Annual and Diurnal Temperature Variations in the Upper Atmosphere

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Abstract

An attempt is made to determine the annual and diurnal variations of temperature to a height of 80 km by knitting together available observations and theoretical results. The observations include radiosonde data, meteor-determined densities, V-2 rocket observations, anomalous propagation of sound, and infrared spectroscopy. The theory most heavily relied upon is the radiative computation made by Gowan. For the annual temperature range there is reasonably good agreement between theory and observations, such that in the upper ozonosphere (30 to 55 km), summer temperatures are higher than winter temperatures by amounts as large or larger than than these at the surface. As in lower levels, the annual temperature range aloft increases with latitude. For the diurnal temperature variations there is a large disagreement between the computations of Gowan and Penndorf. However, available evidence appears to favor Gowan's computations, which, together with other considerations, indicate a diurnal temperature range of the order of 10 to 15° C.

The magnitude of the annual and diurnal variations of air temperature below about 20 km are well-known from years of observations gathered from a widespread aerological network. For altitudes above 20 km there exist only sporadic observations, both direct (sounding balloons, rockets) and indirect (meteors, sound, spectroscopy, etc.) taken at a few places. Despite the scarcity of data there have been attempts (MARTYN and PULLEY, 1936), (WARFIELD, 1947) based on indirect observations to construct the average temperature-height curve to over 100 km, which compared quite well with later direct observations taken by the V-2 rocket soundings in New Mexico (DURAND, 1949). This verification of a result derived from speculative reasoning based on a few observations, encourages a similar attempt to arrive at some idea of the annual and diurnal temperature variations of the upper atmosphere to a height of 80 km. This height is chosen because it represents the rough upper limit of the most abundant observations, both direct and indirect.

The problem of determining the annual and diurnal temperature variations of the upper atmosphere is more than just an attempt to extend aeroclimatological statistics to greater heights than attained by routine soundings. For, just as many important dynamical consequences follow from the annual and diurnal march of temperature in the surface layer of air, so it follows that significant dynamical influences may also result from annual and diurnal heating and cooling of the upper portion of the ozonosphere (30 to 55 km) if it can be shown that the magnitudes of the temperature variations are appreciable. For example, in a search for possible non-cyclic solar controls of sea-level weather it has been suggested that the ozone layer, through its strong absorption of certain wave-lengths of ultraviolet radiation, may serve as the

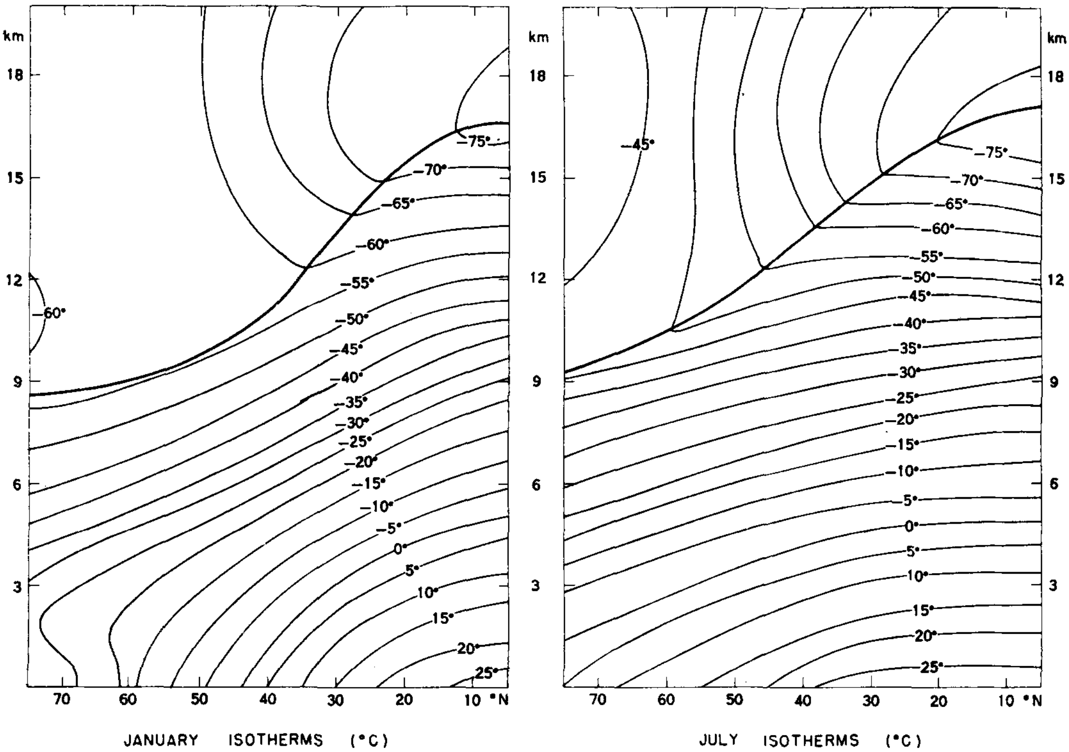

Fig. 1. January and July vertical cross-sections of the average temperature field, Northern Hemisphere, in ° C.

connecting link between sporadic bursts of intensified ultraviolet radiation from the sun and the changes in the circulation pattern of the terrestrial atmosphere. If it can be shown that the absorption of normal intensity ultraviolet solar radiation by the top layers of the ozonosphere yields only insignificantly small annual and diurnal temperature ranges, then it would be difficult indeed to expect that a short burst of ultraviolet radiation of double, or even triple, normal intensity could cause a significant warming up of these same layers and resulting important dynamical consequences on the general circulation. On the other hand, if annual and diurnal temperature variations turn out to be large, then the prospect for demonstrating an important sporadic ultraviolet effect on sea level weather is more promising.

