Night cooling of the ozonosphere

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Some estimate of the cooling of the ozonosphere can be made using the calculated solar energy absorbed in various layers. After sunset this source of energy is removed and cooling occurs mainly by radiation up and down. The conclusion is reached that during a night the cooling is about 30° C at 50 km. and only a fraction of a degree below 25 km. This means that the ozonosphere temperature inversion around 40 km. will persist throughout the night.

Ballard (1941) finds that at about 25 km. height during July over Nebraska the air temperature drops by about 10° C during the eight hours from 6 p.m. to 2 a.m., and ordinarily by less than 2° C around 15 km. On the other hand, Penndorf (1936) obtained a much smaller diurnal variation of temperature in the stratosphere, the maximum being 0-3° C per day around 50 km. height.

The amount of radiation energy being absorbed by various layers of the ozonosphere was obtained by Gowan (1936, 1947) during calculations of temperature on the assumption of radiation equilibrium. At sunset one source of this energy is cut off but the others remain. The temperature should therefore fall during the night, and an estimate of how much can be made with the numerical values available. The results which follow lie generally between the two sets of data referred to above, for comparable heights.

The assumptions of two conditions—radiation equilibrium before sunset, and that sunset is instantaneous—gives a preliminary basis for calculation. Let the mass of air in a 1 sq.cm. column of any layer be \( m \) g.; let the specific heat at constant pressure be \( c_p \); let \( R \) be the net radiation from the layer in g.cal./sq.cm./hr.; and let \( dT \) degrees be the fall in temperature in \( dt \) hours. Then the conditions are represented by the equation

\[
-mc_p dT = 2Rdt.
\]

There is no net exchange of radiation horizontally in any layer, so \( 2R \) is the total net radiation from a sample column. This may be considered constant as a first approximation, or may vary according to several plausible laws. Below are outlined five different ways of treating this net radiation.

(1) \( 2R = \) a constant, namely, the solar energy which was being absorbed, and after sunset must be radiated, no account being taken of the change in temperature of the layer. Thus \( \frac{dT}{dt} = -\frac{2R}{mc_p} \); and \( t(\frac{dT}{dt}) \) gives the temperature change in \( t \) hours.

(2) \( 2R = \beta T \), where \( \beta \) is a constant, and \( T \) is the absolute temperature of the layer \( t \) hours after sunset. Thus \( -mc_p dT = \beta T dt \), which gives on integration

\[
\log_{10} T = \log_{10} T_0 - \frac{0.434\beta}{mc_p}.
\]
(3) \(2R = \beta(T - T_s)\), where \(T_s\) is the estimated effective temperature of the other adjacent layers. This is Newton's law of cooling if \(T_s\) is constant, as here provisionally assumed. The actual decrease in \(T_s\) during the night is small because it mostly represents the lower layers containing the bulk of the absorbing material, and these are found to change relatively little in temperature. The factor \((T - T_s)\) tends to remain large if \(T_s\) decreases with \(T\), and the calculated cooling would be larger than if \(T_s\) is constant. By considering \(T_s\) constant some compensation is obtained for the maximizing effect of the assumptions of instantaneous sunset with midday equilibrium temperature. By substitution \(-mc_p dT = \beta(T - T_0)dt\), which gives

\[
\log_{10}(T - T_s) = \log_{10}(T_0 - T_s) - \frac{0.434 \beta t}{mc_p}
\]

on integration

\[
\beta = \frac{2R}{(T_0 - T_s)}.
\]

(4) \(2R = \beta(T - T_\infty)^n\), where \(T_\infty\) is the equilibrium temperature for night, i.e. with no solar energy reaching the layer. Substituting in the original equation, and using \(\beta = 2R/(T_0 - T_\infty)^n\) gives

\[
(T - T_\infty)^{n-1} = \frac{mc_p(T_0 - T_\infty)^n}{(n - 1) 2Rt + mc_p(T_0 - T_\infty)}
\]

on integration and simplification. This formula was applied in two cases: \((a) n = 2, \) \((b) n = 4.\)

The data used were from the calculation of temperature under radiation equilibrium for midsummer at latitude 50 with 0.280 cm. of ozone, and 10% relative humidity at the tropopause. The absorption of all constituents was corrected for pressure. The cooling in 8 hr. was found by each of the five methods outlined above. The results are summarized for comparison in table 1. The largest values come from method (1) and the smallest from (4b). The variations, however, are less significant than the general agreement of the order of magnitude, tens of degrees above 35 km. and fractions of a degree below 25 km.

<table>
<thead>
<tr>
<th>Table 1. Drop in Temperature (°C)</th>
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The actual course of the temperature change will be a small, slow cooling from shortly after midday to sunset, followed by a more rapid cooling from sunset to sunrise. The night rate will be smaller than the smallest in the table, but with the start of cooling during the afternoon the final result from, say, 1 p.m. to sunrise may not be very different from the values in the table.

