THE DISTURBED CIRCULATION OF THE ARCTIC STRATOSPHERE [1,2]

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(Manuscript received 6 April 1959)

ABSTRACT

The annual cycle of disturbed circulation in the arctic stratosphere (*i.e.*, areas north of the main jet-core of the Ferrel westerlies) is discussed in the light of climatological and synoptic evidence. In summer, the Ferrel westerlies and their wave-trains choke off with height to a zero-level near 50 mb. Because of the warmth of the arctic stratosphere, easterly anticyclonic flow is continuous at higher levels. In August, cooling begins, and by September a barotropic westerly vortex, continuous with the Ferrel vortex, is established. In October and November, this vortex becomes baroclinic and appears distinct from the Ferrel westerlies. Winter circulation has two aspects: (1) warm, barotropic, anticyclonic flow associated with a strengthening and extension of the Alaskan warm ridge, which extends to above 25 mb; and (2) cold-low development near the pole, with a marked tendency for the development of a cold trough over eastern Canada. Polar-night jet-streams and travelling baroclinic waves characterize the outer parts of the cold-lows. The thermal waves associated with these disturbances often have large amplitudes and normally affect the entire stratosphere. The apparent independence of stratospheric disturbances centered above 50 mb is examined, and the probable physical and dynamical processes involved are tentatively discussed.

1. Introduction

Much interest has been aroused in the past few years by the winter circulation of the arctic stratosphere. With the post-war expansion of the northern rawinsonde network, it has become possible to extend climatological and synoptic analysis to levels as high as 25 km (25 mb approximately). This welcome development has shown that the higher isobaric surfaces, especially north of the mid-latitude jet, display an annual régime very different from what one might have inferred from the stratospheric literature of 1950 and earlier. It has emerged that the arctic stratosphere has a very disturbed climate, which displays, in the sudden warming episodes, one of the most dramatic of all types of atmospheric instability.

The sudden warmings have naturally attracted much attention in the literature. First described by Scherhag (1952) seven years ago, they have recently been the subject of excellent case-studies by Godson and Lee (1958), Craig and Hering (1959), Teweles (1958), and Godson (1959). Their great interest has tended to divert attention from broader aspects of the stratospheric behavior. There have been very few attempts at an effective statement of the characteristic annual climatic cycle at these levels. Wexler and Moreland (1958) and Hare and Orvig (1958) have made a start in this direction, but the scarcity of data makes both papers tentative.

[1] The research reported in this article has been sponsored by the Geophysics Research Directorate, Air Force Cambridge Research Center, under Contract No. AF 19(604)-3865.

[2] A condensation of this paper was presented at the 175th National Meeting of the American Meteorological Society held in Chicago, Illinois, 24–27 March 1959.

The writer's objective here is to attempt such a statement, though necessarily at the crudest level of synthesis. Two main sources of material were available to him—(1) rawin and radiosonde summaries for stations scattered through a wide range of longitude and as close to the pole as Alert (82N), and (2) daily synoptic charts and cross-sections at levels up to 25 mb. In view of the great interannual variability at these levels, it seems essential to combine synoptic experience with climatological data if reasonable inferences are to be made from either.

2. Sources of data and terminology

The climatological data employed in the paper consist of (1) upper-wind summaries (from rawin data) prepared by the Air Weather Service and the United Kingdom Meteorological Office, and (2) temperature averages compiled from numerous sources. It is well-known that both types of data contain many errors at the higher levels. No attempt will be made to assess these errors here in detail; recent discussions by Chiu (1958), Reiter (1958), and Teweles and Finger (1959), make the limitations of the data very clear. In brief summary, the main weaknesses are as follows:

(a) The wind parameters are affected by (i) inherent observational errors, including data transmission errors, probably 2 to 3 m sec^{-1} root-mean-square at 20 to 30 km and by (ii) early termination of balloon ascents, which tends to weight all parameters towards the values they usually assume in warm, low-wind conditions.

(b) The temperature values (and determination of

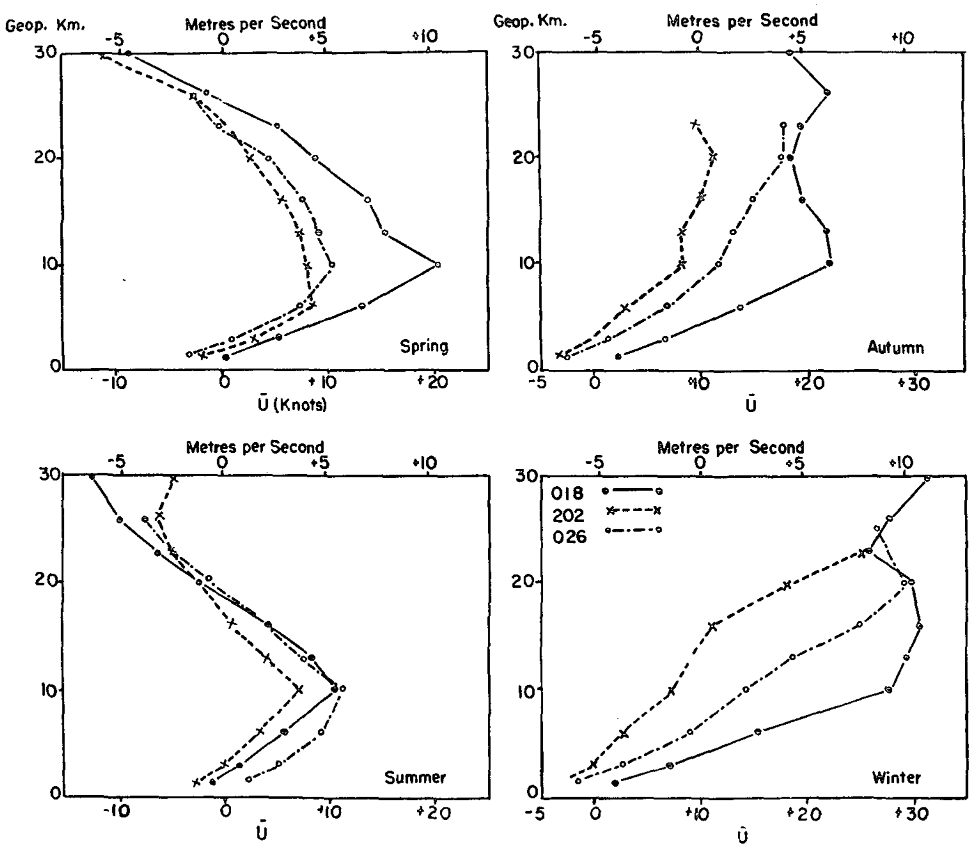

FIG. 1. The zonal component of seasonal resultant observed wind ($\bar{u}$) as a function of height, Barrow, Thule, and Keflavik. From Air Weather Service rawin data tabulations. Size of sample decreases seriously above 16 km.

heights) are affected by random instrumental error, and by systematic radiation errors, which are only partially compensated by applied corrections. United States duct-type equipment shows, at certain seasons, a considerable spurious diurnal variation even after correction, at levels above 100 mb. Selective termination of balloon ascents also weights the published temperature averages in the same sense as that discussed for wind; above 50 mb, the final source of error is prohibitive in winter.

In the circumstances, the writer will make greater use of wind data than of temperatures. As Chiu (1958) has recently shown for more southerly latitudes, it is possible to construct reasonably good pictures of the mean flow at the higher levels from such data.

The synoptic charts and cross-sections used here were drawn under the writer's direction by the Arctic Meteorology Research Group at the Central Analysis Office, Meteorological Branch, Canadian Department of Transport. Since 1 June 1958, daily analysis for 0000 GCT has been carried out as follows:

(a) 200-mb and 100-mb charts for areas north of 30N between the date-line (in mid-Pacific) and the 50E meridian. Since 1 December 1958, the 200-mb analysis has been circumpolar.
(b) 25-mb charts for the North American data transmission area (roughly as in fig. 3).
(c) Cross-sections to 25 mb along the 80W alignment from Charleston (208) to Eureka (917) and along a transarctic path from Whitehorse (964) through Sachs Harbour (051) to Alert (082) and Nord (310).

The group also has access to circumpolar 100-mb analyses prepared by the Arctic Forecast Team of the Meteorological Branch at Edmonton, Alberta. Much use has also been made of the well-known United States Weather Bureau circumpolar 100-mb and 50-mb analyses for 1952–53. General discussions of the problems of stratospheric analysis have been prepared by Muench (1958), the Air Weather Service (1953), and Moreland and Cluff (1955). A report on the methods employed in the project *will* be published elsewhere.

Difficulties concerning data collection and the correction of radiation errors have so far prevented our extension of 25-mb analysis outside the North American sector, except in isolated cases of deliberate historical analysis. Climatological data are also scarce above 16 km (100 mb) on the Russian side of the pole. At the higher levels, therefore, the evidence presented is confined to a restricted sector, a serious limitation in view of the dominance of long-wave systems at these levels.

In the following account, the writer will employ standard usage as follows:

tropopause to 20 km—*lower stratosphere*
20 km to 30 km—*middle stratosphere*
above 30 km—*upper stratosphere*

Apart from the recognition of a middle layer, this usage follows Goody (1954). The layer between 20 and 30 km (roughly 50 to 10 mb) contains the ozone peak and is the site of the main disturbance activity in cooler seasons; it also contains the summer stratospheric easterlies. Hence, its separate recognition as a middle layer is convenient.

Finally, the writer will use the term *Ferrel westerlies* to denote the circumpolar vortex of the troposphere and lower stratosphere. The westerlies that appear in winter at higher levels and closer to the pole (Kochanski, 1955, especially fig. 1) will be called the *polar-night vortex*. The two systems are distinct both climatologically and on the daily map. They are distinguished on fig. 14.

