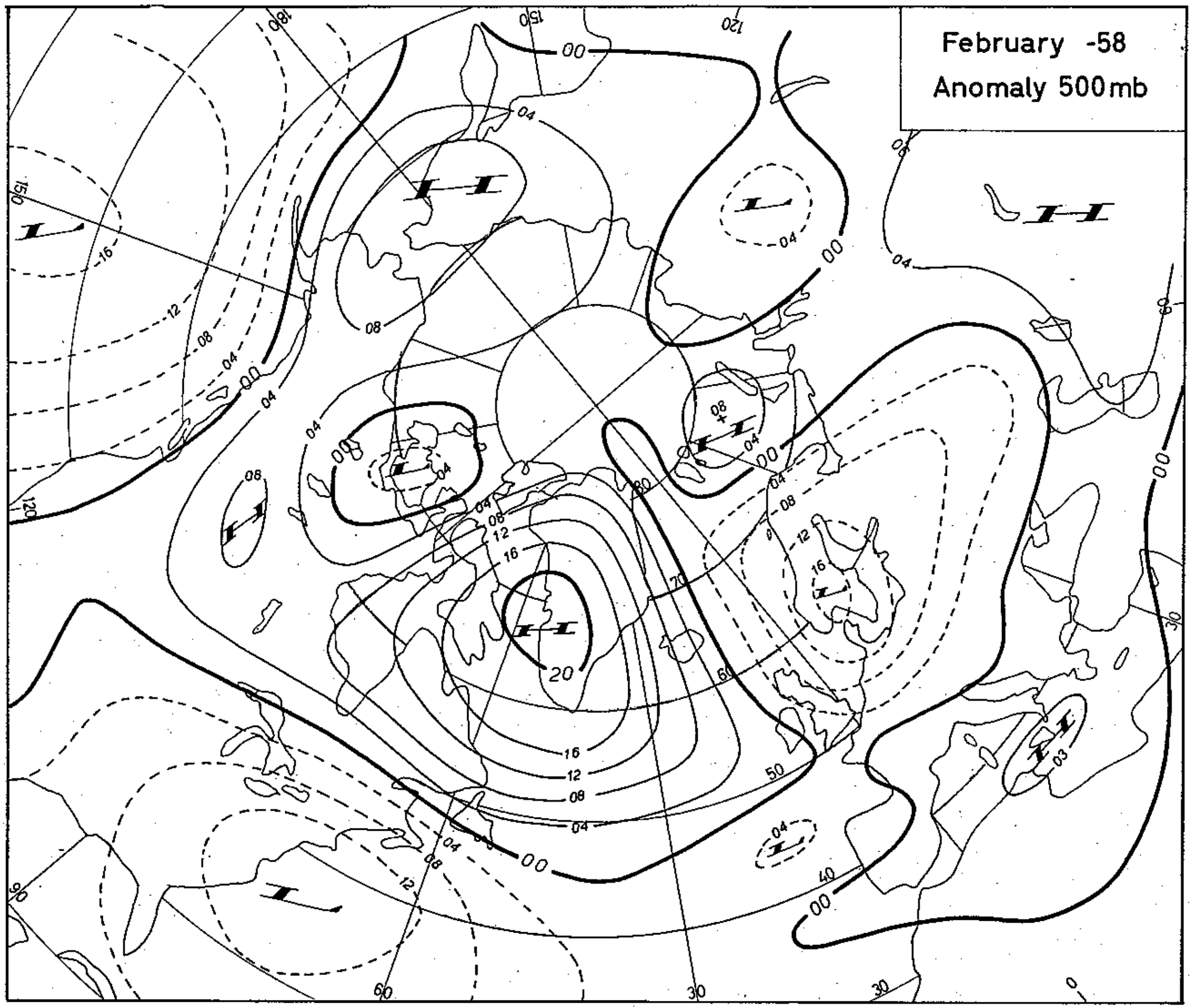


Fig. 22. February 1958. Mean anomalies of the 500 mb heights.

The stratopause, however, is strongly influenced by the stratospheric-mesospheric disturbances.

First, the resemblance between the pattern of the mean temperature changes in Figs. 3—5 and the contour pattern on the daily 10 mb maps in January 1958, previous to the great changes in the last week of the month (BEHR — SCHERHAG — WARNECKE, 1960c), will be pointed out.

It is probable that the extended shape of the circumpolar vortex at 10 mb is a necessary qualification and a forerunner for the great stratospheric occurrences after 20. January. Assuming that the mean 100 mb changes are representative of the stratosphere, then Figs. 3—5 indicate that the mean temperature changes in the stratosphere from late fall to mid-winter have been conducive to the stratospheric circulation pattern in January 1958.

From the mean monthly temperature changes at 100 mb, and from the daily 10 mb maps of the IGY, it appears that stratospheric warmings are generally at sub-tropic to temperate latitudes during the fall. This warming must be due to subsidence.

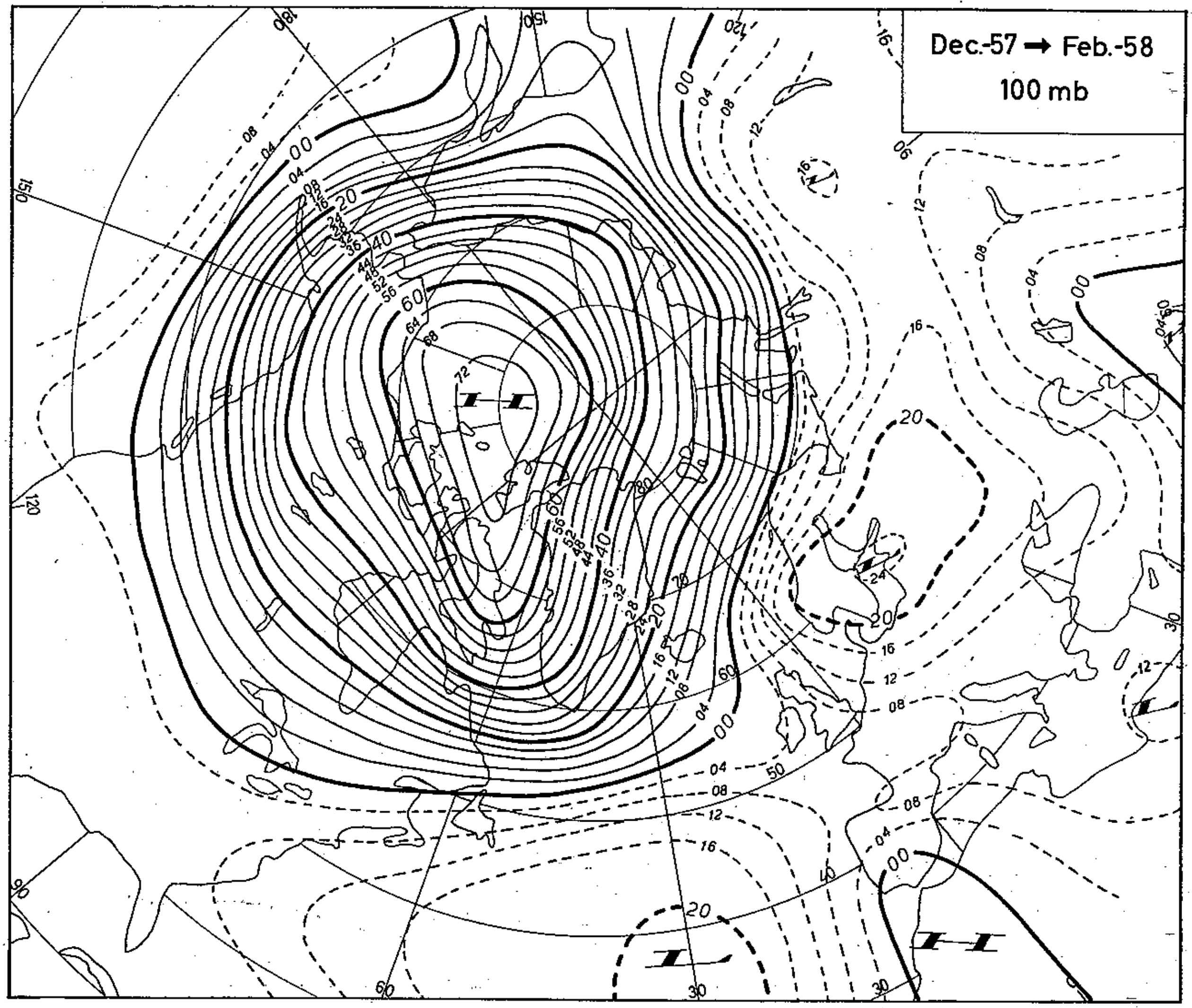


Fig. 23a. Change from December 1957 to February 1958 of the mean monthly 100 mb heights.

It is probable that the regular rapid breakdown of the stratospheric circumpolar vortex just after the spring equinox is not only due to increasing radiation, but is strongly favoured by subsidence. Wexler (1959) assumes that this is the case in the Antarctic, too.

It is reasonable to assume an annual cycle of vertical motions in the lower mesosphere and the stratosphere. Subsidence in these layers, prevailing at sub-tropical to temperate latitudes in the fall through the early winter, is mainly displaced towards polar latitudes during the late-winter, bringing the general winter subsidence in the upper mesosphere downward.

In the undisturbed atmosphere strong subsidence occurs suddenly in the lower mesosphere and the upper stratosphere after the spring equinox at the poles. The subsidence is penetrating to lower levels, also to the troposphere during the spring, and is simultaneously displacing from polar to temperate latitudes and dispersing. In the

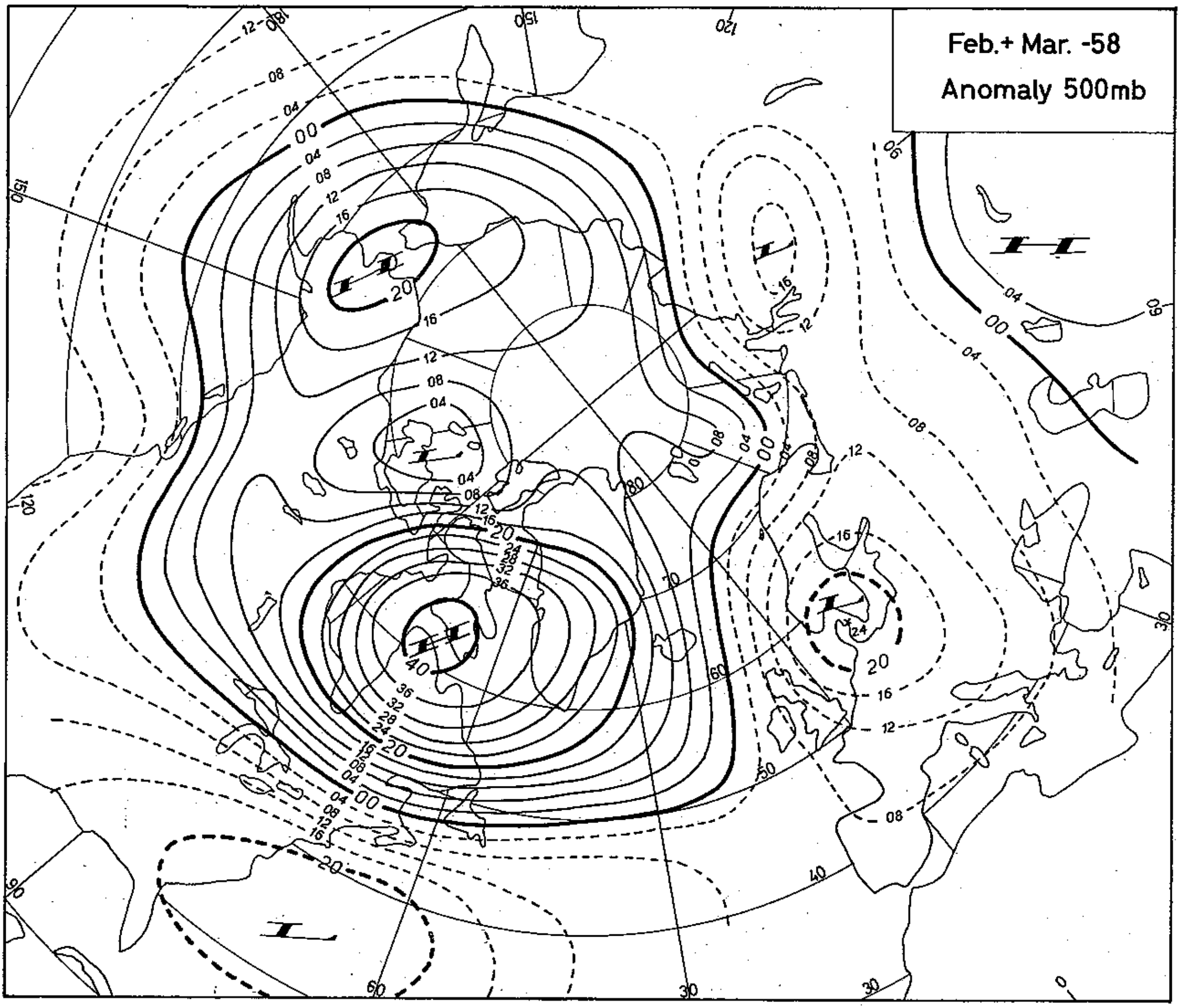


Fig. 23b. The sum February and March 1958 of the mean monthly 500 mb height anomalies.

disturbed circumpolar vortex the strong subsidence warming may occur irregularly and earlier.

Coincident with the stratospheric subsidence in the fall and early winter at subtropical to temperate latitudes, there must be an inflow to the polar regions in the lower stratosphere and the upper troposphere. This flux is branching in subsidence to the middle and lower troposphere, and there are ascending motions through the lower mesosphere in the polar areas. These ascending motions are included in the generation of the circumpolar winter vortex. As the stratospheric winter vortex in the Antarctic is stronger and more regular than in the Arctic, the flux in the lower stratosphere and the upper troposphere to the Antarctic should generally be stronger than to the Arctic.

When looking at the recorded stratospheric winter warmings during the fifties, and especially the warmings at 10 mb during the winter 1957—58, we may divide them into two categories, as pointed out by Teweles (1960). The most sudden and "ex-

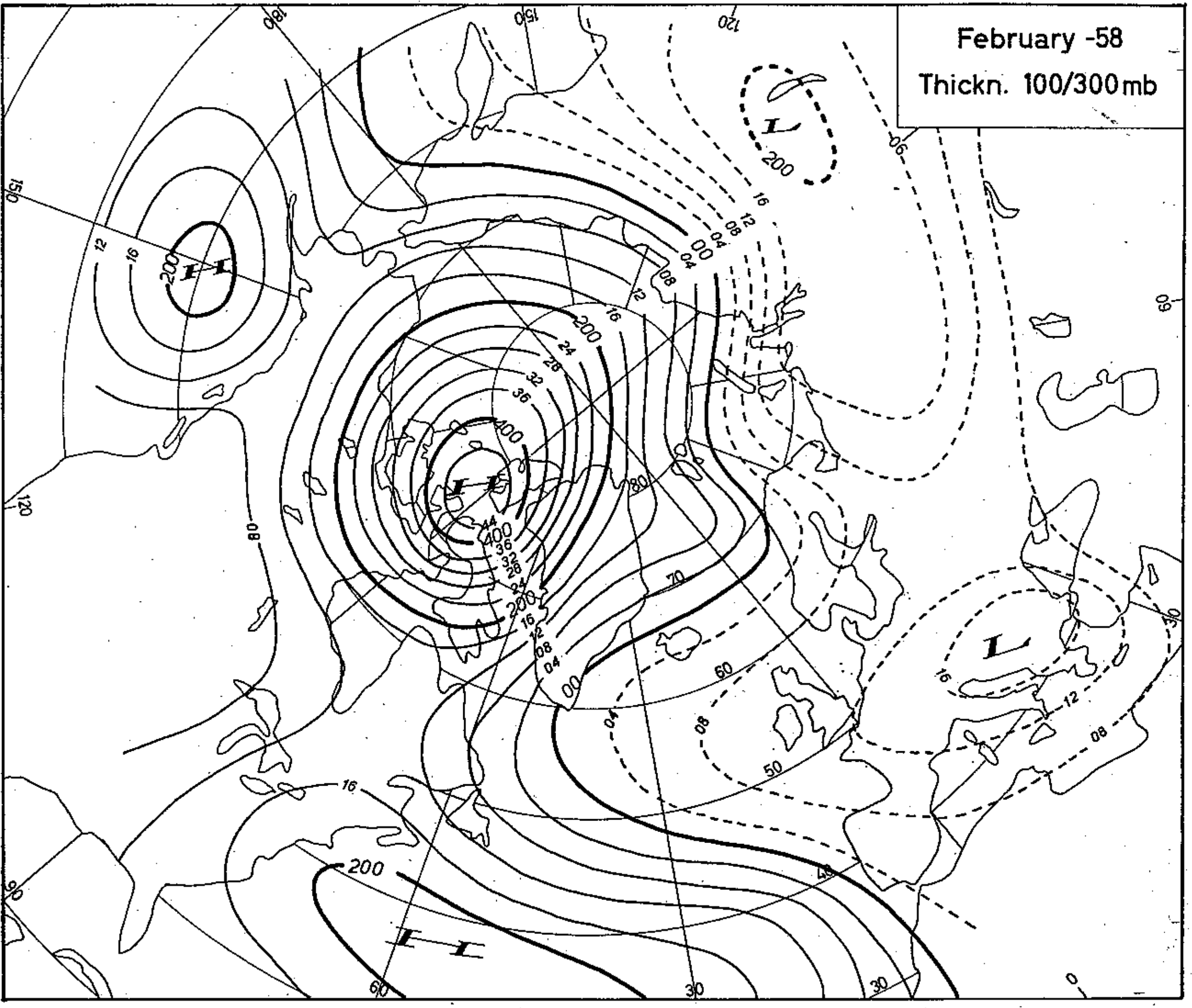


Fig. 24a. February 1958. Mean departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer (main numbers in metres, small numbers in decametres).

plosive" type has peak-character, and the preferred location for the appearance of this type seems to be the southeastern part of westward-moving troughs and lows. The peak-temperature above a station frequently occurs just after the axis of the stratospheric jet has passed the station, sweeping towards west or northwest. Subsequently, the wind speed decreases rapidly and the 10 mb height increases.

Another, as far as we know, more frequent type of warming does not have such a pronounced peak character, may have a rapid increase, but is usually of a longer duration. This type occurs in the northern part of eastward-moving ridges, later being nearly stationary.

Experiences from the monthly mean maps indicate that the strong stratospheric late-winter warmings also differ in their ability to penetrate to lower levels of the stratosphere. This difference seems to have a geographical distribution.

January—February 1958 peak warmings occurred at 10 mb over the Atlantic,

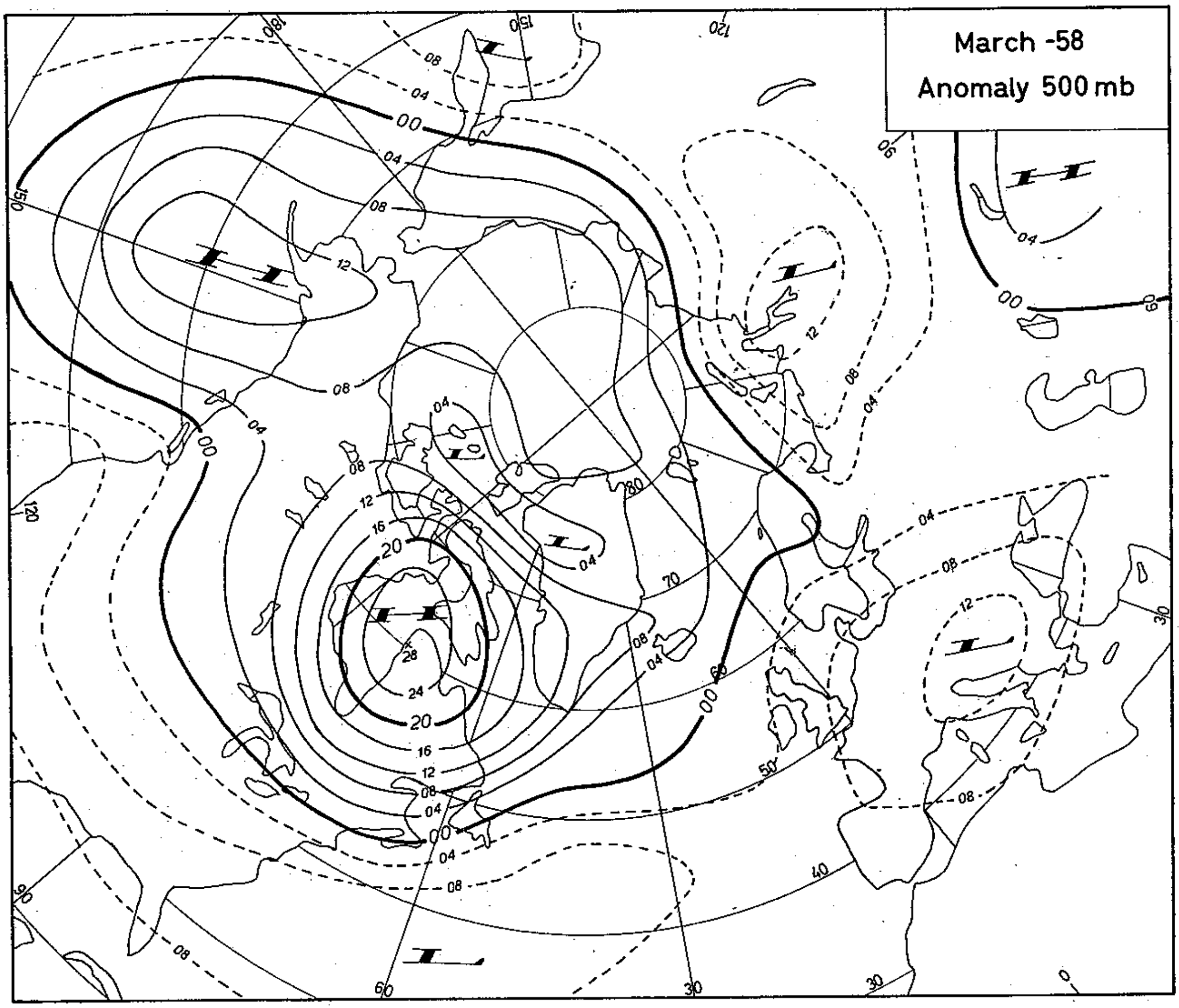


Fig. 24b. March 1958. Mean anomalies of the 500 mb heights (numbers in decametres).

Europe, Western and Southern Asia, but they did not produce lasting effects in the lower stratosphere in the same areas. On the contrary, the warmings migrated towards polar latitudes, where they extended downward. A similar sequence of events occurred in 1952 also, ("Berlin-Thule-Phenomenon").

About 20. January 1957, a high level warming (above 25 mb) of moderate type appeared over the western U.S. (Teweles, 1958). The warming moved eastward, gradually intensified and extended slowly downward. Southeast of Newfoundland the warming center curved northward, later moving towards the northwest. The effect intensified more rapidly, extended downward to lower levels, and appeared as peak warming above Keflavik and Thule in the first days of February. Teweles has also pointed out a stratospheric warming at the same time above Yakutsk in Siberia.

Above Northeastern Asia, the Aleutians, Alaska, Canada, Greenland and the adjacent Arctic, high level warmings frequently have been recorded in the winter during the fifties. Nearly all of these warmings have extended downward to the lower strato-

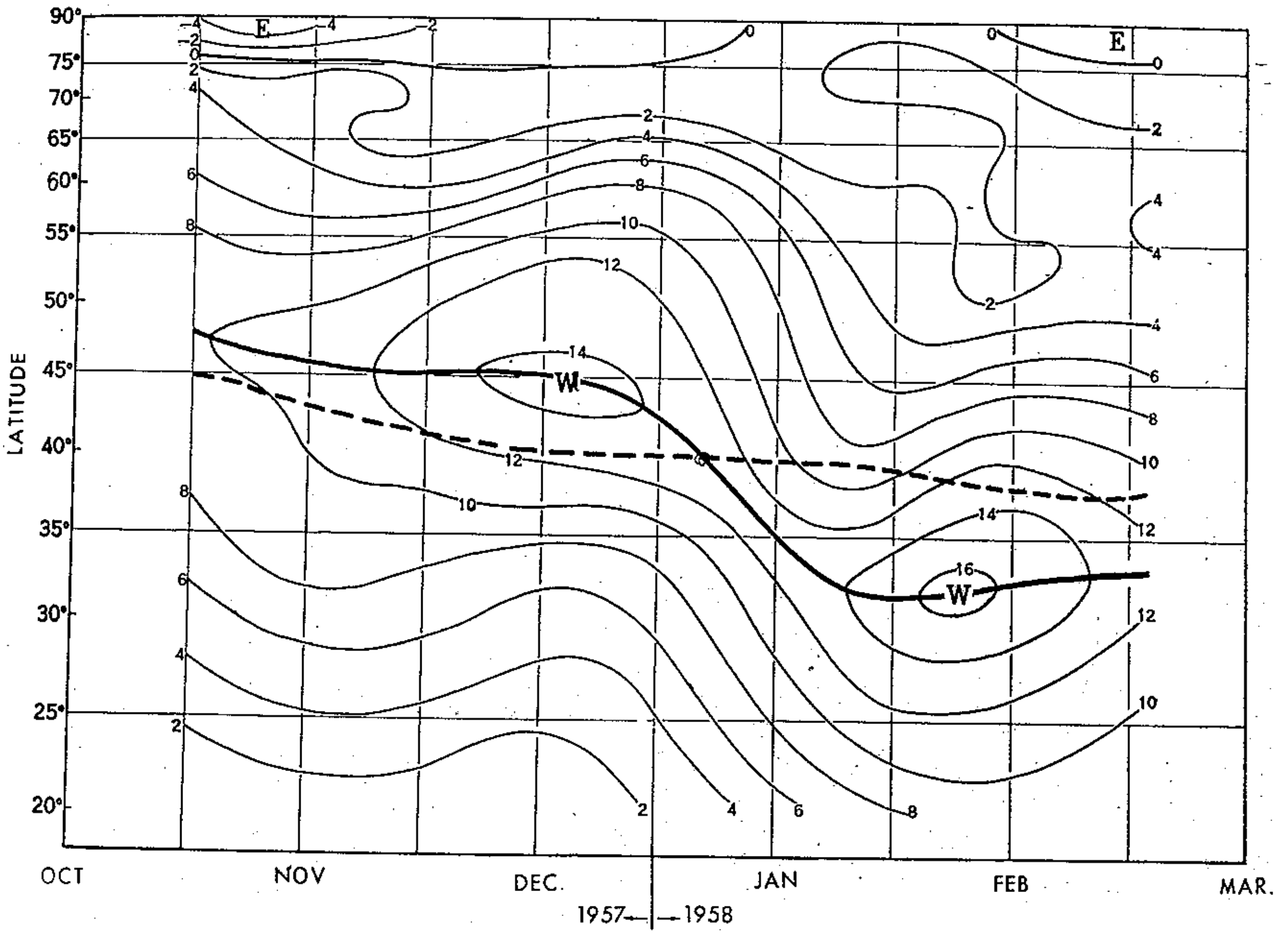


Fig. 25. Time-latitude section of 30-day mean zonal wind components averaged over the Western Hemisphere at 700 mb. Wind speeds were computed twice-monthly in 5° latitude belts from the period mid-October to mid-November 1957 until the period mid-February to mid-March 1958. Isotachs are in metres per second with easterly winds negative. Centers of maximum west wind are labeled W; easterly centers are labeled E. The axis of max. westerlies (heavy solid line) was north of its normal latitude (broken line) during the first half of the period but displaced far southward during the latter half (after W. H. KLEIN).

sphere, spreading, covering great areas and lasting for some time. Some of the warmings have had peak-character, others not.