Annual Temperature Variations in the Lower Atmosphere

First, let us examine the annual course of average temperature in the lowest 19 km of the Northern Hemisphere from latitudes 10° to 70° N, as illustrated in average vertical cross sections for January and July (fig. 1). These data represent the final product of analysis of all available pre-World War II aerological data for the Northern Hemisphere (1945). The surface data were taken from HANN-SÜRING (1939). The work was done during the war at the U. S. Weather Bureau under the leadership of R. C. Gentry with the assistance of Air Weather Service weather officers. A description of these, as well as other war-time meteorological chart products, is given elsewhere (WEXLER and TEPPER, 1947).

In fig. 1 the isotherms for each 5° C are shown as thin lines with thicker lines representing the tropopause. The latter lines represent the latitudinal variation in mean height of the tropopause taken from FLOHN (1947). Their presence is not to be interpreted as implying that the tropopause is a continuous surface extending from tropical to polar regions. Indeed, it is well-known

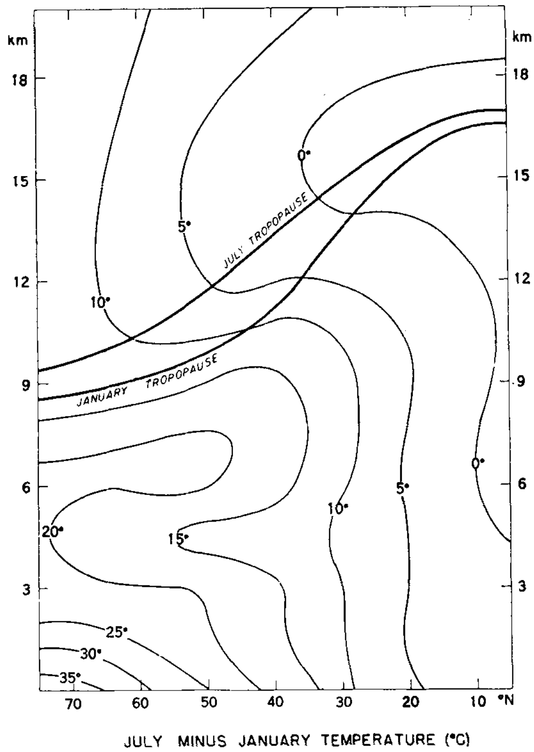

Fig. 2. Range of average temperatures from winter (January) to summer (July), Northern Hemisphere, in ° C.

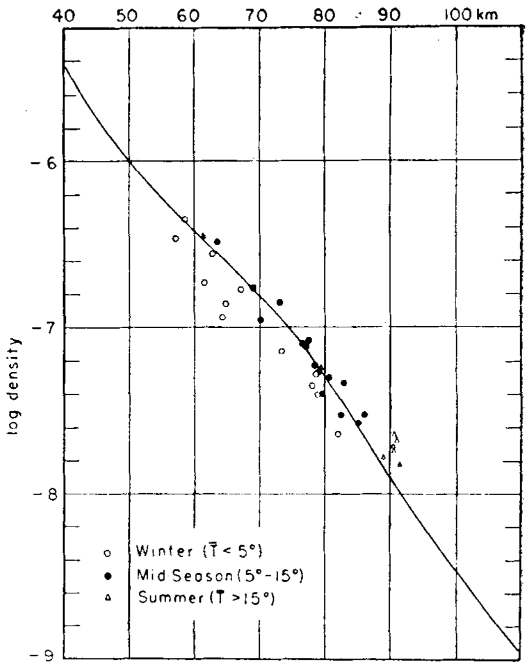

Fig. 3. Points: densities computed from meteor observations (Whipple, Jacchia, Kopal). $\bar{T}$ is the normal surface temperature for the observation day at Boston, Mass. Curve: NACA density-height curve (Warfield).

that there may be more than one tropopause over a given point during certain meteorological conditions and during other types of situations it may be missing altogether.

The annual range of average temperature as approximated by the difference of the average temperatures, July—January, is shown in fig. 2. In general, the temperature range decreases at all altitudes from north to south. There appears to be distinct evidence of two layers where minima in the temperature range occur with increasing altitude. The first is at 4 to 5 km, and is especially distinct at higher latitudes. The second is found in a sloping layer, roughly corresponding to the average height of the tropopause. An attempted explanation of these two interesting minima is outside the scope of this paper; for our purpose, it is significant to note that at great heights (above 15 km) there is some evidence of an increasing temperature range with height at all latitudes. The problem is to see if this is so, and to find the magnitude of the temperature range at heights above the level of direct observations.

Computed Annual Temperature Range for an Isothermal Atmosphere

In seeking to extend the annual range to greater heights, use is made of the seasonal variation of densities as determined by meteors observed over Massachusetts (Whipple, Jacchia, Kopal, 1949), and the observations are reproduced in fig. 3.

Taking the observed annual range of the densities at a mean height of 78 km over Massachusetts as $\Delta \log_{10} \varrho = 0.5$ (loc. cit. pp. 154—5), let us convert this value to that of the absolute temperature of the atmosphere from 0 to 78 km under the assumption that the atmosphere is isothermal at temperature T_0.

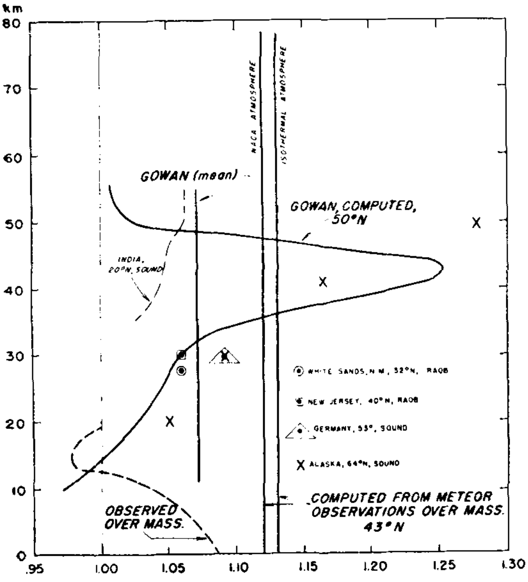

Fig. 4. Annual temperature ranges, observed and computed, expressed as ratios of the absolute temperature, summer to winter.

