Forecasting Minimum Temperatures

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(Manuscript received 14 March 1951)

Abstract

Because of the fact that too many factors influence the nocturnal cooling of the ground surface, simplifications are necessary in order to derive a convenient formula giving the amount of cooling from sunset to sunrise. These simplifications must be justified both theoretically and practically. The solution is finally found by semiempirical methods. For the practical application of this cooling formula a diagram is constructed. A test of the results using observations for Vienna shows a good agreement between the predicted and observed temperature minima.

Introduction

The importance of calculation of minimum temperatures and the prediction of ground frost for agriculture, especially for the so-called frost districts is obvious. For this purpose a vast number of more or less empirical formulae have been produced by various authors. For a discussion of empirical formulae for forecasting the minimum temperature the reader is referred to E. S. ELLISON (1928). Some details can also be found in the textbook by GEIGER (1942). Other formulae and diagrams have been developed from BRUNT'S (1939) expression for the net loss of heat by radiation from ground [see DUFOUR (1938), JACOBS (1940) and LÖNNQVIST (1946)], but actually these formulae have not given a satisfactory solution of the problem. One difficulty is that a particular formula may give excellent results in one locality but not in another. The whole problem is made complex by the many factors which influence the nocturnal cooling. If we consider all these factors in a mathematical treatment the solution, if any can be given at all, would be too complex for practical purposes. If we oversimplify the problem on the other hand we can not expect accurate results.

Theory of nocturnal cooling of the earth's surface

The conditions which favour nocturnal cooling of the ground surface are well known from experience. These conditions are;

a) Clear skies,
b) Absence of wind,
c) Low vapour pressure in the atmosphere, and
d) Low thermal conductivity and specific heat of the ground.

All these factors have to be considered in a mathematical treatment for calculating the drop of temperature at the ground during the night.

A first theoretical approach is due to BRUNT (1939). His well known formula can be written as follows;

[1] Most of the work for this study was done during a visit to the Training and Research Section of the Canadian Meteorological Service at Toronto under an UNESCO fellowship granted by the Canadian Council for Reconstruction.

$$\Delta T_0 = \frac{2}{\sqrt{\pi}} \frac{E}{\sqrt{k_g \varrho_g c_g}} \sqrt{L} \qquad (1)$$

where ΔT_0 is the amount of the cooling at the ground surface from sunset to sunrise during a night of length L. E is the effective outgoing radiation, which is assumed to be constant during the cooling process and k_g, c_g and ϱ_g are the coefficient of heat conductivity, the specific heat and the density of the soil respectively. It is supposed that k_g, c_g and ϱ_g are constant both with respect to time and with depth. In order to derive equation (1) the effect of transport of heat by eddies from the atmosphere to the ground must be neglected. A further necessary simplification is that we consider only the vertical transport of heat and that we assume an isothermy in the ground at sunset as an initial condition.

Although some of these simplifications are rather restrictive and exclude many practical cases, BRUNT's formula has been used with remarkable success for the construction of frost prediction diagrams as has been stated before. The assumption, E = constant, seems to be a rather close approximation, at least for the beginning of the cooling process as was shown by SAUBERER (1936) by direct measurements. If the length of the night increases it might be necessary to consider a variation of E. As GROEN (1947) pointed out it is not too difficult to allow for a linear variation of E. For practical purposes however the assumption, E = constant, can be justified as long as no advection takes place and there is clear sky or no change in the cloudiness. In order to evaluate E we can use empirical formulae which give in general a close fit to the observations obtained by different authors. It is also possible to calculate the downward flux of the atmospheric radiation by means of a radiation chart whenever a radiosonde ascent is available. For different seasons in 1949 CZEPA and REUTER (1950) have evaluated the effective outgoing radiation in Vienna using ELASSER's radiation chart. A comparison of these results with values obtained by the empirical formula of ÅNGSTRÖM showed that the latter frequently overestimated E by as much as 20%. The same result was obtained by SAUBERER (1936) from measurements using a radiometer designed by ALBRECHT. In a recent study, LÖNNQVIST (1950) came to the same conclusions. On the other hand ROBINSON (1950) was able to show that values of E obtained by means of ELSASSER's chart are in general too low compared with measurements. In practice usually no measurements of E at sunset are available so that an empirical formula or a diagram has to be used. The assumption of dealing only with a vertical transport of heat in the ground and of considering k_g, c_g and ϱ_g as constants can be justified in general since only relatively small layers of the ground are affected by the nocturnal cooling.