These events in the winter stratosphere indicate that at least two different developments are contributing to the strong changes.

The semi-permanent waves generated by the topography, and possibly the waves produced by the tropospheric circulation elsewhere, too, are transmitted upward. The winter stratosphere, however, is baroclinically active and the wave-impulses are changed. We may interpret the result of these transmutations as the baroclinic stratospheric waves.

The mutual effect of the baroclinic stratospheric waves and the disturbances of a general yearly subsidence cycle in the lower mesosphere and the stratosphere, may explain some features concerning the peak warmings and the retrogressing troughs.

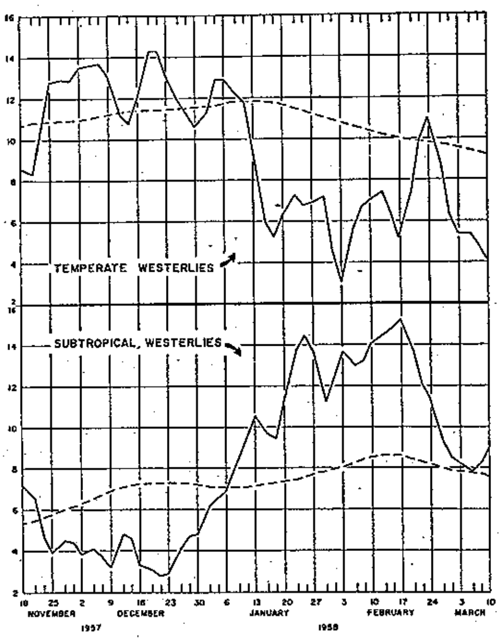


Fig. 26. Time-variation of speed of 700 mb westerlies averaged over the Western Hemisphere, for temperate zone (35°—55°N) above, and subtropical zone (20—35°N) below. Solid lines connect 5-day mean index values (plotted at middle of period and computed three times weekly), and dashed lines show variation of corresponding normal. Note the prolonged period of above normal subtropical westerlies, with record high value from February 15 to 19, 1958 (after W. H. Klein).

Experiences from the varying influences of the stratospheric warmings during the later winters indicate that the ability of the subsidence from upper levels to extend downward is highly dependent upon the circulation type in the lower levels.

Teweles (1958) has made an excellent analysis of the happenings over North America and Greenland during late January and early February 1957. The subsidence warming at high levels was primarily moving eastward, and only slowly extending downward. When the warming reached the southern part of the trough over the Eastern U.S., a pronounced wind maximum was generated.

In the decelerating jet stream, east of the wind maximum, kinetic energy was converted into potential energy through vertical motions associated with horizontal divergence to the left of the stream, and convergence to the right. The conversion of energy can be accomplished in either of two ways: by the elevation of potentially colder air below, or the depression of potentially warmer air above (Riehl and Teweles, 1953).

The temperature pattern of the stratospheric waves frequently shows rising motions from ridge to trough, and sinking motions from trough to ridge. In these cases the subsidence alternative in the decelerating jet stream is favoured, and strongly increasing subsidence and warming are then generated on the southeastern side of the trough. Combined with the subsidence warming a new strong wind maximum is produced. In connection with the subsidence and the increasing heights of the isobaric levels, both the trough and the areas of warming and windmaximum are now retrogressing.

In January 1957, the change from progressing to retrogressing movement of the area of maximum warming occurred southeast of Newfoundland.

Other subsidence-effects may run together with the occurrences described here, particularly the warming impulses from the ionosphere-mesosphere, and by strong oscillations of the stratosphere — lower mesosphere, the events may be repeated too, as happened in late January and early February 1958.

Subsidence and convergence may increase the ozone content, and the heat absorp-

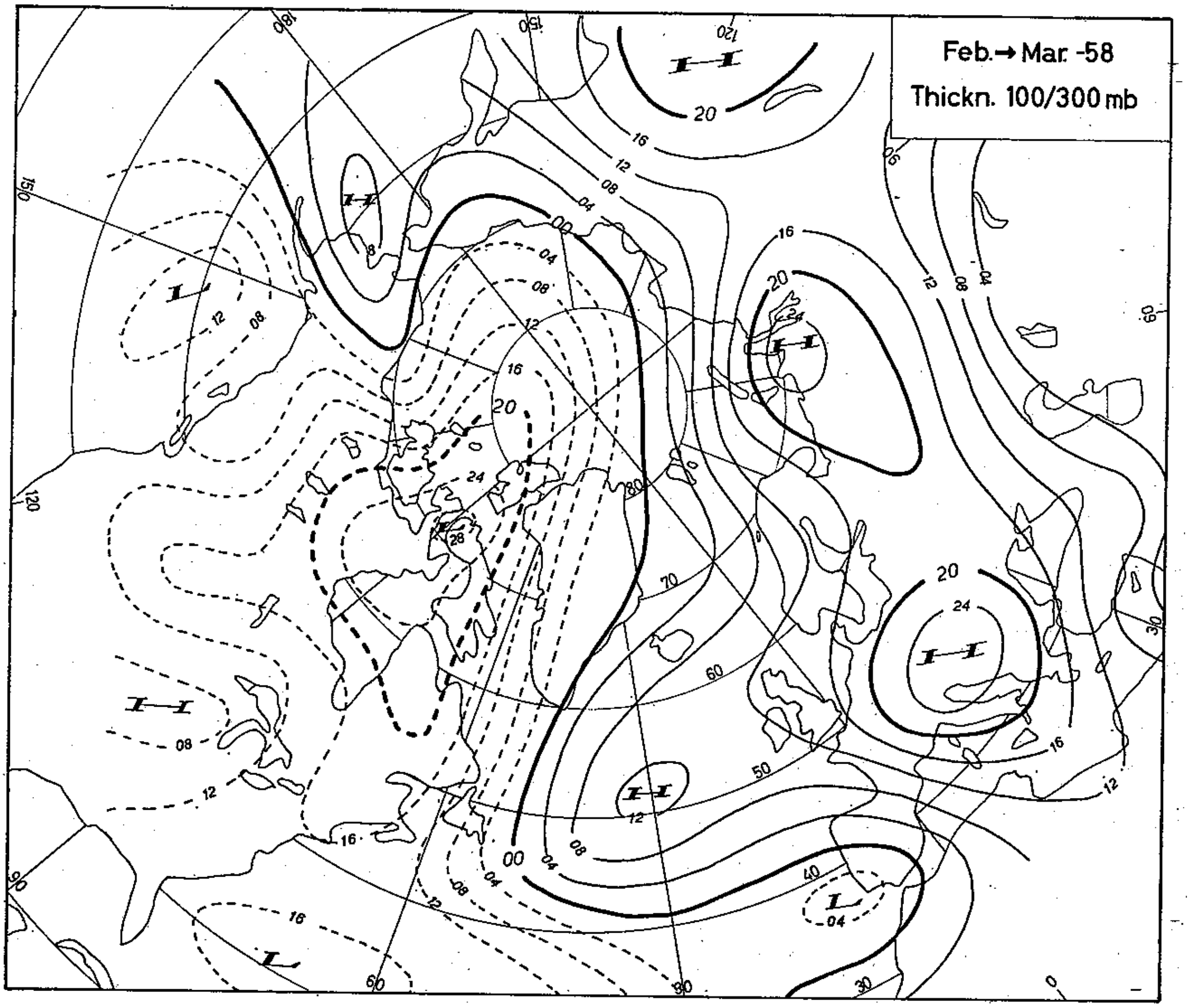


Fig. 27a. Change from February to March 1958 of the mean monthly departures from the 5-year (1949 —53) of the 100/300 mb thickness layer.

tion of the ozone will then additionally intensify the temperature rise. Scherhag (Behr and coll., 1960a) has reported that during the strong stratospheric warmings over the Atlantic and Europe after 20. January 1958, unusually high ozone concentration was observed over Arosa, especially in the layer between the 16 mb and 8 mb levels.

From the rocket-grenade measurements of temperature in the mesosphere over White Sands and Fort Churchill (Stroud and coll., 1960, Fig. 13), we may conclude that on the American side of the Northern Hemisphere, the circumpolar winter vortex is generally intensifying up to 50—70 km, varying with latitude, and hence weakening with height. If the meridional temperature gradient in the upper mesosphere over Siberia is of the same sign as over North America, the circumpolar winter vortex is generally weakening with height above 50—70 km in the mesosphere, otherwise the axis of the vortex will be tilting towards Siberia with increasing height.

Very strong disturbances, however, may occur in the upper mesosphere. On 11.

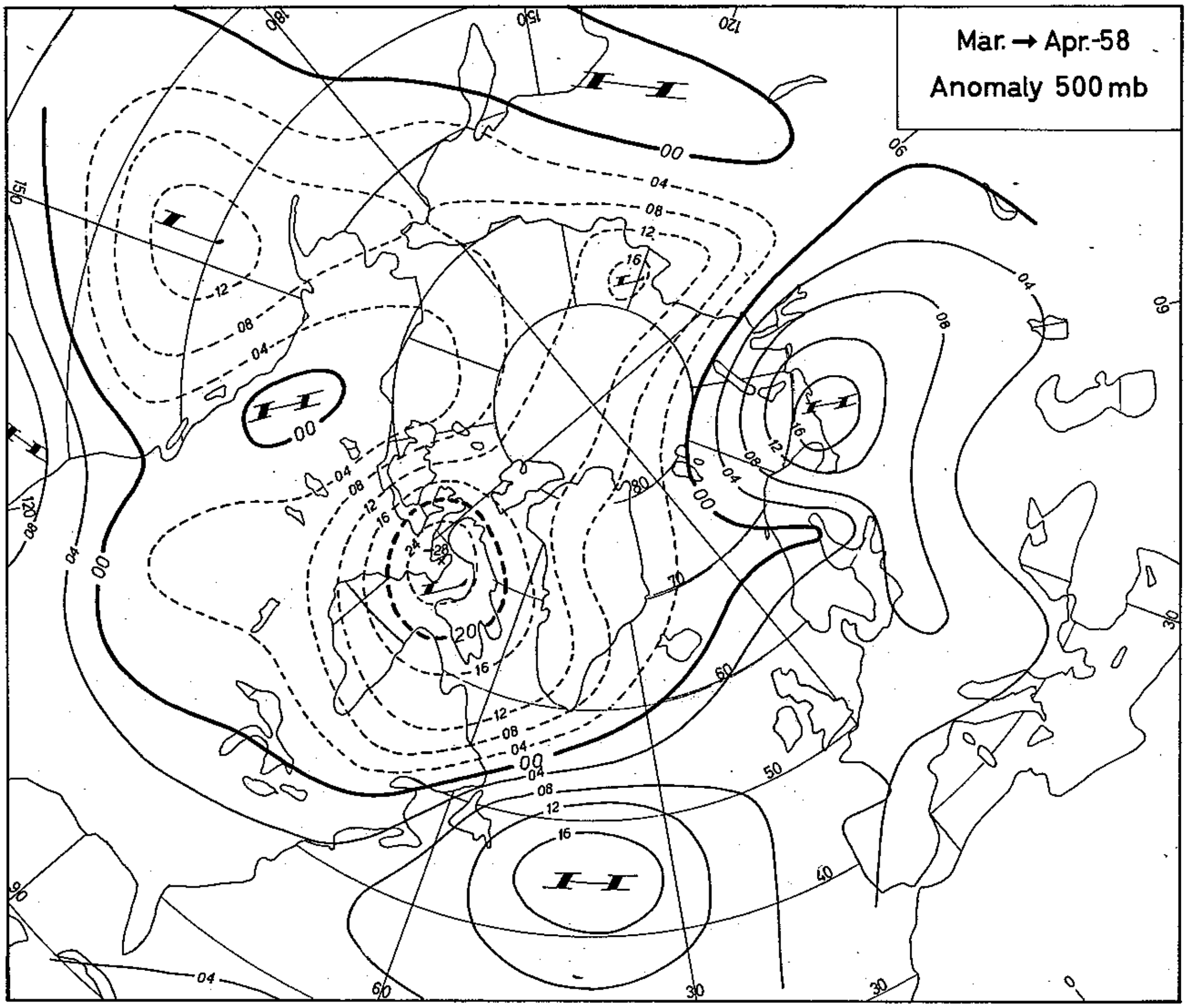


Fig. 27b. Change from March to April 1958 of the mean monthly anomalies of the 500 mb heights.

December 1957, an irregular mesopeak of 290 K. degrees at 72 km was measured over Churchill, with a wind maximum of 360 knots just above the warm peak. On 14. December, the mesopeak was lowered to 66 km and 249 K. degrees.

To the knowledge of the writer, no temperature measurements from the layer near 50 km height within the polar night are available, but the warm ozone layer probably cannot exist there. Near the boundary of the polar night in the layer at 50 km height, large temperature gradients exist and extremely strong winds are generated. On 12. November 1956, 330 knots of wind from the west was measured over Churchill at 58 km.

As the temperature gradients between White Sands and Churchill are very small in the layer near 50 km, the wind forces must be concentrated near the limit of the polar night in this layer. In this way the jet stream in the circumpolar winter vortex probably is intensifying and contracting with height up to 50 km, and expanding higher up.

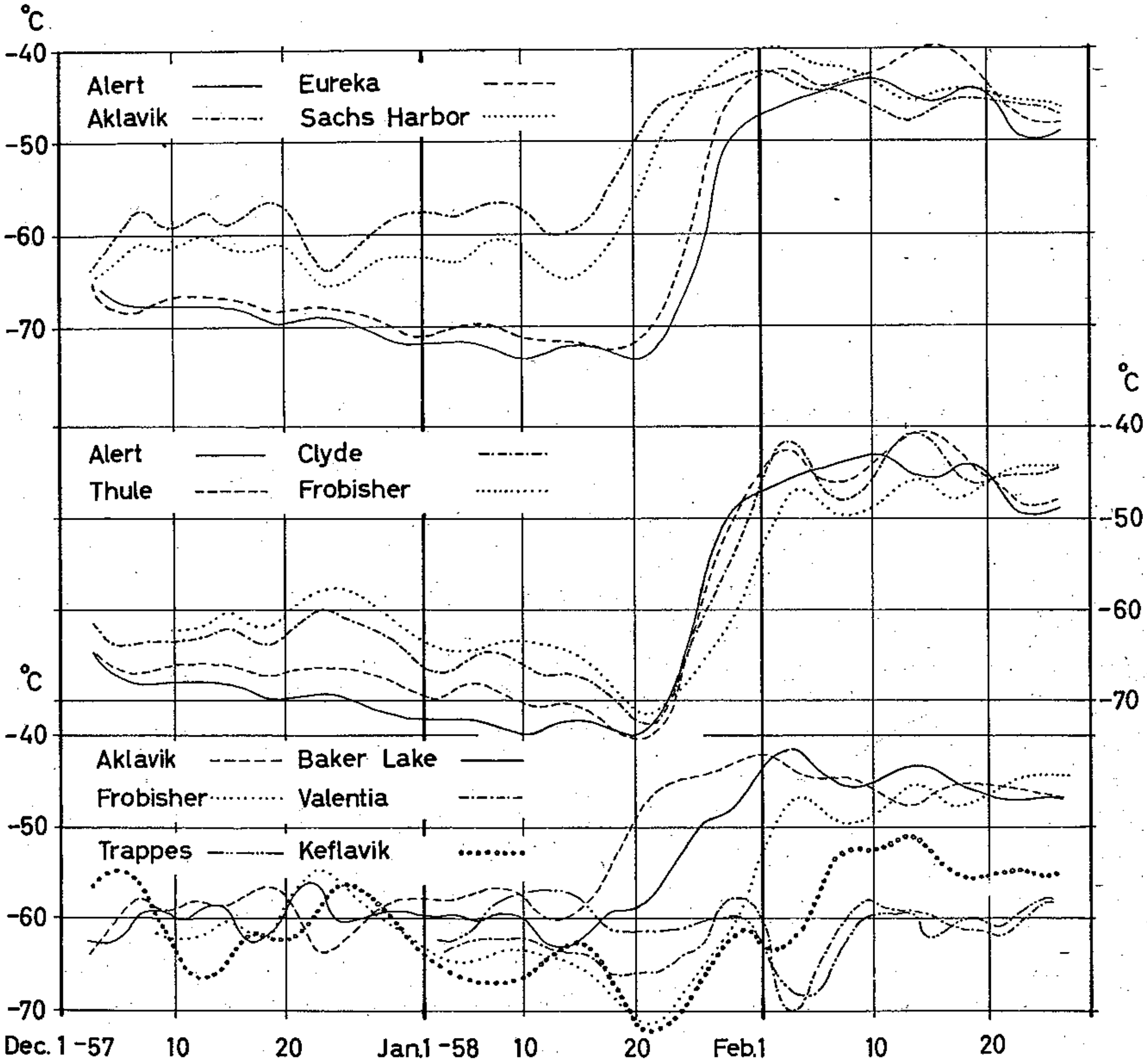
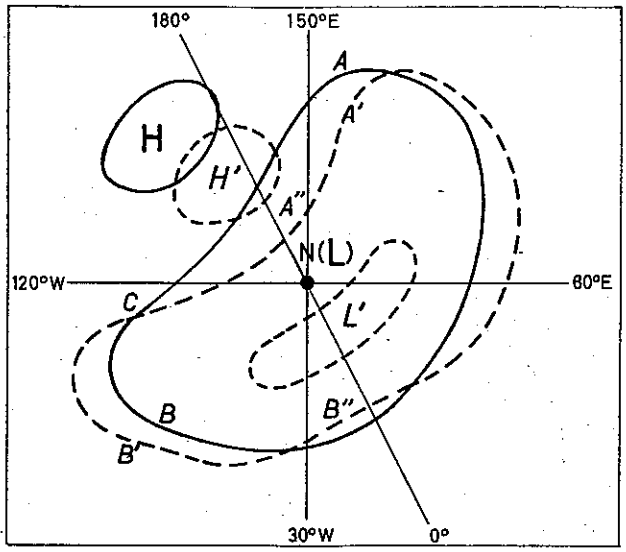


Fig. 28. 5-day running means of 100 mb temperatures. Above: Aklavik — Sachs Harbour — Eureka — Alert. In the middle: Alert — Thule — Clyde — Frobisher. Below: Aklavik — Baker Lake — Frobisher — Keflavik — Valentia — Trappes.

Several years of data with radiosonde-measurements from the Antarctic are now under consideration, but we have no records of any breakdown of the circumpolar winter vortex prior to the spring equinox. This implies that the vortex is stable, also in the layer near 50 km, and is not brought out of balance neither by the tropospheric-stratospheric perturbations, nor by ionospheric-upper mesospheric disturbances.

The writer is estimating the polar jet in the layer at 50 km as a kind of stabilizer for the whole circumpolar vortex in the winter, not due to the kinetic energy, but to the radiational stability of the mesopeak and the regulation of the vertical motions through the vortex in the lower mesosphere.

Fig. 29. Sketch illustrating the cumulative subsidence effect just prior to the breakdown of the circumpolar vortex. *ACB* is a contour with ridge-high *H* at about 10 mb, generated by the influence of the Himalaya Mountains. The vortex center is near the Pole. The general subsidence from the lower mesospehre is favoured from *A* intensifying and running up to *A*'' moving the vortex center to *L*', and influencing the vortex near the mesopeak at *D* (fig. 30). Due to the displacement of the vortex center a meridional-vertical oscillation is generated. The subsidence is increased at *F* (fig. 30), and may influence both at *B*' and *B*''. The subsidence effect at *B*'' is cumulatively intensified due to the generated strong wind field at *B*' (see the text).



The final breakdown of the circumpolar winter vortex on the Northern Hemisphere prior to the spring equinox is judged to start when the stratospheric disturbances are reaching up towards the mesopeak-level near the polar-night boundary.

Fig. 29 outlines contours at 10 mb. The contour line *AB* with the high *H* is meant as the pattern produced by the effect of the Himalaya Mountains on the jet. The vortex center (*L*) is near the North Pole. In the areas *A* and *B* the subsidence from above is taking effect, but is more frequently favoured at *A* (owing to the mountain-effect) than at *B* due to sweeping cold troughs from the downstream at *C*).

The stage *A*'*A*'' with the high *H*' and the low center *L*', is reached after the effects of the Himalaya Mountains and the subsidence from above are combined, which makes the trough retrograde and also, later, gradually weakens the jet westward over Asia. At 10 mb, the wind is now blowing across the polar area and the subsidence reaches the mesopeak level at *D* (Fig. 30).

From this stage, the events may occur rapidly. The subsidence at *D* induces rising motions at *E*, inclining the orbits in the polar vortex, marked *IK*. A meridional disturbance in the mesopeak level and

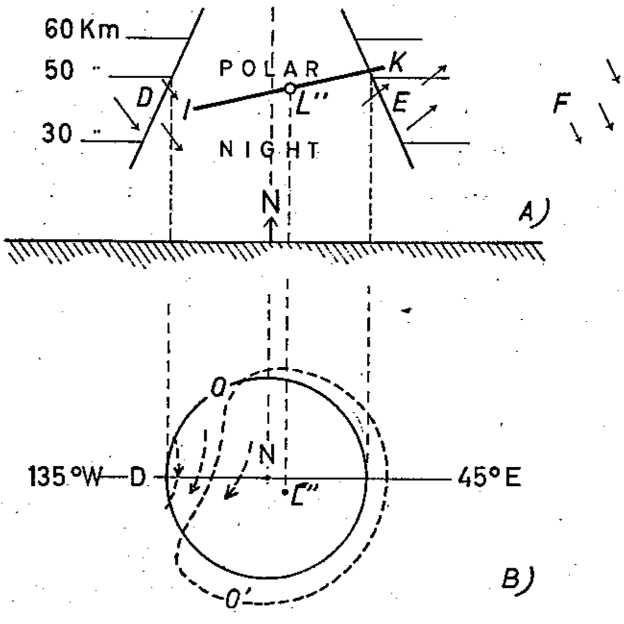


Fig. 30. Sketch A) shows the inclined orbit and the meridional-vertical oscillation generated in the lower mesosphere when the vortex is disturbed by subsidence at *D*. B) presents the change of the contours near the mesopeak level. In the area *DO* the warming will occur both due to subsidence and advection, and the polar-night cone will rapidly be filled by warmer air.

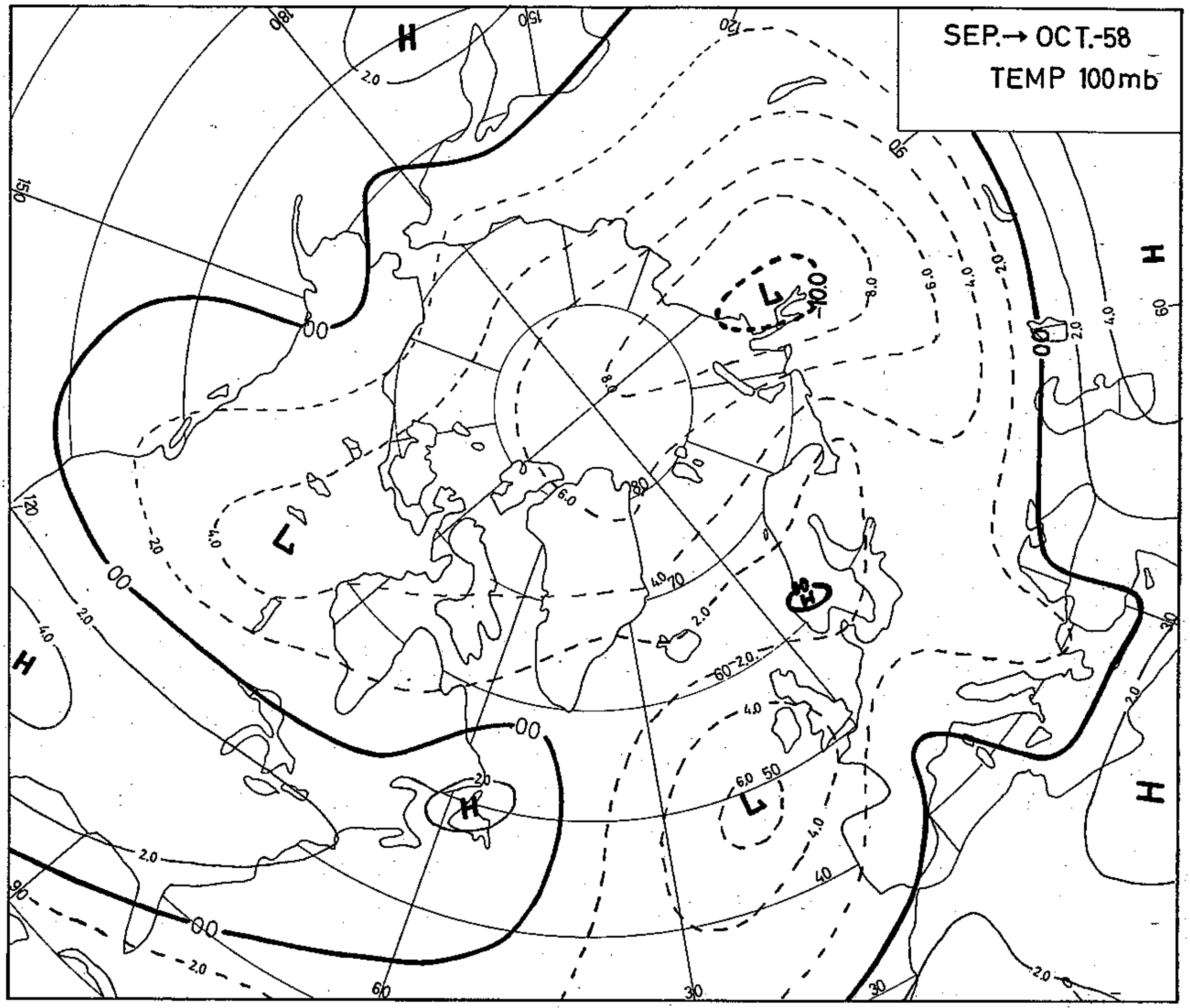


Fig. 31. Change from September to October 1958 of the monthly mean temperatures at 100 mb. Dashed lines negative values. C°.

the lower mesosphere is created, resulting in increasing subsidence at *F*. This subsidence may be pendulating between temperate and subtropical latitudes owing to the polar disturbances. When extending downward, the subsidence influences the baroclinic stratospheric wave at *B*' in the 10 mb surface (Fig. 29). A cumulative effect is now produced at *B*", due to the pre-existing temperature field in the contour pattern. The intensified anticyclonic development at *B*" and *A*" meet across the low (*L*'), splitting it into two parts.