Using the well-known hydrostatic formula

$$p = p_0 e^{-\frac{mg}{R}\frac{z}{T_0}} \tag{1}$$

$$\varrho = \frac{mp_0}{RT_0} e^{-\frac{mg}{R}\frac{z}{T_0}} \tag{2}$$

where p is barometric pressure at height z,
p_0 is pressure at $z = 0$,
R is the universal gas constant,
g is the acceleration of gravity,
m is the molecular weight of air taken equal to 28.97, the sea-level value, since the composition of air at 70 km is the same as that at sea level (CHACKETT, PANETH, and WILSON, 1949),
ϱ is air density.

Now since average sea level pressure at Massachusetts is exactly the same (1016 mb) in summer as in winter, differentiate (2) logarithmically holding p_0 and z (78 km) constant, thus obtaining

$$\frac{d\varrho}{dT_0} = \frac{\varrho}{T_0}\left(\frac{z}{H_0} - 1\right) \tag{3}$$

where $H_0 = RT_0/mg$ is the height of the homogeneous atmosphere corresponding to surface temperature, T_0.

From (3) it is seen that $\frac{d\varrho}{dT_0} < 0$ for $z < H_0$

and $\frac{d\varrho}{dT_0} > 0$ for $z > H_0$

Since $H_0 \sim 8$ km and since Whipple's densities are given for an average height of $z = 78$ km, it is thus seen that $d\varrho/dT_0 > 0$. This means, if we accept Whipple's result that the summer density at 78 km is higher than that for the winter, that the temperature of the lowest 78 km of the summer atmosphere is higher than that of the winter atmosphere. Let us find this difference in temperature. Rewriting (3) as

$$\frac{dT_0}{T_0} = -\frac{d\ln\varrho}{1 - \frac{z}{H_0}} = -\frac{2.3 \cdot d\log_{10}\varrho}{1 - \frac{z}{H_0}} \tag{4}$$

and taking $d\log_{10}\varrho = 0.5$ at $z = 78$ km from Whipple's data, $T_0 = 264°$ K[1] ($H_0 = 7.72$ km), we find that $dT_0/T_0 = 0.1265$, $dT_0 = 33.4°$ and $T_{os}/T_{ow} = 1.13$, where T_{os}, T_{ow} are the summer and winter values, respectively, of the temperature of an assumed isothermal atmosphere extending from 0 to 78 km over Massachusetts.

In fig. 4 the ratio of 1.13 is drawn as a straight line from 0 to 78 km. Also in this figure are drawn two curved lines: (a) a dotted curve extending from 0 to 19 km over Massachusetts and showing the ratio T_s/T_w as computed from aerological observations (Anonymous, 1945); and (b) a solid curve showing the ratio T_s/T_w calculated from Gowan's computations (GOWAN, 1947 a) of the vertical temperature distribution satisfying radiative equilibrium for an atmosphere at 50° N characterized by 0.03 % by volume of CO_2, a constant water vapor mixing ratio corresponding to 10 % humidity at the tropopause, and the average distribution of ozone at that latitude, with winter having a greater ozone

[1] 264° K is the harmonic mean temperature of the NACA tentative standard atmosphere (WARFIELD, 1947 from 0 to 78 km. If the arithmetic mean (273° K) is used, then $dT = 36°$ and $T_{os}/T_{ow} = 1.14$.

content, 0.28 cm (NTP), compared to summer, 0.20 cm. In both seasons the atmosphere is considered to be irradiated by a 6,000° K black-body sun.

Gowan's computations were based upon division of the atmosphere into 5 km layers from 55 to 15 km, and a bottom layer from 15 km to 11 km (the latter figure being the average height of the tropopause at 50° N). He gives as his final result the mean temperature of each layer separately for winter and summer, under conditions of radiative equilibrium. It was a simple matter, then, for the present writer to draw a smooth curve through the ratios of these various mean temperatures to achieve the result shown in fig. 4.

The ratio, T_s/T_m, as found from Gowan's results, increases to unity at 14 km, a height 5 km lower than the increase found in the observed curve over Massachusetts. This discrepancy which may be due to a difference in the heights of the tropopause is not important for the present discussion. What is important, however, is that above a certain altitude, both curves, the computed and observed, increase to unity, an indication that at the higher altitudes summer temperatures are probably greater than winter temperatures.

Since some observational evidence has been found indicating that the ultraviolet intensity is lower than that expected from a 6,000° K sun, Gowan also computed the radiative equilibrium temperatures for a summer atmosphere irradiated by a 4,000° K sun. In this latter case the temperature at 50—55 km is 347° K, or 100° less than for the 6,000° K sun; while at 11—15 km the temperatures are almost equal, 215° K vs. 217° K (see Table II). Unfortunately, no computations are at hand for a winter atmosphere irradiated by a 4,000° K sun, so it is not possible to prepare a curve showing T_s/T_w against height for this case. It is likely, however, that the ratio curve would be quite similar to that for a 6,000° K sun.

Computed Annual Temperature Range for the NACA Atmosphere

Let us now perform new calculations involving use of the seasonal density variations observed by Whipple, and operating not in an isothermal atmosphere assumed previously but in an atmosphere whose temperature distribution is that of the NACA tentative standard atmosphere (WARFIELD, 1947). (See fig. 7 in this paper.)