To assume isothermal conditions in the ground at sunset seems to be rather arbitrary, but actually it has been shown by several investigations (BRACHT 1949) that near isothermy is observed in the highest layers of the ground at sunset. It is however not difficult to allow for a linear variation of temperature in the ground as will be seen later. In most cases the resultant cooling of the surface is only slightly modified.

As may be seen from (1) the cooling of the surface depends essentially upon the soil constants. Therefore we must expect a variation of ΔT_0 from one place to the other. Since furthermore the conductivity of heat changes to a high degree with the water content of soil, ΔT_0 at a given station will depend upon the moisture conditions in the soil. Since the value of k_g in snow is very small we get a much larger cooling at a snow surface than on ground even if the radiation is the same. For details the reader is referred to a paper by the author (REUTER 1948).

As we have seen most of the assumptions made by BRUNT in deriving equation (1) can be justified. As BRUNT has stated, equation (1) cannot be a final solution of the problem as long as we neglect the effect of heat transport from the air to the ground. Therefore the diagrams based upon (1) can be used for clear calm nights only. Following a study by PHILIPPS (1940) the author (REUTER 1947) has introduced the eddy transport of heat in the atmosphere into the theory. In so doing the whole problem becomes more complex and further simplifications are necessary. The solution becomes relatively simple under the assumption that the coefficient of eddy conductivity may be considered as constant.

If we allow for a linear variation of the temperature T as an initial condition both in the ground and in the atmosphere and if we make the same simplifications as regard to E and the soil constants as before, our problem can be expressed by the following differential equations, where z is the vertical coordinate and t the time.

$$z < 0 \quad \frac{\partial T_1}{\partial t} = \frac{k_g}{c_g \varrho_g} \frac{\partial^2 T_1}{\partial z^2} \text{ soil} \qquad (2\text{ a})$$

$$t = 0, \; T_1 = T_i - B(-z);$$

where B is the lapse rate of temperature in the soil and T_i the temperature of the surface at sunset.

$$z > 0 \quad \frac{\partial T_2}{\partial t} = \frac{A}{\varrho} \frac{\partial^2 T_2}{\partial z^2} \text{ atmosphere} \qquad (2\text{ b})$$

$$t = 0, \; T_2 = T_i - \gamma z;$$

where γ is the lapse rate of temperature in the atmosphere at sunset and ϱ the density of the air which is assumed to be constant.

The boundary conditions are;

$$(T_1)_{z=0} = (T_2)_{z=0} \text{ for any time} \qquad (3\text{ a})$$

and

$$E + k_g \left(\frac{\partial T_1}{\partial z}\right)_{z=0} - c_p A \left(\frac{\partial T_2}{\partial z} + \gamma_d\right)_{z=0} = 0; \qquad (3\text{ b})$$

where A is the coefficient of eddy conductivity in the air ($A = K\varrho$ where K is the coefficient of eddy diffusivity), c_p is the specific heat of the air and γ_d the dry adiabatic lapse rate. The condition (3 b) states simply that the sum of the effective outgoing radiation, the flow of heat from below to the surface and the flow of heat from the atmosphere to the surface is equal to zero. The solution of (2 a) and (2 b) with (3 a) and (3 b) can be found following a study by the author (REUTER 1947) in which the solution of a similar system of equations with slightly different initial and boundary conditions was given.[1] For the surface it yields;

$$\Delta T_0 = T_i - T_0 =$$

$$= \frac{2}{\sqrt{\pi}} \frac{E + B k_g + (\gamma - \gamma_d) c_p A}{\sqrt{k_g \varrho_g c_g} + c_p \sqrt{A \varrho}} \sqrt{L}, \qquad (4)$$

where the meaning of the different quantities is the same as in equation (1). Putting $A = 0$ and $B = 0$ in (4) we arrive at BRUNT's formula (1).

