Upper Air Temperatures over an Antarctic Station

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Abstract

Typical features of antarctic upper air temperature distributions are demonstrated by means of average values and selected ascents. A great seasonal change in the height of the tropopause is pointed out. Attention is drawn to the remarkable agreement between observed stratospheric temperatures in winter and Dobson's estimated values. The causes for the heating of the stratosphere in late winter and spring are discussed. The difference between the tropospheric temperatures in summer between Arctic and Antarctic is assumed to be due to the different value of the surface albedo. Rough estimates indicate that the absorption in the 9.7 μ band by ozone is of great significance to the observed stratospheric temperature contrast between the two polar regions in winter.

The aerological observations at Maudheim (lat. 71° 03′ S., long. 10° 56′ W.), the base of the Norwegian-British-Swedish Antarctic Expedition, 1949—52, started on March 15, 1950, well over a month after the landing on February 10. During March, quite a few soundings were carried out, but the quality of these first soundings was rather unsatisfactory, due to instrumental and operational difficulties. As from April 1, however, the programme of one sounding a day was steadily adhered to, whenever weather permitted launching. The daily ascents were suspended for 10 days in May, 1950, when weather conditions enforced a radical change in the launching technique hitherto used. The last sounding was made on January 10, 1952.

During the entire period some 650 sounding balloons were launched, nearly all carrying a radiosonde of Finnish design (Väisälä type) and a rawin transmitter for upper wind measurements.

In this note we shall deal with the temperature measurements only, which are now partly worked up. The upper wind data are being examined by Gordon de Q. Robin, phycisist of the expedition, who was responsible for this part of the investigations.

Table I contains the numbers of temperature data at stated pressures for the different months, both years combined. Thus, the table shows the number of observations upon which our mean values are based. All unreliable ascents have been disregarded, and there are only very few cases with more than one sounding a day.

All ascents have been recalculated after our return, and for higher levels corrections for radiation error have been applied according to the latest method introduced by VÄISÄLÄ

Table 1. Number of temperature readings at different standard pressure surfaces

	Jan.	Febr.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
850	38	23	31	51	46	58	60	60	56	58	57	58
700	37	23	31	51	44	58	59	60	56	58	57	58
600	37	23	31	51	43	58	59	60	56	58	57	57
500	37	23	30	51	42	58	58	60	56	57	56	57
400	37	23	30	51	42	57	55	60	55	57	54	57
300	37	23	30	51	42	57	53	59	54	57	52	57
250	36	23	29	51	41	56	53	59	54	57	52	56
200	36	23	29	46	38	56	52	59	54	57	51	56
150	35	23	28	42	36	54	46	57	53	56	50	53
100	35	23	25	38	32	45	27	46	49	54	45	48
85	27	21	19	30	25	30	9	23	40	49	38	42
70	22	19	18	22	21	9	3	18	35	44	34	39
60	20	17	17	18	11	4	1	6	26	38	25	35
50	16	14	12	12	10	2	0	2	21	29	17	23
40	9	9	8	7	2	0	0	0	12	19	10	14
30	3	3	2	3	0	0	0	0	6	13	2	2
25	2	0	0	1	0	0	0	0	4	7	1	1
20	1	0	0	0	0	0	0	0	1	1	0	0

(1941) and RAUNIO (1951). In this connection it should be mentioned that in summer the ascents were made in the evening with low solar altitude. In this manner we tried to reduce the radiation error to a minimum.

Some features of the upper air temperature distribution, according to selected ascents

Figure 1 shows six selected ascents from Maudheim, taken at different seasons. An ordinary log p, T diagram has been used. Ascent I shows typical summer conditions. At this time of the year there exists no ground inversion during the day, though at night a rather small inversion may develop. Even in the lowest layers the temperature is generally below 0° C, a value barely exceeded in very few cases.

A marked tropopause is usually found in summer, often rather close to the 300 mb level. Above the tropopause there is found a layer of variable thickness with a pronounced negative lapse-rate. Higher up the temperature increases slightly up to the highest levels reached.