Let T_m be the NACA temperature at a given height, z, and let T_s and T_w be the summer and winter temperatures, respectively, at that same height in the atmosphere over Massachusetts where it is assumed that

$$T_s = T_m + \frac{\Delta T}{2},\quad T_w = T_m - \frac{\Delta T}{2}, \tag{5}$$

ΔT is the annual range of temperature at z.

The hydrostatic relationships for the summer and winter atmospheres, respectively, are

$$p_s = p_{os}\, e^{-\frac{mg}{R}\int\limits_0^{78\text{ km}} \frac{dz}{T_s}},\quad p_w = p_{ow}\, e^{-\frac{mg}{R}\int\limits_0^{78\text{ km}} \frac{dz}{T_w}} \tag{6}$$

where p_s, p_w are summer and winter pressures at 78 km, and p_{os}, p_{ow} are summer and winter pressures at 0 km. Dividing equations (6) and eliminating p_s and p_w by the respective Equations of State we have

$$\frac{\varrho_s}{\varrho_w} = \frac{T_w}{T_s}\, e^{-\frac{mg}{R}\int\limits_0^{78\text{ km}} \left[\frac{1}{T_s} - \frac{1}{T_w}\right] dz} \tag{7}$$

where ϱ_s, T_s are summer values of atmospheric density and temperature, respectively, at 78 km,

ϱ_w, T_w are corresponding values for winter.

Now from (5)

$$\frac{1}{T_s} - \frac{1}{T_w} = -\frac{\Delta T}{T_m^2 - \left(\frac{\Delta T}{2}\right)^2} \sim -\frac{\Delta T}{T_m^2} \tag{8}$$

since $\Delta T \ll T_m$.

Substituting (8) in (7) and solving for the "mean value" of ΔT, $\overline{\Delta T}$, for the layer 0 to 78 km we have

$$\overline{\Delta T} = \frac{-\ln\frac{T_w}{T_s} - \ln\frac{p_{os}}{p_{ow}} + \ln\frac{\varrho_s}{\varrho_w}}{\frac{mg}{R}\int\limits_0^{78\text{ km}} \frac{dz}{T_m^2}} \tag{9}$$

Evaluating the numerator of this expression for the atmosphere over Massachusetts, we have seen that $\ln p_{os}/p_{ow} = 0$ and $\ln \varrho_s/\varrho_w = 0.5 \cdot 2.3 = 1.15$. For lack of definite information of the value of $\ln T_w/T_s$ we shall assume it to be near 0 since it is likely that at 78 km, where a minimum point in the temperature height curve occurs, there exists only a very small annual range of temperature because of lack of significant solar radiation absorption. However, even if T_s were 20 % larger than T_w, then $\ln T_s/T_w = 0.18$ which is much smaller than $\ln \varrho_s/\varrho_w = 1.15$.

The integral in the denominator of (9) is computed from the NACA tentative standard atmosphere. The final result is $\Delta \overline{T} = 29°$ C and

$$\frac{\overline{T}_s}{\overline{T}_w} \sim \frac{\overline{T}_m + \frac{\Delta \overline{T}}{2}}{\overline{T}_m - \frac{\Delta \overline{T}}{2}} = 1.12 \qquad (10)$$

The value of $\overline{T}_s/\overline{T}_w$ given above is drawn as a straight line in fig. 4. It is seen that this line lies slightly to the left of that computed for an isothermal atmosphere and is thus closer to the line showing the ratio of the harmonic mean temperatures from 11 to 55 km of Gowan's summer and winter stratospheres. However, the fact that the line does lie appreciably to the right of Gowan's mean indicates that the annual variation of temperature in the lower 78 km of an NACA atmosphere, using Whipple's observations, may be even larger than Gowan's computations show[1].

Comparison with Temperatures Determined by Sound Measurements

The annual variation of temperature in the ozonosphere described here is in sharp disagreement with results found by sound measurements over N. Germany (GUTENBERG, 1946). In figure 2 of Gutenberg's paper, curves are presented showing that in the layer 20 to 60 km, the summer temperatures are *lower* than those for winter, with the maximum difference of 90° C occuring at 40 km. It is likely that this is a false result, arising from failure to take into adequate account the wind difference between summer and winter in the European stratosphere.

Wind observations of smoke from artillery shell bursts at 100,000 ft. (30 km) taken over southeast England in 1944—45 (JOHNSON, 1946; MURGATROYD and CLEWS, 1949) show that there is a monsoonal reversal from westerly winds in winter (+ 37 m/sec.) to easterly in summer (— 13 m/sec.), a difference of 50 m/sec. Assuming that this seasonal difference in wind is characteristic of those winter and summer days on which sound measurements were made in Germany and which form the basis of Gutenberg's curves, it will be shown that the wind difference can more than account for the apparent excess temperature of winter over summer at 30 km, which in Gutenberg's figure 2 amounts to nearly 50° C.

In figure 3 of Gutenberg's paper the observed sound velocities are given for the layer 20 to 60 km; it is not clear whether these sound velocities were corrected for wind, and if so, how the winds were measured. At 30 km the winter sound speed is 37 m/sec. greater than the summer speed, a value, however, which is less than the 50 m/sec. wind speed seasonal difference. This may signify that the greater winter sound speed observed at 30 km need not necessarily be attributed to higher temperature, but may very readily be caused by wind, provided that Gutenberg's sound speed measurements were made to the east of the sound source during both winter and summer. Information on this latter point could not be found.