3. The summer circulation

The writer has discussed the summer circulation elsewhere. It is well-known (Kochanski, 1955) that the arctic stratosphere is warm and almost isothermal ($T \sim -40$C) between late May and mid-August, with temperature gradients directed equatorward at all levels. This régime is ushered in rapidly on widely different dates from year to year. After mid-August, rapid cooling destroys both the warmth and the high-level easterlies that dominate the circulation of the middle stratosphere. There is no true spring, since the summer régime usually begins impulsively. For the sake of convenience, however, the familiar terms are both retained in this account.

The main characteristics of the summer circulation are as follows:

(a) anticyclonic easterly circulation above a level between 15 and 20 km (*i.e.*, in the middle stratosphere) round a center near the geographical pole;
(b) a progressive upward decrease in strength of the Ferrel westerlies, from the level of their maximum intensity at the arctic tropopause to the zero-level at the base of the middle stratospheric easterlies;
(c) a virtual absence of truly stratospheric disturbances. The thermal field between 250 and 100 mb is disturbed by the wave systems of the Ferrel westerlies; above 100 mb, the temperature régime seems almost free of short-period oscillations, and the flow is almost wholly barotropic.

Since the present paper is concerned with disturbed circulation, and since the summer circulation is treated elsewhere, no further discussion will be presented here (Hare, 1960).

4. The autumnal cooling

Strong cooling begins in the 20- to 30-km layer in early August, extending to the lower stratosphere by the end of the month (when daily cooling of the order of magnitude of 1C in three days is typical). By the beginning of October, temperatures have often fallen near the pole to -60C at 25 mb. The cooling affects all latitudes north of 40N.

The height falls associated with this cooling quickly destroy the easterly anticyclonic circulation of the middle stratosphere. By late August, the synoptic chart at 25 mb and 50 mb is completely disorganized. Very rapidly, however, as the cooling continues, a westerly circumpolar vortex appears at the levels near the pole and spreads laterally to lower latitudes. The early appearance of these upper westerlies—long before, in fact, the ozone layer is in darkness—seems to be one of the more reliable climatological facts about these highly variable upper layers. The autumn (September–November) profiles of zonal wind component (fig. 1) mostly show (a) an increase to a strong westerly maximum at tropopause level and (b) fairly strong but steady circumpolar flow in the layer 13 to 26 km.

Initially, this vortex is barotropic. The stratosphere in mid-September is usually quasi-isothermal from the pole to 30N through a considerable depth. Fig. 2 gives a time cross-section of temperature of the 25-mb surface along 80W for the autumn of 1958. It shows clearly that the entire surface was quasi-isothermal in mid- and late September. The 200-mb surface is also quasi-isothermal at this time of year, but the 100-mb surface retains an equatorward temperature gradient related to the extremely cold tropical tropopause layer at that level south of about 40N. The perturbations of the jet-stream belt, and the highs and lows of the polar cap, retain the characteristic temperature fields of summer—*i.e.*, cold ridges and warm troughs in the lower stratosphere. The combined effect of these influences is (a) to permit the Ferrel circumpolar vortex to extend barotropically to the middle stratosphere, probably to well above 30 km, with low wind speeds and a much lower meridional shear than at 200 mb; (b) to remove all the strongly baroclinic short-wave disturbances of the lower stratosphere below 20 km; and (c) to permit the long-wave structure

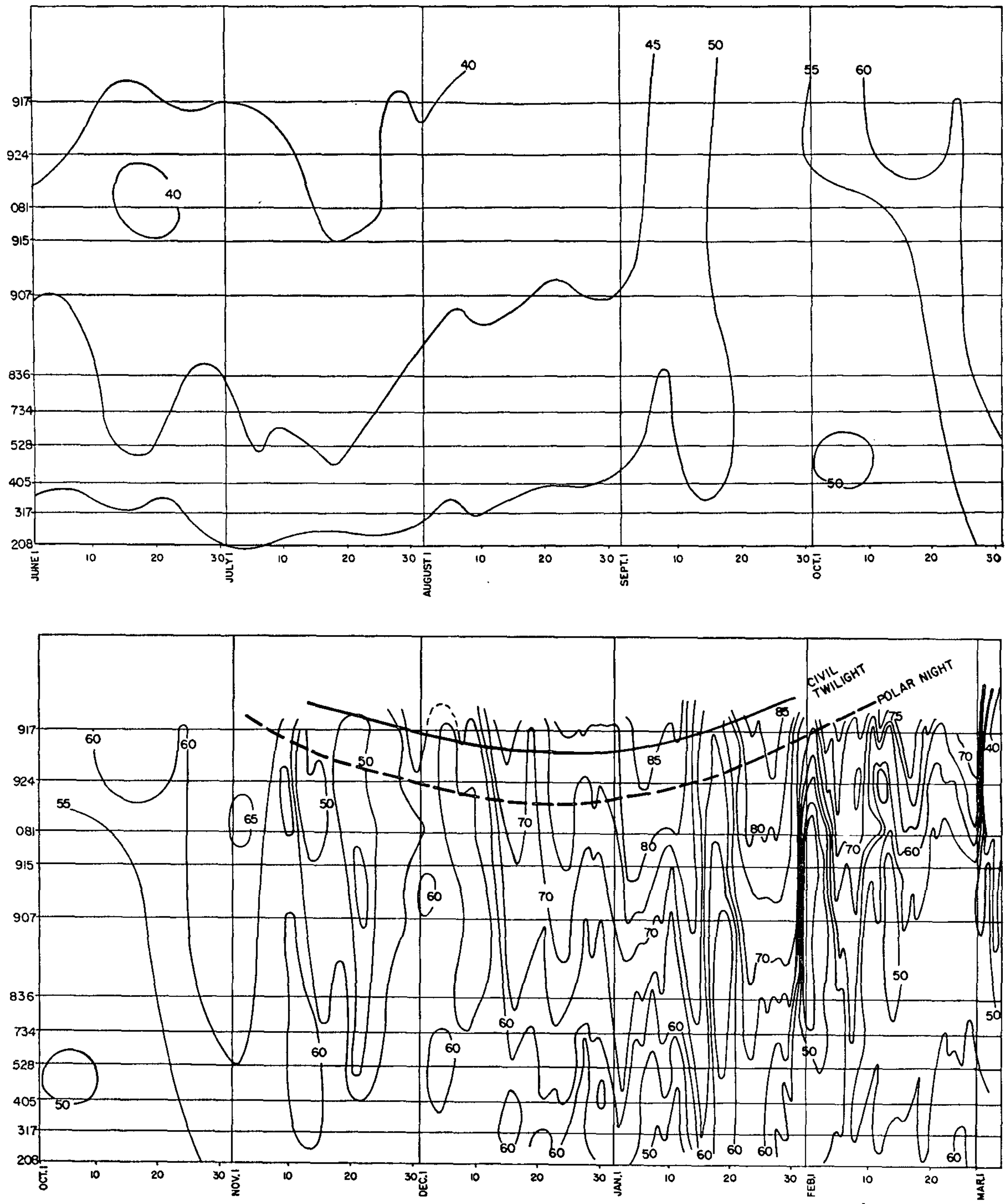

Fig. 2. Temperature of the 25-mb surface along the 80W meridian from Charleston, S. C. (208) to Eureka, Ellesmere Island (917), as a function of time, 1 June 1958 to 4 March 1959. The dashed polar-night curve is the extreme northern limit of direct solar irradiation at 25 mb. Civil twilight curve is the line along which the solar elevation angle at noon is −11 deg with respect to the 25-mb tangent horizontal plane. Note (i) the conservative, equatorward gradient of summer; (ii) isothermal conditions in September; (iii) 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.

of the Ferrel westerlies to extend through the entire stratosphere in a subdued form. At this time, then, the Ferrel vortex and polar stratospheric vortex are continuous and inseparable.

In 1958, the polar stratospheric vortex became slowly more baroclinic in the six weeks following 15 September, due to cooling of the middle stratosphere. Temperatures fell below −60C at 25 mb in lat 80N in early October (see fig. 2) and then stabilized on the Canadian side of the pole. A weak, though definite, poleward temperature gradient existed by mid-October. The westerly circumpolar flow continued essen-

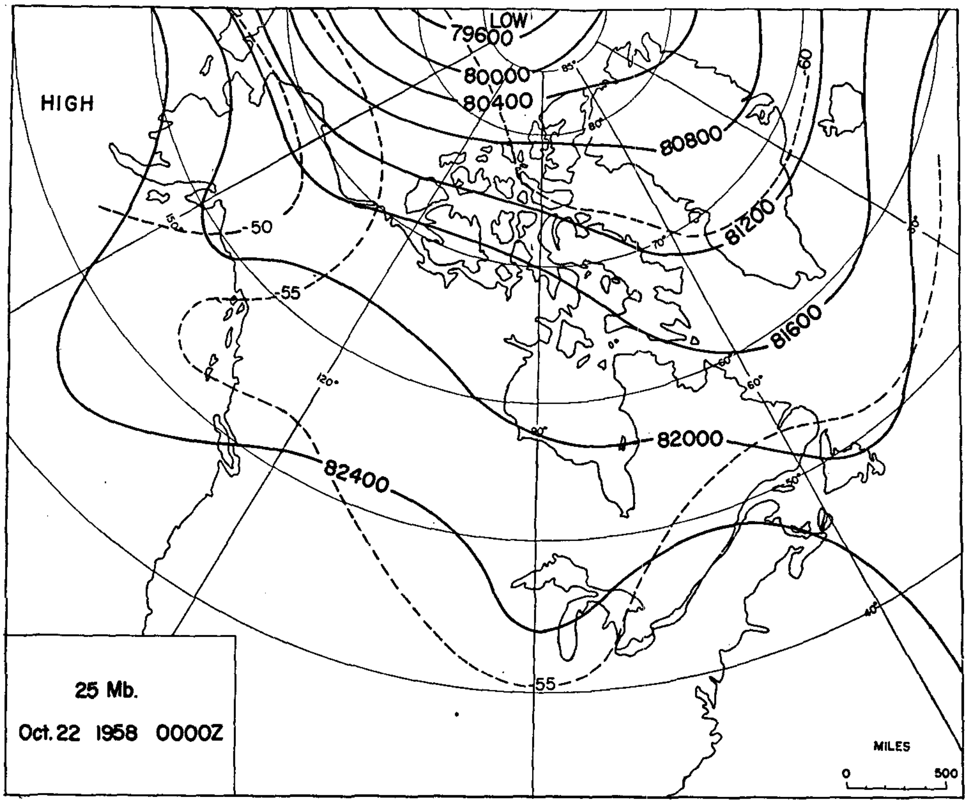

FIG. 3. 25-mb chart for 0000Z 22 October 1958, at main phase of warming over Alaska. The thermal field so established was very persistent.

