CHAPTER 17

TEMPERATURE

CHAPTER 17

TEMPERATURE

Contents Page

LIST OF DIAGRAMS

Figure | Page

CHAPTER 17

TEMPERATURE

17.1 RADIATION - GENERAL LAWS

All matter which is not at the absolute zero of temperature emits energy in the form of electromagnetic waves into surrounding space. Radiation affecting the earth and its atmosphere originates from the sun, the ground (or sea) or the atmosphere itself. In order to understand the effects of radiation on meteorological problems it is necessary to have a knowledge of the main laws of radiation. These laws are stated briefly below without any elaboration of their derivation or general properties. If a more complete treatment is required the reader should consult the reading references quoted at the end of this chapter.

A body which emits, for all wavelengths, the maximum amount of radiation possible at a given temperature is said to radiate as a 'black body'. The amount of energy, $E_\lambda \, d\lambda$, radiated each second per unit area of a black body in the portion of the spectrum between the wavelengths λ and $\lambda + d\lambda$ is given by Planck's Law:

$$E_\lambda = \frac{c_1 \lambda^{-5}}{\exp(c_2/\lambda T) - 1} \qquad \ldots \quad (17.1)$$

where $c_1 = 3.741\,832 \times 10^{-16}\ \mathrm{J\,m^2\,s^{-1}}$ and $c_2 = 1.438\,786 \times 10^{-2}\ \mathrm{m\,K}$.

E_λ is known as the 'spectral radiant exitance' (or, formerly, as the 'spectral radiant emittance'). The form of the variation of E_λ with λ is shown in Figure 1.[1] The total amount of radiation emitted each second by unit area of the body is given by the total area under the curve,

$$\int_0^\infty E_\lambda \, d\lambda = \sigma T^4 \qquad \ldots \quad (17.2)$$

This is Stefan's Law; T is the temperature in kelvins and σ is a constant, equal to $5.670\,32 \times 10^{-8}\ \mathrm{W\,m^{-2}\,K^{-4}}$. The wavelength, λ_{max}, at which E_λ is greatest is given by

$$\frac{dE_\lambda}{d\lambda} = 0$$

giving

$$\lambda_{max} = \frac{c_3}{T} \qquad \ldots \quad (17.3)$$

where c_3 is a constant. Equation 17.3 is Wien's Law. If λ_{max} is in units of μm and T in kelvins, then $c_3 = 2898$.

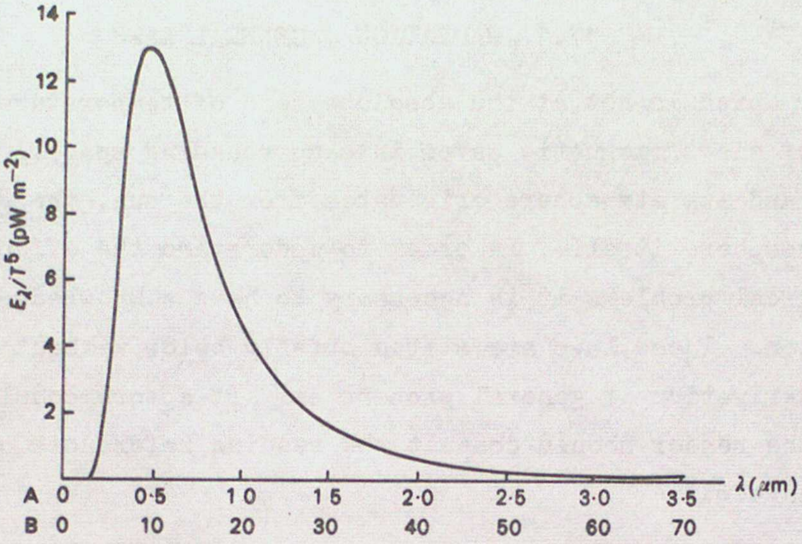


FIGURE 1. Theoretical distribution of black-body radiation
A: $T = 6000$ K B: $T = 300$ K

Matter both absorbs and emits radiation. The absorptivity, $\alpha_{\lambda T}$, of a body is defined as that fraction of incident radiation which it absorbs. The emissive power, $e_{\lambda T}$, of a body is the energy radiated from unit surface in unit time. The subscript λT has been used to indicate the dependence of absorptivity and emissive power on both wavelength and temperature. Kirchhoff's Law states that at a given temperature the ratio of emissive power to absorptivity $e_{\lambda T}/\alpha_{\lambda T}$ for a given wavelength is a constant for all bodies. It follows from Kirchhoff's Law that a poor absorber of radiation of a given wavelength is a poor emitter. Also a good absorber is a good emitter and a black body absorbs all incident radiation. If the ratio $e_{\lambda T}/E_\lambda$ is a constant for a body for all wavelengths that body is known as a 'grey' body, that is, it absorbs and emits a constant fraction of black-body radiation for all λ. For most substances absorptivity varies with wavelength and temperature. This variation is an important property of gases. For incident radiation of a given wavelength a gas may be nearly completely transparent, that is, it absorbs, and by Kirchhoff's Law emits, practically no radiation of that wavelength. Yet for radiation of a neighbouring wavelength (or fairly narrow range of wavelengths) the gas may be nearly opaque, that is, it may act almost as a black body absorbing all incident radiation and emitting as a black body in those wavelengths. Some of the gaseous constituents of the atmosphere exhibit these selective properties to a marked degree and thereby exert a fundamental control on the radiation balance of the earth.

Beer's Law of absorption relates the intensity of radiation I after traversing a layer containing mass m to the incident radiation I_0 by the equation

$$I = I_0 \exp(-k_\lambda m) \qquad \qquad (17.4)$$

where k_λ is known as the absorption coefficient which varies with λ.

With a knowledge of these laws it is possible to deduce some of the effects of radiation on the earth and its atmosphere.

17.2 SOLAR RADIATION

The prime source of energy for the earth and atmosphere is solar radiation. For the purposes of this Handbook the sun may be assumed to radiate as a black body at about 6000 K, although there are some differences from the black-body curve, mainly in the ultra-violet, arising from absorption within the outer layers of the sun itself. From Figure 1 it can be seen that almost all the energy (99 per cent) is contained with the region 0.15 to 4 μm, and about half the total energy is in the form of visible light, that is between 0.4 and 0.8 μm. In accordance with Wien's Law, the wavelength of the maximum intensity of solar radiation is at about 0.5 μm. The flux of solar radiation through unit area, perpendicular to the beam, at the mean distance of the earth from the sun, is known as the 'solar constant'. Its value is difficult to measure or estimate accurately, and it may in fact vary slightly, but it probably lies between 1.353×10^3 and $1.395 \times 10^3\ \mathrm{J\,m^{-2}\,s^{-1}}$ ($\approx 5\,\mathrm{MJ\,m^{-2}\,h^{-1}}$). The amount of radiation falling on a horizontal surface of unit area at the outer limits of the earth's atmosphere varies with latitude and with season: at latitude 50°N the daily amount varies from about 7.5 MJ $\mathrm{m^{-2}}$ at the winter solstice to about 42 MJ $\mathrm{m^{-2}}$ in late June, with values of about 24 MJ $\mathrm{m^{-2}}$ at the equinoxes. (See 'Smithsonian Meteorological Tables' [2] for further details.)

On passage through the earth's atmosphere, some of the solar radiation is absorbed, scattered and reflected before the remainder reaches the ground as the 'direct' solar beam. Absorption of solar radiation is the main factor that determines the temperature profile throughout almost the whole depth of the atmosphere. By the time the beam has penetrated to about 100 kilometres radiation of wavelength less than about 0.2 μm has almost all been absorbed, leading to the high temperatures observed in this region, the thermosphere. Below this level and above the ozone layer is a region in which comparatively little absorption takes place and temperatures are low, with a minimum known as the mesopause, at around 80 kilometres. Between 70 and 35 kilometres, ozone absorbs strongly, particularly for wavelengths between 0.22 and 0.29 μm leading to the high temperatures of the upper stratosphere and lower mesosphere, the maximum temperature being found at the stratopause, at about 55 kilometres. Ozone also absorbs some radiation in the near ultra-violet and visible ranges of the spectrum, resulting in some heating of the layer between 15 and 35 kilometres, but in the region around the tropopause, absorption, mainly by water vapour and carbon dioxide, is again weak and temperatures are low. Although absorption by water vapour becomes increasingly important with decreasing height in the troposphere, the most important effect by far is the absorption of solar radiation by the earth's surface, leading to the upward transport of sensible and latent heat by turbulence and convection.

On its passage through the earth's atmosphere, a proportion of the energy in the solar beam undergoes scattering by density fluctuations not much larger in scale than the constituent molecules and by suspended matter in the atmosphere. For particles much smaller than the wavelength of the light, the amount of energy scattered is inversely proportional to the fourth power of the wavelength (Rayleigh scattering), but for larger particles the laws are more complicated: the theory has been worked out by Mie.[3] Some of the energy so scattered is lost into space; about 5-10 per cent of the incident energy reaches the earth's surface as diffuse solar radiation.

On a clear day in the latitudes of the southern half of the British Isles about 50-70 per cent of the energy passing through unit area at the top of the atmosphere reaches the surface as the direct plus diffuse components of solar radiation. Figure 2 shows, month by month, the solar radiation incident in one day on unit horizontal surface at the top of the atmosphere,[4] and the global (direct plus diffuse) radiation received on unit area of the surface during a very clear day. The latter curve is derived from radiation data for Kew for the years 1950-54, taking hourly radiation values for complete hours of sunshine and estimating for each part of the year the maximum radiation likely to be received during each hour.

Solar radiation is further depleted by reflection from clouds and from the earth's surface. Recent satellite observations (Vonder Haar and Suomi[5]) have shown a mean albedo in the region of the British Isles of about 35-40

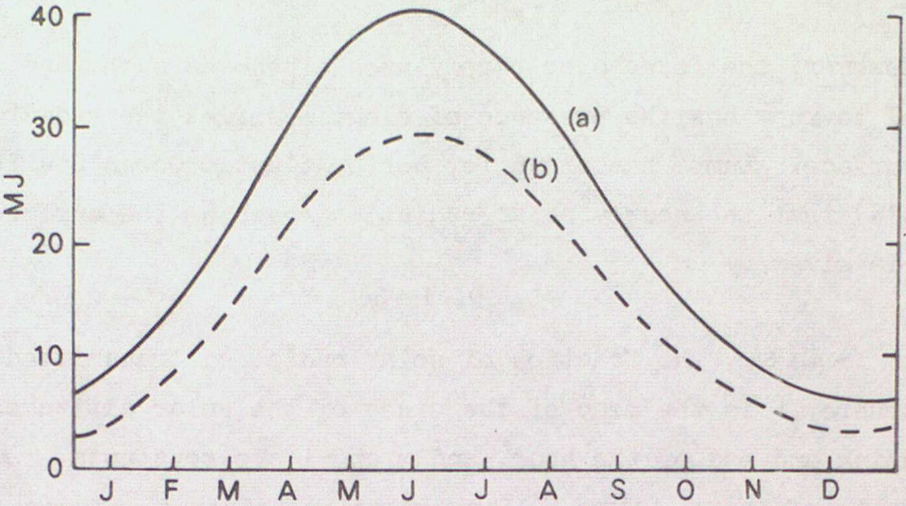


FIGURE 2. Total solar energy incident on one day on one square metre: (a) at the top of the atmosphere, and (b) at the surface on a clear day

Data for southern England (Kew)

per cent, the most important contribution to the mean albedo coming from clouds. A rough indication of the albedo of various types of surface and of cloud is given below in Table 17.1 (adapted from Table 154, 'Smithsonian Meteorological Tables'.[2]

TABLE 17.1 Albedo of various surfaces

Land Type	Land Albedo %	Water (direct sun only) Z	Water Albedo %	Cloud Type	Cloud Albedo %
Desert	24-28	0 (i.e. sun overhead)	2.0	Stratus overcast 0-150 m thick	05-63
Fields (green, wheat etc.)	03-15	20	2.1	Stratus overcast 150-300 m thick	31-75
Fields (dry, ploughed)	20-25	40	2.5	Strato overcast 300-600 m thick	59-84
Grass (various conditions)	14-37	50	3.4	Stratocumulus overcast	56-81
Ground (bare)	10-20	60	6.0	Altostratus, occasional breaks	17-36
Mould (black)	08-14	70	13.4	Altostratus overcast	39-59
Sand (dry)	18	80	34.8	Cirrostratus overcast	44-50
Sand (wet)	09	85	58.4	Cirrostratus and altostratus overcast	49-64
Snow or ice	46-86	90	100		

Z = sun's distance in degrees from the zenith

However, the forecaster rarely uses albedo as such, and it is more useful to know how the presence of cloud modifies the radiation receipt at the surface. Lumb[6] has shown for North Atlantic Ocean Station 'J' (52°30'N, 20°00'W) that the hourly solar radiation reaching the surface on cloudy days is given by

$$Q = 135 f s$$

where $f = a + bs$, the fraction of solar radiation transmitted through the atmosphere, s is the mean of the sines of the solar altitude at the beginning and end of the hour, and a and b are constants. Lumb assumes the value of 1.353×10^3 W m^{-2} for the solar constant. Figure 3 summarizes the results of this study; the results showed good agreement with observed values from North Atlantic Ocean Station 'A' (62°00'N, 33°00'W).

The values of a and b used to calculate f are given opposite in Table 17.2.

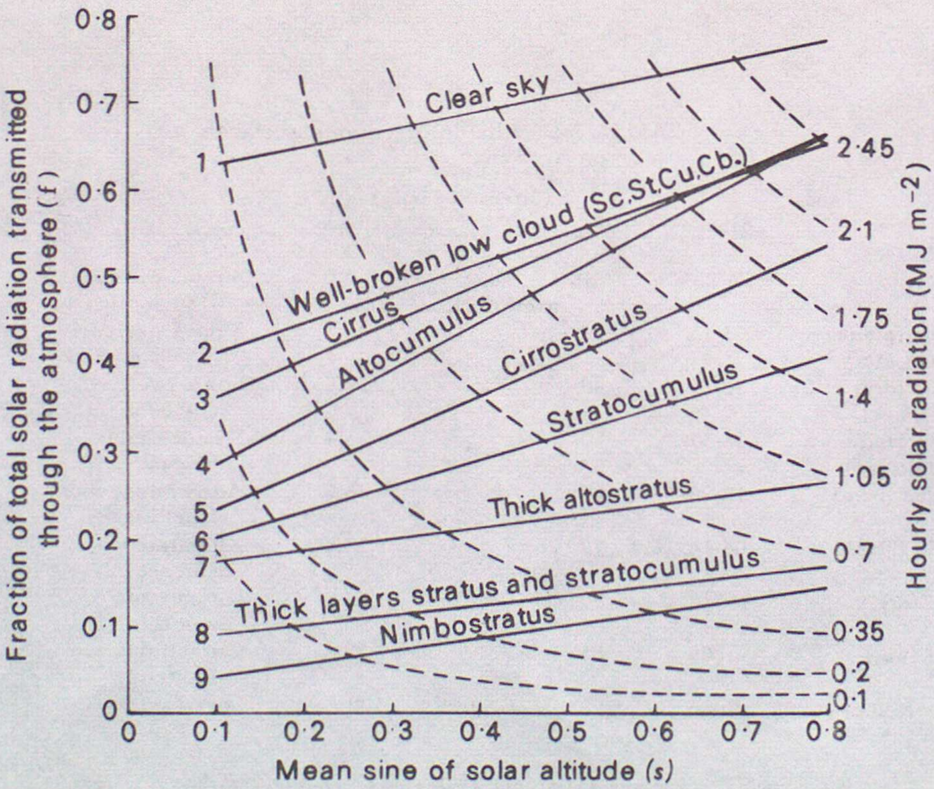


FIGURE 3. Isopleths (curved lines) of the hourly amounts of total solar radiation Q

The straight lines represent the empirical relationships between the fraction of solar radiation transmitted through the atmosphere (f) and the sine of the solar altitude (s) for nine categories of cloud type and amount.

TABLE 17.2

Category	Cloud	a	b
1	Clear sky	0.61	0.20
2	Well-broken low cloud, little or no medium or high cloud	0.38	0.34
3	Large amounts of cirrus cloud	0.32	0.43
4	Thin layers of medium cloud	0.24	0.45
5	Veil of cirrostratus covering whole sky	0.19	0.43
6	Large amounts of stratocumulus with little or no medium cloud	0.17	0.30
7	Thick layers of medium cloud	0.16	0.13
8	Thick layers of low cloud perhaps with layered medium cloud (usually accompanied by drizzle)	0.08	0.11
9	Thick layers of low cloud, probably also thick layers of medium cloud (usually accompanied by rain)	0.03	0.14

Values of the solar elevation for use in deriving s can be read off from the diagrams in Table 170 of the 'Smithsonian Meteorological Tables'.[2]

The values of hourly radiation derived from Lumb's formulae agree reasonably well with those observed at Kew, but are some 30 per cent higher in winter months, when atmospheric pollution may have had a marked influence in the period during which the Kew data were gathered.

17.3 TERRESTRIAL AND ATMOSPHERIC RADIATION

The surface of the earth radiates very nearly as a black body over a range of wavelengths appropriate to its temperature; the distribution of energy within its radiation, assuming the average temperature of the earth's surface to be 300 K, is given in Figure 1 if scale B is used. It can be seen that almost all the terrestrial radiation, as it is called, lies within the range 3 to 100 μm, and that there is very little overlap with the solar radiation. This enables a convenient nomenclature to be used, solar radiation being termed short-wave radiation, terrestrial radiation being termed long-wave radiation. The amount of energy radiated by unit area of the earth's surface at 300 K is about 6×10^{-6} of that radiated by unit area of the sun's surface. For a radiator at 200 K, the intensity of radiation is reduced still further by a factor of about 5.

The absorption properties of the constituents of the atmosphere for terrestrial long-wave radiation are widely different from those for short-wave radiation. Carbon dioxide and water vapour are the active absorbers of terrestrial radiation. Carbon dioxide has a wide band of strong absorption between about 13 and 17 μm. Within the wavelengths of terrestrial radiation the absorption by water vapour is very complex. Water vapour has strong absorption bands between 5.5 and 7 μm and from about 14 μm upwards. It is partially absorbent within the ranges 7 to 8.5 μm and 11 to 14 μm. Between 8.5 and 11 μm there are some very weak absorption lines but water vapour is virtually transparent in this range. These bands span the region where by far the greater part of the energy in terrestrial radiation is contained and it is clear that carbon dioxide and water vapour will have a considerable effect on terrestrial radiation. Simpson[7] advanced the understanding of the effect of long-wave radiation in a series of classical researches.

One of Simpson's hypotheses was that a column of air containing 0.3 millimetres of precipitable water is capable of absorbing completely terrestrial radiation in the wavelength from 5.5 to 7 μm and from 14 μm upwards. For air with a dew-point of 10°C and temperatures between 10 and 20°C, this amount of water is present at ground level in a column of air about 30 metres in depth. Thus the lowest layer of the atmosphere, even with cloudless skies, absorbs a considerable part of terrestrial radiation and radiates in these wavelengths back to the earth. It is only that part of terrestrial radiation to which the atmosphere is wholly or partly transparent which leaves the earth's atmosphere and returns to space. This is the so-called 'greenhouse effect' and it is of fundamental importance in maintaining the temperature of the earth. If the atmosphere were completely transparent to terrestrial radiation it has been calculated that the earth would be in general equilibrium with incoming solar radiation if its temperature was about 240 K. Thus the absorption bands of water vapour are vitally important to life on the earth.

The effect of cloud on long-wave radiation is also very important. Cloud layers can be regarded as black-body radiators and absorbers. A cloud layer of adequate thickness may be regarded as a perfect black body in these wavelengths and will absorb all radiation emitted from the ground and re-radiate as a black body at its own absolute temperature. This is usually lower than the ground temperature but for low clouds the temperature difference will be only small. Thus the presence of layer cloud has a moderating effect on ground temperatures. By day the cloud sheet prevents a large part of solar radiation from reaching the ground and so reduces the day maximum temperature. The same cloud returns to the earth a considerable proportion of the outgoing terrestrial radiation both by day and by night.

During the night when solar radiation is absent this returned radiation considerably decreases the radiative cooling, possibly even to zero, and so prevents the temperature from falling to the value it would reach with clear skies.

Another factor which must be considered is radiation from the atmosphere. Carbon dioxide and water vapour absorb long-wave (and, to a much more limited degree, short-wave) radiation. By Kirchhoff's Law they are also emitters of radiation of the same wavelength. Thus the atmosphere itself radiates. Considering a 'layer' of infinite lateral extent it will radiate both downwards and upwards to its neighbouring lower and upper layers. These layers in turn will radiate to each adjacent layer and so on until some radiation reaches the ground. At the other extreme the outermost layer of the atmosphere radiates to space - energy which is then lost to the system of earth and its atmosphere.

With clear skies, the downward component of long-wave radiation, $R_L\downarrow$, is about 60-80 per cent of the upward component, $R_L\uparrow$, the actual value depending upon the distribution of temperature, water vapour and carbon dioxide with height. For radiation received at the ground, the lowest 100 metres contributes strongly to $R_L\downarrow$, the contribution from higher layers having a decreasing effect with increasing height, and conditions above 1000 metres have little influence when the sky is clear. The water-vapour and temperature distributions in the lowest few hundred metres are often quite highly correlated with the temperature and humidity at screen level, and because of this a number of workers have been able to express $R_L\downarrow$ at the ground in terms of T, the screen temperature, and e, the water vapour pressure at screen level.

Ångström[8a,b] proposed the following formula for the downward component of long-wave radiation received at the ground:

$$R_L\downarrow = \sigma T^4 (A - B \times 10^{-\gamma e}) \quad \dots \quad (17.5)$$

where $A = 0.81$, $B = 0.24$ and $\gamma = 0.052$; $R_L\downarrow$ is in mW/cm^2 if σ is expressed in mW cm^{-2}K^{-4}; e is in millibars. (Ångström's formula is sometimes quoted as

$$R_L\uparrow = \sigma T^4 (A_1 + B \times 10^{-\gamma e}) \quad \dots \quad (17.6)$$

where $A_1 = 1 - A$ and the expression gives the net upward radiation.) (1 mW/cm^2 = 10 W/m^2.)