Let us use the above data in finding T_s/T_w at 30 km over western Europe, T_s and T_w being the absolute temperature in summer and winter, respectively. The velocity of sound, according to Gutenberg, is given by $20.1 \sqrt{T}$ m/sec. Now according to our assumptions, the observed velocity of sound is given by a summation of the above temperature term and a wind speed. The winter-summer

[1] From additional meteor observations the seasonal density variation turns out to be $\Delta \log_{10} \varrho = 0.39$ instead of 0.5 (see Harvard College Observ. and Center of Analysis, MIT Tech. Rep. No. 2, *Ballistics of the upper Atmosphere*, L. G. Jacchia, p. 17, Cambridge, Mass. 1948). This will decrease $\Delta \overline{T}$ to 23° and T_s/T_w to 1.09, which is thus very close to Gowan's value.

difference in observed sound velocity is then given by

$$W_w - W_s + 20.1\,(\sqrt{T_w} - \sqrt{T_s}) = 37 \text{ meters/sec.}$$

where $W_w - W_s = 50$ meters/sec., the observed winter-summer wind difference.

Substituting this value in the preceding equation, we have

$$\sqrt{T_s} - \sqrt{T_w} = 0.65,$$

showing that the summer temperature at 30 km is greater than the winter temperature.

If we take $T_w = 218°$ K (the NACA value), then we find $T_s/T_w = 1.09$, which when plotted in fig. 4 falls slightly to the right of Gowan's curve. Again, because of various uncertainties, the close agreement should not be taken too seriously, but at least it shows that the annual temperature range as determined by sound measurements over Germany is not irreconciliable with our earlier conclusions if account is taken of the wind in the manner described above.

After the original manuscript of this paper was sent to the Editor, additional observations of temperature at great heights came to the writer, and have been plotted on the original figure 4: (1) "Winds and Temperatures in the Lower Strathosphere" by C. J. Brasefield, *Journal of Meteorology*, February 1950, pp. 66—69; (2) „Investigation of Stratosphere Winds and Temperatures from Acoustical Propagation Studies" by J. A. Peoples, Jr and A. P. Crary which will be published shortly in the *Journal of Meteorology*; (3) "Reflection of Sound Waves from the Stratosphere over India in Different Seasons of the year" by L. S. Mathur, *Indian Journal of Meteorology and Geophysics*, Vol. 1, No. 1, January 1950, pp. 24—34. The Brasefield data included two typical soundings, one made in July 1948 (his figure 5). These two soundings were taken as representatve of summer and winter conditions, respectively, for New Jersey. The second set of observations was taken at 64° latitude over Alaska in the summer of 1949 and winter of 1948. The Indian observations were made at latitude 20° N during the period 30 May—2 June 1946, 16—20 October 1946, and 29 January—1 February 1947. The temperatures for the first period were considered to be representative of summer while those of the third period were taken as winter temperatures. The October temperatures were slightly ($\sim 3°$ C) warmer than the winter temperatures. An attempt was made to eliminate the wind effect on temperatures found from the Alaskan and Indian sound observations by having the recording stations arrayed on perpendicular lines.

These additional data are plotted on fig. 4, and form some interesting comparisons with the material originally there. The new Jersey point falls directly on the Gowan curve. The Alaskan data show a steady increase of the temperature ratio, T_s/T_w, with height to value of 1.28 at 50 km. The Indian curve starts from a much smaller value of T_s/T_w and increases very slowly with height to a value of 1.06 at 50—55 km.

Comparison of the data in fig. 4 for low, middle, and high latitudes reveals much of interest. Taking the observations as representative of their latitudes, it is seen that above 20 km, at all latitudes, the temperature ratio increases with height, but much slower at 20° latitude than at 64° latitude. At 50 km the temperature ratio at 64° latitude is larger (1.28) than at 20° latitude (1.06). A similar increase northward is noted for the surface temperature ratio, 1.02 at latitude 20° and 1.14 at 64° (Hann-Süring, 1939, p. 180). The cause of these parallel variations is the same, namely, a much greater increase in insolation from winter to summer at higher latitudes than at lower latitudes. For example, at the bottom of an atmosphere having a transmission coefficient of 0.8, the ratio of the direct solar radiation received for the warm half of the year to the cold half of the year is 1.4 at 20° latitude and 10 at 64° latitude (Milankovitch, 1930, p. A64).

Although we have been dealing with comparisons of the temperature ratio T_s/T_w, the same conclusions would apply qualitatively to the annual temperature range, $T_s - T_w$. For example, at 50 km the annual range at 64° N latitude in Alaska is 65° C while at 20° N latitude in India, it is 20° C. At the surface the corresponding ranges are 41° C and 6° C, respectively.

The increasing positive temperature difference, $T_s - T_w$, with increased latitude is consistent with the change in winds observed above 17 km in middle and high latitudes from westerly in the winter to easterly in the summer.

Although the summer-winter temperature ratios are larger in the ozonosphere than they are at the surface, this is not the case for the heat energies because of the much smaller densities aloft. For example, the ratio of the heat energy gained from winter to summer in the ozonosphere (pressure at base taken as 50 mb) to that gained by the whole atmosphere below is 0.10 at latitude 64° and 0.22 at latitude 20°.

Diurnal Variation of Temperature

The evidence regarding the diurnal variation of temperature in the upper atmosphere is less consistent than that concerning the annual variation. Ballard found from serial sounding balloons that at 25 km over Nebraska in July the air temperature drops 10° C in the 8 hours from 6 p.m. to 2 a.m. and less than 2° C at 16 km (BALLARD, 1941). It is not possible to assess the significance of these findings since only one sounding a day was taken and not much was known about radiational effects on the early temperature measurements in the stratosphere. There is however some indirect evidence based on infrared spectroscopy regarding the diurnal temperature variations in the ozonosphere. Adel by measuring the absorption in the ozone layer of 9.6 μ radiation coming from the sun at day and from the moon at night, and making a separate measurement of the 9.6 μ emission of the ozone layer alone was able to determine "ozonosphere" temperatures for several months in the spring of 1948 at Alamogordo, New Mexico (ADEL, 1949). Only one of his series of measurements went from night to the following day, thus revealing partial information on the diurnal variation of temperature in the ozone layer. The observations are shown in Table 1.