Even though the day to day variation of the temperature has not been fully investigated so far, it is evident that these variations are comparatively small during the summer, especially in the stratosphere. If we consider the 35 readings obtained from the 100 mb level during January the extreme temperatures recorded are − 38,2° C and − 45,0° C, and in

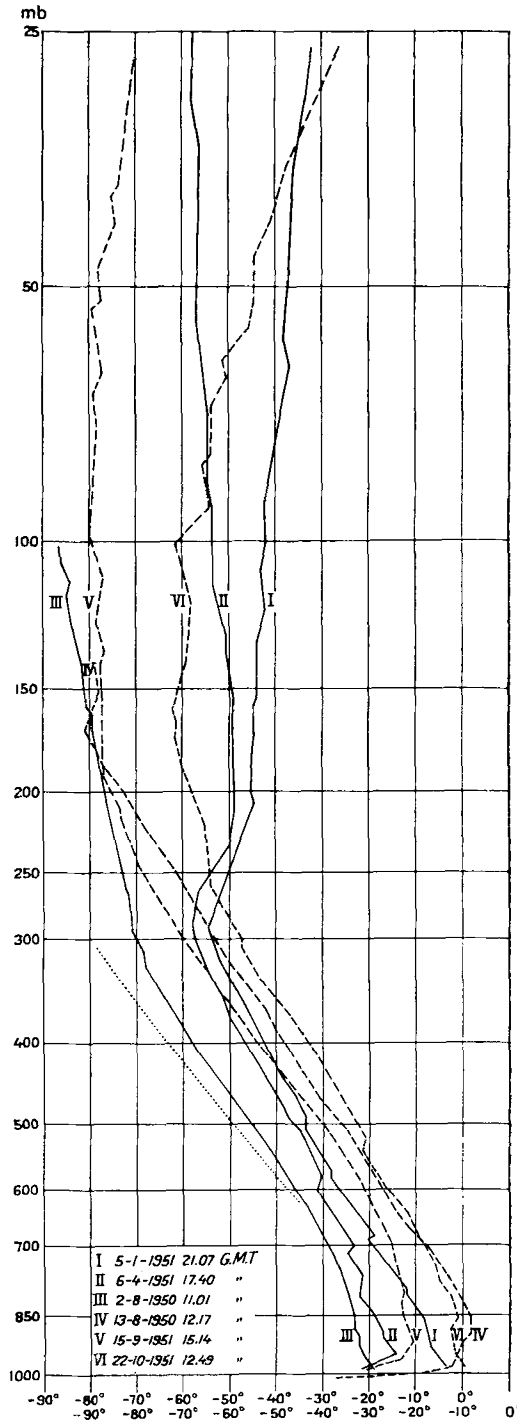

Fig. 1. Selected ascents. Full lines upper temperature scale, dashed lines lower temperature scale. Dotted line dry adiabatic.

21 of the cases the readings range between $-41,0°C$ and $-42,9°C$.

In autumn, a ground inversion is frequently present even in the middle of the day, and the tropopause is usually very marked. The stratosphere has now become considerably colder, especially at the higher levels, and the temperature decreases slightly upwards (Ascent II).

Ascents III and IV show winter conditions: Generally a pronounced ground inversion is found, but a long-lasting gale or a heavy overcast sky may cause the inversion to disappear completely even in the middle of winter.

At this time of the year, temperatures in the lowest part of the troposphere undergo large day-to-day fluctuations, partly caused by vertical motions. At the top of the inversion layer or at a height of about 1,500 m, subsidence above may cause the temperature rise nearly to the highest values that are found in summer time. Ascent IV showed a temperature of $-8,2°C$ at the 850 mb level, whereas the absolutely highest value measured at this level was $-4,9°C$ on January 20, 1951.

One of the most interesting features of the antarctic winter ascents is the indefinite character of the tropopause, frequently observed during the winter months. COURT (1942) who first discussed this fact, goes as far as to say that the tropopause disappears during the antarctic winter. However, this is a matter of definition. In cases where no abrupt change in lapse-rate takes place, the generally adopted definition is: "The tropopause is taken as the point where the mean lapse-rate to the kilometre next above is 2°C or less, provided that it does not exceed 2°C for any subsequent kilometre", (Gibbs, Gotley and Martin, 1950).