Brunt[9] suggested an alternative formula, based mainly on observations by Dines at Benson:

$$R_L \downarrow = \sigma T^4(a + b\sqrt{e}) \qquad . \ . \ . \ . \ (17.7)$$

For Benson, the values of a and b were found to be 0.52 and 0.065 respectively. For observations from other places, a varied from 0.34 to 0.60 and b from 0.042 to 0.110, with mean values for a of 0.44 and b of 0.080: these are the values most frequently quoted in text-books. Monteith[10] has criticized the use of mean values on the ground that they take into account data from six stations in various parts of the world and are subject to varying experimental error and variations arising from different distributions of water vapour with height, and also that the data include some anomalous observations from Robitzsch at Lindenburg. For the British Isles it would appear to be preferable to use the constants appropriate to Benson, and Monteith has in fact shown, using Belasco's air-mass analysis[11] to obtain seasonal values of water-vapour distribution with height, that the Brunt equation with the Benson constants provides reasonable results for all air masses in the region of the British Isles (root-mean-square errors of $\pm 5.4\,\mathrm{mW/cm^2}$ at 0°C and $\pm 8.4\,\mathrm{mW/cm^2}$ at 30°C).

More recently, Swinbank[12] has proposed the use of the equation

$$R_L \downarrow = 5.31 \times 10^{-4} \times T^6\,\mathrm{mW/cm^2} \qquad . \ . \ . \ . \ (17.8)$$

which he claims provides estimates with a probable error of $< 0.5\,\mathrm{mW/cm^2}$; the results imply that there is a fairly high correlation between the screen-level air temperature and the temperature and humidity distributions with height. Although Paltridge[13] has thrown doubts on the validity of the equation for daytime use, values derived from Swinbank's formula are within about 3 per cent except perhaps during summer afternoons when they may be systematically high. Paltridge has also suggested that the increased downward long-wave radiation resulting from the presence of cloud amounts on average to about $0.75\,\mathrm{mW/cm^2}$ per okta mean cloud cover. This value would not necessarily apply on a given day and the actual value would depend greatly on the height of the cloud base.

17.4 HEAT TRANSPORT AND TEMPERATURE VARIATIONS IN AND NEAR THE GROUND

The temperature field in the lowest layers of the atmosphere depends to a very large extent upon the nature of the underlying surface, which absorbs most of the solar energy falling upon it and which eventually re-radiates most of the energy back in a form which can be absorbed by the lower atmosphere. The effect of solar radiation absorbed by the ground is first of all to heat the immediate surface layers, modifying the temperature gradient in the ground beneath. Heat is then conducted down the temperature gradient at a rate which depends upon the thermal properties of the soil. The heat flux, Q, through unit area of a horizontal surface at depth z, is given by

$$Q = -k \frac{\partial T}{\partial z} \quad \ldots . \quad (17.9)$$

where k is the thermal conductivity of the soil. If we consider an infinitesimally thin layer of soil, of thickness dz, the heat entering the upper boundary of the layer will partly be absorbed within the layer, so raising its temperature and the rest will leave through the lower boundary. If Q enters the top, then $Q - \frac{\partial Q}{\partial z} dz$ will leave the layer, that is, $\frac{\partial Q}{\partial z} dz$ will be absorbed per second, and the rise of temperature within the layer will be given by

$$\frac{\partial T}{\partial t} = \frac{\partial Q}{\partial z} \times \frac{1}{\rho c} \quad \ldots . \quad (17.10)$$

where ρ is the density of the material and c is its specific heat. From equation 17.9 above,

$$\frac{\partial Q}{\partial z} = -k \frac{\partial^2 T}{\partial z^2} \quad \ldots . \quad (17.11)$$

and therefore:

$$\frac{\partial T}{\partial t} = -\frac{k}{\rho c} \frac{\partial^2 T}{\partial z^2} \quad \ldots . \quad (17.12)$$

The quantity $\frac{k}{\rho c}$, often denoted by a, is known as the thermal diffusivity of the soil (sometimes called the thermometric conductivity of the soil).

If the temperature at the surface of the soil, T_s, varies with time about the mean, $\overline{T}_s$, according to the formula

$$T_s = \overline{T}_s + A_0 \sin(\omega t + \phi) \quad \ldots . \quad (17.13)$$

where $\omega = \frac{2\pi}{P}$, P being the period of the cycle and A_0 its amplitude; then, if k is constant with depth, the temperature at depth z at time t is given by:

$$T = \overline{T}_s + A_0 \exp(-\alpha z)\sin(\omega t - \alpha z + \phi) \quad \ldots . (17.14)$$

where

$$\alpha = \sqrt{(\pi / a P)} = \sqrt{(\pi \rho c / P k)} \quad \ldots . (17.15)$$

The above equation shows that the amplitude of the temperature variation decreases exponentially with depth, the rate of decrease depending upon the thermal diffusivity and on P. For a short-period change, say one day, A decreases rapidly with depth; for a period of one year, the amplitude decreases much less rapidly with depth (by a factor of $\sqrt{365} \approx 19$). Similarly, the term $-\alpha z$ in the sine expression shows that the sine wave at depth z lags behind that at the surface by an angle αz; again the phase lag for the annual variation is a much smaller proportion of the period than it is for the diurnal variation. The amplitude variations with depth and also those of the phase lag can both be used to estimate a and hence, if ρ and c are known, k. Deacon[14] has given the following values of the amplitude, A, of temperature variation at depth z as a percentage of A_0, and the phase lag for various values of αz:

TABLE 17.3

αz	0.1	0.2	0.6	1.0	2.0	4.0	6.28
A (%)	90.5	82	55	37	14	2	0.2
Phase lag (degrees)	6	11	35	57	115	230	360

The values of k, ρ and c vary considerably for differing types of soil, the condition of the surface and the water content. Soil consists of particulate mineral and vegetable matter with considerable spaces between the particles. The particles are usually of a range of sizes and consist of matter of differing specific gravity. When the soil is dry, gases will fill the interstices but, as the moisture content of the soil increases, water will displace some of these gases until the soil becomes waterlogged, the interstices being largely filled with water. It is clear from first principles that density, specific heat and conductivity will vary with the composition of the solid particulate composition of the soil and the water content. The nature of the surface is also important. Compacted surfaces have different properties from loose irregular surfaces. In general, wet soils have a higher specific heat than dry soils. According to Wedmore,[15] values of specific heats range from 0.83 for clay to 0.27 for sandy loam. Densities are very variable but Sutton[16] gives the following typical values which are approximate:

	$10^3 \times kg\ m^{-3}$
Dry loose loam	1
Wet loam	1.8
Clay	2
Sandy loam	1.5
Loose fine sand	0.7
Compact wet sand	1.8
Light soil containing grass roots	0.3 to 0.5

Conductivities depend largely on water content, but the average values of thermal diffusivity, a, for most soils lie between 10^{-2} and $10^{-3}\,cm^2/s$, for example:

Soil	$a(cm^2\ s^{-1})$	$\rho(10^3 \times kg\ m^{-3})$	$c(J\ kg^{-1}\ K^{-1})$
Clay	2.10^{-3}	1.8	3.3
Light soil with grass roots	3.10^{-3}	0.3	1.3

The heat flux in the soil at any depth is obtained by differentiating equation 17.14 with respect to z, and substituting in equation 17.11:

$$Q = \sqrt{2 k \alpha}\, A_0 \exp(-\alpha z) \sin(\omega t - \alpha z + \phi + \pi/4) \quad . \ . \ . \ . \ (17.16)$$

At the surface:

$$Q_0 = \sqrt{2 k \alpha}\, A_0 \sin(\omega t + \phi + \pi/4) \quad . \ . \ . \ . \ (17.17)$$

i.e. $Q_0 \propto k\alpha$

or $Q_0 \propto \sqrt{(\rho c k)}$ . . . . (17.18)

As an example of the orders of magnitude of the heat flux, Table 17.4 (Priestley[17]) shows the amount of heat transferred to the atmosphere when cold air flows over a warmer surface: in this instance the initial temperature difference was 10 degrees and the surface wind speed was 10 metres/second.

TABLE 17.4

	'Conductive capacity' $\sigma = \rho c\sqrt{k}$	Heat transfer to air first 12h	second 12h
		$MJ\ m^{-2}$	
New snow	0.01	0.2	0.08
Old snow; dry sand	0.05	0.9	0.4
Wet soil	0.17	3.1	1.5
Ice	0.21	3.8	2.1
Ocean (allowing for stirring of water)	42.0	14.2	14.2

As we have seen, heat transfer in the soil is carried out mainly by conduction, but if this process were the only one at work in the atmosphere the diurnal variation of temperature would cease at a height of a few metres above the ground. However, in the atmosphere heat transfer is carried on by very much more effective processes:

1. Radiation.
2. Mechanical turbulence.
3. Convective turbulence.

A great deal of effort has been expended on the development of theories of turbulence but the subject is highly mathematical and the detailed treatment is beyond the scope of this Handbook; particularly the application to day-to-day forecasting is not yet practicable. However, it is essential for the forecaster to have a qualitative idea of the processes at work, and a brief outline is given below.

Mechanical turbulence is a result of the drag exerted by the surface on air moving above it. The flow is broken up into eddies (see Chapter 16 - Wind, and Chapter 23 - Bumpiness in aircraft) or fluctuations imposed upon the mean air flow. The fluctuations occur in all directions and the vertical components are responsible for transporting heat upwards and downwards. Brunt[1] quotes the following expression for the net upward flux of heat across unit horizontal surface resulting from turbulence transfer:

$$\text{Heat flux} = -K_H \rho c_p \partial\theta/\partial z \qquad \ldots . \ (17.19)$$

where

K_H = eddy diffusivity for heat,

θ = potential temperature,

ρ = air density, and

c_p = specific heat of air at constant pressure.

The formula is analogous to that for soil, with θ replacing T since air is compressible and θ constant with height represents the equilibrium condition for unsaturated air (dry adiabatic lapse rate). The effect of the turbulent heat flux is to reduce the difference between the actual lapse rate and the dry adiabatic lapse rate. The difficulty with this treatment, however, is that K_H is not constant with height and varies a good deal with the lapse rate, the wind speed, the nature of the surrounding terrain, and the scale of the flow. Mechanical turbulence is predominant when winds are strong; the height to which it extends depends upon the wind speed and the nature of the surface over which the air is moving, increasing with increasing wind speed and with greater surface roughness. The calculation of heat fluxes, using observations of two heights, is comparatively straightforward (Deacon[14]), but routine observations of this kind are not available, and it is not immediately obvious how the results of such calculations can be applied in day-to-day forecasting. Convection occurs when the near-surface air is heated, as when solar radiation is absorbed by the land surface or when cold air moves over a warm sea, and the air is unstable over a suitable depth. The precise mechanism by which convection is initiated is not well understood; lapse rates up to many hundred times the dry adiabatic lapse rate can exist in the lowest metre or so (Lawrence[18]), and Best[19] found the following values for midday mean-temperature differences over grass at Porton, Salisbury Plain, on clear days.

TABLE 17.5

Layer	Temperature differences June	Temperature differences December	Temperature differences corresponding to dry adiabatic lapse rate
cm	*degC*	*degC*	*degC*
2.5-30	-2.20	-0.80	-0.0030
30-120	-0.96	-	-0.0089

Negative values indicate temperatures decreasing upwards.

In these lowest layers, radiation must play an important part in determining the temperature distribution. Immediately above this layer the turbulence often shows the characteristics of mechanical turbulence and depends upon the nature of the surface and on the wind speed; within this region $\partial\theta/\partial z \propto z^{-1}$. At much higher levels there is a region where the fluctuations are organized into convective bubbles or plumes with comparatively steady conditions in the descending air between them; in this region the lapse rate is near the dry adiabatic value, that is to say $\partial\theta/\partial z$ is small. Between these two regions lies a zone of transition, where apparently random fluctuations are still marked, but the motion shows some signs of organization into convective plumes or thermals, and where $\partial\theta/\partial z \propto z^{-4/3}$.

Pasquill[20] has examined the heat fluxes near the surface on a few selected occasions. On a clear spring day, over clayland pasture, the following values were found for the period near midday:

TABLE 17.6

	$W\ m^{-2}$
Incoming solar radiation	523
Reflected solar radiation	87
Net outward long-wave radiation	134
Heat absorbed in soil	70
Heat used in evaporation	81
Heat removed by turbulence	151

The last quantity was assessed from the others, which were measured directly. On a clear day the heat used in evaporation is probably the most variable of the above measured quantities, with corresponding effects on the amount of heat supplied to the air. Indeed it is not unknown for almost all the available energy to be so used. This was so in an example described by Zobel,[21] which illustrates a further way of adding heat to the lowest layers of the atmosphere. The inversion layer in this instance was bombarded by thermals from below, with some penetration by the thermals into the inversion layer, and compensating motion of air with higher potential temperature from above the base of the inversion. This led to raising of the base of the inversion, cooling of the air above the initial base of the inversion and below the new level, and warming of the air below the original inversion base. How frequently this process occurs is not known; detailed measurements are required on the few occasions when one can be sure that advection has not played a significant part in changing the temperature profile.

On a clear night with light winds it is radiation which plays the dominant role in establishing the temperature structure in the lower atmosphere. Long-wave radiation is emitted from the ground: some of this radiation is absorbed by the atmosphere and some passes through to be lost to space. Some of the heat absorbed by the lower atmosphere is re-radiated, some downwards back to the ground and some upwards. On the whole, however, there is a net loss of radiation by the ground and the lower atmosphere and cooling occurs. The precise rate of cooling depends not only upon the temperature but upon the distribution of water vapour with height in the atmosphere, a very variable quantity. Most methods of forecasting night cooling make allowances in some way for the water-vapour content of the lower atmosphere. The result of the cooling of the surface layers is that an inversion is established, the atmosphere becomes stable and turbulence is damped down. One result of this is that the wind speed very near the surface becomes lighter, since momentum is no longer brought down from above by turbulence to compensate for that lost at the surface by drag.

The wind near the top of the inversion layer is hardly affected, and may even increase, so that the vertical wind shear becomes greater. The vertical wind shear may reach such a magnitude that it breaks down, temporarily re-establishing turbulence in the lowest layers, and leading to a short-lived increase of temperature before cooling starts again.

On clear nights the minimum temperature is often at a height of a few centimetres above the ground, where the temperature may be about 1 degree lower than the temperature at the surface, although the difference may occasionally reach 3 degrees. The precise reason for this phenomenon is not known, but it has been suggested that it may be produced by radiation from a very tenuous ground mist. Oke[22] has described the raised minimum and has shown that it may also be brought about by back radiation from cloud when the base of the cloud is warmer than the surface.

17.5 VARIABILITY OF TEMPERATURES

When forecasting temperatures it is useful to have some knowledge of the variability of temperatures, both in time and space. The range, scale and periods of the fluctuations and oscillations vary widely. In the large scale of major frontal depressions the range may be some 5-10 degrees, the scale of the order of thousands of kilometres and the period a few days; such features are readily recognizable from observations in the normal synoptic network. In the medium scale, there are fluctuations of a few degrees on a horizontal scale of some hundreds of kilometres which persist for a few hours and are normally revealed by the surface synoptic network in the more populated parts of the world. Such features may slip through the upper-air network, since their horizontal distance and time scales are of the same orders as the separation of the stations and the interval between observations. These variations within an air mass may arise from mesoscale systems, local variations of topography, precipitation or surface conditions, and, when observed, some allowance can be made for the advection of colder or warmer air.

Variations on a smaller scale, of the order of kilometres and minutes, are normally observable only by special instrumentation or networks - not easy to study and impossible to forecast in detail. Nevertheless, it is important that the forecaster should know something about the behaviour of these fluctuations, since this determines the precision and accuracy at which he should aim in his forecasts. The small-scale inhomogeneities have been studied from this viewpoint by, among others, Best,[23] who placed three thermometer screens at equal intervals along a line 15 metres in length. During lapse conditions and low wind speeds there were differences of 0.5-1 degree from minute to minute and from one screen to another, the differences decreasing with increasing wind speed. In neutral conditions the differences were generally less, while in inversions differences between screens were often as much as 0.8 degree, the differences persisting for 10-30 minutes. Thus, while it is reasonable to try to forecast temperatures to the nearest degree, it would be pointless to attempt greater precision.

Small-scale fluctuations are also observed in the free atmosphere well above the surface. Frith[24] described 'closed temperature and humidity patterns, on a scale measured in tens of miles' at 700-800 millibars, with temperature variations of the order of 0.5-1 degree, sometimes greater, and variations in frost-point of more than 15 degrees. Grant[25] and Warhaft[26] have found similar fluctuations of temperature in convective regimes.

Temperature inversions may also give rise to variations of temperature (and humidity) in a horizontal plane. Recent investigations (for example, Hooper,[27] Barbé,[28] Readings, Golton and Browning[29]) have found that temperature inversions of a few degrees can occur over a vertical interval of the order of several metres to a few tens of metres. If the inversion layers are sloping, or undulating, large temperature differences may occur over quite short horizontal distances - several degrees per kilometre on occasion.

Small- and meso-scale variations of temperature arising from the effects of topography are discussed in later sections.

The diurnal range of temperature varies widely with time of year, location, cloudiness, state of ground, etc. A knowledge of the temperature ranges which occur in various conditions is often a help to the forecaster in deciding whether his forecast temperatures are reasonable values. A good deal of information is available on the monthly means of the daily range of temperature for a number of places, but much less has been published on the diurnal variation in different synoptic situations. Belasco[11] has given mean values of maximum and minimum temperatures for Kew, in south-east England, for different air masses for each month of the year for the period

1871-1940, the basis of the air-mass classification being the source region and the nature of the trajectory from the source region to the British Isles. Although in many respects this type of air-mass classification is rather unsatisfactory, further grouping of the data can be carried out to show some interesting and important features. Firstly, many of the types can be put into two groups according to whether the trajectory, on approaching the British Isles, is cyclonically or anticyclonically curved; the diurnal ranges within each group show little variation with source area, almost all falling well within 1-1.5 degrees of the mean for the group. The largest difference (2.5 degrees) between members of a group is between tropical air and polar air from the north with anticyclonic trajectories in winter, presumably because of the very persistent cloudiness of the tropical air. The values for the two groups are shown in Table 17.7; it should be remembered that the diurnal variation on any occasion may differ considerably from the mean values quoted in the table. Also given in the table are the mean values for tropical air with a straight trajectory reaching the British Isles, (a) over the sea - maritime tropical, and (b) from the Continent - tropical continental. It will be noted that there is little difference between the ranges for tropical maritime air with a straight track and cyclonic polar air for much of the year: again the persistent cloudiness of the tropical air in winter shows up in the smaller diurnal variations in those months.

The data for Belasco's remaining types, anticyclonic air with a rather indeterminate track, but possibly with a slight drift from a given direction, have not been quoted. For most of these classes, the ranges are somewhat similar for air masses with a more definite, anticyclonic, trajectory, but for stagnant anticyclonic air, and for air with a light drift from a south-easterly direction in summer, the ranges are increased by about 2-3 degrees.

Table 17.7 includes Belasco's data for all occasions at Kew, and these may be compared with the values in the next column, for Rye, Sussex, for the period July 1945 to June 1948.[30] The greater ranges at Kew are probably mainly because it is an inland station, whereas the Rye site was only 5 kilometres from the coast. The next two columns give the summer and winter mean ranges for cloudy and for clear conditions at Rye. The final columns show the average and the extreme daily ranges at Kew during the period 1950-54 for clear days. The large variability is mainly a result of varying temperature structure of the lowest layers on individual occasions.

TABLE 17.7 Diurnal variation of temperature at Kew and Rye in different conditions (degC)

Month	KEW, 1871-1940 (after Belasco[11]) Polar[1] (cyclonic)	Polar and tropical[2] (anticyclonic)	Tropical (straight trajectory) maritime[3]	Tropical (straight trajectory) continental[4]	All occasions	RYE, 1945-48 (Best *et alii*[30]) All occasions	Clear days	Cloudy days	KEW, 1950-54 Clear days Average	Maximum	Minimum
Jan.	3.5	5	2	2	5	3	6	1	5	6.5	4.5
Feb.	4.5	5.5	3	5.5	5.5	3.5			7.5	6.5	7
Mar.	5	6.5	4.5	8.5	7	7			9.5	16.5	5.5
Apr.	6	8.5	4.5	10.5	8.5	8			11.5	18	7
May	7	9.5	8	11.5	9.5	8.5			12	16.5	8
June	7	9	6.5	10.5	9.5	7.5			12.5	17.5	8.5
July	6.5	9.5	6.5	11.5	9	8	13.5	3.5	11	14	8.5
Aug.	6.5	9.5	6.5	11.5	9	8.5			10.5	15	7.5
Sept.	6	8	6	9	8.5	7			10	15	7
Oct.	5.5	6.5	4	7	7	7			9.5	12	5
Nov.	4.5	5.5	2	5	5.5	4			7	11.5	4.5
Dec.	4	5	2	4	5	2.5			6	8	4

Notes: of Belasco's classification:
1. P_1, P_3, P_5 and P_7
2. T_2, P_2, P_4 and P_6
3. T_1
4. T_3

17.6 FORECASTING TEMPERATURES NEAR THE GROUND

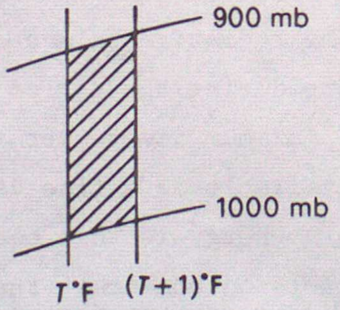
Although there is hope that the equations governing the physical processes described in 17.4 (page 11) will be used in due course in a computer model of the boundary layer, the practical forecaster must at present rely, for his forecasting of temperature, on a more empirical approach. The essence of this approach is an appreciation of the temperatures accompanying various synoptic situations; diurnal and advective temperature changes can be assessed by consideration of the physical processes involved and by use of empirical formulae which have been devised for simple meteorological situations. The variations of temperature near the ground are usually forecast by applying a formula to an extreme case. For example, methods for calculating day maximum and night minimum temperatures at screen level usually commence with the assumption of cloudless skies. On this simple approach a possible extreme value of the temperature is obtained. Depending on the forecast of the other meteorological elements for the period, for example, direction and speed of wind, cloud amounts and thicknesses, precipitation etc., that extreme value is modified, sometimes by objective methods but more often the value has to be adjusted in the light of a forecaster's experience and interpretation of the physical processes at work.

17.6.1 Day heating

17.6.1.1 Maximum day temperature. The method most widely used to forecast the maximum day temperature in the neighbourhood of the British Isles is based on Professional Note No. 63 by Gold.[31] In this Note, Gold makes broad assumptions regarding the radiative balance at the ground and in the lowest layers of the atmosphere during the period from dawn to 1500 local time (the approximate time of maximum day temperature) and calculates the maximum energy likely to be available on a fine day for heating the lowest layers of the atmosphere. This energy is then equated to areas on a tephigram and from this it is possible to construct the probable distribution of temperature in the vertical at the time of maximum temperature.