In the mean time, the vortex at the mesopeak-level may nearly have vanished. When the subsidence reaches the polar vortex at *D* (Fig. 30), the orbits are made concave at *D*, and are gradually changed to the pattern *OO*' with the inclined position *IK*, and the center at *L*". At *O*, the warming now occurs due to both subsidence and advection. The vortex is rapidly warmed up from the *OD* area, and after a short time warm mesopeak air is blowing across the polar area. The vortex center at the mesopeak

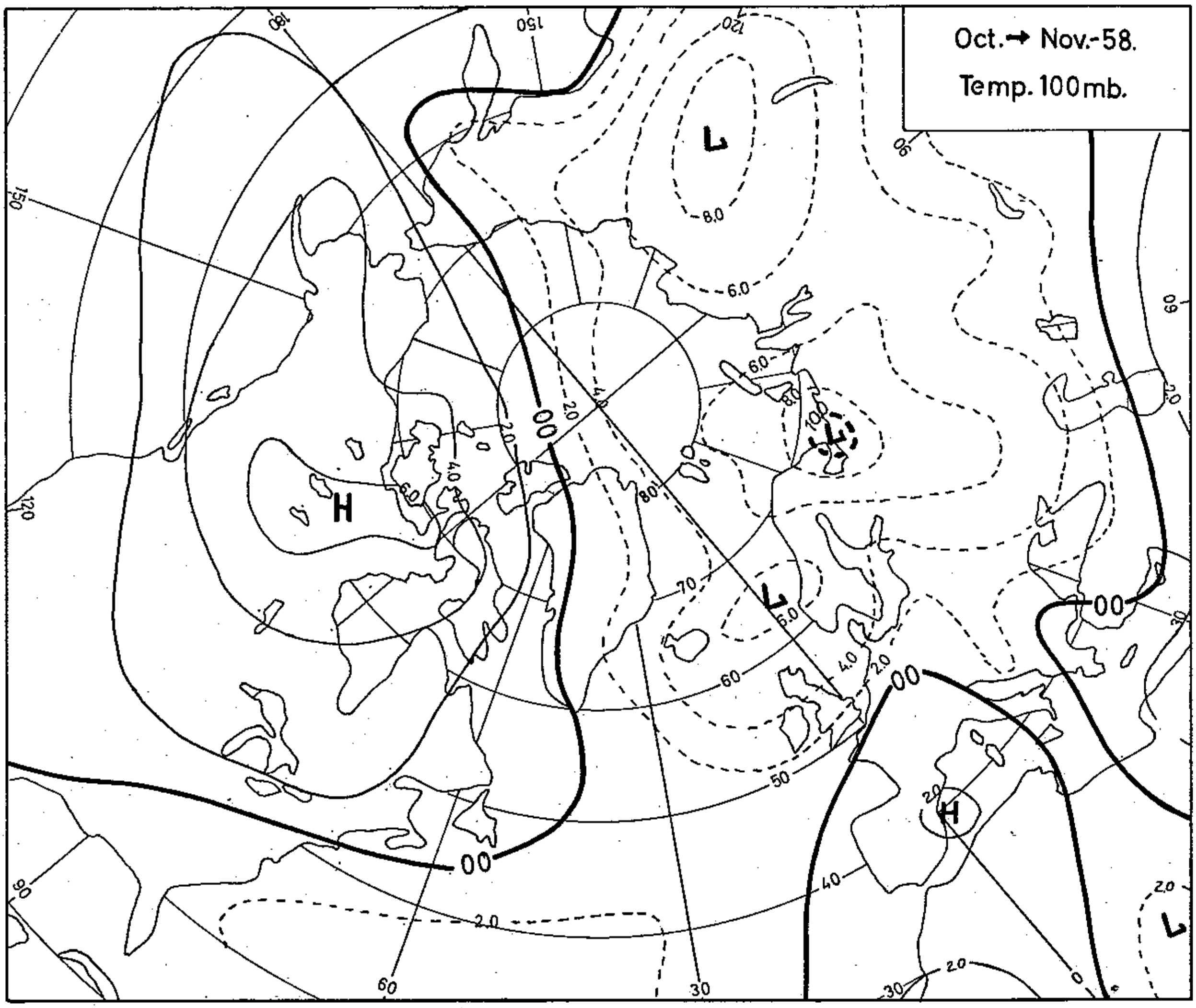


Fig. 32. Change from October to November 1958 of the monthly mean temperatures at 100 mb.

level probably has no qualification to exist out of the polar area, and is rapidly filling. This development is facilitated by the general winter subsidence in the upper mesosphere at high latitudes.

After the ridge formation over the Arctic Basin on 25. January 1958, a systematic westward movement in the stratosphere was observed, not only of troughs, but also of low and high circulation centers (BEHR and coll., 1960a). It is reasonable that the recorded west drift was due to the fact that the subsidence high in the Arctic was stronger and more wide-ranging in levels higher up than 10 mb, the upper level high conducting the slow rotation of the stratospheric systems. But the effect in the retrogressing troughs may also have been contributing.

In connection with the breakdown of the polar vortex at the mesopeak level, marked changes in the regime of vertical motions probably occur in the whole mesosphere at polar latitudes, but the subsidence is assumed to be strongest near the mesopeak level. At Fort Churchill, rocket-grenade and falling sphere measurements were made

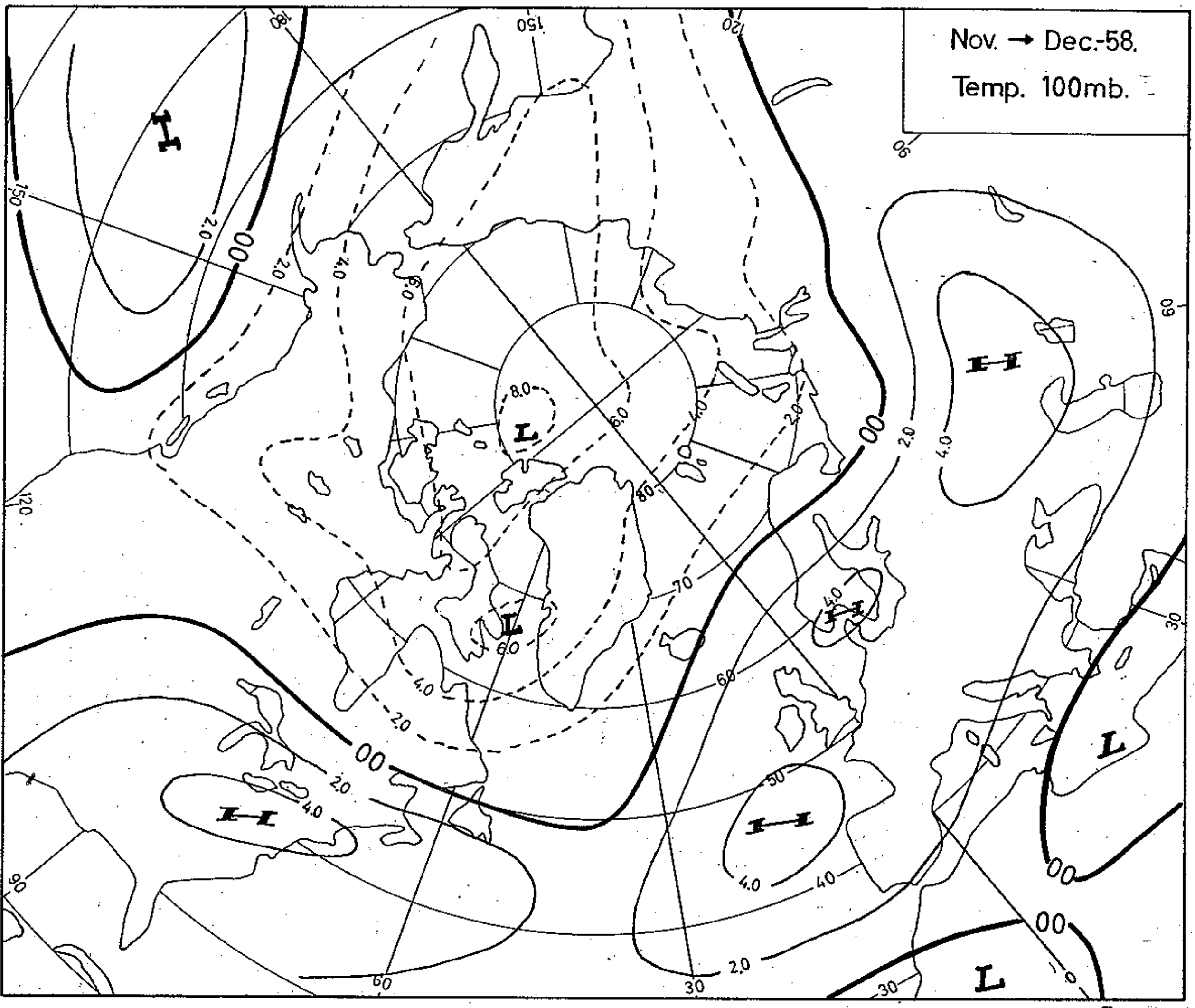


Fig. 33. Change from November to December 1958 of the monthly mean temperatures at 100 mb.

at both midnight and mid-day 27. January 1958, and at mid-day 29. January (Stroud and coll., 1960). The writer is judging both the temperature and the wind changes as a consequence of the westward movement of a sizable warm ridge in the mesosphere with the front between 50 and 60 km, and the maximum temperature effect between 40 and 50 km. This warm ridge was probably created over the North Atlantic-European Arctic in connection with the breakdown of the vortex. The fact that the maximum wind speed appeared above the level of maximum temperature is also in accordance with the movement of such a ridge. The northeast wind up to 47 km just after midnight which lowered to 43 km at mid-day 27. January, belonged to the westward moving part-vortex from the previously divided polar vortex.

In the view given above, two different developments are pointed out which can intensify sufficiently to break down the circumpolar winter vortex. The development *A* —*A*' — *A*'' (Fig. 29), starting as a ridge and supplied by the effect in a retrogressing trough, we may call the *A*-type. The occurrences in the *B* — *B*' — *B*'' area which

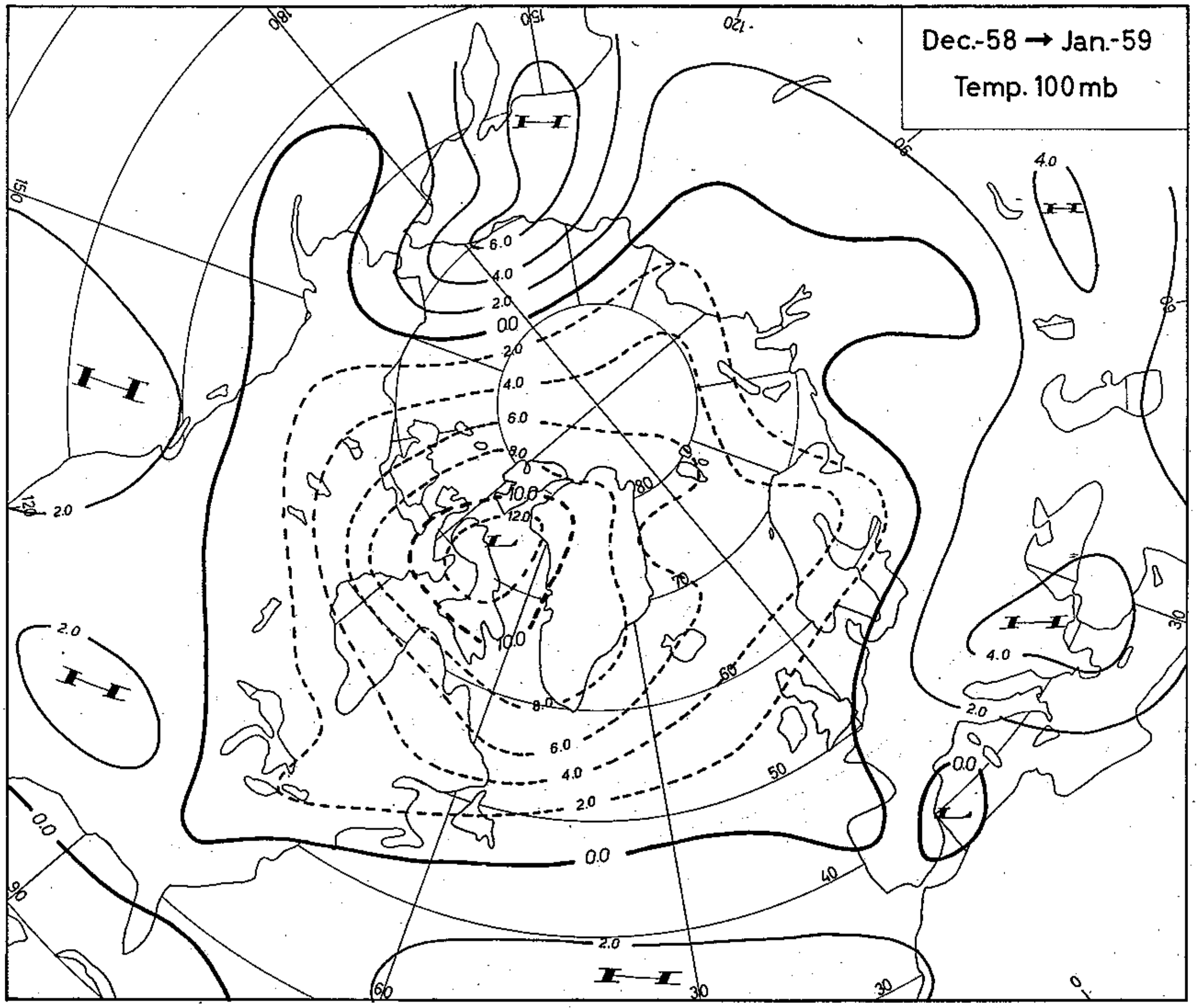


Fig. 34. Change from December 1958 to January 1959 of the monthly mean temperatures at 100 mb.

start as rapidly passing effects from trough and ridge developments of moderate intensity, but later, influenced by subsidence from above, generate increasing subsidence combined with the retrogressing trough, and ending with strong ridge development, we may call the *B*-type.

Above, it is supposed that the two types of developments co-operate in the breakdown of the circumpolar vortex, but this combined effect is probably not quite necessary. In 1958, 1957 and 1952, the breakdown of the circumpolar vortex started in last half of January. All three years had peak warmings of the *B*-type at temperate latitudes, and probably, like in 1958, effects from the *A*-type in the Arctic, too.

In 1959 and 1956, the close of the stratospheric circumpolar vortex occurred in the beginning and at the end of March, respectively, and was produced by the *A*-type only. In 1955, the stratospheric warming in the Arctic occurred in the beginning of January. The warming type is doubtful, but was probably of the *A*-type.

It is yet doubtful how much the ionospheric-upper mesospheric disturbances may

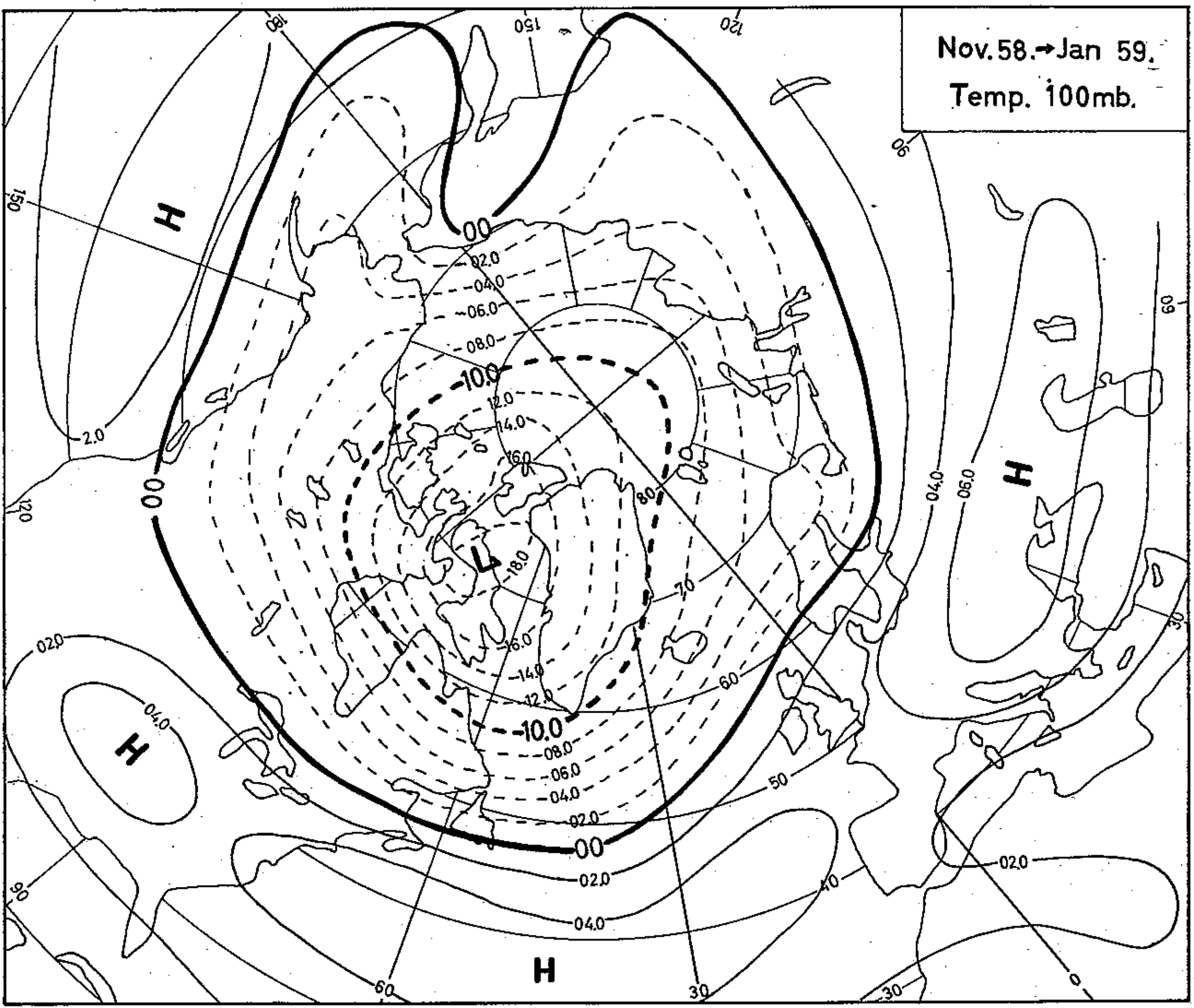


Fig. 35. Change from November 1958 to January 1959 of the monthly mean temperatures at 100 mb.

be engaged in the breakdown of the circumpolar winter vortex. LETHBRIDGE, PANOFSKY and NEUBERGER (1958) have found a statistical relation between the date of the beginning of the breakdown in the Arctic and magnetic activity. PALMER (1959) points out the correspondence in time between a change in drag on the satellite 1957 $\beta 1$ (Sputnik II) and the great stratospheric changes in last part of January 1958. In the opinion of the writer, however, the high level impulses are not sufficient to break down the circumpolar winter vortex, perhaps not necessary. When comparing with the Antarctic, it is reasonable to assume that the baroclinic stratospheric waves are essential for transmitting subsidence effects downward during wintertime.

At Fort Churchill (STROUD and coll., 1960), strong disturbances occurred in the upper mesosphere from 11. to 14. December 1957, but no changes followed in the stratosphere. It is reasonable to assume that ionospheric-upper mesospheric disturbances are more frequent than the observed peak warmings during wintertime in the stratosphere.

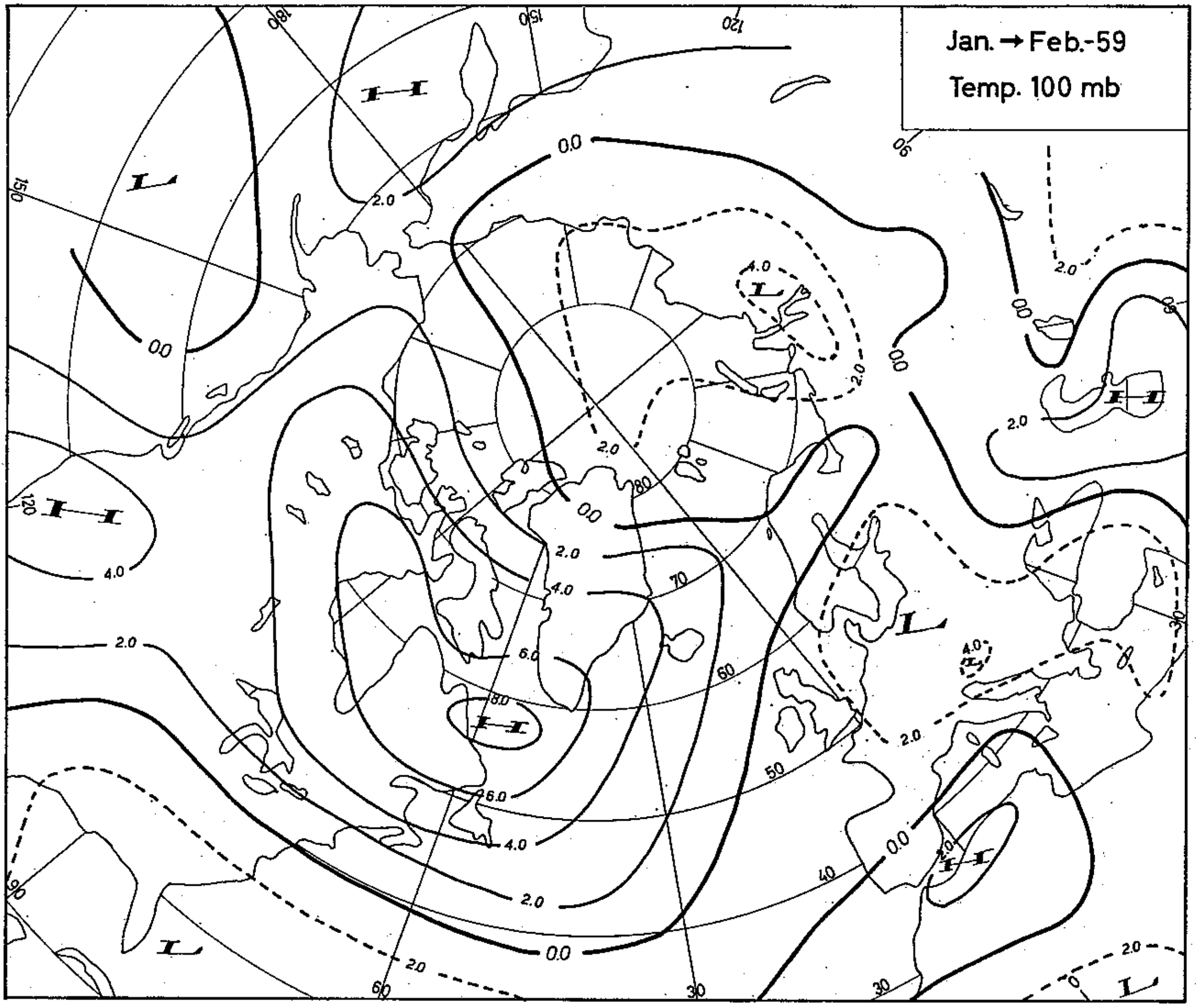


Fig. 36. Change from January to February 1959 of the monthly mean temperatures at 100 mb.

Nevertheless, the impulses from higher levels may occasionally affect and intensify the occurrences in the stratosphere when density changes and warmings from above coincide with strong disturbances in the stratosphere and the lower mesosphere.

e. The annual variation of the ozone. The yearly cycle of vertical motions and fluxes in the stratosphere and the lower mesosphere mentioned above (p. 34) may explain the annual variation of the total ozone content and the extremes.

The ozone minimum in the fall, and the increasing ozone content in the atmosphere during the later months of the year at temperate to sub-tropical latitudes, is probably caused by increasing subsidence in the lower mesosphere and the stratosphere during the fall and early winter. The ozone minimum at polar latitudes occurs a little later in the year. This may be explained by the fact that the ozone content in higher latitudes in this part of the year is mainly controlled by advection from temperate latitudes.

The northward delayed date of the ozone minimum is based on records from Arosa and Tromsø. The writer is otherwise of the opinion that time-sections, showing the

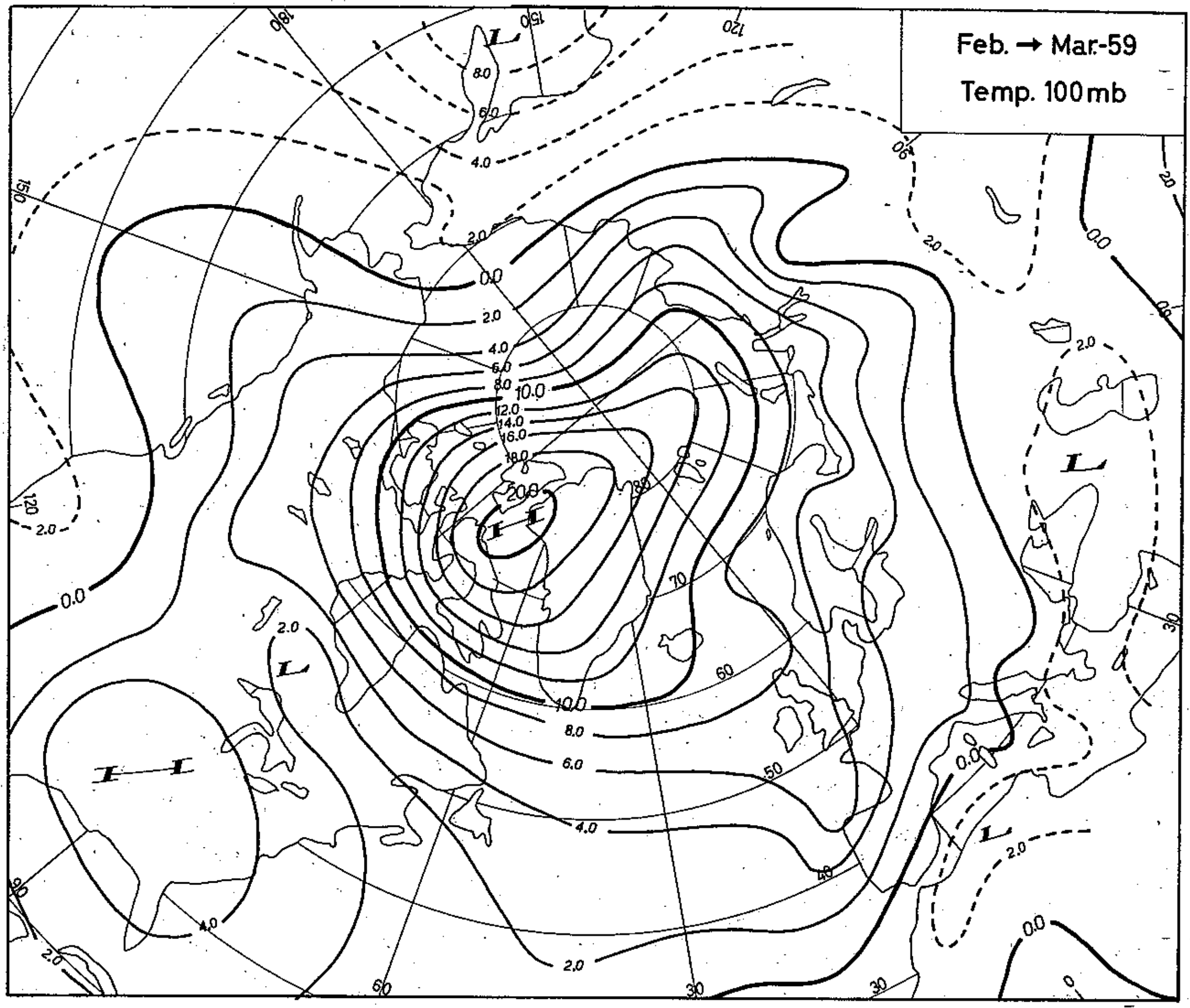


Fig. 37. Change from February to March 1959 of the monthly mean temperatures at 100 mb.

variation of the total ozone content with season and latitude, ought to be differentiated as to the longitude. This idea is based on the observation that the regime of vertical motions in the northern-hemispheric stratosphere strongly varies with the longitude.