Although the temperature varies from — 37° to — 41° C there is no systematic trend indicating a diurnal tendency. In the same table for comparison is shown Gutenberg's summary (GUTENBERG, 1949) of the White Sands, New Mexico, radiosonde observations at the highest elevation, 27.5 km, capable of

Table 1.
Adel's Ozonosphere temperature over Alamogordo, New Mexico
(Observed by radiation measurements at 9.6 μ)

Date	Time	Temperature
4/24/48	23: 43 Mst	— 39° C
	23: 55	— 41
4/25/48	00: 10	— 41
	12: 15	— 39
	12: 35	— 37
	12: 53	— 41

Gutenberg's Summary of White Sands, New Mexico, Raob data at 27.5 km
(April 1947—May 1948)

Month	Jan./Febr.	March	April
Temperature	—48 ± 2° C	—50 ± 4	—44 ± 6
No. Obs.	6	7	24
Month	**May**	**July**	**September**
Temperature	—38 ± 7	—37 ± 6	—42 ± 4
No. Obs.	7	12	5

summarization. White Sands is 20 miles southwest of Alamogordo. The average monthly temperatures and standard errors are shown, and it is seen that Adel's late April data fit in very nicely between the April and May means. This good agreement brings up the question as to what atmospheric layer Adel's temperatures really refer to. Since most of the ozone is found between 15 and 30 km, one might think that Adel's temperatures should be indicative of the mean temperature of the layer 15 to 30 km. However, the White Sands mean seasonal temperature-height curves plotted in fig. 1 of Gutenberg's paper show that the mean temperature of this layer should be — 50° C or about 10° colder than Adel's values.[1] On the other hand the close agreement between Adel's temperatures and the White Sands mean temperatures at 27.5 km may be

[1] Average value of Adel's temperatures is —40° C, with extremes of —33° and —48°C. An independent comparison by the writer of the White Sands soundings taken within a few hours of Adel's measurements verified the relative warmth of Adel's temperatures.

interpreted as an indication that Adel's values refer to the layer 25 to 30 km.

The White Sands temperatures at 27.5 km indicate a definite annual variation, March being coldest and July warmest of the months shown. The ratio of the absolute temperatures[1] of these two months is 1.06, which when plotted on fig. 4 falls almost exactly on the curve drawn from Gowan's computations described previously. In view of the large standard error of the temperatures at these extreme altitudes where data are sparse, too much significance should not be given to this close agreement.

Returning to the diurnal temperature variation it is indicated by the single case from Adel that the variation, if any, is very small in the layer from 25 to 30 km. This result is in agreement with Gowan's computations of the nocturnal cooling of the ozonosphere shown in fig. 5.

Two cooling curves are given in this figure: A and B. They both apply to the 8-hour cooling (summer night) of a stratosphere of constant mixing ratio corresponding to 10 % relative humidity at the tropopause, and containing 0.280 cm of ozone. The initial temperature distribution of each stratosphere varies, however. Stratosphere A has the initial temperature distribution corresponding to radiative equilibrium under a 6,000° K sun while Stratosphere B had the initial temperature distribution corresponding to radiative equilibrium under a 4,000° K sun. These initial temperature distributions are shown in Table II.

Table II. Initial Vertical Temperature Distributions, Gowan, Penndorf (in °K)

Ht. in km	Gowan A	Gowan B	Ht. in km	Penndorf
50—55	448	347	50	300
45—50	424	333	47	280
40—45	397	319	41.5	261
35—40	332	301	36	253
30—35	295	282	30.5	250
25—30	272	266	25.5	242
20—25	249	245		
15—20	232	229	17	233
11—15	217	215		

[1] Solar radiational errors were eliminated to within 1° C by use on many of the White Sands radiosondes of the new unshielded white-coated thermistor (Brasefield, 1948), so that the annual variation indicated is believed to be close to that of the true air temperature.

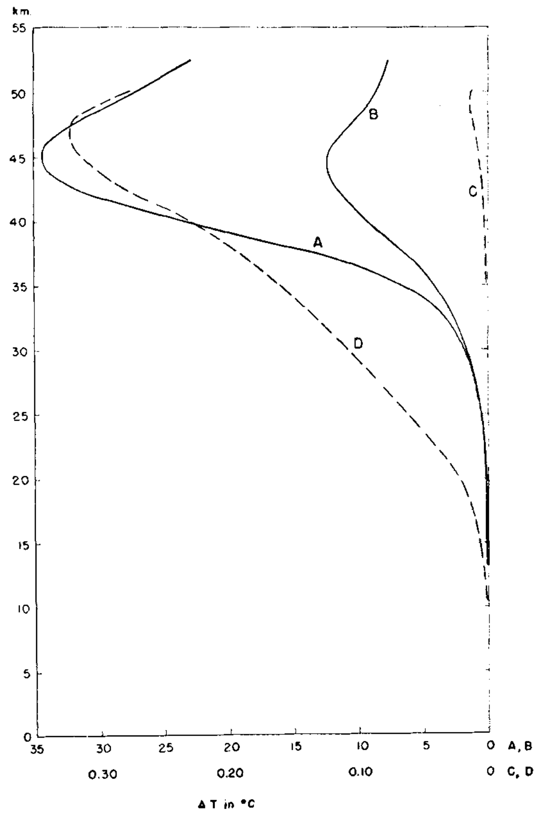

Fig. 5. Full curves: Ozonosphere cooling after an 8 hour summer night (Gowan). Curve A applies to a warmer initial atmosphere than curve B. Upper scale in abscissa applies to these curves.
Dashed curves: Daily cooling (curve C) and heating (curve D) in the ozonosphere (Penndorf). Lower scale in the abscissa applies to these curves.