The calculation starts with the amounts of energy received on a unit horizontal surface at the outer edge of the earth's atmosphere at latitude 50°N in one day near the middle of each month. Gold then assumes that 50 per cent of the incident energy amounts will reach the earth's surface by 1500 local time as direct solar radiation; he makes an allowance, based on experiments by Dines and Dines[32] for the diffuse short-wave radiation from the sky to give the maximum possible total amount of short-wave radiation falling upon unit area of the earth's surface up to that time. He assumes that the albedo of the surface is 20 per cent, i.e. that 20 per cent of the incident short-wave radiation is reflected back into space. As seen in 17.3 (page 7), the surface loses heat in the form of long-wave radiation. Some of this radiation is absorbed by the lower atmosphere, while the rest is lost to the higher levels of the atmosphere or to space. Some of the radiation absorbed by the lower atmosphere is re-radiated back to the earth's surface. Gold assumed that 30 per cent of the long-wave radiation from the earth's surface would not go towards heating the lower atmosphere. One further allowance needs to be made, that is for evaporation. The values of daily evaporation used by Gold were derived mainly from mean data for London (Camden Square), given on page 206 of Shaw's 'Manual of Meteorology', Volume II. The net energy available for heating the lower atmosphere was then calculated as the total short-wave radiation received at the surface, less 20 per cent reflected, less the heat required for evaporation, less 30 per cent of the long-wave radiation from the earth's surface from one hour after sunrise to 1500 local time.

By finding the amount of heat required to raise the temperature of a column of air, of unit cross-section, extending from 1000 to 900 millibars by 1°F, and equating it to the area enclosed by the 1000- and 900-millibar isobars and two isotherms 1°F apart on the tephigram, Gold was able to express the energy values he obtained as areas on the tephigram:



The estimation of areas on a tephigram is not an easy or an accurate process, and Gold went further to express the energy as the thickness in millibars of the layer which would be changed from an isothermal to a dry adiabatic state if the heat were added at the bottom of the layer. Gold's conversions from energy to millibars, as quoted in Professional Note No. 63, were not all correct, but the correct values have been given by Johnston[33] and these are given in Table 17.8, together with the energy values given by Gold. One further advantage of using millibar values is that they are independent of the scale of the tephigram used.

TABLE 17.8 Energy available for heating the lower atmosphere by 1500 local time on a clear day near the middle of each month, and depth of layer in millibars, Δp_1, converted from an isothermal to a dry adiabatic state.

Month	J	F	M	A	M	J	J	A	S	O	N	D
Energy (MJ m^{-2})	1.7	2.9	4.2	5.9	7.3	7.5	6.9	6.3	4.8	3.3	1.7	1.3
Depth of layer (mb)	62	82	97	115	128	130	124	119	105	88	62	55

Johnston[33] has described a simple way in which the above figures can be applied. From Table 17.8 a layer thickness, Δp_1, for the month concerned is obtained. The isobars p_0 and $(p_0 - \Delta p_1)$ are marked on the tephigram (see Figure 4). The point I is selected on the $(p_0 - \Delta p_1)$ isobar such that the area of triangle I I' F is equal to that enclosed by the dry adiabatic through I, the surface isobar, and the environment curve, i.e. until area A B D in the example is equal to the sum of areas E' A I' and B E I. Then F (the intersection of the dry adiabatic through I and the surface isobar) represents the maximum temperature. A scale can readily be made to assist with the assessment, consisting simply of two lines at right angles, as I I' and C I F. In use I is kept on the isobar, I I' and C I F being parallel to the isotherms and dry adiabatics respectively.

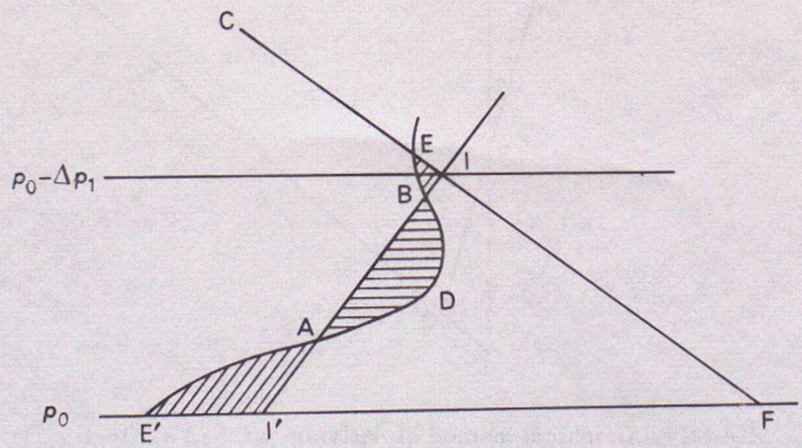


FIGURE 4. Estimation of maximum day temperature (Johnston[33])

Inglis[34] has suggested three modifications to Johnston's method:

(a) Having found the isothermal T' through I, such that equal areas are enclosed on either side of the environment curve, the mere addition of a temperature rise, r_1, to the value of T' would give a maximum temperature. This is so, but since the isothermal I I' and consequently the dry adiabatic C E I F need to be found anyway, the method is no simpler than the version given by Johnston. Values of r_1 are given in Table 17.9.

TABLE 17.9 Rise in temperature at the surface when the layer from p to $(p_0 - \Delta p_1)$ is heated from an isothermal to a dry adiabatic state.

Month	J	F	M	A	M	J	J	A	S	O	N	D
r_1 (degC)	5	6.5	8.5	9.5	10.5	11	10.5	10	9	7	5	4.5

(b) If the environment curve at dawn is more nearly a saturated adiabatic than an isothermal, it is easier and more accurate to use a scale with a line drawn at an angle to the isothermal: Inglis calls this the 'oblique' method. The method of application of the scale is similar to that for the 'isothermal' method, but the depth of the layer, Δp_2, over which the equality of areas is assessed will be different, and so will the increment of temperature, r_2, to be added to the temperature at the point where the oblique line intersects the surface isobar on the tephigram. The derivation of Δp_2 and r_2 are illustrated in Figure 5. For any month let I' F be the mean surface isobar and let F correspond to the mean daily maximum temperature for, say, southern England. Subtract r_1 from the maximum temperature to get the point I' and let the isothermal and the dry adiabatic meet at I.

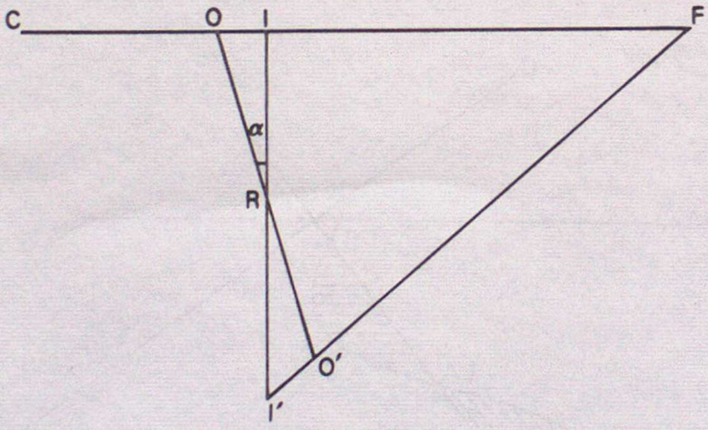


FIGURE 5. Graphical method of deriving Δp_2 and r_2 (Inglis[34])

Draw the oblique O O' at the required angle of inclination α so that O R I and O' R I' are equal in area. Then F O' would give the rise in temperature r_2, and the difference in pressure between O and the surface will give the depth of the layer Δp_2. For $\alpha = 45^\circ$, values of Δp_2 and r_2 are given in Table 17.10.

TABLE 17.10 Depth of layer (Δp_2) which is changed by heating from an oblique at 45° to a dry adiabatic, and associated rise in temperature r_2

Month	J	F	M	A	M	J	J	A	S	O	N	D
Δp_2 (mb)	85	115	135	155	170	175	170	155	140	120	85	70
r_2 (degC)	3.5	4.5	6	6.5	7.5	8	7.5	7	6.5	5	3.5	3

(c) Inglis also described a further variation in which the oblique line is replaced, in principle by a saturated adiabatic, but in practice by a set of oblique lines, one for each month, with slopes which vary from month to month. This method offers little advantage over the oblique method, and is more complex, so it will not be described here. Readers who wish to learn more of this variation should consult the original paper.[34]

17.6.1.2 <u>Maximum temperatures and thicknesses - 1000-500-mb thicknesses.</u> Boyden[35] has used the 1000-500-mb thickness pattern in an attempt to forecast the maximum temperature in certain months, on the grounds that the surface temperature should bear some relationship to the mean temperature, or thickness, of a layer between two isobaric surfaces based at or near the ground. The relationships he found were fairly complicated, and often rather vague, but some useful results were obtained. The basic relationship is shown in Figure 6; the temperature to be read off for the appropriate thickness is denoted by S_0, S_3, S_6, S_9 and S_{12}, where the suffix denotes the total number of hours of sunshine.

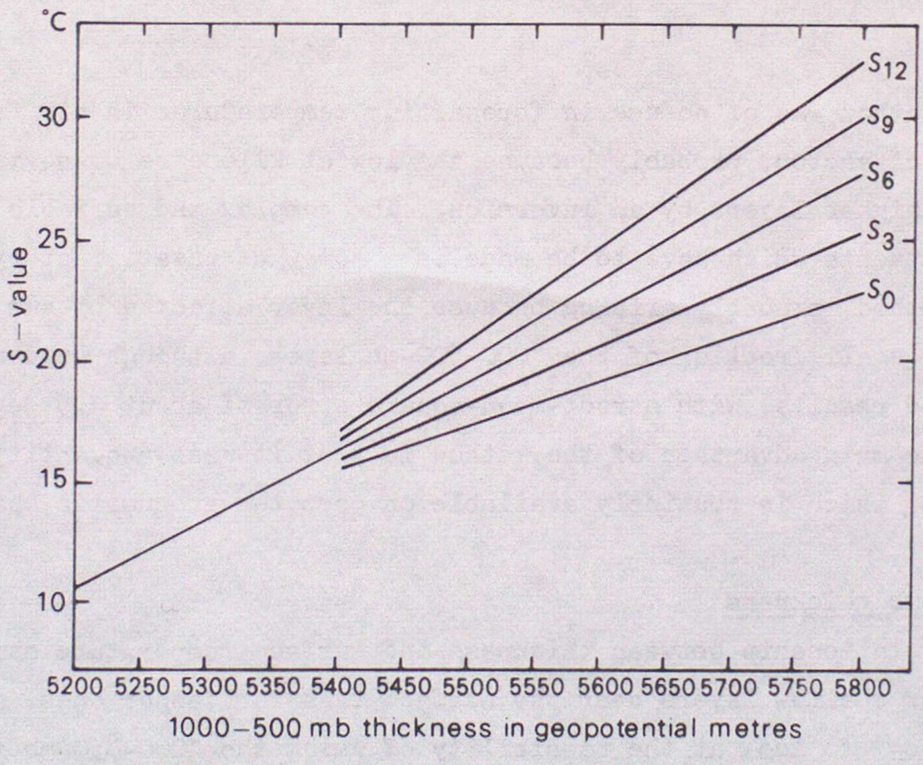


FIGURE 6. Forecasting diagram for maximum temperature

The basic forecast is then modified for different wind directions (the classification of which is given in Table 17.11), for different months and a further allowance has sometimes to be made, in terms of *N*, the total number of hours of sunshine. The final results are given in Table 17.12.

TABLE 17.11 Classification of trajectories

NW (north-west)	Trajectories from between north and west.
NE (north-east)	Trajectories from between north and about east-north-east, thus excluding air which crossed the Netherlands and northern Germany.
SW (south-west)	Trajectories from between about south-west and west.
C (Continental)	Trajectories from the mainland of Europe, disregarding Denmark, Norway and Sweden.
St (stationary)	Trajectories having no significant direction because the wind was light, or occasionally because the trajectory was more nearly circular than straight.

TABLE 17.12 Maximum temperature at Kew in °C

	April	*May*	*June, July and August*	*September*
NW	-	Appropriate S-value less 1°	Appropriate S-value	S
NE	$S_6 + \frac{N}{2} - 6$	$S_6 + \frac{N}{2} - 5$	Appropriate S-value but $S_6 - 5$ if sunshine zero	-
SW	$S_6 + \frac{N}{2} - 5$	$S_6 + \frac{N}{2} - 4$	$S_6 + \frac{N}{2} - 3$	$S_6 + \frac{N}{2} - 4$
C	-	-	-	-
St	-	-	S_{12} but 3° lower if continuous rain occurs	

The method was of no use in forecasting temperatures in air from the continental sector, probably because the lowest kilometre or so is 'isolated' from the higher layers by an inversion. The complex and variable nature of the adjustments which have to be made is a somewhat unsatisfactory feature of the method, probably arising because the layer affected by surface heating is only a small fraction of the 1000-500-mb layer, although Boyden claims quite good results, with a root-mean-square error of about 1.5 degrees or less. The main advantage of the method is that it uses a quantity, 1000-500-mb thickness, which is routinely available on computer prognostic charts.

1000-850-mb thickness

The relationship between thickness and surface temperature should be better for shallow layers near the surface than for deeper ones, and this led Inglis [34] to look at the possibility of using the 1000-850-mb thickness. There was also the consideration that the 1000-850-mb layer corresponds well with the layer within which surface heating occurs in summer (see 17.6.1). Gold's values were used by Inglis to determine the maximum temperature from the mean temperature of the 1000-850-mb layer with certain assumptions about the lapse rate and humidity within the layer. The results are given below in Table 17.13. For further details the reader should refer to Inglis's paper.

TABLE 17.13 1000-850-mb thickness values and corresponding Gold maxima for August

Thickness	0	1	2	3	4	5	6	7	8	9
gpm					degrees Celsius					
1280	7.3	7.5	7.7	7.9	8.1	8.3	8.4	8.6	8.8	9.0
1290	9.2	9.4	9.6	9.8	10.0	10.1	10.3	10.5	10.7	10.9
1300	11.1	11.3	11.5	11.7	11.9	12.1	12.2	12.4	12.6	12.8
1310	13.0	13.2	13.4	13.6	13.8	13.9	14.1	14.3	14.5	14.7
1320	14.9	15.1	15.3	15.5	15.7	15.9	16.0	16.2	16.4	16.6
1330	16.8	17.0	17.2	17.4	17.6	17.7	17.9	18.1	18.3	18.5
1340	18.7	18.9	19.1	19.2	19.4	19.6	19.8	20.0	20.1	20.3
1350	20.5	20.7	20.9	21.0	21.2	21.4	21.6	21.8	21.9	22.1
1360	22.3	22.5	22.7	22.8	23.0	23.2	23.4	23.6	23.7	23.9
1370	24.1	24.3	24.5	24.7	24.9	25.1	25.2	25.4	25.6	25.8
1380	26.0	26.2	26.4	26.5	26.7	26.9	27.1	27.3	27.4	27.6
1390	27.8	28.0	28.2	28.3	28.5	28.7	28.9	29.1	29.2	29.4

	April	May	June	July	September
Correction (degC)	-0.9	+0.2	+0.7	+0.5	-1.0

Inglis found a standard error roughly equal to that for the tephigram method: the results are compared in Table 17.14:

TABLE 17.14 Comparison of tephigram and 100-850-mb thickness methods as predictors of maximum temperature using data for Aughton 1966-69

	Number of observations	Root-mean-square error Tephigram method	Thickness method
		degC	degC
April	32	1.67	1.56
May	29	1.39	1.44
June	43	1.65	1.66
July	45	1.94	1.82
August	28	1.55	1.68
September	27	1.13	1.10
Apr. - Sept.	204	1.62	1.59

17.6.1.3 Rise of temperature on clear mornings. For a number of purposes it is useful to know the expected temperature at various times before the maximum is reached. Wallington[36] has given values of the energy available, in terms of areas on the tephigram, up to various local times in each month of the year. However, there are some inconsistencies in his tables and no indication is given in the original publication of how the values were obtained. Table 17.15 shows, for each month of the year, the thickness of the layer which is changed from an isothermal to a dry adiabatic state by insolation over southern England for various times after sunrise. The table is based on unpublished data available in the Meteorological Office and has been attributed to Johnston. The data agree reasonably well for most months with the data given by Wallington when inconsistencies in the latter have been removed by smoothing.

TABLE 17.15 Thickness of layer which is changed from an isothermal to an adiabatic state by insolation (southern England)

Month	Hours from sunrise 1	2	3	4	5	6	7	8	9	Max.
	millibars									
Jan.	16	29	41	50	58	-	-	-	-	61
Feb.	18	33	46	57	65	73	-	-	-	81
Mar.	20	37	52	63	73	82	90	-	-	97
Apr.	22	40	56	69	80	89	98	106	-	115
May	23	42	59	72	83	93	102	110	118	127
June	23	42	60	73	84	94	103	112	119	130
July	23	42	59	73	84	94	103	111	118	125
Aug.	22	41	57	70	81	91	99	108	115	119
Sept.	21	38	54	66	76	85	93	100	-	104
Oct.	19	35	49	60	69	77	85	-	-	87
Nov.	17	31	43	53	61	-	-	-	-	61
Dec.	15	28	40	49	-	-	-	-	-	53

Note

Over damp soils, e.g. clay, the energy corresponding to the values given is not available for heating the lowest layers of the atmosphere during the first few hours of sunshine. In such localities the values to the left of the lines marked A in Table 17.15 should be reduced by one-third and the values between the lines marked A and B should be reduced by one-fifth.

The method of use is similar to that described for the Johnston method in 17.6.1.1 (page 21).

Jefferson[37] studied the temperature rise on a number of mornings at Northolt when there was little or no cloud or fog to interfere with the receipt of solar radiation. By taking, or assessing, a representative upper-air ascent he was able to calculate the expected temperature rise at the times when energy corresponding to 0.27, 0.54, 1.09 and 2.18 cm^2 on Metform 2810 (1963 edition); by comparing these calculated temperature rises with the actual temperature rise during the morning, he was able to deduce the times by which the given amounts of energy had been received. The results are given below in Table 17.16.

If Jefferson's data are compared with those given in Table 17.15 (and the conversion of the data from one form to another to enable this comparison to be carried out is an interesting exercise in itself) it is seen that there is only fair agreement. It is not possible to say, on the basis of studies so far, which set of results is more correct. The data of Table 17.15 appear to be an almost linear interpolation of energy receipt from shortly after sunrise to the time of maximum temperature, whereas it seems likely that the rate of energy receipt should increase from sunrise to noon as the sun climbs in the sky.

TABLE 17.16 Average time in hours from sunrise

	Equivalent areas on tephigrams *cm*			
Large-scale insert on Metforms 2810A and 2810B (1956 editions)	0.49	0.97	1.95	3.9
Metform 2810 (1963 edition)	0.27	0.54	1.09	2.18
Small-scale section of Metforms 2810A and 2810B (1956 editions)	0.12	0.24	0.49	0.97
	Hours from sunrise			
January	2.6	4.2	4.7	-
February	2.3	3.3	4.2	5.8
March	2.0	2.7	3.5	4.5
April	1.8	2.3	3.0	3.7
May	1.6	2.1	2.8	3.3
June	1.5	1.95	2.7	3.1
July	1.4	1.9	2.75	3.2
August	1.5	1.95	2.9	3.4
September	1.6	2.1	3.1	3.6
October	1.9	2.6	3.5	4.0
November	2.4	3.3	4.1	-
December	3.0	4.2	4.7	-

Jefferson's approach seems a logical way to tackle the problem but the time taken to go from an energy receipt corresponding to 1 cm^2 to that corresponding to 2 cm^2 is in many months suspiciously short. There is a need for more research in this area, but on the whole it is suggested that the values in Table 17.15 be used as a guide, bearing in mind that the rate of rise of temperature may be rather less than that given by the table during the first hour or two after sunrise.

All the above methods are based on the assumption that temperature changes in the lowest layers of the atmosphere are a result of heating of the surface of the earth by solar radiation. The process described by Zobel[21] is not taken into account; however, on the day in question, the rise in temperature would have been predicted quite well by the Gold method. On the other hand, more detailed calculations of the energy balance, using experimental data on evaporation and net long-wave radiation loss, suggest that no energy is available for heating the air during some winter months, yet rises of 5 degrees or so on clear days in December and January are not rare. As stated before, there seems to be a need for more work on this topic, and it should be well within the scope of the outstation forecaster to make a useful contribution. Something along the lines of Jefferson's work, extending to the time of maximum temperature, would be valuable.

17.6.1.4 <u>Rise of temperature on days of fog or low cloud</u>. Jefferson[38] has described a method for estimating the temperature rise on a day when fog or low cloud is initially present:

(i) Using Table 17.15 and the method of 17.6.1.1 (page 21), draw a curve showing rise of temperature from sunrise, assuming a clear day (full line in Figure 7).

(ii) Applying constant 'delay factor' f, plot points C, F ... , such that

$$AB/AC = DE/DF = \dots f,$$

where f is 0.25 for thick stratus or deep fog and 0.35 for thin stratus or shallow fog. The broken line then represents the rise of temperature on a foggy day.

(iii) If this curve reaches temperature T_c, at which fog can be expected to disperse,* the curve then rises more steeply and is more in line with clear-day characteristics (e.g. dotted line in Figure 7).

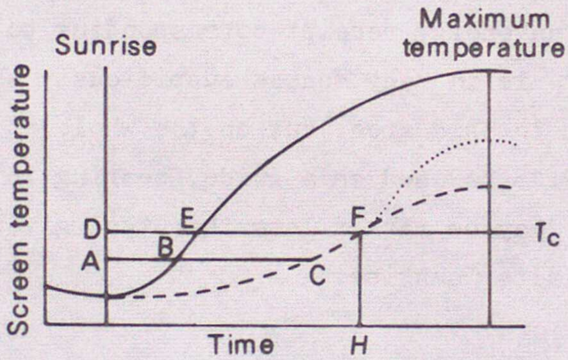


FIGURE 7. Construction of temperature curves

——— Forecast temperature curve for clear day.
– – – Forecast temperature curve for fog, assuming $f = 0.35$ and T_c is not reached.
.... Forecast temperature curve after dispersal of fog, assuming T_c is reached at time H.

Notes

(a) On a particular occasion, f can be evaluated from an observation of temperature made not less than 3 hours after sunrise. If t_2 is the time at which the temperature was observed, t_1 the time estimated by method (i)) at which the same temperature would have been reached on a clear day, and t_0 is the time of sunrise, then $f = (t_1 - t_0)/(t_2 - t_0)$.

(b) See sections 20.8.1 in Chapter 20 - Visibility, and 19.6.3.2 in Chapter 19 - Clouds and precipitation, for the use of this method in forecasting the dispersal of fog and low stratus.

* See section 20.8 (page 52) in Chapter 20 - Visibility.

17.6.2 Night cooling

17.6.2.1 General principles. The temperature which is attained by the air at the end of a clear night depends on several factors. First there is the loss of heat by long-wave radiation from the surface of the earth. There is also some loss of heat from the lower layers of the air.[39] There is also absorption and radiation of long-wave radiation by the water vapour present in the air. There is the further complication of absorption and radiation of long-wave radiation by the cloud or fog, when either of these is present.