The assumption under which (4) is derived, namely A = constant, is only a rough approximation. Attempts were made by J. G. JAEGER (1945) and R. FROST (1948) to allow for a variation of A in the form of a power law according to SUTTON's well known theory. The treatment then becomes so complicated that it cannot be successfully used for practical purposes. As JAEGER pointed out, it is only possible to describe in this way the temperature distribution in atmospheric layers near the ground assuming that the surface temperature is kept at a temperature given by equation (4). Starting from isothermal conditions both in the ground and in the atmosphere, R. FROST neglects the transport of heat from below to the ground surface which seems to be in many cases a rather crude approximation. Furthermore the assumption of an increasing A with height holds only for layers near the ground; in higher levels the eddy coefficient will again decrease. Actually, in many cases, layers up to several hundred meters are influenced by the nocturnal cooling of the earth's surface. The author (REUTER 1947 and 1948 b) was able to show that, with the assumption A = constant, the formation of the ground inversion can be described sufficiently. Therefore it seems to be more convenient for practical purposes to take a mean value of A estimated from the observed horizontal wind velocity. As PEASCHKE (1937) showed, the value of A at a certain level can be expressed as a linear function of the horizontal wind velocity. A final justification of the simplifying assumption A = constant can be given only by a verification in practice as it will be seen later.

[1] In the previous investigation the third term in (3 b) was assumed to be $-c_p A \left(\frac{\partial T_2}{\partial z}\right)_{z=0}$. The flow of heat from the atmosphere to the surface however depends on the gradient of the potential temperature i.e. $\frac{\partial \Theta}{\partial z} = \frac{\partial (T_2 + \gamma_d z)}{\partial z} = \frac{\partial T_2}{\partial z} + \gamma_d$, so that the third term of the boundary condition (3 b) may be included in its present form.

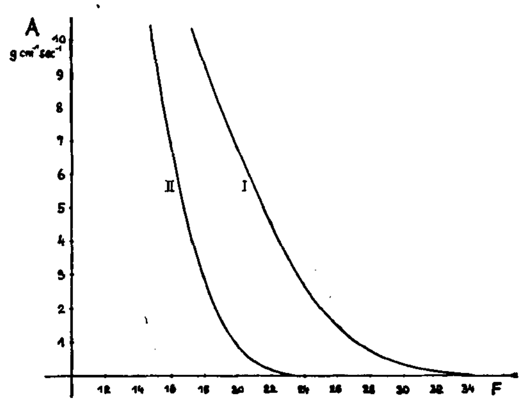

Fig. 1. The characteristic function F plotted against the eddy coefficient A. Curve I — normal ground; Curve II — after remarkable rainfall.

For a detailed study of the *temperature distribution* near the ground the assumption of a constant coefficient of eddy conductivity is of course not a close approximation. We are mostly interested however in the drop of temperature at the *surface* during the night so we may use equation (4) with some hope of success. Before we deal with practical applications of equation (4) it must be emphasized that the method outlined above gives only the static drop of temperature from sunset to sunrise and cannot include temperature changes produced by advection; upslope or downslope motion; and the effect of freezing or melting processes, evaporation and the formation of dew.

Practical application of the cooling formula

In order to apply equation (4) for practical purposes it is convenient to separate three factors on the right side. Equation (4) may then be written;

$$\Delta T_0 = F \cdot [E + Bk_g + (\gamma - \gamma_d)\, c_p A] \cdot \sqrt{L}, \tag{5}$$

where F is given by

$$F \equiv \frac{2}{\sqrt{\pi}} \frac{1}{\sqrt{k_g \varrho_g c_g} + c_p \sqrt{A \varrho}}. \tag{6}$$

We will denote the characteristic function of the ground as F and plot F against A for given soil constants (Fig. 1). The function F varies from place to place and depends, to a certain extent, on the water content of the soil. The curves shown in Fig. 1 are plotted, using soil constants given in the textbok of Meteorology by HANN-SÜRING (1937). The effect of water content on the variation of heat conductivity in the ground has not yet been investigated sufficiently, although some progress has been made in recent studies by BRACHT (1949) and ALBRECHT (1948). Therefore Curve II in fig. 1 represents only a rough approximation showing how F will change after a remarkable rainfall.