Prior to, or combined with, displacements of the circumpolar winter vortex, stratospheric subsidence ridges are advecting ozone from temperate latitudes across parts of the Arctic in preference to the American and European Arctic.

As pointed out by Godson (1960), the pronounced ozone maximum in the Arctic in the early spring is caused by sudden and strong subsidence, which closes the stratospheric winter circulation. Due to displacing southward of the subsidence, the ozone maximum is delayed southward. In connection with the simultaneous dispersion of the subsidence, the ozone maximum is declining.

The fallout of long-lived radioactive debris from nuclear test explosions shows an annual variation similar to that of the ozone, with minimum in October and maximum in the spring (Bleichrodt, Blok and Dekker, 1961). Both features have probably the same cause, too.

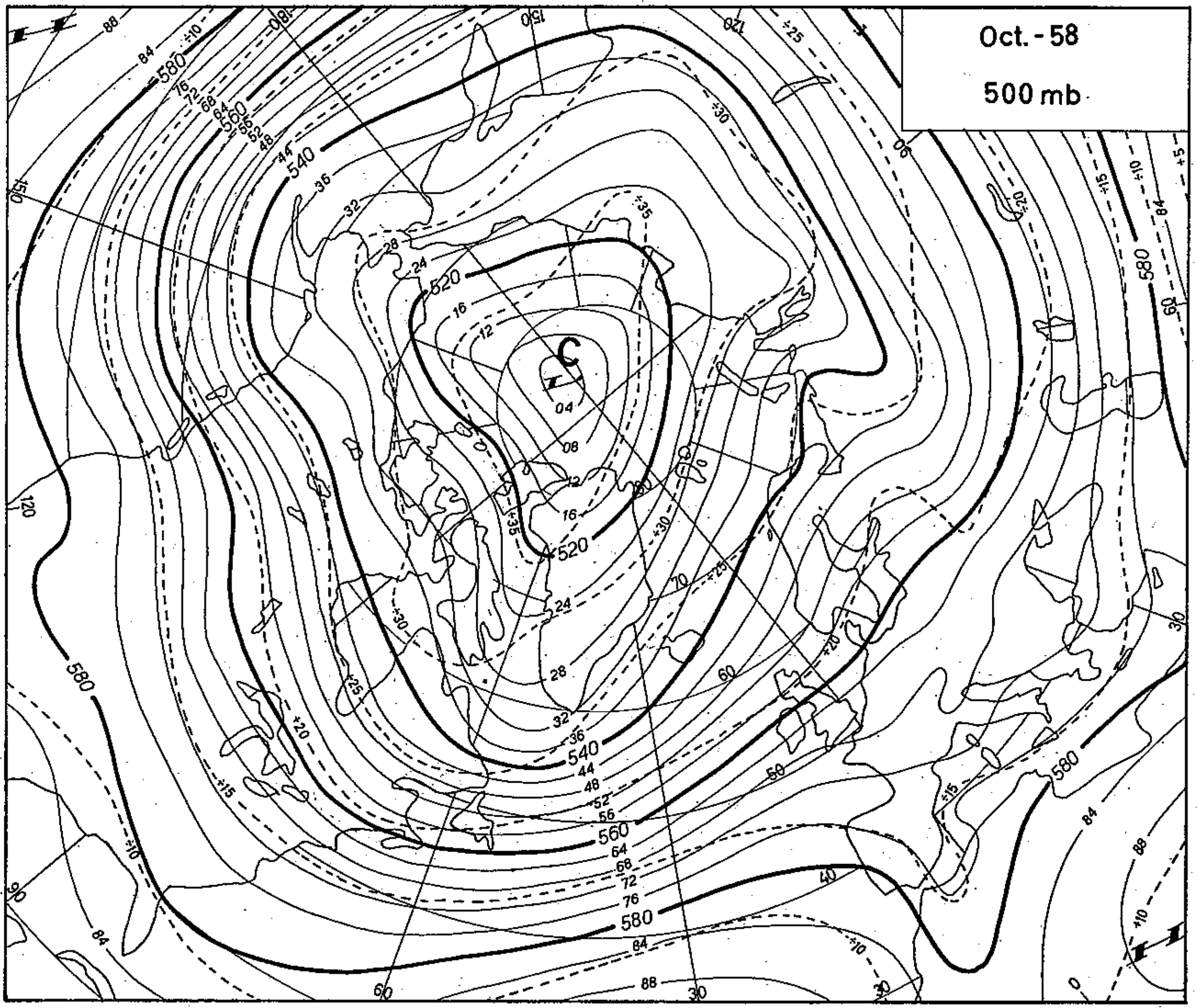


Fig. 38. October 1958. Mean contours (decameters) and isotherms (C°) of the 500 mb surface.

The increasing fallout from October may be due to the subsidence from the lower mesosphere and the upper stratosphere to the lower stratosphere at sub-tropical to temperate latitudes, bringing the radioactive debris down in the troposphere. Coincident with the increasing content of the debris in the troposphere during the fall and winter, a northward flux takes place in the lower stratosphere, partly accumulating radioactive particles in the stratosphere and mesosphere at high latitudes. The spring maximum of the fallout is then reached after the strong stratospheric subsidence maximum has occurred at high latitudes.

f. Frequency of the great winter warmings. In connection with the strong stratospheric warmings, we may ask: Is the observed frequency in the Northern Hemisphere during the fifties a normal feature? Any direct answer to this question is not available, but we may get an indirect answer through the stratospheric warming effects on the tropospheric circulation in North Europe.

Owing to the geographical position, Norway has colder winters than normal when

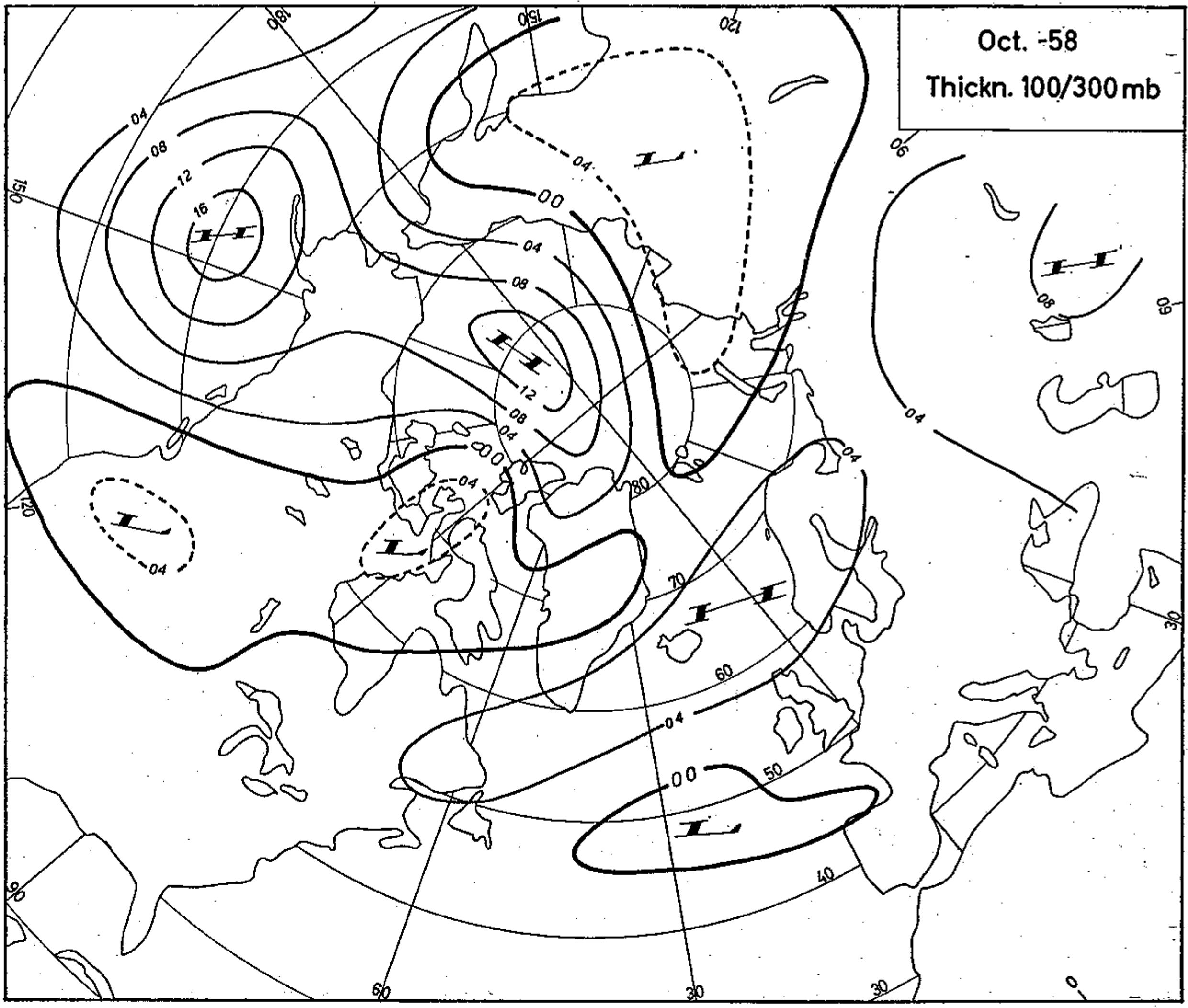


Fig. 39a. October 1958. Mean departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer.

the zonal index in the Northeastern Atlantic is sub-normal. The high-stratospheric winter warmings in the fifties have occurred in different ways, but the final result in the lower stratosphere, influencing the troposphere, has not been very different. In broad features, the late-winter warmings in the lower stratosphere have prevailed in North America (Canada), Greenland and the Arctic, that is, there has been a tendency towards low zonal index in the North Atlantic and cold late-winters in Norway, except the east coast of Finnmark.

The effect may be illustrated by the Oslo temperatures, which in this respect are representative of most of Norway. The table below presents the annual variation of the mean surface temperature anomalies at Blindern (Oslo) during the period (1945—59) as departure from the 30-year normal (1901—30).

Blindern (*Oslo*). Mean surface temperature anomalies (1945—59). Departures from the normal (1901—30), °C.

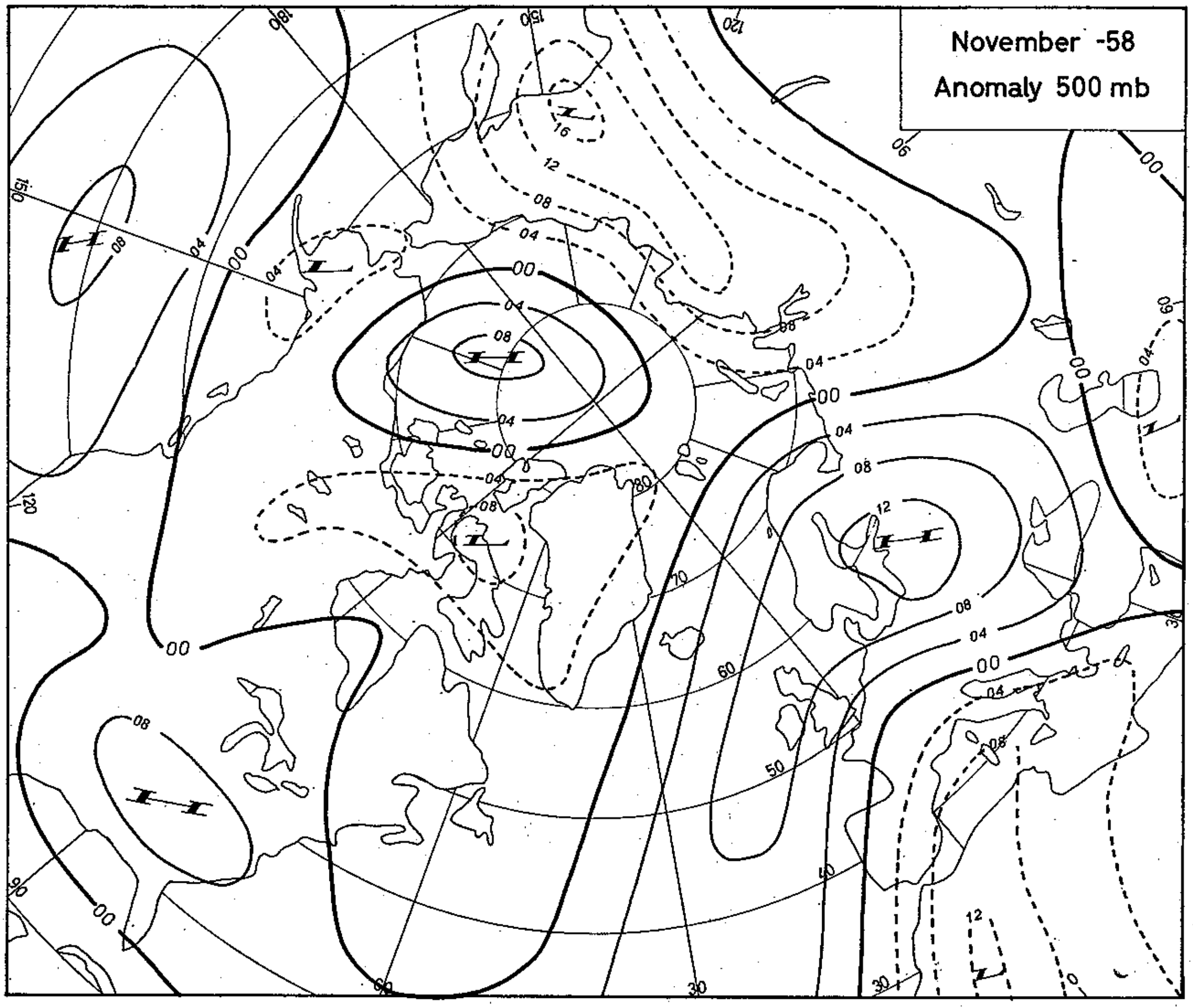


Fig. 39b. November 1958. Mean anomalies of the 500 mb heights.

Jan.	Febr.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
–0.6	–1.2	–0.4	0.8	0.9	0.3	0.5	1.2	0.8	0.8	0.9	1.5

January, February and March have been colder, and the rest of the year has been warmer, most pronounced in December and in the fall.

January is somewhat out of phase in this connection. The colder January may be due to local cold troughs above Norway, or to the fact that the tropospheric zonal index in the Western Hemisphere may begin to decline prior to the stratospheric warmings, as happened in the winter 1957—58. As mentioned above (p. 24), this effect may be explained by a direct influence of the Himalaya Mountains on the jet stream. NAMIAS (1958) has pointed out that the 700 mb mean January (1948—55) shows an area with positive anomaly heights, centered over the Davis Straits and an area with negative anomaly heights, centered over North Scandinavia compared with the earlier U.S. Weather Bureau normal.

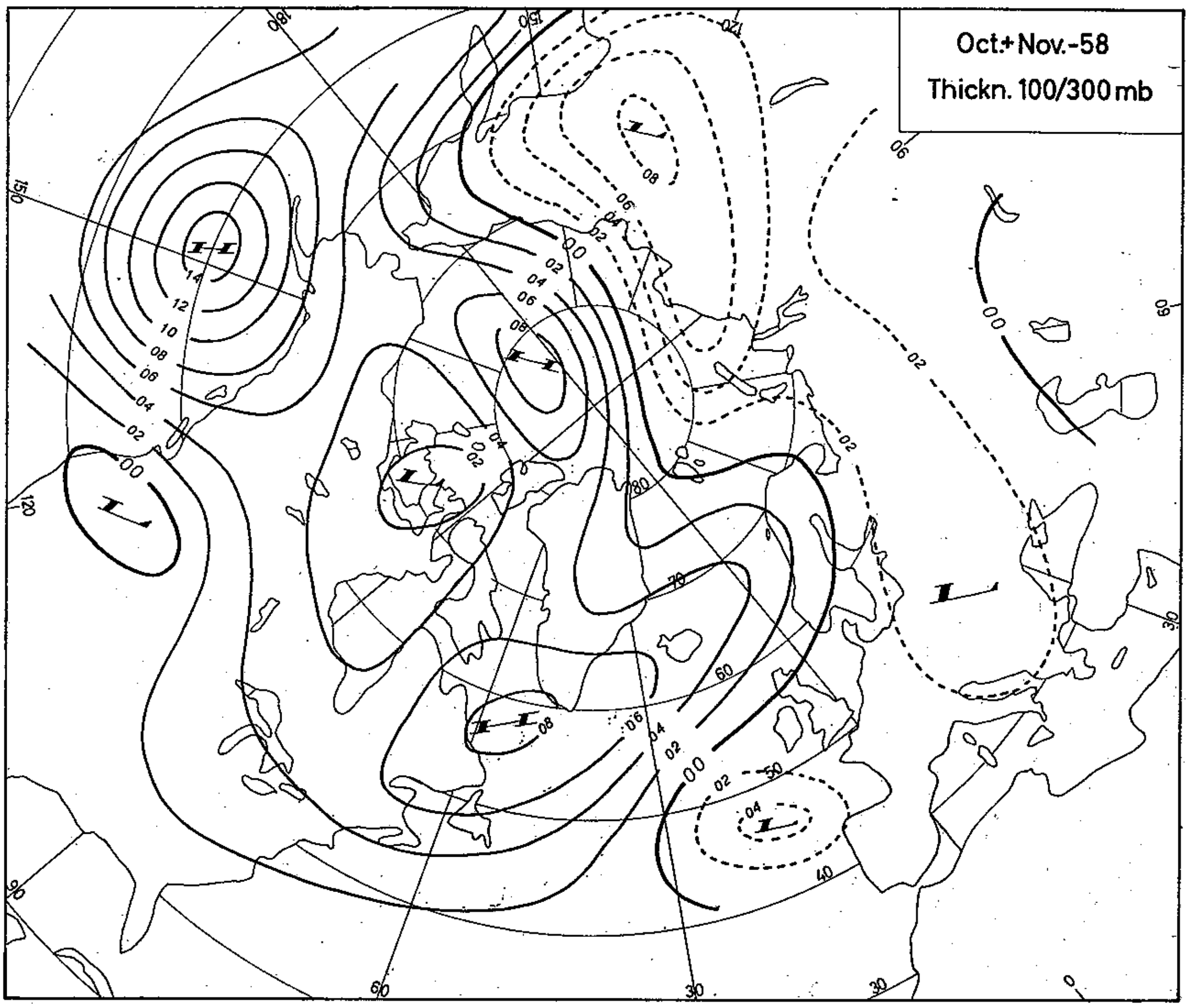


Fig. 41a. The sum October and November 1958 of the mean monthly departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer.

Another feature is worth mentioning regarding the mean temperature changes at 100 mb during the later years. From the Northeastern Atlantic over North Europe and Siberia, the stratospheric cooling from the fall to December has been greater than normal. This change will produce a stronger jet above North Europe to Asia, which means milder falls than normal in Norway. This result is in accord with the table above. If the Siberian cooling is combined with Canadian warming in the stratosphere (as happened in November 1958), negative temperature anomalies will follow in Norway (December 1958 and January 1959).

Greater stratospheric cooling than normal during the fall above Siberia produces a stronger jet above Himalaya in the early winter and the orographic effect will be intensified, followed by greater disturbances in the stratosphere, which again will favour an early breakdown of the circumpolar vortex.

In this way, by estimation of the reciprocal action between the jet stream and the

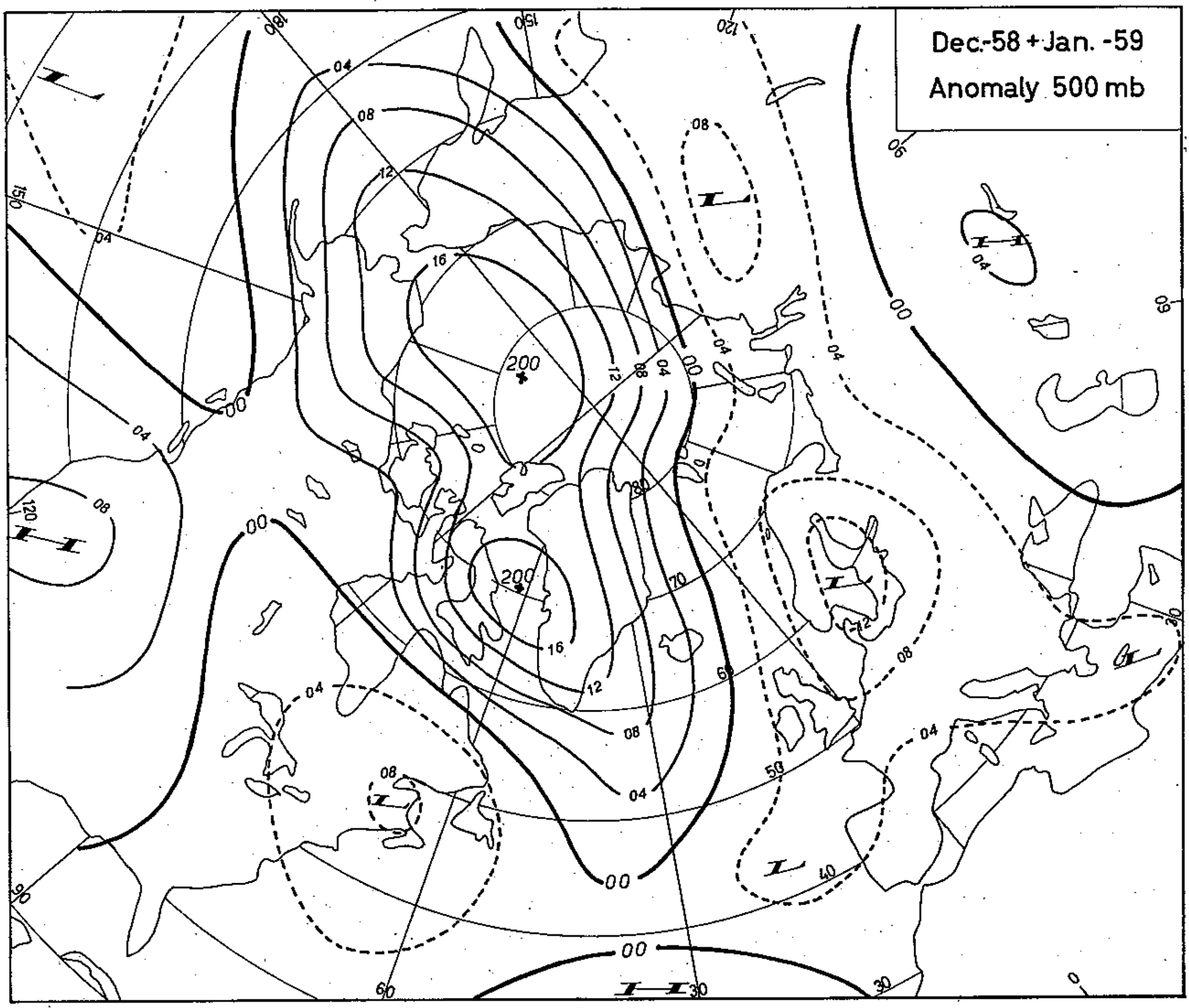


Fig. 41b. December 1958 and January 1959. Mean anomalies of the 500 mb heights.

Himalaya Mountains, and of the stratospheric disturbances and their effects on the troposphere, we may get an explanation of the circulational connection between mild fall and cold late-winter in Norway.

There has been a climatological trend in the precipitation changes in Norway, also, consistent with the above mentioned stratospheric circulation type.

This only indicates that the annual temperature changes (by monthly means) in the lower stratosphere at temperate and polar latitudes of the Northern Hemisphere probably have been more irregular and non-symmetrical in the later years than normal. It seems, that the yearly variation of the general circulation has been subjected to a rhythmical change embracing the whole atmosphere of the Northern Hemisphere.