Theoretical attacks on the problem by Brunt,[9] Frost,[39] Knighting[40] and others have concentrated on idealized models of clear nights in June and the outcome of their researches has not yet found any real application in practical forecasting.

A number of empirical methods of forecasting night minimum temperature in the screen have been developed and those in common use in the United Kingdom all assess the water vapour present in terms of dew-point. These methods all depend on solving an equation of the form

$$T_{min} = aT + bT_d + C \qquad \ldots \quad (17.20)$$

where T is a temperature at some particular time in the previous afternoon, T_d is the dew-point at a particular time or the mean over the cooling period, a and b are constants, and C is a quantity which varies only with wind speed and cloud amount. Temperature data for nights of clear sky and no wind are examined first and values of a, b and C derived. Different values of C are then obtained for various categories of wind speed and cloud amount from the data for the remaining occasions, the assumption generally being made that a and b remain constant.

There are three ways of determining the constants of equation 17.20. Reasonable values of a, b and C may be assessed by inspection of a number of examples; this was done by Boyden[41] and McKenzie.[42] A regression may be set up with a, b and C calculated statistically; this was done by Craddock and Pritchard,[43] by Gordon, Perry and Virgo[44] and by Vince.[45] Lastly, a set of cooling curves may be drawn; this was Saunders' method.[46]

The methods described in this section are applicable only when the ground is not snow-covered: see 17.7.4 (Effect of snow cover) on page 59. These methods will now be discussed in greater detail.

17.6.2.2 Night minimum temperature (Boyden). Boyden[41] put a and b each equal to $\frac{1}{2}$, and he used a wet-bulb temperature; no dry-bulb temperature appears in his formula, which is:

$$T_{min} = \tfrac{1}{2}(T_w + T_d) - C \qquad \text{. . . . (17.21)}$$

where T_w is the wet-bulb temperature at the time of maximum dry-bulb temperature, T_d is the dew-point at the same time, and C varies with cloud amount and wind speed.

Boyden obtained some data for Kew Observatory from which allowances could be made for mean low-cloud amounts, wind speed and occasions of fog. His results are given in Table 17.17.

TABLE 17.17 Values of C at Kew

Surface wind speed	Low cloud amount (oktas) < 2	2-6	> 6
knots		*degC*	
0	+2	+1	-0.5
1-2	+1.5	+0.5	-0.5
3-5	+1	0	-1
6	+0.5	-0.5	-1
7	-0.5	-0.5	-1
8	-1.5	-1.5	

The method is simple to use, as follows:

(i) Forecast mean wind speed (at anemometer height) and mean amount of low cloud.

(ii) Determine the night minimum from the formula, giving C the value appropriate to the forecast mean wind speed and cloud amount.

(iii) If T_{min} is less than T_d as determined in (ii), take the night minimum at $(T_{min} - 2)$°C.

To use Boyden's formula in any particular locality it is necessary to determine the extent to which the corrections given by the 'curves' in Table 17.15 apply and to determine the values of C by comparison of predictions with a sufficient number of actual observations of night minima. Boyden also suggests that the correction for fog may be less than 2.0 degrees at most stations as the thermometers at Kew were exposed in a screen on a north wall at a height of 10 feet (3.05 metres) which differs somewhat from the standard exposure.

17.6.2.3 Night minimum temperature (McKenzie). McKenzie's formula[42] is similar to Boyden's except that he used a dry-bulb temperature directly in his formula, which is

$$T_{min} = \tfrac{1}{2}(T_{max} + T_d) - C \qquad \text{. . . . (17.22)}$$

where T_{max} is the maximum temperature during the afternoon, and T_d is the average dew-point of the air mass expected during the night.

From a large number of suitable examples at Dyce, McKenzie found the constants for use with varying cloud amounts and surface wind speeds. In Table 17.18 his original table has been converted into one with cloud amount at intervals of 2 oktas and temperatures in °C. All constants were found to be negative.

TABLE 17.18 Values of constants (C) for Dyce

Average surface wind speed in knots	Average low cloud amount in oktas 0	2	4	6	8
			degC		
0	-8.5	-6	-6	-4.5	-4.5
1-3	-7	-5	-5	-4	-3
4-6	-4.5	-4	-3.5	-3	-1.5
7-10	-4	-3.5	-3	-3	-1.5
11-16	-3.5	-3	-3	-1.5	-1
17-21	-2	-2	-2	-1.5	-1
22-27	-2	-1.5	-1.5	-1.5	-1

Values of the constant for varying wind speeds and cloud amounts for other localities may be obtained as follows:

Select a suitably large number of occasions from the records for which the dew-point remained constant within 1-1.5 degrees from 1200 GMT one day to 0600 GMT the following day. From the main observations over that period calculate the mean dew-point, the average cloud amount and the average wind speed (or force). Using known values of T_{max} and T_{min} compile a series of constants for various combinations of low-cloud amount and wind speed. Compile for that locality a table similar to Table 17.18, using average values of the constants found for each combination of cloud amount and wind speed.

McKenzie's method is very popular and many forecasters apply to their own stations the table which was originally devised for Dyce. By an accident of good fortune McKenzie's constants for Dyce are roughly the average of constants which have been calculated for a number of stations mainly in eastern England, but this does not necessarily make them valid for other stations. Anyone who contemplates using McKenzie's table should therefore first ensure that it is applicable to his particular locality. It would be better still if he started from scratch and worked out a McKenzie's table from data for his own station.

17.6.2.4 Night minimum temperature (Craddock and Pritchard). From a statistical investigation of observations from 16 stations in eastern England not close to the sea Craddock and Pritchard[43] obtained the regression equation:

$$T_{min} = 0.316\,T_{12} + 0.548\,T_{d12} - 1.24 + k = X + k \qquad \ldots\ (17.23)$$

where T_{12} is the observed temperature at 1200 GMT in °C, and T_{d12} is the observed dew-point at 1200 GMT in °C. Values of X are obtained from Table 17.19; k is an allowance for mean geostrophic wind speed and cloud amount, made by adding the appropriate corrections from Table 17.20.

This formula is useful for forecasting night minimum air temperatures over a wide area of eastern England. Anyone who proposes to apply it to other parts of the country must ensure, by examining past records, that its use is appropriate to his area. The method could be applied to, and regression equations worked out for, an individual station.

17.6.2.5 Night minimum temperature by graphical methods. The basic idea behind the graphical method is due to Saunders,[46] who devised methods whereby an attempt can be made to estimate curves indicating the rate of cooling during the night. Thus with an estimate of the night minimum it is possible to fill in the temperature curve from early evening to night minimum.

The basis of Saunders's work[47] rests on a discontinuity in the rate of cooling at grass level which occurs within about an hour after sunset and is particularly sharp and well defined on clear nights with light winds. The discontinuity appears to be associated with the deposition of dew on the grass. A similar but somewhat less sharply defined discontinuity in the rate of cooling at screen level is usually shown on thermograms in the British Isles on clear evenings at times varying between 1600 and 2230 GMT according to the time of year. Saunders's earlier work was done at Abingdon and screen thermograms for Abingdon for 24-26 March 1944 are reproduced in Figure 8. The discontinuity at 2000 GMT is very clear on all three evenings.

TABLE 17.19 Values of X

T_{d12} / T_{12} °C	-3	-2	-1	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16°C
										degC										
3	-1.9	-1.4	-0.8	-0.3	+0.3	0.8	1.4													
4	-1.6	-1.1	-0.5	+0.0	0.6	1.1	1.7	2.2												
5	-1.3	-0.8	-0.2	0.3	0.9	1.4	2.0	2.5	3.1											
6	-1.0	-0.4	+0.1	0.7	1.2	1.8	2.3	2.8	3.4	3.9										
7	-0.7	-0.1	0.4	1.0	1.5	2.1	2.6	3.2	3.7	4.3	4.8									
8	-0.4	+0.2	0.7	1.3	1.8	2.4	2.9	3.5	4.0	4.6	5.1	5.7								
9	-0.0	0.5	1.1	1.6	2.2	2.7	3.2	3.8	4.3	4.9	5.4	6.0	6.5							
10	+0.4	0.8	1.4	1.9	2.5	3.0	3.6	4.1	4.7	5.2	5.8	6.3	6.9	7.4						
11	+0.7	1.1	1.7	2.2	2.8	3.3	3.9	4.4	5.0	5.5	6.1	6.6	7.2	7.7	8.3					
12	1.0	1.5	2.0	2.6	3.1	3.6	4.2	4.7	5.3	5.8	6.4	6.9	7.5	8.0	8.6	9.1				
13	1.3	1.8	2.3	2.9	3.4	4.0	4.5	5.1	5.6	6.2	6.7	7.3	7.8	8.3	8.9	9.4	10.0			
14	1.6	2.1	2.6	3.2	3.7	4.3	4.8	5.4	5.9	6.5	7.0	7.6	8.1	8.7	9.2	9.8	10.3	10.9		
15	1.9	2.4	3.0	3.5	4.0	4.6	5.1	5.7	6.2	6.8	7.3	7.9	8.4	9.0	9.5	10.1	10.6	11.2	11.7	
16	2.3	2.7	3.3	3.8	4.4	4.9	5.5	6.0	6.6	7.1	7.7	8.2	8.7	9.3	9.8	10.4	10.9	11.5	12.0	12.6
17	2.6	3.0	3.6	4.1	4.7	5.2	5.8	6.3	6.9	7.4	8.0	8.5	9.1	9.6	10.2	10.7	11.3	11.8	12.4	12.9
18	2.9	3.4	3.9	4.4	5.0	5.5	6.1	6.6	7.2	7.7	8.2	8.8	9.4	9.9	10.5	11.0	11.6	12.1	12.7	13.2
19	3.2	3.7	4.2	4.8	5.3	5.9	6.4	7.0	7.5	8.1	8.6	9.1	9.7	10.2	10.8	11.3	11.9	12.4	13.0	13.5
20	3.5	4.0	4.5	5.1	5.6	6.2	6.7	7.3	7.8	8.4	8.9	9.5	10.0	10.6	11.1	11.7	12.2	12.8	13.3	13.8
21	3.8	4.3	4.8	5.4	5.9	6.5	7.0	7.6	8.1	8.7	9.2	9.8	10.3	10.9	11.4	12.0	12.5	13.1	13.6	14.2
22	4.1	4.6	5.2	5.7	6.3	6.8	7.4	7.9	8.5	9.0	9.5	10.1	10.6	11.2	11.7	12.3	12.8	13.4	13.9	14.5
23	4.5	4.9	5.5	6.0	6.6	7.1	7.7	8.2	8.8	9.3	9.9	10.4	11.0	11.5	12.1	12.6	13.2	13.7	14.2	14.8
24	4.8	5.2	5.8	6.3	6.9	7.4	8.0	8.5	9.1	9.6	10.2	10.7	11.3	11.8	12.4	12.9	13.5	14.0	14.6	15.1
25	5.1	5.6	6.1	6.7	7.2	7.8	8.3	8.9	9.4	9.9	10.5	11.0	11.6	12.1	12.7	13.2	13.8	14.3	14.9	15.4
26	5.4	5.9	6.4	7.0	7.5	8.1	8.6	9.2	9.7	10.3	10.8	11.4	11.9	12.5	13.0	13.6	14.1	14.6	15.2	15.7
27	5.7	6.2	6.7	7.3	7.8	8.4	8.9	9.5	10.0	10.6	11.1	11.7	12.2	12.8	13.3	13.9	14.4	15.0	15.5	16.1

T_{12} temperature at 1200 GMT

T_{d12} dew-point at 1200 GMT

TABLE 17.20 Correction k (degC) to allow for mean geostrophic wind speed and cloud amounts

Mean cloud amount in oktas*	Mean geostrophic wind speed during night in knots			
	0-12	13-25	26-38	39-51
	degC			
0-2.0	-2.2	-1.1	-0.6	+1.2
2.1-4.0	-1.7	0.0	0.0	+1.7
4.1-6.0	-0.6	+0.6	+0.6	+2.8
6.1-8.0	0.0	+1.1	+1.1	-

* Mean of values at 1800, 0000 and 0600 GMT

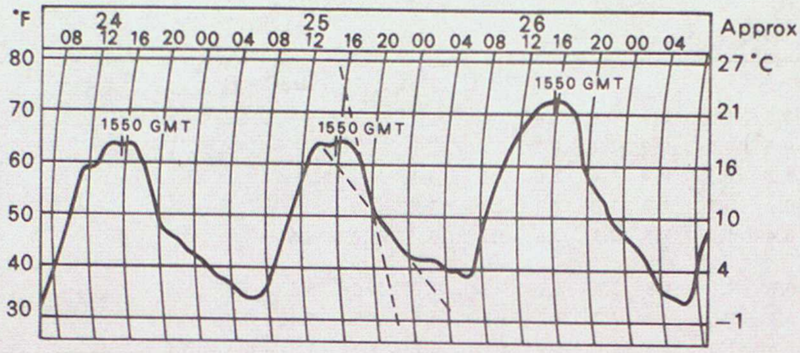


FIGURE 8. Screen thermograms for Abingdon, 24-26 March 1944

The investigation was continued at Northolt during 1948-50, as a result of which Saunders was able to derive a method for determining a cooling curve in three stages:

(a) He constructed a graph showing, throughout the year, the time of occurrence of the discontinuity in the rate of cooling.

(b) He obtained regression equations relating the temperature, T_r, at the time of the discontinuity, to the maximum temperature during the preceding afternoon, T_{max}, and the dew-point at the time of maximum temperature, T_d. The equations were of the form:

$$T_r = \tfrac{1}{2}(T_{max} + T_d) - D \qquad \text{. . . . (17.24)}$$

where the value of D depended upon whether or not there was a low-level inversion during the afternoon.

(c) He constructed graphs showing how the subsequent cooling, after T_r had been reached, depended upon the geostrophic wind speed. It was found necessary at this stage to apply a correction during the summer months when the period of cooling after T_r was reached was less than eight hours.

Saunders's technique aroused considerable interest, and a number of articles on the application of this method are listed in the bibliography at the end of this chapter. Barthram[48] devised a diagram which provides a convenient way of performing the necessary operations, and Tinney and Menmuir[49] carried out a particularly thorough investigation based on observations from 13 outstations in the Midlands and eastern England over a period of four years - a total of about 14 000 observations. They showed that the differences between stations were small enough to be ignored in practice and produced composite curves which can be applied for forecasting night minimum temperatures at most places within the area studied. The diagram and method of use are shown in Figure 9.

(i) The maximum temperature, T_{max}, is plotted at the appropriate time on the upper centre portion of the diagram. A horizontal line (isotherm) is drawn to the left until it reaches the appropriate dew-point in the upper left-hand part of the diagram; for clear skies in summer, for example, the top line of dew-point values is to be used if an afternoon inversion exists with its base below 900 millibars, the bottom line if no such inversion exists. From this point a line is drawn parallel to the sloping grid lines until it reaches the main diagonal. The vertical co-ordinate of this point of intersection is the temperature, T_r, at the time of discontinuity, given by the left-hand curve below the central portion of the diagram. This is a convenient way of solving the regression equation (17.24):

$$T_r = \tfrac{1}{2}(T_{max} + T_d) - D$$

with D = 3.0 when an inversion is present and D = 1.0 when no low-level inversion exists.

(ii) Subsequent cooling so calculated by means of the upper right-hand portion of the diagram, by moving horizontally to the right along the T_r isotherm until the appropriate gradient wind speed curve is reached, dropping vertically to the main diagonal, and then drawing to the left until the time of minimum temperature (taken as sunrise) is reached. A correction to T_{min} is then applied, if necessary, for cloud amount (using diagrams of the type (a)) and length of cooling period (diagram (b) of Figure 9).

Figure 9 is only schematic and fuller details, given below, are needed before the method can be used.

Figure 10 shows the variation throughout the year of the time of discontinuity: the data used were for all stations over a two-year period for

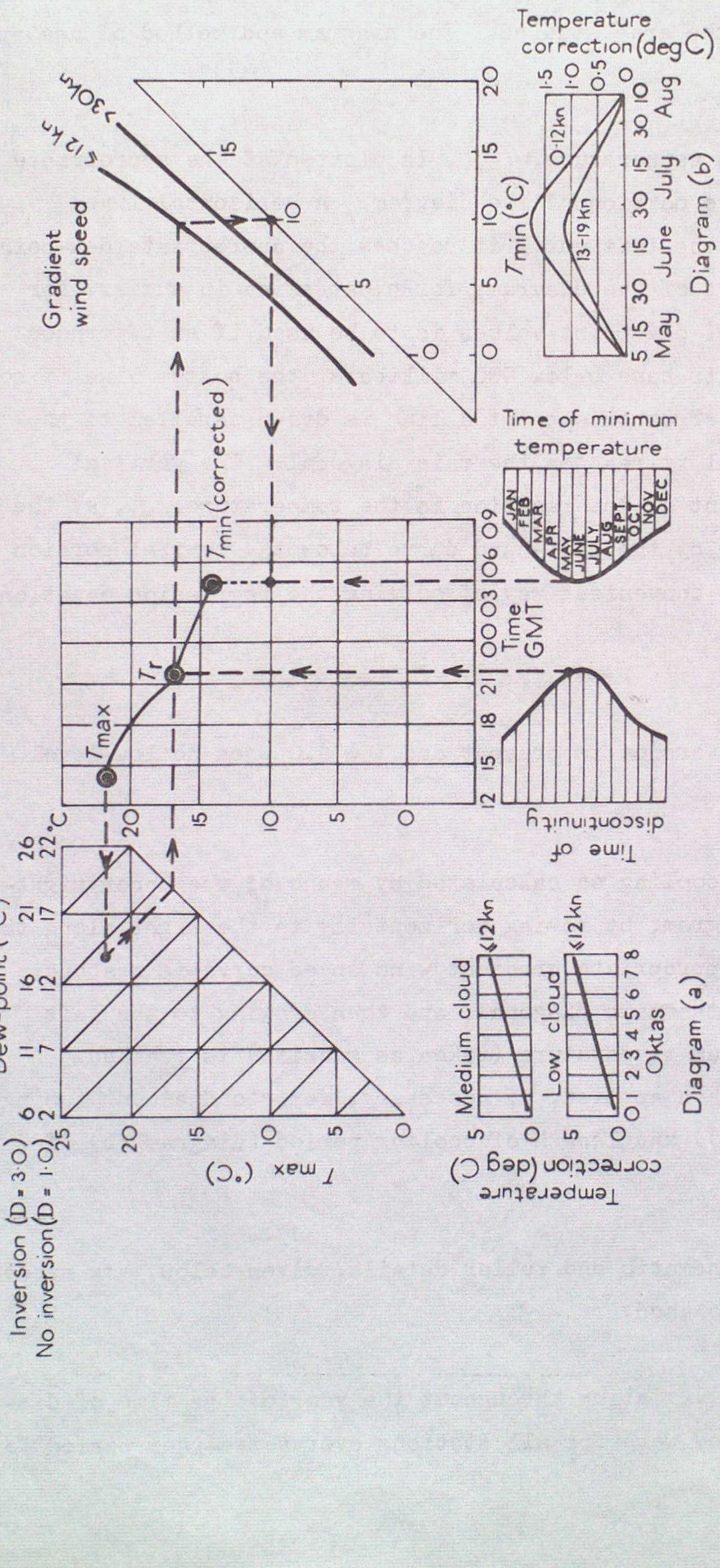


FIGURE 9. Diagram for forecasting a cooling curve

Example: date 30 May, $T_{max} = 22°C$, $T_d = 14°C$, no inversion, wind speed = 10 kn, cloud cover 6/8 low cloud, T_{min} (nil cloud) = 10°C, T_{min} corrected for cloud and reduced cooling = 14°C.

Note. For clarity, only two ranges of gradient wind speed are shown.

Diagram (a): correction to T_{min} for average cloud amount (see also Figure 11).

Diagram (b): correction for reduced length of subsequent cooling (see also Figure 12).

nights with less than two oktas of cloud. The time of T_{min} is also given; the line A, at 0200 GMT, gives the time of minimum temperature over a non-thawing snow cover.

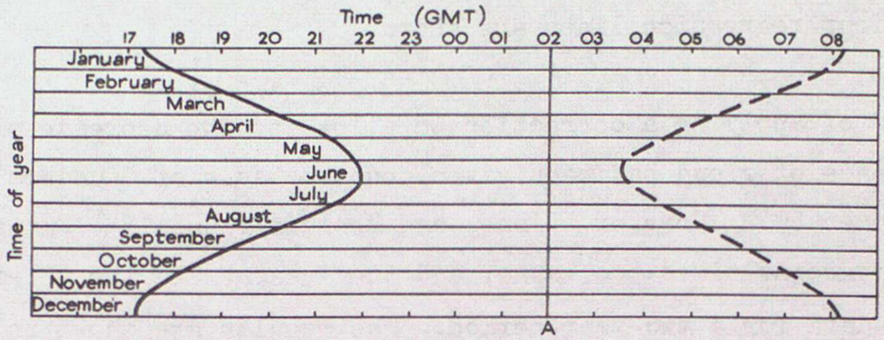


FIGURE 10. Monthly values of times of temperature discontinuity T_r and minimum temperature T_{min}

——— Monthly variation of T_r

— — Monthly variation of T_{min}

A - Time of T_{min} for non-thawing snow cover

Values of D for use in various conditions are given in Table 17.21.

TABLE 17.21 Mean values of the constant D

		Summer	*Winter*
Inversion with base below 900 mb	Radiation nights	3.0	2.5
	8/8 cloud nights	2.5	2.0
No inversion with base below 900 mb	Radiation nights	1.0	0.5
	8/8 cloud nights	0.5	0.0

Regression equations were calculated by Tinney and Menmuir,[49] relating T_{min} and T_r for various classes of gradient wind speed in winter and in summer. The equations were of the form:

$$T_{min} = a + b\,T_r + c\,T_r^{\,2} \qquad \ldots\ldots (17.25)$$

The values of a, b and c are given in Table 17.22.

TABLE 17.22 Values of a, b and c for various classes of gradient wind speed in summer and winter

Gradient wind	Summer			Winter		
speed (V_g) (kn)	a	b	c	a	b	c
13	-6.1	1.19	-0.0141	-5.7	0.92	-0.0091
13-19	-3.3	0.82	0.0015	-4.7	0.86	-0.0051
20-30	-2.3	0.82	0.0015	-4.2	0.86	-0.0051
30	-1.3	0.82	0.0015	-3.2	0.86	-0.0051

Note The values of b and c for $V_g \geqslant 20$ knots have been assumed equal to those for $V_g = 13\text{-}19$ knots, and adjustments made to a only to give the best fit to the data.