As in the coldest months the lapse-rate usually exceeds 2°C at the highest levels of the soundings, the tropopause certainly disappears completely according to this definition. On the other hand, it is usually possible to point out a level above which the lapse-rate is quite different from the values found below, especially when mean values are considered. Therefore, to suit antarctic conditions, the quoted definition should be modified, and this can be done by deleting the subsidiary clause, the definition thus being: "The tropopause is taken as the lowest point in the transitional zone, where the mean lapse-rate for the kilometre next above is 2°C or less." This definition has been adopted for the Maudheim ascents, even when a real inversion exists. In this way it has been possible to define a tropopause in nearly all cases, often, however, with poor distinctness (Ascent III). Still, in some few cases the lapse-rate exceeds the limit mentioned for all layers between the 700 and the 100 mb surface, the tropopause thus completely disappearing even according to our definition.

On the other hand, the tropopause may occasionally be very distinct, even in the middle of the winter. In such cases it is usually found at considerable heights, and with extremely low temperatures, possibly the lowest values ever recorded in the atmosphere, whereas the lower part of the troposphere is warm and dry. In other words, these ascents show conditions similar to those observed in warm anticyclones on middle latitudes, the absolute temperature values being disregarded.

Although the extremely low temperatures in the stratosphere usually are connected with high temperatures in the lower troposphere, this is not always the case, as seen from ascent V. Furthermore, this ascent shows that very low temperatures may occur as late as in the middle of September. In late spring the stratosphere gradually becomes warmer, and as the heating starts at the highest levels, the lapse-rate may obtain comparatively great negative values. Thus, the temperature may rise as much as 30 to 40 degrees centigrade from the 100 mb to the 25 mb level, Ascent VI.

Mean temperature distribution

Mean ascent curves have been derived for half monthly as well as for monthly periods. In the lowest layer, the mean temperatures are determined for heights of 250, 500, 750 and 1,000 metres. Higher up the means refer to isobaric surfaces. Starting at 850 mb, the distances used between successive reference surfaces were 30, 20 and 10 mb for the layers 850 to 640 mb, 640 to 300 mb and 300 to 100 mb, respectively. Above the last mentioned surface a constant distance of 5 mb has been used. In this way, a smooth mean ascent curve has been obtained for each period. Furthermore, it has been possible to make use of additional readings above the highest standard surface reached by the balloon.