tially without disturbance above 100 mb, with a slight increase in strength. Fig. 4 indicates that the polar temperature field from 100 mb upwards usually remains comparatively stable for several weeks after the resumption of westerly flow at these levels. This result implies that the early autumnal circulation (above the disturbed lower layers) is undisturbed because it is barotropic and that the flow continues without disturbance until some critical value of baroclinicity or meridional shear is attained (Hare and Orvig, 1958; Godson and Lee, 1958). In 1958, the onset of the characteristic winter-type disturbance was very sudden, beginning with a rapid warming over Alaska in mid-October, with accompanying anticyclogenesis, and a simultaneous shift of the cold polar vortex towards Europe and western Siberia. Hence, it is possible to test the baroclinic hypothesis.

Fig. 3 shows the 25-mb map for 0000 GCT 22 October, when the strong Alaskan warming was established. It is apparent that the initial warming took place far west of Alaska, and a considerable part of the rise of temperature was due to simple advection. North of Alaska, the anticyclonic shear on fig. 3 can be estimated as being of the order of 3×10^{-5} sec^{-1}, a value that had been exceeded several times in the previous two weeks without apparent disturbance of flow. The horizontal poleward temperature gradient was about 10C per 1000 km (10^{-7}C cm^{-1}). Moreover, even this very moderate degree of baroclinicity was as much the result as it was the cause of the sudden warming, for most of the previous two weeks gradients had not exceeded 5C per 1000 km.

Thus, the breakdown of stability in the 1958 vortex took place at a time of only very moderate baroclinicity and meridional shear of the westerlies. It seems unlikely that the observed shear and baroclinicity can have exceeded any reasonable stability criterion. Hence, the onset of disturbance must either be referred to a much modified dynamical theory, or else the disturbance must be assumed to have had its origin over Siberia or northern Europe. It will be noted in fig. 4 that the dispersion of 100-mb temperature rises rapidly at Barrow in October, whereas at Thule, 80 deg of longitude further east, the rise to winter levels is deferred into November, in spite of the fact that Thule's values later in the winter are much larger than Barrow's. Hence, the 1958 pattern and the climatology of 100-mb temperature variations both indicate an initial breakdown of stability over northern Siberia and a subsequent eastward movement of the effects.

5. The midwinter régime

The mean charts for all stratospheric levels in midwinter (fig. 5, December–February) show strong and

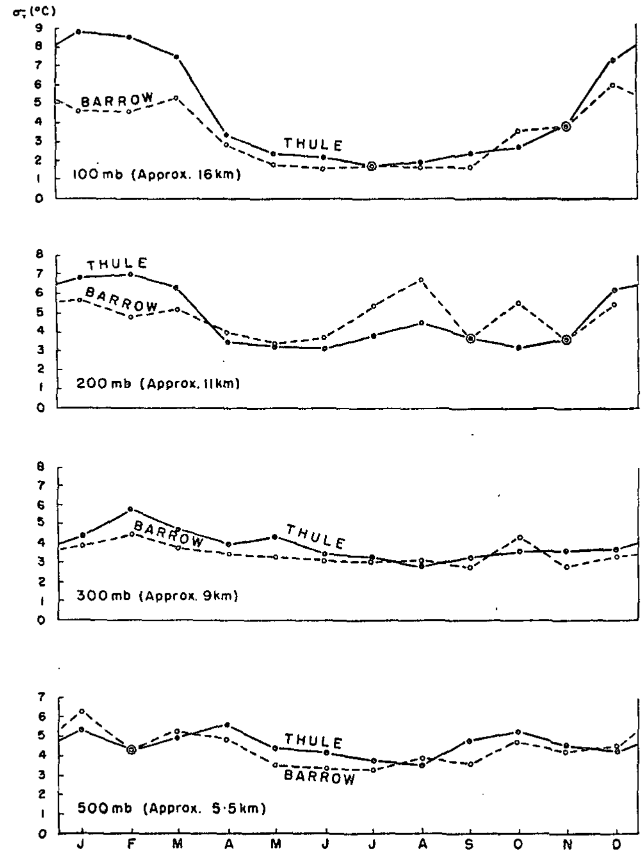

FIG. 4. Seasonal variation of dispersion of observed daily temperature values, 1951–55, Barrow and Thule, in mid-troposphere (500 mb), near the tropopause level (300 mb), basal stratosphere (200 mb) and lower stratosphere (100 mb). Parameter is the standard deviation (σ_T) in deg Celsius. Computed from grouped data published by Tolefsen (1957).

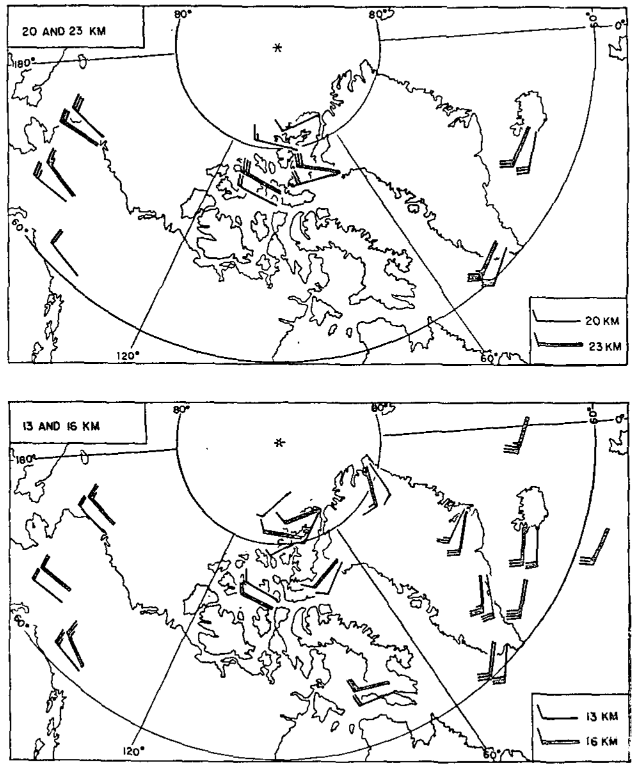

FIG. 5. Resultant observed winds at selected stratospheric levels, winter season, from rawin-data tabulations by the Air Weather Service. Weather Ship data are for January and were compiled by the U.K. Meteorological Office.

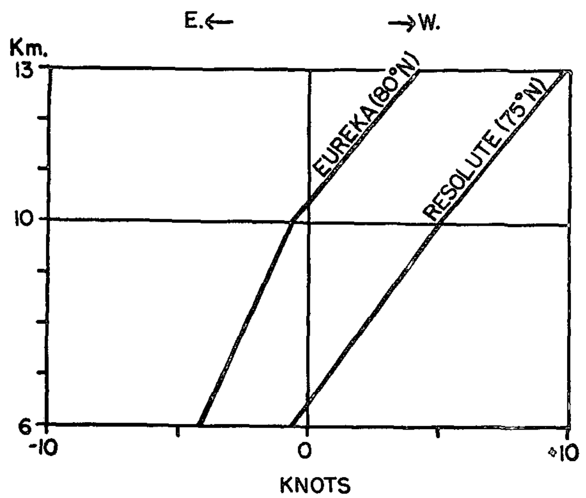

FIG. 6. Profiles of winter mean zonal wind component ($\bar{u}$) at Resolute and Eureka, to illustrate the poleward shift of the westerly vortex center in the stratosphere.

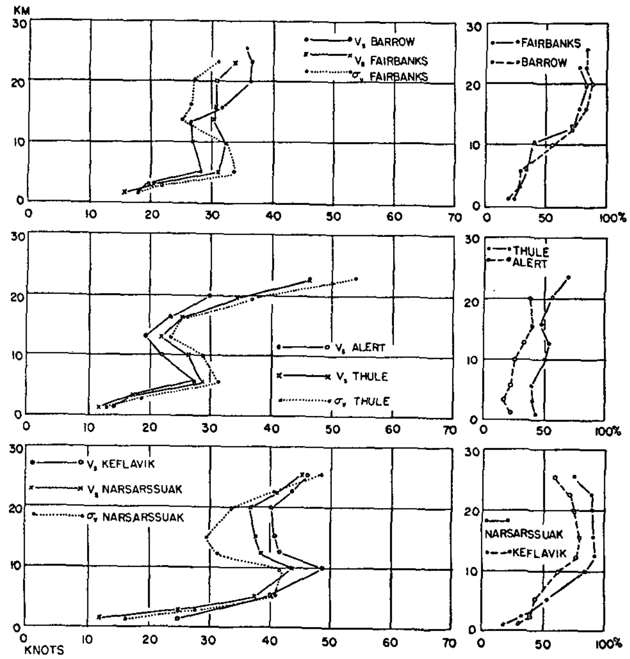

FIG. 7. Mean scalar wind speed (V_s) and constancy ($100\ V_r/V_s$, where V_r = resultant wind) as a function of height, winter season. For selected stations, vector standard deviation (σ_v) is also plotted.