These regression equations form the basis of the upper right-hand portion of Figure 9, T_r being the vertical co-ordinate and T_{min} the horizontal co-ordinate; two diagrams are needed, one for each season. In Figure 9 only two of the four regression lines are shown.

A method of applying a correction to allow for the decrease of cooling in the presence of cloud had been given, on the basis of Saunders's original work, by Summersby.[50] However, Tinney and Menmuir found that Summersby's corrections were at times too large, and they reassessed the effect of cloud cover using data for a two-year period. The results are shown in Figure 11, which gives the detailed corrections shown schematically at (a) in Figure 9. If both medium and low cloud are present, only one correction, the greater of the two, should be applied.

A further correction which must be added to T_{min} takes into account the reduced time available for cooling after T_r has been reached: this correction, illustrated at (b) in Figure 9, is shown in greater detail in Figure 12.

17.6.2.6 Night cooling under cloudy skies. The methods described in the preceding section all make some allowance for the reduction in the rate of cooling as the amount of cloud increases, and some make allowance for the type of cloud - the lower the cloud base in general the warmer it is and the greater the 'blanketing' effect. Mizon[51] also has published a diagram, as in Figure 13, which gives values of the factor k for varying cloud amounts of different types, where k is the ratio of the actual cooling on a given night to that which would occur if skies were clear of cloud.

17.6.2.7 Night cooling after fog formation. Often cooling at and near the ground continues after the formation of fog, either because the fog is not completely opaque to the long-wave radiation from the surface and near-surface layers, or because cooling near the fog top is spread downwards by slight turbulence. Barthram[48] produced a diagram, based on two years' observations at Wyton, Huntingdonshire, showing the likely fall of temperature after fog formation as a function of the number of night hours remaining (Figure 14). The diagram may be applicable to other places in low-lying, flat areas of eastern England, but it might be unwise to use it blindly for different types of topography.

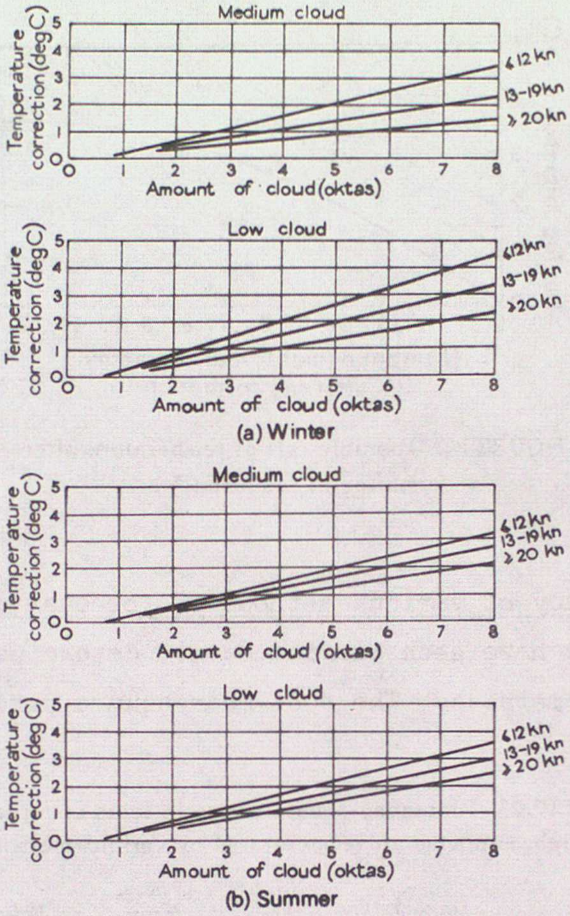


FIGURE 11. Graphs showing relation between average cloud amount and decrease in cooling in (a) winter and (b) summer

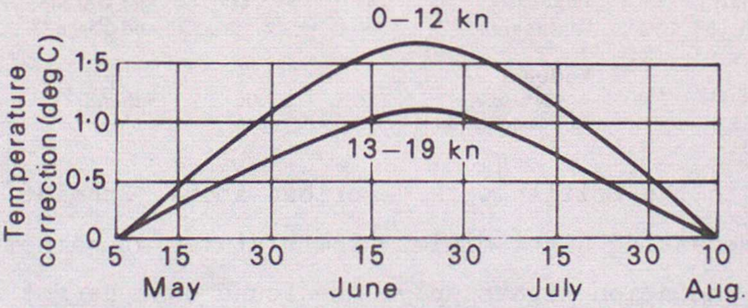


FIGURE 12. Correction for reduced length in summer of the period of cooling between times of T_r and T_{min}

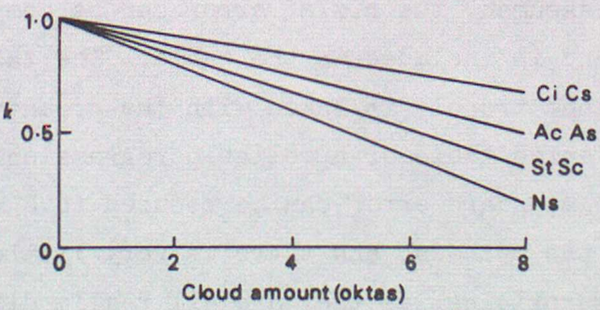


FIGURE 13. Values of *k* for various cloud types and amounts

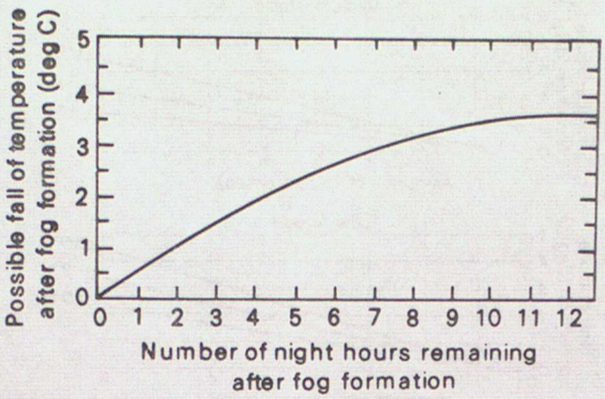


FIGURE 14. Possible fall of temperature after formation of radiation fog

17.6.2.8 <u>Accuracy of various methods of forecasting night minimum temperatures</u>. Tests have been carried out to determine the order of accuracy obtained by various methods. The root-mean-square errors are listed in Table 17.23.

TABLE 17.23 Root-mean-square errors in tests of forecasting night minimum air temperatures by various methods

Station or area	Method	No. of cases	r.m.s. error *degC*	Reference
10 stations in eastern England	Saunders	1427	1.89	Steele, Stroud and Virgo[52]
Mildenhall	Saunders McKenzie Regression	365 731 366	2.16 2.11 2.30	Gordon and Virgo[53] and Gordon, Perry and Virgo[44]
Warnemünde	Various according to season	756	2.06	Schmidt[54]

In addition to the British work described above, Schmidt[54] has also done some work on forecasting night minimum temperatures at Warnemünde on the Baltic coast on radiation nights only. He found that he got the best results with different methods at different times of year. He has not given his results in terms of root-mean-square (r.m.s.) errors, but, if normal distribution of errors is assumed, the r.m.s. error can be computed from the figures given in his paper and is included in the table. The table as a whole indicates that if sufficient trouble is taken with the groundwork (either in obtaining a good McKenzie table or a reliable regression equation or good cooling curves), the standard error can be reduced to a value of about 2 degrees by any of the methods, and there is very little to choose between them. This is not surprising, as they are all really different methods of solving the same basic equation.

If Warnemünde (on the Baltic coast of Germany) can be compared with inland stations in eastern England, it is reasonable to conclude that there is very little to be gained by varying the method according to the time of year. It also seems probable that the standard error of forecasting a night minimum air temperature is unlikely to be reduced much below 2 degrees. This means that a forecast night minimum air temperature will be within $\pm$ 2 degrees of the true value on about 68 per cent of occasions.

17.6.3 Grass minima and ground frost

From 1906 to 1960 inclusive it was Meteorological Office practice to record a 'ground frost' when the grass minimum thermometer reached 30°F (-1°C) or below (30.4°F for thermometers read to tenths) and statistics were compiled on this basis. From 1 January 1961 statistics have referred to a grass minimum temperature below 0°C and in forecasts the use of the term 'ground frost' has signified a grass minimum temperature below 0°C. This change in practice must be borne in mind by anyone who is comparing statistics for the two periods.

The grass minimum temperature is usually several degrees below the screen minimum temperature and, on radiation nights particularly, it is also a few degrees below the bare soil temperature. Since cooling at night takes place by radiation from the surface of the earth it might seem logical to forecast the bare soil minimum temperature directly and then relate the grass minimum to it. All practical methods, however, start from a forecast screen minimum temperature (T_{min}) and then proceed to grass minimum temperature (T_g). There are several reasons for this. Screen temperatures can be obtained at any time of the day or night, but bare soil temperatures and grass temperatures are very difficult, if not impossible, to obtain during the afternoon when the forecast has to be made; normally, therefore, only one grass temperature - the grass minimum temperature - is observed every 24 hours. Moreover the rules for forecasting T_{min} are now fairly well established and provide a good starting point.

17.6.3.1 Ground frost (Faust). Faust[55] conducted an investigation at Münster to determine the conditions under which ground frost would occur on radiation nights with less than 2 tenths of cloud and surface wind speed less than 4 knots during the cooling period. This investigation did not lead to a forecast of the expected night minimum grass temperature but, instead, Faust stated the condition for the occurrence of a ground frost as

$$T_{14} + \tfrac{1}{2}T_{d14} < 17.2 \qquad \dots . (17.26)$$

where T_{14} and T_{d14} are the screen and dew-point temperatures respectively in °C at 1400 local time. Münster is a lowland station. Jefferson[56] tested the formula at Hullavington (on the south-west of the Cotswold Hills, 98 metres above sea level) for all radiation nights which fulfilled the requirements during the period September 1950 to June 1951. From a plot of $(T_{14} + \tfrac{1}{2}T_{d14})$ against grass minimum temperature, Jefferson found that the curve intersected the 0°C isotherms at about 17.2. James[57] also tested the formula at St Athan, South Glamorgan, (46 metres above sea level) for the whole of 1945 and the period November 1949 - May 1951 and found a critical value of 16.7.

From these results it would seem that for radiation nights, as defined, the formula could be applied with fair success at most stations in reasonably level country within 100 metres or so of sea level, but both Jefferson and James found a fair amount of scatter about their curves and from the published graphs it would appear that ground frosts were recorded with values of $(T_{14} + \tfrac{1}{2}T_{d14})$ as high as 19.

Lawrence[58] has refined Faust's method by taking account of the number of consecutive preceding days up to a maximum of 10 on which the reported rain was nil, a trace or 0.01 inch (0.25 mm). He also found he could improve his results if some allowance were made for soil temperature at a depth of 4 inches (102 mm). As very few stations possess soil thermometers (or have soil thermometers in the neighbourhood), this refinement is of limited practical value.

17.6.3.2 Frost forecasting (Smith). L.P. Smith[59] developed a simple diagram for use by agriculturists for frost forecasting and an account has also been given by Meads.[60] The method depends on plotting a large number of past observations taken at a particular time of day - Smith chose 1800 GMT - on a graph in which the abscissae are dew-points (or wet-bulb temperatures, if preferred) and the ordinates are values of the quantity $(T + W + C - R)$, where T is the dry-bulb temperature ($^{\circ}$F)

W is the wind strength in m.p.h.

C is cloud amount in eighths

and R is the number of days since the last rainfall up to a maximum of eight.

Figure 15 shows the diagram for Bristol. To use the diagram the forecaster works out $(T + W + C - R)$ and inserts it on the diagram to accord with the appropriate value of the dew-point. According to where the point falls, he then predicts frost, ground frost or no frost.

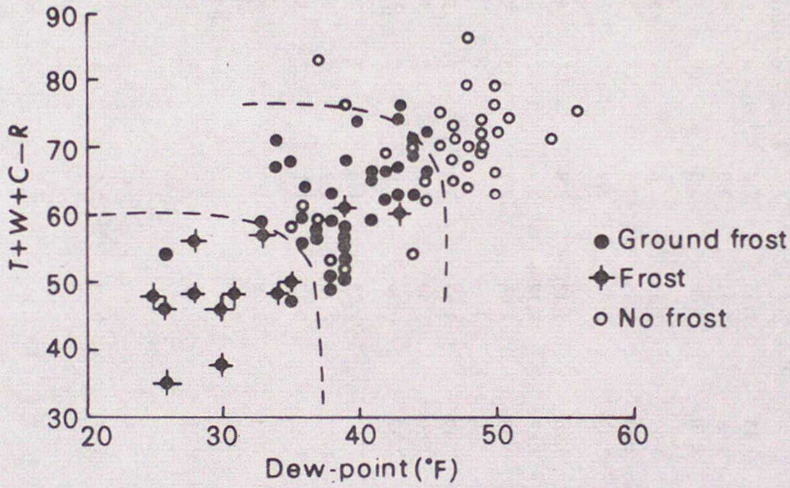


FIGURE 15. Frost-forecasting diagram for Bristol

Halim[61] obtained good results with the method in Pakistan, and Gordon[62] tested the method using 1500 GMT data for Mildenhall.

The method was devised for agriculturists rather than for professional forecasters (hence the use of Fahrenheit and miles per hour); it is worth recommending to any grower who has sufficiently long meteorological records on his own farm.

17.6.4 Depression of the grass minimum temperature below the air minimum temperature

Craddock and Pritchard[43] in their statistical investigation used actual values for 38 cases from 16 stations to construct a table of mean values of the depression $T_{min} - T_g$ under various conditions of cloud amount and geostrophic wind speed. Saunders[48] compiled a similar table for clear nights

TABLE 17.24 Summary of depressions of grass minimum temperatures below air minimum temperatures for 10 stations in eastern England (October 1961 - September 1965)

Weather	State of ground	Cloud amount	Geostrophic wind speed	Number of cases	Mean depression	σ	Median	25 per cent limits		45 per cent limits		Category
		oktas	*knots*		*degrees Celsius*							
No precipitation, No fog, No snow cover, $T_{min} \geqslant$ °C	dry	2 or less	0-12	366	3.35	1.33	2.8	2.0	4.0	1.3	4.7	A
			13-24	584	3.08	1.31	2.8	2.0	3.7	1.4	5.0	B
			⩾ 25	182	2.70	1.06	2.3	1.9	3.1	1.2	3.9	C
	moist or wet	2 or less	0-12	69	3.09	1.33	2.6	1.7	3.5	1.2	4.7	D
			13-24	380	2.63	1.15	2.5	1.7	3.4	1.4	4.3	E
			⩾ 25	370	2.43	0.87	2.2	1.7	2.9	1.3	3.3	F
	dry	3-5	0-12	228	3.26	1.13	2.9	2.1	4.2	0.9	5.3	G
			13-24	379	2.67	1.32	2.4	1.7	3.6	1.0	5.1	H
			⩾ 25	143	2.36	1.06	2.1	1.5	3.0	0.7	4.8	I
	moist or wet	3-5	0-12	118	2.89	1.07	3.0	1.7	3.7	1.2	4.7	J
			13-24	290	2.50	1.59	2.1	1.5	3.4	0.9	4.8	K
			⩾25	287	2.32	0.96	2.0	1.5	3.1	0.8	4.1	L
	dry	6-8	0-12	366	1.67	1.25	1.4	0.7	2.4	0.2	3.5	M
			13-24	544	1.52	1.12	1.2	0.6	2.1	0.2	3.0	N
			⩾ 25	214	1.37	0.87	1.1	0.7	1.8	0.3	2.5	O
	moist or wet	6-8	All	893	1.63	1.08	1.3	0.8	2.1	0.4	2.9	P
Rain, No snow cover, $T_{min} \geqslant$ °C	moist or wet	6-8	All	863	0.73	0.49	0.5	0.4	1.3	0.0	3.4	Q
Falling snow	snow-covered	6-8	All	46	0.57	0.18	0.5	0.4	0.8	0.0	1.5	R
No precipitation, No fog	snow-covered	0-5	All	142	2.47	1.30	2.3	1.0	4.1	0.2	5.6	S
		6-8	All	61	1.47	1.29	1.2	0.4	3.0	0.1	4.7	T
Fog by 21 GMT, $T_{min} \geqslant$ °C	no snow cover	Sky obscured	All	74	0.42	0.24	0.2	0.0	0.6	0.0	1.6	U

at Northolt between August 1948 and April 1950 and, in addition, expressed the scatter of observations about the mean in terms of the standard deviation. The most comprehensive tabulation of this sort has, however, been carried out by Steele, Stroud and Virgo[63] who derived Table 17.24 from well over 6000 observations made at 13 inland stations in eastern England. As the distributions are not normal, there is no simple way of expressing in terms of the standard deviation the percentage of observations lying within various limits on each side of the mean. They therefore calculated the 25 per cent limits and 45 per cent limits on each side of the median and included them in the table; these limits correspond to ranges of 50 per cent and 90 per cent respectively. When ground temperatures fall below 0°C and the ground itself is not frozen, significant quantities of heat are released during the freezing of the water in the ground, but Steele, Stroud and Virgo found too few cases to make separate categories. They also developed a method of combining the scatter of forecasts of T_{min} about the mean with the scatter of $T_{min} - T_g$ to enable them to calculate the percentage probability of a ground frost corresponding to a given forecast night minimum air temperature (T_{min}) and the results are summarized in Table 17.25.

TABLE 17.25 Percentage probability of a grass minimum below 0°C for certain forecast night minimum air temperatures

Category*	Forecast air minimum (°C)												
	+8	+7	+6	+5	+4	+3	+2	+1	0	−1	−2	−3	−4
								per cent					
A	1	3	8	16	29	45	62	77	88	95	98	>99	
B	1	3	8	16	29	45	63	77	88	95	98	>99	
C	<1	2	5	11	21	37	55	72	86	94	98	>99	
D	1	3	7	14	26	40	57	73	87	94	97	>99	
E	1	2	6	13	24	39	56	73	85	93	97	>99	
F	<1	1	3	8	17	32	51	69	84	93	97	>99	
G	<1	3	8	15	28	44	61	76	87	94	98	99	>99
H	<1	2	5	11	21	35	53	70	84	92	97	99	>99
I	<1	1	4	8	17	31	48	66	81	91	97	99	>99
J	<1	2	5	12	23	39	57	73	86	94	98	99	>99
K	<1	1	4	9	19	33	50	67	82	91	97	99	>99
L	<1	1	3	7	15	29	47	65	81	91	96	99	>99
M	<1	1	3	6	13	24	38	56	72	85	93	98	>99
N		1	1	4	10	20	34	52	70	86	93	97	>99
O		1	1	3	8	17	32	50	69	84	93	97	>99
P		1	2	4	11	21	34	55	72	86	94	98	>99

* The categories are here assumed to be defined by forecasts of the various variables including air minimum.

The categories denoted by letters of the alphabet correspond with those in the right-hand column of Table 17.24. Rounded off to the nearest 10 per cent or any other convenient value, the figures in this table could be used by forecasters who require a statement of probability of ground frost over the general area of eastern England.

Steele, Stroud and Virgo found that there were statistically significant differences between the mean values in categories A - O inclusive at most of the stations in their investigation and they were unable to find any simple way of transferring results obtained at one station to somewhere else in the neighbourhood. They therefore concluded that, unless actual observations are available from the place in question, the best that can be done is to use mean values for the area.

Hogg[64] investigated the values of $T_{min} - T_g$ which actually produced ground frosts on radiation nights at seven inland places in south-west England, and concluded that on those particular occasions the mean depression was 4 degrees (Celsius). This is about 1 degree greater than the corresponding figure which Steele, Stroud and Virgo[63] found for the 13 stations they examined in eastern England. Hogg found a slightly larger depression in spring and summer than in autumn and Steele, Stroud and Virgo confirm this; they estimate, however, that the seasonal means are within 10 per cent of the annual means. Although the seasonal variation is not large, the variation from one site to another may be considerable; slope of the ground, type of soil and compactness of the surface layers may all be contributing factors, although their precise contributions are not yet known. Therefore it cannot be too strongly emphasized that results obtained at one place should not be transferred elsewhere without good reason.

17.6.4.1 Graphical method. Sills[65] plotted a scatter diagram for Cottesmore with 1800-0900 GMT mean cloud cover and mean gradient wind speed as parameters; he based his diagram on one year's observations. Only low cloud (base 2400 metres or less) was considered and sky obscured was counted as 8/8. Isopleths of best fit of grass-minimum depressions were drawn and Figure 16 obtained.

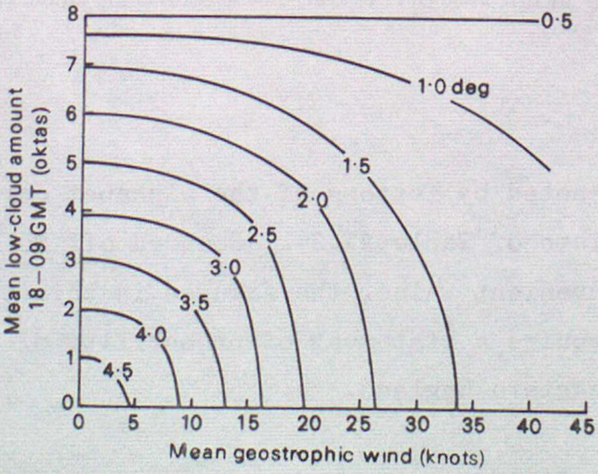


FIGURE 16. Isopleths of grass minimum depression below air minimum temperature at Cottesmore

Cottesmore was one of the stations used by Steele, Stroud and Virgo, and Sills was therefore able to compare the results obtained from his diagram with those obtained from their tabulations for Cottesmore; the overall results obtained by his diagram were slightly better.

In a similar diagram for West Raynham, mean surface wind speed was used instead of mean gradient wind speed. Until comparative statistics exist, the choice of the wind parameter remains a matter of local preference.

17.6.5 Minimum temperatures at or over various surfaces

17.6.5.1 Minimum temperature at a soil surface (Gloyne). Minimum temperatures over bare soil have been investigated by Gloyne[66] for radiation nights only at Starcross, Devon, during the two-year period 1949-50. The site was 8.8 metres above mean sea level a few hundred metres from the west bank of the Exe estuary; the soil was sandy loam with a good deal of silt. Following Hogg,[64] Gloyne defined radiation nights as those nights on which the grass minimum was 7 degF (approximately 4 degC) or more below the screen minimum temperature, and on the vast majority of these nights it was found that the screen minimum was higher than the bare soil minimum which was in turn higher than the grass minimum temperature. The table of mean values given by Gloyne, converted to degrees Celsius, is given in Table 17.26.