The cooling curve for Stratosphere A was taken from Table 1, Column 1 of Gowan's paper (GOWAN 1947 b) while the cooling curve for Stratosphere B was obtained in a private communication from Gowan (7 March 1950). They both show cooling of 1° C at 25—30 km, a result in good agreement with Adel's observations.

The lack of appreciable diurnal temperature variations at 25—30 km may also be considered in good agreement with Penndorf's cooling rates (PENNDORF, 1936) computed in a different manner from Gowan and also shown in fig. 5 as Curve C. This cooling curve represents the cooling in ° C per day for an atmosphere containing 0.26 cm of ozone of vertical distribution corresponding to the summer atmosphere over Arosa (lat. 45° N). The vertical temperature distribution on which Penndorf based his cooling calculation shown in Curve C, increases from 233° K at 17 km to 300° K at 50 km.

The two cooling curves of Gowan and Penndorf, while agreeing in shape and showing relatively little cooling in lower levels, depart widely aloft; in the layer 40 to 50 km where both computations show the maximum values, the cooling rates differ by a factor of more than 1,000. Many reasons may be advanced to account for this major discrepancy; for example, both computations are quite sensitive to the vertical distribution of ozone, water vapor and temperature. The two distributions of ozone adopted are apparently not much different. Although Gowan took into account the loss of heat in the infrared bands of water vapor and carbon dioxide, Penndorf did not; however, Gowan states that the effect of the latter gas is negligible. Most important appears to be the effect of the vertical temperature distribution assumed at the beginning of the cooling period; the results of two differing temperature distributions (Gowan A and B, Table II) are illustrated by comparing curves A and B in fig. 5. As may be seen from Table II, Penndorf's temperatures are appreciably lower than Gowan's throughout the layer 30 to 50 km. A cooler atmosphere of course should be expected to cool less rapidly than a warmer atmosphere, other things being equal. But it is not likely that this difference in initial temperature distributions can account for the large difference in cooling rates. Even if Penndorf had assumed an initial temperature of 500° K at 50 km, his cooling would still be 100 times smaller than Gowan's value at that height as explained below.

In fig. 5 is plotted Curve D which is the *heating* curve in °C per day for the same Penndorf atmosphere described previously and irradiated by a 5,910° K sun at lat. 45° on June 22. The heating values shown by Curve D are much larger in magnitude than the cooling values shown in Curve C, a result which indicates lack of radiative equilibrium. However, for radiative equilibrium to exist, Penndorf states that a temperature of at least 500° K would be required at 50 km (loc. cit. p. 267). This would then mean that at 500° K Penndorf's cooling rate would be equal to the heating rate, but would still be too small compared to Gowan's results by a factor of at least 100.

In absence of a more positive explanation of the discrepancy, we may try to see which of these two theoretical results agrees better with observations. The observations will be of two types: Those based on what is believed to be known of the annual temperature variations at the top of the ozone layer (40 to 55 km) and those based on the temperature lapse-rate between 55 and 80 km.

Referring to Gowan's nocturnal cooling values in fig. 5, it may first be said that for his results to be physically reasonable, the daytime heating values must be of the same magnitude, but of opposite sign, as the cooling values; for otherwise the air at these levels would be rapidly cooling or heating, and would soon reach impossibly low or high values.

The net warming up from winter to summer occurs as the days grow longer and the nights shorter, thus allowing a small positive temperature gain to accumulate each day. Adopting the Gowan computations — both seasonal and diurnal — one would expect at 40 to 45 km an annual temperature cycle of maximum range of 75° C on which is superimposed daily oscillations of range at least 10 to 12° C. In contrast to Gowan's values, Penndorf's results if plotted on a similar graph would show practically no diurnal temperature variation at any place in the ozonosphere. Let us estimate from the full value of Penndorf's daily heating rate the maximum annual range of temperature to be expected in the layer 20 to 50 km. Penndorf's average daily heating for this layer at the spring equinox is about 0.12° C per day (loc. cit. p. 266). In 180 days (i.e., from midwinter to midsummer) this layer would then warm about 22° C. From the earlier discussion on annual temperature variations, as illustrated in fig. 4, it appears likely that the average annual range for the layer 20 to 50 km is larger than 22°. Thus even if non-radiative heat dissipation, such as advection and convection, are neglected and the *full* amount of Penndorf's daily radiative heating value is used, the expected annual temperature variation cannot be accounted for. It appears from this consideration that Penndorf's heating rates are too low.

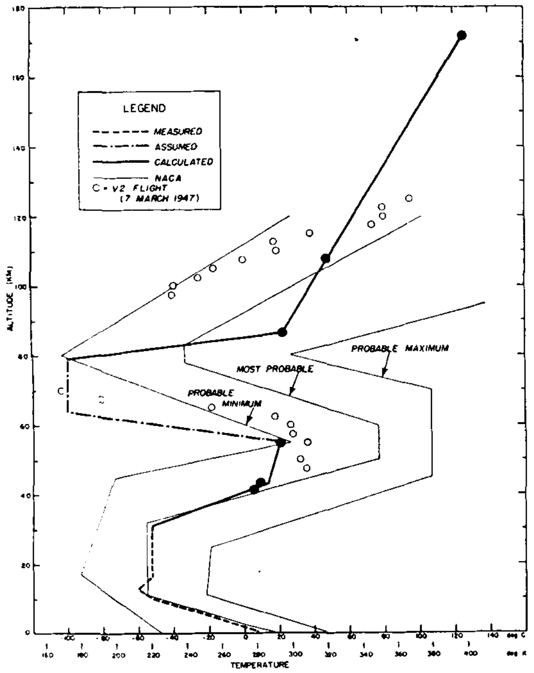

Fig. 6. Comparison of measured upper atmosphere temperatures (Helgoland blast), V-2 rocket results, and NACA tentative standard (after Cox).