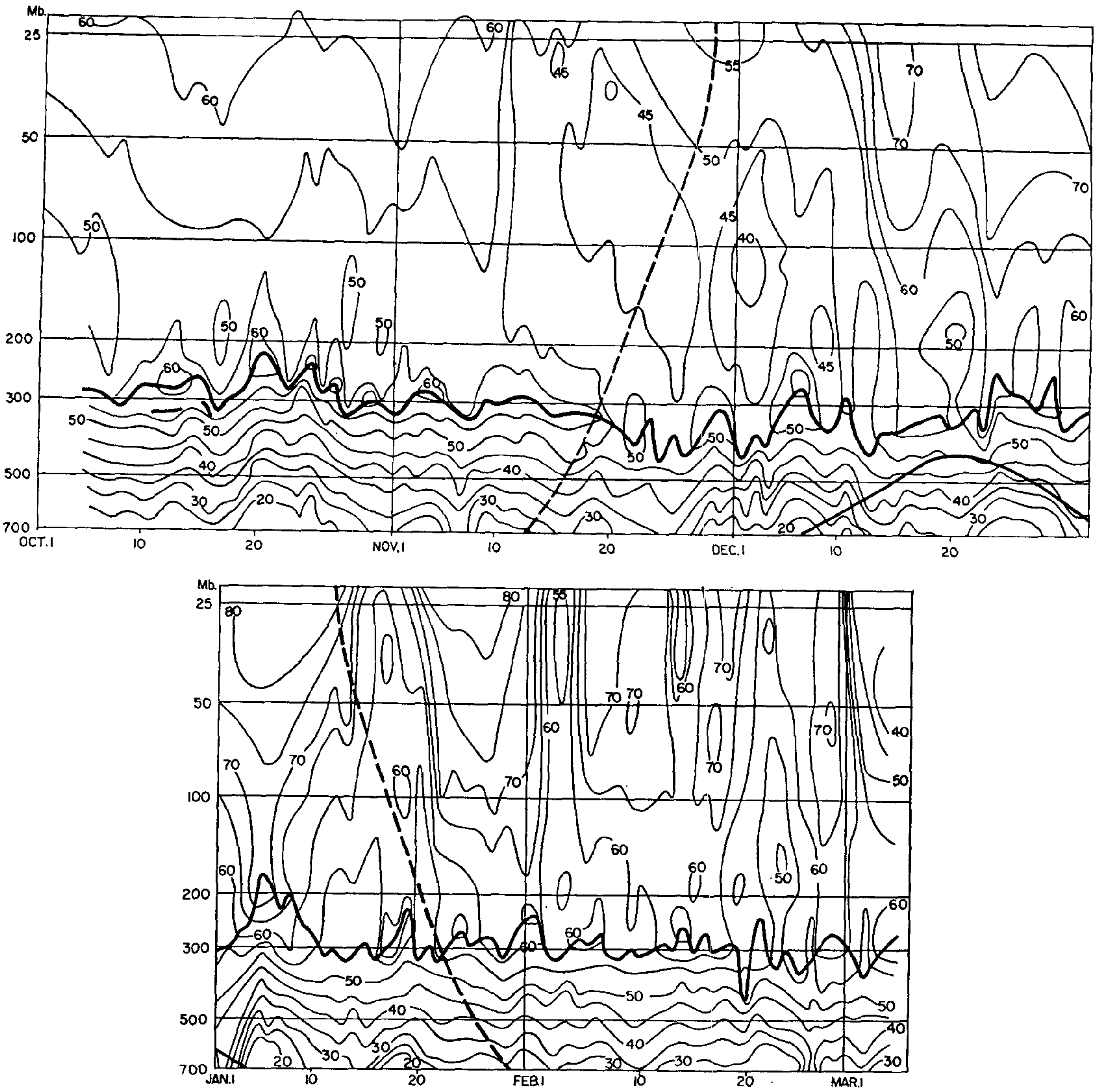

FIG. 8. Time-height section of temperature (deg Celsius) at Resolute (924), 6 October–5 March, 1958–59 winter, from 700 to 25 mb. Dashed curve is limit of polar night (sun below horizon). Solid line is civil twilight (solar elevation: −6 deg with respect to the apparent horizon). Heavy black line is arctic tropopause. Note that the impulsive temperature changes are of either sign.

persistent westerlies round the polar-night vortex, which appears at first sight to continue upwards the Ferrel vortex. The latter, however, is multiple-cored, with centers over eastern Siberia, Cape Chelyuskin, and arctic Canada. As fig. 6 shows, the Canadian center shifts polewards in the upper troposphere, and at stratospheric levels there seems to be a single center near the pole.

The highly unstable nature of the polar-night westerlies is well shown by fig. 4, which gives the seasonal variation of the standard deviation of temperature at Barrow and Thule. At 200 mb, variability is high throughout the year, because of vertical motion associated with travelling waves in the Ferrel westerlies. At 100 mb and 25 mb, however, the curves show two distinct seasons. From October until March, dispersions are very high, especially at Thule. In April, there is an abrupt fall to very small values, which persist through the following summer. It is thus clear that changes due to advection or vertical motion (or both) are of large amplitude within the polar-night circulation. These inferences from the thermal field may be compared with the evidence of upper-wind climatology. Fig. 7 gives vertical profiles of scalar wind speed and constancy for selected arctic stations in winter (December–February). Zonal wind profiles have already been presented in fig. 1. From the two figures, one can generalize as follows:

(a) Scalar wind speed is a maximum near the tropopause and again in the middle stratosphere or above. Constancy increases rapidly in the upper

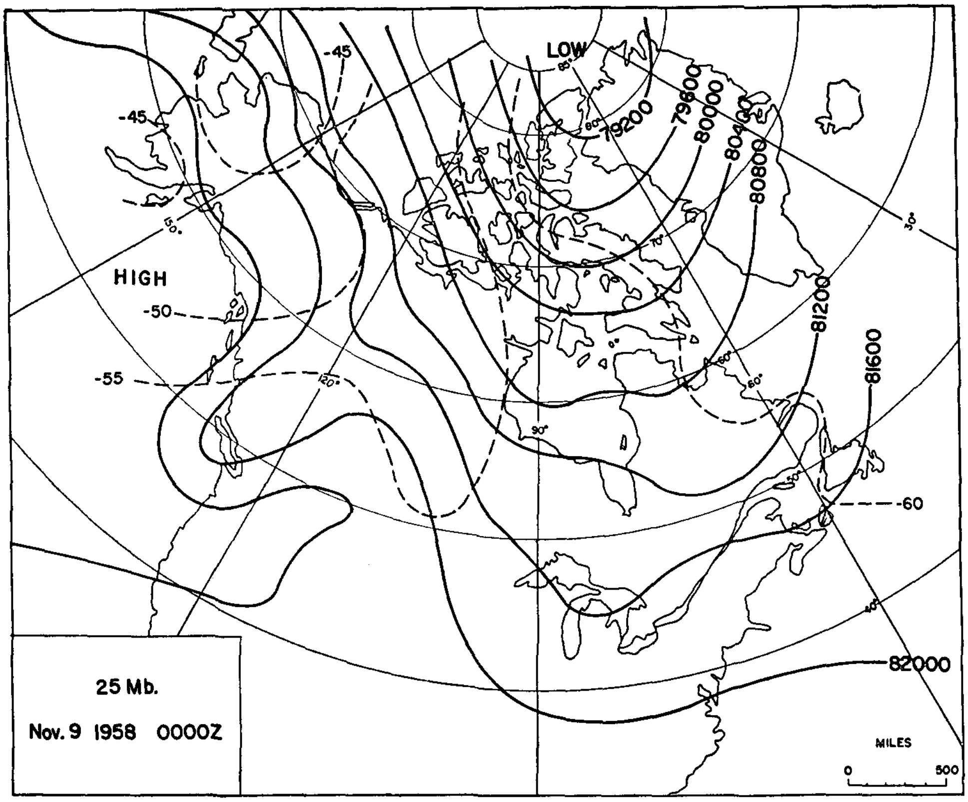

FIG. 9. 25-mb chart for 0000 GCT 9 November 1958, at beginning of impulsive warming at Resolute (924).

troposphere and is very high (70 to 80 per cent) in the lower stratosphere, with some decrease above.

(b) The zonal component of the resultant wind increases uniformly with height throughout the lower stratosphere at Barrow and Thule, close to the pole, but there is a deep barotropic layer in the lower stratosphere at Keflavik, in lat 60N. No tropopause-level maximum occurs.

The climatology thus indicates a rather stable westerly flow which nevertheless permits an extremely high variability of temperature.

In this paper, attention will be focussed on the following aspects of the winter circulation: (a) the depth, geographical extent, and time-scale of the observed temperature changes, both positive and negative; and (b) the apparent independence of the middle stratospheric motion systems. In particular, it will be shown that the sudden warmings have counterparts in sudden coolings, almost equally spectacular. The case studies will be drawn from the 1958–59 winter.

Fig. 8 shows a time cross-section for the period 6 October 1958 to 5 March 1959 for Resolute (924), from 700 mb up to the 25-mb surface. It was constructed from observed values, supplemented by interpolated values at 25, 50 and 100 mb from synoptic analysis of the temperature field. Even at 25 mb, the ratio of interpolated values to observed values is below 25 per cent, Resolute being one of the most reliable of the arctic rawinsonde stations. Fig. 2 presents, for a longer period, a 25-mb time-section along the 80W meridian from Eureka (917) to Charleston (208).