TABLE 17.26 Values of means of bare-soil minimum temperature minus grass minimum temperature at Starcross, Devon on radiation nights in 1949 and 1950

	J	F	M	A	M	J	J	A	S	O	N	D
1949	1.4	2.1	2.4	2.9	2.3	2.7	2.4	1.9	2.2	2.4	1.3	1.1
1950	-	0.8	1.3	2.2	2.3	2.7	2.2	2.5	2.1	2.6	2.0	2.1

For all nights (including radiation nights) in one year Ritchie[67] derived Table 17.27 for Wyton which is on clay at 39.6 metres above mean sea level.

TABLE 17.27 Values of monthly means of bare-soil minimum temperature minus grass minimum temperature at Wyton, Hunts. for 358 nights between October 1967 and September 1968

J	F	M	A	M	J	J	A	S	O	N.	D
0.6	1.3	1.0	0.9	0.7	0.9	0.9	0.4	0.1	1.1	1.2	0.6

Although Tables 17.26 and 17.27 are both based on fairly short periods, they do show the order of magnitude of the quantities involved.

17.6.5.2 Minimum temperature at a concrete surface. Both Ritchie[67] at Wyton and Parrey[68] at Watnall have investigated minimum temperatures over large surfaces of concrete comparable with roads. One of Ritchie's surfaces was ordinary concrete and the other was similar concrete of the same texture and thickness coated with a layer of black bitumastic paint; the blackness of the surface seemed to have very little effect on the minimum temperature. Ritchie also concluded that minimum temperatures are less erratic over concrete than over grass and also almost always higher. Parrey concluded that 'for a given road the depression of the road temperature at night below the air temperature at four feet (above surface level) depends largely on the length of time available for outgoing radiation.' As he confined his researches to the colder months of the year, this thesis is illustrated in this handbook by the annual trend given by Ritchie in Figure 17.

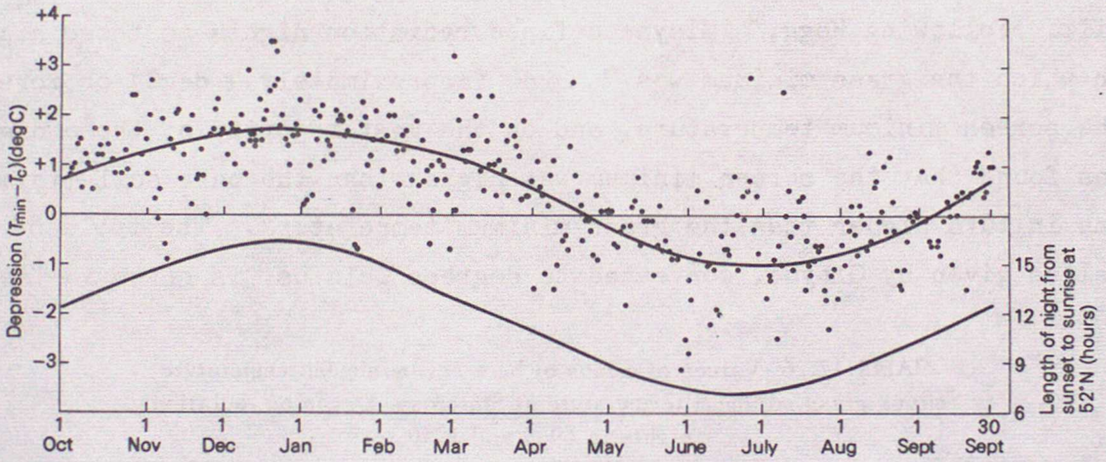


FIGURE 17. Depression of road minimum below air minimum temperature (upper curve) compared with length of night (lower curve)

The forecast concrete surface minimum temperature, T_c, is obtained from the forecast air minimum, T_{min}, by applying the difference $T_{min} - T_c$ read from the curve, or by use of the equation derived by Parrey:

$$T_{min} - T_c = 0.28t - 2.9 \qquad . \; . \; . \; . \; (17.27)$$

where t is the interval between sunset and sunrise in hours. The standard deviation of individual values from the curve, or from the value given by the regression equation, is about 1 degree.

Clark,[69] using temperature data at various depths in thick concrete slabs, tried an alternative approach. He found that the thermal diffusivity (see 17.4, page 11) of concrete was very much greater than that of any kind of soil, with consequent penetration of the diurnal heating to greater depths. Clark further found that the cooling of the surface from sunset to sunrise depended only upon the time of year and the overnight cloud cover: the values are given below in Table 17.28: the method showed a root-mean-square error of about 1.5 degrees when tested on independent data.

TABLE 17.28 Cooling of surface of a thick concrete slab overnight for each winter month for various ranges of cloud cover

Cloud cover	Nov.	Dec.	Jan.	Feb.	Mar.
oktas			*degC*		
0-2	5.5	4	3	5	8.5
3-6	4	3	2.5	4	6.5
7-8	2	1	1	2	4

In effect, what Clark found was that the properties of the concrete, and the rate at which the surface could lose heat by radiation, dominated the cooling: the state of the atmosphere near the surface played only a minor role.

Thornes,[70] using data similar to that of Clark, but with a more rigorous approach, showed that the water vapour content of the lowest layers and turbulence were also important. For full details the reader should refer to the original paper; the method is not described here as it (like Clark's method) needs a concrete-surface temperature, a quantity not normally measured, at sunset, a time too late for any forecast to be of much value.

At a number of meteorological offices, observations are made of the minimum temperature of a concrete slab, size 90 by 60 centimetres, and 5 centimetres thick. The slab is laid on a layer of sand, with soil beneath. Although the observations are of value in that they are more consistent than grass minimum temperatures, the relative thinness of the slab leads to differences in behaviour of the temperature at the slab surface and that at a road surface. In particular, the difference between the air minimum temperature and the slab minimum temperature does not show the sinusoidal variation with time of year of Figure 17, and the mean differences for a given month vary considerably from place to place. It might be unwise, therefore, to try to devise a forecasting technique for road-surface minimum temperatures using slab minimum temperatures, although Parrey, Ritchie and Virgo[71] have used the latter for verification purposes.

17.6.6 Forecasting temperatures two or more days ahead

Most of the methods described in the preceding sections cannot be used for forecasting temperatures more than a day ahead, the exception being that of Boyden, described in 17.6.1.2 (page 25), relating the 1000-500-millibar thickness to the surface maximum temperature. Parrey[72] has extended Boyden's method to the forecasting of minimum temperatures also: he has derived eight diagrams, four for minimum temperature and four for maximum temperature, relating the appropriate temperature at Watnall for each month of the year to the 1000-500-millibar thickness over the area: each of the set of four diagrams is for geostrophic wind direction within a given quadrant. A sample diagram is reproduced at Figure 18; it applies strictly only to Watnall, and for the method to be used at any other station a similar set of diagrams would need to be derived from the data for that station.

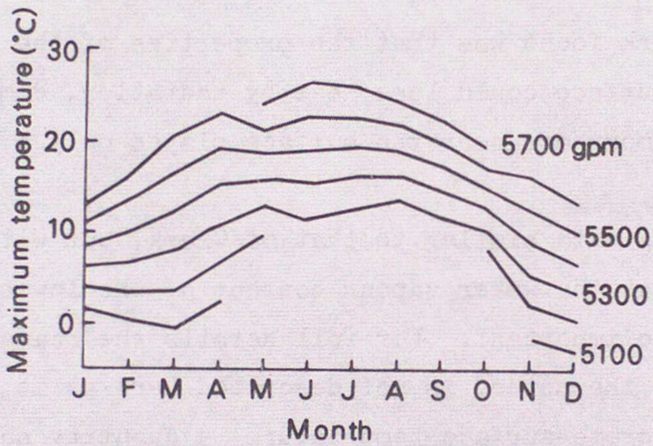


FIGURE 18. Maximum temperature related to 1000-500-mb thickness for geostrophic winds in the north-west quadrant for each month of the year

Obviously the period for which the temperature forecasts can be given depends upon the period for which thickness prognoses are available, but Parrey found that by taking climatological values into account he could extend the period over which useful forecasts could be made. For example, he forecast the maximum temperature on the third day as the mean of the forecast maximum temperature for the second day and the climatological normal.

As mentioned in 17.6.1.2 (page 25) the use of the 1000-500-millibar thickness has its disadvantages, and the use of shallower layers from the surface would probably produce an improvement, provided the prognoses for that layer are as good as those for the 1000-500-millibar layer.

17.7 SOME FACTORS INFLUENCING TEMPERATURE

17.7.1 Effect of surface wind

The existence of a surface wind reduces the diurnal variations of temperature, the associated turbulence spreading heat upwards from the surface when heating is taking place, and transporting heat downwards at night to offset the radiative cooling to some extent. The stronger the wind, the greater the mechanical turbulence set up and the smaller the diurnal variations. The effects of surface winds have been included in many of the formulae and techniques described above.

In coastal regions onshore winds exert considerable control on temperatures both in winter and summer and by day and night. The annual variation of sea temperatures around the British Isles is considerably less than that of the land and the diurnal variation is but a small fraction of that of the land. Temperatures of the air near and in contact with the earth's surface are closely controlled by the temperature of the underlying surface. It is widely known that in coastal regions with onshore winds the diurnal variation of temperature is markedly reduced. The extent to which this effect penetrates inland varies widely with topography and meteorological conditions. The seasonal effect on temperatures is largely responsible for the temperate climate of the British Isles. In regimes of easterly wind the passage of air across the North Sea materially modifies the characteristics of the lower-level airstream and some rules for the computation of temperatures are given in 17.7.8.1 (page 67).

Apart from the general effect of surface winds on temperatures, local winds exert a powerful effect on temperatures in certain localities under given meteorological conditions. Local winds of most importance for forecasting of temperature in the United Kingdom are land- and sea-breezes and katabatic winds.

17.7.1.1 Effect of land- and sea-breezes. It is sufficient for the purpose of this chapter to regard land- and sea-breezes as being caused by the differential heating and cooling of land and water surfaces by day and by night. In the absence of a pressure gradient, insolation will cause temperatures to rise to higher values over the land than over the water. The resulting difference in density leads to the isobaric surfaces in the free atmosphere being higher over land than over the sea, thereby causing a flow from land to sea at those levels and this in turn induces air in the lower layers to move from coastal waters across the coastal region. The sea-breeze usually commences to blow at right-angles to a flat coast but, as the geostrophic control gradually exerts its effect, the sea-breeze gradually changes its direction, tending to blow parallel to the coast as though there were a depression over the land. Geographic features cause sea-breezes in some localities to differ materially from this general rule. The sea-breeze advects cooler (and moister) air to some distance inland and this reduces day temperatures along a coastal strip. In practice, a pressure gradient normally exists and the associated general wind must be vectorially compounded with the sea-breeze component to obtain an estimate of the wind direction and speed for any locality on any one day. It is obvious that a sufficiently strong pre-existing wind from land to sea may inhibit sea-breezes.

On the coast the sea-breeze may set in well before noon and the diurnal rise of temperature is then severely checked. The penetration of the sea-breeze inland appears to depend on the pressure gradient, the stability of the air and on topography. No general rules can be given which apply to all localities. The sea-breeze may set in in one of two characteristic ways. When the sea-breeze reinforces the gradient wind (that is, the gradient wind blows from sea to land) the onset of the sea-breeze is marked by no abrupt effect on surface temperatures and possibly only by a strengthening of the surface wind with little change in direction. When the gradient wind opposes the sea-breeze, the onset of the sea-breeze may be quite sudden, wind direction then changes abruptly - sometimes by 180° - and there is often a marked drop in temperature. This latter type of sea-breeze is most readily recognized with limited instrumental observations and the sudden cooling and change of wind are often quite noticeable without instruments.

The extent of the penetration of a sea-breeze inland is variable and for a discussion of this the reader is referred to Chapter 16 - Wind.

The effect of sea-breezes on temperatures is also quite variable. In an investigation at Worthy Down, Peters[73] found that about half the sea-breezes were accompanied by a sudden drop in temperature of 1 - 4 degF (0.5 - 2 degC) at their onset or an abrupt commencement of the usual evening fall of temperature. (Worthy Down is about 40 kilometres inland from the Solent and nearly 64 kilometres from the northern extremity of the Isle of Wight. Sea-breezes attain a maximum frequency between 1600 and 1800 GMT at Worthy Down.) He did not say what proportion of the temperature changes was associated with a sudden and readily recognizable onset of a sea-breeze.

For 58 occasions on which sea-breezes were positively identified at Worthy Down, Peters found that on only two of these occasions had Worthy Down a maximum temperature no higher than that at St Catherine's Point, Isle of Wight, while on 12 occasions the maximum temperature inland was more than 5 degrees higher than that on the coast.

Land-breezes are less well marked and less extensive in both horizontal and vertical extent than sea-breezes. They probably exert little effect on temperatures on coasts except when a light onshore wind is reversed.

17.7.1.2 Effect of katabatic and anabatic winds. Katabatic and anabatic winds are caused by the fact that air over a cooled slope is normally colder (air over a warmed slope normally warmer) than air at the same level over neighbouring lower ground. In the presence of such differences air does not normally remain in equilibrium and local circulations are established in which there is a downward component of surface wind over a cooled slope and an upward component over a warmed slope. It follows that katabatic and anabatic winds are closely related to local topography.

In general, the effect of katabatic winds is to cause valleys to fill with air which is somewhat colder than that at higher levels on the hillsides thereby causing rather lower minimum screen temperatures at the lower-lying stations. Where the local topography is in the form of a basin or hollow from which the cold air cannot flow, the area is often said to be in a frost pocket or frost hollow since these areas tend to experience more frequent and severe frosts and a shorter frost-free season than the neighbouring areas. Forecasts of frosts for horticultural and commercial purposes may be very important and the extent to which katabatic winds may contribute to lower minimum temperatures at localities at the base of hills or in frost pockets must be considered for each locality.

Where coastal regions are backed by a range of hills or a high land mass, katabatic and anabatic winds reinforce land- and sea-breezes respectively and the combined effect on temperature is rather greater than if only one 'local' wind was blowing.

The modifications to temperatures due to local winds are so variable and so intensely local tnat general rules for forecasting the variations cannot yet be given. In addition, these variations depend not only on the local wind but also on the type of soil, the moisture content and the distribution and type of vegetation in the locality. In spite of the complexity of the problem, forecasts have to be given and the only sound procedure seems to be to forecast from physical reasoning and modify the method as experience is gained of the locality. Forecasters may derive a further useful insight into the problems from 'Nocturnal winds' by Lawrence[74] and from some of the references therein, particularly 'The climate near the ground', by Geiger.[75] This is a full and easily readable account of climate in the lowest two metres of the atmosphere.

17.7.2 Effect of cloud

The effect of cloud on diurnal heating and cooling has been taken into account in most of the methods discussed so far. During the day, a cloud top reflects most of the solar radiation falling upon it. A complete cover of cloud which persists all day reduces considerably the diurnal rise of temperature. If the cloud is of suitable form and not too thick, it may be 'burned off' during the day; the forecasting of maximum temperature may be carried out by the method of 17.6.1.4 (page 29). At night, the presence of cloud leads to a reduction in the fall of temperature, the loss of energy by long-wave radiation from the ground being offset by back-radiation from the base of the cloud. Both the cloud amount and the temperature of the base are important. Even thin cloud will radiate almost the full black-body radiation downwards. The amount of downward radiation therefore depends largely on the height of the cloud base: the lower the cloud, the greater the 'blanketing' effect.

17.7.3 Effect of fog

Fog at low levels may be regarded as a layer of cloud in contact with the ground and the effect of fog on temperatures could be treated quite generally by discussing the differences in the radiation balance on occasions of no fog and those of fog. Detailed information on temperatures near the ground is available and this justifies a more complete discussion of the effect of fog on temperatures. Individual fogs show marked deviations from the temperature regimes described below but, as very precise and frequent readings of temperatures in the vertical are unlikely to be available at other than a very few specially selected sites for research purposes, or at a few major outstations, a description will be given of the sequence of events in the temperature field on a typical radiation night with clear skies on which fog forms after sunset.

The afternoon preceding a typical radiation night may be regarded for this explanation as one of light winds and little or no cloud with insolation dominating temperatures and the lapse rate in the lowest hundred metres or so. The surface of the ground will be warm and the lapse rate in the lowest layers will be generally super-adiabatic or at least dry-adiabatic. As the sun sinks towards the horizon the out-going radiation from the earth's surface exceeds the incoming solar radiation by a large amount; the warm ground cools rapidly and soon reaches a temperature several degrees cooler than tne air just above the ground. The air in immediate contact with the ground is cooled and Best[19] found that in the lowest 1.2 metres the afternoon super-adiabatic lapse had been converted into an inversion approximately 100 minutes before sunset. The air in the lowest 30 metres or so loses heat by long-wave radiation to the ground and by eddy transfer of heat towards the ground. This rapid cooling of the ground and the air in the lower levels is not prolonged and normally between sunset and one or two hours later the rate of cooling shows a discontinuity (see 17.6.2.5, page 34). Subsequently temperatures continue to fall but at an appreciably slower rate. Although the fall of temperature at screen level is not always steady and minor fluctuations of temperature are superimposed on the trace of the thermograph, for forecasting purposes it may still be regarded as a slow, steady fall. According to Stewart[76] the level of most rapid fall of temperature at sunset is close to the ground but, a few hours later, it seems to be near the level of maximum temperature - probably a hundred metres or so above the ground. Temperature gradients near the ground thus increase to maximum inversion conditions shortly after sunset and then tend to decrease slowly as cooling at higher levels becomes greater than near the ground. On a clear night, cooling in the lower levels continues and we shall assume that fog forms. Fog usually forms in a thin layer at ground level and then grows vertically. After a period of growth which may be quite rapid the fog becomes sufficiently dense and deep to affect the radiation balance materially. Marked changes then occur in the temperature regime and these changes are particularly well marked both below and above screen level. Temperatures at ground level increase sharply to about the screen-level value, partly due to the fog acting as a 'black-body' radiator and partly due to heat conducted upward through the soil. The fog, now acting as a black body, will also be radiating strongly in an upward direction from some level below the fog top. This outward radiation will cause temperatures near the fog top to decrease. This decrease can be quite rapid. In an account of temperatures on a night of radiation fog, Johnson and Heywood[77] observed that the temperature at 12.4 metres began to fall rapidly (at 0045 GMT) as the fog thickened. The fall amounted to 4 degF (2.2 degrees Celsius) in 25 minutes and by 0130 GMT the fall had been sufficient to convert the inversion into a lapse of temperature below 12.4 metres. At 0055 GMT the temperature at the 30-metre level also began to drop rapidly and by 0130 GMT it had fallen by about the same amount. Similar falls occurred in succession at higher levels as the fog became deeper. At heights above the maximum level to which the fog extended, the fall in temperature was only about 1 or 2 degF (0.5 or 1 degree Celsius).

As the radiative processes continue, temperatures at or near screen level show only small falls but radiation from the fog top cools the air at these upper levels and diminishes the inversion of temperature from the ground to this level. If the fog persists for a few hours this differential rate of cooling transforms the inversion in the fog layer to a lapse. Stewart[76] states that in deeper fogs the lapse rate usually seems close to the saturated adiabatic value 0.6 degrees per 100 metres. Above the fog layer is a steep inversion.

This type of temperature distribution persists until dawn. As the sun rises and incoming solar radiation increases, temperatures in the lower levels rise and, except for persistent winter fogs, the temperature inversion above the fog is normally broken down and after some time the fog disperses. For more detailed discussions of temperatures in specific radiation fogs reference should be made to papers 76, 77 and 30 in the Bibliography.

17.7.3.1 Sea fog. In coastal areas the normal diurnal variation of temperature may be markedly affected by the presence of sea fog. Such fog may be carried onshore by the general wind or on occasions by the sea-breeze. The fog has a high albedo so that only some 20 per cent of the incoming solar radiation is available for warming the ground and fog layer. At night sea fog may drift 160 kilometres or more inland but during the day, except in winter, it is usually cleared by solar radiation except from a coastal strip of less than 24 kilometres. While the fog lasts the temperature will not vary much from the sea temperature, and the return or onset of fog with the sea-breeze may bring a drop of 5 degrees or more in temperature.

The advection of sea fog from the North Sea across the east coast of Great Britain and the extent of its penetration inland always seem to present a difficult problem. A detailed account of a number of occurrences has been given by Lamb.[78]

17.7.4 Effect of snow cover

The high albedo of snow, particularly when fresh, its efficiency in allowing energy to escape from the surface as long-wave (nocturnal) radiation, and its poor conductivity (only about one-tenth of the value for dry soil) lead to pronounced differences between the behaviour of temperatures over snow and over snow-free surfaces. During the day-time, only a small proportion of the short-wave radiation falling on a snow surface is absorbed, so little heat is available for warming the air near the surface, and temperatures remain low. Maximum temperatures are most often within the range +1 to +3°C, and only rarely below -2°C. If the ground is only partially covered by snow, or if the snow is thawing, maximum temperatures are mainly within the range +1 to +5°C.

At night, the snow surface rapidly emits long-wave radiation; if skies are clear and winds are light, the surface cools very rapidly, and the temperature of the air at screen level is likely to fall 2 - 4 degrees below the minimum calculated by the methods of 17.6.2 (page 31). The minimum temperature is generally reached earlier in the night than when the surface is snow-free (see 17.6.2.5, page 34). If cloud is present, however, back-radiation prevents such a rapid fall of temperature, and the existence of anything stronger than a very light breeze, with the associated turbulence bringing down heat from the higher layers, has the same effect. Temperatures are likely to fall to about 1 degree lower than over a snow-free surface over a fairly wide range of cloudiness and snow cover until the wind speed reaches 15 knots or so, when the differences become very small.

17.7.5 Effects of topography - airflow over hills

The great complexity of airflow over hills was described in Chapter 16 - Wind, and it is obvious that the vertical components of these motions will modify the vertical distribution of temperature. Many of these local currents are so limited in scale, so essentially disorganized or turbulent that the effect of most of them on temperature is usually ignored in forecasting for flights by powered aircraft. Flights in gliders may require rather more detailed information but forecasts for gliding are usually provided for selected areas (at least at the commencement of the flight) and both pilot and forecaster tend to be 'specialists' in the type and scale of phenomena of particular interest for gliding. No attempt is made in this handbook to discuss this specialized aspect of forecasting.

There are certain types of air flow across hill barriers which are fairly well organized or follow a set pattern and the effect of three of these types on temperatures will be described.

17.7.5.1 <u>Föhn effect</u>. One of the most spectacular effects of topography on temperature is the 'föhn effect'. The term 'föhn wind' originally applied to 'a warm, dry wind that sometimes blows down Alpine valleys' (Lockwood [79]), but it is now used to mean any wind which is blowing down a mountain, and which is warmer and drier than the air at the same height on the upwind side of the mountain. The classic explanation of the phenomenon was given by Hann[80] in 1866, who attributed it to the ascent of moist air on the windward side of a mountain, with condensation and eventually precipitation of some of the water, and cooling at the saturated adiabatic lapse rate after condensation had started. On the descent, evaporation of the water droplets left in the cloud would take place fairly quickly, and warming thereafter would be at the dry adiabatic lapse rate, resulting in the air at a given level being warmer and drier on the leeward side than on the windward side.