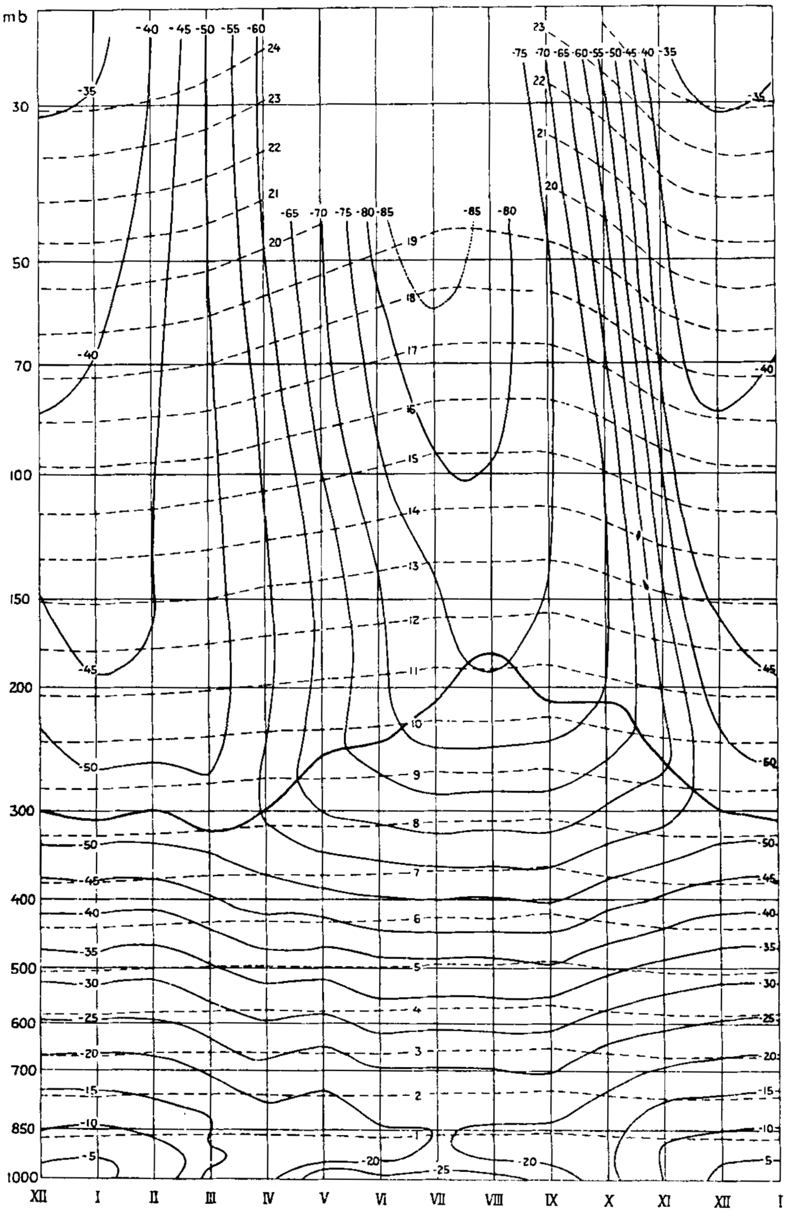

Fig. 2. Isopleth diagram. Full lines isotherms. Dashed lines heights in geopotential kilometres. Heavy full line the tropopause.

On the basis of the monthly mean curves for both years combined an isopleth diagram has been constructed, showing the yearly course of the mean temperature at any level as well as the course of the tropopause, as derived from the same curves (figure 2). The heights entered on the diagram have been calculated from the mean temperature values. For the highest levels the mean temperatures are based on the well-known method of mean differences, although different authors have raised strong objections against this method. (SCHERHAG, 1948). This question will be discussed more closely in the final expedition report. Possible errors arising from the method used are of minor importance to the conclusions in the present note.

The troposphere. In the troposphere the mean temperature in summer decreases all the way from the ground up to the tropopause. Strictly speaking, this applies to day time conditions only, because a considerable daily variation is present in the lowest hundreds of metres. In the months of January and February the mean temperature near the surface was —4° to —5° C, that means considerably lower than in the corresponding months in the Arctic. Thus, over the north polar pack ice, SVERDRUP (1933) found in the warmest month a mean temperature of +3° C at 600 metres. Apart from the very highest layers, the whole troposphere is in summer considerably warmer over the Arctic than over the Antarctic. (Figure 3.)

During the autumn the troposphere is gradually cooled, especially in the lowest kilometre in connection with the formation of a surface inversion. In winter, when the inversion becomes persistent, there is in the lowest kilometre a mean temperature rise upwards of some 7° C. Thus, close to the ground a considerable annual temperature variation is found, although the range is somewhat smaller at Maudheim than at Arctic Bay. (Figure 3.)

Above the inversion layer the range rapidly decreases, and at 700 mb the difference between the highest and lowest monthly mean temperature is only 7° C, less than half the corresponding value from Arctic Bay.

This small annual temperature change of the middle troposphere seems to take place mainly in spring and autumn, whereas the month to month variation during summer and winter is extremely small. Thus, no distinct maximum and minimum can be pointed out, at least not from our comparatively short period of observation.