These cross-sections show that the winter of 1958–59 fell into the following three distinct phases: (a) a period of stable, slightly baroclinic westerly flow from October 6 until November 9, with weak disturbances (amplitude 2C to 3C at 50 mb) in a layer from 150 mb to above 25 mb; (b) a warm, anticyclonic phase from November 9 until December 10, ushered in by an impulsive warming; and (c) a midwinter cold phase from December 11 onwards. The cold of winter (and the polar-night vortex that it accompanies) thus came into being with extreme irregularity. There was no gradual generation of a cold polar stratosphere, as might be expected from a radiative hypothesis. Godson and Lee (1958) have presented results showing that irregularity is the rule rather than the exception.

The temperature waves of figs. 2 and 8 were visible throughout the cooler months, but they grew in amplitude as the winter progressed. The following properties manifested themselves:

(a) Power spectrum analysis of the 25-mb- and 100-mb-temperature series at Resolute indicated that most of the variance was associated with a sixteen-day period, with a smaller contribution from thirty-two days. A two- to three-week period thus seems typical of the record, shorter-period changes being negligible; the 500-mb and 300-mb series for the same period (within the Ferrel levels) showed no such preference for long-period oscillation.

(b) The geographical range of the disturbances (fig. 2) increased to a maximum in early January and then decreased rapidly; at their maximum extent, they were attaining 35N.

(c) The thermal waves (fig. 8) at Resolute normally extended through the entire stratosphere, but were not obviously related to tropospheric changes. Thermal amplitudes were greatest at the highest levels.

(d) In detail, the waves appeared as impulsive changes of temperature (of over 30C in 48 hr at their most intense) of either sign, interspersed with long periods of stability. Impulsive coolings were as frequent as the celebrated warmings but were normally less rapid.

These sections strongly suggest the existence of a distinct suite of stratospheric disturbances centered at levels above 20 km. The existence of such disturbances has already been inferred by Austin and Krawitz (1956), Scherhag (1958), and Wexler and Moreland (1958). Synoptic evidence for the distinctness of these systems will now be presented.

We may begin with phase (b), marked by the abrupt warming at Resolute on November 9 to 11. For a week previously, the thermal field had remained almost static—warm (> -50C) over Alaska and the Mackenzie Valley, cold (< -60C) over Ellesmere and North Greenland. Pressure had been high south of Alaska and low over the pole, so that a west to north-west flow, moderately baroclinic in structure, was established over arctic Canada.

Figs. 9 and 10 present 25-mb maps for 9 and 11 November, at either end of the main warming at Resolute. On November 9, a cold trough (below -65C at the floor) passed Resolute, moving south to a line between Frobisher and Pittsburgh by November 11. Behind the trough, the north-westerlies soon backed over the Canadian islands and Alaska, where temperatures were about -45. In this current, the isotherms advanced eastwards, the -50C isotherm passing Resolute on the 10th. At first sight, this looks like an advective process, but at no time did the motion of isotherms exceed about 60 per cent of the observed- or geostrophic-wind components normal to them.

The warm epoch of November and early December was accompanied by the development of a strong, warm ridge south of Alaska at the 25-mb level. The two peaks of the warmth—on November 20–21 in the 25- to 50-mb layer, and on December 2 (in the darkness of the polar night) between 175 and 50 mb (chiefly near 125 mb)—were marked by strong ridges extending north and east from the Alaskan focus. The later warm phase, culminating on 2 December, was more spectacular. A closed anticyclone was centered west of Sachs Harbour (051)—see fig. 11—at 25 mb, and even at 100 mb the flow was still weakly anticyclonic. Highest temperatures on this date were near the 100-mb surface over the eastern Canadian arctic (-35C to -40C).

On 11 December, the 25-mb map still showed (fig. 12) a warm ridge extending north-eastwards across arctic Canada from a center somewhere south of Alaska. In the previous 48 hr, however, a strong cold trough had approached Alaska from the west; south-westerlies of 45 to 70 kn extended from the Bering Sea to the Queen Elizabeth Islands, and some fall of temperature had occurred. In the next 48 hr, the trough ushered in the polar-night vortex by moving rapidly east, lying roughly through Thule (202), Frobisher (909), and Moosonee (836) by the 13th. At the same time, the vortex center itself moved rapidly towards Canada (having been located east and north-east of Greenland for the previous month). Between 11 and 12 December, Alert's 50-mb height (measured) dropped 1400 ft, with an estimated 25-mb drop of 1830 ft. On fig. 16, the height difference at 50 mb between Whitehorse (964) and Alert (082) is plotted. It increased from 1050 ft on December 10 to 3510 ft on the 12th and 4760 ft on the 14th.

The cross-sections for 13 December (figs. 14 and 15) show the corresponding generation of strong zonal circulation at middle stratospheric levels. Both sections show that the flank of the polar-night vortex was essentially a narrow system of strong west-north-westerlies with a jet core well above the 25-mb surface. The isotachs were concentric with this core and were clearly independent of the systems at tropopause level (the more familiar wave systems of the Ferrel vortex). On both sections, the two disturbance-levels were separated by a quasi-barotropic layer of minimum specific kinetic energy at 100 to 200 mb.

The 25-mb map for 18 December (fig. 13) may be regarded as the culmination of this initial cooling. The vortex center was now south of Frobisher (909), and the strongest circulation was a north-northwesterly jet in the strong baroclinic zone west of the cold, barotropic core extending from North Greenland to northern Québec.

6. The late winter instability

The polar-night vortex established over the North American sector of the arctic in mid-December of 1958 persisted until the beginning of March, when an im-

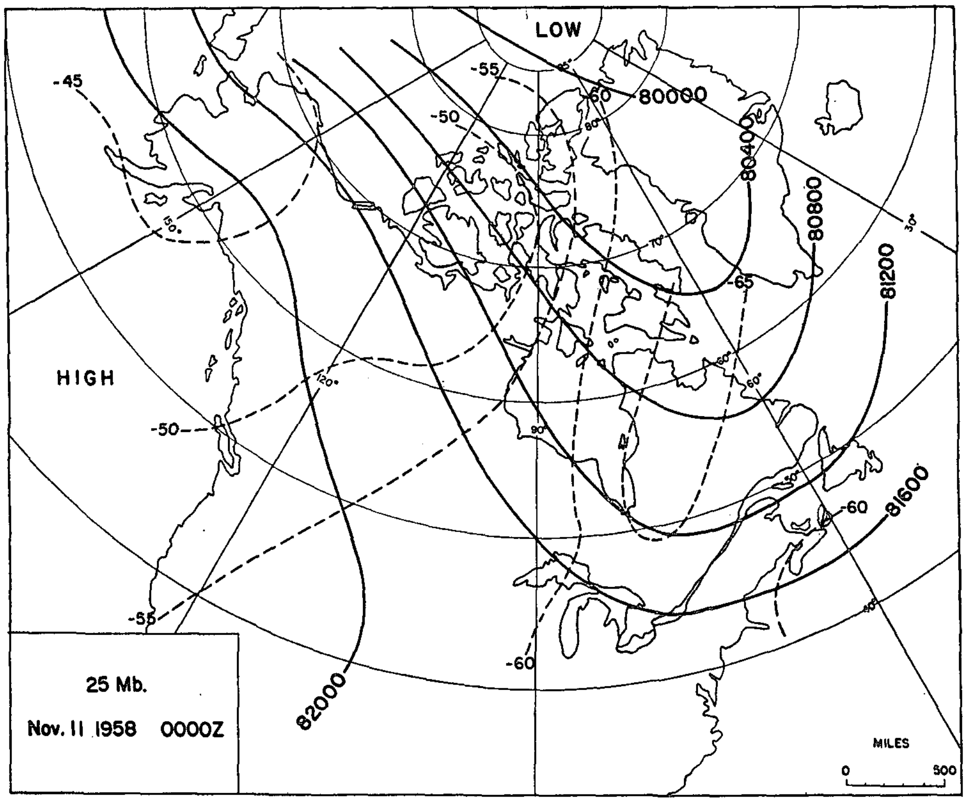

FIG. 10. 25-mb chart for 0000 GCT 11 November 1958, after warming had passed Resolute.

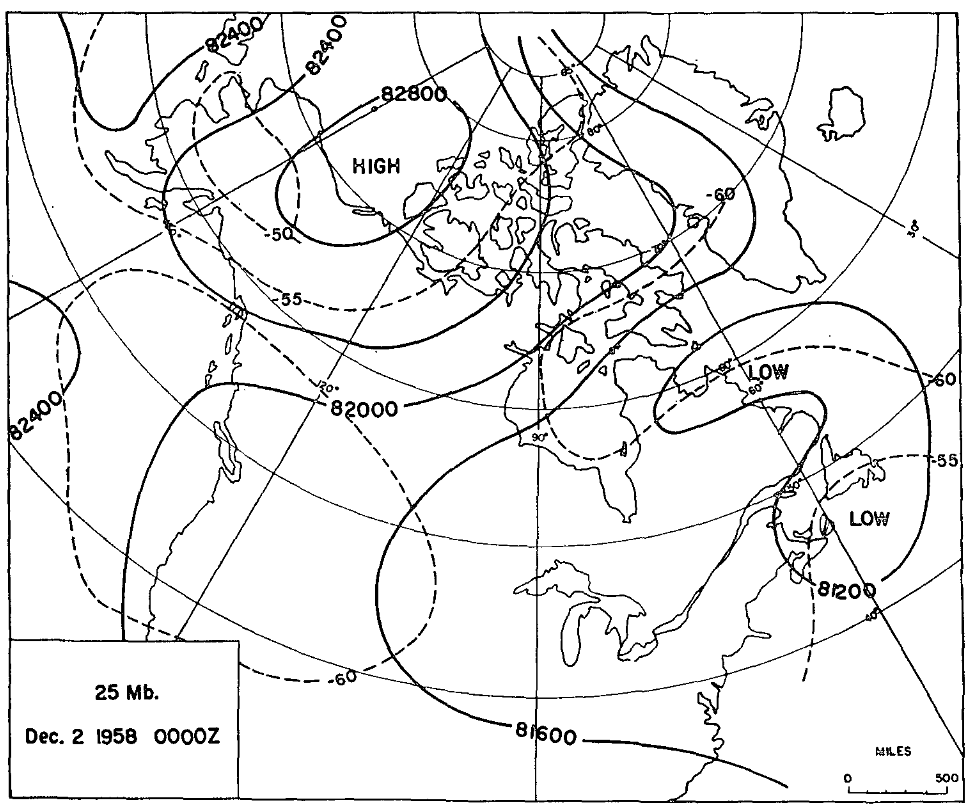

FIG. 11. 25-mb chart for 2 December 1958, at climax of early winter warm phase. This map is typical of the middle stratosphere during warm periods of winter.