It is now thought that this explanation is not entirely correct: it does not explain why the warm air should descend on the lee side, and in any case föhn winds and the associated temperature rise have been observed without the formation of precipitation on the windward side. It now seems more likely that the descent of air on the lee side is often a result of subsidence of upper-level air, airflow at lower levels being blocked by the mountains, and the subsidence being materially assisted by the descending portion of large amplitude lee waves (see Figure 19). The upper-level air, in the stable conditions in which the lee waves occur, would have a higher potential temperature and lower water-vapour content than that originally at the surface, and on arrival at the surface would therefore be warmer and drier than that which it replaced.

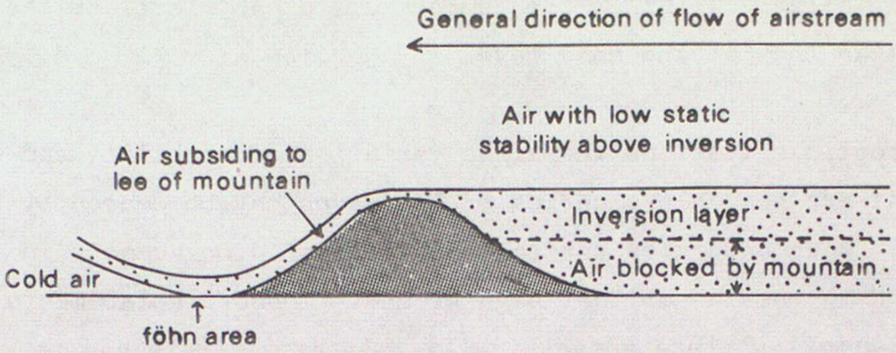


FIGURE 19. A typical föhn airstream

Examples of the föhn effect have been given by a number of authors, including McCaffery,[81] Lawrence,[82] Lockwood[79] and Girdwood.[83] In the instance quoted by McCaffery, south-easterly airflow across the Cairngorms resulted in a temperature at Kinloss about 3 degrees higher than that at Leuchars. Lawrence drew attention to a particularly well-marked example which occurred on 18 February 1945. An unusually warm and moist south-westerly airstream had been advected quite rapidly from the region of the Azores towards Scotland. Air which had passed Aldergrove with a screen temperature of 11°C at 1700 GMT on the 17th reached Dallachy (near Buckie) around 0300 GMT on the 18th with a temperature of 15-16°C. Lockwood reported an occasion, among others, when a mild, dry southerly airstream covered much of the British Isles, with temperatures generally about 10-13°C, rather above normal for mid-March. At both Elgin and Gordon Castle (near the Moray Firth) temperatures of 22°C were achieved, while in Northumberland, to the lee of the Pennines, 23°C was recorded. At Kinloss the start of the föhn wind was accompanied by a temperature rise of nearly 7 degrees in less than half-an-hour. Rapid fluctuations of temperature have also been reported by Girdwood: on one occasion in May 1960 at Turnhouse (Edinburgh) the temperature rose from 15°C to about 21°C, and then fell to 12°C, all within two hours, the main part of the rise and fall occupying less than an hour. This event was explained as the temporary replacement of cool moist air from the North Sea by air which had passed over the Pentland and Moorfoot Hills and had been warmed by the föhn effect.

However, the föhn wind and associated temperature rises do not always occur when air flows over a range of hills.

17.7.5.2 <u>Effect of forced ascent of air</u>. When air with a lapse rate which is less than the dry-adiabatic rate is forced to ascend bodily as in surmounting a range of hills the temperature at any level on the hillside is lower than that at the same level in the free air.

In synoptic situations where the variation of stability and wind flow with height are suitable, a series of lee waves may be caused by high ground and the air then undergoes substantial vertical displacement in the neighbourhood of and to the lee of that high ground. These displacements may be regarded as taking place adiabatically. If an estimate can be made of the vertical extents of these displacements the effect on the vertical temperature distribution may be readily estimated from a representative tephigram upwind of the high ground by using dry-adiabatic cooling until saturation is reached and the wet-adiabatic rate thereafter.

Although all temperatures are affected both in ascent and descent, attention is drawn particularly to the lowering of the 0°C isotherm during ascent. The pattern of vertical motion in mountain waves varies with height in a way which cannot be readily determined in operational forecasting and some working rule must be adopted. According to Corby[84] a reasonable basis for the purpose of calculating the lowering of the freezing level is to assume that the air at all levels follows the shape of the ground. Thus for an onshore wind passing over mountains rising from the coast to 600 metres above mean sea level it may be assumed that the air may at times over the high ground be lifted at all levels 600 metres above its undisturbed level. The maximum effect on the level of the 0°C isotherm may be assessed from the representative upwind tephigram. The choice of this sounding should be such that it is reasonably sure that the sounding itself has not been affected by similar wave motions. It must be remembered that high ground can cause waves having an amplitude at least as great as that of the topography even at heights several thousands of metres above the surface. When a suitable airstream flows perpendicularly across a long ridge Corby considers that the amplitude of the waves may exceed that of the ground (by a factor of perhaps $1\frac{1}{2}$). He remarks that if the height of the highest ground including that of the individual peaks is used in applying the procedure, some additional tolerance for the amplitude of the waves exceeding that of the ground will automatically be included, since actual lee waves from the high ground will be smaller than those calculated for an infinite ridge of the height of the highest peak. Since the largest amplitude lee waves commonly take place in layers of the atmosphere having great static stability the lowering of the 0°C isotherm is particularly important since the effect of vertical motions on the temperature is greater for greater stabilities. If lee waves seem likely some suitable value for the possible lowering of the 0°C isotherm should be included in the forecast if the route is over or near to high ground.

When the temperature and wind distribution are unfavourable for the formation of lee waves, the disturbance to the flow over the hill does not usually extend to such great heights. However, since it is difficult at present to determine with reasonable certainty whether or not lee waves can exist, it is prudent to allow for the same lowering of the 0°C isotherm as indicated in the preceding paragraph. If the level of the 0°C isotherm is near the hill tops it may be lowered locally below them. Although it would not normally be prudent in such circumstances to plan a flight at an altitude near that of the hill tops, some such flights may be made intentionally and, if this is known, forecasters should make a reference in the written forecast to the possibility of the 0°C isotherm being lowered below that of the hill tops.

17.7.5.3 Effect of a range of hills. In winter a range of hills may effectively prevent warm air from penetrating to ground level in the low-lying ground to the lee of the hills in certain synoptic types in winter. The following sequence of events seems to occur. The area is under the influence of an anticyclone with strong radiative cooling, and light winds for one or more days. In consequence the ground becomes very cold and the air in the lowest hundred metres or so is also cold and very stable. If a warm front then moves across the area from the direction of the hills with a gradient wind behind it of not more than about 20 knots that part of the front above the tops of the hills moves at about the average speed. The hills are an effective barrier to the warm air in the lower levels and the warm air is dammed up to windward of the hills. Some warm air tends to follow the lee contour of the hills towards lower ground but the penetration of warm air down the slope is effectively opposed by the strong thermal stability of the dome of cold air of greater density which lies in the valley. Mechanical stirring due to turbulence is a fairly slow process particularly as winds are not strong and several hours may elapse between the passage of the free warm-front surface across the area at the level of the hilltops and the time when warm air fills the valleys. Judgement is needed in forecasting the time of arrival of warm air at ground level. This is important for forecasting temperatures but generally more so for the accurate forecasting of visibilities. The Vale of York is prone to this type of occurrence with high pressure to the east, a light southerly gradient and a warm front crossing the Pennines with a light south-westerly wind; the Forth and Clyde valleys are also susceptible.

17.7.6 Effects of shelter - valleys

The climate and temperature regimes of valleys are the results of a complex interaction of a number of factors, such as varying exposure to solar radiation, local wind circulations (see 17.6.6, page 51), etc. General rules for forecasting cannot be given, but some case studies are worth mentioning. Dight[85] has shown that the diurnal range of temperature in Scottish glens shows a higher frequency of high values than does a station in a relatively flat area. Part of this he attributed to lower minimum temperatures as a result of cold air pooling in the valley bottom, and part to the breaking down during the day of the stable stratification in the valleys by turbulent flow over the hills or by lee waves, bringing down warm dry air from aloft in much the same way as the föhn wind. Harrison,[86,87] studying the night minimum temperatures in a valley in Kent, found that the night minimum at a given station depended largely on the height of that station above the valley floor (or above the area into which cold air could drain), rather than on the height of the station above sea level (see Figure 20(a)). In fact, the night minimum temperatures at four stations on the valley floor were very similar, although their heights above sea level varied considerably (see Figure 20(b)).

Examples of the part played by katabatic drainage in producing very low minimum temperatures have been reported by Smith[88] and by Oliver.[89]

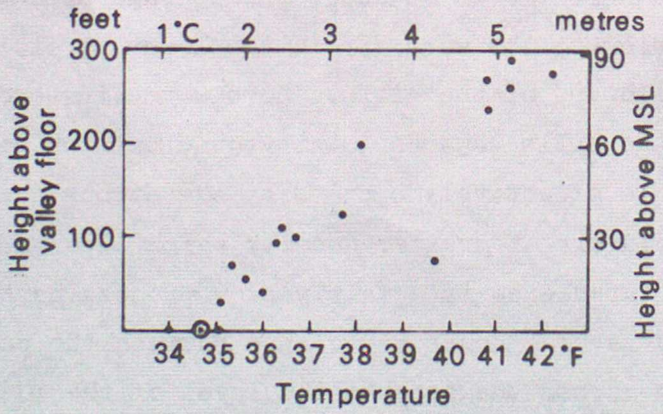


FIGURE 20(a). Night minimum temperatures plotted against height above valley floor for nights when radiation effects outweighed others

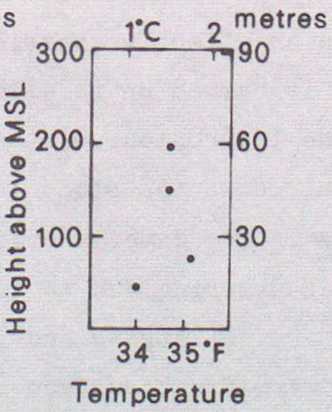


FIGURE 20(b). Valley-floor night minimum temperatures plotted against height above MSL for nights when radiation effects outweighed others

28 selected radiation nights, 19 stations.

17.7.7 Effects of built-up areas

The replacement of natural features and vegetation by roads and buildings affects the airflow over the area, the thermal properties of the 'surface', and reduces the heat losses by evaporation; as a result the temperature regime is changed. In addition, energy is also added to the urban atmosphere by human activities. The effect of all these changes is to increase the temperature, so that in general an urban area is warmer than a nearby country area. This applies both by day and by night, but the magnitude of the temperature change is usually greater at night than during the day, the buildings and roads absorbing heat during the day and slowly releasing heat at night, acting as a vast reservoir. It appears that the character of the area within a radius of 500 metres or so largely determines the magnitude of the temperature effect, the changes being greatest in areas where buildings are predominantly low. Where buildings are mainly tall there is a reduction in the amount of radiation reaching street level, thus reducing the magnitude of the temperature effect. Chandler[90] has reported on the urban 'heat island' effect, as it is called, in London and Leicester; the effect, as indicated above, seems to depend more upon the character of the immediate surroundings and not on the size of the city. The magnitude of the temperature differences between town and country decreases with increasing wind speed, but in suitable conditions can be as much as 5 degrees (see Figure 21).

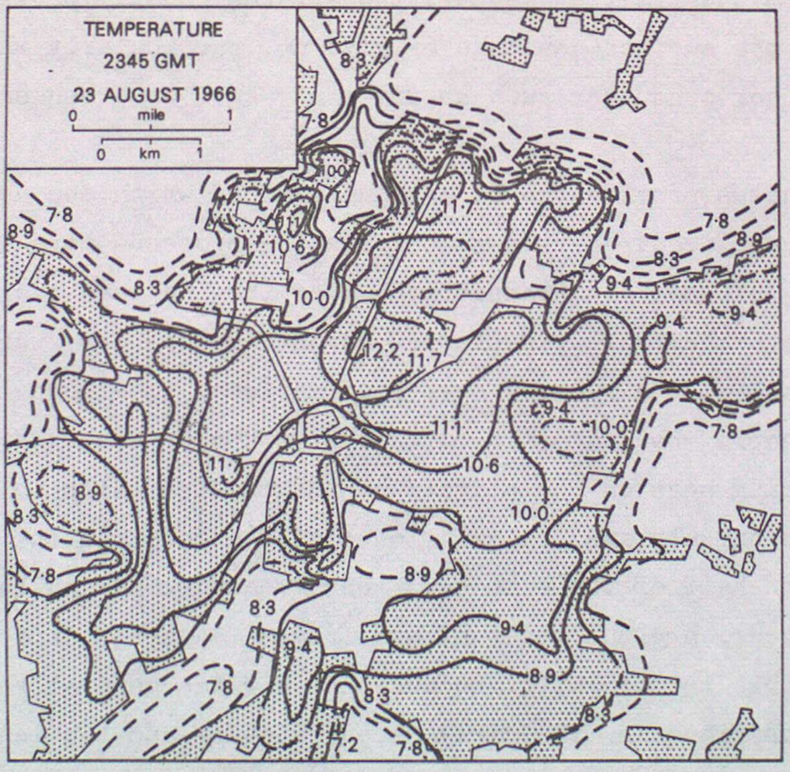


FIGURE 21. Temperature pattern for Leicester at 2345 GMT, 23 August 1966

Isotherms are labelled in degrees Celsius. Broken lines indicate isotherms whose exact position is uncertain.

A WMO Technical Note, No. 108, 'Urban Climates',[91] contains a number of interesting papers on the subject.

17.7.8 Effects of contact with the earth's surface

In 17.4 (page 11) the factors controlling heat transport near the surface were discussed. Briefly, the heat-exchange processes are largely dependent upon σ, the 'conductive capacity' (Priestley[17]) of the air and that of the underlying surface. The values of σ for various surfaces are given in Table 17.4 (page 13); that for air varies usually between 0.1 and unity, depending upon the intensity of turbulence in the surface layers. It can readily be seen that σ for water is very much greater than that for air, which in turn is a good deal larger than the conductive capacity of solid surfaces. In effect this means that the ocean acts as an almost infinite reservoir when giving up heat to, or abstracting heat from, the atmosphere; the rate of heat transfer depends therefore largely on the state of the atmosphere. If the atmosphere is stable, and the heat exchange extends through only a shallow layer of the atmosphere, equilibrium is soon reached, and only small amounts of heat are transferred from one medium to the other. If, however, the atmosphere is unstable and convective mixing extends through a considerable depth, equilibrium is achieved more slowly, and greater amounts of heat are transferred. Over land, on the other hand, the properties of the surface form the limiting factor; the air can usually provide or absorb more heat than the surface can take or give up, particularly when the surface consists of a poor conductor such as dry sandy soil or fresh snow.

The above notes provide the background against which the next two subsections should be read. Air-mass modification proceeds more quickly, and greater amounts of heat are transferred, over the sea than over the land; greater amounts of heat are transferred across the boundary when the air is unstable, as when cold air moves over a warmer sea, than when the air is stable, for example when warm air moves over a cooler sea. The importance of the surrounding seas on the climate of the British Isles is well known, and forecasters quickly acquire experience of the modification of air masses from the western half of the compass. Less frequent, but no less important when they do occur, are outbreaks of air from an easterly direction. They are more difficult to deal with, because the air has generally had a continental land track in its fairly recent history and the length of the sea track varies considerably with the overall direction of approach. The next two subsections will concentrate mainly on this problem.

17.7.8.1 <u>Flow of cold air over a warmer sea</u>. Frost[92] has derived some simple working rules for calculating the changes in air temperature and dew-point occurring in cold air masses whilst crossing the North Sea in winter. The work is based on the turbulent transfer of properties and assumes that:

(a) the variation of wind direction with height may be neglected,

(b) the air in contact with the surface of the sea is saturated at the temperature of the sea surface, and

(c) the mass of water vapour per unit mass of air is originally constant with height (this would be satisfied if the air over the land had been well stirred).

If r_o and T_o are the humidity mixing ratio and temperature of the original air, and r_m and T_m, are the values corresponding to the sea-surface temperature, then the humidity mixing ratio, r, and temperature, T, of the air after a trajectory over the North Sea in excess of about 100 kilometres are given by the following formulae:

$$T = T_o + 0.6(T_m - T_o) \qquad \ldots\ldots (17.28)$$

$$r_o = r_o + 0.6(r_m - r_o) \qquad \ldots\ldots (17.29)$$

These formulae apply to all cold airstreams crossing a warmer sea surface. They may be used to calculate modifications to temperature and moisture content of cold air crossing any sufficiently extensive warmer water surface, for example, a cold northerly outburst reaching north Scotland or a cold westerly current reaching Norway. They are particularly valuable for forecasting temperatures in the United Kingdom in easterly situations.

Application of the formulae is extremely simple. Estimate the trajectory of the air over the North Sea, and make the best estimate of the surface temperature of the North Sea along the trajectory from the latest available ships' observations and by comparison with mean values. Determine r_o and r_m from the tephigram and calculate T and r. The dew-point corresponding to r is obtained from the tephigram.

Blackall[93] has criticised Frost's arguments on the grounds that the factor 0.6 applies only to sea crossings greater than 300 nautical miles (550 km) in length, that the presence of any inversion is ignored, and that sometimes the value of r is greater than the saturation humidity mixing ratio at T. Instead, Blackall put forward a more empirically-based method, using a simple construction on the tephigram, which takes into account the depth of convection and the duration of the sea crossing. The method is based on the equation

$$T = T_m - (T_m - T_o)\exp(-12\,t/d) \qquad \ldots . \quad (17.30)$$

where t is the duration of the sea crossing, in hours, and d is the depth of convection in millibars. The procedure is as follows and is illustrated in Figures 22 and 23:

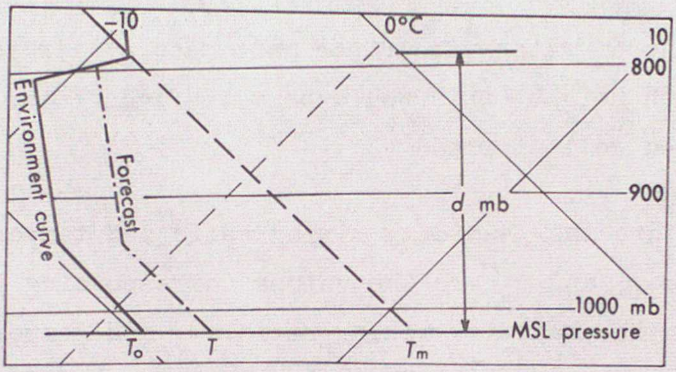


FIGURE 22. Steps in the preparation of a forecast of temperature when the environment curve needs no modification and step (c) may be omitted

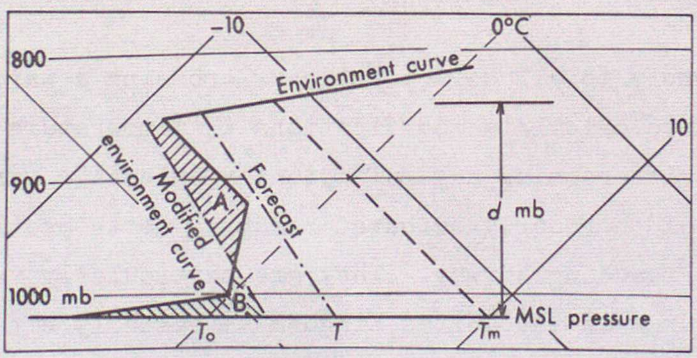


FIGURE 23. Steps in the preparation of a forecast of temperature when a modification of the environment curve is necessary

(a) On the sounding in the air upwind of the sea crossing, draw in the sea-level isobar and, if necessary, extend the ascent downwards to meet this isobar at coastal temperatures.

(b) Draw the dry adiabatic through the sea temperature T_m (see (g)). The pressure at which this line meets the environment curve is subtracted from the surface pressure to give the depth of convection d in millibars.

(c) Establish the expected lapse rate; if the air already has a lapse rate implying convection throughout the layer d (see Figure 22) the environment curve should not be changed as any change is unlikely to lead to more-accurate results. If a lapse rate needs to be forecast construct a modified environment curve as follows: draw a line through the layer d, with a lapse appropriate to d (see side of Table 17.29) such that the environment curve encloses equal areas (A and B in Figure 23) on each side of the line. Where this line meets the surface isobar is T_o. This step represents complete mixing without addition of heat.

(d) Determine the time, t, that the air will spend over the sea from the available wind information for the layer d and the fetch. Note that because of mixing the wind velocity in the convection layer is more or less uniform and great care is required if actual winds are to be used from upwind stations when the soundings from these stations indicate stability over land but instability over the sea. Gradient winds from surface and 850-millibar analyses and prognoses should be sufficiently accurate in most cases.

(e) From Table 17.29 for the appropriate values of t hours and d millibars find the value of $\exp(-12t/d)$ and thence T from equation 17.30.

(f) From temperature T on the surface isobar, draw a line parallel to the environment curve produced in step (c); this is the predicted environment curve for when the air finishes its crossing.

(g) If the sea temperature is far from uniform a mean value will often give good answers; however, if the changes in T_m will mean changes in d such that $\exp(-12t/d)$ alters significantly then it will be necessary to proceed by steps - as will also be necessary when the passage is expected to take more than 24 hours.