From sunrise to about 1 p.m. the ozone layers are warming up again. During spring the period of warming is lengthening, and that of cooling shortening, producing an all-round increase in temperature. During autumn the reverse process takes temperatures down to their winter values. The cooling during the 16 hr. of a midwinter night was calculated by (4b). This was plotted along with the corresponding summer cooling for comparison as shown in figure 1. Maximum absorption of solar energy in ozone occurs lower down in summer when the sun’s rays have a short path through the atmosphere. This gives more energy to be radiated after sunset, and hence a maximum in the estimated cooling around 45 km. The winter curve would probably show a similar maximum at some greater height than 55 km, where the ozone concentration is small and unknown, making calculation impossible. Möller (1935) has calculated the heat balance of the free atmosphere over Lindenberg in June up to 10 km. The result at 10 km, is about 0·5°C cooling per day, or about 0·2°C per 8 hr., and if extrapolated to 13 km. would be less than 0·1°C per day. Tanck (1940) calculated the heating effect due to absorption of sunlight, and found a similar value. Both these observers agree that the temperature change decreases with height in the high troposphere, and then increases again in the upper stratosphere. Elsasser (1942) also discusses this question and comes to a similar conclusion through the use of his radiation chart.

\[ AT \] in °C

**Figure 1.** Graph of ozonosphere cooling.
Baur & Phillips (1934, 1935, 1936) have studied the heat and radiation balance in the atmosphere. This was done for clear skies, and also including the effects of turbidity and cloudiness, for summer, winter and equinoctial seasons. The distribution with height is quite different with and without cloud, but the values agree at about 10 km., and are substantially the same as Möller’s figure.

Barbier, Chalonge & Vassy (1939) discuss the spectroscopic evidence for a low temperature of the ozonosphere at night. The method is based on the contrast in the absorption bands of ozone, and naturally refers to the region where the bulk of the ozone is found, i.e. below 30 km. Vassy & Vassy (1939) also discuss the question, and show that the day and night readings are not different within the precision of measurement. At greater heights Bartels (1936) has estimated that the change in temperature between day and night around 200 km. is certainly less than 100°C.

All this auxiliary evidence is confirmatory, or at least not contradictory, within its range. The calculations in this paper show less than 2°C per night under 30 km., yet only up to 30°C per night around 50 km. The direct evidence of a temperature change from day to night is sparse and contradictory. It seems from the work of Ballard (1941) that the insolation error during daytime flights is not large, but it may be important when the differences are required for evaluation of the cooling. There are known to be large and relatively sudden changes in the height of the tropopause, and consequently in the temperature of the lower stratosphere. These have no connexion with day and night, but are the result of air mass movements which seem to extend some way into the stratosphere. The great discrepancies which appear in the average observations of different observers, are mainly due to geography and climate. It seems certain that the cooling of the lower stratosphere, perhaps up to 30 km., is not governed by radiation. Above 30 km. it is more probable that radiation plays a dominant role in any diurnal temperature changes, though winds have been observed during observations of noctilucent clouds and meteor trails.

If the cooling of the lower stratosphere depends on convection and advection it follows that radiation equilibrium may never be attained, as assumed in the original calculations (Gowan 1947). For those it was assumed that there is sufficient mixing to keep the composition nearly constant, but not enough convection to interfere with the attainment of radiation equilibrium. The agreement between calculated temperatures and balloon observations is good for the lowest layer, 11 to 15 km., but around 15 to 25 km. is only fair, the calculated temperatures being 10°C too high. This disagreement might be due to any one or all of the following factors: (1) Radiation equilibrium may not be attained, hence the observed temperature would never be as high as the calculated temperature. (2) Insufficient water vapour may have been assumed for these layers, since for higher amounts the calculated temperatures are lower. (3) If more carbon dioxide were assumed its extra radiation would lead to a lower calculated temperature. (4) The relatively weak ozone band at 14µ was neglected. This overlaps with both carbon dioxide and water vapour but could have more effect from 15 to 25 km. than from 11 to 15 km., due to the
maximum concentration of ozone around 22 km. Some adjustment among these factors might be found by trial which would give agreement at all heights observed directly. It seems likely that more observations of composition at greater heights may soon be available from experiments with rockets. Then perhaps fewer trial calculations will be necessary to reach an agreement over all the layers within the ozonosphere.

Although these results are tentative it seems clear that even the temperature change above 35 km. is not sufficiently great to destroy the ozonosphere temperature inversion during a summer or a winter night. Hence the existence of a temperature inversion is a valid explanation of the minimum in number of meteor disappearances, and of the abnormal audibility of explosions extending throughout the night.

References

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