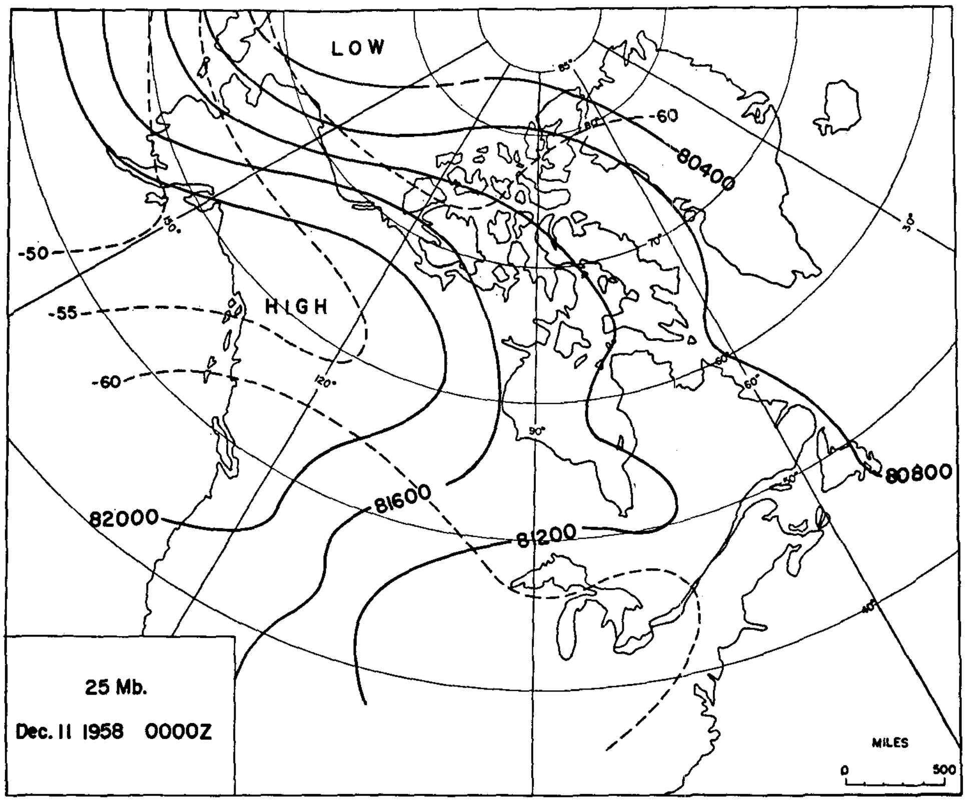

FIG. 12. 25-mb chart for 0000 GCT 11 December 1958, at onset of first impulsive cooling of 1958–59 winter. The trough west of Alaska moved rapidly eastward and deepened. In the following 24 hr, Alert's (082) height fell an estimated 1830 ft.

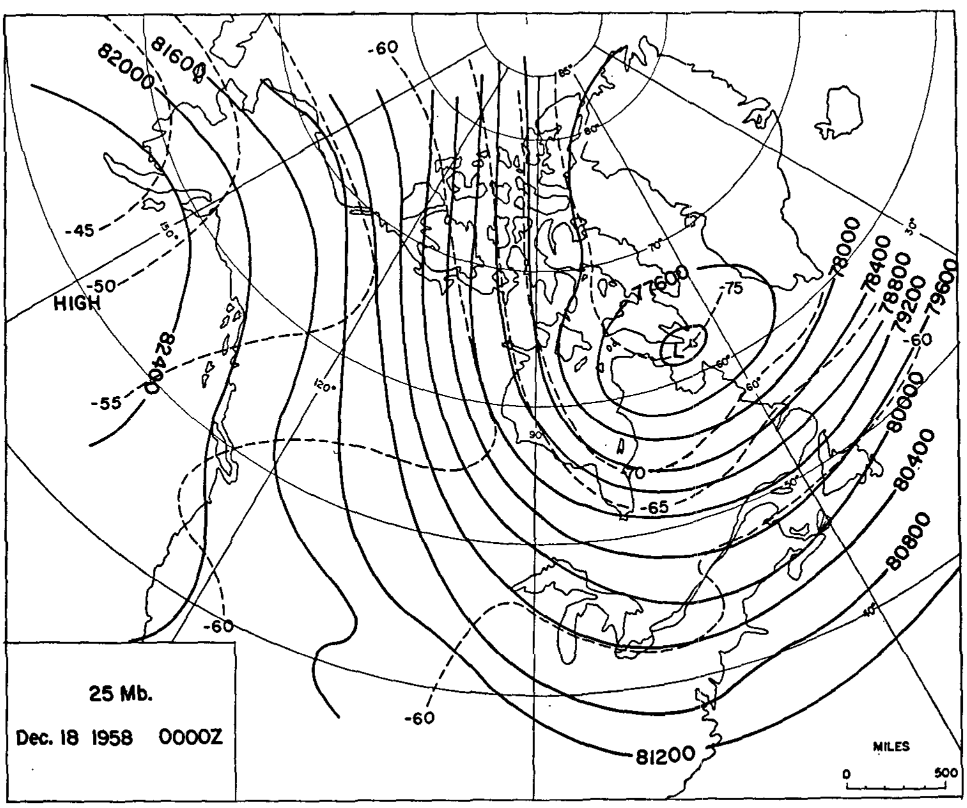

FIG. 13. 25-mb chart for 0000 GCT for 18 December 1958, with fully established polar-night vortex. The warm ridge over Alaska and cold trough over eastern Canada are quasi-permanent features of the vortex.

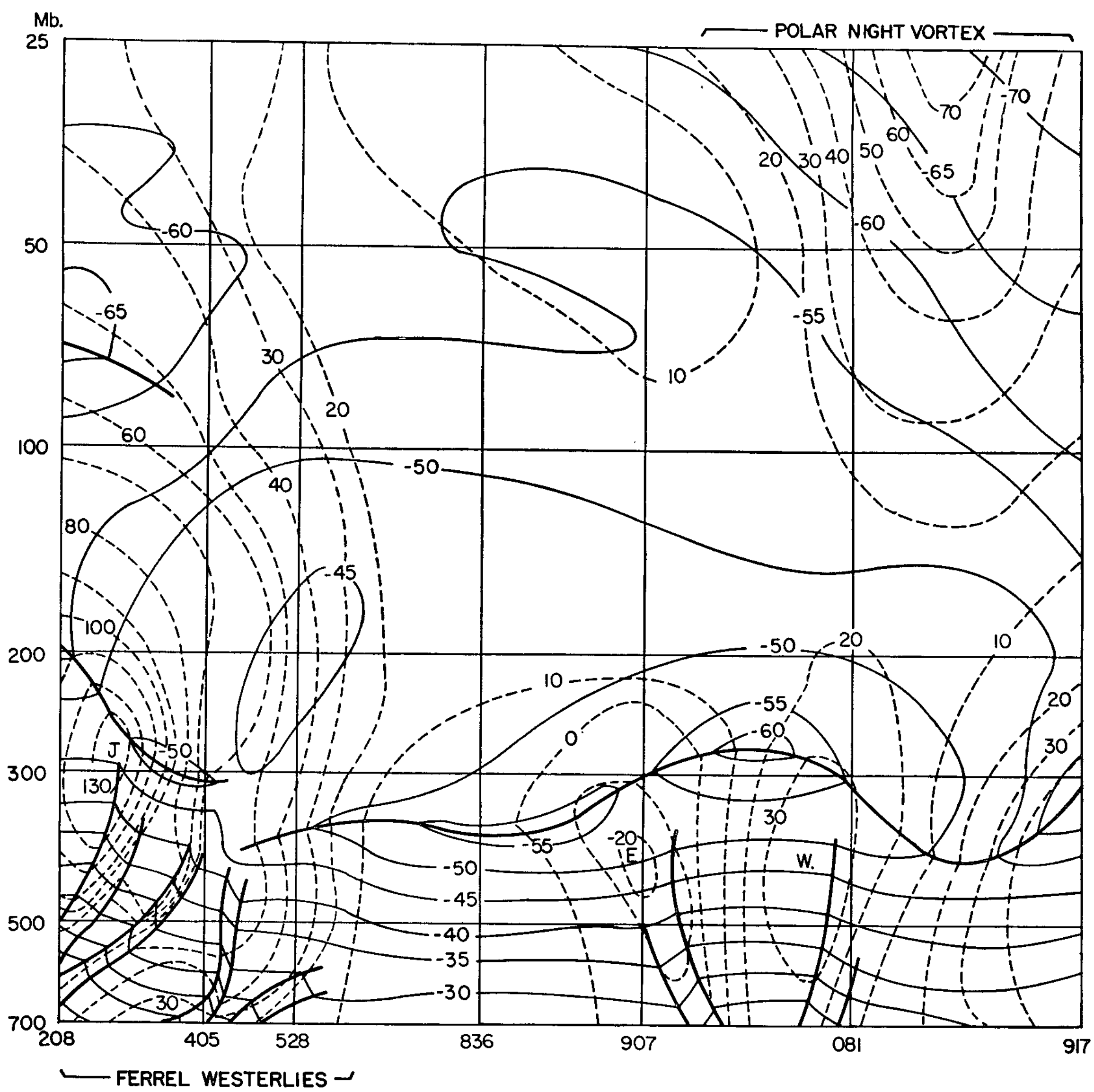

FIG. 14. Cross-section from Charleston (208) to Eureka (917) along 80W meridian for 0000 GCT 13 December 1958. Isotherms (deg Celsius)—thin solid lines, isotachs (knots) of zonal wind component—dashed lines, tropopauses and fronts—heavy lines. Note polar-night jet associated with developing cold of middle stratosphere over Canadian arctic.