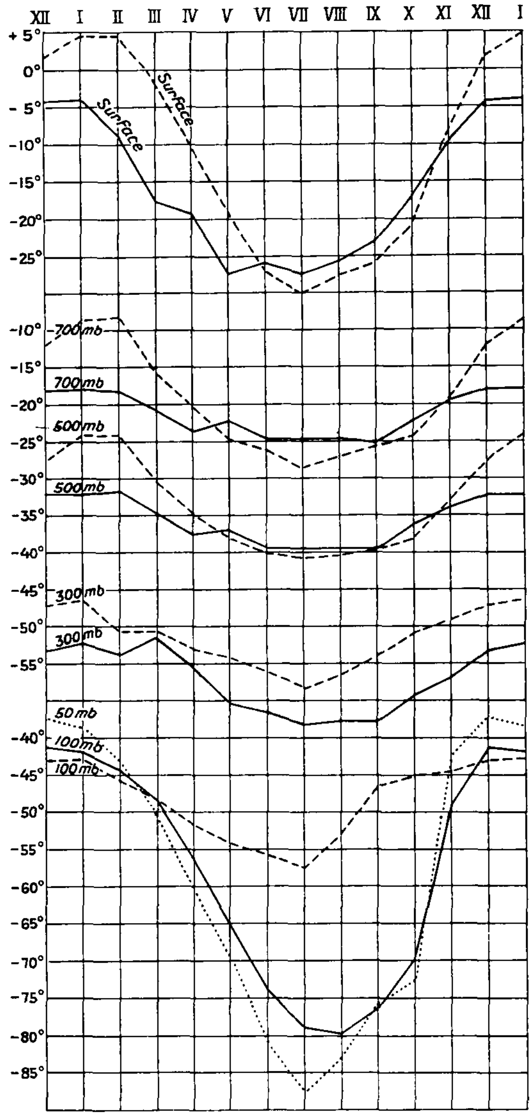

Fig. 3. Mean temperatures at some standard pressures for different months. Month 1 January and July for Maudheim and Arctic Bay respectively. Full lines Maudheim values. Dashed lines Arctic Bay values. Dotted line Maudheim temperatures at 50 mb.

The stratosphere. The great seasonal variation in the stratosphere is the most striking feature of the mean temperature distribution, as shown in figure 2. Starting with the highest levels, we find a distinct maximum in December, i.e. when the sun has its highest altitude, and the same applies to all the layers above the 100 mb surface. At this time of the year the lapse-rate is negative through the whole stratosphere, from the tropopause up to at least 25 mb. Soon after summer-solstice the highest layers start to cool off, at first slowly, but more rapidly in the early autumn. The stratification, therefore, becomes gradually more isothermal, and is almost exactly isothermal in March. As the cooling continues and remains greatest in the uppermost layers, a temperature decrease with height above the tropopause becomes more and more pronounced during the autumn and the first part of the winter. At winter-solstice, or possibly a few weeks later, the stratospheric lapse-rate attains its highest positive value, exceeding 2°C per 1,000 metres from a height of 13 kilometres in July. In this month the lowest mean value at the 50 mb surface is found, the minimum being very distinct. In August, the warming of the highest layers starts, but in that month the temperature in the lowest part of the stratosphere reaches its minimum value.

This time lag in the occurence of the yearly minimum at different levels, produces a marked skewness in the diagram shown in figure 2. This marked feature in the temperature distribution is fairly well established, although in the coldest months the absolute temperature values are rather unreliable above 85 mb, due to very few readings. In July, all ascents above 100 mb showed a marked temperature decrease with height, but in August the stratification was more isothermal.

A heating of the highest layers is noticeable in late winter, but the main warming takes place after spring equinox. At 50 mb, the most rapid increase in temperature is found between October and November, or in the middle of October according to the half monthly mean values. From the first to the second half period of October, the temperature at that level rose from −69.5 to −57.4°C in the first year and from −71.4 to −51.0° C in the second, with a mean temperature rise of 16°C for both years combined, i.e. one degree per day.

Due to this pronounced heating of the highest layers in early spring, the stratospheric lapse-rate undergoes a rapid change in this period. In fact, the mean lapse-rate changes from its highest positive to its highest negative value in about three months. In the lowest part of the stratosphere the main heating takes place in late spring and early summer.

Comparing our results with arctic conditions, we find remarkable contrasts in winter, whereas the summer conditions are quite similar, as previously pointed out by COURT (1942). Unfortunately, aerological data from the Arctic are still scanty, especially for the higher layers in winter. To obtain the values at the highest level for Arctic Bay, figure 3, the mean ascent curve for the coldest months had to be extrapolated (HENRY and ARMSTRONG (1949)).