In practice however we usually do not know the value of the soil constants. Therefore it might be possible and convenient to find the proper function F for a particular place and for certain ground conditions by using the observed values of ΔT_0 and computing F as will be shown later. Furthermore we can try to plot F against the horizontal wind velocity rather than against the mean value of A.

The diagram in fig. 2 shows $\Delta T_0 = F^\star \sqrt{L}$ for different seasons and for different values of

$$F^\star = F \cdot [E + Bk_g + (\gamma - \gamma_d)]\, c_p A \tag{7}$$

using for L the length of the night for a latitude of 48 degrees. As the radiation is usually measured in cal per cm² per minute we will for convenience express the term $[E + B \cdot k_g + (\gamma - \gamma_d)\, c_p A]$ in this unit. If furthermore we express the length of the night L in hours we express F in the same units as in Fig. 1 (i.e. the c.g.s.system).

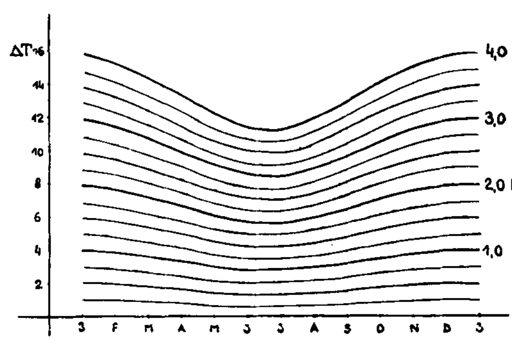

Fig. 2. Seasonal variation of nocturnal cooling for various values of $F^\star$ (latitude 48 degres).

In order to find the amount of nocturnal cooling at the surface in the course of a night we have to start with fig. 1. We obtain the amount of F for a certain value of A. Multiplying F by the observed or calculated effective outgoing radiation E (expressed in cal/cm² min), which should be corrected by the usually small quantity $Bk_g + (\gamma - \gamma_d)\, c_p A$, we can immediately read off from the diagram (Fig. 2) the possible drop in temperature for the particular case.

The order of magnitude of $(\gamma - \gamma_d)\, c_p A$ can be illustrated by assuming relatively high values. Let

$$\begin{aligned} A &= 10 \text{ g cm}^{-1}\text{ sec}^{-1} = 600 \text{ g cm}^{-1}\text{ min}^{-1} \\ \gamma &= 0 \\ \gamma_d &= 10^{-1} \text{ deg cm}^{-1} \\ c_p &= 0.24 \text{ cal } g^{-1} \text{ deg}^{-1} \end{aligned}$$

then $\gamma_d c_p A = 0.014$ cal cm^{-2} min^{-1} whereas E is of the order of 0.10—0.15 in the same units. Therefore $(\gamma - \gamma_d)\, c_p A$ is negligible unless the cooling period is very long.

The order of magnitude of $B \cdot k_g$ can be illustrated by assuming values for B and k_g as obtained by BRACHT (1949). Let

$$\begin{aligned} k_g &= 0.0018 - 0.0025 \text{ cal cm}^{-1}\text{ sec}^{-1}\text{ deg}^{-1} = \\ &= 0.11 - 0.15 \text{ cal cm}^{-1}\text{ min}^{-1}\text{ deg}^{-1} \\ B &= 0.1 - 0.2 \text{ deg cm}^{-1}, \end{aligned}$$

then $B \cdot k_g = 0.011$—0.015 which is of the same order of magnitude as $\gamma_d c_p A$. In some cases it would be necessary to consider this term if observations of the temperature in the ground were available. If a ground inversion exists at sunset γ becomes negative. This means that the cooling is diminished. For a detailed discussion the reader is referred to REUTER (1947).