pulsive warming over the Canadian arctic drove it across the pole into Siberia and northern Europe. It is well-known that a similar breakdown occurs every winter, though the date and pattern vary greatly.

As fig. 2 shows, 25-mb temperatures near the vortex center (which seems to have been close to Alert (082) for much of the winter) fell below −85C in January. The remarkable thing about fig. 2 at this time is the marked instability of the entire thermal field. Temperature waves of a range of 15 to 20 deg Celsius occurred at 15- to 20-day intervals. Each of these waves affected the entire vortex down to Charleston (208) in December and January. In February and early March, however, the waves changed markedly in character as follows:

(a) The geographical amplitude became more restricted; the warming of 31 January–1 February, for example, was hardly effective at Buffalo (528) and those of later February barely attained 60N.

(b) The thermal amplitude became much greater. The impulsive warming of 31 January–1 February raised 25-mb temperature at Coral Harbour (915) by 36C in 48 hr. The waves of February were less impressive, but the final warming of 1–2 March again produced 30C changes in 48 hr.

Fig. 8 shows that all the spatial patterns of fig. 2 were reflected in similar patterns in depth; every major wave crossing the 80W meridian at 25 mb was associated with changes of similar sign in Resolute's

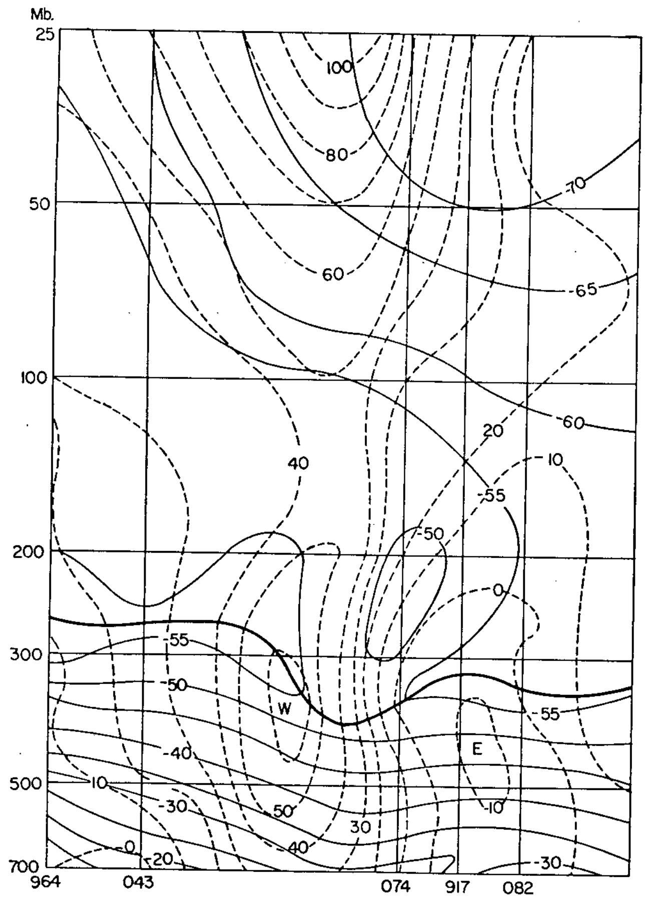

FIG. 15. Cross-section along line Whitehorse (964) to Alert (082) for 0000 GCT 13 December 1958. The line of section crosses the polar-night jet at right angles. Isotachs are for northwesterly component.

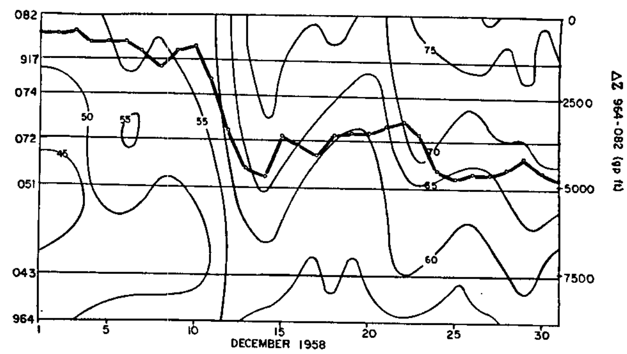

FIG. 16. Temperature of the 50-mb surface between Whitehorse (964) and Alert (082), December 1958. Solid line shows the height difference between the two stations. Note the large increase in this difference associated with the cooling of December 11–13.

column right down to the tropopause. The instability of later winter thus followed the pattern of earlier waves.

The impulsive warmings of 31 January–1 February and 1–2 March closely resembled one another and also resembled that of 29 January 1957, as discussed by Teweles (1958). They displayed the following characteristics:

(a) Both occurred immediately behind the trough-line of eastward-moving baroclinic waves of short wavelength and high spatial amplitude. Fig. 17 shows the sequence of 25-mb temperature fields associated with the 31 January–1 February warming. It is apparent that the

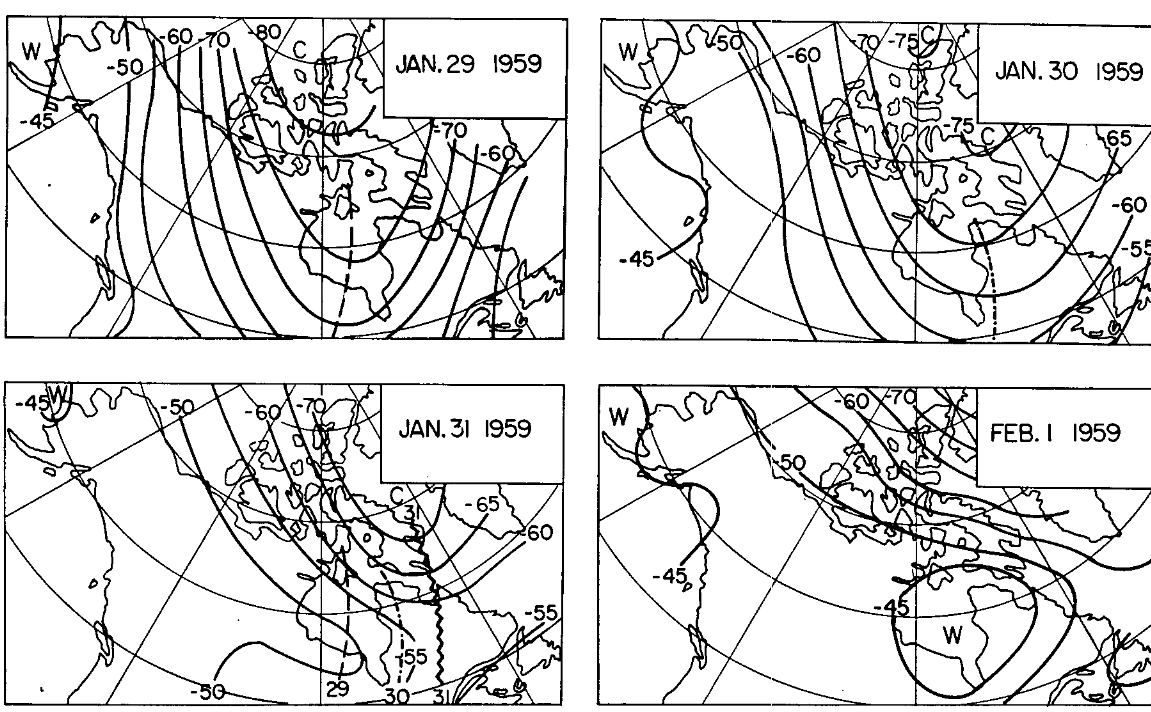

FIG. 17. 25-mb temperature fields during sudden warming of 29 January–1 February 1959. A cold trough over Hudson's Bay on 29 January moved east at about 15 kn, successive daily positions being shown on the map for 31 January. The main warming occurred between 31 January and 1 February, associated with subsidence behind the trough-line. The main polar-night vortex was little affected.