Figure 3 shows that in summer Maudheim is slightly warmer than Arctic Bay at 100 mb, whereas in the coldest months it is some 20 to 25°C colder. The yearly range at this level is 39°C and 15°C, respectively, the values thus differing by some 25° C. This difference probably increases with height.

Tropopause layer. Figure 2 demonstrates a great annual variation in the height of the tropopause over Maudheim. In the summer months, when the temperature of the lower stratosphere is comparatively constant, the tropopause is found near the 300 mb surface. In the autumn, its mean height increases as the temperature of the lower stratosphere decreases, reaching its highest value in winter when the temperature in the layer just above has its minimum. As this part of the stratosphere becomes warmer again in spring, the height of the tropopause gradually decreases, reaching "normal" summer value at the same time as the temperature of the 250 mb surface reaches its summer value.

It is of interest to examine if this intimate connection between the height of the tropopause and the stratospheric temperature means that the tropospheric conditions are without significance. For this purpose, the pressure at the tropopause (y) has been correlated with the temperature at the 600 mb (x_1) and the 150 mb (x_2) levels, all three values taken from each of the 43 half-monthly mean ascent curves.

Calling the correlation coefficient between y and x_1 r_{yx_1}, and that between y and x_2 r_{yx_2}, we found $r_{yx_1} = +0.48$ and $r_{yx_2} = +0.91$. The correlation coefficient between x_1 and x_2 was rather high, $r_{x_1x_2} = +0.74$, and the partial coefficients were found to have the following values $r_{x_1y \cdot x_2} = -0.73$ and $r_{x_2y \cdot x_1} = +0.95$.

These figures clearly show how the correlation coefficient r_{yx_1} gives a wrong impression of the connection between the height of the tropopause and the temperature of the lower troposphere. The linear regression equation will be:

$$y = 331 - 7.0x_1 + 4.5x_2$$

This shows that a given change in temperature at one of the two layers under consideration will be followed by the greatest change in the mean height of the tropopause if it occurs at the 600 mb level. However, the greater variation in the stratospheric mean temperature will make the last term of the equation dominate when seasonal changes are considered.

On the cause of the observed temperature distribution

In summer the troposphere is much colder in the Antarctic than in the Arctic. This striking difference is probably caused by the unequal values of the albedo in the two regions. At Maudheim, LILJEQUIST (1954) carried out extensive radiation measurements and found remarkably high albedo values even in summer. Corresponding measurements in the Arctic show much lower values, meaning that there the surface will receive more energy from the sun and accordingly attains higher summer temperatures. By various thermodynamical processes parts of this energy are transferred to the troposphere.

Before discussing the observed stratospheric temperatures, a few general remarks about the radiation in this part of the atmosphere are needed. According to DOBSON *et collab.* (1946) the main absorbing gases are carbon dioxide, water vapour and ozone. The amounts of water vapour as well as carbon dioxide generally decrease from the ground upwards, whereas ozone is found mainly in the stratosphere, the amount increasing with height in the layers under consideration.

To our discussion the absorption power of ozone in the infrared is of special interest. The main absorption band of the ozone lies at about 9.7 μ, that is in a part of the spectrum where both water vapour and carbon dioxide are nearly transparent. Therefore, most of the energy that is absorbed by the ozone comes from the surface or from the tops of the clouds, and depends particularly on the surface temperature (DOBSON 1946).

At Maudheim, the sun is below the horizon for approximately two months. On the other hand, according to a diagram by COURT (1942) the direct solar beams reach all the layers of the stratosphere for a short period during the day even in the darkest season. Still, the influence of the direct solar radiation may safely be neglected during this part of the year because of the great negative elevation of the sun and the short duration of sunshine. Furthermore, the absorption in the infrared by ozone must be of minor importance, as the surface temperature of the antarctic continent is extremely low. Due to strong circulation and slow adjustment to radiative equilibrium in the stratosphere in winter the observed temperature values at Maudheim can be expected to be highly dependent on the radiative conditions over the continent.