4. The winter 1958—59.

a. Mean temperature changes at 100 mb. The changes from month to month through the period September 1958 to March 1959 are presented by the figures 31—37. Fig. 32

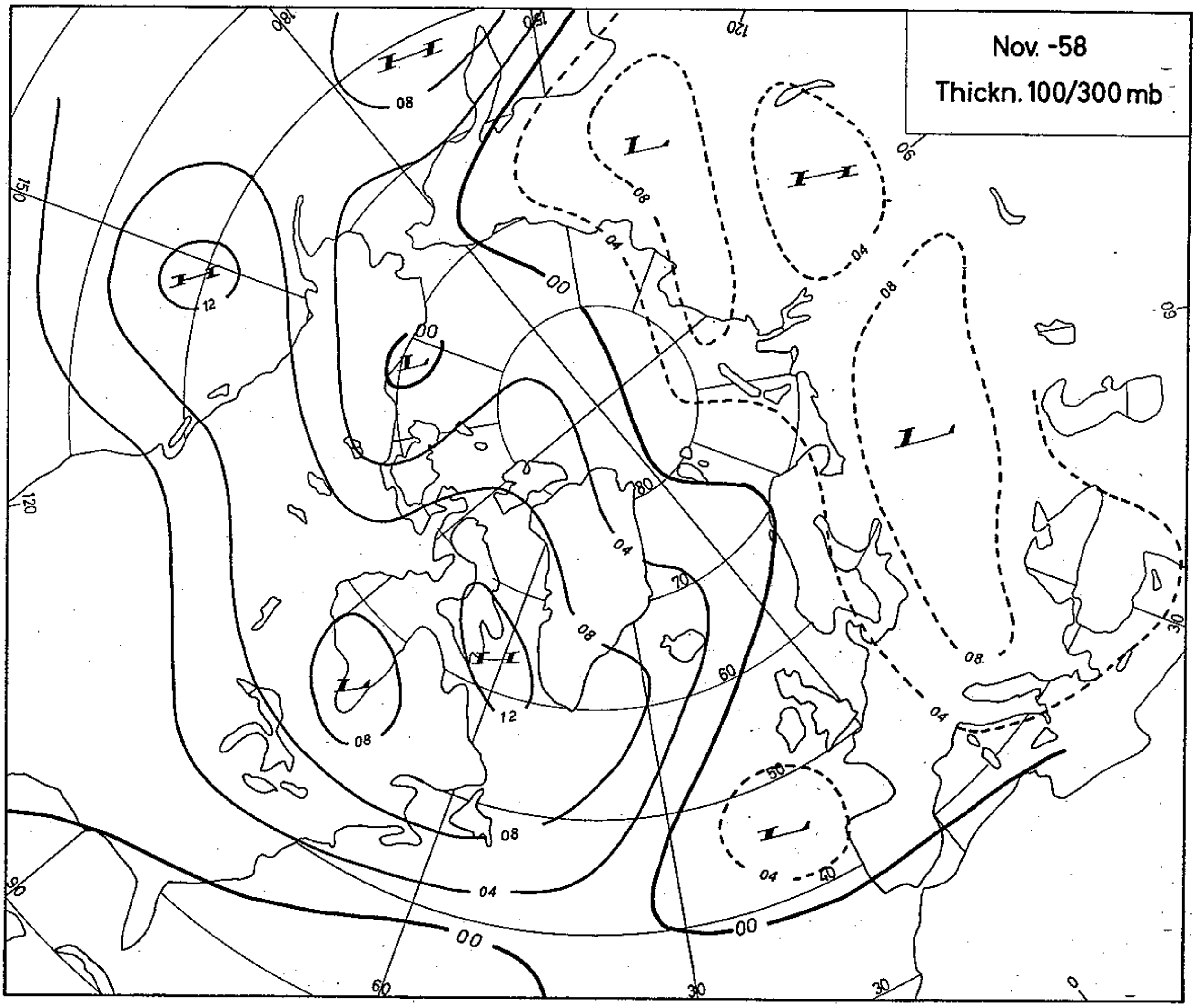


Fig. 40. November 1958. Mean departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer.

illustrates the great difference between the mean stratospheric temperature changes from October to November on the western and eastern part of the Northern Hemisphere at polar and temperate latitudes, the eastern part having strong cooling at 100 mb with the maximum zones above the northern part of Eurasia, and most of the western part having pronounced warming at 100 mb with the maximum zone above mid-Canada. Another warming belt is found at sub-tropical latitudes over Africa and Western Asia. It should be noted that North Siberia had the strongest cooling at 100 mb in the Northern Hemisphere from September to October 1958, with an area southwest of Mys Cheliuskin having a mean cooling at 100 mb of about 10°C (Fig. 31).

Fig. 32 is a clear interpretation of a pronounced and continual regime of vertical motions in the northern stratosphere in November 1958. At 100 mb, the air must have been ascending from Eastern Canada to mid-Siberia, with a maximum along 60°N, and descending from the northeastern coast of Asia across the Bering Sea and Alaska

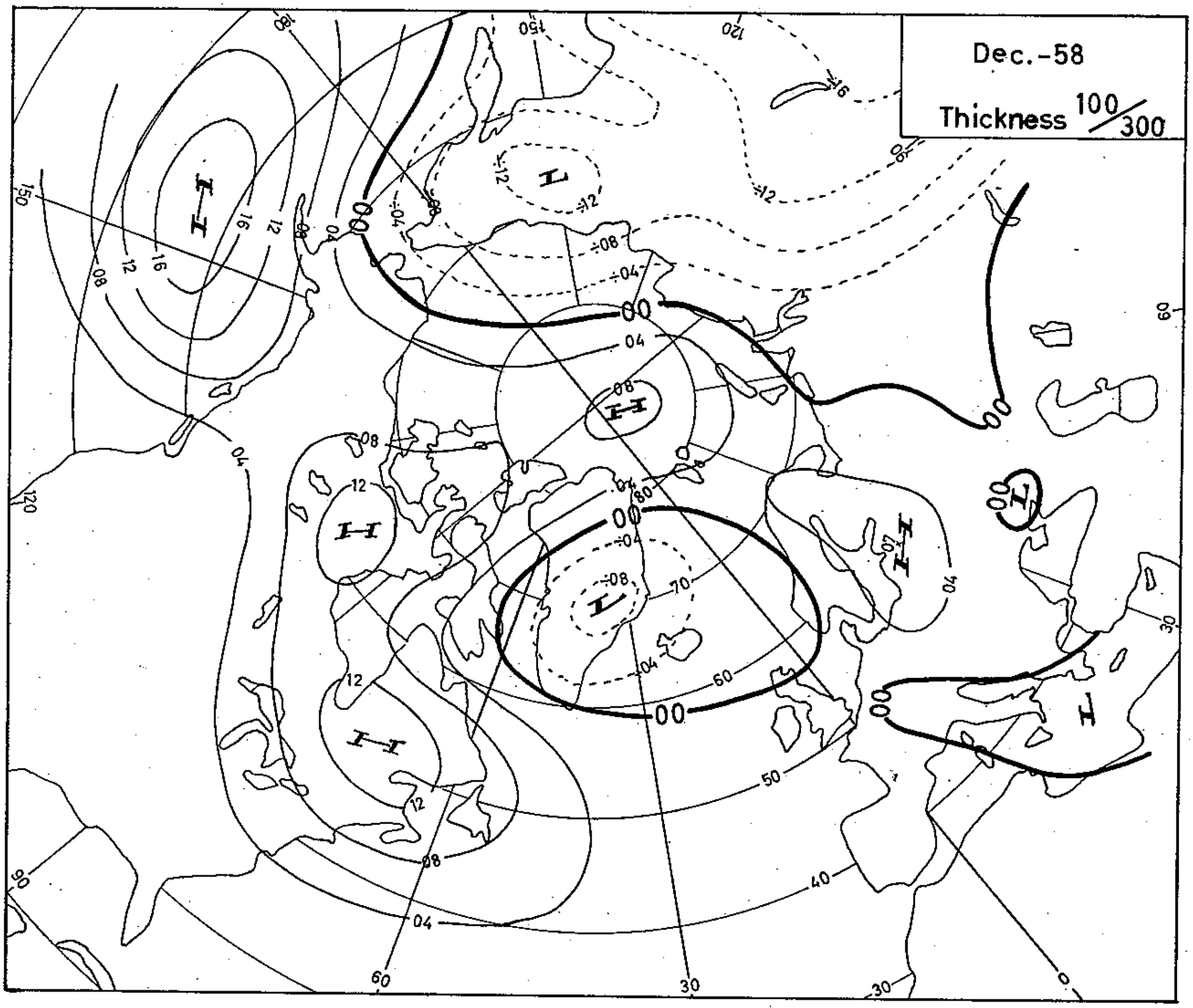


Fig. 41c. December 1958. Mean departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer.

to mid-Canada. The subsidence must also have covered most of the Arctic Basin. Fig. 54 verifies the temperature rise in November 1958 at 25 mb, along the 80°W meridian. **Hare** has also shown that the warming started suddenly in mid-October at 25 mb over Alaska.

The geographical distribution of the mean stratospheric temperature changes from November to December (Fig. 33) is quite different from the previous month. The strongest cooling at 100 mb is now in the Alaskan and Canadian Arctic, with a warming belt from the Southeastern U.S. across the Atlantic and Europe to temperate and subtropical latitudes in Asia.

Fig. 34 shows that the mean cooling from December 1958 to January 1959 at 100 mb had its maximum area over Baffin Land. There is a warming band south of the cooling area round nearly the whole Northern Hemisphere. The warming belt had its maximum values and its most northern position above East Siberia, where the warming extended

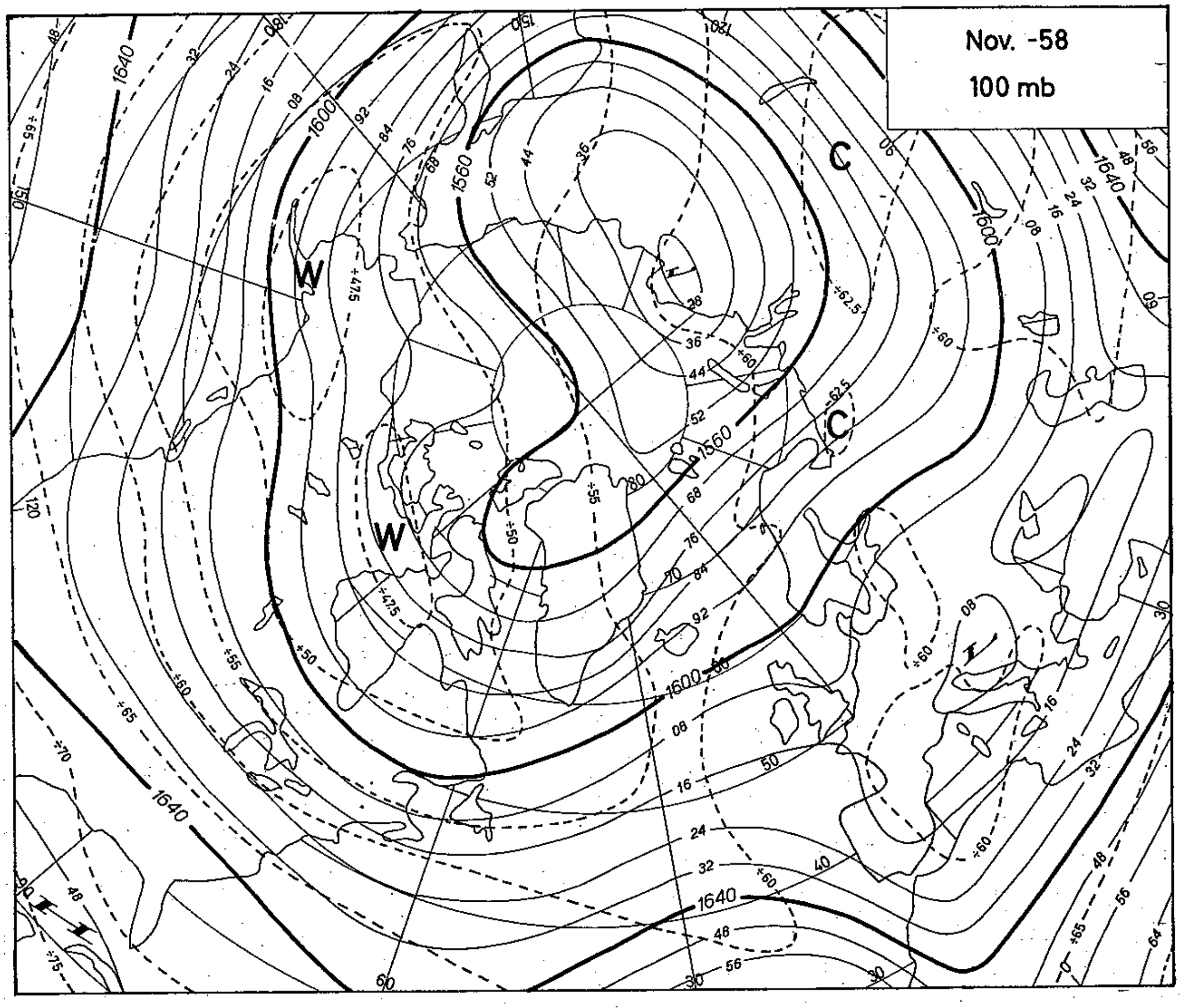


Fig. 42. November 1958. Mean contours and isotherms of the 100 mb surface.

partly into the Arctic Basin. Both figures 33 and 34 are included in Fig. 35, presenting a nearly circular cooling area with the center over Baffin Land, and a warming belt round the whole Hemisphere.

Fig. 54 unveils two sudden and strong temperature rises at the 80°W meridian, the first one at the end of January, and the second one in the beginning of March. Although these two impulsive warmings resembled one another (HARE 1960), they had quite different influences on the circumpolar vortex. The warming at the end of January had its maximum zone over North America near 60°N. The result is a weakening of the stratospheric westerlies in the southern part of North America, and a strengthening of the stratospheric circumpolar vortex in the Arctic. In the beginning of March, the maximum warming was situated over Baffin Land and Ellesmere Land, and this warming marked the closing of the stratospheric winter vortex on the American side of the pole and the displacing of the vortex center towards Siberia.

The mean temperature changes in the 100 mb surface, following the two warming

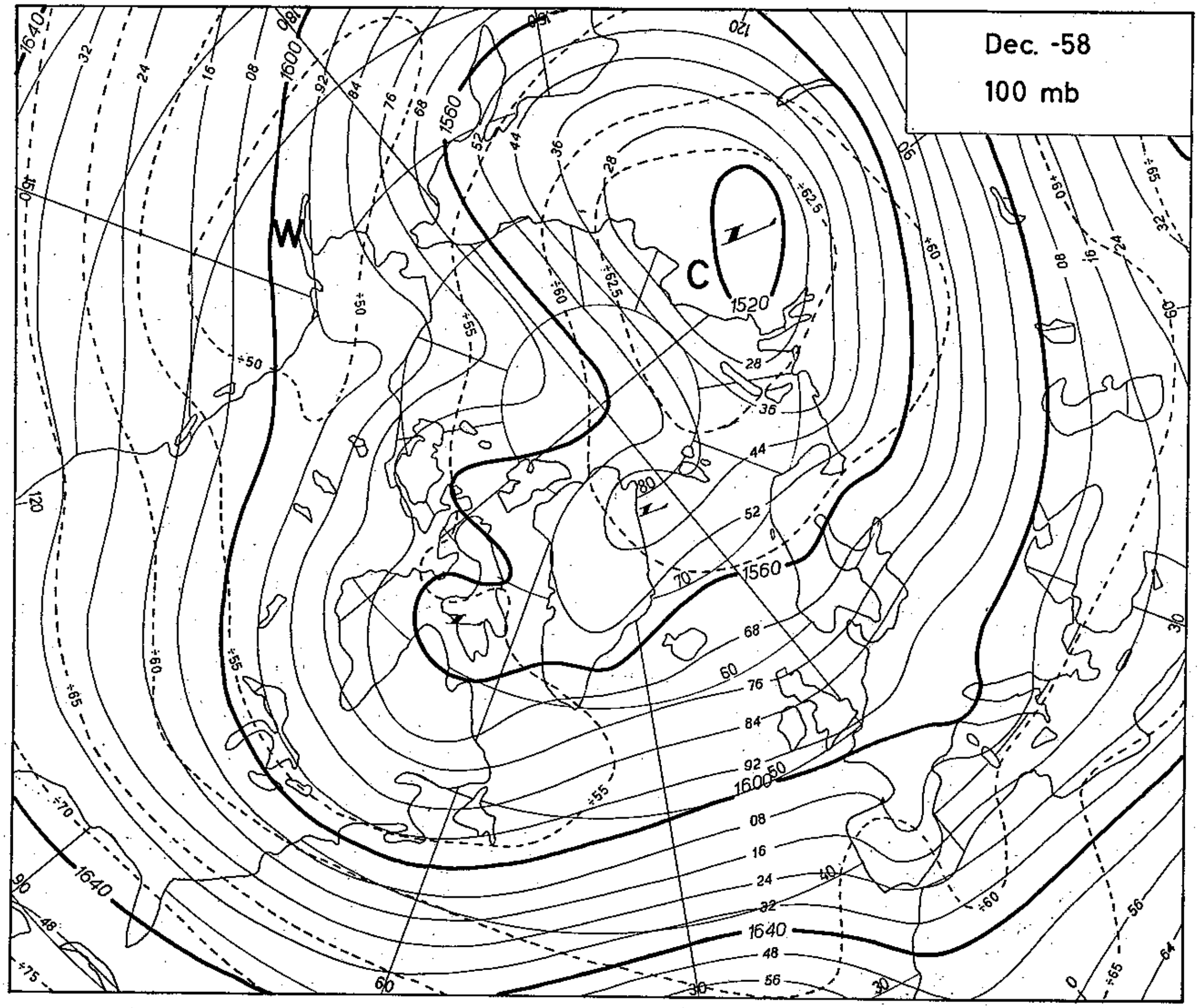


Fig. 43. December 1958. Mean contours and isotherms of the 100 mb surface.

epochs at 25 mb, are presented in Figs. 36 and 37. Fig. 36 shows warming over North America with the maximum zone near 60°N. Most of the Arctic Basin, North Siberia and Europe had weak cooling. In Fig. 37, the pronounced warming area, with the maximum zone near Thule, is dominating. A cooling belt stretches from the North Pacific, across Asia, the Mediterranean and North Africa to the South Atlantic and South America.

b. Anomaly maps of the 100/300 mb thickness patterns and the 500 mb heights. The mean tropospheric circulation in October 1958 had only small departures from normal (ANDREWS, 1958). In the 500 mb surface (Fig. 38), the Greenlandian and Siberian troughs were a little deeper than normal and small anticyclonic deviations existed over North America and Europe. The mean tropospheric temperature distribution was also near normal in October.

Looking at the mean temperature pattern in the lower stratosphere (Fig. 39a), October shows a marked warm area with two centers over the Alaskan Arctic — the

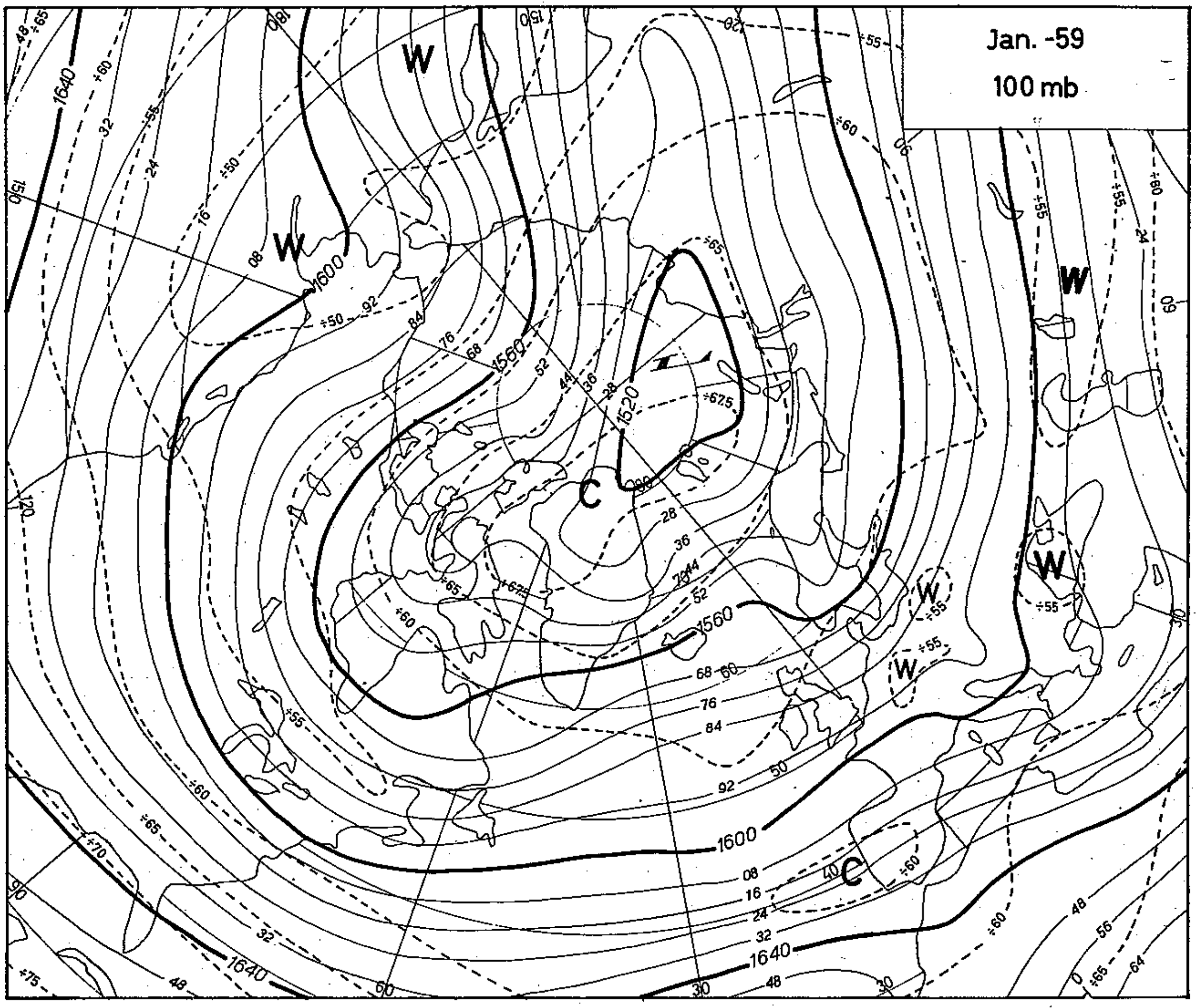


Fig. 44. January 1959. Mean contours and isotherms of the 100 mb surface.

Alaskan Gulf, and a weaker warm area above the North Atlantic and Europe. Further, a weak cold area extends from the Western U.S. to Northeastern Canada, and a stronger cold area is above East Siberia. As mentioned above, the strongest cooling at 100 mb from September to October 1958 in the Northern Hemisphere was in Siberia. Fig. 39b presents the mean anomaly picture of the 500 mb contours in November 1958. There are several qualitative conformities between Figs. 39a and 39b, and it is reasonable to assume that a stratospheric influence has been active in forming the mean anomaly patterns of the tropospheric circulation in November 1958. According to Paroczay (1958), the changes in the tropospheric circulation from the October regime occurred about the middle of November, and his 700 mb map, November 16.—30., minus November 1.—15., 1958, has pronounced qualitative conformities with Fig. 39a.

During November 1958, the stratospheric warming moved to Eastern North America (Hare, 1960). Fig. 40 presents the monthly departure from the 5-year mean (1949—53) of the 100/300 mb thickness pattern, and shows positive anomalies from the North

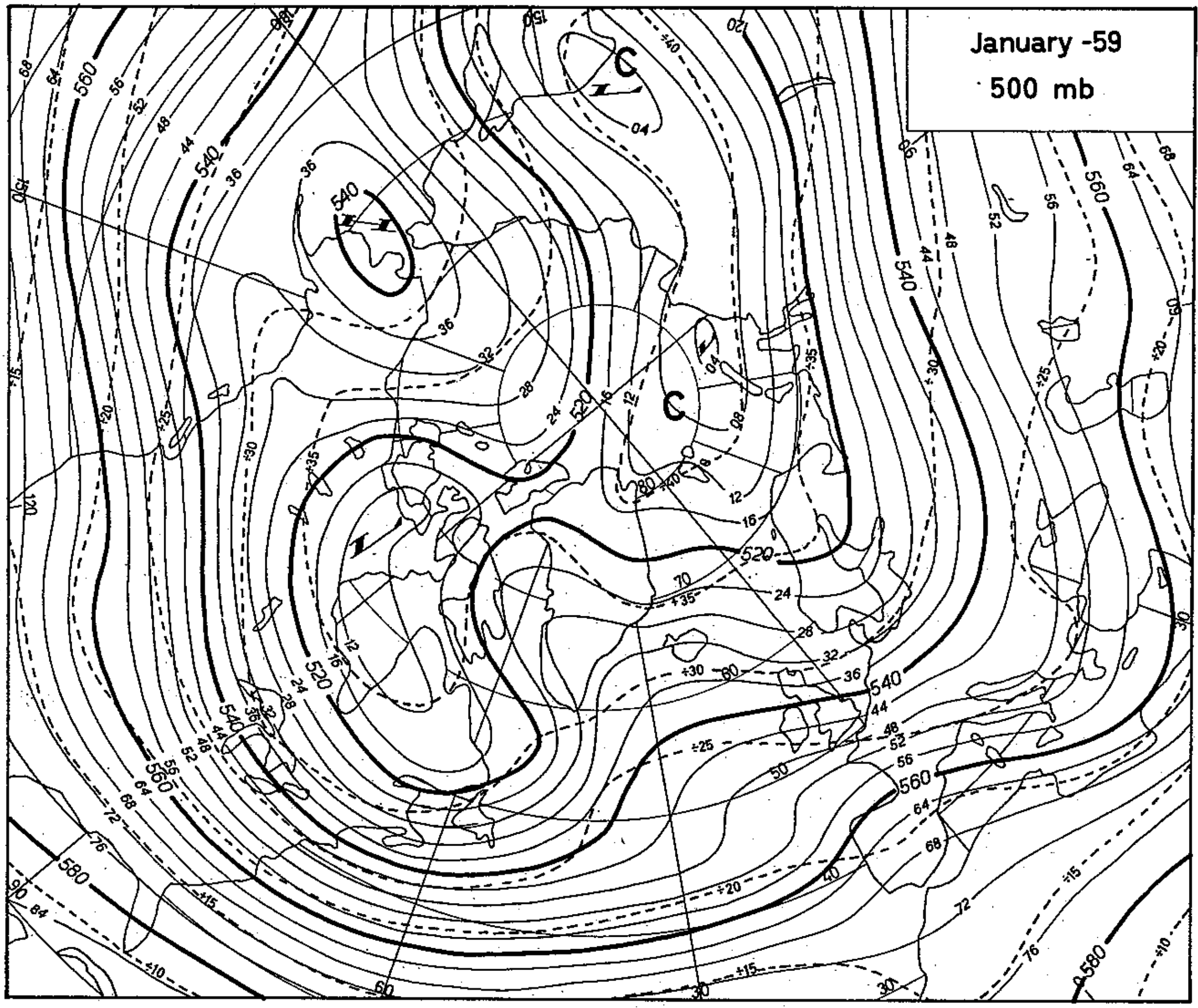


Fig. 45. January 1959. Mean contours and isotherms of the 500 mb surface.

Pacific across North America to the North Atlantic, negative anomalies above Europe and Siberia. This is in accordance with the temperature change at 100 mb from October to November (Fig. 32).

As the difference between the stratospheric temperature changes on the eastern and western Hemisphere started early in October 1958, the mean departures of the 100/300 mb thickness pattern in October and November are added in Fig. 41a, showing negative anomalies from Southwestern Europe to Siberia, positive anomalies from the North Pacific across the American Arctic and Greenland to the North Atlantic. The 100/300 mb anomaly thickness pattern changed only little from November to December (Fig. 41c), and it is reasonable to assume that the stratospheric anomally thickness pattern during October-November-December 1958 in the Northern Hemisphere has influenced the troposphere and was a contributing cause of the tropospheric circulation pattern during December 1958—January 1959 (Fig. 41b). There is no doubt that the stratospheric temperature anomalies in November 1958 are of stratospheric origin.