But what proof have we that Gowan's large daily temperature oscillations of magnitude at least 10° C exist at the top of the ozonosphere, from 40 to 55 km? In other words, is it not possible for the winter to summer warming of the top layers of the ozonosphere to occur as a day-to-day accumulation of small differences of two much smaller quantities, say 1/10th of Gowan's values as shown in fig. 5. Arithmetically, of course, this solution is possible, but there is additional evidence which lends some support to the magnitude of Gowan's cooling values.

If a diurnal temperature variation of at least 10° C is observed at 55 km and zero temperature variation exists at 25 km, this would mean that the temperature inversion found between 30 and 55 km, and illustrated in fig. 6 (Cox, 1948), would be strengthened by day and weakened by night, but remain as an inversion day and night. What can one say of the layer from 55 to 80 km? First, it appears that in the upper portion of this layer, say, from 70 to 80 km, there is relatively little known absorption of solar radiation. Below 70 km absorption by ozone heats the air with the maximum heating between 40 and 55 km, as described previously. Thus the layer 55 to 80 km is analogous to the layer of air above the ground which is heated by insolation at day, forming a steep, or even a superadiabatic lapse rate in the surface layer, and cooled at night, forming an inversion. The absorbing layer at the top of the ozone layer is not nearly so sharply bounded as the ground in its absorptive properties, but the analogy is fairly good. Thus, if during the daytime heating period the temperature at 55 km should increase by at least 10° while the temperature at 70 to 80 km remains unchanged, one might expect that a steep, possibly even a superadiabatic lapse rate, might be established in the lower portion of this layer.

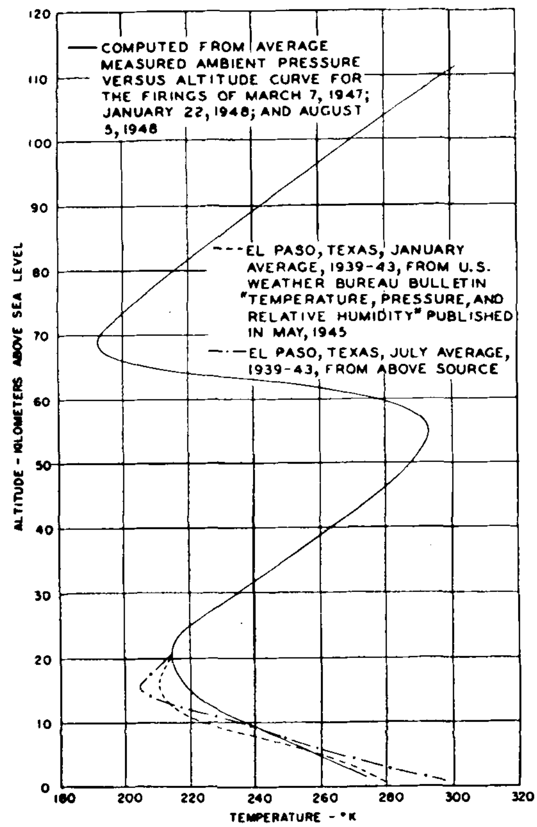

Fig. 7. Relation of temperature to altitude above White Sands, New Mexico (after Newell).

To support the contention of the existence of a steep or superadiabatic lapse rate in the layer 55 to 80 km, we can cite temperature measurements made by the V-2 rocket soundings released at 1123 MST on 7 March 1947, at 1313, 22 January 1948, and 1837, 5 August 1948 at White Sands, New Mexico; the mean sounding of which is shown in fig. 7 (NEWELL, 1950). These observations show a temperature fall of 75° C between 60 and 66 km, or a superadiabatic lapse rate amounting to 125% of the dry adiabatic rate. For this lapse rate to become adiabatic, the temperature at 60 km would have to be decreased by 15° C. No night rocket soundings at which air temperatures have been measured are available for comparison with those made at day. Cox, in his determination of stratospheric temperatures from the Helgoland blast of 1,100 GCT, 18 April 1947, was forced to assume the existence of a superadiabatic layer from 57 to 62 km to render his data interpretable (fig. 6).

It should be pointed out that there exists considerable uncertainty in the determination of the air temperatures from the V-2 rockets. Because of the high speed of the rocket, temperatures are not measured by a temperature element, but by two methods, not independent, which are based on air pressure. The first method uses the slope of the *log p* vs. altitude curve to compute the mean temperature of 5 or 10 km thick layers. The second method is to determine the density from the speed of the rocket and the ram pressure (pressure measured at the nose of the rocket), and then by use of atmospheric pressure to find the air temperature by the gas law. For a further discussion of the fairly large probable errors (~20° C) involved in the temperature values, see a paper soon to be published in the *Journal of Geophysical Research* (HAVENS, KOLL, LA GOW, 1950) and kindly loaned to the writer by the authors.

In concluding this section on diurnal temperature variations, it may be stated that while evidence points to practically zero diurnal temperature change in the lower part of the ozonosphere (up to 30 km), evidence is not so convincing as to the magnitude of the change in the upper portion of the ozonosphere (from 30 to 55 km). It is likely that Penndorf's values are too low and that the much higher values of 10 to 15° C, as deduced from Gowan's computations and considerations of lapse rate and annual temperature range, are nearer the truth. It would be most helpful to the study of the diurnal temperature variations in the ozonosphere if serial day and night rocket soundings could be made.

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