Thus, in the coldest months the temperatures of the lower stratosphere must be determined mainly by absorption and radiation in the infrared of carbon dioxide and water vapour. Dobson's estimates of the radiative equilibrium temperatures in the lower stratosphere of each of the two gases are 200° A. and 190° A., respectively, in remarkable agreement with the observed values. Furthermore, with decreasing amount upwards of the absorbing gases, the temperature has to decrease with height, again in accordance with our observations.

The heating of the highest layers which starts soon after winter solstice is most likely due to increasing intensity of the direct solar radiation. The flux of the long-wave radiation from below must remain comparatively unchanged, because at this time of the year no rise in temperature can be noticed in the troposphere or at the surface, and accordingly no heating of the stratosphere can arise from absorption of this radiation.

On the basis of these assumptions the skewness on the isopleth diagram (fig. 2) is easily explained. Firstly, the duration and intensity of sunshine increase with height, meaning that the radiation from the returning sun must be of greater importance at higher levels than further down. Secondly, because of the downwards increasing density of the air a greater amount of absorbed energy is needed to produce a noticeable temperature rise in the lowest part of the stratosphere. The relatively small amount of ozone in the lowest part of the stratosphere lends added importance to the last named factor. Altogether it is evident that a temperature rise caused by increasing sun radiation must be observed first at higher levels.

As the altitude of the sun increases in spring, the warming of the stratosphere takes place with increasing rapidity. However, as the lowest part of the atmosphere including the surface simultaneously is heated, the long-wave radiation from below attains greater significance, especially to the conditions in the lowest parts of the stratosphere. Furthermore, if the ozone distribution shows variations that are similar to those observed in the northern hemisphere, the amount of ozone increases rapidly in late winter. There exist, therefore, valid reasons for assuming that the observed heating in spring as well as the rapid change in the lapse-rate are caused by the growing significance of ozone. Moreover, there are strong indications that at least above 100 mb, the absorption of the direct solar radiation is the dominating factor, even in spring and late summer. The rate of warming observed in October, exceeding 1°C day in some layers, is much greater than the maximum value to be expected from absorption of the outgoing radiation.

It should be added, that in the layers just above the tropopause, other factors, such as vertical motion extending from the troposphere, may be of considerable importance to the temperature distribution.

Next, we turn our attention to the observed difference between arctic and antarctic temperature in the stratosphere during winter. Court, who first discussed this marked contrast, concludes that a stronger meridional exchange of air in the northern hemisphere is the main reason for the higher temperature. Rubin, dealing with the same problems, shows that outgoing radiation may account for the observed cooling of the antarctic stratosphere from autumn to winter, and agrees with Court in his explanation of the higher values over the Arctic (RUBIN, 1953). Both authors arrive at this conclusion, as they consider the available outgoing radiative energy from the surface too small to account for the difference.

When the upper wind data from Maudheim have been examined, more light will be thrown upon this matter. Therefore, a more detailed discussion will be presented in our final expedition report. At present, it will only be showed that there really exists radiative effect which may account for most, perhaps for all of the observed temperature difference.

A rough estimate can be made of the amount of energy needed to produce the difference. For comparison, the station Tromsø (69° 39′ N., 18°57′ E.) had to be used (JOHANSEN, 1949), as no winter data for the highest layers were available from Arctic Bay.

Table II. Mean temperatures of the air mass between the 250 and the 50 mb surfaces

	I	II	III	IV	V	VI
Maudheim	- 43.6	- 45.0	- 47.6	- 54.6	- 63.9	- 71.4
Tromsø	- 45.0	- 46.9	- 49.9	- 53.8	- 59.3	- 59.7

	VII	VIII	IX	X	XI	XII
Maudheim	- 76.2	- 77.0	- 74.2	- 68.9	- 53.9	- 44.6
Tromsø	- 63.0	- 60.9	- 56.7	- 53.1	- 46.1	- 45.0

Table II shows mean temperatures in corresponding months for the air mass between the pressure surfaces 250 and 50 mb. The difference between the two figures for each month is proportional to the net amount o heat that is needed for raising the temperature from the lower to the higher value.