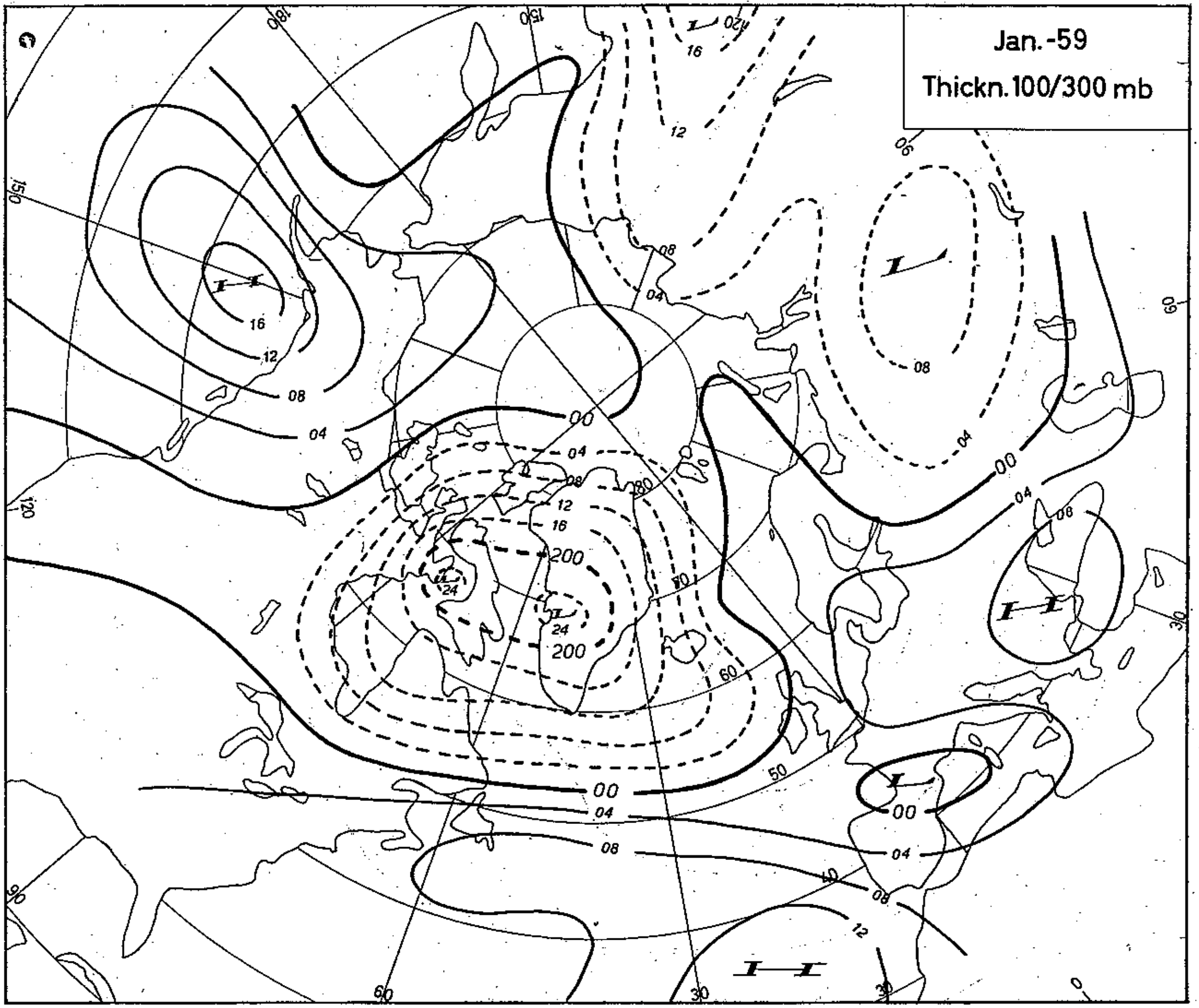


Fig. 46a. January 1959. Mean departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer.

The temperature pattern at 100 mb (warm ridge and cold trough, Fig. 42), is consistent with impulses originating in the stratosphere (AUSTIN and KRAWITZ, 1956), and can only be explained by a continual regime of vertical velocities. The time-height section of temperature at Resolute (Fig. 55) also shows that the warming extends downward with time from 25 mb to 200 mb.

In accordance with the mean stratospheric temperature changes during September-October-November 1958 (Figs. 31—32), the circumpolar vortex center at 100 mb moved from the central Arctic to North Siberia (Fig. 42). During December 1958, the stratospheric cooling invaded the Arctic, Greenland and North America. However, the 100 mb vortex center persisted in North Siberia with a trough extending across Greenland to Labrador (Fig. 43). In January 1959, the stratospheric cooling strengthened in the Baffin—Greenland area and the vortex center at 100 mb moved to the Eurasian Arctic with troughs above Siberia and across Greenland to Hudson Bay (Fig. 44).

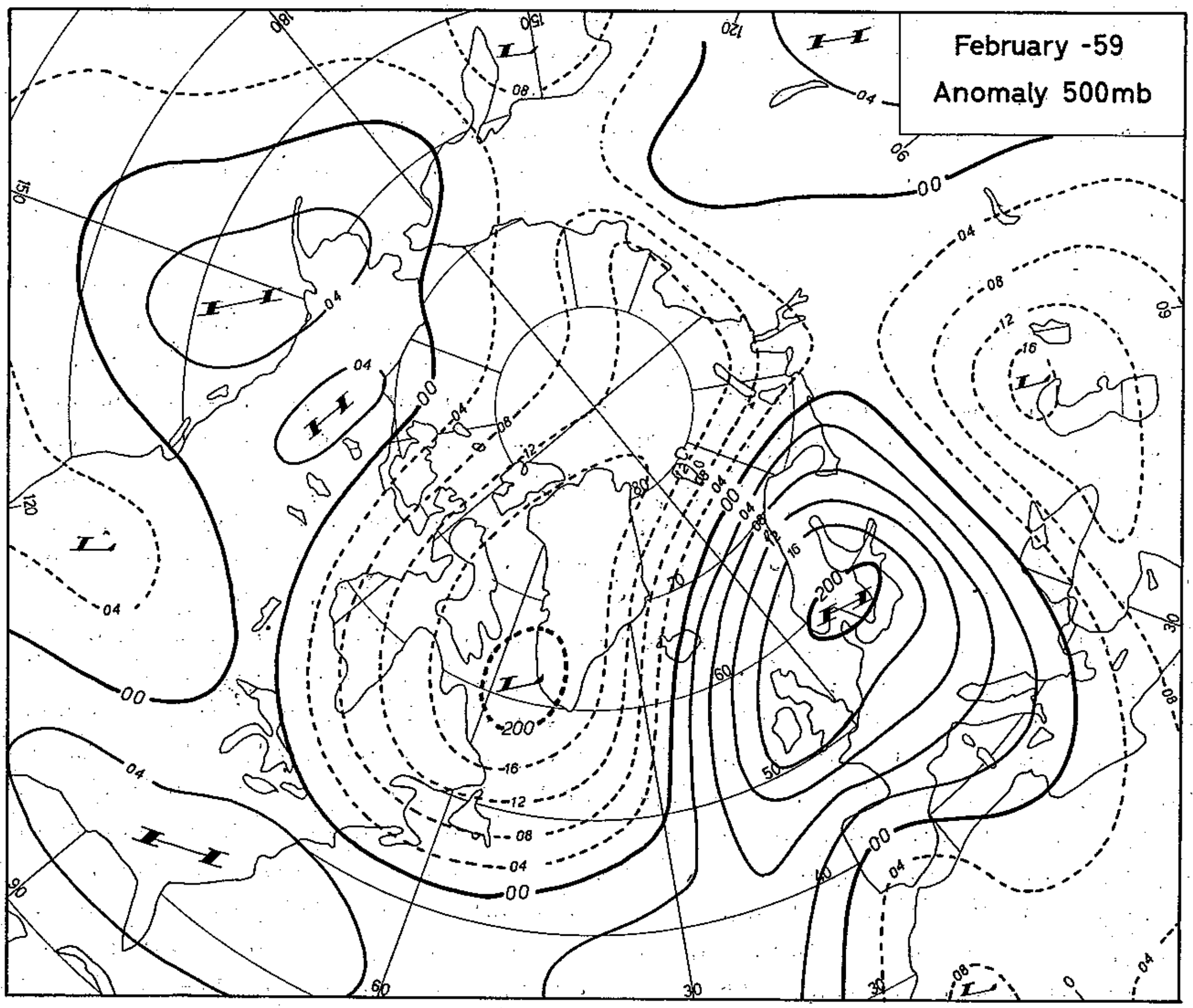


Fig. 46b. February 1959. Mean anomalies of the 500 mb heights.

The nearly normal tropospheric circulation pattern in October 1958 (Fig. 38) was changed to a regime with the circumpolar vortex center displaced to North Siberia and a ridge from the Bering Sea across the Arctic and Greenland to the North Atlantic. This regime persisted in its broad features to January 1959 (Fig. 45).) Based on the shown material, it is a reasonable postulate that the mean tropospheric circulation in the Northern Hemisphere during November-December 1958—January 1959 has been influenced by the mean stratospheric temperature changes September-October-November 1958, that is, the trend of the mean circulation change occurred primarily in the stratosphere.

A few points are worthy of note. The development of the Labrador low in the troposphere during December 1958 seems to have no special indication in the stratospheric temperature pattern, and it is probably an usual adjustment in the tropospheric circulation pattern.

During December 1958—January 1959, cold air was evolving in the Arctic. As in the previous winter 1957—58 it seems that the cold stratospheric core in the Arctic at first has not affected the troposphere dynamically, but on the contrary, has intensified the mean anticyclonic anomaly in the arctic troposphere.

However, during December 1958, and especially during January 1959, strong perturbations developed in the stratospheric polar vortex above North America. Hare (1960) has pointed out two cases of probable inter-action between stratospheric and tropospheric waves in the American Arctic, namely 20.—24. December 1958 and 2.—10. January 1959 (Fig. 54). The writer will give attention to the strong stratospheric oscillations during the period 10.—25. January 1959. It is most probable that the stratospheric circulation and its large disturbances in North America and the American Arctic during these periods have co-operated with and influenced the troposphere, and especially in the days 15.—25. January have contributed to the displacing and releasing of the tropospheric energy in the area Labrador—Greenland—the North Atlantic—North Europe and the European Arctic. A discussion of the stratospheric temperature changes will be given in the next section, while only the changes in the tropospheric circulation will be mentioned briefly here.

On 14. January 1959, a tropospheric frontal wave began to develop south of the Great Basin. The cyclogenesis increased rapidly, the low curved northward and later towards north-west and west. On 19. January, the low center (965 mb) had reached the northern coast of Hudson Bay, where the low stagnated and filled rapidly. The low development was favoured by upper waves (Bjerknes, 1951). Fig. 54 shows that oscillations at 25 mb were running from NW to SE. On 21. January 1959, a new frontal low was in strong development south of the Great Basin. The low moved northward, reached Baffin Land on 24. January as a strong and deep low (955 mb), stagnated and began to fill. This low brought warm tropospheric air over Labrador and the North Atlantic to South Greenland. The generated shear by South Greenland produced a new wave which managed to penetrate the tropospheric blocking ridge over Greenland and the Northeastern Atlantic, a blocking which had existed a long time (Fig. 45). The wave-low moved eastward north of Norway. In connection with the penetration of this low, changes occurred on a large scale in the troposphere over the Northeastern Atlantic, Europe and the adjacent Arctic, and was followed by unusually strong and frequent cyclonic developments in the area Labrador—Greenland—European Arctic towards Novaja Zemlja. These tropospheric developments were favoured by the stratospheric circulation (Figs. 44, 46a, 47 and 49b) and lasted a long time. The warmest February and March ever recorded in North Scandinavia, occurred during this period.

Fig. 46a presents the stratospheric favouring of cyclonic developments in the troposphere in the region Labrador—the North Atlantic—East Greenland during January 1959, and Fig. 46 b shows the anomaly pattern of the mean 500 mb heights during February 1959. Fig. 47 shows that the stratospheric favouring of the cyclonic developments over the Greenland—Eurasian Arctic continued in February.

During February 1959, an intense and persistent blocking high existed in the troposphere over Western Europe (Fig. 46b). This strong anticyclonic development had

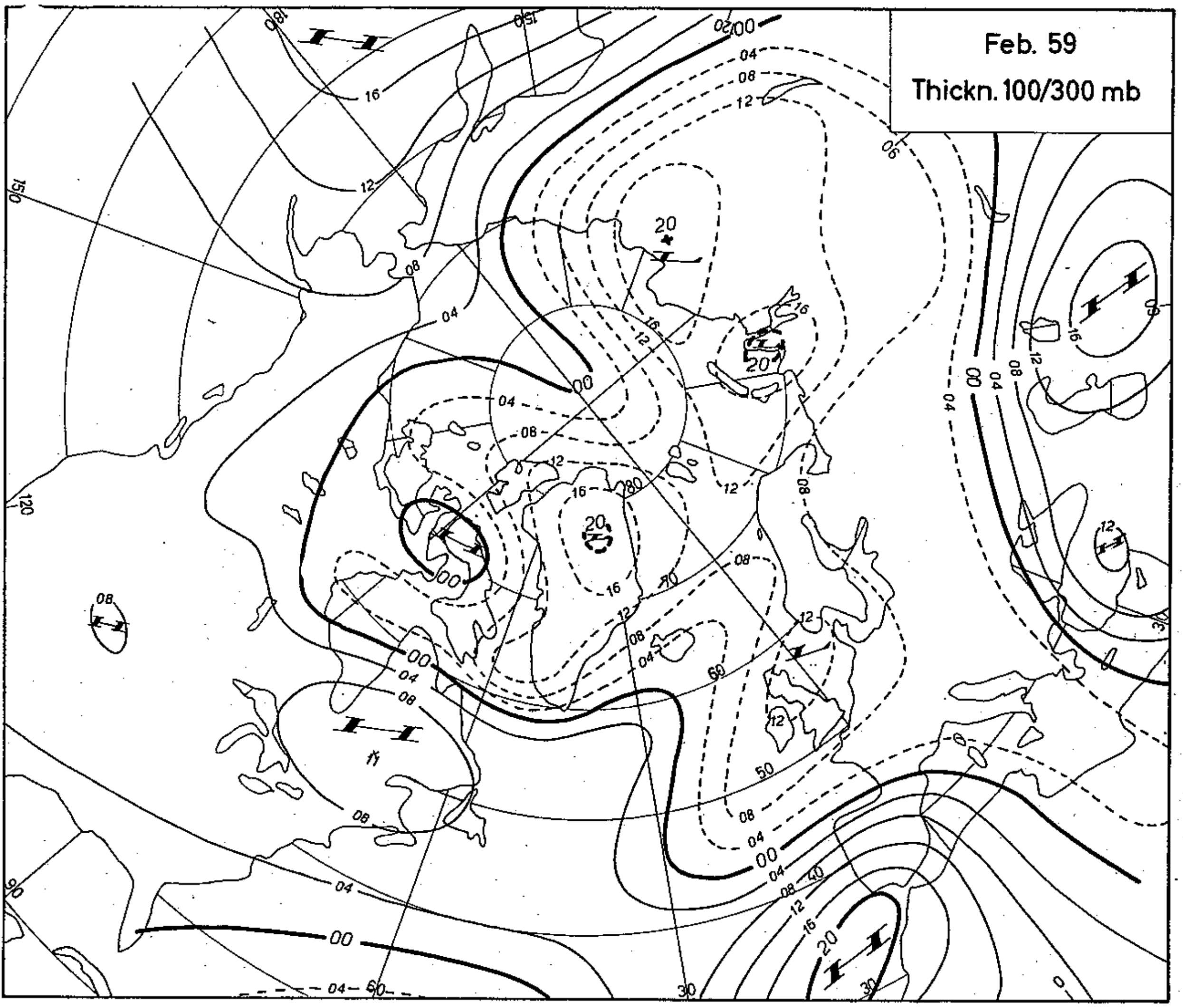


Fig. 47. February 1959. Mean 100/300 mb thickness departures from the 5-year mean (1949—53).

only a weak previous indication in the stratosphere (Fig. 46a), and must be taken for a tropospheric feature solely, perhaps as a kind of adjustment to the unusually strong cyclonic developments in the Atlantic—European Arctic. Fig. 47 also shows ascending motions in the lower stratospheric layer above the tropospheric blocking high in Western Europe.

As pointed out by O'Connor (1959) and Hofman (1959), the change of the mean tropospheric circulation pattern from January to February 1959 was extremely large in the Northern Hemisphere. Fig. 48a presents the change from January to February 1959 of the anomalies of the mean 500 mb contour heights. Fig. 48b shows the change from November 1958 to January 1959 of the mean 100/300 mb thickness pattern. The great conformity between these two figures confirms the postulate that the change of the tropospheric circulation pattern from January to February 1959 to some extent was prepared, or facilitated, by the stratospheric circulation and the previous change of

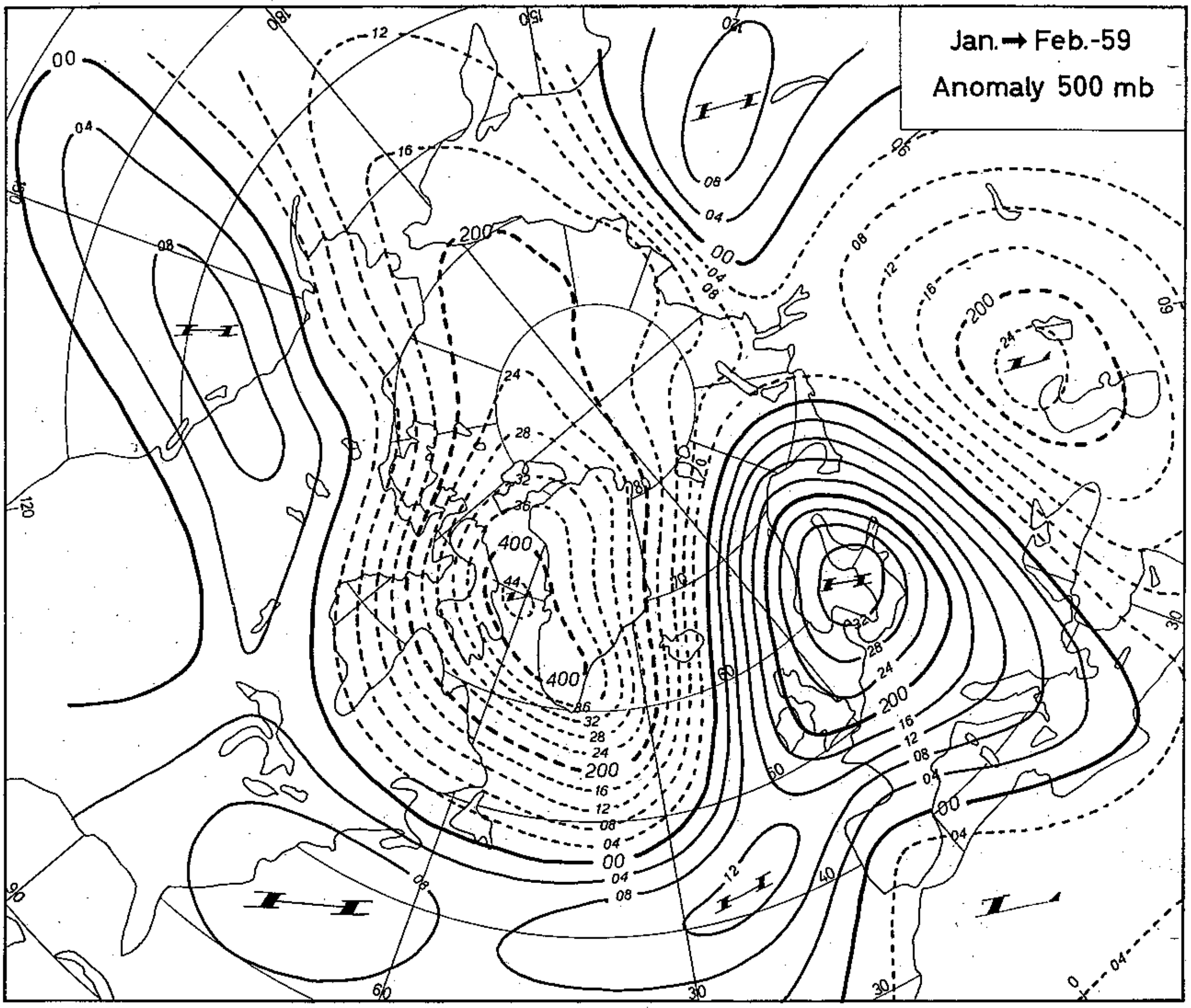


Fig. 48a. Change from January to February 1959 of the mean monthly 500 mb height anomalies.

the stratospheric temperature pattern. As in Figs. 46a and 46 b, the European high in Fig. 48a has no counterpart in Fig. 48b.

The change of the tropospheric circulation over the North Atlantic, Northwestern Europe and the adjacent Arctic (Fig. 48a) was explosive and occurred during a few days in the last week of January. Looking at the atmospheric happenings in the Northern Hemisphere during December 1958, and especially during January 1959, one gets the impression that the atmosphere was in a transitional state, chiefly manifested in the stratosphere. The tropospheric circulation type was mainly conservative and only locally affected by the stratospheric fluctuations (North America). When the tropospheric circulation type changed in the last part of January and became consistent with the stratospheric conditions, this may explain the explosive character of the tropospheric changes, but only an extensive, static and dynamic, analysis of the whole atmosphere can clarify the stratospheric participation in the tropospheric overturnings themselves.

Figs. 49a and 49b present the mean flow and the mean temperature pattern in

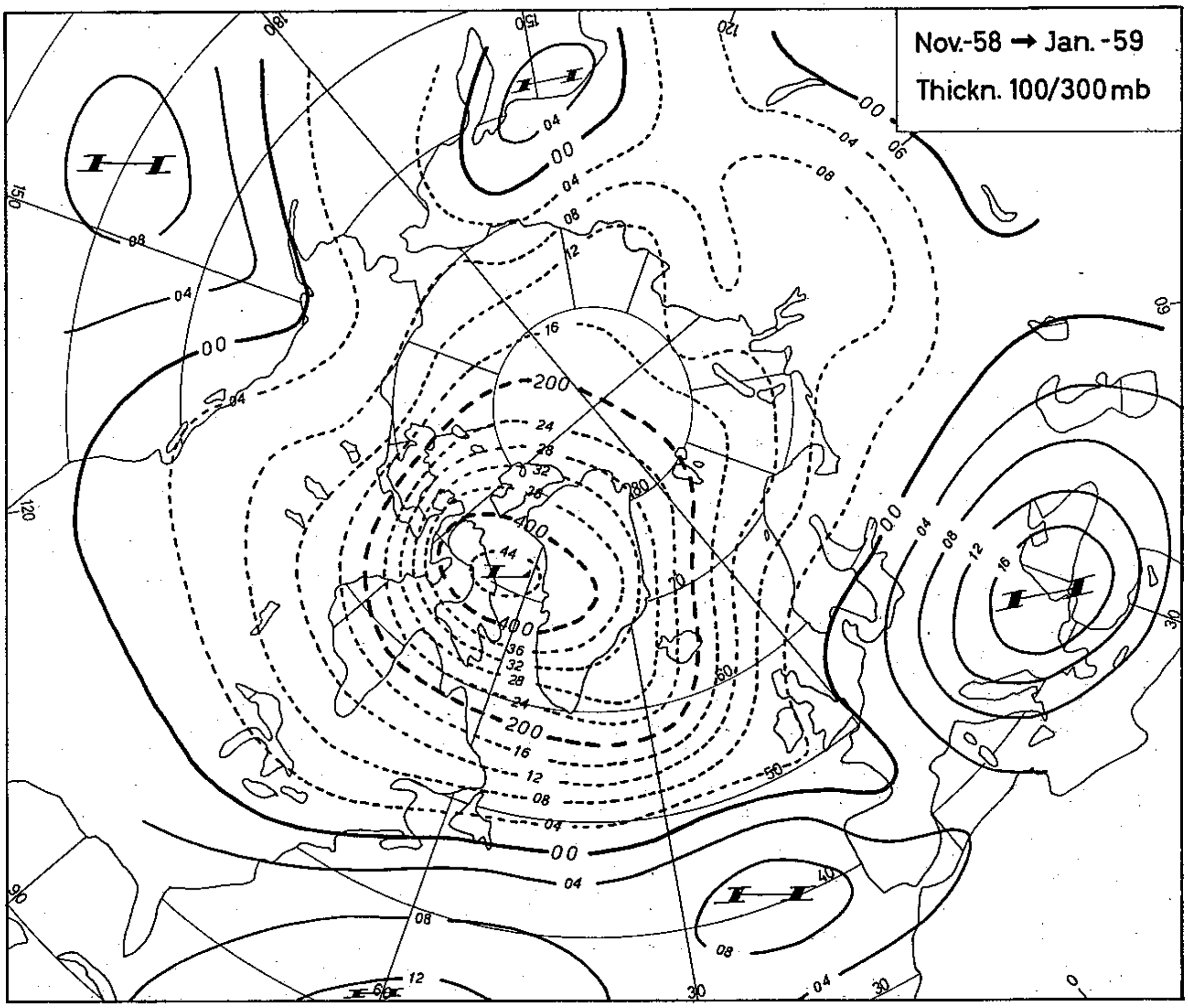


Fig. 48b. Change from November 1958 to January 1959 of the mean monthly 100/300 mb thickness layer.

February 1959 at 500 mb and 100 mb, respectively, and show the joint action of the two levels, especially in the area of cyclogenesis, from Davis Strait to the northern Norwegian Sea.

The mean tropospheric circumpolar vortex changed only little in the Arctic from February to March 1959 (Figs. 46b and 50). In the sub-Arctic, however, some alterations occurred. The flow at 500 mb weakened in the North Atlantic and the Norwegian Sea, and the vortex strengthened in Western Siberia. The February trough over Eastern Siberia moved to Alaska, an anticyclonic departure appeared over Kamchatka, and over the western U.S. a weak cyclonic anomaly turned anticyclonic in March. The Labrador trough moved to the northeastern Atlantic, and the blocking high over Western Europe moved eastward and weakened.