Summarizing the results of a practical application of the cooling formula we come to the following conclusion; because of our lack of knowledge of the physical constants of soil we have to use first of all the observational material to find the function F for a particular place. We may then use a graph like fig. 1 and a diagram like fig. 2 to predict the temperature fall (we will illustrate this method in the next section which deals with a practical test of the theoretical formula).

Practical test of the theory

In order to test the theory outlined above for a particular case, temperature observations for Vienna have been used. According to the strict theory, the ground surface temperatures should be used; in many cases however the amount of ΔT_0 at the screen level will not differ too much from that at the surface.[1] The difference between air and ground surface temperature at sunset may occasionally be noticeable, especially in summer.

Table 1. Nocturnal cooling for days with clear skies at Vienna — March, August and September 1949

Date	Temperature difference Sunset to Sunrise	F^*	E cal/cm² min.	F	Mean wind velocity km/h
March					
9—10	4.9	1.37	0.174	7.9	16.3
21—22	3.5	1.00	0.169	5.9	24.1
23—24	7.9	2.33	0.179	13.0	10.0
24—25	10.4	3.08	0.178	17.3	3.7
25—26	10.0	2.93	0.180	16.3	7.8
26—27	12.0	3.55	0.187	19.0	4.1
27—28	9.8	2.90	0.184	15.8	5.1
August					
6— 7	8.3	2.75	0.150	18.3	4.3
7— 8	10.1	3.33	0.154	21.6	2.6
Sept.					
24—25	6.7	1.68	0.152	11.1	11.6
25—26	8.3	2.42	0.169	14.3	8.0
26—27	10.5	3.00	0.158	19.0	3.5
27—28	9.5	2.62	0.151	17.3	3.1
28—29	10.0	2.92	0.153	19.1	2.5

In Table 1 a selection of observations for nights with clear sky conditions only, in different seasons in 1949, are shown. The amount of F is obtained from fig. 2 with the known value of ΔT_0 (the night time cooling). The effective outgoing radiation was computed with the aid of ÅNGSTRÖM's formula;

$$E = \sigma T_0^4 \,(0.194 + 0.236 \cdot 10^{-0.069\, e_0}) \quad (8)$$

[1] From the complete solution of equations (2 a) and (2 b) the author (REUTER 1948 b) came to the conclusion that for $A \geqq 1$ g. cm^{-1} sec^{-1} the temperature difference between screen and surface must be considered as almost constant during the night. Observations of the temperature difference between screen and surface for Vienna (to be published later) show *in the mean* over a long period the same result.

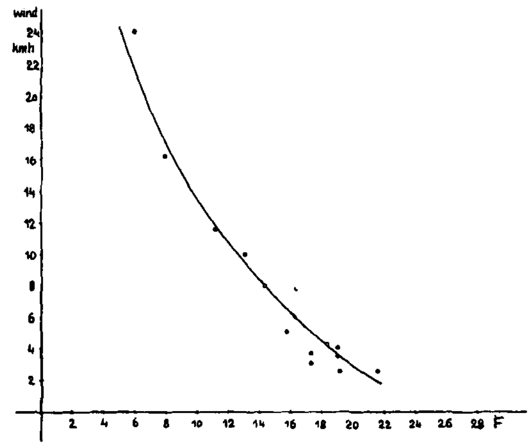

Fig. 3. The characteristic function F plotted against wind velocity. Data are for Vienna for clear nights, March—September 1949.

where σT_0^4 is the black body radiation in cal cm^{-2} min^{-1} and e_0 is the vapour pressure in mm. The value of E may also be obtained graphically using the temperature and relative humidity at sunset.[1]

The value of E obtained from (8) might be incorrect as has been pointed out previously; also it should be corrected by the factor $B \cdot k_g + (\gamma - \gamma_d)\, c_p A$. This factor is neglected here. The error made by too high a value of E is however not too serious so long as we exclude observations where the wind velocity is either too high or is changing in direction (advection). It is important that we use in any one period the same method for evaluating E [for instance equation (8)]. Naturally the best way to obtain E is by direct measurement, but unfortunatly such measurements were not available.