TABLE 17.29 Values of $\exp(-12\,t/d)$

d	Duration of crossing, t (hours)														Most probable lapse
(*mb*)	1	2	3	4	5	6	7	8	9	12	15	18	21	24	(*degC*)
700	0·98	0·97	0·95	0·93	0·92	0·90	0·89	0·87	0·86	0·81	0·77	0·73	0·70	0·66	
600	0·98	0·96	0·94	0·92	0·90	0·89	0·87	0·85	0·84	0·79	0·74	0·70	0·66	0·62	
500	0·98	0·95	0·93	0·91	0·89	0·86	0·84	0·82	0·81	0·75	0·70	0·65	0·60	0·56	40
450	0·98	0·95	0·92	0·89	0·88	0·85	0·83	0·81	0·79	0·73	0·67	0·62	0·57	0·53	35
400	0·97	0·94	0·91	0·89	0·86	0·84	0·81	0·79	0·76	0·70	0·64	0·58	0·53	0·49	31
350	0·97	0·93	0·90	0·87	0·84	0·81	0·79	0·76	0·73	0·64	0·60	0·54	0·49	0·44	26
300	0·96	0·92	0·89	0·85	0·82	0·79	0·76	0·73	0·70	0·62	0·55	0·49	0·43	0·38	22
250	0·95	0·91	0·86	0·83	0·79	0·75	0·71	0·68	0·65	0·57	0·49	0·42	0·36	0·32	17
200	0·94	0·89	0·84	0·79	0·74	0·70	0·66	0·62	0·58	0·49	0·41	0·34	0·28	0·24	14
180	0·94	0·88	0·82	0·77	0·72	0·67	0·63	0·59	0·55	0·45	0·37	0·30	0·25	0·20	13
160	0·93	0·86	0·80	0·74	0·69	0·64	0·59	0·55	0·51	0·41	0·33	0·26	0·21	0·17	12
140	0·92	0·84	0·77	0·71	0·65	0·60	0·55	0·50	0·46	0·36	0·28	0·21	0·17	0·13	10
120	0·90	0·82	0·74	0·67	0·61	0·55	0·50	0·45	0·41	0·30	0·22	0·17	0·12	0·09	9
100	0·89	0·79	0·70	0·62	0·55	0·49	0·43	0·38	0·34	0·24	0·17	0·12	0·08	0·06	8
80	0·86	0·74	0·64	0·55	0·47	0·41	0·35	0·30	0·26	0·17	0·11	0·07	0·04	0·03	6
60	0·82	0·67	0·55	0·45	0·37	0·30	0·25	0·20	0·17	0·09	0·05	0·03	0·01		5
50	0·79	0·62	0·49	0·38	0·30	0·24	0·19	0·15	0·12	0·06	0·03	0·01	—	—	4
40	0·74	0·55	0·41	0·30	0·22	0·17	0·12	0·09	0·07	0·03	0·01	—	—	—	4
30	0·67	0·45	0·30	0·20	0·14	0·09	0·06	0·04	0·03	0·01	—	—	—	—	3

The procedure described above does not allow for the presence of fronts and other dynamic means of heating and cooling.

The increase in surface humidity mixing ratio is taken to be

$$r - r_o = 0.18(T - T_o) \quad \ldots\ldots \quad (17.31)$$

with the dew-point remaining constant with height. For details of the argument leading to this result, and for more information on the method of temperature forecasting the reader is referred to Blackall's paper.[93]

17.7.8.2 Flow over a colder sea surface. Contact with the cooler underlying surface soon cools the lowest layers of the atmosphere and soon establishes a stable lapse rate or an inversion. The cooling is partly offset by turbulent transport of heat from the higher layers, but the effectiveness of this process is limited by the stability of the lower layers; cooling is normally limited to the zone of mechanical turbulence resulting from surface friction.

Belasco[11] has established some figures relating to cooling of air as it moves from the region of the Azores to the British Isles in both winter and summer. These figures were determined from actual values and they include the effect of all the physical processes which were at work during the period of advection of the air. His values are given in Table 17.30. This, together with the actual air and sea temperatures obtaining, should enable an estimate to be made of the low-level stratification and temperatures of warm air masses as they approach the British Isles.

TABLE 17.30 Average decrease of temperature in tropical maritime air between the Azores and the coasts of south-west England

	Average decrease			
	Sea temperature	Air temperature		
Season		Surface	950 mb	900 mb
	°C	°C	°C	°C
Summer	6	5	1.5	1
Winter	5.5	4.5	3	1.5

In an investigation of 'Haars or North Sea fogs on the coast of Great Britain', Lamb[78] developed the theoretical aspects of the cooling of warm air masses over a cool sea. Details of the derivation of his formulae, assumptions and approximations are contained in the original paper and are not reproduced here. One aspect of this work is directly applicable to forecasting screen-level temperatures when warm air over a cold North Sea is being advected to the eastern coastal region. If T_o is the initial temperature of the warm air as it leaves a land surface and T_m is the temperature of the sea (assumed constant) then the temperature, T, of the air after traversing various distances across the sea surface is given by:

$$T = T_m + (T_o - T_m)f(xz) \quad \ldots . \quad (17.32)$$

where x is the distance traversed and z is the height above the surface. Taking z as 1.22 metres (that is, screen height) the following figures are the values of $f(xz)$ for trajectories up to 1000 kilometres in steps of 100 kilometres.

x(km)	100	200	300	400	500	600	700	800	900	1000
$f(xz)$ for $z = 1.22$ m	0.175	0.152	0.141	0.133	0.127	0.123	0.119	0.116	0.113	0.110

17.7.9 Cooling of air by precipitation

As precipitation falls through unsaturated air, evaporation takes place, the latent heat of evaporation being extracted mainly from the air through which the precipitation is falling. The lower limit to which the air can be cooled by evaporation of water into it is the wet-bulb temperature. A temperature very close to the wet-bulb value is reached after about $\frac{1}{2}$-$\frac{3}{4}$ hour of very heavy rain, or after about 1-2 hours of less intense falls.[94,95]

During thunderstorms, the cooling of air by precipitation often leads to a downdraught of air; the temperature in the downdraught as it reaches the surface can be derived by the method of Fawbush and Miller.[96] This temperature is given by the surface temperature of the saturated adiabatic through the intersection of the wet-bulb curve and the 0°C isotherm. Frozen precipitation can, unlike rain, cool the air to below its wet-bulb temperature.

Lumb[97] has studied the cooling effect of falling snow; he came to the conclusion that the surface temperature is unlikely to be reduced below 0°C if the following conditions hold:

(a) in prolonged frontal precipitation if the surface wet-bulb temperature is higher than 2.5°C,

(b) within extensive areas of moderate or heavy instability precipitation if the surface wet-bulb temperature is higher than 3.5°C.

17.8 FORECASTING TEMPERATURES IN THE UPPER AIR

The preceding sections, although dealing nominally with the forecasting of temperatures at screen level or at the surface itself, have perforce included some discussion of the temperature profiles in the lower atmosphere. Forecasting temperatures within this region is often very tricky, particularly in inversion conditions, and the forecaster needs a sound knowledge of the relevant physical processes. Further guidance appears in the sections of Chapter 16 - Wind - dealing with the structure of sea-breezes and the depth of the atmospheric boundary layer.

Above the friction layer, and up to 300 millibars or so, forecasts of temperature are derived mainly from computer-produced prognoses. Such forecasts are sometimes modified subjectively, before they are passed on to outstations, in areas where the computer products are known to be subject to error or where they tend to smooth out detail. A good deal of care has to be taken in adjusting the computer output, or the modified forecast is likely to be more in error than the original. A thorough knowledge of the behaviour of atmospheric systems and of the behaviour of computer models is essential. More detail is given in Chapter 3 - Background to computer models - on the use of computer products in upper-air forecasting. At 200 millibars and above, computer prognoses are less reliable than lower down, and, at present, forecasting temperatures at these levels is carried out on the basis of persistence and/or extrapolation of observed patterns and trends. Some useful climatological information is given in Chapter 8 - Jet streams, tropopause and lower stratosphere.

BIBLIOGRAPHY

1. BRUNT, D.; Physical and dynamical meteorology. 2nd edition. Cambridge University Press, Cambridge, 1939.

2. Washington, Smithsonian Institution; Smithsonian meteorological tables. Smithson Misc Coll, Washington, 114, 1951.

3. MIE, G.; Beiträge zur Optik trüber Medien speziell kolloidaler Metallösungen. Ann Physik, Leipzig. 25, 1908, p.377.

4. MONTEITH, J.L.; Attenuation of solar radiation: a climatological study. Q J R Met Soc, London, 88, 1962, pp. 508-521.

5. VONDER HAAR, T.H. and SUOMI, V.E.; Measurements of the earth's radiation budget from satellites during a five-year period. Part I: Extended time and space means. J Atmos Sci, Lancaster, Pa, 28, 1971, pp. 305-314.

6. LUMB, F.E.; The influence of cloud on hourly amounts of total solar radiation at the sea surface. Q J R Met Soc, London, 90, 1964, pp. 43-56.

7. SIMPSON, G.C.; Some studies in terrestrial radiation. Mem R Met Soc, London 2, No. 16, 1927. Further studies in terrestrial radiation. Mem R Met Soc, London, 3, No. 21, 1928. The distribution of terrestrial radiation. Mem R Met Soc, London, 3, No. 23, 1929.

8a. ÅNGSTRÖM, A.; Radiation of the atmosphere. Smithson Misc Coll, Washington, 65, 1915, p.3.

8b. ÅNGSTRÖM, A.; Über Variationem der atmosphärischen Temperaturstrahlung und ihren Zusammenhang mit der Zusammensetzung der Atmosphäre. Beitr Geophys, Leipzig, 21, 1929, p.145.

9. BRUNT, D.; Notes on radiation in the atmosphere. I. Q J R Met Soc, London, 58, 1932, pp 389-420.

10. MONTEITH, J.L.; An empirical method for estimating the long-wave radiation exchanges in the British Isles. Q J R Met Soc, London, 87, 1961, pp. 171-179.

11. BELASCO, J.E.; Characteristics of air masses over the British Isles. Geophys Mem, London, 11, No. 87, 1952.

12. SWINBANK, W.C.; Long-wave radiation from clear skies. Q J R Met Soc, London, 89, 1963, pp. 339-348.

13. PALTRIDGE, G.W.; Day-time long-wave radiation from the sky. Q J R Met Soc, London, 96, 1970, pp. 645-653.

14. DEACON, E.L.; Physical processes near the earth's surface. Flohn,H. (editor). World survey of climatology, Vol.2, General climatology, 2. Barking, Elsevier Publishing Co. Ltd, 1970, pp. 39-104.

15. WEDMORE, E.B.; Thermal conductivity of soil. Q J R Met Soc, London, 67, 1941, p.38.

16. SUTTON, O.G.; Micrometeorology, London, McGraw-Hill, 1953.

17. PRIESTLEY, C.H.B.; Turbulent transfer in the lower atmosphere. Chicago, University of Chicago Press, 1959.

18. LAWRENCE, E.N.; Large air-temperature lapse-rates near the ground. Weather, London, 27, 1972, pp. 27-30.

19. BEST, A.C.; Transfer of heat and momentum in the lowest layers of the atmosphere. Geophys Mem, London, 7, No. 65, 1935.

20. PASQUILL, F.; Eddy diffusion of heat and water vapour near the ground. Proc R Soc, London, A, 198, 1949, pp. 116-140. (See also p.200, Reference 16.)

21. ZOBEL, R.F.; Temperature and humidity changes in the lowest few thousand feet of atmosphere during a fine summer day in southern England. Q J R Met Soc, London, 92, 1966, pp. 196-209.

22. OKE, T.R.; The temperature profile near the ground on calm clear nights. Q J R Met Soc, London, 96, 1970, pp. 14-23.

23. BEST, A.C.; Horizontal temperature differences over small distances. Q J R Met Soc, London, 57, 1931, pp. 169-176.

24. FRITH, R.; Small-scale temperature and humidity patterns in the free air. Met Res Pap, London, No. 402, 1948.

25. GRANT, D.R.; Fluctuations of temperature in clear air as recorded by an ultra-rapid thermometer. Met Res Pap, London, No. 658, 1951.

26. WARHAFT, Z.; The relation between temperature and humidity in the free atmosphere under conditions of stable stratification and strong thermal intermittency - a case study. Q J R Met Soc, London, 99, 1973, pp. 89-104.

27. HOOPER, A.H.; Some features of the new United Kingdom radiosonde. Geneva, World Meteorological Organization. Upper-air instruments and observations, Geneva, 1970, pp. 364-367.

28. BARBÉ, G.D.; Deux remarquables exemples de cisaillement dans la stratosphère concernant le vent et la température. NATO Advisory Group for Aerospace Research and Development. Aerodynamics of atmospheric shear flows. AGARD Conf Proc No. 48, 1970, pp. 4.1-4.12.

29. READINGS, C.J., GOLTON, E. and BROWNING, K.H.; Fine-scale structure and mixing within an inversion. Boundary-Layer Met., Dordrecht, 4, 1973, pp. 284-300.

30. BEST, A.C., KNIGHTING, E., PEDLOW, R.H. and STORMONTH, K.; Temperature and humidity gradients in the first 100m over south-east England. Geophys Mem, London, 11, No. 89, 1952.

31. GOLD, E.; Maximum day temperatures and the tephigram. Prof Notes, Met Off, London, 5, No. 63, 1933.

32. DINES, W.H. and DINES, L.H.G.; Monthly mean values of radiation from various parts of the sky at Benson, Oxfordshire. Mem R Met Soc, London, 2, No. 11, 1927.

33. JOHNSTON, D.W.; The estimation of maximum day temperature from the tephigram. Met Mag, London, 87, (September) 1958, pp. 265-266.

34. INGLIS, G.A.; Maximum temperature on clear days. Met Mag, London, 99, (December) 1970, pp. 355-363.

35. BOYDEN, C.J.; Forecasting of maximum surface temperature from 1000-500-millibar thickness lines. Met Mag, London, 91,(September), 1962, pp. 242-246.

36. WALLINGTON, C.E.; Tricks with the tephigram - 2. Gliding, London, 6 1955, p. 28.

37. JEFFERSON, G.J.; Temperature rise on clear mornings. Met Mag, London, 79, (February) 1950, pp. 33-41.

38. JEFFERSON, G.J.; A method of forecasting the time of clearance of radiation fog or low stratus. Met Mag, London, 79,(April) 1950, pp. 102-109.

39. FROST, R.; Calculation of night minimum temperature. Prof Notes, Met Off, London, 6, No. 95, 1948.

40. KNIGHTING, E.; A note on nocturnal cooling. Q J R Met Soc, London, 76 1950, p. 494.

41. BOYDEN, C.J.; A method of predicting night minimum temperatures. Q J R Met Soc, London, 63, 1937, pp. 383-392.

42. McKENZIE, F.; A method of estimating night minimum temperatures. S.D.T.M. No. 68, London, 1944, (Unpublished; copy available in the Meteorological Office Library, Bracknell).

43. CRADDOCK, J.M. and PRITCHARD, D.L.; Forecasting the formation of radiation fog - a preliminary approach. Met Res Pap, London, No. 624, 1951.

44. GORDON, J., PERRY, J.D. and VIRGO, S.E.; Forecasting night minimum air temperature by a regression equation. Met Mag, London, 98, (September) 1969, pp. 290-292.

45. VINCE, J.; Minimum temperature at Gatwick airport. (Unpublished)

46. SAUNDERS, W.E.; Night cooling under clear skies. Q J R Met Soc, London, 75, 1949, pp. 154-160.

47. SAUNDERS, W.E.; Some further aspects of night cooling under clear skies. Q J R Met Soc, London, 78, 1952, pp. 603-612.

48. BARTHRAM, J.A.; A method of forecasting a radiation night cooling curve. Met Mag, London, 93, (August) 1964, pp. 246-251.

49. TINNEY, E.B. and MENMUIR, P.; Results of an investigation into forecasting night-minimum screen temperatures. Met Mag, London, 97, (June) 1968, pp. 165-172.

50. SUMMERSBY, W.D.; Night cooling under cloudy skies. Met Mag, London, 82, (July) 1953, pp. 210-213.

51. MIZON, E.A.; An objective aid for minimum temperature forecasts - Alice Springs, N. Territory. Aust Met Mag, Melbourne, 38, 1962, pp. 20-31.

52. STEELE, L.P., STROUD, P.A.J. and VIRGO, S.E.; Accuracy of forecasting night minimum air temperatures by the method due to Saunders. Met Mag, London, 98, (April) 1969, pp. 107-113.

53. GORDON, J. and VIRGO, S.E.; Comparison of methods of forecasting night minimum temperatures. Met Mag, London, 97, (June) 1968, pp. 161-164.

54. SCHMIDT, K.-H.; Ein Betrag zur Vorhersage der Temperaturminima in Nächten mit geringer Bewölkung. Zeit Met, Berlin, 14, Nr 11/12, 1960, pp. 297-307.

55. FAUST, R.; Ein Hilfsmittel zur Nachtfrostvorhersage. Ann Met, Hamburg, Heft 3/4, 1949, pp. 105-111.

56. JEFFERSON, G.J.; Forecasting ground frost. Met Mag, London, 80, (October) 1951, pp. 295-296.

57. JAMES, W.E.; Forecasting ground frost. Met Mag, London, 82, (March) 1953, pp. 91-92.

58. LAWRENCE, E.N.; Forecasting grass minimum and soil temperatures under clear skies and light winds. Met Mag, London, 89, (February) 1960, pp. 33-42.

59. SMITH, L.P.; A simple method of frost forecasting for use by growers. Proc XVI Int Hort Congr, Brussels, 1962. Gembloux, Editions J. Duculot, S.A., Vol 5, 1964, pp. 226-230.

60. MEADS, Dorothy; A simple method of frost-forecasting for use by growers. Weather, London, 18, 1963, pp. 332-334.

61 HALIM, A.; Frost forecasting by growers. Weather, London, 19, 1964, p.228.

62. GORDON, J.; A report on L.P.Smith's simple method of frost forecasting. 1969 (Unpublished)

63. STEELE, L.P., STROUD, P.A.J. and VIRGO, S.E.; An empirical approach to forecasting grass minimum temperatures and the probability of grass minima below 0°C in eastern England. Met Mag, London, 98, (November) 1969, pp. 340-348.

64. HOGG, W.H.; Frequency of ground frost in S.W. England on radiation nights, 1949. (Unpublished; copy available in the Meteorological Office Library, Bracknell)

65. SILLS, A.G.; An investigation into the depression of the grass minimum temperature below the air minimum temperature at Cottesmore. Met Mag, London, 98, (November) 1969, pp. 348-351.

66. GLOYNE, R.W.; Radiation minimum temperature over a grass surface and over a bare-soil surface. Met Mag, London, 82, (September) 1953, pp. 263-267.

67. RITCHIE, W.G.; Night minimum temperatures at or near various surfaces. Met Mag, London, 98, (October) 1969, pp. 297-304.

68. PARREY, G.E.; Minimum road temperatures. Met Mag, London, 98, (September) 1969, pp. 286-290.

69. CLARK, C.M.; Forecasting the night minimum temperature of a concrete surface in winter. London, Meteorological Office, Forecasting Techniques Memorandum No. 17, 1969.

70. THORNES, J.E.; An objective aid for estimating the night minimum temperature of a concrete road surface. Met Mag, London, 101, (January), 1972, pp. 13-24.

71. PARREY, G.E., RITCHIE, W.G. and VIRGO, S.E.; Minimum temperatures at the surfaces of concrete roads and concrete slabs. Met Mag, London, 100, (January) 1971, pp. 27-31.

72. PARREY, G.E.; Forecasting temperature for the gas and electricity industries. Met Mag, London, 101, (September) 1972, pp. 264-270.

73. PETERS, S.P.; Sea breezes at Worthy Down, Winchester. Prof Notes Met Off, London, 6, No. 86, 1938.

74. LAWRENCE, E.N.; Nocturnal winds. Prof Notes, Met Off, London, 7, No. 111, 1954.

75. GEIGER, R.; The climate near the ground. Cambridge, Mass., Harvard University Press. 4th Edition, 1965.

76. STEWART, K.H.; Radiation fog: investigations at Cardington 1951-54. Met Res Pap, London, No. 912, 1955.

77. JOHNSON, N.K. and HEYWOOD, G.S.P.; An investigation of the lapse rate of temperature in the lowest hundred metres of the atmosphere. Geophys Mem, London, 9, No. 77, 1938.

78. LAMB, H.H.; Haars or North Sea fogs on the coast of Great Britain, 1943. (Unpublished; copy available in Meteorological Office Library, Bracknell)

79. LOCKWOOD, J.G.; The occurrence of Föhn winds in the British Isles. Met Mag, London, 91, (March) 1962, pp. 57-65.

80. HANN, J.von; Zur Frage über den Ursprung des Föhnes. Met Zeit, Wien, 1, 1866, p.257.

81. McCAFFERY, W.D.S.; Föhn effect over Scotland. Met Mag, London, 81, (May) 1952, pp. 151-152.

82. LAWRENCE, E.N.; Föhn temperature in Scotland. Met Mag, London, 82, (March) 1953, pp. 74-79.

83. GIRDWOOD, D.; Exceptional temperature and humidity fluctuations at Turnhouse. Met Mag, London, 89, (December) 1960, pp. 336-338.

84. CORBY, G.A.; Air flow over mountains: Notes for forecasters and pilots. Met Rep, London, 3, No. 18, 1957.

85. DIGHT, F.H.; The diurnal range of temperature in Scottish glens. Met Mag, London, 96, (November) 1967, pp. 327-334.

86. HARRISON, A.A.; Variations in night minimum temperatures peculiar to a valley in Kent. Met Mag, London, 96, (September) 1967, pp.257-265.

87. HARRISON, A.A.; A discussion of the temperatures of inland Kent with particular reference to night minima in the lowlands. Met Mag, London, 100, (April) 1971, pp. 97-111.

88. SMITH, K.; A note on minimum screen temperatures in the Houghall frost hollow. Met Mag, London, 96, (October) 1967, pp 300-302.

89. OLIVER, J.; Low minimum temperatures at Santon Downham, Norfolk. Met Mag, London, 95, (January) 1966, pp. 13-17.

90. CHANDLER, T.J.; Night-time temperatures in relation to Leicester's urban form. Met Mag, London, 96, (August) 1967, pp. 244-250.

91. Geneva, World Meteorological Organization. Urban climates. Tech Notes, Wld Met Org, Geneva, No. 108, 1970.

92. FROST, R.; The influence of the North Sea in winter on temperature and dew-point. 1941. (Unpublished; copy available in Meteorological Office Library, Bracknell)

93. BLACKALL, R.M.; Warming of the lower troposphere by the sea. Met Mag, London, 102, (March) 1973, pp. 65-73.

94. London, Meteorological Office; Note on the lowering of cloud base during rain. S.D.T.M. No. 13, 1941. (Unpublished; copy available in the Meteorological Office Library, Bracknell).

95. DOLEZEL, E.J.; Saturation and cooling of air layers by evaporation from falling rain. J Met, Lancaster, Pa, 1, 1944, pp. 89-97.

96. FAWBUSH, E.J., and MILLER, R.C.; A basis for forecasting peak wind gusts in non-frontal thunderstorms. Bull Amer Met Soc, Lancaster, Pa, 35, 1954, pp. 14-19.

97. LUMB, F.E.; Downward penetration of snow in relation to the intensity of precipitation. Met Mag, London, 92, (January) 1963, pp. 1-14.

General reading

JOHNSON, J.C.; Physical meteorology. Cambridge, Mass, MIT Press and New York, John Wiley and Sons, Inc., 1954.

KONDRAT'YEV, K.Ya.; Radiation in the atmosphere. International Geophysics Series (Vol. 12). New York, Academic Press, 1969.

FLEAGLE, R.G. and BUSINGER, J.A.; An introduction to atmospheric physics. International Geophysics Series (Vol.5). New York, Academic Press, 1963.