The mean tropospheric flow pattern in the Northern Hemisphere in March 1959 was a reasonable progression of the strong February circulation and was mainly consistent with the stratospheric anomaly pattern in February (with exception of the European

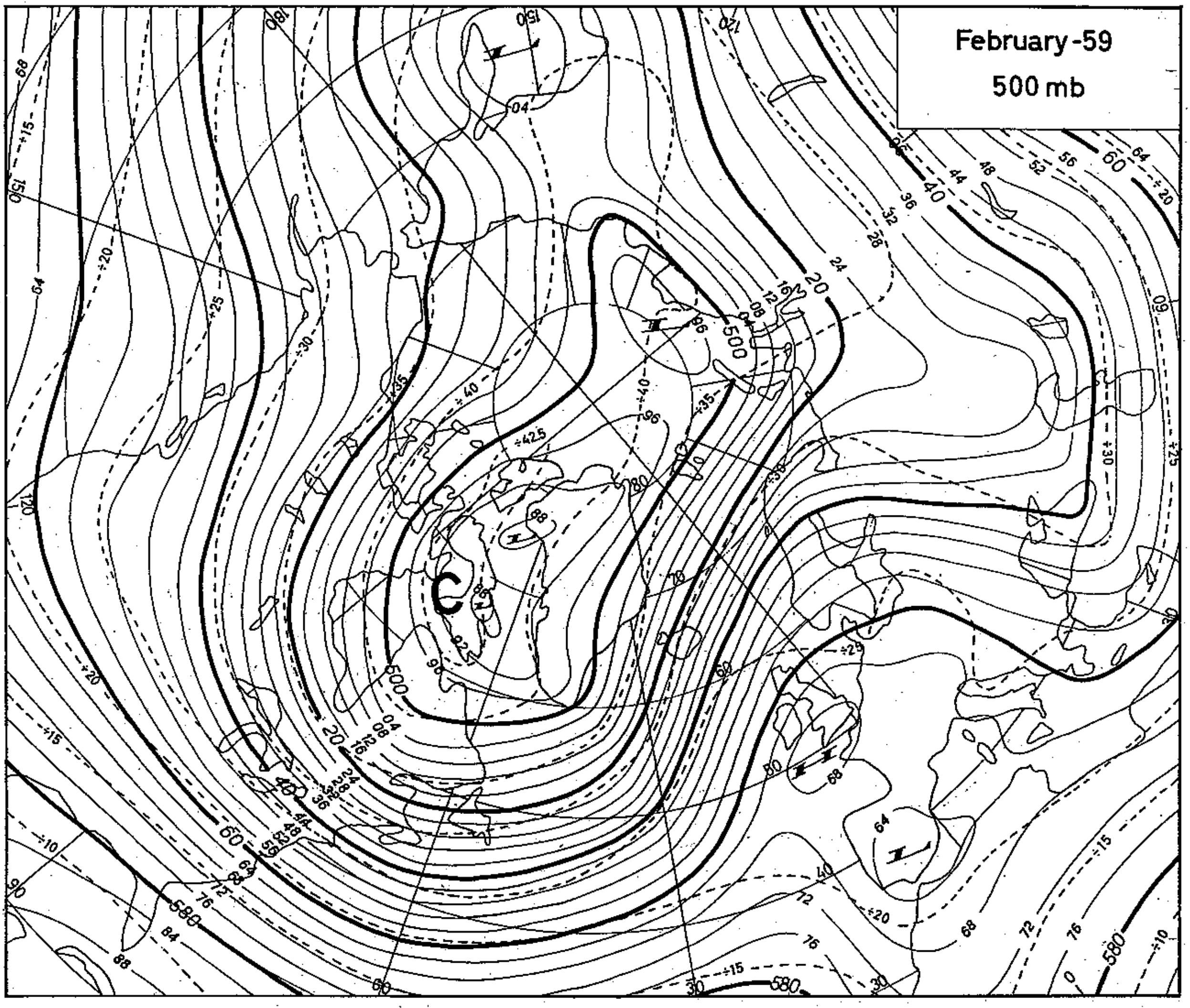


Fig. 49a. February 1959. Mean contours and isotherms of the 500 mb surface. Equidistance 40 m.

blocking high, Fig. 47), but was only slightly in correspondence with the great and sudden stratospheric warming which occurred in the American Arctic and sub-Arctic at the beginning of March (Figs. 54 and 55). The warming extended downward to the lower stratospheric levels (Fig. 37) and displaced the circumpolar vortex center at 100 mb to the Siberian coast (Fig. 51).

The mean temperature pattern at 100 mb in March is evincing an extensive regime of vertical velocities, the Canadian warm core being more than 20°C warmer than the Siberian cold core (at same latitudes).

The strong stratospheric warming in the American Arctic at the beginning of March, a warming which marked the end of the stratospheric winter temperatures and completed the breakdown of the stratospheric jet on the American side of the pole (Hare, 1960), does not seem to have affected the tropospheric circulation before April, and then only weakly. Fig. 52a presents the March departure from the 5-year mean (1949 —53) of the 100/300 mb thickness pattern, and Fig. 52b shows the April anomalies

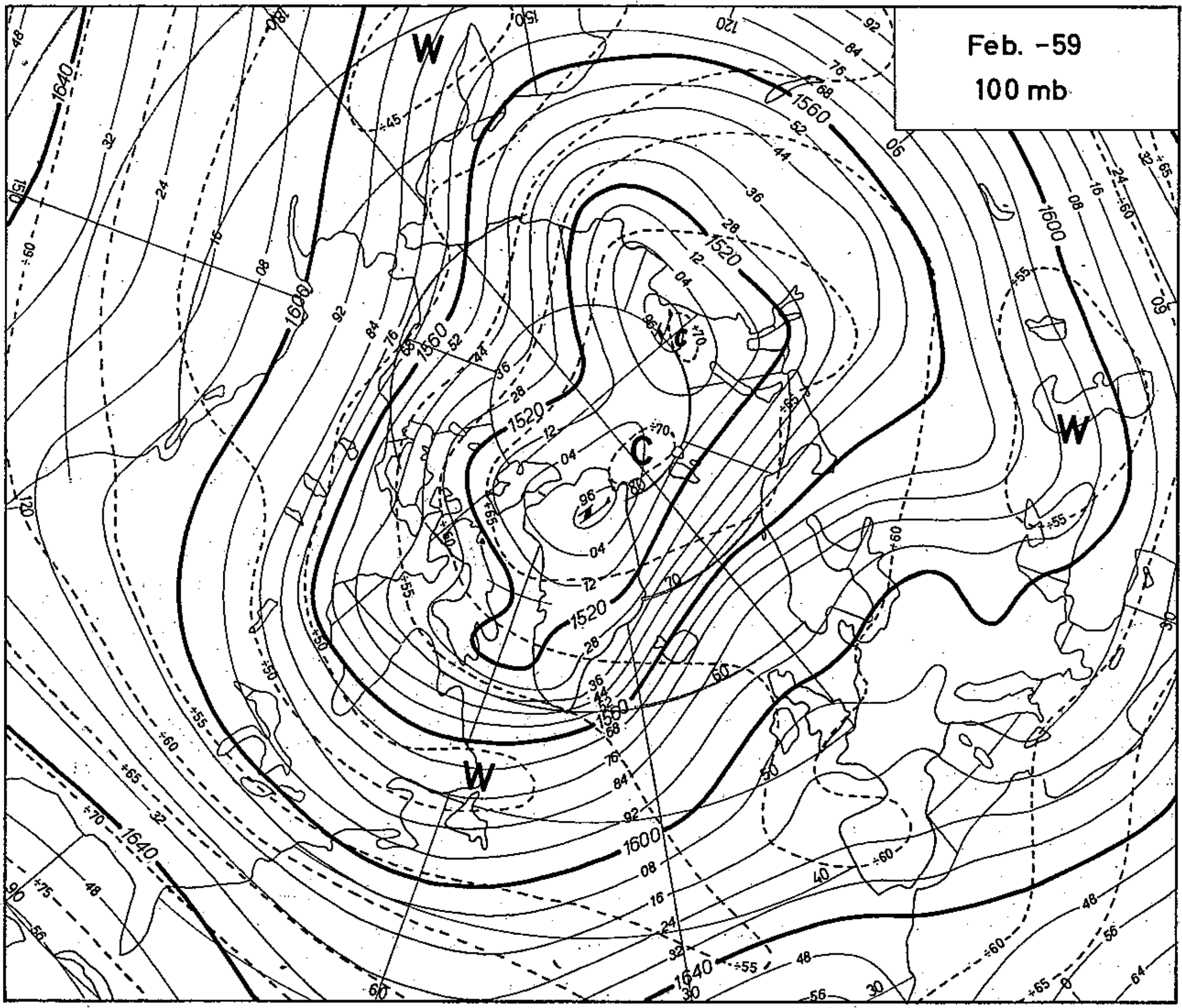


Fig. 49b. February 1959. Mean 100 mb contours and isotherms. Equidistance 80 m.

of the 500 mb contour heights. The two figures agree in the positive areas at Greenland and in the Arctic, and in the negative areas in Siberia. With the exception of Europe (the March blocking high) the similarity between Figs. 53a and 53b are better. Fig. 53a presents the change from February to March 1959 of the 100/300 mb anomaly thickness pattern, and Fig. 53b shows the change from March to April 1959 of the 500 mb height anomalies.

c. *Discussion of the stratospheric temperature changes.* The stratospheric cooling in the fall at temperate and arctic latitudes is usually stronger in the Eastern Hemisphere than in the Western, but in 1958 the difference was much greater than normal. In early October, a stratospheric warming started over Japan, and progressed to the Aleutian area during the month (Fig. 31). During November, the subsidence warming intensified and extended further eastward to the American Arctic (Fig. 32). The stratospheric low center at 100 mb was displaced to North Siberia (Fig. 42). Fig. 54 shows that the warming was pronounced at 25 mb.

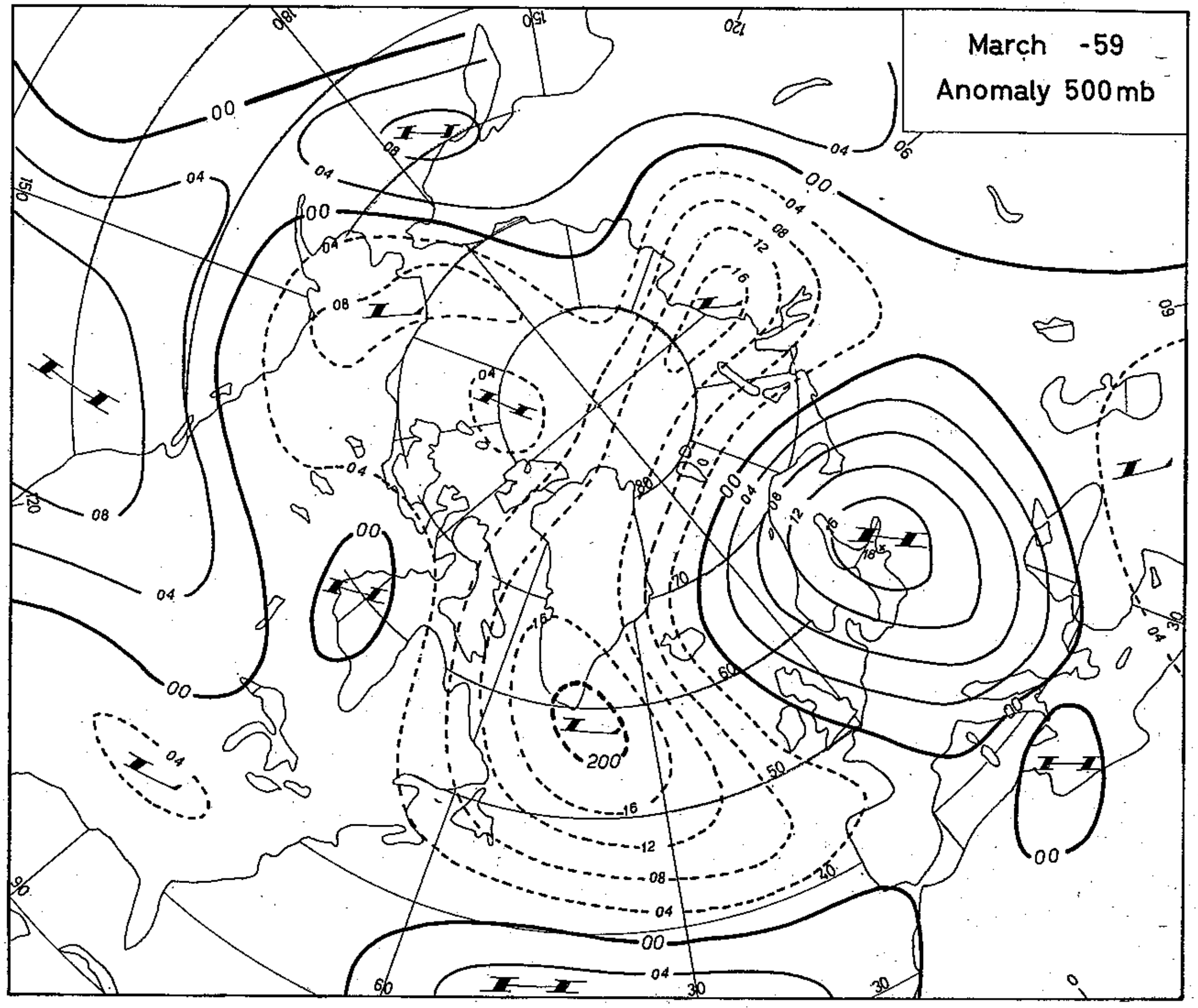


Fig. 50. March 1959. Mean 500 mb height anomalies.

Boville (1960) has discussed the October circulation, and concludes that the warm stratospheric Aleutian high is formed consequent to the tropospheric cyclogenesis. This is probably the case in the first half of the month, but not later. According to Teweles (see above p. 14), it is more reasonable to judge the circulation patterns in this area as orographic effects from the Himalaya Mountains on the jet stream which was intensified by the stratospheric cooling over Siberia (Fig. 31).

It is also unlikely to interpret the temperature pattern in Fig. 42 as an effect of tropospheric cyclogenesis. Fig. 55 shows that the warming over Resolute appeared first at 25 mb. Perhaps we may explain the stratospheric circulation pattern in November 1958 as a result of unstable stratospheric waves from Greenland to Siberia and orographic effects from the Himalaya Mountains. The extension of the subsidence eastward over Canada has probably been a stratospheric effect from higher levels.

Throughout the whole winter 1958—59, the lower stratosphere kept colder than normal above East Siberia and warmer than normal in the Aleutian—Alaskan area.

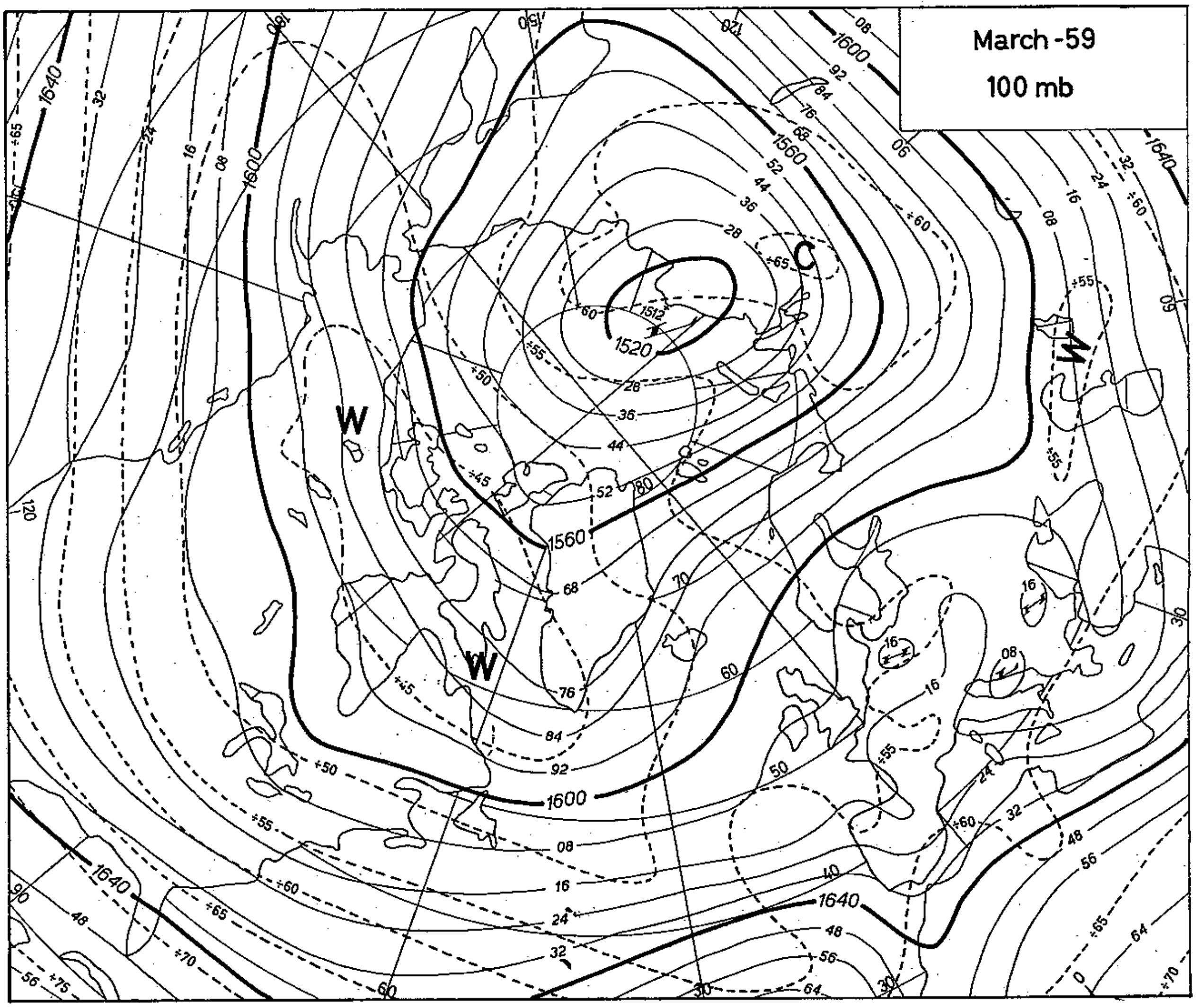


Fig. 51. March 1959. Mean 100 mb contours and isotherms.

The anomaly areas were, however, somewhat varying (Figs. 41c, 46a, 47 and 52a). Otherwise, great disturbances occurred in the stratosphere over the most part of North America, Greenland, the Northeastern Atlantic, Nort Europe and the adjacent Arctic.

During December, a cold trough appeared at 100 mb over Greenland, strongly intensifying and extending towards Hudson Bay during January (Figs. 41c, 43, 44 and 46a). Downstream from the Aleutian-Alaskan ridge to this trough, great stratospheric disturbances were developing. HARE (1960) has analysed these disturbances by a time-section of the temperatures at 25 mb along the 80°W meridian (Fig. 54), showing waves of increasing thermal amplitude from November to the beginning of February. The sloping wave-pattern on the time section indicates that the disturbances have entered from the northwest, with the exception of the peak warming at the end of January, which probably has come from the west with maximum intensity near 60°N (see also Fig. 36).

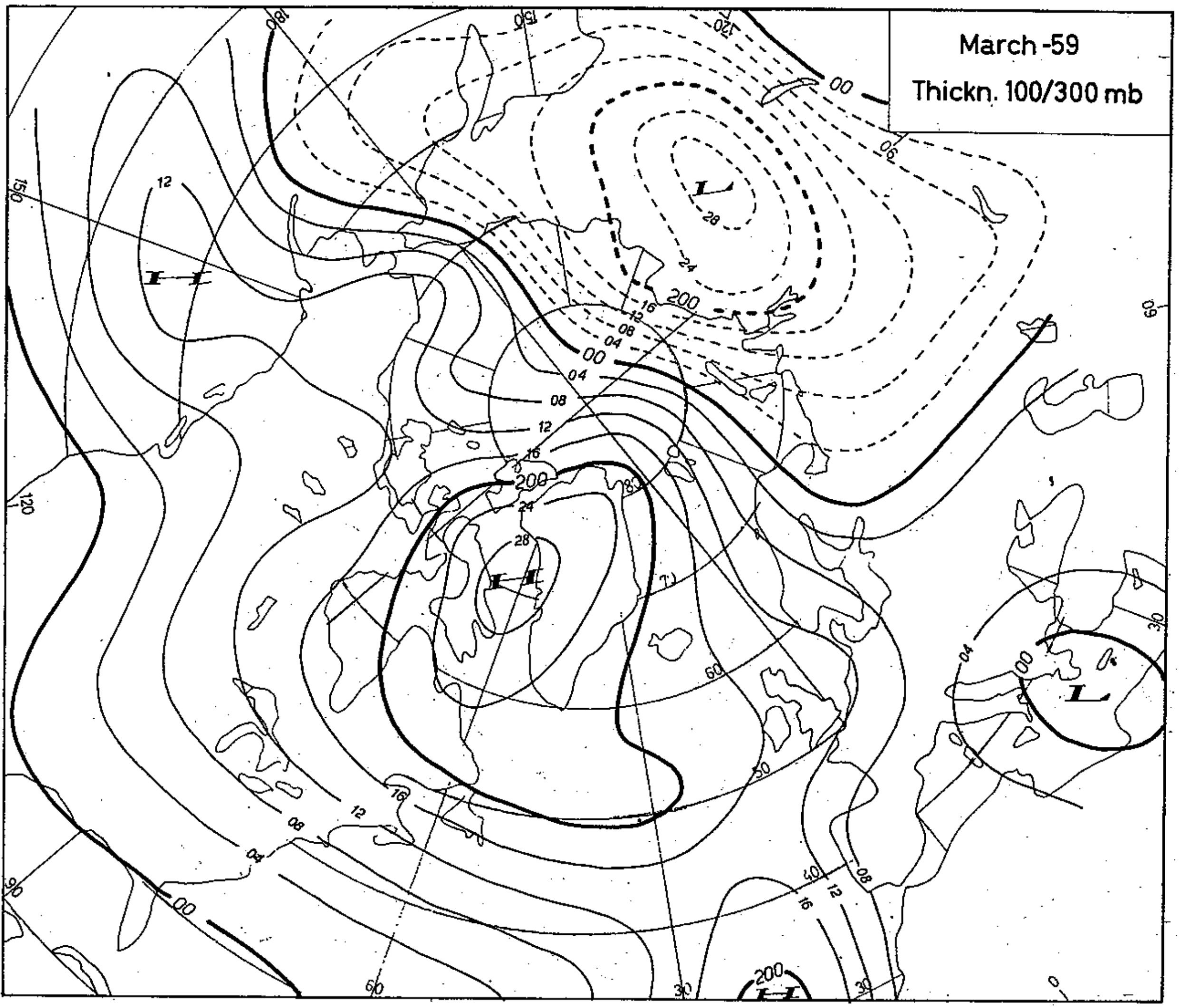


Fig. 52a. March 1959. Mean departures from the 5-year mean (1949—53) of the 100/300 mb thickness layer.

It seems reasonable that the strong stratospheric disturbances during this time have not only favoured and intensified the two tropospheric developments, which started as frontal waves on 14. and 21. January, respectively, near the Great Basin, but have also originated the tropospheric circulation changes in the area from Davis Strait to the Barents Sea during the last week of January.

From the time-section at 25 mb along the 80°W meridian, it appears that in the first half of January the stratospheric disturbances were strongest in the southern part, just in the areas generating the two tropospheric frontal waves which were the forerunners of the great tropospheric circulation changes, as described above (p. 64).

The stratospheric temperature oscillations from mid-January through the first part of February, shown in Fig. 54, are not only latitudinally moving waves, but also interpreting a marked meridional oscillation of the circumpolar vortex. At the time about 18.—19. January, when the warm ridge passed the 80°W meridian, the strato-

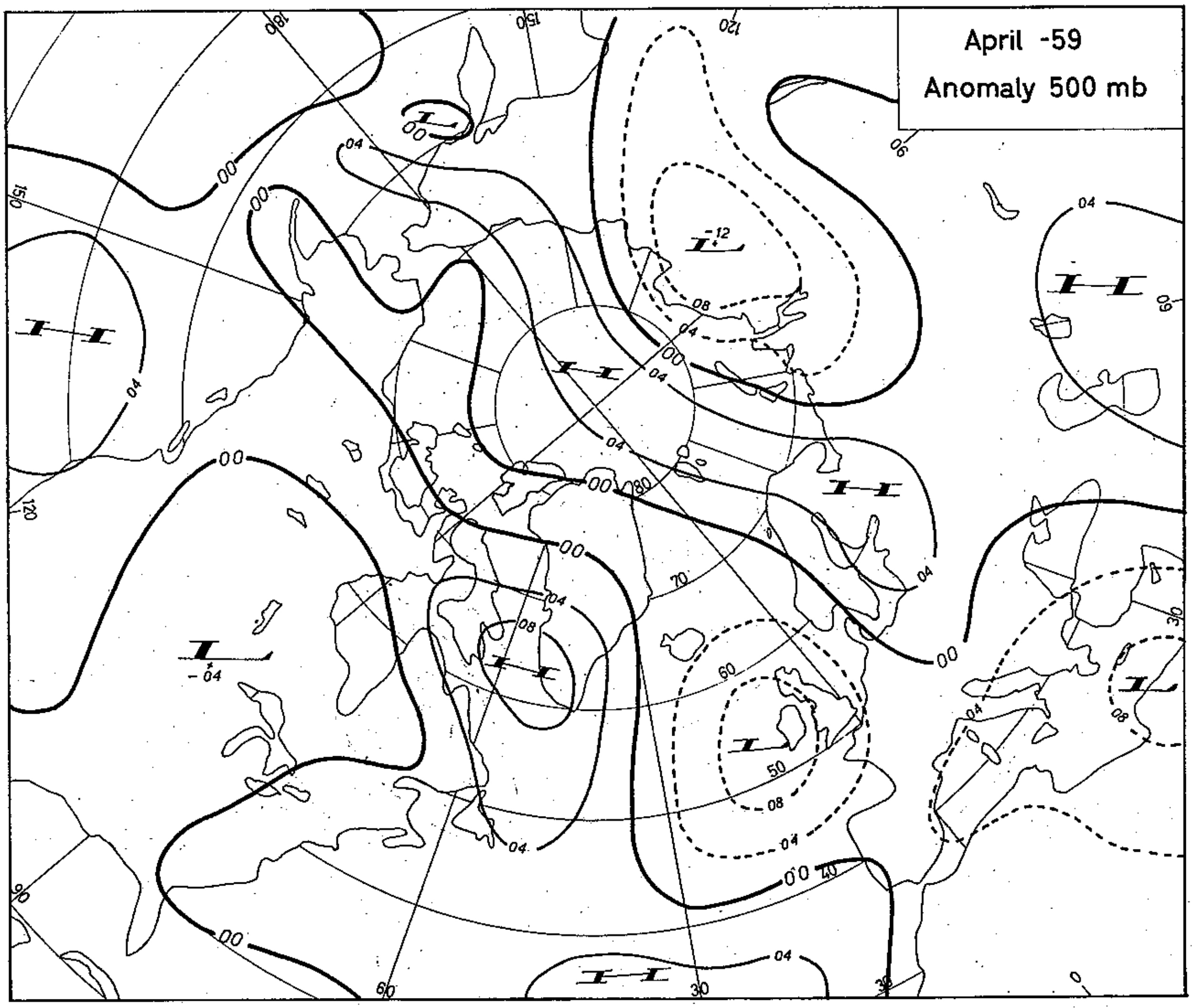


Fig. 52b. April 1959. Mean 500 mb height anomalies.

spheric cold core was situated over the European Arctic. The warm ridge, which had maximum intensity near 50°N, continued across the Northern Atlantic to Scandinavia, coincident with the displacement of the stratospheric cold core to the Alaskan side of the Arctic Basin, and also coincident with the tropospheric changes in the North Atlantic — European Arctic and sub-Arctic.

From thorough analyses only, can we ascertain the stratospheric priority of the tropospheric circulation changes in the Arctic during the last week of January 1959, but there is no doubt that pronounced co-operation existed between the stratosphere and troposphere over North America in January. This co-operation was displaced to the Arctic and continued through February.