We have used the observational material of 1949 to find the function F. The result is illustrated in fig. 3. It must be emphasized that for convenience the wind velocity is expressed in km/h whereas F remains in the same units as in fig. 1. As is apparent from the graph, a close relation exists between wind velocity and the function F. fig. 3 together with the diagram (fig. 2) enables us to predict minimum temperatures for any other period. We have only to read off the amount of F from fig. 3 for a given wind velocity. Multiplying this value with the effective outgoung radiation E, evaluated from (8) in cal/cm² min, we obtain immediately from the diagram in fig. 2 the possible drop in temperature for the particular case. The advantage of this semiempirical method is that the graph of fig. 3 can be modified by continually using the most recent observations. A graph for F should be obtained from observations for both dry and wet ground conditions. Then the value of F used in the forecast could be adjusted

[1] A convenient diagram for computing E can be found in the November issue 1948 of the Periodical *Wetter und Leben* (Zeitschrift für praktische Bioklimatologie, Vienna)

Table 2. Forecast and observed temperature minima at Vienna — August and September 1950

Date	Temp. Sunset C°	Vapour pressure Sunset mm	Cloudiness	E (formulae 8 and 9) cal/cm² min.	Mean wind velocity km/h	Forecast temperature Sunrise C°	Observed temperature Minima C°
August							
20—21	20.4	9.4	clear	0.151	3.0	11.1	12.7
21—22	22.3	9.9	clear	0.158	2.5	12.8	13.1
22—23	25.7	10.0	clear	0.159	2.0	15.0	15.3
23—24	25.8	13.8	clear	0.146	2.0	16.0	15.7
24—25	26.6	12.4	clear	0.151	10.2	20.8	19.2
27—28	24.8	13.3	3/10 As	0.118	2.5	17.3	17.8
28—29	25.6	15.4	clear	0.142	2.0	16.1	16.4
30—31	21.2	11.4	3/10 As	0.115	2.9	14.0	14.0
31— 1	23.5	10.3	4/10 Cs	0.134	4.6	15.7	15.5
September							
3— 4	13.5	8.3	4/10 As	0.109	15.0	12.2	12.0
4— 5	14.5	8.7	2/10 As	0.125	13.2	10.6	10.2
10—11	20.9	14.8	9/10 St	0.066	10.6	18.1	17.8
11—12	22.0	12.6	clear	0.139	1.2	12.0	14.2
13—14	20.2	9.8	clear	0.144	2.7	10.7	11.5

in individual cases by considering the moisture conditions in the ground.

The result of the prediction of minimum temperatures for a period in the summer 1950 can be read off from Table II. Excluded are those nights when precipitation occurred or when the wind velocity was too high. The cloudiness has been considered using a formula due to DORNO, namely;

$$E_w = E\,(1 - k\,w), \tag{9}$$

where E_w is the observed net loss of radiation from the ground when w tenths of the sky are covered with clouds. E being the net loss of radiation (effective outgoing radiation) to a clear sky under the same conditions of temperature and humidity given by equation (8). The factor k has the following values; for Cs — 0.031, for As — 0.063, for St — 0.085 and for Ns — 0.099.

Table 2 shows that the objective method outlined in this study gives rather accurate results. As has been mentioned before, the results of Table 2 might be used to modify the graph of fig. 3 before we commence predictions for another period. Again it must be emphasized that the function F depends essentially upon the special features of a particular place. Therefore the graph of fig. 3 can be used for Vienna only. A similar application for another place was given by the author in a Technical Circular of the Canadian Meteorological Service (1950), using data for Toronto.

Acknowledgments

The author wishes to express his gratitude to Dr Andrew Thomson, Controller of the Canadian Meteorological Service, for many suggestions and his interest during the preparation of the study. Also sincere thanks are due to Mr C. M. Penner of the Training and Research Section of the Canadian Meteorological Service for his help and advice during the course of the work.

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