In month IV the temperatures at both stations are practically equal, but three months later there is a difference of some 13°C. This means that as compared to Maudheim, the air column over Tromsø has gained a net amount of heat of approximately 7 gcal. cm^{-2} day^{-1}. In addition, we must take into account the excess of lost energy due to radiation at higher temperatures. For this rough estimate we will assume the loss of heat by the air column under consideration to be equal to 20 % of the total black body radiation, as previously done by Elsasser in a similar connection (ELSASSER, 1942). Using the mean temperature during the three months considered for both stations, we find an excess in lost energy at Tromsø of some 6 gcal. cm^{-2} day^{-1}. In other words, to produce the observed temperature contrast, the air mass between the 250 and 50 mb levels over Tromsø must have received an amount of heat of some 10—15 gcal. cm^{-2} day^{-1} in addition to the amount received by the same air column over Maudheim.

It is not unlikely that the absorption by ozone in the 9.7 μ band may produce a difference in absorbed energy of the same order of magnitude. Over the antarctic continent, the surface temperatures in winter are extremely low. Temperature measurements in the ice made on the inland plateau by the glaciologist Charles Swithinbank on a weasel journey from Maudheim to the border of the inland ice cap indicated a mean surface temperature for the whole year of some $-40°$C (Robin, 1953). Further inland it must be even colder. Thus, a mean temperature of some 220° A for the months under consideration does not appear unreasonable.

In the Arctic, the winter temperatures are much higher. In addition, the more extensive cloud cover may rise the effective temperatures for the outgoing radiation. With a temperature some 30° higher than the estimated value over the Antarctic, the intensity in the 9.5 μ band increases from 11 gcal. cm^{-2} day^{-1} to 24 gcal. cm^{-2} day^{-1} within a band width of 1 μ.

Dobson states that some 70—80 % of the energy is absorbed by the ozone, and if this absorption takes place in the lower part of the polar stratosphere in winter, it seems to be sufficient to account for the observed temperature difference. This effect may also explain the more isothermal stratification observed in the arctic stratosphere at this time of the year.

REFERENCES

Court, A., 1942: Tropopause disappearance during the antarctic winter. *Bull. Amer. meteor. Soc.*, **23**, pp. 220—238.

Dobson, G. M. B., Brewer, A. W., and Cwilong, B. M., 1946: Meteorology of the lower stratosphere. (Bakerian Lecture) *Proc. Roy. Soc. London, A*, **185**, pp. 144—175.

Elsasser, W. M., 1942: Heat transfer by infrared radiation in the atmosphere. *Harvard meteor. Stud.*, No. 6, p. 97.

Gibbs, W. J., Gotley, A. V., and Martin, A. R., 1950: Heard and Macquary Islands, 1948. *Austral. natl. Antarctic Res. Exped. Rep.*, Ser. D., **1**, part 1 a, p. iii.

Henry, T. J. G., and G. R. Armstrong, 1949: Aerological data for northern Canada. Toronto, Dept. of Transport, pp. 132—145.

Johansen, H., 1949: Mean upper air data obtained from soundings at Tromsø during the years 1941—44. *Geofys. Publ.*, **17**, No. 4, pp. 16—17.

Liljequist, G. H., 1954: Radiation, and wind and temperature profiles over an Antarctic snow-field. A preliminary note. *Proceedings of the Toronto Meteorological Conference* 1953.

Raunio, N., 1951: Amendments to the computation of the radiation error of the Finnish radiosonde. *Geophysica*, **4**, pp. 14—19.

Robin, G., 1952: Queen Maud Land. The scientific results of the international expedition. *The Geographical Magazine*, p. 289.

Rubin, M. J., 1953: Seasonal variations of the Antarctic tropopause. *J. Meteor.*, **10**, 127—134.

Scherhag, R., 1948: *Wetteranalyse und Wetterprognose*, Berlin, Springer-Verlag, p. 59.

Sverdrup, H. U., 1933: Meteorology, The Norwegian North Polar Expedition with the Maud, 1919—25. II, Pt. I, Discussion, pp. 38—39.

Väisälä, V., 1941: Der Strahlungsfehler der finnischen Radiosonde. Univ. of Helsinki, Inst. of Meteor. Mitteilungen No. 47.