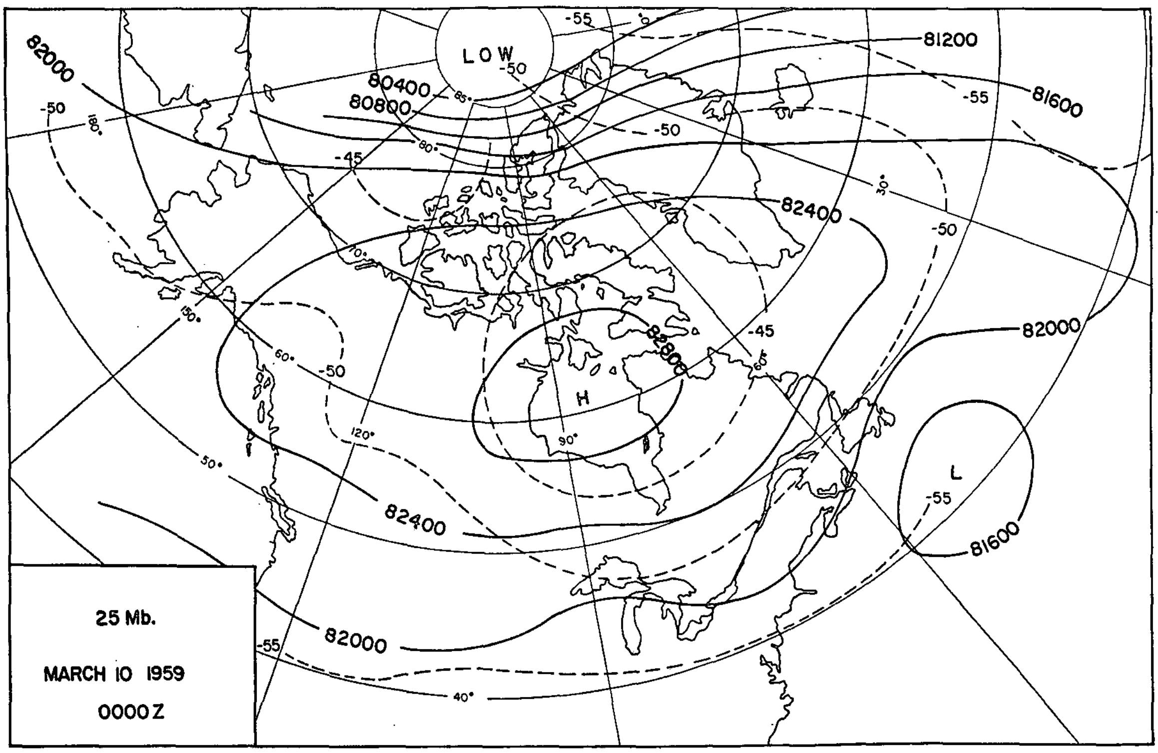

FIG. 18. 25-mb chart for 0000 GCT 10 March 1959, showing a strong anticyclone over Hudson's Bay following final impulsive warming of 1958–59 winter. The polar-night vortex has been displaced to the Eurasian side of the pole.

warming was due to subsidence behind the trough-line.

(b) The maximum amplitude of the warming occurred about 5 deg of lat south of the jet-maximum at 25 mb, but warming extended on either side of the jet.

(c) The warming occurred earliest at and above the 25-mb surface, descending slowly to the lower stratosphere. The base of the warm layer, in which lapse rates were strongly negative (at times as much as -4 deg Celsius km^{-1} through the entire 25- to 50-mb layer), was sharply defined. Teweles (1958) applies the term *stratopause* to this surface.

The effect of the descending warm wave was to trap very cold air below the warm middle stratosphere. For a short period in early March 1959, the 100-mb map displayed full winter cold over the Canadian arctic while temperatures at 25 mb exceeded -45C.

The earlier sudden warming in 1959 was short-lived, the polar-night vortex reappearing with little loss of vigor in February. The warming of 1–2 March, however, completed the breakdown on the American side of the pole, the entire stratosphere assuming summer-like warmth by 10 March. Fig. 18 shows the 25-mb chart for 0000 GCT 10 March 1959. A large anticyclone covered Canada, while the remains of the polar-night vortex had been displaced to the Eurasian side of the pole, where the entire stratospheric column was still very cold (< -65C) on this date. Godson's (1959) results make it clear that this is a typical sequence; warming and anticyclogenesis on the American side tend to precede a later collapse of the polar vortex on the Eurasian flank. On the other hand, the 1957 vortex broke up into separate cells, one of which died out over western North America. It is quite obvious that the breakdown follows no single pattern.

7. Discussion of results

It is by now clear that the arctic stratospheric circulation displays unique and exciting challenges to the meteorologist. The existence of an independent set of motion systems, and of an annual thermal cycle of unexpectedly complex type, are excellent reasons why the theoretician should concern himself with these levels at high latitudes. This paper will be concluded by a review of the evidence presented in the light of unsolved theoretical questions.

The stratospheric wave-domain. It has been shown above that the polar-night vortex is distinct from the Ferrel vortex, from which it is normally separated by a barotropic, sluggish layer in the lower stratosphere. It has also been shown that the zonal flow around the vortex-center is disturbed by standing perturbations—notably by a warm-cored ridge over Alaska and a cold trough over eastern Canada. These quasi-permanent systems are apparently in phase with the similar perturbations of the Ferrel system.

There is strong evidence that the instability of the winter temperature field so graphically displayed in figs. 2 and 8 arises from the eastward motion of baroclinic waves in the polar-night vortex. Each major cooling was associated with approach of a deepening cold trough; each impulsive warming occurred behind the trough-line of a weakening trough. In short, the impulsive temperature changes of the layer above 100 mb seem to be due in the first instance to the vertical-motion systems associated with eastward-moving waves within the zonal flow round the vortex. Advection of the resulting warm air then contributes to the variability. The large sudden warmings that brought the winter of 1958–59 to a close appeared to be of the same type, merely displaying larger amplitudes in the associated thermal waves. There seems no need to invoke external impulses to account for the sudden-warming phenomenon; in actual synoptic analysis, cases of the latter look like typical instability phenomena associated with baroclinic waves and their attendant vertical motion. A study of the stability characteristics of the polar-night wave-trains for 1958–59 is now being prepared, in an effort to demonstrate that they occur at times of critical baroclinicity over the Alaskan ridge.

All the evidence indicates that the stratospheric waves are essentially independent of the Ferrel waves. Instances of apparent interaction are discussed below, but the existence of the barotropic separating layer is a strong argument for suggesting that the two wave domains are normally independent. There is clear evidence that the natural periodicity of the two systems is very different. Reference has already been made to spectral estimates made for Resolute's thermal field. These show that the 100-mb and 25-mb thermal fluctuations are largely due to periodicity centering on 16 days, whereas the 300-mb, 500-mb, and 1000-mb spectra show no such maximum; the 500-mb surface actually has a slight minimum at this frequency. The spectra of the two wave domains—Ferrel and polar-night—are thus quite different.

The 200-mb map, which is almost wholly stratospheric in high latitudes, nevertheless displays the Ferrel waves; its contour patterns are like those of the 500-mb and 300-mb charts. In other words, its vorticity systems are largely the upward extension of tropospheric systems. It is well-known that the perturbation temperature field at this level reflects the vertical motion of the Ferrel wave systems, with subsidence in troughs and uplift over ridges (Kochanski, 1954; Sawyer, 1951). A remarkable fact that emerges from the present investigation, however, is that the broad temperature distributions of the stratosphere, which are dynamically related to the polar-night vortex, descend to the 200-mb surface. In early March 1959 for example, the 100- and 25-mb surfaces saw warmth (−40C to −50C) over North America and the Atlantic and very cold conditions (at least at 100 mb) over the hemisphere beyond the Greenwich meridian and 180W. This remarkable distribution was strongly developed at 200 mb, with temperatures near −45C in the western hemisphere and −60C to −65C over Siberia and Europe. It appears that these great temperature waves of the stratosphere are able to descend to the tropopause through the upper Ferrel systems.

Interaction with Ferrel waves. Precise harmonic analysis of circumpolar charts is required before one can really answer the question whether the polar-night and Ferrel wave domains are interconnected. In the absence of such analysis, one can only point to apparent cases of interaction.

Two such cases occurred in the winter of 1958–59. On 20 to 22 December 1958 (fig. 8), an eastward-moving trough in the polar-night system produced an impulsive cooling over the whole arctic at stratospheric levels. At Resolute, Hall Lake (081) and several other eastern stations, the cooling extended down to the 400-mb surface and was followed on the 23rd by rapid warming at these levels and a rise of the tropopause to 250 mb. A frontal cyclone over Labrador and Baffin simultaneously deepened very intensely, attaining 950 mb by January 24, under conditions that would not normally have suggested it. Much more spectacular (fig. 8) was the period from 2 to 10 January 1959, which saw the waxing and waning of a tropospheric warm high near Resolute beneath the cold core of the polar-night vortex. On 5 January, the tropopause at Resolute had attained 180 mb at a temperature of −74C, and the entire stratosphere was below −70C. It is difficult, looking at fig. 8, to doubt that interaction between the two domains occurred at this time.

It is as yet impossible to prove this connection. It will be suggested here (i) that at certain times, probably by chance phase-correlation, interaction may occur between the two wave domains, producing strong development of either sign at tropospheric levels and (ii) that such interactions are most probable in the later winter, when the thermal waves of the stratosphere have the largest vertical amplitude. Namias (1958), Julian[3] (1959), and others have suggested, in particular, that the Ferrel system may run through its primary index cycle in late February and early March because of some unspecified linkage with the breakdown of the polar-night vortex.

Acknowledgments. The daily analysis on which this paper is based was conducted in space provided at the Central Analysis Office, Meteorological Branch, De-

[3] Julian, P.: personal communication.

partment of Transport, Canada, where the data were collected and plotted by courtesy of the officer-in-charge, Mr. J. Leaver. The author expresses warm thanks to Mr. Leaver and to the members of the Arctic Meteorology Research Group who shared in the analysis. Special thanks are due to Mr. B. W. Boville for good advice and daily assistance and to Mrs. Ulla-Brita Manley who carried out the spectral analyses. Miss Kay Allington drew the maps.

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