The sudden and strong temperature rise at the end of January over North America was probably due to combined effects, in the same way as explained formerly (p. 43). The meridional stratospheric oscillation mentioned above was combined with changes of vertical velocities, and may have induced subsidence over the Pacific or the western

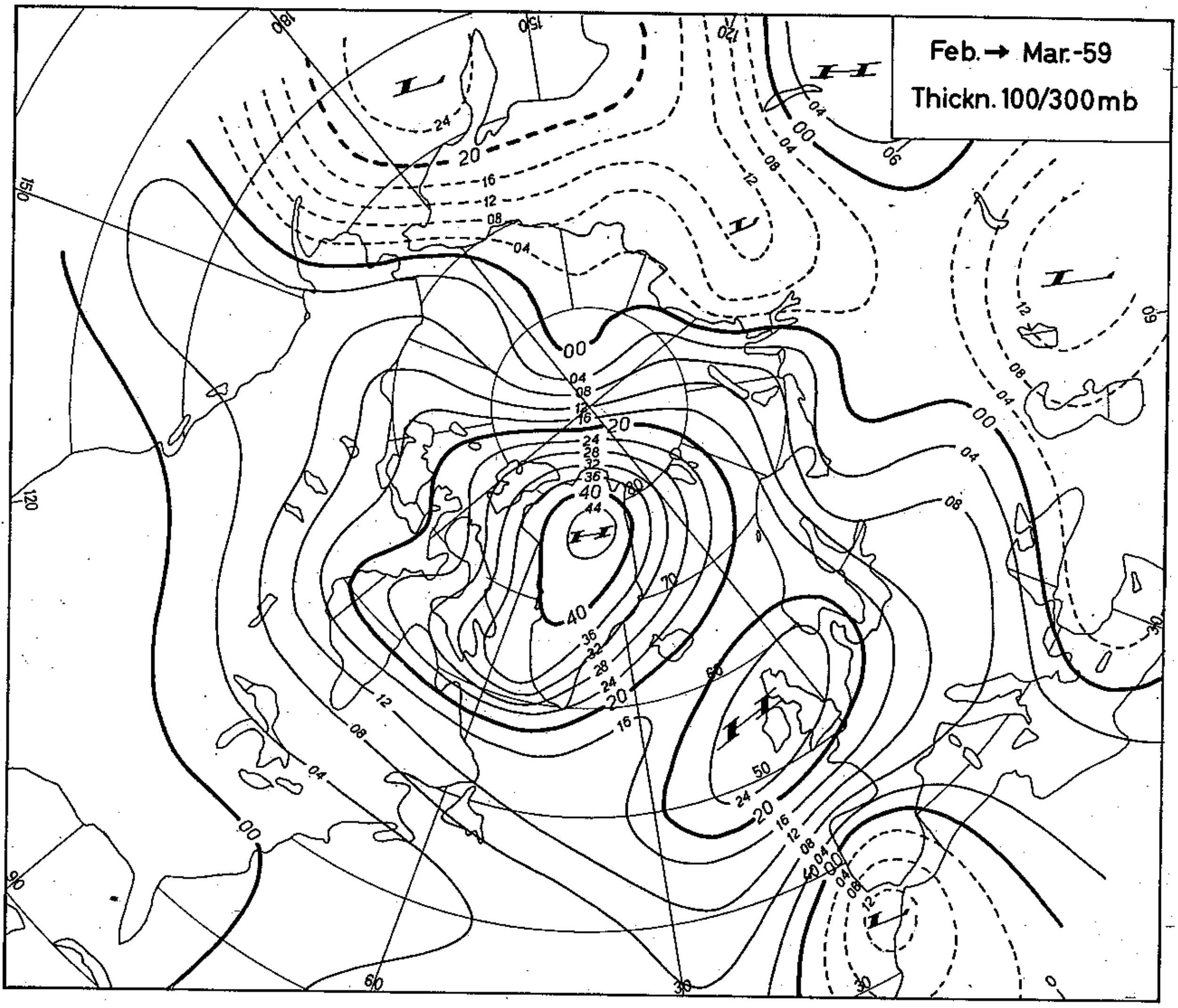


Fig. 53a. Change from February to March 1959 of the mean monthly 100/300 mb thickness departures from the 5-year mean (1949—53).

part of the U.S., intensifying the subsidence in a progressive wave. Fig. 34 presents a warm ridge near Kamchatka about mid-January. The warming effect may also have increased by heat absorption of the ozone. In the lower stratospheric levels, the warming was delayed and less strong.

The warm subsidence ridge, entering from the west, had maximum intensity near 60°N over North America, and stopped over the western part of the North Atlantic (Fig. 36). Fig. 54 shows that the stratosphere over North America south of 60°—50°N after the large changes, became nearly as undisturbed as in summertime. The disturbances continued, however, in the Arctic, especially in the Eurasian Arctic. Figs. 47 and 49b show that the stratospheric vortex strengthened in the Arctic in February.

The next sudden stratospheric warming occurred in the Canadian Arctic in the beginning of March. As pointed out by Hare (1960), the two impulsive warmings resembled one another closely. The last one, however, extended more pronounced to

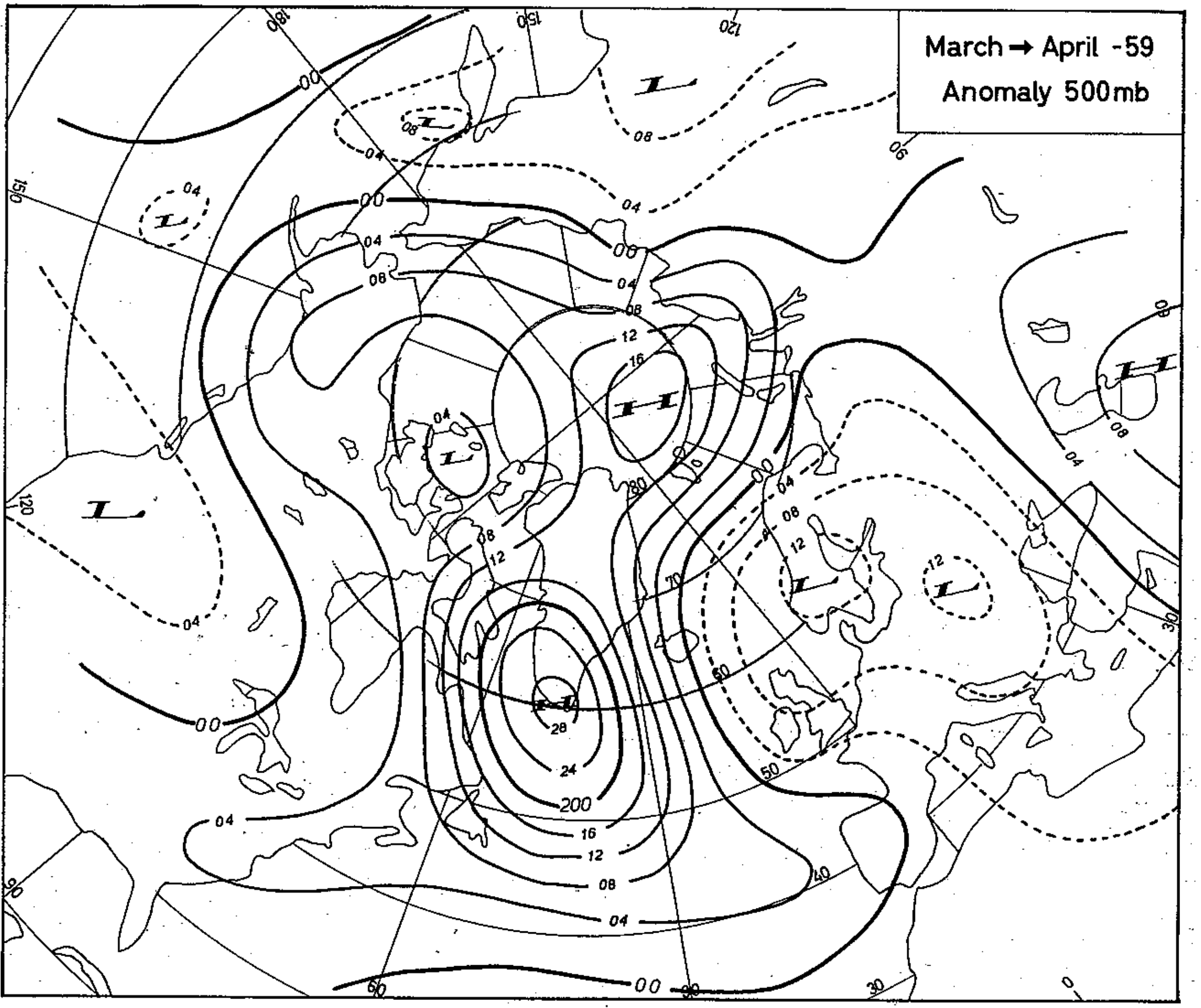


Fig. 53b. Change from March to April 1959 of the mean monthly 500 mb height anomalies.

the lower stratospheric levels and covered greater areas (Figs. 37, 52a and 53a). Not even the March-warming broke down the stratospheric circumpolar vortex. The vortex center at 100 mb was displaced to Mys Cheliuskin at the Siberian coast (Fig. 51).

Worth mentioning is that the cyclonic effect from the stratosphere on the troposphere appeared over Siberia in March (Fig. 50), but that the effect of the stratospheric warming on the tropospheric circulation was delayed until April (Figs. 52a and 53b).

5. Conformities and contrasts between the two winters, 1957—58 and 1958—59.

A comparison of the principal atmospheric occurrences in the Northern Hemisphere during the two investigated winters may illustrate a few features by the great stratospheric changes and their effects on the tropospheric circulation.

The autumnal stratospheric cooling in 1957 had no marked peculiarities, and no special stratospheric effects on the troposphere were to be traced. The September-October anomalies of the 100/300 mb thickness pattern indicated low zonal index at temperate latitudes, and this was in accordance with the tropospheric circulation. The stratospheric cooling in the Arctic during November was a little stronger than normal, and seems to have affected the troposphere (Figs. 12a and 12b).

In 1958, the stratospheric cooling in the Arctic during the late fall was marked non-zonal and this feature was later traced on the tropospheric circulation in a pronounced way (Figs. 41a, 41b and 41c).

During December 1957 and the most part of January 1958, the stratospheric cooling in the Arctic continued stronger than normal (Figs. 13 and 14). In December, the low-stratospheric and tropospheric anomaly patterns resembled one another to some degree in the Arctic, but absolutely not in January. Just the same features existed in December 1958 and January 1959. Really, it seems that both in January 1958 and 1959, the strong and cold stratospheric vortex center in the intensifying and nearly mature stage has increased the pre-existing anticyclonic anomaly in the arctic troposphere.

Otherwise, the mean stratospheric cooling patterns from November to January were rather different in the two winters. Fig. 5 shows an extended shape of the cooling pattern at 100 mb, while Fig. 35 presents a nearly circular distribution. In both winters, the areas of maximum stratospheric cooling were stronger than normal.

The writer is of the opinion that the mean stratospheric cooling pattern in the Arctic from late fall to mid-winter has had a deciding influence on the later progress of the strong stratospheric disturbances.

In January 1958, the circulation pattern prior to the breakdown of the winter vortex favoured the accumulation of disturbances in two areas on each side of the Northern Hemisphere, partly moving meridionally, and exploding in a pronounced climax just prior to, and under, their joining above the Arctic. A complexity of convulsion-like stratospheric changes followed, completing the breakdown of the stratospheric vortex.

The winter stratosphere 1958—59 was more frequently disturbed than in the previous winter, but the disturbances were moving nearly zonally round the Northern Hemisphere, without marked accumulation in mid-winter. A climax occurred over North America between 50° and 60°N at the end of January, but this event only confirmed the displacement of the stratospheric vortex center to the Eurasian Arctic as a lasting feature.

One peculiarity is worth mentioning. The mean tropospheric circulation in February 1958 had pronounced conformity with the mean tropospheric circulation in January 1959 (Figs. 21 and 45). This similarity is also pointed out by O'Connor (1959), based on the mean 700 mb maps. The two months, however, had quite different anomaly patterns in the stratosphere (Figs. 24a and 46a). In both cases, the following month, March 1958 and February 1959, respectively, obtained a tropospheric anomaly which was in accordance with the stratospheric anomaly pattern (Figs. 24b and 46b).

6. Conclusions. Examinations in this paper of the stratospheric influences on the troposphere show that extended forecasts comprising arctic to temperate latitudes, may be improved in wintertime by estimating the stratospheric occurrences in addition

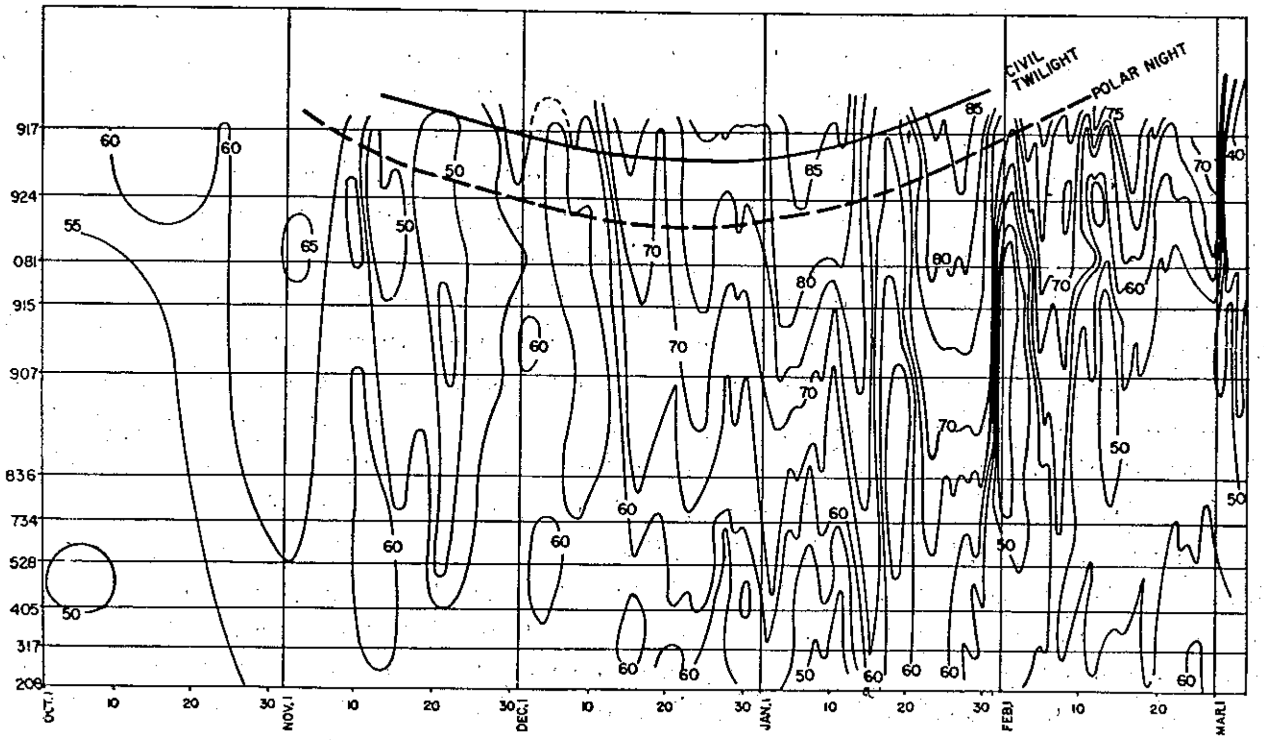


Fig. 54. Temperature of the 25-mb surface along the 80 W meridian from Charleston, S.C. (208) to Eureka, Ellesmere Land (917), as a function of time, 1 October 1958 to 4 March 1959. The dashed polar-night curve is the extreme northern limit of direct solar irradiation at 25 mb. Civil twilight curve is the line along which the solar elevation angle at noon is −11 deg. with respect to the 25-mb tangent horizontal plane. Note the strongly disturbed thermal field of winter, with waves of increasing thermal amplitude culminating in the impulsive warmings of 31 January and 2 March, 1959 (after Hare).

to the tropospheric developments. This improvement is meant to be a consequence, not only of the great stratospheric changes, but also of the co-operation between stratosphere and troposphere on a minor scale through the whole winter season, and partly also during the fall and spring.

In middle and high latitudes of the Northern Winter Hemisphere we may, in broad features, classify the atmospheric disturbances in three categories which are mutually influencing each other: (1) effects from the topography, (2) baroclinic tropospheric developments, and (3) baroclinic stratospheric waves. In this paper, the last term is applied to disturbances previously originating from (1) and (2) by transmission upward, but later transformed by the baroclinicity of the stratosphere and the lower mesosphere (and probably by ionospheric — upper mesospheric impulses, too). These disturbances are, in the terminology applied here, regarded as stratospheric features.

The investigations are simplified by using monthly mean values and their departures from the normal. The examinations only comprise effects in direction from the stratosphere to the troposphere. The large stratospheric changes are, of course, of great interest, but all long-period variations during the two winters are investigated as to the effects on the troposphere.

Two years is a short time for drawing any conclusions from empirical investigations, but when the results are logically affirmed they obtain a greater range of validity.

Experiences indicate that the clue to the understanding of the effects on the troposphere from the stratosphere may be based on a few principles. (1): The tropospheric circulation type has great persistence. (2): Strong impulses from higher levels may instantaneously influence and disturb all atmospheric layers below, but *transitory* impulses can contribute to the producing of longlasting effects on the tropospheric circulation only when the impulses harmonize with the pre-existing conditions in the troposphere. (3): When the circulation type of the troposphere is not disposed to the stratospheric impulse, however stable and longlasting, the case of slowly influencing boundary conditions will occur. A protracted influence from the lower stratosphere may gradually change the regional distribution of the tropospheric activity.

From the winters 1957—58 and 1958—59, the following figures present occurrences with stratospheric effects on the troposphere: Figs. 12a—b, 23a—b, 24a—b, 39a—b, 41a—b, 46a—b, 48a—b, 49a—b, 52a—b, and 53 a—b. Both cyclonic and anticyclonic impulses from the stratosphere are estimated. The material is somewhat scarce to decide any difference between the two types, but it seems that anticyclonic influences from higher levels have greater chances of being effective in the Arctic.

Even if the high-stratospheric warmings may be sudden and strong, the subsidence effects reaching the troposphere are damped and slow. A general effect of the stratospheric subsidence is a lowering of the tropopause. In the Arctic, the cold tropospheric air will be warmed up only a little, and due to the decreasing thickness, the cold air will cover greater areas, spreading and flowing southward. North Europe and West Siberia are regularly reached with cold tropospheric air by stratospheric subsidence in the Arctic. South of the edge of the cold tropospheric air, cyclonic developments will be favoured, increasing the zonal index at temperate to sub-tropical latitudes. During March 1958, these described occurrences reached Mid- and South Europe, and lasted through April.

The stratospheric circulation obviously can not produce energy in the troposphere. Stratospheric disturbances, however, may incite the troposphere in certain areas and contribute to the release of energy. Studying the atmospheric circulation during January 1959, it seems that such stratospheric incitements may have been active on the troposphere over North America, and the later, nearly explosive cyclonic development in the troposphere in the North Atlantic—European Arctic has clearly been encouraged by the stratospheric flow and temperature field.

During October and November 1958, cyclonic and anticyclonic developments in the stratosphere were co-operating, apparently equivalent in each area, and generated pronounced effects in the troposphere. There was little time lag and the effects were long-lasting.

It appears from what is said above, that the time lag involved in the stratospheric-tropospheric changes must be varying. Monthly mean maps are not well suited for such examinations, the maps including both time lag and persistence, but the material

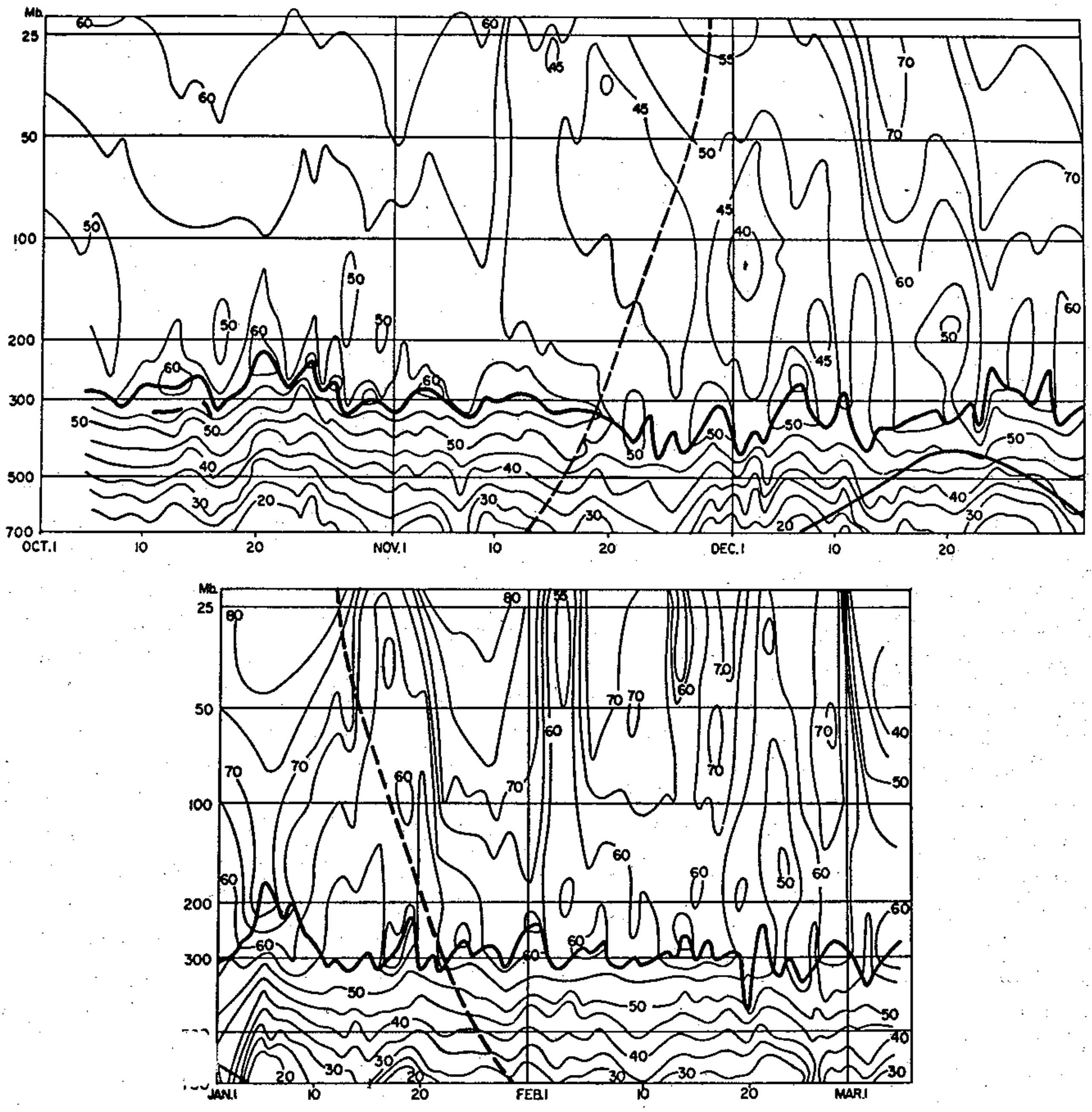


Fig. 55. Time-height section of temperature (deg. Celsius) at Resolute (924), 6 October—5 March, 1958—59 winter, from 700 to 25 mb. Dashed curve is limit of polar night (sun below horizon). Solid line is civil twilight (solar elevation —6 deg. with respect to the apparent horizon). Heavy black line is arctic tropopause. Note that the impulsive temperatures are of either sign (after Hare).

shows that in most cases the 100/300 mb thickness anomaly pattern is followed by a similar 500 mb anomaly in the *next* month. Probably, cyclonic incitements from the stratosphere generally have great time lag to the troposphere in the Arctic.

Experiences show that the stratospheric influence on the tropospheric circulation appears effective and decisive only in certain places, and for isolated periods. This is also a consequence of the previously mentioned principles. When the stratosphere and the troposphere have joined in circulational co-operation, we may expect an extreme tropospheric circulation type of long duration.

The regional distribution of the stratospheric influence on the troposphere in the winter season is estimated to comprise the cap of the Northern Hemisphere down to about 50°N. Occasionally, the effects may reach farther south towards the sub-Tropics, but this is judged to be secondary and chiefly as an circulational adjustment to the happenings in the north.

When estimating the stratospheric influence on the tropospheric circulation, cases with *simultaneous* effects on the lower stratosphere from the topography and tropospheric developments must be excluded. As to the topographic effect, the Himalaya Mountains are dominating, and especially in December and January when the jet stream is strongest. The Himalaya effect regularly covers the North Pacific and the Aleutian area, where meridional-vertical circulations are simultaneously forced, both in the troposphere and the stratosphere but of opposite sign. Occasionally, this forced effect may also comprise greater areas, in first line East Siberia and Canada.

Over the central Arctic in December and January, during the intensification of the stratospheric vortex, there seems to be an anticyclonic influence on the troposphere. This effect is explained from the flux northward involved in the stratospheric vortex, which branches in subsidence to the troposphere and ascends towards the lower mesosphere. The effect, however, must also be estimated in relation to the time lag of cyclonic incitements from the stratosphere.

An other feature of the great stratospheric changes in wintertime is worth mentioning. In spite of the fact that strong warmings occur in temperate and sub-tropical latitudes at high levels, the final result of the great warmings in the lower stratosphere is chiefly found in high latitudes. These occurrences must primarily be due to the effect of the topography, especially the Himalaya Mountains. The effect, however, may also be related to the great changes in the regime of vertical motions, combined with the breakdown of the circumpolar vortex, and with subsidence being established in the Arctic throughout the whole stratosphere and mesosphere. A contributing effect is also the general low tropopause at high latitudes.

In 1955 and 1958, the breakdown of the stratospheric circumpolar vortex occurred early (in January), but the vortex was re-established somewhat later in the winter season. In both years, the changes after the spring equinox occurred only slowly and the summer conditions in the arctic stratosphere were not reached before May. This indicates that the action in the regular rapid breakdown of the winter vortex after the spring equinox is perhaps more dynamical than radiational.

When a stratospheric subsidence ridge is influencing the vortex in the lower mesosphere in the Arctic, a chain of rapid changes and disturbances may occur: Displacement and gradual filling of the vortex center near the mesopeak level is due to subsidence and advection of warmer mesopeak air, combined with meridional-vertical oscillations in the mesosphere. Density changes in the lower mesosphere, combined with variations of the heat absorption by the ozone, may also contribute to the temperature changes.

The baroclinic stratospheric waves are the tool bringing the subsidences in the lower mesosphere — upper stratosphere down to lower atmospheric layers. The writer has previously pointed out (p. 46) two different procedures, which are called A(ridge)-type and B (trough)-type, respectively, and which are active by the conversion of kinetic energy into potential energy.

As to the ionospheric-upper mesospheric disturbances, the writer is of the opinion that *transitory* impulses from the mesosphere can have effective circulational influence on the stratosphere only when the stratospheric disturbances in advance are running consistent with the impulses from above.

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